

**Petrology, Geochemistry and Evolution of the
Labrador Trough Basaltic Suite, Labrador and New Québec**

By

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Abstract

The central Labrador Trough is an Early Proterozoic volcano-sedimentary succession which preserves the remnants of a major magmatic province, the Labrador Trough Basaltic Suite (LTBS). The Menihek gabbros, which are distinctive members of the LTBS, form a sill-sediment complex with turbidites of the Menihek Formation. The Menihek gabbros often exhibit spectacular plagioclase glomeroporphyritic textures. Glomerocryst plagioclase grains are distinct from groundmass plagioclase crystals, but glomerocrysts and groundmass share similar initial ϵ_{Nd} ratios. The glomerocrysts are interpreted as flotation cumulates formed at the roofs of mid-level crustal magma chambers which fed the Menihek sills. The systematic variations in glomerocryst size and frequency with stratigraphic position are consistent with flow differentiation during emplacement. The Menihek gabbros exhibit geochemical variations which indicate that fractional crystallization of plagioclase, hypersthene and augite occurred during boundary layer solidification of the intrusions, largely after cessation of flow.

Tholeiitic basalts of the LTBS (Le Fer and upper Menihek/Willbob basalts) exhibit many textural and geochemical similarities, despite a difference of ~ 260 Ma in the ages of the two principal suites. Both consist of low-MgO basalts with transitional MORB-type compositions. Both evolved from primitive partial melts by fractionation of olivine in deep crustal or sub-crustal magma reservoirs, followed by plagioclase+pyroxene fractionation in mid-level crustal magma chambers. The transitional MORB compositions of the tholeiites, and their initial ϵ_{Nd} ratios, indicate that the parental magmas were the products of partial melting of depleted asthenospheric mantle, with minor contributions from a relatively undepleted mantle source, and minor assimilation (1-10%) of Archean sialic crust. The alkaline rocks of the LTBS (Nimish, lower Menihek and lower Murdoch basalts) exhibit strong compositional diversity, and it is probable that variations in mantle source composition and/or melting depths, crustal assimilation, fractional crystallization and post-crystallization processes influenced the geochemistries of these suites. The Nimish and

lower Menihek basalts exhibit extended trace element patterns which indicate that their parental melts were derived primarily from a relatively undepleted mantle reservoir, but with variable contributions from a depleted source.

The stratigraphic and geochemical characteristics of the LTBS suggest that magmatism accompanied crustal extension and formation of pull-apart basins on the thinned margin of the Superior craton. Initial cycle 1 sedimentation (> 2169 Ma) was fluvial and accompanied by localized alkaline volcanism, but subsequent subsidence and marine transgression led to deposition of shallow to deep water sediments, culminating with the Le Fer turbidites. In the axial portions of the basin, where sedimentation rates were low, turbidite deposition was interspersed with eruption of the Le Fer submarine tholeiitic basalts and intrusion of the Le Fer gabbros. Magmatic activity and sedimentation continued until at least 2142 Ma, at which point crustal extension and basin subsidence ceased, sedimentation was interrupted, and in places pre-existing strata were eroded. Although there is some evidence of an inter-cycle unconformity, it is doubtful that the two cycles are separated by a major orogenic event. Renewed extension led to the formation of the cycle 2 pull-apart basin, which was localized in the same general area as a result of crustal weakness related to cycle 1 extension. Crustal attenuation and basin subsidence again led to a transgressive marine sequence, culminating with the Menihek turbidites. Alkaline volcanism of the Nimish and lower Menihek formations accompanied early cycle 2 sedimentation, while turbidite deposition was accompanied by tholeiitic volcanism in the axial regions of the basin (Doublet terrane), and by emplacement of numerous gabbro sills in flanking regions (Howse zone). The Superior craton is likely to have formed the basement to the Labrador Trough succession throughout most of its evolution, but the eruption of late high-Mg basalts in the Doublet terrane, coupled with the emplacement of high-Mg magmas as sills, may record the late rapture of the continental crust. The continental crust thus acted as a density barrier to primitive partial melts until late in the evolutionary history of the Labrador Trough. The transition from alkaline to tholeiitic magmatism in each volcano-sedimentary cycle is thought to reflect progressive increases in the degree of decompressive partial melting in the upwelling, heterogeneous asthenospheric mantle during crustal attenuation.

Sommaire

La Fosse Labrador centrale est une succession de roches volcano-sédimentaires du protérozoïque inférieur qui préserve les résidus d'une province de magmatisme principal, la Suite Basaltique de la Fosse Labrador (SBFL). Les gabbros Menihek, qui sont des membres distinctifs de SBFL, forment un filon-couche de sédiment complexe avec des turbidités de la Formation Menihek. Les gabbros Menihek exposent souvent des textures de plagioclase glomérophyrique spectaculaires. Les grains glomérocrysts de plagioclase sont distinctifs ou différents des cristaux de matrice plagioclase, mais les glomérocrysts de plagioclase ont les mêmes taux ε_{Nd} . Les glomérocrysts sont interprétés comme étant des flottations cumulées au bout des niveaux moyens de réservoirs magmatiques crustal qui nourrissent les filons Menihek. Les variations systématiques en grandeurs et fréquences des glomérocrysts avec position stratigraphique sont consistantes avec les différenciations de coulée pendant les emplacements. Les gabbros Menihek exhibent des variations géochimiques qui indiquent que des fragments de cristallisation plagioclase, hypersthène et augite ont eu lieu pendant la solidification des couches limites des intrusions, en grande partie après l'arrêt des écoulements.

Les basaltes tholéitiques des SBFL (Le Fer et les basaltes supérieurs Menihek/Willbob) exhibent plusieurs similarités texturales et géochimiques malgré une différence de ~260 Ma dans les âges des deux suites principales. Toutes les deux consistent en un basalte bas en MgO avec des compositions transitionnelles de type Basalte de Chaîne Médio Océanique. Tous les deux ont été développés à partir de produits primitifs partiellement fondus par fractionnement d'olivine en crustal profond avec des réservoirs de magma sub-crustal, suivis par la fractionation de plagioclase+pyroxène en chambres de cristal magma de niveau moyen. Les compositions transitionnelles de Basalte de Chaîne Médio Océanique des tholéitiques, et leurs taux initiaux ε_{Nd} , montrent que les magmas parents étaient les produits de fusion partielle de manteau asthénosphérique épuisé avec contribution mineure d'un manteau terrestre relativement non épuisé, et une

assimilation mineure (1-10%) d'une croûte sialique Archéene. Les roches alcaline du SBFL (Nimish, bas Menihek et les bas basaltes Murdoch) montrent une forte diversité de composition et il est probable que les variations dans la composition du manteau de source et/ou profondeur de fusion, assimilation crustal, cristallisation fractionnelle et les procédés post-cristallisation ont influencé la géochimie de ses suites. Le Nimish et les basaltes du bas Menihek exhibent des modèles de trace d'éléments étendus qui indiquent que les produits fondus parentaux venaient surtout en premier lieu d'un manteau de réservoir relativement non-épuisé mais avec des contributions variables d'une source épuisée.

La stratigraphie et les caractéristique géochimique du SBFL suggèrent que le magmatisme accompagne l'extention et la formation crustal de bassins séparés sur la fine marge du craton Supérieur. La première sédimentation cycle 1 (>2169 Ma) était fluviatile et accompagnée par volcanisme alcalin localisé, mais un affaissement subséquent et une transgression marinière ont conduits à la déposition de sédiments en eau peu profonde et profonde, culminant avec les turbites Le Fer. Dans les portions axiales du bassin, où les taux de sédimentation étaient bas, des dépôts de turbite étaient parsemés d'éruptions de basaltes tholéïtiques du Le Fer sous-marin et l'intrusion des gabbros du Le Fer. Les activités magmatiques et sédimentations continuent jusqu'à 2142 Ma, où les extensions crustals et affaissements de bassins cessent, la sédimentation fut interrompue et en certaines places les strata pré-existants furent érodés. Quoiqu'il y ait évidence de non conformité d'un inter-cycle, il est douteux que les deux cycles soient séparés par un événement orogénique majeur. Des extensions renouvelées ont conduit à la formation du cycle 2 bassin séparés, qui sont localisés dans la même région comme résultat d'une faiblesse crustale allié à l'extension du cycle 1. L'atténuation crustale et l'affaissement du bassin de nouveau amènent à une séquence marinière transgressive, culminant avec les turbites Menihek. Des volcanismes alcalins du Menihek et les formations du bas Menihek accompagnaient la sédimentation ancienne du cycle 2, pendant que la déposition de turbidite étaient accompagnée par le volcanisme tholéïtique dans les régions axiales du bassin (terrain Doublet) et par l'emplacement de nombreux filons de gabbros sur les régions de flancs (zone Howse). Le craton Supérieur a probablement formé la succession de base de la Fosse Labrador d'un bout à l'autre de ses évolutions, mais l'éruption des

dernier basalts du haut-Mg dans les terrains Doublet, associé avec l'emplacement de magmas de haut-Mg comme filons-couche, peut enregistrer la rupture de la croûte continentale. Alors la croûte continentale agit comme barrière de densité de produits primitifs fondus partiellement tard dans l'évolution de l'histoire de la Fosse Labrador. La transition de l'alkaline au magmatisme tholéïique dans chaque cycle volcano-sédimentaire démontre un signe d'augmentation progressive dans le degré de décomposition de produits fondus partiellement dans la remontée d'eau marine dans le manteau terrestre asthénosphérique hétérogène pendant l'atténuation crustale.

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Part I
Introduction and General Geology

Chapter 1

Introduction

1.1 General Introduction

The Labrador Trough is a linear belt of Early Proterozoic igneous and sedimentary rocks bordering the eastern margin of the Superior Province in north-eastern Québec and western Labrador. The belt is one of the most extensively studied of Canadian Proterozoic basins. This is largely due to the discovery of large deposits of iron ore in 1929 which provided impetus for subsequent exploration and regional mapping. A recent resurgence of mineral exploration in the Labrador Trough has been fuelled by the potential for precious metal mineralization, particularly in mafic and ultramafic intrusive rocks. A concomitant renewal of scientific interest reflects new interpretations of regional tectonic elements, additional mapping in hinterland regions, and increasing availability of geochronological data, which have thrown into question traditional views of tectonic setting and event timing.

The Labrador Trough was classically interpreted as a geosyncline developed in a continental setting. In this model, shallow water clastic and chemical sedimentary rocks in the western part of the belt represent miogeosynclinal deposits laid down on Archean basement, while the turbidite-basalt-gabbro assemblage in the east preserves the remnants of eugeosynclinal or ophiolitic rocks (Dimroth *et al.*, 1970). Comparable Proterozoic sequences are intermittently exposed along the northern and north-eastern margins of the Superior Province in the Cape Smith and Hudson Bay fold belts (Fig. 1.1). Lithological and stratigraphic similarities underlie the hypothesis that individual segments are remnants of a once nearly continuous basinal succession, the Circum-Ungava geosyncline (Dimroth *et al.*, 1970). Baragar and Scoates (1981) expanded this interpretation to include fold belts along the western and southern margins of the Superior Province which show some stratigraphic similarities with the northern belts, and are of approximately the same age. In

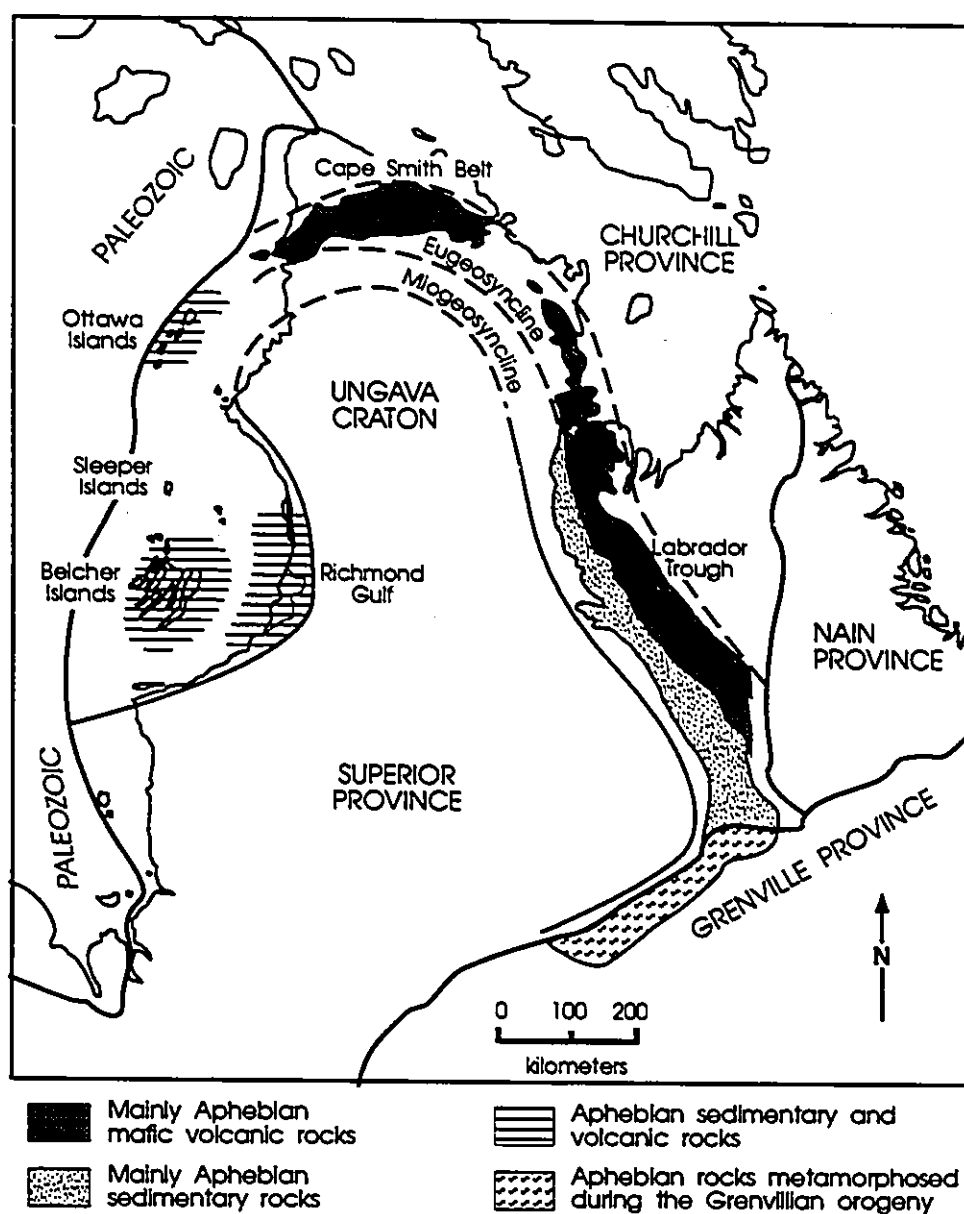


Figure 1.1. Location and subdivision of the principal fold belts of the Circum-Ungava geosyncline as defined by Dimroth *et al.* (1970).

particular, they noted the occurrence of Superior type iron formation, ultramafic flows and sills, and plagioclase-phyric or glomeroporphyritic gabbro in similar stratigraphic positions in many of the fold belts. These authors hypothesized that the basins were initiated by ensialic rifting along the margin of the Superior craton and subsequent development of an annular trough, the Circum-Superior Belt, in which initial sedimentation occurred.

Continued rifting and attenuation of sialic crust led to widespread mafic and ultramafic magmatism, but expansion of the northern rift zones ceased prior to development of a true oceanic basin.

In recent years, alternative and often conflicting tectonic interpretations have been advanced for some of the Circum-Ungava fold belts, including the Labrador Trough. Perhaps the most critical point of difference among the various models is the degree to which the Early Proterozoic basins were underlain by sialic crust throughout their development. Wilson (1968), Gibb and Walcott (1971) and Thomas and Gibb (1977) favoured models in which the Circum-Ungava fold belts represent a more or less continuous suture zone resulting from collisions between the Superior craton and Archean elements of the Churchill Province, following closure of an intervening Early Proterozoic ocean. In contrast, Hoffman (1987) hypothesized that the Labrador Trough and Hudson Bay segments preserve foredeep deposits laid down in flexural basins developed on the Superior craton. A common feature of these models is the early deposition of continental or shallow marine sediments on Archean basement. The models diverge, however, in the interpretation of the subsequent evolution of the basins, particularly in regard to the origin of the mafic magmatism so prevalent in the fold belts. The suture zone model, for example, interprets the igneous succession as oceanic crust, whereas the foredeep model requires that the basal sequences were deposited on continental crust. The nature and origin of mafic magmatism is a significant problem which must be addressed in any interpretation of the tectonic setting of the Labrador Trough.

1.2 Purpose

This study was initially directed at examining and interpreting glomerocryst distributions in the plagioclase-glomeroporphyritic to anorthositic gabbro sills (blotchy gabbro, leopard rock) of the central Labrador Trough. The sills are conspicuous and commonly spectacular members of a major igneous suite which dominates the upper part of the stratigraphic succession for much of the length of the belt. Baragar (1967) termed this succession the Labrador Magmatic Province. Because the current study deals

exclusively with the mafic intrusive and extrusive rocks of the Labrador Magmatic Province, which also includes peridotite, carbonatite and rhyolite, the basaltic rocks of the belt, including intrusive members, are here informally referred to as the Labrador Trough Basaltic Suite (LTBS).

Geochemical and textural similarities suggest that the glomeroporphyritic sills are petrogenetically linked to spatially associated aphyric sills and volcanic rocks of the LTBS. The project thus evolved to a geochemical study of the principal members of this magmatic province. In light of disparate tectonic models, the potential for insight into the source regions and setting of LTBS rocks through application of trace element and isotopic analysis was of particular interest. The principal objectives of the present study were therefore: (1) to determine the petrogenetic relationship between plagioclase glomerocrysts and gabbroic matrices of glomeroporphyritic sills, (2) to determine glomerocryst size and frequency distributions in the sills in order to examine the origins of glomerocryst zonation, (3) to determine the petrogenetic relationship between glomeroporphyritic and aphyric sills (4) to examine compositional variations in the various mafic suites to determine the nature of differentiation processes; (5) to compare the geochemical and isotopic characteristics of the principal members of the LTBS to determine their genetic relationships; and (6) to use the geochemical and isotopic signatures of the suite as an aid to interpretation of tectonic setting.

1.3 Methods

This study was initiated under a program of regional metallogenic studies in Labrador undertaken by the Geological Survey of Canada as part of the Canada-Newfoundland Mineral Development Agreement. Field mapping was conducted in the Howse Lake and Irene Lake areas of the central Labrador Trough (Fig. 1.2) to establish stratigraphic relations amongst various gabbroic sill types and host rocks. Four sills in the Howse Lake area were chosen for detailed mapping and sampling. A bulk sample for geochronological analysis was obtained from one glomeroporphyritic sill, also at Howse Lake. The results of the geochronological analysis are reported in Findlay *et. al.*

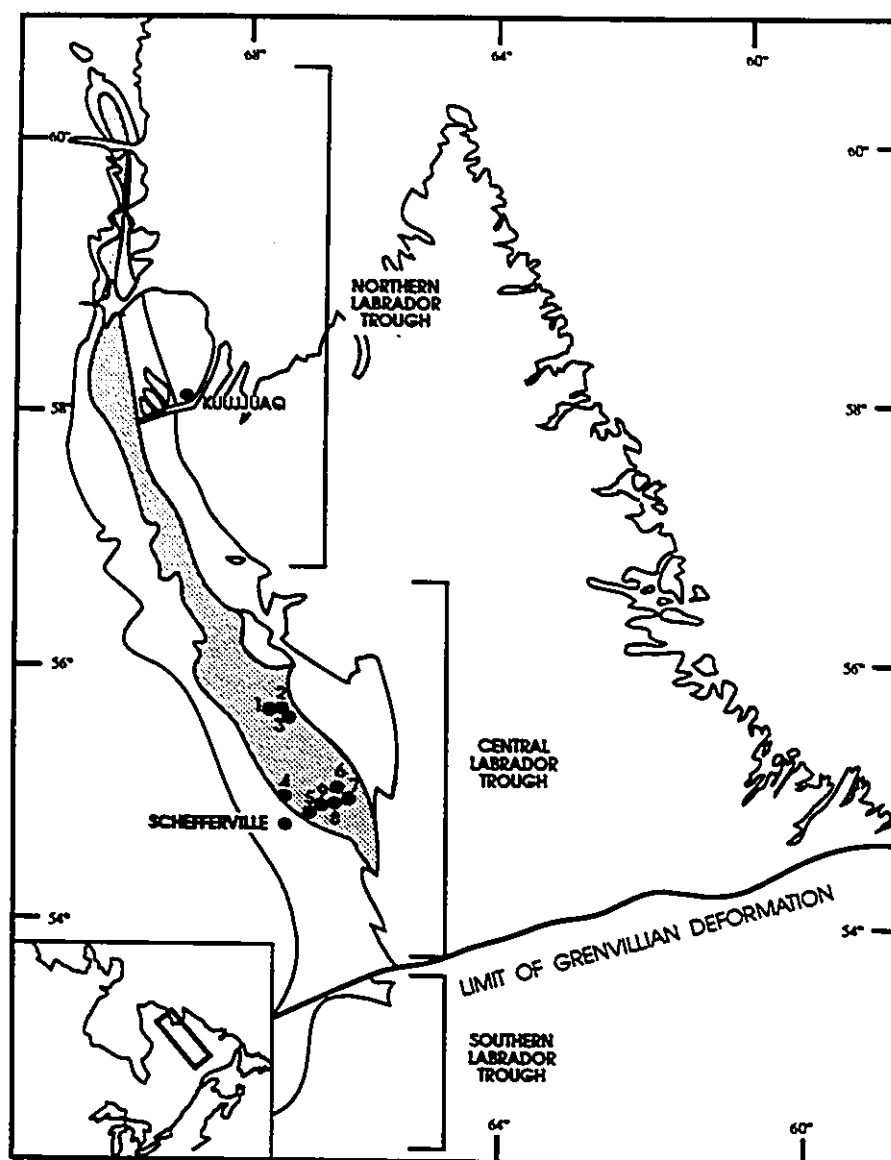


Figure 1.2. General areas of field work and sampling undertaken in this study. (1) Chassin-Outlook lakes, (2) Irene Lake, (3) Murdoch Lake, (4) Tait Lake, (5) Howse Lake, (6) Retty Lake, (7) McNeil Lake, (8) Doublet Lake and (9) Frederickson Lake. The dominantly igneous portion of the Labrador Trough is shaded. Regional subdivision is from Dimroth *et al.* (1970).

1995). Regional sampling of basaltic units was carried out in the Outlook, Murdoch, Tait, Frederickson, Retty, Doublet and McNeil lakes areas (Fig. 1.2). Each of these regions has been mapped in previous studies by various workers. Consequently, field work in the course of this study was focused primarily on lithological examinations and sample

collection. With few exceptions, no attempt was made to further refine stratigraphic or structural subdivisions.

Petrological descriptions are based on field observations and examination of approximately 400 thin and polished thin sections. Geochemical and other types of data were obtained using several analytical techniques. Glomerocryst size, shape and frequency distributions were determined by image analysis of serially sectioned samples using equipment and software made available by the Department of Biology, University of Ottawa. Electron microprobe analyses were obtained on an instrument operated by the Geological Survey of Canada. Major and trace element compositions of whole rock samples and mineral separates were determined by X-ray fluorescence (XRF) and inductively-coupled plasma atomic emission spectrometry (ICAP-AES) in the Department of Geology, University of Ottawa, and by instrumental neutron activation analysis (INAA) at the Institut de génie énergétique, École Polytechnique, l'Université de Montréal. Sm-Nd isotopic analyses were performed by Reginald Thériault of the Geochronological Laboratory of the Geological Survey of Canada. Rb-Sr and additional Sm-Nd analyses were performed by Dr. John Blenkinsop of the Department of Earth Sciences, Carleton University.

1.4 History of Investigation

The extensive literature on all aspects of the geology of the Labrador Trough precludes an exhaustive review, but a number of important contributions deserve mention in the context of this study. The history of early exploration in the region was documented by Le Bourdais (1957) who noted that the first geological survey in the region was conducted by A.P. Low of the Geological Survey of Canada in the years between 1872 and 1895. Much of the early mapping in the southern and central portions of the Trough was completed in the 1940s and 1950s by two related exploration companies, the Hollinger-North Shore Exploration Company Ltd. and the Labrador Mining and Exploration Company Ltd. This work was initially stimulated by the discovery, and subsequent development, of iron ore deposits in the Schefferville area. In addition, base

metal exploration programs were initiated by the two companies in the late 1930s and continued until 1986, when operations ceased. The field assets and much of the claim territory of these companies were subsequently acquired by La Fosse Platinum Group Inc., which conducted a resampling program targeting known massive sulphide showings. The announcement of anomalous platinum group element (PGE) concentrations by La Fosse in 1986 stimulated intense exploration activity by numerous companies in the late 1980s.

Government agencies also have a long history of activity in the Labrador Trough. Regional mapping programs pertinent to this study have been carried out by the Geological Survey of Canada (Baragar, 1958; 1967; Donaldson, 1966; Frarey, 1967; Wynne-Edwards, 1960; 1961), the Ministère de l'Énergie et des Ressources, Québec (Bérard, 1965; Clark, 1978; 1979; 1980; 1984; Dimroth, 1964; 1965; 1967; 1969; Dressler, 1979; Sauvé and Bergeron, 1965) and the Newfoundland Department of Mines (Evans, 1978; Wardle, 1976; 1979). Material from government and industry sources has been incorporated into regional compilation maps by Dimroth (1978), Wardle (1982) and Clark (1989), which together cover most of the Labrador Trough. These maps formed the basis for area selection in the regional sampling program completed in this study.

The stratigraphy, structure and evolution of the central portion of the Trough have been studied in considerable detail. Geological summaries are given by Frarey and Duffell (1960), Harrison *et al.* (1970), Dimroth (1978), Wardle and Bailey (1981), Le Gallais and Lavoie (1982) and Wardle *et al.* (1990a). The petrology of igneous rocks in the Labrador Trough has been described by Baragar (1960; 1967), Fahrig (1962), Dimroth *et al.* (1970), Dimroth (1978), Evans (1978), Wardle and Bailey (1981), Wares and Goutier (1989), and St. Seymour *et al.* (1991). Sm-Nd Isotopic results from the igneous rocks have recently been published by Rohon *et al.* (1993) and Skulski *et al.* (1993). Some of the interpretative models dealing with tectonic setting and evolution have already been mentioned; these models, and others, will be described in more detail in subsequent sections. It should be mentioned here, however, that the Labrador Trough is now considered by most workers to form part of an extensive Early Proterozoic orogenic system informally termed the Pan-American System (Lewry and Collerson, 1990). The dominant component of this system is the Trans-Hudson Orogen (Hoffman, 1981), the various components of which are described in detail in a comprehensive Geological Association of Canada special paper,

"The Early Proterozoic Trans-Hudson Orogen of North America" (Lewry and Collerson, 1990). The north-eastern segments of the Pan-American System, including the Labrador Trough, are similarly described in a special volume of Geoscience Canada in which the proceedings of a conference entitled "Recent Advances in the Geology of the Eastern Churchill Province" are published (Geoscience Canada, 1990).

The hypothesis that the glomeroporphyritic sills of the Labrador Trough formed by the mechanical process of plug flow was advanced by Baragar (1960). Flow differentiation during emplacement was also invoked as the origin of zoned ultramafic sills in the central Trough (Fahrig, 1962). The same process was postulated for the development of olivine rich zones in picritic sills of the Isle of Skye (Simkin, 1967; Drever and Johnston, 1967), and for lithological variation in the feeder dyke to the Muskox intrusion (Bhattacharji, 1967). Bagnold (1954) theorized that mechanical interactions played a role in the transport of stream sediments and derived empirical equations to describe the driving force of the process, which he termed grain dispersive pressure. Komar (1972a; 1972b; 1976) applied Bagnold's results to the case of mafic magmas carrying suspended crystal loads and showed theoretically that flow differentiation could result in the observed grain concentration profiles obtained from small dykes. Since Komar's work, the process has been the subject of little further investigation.

1.5 Course of Treatment

This study of the LTBS has two principal components: (1) an examination of the petrology, geochemistry and origin of the Howse Lake sills, and (2) an examination of the petrology, geochemistry and origin of the LTBS as a whole. The Howse Lake sills are predominantly gabbroic bodies which commonly exhibit spectacular plagioclase-glomeroporphyritic texture. In order to formulate a model for the origin of the texture, four sills in the Howse Lake area were mapped in detail. The remaining mafic igneous rocks of the LTBS were examined on a regional basis with the aim of establishing the geochemical characteristics of individual members. Part I of this study provides an overview of the geology of the Labrador Trough and a definition of the LTBS. Part II deals with the petrology and geochemistry of the Howse Lake sills, and the origin of the

glomeroporphyritic texture. Part III describes the petrology and geochemistry of the basaltic members of the LTBS, and possible genetic relationships among them. The final section examines the tectonic setting and evolution of the LTBS.

Chapter 2

General Geology

2.1 Regional Setting

Traditionally, the term "Labrador Trough" has been applied to a north-westerly trending Proterozoic fold and thrust belt situated along the north-eastern margin of the Superior Province. The term originated as a reference to the basinal setting of sedimentation but, as pointed out by Frarey (1967) and Hoffman (1988), the name is somewhat misleading in that the area is not a geographic or structural depression, and is situated largely in Québec rather than Labrador. Hoffman (1988) proposed "the Proterozoic foreland of the New Québec Orogen" as an alternative, whereas Wardle *et al.* (1990a) and Lewry and Collerson (1990) employed the term "the western division of the Labrador segment of the Trans-Hudson Orogen". The rationale behind both suggestions was, in part, the seeming inappropriateness of "Labrador Trough". The terms, however, also reflect the authors' interpretations of the relationship between the Labrador Trough (and adjacent terranes) to other tectonic associations formed in the Early Proterozoic. More recently, the Labrador Trough has been referred to simply as the New Québec orogen (e.g., Skulski *et al.*, 1993; Rohon *et al.*, 1993). This thesis adopts the terminology of Hoffman (1990a) and Lewry and Collerson (1990) for regional tectonic elements (e.g., Trans-Hudson Orogen), but retains "Labrador Trough" to refer to a specific assemblage of Early Proterozoic rocks isolated from adjoining elements of the New Québec orogen. Despite its geographical unsuitability, the name has the advantage of brevity over other terminology, as well as a long standing history of usage. Given the speculative nature of orogenic boundaries and interconnections, the term has the additional benefit that it presupposes no genetic relationships.

Stockwell's (1964) definition of the Churchill Structural Province was based on a common period of deformation and metamorphism which he termed the Hudsonian Orogeny. The increasing level of detail in geological mapping and the widespread

application of precise geochronology has led to the recognition of the fundamental duality of the Churchill Province (cf. Hoffman, 1990a). It includes a variety of variably metamorphosed, volcano-sedimentary fold belts of Early Proterozoic age, as well as adjacent metastable or remobilized Archean terranes which often contain erosional remnants of infolded and highly deformed Proterozoic cover. This recognition led to the subdivision of the Churchill Province into a series of remobilized Archean terranes and intervening orogenic belts. One such belt, the Trans-Hudson Orogen (Hoffman, 1981; 1990a; 1990b), encompasses all of the fold belts of the Circum-Ungava Geosyncline of Dimroth *et al.* (1970). The Trans-Hudson Orogen (Fig. 2.1) is itself part of a larger Early Proterozoic orogenic association, the Pan-American System (Lewry and Collerson, 1990).

The extent and subdivision of the Trans-Hudson Orogen in the northern and eastern segments are the subjects of some debate. As generally defined, the orogen extends northward from South Dakota between the Wyoming and Superior cratons, and curves north-eastward across the Hudson Bay basin between the Superior craton and remobilized Archean crust of the Hearne Province. At this point, interpretations of the extent of the system, and the locations of collisional sutures, diverge. Lewry and Collerson (1990) suggest that it continues through the Cape Smith Belt, across Baffin Island and into Greenland. These authors argue that an arm of the orogen extends southward between the Superior and Nain cratons (Fig. 2.1A), and the high grade gneissic hinterland east of the Labrador Trough represents partially dismembered segments of the Nain craton remobilized during Superior-Nain collision. Hoffman (1988; 1990a; 1990b) identified this high grade terrane as a southern splay of the Rae Province, and postulated the existence of two orogens, the New Québec and Torngat, between the Superior craton and Rae Province, and Rae Province and Nain craton, respectively (Fig. 2.1B). While the hinterland to the Labrador Trough has yet to be conclusively identified as part of the Rae or Nain provinces, the terminology of Hoffman (1990b) and Wardle *et al.* (1990b) will be followed here. Irrespective of the provenance of the hinterland, the Labrador Trough represents a foreland thrust and fold belt bounded by undisturbed Superior crust to the west, and by remobilized Archean crust, infolded Proterozoic supracrustal rocks and orogenic granites to the east.

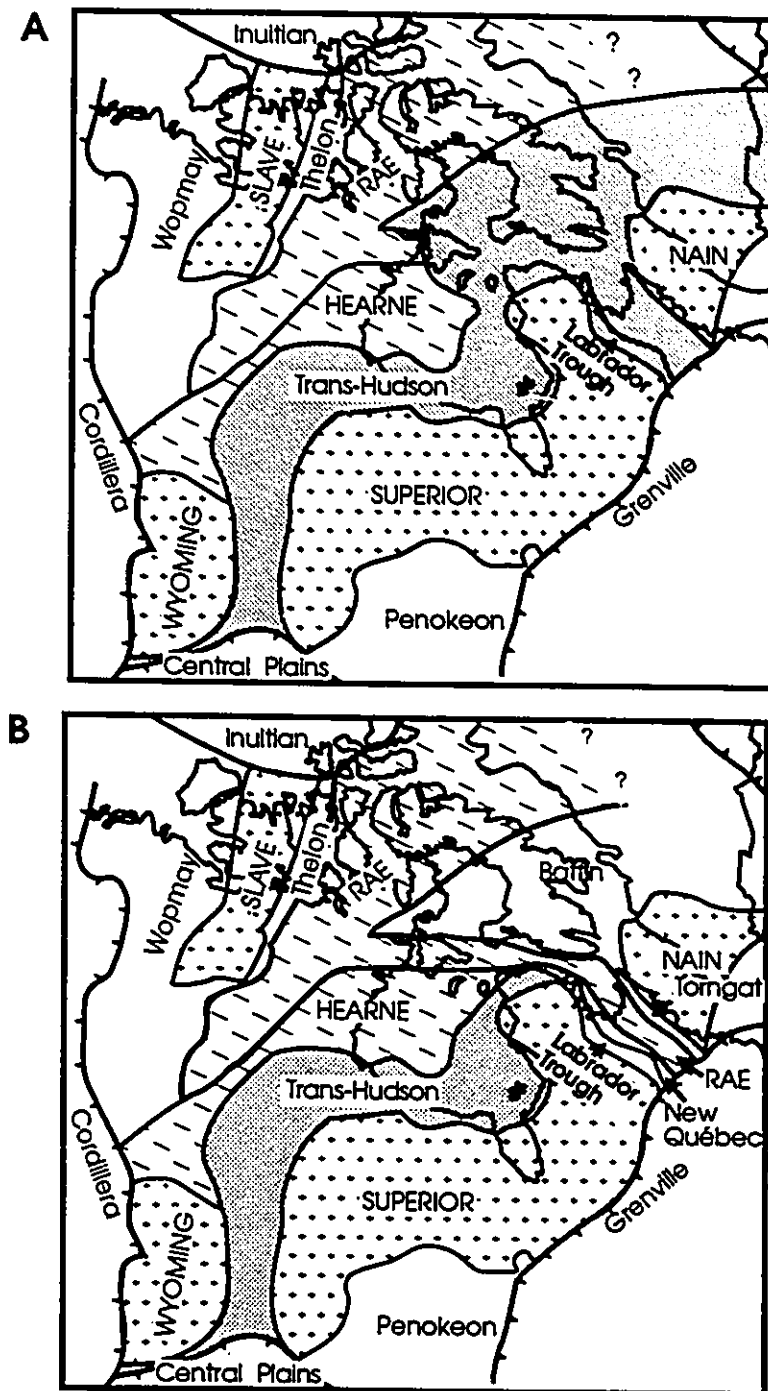


Figure 2.1. Principal components of the Pan-American orogenic system and the Trans-Hudson Orogen after (A) Lewry and Collerson (1990) and (B) Hoffman (1990a). Uppercase print - structural province; lowercase print - orogenic belt.

2.2 Stratigraphy and Structure of the Central Labrador Trough

The Labrador Trough was subdivided into southern, central and northern segments (Fig. 1.2) by Dimroth *et al.* (1970). The central Trough contains the most complete and least deformed stratigraphic sequence, while much of the lower portion (cycle 1; see below) of the stratigraphic succession is absent in the northern Trough. The rocks of the southern Trough have undergone relatively intense metamorphism and deformation during Grenvillian orogenesis. The central Trough exposes rocks of the Kaniapiskau Supergroup, which consists of four principal lithostratigraphic units: the Knob Lake, Doublet, Laporte and Montagnais groups. A simplified table of formations is provided in Table 2.1, and a summary of the principal components of the Kaniapiskau Supergroup follows. This summary is drawn primarily from the works of Dimroth (1978), Wardle and Bailey (1981) and Wardle *et al.* (1990a; 1990b). For detailed lithological and stratigraphic descriptions these publications should be consulted.

Table 2.1. Simplified table of formations for the central Labrador Trough. Modified after Wardle and Bailey (1981).

Kaniapiskau Supergroup	Montagnais Gp.			Doublet Gp.	Laporte Gp.	
	Tamarack R. Fm.					
	Unconformity ?					
	Upper Knob Lake Gp.	Fenniman subgroup	Menihok Fm.			
			Sokoman/Nimish Fm.			
			Wishart Fm.			
	Unconformity					
	Lower Knob Lake Gp.	Attikamagen subgroup	Fleming/Dolly Fm.			
			Denault Fm.			
			Le Fer/Bacchus Fm.			
		Swampy Bay subgroup				
		Pisoleit subgroup				
		Seward subgroup	Du Portage/Sawyer Lake Fm.			
	Chakanipau Fm.					
	Unconformity					
	Archean Basement (Ashuanipi Complex)					

Archean gneissic and granitic rocks of the Superior craton (Ashuanipi metamorphic complex) form the western basement to the Early Proterozoic rocks of the Knob Lake Group in the central Trough (Fig. 2.2). The oldest members of the Knob Lake sequence

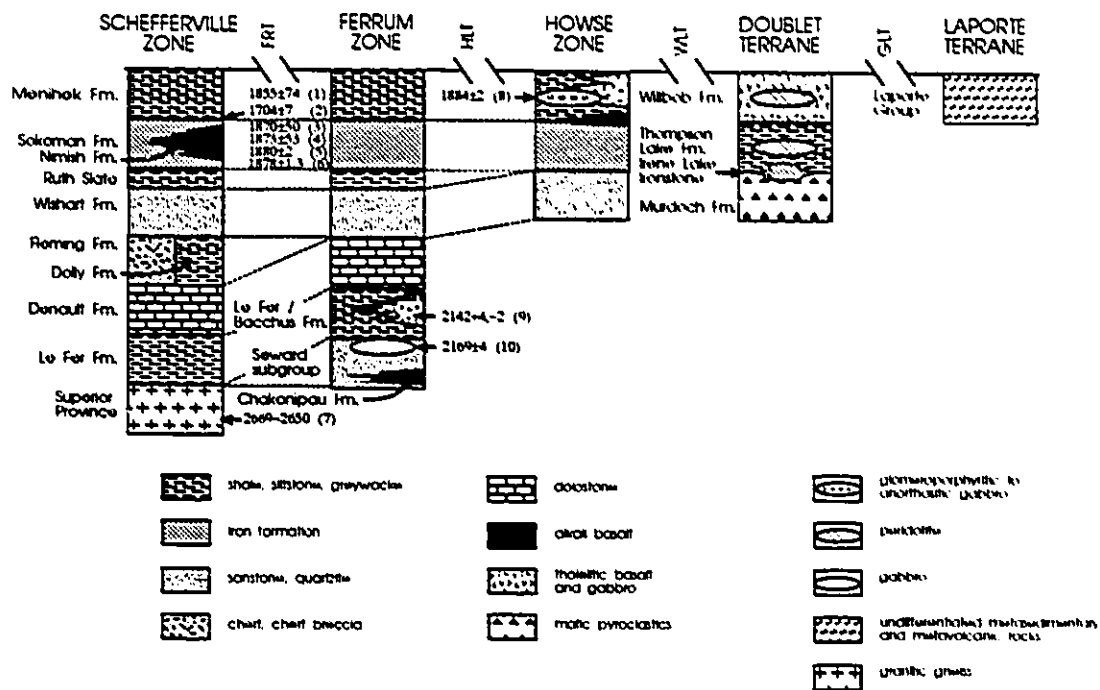


Figure 2.2. Generalized stratigraphy of the central Labrador Trough from west to east (unit thickness arbitrary). Names in uppercase print refer to lithotectonic domains (see text). FRT-Ferrum River Thrust, HLT-Howse Lake Thrust, WLT-Walsh Lake Thrust, GLT-Gill Lake Thrust. Pertinent ages in millions of years are indicated (see section 2.3 for discussion). Sources: (1) and (3) Fryer (1972), (2) Bell and Thorpe (1986), (4) Dressler (1975), (5) Chevé and Machado (1988), (6) and (8) Findlay *et al.* (1995), (7) Mortensen and Percival (1987), (9) Clark (1984), (10) Rohon *et al.* (1993).

consist of continentally derived sandstone, arkose and pebble conglomerate of the Seward subgroup, which unconformably overlies the Ashuanipi complex. Strata of the Seward subgroup are conformably overlain by moderate to deep water marine siltstone-shale and basalt of the Le Fer Formation, which is known as the Bacchus Formation in the north-central Trough. This unit is in turn overlain by shallow water stromatolitic carbonates of the Denault Formation followed by the Fleming chert breccia. The Fleming Formation is transitional eastward to a deeper water facies known as the Dolly shale. Dolostone, arkose and turbidites of the Pistolet and Swampy Bay subgroups represent second order basin sediments deposited between the Seward and Attikamagen subgroups in the north-central Trough (Wardle and Bailey, 1981). The Seward subgroup and Le Fer, Denault, Fleming and Dolly formations of the lower Knob Lake Group preserve an early

second depositional cycle (upper Knob Lake Group) by a regional unconformity (cf. Dimroth *et al.*, 1970; Le Gallais and Lavoie, 1982). In some localities, however, there is an apparent gradational transition from the Denault dolostone, through the Fleming chert breccia, to the Wishart quartzite (Birkett *et al.*, 1991). This may indicate that the two cycles are separated by local unconformities which are diachronous throughout the central Trough.

The upper Knob Lake Group is a transgressive marine sequence which progressively overlaps earlier sediments. The succession commences with mature sandstone and arkose of the Wishart Formation which overly the Denault Formation at the western margin of the central Trough and locally rest directly on Archean basement. Superior type iron formation (Sokoman Formation) conformably overlies the Wishart Formation and is in turn overlain by turbidites and basalts of the Menihek Formation. The Sokoman Formation includes a basal carbonaceous shale member (Ruth shale), and mildly alkaline basalts of the Nimish Formation are intercalated with the ironstone in the south-central Trough. Volcanic rocks of this suite have also been identified at the base of the Menihek Formation in the Andre Lake area (Wardle, 1979). The Nimish Formation includes a distinctive conglomerate which contains cobbles of Nimish basalt, Sokoman ironstone and syenite. The syenitic cobbles are geochemically comparable to the felsic volcanics of the Nimish suite, and are thought to be comagmatic. (Findlay *et al.*, 1995). If this is the case, then the conglomerate represents the product of local uplift and erosion during deposition of the Sokoman Formation and eruption of the Nimish volcanics.

The Menihek Formation is overlain by the Tamarack River Formation, a sequence of arkose, dolostone and argillite present only in the south-central Trough. This unit was originally considered part of the Menihek Formation by Fahrig (1967), but the presence of upper Knob Lake clasts in arkosic rocks of the formation led Wardle and Bailey (1981) to suggest the unit was a syntectonic molasse, possibly formed during the initial stages of Hudsonian uplift and deformation. Because of the uncertain relationship with underlying rocks, the formation has not been included in the Knob Lake Group in Table 2.1.

The Doublet Group is in fault contact with the Knob Lake Group along the Walsh Lake thrust. The stratigraphically lowest rocks of this sequence are those of the Murdoch

Formation, an assemblage predominantly composed of mafic schists, mafic pyroclastic rocks, basalts and argillites. The overlying siltstones of the Thompson Lake Formation are commonly pyritic, and minor lenses of ironstone occur at the Murdoch-Thompson Lake contact near Irene Lake (Baragar, 1967). The relationship between the Murdoch and Thompson Lake formations is generally considered conformable (Baragar, 1967; Frarey, 1967), but the contact is unexposed. The Willbob Formation, a thick sequence of massive to pillowed basalt, caps the Doublet Group. This formation includes plagioclase glomeroporphyritic flows which are similar to those of the upper Menihek Formation.

The Gill Lake thrust forms the boundary between the Doublet and Laporte groups, the latter sequence occupying the easternmost regions of the central Trough. The Laporte Group consists largely of metasedimentary and metavolcanic rocks interpreted by van der Leeden *et al.* (1990) as typical of deposits formed in distal, deep water basins. Relationships with other Labrador Trough sequences are poorly constrained, but Dimroth (1978) identified metasedimentary lithologies which he considered equivalent to strata of the Seward subgroup, Le Fer/Bacchus Formation and Murdoch Formation in the vicinity of the Wheeler basement dome. Dimroth (1978) concluded that the Laporte Group was largely equivalent to the Knob Lake Group. Metabasalt and metagabbro of the Laporte Group have tholeiitic geochemical characteristics typical of ocean floor basalts (Wardle and Bailey, 1981; van der Leeden *et al.*, 1990). The Laporte Group strata have undergone increasing degrees of metamorphism and deformation along the eastern boundary of the terrane and are gradational into amphibolite to granulite grade gneisses of the remobilized Archean hinterland. This is particularly true in the vicinity of the Lac Tudor shear zone, a complex belt of deformation involving lithologies of both the Laporte Group and hinterland gneiss (van der Leeden *et al.*, 1990). Wardle *et al.* (1990b) have speculated that the Lac Tudor shear zone is part of the suture zone(s) between the Rae Province and the Labrador Trough and underlying Superior basement.

The term Montagnais Group has been used to refer to a sequence of mafic and ultramafic rocks which intrude the sedimentary strata of the Knob Lake and Doublet groups. The intrusions are generally sill-like bodies, and cross-cutting relationships with the host sediments are relatively uncommon. The Montagnais intrusions have been subdivided on the basis of predominant composition into two groups, the Wakuach Gabbro

and the Retty Peridotite. The Wakuach Gabbro consists largely of aphyric to plagioclase-glomeroporphyritic sills in which ultramafic differentiates are relatively uncommon. These sills intrude both Knob Lake and Doublet group sediments where they constitute a significant proportion of the total stratigraphic thickness. In contrast, the ultramafic sills of the Retty Peridotite are commonly well differentiated into peridotite, pyroxenite and gabbro (Beaudoin and Laurent, 1989), and were emplaced only within the strata of the Doublet Group.

Wardle *et al.* (1990a) have interpreted the overall structure of the central Labrador Trough as that of a west-verging thrust stack with a narrow autochthonous zone preserved in the western portion of the foldbelt between the Stakit Lake thrust and Superior basement (Fig. 2.3). These authors interpret the Stakit Lake thrust as a basal décollement marking the westward approach of the allochthonous thrust package. Variations in the lithological assemblage, metamorphic grade and tectonic style led Wardle *et al.* (1990a) to subdivide the central Trough into four discrete lithotectonic domains delineated by thrust faults. From west to east these are the Schefferville and Howse zones, and Doublet and Laporte terranes. Based on observations in the Howse and Outlook Lakes areas as described below, this thesis adopts a five-fold subdivision. From west to east these are the Schefferville, Ferrum and Howse zones, and Doublet and Laporte terranes (Fig. 2.3). The delineating structures are the Ferrum River, Howse Lake, Walsh Lake and Gill Lake thrust faults. The partially autochthonous Schefferville zone is dominated by a series of imbricate thrust faults rooted in the Stakit Lake thrust. The structural style of the allochthonous thrust slices to the east is dominated by long wavelength folding. Wardle *et al.* (1990b) suggest this change is due to the deep-rooted nature of the Ferrum River, Walsh Lake and Gill Lake thrusts, which these authors suggest extend to the basement-cover interface. Metamorphic grade increases eastward from sub-greenschist in the Schefferville, Ferrum and Howse zones, through greenschist in the Doublet terrane, and reaches amphibolite facies in the Laporte terrane.

Changes in structural style and metamorphic grade in the lithotectonic domains are accompanied by considerable variation in lithological compositions. As indicated by their names, the Doublet and Laporte terranes expose rocks of the Doublet and Laporte groups respectively. The Schefferville, Ferrum and Howse zones are underlain by rocks of the

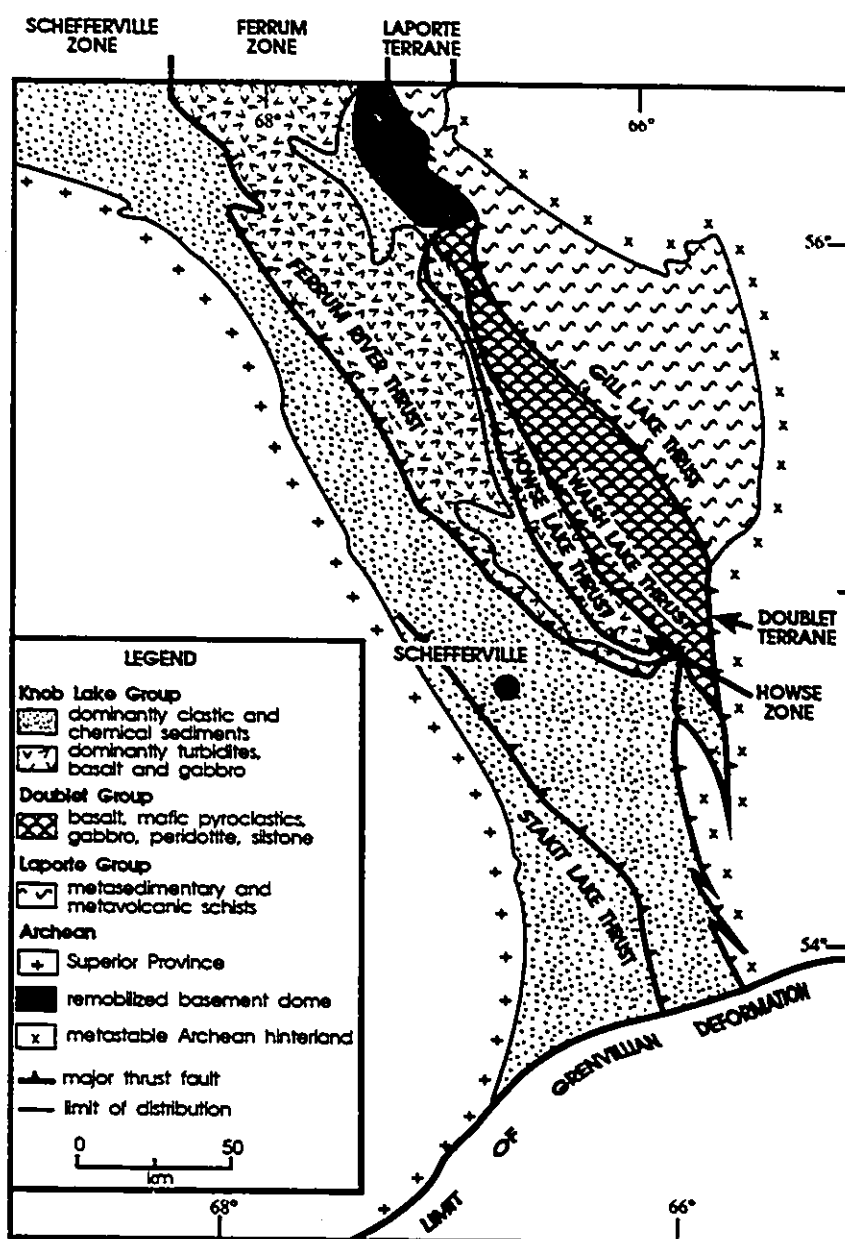


Figure 2.3. Lithotectonic subdivision of the central Labrador Trough. Modified after Wardle *et al.* (1990a).

Knob Lake Group, with significant facies changes evident from one zone to the next (Fig. 2.2). The most prominent change relates to the type and abundance of igneous rocks in the domains. In the Schefferville zone, the principal episode of magmatic activity produced the localized and relatively limited amounts of Nimish volcanic and intrusive rocks. The volcanic rocks range from basalt to trachyte and rhyolite, while the intrusive rocks

emplaced in the Wishart and Sokoman formations range from peridotite and pyroxenite to diorite in composition (Evans, 1978; Watanabe and Fowler, 1989). As noted, syenitic cobbles in the Nimish conglomerates probably represent unroofed and eroded Nimish stocks, and presumably these were the intrusive equivalents of the trachyte and rhyolite flows. Magmatism in the Schefferville zone was thus largely confined to localized eruption and emplacement of mafic to felsic Nimish melts during deposition of the upper Knob Lake Group.

The Ferrum River thrust marks the transition to deeper water sedimentary conditions which prevailed in the Ferrum and Howse zones. With the exception of the Seward subgroup, the shallow water clastic and chemical sediments which characterize the Schefferville zone (e.g., Denault, Dolly, Fleming, Wishart and Sokoman formations) are reduced in thickness or wedge out completely moving eastward, and the rhythmically bedded nature of the greywacke-siltstone-shale sequences of the Le Fer and Menihek formations indicate a turbidite origin. The greater thickness of the Seward subgroup in the Ferrum zone has been attributed to its deposition in a narrow rift basin largely coincident with this domain in the initial stages of sedimentation (Wardle and Bailey, 1981)

A more prominent difference between the stratigraphic successions of the Schefferville and Ferrum/Howse zones is the widespread appearance of gabbro sills and basalts to the east of the Ferrum River thrust. The intrusions occur throughout the Knob Lake Group strata, but are concentrated within the Le Fer (cycle 1) and Menihek (cycle 2) formations. The principal volcanic horizons similarly appear in the Le Fer and Menihek formations. Since the lower and upper Knob Lake group are separated by an unconformity, the two basaltic sequences represent distinct magmatic episodes.

Cycle 1 basalts and associated gabbros are confined to the Ferrum zone of the central Labrador Trough. The basalts occur throughout the distribution of the Le Fer Formation in the Ferrum zone, both in a geographic and stratigraphic sense, although the thickest volcanic horizons appear to be localized in the upper part of the sequence. In contrast, the principal basaltic member of the Menihek Formation occurs at the top of the sequence in a narrow band adjacent to the Walsh Lake thrust. Cycle 2 gabbroic sills, including plagioclase-glomeroporphyritic to anorthositic varieties, also appear to be confined to this band. The Menihek Formation to the west of the most westerly

glomeroporphyritic sill is poorly exposed and the area contains numerous elongate lakes coincident with a series of faults extending most of the length of the Howse zone (cf. Baragar, 1967; Wardle, 1982). The Wishart-Sokoman-Menihek sequence is locally repeated to the east of this fault zone. Based on these features, it is probable that eastern portion of the Howse zone of Wardle *et al.* (1990a) represents a discrete tectonic slice with a distinct lithological assemblage (i.e., cycle 2 aphyric and glomeroporphyritic gabbros and Menihek basalts). As previously indicated, this area is referred to as the Howse zone in this thesis. The boundary between the Howse and Ferrum zones is thought to be a thrust fault (Howse thrust) coincident with the concentration of faults in the Menihek Formation which mark the most westerly occurrence of the cycle 2 glomeroporphyritic sills (Fig. 2.3).

The basaltic rocks of the Le Fer and upper Menihek formations, which constitute the most significant successions of extrusive rocks in the Knob Lake Group, were classified as a tholeiitic suite by Baragar (1967) and Dimroth (1978). The voluminous sills intruding these formations were also classed as tholeiitic by these authors and considered comagmatic with the spatially associated extrusions. Minor amounts of alkaline rocks, however, also occur in the succession. Slightly alkaline (Dimroth, 1978; Clark, 1984) trachybasaltic flows and pyroclastic rocks (Chakonipau Formation; Fig. 2.2) are intercalated with clastic sediments at the base of the Seward subgroup in the Lac Musset area (Baragar, 1967). Alkaline mafic volcanic and pyroclastic rocks also occur in the lower Menihek Formation immediately east of the Howse Lake thrust (Findlay *et al.*, 1989).

The Walsh Lake thrust marks the boundary between the Doublet terrane and the Howse zone. The stratigraphy of the Doublet terrane is dominated by mafic volcanic and pyroclastic rocks, some of which may be correlative with the mafic rocks of the Menihek Formation in the Howse zone. Clastic sediments, however, are much less abundant and, with the possible exception of the Irene Lake ironstone, there are no chemical sedimentary rocks equivalent to those of the Knob Lake Group. The Doublet terrane is in fault contact with the Laporte terrane along the Gill Lake thrust. As noted earlier, specific correlations between the Laporte Group and the Doublet or Knob Lake groups are poorly known due to the higher degrees of metamorphism and deformation in the former. Dimroth (1978), however, identified several lithologies which he considered correlative with certain strata

of the upper Seward subgroup, the Le Fer/Bacchus Formation and the Murdoch Formation. If these correlations are correct, the Laporte Group is essentially the stratigraphic equivalent of the Knob Lake Group.

2.3 Age Relationships

The Knob Lake Group can be divided into upper and lower members separated by a regional unconformity or a series of less extensive local unconformities. This fact, coupled with the repetitive and transgressive nature of sedimentation, has been used as evidence for the presence of two distinct depositional cycles. The lower Knob Lake Group corresponds to cycle 1 sedimentation, while the upper Knob Lake Group was formed during the second depositional cycle (Dimroth, 1978; Le Gallais and Lavoie, 1982). Cycle 1 began with deposition of the Seward subgroup upon basement rocks of the Ashuanipi Complex (Superior craton). In a U-Pb isotopic study of the Ashuanipi complex, Mortensen and Percival (1987) concluded that felsic plutonism occurred during granulite facies metamorphism between 2669 and 2650 Ma, with local retrogression after 2642 Ma. The latter date places a rough limit on the absolute age of the oldest sedimentary rocks of the Labrador Trough.

The only precise age available for cycle 1 sedimentation and volcanic activity was obtained from rhyolite of the Mistamisk Formation in the Ferrum zone, which provided a U-Pb zircon date of 2142 ± 2 Ma (Dressler and Krogh, cited by Clark, 1984). Clark and Thorpe (1990) have suggested that the Mistamisk Formation is correlative with the upper part of the Bacchus Formation, and therefore 2142 Ma is the minimum age of Le Fer/Bacchus volcanism. Unfortunately, there are no reported instances of rhyolite in the Le Fer Formation, and the volcanic rocks of the latter examined in this study are homogeneously basaltic. A better estimate for the age of Le Fer magmatism is provided by the U-Pb zircon date of 2169 ± 4 Ma obtained from a gabbroic sill emplaced within sediments of the Seward subgroup near Lac Cramoilet (Rohon *et al.*, 1993). Although there have been no detailed geochemical studies of Le Fer basalts or the gabbroic sills, Rohon *et al.* (1993) concluded that the Lac Cramoilet sill was coeval with the Le Fer/Bacchus basalts. The 2169 Ma age of the intrusion is significantly older than the 2142 Ma age of

the Mistamisk rhyolite. If the Mistamisk-Le Fer correlation suggested by Clark and Thorpe (1990) is correct, and if the Cramolet Lake sill is related to Le Fer volcanism, then this magmatic episode continued over an interval of approximately 30 Ma. Alternatively, the Mistamisk rhyolite may represent a later magmatic episode, possibly related to uplift during the 260 Ma interval between cycle 1 and 2 tholeiitic magmatism.

Fryer (1972) obtained whole rock Rb-Sr ages of 1870 ± 50 Ma and 1855 ± 74 Ma for the cycle 2 Sokoman and Menihek formations respectively. An age of 1877.8 ± 1.3 Ma was obtained by Findlay *et al.* (1995) for syenitic cobbles in conglomerate of the Nimish Formation. The cobbles are interpreted as erosional remnants of felsic members of the Nimish igneous suite and consequently 1878 Ma is also the age of the Sokoman Formation in the Schefferville zone. Comparable estimates for the age of the Sokoman Formation were obtained from the Castigon Lake carbonatite complex of the north-central Trough which includes pyroclastic members interbedded with the ironstone. Dressler (1975) reported a K-Ar age of 1873 ± 53 Ma for a lamprophyre dyke related to the complex and also interpreted to be of Sokoman age. Similarly, Chev  and Machado (1988) obtained a U-Pb zircon date of 1880 ± 2 Ma for a carbonatite dyke forming part of the complex. The similarity of these estimates indicates that the 1704 ± 7 Ma Pb-Pb age reported by Bell and Thorpe (1986) for phosphorites at the base of the Menihek Formation in the Schefferville zone does not reflect primary deposition.

The age of cycle 2 glomeroporphyritic gabbro is given by the U-Pb zircon date of 1884 ± 2 Ma derived from an intrusion at Howse Lake in the Howse zone (Findlay *et al.*, 1995). The sill was emplaced within turbidites of the Menihek Formation and hence provides a minimum age for deposition of the Menihek strata in the Howse zone. The fact that the Nimish suite (and therefore the Sokoman Formation) formed at 1878 Ma in the Schefferville zone confirms the transgressive nature of the Menihek turbidites. The local occurrence of glomeroporphyritic texture in Menihek basalts (Dimroth, 1978), and the geochemical similarities between basalts and sills (Baragar, 1960) suggest that the two are genetically linked. If this is the case, then basaltic volcanism of the first and second sedimentary cycles are separated by an interval of approximately 260 Ma.

The rocks comprising the Doublet Group are generally unsuitable for precise geochronological work using available methods, and therefore timing of sedimentation and

igneous activity in this terrane is problematic. Baragar (1967) noted the textural similarities between glomeroporphyritic rocks of the Howse zone (upper Menihek Formation) and Doublet terrane (Willbob Formation), and concluded they were comagmatic. Limited support for temporal equivalence of upper Menihek and Willbob magmatic activity is found in the 1885 ± 67 Ma Pb-Pb age derived from the latter basalts by Rohon *et al.* (1993). The large analytical uncertainty associated with this date, however, prevents positive correlation. The same authors obtained a Pb-Pb date of 1930 ± 49 Ma for cycle 2 gabbroic rocks. Although the two Pb-Pb ages are equivalent within error, there is considerable room for temporal variation in the data. Potential correlations and temporal relationships among the Knob Lake and Doublet group strata are addressed in subsequent chapters.

No geochronological data are available for the Laporte Group, and the higher grade of metamorphism and greater deformation makes lithological comparisons with the western sequences difficult. Dimroth *et al.* (1970) considered the Laporte Group to be younger than the rest of the Labrador Trough sequence. Subsequent work led Dimroth (1978) and Wardle and Bailey (1981) to conclude that this sequence is, at least in part, the lateral equivalent of the Knob Lake Group, with the apparent younger stratigraphic relationships resulting from westward overthrusting onto the Doublet Group. The similarity in geochemical characteristics of the Laporte metabasalts and the basaltic rocks of the Knob Lake Group supports this conclusion, although the absence of demonstrably equivalent marker horizons introduces some ambiguity to proposed correlations.

The youngest available date from the Labrador Trough sequence is that of 1870 ± 4 Ma obtained from zircons in a rhyodacite in the Lac Le Moyne area of the north-central Trough (Machado *et al.*, 1990). The stratigraphic position of the rhyodacite is poorly constrained, although Dressler (1979) considered it to be part of the Murdoch Formation. This date is equivalent, within error, to the U-Pb zircon date of 1874 ± 3 Ma obtained by Machado (1990) from glomeroporphyritic gabbro of the northern Labrador Trough. The glomeroporphyritic sill was emplaced near the top of the Hellancourt Formation of the northern Trough, and this basaltic sequence is generally considered correlative with the Willbob Formation of the central Trough (cf., St. Seymour *et al.*, 1991; Skulski *et al.*, 1993). Given that glomeroporphyritic gabbro in the central Trough is 1884 Ma, and

assuming the equivalency of this lithology in the two regions, then cycle 2 magmatism continued over an interval of at least 10 Ma. It is therefore possible that the Lac Le Moyne rhyodacite is the product of the waning stages of cycle 2 magmatism. Alternatively, this activity may be related to the initial stages of uplift and deformation accompanying continental collision.

The timing of Hudsonian deformation and metamorphism is bracketed by the 1870 Ma date from the Lac Le Moyne rhyodacite, and a mean age of 1830 Ma for metamorphic zircons in the Kuujuaq Batholith of the northern Trough hinterland (Perreault and Hynes, 1990). Perreault and Hynes (1990) also reported U-Pb dates of 1845-1833 Ma for igneous zircons in the batholith. These authors interpret these data to reflect the emplacement of the batholith subsequent to granulite facies metamorphism in the northern hinterland since mineral assemblages appropriate to this metamorphic grade are absent from the intrusion. The 1830 Ma mean age of metamorphic zircons was thought to reflect a second metamorphic event which reached amphibolite facies conditions at about 1830 Ma (Perreault and Hynes, 1990). If these interpretations are correct, Hudsonian metamorphism reached granulite facies conditions between 1870 Ma and 1845 Ma, and this initial metamorphic episode was overprinted by a second deformational event at *ca.* 1830 Ma.

2.4 Tectonic Models

The tectonic setting of the sedimentary and volcanic rocks of the Labrador Trough has been the subject of study for some fifty years. It is generally accepted that the succession reflects cyclic sedimentation in an essentially marine basin. In the western, predominantly sedimentary region there is evidence for derivation of clastic detritus from a westerly provenance in the initial stages of a cycle, and from both west and east during pelitic sedimentation in the later stages (cf. Dimroth *et al.*, 1970; Wardle and Bailey, 1981; Le Gallais and Lavoie, 1982). It is also accepted that during the first cycle the basin was relatively restricted in size, with the continental sandstones and alkaline volcanic rocks of the Seward subgroup deposited in a narrow central depression along the eastern margin of the Superior Province. Increasing subsidence and probable widening of the basin provided the environment for deposition of the Pistolet and Swampy Bay subgroups. Subsidence

culminated with accumulation of the deep water siltstones and shales of the Le Fer Formation. The return to relatively shallow water conditions over a broader area led to the precipitation of the Denault dolostone which caps the first depositional cycle. In the central Trough, cycle 2 sedimentation commenced with deposition of the Wishart quartzites which onlapped the Superior basement and extended northward (Allison Formation of the northern Trough).

There are two principal points of contention regarding the evolution of the Labrador Trough, and both stem from interpretations of the rocks preserved in the eastern part of the belt. The first pertains to the spatial extent of the basin; i.e., do the Labrador Trough strata preserve an essentially *in situ* basin, or were parts of the sequence transported from primary depositional sites to the east of the present boundaries? The second problem is indirectly related to the first. Was the entire Labrador Trough sequence deposited on sialic basement, or did a continental margin rift develop into a true, albeit perhaps narrow, oceanic basin? In a comprehensive review of the stratigraphy and evolution of the Trough, Dimroth *et al.* (1970) interpreted the belt as an ensialic rift which developed into a geosyncline along the eastern margin of the Ungava craton (Superior Province; Fig. 2.5A). In this model, the western, predominantly sedimentary portion of the belt preserves miogeosynclinal rocks, and the turbidites and mafic igneous rocks to the east represent eugeosynclinal deposits (ophiolites). The axis of the developing basin was thought to have migrated westward during infilling thus causing onlapping of cycle 2 sediments. Continued rifting led to broadening and lengthening of the basin, and medial upwelling split the rift into shallow western and deeper eastern subbasins. The metasedimentary and metavolcanic schists (Laporte terrane) that border the eugeosynclinal belt were not considered part of the geosynclinal suite. Dimroth *et al.* (1970) also proposed that the Abner, Baby and Hellancourt formations in the northern Trough, and the Irene Lake ironstone, Thompson Lake Formation and Willbob Formation in the central Trough, formed a third depositional cycle stratigraphically above the upper Knob Lake Group. The entire sequence was thought to represent *in situ* deposition in a basin which subsequently underwent only minor shortening during orogenesis.

Baragar and Scoates (1981) and Wardle and Bailey (1981) also preferred an *in situ* origin for the Trough sequence. Both models proposed rifting of the passive margin of the

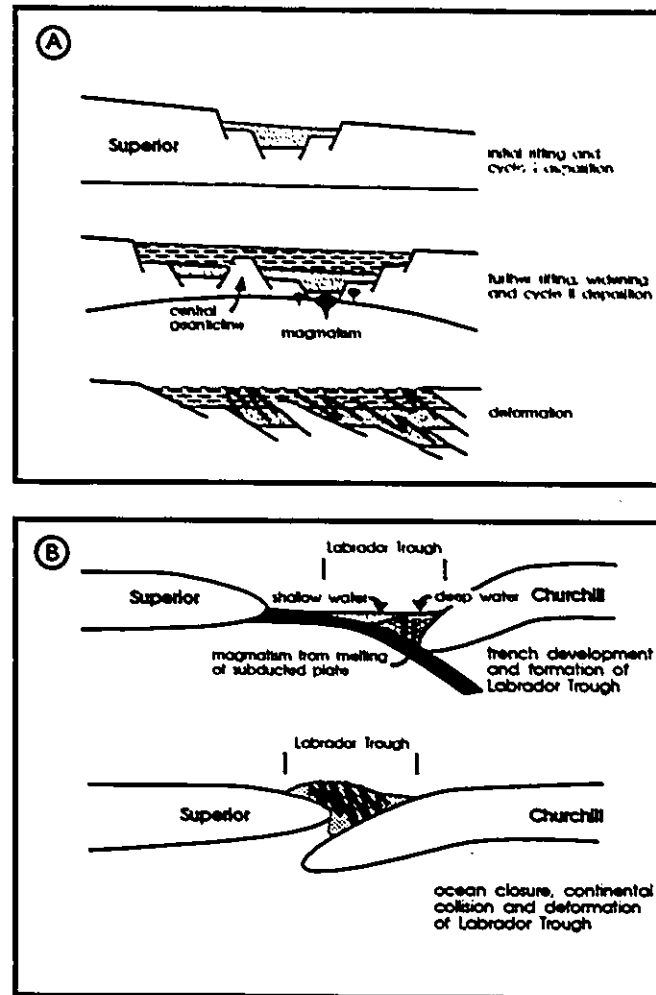


Figure 2.5. Schematic illustration (not to scale) of tectonic models proposed for the evolution of the Labrador Trough. (A) Ensisialic geosyncline (Dimroth *et al.*, 1970). (B) Oceanic basin with subduction and continental collision (Gibb *et al.*, 1978).

Superior province as the mechanism of basin initiation. The development of the sedimentary succession was thought to occur generally as envisaged by Dimroth *et al.* (1970). Baragar and Scoates (1981) suggested that continued rifting led to attenuation of sialic crust and subsidence of the rift basin. Basaltic magmas were produced by partial melting of upwelling mantle material, and this upwelling eventually led to rupture of the continental crust. These authors speculated that incipient oceanic crust was formed, but further opening was halted when spreading in the oceanic basin along the southern margin of the Superior Province assumed the total plate movement involved in the rifting process.

A similar evolution was postulated by Wardle and Bailey (1981) with the exception that rifting was slow and sporadic, and the locus of rifting migrated eastward during much of cycle 1 and 2 sedimentation. During the final stages of the second depositional cycle, more rapid rifting engendered rapid subsidence and eventually voluminous basaltic magmatism (Willbob and Hellancourt formations). In both models, the absence of true oceanic crust (i.e., ophiolites) was explained by the failure of the embryonic oceanic rift to further develop. Le Gallais and Lavoie (1982) concurred with the incipient oceanic rift model but noted that although the cycle 2 turbidite-gabbro-basalt assemblage was best interpreted as the product of oceanic rifting, it is allochthonous and was transported westward and thrust over continental shelf deposits.

An alternate interpretation was proposed by Wilson (1968) and Gibb and Walcott (1971) who suggested that the Circum-Ungava Proterozoic belts, including the Labrador Trough, represent collision zones between the Superior Province and Archean microcontinents of the Churchill Province. The Proterozoic belts were thought to be oceanic deposits which were thrust onto the Superior Province during collision. Dewey and Burke (1973) adopted the collisional model and interpreted the western sedimentary belt of the Trough as continental shelf deposits, and the eastern igneous sequence as obducted oceanic crust containing upthrust mantle slices represented by the Retty Peridotites. According to these authors, not only were the belts developed in a passive margin-oceanic basin setting rather than an ensialic rift, but much of the preserved strata were transported to their present positions during ocean closure and collision.

Support for the collision hypothesis came from gravity profiles across the Labrador Trough and adjoining terranes which suggest that the Archean foreland and hinterland were not equivalent in terms of gravity signatures, as might be expected if the Trough was an ensialic basin. Kearey (1976, 1977), Gibb *et al.* (1978) and Thomas and Kearey (1980) proposed that the Trough preserves oceanic trench deposits formed adjacent to the Churchill proto-continent (Fig. 2.5B). The trench was thought to have developed as a result of subduction of oceanic crust beneath the Churchill Province. Early deposits were laid down in shallow water during the initial stage of trench formation. Continued subduction led to a deeper trench and hence deeper water sedimentation. The cyclic nature of sedimentation was explained by periodic halting of the subduction process, which led to

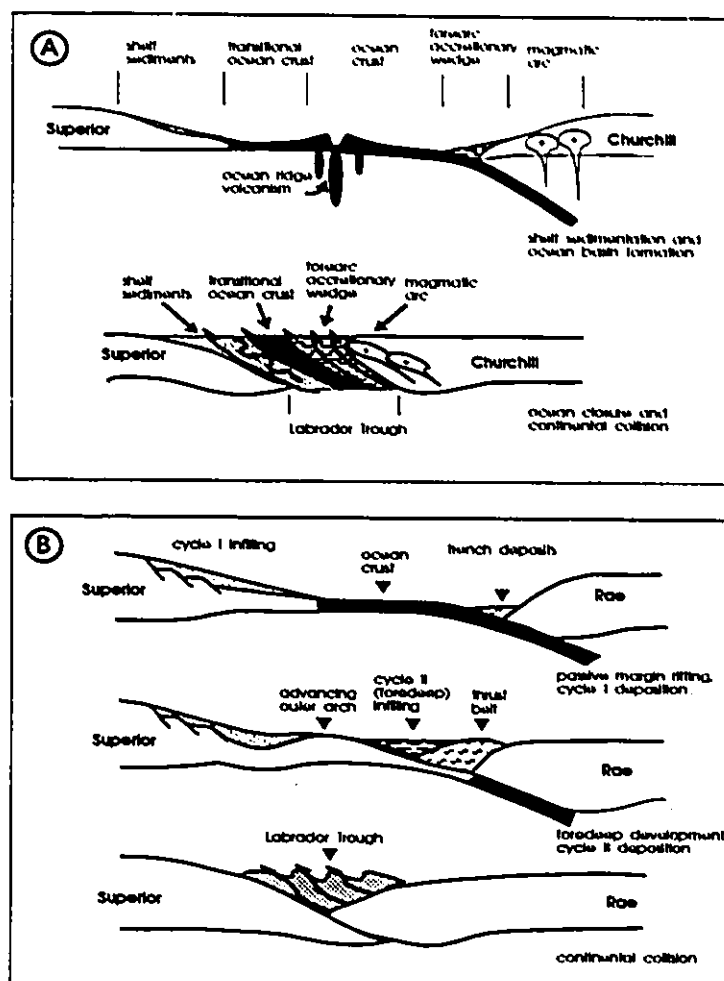


Figure 2.6. Schematic illustration (not to scale) of tectonic models proposed for the evolution of the Labrador Trough. (A) Oceanic basin with subduction, arc development and continental collision (van der Leeden *et al.*, 1978). (B) Foredeep (Hoffman, 1987; 1988).

reduction of water depth through infilling. In this model, the thick igneous succession was derived from melting of the down-going oceanic slab.

Recent work in the hinterland of the Labrador Trough has led to a slightly more complicated tectonic model of ocean closure and continental collision (Fig. 2.6A). van der Leeden *et al.* (1990) have suggested that granitic batholiths emplaced within gneissic rocks in the western part of the Churchill hinterland (Rae Province; Hoffman, 1990a) represent a fossil magmatic arc developed over a subduction zone in which oceanic crust was subducted beneath the Rae Province. In this model, the eastern volcano-sedimentary belt

represents early or transitional oceanic rift deposits, whereas true oceanic crust has been removed by subduction. van der Leeden *et al.* (1990) interpreted the Laporte terrane as a fore-arc accretionary wedge derived from the down-going oceanic slab.

Hoffman (1987, 1988) also argued the allochthonous nature of much of the Trough sequence, but postulated that significant portions of the stratigraphy were deposited in one, or possibly two, foreland basins or foredeeps rather than in an ensialic or oceanic rift basin. Development of the foredeeps was considered to result from flexure of the margin of the Superior Province in response to loading along the leading edge of the craton. Flexure would result in a foredeep-outer arch pair which would move westward with progressive loading during encroachment of overriding thrust sheets from the east; the thrust sheets presumably represented by the schistose rocks of the Laporte terrane (?) and adjacent hinterland gneiss. Migration of the outer flexural arch would cause progressive uplift and possibly erosion of earlier platformal sediments, while foredeep migration resulted in a transgressive marine succession.

According to the foredeep model, cycle 1 sedimentation (Seward, Pistolet, Swampy Bay and Attikamagen subgroups) records rifting and subsidence of an east-facing, passive continental margin (Fig. 2.6B). The unconformity between the Denault and Wishart formations of the central Trough represents uplift associated with the passage of the outer arch across the platform. Cycle 2 was subsequently deposited in the prograding foredeep, with shallow to moderate water sequences laid down on the outer ramp, and turbidites developed in the axial zone. Hoffman (1987, 1990a) suggested that the mafic igneous succession associated with turbidites may represent either axial zone foredeep magmatism in front of the encroaching thrust belt, or supra-subduction volcanism behind the thrust belt. In the latter case, Hoffman (1990a) proposed that the volcano-sedimentary belt was subsequently thrust over foredeep deposits, and noted that the volcanic rocks should display geochemical characteristics consistent with subduction related magmatism. Hoffman (1987) also raised the possibility that the cycle 1 sub-basin deposits of the Pistolet and Swampy Bay subgroups were also laid down in a foredeep. Wardle *et al.* (1990a), however, noted that the absence of cycle 1 strata in the northern Trough, the lack of an obvious early thrust belt or evidence of a deformational episode, and the fact that no intra-cycle 1 unconformity has been recognized make a second foredeep unlikely.

Rohon *et al.* (1993) interpreted the large ion lithophile (LIL) element enriched nature of the Willbob basalts as evidence of formation in a back-arc environment, with magmatism presumably related to subduction processes. This interpretation is in accord with the supra-subduction zone model of Hoffman (1990a) in which volcanism occurred in pull-apart basins developed above an east-facing subduction zone. Skulski *et al.* (1993), however, concluded that there was no geochemical evidence from cycle 2 volcanics in the northern Trough for subduction related magmatism. These authors suggest that the alkaline geochemical compositions of early cycle 2 volcanism (Nimish Formation) reflect derivation from subcontinental mantle, with thinning and possibly rupture of the overlying Superior crust in the late stages of magmatism leading to voluminous and widespread eruption of tholeiitic basaltic. Cycle 2 sedimentation and volcanic activity was thought to have occurred in pull-apart basins developed on Superior continental crust, possibly in response to oblique convergence of the Superior Province and southern arm of the Rae Province prior to continental collision. The absence of cycle 1 strata in the northern Trough was explained by the presence of a crustal promontory in the north. Evolution of the Labrador Trough was thought to have culminated with the collision between the Superior Province and the southern splay of the Rae Province (Skulski *et al.*, 1993).

From the preceding discussion it is apparent that opinion is divided regarding the tectonic setting and evolution of the Labrador Trough. Structural studies have demonstrated that much of the succession is allochthonous and has been transported westward, probably as a result of continental collision. As noted earlier, one of the principal differences among tectonic models revolves around an ensialic versus oceanic depositional environment. Of particular importance to these arguments is the interpretation of the basalt-gabbro-turbidite sequences of the two depositional cycles. The geochemical nature of the mafic rocks has generally been described as typical of ocean floor basalts, and therefore inconsistent with magmatism in an intracratonic basin, continental rift or foredeep environment. More recent analytical work suggests that the mafic rocks have transitional geochemical affinities which are distinct from modern MORB (Rohon *et al.*, 1993; Skulski *et al.*, 1993). Arguments against an oceanic setting include the lack of demonstrably ophiolitic sequences, the presence of internal continental basement domes, and the transitional MORB geochemistry of the basaltic rocks. Several authors, however,

have countered these arguments by suggesting that the lack of ophiolite structure and transitional geochemical signatures are due to the embryonic nature of the oceanic rift.

Speculation regarding tectonic setting is further complicated by the relatively unknown nature of the eastern boundary of the Labrador Trough. Only recently have detailed studies been conducted in the metamorphosed and highly deformed hinterland, and it is as yet unclear what position these rocks fill in the overall tectonic environment. Even the relationship between the relatively undeformed sequences of the Trough and the schistose rocks of the Laporte terrane is somewhat obscure. It is unlikely that a comprehensive tectonic model can be developed without conclusive resolution of the setting of the tholeiitic suite, determination of the relationship between the Laporte Group and less deformed rocks to the west, and an understanding of the relationship between the Trough succession and the gneissic terranes to the east.

2.5 Summary

The Labrador Trough is an Early Proterozoic fold and thrust belt which forms the foreland of the New Québec Orogen, which is itself part of the Trans-Hudson orogenic system. The central Labrador Trough consists of an autochthonous western sequence overlain by a series of allochthonous thrust slices derived from the east. Five lithotectonic subdivisions delineated by thrust faults can be recognized on the basis of lithology, metamorphic grade and structural style. The westernmost domain (Schefferville zone) is characterized by shallow water clastic and chemical sedimentary rocks and limited quantities of mafic to felsic alkaline igneous rocks. Moving eastward, the allochthonous assemblages (Ferrum and Howse zones) are increasingly dominated by strata formed in a deeper water basinal environment and mafic igneous rocks. The Walsh Lake thrust marks the appearance of a succession, dominated by mafic and ultramafic igneous rocks (Doublet Group), which is not readily correlated with those of the more westerly domains (Knob Lake Group). The easternmost lithotectonic subdivision of the central Trough is characterized by amphibolite grade metasedimentary and metavolcanic schists (Laporte Group) whose relationships to the lower grade rocks to the west are also uncertain. The hinterland to the central Labrador

Trough is formed by high grade gneissic rocks of the Rae Province which locally adjoins the Laporte terrane along the Lac Tudor shear zone. In other areas, the Laporte Group is gradational into the hinterland gneisses through increasing metamorphism and deformation.

The Knob Lake Group preserves two sedimentary cycles, both of which include thick successions of mafic igneous rocks. Deposition of cycle 1 rocks began between 2642 Ma and 2169 Ma, the latter date reflecting the age of cycle 1 gabbroic intrusions. Cycle 1 sedimentation may have continued until 2142 Ma, which is the age of Mistamisk/Le Fer rhyolitic volcanism. Cycles 1 and 2 are separated by an unconformity which reflects a significant depositional hiatus. Emplacement of gabbroic rocks related to the second sedimentary cycle had begun by 1884 Ma and continued until at least 1874 Ma, the two dates representing the ages of glomeroporphyritic gabbro in the central and Northern Troughs respectively. Cycle 2 sedimentation ceased and Hudsonian orogenesis began between 1870 Ma and 1845 Ma, the ages of the youngest igneous rocks of the central Trough and post-metamorphic granite respectively. A second metamorphic event overprinted initial deformation at *ca.* 1830 Ma.

The Labrador Trough preserves the remnants of a larger marine basin which developed along the eastern margin of the Superior Province. The tectonic setting of the Trough succession is somewhat controversial. Most workers agree that initial sedimentation occurred within a basin situated on sialic crust, but interpretations of subsequent evolution diverge markedly. The principal models involve deposition in: (1) an ensialic rift basin developed on the passive margin of the Superior craton, (2) an oceanic basin formed between the Superior and Rae provinces, (3) a foredeep basin developed on the margin of the Superior Province in response to encroachment of and easterly-derived fold and thrust belt in front of the converging southern arm of the Rae Province, (4) a back-arc basin situated above a subduction zone, and (5) pull-apart basins formed on the margin of the Superior craton in response to convergence of the southern arm of the Rae Province. In all models, collision between the Superior and Rae Provinces led to significant lateral shortening of the Trough sequence through folding and overthrusting.

Chapter 3

The Labrador Trough Basaltic Suite: Definition and Distribution

3.1 Definition

Mafic igneous rocks constitute a significant proportion of the stratigraphic succession of the Labrador Trough. In the central Trough, the principal intervals of mafic rocks occur in the Le Fer, Menihek, Murdoch and Willbob Formations. Despite variations in age and spatial distribution, these intervals show many similarities. Each consists of interstratified basalts, mafic pyroclastic rocks and moderate to deep water marine sediments. Each has been intruded by gabbroic sills, many of which are indistinguishable from the thicker flows. With few exceptions, the extrusive rocks of each sequence are basaltic and exhibit similar petrological characteristics, a fact noted by Dimroth (1978, p.158) who observed "All these basalts are lithologically indistinguishable". The gabbroic sills emplaced within each sequence have similar geochemical characteristics, although some show evidence of *in situ* crystal accumulation. The primary lithological differences among individual intervals arise from variation in the relative proportions of the main rocks types, and the occurrence of other types of igneous rock. Several such differences are notable, and pre-eminent among these is the predominance of mafic pyroclastic rocks in the Murdoch Formation, and the occurrence ultramafic sills in the Doublet terrane. Also conspicuous is the occurrence of strongly alkaline basalt and mafic tuff at the base of the Menihek Formation.

This study deals primarily with the petrogenesis of the mafic volcanic rocks of the Le Fer, Menihek, Murdoch and Willbob formations, and the spatially associated intrusions. Minor intervals of mafic igneous rocks are locally present in the Chakonipau and Nimish formations. The Chakonipau volcanics were excluded from the study because they have a limited geographic distribution, are poorly exposed, and are highly altered and therefore unsuitable for geochemical work. The Nimish rocks were not sampled in the present study because they are the subject of an ongoing study by Watanabe (in preparation.) Data from

this sequence, however, has been provided by Watanabe (personal communication, 1993) such that comparisons with the Nimish rocks have been included in several of the interpretative sections.

The extrusive rocks of the Le Fer, Menihek, Murdoch and Willbob formations, and the associated gabbroic sills, constitute a major igneous succession here referred to informally as the Labrador Trough Basaltic Suite (LTBS). This term is introduced to identify specific intrusive and extrusive mafic rocks in the central Trough which are the focus of the study, and to distinguish these from other igneous rocks (e.g., carbonatites, rhyolites) whose stratigraphic relationships are unknown. Consequently, usage of the term is one of convenience and carries no petrogenetic implications. For example, the metabasalts of the Laporte terrane may be genetically related to the mafic rocks of the LTBS. But the metamorphism and deformation to which these rocks have been subjected, and the paucity of pertinent petrological and geochronological data, precludes correlations. Hence these rocks are not included in the LTBS as defined here.

The basalts of the LTBS are conveniently grouped according to the formations in which they occur. Subdivision of the associated gabbros is less obvious. Under existing stratigraphic nomenclature all gabbroic sills are included in the Wakuach Gabbro of the Montagnais Group (Frarey and Duffel, 1964). Baragar (1967) used a subdivision of the Wakuach Gabbro based primarily on petrological features. Gabbros found to the west of the Walsh Lake thrust were termed glomeroporphyritic or normal depending on the presence or absence of plagioclase glomerocrysts. Gabbros to the east of the thrust, across which metamorphic grade increases to greenschist facies, were termed metagabbro. Metagabbro was considered the metamorphosed counterpart of normal gabbro (Baragar, 1960; 1967).

Geochronological data confirm the existence of at least two, temporally distinct, sets of gabbroic sills intruding the Knob Lake Group (Rohon *et al.*, 1993; Findlay *et al.*, 1995). The first set of sills was emplaced at approximately 2170 Ma and is spatially associated with volcanic rocks of the Le Fer Formation. The second set was emplaced at approximately 1880 Ma and is spatially associated with the Menihek volcanics. This information invalidates the inclusion of all gabbros under the single term Wakuach Gabbro. Accordingly, sills which occur west of the Walsh Lake thrust are subdivided on

the basis of temporal and stratigraphic relationships into the Le Fer and Menihék gabbros. The metagabbros of Baragar (1960) constitute a structurally bounded set of intrusions which were emplaced within the strata of the Doublet Group. Because no geochronological data are available for these sills it is unknown whether they represent more than one episode of gabbro generation. Consequently no further subdivision is presently possible, and these sills are referred to as the Doublet gabbros.

3.2 Distribution and Areas of Sampling

Samples from the eight members of the LTBS were obtained from the general areas illustrated in Figure 1.2 and described below. Individual sample locations are indicated on larger scale maps of each area examined in the course of regional sampling, which are provided in Appendix A1. The locations of the Menihék sills sampled in detail at Howse Lake are indicated in Figures 4.1 and A1.1. The locations of individual samples from which geochemical and lithological data were obtained for the Menihék sills are denoted by stratigraphic height above the basal contact in Appendices A3, A4 and A6.

Fer Basalts

The Le Fer Formation of the Ferrum zone includes significant quantities of basalt, particularly in the northern half of the zone (Fig. 3.1). The volcanic rocks are interbedded with turbidites, and the formation is intruded by a profusion of gabbroic sills. Discrimination between intrusive and extrusive rocks is often difficult due to the paucity of diagnostic textures in exposed sequences. Based on examinations of several thick flows, Dimroth (1978) concluded that many of the gabbroic rocks of the formation were of extrusive origin. Given the intimate association of the two rock types, and the large areas they cover, only detailed mapping could resolve individual sills and flows and avoid misidentification. Consequently, it is likely that Figure 3.1 is an under-representation of the distribution of Le Fer basalts, while the opposite is true for the Le Fer gabbros.

Baragar (1967) noted the presence of basalts at several levels of the Le Fer (formerly Attikamagen) Formation, and identified a 300 m thick sequence near the base of

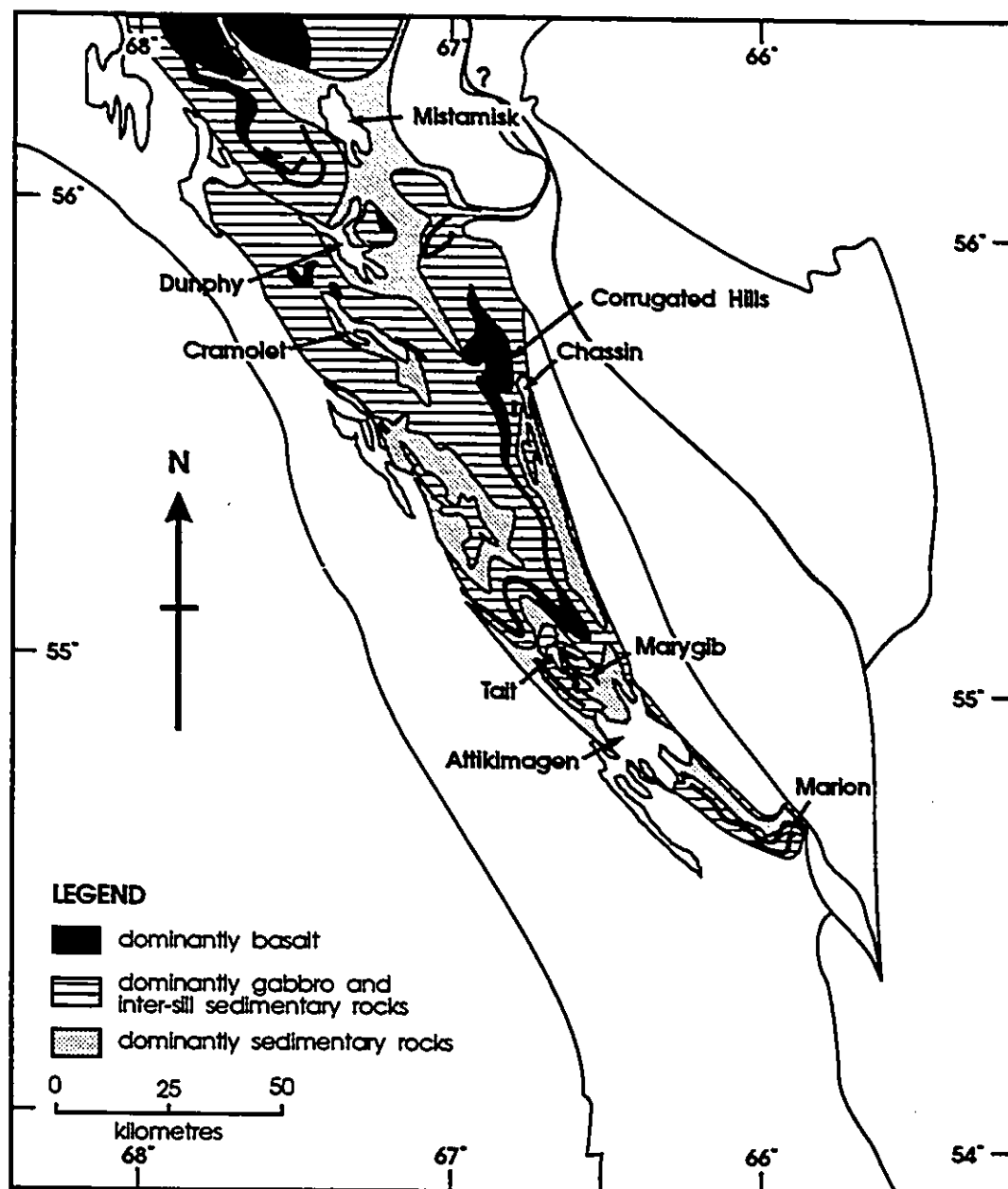


Figure 3.1. General distribution of the Le Fer Formation and Le Fer gabbros in the Ferrum zone of the central Labrador Trough. Names refer to geographical locations mentioned in the text.

the succession in the Tait Lake area. This sequence, which is developed near the contact with the underlying Seward sandstones, was considered the major volcanic member of the formation. In the Cramoiet-Dunphy lakes region to the north of Tait Lake, Dimroth (1978) subdivided the Le Fer Formation into three principal members, the central of which consists predominantly of basalts. Dimroth (1978) also mapped a 1600 m thick basaltic sequence in the north-central Trough (Mistamisk Formation) which he considered correlative with either the upper Le Fer or the upper Menihek Formations. Subsequent geochronological work led Clark and Thorpe (1990) to correlate the Mistamisk basalts with the upper part of the Le Fer Formation. The presence of a significant interval of basalts near the top of the Le Fer Formation is consistent with observations in the Marion Lake area where 800 m of pillow basalt was identified near the top of the Le Fer (formerly Attikamagen) Formation (Wardle, 1979). In light of these observations it seems probable that basalts were erupted during the entire interval of Le Fer deposition, although such eruptions were likely episodic.

The best exposures of Le Fer basalt occur in the Mistamisk Lake area where Dimroth (1978) estimated the aggregate thickness of basalts to be approximately 1600 m, and in a region north-west of Chassin Lake known as the Corrugated Hills where at least 1500 m of basalt are present (Fig. 3.1). Part of the Corrugated Hills sequence extends southward in a more or less continuous band, roughly 2 km in width, to Attikamagen Lake. The total stratigraphic thickness of basalts in the Le Fer Formation is impossible to determine due to structural complexity, the ubiquity of inter-flow material, and the problem of extrusive versus intrusive origins for gabbro sheets. Available estimates suggest that the formation includes an aggregate thickness of at least 1000 m of basalt in the southern half of the Ferrum zone. In the northern half of the zone, the minimum thickness of the basaltic component of the formation is in the range of 1500-3000 m, depending on the relative stratigraphic position of the upper (Mistamisk) volcanics.

Sampling of Le Fer basalts was carried out in the Corrugated Hills west of Chassin Lake, and across an anticlinal structure between Tait and Marygib lakes (Fig. 3.1). Large areas of basalt are exposed in a broad anticline developed a few kilometres west of Chassin Lake. The core of the structure is composed of a thick sequence of Le Fer shales, while the uppermost flows are separated from dolostone of the Denault Formation by only a few

metres of gabbro. Consequently the flows present in the area are part of the upper volcanic member of the Le Fer Formation. The Tait Lake anticline is cored by Seward sandstones, and basalts exposed in this structure are thought to belong to the lower volcanic member as described by Baragar (1967). In both locations, specimens were collected from aphanitic or fine grained flows to ensure a sample suite representative of the extrusive rocks of the formation.

Le Fer Gabbros

The gabbroic sills of the Ferrum zone intrude strata of both the Seward subgroup and the Le Fer Formation, but are concentrated in the latter. Basalt and gabbro are intimately associated in the Le Fer Formation, with virtually every volcanic horizon bounded by, or in close proximity to, gabbroic sheets. The sills are locally in structural contact with dolostone of the Denault Formation, but intrusive contacts with the latter have not been recognized. This indicates that the Le Fer gabbros were emplaced prior to deposition of the dolostone. The sills are thus linked to Le Fer volcanism by their spatial distribution and stratigraphic position (i.e., pre-Denault).

Gabbro, as mapped, constitutes a much greater stratigraphic thickness and covers a much larger area in the Ferrum zone than either basalts or sedimentary rocks (Fig. 3.1). The distribution of the sills is similar to that of the basalts in that they are most abundant in the northern part of the zone, whereas sediments predominate in the south. A general exception appears to be the abundance of gabbro and paucity of basalt in the western part of the domain. This feature may reflect the difficulty of distinguishing the two rock types rather than a true difference in their distributions. Alternatively, because the overall younging direction of the Le Fer Formation is to the east, the pattern could reflect a primary stratification with basalts concentrated in the upper parts of the sequence, but gabbros present throughout. This interpretation is consistent with the greater thickness of basalt identified at the top of the Le Fer Formation by Dimroth (1978) and Wardle (1979). As with the Le Fer basalts, the total thickness of gabbro is difficult to estimate, but these rocks may constitute as much as 70% of the stratigraphy in the northern part of the Ferrum zone.

Despite the widespread distribution and the often well exposed nature of the Le Fer gabbros, it was generally difficult to establish sill contacts. In part this reflects the regional scale of the sampling program undertaken in this study; individual sills were not examined in detail as was the case for the Menihek gabbros. But the lithological similarities and close association of basalt and gabbro, the often massive character of both rock types, and the poorly exposed nature of intervening sediments all contributed to the problem of distinguishing sills and flows. Intrusive contacts in which quenched gabbro passed into subophitic, medium grained gabbro within a few metres of the contacts were observed in two sills. From these intrusions, one located to the north-west of Chassin Lake and the other in the Tait Lake area, nine samples were obtained.

Menihek Basalts

Volcanic rocks form part of the Menihek sequence only in the Howse zone of the central Trough and, as with the Le Fer basalts in the Ferrum zone, the Menihek basalts are most prominent in the northern half of the domain. The volcanic rocks occur at two stratigraphic levels in the formation. The main volcanic member occurs at the top of exposed portion of the formation adjacent to the Walsh Lake thrust (Fig. 3.2). It is approximately 1000 m thick (Baragar, 1967), and extends more or less continuously from the northern end of the zone to Kathleen Lake, a distance of about 45 km. Small lenses occur sporadically for a further 50 km southward to Martin Lake near the southern end of the domain. These lenses are generally less than 100 m thick, have strike lengths of a few kilometres at most, and occur at the same stratigraphic level as the upper volcanic sequence.

The lower basaltic horizon was observed only in the Howse Lake area, where the flows are intercalated with pyroclastic rocks and volcanoclastic sediments. The aggregate thickness of the Howse Lake volcanic sequence is estimated to be approximately 100 m, of which basalt comprises somewhat less than half. The sequence outcrops close to exposures where shale and red cherty iron formation are finely interbedded. Similar interstratification in the Schefferville area was interpreted by Gross and Zajac (1983) as a gradational contact between the Sokoman and Menihek Formations. Hence the Howse Lake volcanic

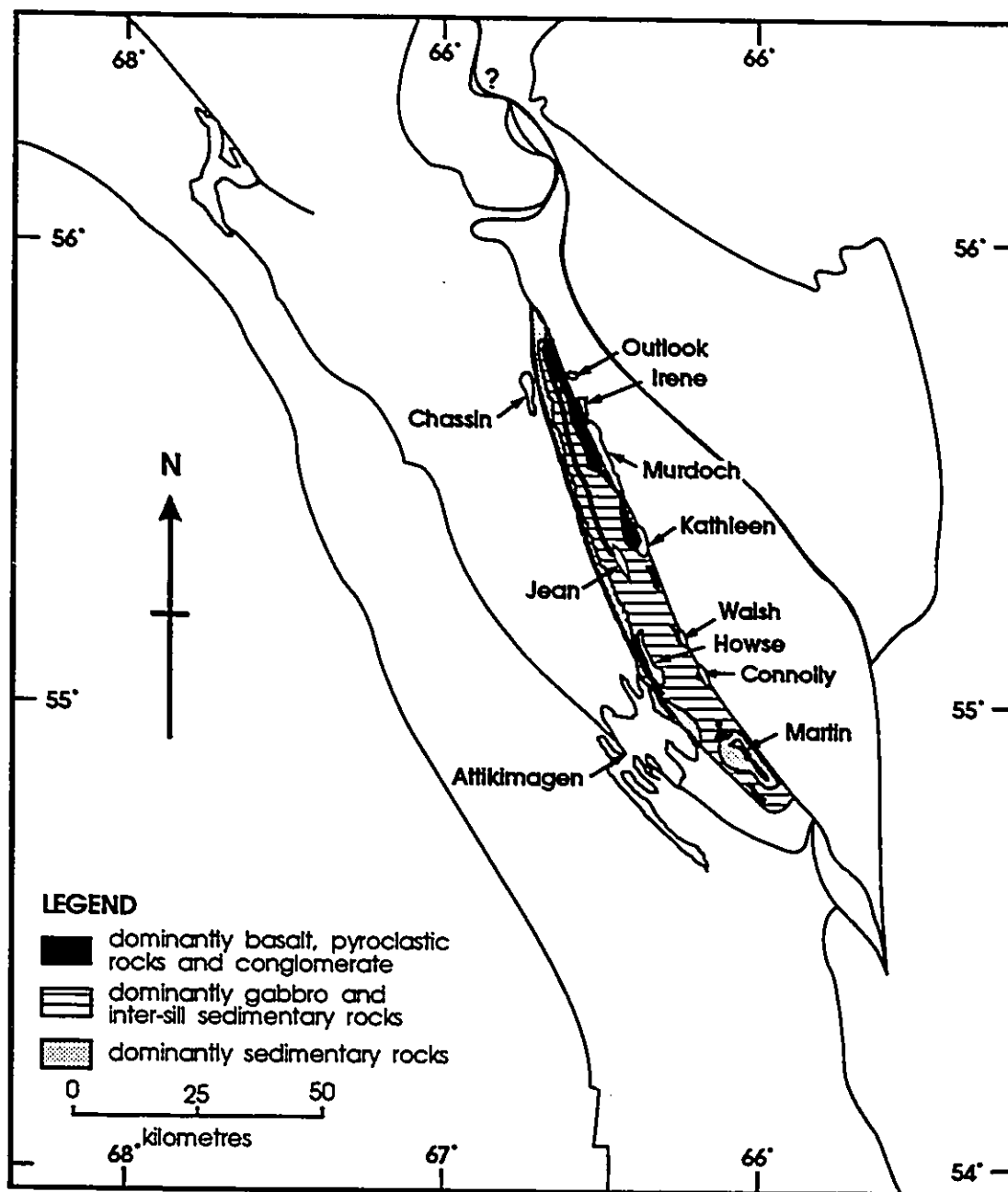


Figure 3.2. General distribution of the Menihek Formation and Menihek gabbros in the Howse zone of the central Labrador Trough. Names refer to geographical locations mentioned in the text.

horizon occurs near the base of the Menihek formation. The continuity of the horizon is unknown; it is exposed along the southern shores of Howse Lake for a distance of approximately 10 km, but outcrop disappears to the north and south.

In the northern part of the Howse zone, between Outlook and Chassin lakes, Baragar (1967) mapped a thin band of pyroclastic rocks below the main volcanic member. The band continues southward for approximately 50 km to Jean Lake, and has a maximum thickness of less than 100 m. Baragar (1967) described this interval as consisting predominantly of pyroclastic rocks, and if this is the case then it could be correlative with the pyroclastic horizon at Howse Lake. The exposures of the Outlook Lake horizon examined in this study, however, were composed of conglomerates, with large, well rounded cobbles of basalt and jasper. The Outlook Lake band also appears to be much higher in the stratigraphy of the Menihek Formation than the pyroclastic horizon at Howse Lake, which occurs at the base of the formation. In addition, clasts in the conglomerates are compositionally similar to the Howse Lake flows (see Chapter 8). This suggests that, while the Howse and Outlook lakes pyroclastic/conglomeratic horizons are not correlative, the equivalent to the Howse Lake sequence was originally present to the north. The absence of such rocks may be a function of exposure, or the sequence may have been removed by erosion soon after deposition, thus providing cobbles to the conglomerates. Penecontemporaneous eruption and erosion of volcanic flows resulting in intraformational conglomerates also occurred in the Nimish/Sokoman formation, and in the lower Menihek sequence at Howse Lake.

Samples of the main volcanic member of the Menihek Formation were obtained from well exposed areas west of Murdoch Lake. Three additional samples were collected across a small lens of basalt present west of Connolly Lake. Samples were also collected from the Outlook Lake-Howse Lake pyroclastic/conglomeratic horizons. These consist of basaltic clasts from the Outlook Lake conglomerates, and basaltic flows and pyroclastic rocks at Howse Lake.

Menihek Gabbros

The distribution of gabbroic sheets in the Howse zone is similar to that of the Le Fer gabbros in the Ferrum zone, with sills predominating over basalts except in the north-eastern part of the domain, and extending across the entire zone (Fig. 3.2). Although some sheets may have extrusive origins, the problem of differentiation is somewhat alleviated by the common occurrence of features which clearly indicate intrusion. Chilled margins are often preserved at the contacts with underlying sediments and partially recrystallized sedimentary xenoliths are not uncommon. The sills usually show some degree of differentiation which may be manifest in rhythmic layering, cumulate oxide horizons, felsic differentiates or variations in glomeroporphyritic contents. The occurrence of strongly plagioclase glomeroporphyritic sills is a characteristic feature of the Howse zone and although glomeroporphyritic texture is not confined to sills of this domain, no other rocks exhibit the extreme concentrations or ubiquity of glomerocrysts. This fact forms part of the basis for subdivision of the Howse zone of Wardle *et al.* (1990a) into two distinct domains (Le Fer and Howse zones).

In the Howse Lake area, the Howse zone preserves a stratigraphic sequence approximately 6-8 km thick, of which 50-80% is gabbro. Mapping in this area indicates that as much as 60% of the gabbroic rocks are porphyritic or glomeroporphyritic. Although these proportions are variable within the Howse zone, they provide a rough measure of the predominance of gabbroic material in the domain. The Howse Lake area was the site of a detailed examination of the Menihek gabbros, and the majority of samples were collected in this area.

Murdoch Basalts

The Murdoch Formation is composed predominantly of mafic pyroclastic rocks and basalts, with relatively minor amounts of acidic rocks and sediments (Baragar, 1967; Frarey, 1967; Dimroth, 1978). The formation extends for much of the length of the Doublet terrane in a single, poorly exposed but well defined band adjacent to the Walsh Lake thrust (Fig. 3.3). Due to poor exposure, generally well developed schistosity and discontinuous nature of individual rock types, the formation is undifferentiated on

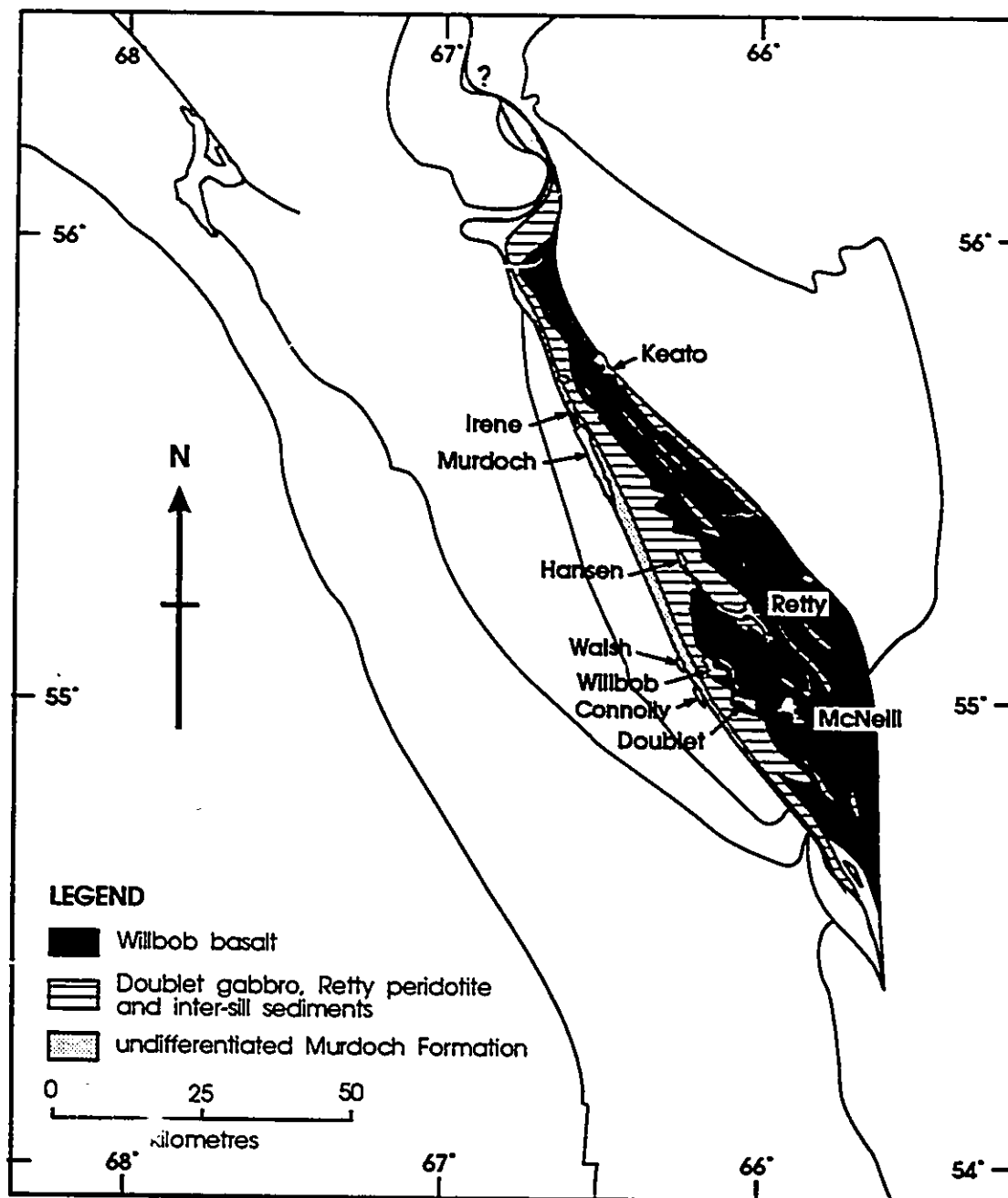


Figure 3.3. General distribution of the Murdoch and Willbob formations and Doublet gabbros in the Doublet terrane of the central Labrador Trough. Names refer to geographical locations mentioned in the text.

geological maps. Consequently the amount and distribution of basalt is difficult to determine. Baragar (1967) noted several localities in the Murdoch Lake area where massive basalt was exposed, but considered such rocks to be a minor component of the formation, which was estimated to be approximately 2000 m thick. In the Willibob Lake area, Frarey (1967) estimated the thickness of the formation to be 700-1500 m, to which basalts made a minimal contribution. Dimroth (1978), however, observed abundant exposures of pillowed basalt north of 56° latitude, and estimated that these comprised a minimum of 50% of the estimated 1200 m thickness of the formation in this area. These observations indicate that basalt flows are most abundant in the northern part of the Doublet terrane where they may have an aggregate thickness of as much as 600 m, but are an increasingly minor component of the Murdoch Formation moving south.

The undifferentiated nature of the Murdoch Formation on geological maps gave rise to problems in obtaining samples of the basalts. Several large exposures of the formation were investigated west and north of Murdoch Lake, but these were found to consist of either mafic pyroclastic rocks or chlorite schists. Consequently, most of the limited number of samples of Murdoch basalt used in this study were obtained from a single area located north of Irene Lake where pillowed basalt is relatively well exposed in a zone of outcrop approximately 800 m wide. Although the stratigraphic continuity of this exposure is unknown, one area of relatively continuous outcrop extends for approximately 150 m in which no non-basaltic rocks were identified. This represents a minimum stratigraphic thickness of approximately 100 m, and it is probable that the entire zone represents a 200-400 m thick interval composed predominantly of basalt. This interval is centrally located in terms of the geographic distribution of the Murdoch Formation, and presumably occupies a stratigraphic position near the middle of the Murdoch Formation. A few additional samples were acquired from isolated outcrops in a poorly exposed area east of Connolly Lake. These outcrops occur near the lowest exposed portion of the Murdoch Formation, and hence represent a lower stratigraphic level than the Irene Lake occurrence.

Willbob Basalts

The Willbob Formation occupies the eastern half of the Doublet terrane and includes the thickest, essentially continuous, sequence of basalts in the central Labrador Trough. At its widest point, the formation underlies an area extending for 30 km from the contact with Thompson Lake Formation to the eastern border of the terrane where it is truncated by the Gill Lake thrust (Fig. 3.3). The stratigraphic thickness of the formation in the central part of the domain is estimated to be greater than 5000 m (Baragar, 1967; Frarey, 1967), but because the top of the sequence is not exposed this represents a minimum thickness. Volcanic rocks constitute roughly 90% of the formation, with the remainder consisting of inter-flow sediments and minor mafic pyroclastic rocks.

The Willbob Formation is probably the best exposed stratigraphic unit in the central Trough, with more or less continuous outcrop available throughout much of its distribution. Sampling of the Willbob basalts was undertaken in the southern half of the Doublet terrane where the formation is thickest. Individual sample traverses were conducted from Thompson and Retty lakes to the eastern border of the terrane, between Retty and Doublet lakes, and to the south of the western arm of McNeil Lake.

Doublet Gabbros

In contrast to gabbros of the Le Fer and Howse zones, those of the Doublet terrane exhibit a relatively restricted distribution and are subordinate in abundance to basalts (Fig. 3.3). A thick sequence of gabbro occurs within the Murdoch Formation in the northern part of the Doublet terrane, but only a few small pods and lenses of gabbro intrude this unit to the south (cf. Baragar, 1967; Wardle, 1982). Gabbro is more abundant in the Willbob Formation, commonly forming sills less than a few hundred metres thick and generally continuous for only short distances along strike. Most of these sills are too small to be shown in Figure 3.3, and many may, in fact, represent thick flows. Several thicker and more extensive zones of gabbro occur within the Willbob volcanics, although these may represent more than one sill. The largest of these occurs adjacent to the Walsh Lake thrust south of Keato Lake. This band has an apparent stratigraphic thickness of approximately 3000 m, and is continuous for a strike length of over 30 km.

The Doublet gabbros are most voluminous in the Thompson Lake Formation where they are associated with, or intrinsic parts of, ultramafic sills (Retty Peridotite; cf. Fahrig, 1962; Beaudoin and Laurent, 1989). The ultramafic sills are generally bracketed by gabbroic bodies, and the two types of intrusions may be separated by thin horizons of Thompson Lake shale. The gabbro-shale-peridotite assemblage occupies a structurally contorted belt which extends for most of the length of the Doublet terrane. The belt has a maximum stratigraphic thickness of roughly 2000-2500 m in the southern part of the terrane, of which gabbro constitutes at least 50%. The close association of the three principal lithologies, and the tendency toward numerous narrow bands of gabbro, peridotite and shale, prevent illustration of individual gabbro sills in Figure 3.3.

Most of the gabbroic rocks examined in the Doublet terrane lack features indicative of *in situ* differentiation. As with the Le Fer basalts, the Willbob basalts can be massive, fine to medium grained rocks which are lithologically similar to the Doublet gabbros. It is therefore possible that many of the small gabbroic bodies are actually thick flows. There is no obvious lithological difference between gabbros emplaced within the Murdoch, Thompson Lake and Willbob Formations other than the fact that some of the mafic intrusives in the Thompson Lake Formation are associated with ultramafic sills. Therefore it seems unlikely that more than one generation of gabbro is present in the Doublet terrane.

Two sequences of gabbro situated at different stratigraphic levels in the Doublet terrane were sampled. East of Retty Lake, samples were collected across a 250 m thick gabbro sill emplaced within the Willbob Formation. The sill is separated from underlying peridotite by a narrow interval of black shale, and hence appears to be a separate intrusion. The second suite of gabbro samples was obtained from a 200 m thick sill which also intrudes the Willbob Formation to the north-east of Doublet Lake.

3.3 Summary

Major sequences of intrusive and extrusive mafic igneous rocks occur in the Knob Lake and Doublet groups of the central Labrador Trough. Together, the intrusive and

extrusive mafic rocks define the LTBS for the purposes of this study. The principal intervals of basalt occur in the Le Fer Formation of the Ferrum zone, the upper Menihek Formation of the Howse zone, and in the Murdoch and Willbob formations of the Doublet terrane. The basaltic members of the Le Fer, upper Menihek and Murdoch formations are thickest in the northern half of the central Trough, while the opposite appears to be true for the Willbob basalts. In addition, basalts appear to be more abundant in the upper part of the Le Fer and Menihek formations, although limited quantities of volcanic and pyroclastic rocks are present in the lower Menihek Formation at Howse Lake. Conglomerates in the Outlook Lake area in the northern part of the Howse zone contain clasts which are compositionally identical to the lower Menihek volcanics. This suggests that eruption of comparable volcanics may have occurred locally throughout the Howse zone, but these were largely removed by penecontemporaneous uplift and erosion.

The basaltic suites are spatially associated with three sets of mafic intrusions, here referred to as the Le Fer, Menihek and Doublet gabbros, which occur in the Ferrum zone, Howse zone and Doublet terrane respectively. In comparison to the associated basalts, the gabbroic rocks are more extensively and uniformly distributed in their respective lithotectonic domains. The contrast in distribution is particularly evident in the Ferrum and Howse zones, where the basalts tend to predominate in the northern and eastern areas. This difference may, in part, reflect the difficulty of distinguishing the two rock types. It is also possible, however, that gabbros were emplaced throughout the sedimentary horizons, while basalts were concentrated in the upper stratigraphic levels such that they are now more abundant in the eastern areas of the domains. This is clearly the case for the upper Menihek basalts, since the Menihek gabbros commonly exhibit features indicative of intrusion which allows differentiation between flows and sills. But it is often difficult to distinguish the Le Fer and Doublet gabbros from closely associated volcanic rocks.

Part II
Petrology and Geochemistry of the Menihek Gabbros

Chapter 4

Petrology And Mineral Chemistry Of The Menihek Gabbros

4.1 Geology of the Howse Lake Area

Howse Lake lies near the western boundary of the Howse zone in the south-central Labrador Trough (Fig. 3.2). The southern part of the Howse zone is composed predominantly of turbiditic sedimentary rocks of the Menihek Formation and intercalated sills of the Menihek gabbros. The Sokoman and Wishart formations, which form relatively thick sequences underlying the Menihek Formation in the Schefferville and Ferrum zones, are limited to thin and discontinuous wedges at the western edge of the Howse zone. Most, if not all, of the sills in the Howse zone are hosted by the Menihek Formation. The sedimentary rocks show consistent north-westerly strikes and north-easterly dips, and this general attitude is consistent with the long wavelength folding which dominates the structural style of the medial part of the central Trough. The Menihek sills typically show the same structural attitudes as their host rocks.

The stratigraphically lowest rocks exposed in the Howse Lake area are dolostones of the Denault Formation. This unit outcrops in two narrow bands separated by horizons of Sokoman iron formation and Menihek turbidites. The first band occurs along the eastern shore of Attikamagen Lake and the second at the north-eastern end of a small lake between Attikamagen and Howse lakes (Fig. 4.1). The two intervals of dolostone are separated by cherty iron formation of the Sokoman Formation and pelitic sediments of the Menihek Formation. The inferred position of the Howse thrust, which delimits the Howse zone and marks the western edge of glomeroporphyritic sills, lies within the Menihek Formation between the two bands of dolostone.

The Menihek Formation is composed predominantly of interbedded siltstone, shale and lesser arkose. The shales typically contain discontinuous laminae and lenses of disseminated to massive pyrite, and in places the sulphide content is sufficiently high for the sediments to be considered sulphide facies iron formation. Intercalation of shale and jasper

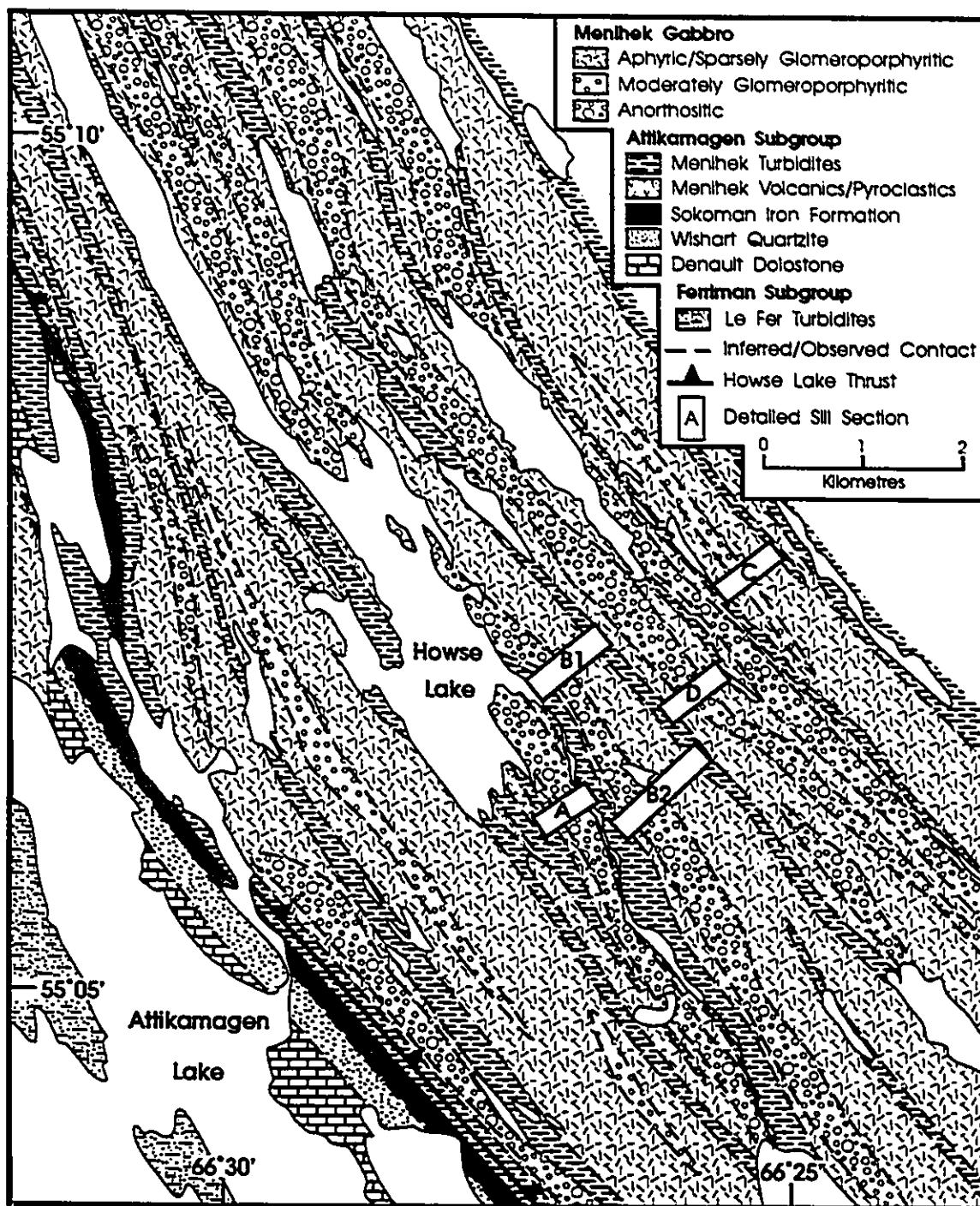


Figure 4.1. Simplified geological map of the Howse Lake area illustrating the distribution of the principal gabbroic lithologies, and the sedimentary and volcanic country rocks. The locations of detailed sampling of sills A through D are indicated.

which occurs near the base of the formation have been interpreted as the transition from the chemical sediments of the Sokoman iron formation to the pelitic sediments of the Menihek Formation (Gross and Zajac, 1983). The lower Menihek Formation also contains several thin horizons of mafic pyroclastic and volcanic rocks which occur immediately above the transition zone.

Bedding in the Menihek Formation at Howse Lake gives relatively consistent NNW strikes ranging from 315° to 350° , and easterly dips between 35° and 65° . These values are comparable to those reported from other sections of the Howse zone (cf. Baragar, 1967; Frarey, 1967; Wardle, 1982). Although numerous local structural features were observed, their effect on stratigraphy are minor; faulting may cause limited displacement of layering in the Menihek sills and small scale folding is apparent in the local deformation of the Menihek turbidites. The secondary nature of such structures, the absence of significant internal faulting and the consistency of stratigraphic attitudes refute the suggestion that the multiplicity of glomeroporphyritic sills is the result of structural repetition (Gebert, 1988). Although some structural repetition is possible, there is little or no evidence to suggest that the numerous glomeroporphyritic sills in the area represent a single, or limited number, of highly imbricated intrusions.

The effects of contact metamorphism on the Menihek sediments are relatively minimal. The most pronounced changes are evident in the country rocks immediately adjacent to sill contacts. These have retained most primary sedimentary features but have been transformed into relatively competent hornfelsic rocks. In many cases, the onset of recrystallization has led to a spotted appearance due to the development of poorly defined aggregates of semi-opaque material. Recrystallization, however, was arrested prior to development of porphyroblasts, and the identity of the minerals comprising the aggregates could not be determined. There is no evidence of anatexis in the host rocks, and sedimentary xenoliths within the sills retain their original textures and are typically angular in shape (Fig 4.2A). This suggests that contamination of the gabbroic magmas by *in situ* assimilation of host rocks was not extensive.

The Menihek sills at Howse Lake range from less than 100 m to greater than 500 m in thickness. Thicker bodies of gabbro occur (Fig 4.1), but since inter-sill sedimentary horizons tend to be narrow, recessively weathered and poorly exposed, these may represent

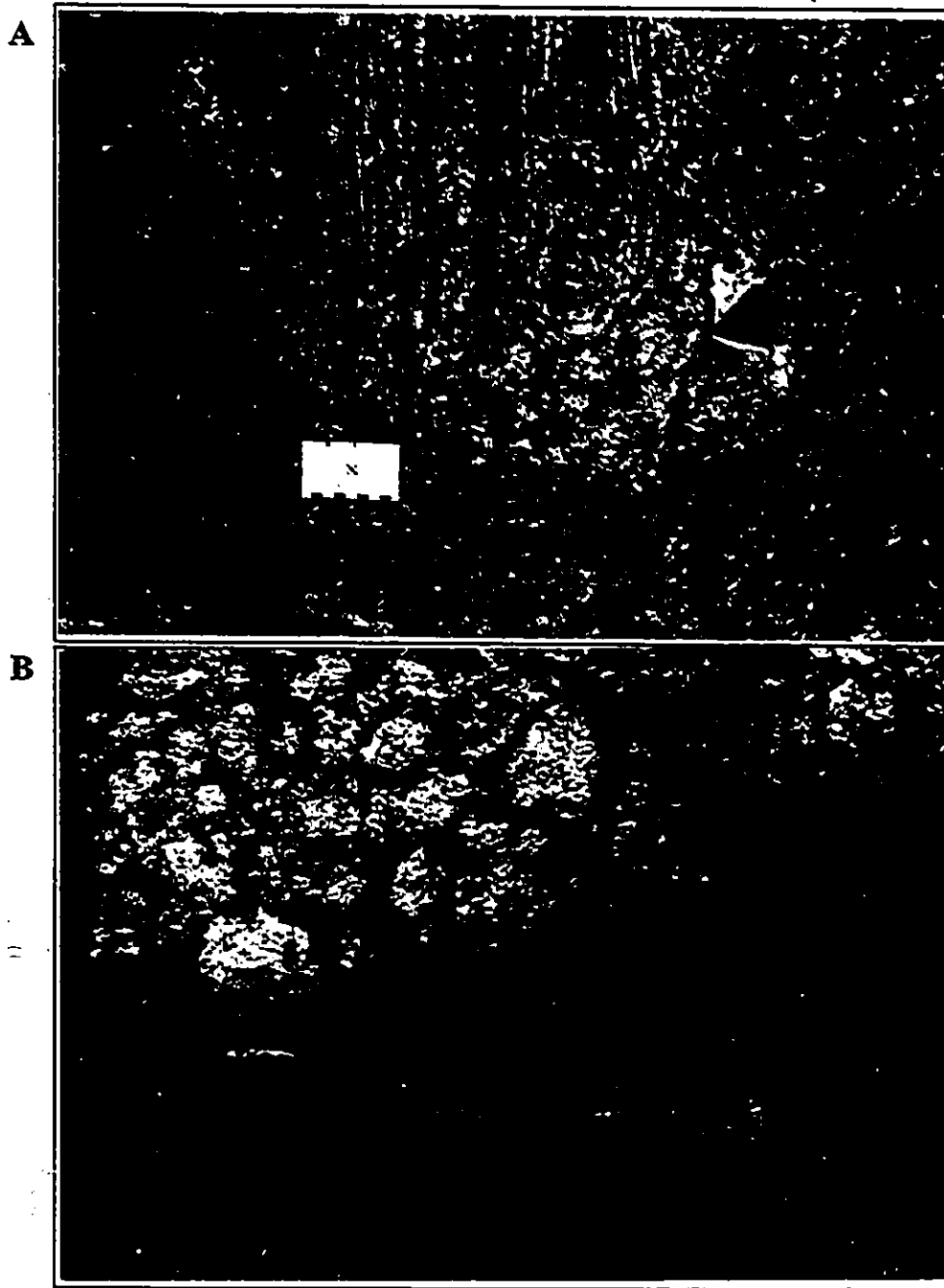


Figure 4.2. (A) Angular enclave of Menihek turbidites with undisturbed primary bedding in a glomeroporphyritic sill at Howse Lake. Scale card is 8 cm in length. (B) Melanocratic noritic dyke (bottom) emplaced at the contact between the core and anorthositic phases of a glomeroporphyritic sill at Howse Lake. The dyke is medium grained and unchilled at the contact with the core phase. Lens cap is 5 cm in diameter.

more than one intrusion. At least 12 individual sills are present in the 10 km wide zone mapped at Howse Lake, but again this number may be a minimum because of imperfect exposure. Discontinuous exposure also prevents determination of the strike lengths of individual sills, but it is clear that the intrusions have considerable lateral continuity, probably on the order of several tens of kilometres and possibly more. Sill contacts and internal linear features are usually parallel to bedding in the host sediments at Howse Lake. Obvious cross-cutting relationships are uncommon, although the pinch and swell of sill thickness may truncate sedimentary horizons or leave islands of Menihek turbidites enclosed by gabbroic material.

The only clearly discordant bodies are thin dykes of relatively melanocratic, medium to coarse grained noritic rocks which are exposed in some of the Menihek sills. The dark, reddish brown colour of the noritic rocks on weathered surfaces contrasts with the greyish or blotchy white colours of the Menihek gabbros. In addition, the dykes contain no glomerocrysts or phenocrysts. In some sills, the dykes cross-cut gabbroic rock at oblique angles to sill strike. In others, the dykes are approximately conformable and seem to have been emplaced along internal planes of the host intrusion (Fig. 4.2B). The dykes are not chilled at the contacts with enclosing gabbro and this suggests that their emplacement occurred prior to significant cooling of the host intrusions. This may indicate that the dykes are a slightly younger manifestation of the same magmatic episode which produced the sills.

4.2 Petrology of the Howse Lake Sills

The most conspicuous lithological variations evident in the Menihek sills arise from differences in the size and concentration of plagioclase glomerocrysts. There is a wide range in the average glomerocryst content of the gabbroic rocks, from those that are essentially aphyric to those which are highly glomeroporphyritic. In addition, individual sills may exhibit considerable variation in the quantity of aggregates at different stratigraphic levels. Despite this variation, three rock types are developed with such regularity that they provide a useful subdivision for descriptive purposes. The volumetrically most important lithology, here termed the gabbroic phase, consists of

aphyric to sparsely glomeroporphyritic gabbroic rock. The gabbroic phase constitutes the entirety of some sills, while in others it forms horizons of variable thickness adjacent to sill contacts (Fig. 4.3). The second principal lithology, the anorthositic phase, consists of moderately to highly glomeroporphyritic gabbro in which glomerocrysts constitute 20-50% (by area) of the rock. This phase occurs as discontinuous lenses within the gabbroic phase, and as thick horizons in the central regions of the sills. The latter type of occurrence is invariably bounded by marginal zones of the gabbroic phase. The third lithology is termed the core phase in reference to the fact that it is usually found enclosed by the anorthositic phase in the centres of the sills. This unit is the most spectacular of the three principal lithologies and consists of large, roughly spherical glomerocrysts set in an almost close-packed arrangement within a gabbroic matrix .

Gabbroic Phase

All sills in the Howse Lake area include some thickness of aphyric to sparsely glomeroporphyritic gabbro, and in most intrusions this lithology is dominant. In aphyric or sparsely glomeroporphyritic intrusions, the entire body is composed of this lithology. In highly glomeroporphyritic sills, the gabbroic phase usually constitutes 40-60% of the total thickness as marginal zones surrounding a central strongly glomeroporphyritic horizon (Fig. 4.3). In a few sills, the gabbroic phase is restricted to thin marginal zones adjacent to contacts, while the bulk of the sill consists of anorthositic gabbro. Neglecting glomerocryst content, the textural and mineralogical variations evident in the thicker horizons of sparsely glomeroporphyritic gabbro are similar to those found in essentially aphyric sills or those which contain no highly glomeroporphyritic interiors. The following descriptions apply generally to sills in which the gabbroic phase constitutes a large portion of the intrusion.

The sills exhibit relatively narrow chill zones at basal contacts with country rocks. Immediately adjacent to the contact, magma has been quenched to aphanitic basaltic rock. In the quenched material plagioclase is the only recognizable groundmass phase, forming tiny, elongate to acicular grains set in a cryptocrystalline, semi-opaque matrix which appears to be divitrified glass. With increasing crystallinity, augitic clinopyroxene appears

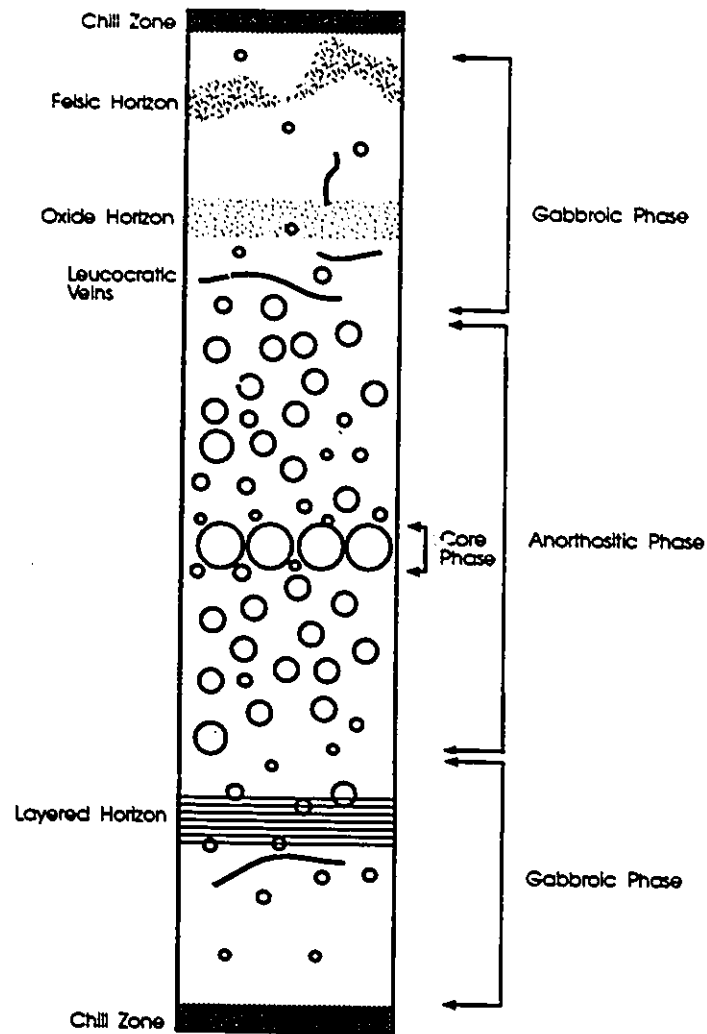


Figure 4.3. Schematic illustration of the principal lithological variations in an idealized Menihek glomeroporphyritic sill. Circles represent glomerocrysts (size exaggerated). Sill thickness is arbitrary.

as equant anhedral intergrown with plagioclase and numerous, very small grains of iron oxides are disseminated throughout the rock. Moving away from the contact, the amount of devitrified glass decreases and augite becomes increasingly intergranular to plagioclase. Within 30 cm to 1 m, a fine grained, holocrystalline gabbro has developed, with clinopyroxene and Fe-Ti oxides forming angular grains filling interstices between decussate plagioclase laths. Orthopyroxene generally comprises less than 10% (as measured by volume) of the fine grained gabbro, and orthoclase and quartz are ubiquitous minor phases. Where unaltered, the orthopyroxene is typically pleochroic in pale green and

pink, and was generally identified as hypersthene. Electron microprobe analyses (section 4.4) generally confirm this identification, although the orthopyroxene is locally more magnesian than hypersthene (i.e., bronzite), typically in the cores of crystals.

Plagioclase glomerocrysts showing strong or complete alteration are usually present in the chilled margins of glomeroporphyritic sills, although they typically comprise a relatively small proportion (< 2%) of the rock. The distribution of glomerocrysts in the chilled margin is usually quite heterogeneous. Plagioclase, clinopyroxene and orthopyroxene are often present as euhedral to subhedral phenocrysts or microphenocrysts. These are also highly altered, but fresh material occasionally occurs in the cores of grains. Olivine phenocrysts, although less abundant than other phases, occur in the chilled margins of a few sills. In one sample (sill A, sample F90-55), however, an anhedral olivine microphenocryst has been overgrown by orthopyroxene, and this suggests that olivine became unstable in the melt prior to emplacement.

Within 1 m to 15 m of the contact, the chill zone grades into the characteristic medium grained subophitic gabbroic rock which typifies the gabbroic phase. This gradation is marked by an increase in the grain size and degree of crystallinity, a change from granular to subophitic texture, and a decline in the extent of alteration of the principal silicates. The mineralogy of subophitic rocks consists of plagioclase and augite, lesser hypersthene, minor but ubiquitous interstitial iron oxides, and accessory amounts of various phases including orthoclase, quartz, pyrrhotite, zircon and apatite. The latter are less abundant than in the fine grained rocks of the chill zone, while hypersthene is more abundant. Plagioclase typically composes 40-50% of the mode, augite 20-40%, and hypersthene 5-35% (Fig. 4.4). The iron oxides are generally composed of magnetite with lamellar intergrowths of ilmenite. As with orthoclase and quartz, the iron oxide content of the subophitic rocks is generally less than that of the finer grained chilled rocks. Only one sill examined contains significant amounts of olivine and, although occasional grains are present in some samples from other sills, the Howse Lake gabbros are notable for the general scarcity of this phase. In the intrusion which contains significant olivine (Howse Lake sill C) this phase is concentrated in the central region of the sill where it comprises a maximum of 15% of the rock.

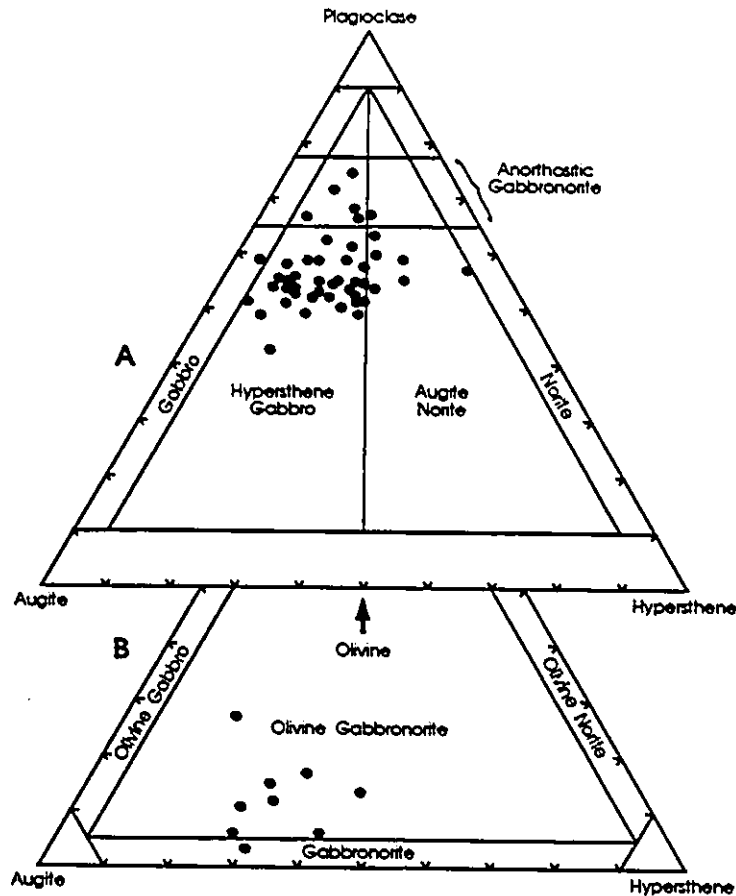


Figure 4.4. Modal classification of gabbroic and anorthositic phase samples from Howse Lake sills B and C. Gabbroic phase samples contain less than 65% plagioclase. Field boundaries and rock nomenclature from Streckeisen (1976).

Modal analyses of selected gabbroic and anorthositic phase samples are plotted in the ternary classification diagrams of the IUGS (Streckeisen, 1976) in Figure 4.4. Because of the continuous variation in glomerocryst content and the fact that the distinction between gabbroic and anorthositic phase lithologies used here is textural as well as modal, the samples are not distinguished by lithological type in the diagram. All gabbroic phase samples contain less than 65% plagioclase, however, and consequently are classified as hypersthene gabbros or, less commonly, as augite norites (Fig. 4.4A). The relatively olivine-rich rocks of the central gabbroic phase of sill C are classified as olivine-hypersthene gabbros or olivine-bearing hypersthene gabbros (Fig. 4.4B). In light of the minor quantities and intergranular nature of phases other than the principal silicates, and

the pattern of mineral zonation (see section 4.4), the medium grained gabbroic phase rocks could also be classified as plagioclase-augite-hypersthene mesocumulates or plagioclase-augite-hypersthene-olivine mesocumulates.

The presence of phenocrysts in the chilled margins of the Menihek sills indicates pre-emplacement crystallization of plagioclase, clinopyroxene, orthopyroxene and olivine. The predominance of plagioclase and pyroxenes, both as microphenocrysts and as the principal gabbroic phase silicates in most sills, suggests that olivine was replaced by orthopyroxene as the Ca-poor ferromagnesian mineral on the pre-emplacement liquidus. This interpretation is supported by the apparent instability of olivine suggested by the reaction relationship between this phase and orthopyroxene in the chilled margins of some intrusions.

Unlike the other sills examined, sill C contains relatively abundant olivine. If this phase is an *in situ* crystallization product, its presence indicates a difference in the composition of the sill C magma and hence either a different magma source or pre-emplacement evolution of a common magma source. If this phase was a pre-emplacement crystallization product, however, differences in olivine content among the sills may have arisen through differences in the timing or location of melt extraction from a source reservoir. The intimate spatial association of the sills, and the occasional occurrence of olivine microphenocrysts in the chilled margins of intrusions other than sill C, suggest that a distinct magma source for sill C is unlikely. It is therefore probable that sill C represents: (1) a less evolved (relative to other sills) melt composition, such that *in situ* crystallization of olivine occurred, or (2) a magma batch extracted from a common source chamber during a period of olivine fractionation such that this phase was present in the melt prior to emplacement.

Ophitic texture is often developed where the gabbroic phase constitutes a significant proportion of the sill stratigraphy. The mineralogy of ophitic textured material is the same as subophitic gabbro, and the principal lithological difference lies in the size of pyroxene grains. The texture usually occurs in the interior parts of gabbroic phase horizons, where scattered 5-25 mm sized, augite and hypersthene oikocrysts impart a mottled appearance to the rock. The development of large pyroxene oikocrysts in the central regions of sills is presumably related to the lower rate of heat loss, and smaller

degrees of undercooling, in comparison to marginal zones. Plagioclase and, where present, olivine may be enclosed by the pyroxene oikocrysts. Typically, plagioclase crystals in the interior of the oikocrysts are considerably smaller than those at the margins of, or outside, the pyroxene grains. This suggests that growth of the latter cut off the supply of appropriate components to the enclosed plagioclase crystals.

Ophitic textured gabbro grades back into subophitic material in the upper parts of the sills, but the level at which subophitic texture returns is highly variable. The upper contacts of the sills are generally unexposed, and in only one Howse Lake sill was an upper chilled margin observed. In this sill, subophitic gabbro grades into aphanitic chilled material similar to that present at the lower contact. Two sills (A and C) exhibit fine grained zones near the inferred upper contacts and although no quenched material is exposed, these appear to be equivalent to the fine grained granular gabbro which is present near the basal contacts. Consequently, it is likely that chilled margins are present at the upper contacts of most intrusions.

The primary mineralogy of the gabbroic phase has undergone variable degrees of alteration, but no sample examined was completely free of secondary phases. In aphanitic and fine grained samples from the chilled margins, plagioclase has generally been replaced by albite and epidote, and pyroxenes have almost invariably been pseudomorphically replaced by chlorite, actinolite and serpentine. As noted, the degree of alteration tends to decline away from sill contacts, and in the majority of medium grained samples plagioclase and augite are relatively well preserved. Groundmass plagioclase may show minor to moderate replacement by epidote and albite in the cores of crystals, and augite may be partially replaced by actinolite and chlorite. In most samples, hypersthene is partially to completely replaced by serpentine and chlorite. Olivine, where present, is commonly completely pseudomorphed by serpentine and magnetite. Iron oxides are often pristine, but in the more altered samples magnetite-ilmenite intergrowths have broken down to a semi-opaque mass of sphene or leucosene containing unaltered ilmenite lamellae arranged in a triangular pattern.

The gabbroic phase may be aphyric to sparsely plagioclase-glomeroporphyritic. In sills which contain the anorthositic phase, the gabbroic phase contains relatively homogeneously distributed glomerocrysts which are typically irregular in shape, 1 mm to

3 cm in diameter, and comprise 1-10% of the rock (Fig. 4.5A). Although glomerocrysts are not uncommon in chilled margins, the total proportion of aggregates is generally low, and these tend to be heterogeneously distributed. Thus the chilled margin at one location may be devoid of glomerocrysts, while numerous small aggregates may be present in the quenched rock a short distance along strike. The number of aggregates in the uppermost levels of the sills is also generally lower than in the main body of the gabbroic phase. In contrast, elevated glomerocryst contents are found in the gabbroic phase adjacent to anorthositic phase horizons. In glomeroporphyritic sills in which the anorthositic phase is absent, glomerocryst content is greatest in the central regions, and the remainder of the intrusion may be aphyric. The glomerocrysts of this type of sill are typically much smaller than in those containing the anorthositic phase (Fig. 4.5B).

Anorthositic Phase

The anorthositic phase is commonly developed in the central regions of glomeroporphyritic intrusions, where it characteristically forms relatively thick and continuous intervals parallel to sill contacts. This lithology may also occur as isolated lenses and bands which are elongated parallel to sill strike (Fig. 4.5A). The anorthositic phase is characterized by abundant glomerocrysts set in a medium to coarse grained groundmass (Fig. 4.6A). The thicker horizons of anorthositic phase gabbro usually show the coarsest groundmass grain size, whereas in thinner lenses and bands the groundmass is normally medium grained. Contacts with the enclosing gabbroic phase are generally gradational. Typically, glomerocrysts increase in size and abundance (Fig. 4.7) across a transition zone, and this is accompanied by a coarsening of groundmass grain size. The transition may take place over a few tens of centimetres, as is the case in most of the thinner lenses, or over several tens of metres as is more common in the thicker horizons. In only one sill at Howse Lake was a sharp contact observed between the gabbroic and anorthositic phases. In this sill there is an abrupt change in glomerocryst content and matrix grain size at the contact, with a further gradational increase in glomerocryst content in the anorthositic phase moving away from the contact. Neither lithology, however, shows evidence of quenching or disruption which would suggest an intrusive relationship.

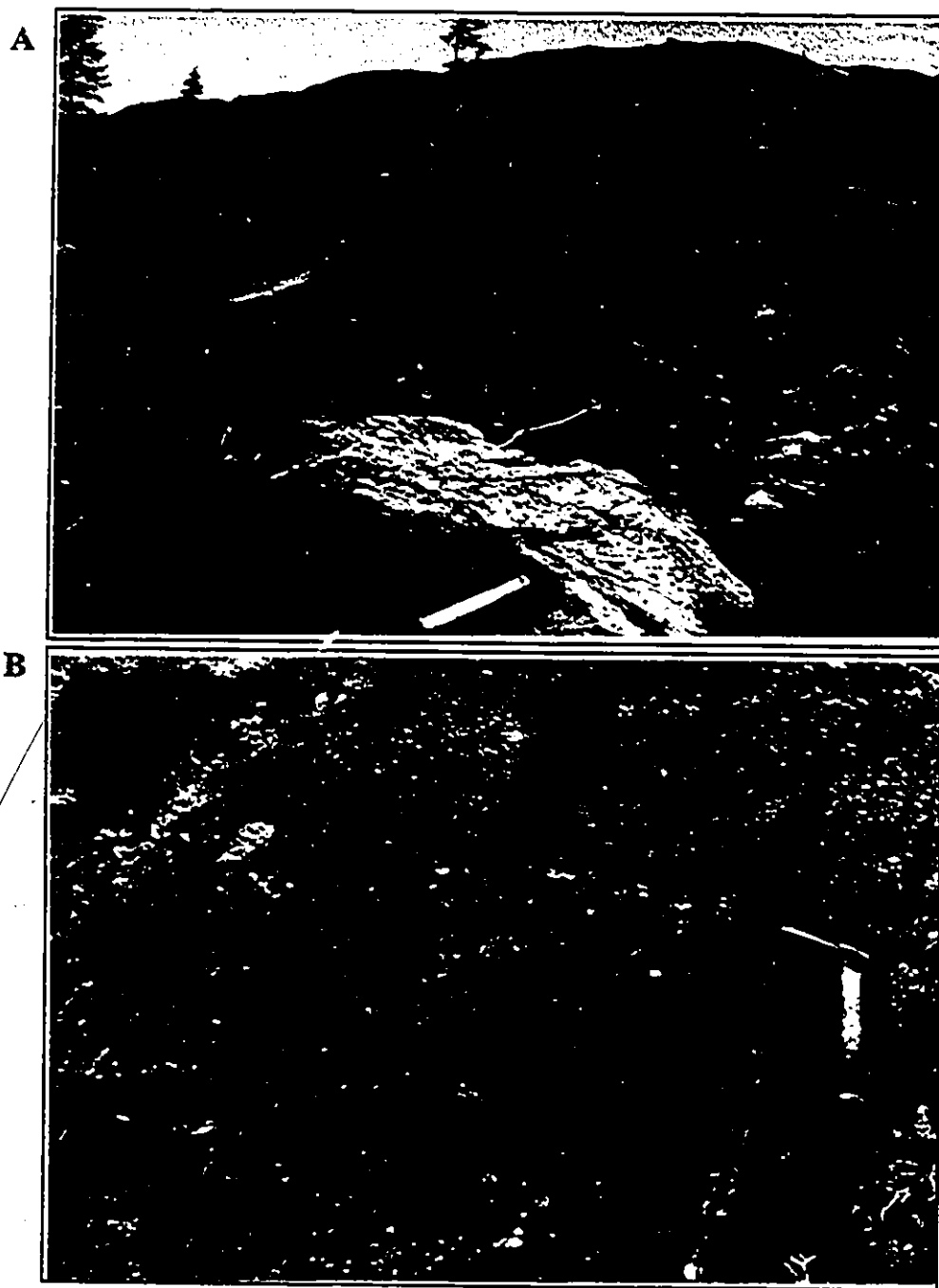


Figure 4.5. (A) Sparse plagioclase glomerocrysts (white) in the gabbroic phase of a glomeroporphyritic sill. A thin horizon of anorthositic gabbro is also present (immediately above hammer), and the exposure is cut by several leucocratic veins. Hammer for scale. (B) Abundant cm-sized plagioclase glomerocrysts in the central gabbroic phase of a sparsely glomeroporphyritic sill lacking the anorthositic phase. Marginal zones in the sill are aphyric. Hammer for scale.

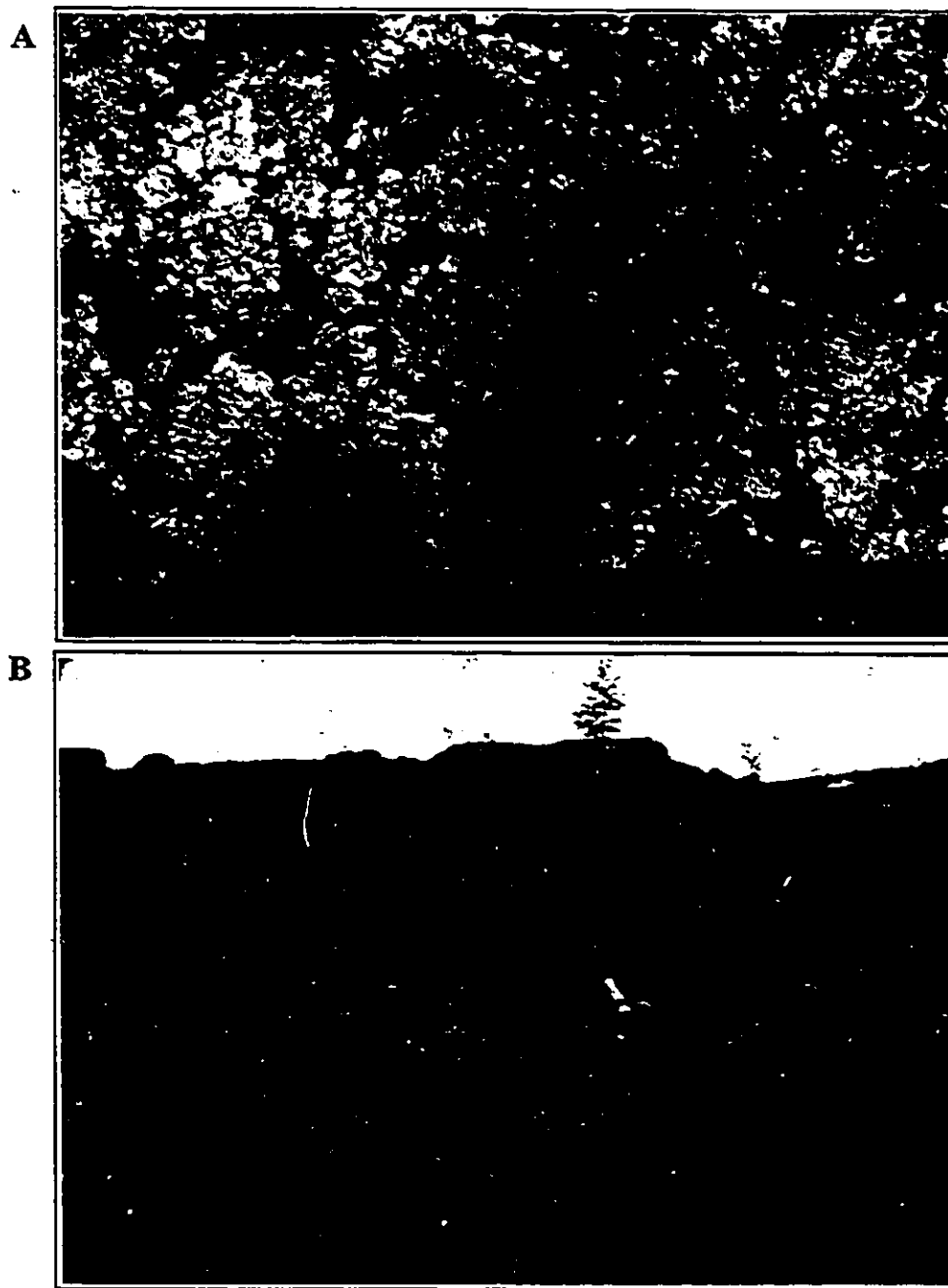


Figure 4.6. (A) Polished slab of typical anorthositic phase gabbro showing irregular plagioclase glomerocrysts set in a coarse grained groundmass composed primarily of augite and hypersthene. The sample is from a thick central interval of anorthositic phase gabbro in a glomeroporphyritic sill. (B) Abundant large plagioclase glomerocrysts (white) in the well developed core phase of a glomeroporphyritic sill. Note the large size, circular shape and close packing of the glomerocrysts. Hammer for scale.

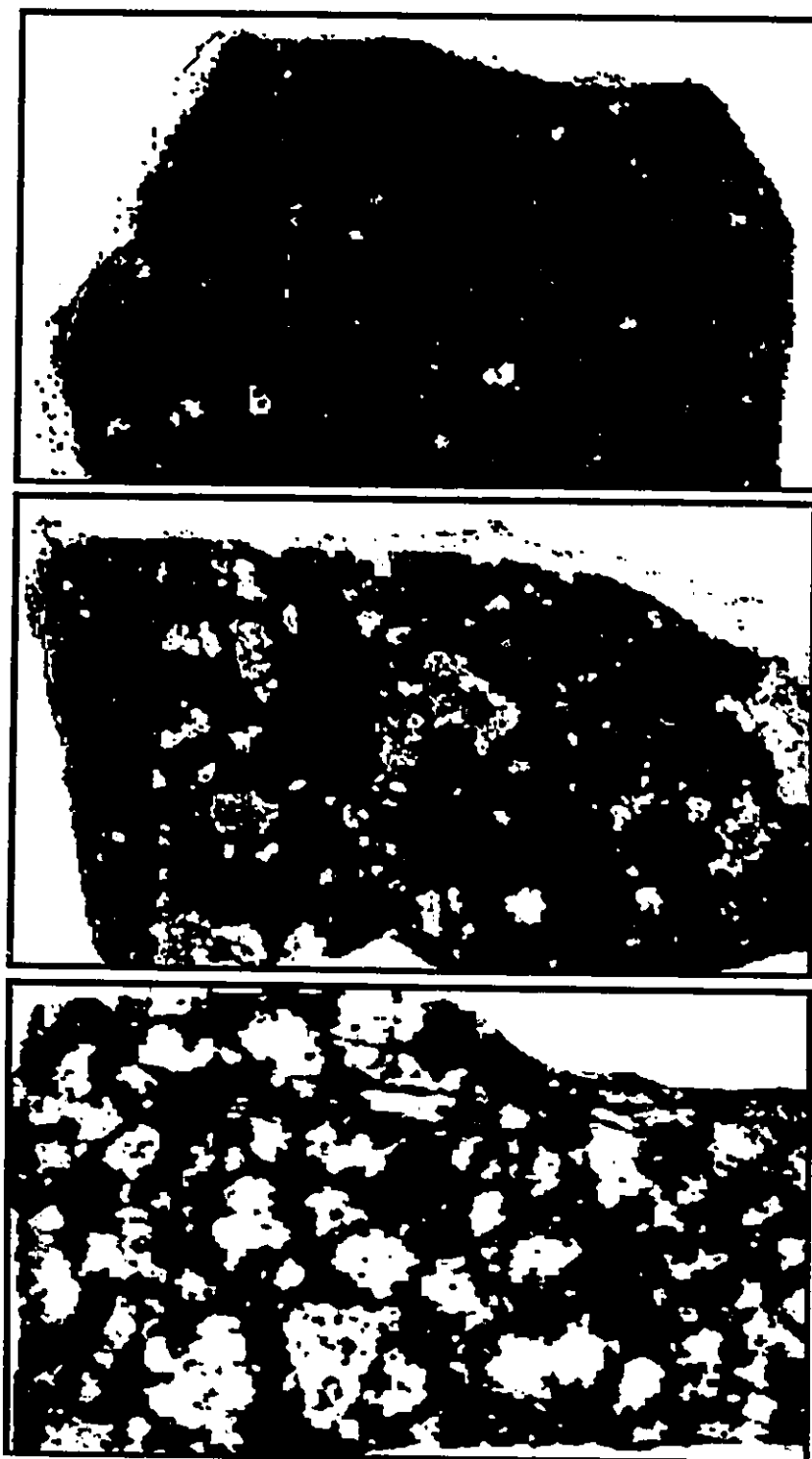


Figure 4.7. Digital images illustrating the increase in glomerocryst size and abundance in the transition from gabbroic to anorthositic phase lithologies. Light areas are glomerocrysts and dark areas are groundmass. Very small light areas are lustrous hypersthene grains, and light areas at margins of the slabs are weathered surfaces. The samples are from Howse Lake sill A. Each frame encloses an area 15 cm in width.

The glomerocrysts of the anorthositic phase are typically 1-5 cm in diameter, irregularly shaped, and comprise from 20-50% of the rock. They are more resistant to weathering than groundmass material and form bleached nodules on exposed surfaces. The matrix of the anorthositic phase consists largely of medium to coarse grained plagioclase, augite, hypersthene and magnetite-ilmenite. As with the gabbroic phase, olivine is rarely present. Where the anorthositic phase exhibits a medium grained groundmass, the relative proportions of the principal silicates are comparable to those of the more plagioclase-rich varieties of the gabbroic phase, and these rocks are hypersthene leucogabbros according to the IUGS modal classification. The total plagioclase content in fully developed anorthositic phase samples, however, is normally greater than 65% (Fig 4.4). Augite is generally more abundant than hypersthene in these rocks, which are classified as anorthositic hypersthene gabbros according to modal proportions.

The groundmass texture in medium grained anorthositic gabbro is subophitic to ophitic, with equant to elongate plagioclase grains partially to completely enclosed by pyroxene. Iron oxides and lesser sulphides are common accessory minerals. With increasing grain size, the proportions of the principal silicates becomes difficult to establish; a single large (7.5 x 4 cm) thin section may only reveal portions of individual groundmass crystals and glomerocrysts. In such rocks, however, plagioclase often appears to be subordinate to pyroxene in the groundmass. In part, this may reflect the coarser nature of groundmass crystals which often seem to be extensions of the glomerocrysts. It is also possible that in coarse grained lithologies a greater proportion of the groundmass plagioclase has been incorporated onto the glomerocrysts as overgrowths.

The primary mineralogy of coarse grained anorthositic phase material has suffered more intense alteration than is normal in the gabbroic phase. Augite is partially to completely replaced by actinolite and chlorite, and hypersthene is usually completely pseudomorphed by serpentine and chlorite. Groundmass plagioclase shows strong alteration except in outer, more sodic, rims. Glomerocrysts are normally completely replaced by epidote, albite or sericite. The stronger alteration of the anorthositic phase presumably derives from its coarse grained and therefore more permeable nature, which would permit greater penetration of alteration fluids.

Core Phase

The core phase is an unusual rock type that is sporadically and discontinuously developed in the central regions of many glomeroporphyritic sills. This lithology was only observed within thick horizons of the anorthositic phase where it forms irregular lenses and horizons which range from 5 m to 25 m in thickness. In many respects, the core phase is similar to anorthositic gabbro, but with the average size of glomerocrysts much larger (Fig. 4.6B). The diameters of glomerocrysts in this lithology often exceed 10 cm and may reach 25 cm, and the shape of the aggregates appears more regular than those of the anorthositic phase. The total glomerocryst content is usually on the order of 40-55%. Contacts with the anorthositic phase are gradational but take place over short distances, commonly less than 1 m. The groundmass grain size ranges from medium to coarse and, although the size and abundance of the glomerocrysts prevented modal analysis, the mineralogical composition of the groundmass of the core phase appears to be similar to that of the anorthositic phase. The nature and extent of alteration of groundmass minerals is also similar to that found in the anorthositic phase. Glomerocryst plagioclase is usually less altered where hosted by a medium grained groundmass, and again this presumably relates to the greater permeability of coarser grain sizes.

Glomerocrysts

The glomerocrysts of the Menihek sills are roughly spherical to irregularly shaped aggregates of plagioclase crystals. They range in size from less than 5 mm to more than 25 cm in diameter. The three principal gabbroic lithologies of the sills are characterized by glomerocrysts of different sizes, and this relationship forms part of the basis for the lithological subdivision. The largest aggregates occur in the core phase, the smallest in the gabbroic phase, and intermediate sizes in the anorthositic phase. Despite this generality, glomerocrysts of all sizes can be found in each lithology. On a gross scale, the largest aggregates tend to be more or less circular in cross section (Fig. 4.6B), while smaller aggregates are more irregular in shape. In detail, however, the large aggregates also have irregular outlines.

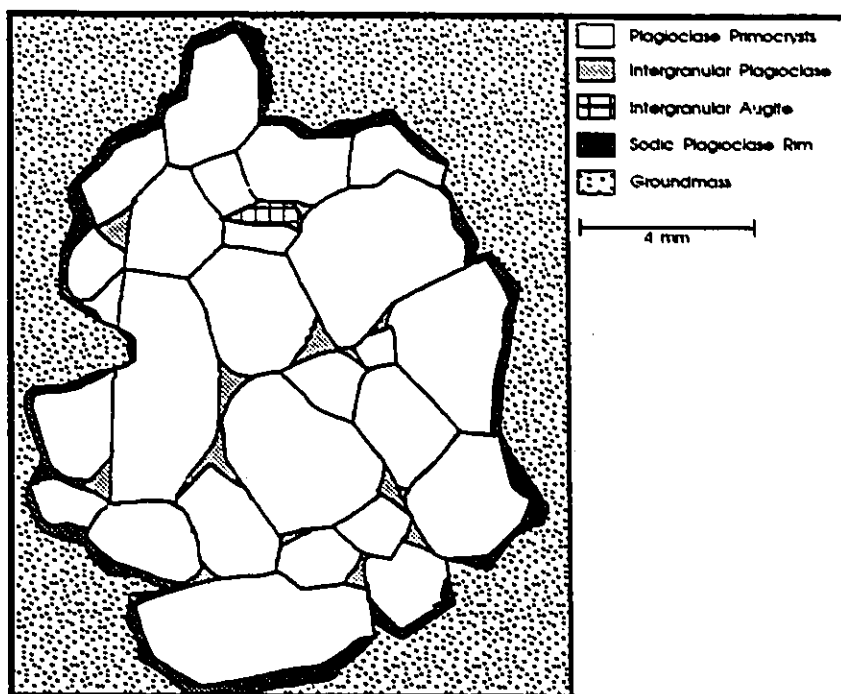


Figure 4.8. Thin section sketch of a small (~1.5 cm in diameter) glomerocryst illustrating the morphology of individual grains, and the nature of their intergrowths. The glomerocryst is overgrown by a peripheral zone of sodic plagioclase, and intergranular spaces between primocrysts are filled by plagioclase and augite. Sample F14-8.

The size of the aggregates depends principally on the number rather than the size of primocrysts. Small aggregates are composed of from 3 to 20 or 30 crystals while large glomerocrysts of the core phase may be composed of several hundred grains. Individual crystals tend to be subhedral and equant to slightly elongate (Fig. 4.8). They range in size from less than 0.5 mm to greater than 6 mm in their longest dimension. Primocrysts may adjoin at triple point junctions or may be separated by intergranular material (Fig. 4.8). Intergranular spaces are most commonly filled by smaller plagioclase crystals and overgrowths on the larger crystals. Less commonly iron-oxides and augite (Fig. 4.9A) fill intergranular spaces, but these phases usually comprise less than 5% of the volume of the aggregate.

A common feature of the aggregates is the development of a peripheral zone of sodic plagioclase (Fig. 4.8). This overgrowth is most apparent where the main body of the crystals are completely altered, leaving only the outer zone of sodic plagioclase clear. The

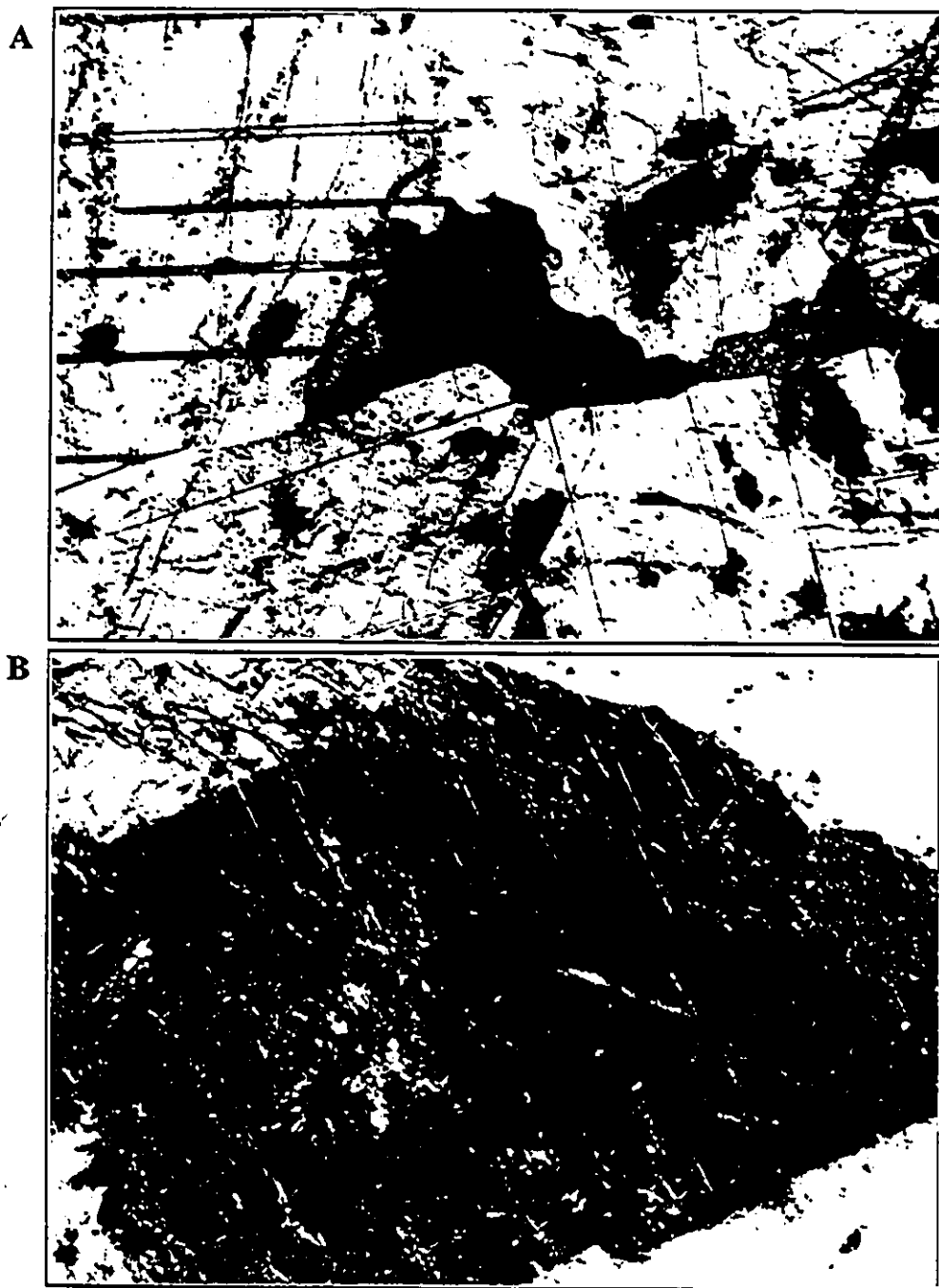


Figure 4.9. (A) Photomicrograph of intergranular augite (brightly coloured) in a large, core phase glomerocryst (pale yellow, white). Crossed polarized light. Field of view is 3 mm in width. Sample F15-4. (B) Photomicrograph of oscillatory zonation in glomerocryst plagioclase. Note the euhedral growth zones and stronger alteration in the core of the crystal. Cross polarized light. Field of view is 3 mm in width. Sample F15-4.

rim is confined to the periphery of the glomerocrysts and hence reflects post-aggregation crystallization. The outer crystals in a glomerocryst may also show resorption textures and embayments filled by groundmass material. In these areas the sodic rim may be present or groundmass crystals may abut directly against the glomerocryst primocrysts. These two features suggest that: (1) the glomerocrysts were formed prior to groundmass crystallization, and (2) glomerocryst primocryst compositions were not in equilibrium with melt compositions during groundmass crystallization.

Glomerocryst crystals usually show compositional zonation where they have not been completely pseudomorphed by alteration minerals. Usually compositional variation takes the form of normal continuous or oscillatory zonation, but complex patterns are not uncommon. The zonation often reveals euhedral growth zones in the central parts of the grains (Fig. 4.9B). Compositional variations are also manifest in alteration patterns found in the glomerocryst crystals. Although some glomerocrysts are relatively free of alteration and others show total replacement except for peripheral sodic zones, the most common pattern is strong alteration of the cores of crystals with lesser degrees of replacement moving towards the rims. This pattern is consistent with a general decline in Ca content from core to rim since Na-rich plagioclase is generally less susceptible to alteration (Deer *et al.*, 1985). The glomerocrysts show two distinct types of alteration, epidotization and sericitization. Epidotization is more common and results in cryptocrystalline semi-opaque material which probably includes some albite and clay minerals, and which appears as dark patches within glomerocryst crystals in thin section. In contrast, crystals which have been sericitized are clouded but remain colourless in thin section.

The morphologies of glomerocryst plagioclase crystals and those found in the groundmass of the gabbroic phase are substantially different. Whereas glomerocryst plagioclase tend to be large and stubby or equant, groundmass crystals are typically lath-like and usually much smaller. The groundmass crystals also display generally lower degrees of alteration, and this rarely involves sericitization. In addition, oscillatory zonation was not observed in groundmass crystals. These differences indicate the presence of two distinct generations of plagioclase in the Menihek sills, an interpretation supported by the occurrence of embayments, resorption textures, and peripheral sodic rims in the glomerocrysts. The morphological differences are much less pronounced in the

anorthositic phase, where groundmass plagioclase forms larger grains which often appear to be part of the glomerocrysts. The strong alteration of all phases in this lithology compounds the difficulty of differentiating between the two generations of plagioclase.

In some samples of the sparsely glomeroporphyritic gabbroic phase, there is a distinct pattern to the distribution of groundmass plagioclase laths in the vicinity of glomerocrysts. Elongate groundmass crystals of the gabbroic phase occasionally exhibit a concentric arrangement around the aggregates (Fig. 4.10). This texture is thought to reflect relative movement between glomerocryst and groundmass feldspar. The forces which may have driven this movement are discussed in Chapter 6, but the fact that movement occurred offers insight into the relationship between the two generations of plagioclase. Since relative movement after groundmass solidification is impossible, the texture could only develop when both glomerocrysts and groundmass plagioclase crystals were present in a semi-crystalline melt. The fact that groundmass crystals are arranged concentrically

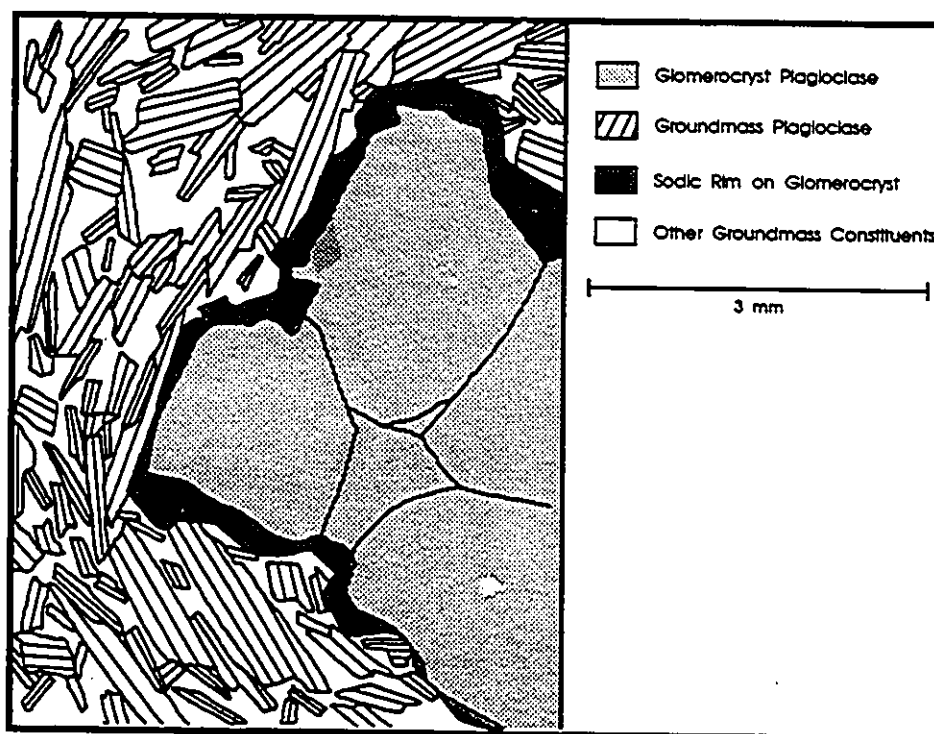


Figure 4.10. Thin section sketch illustrating the alignment of elongate groundmass crystals in a concentric pattern around part of a small glomerocryst. The glomerocryst shows an outer overgrowth of sodic plagioclase. Sample F14-1.

around the outline of the glomerocrysts indicates that the latter were more or less completely formed as groundmass crystallization occurred. It also indicates that either the glomerocryst moved through the semi-crystalline groundmass mush, or the latter flowed around the aggregate. Consequently the glomerocrysts represent a period of plagioclase precipitation and aggregation which predates groundmass crystallization. This confirms the conclusion drawn from the peripheral sodic rims and resorption textures described earlier. The textural contrasts between the two generations of plagioclase suggest that glomerocryst formation also predates sill emplacement (i.e., the glomerocrysts are not *in situ* crystallization products). This would allow for relative movement during or after intrusion.

Other Lithological Variations

In the upper parts of many sills, gabbroic phase material grades into local areas of more felsic rock. The felsic rocks form poorly exposed and irregular lenses and pods which usually cannot be traced between outcrops. The composition of this material ranges from quartz gabbro to granite (Fig. 4.11). The transition from the gabbroic phase to the felsic horizons is gradational and involves increasing orthoclase and quartz contents, typically as patchy granophyric intergrowths, and development of hornblende at the expense of pyroxenes. The felsic rocks also contain greater quantities of accessory zircon, apatite and sphene. Irregular granular textures predominate in the felsic horizons, and these rocks are generally coarser grained than surrounding gabbro. In some locations, the felsic rocks exhibit pegmatitic grain size, with extremely large, bladed pyroxene and hornblende crystals, and coarse, graphic intergrowths of quartz and alkali feldspar. The fact that the felsic horizons exhibit gradational contacts with enclosing gabbro indicates that they are intrinsic parts of the sills.

Centimetre-scale rhythmic layering is locally developed in several glomeroporphyritic sills in the Howse Lake area (Fig. 4.12A). The layered horizons were observed only in the gabbroic phase, either below or above a thick interval of the anorthositic phase. These horizons range from 1 m to 10 m in thickness and tend to be intermittent along strike. Layering is both modal and textural, but is defined principally by

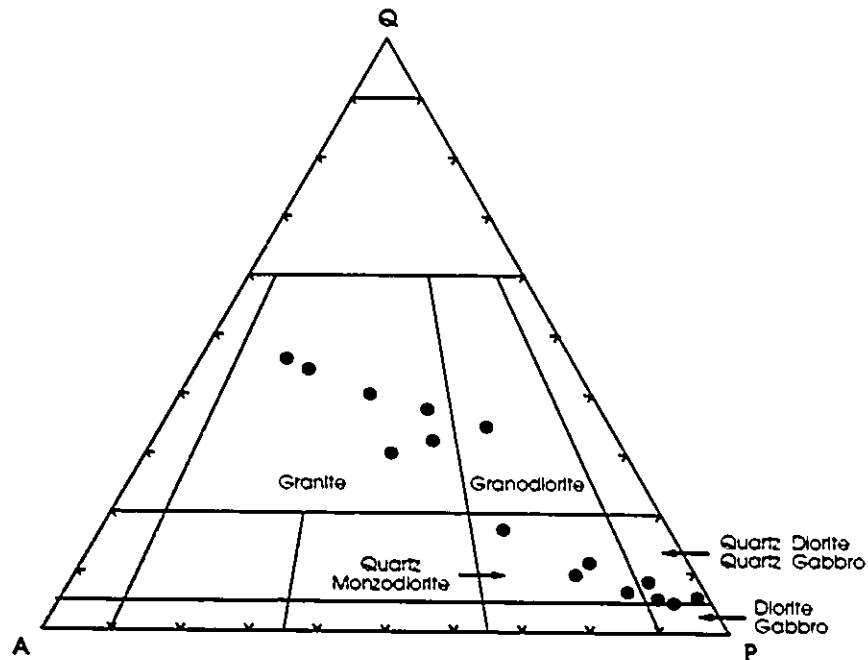


Figure 4.11. Modal classification of felsic samples from Howse Lake sills A, B and C. Field boundaries and rock nomenclature after Streckeisen (1976).

variations in hypersthene content. This phase is concentrated into strings of large blocky or equidimensional grains separated by thicker layers of finer grained plagioclase and augite. Augite tends to form elongate grains which are oriented parallel to intergrown plagioclase laths, and both phases are oriented parallel to the hypersthene-rich bands. Individual layers, and the layered horizon as a whole, are oriented parallel to the attitude of the anorthositic phase and sill contacts. In thin section the parallel arrangement of augite and plagioclase is the most conspicuous aspect of the layering. In outcrop, however, modal differences are more prominent due to differential weathering which has accentuated the contrast between plagioclase-rich and hypersthene-rich layers (Fig. 4.12A). The parallel arrangement of elongate crystals, and the concentration of the larger, equidimensional hypersthene grains, suggests that the layering may have a mechanical rather than chemical origin

Larger scale modal layering due to elevated iron oxide or iron sulphide contents is a common feature of the gabbroic phase in many Howse Lake sills. The sulphide enriched bands are most common in the gabbroic phase, but are occasionally present in the

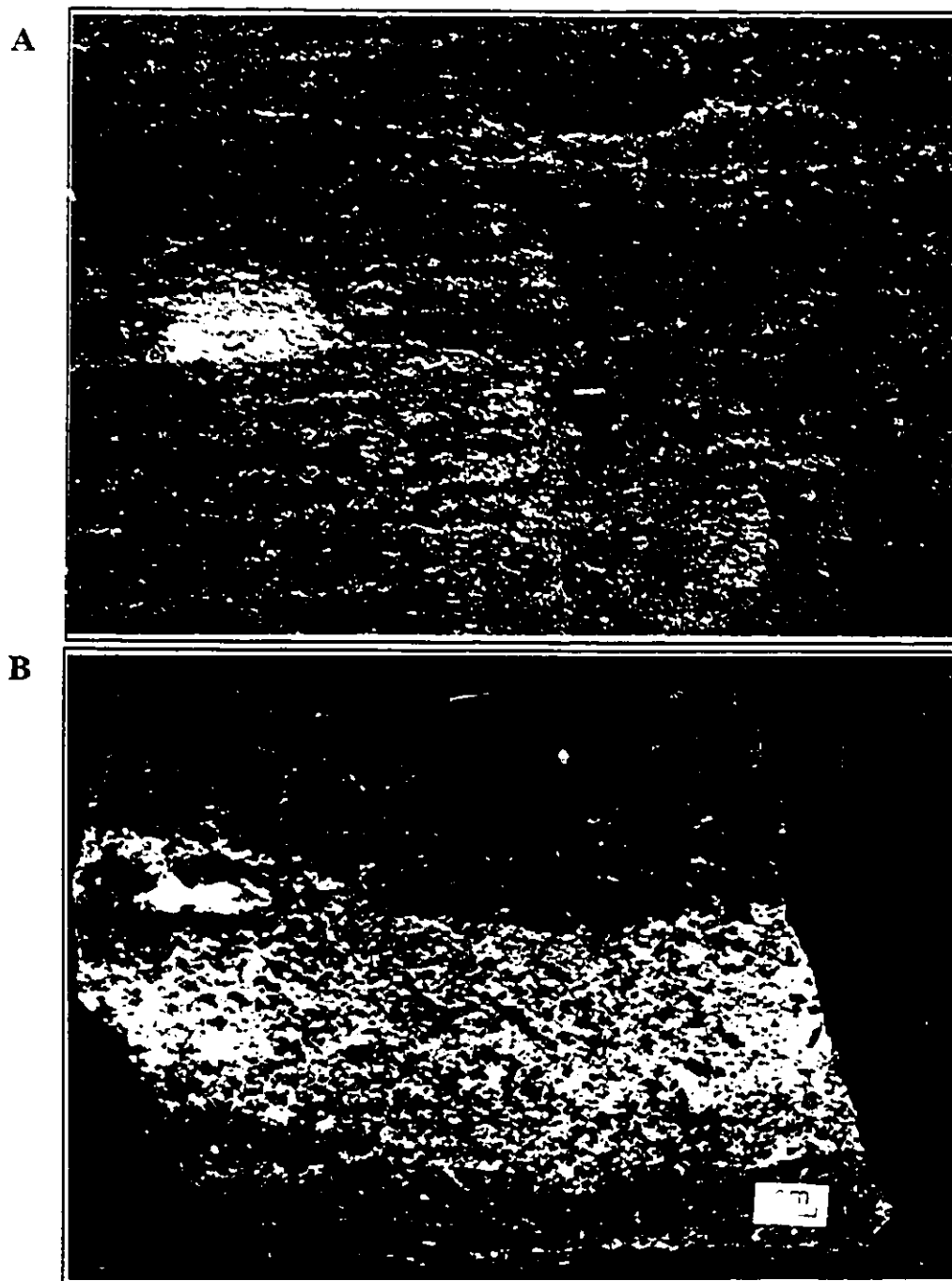


Figure 4.12. (A) Rhythmic layering in the gabbroic phase of a glomeroporphyritic sill. Lens cap is 5 cm in diameter. (B) Sharp contact between a felsic vein and enclosing gabbro.

anorthositic phase. They tend to be sharply defined and relatively narrow, usually ranging from 10 cm to 1 m in thickness. The generally recessive weathering pattern of the sulphide bands made establishment of lateral continuity difficult. Pyrrhotite is the predominant sulphide phase, but minor chalcopyrite is usually present. The sulphides occur as disseminated grains filling interstices between plagioclase and pyroxenes, and the total sulphide content may reach 25%. The magmatic textures of the sulphides indicate that sulphur saturation and formation of immiscible sulphide droplets occurred prior to, or during silicate crystallization. Isotopic analysis of sulphides from these horizons gave δ^{34} values ranging from +4 to +12 per mil, significantly higher than magmatic values (Findlay *et al.*, 1990). This suggests some sulphur from the host rocks ($\delta^{34} = +10$ to +23; Gebert, 1988) has been incorporated into the melts.

The oxide horizons are thicker and more diffuse than the sulphide bands, and have significant lateral continuity; such horizons can often be followed for several kilometres in the Howse Lake sills. The oxide zones occur within the gabbroic phase in the upper parts of the sills, range from a few metres to several tens of metres in thickness, and are gradational into surrounding gabbroic rock. In contrast to the intergranular nature of accessory magnetite-ilmenite in other lithologies, this mineral forms equant, subhedral cumulate grains which comprise 5-20% of the rocks. The abundance and mode of occurrence of Fe-Ti oxides in this lithology indicates that the melt had become sufficiently saturated in iron and titanium for the precipitation of oxide minerals.

A ubiquitous feature of the Menihek sills is the occurrence of bands and veins of extremely leucocratic material of uncertain relationship to enclosing gabbro. These occur throughout the sills, although they seem to be more abundant in the upper half of the intrusions. They also may be present in any lithological variant (i.e., gabbroic, anorthositic and core phases). The leucocratic zones may be cross-cutting or conformable in relation to sill stratigraphy (Fig. 4.5A). In contrast to the diffuse boundaries of the felsic horizons described earlier, contacts between leucocratic rock and enclosing gabbro are invariably sharp (Fig. 4.12B). Although abrupt, the contacts are not discontinuous; mineral grains often extend from enclosing gabbro into the leucocratic zones. The conformable bands

range from a few centimetres to greater than a metre in thickness, whereas the cross-cutting veins are usually less than 10 centimetres thick.

The leucocratic material is composed of an irregular and generally ragged intergrowth of albite, orthoclase, quartz, epidote, actinolite, chlorite, calcite and occasionally minor sulphides including galena. Some of the veins and bands consist of an irregular, fine grained granular intergrowth of quartz and feldspar with intergranular shreds of actinolite and chlorite. In others, coarse granophyric intergrowths of quartz and orthoclase are intergrown with blocky albite, epidote, actinolite and chlorite. The mineralogy and texture of the latter type of leucocratic material is somewhat similar to that found in the felsic horizons. Calcite is often present as intergranular material and occasionally as stringers within the veins. Some veins show relict subophitic textures and contain occasional highly altered glomerocrysts. This indicates that the leucocratic bands/veins are fracture zones which acted as conduits for hydrous fluids which may have had magmatic or metamorphic origins. Such fluids would cause hydrous alteration of the primary gabbroic mineralogy and probably precipitated additional phases such as quartz and alkali feldspar.

4.3 Lithological Variations in Individual Sills

Four sills in the Howse Lake area were mapped and sampled in detail to provide data on the mineralogy and geochemistry of the intrusions, and to determine various physical aspects of glomerocryst distributions. The locations of the sample traverses across the sills, labelled A through D, are indicated in Figure 4.1. The four sills are representative of the predominant types of intrusion found in the Howse Lake area (i.e., dominantly aphyric, dominantly anorthositic, and composite). Modal analyses were performed on selected thin sections from sills B and C, and this data is provided in Appendix A2.

Sill A

Anorthositic phase gabbro is the predominant lithology in sill A, forming a central interval bounded by relatively narrow marginal zones of the gabbroic phase (Fig. 4.13). At the lower contact of the sill, the gabbroic phase is chilled to aphanitic gabbro containing a

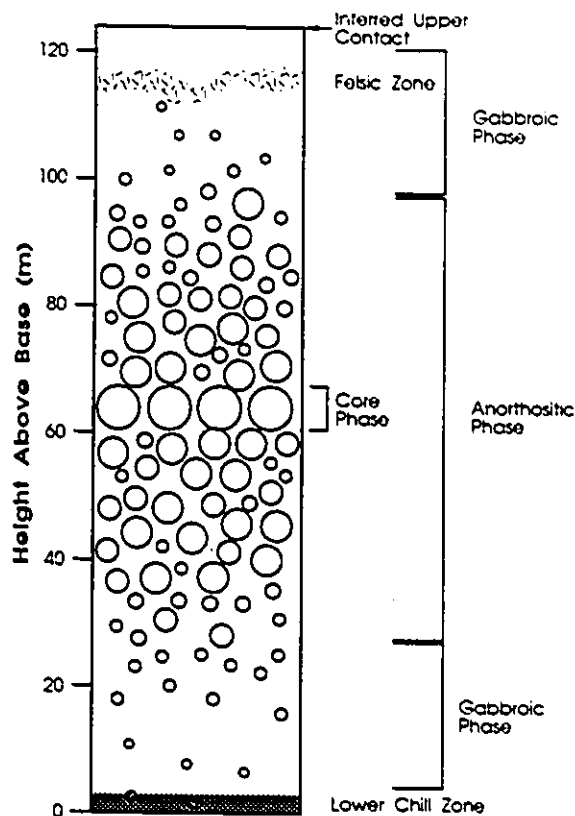


Figure 4.13. Schematic illustration of the lithological variation in sill A. Traverse location is indicated in Figure 4.1. Circles represent glomerocrysts. The size of glomerocrysts and the thickness of the chill zone and felsic horizon are exaggerated for illustrative purposes.

few scattered and highly epidotized glomerocrysts. Within 1 m of the contact, the chill zone has given way to medium grained subophitic gabbro which also contains few glomerocrysts. The groundmass consists predominantly of subequal proportions of augite and plagioclase, with lesser hypersthene and minor intergranular Fe-Ti oxides. The glomerocryst content is fairly constant for the next 10 m, and then gradually increases over a 20 m transition zone into the anorthositic phase. Hypersthene increases in abundance over this interval until it comprises slightly less than half of the total pyroxene, and the iron oxide content declines to 1-2%. The grain size of the groundmass coarsens as glomerocryst content increases through the gabbroic phase, and the texture becomes ophitic with large pyroxene crystals enclosing smaller plagioclase laths. The dominant glomerocryst size increases from less than 1 cm in diameter at the contact to approximately 3 cm in diameter in the anorthositic phase.

The anorthositic phase is completely developed at about 30 m above the base, with an average glomerocryst content of about 40%. This lithology occupies an interval approximately 70 m thick in the central part of the sill. The groundmass in this zone consists of very coarse and highly altered augite and hypersthene grains enclosing smaller plagioclase laths. Fe-Ti oxides are also coarse grained and form irregular crystals filling intergranular spaces between pyroxenes and plagioclase. The core phase is intermittently developed at a stratigraphic height of approximately 65 m above the base of the sill, and is completely enclosed by the anorthositic phase. This horizon has a stratigraphic thickness of about 3 m along the line of the section, but pinches and swells along strike. The transition from the anorthositic phase to the core phase is gradational but rapid. In the core phase the glomerocryst content increases to about 50%, and the predominant size of the aggregates is ~ 8 cm in diameter. The groundmass in the core phase is similar to that of the anorthositic phase, with coarse augite and hypersthene enclosing smaller plagioclase grains.

The anorthositic phase grades back into the upper gabbroic phase with the same type of changes in texture and glomerocryst content as occur in the lower part of the sill. The principal difference in the two transition zones lies in their thickness; the lower transition occurs over an interval of 20 m, while the upper transition occurs over 5 m. At 105 m above the lower contact, the sill consists of medium grained subophitic gabbro containing a few small glomerocrysts. An irregular zone of quartz diorite to quartz monzodiorite occurs at approximately 115 m above the base of the sill. This horizon is poorly exposed and does not appear to be more than a few metres in thickness. Its lateral continuity and relationship to enclosing gabbro are unknown. The felsic rocks are composed of large grains of albitic plagioclase enclosing, or intergrown with, smaller grains of hornblende, fibrous or shredded aggregates of actinolite and chlorite, and lesser amounts of alkali feldspar and quartz. Subophitic gabbro is exposed in a single outcrop between the felsic zone and the inferred upper contact of the sill. This gabbro is finer grained than that found below the felsic horizon and may form part of an upper chilled margin. The upper contact of the sill was not exposed but its presence was inferred from the occurrence of narrow (10 m wide) swampy depression between the uppermost gabbroic outcrop and exposures of an adjacent sill.

Sill B

Two traverses were conducted across sill B at the locations indicated in Figure 4.1, and the distribution of lithologies in the two profiles is illustrated in Figure 4.14. The lower contact of sill B is exposed only in profile 1 where the lower gabbroic phase has been chilled to aphanitic rock containing a few glomerocrysts. This material has undergone strong alteration with essentially complete albitization of plagioclase and chloritization of augite. Over the next metre, the aphanitic rock grades into fine grained granular hypersthene gabbro which occupies at least 10 m in the lower part of the sill. The fine grained gabbroic rock consists of a subhedral granular intergrowth of prismatic plagioclase, anhedral equant augite and smaller amounts of irregularly shaped to prismatic hypersthene. Intergranular spaces are filled by augite, quartz, orthoclase and Fe-Ti oxides. Quartz and orthoclase each comprise 3-5% of the rock, and oxides 4-5% (Fig. 4.15). Accessory phases include sphene, apatite, zircon and pyrrhotite. Alteration in this interval is also fairly strong, with partial to complete replacement of augite and hypersthene by chlorite, actinolite and serpentine. Groundmass plagioclase may be replaced by albite although it is generally less altered than the pyroxenes. Fine grained gabbro grades into medium grained subophitic hypersthene gabbro at the 10-12 m level. The textural change is accompanied by an increase in hypersthene content and a decline in intergranular orthoclase, quartz and oxides (Fig 4.15).

The glomerocryst content of the lower gabbroic phase interval in profile 1 is generally less than 5%. At 40 m above the lower contact there is a noticeable increase in the number and size of aggregates. The concentration continues to increase moving toward the centre of the sill, and at approximately 60 m above the contact the anorthositic phase with a coarse grained groundmass is fully developed. The 20 m transition zone is marked by increasing groundmass grain size and development of occasional large pyroxene oikocrysts. The orthoclase and quartz content of the gabbroic phase in the transition zone also declines to a combined total of less than 2%. The anorthositic phase is approximately 170 m thick in profile 1 and reaches a maximum glomerocryst content of about 40%. This lithology is mineralogically and texturally identical to the anorthositic phase of sill A. The total plagioclase content (groundmass+glomerocrysts) ranges from

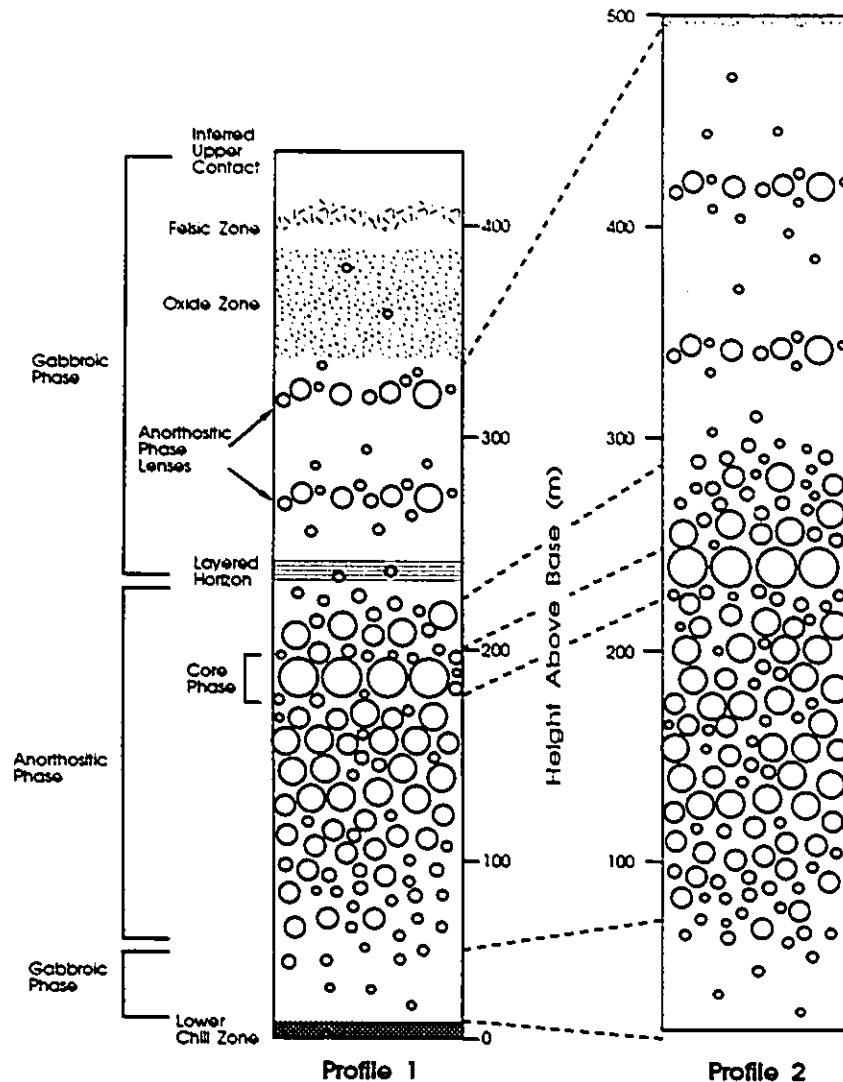


Figure 4.14. Schematic illustration of lithological variation in two profiles across sill B. Traverse locations are indicated in Figure 4.1. Circles represent glomerocrysts. The size of glomerocrysts, and the thickness of the chill zone and felsic horizon are exaggerated for illustrative purposes. Inferred position of the upper contact in profile 2 is 590 m above the base.

60% in the transition zones to as much as 77% in coarse grained anorthositic gabbro (Fig. 4.15). Subequal amounts of augite and hypersthene, and minor amounts of Fe-Ti oxides comprise the remainder of the mode. Orthoclase and quartz are accessory phases in some samples.

A zone of core phase gabbro is present within the anorthositic phase at 190 m above the lower contact of the sill. This horizon is approximately 15 m thick in profile 1,

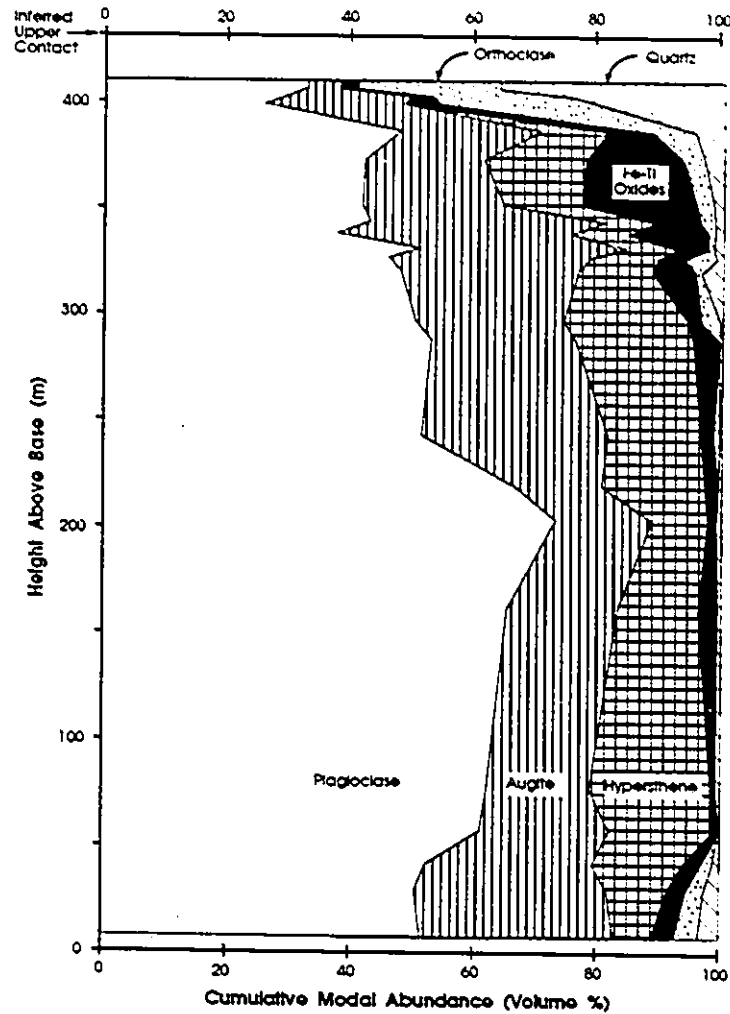


Figure 4.15. Variation in mode with stratigraphic height in profile 1 across sill B.

and is gradational into the overlying anorthositic phase. The lower boundary of the zone is marked by a 5 m thick layer of aphyric norite (Fig. 4.2B), which appears to represent an intrusive horizon emplaced along the anorthositic phase-core phase contact. The noritic horizon contrasts with the gabbroic phase of the sill in its complete lack of glomerocrysts, high hypersthene to augite ratio, and melanocratic nature. Similar noritic rocks form cross-cutting dykes, occasionally cementing breccias of anorthositic phase gabbro, in other sills at Howse Lake, and it is apparent that these rocks post-date sill emplacement. The noritic layer in sill B exhibits no obvious chilled margins against the overlying core phase horizon. This suggests that the noritic layer represents a slightly later injection emplaced

along the anorthositic phase-core phase transition prior to significant cooling of the host sill.

The core phase is characterized by large glomerocrysts, which comprise up to 53% of the rock, set in medium to coarse grained subophitic hypersthene gabbro. The number and size of glomerocrysts declines rapidly above the core phase, and at 230 m the anorthositic phase has given way to the upper gabbroic phase. This horizon locally contains up to 10% glomerocrysts immediately above the anorthositic zone, but glomerocryst content drops to a few percent by the 300 m level. Several minor lenses of anorthositic phase gabbro, in which glomerocryst content reaches 20%, are present between 230 m and 325 m. The lenses range from less than 10 cm to about 1 m in thickness, and rarely extend along strike for distances exceeding 40 m. Anorthositic lenses are not present above the 350 m level in profile 1, and few glomerocrysts were observed above 390 m.

The upper gabbroic phase of profile 1 is composed predominantly of medium grained subophitic hypersthene gabbro containing occasional large pyroxene oikocrysts. A 15 m thick layered horizon, in which granular hypersthene gabbro contains alternating pyroxene-rich and plagioclase-rich layers, occurs a few metres above the anorthositic phase. The lower part of the layered horizon includes a 1 m thick zone of gabbroic rocks which contain up to 15% intergranular sulphides, principally pyrrhotite with lesser chalcopyrite.

The gabbroic phase between 335 m and 390 m is characterized by an enrichment in Fe-Ti oxides. The oxides form subhedral equant grains intergrown with, rather than intergranular to, the silicate phases (Fig. 4.16). The oxide content of the gabbroic rocks reaches a maximum of almost 20% in the zone (Fig. 4.15). At the top of this horizon, the gabbroic phase also contains abundant euhedral apatite grains enclosed by oxides, plagioclase or pyroxene (Fig. 4.16). The morphology of apatite and oxide grains, and the textural relationships between these phases and the silicate minerals, indicates that they are cumulus crystals. Above the oxide horizon, at the 400 m level, gabbroic rocks grade into an 8 m thick felsic zone. This unit consists of coarse grained granitic rocks in which granophyric and graphic textures are well developed. No exposures occur between the felsic zone and the inferred upper contact of the sill.

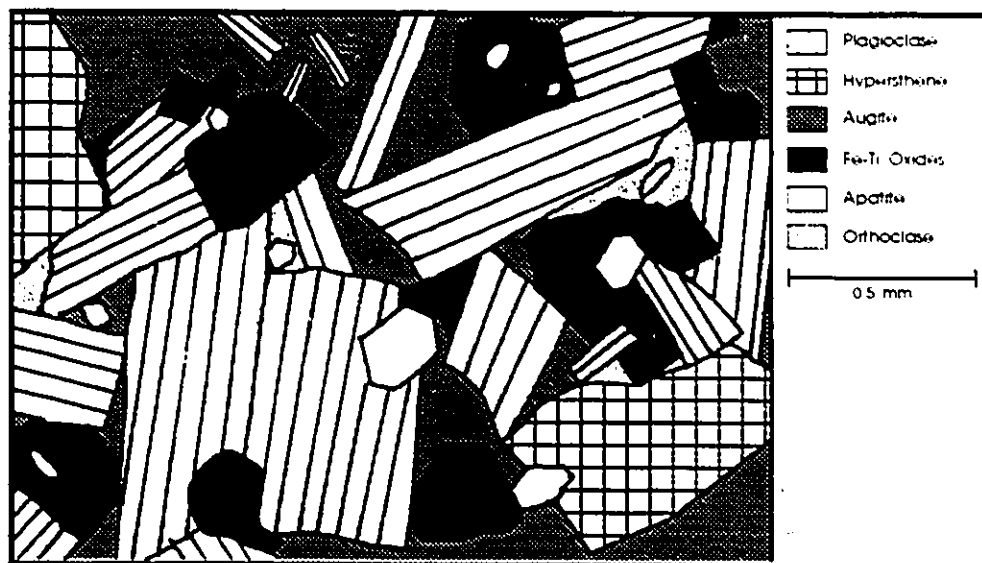


Figure 4.16. Thin section sketch illustrating the morphology of cumulus magnetite-ilmenite and apatite in the oxide zone of sill B. Sample F90-61.

The second profile across sill B is less well exposed, and several of the lithological variations evident in profile 1 were not observed. The overall structure of the profiles is similar, however, and the principal differences lie in the thickness of individual units (Fig. 4.14). The inferred upper contact of the sill occurs at 590 m, and hence sill B is approximately 160 m thicker in profile 2 than in profile 1. The basal contact is not exposed, but the lowest outcrops consist of medium grained subophitic hypersthene gabbro which is identical to the lower gabbroic phase of profile 1. The anorthositic phase is slightly thicker in profile 2, occupying the interval between 85 and 295 m. As in the first section, the core phase is present near the top of the anorthositic phase. In profile 2, however, both upper and lower contacts of this lithology are gradational, and the noritic layer is not present. The anorthositic phase grades back into medium grained hypersthene gabbro with a glomerocryst content comparable to the upper gabbroic phase of profile 1. No layering or sulphide enrichment was observed in the gabbroic phase above the anorthositic zone, although several anorthositic lenses are present. The uppermost exposure in the second profile occurs at 495 m and consists of oxide enriched material equivalent to the oxide horizon of profile 1. The lack of additional outcrop precluded determination of the thickness of the oxide horizon in this section.

Sill C

Sill C is composed entirely of aphyric to sparsely glomeroporphyritic and plagioclase-phyric gabbroic phase rock. The lower contact of the intrusion (Fig. 4.17) is marked by a chill zone of aphanitic gabbro in which glomerocrysts are rare. The chilled material, however, does contain phenocrysts and microphenocrysts of plagioclase, olivine and an unidentified pyroxene, all of which exhibit strong alteration. In the quenched rock, plagioclase is the only identifiable groundmass phase, forming acicular to prismatic crystals. The aphanitic rock grades into fine grained hypersthene gabbro composed of a granular intergrowth of plagioclase, augite and lesser hypersthene. Fe-Ti oxides, orthoclase and quartz each comprise 4-5% of the rock, and sphene, zircon and sulphides are accessory phases.

Within 3 m of the contact, the gabbroic phase has developed a medium grained subophitic texture in which augite encloses, or fills intergranular spaces between, small prismatic plagioclase grains and larger equidimensional hypersthene crystals. The pyroxene grains are consistently much larger than those of plagioclase, and this is a characteristic feature of most samples from the interior of the sill. The mode at the 5 m level is approximately 50% plagioclase, 30% augite, 10% hypersthene, 4% each of Fe-Ti oxides and orthoclase, and 2% quartz (Fig. 4.17). Olivine was not observed in this lithology. Over the next 5 m, hypersthene becomes slightly more abundant, the amount of oxides declines to about 2%, and orthoclase and quartz become accessory phases.

Olivine reappears in the gabbroic rocks of sill C at 20 m above the lower contact, and disappears again at 170 m. Increasing olivine content is balanced by declines in the modal abundance of plagioclase, hypersthene and augite (Fig. 4.17). Olivine is most abundant in the interval between 60 and 130 m, with the maximum content of 15% occurring at approximately 105 m, which is slightly below the stratigraphic centre of the sill. Outside of this interval, olivine occurs as occasional grains which generally comprise less than 5% of the rock. The olivine-bearing zone typically displays ophitic textures, with large pyroxene oikocrysts enclosing plagioclase and olivine. The olivine grains are subhedral to anhedral and much larger than associated plagioclase crystals. Despite the difference in grain size, olivine and plagioclase do not exhibit a poikilitic relationship such

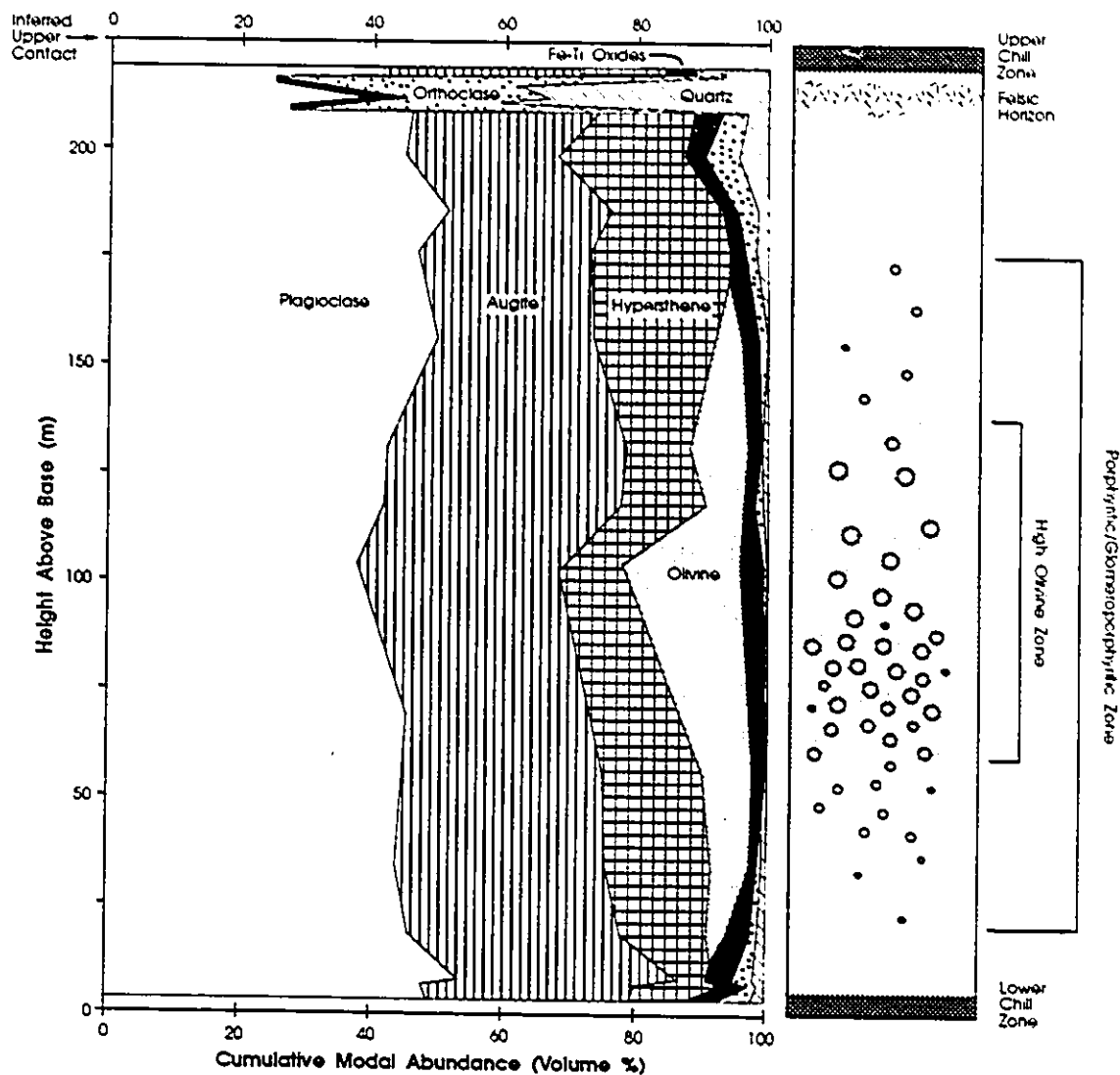


Figure 4.17. Lithological and modal variations in sill C. Traverse location is indicated in Figure 4.1. Circles represent glomerocrysts. The size of glomerocrysts, and the thickness of the chill zones and felsic horizon are exaggerated for illustrative purposes.

as is common between plagioclase and pyroxene. This suggests that olivine crystallized prior to plagioclase. Olivine grains do, however, exhibit reaction rims of fine grained indeterminate material, and in many instances they are overgrown by hypersthene (Fig. 4.18).

It was concluded earlier that the presence of abundant olivine in sill C indicates that either the magma was more primitive than that of other intrusions, or that olivine fractionation was occurring at the time of magma extraction, or both. In either case, the



Figure 4.18. Thin section sketch illustrating reaction rims and hypersthene coronas on olivine in ophitic olivine-hypersthene gabbro in the central regions of sill C. Sample F90-80.

corona textures indicate that this phase became unstable at some point during groundmass crystallization. If olivine formed *in situ*, it must represent an early product of crystallization which became unstable as solidification progressed. This might explain the absence of poikilitic textures involving olivine. Alternatively, if olivine was a pre-emplacement crystallization product, it may have become unstable as a result of the changing P-T conditions accompanying ascent and emplacement. The presence of an orthopyroxene corona on an olivine phenocryst in sample F90-55 indicates that olivine was unstable in the magma of this sill prior to emplacement. But since evolution of the parental melt would seem likely given the numerous phenocrysts and microphenocrysts of different phases in chilled gabbro, this does not necessarily indicate the pre-emplacement presence and instability of olivine in sill C. These questions are discussed further in Chapter 5, where whole rock geochemical data are presented.

Above the olivine-bearing ophitic zone, the gabbroic rocks return to subophitic hypersthene gabbros similar to that below this zone. The principal difference lies in the elevated hypersthene content of the gabbroic phase between 130 and 200 m. The orthopyroxene content peaks at approximately 25% before declining in the upper part of the sill. A 5 m thick felsic horizon is present at the 212 m level, which is approximately 15

m below the inferred position of the upper contact. The felsic rocks include quartz diorite, granodiorite and granite, all of which exhibit coarse anhedral granular textures. The contacts with enclosing gabbro are gradational, and immediately above the felsic horizon the sill has returned to medium grained subophitic hypersthene gabbro. The highest level exposed in the sill consists of fine grained gabbro comparable to the fine grained rock near the base of the intrusion. The inferred contact is approximately 5 m above this outcrop, and the fine grained gabbroic phase material is thought to represent part of the upper chill zone.

In comparison to most glomeroporphyritic intrusions at Howse Lake, the glomerocryst content of sill C is low. In addition, the glomerocrysts are small and plagioclase phenocrysts are quite abundant. Glomerocrysts and plagioclase phenocrysts first appear in the stratigraphy of sill C at 25 m above the contact (Fig. 4.14), and these increase in abundance to a maximum of about 8% between 65 and 90 m. Above this interval the glomerocryst content gradually declines until at 175 m no aggregates are present. The predominant glomerocryst size in the zone of highest concentration is less than 1 cm in diameter, which is considerably smaller than the predominant size in strongly glomeroporphyritic sills.

Sill D

Sill D is similar in structure and lithological variations to sill B, but is less well exposed. The lowest outcrop occurs 10 m above the inferred lower contact (Fig. 4.19) At this point the intrusion consists of medium grained subophitic hypersthene gabbro similar to the lower gabbroic phase of sill B. The average glomerocryst content is approximately 5% and, with minor variations in the abundance of glomerocrysts, this lithology occupies the lower 95 m of the sill. A narrow lens of anorthositic phase with up to 20% glomerocrysts occurs at the 60 m level. This material is essentially identical to the anorthositic lenses found in the upper part of sill B.

The lower gabbroic phase is in sharp contact with the anorthositic phase at 95 m. The contact is abrupt, but there is otherwise no evidence of an intrusive relationship between the two lithologies. At the contact the anorthositic phase consists of approximately 20% glomerocrysts set in a medium grained matrix of hypersthene gabbro.

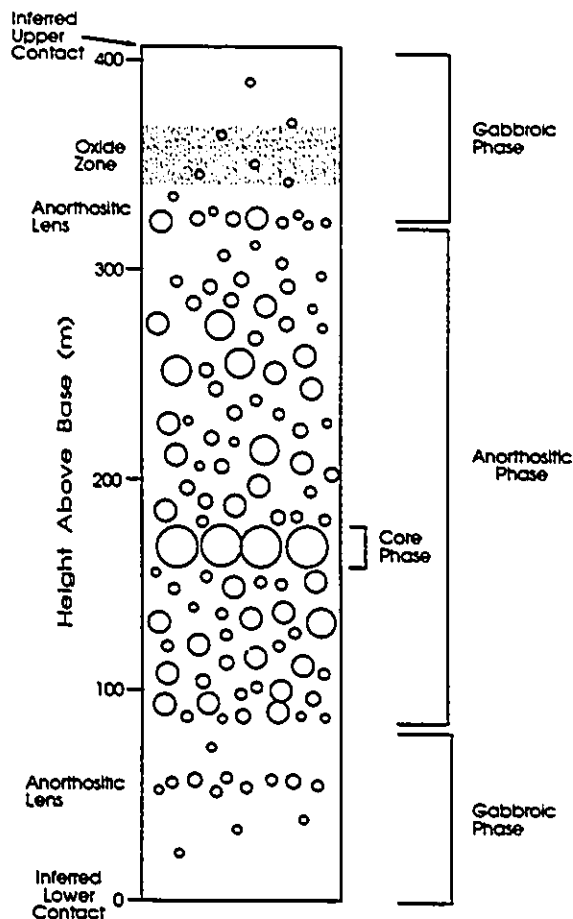


Figure 4.19. Schematic illustration of lithological variations in sill D. Location of traverse is indicated in Figure 4.1. Circles represent glomerocrysts. Size of glomerocrysts and thickness of the anorthositic lenses and core phase are exaggerated for illustrative purposes.

The matrix grain size, and the size and abundance of glomerocrysts increases moving away from the contact, and at the 110 m level, coarse grained anorthositic gabbro with ~30% glomerocrysts is present. This material occupies much of the central portion of the sill between 110 m and 275 m. A 5 m thick band of core phase gabbro is present in the lower part of this interval at 165 m. The core phase is gradational into surrounding anorthositic phase gabbro through an increase in the predominant size and proportion of glomerocrysts. In the anorthositic phase, the predominant glomerocryst size is 2-3 cm in diameter, and this rises to approximately 7 cm in the core phase. Similarly, the maximum glomerocryst content increases from 36% in the anorthositic phase to 44% in the core phase.

The transition between the anorthositic phase and the upper gabbroic phase occurs over an interval of approximately 40 m. In this zone, the size and abundance of glomerocrysts decline to levels comparable to those in the lower gabbroic phase. The upper part of the sill is composed of subophitic hypersthene gabbro which is largely indistinguishable from that which underlies the gabbroic phase. Two lithological variations occur in the upper gabbroic phase. The first consists of a 2 m thick anorthositic phase layer at the 330 m level which is gradational into the enclosing gabbro. This horizon can be traced intermittently along strike for at least 150 m. The glomerocryst content of the layer reaches a maximum of 18%. Several thin lenses of similar material are present in the gabbroic phase between 300 m and 350 m, but these are continuous for distances of only a few metres. The second lithological variation consists of a zone of elevated Fe-Ti oxide content which occurs in the interval between 345 m and 370 m. This horizon is mineralogically and texturally similar to the oxide zone of sill B, and is gradational into surrounding material through the appearance or disappearance of cumulus Fe-Ti oxides. The highest exposure of the sill occurs at approximately 30 m below the inferred upper contact. At this level the sill consists of subophitic hypersthene gabbro and there is no evidence of felsic rocks or an upper chilled margin. Such rocks, however, may be unexposed in the upper part of the intrusion.

4.4 Mineral Chemistry

The compositions of plagioclase, clinopyroxene and orthopyroxene were determined by electron microprobe in selected samples from sills B and C at Howse Lake. The analyses were obtained primarily to test for compositional variations across the main body of an individual sill, and to compare groundmass and glomerocryst plagioclase compositions. Because of the generally strong alteration of orthopyroxene, only a limited number of analyses were obtained for this phase. For the same reason, the composition of olivine was not determined.

Sill C was chosen to examine compositional variations in plagioclase and clinopyroxene relative to stratigraphic position because this intrusion lacks the typically highly altered anorthositic phase and therefore offers a more complete suite of suitable

samples. Because the olivine content of sill C makes this intrusion mineralogically anomalous in comparison to the others examined, additional analyses were obtained from the core and gabbroic phases of sill B for comparison. The generally strong alteration of all minerals in the anorthositic phase made compositional determinations in this lithology impossible. Analytical procedures, estimates of analytical uncertainty, and individual analyses are provided in Appendix A3. The major and trace element compositions of groundmass and glomerocryst plagioclase were also determined by analysis of mineral separates. These data are tabulated with whole rock geochemical analyses in Appendix A4.

Orthopyroxene

The number of orthopyroxene analyses obtained from sills B and C was limited, and consequently compositional variation relative to stratigraphic position could not be examined. Where observable, optical properties, particularly pleochroism, suggest that most orthopyroxene grains have compositions falling within the range of hypersthene. This is confirmed by the microprobe analyses, the majority of which lie in the field of hypersthene in the pyroxene quadrilateral (Fig. 4.20). The enstatite (En) and wollastonite (Wo) contents of hypersthene range from En_{51} to En_{69} (mol. %) and from Wo_1 to Wo_6 respectively, and there is no apparent difference in the composition of this phase in the two sills. Several analyses fall within the field of bronzite, with compositions ranging up to Wo_5En_{78} . Most of these analyses are derived from samples from sill C, but the paucity of bronzite analyses in sill B is probably a reflection of the small sample population rather than a real difference in orthopyroxene compositions of the two intrusions. Mg#s ($=Mg/(Mg+Fe^{+2})$) range from 0.51 to 0.79, which is comparable to the range of this ratio in the clinopyroxene of the sills (Fig. 4.21).

There is some degree of compositional variation within individual orthopyroxene grains, although the limited number of analyses makes interpretation of the overall pattern of zonation somewhat uncertain. All of the crystals from which multiple analyses were obtained exhibited normal zonation, with En contents differing by as much as 11 mol. %. In most analyzed grains, however, the decline in En content is only 5-6 mol. %, with most of this change occurring at crystal margins. The analyses which fall within the field of bronzite were obtained from the central regions of grains. The decline in En content (and

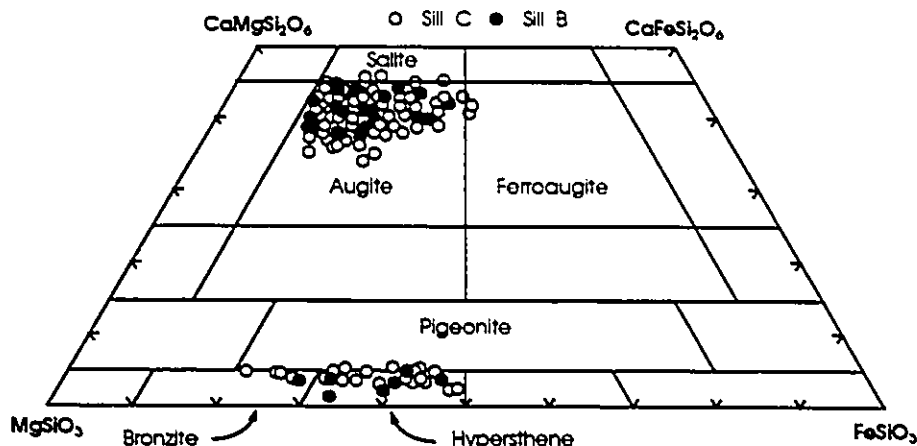


Figure 4.20. Pyroxene quadrilateral illustrating the compositions of orthopyroxene and clinopyroxene in sills B and C as determined by electron microprobe. All available analyses of orthopyroxene are plotted, but intermediate compositions within individual clinopyroxene grains are omitted for clarity. Average 2σ analytical uncertainty in En (MgSiO_3) and Wo (CaSiO_3 - apex of triangle) content is approximately twice the diameter of the plot symbols; uncertainty in Fs (Ferrosilite- FeSiO_3) content is less than the diameters of plot symbols.

Mg#) of orthopyroxene in the sills is accompanied by slight decreases in SiO_2 and Al_2O_3 contents (Fig. 4.21). Conversely, MnO and TiO_2 increase with decreasing Mg#, while CaO and Cr_2O_3 remain relatively constant.

Clinopyroxene

The majority of clinopyroxene analyses lie within the compositional field of augite in Figure 4.20, with a few falling into the fields of ferroaugite and salite. The wollastonite content of the augite ranges from Wo_{34} to Wo_{46} , while the enstatite content ranges from En_{29} to En_{51} . There is no obvious difference in the composition of augite from sills B and C, although no salite or ferroaugite analyses were obtained from sill B. The explanation for this difference probably lies in the fact that the latter compositions were obtained from the rapidly crystallized, iron-rich rocks of the basal chill zone of sill C, and no microprobe analyses were obtained from the stratigraphically equivalent horizon in sill B. The Mg#s of augite grains in the two sills are comparable, with those of sill B ranging from 0.55 to 0.80 and those of sill C ranging from 0.51 to 0.81. These ratios are comparable to the Mg#s of

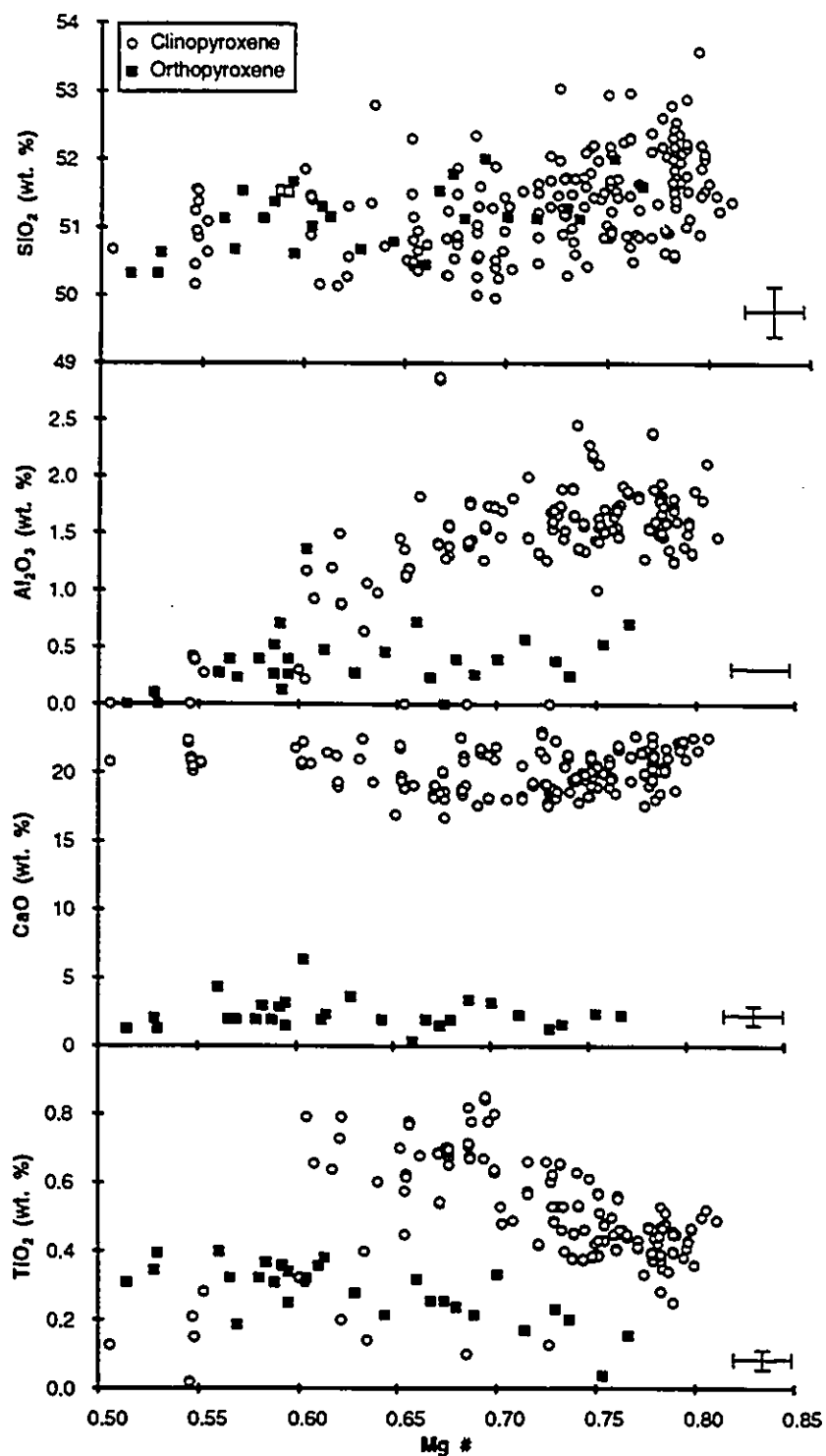


Figure 4.21. Variation in selected oxides with Mg# in clinopyroxene and orthopyroxene from sills B and C. $\text{Mg\#} = \text{Mg}/(\text{Mg} + \text{Fe}_T)$, where Fe_T is total iron expressed as Fe^{+2} . Average 2σ analytical uncertainty is illustrated by error bars where greater than the size of plot symbols.

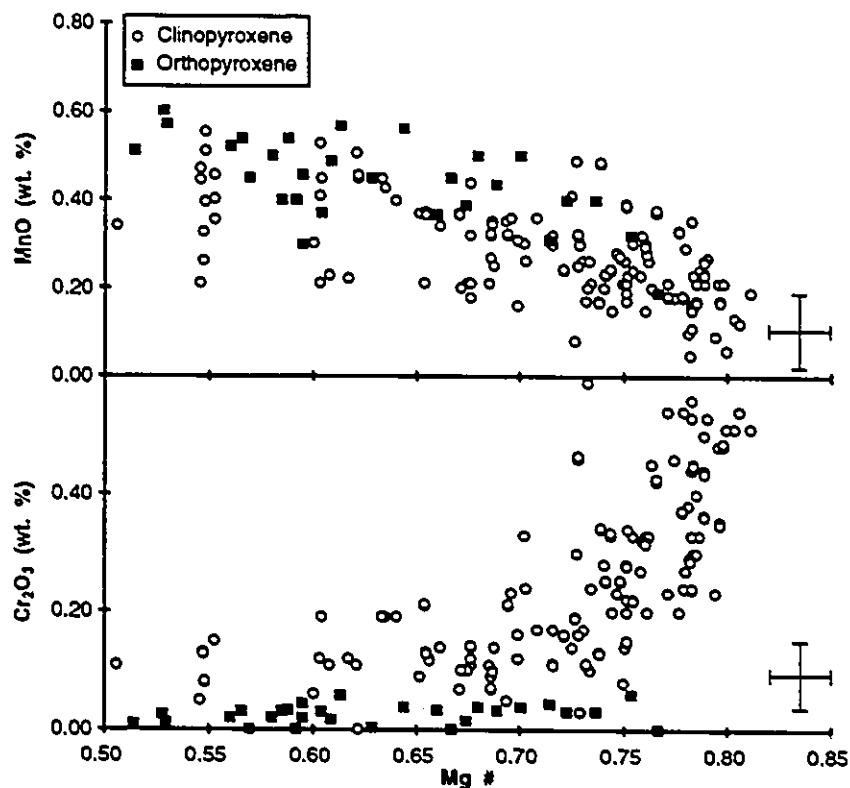


Figure 4.21 (continued). Variation in selected oxides with Mg# in clinopyroxene and orthopyroxene from sills B and C. Mg# = $Mg/(Mg+Fe_T)$, where Fe_T is total iron expressed as Fe^{+2} . Average 2σ analytical uncertainty is illustrated by error bars where greater than the size of plot symbols.

orthopyroxene in the sills.

Augite is usually zoned, and the pattern of compositional variation is equivalent to that of orthopyroxene. Typically, En content is relatively constant, or declines by only a few percent, in the central regions of the augite grains, and then declines more rapidly near the margins (Fig. 4.22). In several crystals the En content declines by 15-17 mol. % from interior to margin, and this may indicate that the extent of zonation is generally greater than is suggested by the average change in En content. Textures of most of the samples are medium grained subophitic to ophitic, with unconnected, angular areas of augite intergranular to plagioclase. Consequently determination of the position of pyroxene grain margins in the two dimensions of a thin section was often difficult. Thus it is possible that some of the analyses classified as "rim" (Appendix A3) were in fact derived from the interiors of the grains. This would result in the general underestimation of the degree of

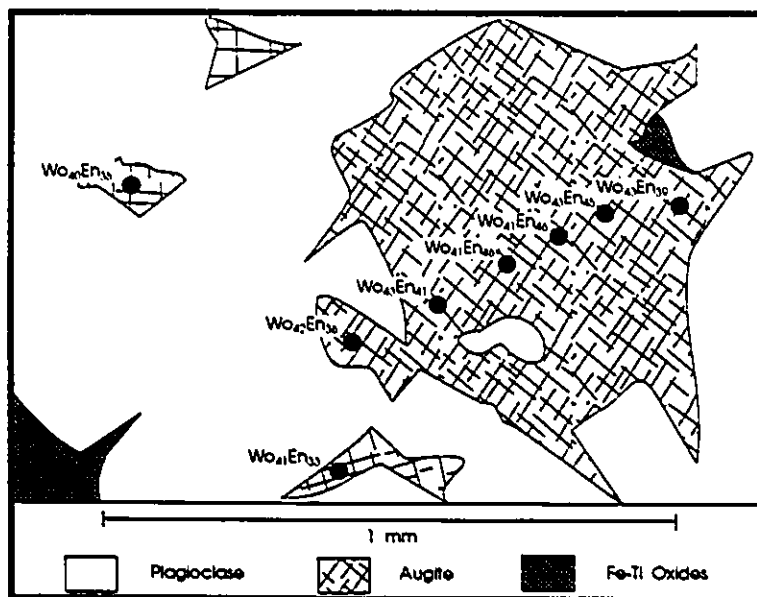


Figure 4.22. Thin section sketch illustrating the pattern of zonation in typical augite grains from sill C. The large crystal exhibits a central zone of relatively constant composition, but the enstatite (En) content declines rapidly at the margins of the grain. Wollastonite (Wo) content is also variable, but does not show a clear trend relative to crystal boundaries. The En contents of the smaller intergranular augite grains are commonly comparable or less than the marginal zones of larger crystals.

zonation. If this is the case, then the zonation in orthopyroxene may also be underestimated by the microprobe data. This problem may also explain the range of compositions found in the central regions of different augite grains within a single thin section, which may differ by as much as 18 mol. % En.

Despite the difficulties involved in determining the total range in compositions of individual grains, it is apparent that there is a general pattern of normal continuous zonation with the largest change in compositions occurring near grain margins. This is apparent in a histogram of analyses in which the En content of interior and marginal areas is compared (Fig. 4.23A). Although there is a large variation in the compositions of both groups of analyses, the outer zones of crystals (labelled "rim" in Fig. 4.23) are generally more iron-rich than the central regions. The histogram of Wo content does not reveal any obvious distinction between the compositions of rims and interior areas (Fig. 4.23B). The decline in Mg# towards marginal zones in augite grains is accompanied by slight decreases

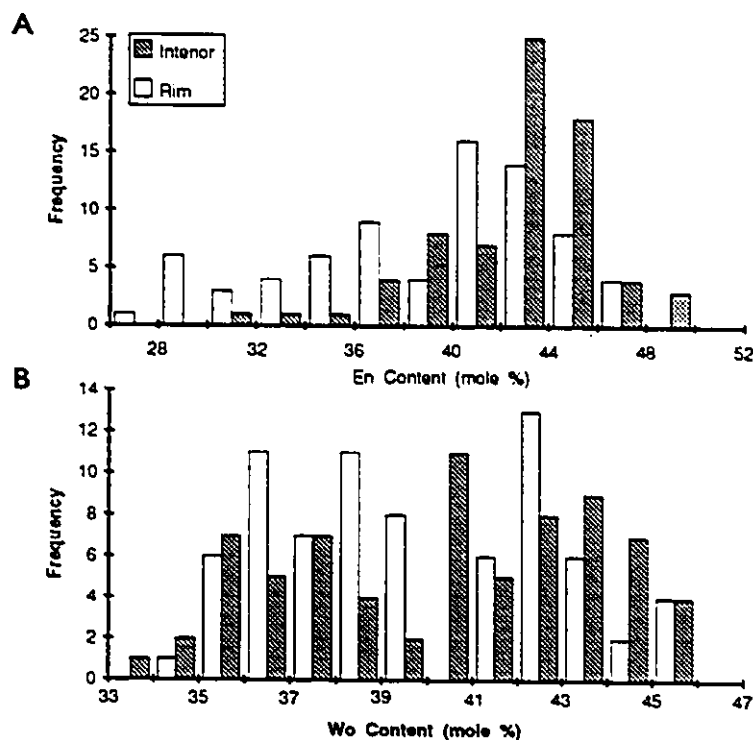


Figure 4.23. Frequency distribution of En (A) and Wo (B) contents of the interiors and rims of augite grains in sills B and C. The En content of interior areas is typically higher than that of the rims of grains, while there is no consistent difference in the Wo content of the two groups of analyses.

in SiO_2 and Al_2O_3 contents, and a much larger drop in Cr_2O_3 (Fig. 4.21). Conversely, the TiO_2 and MnO contents are elevated in iron-rich augite compositions. In keeping with the absence of coherent variation in Wo content, CaO content shows no consistent trend relative to changes in the Mg/Fe ratios.

The composition of augite in sill C is relatively constant across the main body of the intrusion (Fig. 4.24), and only near the lower and upper contacts do significant variations occur. In the fine grained gabbroic phase rocks immediately above the basal chilled margin, the maximum Mg# obtained from clinopyroxene analyses is 0.72 compared to the average of approximately 0.80 in the main body of the sill. This ratio increases to 0.76 in the next sample, and above this the Mg# is relatively constant. The maximum Mg# in augite from fine grained gabbroic phase material at the top of the sill is 0.73, again slightly lower than the main body of the intrusion. The relative constancy of the most Mg-rich augite compositions in the main part of the intrusion indicates that the

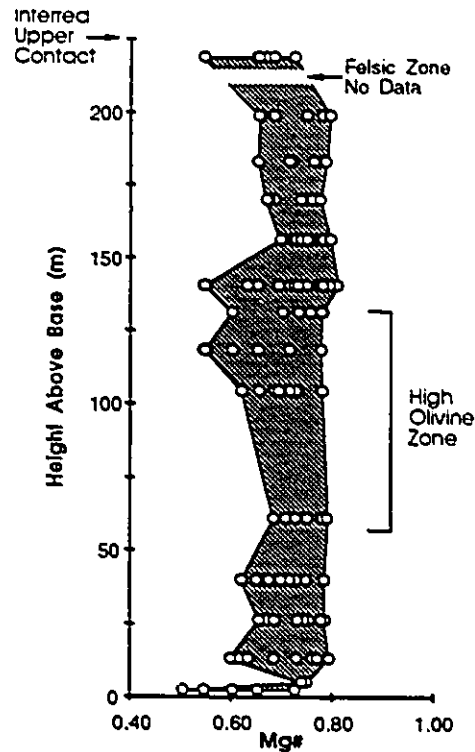


Figure 4.24. Variation in Mg# of augite with stratigraphic position in sill C. $Mg\# = Mg/(Mg + Fe_T)$ where Fe_T is total iron expressed as Fe^{+2} . The average 2σ analytical uncertainty in Mg# is less than the diameter of plot symbols.

initial stages of augite crystallization in these samples occurred under similar physical conditions from melts of similar compositions.

The difference in the maximum Mg# of clinopyroxene in fine grained zones and the main part of the intrusion may be related to two possible processes. The fine grained rocks have undergone more intense alteration and, as discussed in the next section, this has probably led to the Na-rich plagioclase compositions through albitization. Since Mg-rich pyroxene is more susceptible to replacement, higher degrees of alteration in the fine grained zones may have skewed the analytical data toward iron-rich compositions. Alternatively, the low Mg content of clinopyroxene in the fine grained zones may reflect primary compositional differences. The fact that augite typically shows comparatively little zonation in the interiors of grains suggests that the relative rates of chemical diffusion and crystal growth were initially appropriate for isocompositional crystallization and rejection of inappropriate components. The presence of large pyroxene grains in subophitic

and ophitic gabbro provides evidence in support of low degrees of undercooling and relatively slow crystallization rates in the main part of the intrusion. Because of the more rapid cooling and crystallization in the fine grained zones, however, diffusional exchange may have been less efficient, thereby resulting in more iron-rich pyroxene compositions (i.e., closer to initial melt Mg/Fe ratios).

Groundmass Plagioclase

Groundmass plagioclase in sill C ranges in composition from bytownite (An_{84}) to albite (An_4) with no compositional gaps (Fig. 4.25). The feldspars are invariably low in K_2O , with Or contents rarely exceeding 3 mol. %, even in the most Na-rich grains. The majority of albite analyses (five of six) were obtained from the fine grained zone near the base of the intrusion. The analyses were obtained from the interiors of the crystals, and since these are unlikely to represent primary compositions they are interpreted to reflect albitization of the original feldspar during alteration and metamorphism. One albite analysis was obtained from the main part of the gabbroic phase in sill C, but this was derived from the outer margin of a continuously zoned grain and is interpreted to represent a primary composition because of the continuous zonation from core to rim.

Groundmass plagioclase compositions in sill B are comparable to those of sill C, ranging from bytownite to oligoclase with some compositional gaps that probably reflect the smaller sample population. No albite analyses were obtained from sill B samples. Despite the slightly smaller range in compositions, the fact that the An content of the most calcic grains are similar suggests that there is no significant difference between the feldspar of sills B and C.

Groundmass plagioclase crystals of both sills are invariably zoned, but the extent of the zonation is highly variable. Oscillatory zonation was not observed, although reverse zonation without discontinuities may be present. Typically, groundmass plagioclase exhibits continuous or discontinuous normal zonation, with relatively calcic cores and sodic rims (Fig. 4.26). Crystals which are not enclosed ophitically by pyroxenes commonly show asymmetrical outer zones of continuously zoned material separated from an inner core of more calcic feldspar by a compositional discontinuity (Fig 4.26A). The

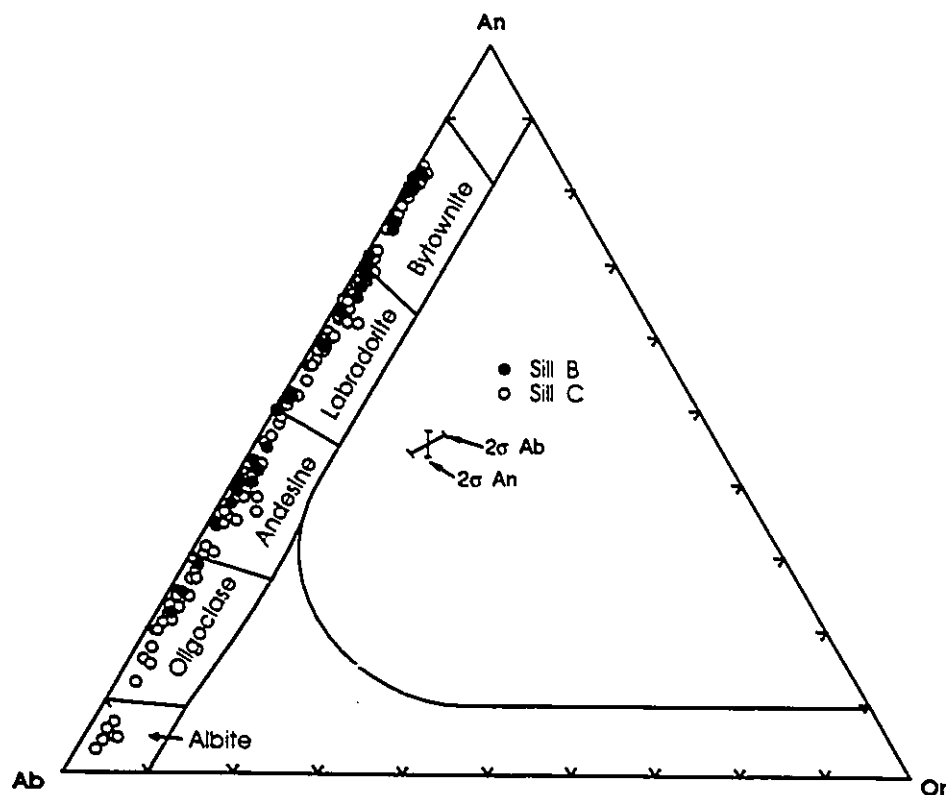


Figure 4.25. Compositions of groundmass plagioclase in sills B and C. An-anorthite, Ab-albite, Or-orthoclase. Average 2σ analytical uncertainty is illustrated by error bars for An and Ab content. Uncertainty in Or content is less than the diameter of plot symbols.

cores of such grains may be relatively homogenous, but are more often also normally zoned. In some grains, a second obvious discontinuity separates a thin rim of highly sodic plagioclase from the interior of the crystal. Other crystals exhibit continuous compositional variation from core to rim, while some are continuously zoned except for a narrow sodic overgrowth.

The variation in An content from core to rim in unenclosed crystals may be as little as 5 mol. %, and as much as 50 mol. %, but is usually between 15 and 25 mol. %. The range of core compositions of individual crystals in a single thin section is often on the same scale as the extent of zonation within individual crystals. Variations of as much as 40 mol. % An are not uncommon, and the total variation in core compositions among all crystals examined, excluding those from the albitized fine grained zones, is more than 60

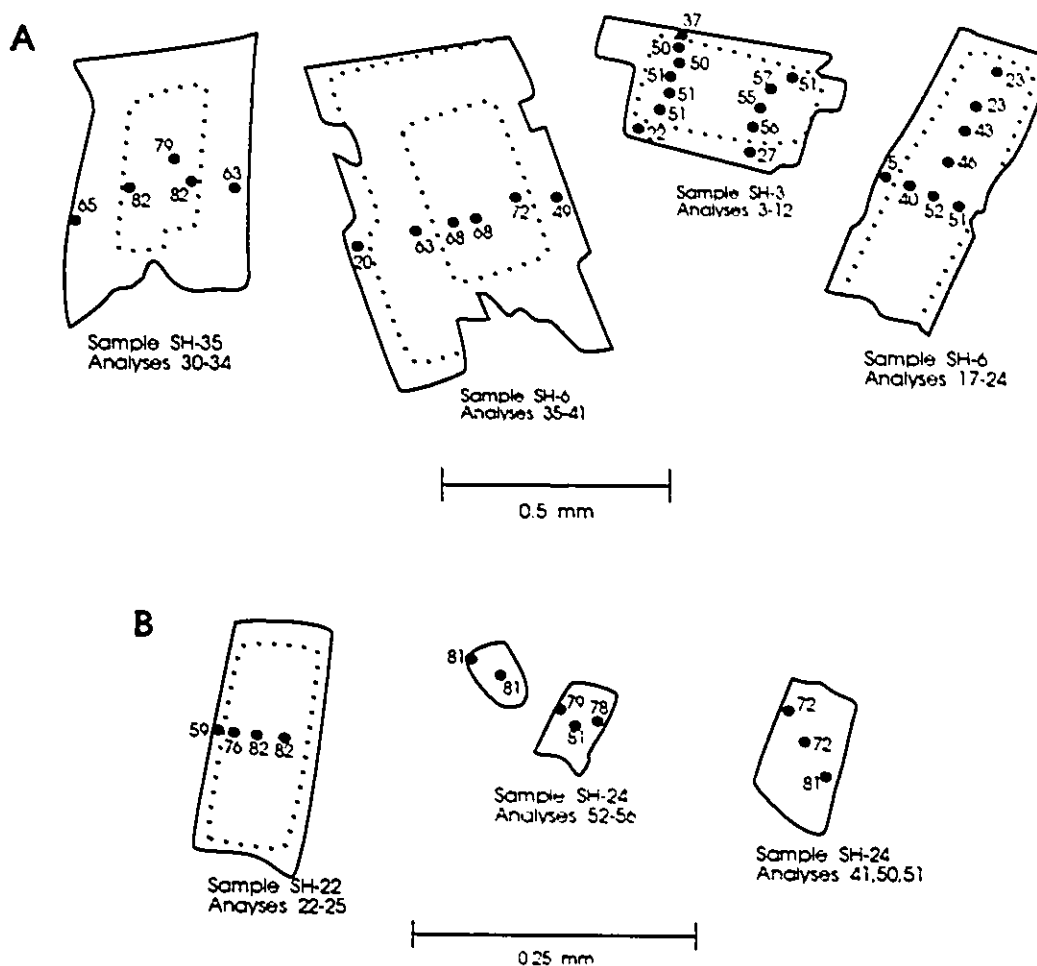


Figure 4.26. Thin section sketches of groundmass plagioclase grains in sill C illustrating the nature of compositional zonation in the feldspar. (A) Larger groundmass grains not enclosed by pyroxene. (B) Small grains completely enclosed by pyroxene. Optical discontinuities are indicated by dotted lines. Anorthite (An) content (mol. %) is indicated at the point of microprobe determination. Labels refer to sample and analysis number (see Appendix A3 for explanation).

mol. % (Fig. 4.27). In part this reflects acquisition of the microprobe data from essentially two dimensional crystals in thin section. Many crystals, however, fill intergranular spaces between larger groundmass grains, and these would be expected to have more sodic cores than earlier, framework grains. The more sodic compositions of intergranular plagioclase is consistent with the iron-rich nature of small intergranular augite crystals in comparison to the interiors of larger grains.

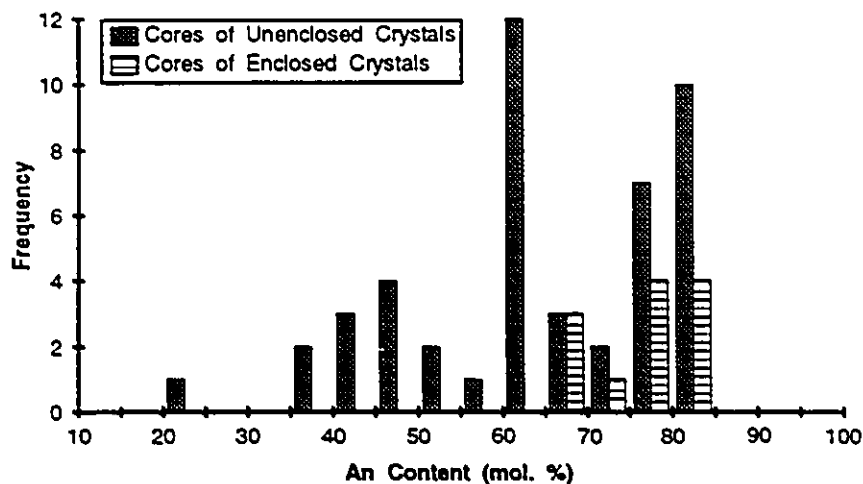


Figure 4.27. Histogram illustrating the range in core compositions among enclosed and unenclosed groundmass plagioclase crystals in samples from sills B and C.

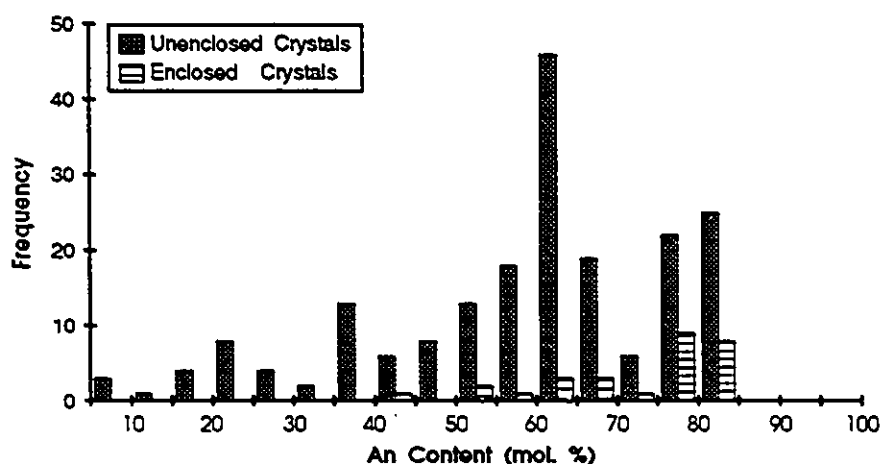


Figure 4.28. Histogram illustrating the total range in anorthite (An) content of enclosed and unenclosed groundmass plagioclase crystals in sills B and C.

Plagioclase grains which are completely enclosed by pyroxene are generally smaller than other groundmass crystals, and usually exhibit less pronounced zonation (Fig. 4.26B and Fig 4.28). Very small grains may be virtually unzoned, while the larger crystals typically show continuous variations of 5-10 mol. % An from core to rim. Less commonly, relatively homogeneous cores are discontinuously overgrown by more sodic feldspar. The range in core compositions (An_{70} - An_{84}) is also much more restricted than in those grains not enclosed by pyroxene (Fig. 4.27). The more restricted compositional variation and

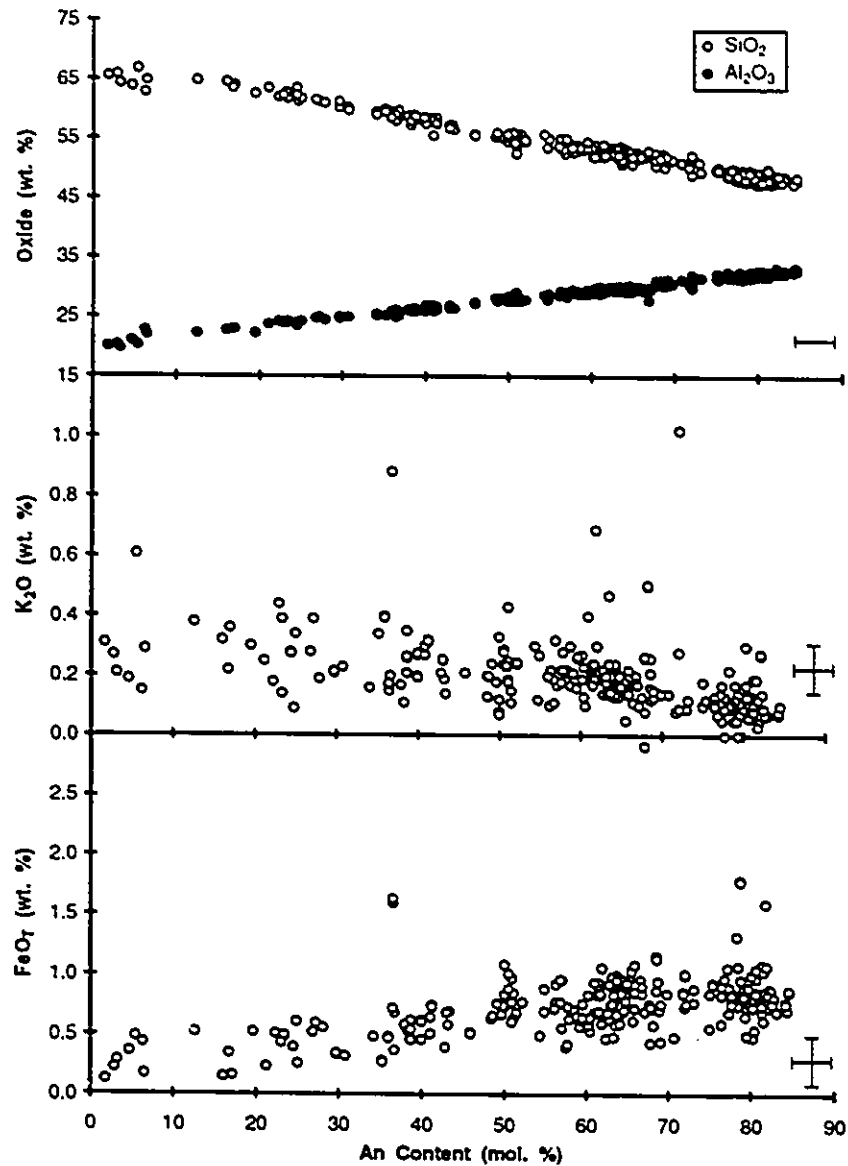


Figure 4.29. Variation in selected oxides with anorthite (An) content in groundmass plagioclase of sills B and C. Average 2σ analytical uncertainty is indicated by error bars where greater than the diameters of plot symbols.

smaller size of plagioclase chadacrysts indicate that their growth was chemically impeded by enclosure in the pyroxene grains.

Variations in the CaO and Na_2O contents of groundmass plagioclase are accompanied by stoichiometric changes in SiO_2 and Al_2O_3 , and thus the latter are well correlated with An content (Fig. 4.29). With decreasing An content, Al_2O_3 declines while

SiO_2 increases. There is also a slight increase in K_2O with decreasing An content, but this increase is considerably less than might be expected from the expansion of sodic plagioclase solid solution fields at low An contents. Whole rock geochemistry discussed in succeeding chapters indicates that the Howse Lake sills are uniformly impoverished in K_2O . Thus it is possible that the limited concentrations of the Or molecule in the more calcic plagioclase compositions (1-3 mol. %) was sufficient to prevent extensive build-up of this component in residual liquids until the last stages of pore crystallization when it was largely accommodated in intergranular orthoclase rather than plagioclase. In contrast to the slight increase in K_2O , FeO_T declines slightly as An content decreases. This is typical of plutonic plagioclase where iron becomes increasingly incompatible in the more sodic compositions due to declining temperature and Al_2O_3 content.

The maximum An content of groundmass plagioclase varies little across much of the stratigraphy of sill C (Fig. 4.30). As with the En content of clinopyroxene, however, the An content of the feldspar in the fine grained zones at the upper and lower contacts is considerably less than in the main body of the sill. As noted, plagioclase compositions near the basal contact of the intrusion are extremely sodic, and this is thought to reflect albitization during alteration and metamorphism. The succeeding sample, however, also exhibits somewhat sodic compositions in comparison to the average An content in samples from higher in the stratigraphy. This may also reflect alteration, although a comparison with compositional variations in augite suggests that the low An content may be the result of more rapid crystallization near the base of the sill. Possibly rapid crystallization at the upper and lower contacts led to less primitive feldspar and pyroxene compositions than in the more slowly cooled interior of the intrusion. The primary composition of plagioclase in the lowest sample, and possibly the highest sample, may then have been overprinted by albitization.

The small, poorly zoned plagioclase chadacrysts are best interpreted as cumulus grains whose growth was curtailed by enclosure in pyroxene oikocrysts. Core compositions in these grains are generally in the range An_{70} to An_{82} , and unzoned examples have compositions of An_{78} - An_{84} . These values are comparable to the more calcic cores in unenclosed grains, and thus initial crystallization of plagioclase in sills B and C produced material with an average composition of approximately An_{80} . As with

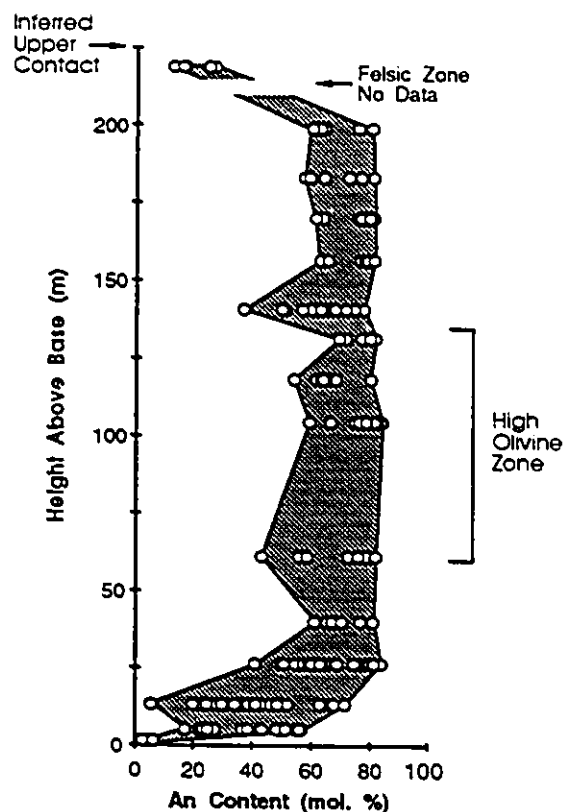


Figure 4.30. Variation in the anorthite (An) content of the cores of groundmass plagioclase crystals with stratigraphic position in sill C. Average 2σ analytical uncertainty in An content is less than the diameter of plot symbols.

clinopyroxene, the constancy of maximum An content in samples from the main body of sill C indicates that plagioclase initially crystallized from liquids whose compositions and physical properties were similar. The difference in the size and number of crystals of pyroxene and plagioclase suggests that the degree of undercooling during crystallization was appropriate for more rapid nucleation of the feldspar. The larger degree of compositional zonation of plagioclase may then reflect the growth of numerous, closely spaced grains in competition for the same chemical constituents in surrounding melt. The smaller number of pyroxene grains enabled individual crystals to grow to larger sizes with lesser zonation due to the lack of interference, at least initially, from other pyroxene grains.

Glomerocryst Plagioclase

Microprobe data were obtained from seven glomerocrysts and one phenocryst in sills B and C. Compositional data are plotted in Figure 4.31, and analyzed grains and surrounding glomerocryst crystals are illustrated in Figure 4.32. Because of the often complex pattern of zonation, analyses are classified as central (within the main body of the primocryst), intergranular (between primocrysts) and rim (peripheral zone at the glomerocryst margin). The main portions of individual primocrysts are generally bytownite or labradorite in composition, with compositions ranging from An_{87} to An_{63} (Fig. 4.31). Typically, the central portions of primocrysts are more calcic than the margins, with most grains examined exhibiting cores whose An content is 80 or higher (Fig. 4.32). A primocryst in sample F16-9 is an exception to this statement in that the crystal exhibits reverse zonation, with the An content increasing from 76 in the core to 81 in outer areas (Fig. 4.32).

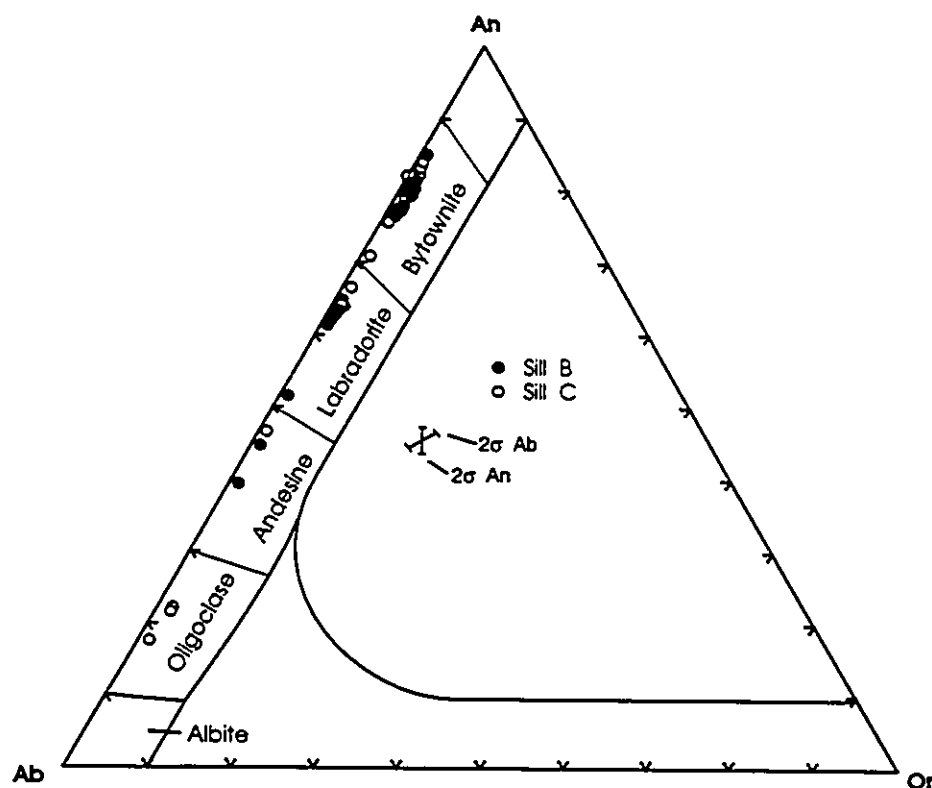


Figure 4.31. Compositions of glomerocryst plagioclase in sills B and C. An-anorthite, Ab-albite, Or-orthoclase. Average 2σ analytical uncertainty in An and Ab content is indicated by error bars. Uncertainty in Or content is less than the diameter of plot symbols.

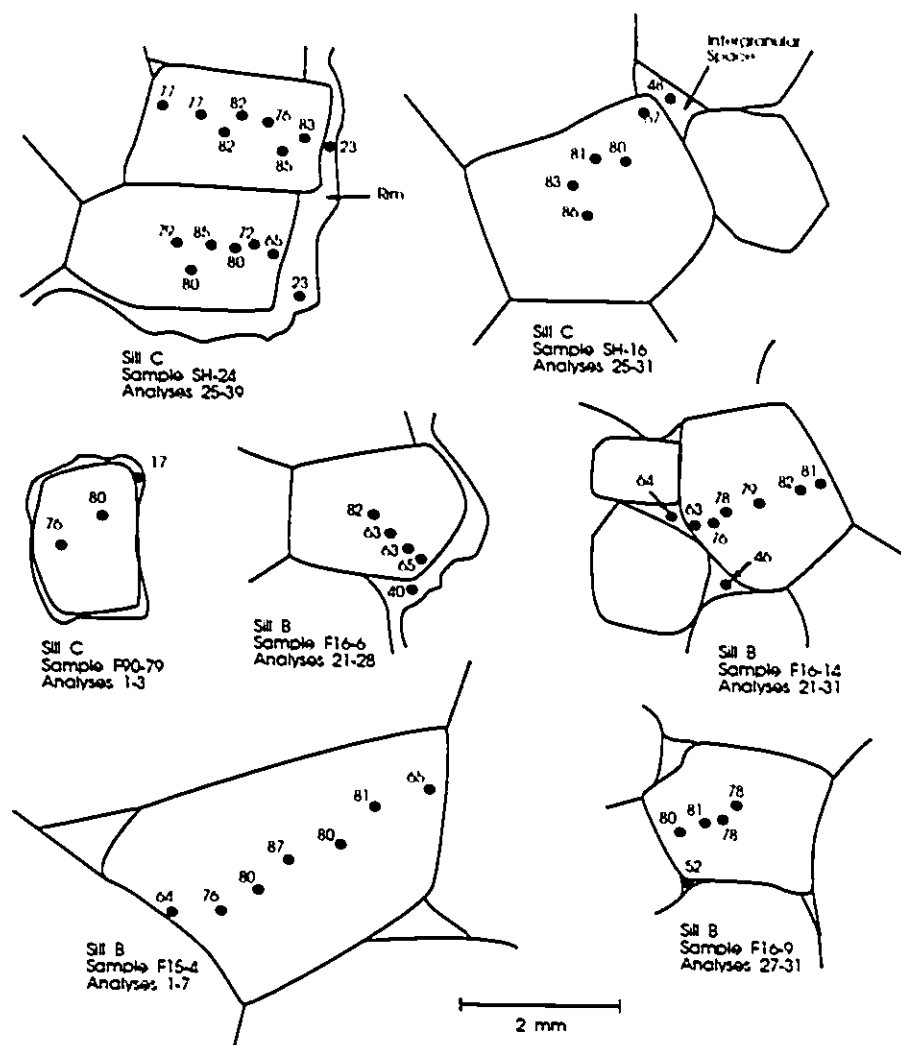


Figure 4.32. Thin section sketches illustrating the nature of compositional variation in glomerocryst plagioclase from sills B and C. Analyzed grains are shown along with portions of bordering crystals. Anorthite (An) content (mol. %) is indicated at the location of each analysis. Labels refer to sample and analysis number (see Appendix A3 for explanation).

Most primocrysts are compositionally zoned to some extent. This usually involves oscillatory or normal zonation but, as described in sample F16-9, reverse zonation is occasionally present. The An content of normally zoned crystals may decline by as much as An_{23} with the largest declines usually found adjacent to intergranular material. This indicates that pore liquids partially crystallized as overgrowths on the primocrysts.

Oscillatory and reverse zonation rarely result in changes in An content of more than 5-6 mol. %.

Intergranular plagioclase in glomerocrysts is invariably more sodic than central areas of primocrysts, although compositions may be similar to the margins of zoned primocrysts. The composition of intergranular plagioclase ranges from An_{64} to An_{31} . These compositions are within the range of groundmass crystals, and so provide no indication of the origin of pore liquids in the glomerocrysts. Intergranular material may therefore have crystallized from liquids trapped during aggregation, or from liquids which infiltrated into the pores at the time of groundmass solidification. The outer sodic rims which commonly occur at the periphery of glomerocrysts may be even more sodic, with compositions ranging from An_{40} to An_{17} . These values are comparable to the compositions of the margins of some groundmass plagioclase crystals, and it is apparent that the glomerocryst rims were added during the later stages of groundmass crystallization.

The glomerocrysts of sills B and C show similar compositional ranges and zonation patterns, despite the generally small size of the aggregates in sill C. Additionally, the composition of a phenocryst in sill C is comparable to that of the glomerocrysts, and this crystal also shows a peripheral zone of sodic plagioclase. Consequently, there is no evidence that phenocrysts and glomerocrysts in the two intrusions are derived from different sources.

The SiO_2 content of glomerocryst plagioclase increases with decreasing An content, while the Al_2O_3 content declines (Fig. 4.33). Decreasing An content is also accompanied by slight increases in K_2O content but, as with groundmass crystals, the concentration of this component is usually quite low. The variation in total iron content also mirrors the trend of groundmass plagioclase, with higher FeO_T in the more sodic compositions.

4.5 Implications of Mineral Compositional Data for Solidification Histories

Hypersthene, augite and plagioclase grains tend to be normally zoned from core to rim, and usually there are central regions in which compositions are fairly uniform. This

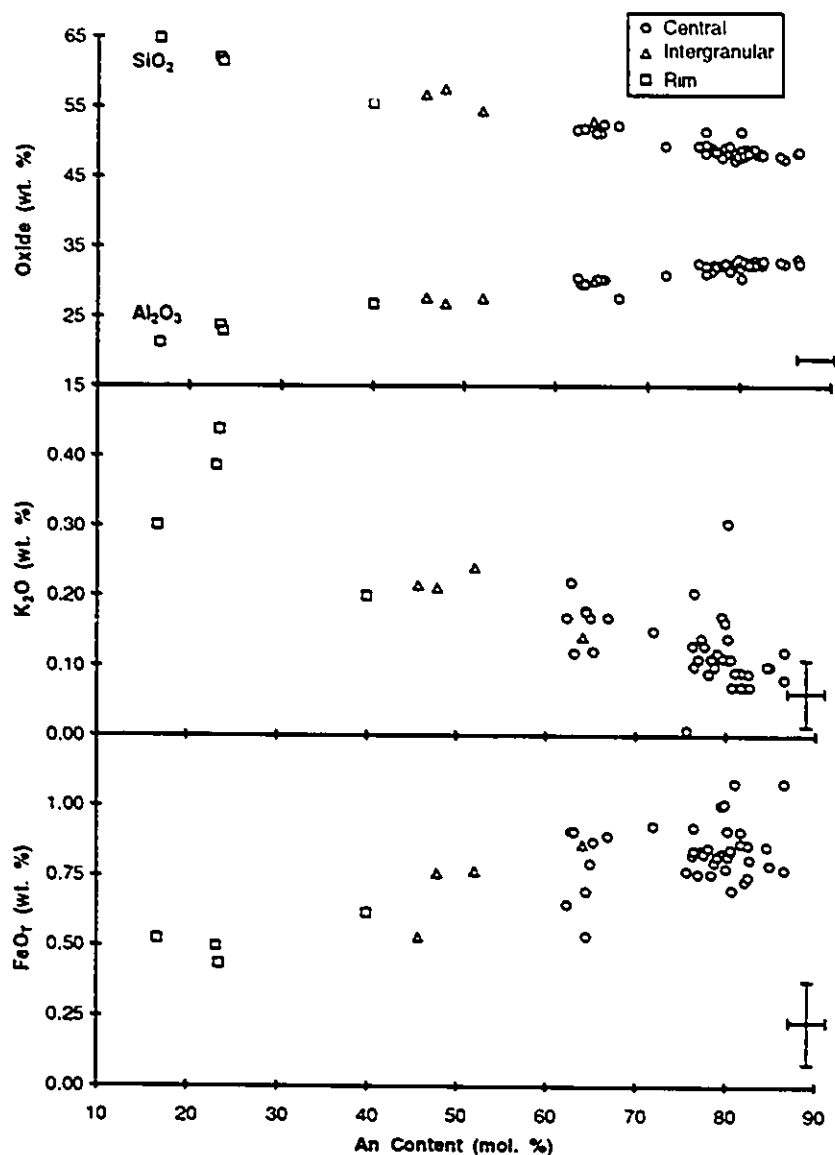


Figure 4.33. Variation in selected oxides with anorthite (An) content in glomerocryst plagioclase from sills B and C. Average 2 σ analytical uncertainty is indicated by error bars where greater than the diameters of plot symbols.

indicates that in the early stages of crystallization the required chemical constituents were exchanged for rejected solute at the growth surfaces of the crystals. The compositional zonation in the grains, and the local presence of minor intergranular Fe-Ti oxides, quartz, orthoclase, and accessory phases indicates that chemical exchange was impeded and eventually curtailed as the crystal framework grew. Further crystallization entailed lower

temperature growth of framework crystals from trapped pore melts, and solidification of low temperature phases in intergranular spaces. The gabbroic phase of the main part of sill C, and the marginal gabbroic phase horizons of other intrusions, thus consist essentially of plagioclase-augite-hypersthene ($\pm$ olivine) mesocumulates. Although no mineral composition data are available from the anorthositic phase, it is probable that this lithology also represents a mesocumulate rock. In fine grained and quenched gabbro at sill margins, rapid crystallization limited the effectiveness of chemical diffusion, and mineral compositions more closely reflect average liquid compositions.

The constancy of the maximum En content of augite and maximum An content of plagioclase across the main part of sill C may indicate that (1) initial crystallization was simultaneous throughout the intrusion (i.e., prior to significant melt evolution), (2) residual melts were replaced by undifferentiated magma (i.e., crystallization occurred in an open system), (3) the early stages of crystal growth (i.e., prior to crystallization from pore fluids) had little effect on residual melt compositions over much of the solidification history of the sill, (4) the sample profile across sill C does not represent a solidification profile, or (5) the microprobe data do not accurately portray variations in mineral compositions. To some extent, the large range in mineral compositions within a single thin-section gives weight to the latter possibility. Nevertheless, the highest temperature plagioclase and augite mineral compositions appear to be relatively constant across sill C.

Since heat loss, and therefore degree of undercooling and crystallization, varies with position in an intrusion (Irvine, 1970), simultaneous crystallization is unlikely. A more probable solidification scenario is crystallization in solid-liquid boundary layers which advanced inward from sill contacts. In the early stages of growth of framework crystals, diffusional or convective exchange of material between pore spaces in the boundary layers and the main body of magma allowed isocompositional crystallization. With increasing degree of crystallinity in the boundary layer, mass exchange became restricted and eventually stopped. Trace element data discussed in the next chapter indicate that the trapped liquid content of most samples from the main body of sill C was roughly 40-50%. The presence of oxide and felsic horizons in the upper portions of many sills indicates that solidification progressed inward predominantly from the floor and walls, while advance of the boundary layer at the roof was relatively limited.

If solidification progressed inward from sill contacts, an alternative explanation for the lack of variation in primitive mineral compositions in sill C involves wall crystallization during flow, such that residual melts were continuously replaced by more primitive magma. As discussed in Chapter 6, however, the proportional thickness of a sill which may have crystallized during emplacement is relatively minor. In addition, the presence of felsic rocks in many intrusions, including sill C, and the development of cumulate oxide horizons, indicate that melt evolution did occur. Thus while some amount of wall crystallization is probable, it is clear that the sills became closed to magma replenishment in the relatively early stages of solidification.

The development of Fe-Ti-rich liquids followed by alkali-rich compositions is a typical pattern for the crystallization differentiation of basic tholeiitic magmas (e.g., the Skaergaard intrusion; Wager and Deer, 1939). This differentiation trend is apparent in many of the Menihek sills where evolving magmas gave rise to oxide and felsic horizons. Although a cumulate oxide horizon was not identified in sill C, it is reasonable to conclude that fractionation of plagioclase and pyroxene during boundary layer solidification influenced residual melt geochemistry, and melt differentiation eventually led to the development of the siliceous rocks near the top of the intrusion. Melt evolution through fractional crystallization is also supported by the compositional zonation apparent in the outer zones of most crystals, which obviously reflects changes in composition of trapped pore fluids due to crystallization.

If it is accepted that melt evolution occurred during closed system solidification, the lack of systematic variation in the major element compositions of plagioclase and augite in sill C can be interpreted in two ways. First, fractional crystallization in the boundary layers did not effect the compositions of later minerals because the effect on residual melts was negligible until fairly late in the solidification history of the intrusion. This might reflect the fact that the plagioclase-augite-hypersthene mesocumulates, which contain ~50% trapped pore fluids, are compositionally similar to rocks which would have formed had crystal fractionation not occurred. Alternatively, it is possible that the sample profile does not record the inward progression of the solid-liquid boundary layer during solidification. For example, if solidification proceeded inward from the walls of an intrusion rather than from the floor and roof, a cross section through the sill will intersect a

series of rocks formed at approximately the same time, or in a restricted time interval relative to the entire crystallization history. Thus the sample suite would not represent a time profile recording successive stages of solidification, and successive melt compositions. Given the large range in mineral compositions, and the poor constraints on sill boundaries, the two interpretations cannot be further evaluated. But it seems likely that, to some degree, both are correct.

4.6 Origin of Glomerocrysts

Comparison Of Groundmass And Glomerocryst Plagioclase Compositions

The maxima in An content of groundmass (An_{84}) and glomerocryst (An_{87}) plagioclase as determined by microprobe analysis are indistinguishable within error ($2\sigma \sim \pm 5$). The cores of groundmass crystals, however, exhibit a much wider range in compositions than the equivalent areas in the primocrysts of glomerocrysts. Additionally, the total range of compositions in glomerocryst plagioclase, excluding sodic peripheries, is generally less than that of the groundmass. Analysis of mineral separates confirms the more calcic nature of the glomerocrysts (Fig. 4.34). The composition of glomerocryst plagioclase obtained from XRF analyses of mineral separates ranges from An_{79} to An_{72} , with an average value of An_{76} for seven samples. Similar analyses of groundmass plagioclase give compositions in the range An_{74} to An_{56} , with an average of An_{65} for six samples.

In addition to CaO, glomerocryst plagioclase is enriched in Al_2O_3 , K_2O , Ba, Sr and Rb, and impoverished in SiO_2 and Na_2O relative to that of the groundmass. The higher Ba and Sr probably reflect greater substitution of these elements into the higher temperature, CaO-rich glomerocryst plagioclase. The higher K_2O and Rb contents are thought to reflect the effects of sericitization, since K_2O contents as determined by microprobe analysis in areas completely free of alteration are considerably lower than those obtained from the mineral separates. Further compositional differences between the two generations of feldspar are found in the concentrations of the rare-earth elements (REE; Fig. 4.35). The

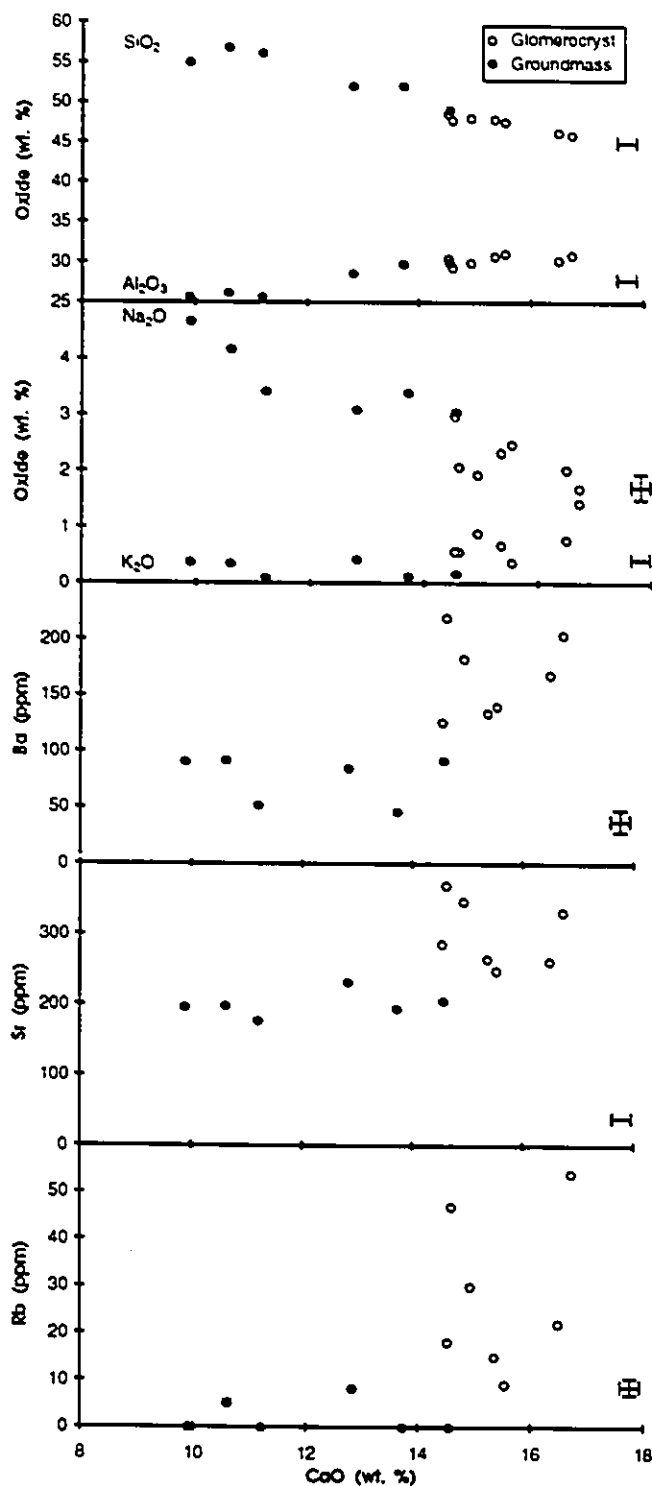


Figure 4.34. Comparison of the concentrations of selected elements in mineral separates of groundmass and glomerocryst plagioclase from sills B and C. The mineral separates were prepared from samples obtained from glomeroporphyritic sills at Howse Lake. Analytical data are tabulated in Appendix A4. Average 2σ analytical uncertainty is indicated by error bars where larger than the diameters of plot symbols.

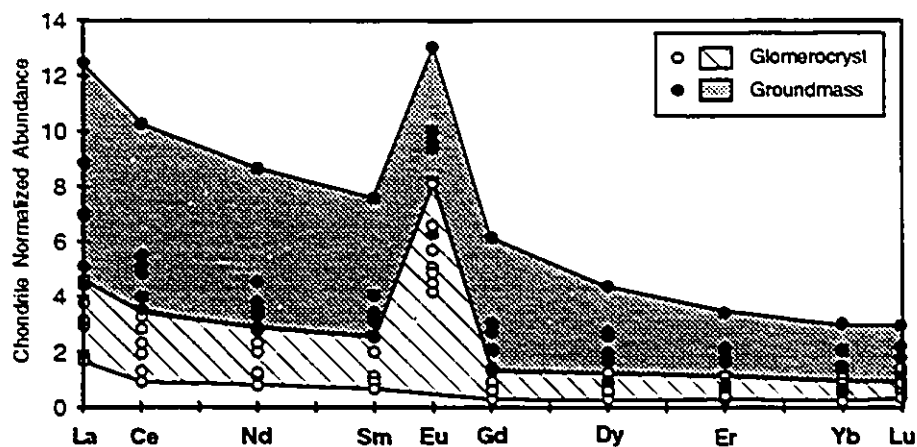


Figure 4.35. Comparison of the REE contents in mineral separates of groundmass and glomerocryst plagioclase from sills B and C. The mineral separates were prepared from samples obtained from glomeroporphyritic sills at Howse Lake. Analytical data are tabulated in Appendix A4. Normalizing values are from Nakamura (1974).

glomerocrysts are generally impoverished in all of the REE relative to groundmass feldspar, and exhibit slightly smaller light REE enrichment trends. The lower REE concentrations of the glomerocrysts are consistent with their generally more primitive compositions.

Origin

In the terminology of cumulus theory, the glomerocrysts show adcumulate textures (Wager *et al.*, 1960). The common mutual interference relationships among primocrysts, the infilling of intergranular spaces by overgrowths on primocrysts and discrete, less calcic crystals, and the paucity of other intergranular minerals, indicate that the glomerocrysts developed by crystallization and aggregation of primocrysts, with subsequent adcumulate growth. Adcumulate rocks are thought to develop by isocompositional growth of primocrysts by diffusive or convective exchange of material between pore spaces and host magma. Most glomerocryst crystals, however, show some degree of zonation. Despite this compositional variation, the glomerocrysts clearly represent the products of accumulation and subsequent enlargement of plagioclase primocrysts to the almost complete exclusion of other phases.

Morphological and compositional differences indicate that glomerocryst and groundmass plagioclase represent two generations of feldspar. The occurrence of resorption textures, the presence of unaltered sodic rims on the peripheries of outer primocrysts alone, and the occasional concentric arrangement of groundmass crystals around glomerocrysts indicate that aggregation predates groundmass crystallization. Since there is no likely mechanism by which these differences in texture and compositions could arise during the crystallization of the sills, formation of the glomerocrysts must have occurred prior to sill emplacement. Enlargement of primocrysts and expulsion of pore melt, however, may have occurred prior to, during, or after emplacement.

The zonation of primocrysts indicates that either adcumulate growth was not isocompositional or that the primocrysts were zoned prior to accumulation. While a degree of pre-accumulation zonation is possible (e.g., plagioclase phenocrysts in volcanic rocks are typically zoned), the fact that zonation extends to primocryst margins suggests that post-aggregation growth was not isocompositional. This may reflect physical or chemical changes in the host magma, which would change the composition of liquidus plagioclase, or inefficient exchange of components between host magma and pore fluids. If adcumulate growth and expulsion of pore fluids predates sill emplacement, inefficient diffusional exchange is the more likely source of the zonation. If adcumulate growth occurred, or continued, after sill emplacement, the change in P-T conditions may be partially responsible for the zonation. As noted in the previous section, the composition of intergranular material provides no evidence in support of either possibility. It would seem likely, however, that unconsolidated aggregates of primocrysts would tend to separate into individual grains as a result of collisions occurring during transport. Since phenocrysts are relatively uncommon in strongly glomeroporphyritic sills, it is probably that the glomerocrysts are essentially xenoliths derived from anorthositic rock which was disrupted and entrained in the Menihek sill magmas.

The fact that some individual groundmass plagioclase grains, particularly chadacrysts, have compositions which are only slightly more sodic than glomerocryst plagioclase suggests that the aggregates may be cognate xenoliths formed in the source reservoir(s) to the Menihek sills. Two lines of evidence support this hypothesis. First, the Nd isotopic composition of glomerocrysts and groundmass are identical. Sm-Nd data are

discussed more fully in Chapter 9, but the data pertaining to the glomerocrysts should be mentioned here. A glomerocryst plagioclase separate (sample F26-1A(1)) from the core phase of a Howse Lake sill has an ϵ_{Nd} value of +3.6 (Table 8.4) at 1884 Ma, which is the age of the sill as determined by U-Pb isotopic analysis of zircons in the matrix of the same sample (Findlay *et al.*, 1995). The ϵ_{Nd} ratio of the matrix (F26-1A(2)) is an almost identical +3.5. A similar ϵ_{Nd} value of 3.2 was obtained from the chilled margin of an aphyric Howse Lake sill (sample F20-7B). These data indicate that the glomerocrysts and groundmass gabbro were derived from source material with very similar isotopic compositions at 1884 Ma. The most logical explanation of this similarity is that glomerocrysts and matrix were both derived from the same source.

The second line of evidence in support of a comagmatic origin comes from the analysis of compositional variation in the Menihek sills and basalts presented in Chapters 5 and 8. There it is concluded that compositional differences between the chilled margins of highly glomeroporphyritic and aphyric sills derives from the pre-emplacement fractional crystallization of plagioclase and pyroxene in the source reservoir(s) to the Menihek magmas. Consequently, it is probable that the glomerocrysts developed by fractional crystallization and accumulation of plagioclase in magma chambers which subsequently served as the source of the Menihek melts (Fig. 4.36). The absence of other phases in the glomerocrysts indicates efficient separation of plagioclase and any cotectic phases.

According to Kushiro (1980), calcic plagioclase would be less dense than basaltic melts at crustal depths greater than about 10-15 km (~5 kbar). Thus flotation accumulation in the roof zones of mid-level crustal magma chambers would provide an appropriate mechanism for separation of co-precipitated plagioclase and pyroxene. Disruption of the plagioclase cumulate during melt extraction might then lead to glomeroporphyritic magmas through entrainment of anorthositic xenoliths. The effects of collisions and subsequent peripheral resorption during ascent and emplacement would tend to smooth out irregularities in the outlines of the anorthositic xenoliths, thereby producing the somewhat rounded glomerocrysts observed in the sills.

If the glomerocrysts originated through crystallization and accumulation of plagioclase in crustal chambers, aphyric to sparsely glomeroporphyritic sills may represent

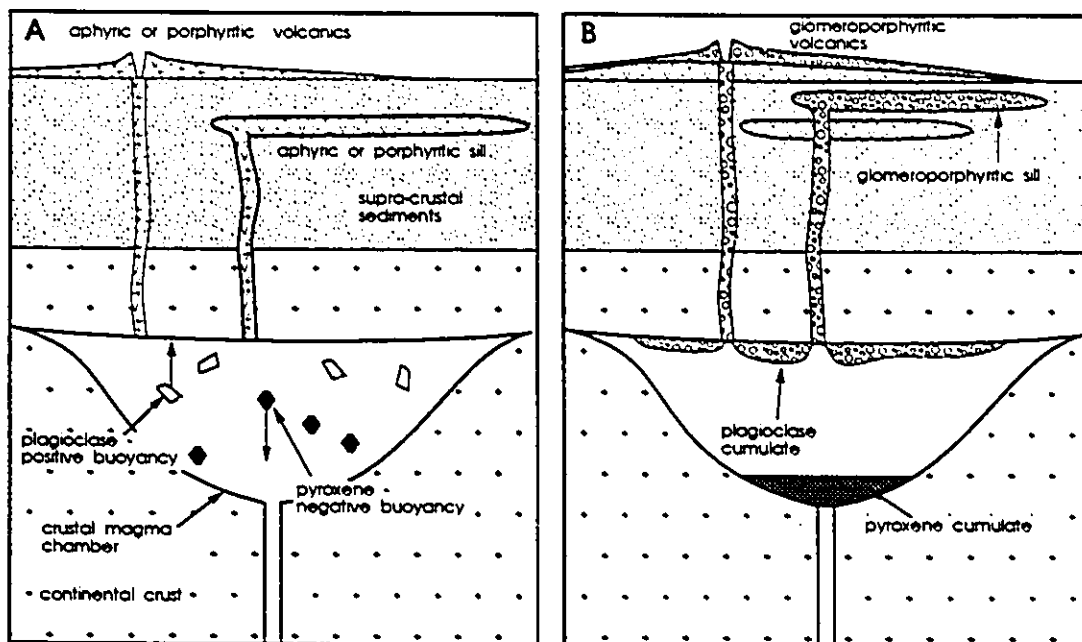


Figure 4.36. Schematic illustration (not to scale) of the development of glomeroporphyritic intrusive and extrusive rocks as a result of fractionation and accumulation of plagioclase. (A) Fractionation of plagioclase + pyroxene in a crustal magma chamber. Early extraction of melt will lead to aphyric or sparsely porphyritic lithologies. Plagioclase accumulates by flotation at the top of the chamber while pyroxene accumulates by settling at the base of the chamber. (B) Extraction of melt through the anorthositic cumulates at the roof of the chamber leads to glomerocryst laden-magmas which are emplaced as sills or erupted as basalts. Repetition of the cycle through replenishment of the chamber will produce multiple horizons of glomeroporphyritic lithologies, as well as locally cross-cutting relationships in which aphyric rocks cross-cut glomeroporphyritic units and vice-versa.

magmas extracted prior to, or between, intervals of plagioclase fractionation and accumulation. The occurrence of at least three horizons of glomeroporphyritic basalt in the Willbobo Formation (Baragar, 1960) suggests that fractionation and accumulation of plagioclase was cyclic. Repeated magma chamber replenishment would allow repetition of the cycle. This would not only produce multiple episodes of glomeroporphyritic volcanism, but would also result in locally cross-cutting relationships among aphyric and glomeroporphyritic intrusions in which either lithology may be the older. This would explain the contrasting observations of various authors regarding which lithological type

represents the earliest magmatic event (cf. Gebert, 1988; Wares and Goutier, 1989; Skulski *et al.*, 1993).

4.6 Summary

The most obvious lithological variations evident in the Menihek sills of the Howse Lake area involve differences in glomerocryst content. Variations in the size and abundance of glomerocrysts, as well as textural variations in the groundmass, give rise to three basic lithological types; the gabbroic, anorthositic and core phases. In a typical glomeroporphyritic sill, the core phase occurs within the anorthositic phase which forms a relatively thick horizon in the central part of the sill. The anorthositic interval is bounded by marginal zones of the sparsely glomeroporphyritic gabbroic phase, which is generally composed of subophitic to ophitic hypersthene gabbro. The marginal zones of gabbroic phase material may contain lenses of anorthositic phase gabbro, as well as modally and texturally layered horizons. Limited quantities of felsic rocks, ranging from quartz diorite to granite, are often present in the upper gabbroic phase, and felsic bands and veins occur throughout the sill.

Groundmass plagioclase, augite and hypersthene in sill C exhibit textures and compositional variations that suggest they are cumulus crystals. The compositional and textural similarities of groundmass minerals in sills B and C indicate that magma compositions and crystallization processes were broadly similar in the two intrusions. The abundance of olivine in sill C, however, indicates that either the magma from which this intrusion developed was slightly more primitive, or that the olivine was present in the melt before emplacement. Mineral zonation patterns and the local presence of low temperature phases in intergranular areas indicate that the gabbroic phase in the sills consists of mesocumulate rocks formed by fractional crystallization of plagioclase and pyroxene in boundary layers adjacent to sill walls. The occurrence of felsic rocks and cumulate oxide horizons indicate that the intrusions became closed to magma replenishment at some point in their development, such that melt evolution occurred. The absence of cryptic variation in mineral compositions across sill C suggests that the sample suite represents a single

interval in the crystallization history of the intrusion rather than a temporal cross-section reflecting various stages of solidification.

Glomerocryst and groundmass plagioclase exhibit differences in morphology, alteration and compositional zonation which indicate that they form two distinct populations of crystals. The maximum An content of groundmass and glomerocryst plagioclase in the sills is indistinguishable given analytical uncertainty, but the average composition of glomerocryst plagioclase is considerably more calcic. Additionally, the extent of compositional zonation in glomerocryst primocrysts is much less than that of groundmass crystals. These observations, and textural relationships between glomerocrysts and groundmass, indicate that the aggregates were formed prior to emplacement. Isotopic data indicate that glomerocrysts and groundmass were derived from sources with identical isotopic compositions. The best interpretation of these features involves formation of the aggregates through fractional crystallization and accumulation of plagioclase in the reservoir from which sill magmas were subsequently extracted.

Chapter 5

Whole Rock Geochemistry of the Menihek Gabbros

5.1 Nature of the Data Set

The geochemical characteristics of the Menihek gabbros were determined through analysis of the major and selected trace element compositions of 155 samples. These include samples collected from measured profiles across sills B and C at Howse Lake, and others obtained from several additional sills in the area, including sills A and D. The suite from sill C is representative of the dominantly aphyric to sparsely glomeroporphyritic intrusions present in the Howse zone, while that of sill B is typical of the more complex sill type in which several lithologies are present. Together, the two suites encompass all of the lithological variants described in the preceding chapter. The geochemical data are used to examine the nature of geochemical variation in sills B and C, and to determine the source of this variation. The relationship between aphyric to sparsely glomeroporphyritic sills and their highly glomeroporphyritic counterparts in terms of magma compositions is also discussed. Sample locations are described in Appendix A1, and geochemical data, details of analytical techniques and estimates of analytical uncertainty are provided in Appendix A4.

5.2. Potential Sources of Geochemical Variation

The Menihek Gabbros are a relatively homogeneous group of intrusions in which the greatest lithological differences among individual sills arise from the distribution of plagioclase glomerocrysts. The dominant rock type of the sills is hypersthene gabbro, but it is apparent from mineralogical variation in the gabbroic lithologies, and in the occurrence of more evolved rock types (e.g., ferrogabbro, granite), that some process has imparted a degree of compositional variability to the intrusions. As discussed in the preceding chapter, the sills are largely composed of mesocumulate rocks, and hence fractional crystallization

has contributed to compositional variations. Other magmatic differentiation mechanisms which can reasonably be hypothesized to have contributed to compositional variation within and among the Menihek sills are magma mixing, contamination and flow differentiation. In addition, compositional diversity resulting from magmatic processes may have been overprinted by post-crystallization alteration and metamorphism, the mineralogical effects of which have been described previously.

Fractional Crystallization

Fractional crystallization may have influenced sill compositions prior to emplacement as well during *in situ* solidification. In the case of the Menihek glomeroporphyritic gabbros, crystallization of plagioclase clearly occurred prior to emplacement of the sills. The presence of phenocrysts and microphenocrysts in chilled margins indicates that clinopyroxene, orthopyroxene and olivine were also pre-emplacement liquidus phases. The effect of pre-emplacement fractional crystallization on the compositions of sequential magma batches derived from the source reservoir are discussed in the final section of this chapter and in Chapter 8, wherein basalt geochemistry is shown to provide a better measure of batch variation.

In situ fractional crystallization is an almost inescapable process in basic sills which solidify over some interval of time. Although the extent of any resultant compositional variation is dependent on crystallization time and fractionation mechanism, it is considered a more plausible source of variation in the Menihek sills than either magma mixing or contamination. Evidence in support of *in situ* fractional crystallization is found in the mesocumulate textures which predominate in the sills, in the nature of compositional zonation in groundmass crystals, and in the occurrence of cm-scale layering and cumulate oxide horizons. Given appropriate conditions, *in situ* fractional crystallization may have also produced the limited amounts of felsic rocks which occur in many of the sills.

Magma Mixing

Part of the observed geochemical variation in the Menihek sills may be due to mixing of magmas. Such a process may occur on a regional scale (i.e., affecting the compositions of different sills or flows in a regional suite), or on a local scale (i.e.,

affecting the compositions of individual sills or flows). In most situations where mixing of dissimilar magmas has been invoked to explain the geochemical or petrographic features of a magmatic suite, the evidence is somewhat ambiguous. In many cases, the intimate juxtaposition and intermingling of contrasting rock types has been cited as evidence to support magma mixing, which then provides a mechanism for the formation of intermediate rock types (e.g., Askja, Alaska; Sigurdsson and Sparks, 1981 and Medicine Lake, California; Gerlach and Grove, 1982). The presence of contrasting magma compositions, however, and their subsequent crystallization as distinct rock types would seem to present as much evidence to support non-mixing of magmas as to support melt hybridization (cf. Gamble, 1979). In a few instances, such as the Gardiner River basalt to rhyolite complex in Wyoming (Wilcox, 1979), and the ferrodioritic to granitic intrusive rocks of Mt. Marsco on the Isle of Skye (Wager *et al.*, 1965), the complete gradation in rock types coupled with the lack of evidence for extensive crystal fractionation support the mixing hypothesis. Irrespective of whether or not magma mixing took place in these complexes, a feature common to each seems to be the juxtaposition of different magma compositions represented by distinct rock types.

On a regional scale, the igneous rocks of the Labrador Trough are predominantly of basic tholeiitic compositions, and the principal members of the LTBS are lithologically and geochemically similar (cf. Dimroth, 1978). There are no significant occurrences of the products of dissimilar magma types in the Labrador Trough, although minor quantities of alkaline rocks occur in the Chakonipau, lower Murdoch, lower Menihek and Nimish formations (see Chapter 8). If mixing of different magmas contributed to geochemical variation in the parental magmas of the LTBS, including the Menihek sills, then the degree of mixing was sufficiently high to result in melt homogeneity, and the site or timing of mixing was such that the endmember magmas were not erupted. Since the dominant composition of the LTBS is subalkaline basalt which can be considered as the direct derivative of a primary mantle partial melts (see Chapter 8), and since no alternative magma types are widespread in the Trough, mixing of dissimilar magma types on a regional scale is unlikely.

Mixing of magmas on a local scale (i.e., individual intrusions) offers a potential explanation for some of the geochemical variation present in the Menihek sills. Multiple

injection and mixing of magmas has been invoked to explain numerous features in large layered intrusions which are inconsistent with single stage fractional crystallization. The process may involve repeated injection of similar melts into a magma chamber, with either mixing of the fresh injection with residual liquids, or displacement of residual liquids by the unfractionated melt. The Muskox intrusion, for example, is thought to have formed from at least 25 separate influxes of magma (Irvine and Smith, 1967). Supporting evidence includes the presence of 25 cyclic units, each consisting of layers of cumulate rocks showing mesocumulate to adcumulate textures. The layers progress from dunite to pyroxenite and gabbro moving upward. The cyclic units also display cryptic layering in which the cumulus phases trend toward lower temperature compositions moving upward through the sequence. Similar theories have been advanced for variations exhibited by the Bushveld and Stillwater complexes, with the exception that sequential injections are thought to represent two distinct magma compositions (Irvine *et al.*, 1983). In each of these intrusions, the characteristic effects of multiple injection and magma mixing are the development of repetitive cumulate stratigraphy, including layers of essentially monomineralic rock, and the generation of cryptic variation in mineral compositions.

The Menihek sills show no evidence of this type of process. Most of the rocks are mesocumulates consisting of at least three silicate phases; there are no monomineralic or pseudo-monomineralic layers. Layering, where present, is on a scale more compatible with an origin by *in situ* crystal fractionation and mechanical alignment than with multiple magmatic injection (cf. Maaloe, 1978; McBirney and Noyes, 1979). There is also no evidence of cryptic layering; the compositions of augite and plagioclase in sill C, for example, are essentially constant across the main part of the intrusion (Fig. 4.24 and 4.30). Finally, geochemical evidence discussed in Chapter 8 indicates that the parental melts to the Menihek gabbros and comagmatic volcanics underwent a common period of differentiation prior to emplacement or eruption. This precludes mixing of magmas in the sills themselves. Multiple injection and mixing of similar or dissimilar magmas is therefore an unlikely source of variation in the sills. If mixing did not occur, some of the sills might still represent composite intrusions developed through multiple injection and independent solidification of different melts. Glomeroporphyritic sills usually contain intervals of coarse grained anorthositic phase material bounded by medium grained hypersthene

gabbro, and some include horizons of the core phase within the anorthositic phase. It is possible that the three lithologies represent separate injections of different magmas, or at least separate injections of magmas with variable glomerocryst loads. Baragar (1960) suggested that some sills in which central zones of aphyric gabbro are bounded by glomeroporphyritic lithologies represent composite intrusions. The noritic layer which underlies the core phase in sill B at Howse Lake also seems to represent a second magmatic injection emplaced along a plane of weakness in the original intrusion. Since no chilled contacts are present, it is likely that the dyke was injected during or immediately following crystallization of sill B. There are several lines of evidence, however, which indicate that the highly glomeroporphyritic rock types in the sills are products of mechanical processes operating on the crystal-laden magmas during emplacement. Notable among these is the ubiquity of development of the central anorthositic phase-marginal gabbroic phase pattern in the Menihek sills. This would seem to preclude an origin in which the anorthositic and gabbroic phases represent the crystallization products of separate magmas. Discussion of the relationships among the various lithologies is continued later in this chapter and in Chapter 6, where details of glomerocryst distributions are presented.

Contamination

The most likely source of material for post-emplacement contamination of the Menihek sills is the Menihek Formation, which hosts most, if not all, of the intrusions in the Howse zone. The extent to which assimilation of country rock has influenced the geochemistry of the sills is difficult to evaluate. As described previously, significant melting of sedimentary xenoliths is unlikely given that they retain primary structures and angular shapes. Where contacts are exposed, the Menihek strata have been converted to spotted hornfels, but these also exhibit primary bedding, and there is no obvious evidence of anatexis. Unfortunately, upper contacts, across which the highest thermal gradients would be established, are rarely exposed. The degree of contamination caused by melting at the contacts, however, would be limited by rapid crystallization of the chilled margins and development of thermal boundary layers (cf. Huppert and Sparks, 1985).

Consequently, the potential for wall rock assimilation in the Menihek sills was probably less than the potential for assimilation of sedimentary xenoliths found within the sills.

Evidence in support of at least some degree of contamination comes from the composition of sulphur in the Howse Lake sills. The $\delta^{34}\text{S}$ values of interstitial sulphides range from +5 to +22 (Findlay *et al.*, 1990). These ratios indicate a sedimentary origin for the sulphur, and the most likely source is the often sulphide-rich Menihek Formation. This evidence, however, is tempered by the fact that sulphur would be one of the most volatile components of the country rocks, and may have been incorporated in the melts by ionic diffusion across the solid-liquid interface (cf. Mohr, 1987). Therefore the presence of sedimentary sulphur in the gabbroic rocks does not necessarily require bulk contamination by assimilation of wall rock.

In contrast to the sulphur data, the Sm-Nd isotopic compositions of the sills suggest that contamination by any type of crustal material was limited. Isotopic data are provided in Chapter 9, which includes an analysis of the genetic relationships among members of the LTBS. A few of the conclusions have a bearing on the potential for *in situ* contamination of the Menihek sills, and these are briefly noted here. The geochemical characteristics of the upper Menihek and Willbob basalts, and the Menihek gabbros indicate that these rocks were derived from the same magma source. The initial ϵ_{Nd} ratios of the basaltic rocks range from +2.7 to +4.7 at 1884 Ma (Table 9.2). The equivalent ratios are +3.5 and +3.2 in the groundmass of the core phase of a glomeroporphyritic sill (sample F26-1A(1)) and the chilled margin of an aphyric sill (sample F20-7B), respectively. The ϵ_{Nd} ratios of the Baby Formation turbidites of the northern Labrador Trough, which are probably correlative with those of the Menihek Formation, range from -3 to -10 (Skulski *et al.*, 1993). Since *in situ* contamination of the extrusive rocks is unlikely, the fact that the ϵ_{Nd} ratios of the sills fall within the basalt range indicates that widespread *in situ* contamination of the intrusions is also unlikely given the isotopic composition of the host rocks. It is possible, however, that localized contamination of the sills took place, possibly contributing to the compositions of the limited amounts of felsic rocks present in many sills.

Flow Differentiation

Flow differentiation is the process by which suspended solids become concentrated in the centres of moving bodies of fluids. In magmas, this results in the concentration of phenocrysts in the central portions of sills and dykes during emplacement, a process previously proposed as the origin of the highly glomeroporphyritic anorthositic gabbros of the Labrador Trough (Baragar, 1960). Flow differentiation as it relates to the distribution of glomerocrysts in the Menihek sills is discussed in Chapter 6, but the potential geochemical effects of heterogeneous glomerocryst distribution should be noted here.

The glomerocrysts are interpreted as the products of plagioclase fractionation in the source reservoir to the Menihek gabbros. Although the glomerocrysts are thus comagmatic with the sills, heterogeneous distribution will still produce variation in the whole rock compositions of samples obtained from glomeroporphyritic intrusions. The extent of geochemical variation imparted by the glomerocrysts is dependent on their composition and concentration. The compositional differences within or among sills is thus a function of variation in host melt chemistry coupled with the variation in glomerocryst content and composition. To examine groundmass variations in a given sill, and to compare groundmass compositions among sills, the effect of the glomerocrysts must be removed. For sparsely glomeroporphyritic samples, this was achieved by serial sectioning of the samples and the use of only glomerocryst-free material for analysis. There is the possibility, however, of re-equilibration between glomerocrysts and host melt, prior to sill crystallization. But chemical exchange between glomerocrysts and melt can be considered minimal for two reasons. First, the glomerocrysts have not re-equilibrated with host melts, as documented by the differences between average groundmass and glomerocryst plagioclase compositions and by the retention of strong zonation in glomerocryst primocrysts. Second, the compositional differences between the two generations of plagioclase are relatively minor, and even complete re-equilibration would have little impact on melt geochemistry. Consequently, the contaminating effect of glomerocryst plagioclase on the groundmass composition of sparsely glomeroporphyritic samples is considered minimal, and geochemical variations must reflect other processes.

Separation of glomerocrysts and groundmass in samples of the anorthositic phase was impossible due to their intimate intergrowth, and analytical data from this lithology

represent both groundmass and glomerocryst compositions. Consequently, geochemical differences among anorthositic phase samples derive from variations in glomerocryst content and composition, in addition to variation in groundmass composition. Since the range in the average composition of glomerocryst plagioclase is not large, compositional diversity among samples of the anorthositic phase must principally reflect variations in glomerocryst content and groundmass composition.

Alteration and Metamorphism

The widespread use of geochemical data for modelling and discriminant purposes has generated considerable interest in the mobility of elements during alteration and metamorphism. Some generally recognized trends include the immobility of Al, Ti, P, Cr, Zr, Y, Nb, Sc, V, Ga, Hf and the rare earth elements, and the mobility of Si, Ca, Na, K, Sr and Ba during breakdown of primary phases. Such generalizations, however, are notoriously prone to exception depending on specific circumstances (Vallance, 1974). For example, Cann (1970), Pearce and Cann (1971) and Winchester and Floyd (1977) concluded from studies of seafloor basalts that Ti, Zr, Y and Nb are immobile during greenschist metamorphism and hydrothermal alteration. In contrast, Beswick (1983) identified Y and Zr as mobile elements during metamorphism and alteration of a komatiitic flow in the Abitibi greenstone belt.

The primary minerals of the Menihek gabbros have suffered variable degrees of replacement as a result of alteration and sub-greenschist grade metamorphism. The most serious effects are serpentization and chloritization of groundmass hypersthene and olivine, and epidotization, albitization or sericitization of glomerocryst plagioclase. Although compositional variations due to element mobility have been minimized by choosing the most pristine samples available for analytical work, no sample is completely free of secondary phases. Consequently element mobility must be a concern in the interpretation of geochemical data.

Whatever the primary differentiation mechanism, the degree of mobility of incompatible elements (i.e., those whose absolute abundance in the system is unchanged

during differentiation) can be established by comparison of their ratios. For compatible elements, however, this approach can not be applied. Without completely unaltered equivalent rocks, there is no way to quantitatively determine which, if any, elements in an altered rock have been mobilized. Fortunately, element redistribution during alteration or low grade metamorphism of plutonic rocks in relatively dry environments generally occurs on a relatively small scale and will not affect whole rock sample compositions. Consequently, with the judicious choice of samples, the effect of element mobility on sample compositions can reasonably be expected to be minor in comparison to analytical uncertainty. In the subsequent analysis of compositional trends in the Menihek sills, coherent and petrologically plausible results indicate that this a valid assumption.

5.3 Pearce Ratios

The interpretation of compositional variations in the Menihek sills relies on the use of Pearce ratio diagrams, which were introduced by Pearce (1968) as a quantitative method of interpreting and illustrating geochemical data. The statistical theory behind, and use of, Pearce ratios have been investigated by Pearce (1978), Stanley and Russell (1988), Russell and Nicholls (1988) and Ernst *et al.* (1988). The development of the approach was a response to the difficulty of deriving quantitative interpretations of geochemical trends due to the nature of the raw data. Because geochemical data reflect concentration (e.g., wt. % $\text{SiO}_2 = 100 \times \left(\frac{\text{mass SiO}_2}{\text{system mass}} \right)$), values are reported with respect to a constant total (e.g., system mass = 1 or 100%). This procedure may introduce spurious correlations among parts of a non-constant total due to closure (Chayes, 1964). Given that the mass of an igneous system, for example, is generally non-constant, comparison of element concentrations as a proportion of the system mass may prove misleading.

One solution to this type of problem is to present data in comparison to a compositional variable that can be considered constant in the system under consideration. This is the approach of Pearce ratios which have the form $\frac{A}{B}$, where A and B are compositional parameters with the same units, such as wt. % SiO_2 and wt. % P_2O_5 . This removes the constant total implied in geochemical data:

$$R = \left(\frac{\text{mass SiO}_2}{\text{total mass}} \right) / \left(\frac{\text{mass P}_2\text{O}_5}{\text{total mass}} \right) = \frac{\text{SiO}_2}{\text{P}_2\text{O}_5}$$

where R is the Pearce ratio. A conserved element, one present in a constant amount throughout the evolution of the system, is used as the denominator in the ratio. As discussed by Ernst *et al.* (1988), the nature of conservation is model dependent. Hence for a magma fractionating plagioclase, CaO is a non-conserved element whereas P₂O₅, which does not partition into the feldspar, is conserved. The mass of P₂O₅ in the system does not change during plagioclase fractionation, although clearly its concentration in the liquid will.

The Pearce ratio approach allows comparison of sample composition in terms of material added or removed during sample formation, and provides a method of monitoring the proportions of material added or removed. Consequently, it can be used to test hypotheses regarding the process responsible for sample variation. This is accomplished by plotting ratios of interest for a sample suite and determining the equation for the line of best fit by regression analysis. The slope of the regression line can then be used to test suspected differentiation mechanisms (cf. Russell and Nicholls, 1988).

To illustrate the information that can be derived from a Pearce ratio diagram, consider a hypothetical intrusion in which a picritic parental magma has differentiated by fractional crystallization of olivine. Fractional crystallization and accumulation of olivine produces dunite cumulates and a series of residual liquids which are periodically erupted as basaltic flows. In appropriately constructed Pearce ratio diagrams, the relationship among the solidified products of the system are readily apparent. One such diagram is shown in Figure 5.1 in which the ratios (Mg+Fe²⁺)/Ti and Si/Ti of hypothetical samples from the system are plotted. Because the samples are comagmatic and related solely by olivine crystallization, they define a strongly correlated trend whose slope is 2. The slope derives from the fact that the cation ratio of (Mg+Fe²⁺) to Si in olivine, (Mg+Fe)₂SiO₄, is 2.

The initial magma, represented by the chilled margin of the intrusion, plots in the central region of the sample array. The basalts, which are derived from the parental melt by removal of olivine, plot along the trend line closer to the origin. Their lower

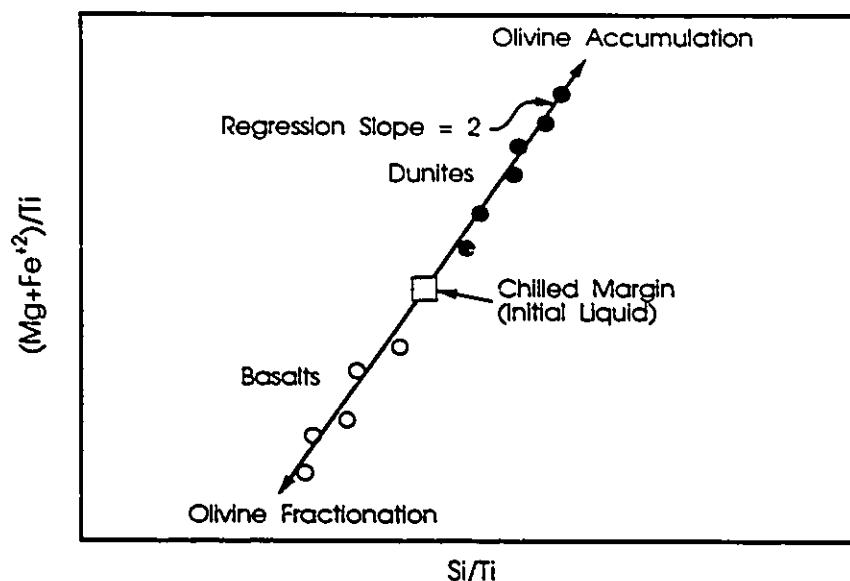


Figure 5.1. Pearce ratio diagram depicting olivine fractionation in a hypothetical magma chamber. Fractionated olivine accumulates to form dunitic cumulates while residual liquids are periodically erupted as basalts. The composition of the initial magma is inferred from the chilled margin of the magma chamber

$(\text{Mg}+\text{Fe}^{+2})/\text{Ti}$ and Si/Ti ratios are the result of higher Ti concentrations in residual melts. This reflects the mass conservation of Ti, and the reduction in system mass by removal of Mg, Fe^{+2} and Si in olivine. The converse relationship holds for the cumulate samples, which plot above and to the right of the chilled margin, and have been derived from the parental liquid by addition of olivine. Since Ti is not a constituent of olivine, its concentration in the cumulate is lower than in the original liquid, and therefore the $(\text{Mg}+\text{Fe}^{+2})/\text{Ti}$ and Si/Ti ratios are higher. The change in ratios thus reflects deviation from the starting composition, and the extent of this deviation is a measure of the extent of the process. The composition of basalt sample B, for example, reflects twice as much olivine fractionation as that of sample A. One further feature of this diagram deserves mention. The composition of the fractionating olivine, which would change during evolution of the system, has no effect on this diagram. This can be illustrated by representing olivine compositions as: $\text{Mg}_x\text{Fe}_{(2-x)}\text{SiO}_4$ where $0 \leq x \leq 2$; irrespective of the value of X, the ratio of $(\text{Mg}+\text{Fe}^{+2})$ to Si is always 2.

The validity of Pearce ratios is dependent on the choice of the denominator (conserved element) for the ratio. As Nicholls (1988) points out, the validity of the ratios can be evaluated by considering three principal criteria:

(1) The denominator must be unaffected by potential sources of sample variation. Therefore only elements that can reasonably be expected to be excluded from hypothetical differentiation processes should be considered. For example, MgO would be a poor choice of conserved element if the process being investigated was thought to involve partial melting or crystallization of olivine. In this instance K_2O or Sr could be appropriate. However, the effects of subsequent processes must be considered; e.g., strong hydrous alteration or metamorphism may disqualify K_2O and Sr due to the mobility of these elements under such conditions.

(2) A Pearce ratio in which both the numerator and denominator are conserved elements will have a constant value for all samples. For practical purposes, this means that the standard deviation of the ratio for all samples will be less than the analytical uncertainty propagated through the ratio. A scatter plot of Pearce ratios in which both numerator and denominator are conserved will result in all samples plotting in the same location (i.e., a single point), within analytical uncertainty.

(3) Pearce plots in which the regression line passes through the origin are invalid, since this can only occur when the denominator is not conserved (Pearce, 1987; Russell and Nicholls, 1988). Consequently, the intercept must differ from zero given the propagated analytical uncertainty associated with the intercept.

The choice of numerators for Pearce ratios is also dependent on the hypothesized process involved in sample variation. If a series of liquids (represented by samples from a suite of basalt flows) is thought to be related by fractionation of olivine, a plot of Pearce ratios in which the numerators are $(Mg+Fe^{+2})$ versus Si in cation proportion can be used to

test the hypothesis. As in the earlier example, if olivine alone is responsible for compositional differences among samples, the line of best fit of the data will have a slope of 2, since this is the stoichiometric ratio of these elements in olivine. If, however, differentiation involved an additional silicate phase, the slope of the regression line will differ from 2. More complex numerators can be defined to test hypotheses regarding multi-phase fractionation or other processes (cf. Ernst *et al.*, 1988).

5.4 Geochemical Variation in Sill C

The geochemical variation in sill C was examined through analysis of 34 whole rock samples obtained from the approximately 225 metres of exposed stratigraphic thickness. These include four aphanitic samples and one fine grained sample from the chilled margin at the base of the sill, one fine grained sample from the upper chilled margin, and four samples of felsic rocks from an irregular siliceous zone located near the top of the sill. The remainder of the samples represent the main body of the intrusion, and these were subdivided into two suites representing aphyric and glomeroporphyritic material. The glomeroporphyritic samples were serially sectioned and only slabs lacking obvious glomerocrysts used for analysis. This procedure allows comparison of samples from across the sill on the basis of groundmass composition alone.

General Features

An examination of geochemical trends in sill C using plots of concentration versus stratigraphic position (Fig. 5.2) reveals that, with the exception of the felsic samples, the magnitude of the variation in major element concentration is not large. The most variable oxide in the main part of the sill is Al_2O_3 (12-18 wt. %), and the most abundant oxide, SiO_2 , has a total range of only about 5 wt. %. The geochemical consistency confirms petrographic observations of sample homogeneity; the essential mineralogy in all non-felsic samples is plagioclase, augite and hypersthene. Olivine occurs in many samples, but reaches significant proportions only in the central part of the sill.

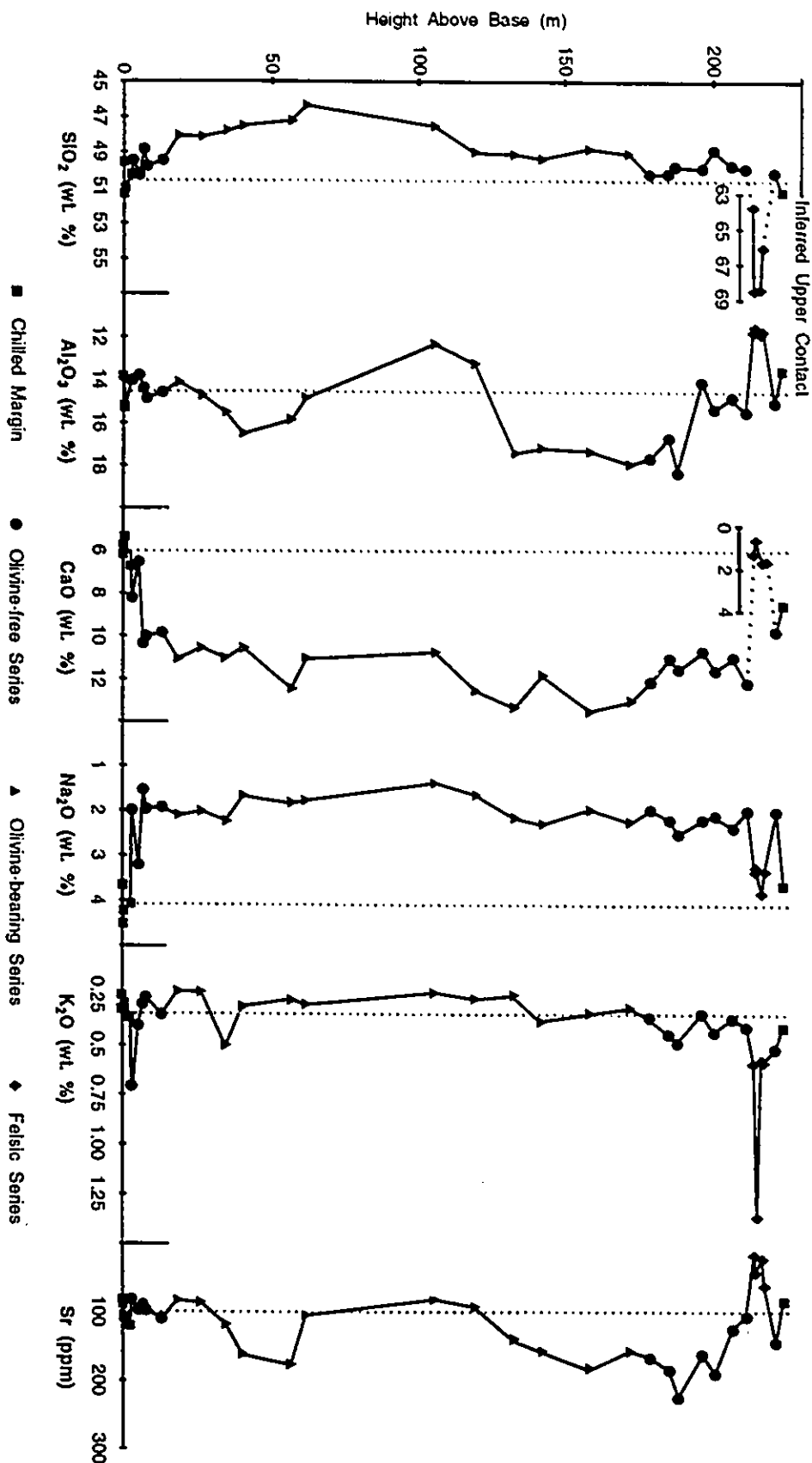


Figure 5.2. Variation in concentration of selected elements across Howse Lake sill C. Vertical broken lines indicate average concentrations in four samples from the lower chilled margin. Average 2σ analytical is less than the size of plot symbols except where indicated.

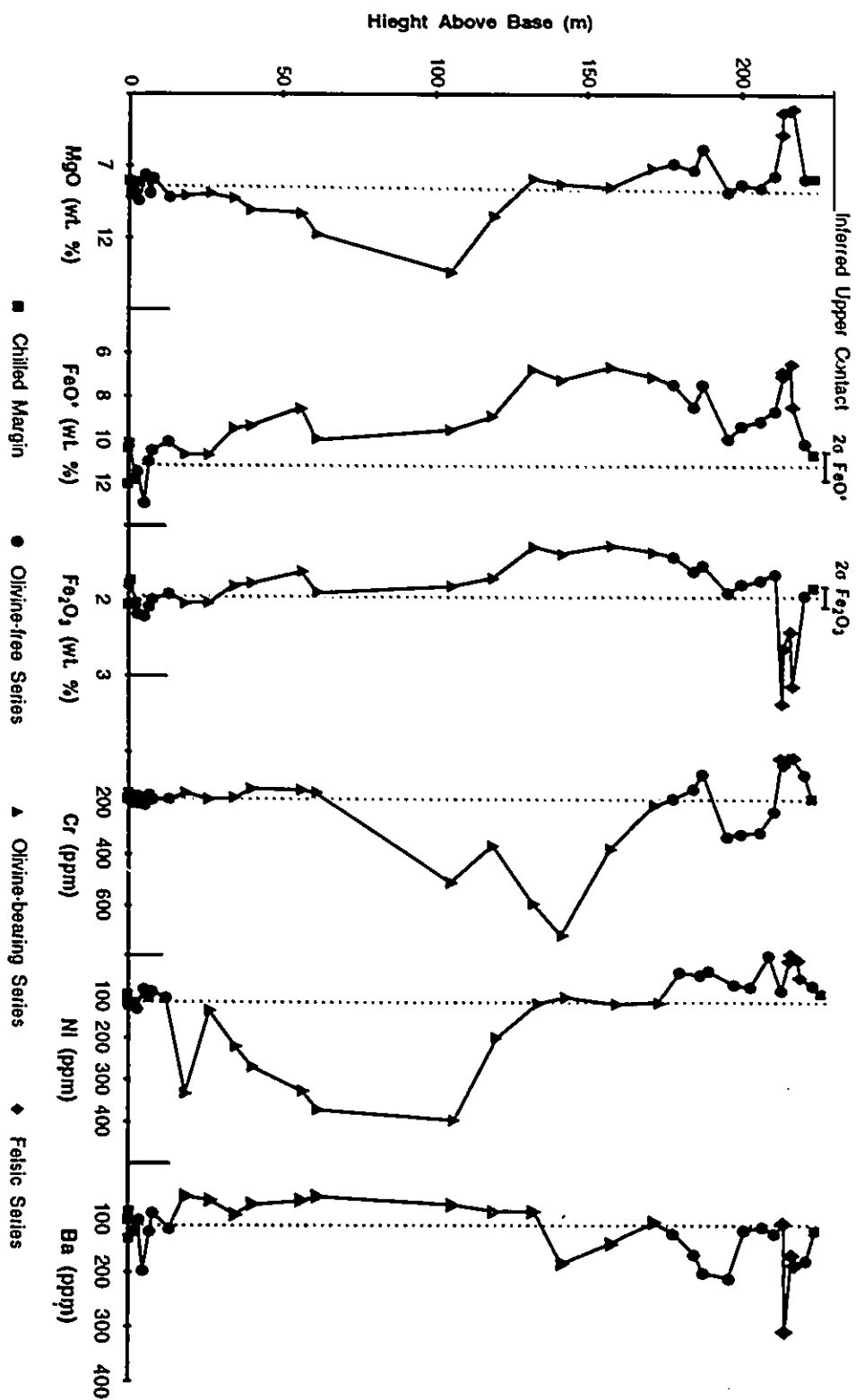


Figure 5.2 (continued). Variation in concentration of selected elements across Howse Lake sill C. Vertical broken lines indicate average concentrations in four samples from the lower chilled margin. Average 2σ analytical is less than the size of plot symbols except where indicated.

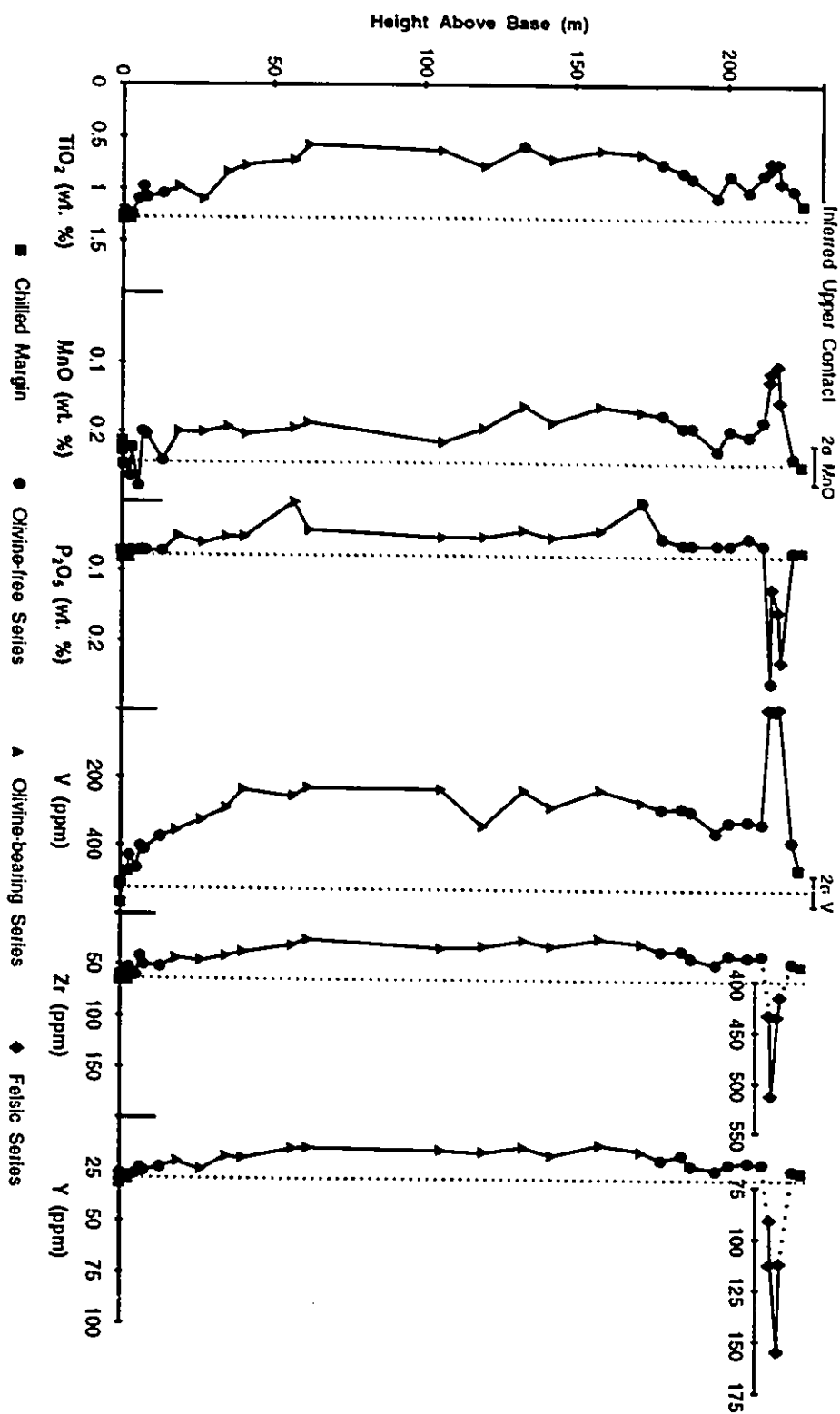


Figure 5.2 (continued). Variation in concentration of selected elements across Howse Lake sill C. Vertical broken lines indicate average concentrations in four samples from the lower chilled margin. Average 2σ analytical is less than the size of plot symbols except where indicated.

Despite the lithological homogeneity, four geochemical subtypes can be recognized. The aphanitic and fine grained samples from the basal contact define one series. Most major element concentrations diverge from marginal values starting at about the sixth sample (5 m) above the contact, and the main body of the sill shows higher CaO and lower SiO_2 , Na_2O , FeO^* (calculated FeO , see Appendix A4), Fe_2O_3 , TiO_2 , P_2O_5 and MnO. Compositional differences are also manifest in trace element concentrations, with higher Sr and lower V, Zr and Y in the main part of the sill. The reverse trend is apparent near the top of the sill where the fine grained sample from the upper chilled margin is compositionally similar to those of the lower margin. The most significant difference between marginal samples and those from the main portion of the intrusion lies in the high Na_2O and low CaO contents of the former. This difference is clearly a reflection of the albitization of plagioclase in marginal samples.

The main body of the sill comprises two compositional types which are most clearly delineated by Ni content. Elevated MgO, FeO^* , and Ni, and low Al_2O_3 and CaO characterize the samples between 60 m and 130 m above the lower contact. This interval coincides with the central zone of elevated olivine content in the intrusion. The high olivine zone also has slightly elevated Fe_2O_3 , and lower SiO_2 , TiO_2 and Sr than the rest of the sill. The upper part of this interval is enriched in Cr, but the enrichment continues in samples immediately above the zone defined by the other elements. Petrographically, the main body of the sill can be subdivided into olivine-bearing and olivine-free units. Of the samples which contain olivine, this phase is a significant modal constituent only in the central part of the intrusion. This is apparent in the geochemical similarity of olivine-bearing samples outside of the high-olivine zone and those which lack olivine.

The fourth and most prominent geochemical type evident in Figure 5.2 is represented by the felsic rocks which occur near the top of the sill. These samples are characterized by elevated SiO_2 , Na_2O , K_2O , P_2O_5 , Zr, and Y, and low Al_2O_3 , CaO, FeO , MgO, TiO_2 , MnO, Cr, Ni and V. The compositions of the felsic samples reflect their mineralogies, which consist largely of quartz, sodic plagioclase and orthoclase. Secondary chlorite and actinolite are the only ferromagnesian phases present except for occasional grains of partially replaced green hornblende. In addition, the amount of accessory apatite and zircon are significantly higher than in the gabbroic samples.

Choice of Conserved Elements for Pearce Ratios

The choice of denominator element for Pearce ratios constructed from sill C data must consider the potential sources of compositional variation discussed previously. As noted earlier, there is considerable textural and compositional evidence which indicates that the Menihek sills underwent *in situ* fractional crystallization during solidification. The low incompatible element concentrations in the main body of the sill (Fig. 5.2) confirm textural observations which indicate that these are cumulate rocks. A consideration of the principal mineralogy of the gabbroic samples suggests that elements with low liquid-crystal partition coefficients for plagioclase, augite, hypersthene and olivine, the probable cumulus phases during solidification of the main part of the sill, might be suitable choices as denominator elements. In consideration of the late felsic differentiates, quartz and alkali feldspar must be included as potential cumulus phases. Given the stoichiometry of these minerals, Ti, P, Mn, Zr and Y were selected from available data as potentially conserved elements. These choices are based on the relatively low partition coefficients of the elements for suspected fractionating minerals (Henderson, 1982), and their comparative immobility during alteration. From microprobe data it is clear that Mn and Ti are not completely incompatible in clinopyroxene, and small amounts of Y also may enter the structure of this phase. No element, however, would be completely incompatible in the cumulate, and the selected constituents are considered the most likely to have been conserved. Constant ratios among these elements would indicate conservation during differentiation and alteration, while variable ratios indicate non-conservation. Non-conservation could reflect inclusion of one or more elements in a fractionating phase, mobility during alteration, or operation of another differentiation mechanism.

Table 5.1 gives the standard deviations and average analytical uncertainties for ratios calculated from the potentially conserved elements using concentration data expressed as cation proportion. If all samples are included in the analysis, only the P/Zr ratio is constant. Since other ratios involving these elements are non-constant, it is unlikely that both elements are conserved for the entire suite. Several additional elements, including K, Cr, Ni, and Sr were tested in an attempt to find conserved elements common to all samples, but none met the requirements of constancy. In contrast, the data indicate that Ti, P, Mn, Zr and Y are conserved if the felsic samples are excluded. For the non-felsic

Table 5.1. Average, standard deviation and average analytical error for ratios constructed from potentially conserved elements in the differentiation of sill C. Standard deviation and analytical error are in bold typeface for constant ratios.

Including Felsic Samples							
Ratio	Ti/Zr	Ti/Y	P/Zr	P/Y	Mn/Zr	Mn/Y	Y/Zr
Average	234	453	17.7	33.7	60.0	115	0.50
Standard Deviation	81.0	152	5.19	8.75	24.1	44.5	0.078
Analytical Error	18.2	64.1	5.21	6.58	15.0	31.9	0.077
Excluding Felsic Samples							
Average	263	505	19.3	31.1	67.6	129	0.52
Standard Deviation	20.0	47.9	2.71	5.76	13.4	23.4	0.046
Analytical Error	20.4	71.4	5.68	7.03	16.9	35.7	0.081

samples, it can be concluded that the variation in the abundance of these elements due to magmatic differentiation and alteration is less than the variation due to analytical imprecision. It can also be concluded that fractional crystallization alone does not adequately explain the composition of the felsic samples unless all of the selected elements were involved in the process. This necessitates fractionation of phases such as apatite, zircon, magnetite and ilmenite.

Simple Numerator Pearce Ratio Diagrams

Figure 5.3 presents Pearce ratio diagrams of sill C data in which the major elements have been plotted against Si, the most abundant element in all samples. The ratios have been calculated using Y and Zr as denominator elements since these (1) have the lowest partition coefficients for the suspected cumulate, (2) are the least mobile during alteration, and (3) have the lowest associated analytical uncertainty relative to their average abundance. Only ratios using Zr as the denominator are illustrated, but Y produces similar plots. Statistical data for ratios involving both Zr and Y for each plot are given in Table 5.2. The four felsic samples are shown in the plots, but since the denominator elements can not be considered conserved, the samples were not included in regressions of the data. The chilled margin samples have also been excluded from the regressions because of potential alteration effects. With a few exceptions, (e.g., Na versus Si), regressions in which the marginal samples are included are essentially identical to those in which they have been excluded. Although the main part of the sill is divided into olivine-bearing and olivine-free

Table 5.2. Slope (m), intercept (b), standard error of the intercept (SE_b), correlation coefficient (r) and t-statistic (t) for Pearce ratio diagrams of major elements versus Si using Y and Zr as denominator elements. Ratios using Zr as the denominator are plotted in Figure 5.3. The critical value of t at the 95% confidence level for 22 degrees of freedom ($\nu=n-2$) is 2.074 (Davis, 1986; Table 2.11), where the number of samples (n) is 24.

Denominator = Zr	m	b	SE_b	r	t
Al-Si	0.49	-2258	595	0.96	16.290
Fe ³⁺ -Si	0.02	129	57.0	0.68	4.350
Fe ²⁺ -Si	0.11	842	370	0.68	4.350
Mg-Si	0.40	-2665	1160	0.75	5.380
Ca-Si	0.36	-2266	254	0.99	28.970
Na-Si	0.09	-206	182	0.89	9.060
K-Si	0.01	-45.8	32.2	0.73	5.050
Ti-Si	0.00	223	14.5	0.52	2.830
P-Si	0.00	14.6	1.9	0.46	2.460
Mn-Si	0.00	18.8	4.6	0.91	10.220
Denominator = Y					
Al-Si	0.48	-4020	1246	0.95	14.090
Fe ³⁺ -Si	0.02	228	110	0.67	4.190
Fe ²⁺ -Si	0.12	1486	713	0.67	4.190
Mg-Si	0.43	-6154	2272	0.77	5.590
Ca-Si	0.36	-4408	506	0.99	27.290
Na-Si	0.09	-325	360	0.87	8.110
K-Si	0.01	-83.0	63.5	0.68	4.380
Ti-Si	0.00	423	32.6	0.42	2.150
P-Si	0.00	27.5	3.7	0.42	2.100
Mn-Si	0.00	27.3	10.2	0.88	8.840

suites in Figure 5.3, trends defined by the two groups are similar and the samples are therefore treated as a single data set for regression purposes.

Figure 5.3 and Table 5.2 indicate that all plots meet the requirements of non-zero intercepts except in the case of Na versus Si using Y as the denominator element. This suggests processes other than fractional crystallization may have influenced the Na content of some samples. Given the albitization of plagioclase in the chilled margin, it is probable that the failure of this plot to meet the requirement of non-zero intercept reflects alteration and mobilization of sodium. In addition to non-zero intercepts, most plots show relatively strong correlations between the ordinate element and Si. These can be tested for statistical significance using a two-tailed student's t test in which the null hypothesis is that the

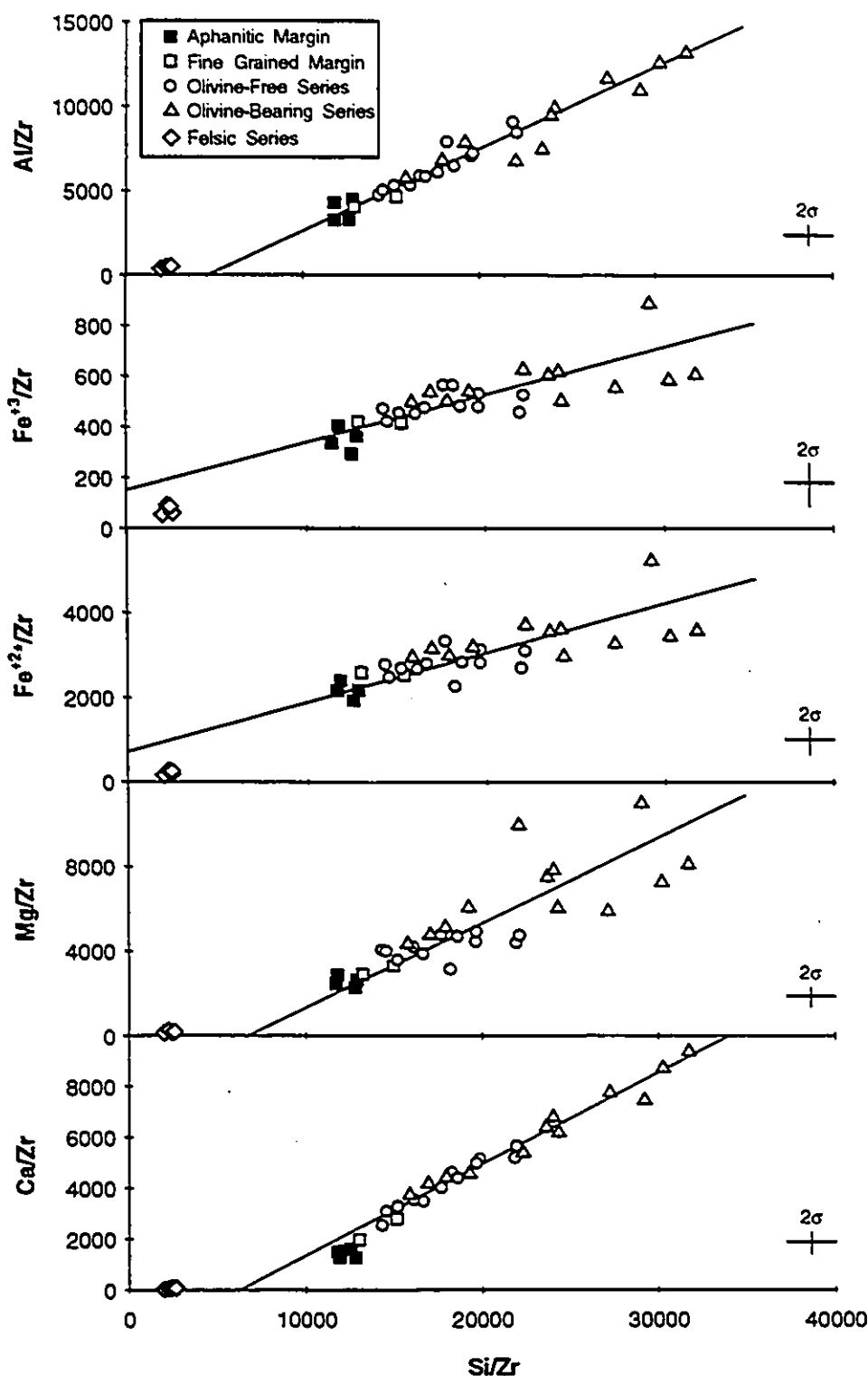


Figure 5.3. Pearce ratio plots of selected major elements versus Si using Zr as the denominator (conserved) element. The reduced major axis regression line for olivine-bearing+olivine-free samples is shown, and regression statistics are provided in Table 5.2. Average 2 σ analytical uncertainties for all samples are indicated by error bars.

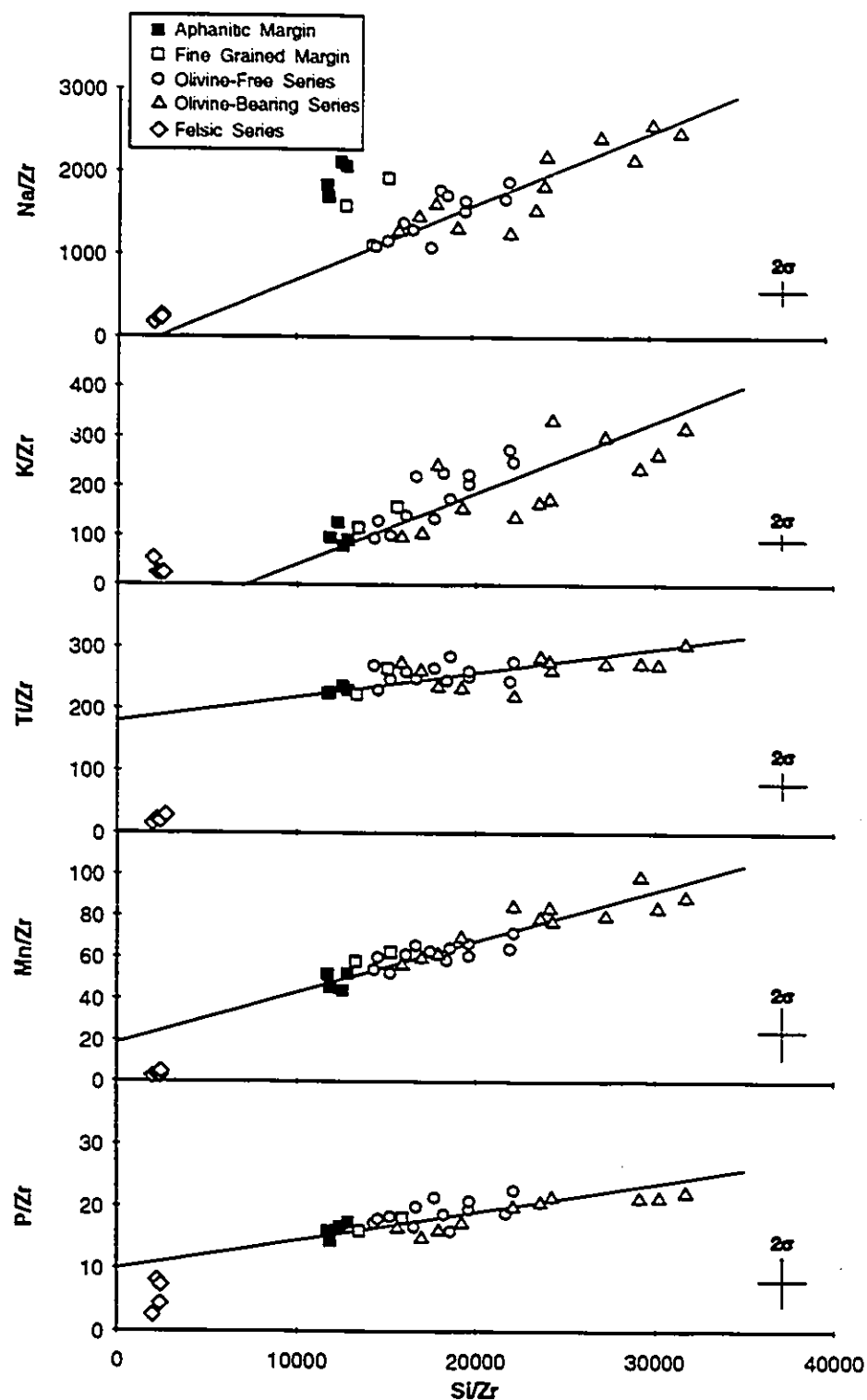


Figure 5.3 (continued). Pearce ratio plots of selected major elements versus Si using Zr as the denominator (conserved) element. The reduced major axis regression line for olivine-bearing+olivine-free samples is shown, and regression statistics are provided in Table 5.2. Average 2σ analytical uncertainties for all samples are indicated by error bars.

correlation coefficient is not significantly different from zero, and there is no relationship between the two variables. Values of the t-statistic are given in Table 5.2, and it can be seen that the null hypothesis can be rejected for all elements at the 95% confidence level.

Positive slopes in the diagrams indicate that the samples are related by gains or losses in both numerator elements, while zero slopes indicate the ordinate element was conserved. Whether a sample has gained or lost material depends on its position in the plot. For example, a sample located in the middle of the array may represent the residual liquid after removal of the fractionate, or it could represent the original liquid plus accumulated fractionate. Without a reference composition these alternatives cannot be resolved.

For most elements, the aphanitic marginal samples plot as a fairly coherent group along the regression line closer to the origin than the other gabbroic samples, while the fine grained samples are generally transitional between the two groups. Since the aphanitic samples represent rapidly crystallized magma, it is reasonable to assume that they, of all available samples, best reflect the composition of the melt prior to onset of differentiation (i.e., they can be used as a reference composition). Given the relative positions in the plots of the chilled margin and the gabbroic samples, the latter can only be interpreted as combinations of the reference composition (starting liquid) and fractionated solids. Consequently, the main portion of the intrusion reflects accumulation of minerals containing Si, Al, Fe^{2+} , Mg, Ca and Na, and to a lesser degree K and Fe^{3+} . The zero slopes in the plots for Ti, Mn and P indicate that these elements occur in constant proportions relative to Zr and Y in all gabbroic samples, and therefore have not taken part in the process. It is noteworthy that in each diagram the samples which plot furthest from the chilled margin samples, and therefore contain the greatest proportion of fractionated solid, are members of the olivine-bearing series

Residual liquids produced by fractional crystallization of some proportion of the original melt after closure of the magmatic system will be impoverished in Si, Al, Fe^{2+} , Mg, Ca, Na, K and Fe^{3+} . Crystallization of such liquids would produce rocks which plot along the trends defined by the cumulate samples, but on the opposite side of the reference composition. The felsic samples do, in fact, plot to the left of the chilled margin but their positions do not coincide with the trends defined by the gabbroic samples. Consequently,

their compositions must reflect either fractionation of additional phases or operation of another process. Because none of the elements in the data set can be considered conserved in all samples, evaluation of these possibilities is difficult.

Complex Numerator Pearce Ratio Diagrams

The principal silicate minerals found in the main body of the sill are plagioclase, augite, hypersthene and olivine and, in light of the above data, it is logical to hypothesize that geochemical variation among samples is due to fractional crystallization and accumulation of one or more of these phases. Such an hypothesis is supported by the textural and compositional characteristics of the principal silicates discussed previously. The relationship among samples in terms of fractional crystallization can be examined using complex numerator Pearce ratios which consider the stoichiometries of postulated cumulus minerals (see Appendix A5 for derivation of numerators). Using the chilled samples as the starting liquid composition, ratio diagrams can be employed to determine the identities and proportions of accumulated minerals. The appropriate complex numerator Pearce ratio diagrams in which Zr is the conserved element are illustrated in Figure 5.4. Similar diagrams are obtained using Y as the conserved element. Regression data for plots using both elements as denominators are provided in Table 5.3.

Figure 5.4A illustrates the relationship between clinopyroxene and plagioclase, the principal Ca-bearing phases. It is clear that the variation in Ca cannot be due to fractionation of either plagioclase (Pl) or clinopyroxene (Cpx) alone, since the regressions of both sample suites exhibit slopes intermediate to those of the two Ca-bearing phases. Regressions of samples with and without olivine are similar, and it can be concluded that both sets of samples have accumulated plagioclase and clinopyroxene in approximately the same ratio. The average slope of the regressions (using Zr and Y as denominators) of the olivine-free samples is 0.68, while that of the olivine-bearing samples is 0.61. These values, which are identical within error, give the average Cpx/Pl ratio in the cumulate of each series.

Figure 5.4B indicates that plagioclase (albite + anorthite) is largely responsible for variations in the Al content of the gabbroic samples, while alkali feldspar (Kfs) was a very minor component of the cumulate. The trends for samples with and without olivine are

Table 5.3. Slope (m), standard error of the slope (SE_m), correlation coefficient (r) and test statistic (t) for complex numerator Pearce ratio plots using Zr and Y as denominator elements. Ratios using Zr as the denominator are plotted in Figure 5.4. The critical value of t is 2.228 at the 95% confidence limit and $\nu=10$ ($n=12$ for both sample sets)

Diagram	Denominator = Zr				Denominator = Y			
	m	SE_m	r	t	m	SE_m	r	t
Figure 5.4A: Cpx vs. Pl								
Olivine-free	0.66	0.10	0.74	4.999	0.69	0.11	0.66	3.961
Olivine-bearing	0.58	0.09	0.78	5.712	0.63	0.09	0.74	5.096
Figure 5.4B: Pl vs. Kfs								
Olivine-free	0.05	0.01	0.85	7.719	0.06	0.02	0.84	7.294
Olivine-bearing	0.04	0.01	0.96	17.340	0.04	0.01	0.95	14.816
Figure 5.4C: Cpx vs. Opx vs. Ol								
Olivine-free	0.68	0.05	0.90	9.698	0.73	0.09	0.81	6.359
Olivine-bearing	1.14	0.13	0.87	8.370	1.14	0.13	0.88	8.668
Figure 5.4D: Opx vs. Ol								
Olivine-free	0.89	0.13	0.79	5.923	0.93	0.13	0.81	6.401
Olivine-bearing	1.86	0.21	0.90	10.108	1.75	0.19	0.91	10.333
Figure 5.4E: Mt vs. Cpx+Opx+Ol								
Olivine-free	0.02	0.01	0.62	3.519	0.03	0.01	0.66	3.922
Olivine-bearing	0.03	0.01	0.67	4.069	0.03	0.01	0.66	3.884
Figure 5.4F: An vs. Ab								
Olivine-free	1.98	0.37	0.77	5.544	1.85	0.34	0.77	5.500
Olivine-bearing	2.13	0.24	0.92	11.647	2.20	0.28	0.90	9.662
Figure 5.4G: Mg vs. Fe²⁺								
Olivine-free	2.65	0.51	0.69	4.352	2.19	0.52	0.46	2.176
Olivine-bearing	3.31	0.47	0.87	8.498	3.66	0.50	0.88	8.850
Figure 5.4H: Cpx+Opx+Ol vs. Pl+Kfs								
Olivine-free	1.03	0.18	0.63	3.638	0.96	0.11	0.82	6.573
Olivine-bearing	0.86	0.12	0.73	4.915	0.89	0.12	0.74	4.985

again similar, with average Kfs/Pl ratios in the cumulate of 0.06 and 0.04 respectively. If alkali feldspar independent of plagioclase formed part of the cumulate, it contributed very little to the total solid. Assuming the average Or content of cumulus plagioclase was 1 mol. %, which is a typical value for groundmass plagioclase as determined by microprobe analysis, the remaining variation on K can be accommodated by approximately 2% alkali feldspar in cumulate. Since alkali feldspar would be an unlikely liquidus phase during crystallization of the other silicates, the best interpretation of these data is that the variation in K is a function of the Or content of cumulus plagioclase rather than the accumulation of alkali feldspar.

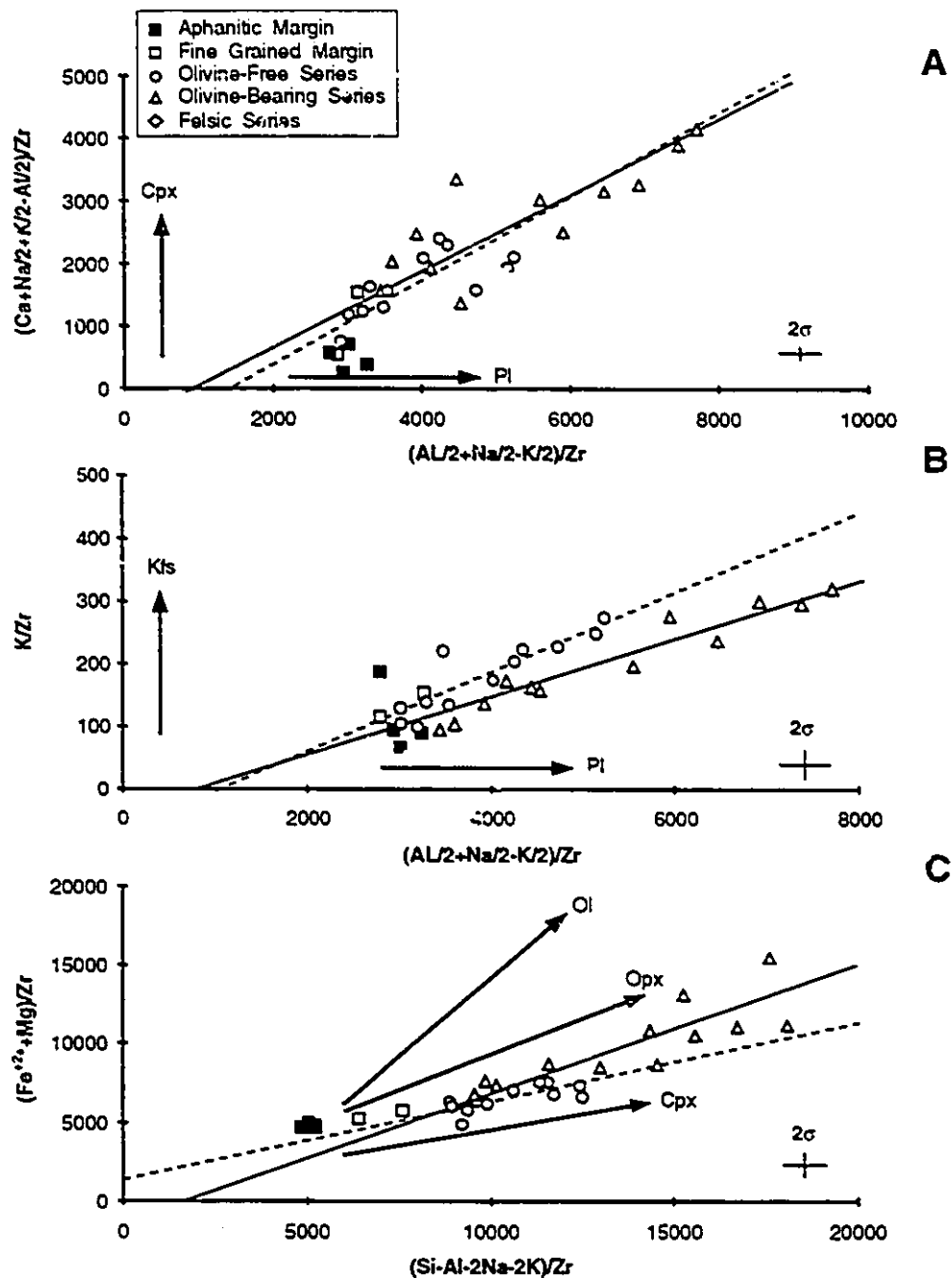


Figure 5.4. Complex numerator Pearce ratio diagrams illustrating the effects of different minerals on sample composition during fractional crystallization in sill C. (A) clinopyroxene (Cpx) versus plagioclase (Pl), (B) alkali feldspar (Kfs) versus plagioclase, and (C) $Fe^{2+}+Mg$ versus total Si less that in feldspar. Reduced major axis regression lines for olivine-bearing (solid) and olivine-free (dashed) samples are shown, and regression statistics are provided in Table 5.3. Average 2σ analytical errors are indicated by error bars.

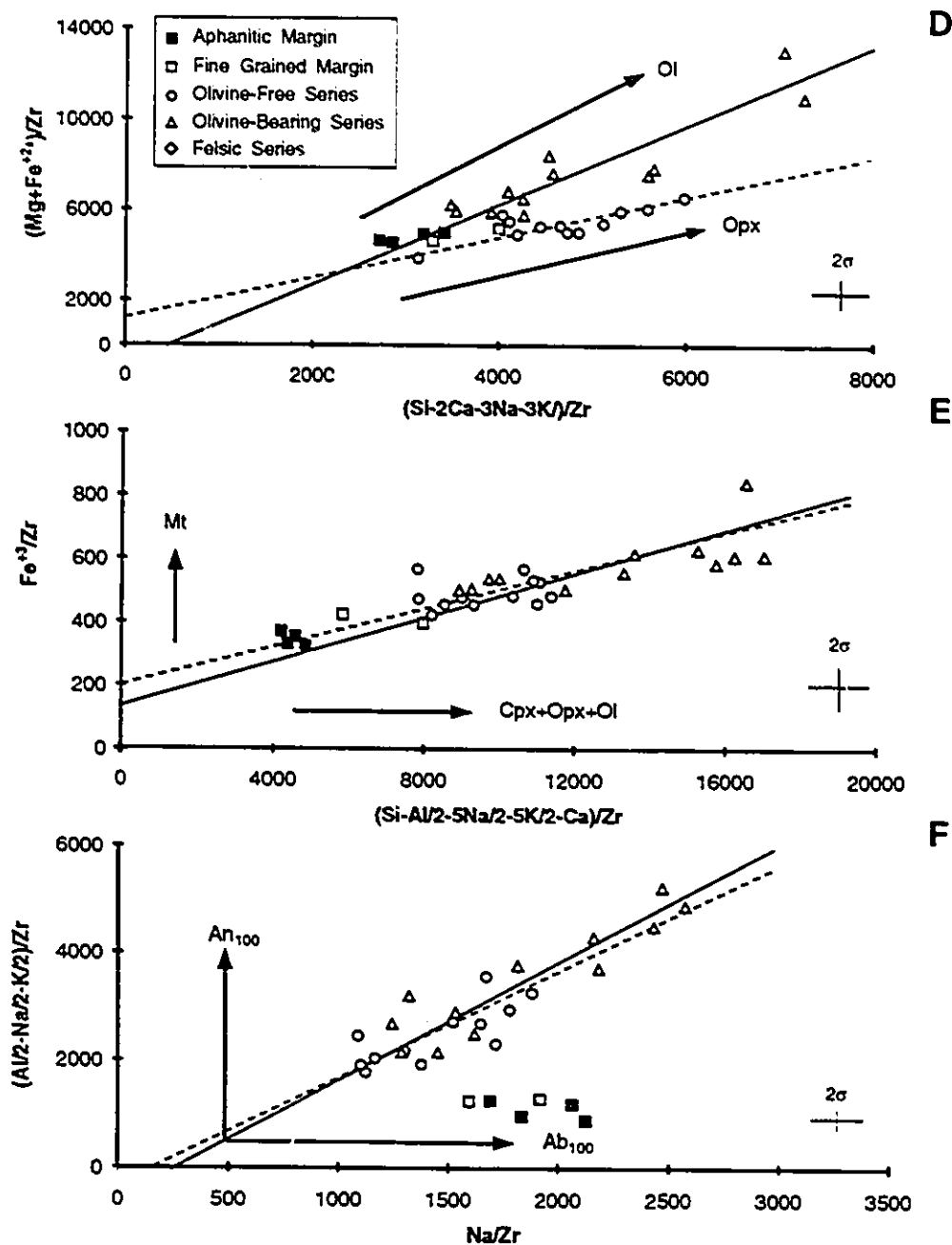


Figure 5.4 (continued). Complex numerator Pearce ratio diagrams illustrating the mineral relationships during fractional crystallization in sill C. (D) $\text{Fe}^{2+}+\text{Mg}$ versus total Si less that in feldspar+clinoproxene, (E) magnetite (Mt) versus Si in clinopyroxene+orthopyroxene (Opx)+olivine (Ol), and (F) anorthite (An) versus albite (Ab) in plagioclase. Reduced major axis regression lines for olivine-bearing (solid) and olivine-free (dashed) samples are shown, and regression statistics are provided in Table 5.3. Average 2σ analytical errors are indicated by error bars.

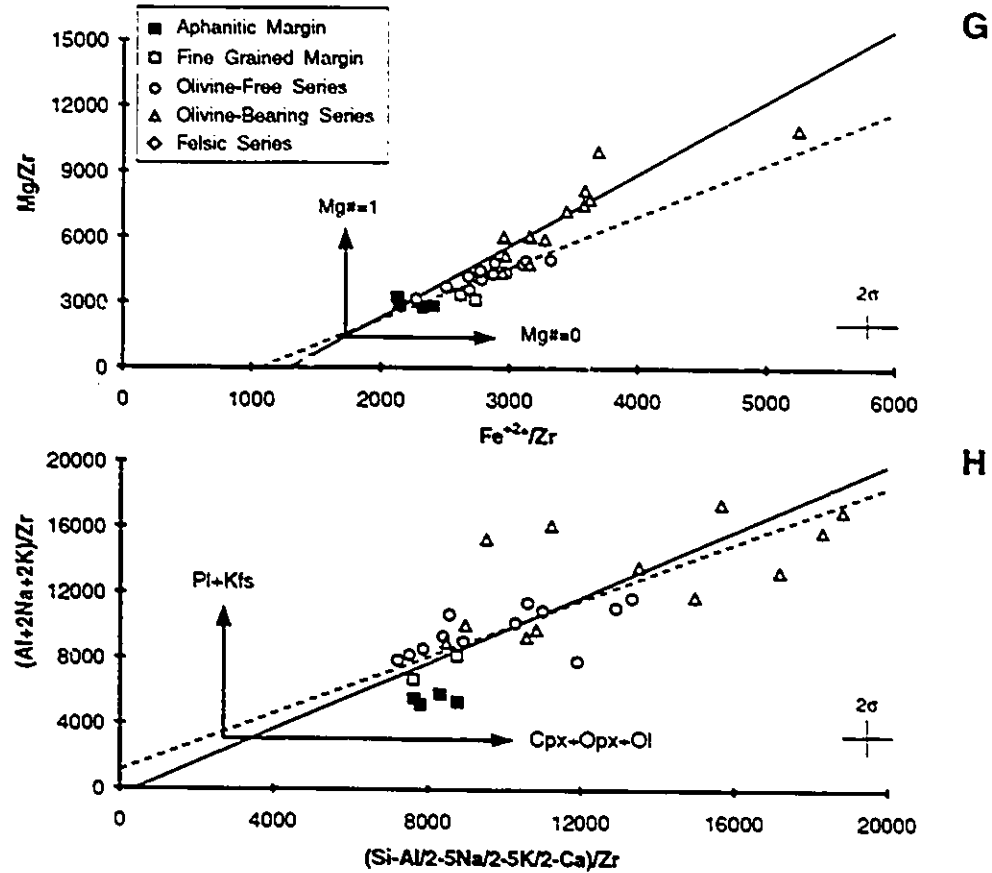


Figure 5.4 (continued). Complex numerator Pearce ratio diagrams illustrating the effects of different minerals on sample composition during fractional crystallization in sill C. (G) Mg versus Fe^{2+} in ferromagnesian silicates and (H) plagioclase+alkali feldspar versus clinopyroxene+orthopyroxene+olivine. Reduced major axis regression lines for olivine-bearing (solid) and olivine-free (dashed) samples are shown, and regression statistics are provided in Table 5.3. Average 2σ analytical uncertainties are indicated by error bars.

If all of the variation in potassium is assumed to arise from substitution into plagioclase, the Kfs/Pl ratios could be accommodated by the fractionation of feldspar having a composition of $(\text{An}+\text{Ab})_{96}\text{Or}_4$ for the olivine-free series, and $(\text{An}+\text{Ab})_{94}\text{Or}_6$ for the olivine-bearing series. The Or content of the cumulus feldspar derived from ratio analysis is thus slightly higher than the average Or content of sill C plagioclase ($\text{Or}_{1.1}$) as determined by microprobe analysis (Fig. 5.5). The 2σ analytical uncertainties associated with the two methods are $\pm\text{Or}_3$ for ratio analysis and $\pm\text{Or}_{0.6}$ for microprobe analysis.

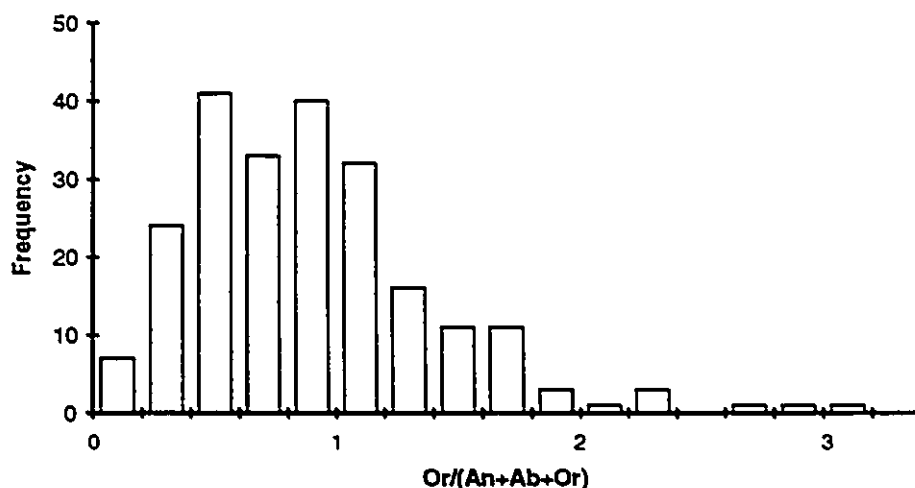


Figure 5.5. Frequency distribution of microprobe analyses of orthoclase (Or) content in plagioclase from sill C. Average Or content is $1.13 \pm 0.56\%$.

Consequently, the small compositional differences may reflect analytical imprecision in the case of the olivine-bearing samples, but the discrepancy is larger than the 2σ confidence limits on the olivine-free data. The magnitude of the discrepancy, however, is such that it may have arisen from minor substitution of K and Al in other phases, particularly clinopyroxene, or mobility of K during alteration. Irrespective of the nature of the K-bearing phase (plagioclase versus alkali feldspar), ratio analysis indicates that the total amount of alkali feldspar in the cumulate was very minor, and this interpretation is in agreement with both microprobe data and petrographic observations.

The accumulation of plagioclase and augite explains the variation observed in Ca, Al, Na and K, and it is possible that augite fractionation may also be responsible for the observed variations in Mg and Fe^{2+} . This possibility is examined in Figure 5.4C, where the hypothetical trends for olivine, orthopyroxene and clinopyroxene fractionation are compared to observed trends. The average slope of the regression lines for the olivine-bearing series is 1.14, while that for the olivine-free samples is 0.71. Because accumulation of clinopyroxene would produce a slope of 0.5, this phase alone cannot be responsible for the total variation in Fe^{2+} and Mg in either series. Figure 5.4C indicates that both clinopyroxene and orthopyroxene were present in the cumulate of the olivine-free series, and that the Opx/Cpx ratio of this cumulate was 0.71.

The relative importance of the three ferromagnesian phases in the compositional variation among olivine-bearing samples cannot be determined from Figure 5.4C. In Figure 5.4D, however, the effect of clinopyroxene fractionation has been removed, and it can be seen that both olivine and orthopyroxene accumulated in the olivine-bearing series. The average slope of 1.81 for these samples corresponds to an Ol/Opx ratio of approximately 4. The slope of the regression line for the olivine-free samples in this diagram is within error of the expected value of 1.0 (i.e., no olivine in the cumulate).

Fe^{3+} is the only other major element which is not constant among the gabbroic samples. Petrographically, the Fe-Ti oxides found in the sill usually appear to be intergrowths of magnetite and ilmenite. The fact that Ti and V, both of which would be expected to partition into an oxide phase, are conserved suggests that oxides did not form part of the cumulate. An alternative explanation for Fe^{3+} variability is substitution into the structure of the ferromagnesian silicates. This possibility is examined in Figure 5.4E where Fe^{3+} is plotted against the silica content of clinopyroxene+orthopyroxene+olivine. The average slope of the regressions is 0.03, with the trends of olivine bearing and olivine-free samples identical within error. This represents an average combined concentration in the mafic silicates of 2.5 ± 1.2 wt. % Fe_2O_3 . The Fe_2O_3 content of augite determined by microprobe analysis ranges from 0.25 to 5.6 wt. %, and averages 2.4 wt. % (Fig. 5.6). The limited amount of orthopyroxene data reveals slightly higher concentrations in this phase, with an average of 3.8 wt. %. No data are available for olivine, but it is probable that the total variation in Fe^{3+} can be attributed to substitution into the three silicates rather than accumulation of oxides.

The relative proportions of plagioclase, clinopyroxene, orthopyroxene and olivine in the cumulate for each series can be determined from the ratios derived from Figures 5.4A-D (Table 5.4). The geochemical variation in the olivine-free samples is the result of the addition to the initial melt of a solid having the average composition of (in mol. %) 48% plagioclase (including 2.7% Or), 30% clinopyroxene, and 22% orthopyroxene. The average composition of the cumulate in the olivine-bearing samples was 51% plagioclase (including 1.8% Or), 30% clinopyroxene, 15% olivine and 4% orthopyroxene. According to this analysis, the presence of olivine has resulted in a reduction in the amount of cumulus orthopyroxene, but has had little effect on clinopyroxene or plagioclase. The

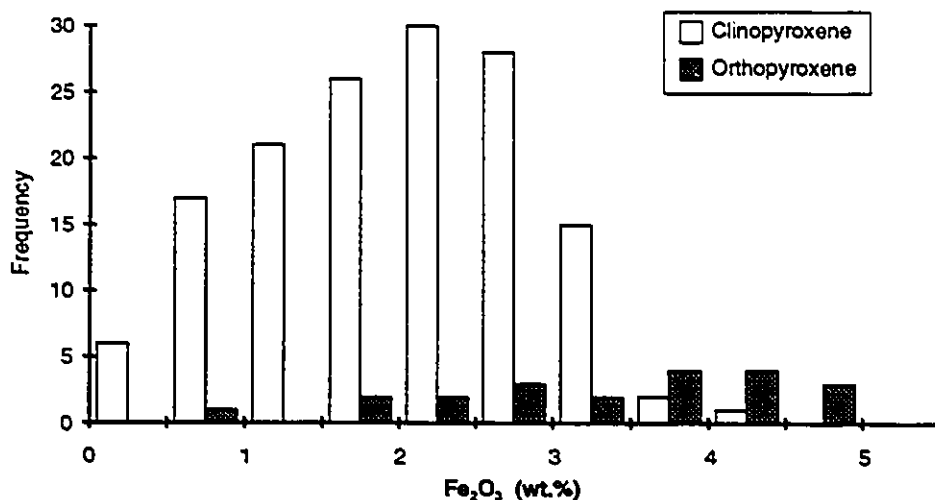


Figure 5.6. Frequency distribution of the Fe_2O_3 content (wt. %) of augite and hypersthene in sill C calculated from microprobe data. Average Fe_2O_3 contents of clinopyroxene and orthopyroxene are 2.4 wt. % and 3.8 wt. % respectively.

substitution of olivine for orthopyroxene in the cumulate suggests *in situ* crystallization of the latter. In contrast, petrographic observations indicate that the olivine has been accommodated by reductions in the modal abundance all of the other silicates, although the abundance of orthopyroxene has been reduced the most. The large uncertainties associated with calculated abundances may explain these differences.

The stoichiometric compositions of the cumulus phases can also be estimated from Pearce diagrams. The average anorthite content of cumulus plagioclase is given by the slope of the trends in Figure 5.4F. The average An/Ab ratio of plagioclase in the olivine-free cumulate was 1.91, and that of the olivine-bearing cumulate was 2.17. Assuming the total variation in K is due to the presence of the orthoclase molecule in plagioclase, the average feldspar compositions of the two suites are $\text{An}_{62}\text{Ab}_{32}\text{Or}_6$ and $\text{An}_{66}\text{Ab}_{30}\text{Or}_4$, respectively. These values are identical given the associated errors (Table 5.4). The plagioclase compositions determined by ratio analysis should reflect the average composition of cumulus grains. While the initial compositions in the cores of cumulus crystals were probably more calcic (i.e., $\sim\text{An}_{80}$), microprobe data clearly indicate strong zonation even within the centres of crystals, as well as a large range in core compositions. This suggests that the average composition of cumulus grains prior to crystallization of

Table 5.4. Mineral proportions, compositions and propagated analytical uncertainty for the cumulate of each sample series calculated from the data of Table 5.3. ¹Total Feldspar was calculated from mineral ratios while ²Total Feldspar was derived from Figure 5.4H.

Olivine-Free Series					
Mineral Ratio		2σ	Mineral Proportion (mole %)		2σ
Pl:Pl	1.00	0	Pl	45.2	5.4
Kfs:Pl	0.06	0.020	Kfs	2.7	1.5
Cpx:Pl	0.68	0.100	Cpx	30.5	7.4
Opx:Cpx	0.71	0.068	Opx	21.6	5.2
Opx:Pl	0.48	0.059	¹ Total Feldspar	47.9	5.6
Pl+Kfs:Cpx+Opx	0.99	0.144	² Total Feldspar	49.8	7.2
An/Ab	1.91	0.354	Mineral Compositions		
Mg/Fe ²⁺	2.42	0.514	An/(An+Ab)	0.66	0.12
			Mg/(Mg+Fe ²⁺)	0.71	0.15

Olivine-Bearing Series					
Mineral Ratio		2σ	Mineral Proportion (mole %)		2σ
Pl:Pl	1.00	0	Pl	49.2	4.7
Kfs:Pl	0.04	0.009	Kfs	1.8	0.9
Cpx:Pl	0.61	0.089	Cpx	30.0	6.3
Ol:Cpx	0.51	0.058	Ol	15.3	3.1
Ol:Pl	0.31	0.055	Opx	3.7	0.7
Opx:Ol	0.24	0.027	¹ Total Feldspar	51.1	4.8
Opx:Pl	0.07	0.003	² Total Feldspar	46.6	6.5
Pl+Kfs:Cpx+Opx+Ol	0.87	0.121	Mineral Compositions		
An/Ab	2.22	0.258	An/(An+Ab)	0.68	0.08
Mg/Fe ²⁺	3.49	0.483	Mg/(Mg+Fe ²⁺)	0.78	0.11

intergranular melt would have been more sodic than the high temperature compositions found in crystal cores. The average calculated cumulus composition ($An_{64}Ab_{31}Or_5$) is thus a reasonable estimate of the average composition of cumulus plagioclase. Considering analytical uncertainty and the potential for Na-metasomatism, the calculated value is in good agreement with the average compositions derived from microprobe data ($An_{60}Ab_{39}Or_1$) and analysis of mineral separates ($An_{65}Ab_{33}Or_2$). This attests to the relatively minor effects on samples from the main body of sill C of the albitization which has occurred at sill margins. The latter effect is readily apparent in Figure 5.4F, where marginal samples have been shifted along the horizontal axis in the direction defined by addition of albite. This shift indicates that these samples, which would have fallen on the

main trend of plagioclase composition at the time of crystallization, have undergone secondary albitization.

The average Mg# of the mafic silicates can be determined in a similar manner (Fig. 5.4G). The average Mg/Fe²⁺ ratio for the olivine-free suite is 2.42, and that of the olivine-bearing samples is 3.49. These ratios correspond to Mg#s of 0.71 and 0.78 respectively, but due to the large associated errors, it is uncertain whether this represents a real difference in the compositions of the ferromagnesian silicates in the two series. The difference could, however, be interpreted to reflect pre-emplacement crystallization of olivine, since the Mg# of this phase would be higher than that of the pyroxenes which crystallized *in situ*. This derives from the fact that pre-emplacement crystallization of olivine would tend to lower the Mg# of the residual melts which were subsequently extracted from the source reservoir to form the sills. The Mg#s of *in situ* pyroxene would thus be lower than that of the previously formed olivine.

The calculated average Mg#s of ferromagnesian minerals in the two sample groups fall within the range of values obtained by microprobe analysis of augite and hypersthene in the sill. The average Mg# of augite as determined by microprobe is 0.72, while that of hypersthene is 0.64. If the compositional difference between these phases is real, then this provides a second possible explanation for the higher average Mg# calculated for the olivine-bearing series. In this case, the average Mg# of the olivine-free series would be lowered by the addition of relatively Mg-poor orthopyroxene. Unfortunately, the paucity of microprobe analyses of orthopyroxene, and the lack of olivine analyses, precludes a more detailed examination of these differences.

The presence of elevated olivine content in the central part of the sill was noted in the description of lithological and geochemical variations in the intrusion. Consequently samples from this zone should depart from trends defined by the remainder of the olivine-bearing samples. This possibility is evaluated in Figure 5.4H, which presents a plot of plagioclase+alkali feldspar versus clinopyroxene+orthopyroxene+olivine. The trends of the two series confirm the average mineral proportions of the accumulated solid calculated from previous diagrams. The slopes of 0.99 and 0.87 translate into total feldspar contents of 47% and 50% for the olivine-free and olivine-bearing series, respectively. These values are equivalent within error to those calculated from ratios obtained from Figure 5.4A-D.

Figure 5.4H also reveals that the three samples from the olivine-rich zone plot well above the average trend of the olivine-bearing samples, while the remaining olivine-bearing samples define a trend similar to that of the olivine-free series. The cumulate of the high olivine samples was therefore enriched in mafic minerals. An examination of Figures 5.4A and 5.4D indicates that, in comparison to the rest of the olivine-bearing suite, it is the amount of olivine in these samples that is elevated rather than the pyroxene content. This is consistent with petrographic observations regarding the elevated modal olivine content of the central part of the sill.

Variation in Cumulate Proportions

An approximate measure of the mass proportion of cumulate in each sample can be calculated from their Zr and Y contents relative to the concentrations of these elements in the initial melt (Fig. 5.7). This requires three assumptions: (1) the chilled margin is a valid representation of the initial melt, (2) the trapped intergranular liquid in each sample was identical (i.e., no melt evolution occurred during solidification), and (3) the bulk partition coefficients of Zr and Y were zero. None of these assumptions are strictly valid. The albitization of plagioclase in marginal samples clearly reflects some degree of compositional change, and this may invalidate assumption (1). Although melt compositions may have changed little until late in the crystallization history of the sill, the felsic horizon indicates that differentiation did occur. This invalidates assumption (2) for at least some samples. Assumption (3) is violated by the fact that neither Y nor Zr will be completely incompatible in the cumulus phases.

The failure of assumptions (2) and (3) would lead to progressive underestimation of the cumulate proportion in samples formed at sequential stages of solidification. For example, fractionation of 50% of the initial liquid mass in a closed system will double the concentration of a perfectly incompatible element in the residual liquid. A rock then formed from 50% fractionated solid and 50% of this residual liquid will have the same concentration of the incompatible element as the starting liquid, and will therefore appear to have no cumulate component. Conversely, if the system is completely open such that the composition of the intergranular liquid is the same in all samples, any rock composed of 50% fractionated solid and 50% intergranular liquid will have an incompatible element

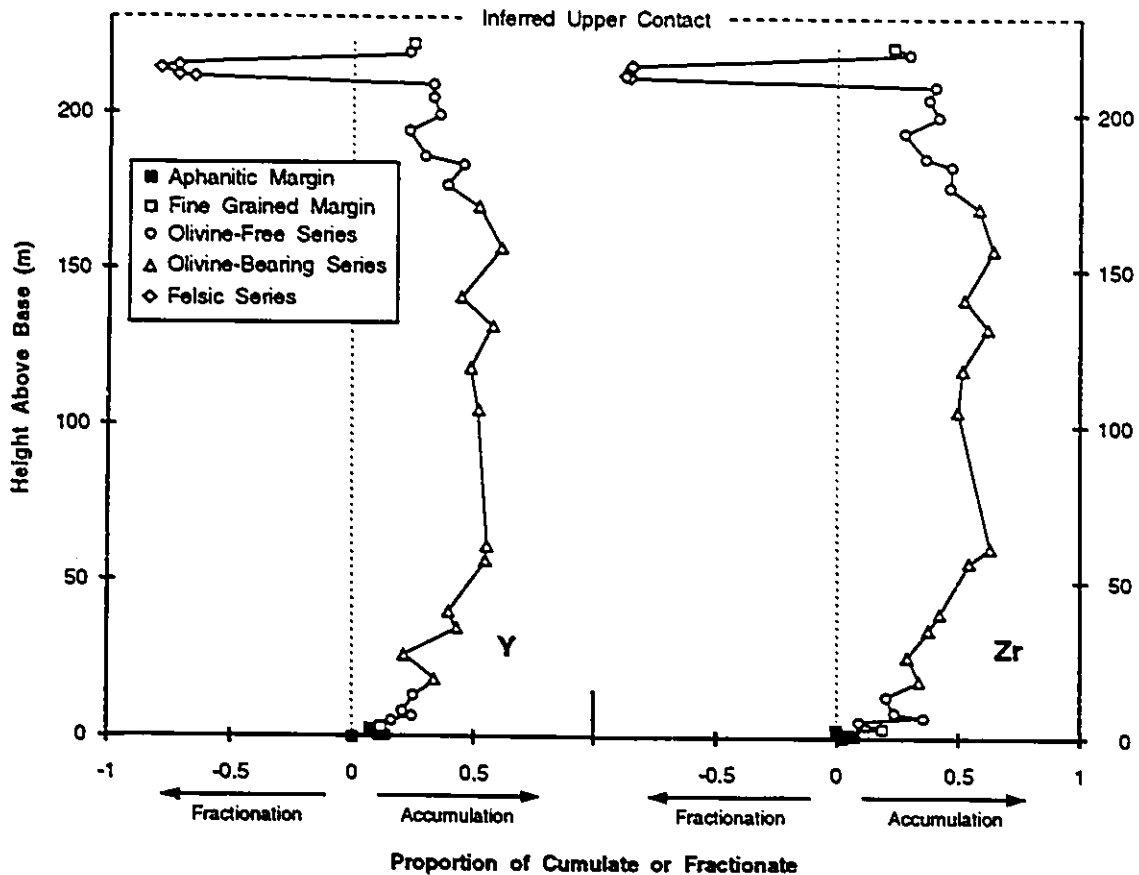


Figure 5.7. Proportions of cumulate added to, or fractionate removed from, the liquids which developed into the principal geochemical series of sill C. Proportions are derived from the sample Y and Zr concentrations relative to the concentrations of these elements in the most primitive (i.e., highest Y-Zr content) chilled margin sample. Dashed lines indicate no gain or loss.

concentration equal to half that of the liquid. Slight compatibility of Y and Zr would have a similar effect, since their inclusion in the cumulus phases would be interpreted as indicating a greater amount of intercumulus melt.

The trend of calculated cumulate proportions in Figure 5.7 is the opposite of that to be expected from failure of assumptions (2) and (3), unless crystallization proceeded from the interior of the sill outward. It is, however, the trend to be expected from crystallization which proceeded from the conduit walls inward, with only minor variations in the compositions of residual melts. This indicates that, although the calculation cannot be considered quantitative, the relative cumulate proportions portrayed in Figure 5.7 are valid.

It should be noted that the maximum cumulate fractions (60+%) occur in samples from the high olivine zone in the central part of the sill.

Nature of the Solidification Process

Consideration of the thermal profiles of plutonic bodies led Jackson (1961), and numerous others since, to the conclusion that boundary layer crystallization is the dominant solidification process during crystallization of intrusive magmas. McBirney and Noyes (1979) reached a similar conclusion based on the analysis of transport mechanisms and nucleation/growth kinetics. In some intrusions, particularly large ones, fractionation has apparently involved crystal transport (e.g., the Skaergaard Intrusion; Wager and Deer, 1939; the Duke Island Complex; Irvine, 1980). In others, there is no evidence of crystal transport and crystallization was an *in situ* process at the margins of the body (e.g., the Palisades Sill; Walker, 1940). *In situ* fractional crystallization was advocated by McCall and Peers (1970) as the differentiation mechanism in the Binneringie Dyke of Western Australia. These authors envisaged fractional crystallization occurring in wall zones with diffusive exchange of material between pore spaces and unfractionated liquid outside of the boundary layer, a process they termed congelation crystallization. Later studies have shown that thermal and compositional convection will act in concert with diffusive exchange and promote the efficiency of the process (Tait *et al.*, 1984; Sparks *et al.*, 1985; Morse, 1986).

The analysis of compositional variation in sill C indicates that the bulk of the intrusion represents some proportion of fractionally crystallized solid and trapped intergranular liquid. The cumulate of most samples in the sill consists of plagioclase, augite and hypersthene, with olivine added in the central part of the sill. With the exception of the latter phase, there is no evidence of sorting or segregation of fractionally crystallized minerals. This indicates that the cumulate was not transported to the site of accumulation. Consequently, it is probable that accumulation occurred at the site of crystallization, with exchange of residual pore liquids and unfractionated melt across the solidification interface. This is consistent with the increasing proportion of cumulate material moving away from sill margins since the decline in the rate of heat loss moving away from the contact would increase the length of time available for fractional

crystallization and chemical exchange between intergranular melt and the main body of magma. This interpretation is also consistent with the crystallization history of the sill as inferred from mineral chemistry data (see Chapter 4).

Origin of Olivine

The distribution of olivine in sill C is inconsistent with *in situ* crystallization alone. If the composition of the main body of magma remained relatively constant, *in situ* crystallization of olivine would result in a uniform distribution of this phase throughout the sill. If the composition of the main body of magma evolved through fractional crystallization, residual melts would be less likely to precipitate olivine. This would result in a distribution in which olivine was found in marginal zones, but declined in concentration towards the centre of the intrusion. With respect to the latter possibility, most data indicate that little melt evolution occurred until late in the solidification history of sill C. It is therefore probable that the concentration of olivine in the central part of the sill reflects operation of another differentiation mechanism.

Figure 5.8 presents ratio diagrams, using Ni as the denominator element, in which the ordinate numerators represent $(\text{Mg} + \text{Fe}^{2+})$ in orthopyroxene and olivine, and the abscissa numerators are Zr. The diagram has no quantitative significance since separation of orthopyroxene and olivine in the numerator cannot be achieved with the available data, and since Ni concentration will vary with olivine and orthopyroxene content and is therefore not conserved. The plot is useful, however, in that the relationship between the two sample suites in terms of olivine content is readily apparent. Because Ni partitions strongly into olivine, accumulation of this phase will produce low Pearce ratios in samples when Ni is used as the denominator. Conversely, fractionation of olivine will greatly reduce the Ni content of residual melts, and rocks formed from such liquids will have high ratios. The effect of orthopyroxene on the magnitude of the ratios will be much less than that of olivine because the solid-liquid partition coefficient of Ni in orthopyroxene is three to ten times less than that in olivine (Irving, 1978; Henderson, 1982).

The felsic samples, which have high ratios due to their low Ni contents, are not illustrated in Figure 5.8 in order to keep the plot at a scale where the relationship between the remaining samples can be determined. The diagram is subdivided into three sections by

the average Zr/Ni ratio of the aphanitic marginal samples and the associated confidence interval calculated from analytical uncertainties. Within the central section of the diagram lie samples whose Zr/Ni ratio cannot be distinguished from the average ratio of the aphanitic samples. To the left lie samples with high Ni contents, while those to the right have low Ni contents. The relationships among samples in the diagrams can be interpreted in two ways. If the chilled margin represents the composition of the initial or parental liquid prior to any solidification (Fig. 5.8A), samples to the left represent rocks formed by addition of olivine (and therefore Ni) to the initial liquid, while those to the right represent rocks which have formed from the initial liquid after removal of olivine (and Ni). This is a reasonable interpretation given the relative positions of the petrographically defined sample suites in Figure 5.8. The occurrence of three olivine-free samples to the left of the initial liquid may reflect the presence of olivine not detected in thin section, or it could reflect elevated orthopyroxene contents which would also tend to raise Ni concentrations. Since the chilled margin samples (representing the parental liquid) occupy the central portion of the diagram, this interpretation is valid only if the system was at least partially closed such that olivine fractionation after emplacement led to Ni-poor residual melt. It also requires a mechanism to concentrate olivine in the central part of the sill as solidification progressed.

The variation in the $(\text{Mg} + \text{Fe}^{2+})_{\text{Ol-Opx}}/\text{Ni}$ ratio with stratigraphic height is illustrated in Figure 5.9. The profile reveals the relative distributions of the olivine-bearing and olivine-free suites. If the aphanitic samples from the lower contact are representative of the initial liquid (Fig. 5.9A), it can be seen that the olivine-bearing series, representing accumulation of olivine, occupies the central part of the sill and is bounded by the olivine-free series. This indicates that the accumulation of olivine was not simply a density driven process. It is also apparent that the degree of olivine fractionation from the parental melt, as represented by the olivine-free samples, increases inward from the contacts. The olivine-free suite consequently exhibits progressively lower Ni concentrations and higher $(\text{Mg} + \text{Fe}^{2+})_{\text{Ol-Opx}}/\text{Ni}$ ratios. If the supposition of boundary layer wall crystallization is correct, this trend is to be expected. The apparent asymmetry in the degree of olivine fractionation may result from gravity acting in aid of the accumulation mechanism at the

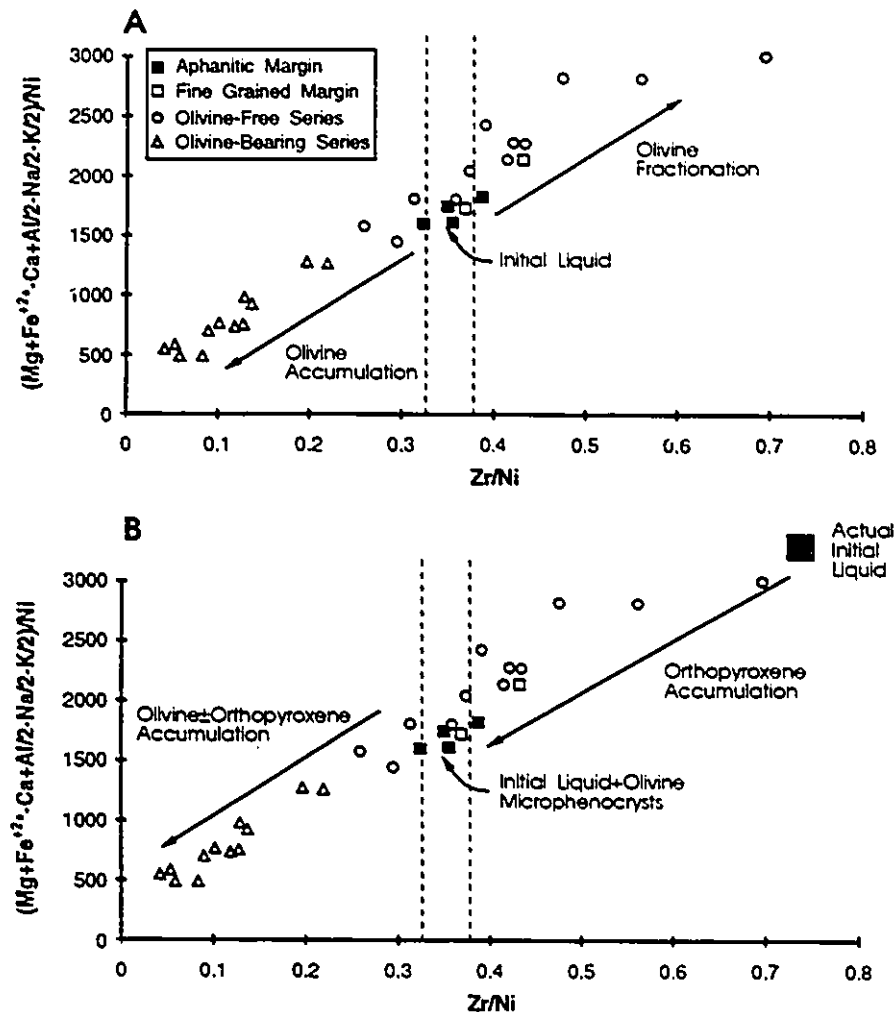


Figure 5.8. Pearce ratio plot of olivine+orthopyroxene versus Zr using Ni as the denominator element. (A) Formation of olivine bearing samples through accumulation of olivine fractionated from the olivine-free series. (B) Formation of olivine-bearing samples through accumulation of pre-emplacement olivine. See text for explanation. Dashed lines indicate the 2σ confidence limits associated with the average Zr/Ni ratio of the aphanitic chilled margin samples.

top of the sill but opposing it at the base. It could also reflect differences in the thermal gradients at the upper and lower contacts.

The alternative interpretation of Figure 5.8 devolves from the possibility that the composition of the chilled margin is not representative of the parental melt, at least in terms of olivine and Ni (Fig. 5.8B). This interpretation requires pre-emplacement crystallization of olivine, intrusion of the olivine-phyric melt, and concentration of olivine

in the centre of the sill during emplacement. This section has been concerned only with variations in groundmass composition and hence the presence of plagioclase glomerocrysts in the sill has been excluded from the discussion. The distribution of glomerocrysts in the sill is described in detail in Chapter 6, but it should be noted here that, as with olivine, they are concentrated in the central part of the sill. The concentrating mechanism is thought to be flow differentiation, and this process would act on any solid particles within the melt during emplacement. Rapid cooling and crystallization near the contacts, however, would trap some olivine crystals and plagioclase glomerocrysts before these solids were transported away from the margins.

Prior to analysis samples were sectioned to provide material free from plagioclase glomerocrysts and phenocrysts. No such separation was undertaken for olivine and pyroxene phenocrysts and microphenocrysts. The compositions of the chilled margin samples therefore represent the parental liquid plus ferromagnesian phenocrysts. Given that olivine microphenocrysts are incorporated in the analyzed chilled margin samples, the true parental melt composition would have a lower Ni content and higher $(\text{Mg} + \text{Fe}^{2+})_{\text{Ol-Opx}}/\text{Ni}$ ratio, and would plot to the right of the chilled margin samples as illustrated in Figures 5.8B and 5.9B. The olivine-free rocks, which would then have higher Ni contents than the true parental magma, could be derived by addition of Ni-bearing orthopyroxene. The olivine-bearing samples, which show much higher Ni content and plot to the left of the olivine-free samples, could be derived by accumulation of pre-emplacement olivine grains.

Rapid crystallization of the upper and lower chilled margins would trap olivine crystals, and therefore samples from these zones plot in the centre of Figure 5.9B. Moving away from the contacts, the number of trapped olivine microphenocrysts declines due to an axial migration of solids, possibly resulting from flow differentiation. Thus Ni concentrations are lower than in the chilled margins, and $(\text{Mg} + \text{Fe}^{2+})_{\text{Ol-Opx}}/\text{Ni}$ ratios are higher. The reduction in Ni content accompanying the decline in the number of olivine phenocrysts is partially balanced by the increasing amount of cumulus orthopyroxene. But since olivine contains much higher Ni concentrations, the loss of this phase is much more effective in controlling Ni content. As the cumulate proportion increases, the addition of orthopyroxene, and hence Ni, eventually counteracts the effect on $(\text{Mg} + \text{Fe}^{2+})_{\text{Ol-Opx}}/\text{Ni}$ of

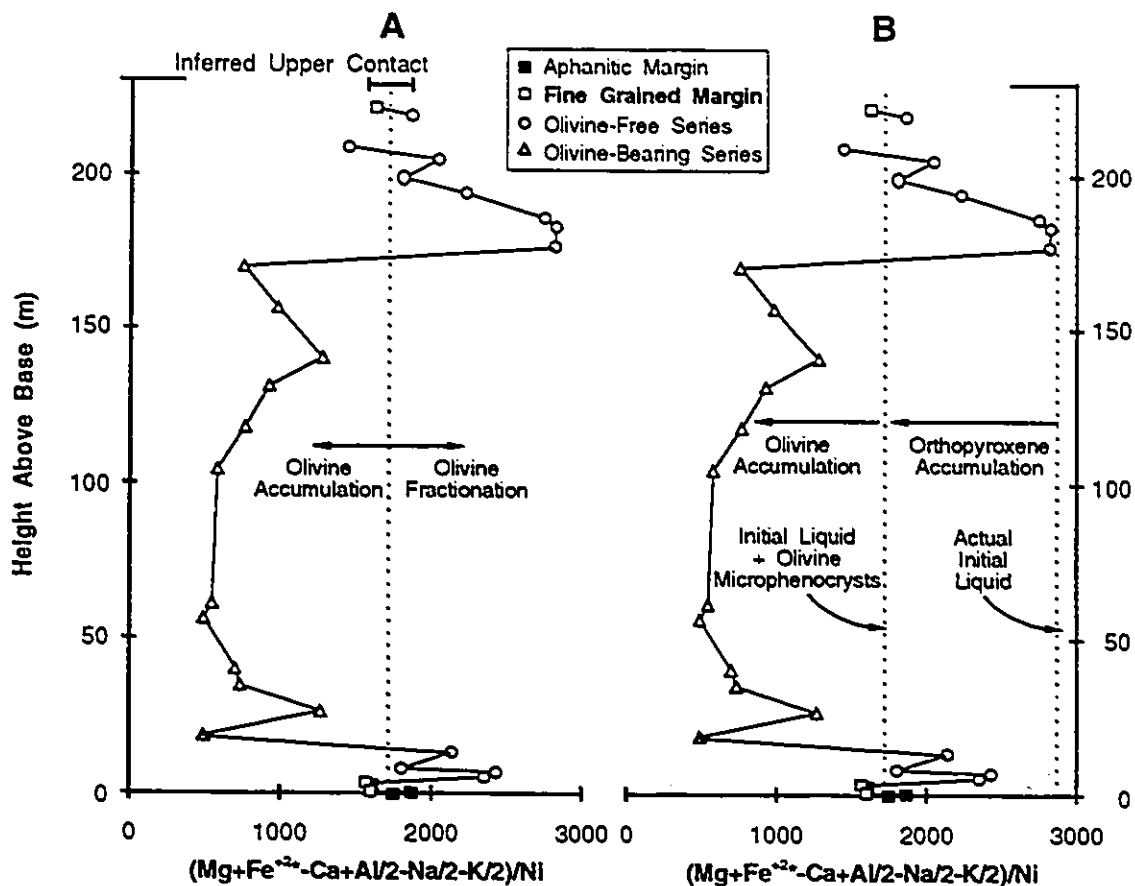


Figure 5.9. Variation in $(\text{Mg}+\text{Fe}^{2+})_{\text{ol+opx}}/\text{Ni}$ with stratigraphic position in sill C. (A) Olivine is fractionated from marginal zones and concentrated toward the centre of the sill during crystallization. (B) Olivine is a pre-emplacement crystallization product which is concentrated toward the centre of the sill during emplacement. Rapid crystallization of the chilled margins traps some olivine grains and therefore these samples represent initial liquid+trapped crystals. The true initial liquid lies to the right of all samples. See text for discussion. The broken vertical line indicates the average ratio for aphanitic chilled margin samples. The average 2σ analytical uncertainty is shown at the top of the diagram.

the loss of olivine, and this ratio declines. In the centre of the sill, olivine accumulates to form the olivine-bearing series, and these rocks contain high Ni concentrations and exhibit low $(\text{Mg}+\text{Fe}^{2+})_{\text{ol+opx}}/\text{Ni}$ ratios. The observed relationships between olivine-bearing, olivine-free and chilled margin samples can thus be explained without *in situ* crystallization of olivine.

The two alternatives discussed above are equally plausible in terms of geochemical considerations. Several observations, however, suggest that the second hypothesis is more probable. (1) Olivine occurs as microphenocrysts in some chilled margin samples. (2) , Where not completely replaced by secondary phases, olivine grains exhibit corona textures which indicate disequilibrium between this phase and the host melt. (3) Olivine is a very minor constituent of all other sills examined at Howse Lake. *In situ* crystallization of olivine in sill C would thus necessitate a unique melt composition in this intrusion. (4) Mineral compositions are essentially constant across the main body of sill C, and this may indicate that there was no significant evolution in melt composition for much of the crystallization history of the intrusion. Thus it is difficult to explain the low Ni contents of the olivine-free series relative to the chilled margins if olivine was a product of *in situ* crystallization. (5) Of the cumulus phases in the main part of the sill, only olivine exhibits a non-uniform distribution. If olivine was not a pre-emplacment crystallization product, this requires that either the accumulation mechanism did not act on other phases, or that sorting occurred prior to the crystallization of the other phases. Since plagioclase and pyroxene occur as microphenocrysts in chilled margin samples, a significant hiatus between the *in situ* crystallization of olivine and that of the other silicates is unlikely. If the olivine crystals were present in the melt during emplacement, however, an accumulation mechanism such as flow differentiation could produce the observed distribution prior to significant *in situ* crystallization. The same process would presumably be responsible for the distribution of plagioclase glomerocrysts.

The strongest argument for *in situ* crystallization of olivine in the sill derives from the fact that the ratio analysis indicates that this phase appears in the cumulate of the olivine-bearing series in place of orthopyroxene. This is consistent with the reaction relationship between the two minerals. But it is more likely that melt evolution would give rise to early crystallization of olivine in the outer zones of the sill, and later crystallization of hypersthene from possibly more siliceous residual melts in the centre of the intrusion. If olivine was a pre-emplacment crystallization product, however, the lower orthopyroxene content of the olivine-bearing samples might reflect the continued growth of olivine as the Ca-poor ferromagnesian silicate at the expense of orthopyroxene. Consequently, it is probable that olivine was present in the magma of sill C prior to emplacement, and

accumulation of this phase in the central regions of the intrusion resulted from a mechanical process such as flow differentiation rather than from melt evolution or other chemical change.

Relationship between Gabbroic and Felsic Rocks

Samples of the felsic series were omitted from Figures 5.8 and 5.9 because of their effect on the scale of the diagrams. The geochemical relationship between these rocks and the rest of the sill is obscured by lack of common conserved elements, but their low Ni contents necessitate considerable loss of Ni-bearing phases if they were derived from the parental melt by fractional crystallization. The unweighted average proportion of cumulate solid in the gabbroic samples is 35% by mass. If the felsic samples were derived by fractional crystallization in the sill, and assuming bulk distribution coefficients of $\ll 1$ and a closed magmatic system, then their Zr concentrations reflect crystallization of 85-88% of the original magma, and their Y concentrations reflect 68-81% crystallization. This degree of fractionation is inconsistent with the range in composition of the gabbroic samples. It is possible, however, that residual liquids derived from fractional crystallization of plagioclase and pyroxene continued to evolve thereby producing rocks of intermediate compositions which are present in unexposed portions of the sill. It is also possible, however, that the felsic series owes its composition to local assimilation of the Menihek country rocks. These possibilities are addressed later in this chapter.

5.5 Geochemical Variation in Sill B

The geochemical characteristics of sill B were determined by major and trace element analysis of 66 samples obtained from two measured profiles across the intrusion. The first is the most complete and includes analyses of 33 samples taken from 410 metres of exposed stratigraphic thickness. The second profile is thicker but less well exposed, and 33 samples were analyzed from approximately 500 metres of exposed stratigraphy. The two profiles are separated by a distance of approximately 1.5 kilometres (Fig. 4.1). The upper contact of the sill is not exposed at either location, but its position is revealed by a narrow, recessively weathered zone, readily apparent in air photographs, which separates

outcrops of Menihék sediments from the uppermost exposures of the sill. The lower contact of the second profile was also unexposed and was arbitrarily placed at the halfway point between nearest outcrops of gabbro and sediments. Sparsely glomeroporphyritic samples from the two profiles were prepared for analysis in the same manner as those of sill C, and consequently the data represent groundmass compositions. Because of the impossibility of separating glomerocrysts and groundmass from samples of the anorthositic phase, these analyses represent whole rock compositions (groundmass+glomerocrysts).

General Features

The geochemical sections across sill B at the two locations are illustrated in Figures 5.10 and 5.11. Comparison of these diagrams reveals that the general geochemical structure of the sill is similar in the two profiles, but the thickness of the second section is significantly greater. Geochemical and lithological variations allow subdivision of the sill into the following compositional types: (1) chilled margin, (2) gabbroic phase, (3) anorthositic phase, (4) oxide series and (5) felsic series. The gabbroic phase is composed of fine to medium grained gabbro-noritic rocks in which sparse glomerocrysts are set in a groundmass composed principally of plagioclase and subequal amounts of augite and hypersthene. This lithology occurs in the interval between the lower chilled margin and the anorthositic phase, and occupies much of the upper part of the sill. The gabbroic phase of sill B is compositionally similar to the olivine-free series of sill C. Its geochemistry is broadly similar to that of the chilled margin, with the most prominent differences found in the lower CaO and Al_2O_3 , and higher FeO^* and incompatible constituents (TiO_2 , MnO, P_2O_5 , Zr and Y) of the latter.

The anorthositic phase consists of medium to coarse grained, highly glomeroporphyritic hypersthene gabbro which occupies the lower half of the sill between 60 and 230 metres in profile 1, and between 85 and 295 m in profile 2. This lithology is underlain by the gabbroic phase in both sections, and the core phase occurs near the top of each interval. As would be expected from the large number of glomerocrysts, samples from the anorthositic phase are characterized by high concentrations of Al_2O_3 , CaO and Sr, and low FeO, MgO and incompatible elements.

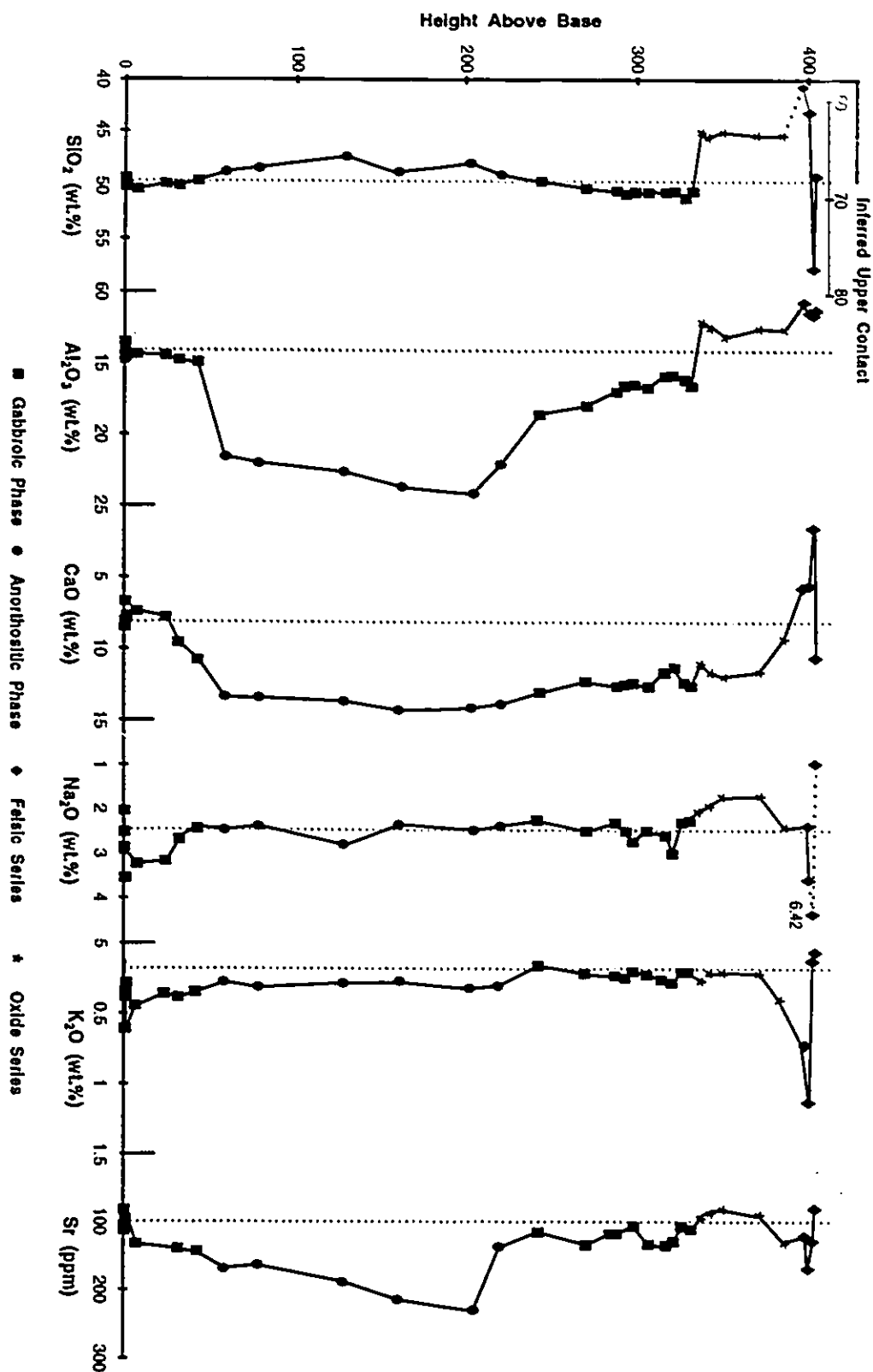


Figure 5.10. Variation in concentration of selected elements across Howse Lake sill B, profile 1. Vertical broken lines indicate average concentrations in three samples from the lower chilled margin. Average 2σ analytical is less than the size of plot symbols except where indicated.

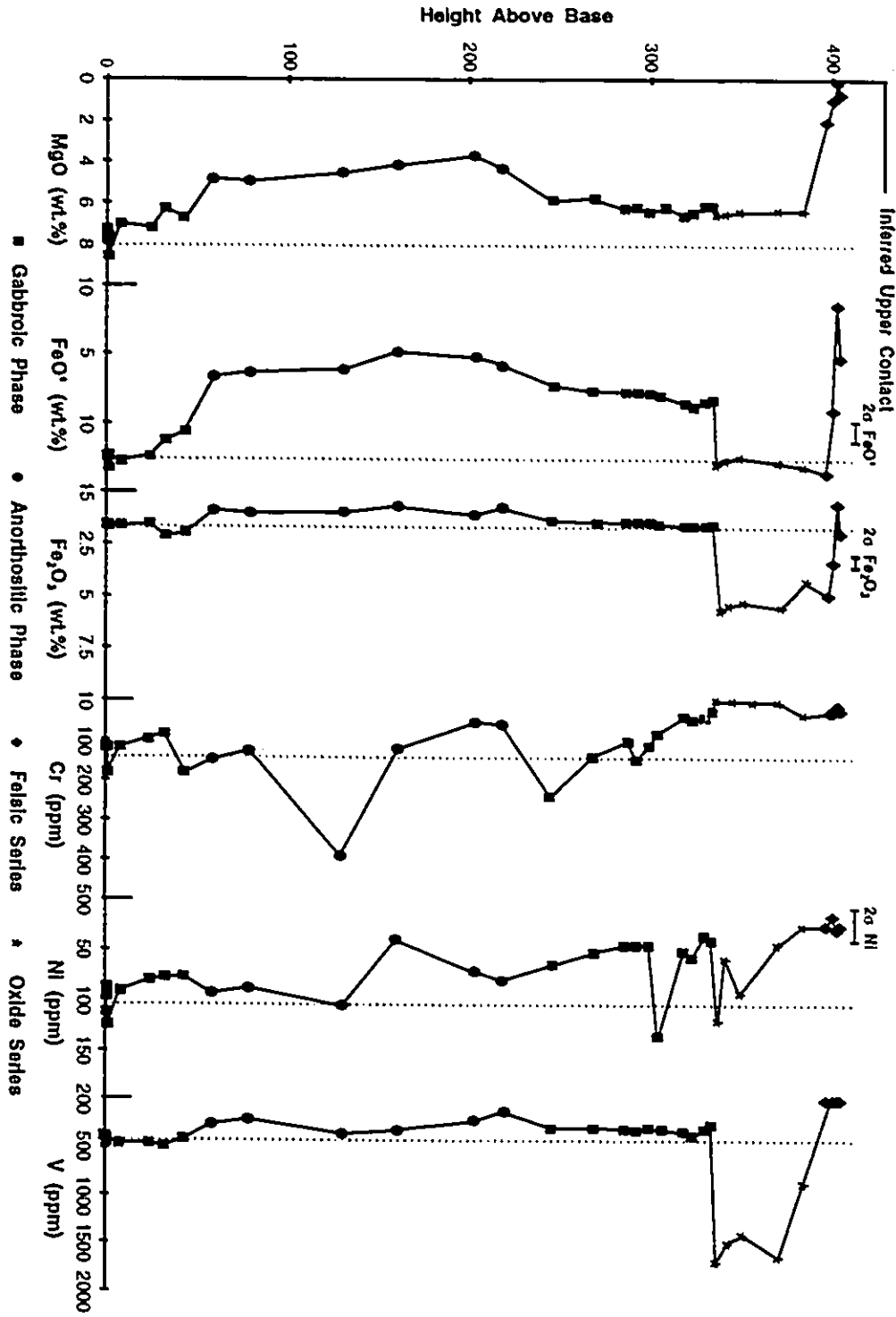


Figure 5.10 (continued). Variation in concentration of selected elements across Howse Lake sill B, profile 1. Vertical broken lines indicate average concentrations in three samples from the lower chilled margin. Average 2σ analytical is less than the size of plot symbols except where indicated.

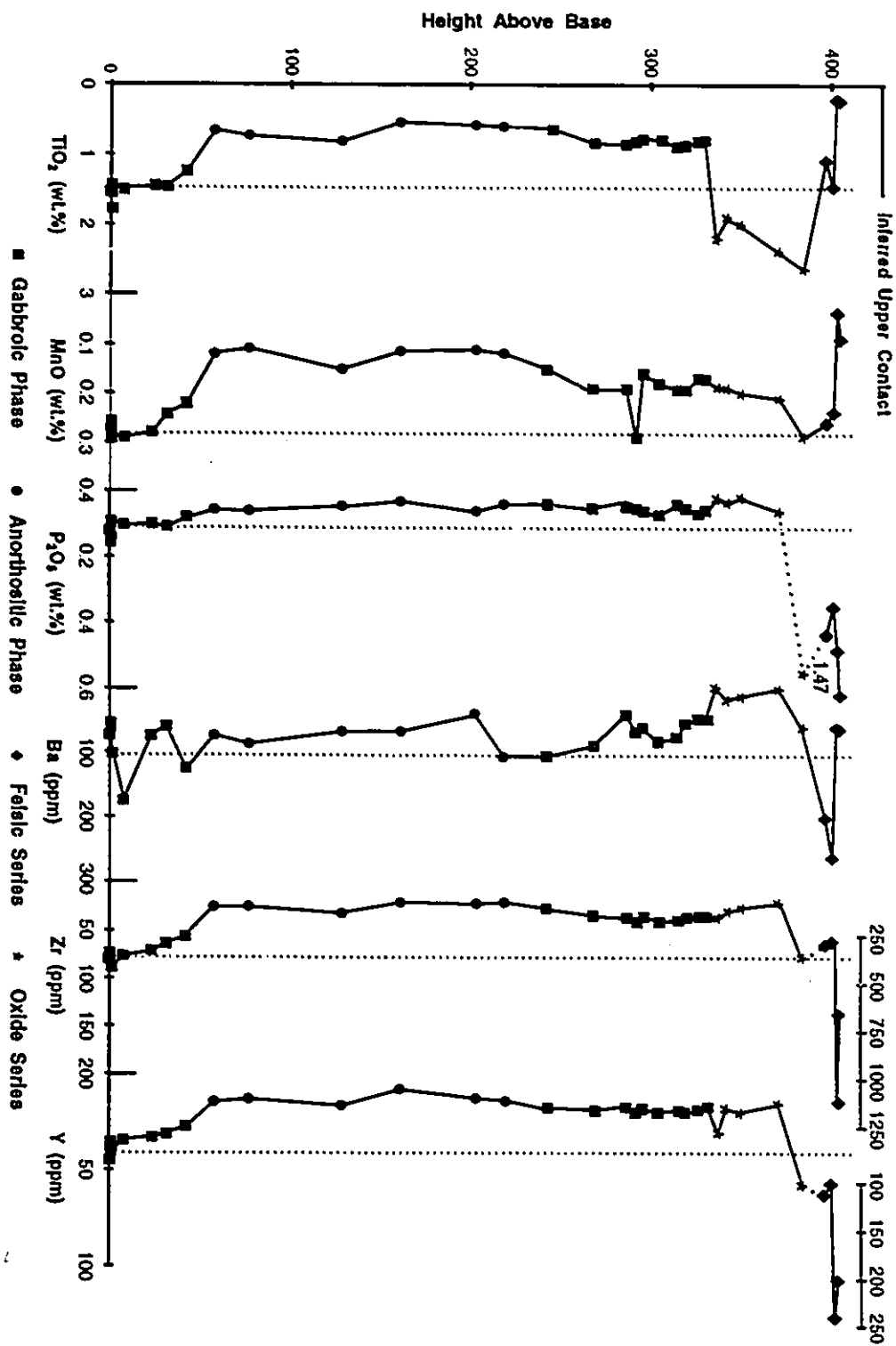


Figure 5.10 (continued). Variation in concentration of selected elements across Howse Lake sill B, profile 1. Vertical broken lines indicate average concentrations in three samples from the lower chilled margin. Average 2 σ analytical is less than the size of plot symbols except where indicated.

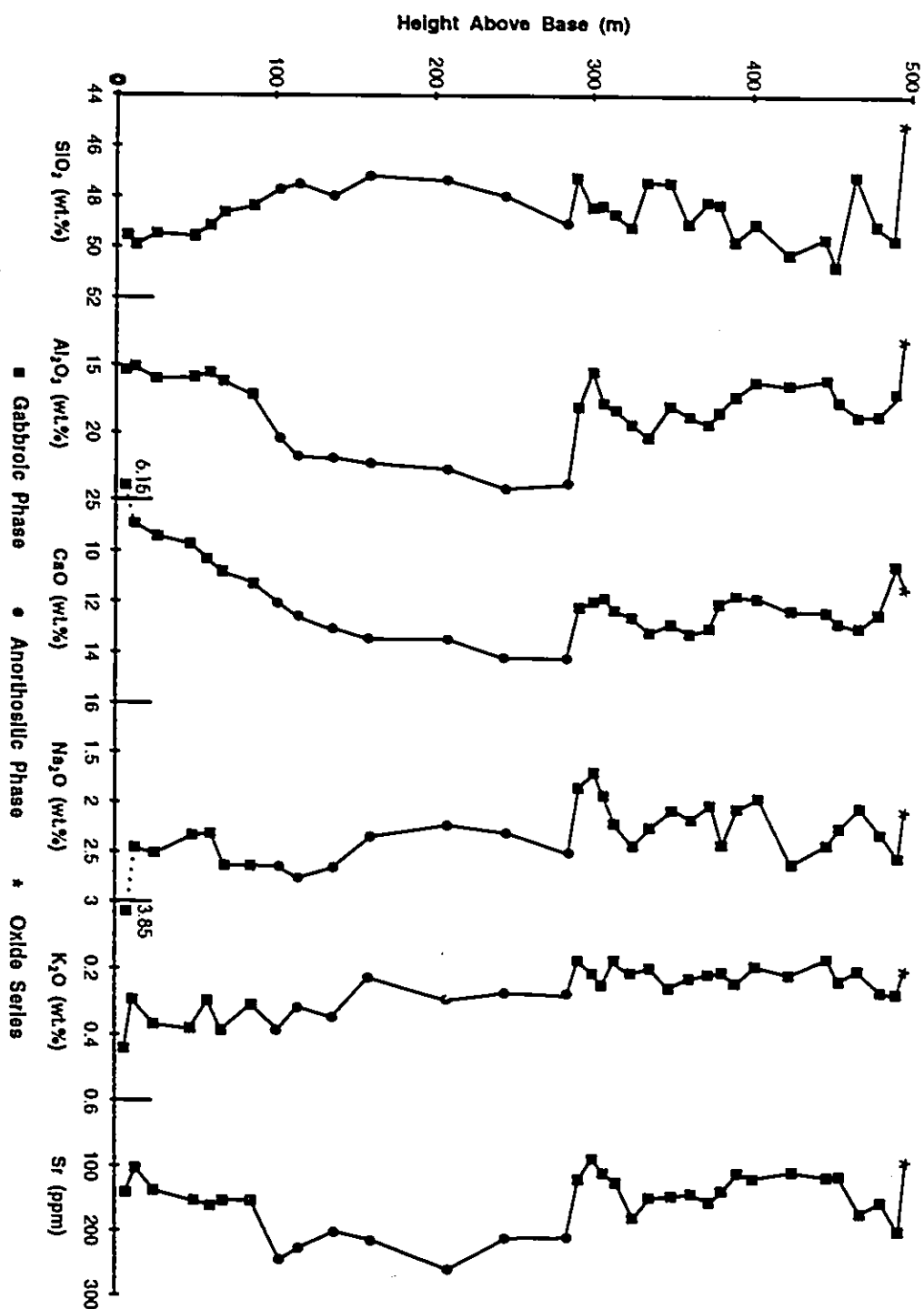


Figure B.11. Variation in concentration of selected elements across Howse Lake sill B, profile 2. The inferred position of the upper contact is 590 m above the basal contact. Average 2σ analytical is less than the size of plot symbols except where indicated.

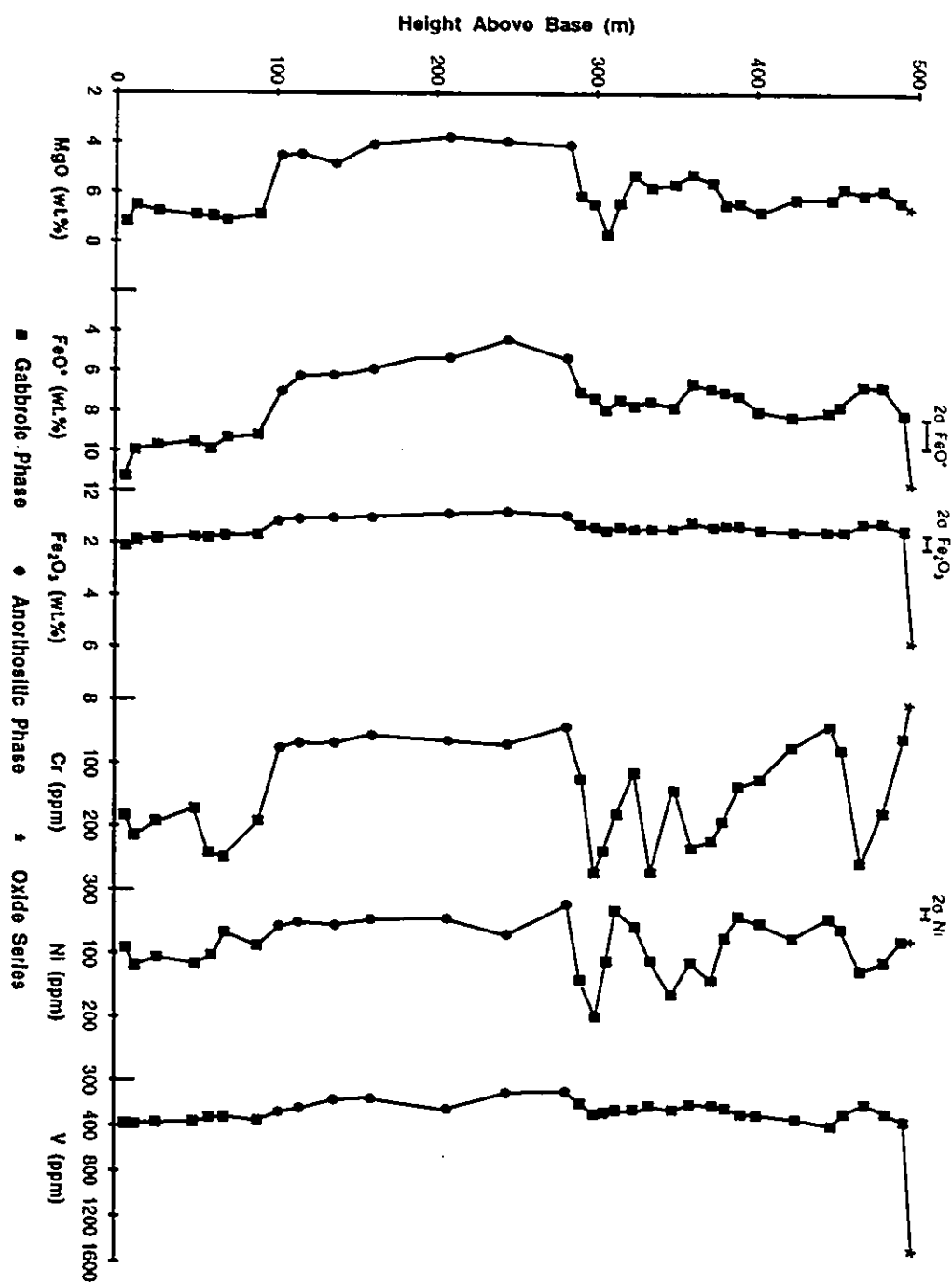


Figure 5.1.1 (continued). Variation in concentration of selected elements across Howse Lake sill B, profile 2. The inferred position of the upper contact is 590 m above the basal contact. Average 2σ analytical is less than the size of plot symbols except where indicated.

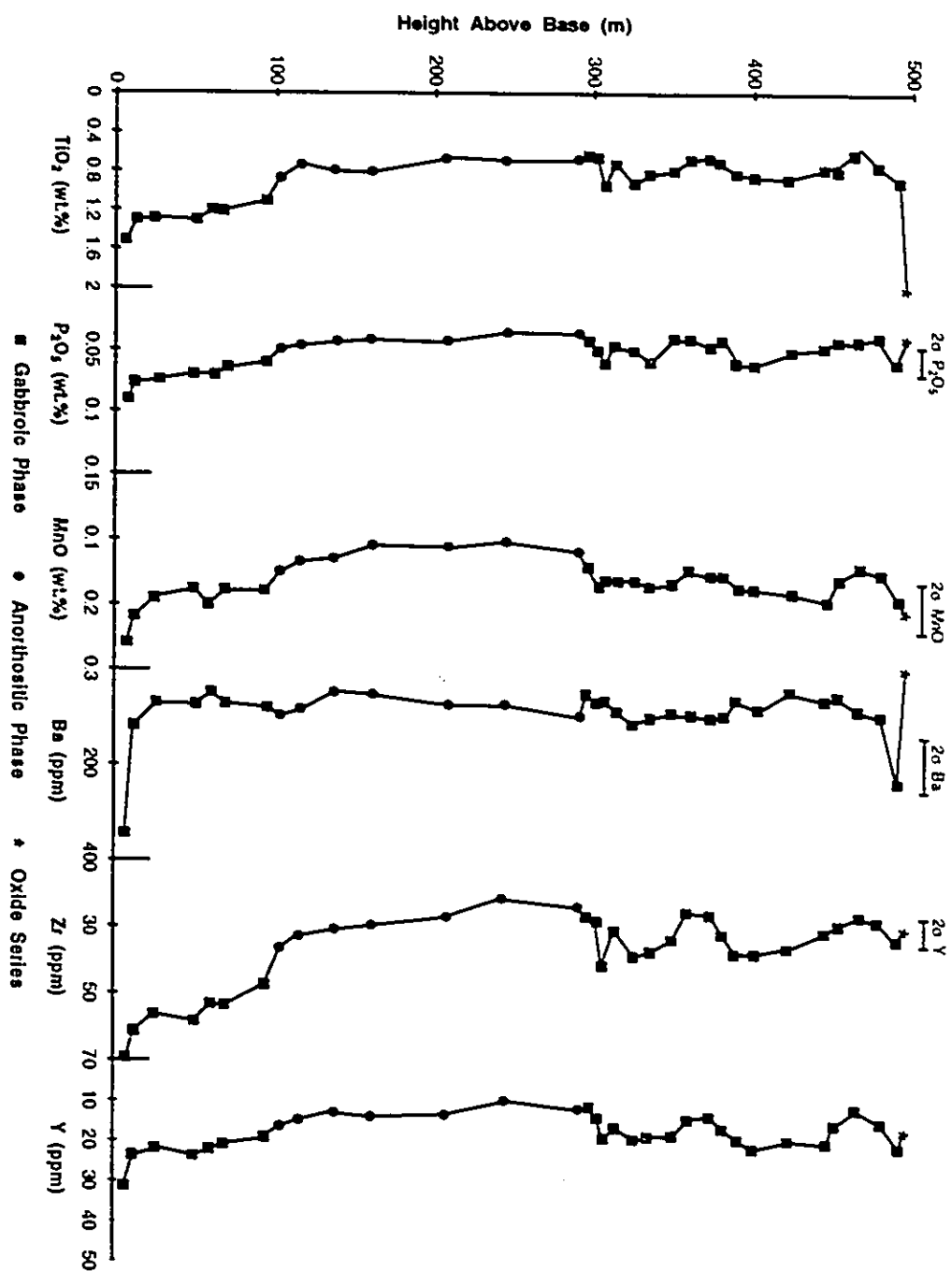


Figure 5.11 (continued). Variation in concentration of selected elements across Howse Lake sill B, profile 2. The inferred position of the upper contact is 590 m above the basal contact. Average 2σ analytical is less than the size of plot symbols except where indicated.

The anorthositic phase grades back into the gabbroic phase between 220 m and 230 m in profile 1 and between 290 m and 295 m in profile 2. The oxide series, which is characterized by high Fe-Ti oxide content, occurs in the interval between 335 m and 390 m in profile 1 and is represented by a single sample from the uppermost exposure in the second profile. The oxide series is distinguished by high concentrations of Fe_2O_3 , TiO_2 , and V, and low SiO_2 and Al_2O_3 . Notable also are the high P_2O_5 , Y and Zr contents of the uppermost oxide sample (F90-61) in profile 1. The high concentrations of these elements reflect elevated modal apatite and, to a lesser degree, zircon in the sample. The relationship between the oxide series and the suite of felsic rocks which occur at the top of profile 1 is unknown, but in the concentrations of MgO, Na_2O , K_2O , P_2O_5 , Ba, Zr and Y the upper oxide sample appears to be gradational into the felsic compositions. The felsic series occurs in an 8 m thick zone located approximately 30 m below the upper contact of the sill in the first section. As in sill C, the mineralogy of the felsic series is dominated by quartz, orthoclase and albite, but in two samples pyroxene comprises up to 25% of the rock. The geochemistry of the samples is variable, but all show high SiO_2 , Zr and Y, and low MgO, FeO^* , Cr, Ni and V.

The felsic series is unexposed in profile 2, as is the lower contact of the sill. The remaining geochemical subdivisions are present and these occur in the same stratigraphic order as in the first section. The principal difference between the two profiles lies in the thickness of the individual lithologies. In profile 1 the anorthositic series occupies an interval 170 m in thickness, while the equivalent horizon in profile 2 is 210 m thick. The overlying gabbroic phase extends from 230 m to 335 m in the first section, and from 295 m to 495 m in the second. The first exposure of oxide-rich material occurs at 335 m above the base of the sill in profile 1, and at approximately 495 m in profile 2. The inferred upper contact of the intrusion occurs at 430 m and 590 m above the base in the first and second locations respectively.

Choice of Conserved Elements for Pearce Ratios

Pearce ratios were employed to investigate the geochemistry of sill B in the same manner used in the analysis of sill C. As in sill C, the main body of sill B exhibits mesocumulate textures and low incompatible element concentrations which indicate that

these are cumulate rocks. The predominance of plagioclase, augite and hypersthene in the sill makes these phases the most likely to have fractionally crystallized. It is apparent from the high Fe-Ti oxide content of the oxide series that magnetite and ilmenite may have also taken part in the process. Given these possibilities, Mn, P, Ba, Y and Zr were chosen as potentially conserved elements. Ti was also examined since this element may have been conserved until oxide precipitation occurred. Non-constant ratios among these elements would indicate that one or more was incorporated into a fractionating mineral, or that some process which introduced independent variability in the concentrations of the elements has influenced the composition of the suite. The high P_2O_5 content of the uppermost oxide series sample in Figure 5.11, for example, is consistent with the elevated modal apatite content, and together these features suggest apatite accumulation occurred in the upper part of the oxide zone.

Table 5.5 summarizes the results of analysis for constancy of ratios derived from the selected elements. As in sill C, if all samples including those of the felsic series are considered, none of the ratios are constant. With exclusion of the felsic samples, only the Mn/Y ratio is constant, although the standard deviation of the Mn/Zr ratios is only slightly greater than the average analytical uncertainty. Variations in the ratios involving Y and P, however, arise largely from the anomalous concentrations of these elements in the apatite rich-oxide series sample (F90-61), and if this sample is excluded Mn, Y and Zr are conserved. If only gabbroic samples are considered, all of the selected elements except Ba are conserved. Consequently, Pearce ratios constructed from Ti, Mn, P, Zr and Y can be used to examine compositional variation in the main part of the sill, but only Mn, Y and Zr are valid denominator elements if the oxide series, excluding sample F90-61, is to be included in the analysis.

Simple Numerator Pearce Ratio Diagrams

Pearce ratio diagrams constructed using Zr as the denominator and the major elements as numerators are illustrated in Figure 5.12. Statistical data for these and for duplicate plots using Y as the denominator element are given in Table 5.6. The diagrams in which Y is the conserved element are not illustrated but are similar to those involving Zr, differing principally in the magnitudes of the ratios. The trends of the data for the

Table 5.5. Average, standard deviation and average analytical error for ratios constructed from potentially conserved elements in sill B. Standard deviation and analytical error are in bold typeface for constant ratios.

All Samples	Ti/Zr	Ti/Y	Mn/Zr	Mn/Y	P/Zr	P/Y	Ba/Zr	Ba/Y	Y/Zr
Average	289	536	58.5	112	22.2	41.9	1.33	2.62	0.52
Standard Deviation	183	238	20.2	32.5	27.9	35.4	0.87	1.59	0.11
Average Error	22.4	75.8	14.60	30.81	2.91	7.39	0.22	0.54	0.08
Excluding Felsic Samples									
Average	306	566	61.8	118	22.8	42.0	1.40	2.74	0.53
Standard Deviation	176	212	15.7	22.2	28.7	36.5	0.85	1.56	0.10
Average Error	23.7	80.0	15.4	32.5	2.98	7.41	0.23	0.56	0.08
Excluding Felsic and Oxide Samples									
Average	256	512	59.4	118	19.3	38.6	1.53	3.00	0.50
Standard Deviation	24.1	60.9	12.2	19.35	2.52	5.57	0.79	1.41	0.06
Average Error	24.5	72.4	14.8	32.5	2.53	6.81	0.26	0.61	0.08

gabbroic and anorthositic phase samples in the diagrams are sufficiently different that individual regression lines were calculated for these suites. The felsic samples do not meet the requirements for valid Pearce ratios and, although illustrated in the diagrams, have not been included in the regressions. Samples from the lower chilled margin of profile 1 were also excluded to prevent biasing of the slopes of regression lines for the remaining samples, although it can be seen from the diagrams that the marginal samples generally plot at or near the confluence of the regressions of the other series. Regressions for the oxide series are not shown since these are distinct from the those of the other series only for plots of Fe^{2+} , Fe^{3+} and Ti versus Si.

All plots meet the requirements of non-zero intercepts. Correlations the major elements and Si can be tested for significance using a two-tailed t test with the null hypothesis that the correlation coefficient is not significantly different than zero. The calculated t-statistic for each regression and the critical values of t for the appropriate degrees of freedom are given in Table 5.6. For most elements the calculated statistic far exceeds the critical value. The only failure of this test at the 95% confidence level is for the correlation between P and Si in the anorthositic phase when Y is used as the denominator element. In addition, the value of the t-statistic for the P-Si correlation in the gabbroic phase, and in both phases where Zr is the denominator, is only marginally greater than the critical value. Given that the concentration of P_2O_5 in the samples is commonly

Table 5.6 Slope (m), intercept (b), standard error of the intercept (SE_b), correlation coefficient (r) and test statistic (t) for ratio diagrams of major elements versus Si using Zr and Y as denominator elements. Diagrams using Zr as the denominator are illustrated in Figure 5.12. Critical values of t at the 95% confidence level are 2.031 (n=38) and 2.201 (n=13; from Davis, 1986, Table 2.11).

Denominator=Zr		m	b	SE_b	r	t
Al-Si	Gabbroic Phase (n=38)	0.50	-1916	427	0.97	33.136
	Anorthositic Phase (n=13)	0.67	-3236	766	0.99	31.782
Fe ⁺¹ -Si	Gabbroic Phase	0.01	150	15.6	0.95	26.463
	Anorthositic Phase	0.01	258	33.2	0.68	4.003
Fe ⁺² -Si	Gabbroic Phase	0.08	1051	86.0	0.96	27.091
	Anorthositic Phase	0.04	1700	219	0.68	4.003
Mg-Si	Gabbroic Phase	0.17	407	172	0.96	27.138
	Anorthositic Phase	0.09	1038	426	0.82	6.392
Ca-Si	Gabbroic Phase	0.34	-1671	193	0.99	50.801
	Anorthositic Phase	0.37	-1748	354	0.99	37.548
Na-Si	Gabbroic Phase	0.07	349	123	0.88	14.872
	Anorthositic Phase	0.08	575	226	0.93	11.642
K-Si	Gabbroic Phase	0.00	83.1	11.8	0.54	4.725
	Anorthositic Phase	0.01	71.9	37.6	0.74	4.871
Ti-Si	Gabbroic Phase	0.00	205	8.23	0.69	7.445
	Anorthositic Phase	0.00	187	27.43	0.68	4.045
P-Si	Gabbroic Phase	0.00	13.6	1.23	0.59	5.493
	Anorthositic Phase	0.00	15.0	2.73	0.60	3.159
Mn-Si	Gabbroic Phase	0.00	15.4	4.51	0.82	11.446
	Anorthositic Phase	0.00	22.5	6.75	0.63	3.471

Denominator=Y		m	b	SE_b	r	t
Al-Si	Gabbroic Phase	0.54	-5070	882	0.97	33.111
	Anorthositic Phase	0.66	-5802	1706	0.99	27.970
Fe ⁺¹ -Si	Gabbroic Phase	0.01	302	46.6	0.88	15.627
	Anorthositic Phase	0.01	397	99.7	0.61	3.216
Fe ⁺² -Si	Gabbroic Phase	0.08	2160	269	0.87	14.807
	Anorthositic Phase	0.05	2620	658	0.61	3.216
Mg-Si	Gabbroic Phase	0.18	382	420	0.94	22.447
	Anorthositic Phase	0.11	1302	996	0.81	6.268
Ca-Si	Gabbroic Phase	0.36	-4151	459	0.98	43.486
	Anorthositic Phase	0.36	-2964	939	0.99	27.423
Na-Si	Gabbroic Phase	0.07	982	253	0.82	11.655
	Anorthositic Phase	0.08	1063	461	0.93	11.708
K-Si	Gabbroic Phase	0.00	184	31.4	0.34	2.554
	Anorthositic Phase	0.01	158	76.3	0.72	4.579
Ti-Si	Gabbroic Phase	0.00	478	35.2	0.42	3.312
	Anorthositic Phase	0.00	358	71.2	0.62	3.378
P-Si	Gabbroic Phase	0.00	30.7	3.41	0.31	2.262
	Anorthositic Phase	0.00	31.4	5.71	0.42	1.831
Mn-Si	Gabbroic Phase	0.00	39.2	9.40	0.73	8.386
	Anorthositic Phase	0.00	53.9	12.7	0.69	4.075

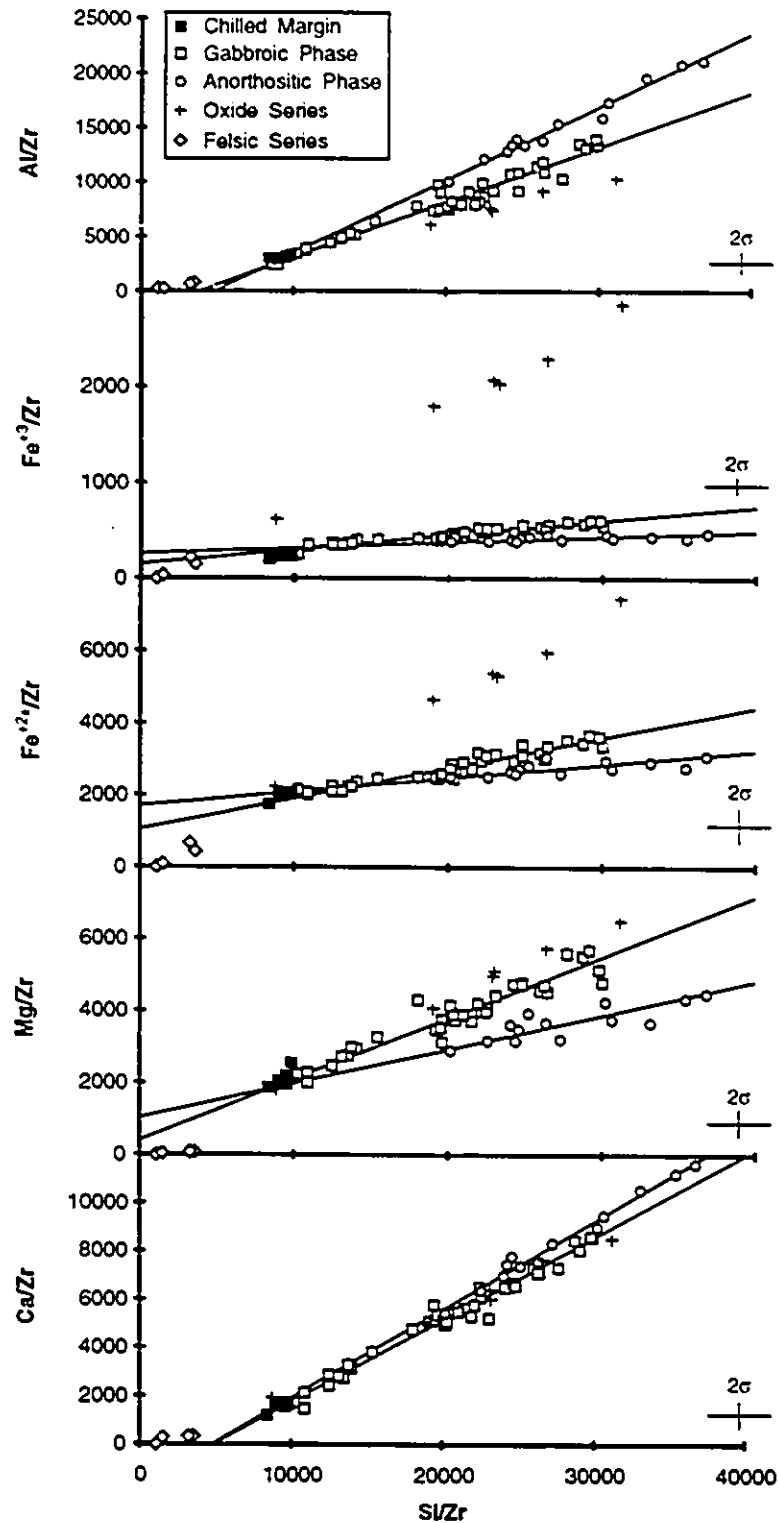


Figure 5.12. Pearce ratio diagrams of major elements versus Si using Zr as the denominator element. The reduced major axis regression lines for the gabbroic and anorthositic phases are illustrated. Statistical data are given in Table 5.6. Average 2σ analytical uncertainties are indicated by error bars.

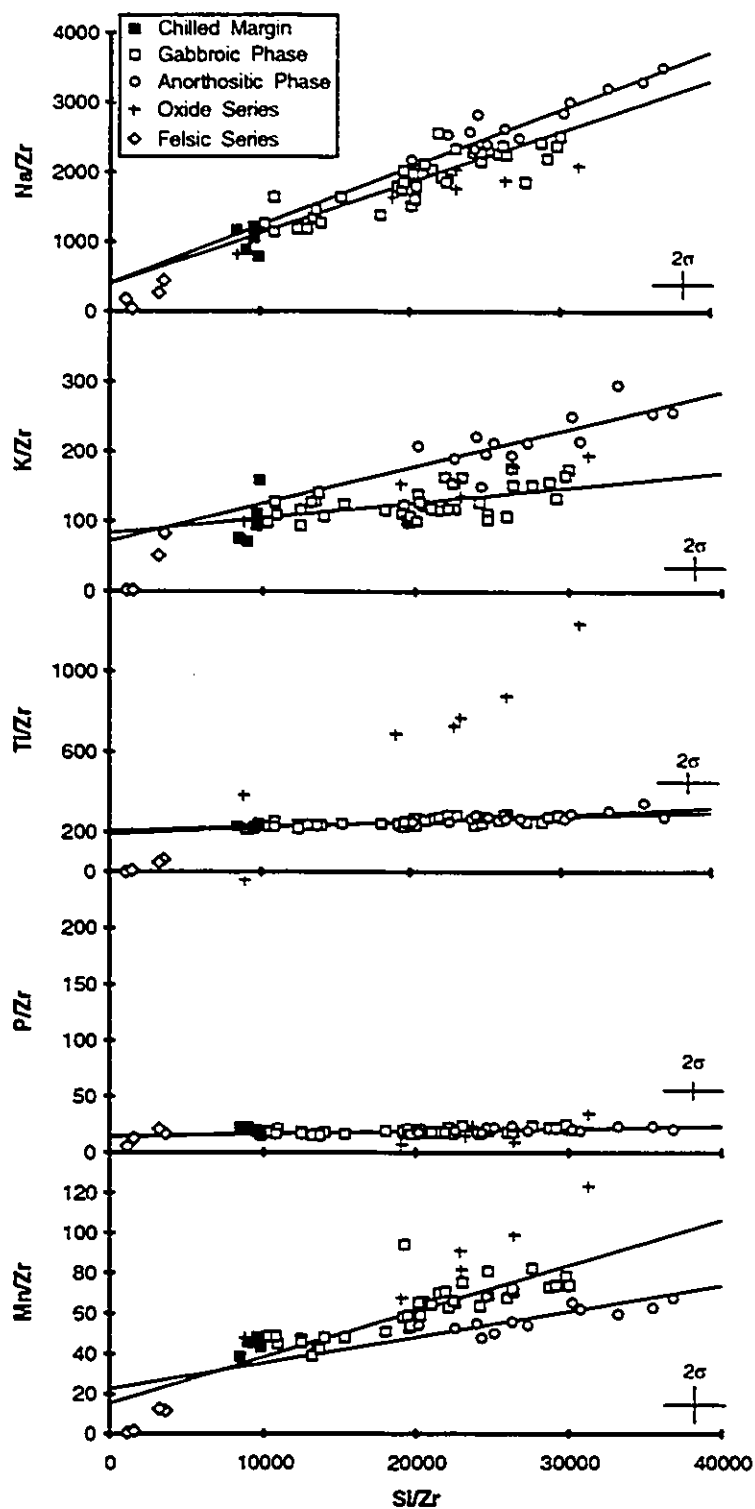


Figure 5.12 (continued). Pearce ratio diagrams of major elements versus Si using Zr as the denominator element. The reduced major axis regression lines for the gabbroic and anorthositic phases are illustrated. Statistical data are given in Table 5.6. Average 2σ analytical uncertainties are indicated by error bars.

close to the detection limit (i.e. $\sigma = \pm 100\%$ of concentration), the poor correlation is not surprising.

The chilled margin samples plot as a cluster to the lower left of the majority of samples from the two profiles in each of the diagrams of Figure 5.12. Fine to medium grained samples of the lower gabbroic phase near the contact plot between the chilled margin samples and those from the main body of the intrusion. Following the reasoning outlined in the analysis of sill C, these relationships are best interpreted as indicative of derivation of the gabbroic and anorthositic phases from the parental liquid (represented by the chilled margin) by addition of variable quantities of fractionally crystallized solids. Examination of Figure 5.12 and Table 5.6 indicate the two series have been enriched in Si, Al, Fe^{2+} , Mg, Ca and Na, with very minor enrichment in Fe^{3+} and K. The remaining elements, which display slopes close to zero, can be considered to have been conserved during differentiation.

Most of the oxide series samples also show enrichment in Si, Al, Fe^{2+} , Mg, Ca, and Na but, as expected, differ from the other suites in relatively strong Fe^{3+} and Ti enrichments. The uppermost oxide sample, however, plots to the left of all but one of the chilled margin samples and therefore may reflect loss of Si, Al, Fe^{2+} , Mg, Ca, and Na, while gaining Fe^{3+} , Ti and P. The felsic samples also plot to the left of the chilled margin, but their relationship to the rest of the samples is uncertain due to the absence of a common conserved element.

Complex Numerator Pearce Ratio Diagrams

The differences in the slopes of the trend lines depicted in Figure 5.12 reflect differences in the composition of the solid component added to the parental liquid. As noted, the principal mineralogy of the gabbroic and anorthositic series consists of plagioclase, augite and hypersthene, with Fe-Ti oxides as additional major phases in the oxide series. The variability in slopes must arise from either the absence of one or more of these phases from the cumulate in a given suite, or to variability in the proportions of these phases in the cumulate. These alternatives are examined in Figure 5.13 using complex numerator Pearce ratio diagrams to determine the identity and proportion of minerals in

Table 5.7. Slope (m), standard error of the slope (SE_m), correlation coefficient (r) and t statistic for complex numerator Pearce ratio plots using Zr and Y as denominator elements. Diagrams using Zr as the denominator element are illustrated in Figure 5.13. Critical values of t at the 95% confidence level are: 2.048 ($n=30$), 2.201 ($n=13$), and 3.182 ($n=5$). The critical value of t at the 90% confidence level with $n=3$ is 2.353 (from Davis, 1986; Table 2.11).

Diagram and Sample Series	Denominator = Zr				Denominator = Y			
	m	SE_m	r	t	m	SE_m	r	t
Figure 5-13A: Cpx+Opx vs. Pl+Kfs								
Gabbroic ($n=30$)	0.88	0.053	0.93	13.127	0.87	0.064	0.89	10.311
Anorthositic ($n=12$)	0.61	0.068	0.71	3.509	0.38	0.08	0.70	3.405
Oxide ($n=5$)	2.31	0.093	1.00	24.815	2.42	0.126	0.99	19.086
Figure 5-13B: Pl vs. Kfs								
Gabbroic	0.01	0.002	0.54	3.356	0.02	0.003	0.44	2.612
Anorthositic	0.02	0.004	0.73	3.672	0.02	0.004	0.69	3.333
Oxide	0.02	0.003	0.94	5.896	0.02	0.001	1.00	22.710
Figure 5-13C: Cpx vs. Pl								
Gabbroic	0.48	0.042	0.84	8.327	0.47	0.046	0.80	7.035
Anorthositic	0.17	0.015	0.94	9.937	0.15	0.017	0.92	8.194
Oxide	0.81	0.044	0.99	16.772	0.86	0.055	0.99	14.239
Figure 5-13D: Opx vs. Pl								
Gabbroic	0.48	0.049	0.79	6.724	0.56	0.069	0.66	6.294
Anorthositic	0.19	0.049	0.32	1.161	0.23	0.060	0.37	4.259
Oxide	0.98	0.028	1.00	32.084	1.02	0.030	1.00	6.910
Figure 5-13E: Mt+Ilm vs. Cpx+Opx								
Gabbroic	0.03	0.003	0.91	11.638	0.03	0.003	0.78	6.516
Anorthositic	0.02	0.004	0.74	3.801	0.01	0.003	0.84	5.270
Oxide	0.24	0.009	0.98	11.657	0.22	0.017	0.98	11.653
Figure 5-13F: An vs. Ab								
Gabbroic	3.00	0.303	0.78	6.654	3.62	0.385	0.76	6.096
Anorthositic	3.78	0.417	0.92	8.002	3.62	0.458	0.89	6.770
Oxide	2.35	0.434	0.89	4.424	2.03	0.363	1.00	41.462
Figure 5-13G: Mg vs. Fe²⁺								
Gabbroic	2.27	0.128	0.94	14.247	2.46	0.158	0.92	12.221
Anorthositic	2.49	0.176	0.87	6.159	2.38	0.389	0.81	4.755
Oxide	1.45	0.064	0.99	20.710	1.44	0.049	1.00	26.889

the accumulated fractionate for each series. Statistical data for the diagrams are summarized in Table 5.7, and interpretations of the slopes in terms of mineral proportions are provided in Table 5.8.

As with the simple numerator Pearce ratio plots, the slopes derived using Zr and Y as denominator elements in Figure 5.13 for the gabbroic and anorthositic suites are essentially identical. The slopes obtained using the two elements for the oxide series show

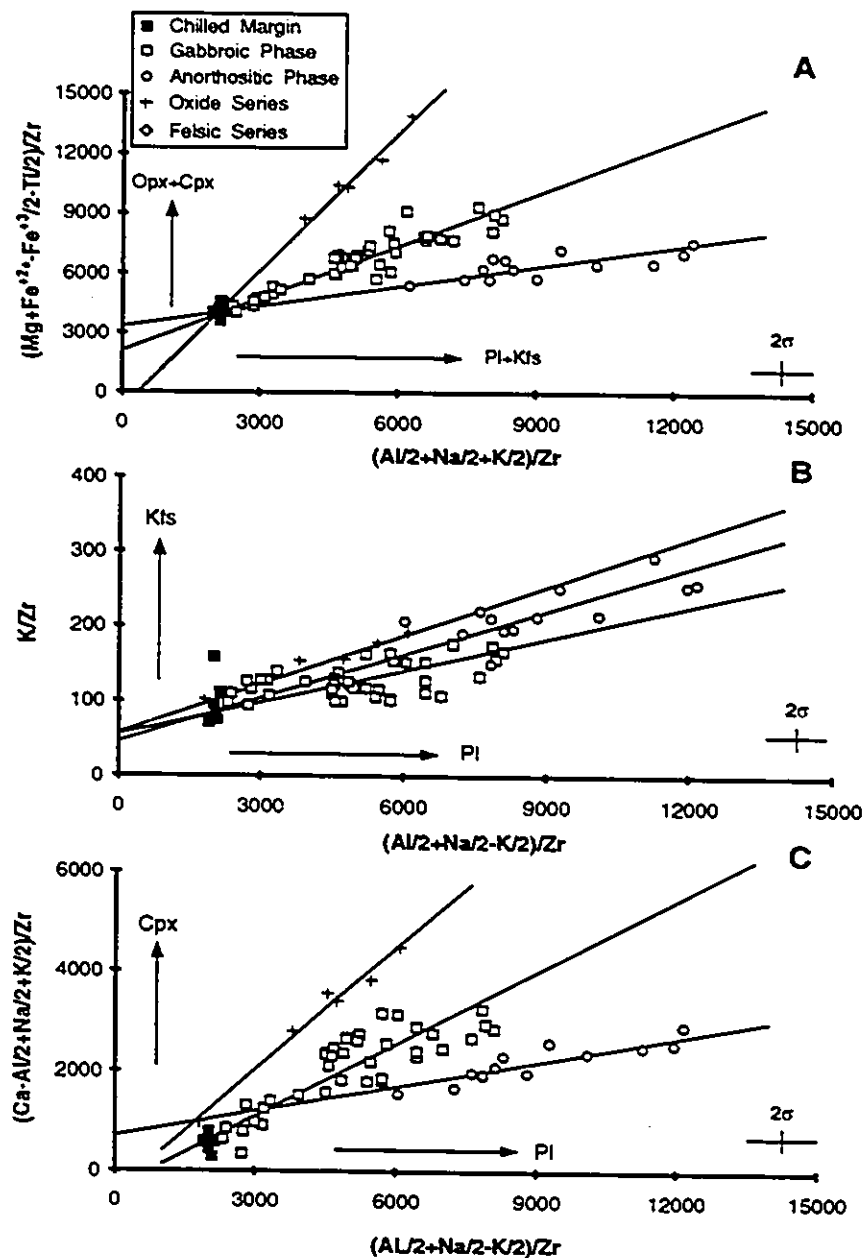


Figure 5.13. Complex numerator Pearce ratio diagrams illustrating the effects of different minerals on sample composition during fractional crystallization in sill B: (A) total pyroxene (Opx+Cpx) versus total feldspar (Pl+Kfs), (B) alkali feldspar (Kfs) versus plagioclase (Pl) and (C) clinopyroxene (Cpx) versus plagioclase (Pl). Reduced major axis regression lines for the three series are shown, and regression statistics are provided in Table 5.7. Average 2σ analytical uncertainties are indicated by error bars.

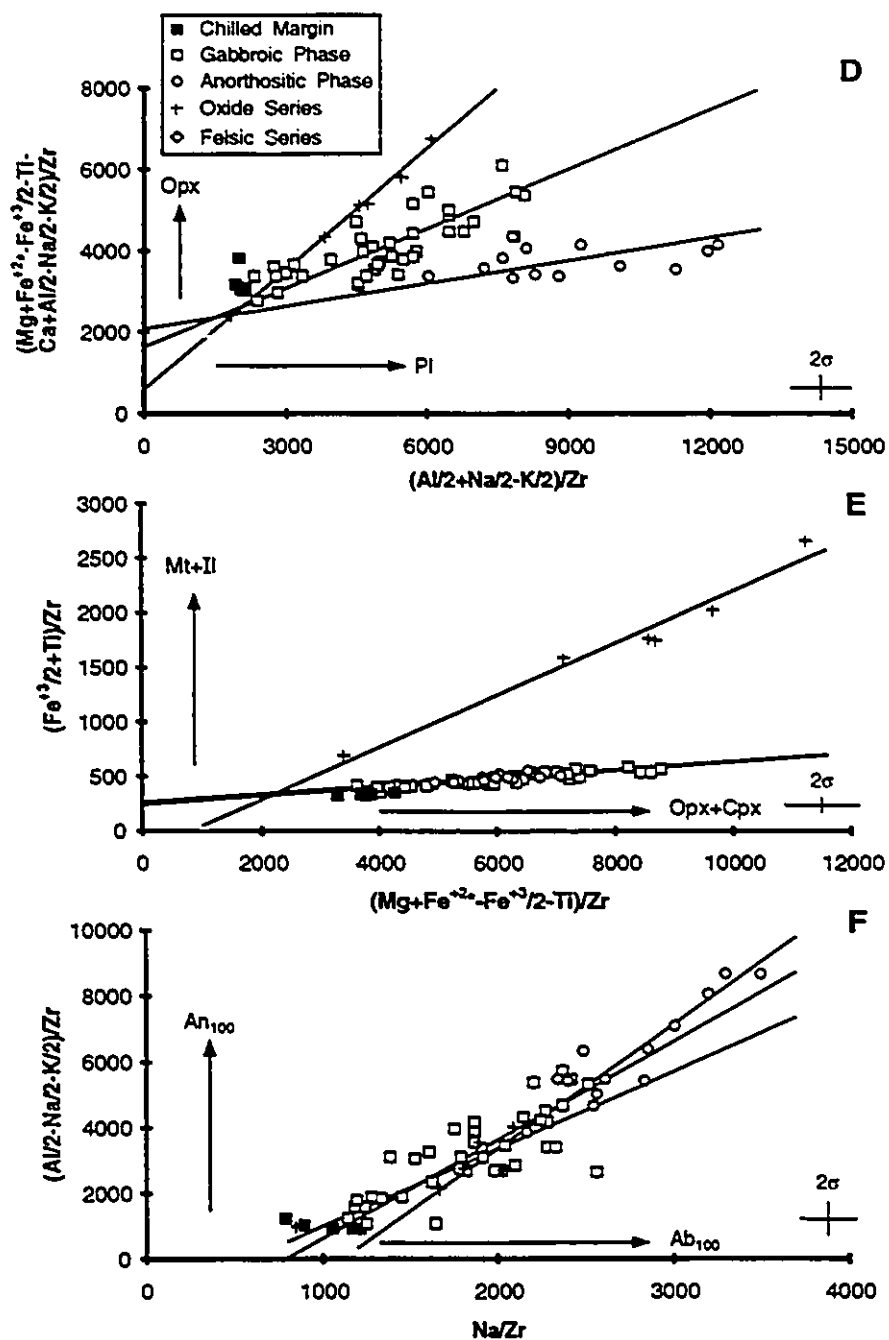


Figure 5.13 (continued). Complex numerator Pearce ratio diagrams illustrating the effects of different minerals on sample composition during fractional crystallization in sill B: (D) orthopyroxene (Opx) versus plagioclase (Pl), (E) magnetite+ilmenite (Mt+Ilm) versus total pyroxene (Opx+Cpx) (F) anorthite (An) versus albite (Ab) in plagioclase. Reduced major axis regression lines for the three sample series are shown, and regression statistics are provided in Table 5.7. Average 2σ analytical uncertainties are indicated by error bars.

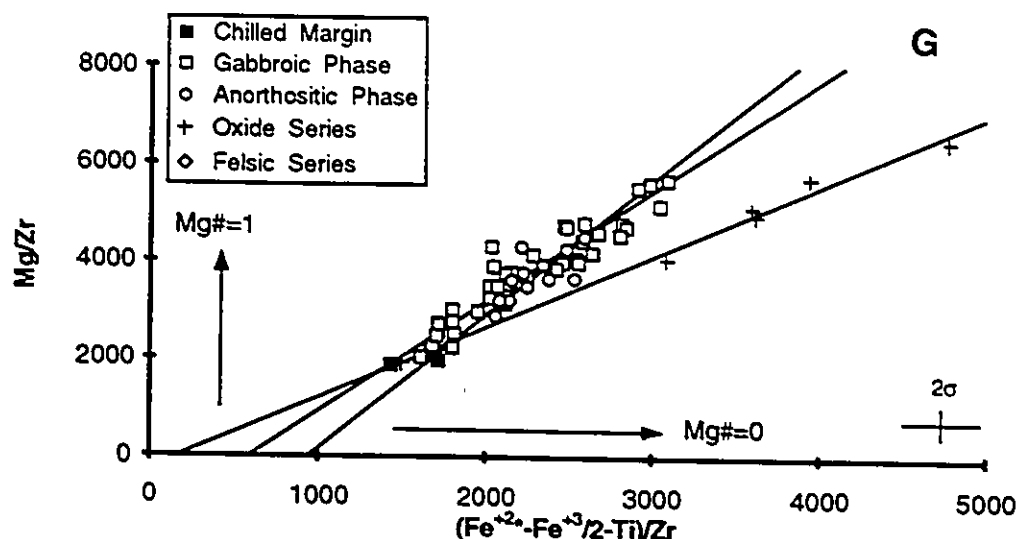


Figure 5.13 (continued). Complex numerator Pearce ratio diagram illustrating the effects of different minerals on sample composition during fractional crystallization in sill B: (G) Mg versus Fe^{2+} in pyroxenes. Reduced major axis regression lines for the three sample series are shown, and regression statistics are provided in Table 5.7. Average 2σ analytical uncertainties are indicated by error bars.

slight discrepancies when the P_2O_5 -enriched sample (F90-61) is included in the regressions. The Y/Zr ratio in this sample is 0.76, which is considerably higher than the average value of 0.52 in the gabbroic and anorthositic series (see Table 5.5 and Fig. 5.10). The P_2O_5 enrichment is consistent with the sample's elevated modal apatite content, and since the apatite-liquid partition coefficient for Y is greater than zero (Morse and Nolan, 1985), it is reasonable to conclude that the discrepancy in the oxide series regressions is the result of non-conservation of Y in the P_2O_5 -enriched sample. Exclusion of the anomalous sample removes much of the discrepancy, but has an adverse affect on the confidence limits associated with regression statistics because of the limited number of samples in this suite.

Figure 5.13A illustrates the relationship between the total feldspar (plagioclase+alkali feldspar) content and the total pyroxene (hypersthene+augite) content in the accumulated solid for the three series. The diagram supports field, petrographic and geochemical evidence that was the basis for compositional subdivision of the sill in that the anorthositic phase clearly reflects accumulation of a plagioclase-rich solid, while the

gabbroic phase and oxide series reflect accumulation of pyroxene and plagioclase in approximately equal proportions. The relative proportions of cumulate plagioclase and alkali feldspar can be derived from Figure 5.13B, from which it is apparent from the small values of the slopes that alkali feldspar was a very minor component of the cumulate in each suite. Individual slopes are identical within analytical error and reflect an average alkali-feldspar/plagioclase ratio of 0.02 ± 0.003 . If the total alkali feldspar content of the cumulate is resident in plagioclase, this translates into an average feldspar composition of approximately $(Ab+An)_{98}Or_2$, (2σ uncertainty in Or content is $\pm Or_{0.5}$) which is within error of values derived from analysis of sill C, and comparable to microprobe results obtained from plagioclase in the two intrusions.

The ratio of clinopyroxene to plagioclase in the cumulate is illustrated in Figure 5.13C. This ratio is 0.48, 0.16 and 0.84 in the gabbroic, anorthositic and oxide series respectively. The relative importance of orthopyroxene and clinopyroxene in the cumulate of each suite can be derived from comparison of Figures 5.13C and D. The latter diagram illustrates the relationship between orthopyroxene and plagioclase, and gives Opx/Pl ratios of 0.52, 0.21 and 1.00 for the gabbroic, anorthositic and oxide series respectively. This indicates that orthopyroxene was more abundant than clinopyroxene in the cumulate of each suite.

The effect of magnetite and ilmenite accumulation on the composition of the suites is examined in Figure 5.13E, which illustrates the relationship between the total oxide (magnetite+ilmenite) and total pyroxene (orthopyroxene+clinopyroxene) contents. The oxide series, with an oxide to pyroxene ratio of 0.23, is the only suite in which Fe-Ti oxides have contributed extensively to the composition of the cumulate. The small slopes for the other suites are consistent with the Pearce ratio plots of Figure 5.12 in which Ti is conserved, and Fe^{3+} shows only a slight positive relationship with Si. If the variability in Fe^{3+} is the result of substitution into pyroxene, the slopes of the regressions in Figure 5.13E translate into average Fe_2O_3 concentrations in these phases of 3.0 ± 0.3 wt. % for the gabbroic phase, and 3.5 ± 0.7 wt. % for the anorthositic series. These values are consistent with microprobe analyses from the sills, and it can be concluded that neither magnetite or ilmenite formed part of the cumulate in the gabbroic or anorthositic series.

The diagrams of Figure 5.13 allow calculation of average mineral proportions in the cumulate for each of the series, while error propagation provides an estimate of the potential variability in the composition of the cumulate in individual samples from a given suite (Table 5.8). It should be emphasized, however, that the overall composition of a sample is dependent on the composition of the cumulate, the degree of accumulation, and the composition of trapped intergranular liquid. Despite the potential variability due to these factors, sample compositions do in fact correspond well to cumulate compositions. The gabbroic rocks of the gabbroic phase have geochemistries intermediate to those of the oxide-rich, plagioclase-poor oxide series and the plagioclase-rich, pyroxene-poor anorthositic phase. This characteristic is reflected in the calculated average cumulate composition which consists of subequal amounts of pyroxene and feldspar, with orthopyroxene and clinopyroxene each accounting for approximately half of the total pyroxene content.

In contrast, the cumulate in the highly glomeroporphyritic, Al-Ca-rich, Fe²⁺-Mg-poor, anorthositic phase consisted of approximately 73% plagioclase, 16% orthopyroxene and 11% clinopyroxene. The composition of the cumulate in the oxide series was approximately 32% plagioclase, 30% orthopyroxene, 25% clinopyroxene, and 16% oxides. From Figure 5.12, it is apparent that the cumulate of the uppermost oxide sample of profile 1 also included apatite, but because only one sample of this type was available, the proportion of apatite cannot be determined. As in Figure 5.12, this sample plots slightly to the left of the chilled margin series in Figures 5.14A-D. This indicates that the principal silicate phases were removed from the melt from which the sample formed, while Fe-Ti oxides and apatite were added.

The average compositions of the silicate phases in the cumulate can be estimated from Figures 5.13F and 5.13G. The An/(An+Ab) ratio in plagioclase was 0.77 ± 0.04 , 0.79 ± 0.09 , and 0.70 ± 0.12 in the cumulate of the gabbroic, anorthositic and oxide series respectively. If all alkali feldspar occurs as Or in plagioclase, these ratios correspond to compositions of $An_{75}Ab_{23}Or_2$, $An_{77}Ab_{21}Or_2$ and $An_{69}Ab_{29}Or_2$. The average Mg# of ferromagnesian silicates was 0.70 ± 0.04 , 0.72 ± 0.13 and 0.59 ± 0.03 in the cumulate of the gabbroic, anorthositic and oxide series respectively. Among these values, only the Mg# of the oxide samples can be considered different from that of the other series given the

Table 5.8. Mineral proportions and mineral compositions in the cumulate of the three principal lithologies in sill B, and propagated analytical uncertainty. Values are calculated from the data of Table 5.7. ¹Total Feldspar was calculated from mineral ratios while ²Total Feldspar was derived from Figure 5.14A.

Gabbroic Phase					
Mineral Ratio		2σ	Mineral Proportion (mol.%)		2σ
Pl:Pl	1	0	Pl	49.5	3.6
Kfs:Pl	0.02	0.002	Kfs	0.8	0.2
Cpx:Pl	0.48	0.04	Cpx	23.7	3.6
Opx:Pl	0.52	0.06	Opx	26.0	4.5
Opx+Cpx:Pl+Kfs	0.87	0.06	¹ Total Feldspar	50.3	3.8
An/Ab	3.31	0.08	² Total Feldspar	53.4	3.1
Mg/Fe ⁺²	2.36	0.04	Mineral Composition		
			An/(An+Ab)	0.77	0.04
			Mg/(Mg+Fe ⁺²)	0.70	0.04

Anorthositic Phase					
Mineral Ratio		2σ	Mineral Proportion (mol.%)		2σ
Pl:Pl	1	0	Pl	72.1	4.1
Cpx:Pl	0.16	0.02	Kfs	1.4	0.4
Kfs:Pl	0.02	0.004	Cpx	11.5	1.5
Opx:Pl	0.21	0.05	Opx	15.0	4.6
Opx+Cpx:Pl+Kfs	0.36	0.07	¹ Total Feldspar	73.5	4.1
An/Ab	3.70	0.44	² Total Feldspar	73.6	5.0
Mg/Fe ⁺²	2.56	0.45	Mineral Composition		
			An/(An+Ab)	0.79	0.09
			Mg/(Mg+Fe ⁺²)	0.72	0.13

Oxide Series					
Mineral Ratio		2σ	Mineral Proportion (mol.%)		2σ
Pl:Pl	1	0	Pl	30.8	2.0
Cpx:Pl	0.81	0.04	Kfs	0.7	0.4
Kfs:Pl	0.02	0.003	Cpx	25.0	3.4
Opx:Pl	0.98	0.03	Opx	30.3	13.2
Opx+Cpx:Pl+Kfs	2.31	0.09	Mt+Ilm	13.2	2.1
Mt+Ilm:Cpx+Opx	0.24	0.01	¹ Total Feldspar	31.5	2.0
Mt+Ilm:Pl	0.43	0.02	² Total Feldspar	28.2	2.8
An/Ab	2.35	0.40	Mineral Composition		
Mg/Fe ⁺²	1.45	0.06	An/(An+Ab)	0.70	0.12
			Mg/(Mg+Fe ⁺²)	0.59	0.03

relatively high analytical uncertainties. The plagioclase and pyroxene compositions of the gabbroic and anorthositic suites are also indistinguishable from those derived from ratio analysis of sill C.

Despite the large uncertainties, the general trends of the calculated mineral compositions are consistent with the principal mineralogical differences among the main lithologies in sill B. The more calcic average plagioclase composition in the anorthositic series, for example, probably reflects the addition of numerous calcic glomerocrysts to the parental magma. Similarly, chemical changes in the melt would seem necessitated by the abundance of Fe-Ti oxides in the oxide series, and therefore more evolved mineral compositions might be expected in this suite. The less calcic plagioclase and less magnesian pyroxene compositions derived from ratio analysis for the cumulate of the oxide series are consistent with such melt evolution.

If this is the correct interpretation of mineral compositions in the oxide series, melt evolution presumably reflects the accumulation of Ca-rich plagioclase and Mg-rich pyroxenes in the gabbroic and anorthositic suites. This would be possible only if the sill became closed to replenishment during crystallization of the silicates. Closure would result in an increasingly iron-rich residual melt, but would also lead to increasing concentrations of incompatible elements, including Y and Zr. Melt trapped among cumulus grains would thus contain elevated incompatible element concentrations. Given equivalent cumulate proportions, a sample from the oxide series should therefore exhibit higher Zr and Y contents than a sample from one of the other series which crystallized from less evolved melt. With the exception of the P_2O_5 enriched sample (F90-61), however, there is no obvious elevation in the Y and Zr contents of the oxide suite (Fig. 5.11 and 5.13). This may indicate that the site of accumulation of the oxide minerals was not the site of crystallization.

Variation in Cumulate Proportions

An approximate measure of the relative proportion of cumulate and trapped liquid can be derived from the concentrations of Zr and Y following the approach outlined in the analysis of sill C, and using the chilled margin samples as representatives of the initial liquid. The variation in cumulate proportion as calculated from profile 1 is illustrated in

Figure 5.14. As in sill C, the cumulate proportion increases moving away from lower contact and the samples in the interior portion of the sill reflect significant crystal accumulation. Most samples show cumulate fractions in the 50-70% range, with a few reaching 75-80%. The cumulate proportion in the oxide series is comparable to the other suites with the exception of the uppermost sample (F90-61). If the liquid trapped in the oxide samples was enriched in incompatible elements, the calculated cumulate proportions would underestimate the true proportions. Since the calculated cumulate fraction reaches 75%, which is equivalent to the highest values from the other series, it seems unlikely that liquids trapped in the lower oxide series samples were substantially more evolved than those trapped in gabbroic or anorthositic phase samples.

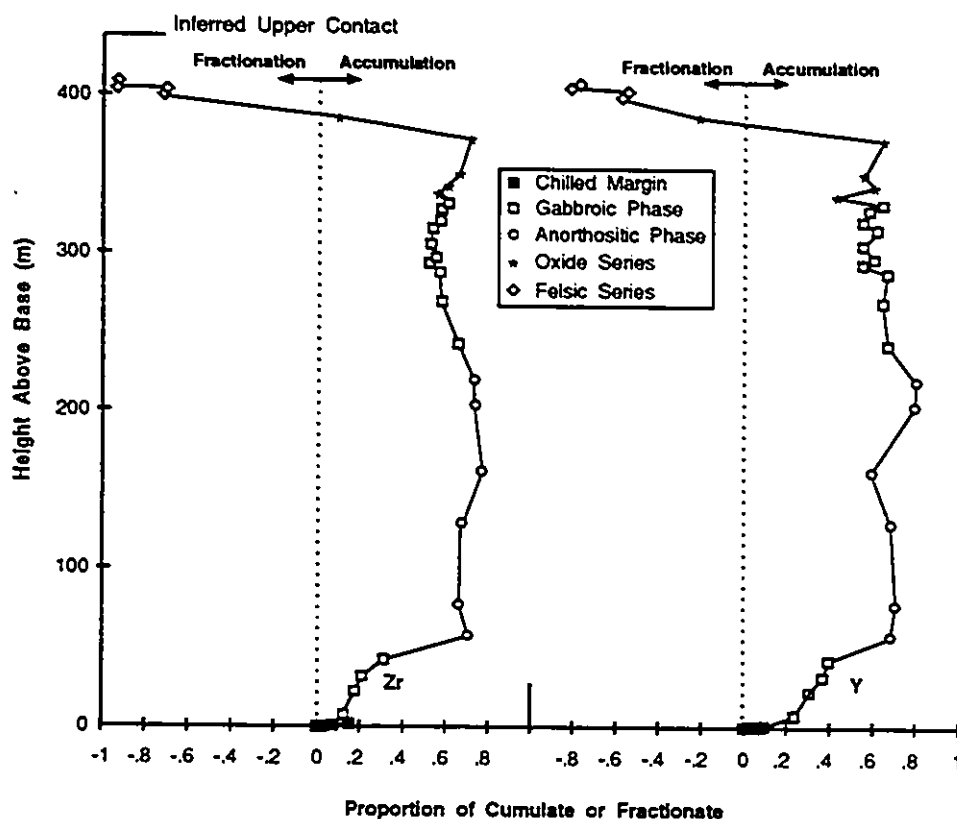


Figure 5.14. Proportions of cumulate added to, and fractionate removed from, the liquids which developed into the principal geochemical series of sill B. Proportions were obtained from the concentrations of Zr and Y in the individual samples compared to the most primitive (i.e., highest Y-Zr content) chilled margin sample. Dashed lines indicate no gain or loss.

In contrast, the low cumulate proportion calculated for sample F90-61, which clearly contains accumulated oxides, must indicate either Zr and Y enrichment in the trapped liquid, or accumulation of zircon and apatite. From the Pearce ratio diagrams of Figure 5.12 it was concluded that apatite formed part of the cumulate in this sample. Apatite accumulation is also suggested by the common textural relationship between this phase and Fe-Ti oxides found in the sample in which the oxide grains enclose euhedral apatite crystals (Fig. 4.16). No such relationship exists in the distribution of zircon, and it is considered unlikely that this phase formed part of the cumulate in the sample. If this interpretation is correct, the elevated Zr content indicates that the trapped liquid had evolved to a more incompatible element-enriched composition. Melt evolution is supported by the elevated Na_2O , K_2O and Ba content (Fig. 5.11) of sample F90-61 relative to the rest of the samples from the series, and by the evidence from Figures 5.12 and 5.13 which indicate that this sample reflects silicate fractionation rather than accumulation. Unfortunately, the uncertainty regarding the proportion of trapped liquid prevents any qualitative assessment of the degree of melt evolution.

Nature of the Solidification Process

The preceding analysis indicates that the compositions of most samples from the gabbroic, anorthositic and oxide series represent combinations of accumulated solid and trapped liquid. The gabbroic phase shows no textures indicative of sorting or separation of fractionally crystallized phases that might suggest transport to the site of accumulation. This indicates that this suite formed through *in situ* fractional crystallization and accumulation of plagioclase, orthopyroxene and clinopyroxene. Assuming that the composition of the chilled margin can be considered representative of the parental melt, the cumulate fraction in the gabbroic phase declines at the lower contact of the sill. A similar relationship was observed at both upper and lower margins of sill C, and this was considered indicative of more rapid crystallization near the margins due to higher cooling rates. The trend of increasing cumulate fraction moving away from the lower contact is consistent with *in situ* fractionation since cumulate proportion would be dependent on the rate of heat loss, with the highest rate of cooling, and consequently the smallest cumulate proportion, found in samples located near the sill contacts. The gabbroic phase is thus

equivalent to the olivine-free samples from sill C both in terms of the nature of the fractionation process and in the composition of the cumulate.

Formation of the Anorthositic Phase and Oxide Series

The compositions of the cumulate in the anorthositic phase and oxide series depart from that of the gabbroic phase. If these also represent the products of *in situ* fractionation, melt compositions must have differed from those from which the gabbroic phase crystallized. Since there is no evidence, based on Y and Zr content, of melt evolution (excluding sample F90-61), fractional crystallization and accumulation could not have been entirely *in situ* phenomena in the anorthositic and oxide series. That this is the case is most obvious in the anorthositic phase, which is both texturally and compositionally distinct from the other suites. The compositional and morphological differences between groundmass and glomerocryst plagioclase were earlier presented as evidence of crystallization of glomerocrysts prior to emplacement. The principal difference between the cumulate of the gabbroic and anorthositic phases lies in the plagioclase content, while the ratios of orthopyroxene to clinopyroxene are similar. The anorthositic series, which contains large quantities of glomerocrysts, can therefore be interpreted as the product of addition of glomerocrysts to liquids undergoing *in situ* fractionation of a cumulate having mineral proportions similar to that of the gabbroic phase. The high concentration of glomerocrysts in the anorthositic series, and the general pattern of glomerocryst distribution in the sill, is thought to result from flow differentiation during emplacement. This process is discussed in Chapter 6.

While pre-emplacement fractional crystallization of plagioclase is reasonable given the chilled margin compositions, pre-emplacement oxide crystallization is unlikely. Consequently this phase must have crystallized within the sill after emplacement. The presence of Fe-Ti oxides in the cumulate of only the oxide series indicates that these minerals were not liquidus phases during solidification of the rest of the sill. Oxides are present in small quantities in all samples, however, and therefore trapped liquids evolved to the point where these phases, along with other accessory minerals, crystallized in the intergranular spaces among the cumulate silicates. Together, these observations indicate that the melts from which oxides were fractionated as cumulus phases were more evolved

than those from which the other series developed, but could have been derived from the latter through silicate fractionation.

Iron enrichment is a characteristic trend of tholeiitic suites in which fractional crystallization was the dominant differentiation mechanism. In many intrusive bodies (e.g., the Skaergaard intrusion; Wager and Brown, 1968; the Kiglapait intrusion; Morse, 1981) this trend is terminated by the fractionation of Fe-Ti oxides, and residual melts subsequently evolve toward more siliceous and alkali-rich compositions. In such bodies, the precipitation of oxides is thought to reflect concentration of iron, particularly ferric iron, in residual liquids from which mafic silicates and plagioclase have been removed. Increased oxygen fugacity due to concentration and dissociation of juvenile water or incorporation of meteoric water will also promote oxide crystallization. In profile 1 of sill B, the elevated incompatible element content and the presence of cumulate apatite in the uppermost sample of the oxide series attest to the presence of more evolved residual liquids in the upper part of the sill, and this liquid is the most likely source of the cumulate oxides in the underlying samples. The relatively low incompatible element content of the latter suggests that the site of oxide precipitation and the site of accumulation were not the same. Given the density of the oxide cumulate, it is probable that oxide grains formed from an evolved liquid near the top of the sill would migrate downward under the influence of gravity. The absence of cumulate apatite in the lower oxide samples may indicate that precipitation of this phase was induced by oxide fractionation, or it could reflect the lower density contrast between the melt and apatite which might prevent sinking. The small grain size of the oxides, and the continued solidification of the sill would limit the effectiveness of density fractionation, and cumulate oxides would be confined to a narrow zone underlying the site of precipitation.

Relationship between Gabbroic and Felsic Rocks

The felsic horizon of sill B is similar to that of sill C in that it consists of quartz rich rocks which occur near the top of the intrusion. The felsic rocks contain high concentrations of incompatible elements (e.g., K_2O , P_2O_5 , Zr, Y) and are depleted in the components which constitute the bulk of samples from other lithologies in the sill (e.g., Al_2O_3 , CaO, MgO, FeO*). This suggests that, as with the oxide series, the felsic rocks may

have developed through fractional crystallization. The fact that the uppermost oxide series sample underlying the felsic horizon is also incompatible element enriched suggests that melt evolution continued during crystallization of Fe-Ti oxides. The fractionation of large quantities of oxide phases would tend to drive residual liquids to more siliceous compositions, and hence may have contributed to the composition of the overlying felsic series. Analysis of trace element ratios indicates that there are no conserved elements common to all of the sill B lithologies in the available data set. This indicates that either all available elements were involved in fractional crystallization or that a different process gave rise to the felsic rocks. Further discussion of the origin of this suite is given in section 5.7.

5.6 Relationship Between Glomeroporphyritic and Aphyric Sills

The most visible lithological variation in the Menihek sills derives from the presence of plagioclase glomerocrysts, and this fact has been used by previous workers as a means of categorizing the intrusions. Although mapping in the Howse Lake area indicates that few sills are completely aphyric, there are clear differences in the size and abundance of glomerocrysts in individual intrusions. This is particularly evident in the contrast in glomerocryst size and concentration between sparsely- and highly glomeroporphyritic intrusions. A useful subdivision of the intrusions can thus be made on the basis of the presence or absence of the anorthositic phase. To discriminate between these sill types in the following discussion, aphyric and sparsely glomeroporphyritic intrusions lacking the anorthositic phase are referred to as aphyric (i.e., sill C), while sills containing the anorthositic phase are termed glomeroporphyritic (i.e., sill B).

The gabbroic phases of aphyric and glomeroporphyritic sills are mineralogically, texturally and geochemically similar, and these similarities have been cited as evidence of a comagmatic origin for the two categories of sills (cf. Baragar, 1967; Skulski *et al.*, 1993). The preceding analysis of differentiation in sills B and C confirms the similar crystallization histories of the two types of sills. The composition of the cumulate in equivalent lithologies in the two intrusions are also similar; the cumulate in the gabbroic phase of sill B consisted of $50 \pm 4\%$ plagioclase, $24 \pm 4\%$ clinopyroxene and $26 \pm 5\%$

orthopyroxene, while the olivine-free series in sill B accumulated a solid with the composition $48 \pm 7\%$ plagioclase, $30 \pm 7\%$ clinopyroxene and $22 \pm 5\%$ orthopyroxene. The other lithological variants in the two sills cannot be directly compared, but the similarity of the gabbroic phases supports a comagmatic origin for aphyric and glomeroporphyritic intrusions. Given a comagmatic origin, incompatible element ratios should be identical in samples from the two types of sill. In addition, the differentiation trends defined by gabbroic phase samples (i.e., those lacking significant quantities of glomerocrysts) from aphyric and glomeroporphyritic intrusions should be similar.

In the preceding section, analysis of geochemical variation in sills B and C indicated that Y, Zr, P, Ti and Mn behaved as essentially incompatible elements during much of the crystallization histories of the two intrusions. A comparison of the ratios of these elements in samples from different intrusions can be used to test the comagmatic hypothesis. Calculated ratios are plotted in Figure 5.15 for all gabbroic and anorthositic phase samples obtained from the Menihek sills, and statistical data are given in Table 5.9. The data set excludes more evolved lithologies as well as oxide- or sulphide-rich samples in which the incompatibility of one or more of the elements is questionable.

In Figure 5.15, only a few samples lie outside of the 2σ confidence limits on the mean values of the ratios. Assuming a normal probability distribution for the analytical data, 5% of the samples could be expected to exceed the 2σ confidence limits, and consequently there is no evidence for non-equivalency of the ratios. Table 5.9 offers support for constancy of all of the calculated ratios except Ti/Zr. The difference, however, between the standard deviation of this ratio and the average analytical uncertainty is minor. Since the Ti/Zr ratios of chilled margin samples fall within the 2σ error bars in Figure 5.15, the non-constancy of the Ti/Zr ratio probably reflects limited compatibility of TiO_2 in pyroxene during sill crystallization rather than a fundamental difference in the TiO_2 content of the initial melts. Limited compatibility would result in slightly higher Ti/Zr ratios in cumulate samples relative to the chilled margins, and such a relationship is evident in Figure 5.15.

Mn/Zr ratios show a similar trend in the diagram, and Mn may also have been slightly compatible during the crystallization of pyroxenes. The small amounts of MnO

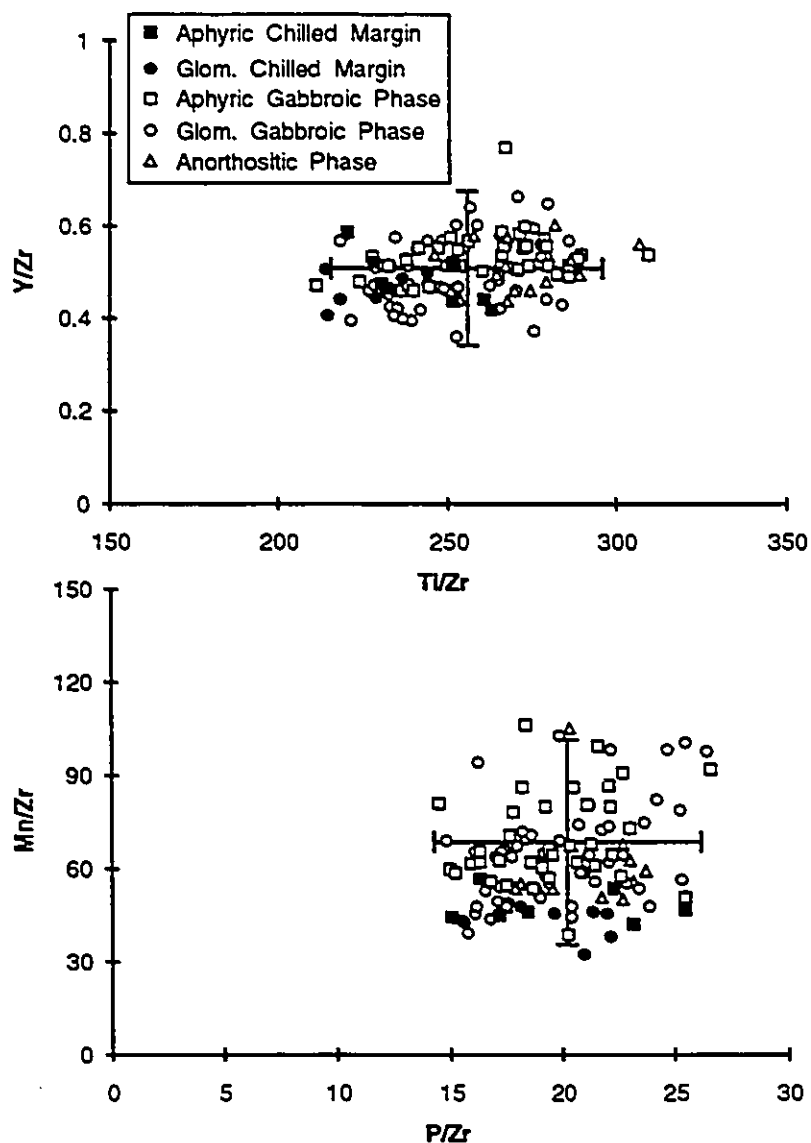


Figure 5.15. Incompatible element ratios for all gabbroic samples from the Menihek sills. Average 2σ analytical uncertainties are indicated by error bars which intersect at the mean for all samples. Note that the error bars embrace the majority of values.

Table 5.9. Average, standard deviation and average analytical error for incompatible element ratios derived from samples of the gabbroic and anorthositic phases in aphyric and glomeroporphyritic Menihek sills. Constant ratios are shown in bold typeface.

Ratio	Ti/Zr	Ti/Y	P/Zr	P/Y	Mn/Zr	Mn/Y	Y/Zr
Average	255	508	19.8	39.4	63.3	124.9	0.518
Standard Deviation	21.2	67.4	2.85	6.91	16.1	31.4	0.067
Average Analytical Error	19.8	71.8	2.94	7.45	16.5	34.5	0.080

and TiO_2 reported in microprobe analyses of hypersthene and augite from the sills confirms this interpretation. Although Ti and Mn may not have been completely incompatible during crystallization, it can be concluded that the ratios of Zr, Y, Ti, P and Mn in the initial magma of each of the Menihek sills sampled were essentially identical. This conclusion holds for the anorthositic phase of the sills, and hence precludes the possibility that this lithology developed from separate injections of chemically distinct magma.

In addition to the constancy of conserved element ratios, samples from aphyric and glomeroporphyritic sills should define similar trends in Pearce ratio plots if derived from a common magma source. In Figure 5.16, the major element compositions of all gabbroic phase samples are plotted against Si using Zr as the denominator element. Statistical data for the diagrams is provided in Table 5.10 for plots using both Zr and Y as denominator elements. Anorthositic phase samples have been excluded since the presence of glomerocrysts distorts the major element trends. Table 5.10 indicates that the correlations depicted in Figure 5.16 are significant for all elements versus Si. The equivalency of slopes for the two groups of gabbroic phase samples can be tested by computing the probability of obtaining the observed difference in slope based on a standardized normal probability distribution. The appropriate test statistic (Z) is given by (Davis, 1986):

$$Z = \frac{m_1 - m_2}{\sqrt{\sigma_{m_1}^2 + \sigma_{m_2}^2}}$$

where:

m_1 and m_2 = slopes of the regression lines

σ_{m_1} and σ_{m_2} = standard error of slopes m_1 and m_2 .

The slopes of the regression lines determined for the two data sets using both Zr and Y as denominator elements are statistically equivalent for all major elements except magnesium (Table 5.10). For the relatively immobile elements Al and Ca, and the conserved elements Ti, P and Mn, the regression lines fitted to the sample suites are virtually identical. The regressions of Na/Zr and K/Zr versus Si/Zr diverge slightly, but considering the variability of these ratios in chilled margin samples, the divergence probably reflects the mobility of Na and K during alteration and/or metamorphism. The

Table 5.10. Slope (m), intercept (b), correlation coefficient (r), standard error of the slope (SE_m), standard error of the intercept (SE_b), and test statistics (t) for regressions of the aphyric, glomeroporphyritic and felsic sample suites shown in the Pearce ratio diagrams of Figures 5.16. The test statistic (Z) for equivalency of slopes between aphyric and glomeroporphyritic samples is also shown. Critical values of t at the 95% confidence level are: aphyric series, n=37, t=2.027; glomeroporphyritic series, n=53, t=2.008, felsic series, n=16, t=2.120 (from Davis, 1986, Table 2.11). Critical value of Z at the 95% confidence level is ± 1.96 .

Denominator = Zr								
Plot	Series	m	b	r	SE_m	SE_b	t	Z
Al-Si	Aphyric	0.50	-2490	0.98	0.017	374	38.83	0.03
	Glomeroporphyritic	0.50	-1708	0.96	0.019	438	34.26	
	Felsic	0.31	-161	0.97	0.018	68.0	21.15	
Fe ⁺³ -Si	Aphyric	0.02	182	0.75	0.002	42.2	8.82	0.02
	Glomeroporphyritic	0.02	98.3	0.93	0.001	20.3	25.42	
	Felsic	0.06	-61.9	0.96	0.004	14.4	17.87	
Fe ⁺² -Si	Aphyric	0.11	1052	0.74	0.011	251	8.69	0.02
	Glomeroporphyritic	0.11	660	0.93	0.005	123	24.51	
	Felsic	0.17	-182	0.95	0.013	48.8	15.53	
Mg-Si	Aphyric	0.33	-1511	0.77	0.034	742	9.56	3.39
	Glomeroporphyritic	0.22	-518	0.93	0.011	261	24.00	
	Felsic	0.16	-228	0.94	0.015	56.4	12.76	
Ca-Si	Aphyric	0.32	-1531	0.97	0.012	267	34.35	0.63
	Glomeroporphyritic	0.31	-1223	0.98	0.009	193	50.01	
	Felsic	0.20	-281	0.92	0.020	75.6	11.65	
Na-Si	Aphyric	0.07	171	0.82	0.007	147	11.27	1.40
	Glomeroporphyritic	0.06	503	0.86	0.005	106	15.76	
	Felsic	0.16	-85.2	0.90	0.018	68.2	9.63	
K-Si	Aphyric	0.004	75.2	0.51	0.001	19.7	4.26	1.82
	Glomeroporphyritic	0.003	66.1	0.50	0.001	21.9	4.97	
	Felsic	0.028	-32.9	0.68	0.005	20.0	4.20	
Ti-Si	Aphyric	0.001	235	0.34	0.001	14.5	2.43	-1.68
	Glomeroporphyritic	0.001	225	0.51	0.001	14.0	5.17	
	Felsic	0.026	-38.6	0.96	0.002	7.26	16.53	
P-Si	Aphyric	0.000	16.0	0.39	0.000	1.91	2.96	-1.90
	Glomeroporphyritic	0.000	15.7	0.44	0.000	3.11	4.14	
	Felsic	0.006	-6.72	0.87	0.001	3.04	8.46	
Mn-Si	Aphyric	0.002	27.9	0.88	0.000	4.27	15.35	-0.14
	Glomeroporphyritic	0.002	20.8	0.88	0.000	5.39	17.74	
	Felsic	0.005	-6.12	0.96	0.000	1.43	15.88	

Denominator = Y

Plot	Series	m	b	r	SE_m	SE_b	t	Z
Al-Si	Aphyric	0.50	-4549	0.98	0.02	720	37.79	1.36
	Glomeroporphyritic	0.51	-3700	0.96	0.02	853	35.29	
	Felsic	0.34	-760	0.95	0.03	257	14.88	

Plot	Series	m	b	r	SE _m	SE _b	t	Z
Fe ⁺³ -Si	Aphyric	0.02	315	0.78	0.00	77.8	10.01	-0.44
	Glomeroporphyritic	0.02	141	0.91	0.00	51.6	20.70	
	Felsic	0.06	-241	0.91	0.01	59.2	10.20	
Fe ^{+2*} -Si	Aphyric	0.11	1757	0.78	0.01	473	9.75	-0.31
	Glomeroporphyritic	0.11	980	0.90	0.01	317	19.76	
	Felsic	0.19	-711	0.91	0.02	185	10.78	
Mg-Si	Aphyric	0.33	-2828	0.78	0.03	1370	10.01	2.86
	Glomeroporphyritic	0.24	-1846	0.93	0.01	552	24.75	
	Felsic	0.19	-909	0.91	0.02	194	10.52	
Ca-Si	Aphyric	0.31	-2583	0.97	0.01	527	32.01	-0.45
	Glomeroporphyritic	0.32	-2548	0.97	0.01	456	41.80	
	Felsic	0.25	-1261	0.86	0.03	312	8.14	
Na-Si	Aphyric	0.07	389	0.81	0.007	276	10.96	1.91
	Glomeroporphyritic	0.06	1181	0.83	0.005	217	13.81	
	Felsic	0.19	-505	0.83	0.028	257	6.87	
K-Si	Aphyric	0.00	154	0.49	0.001	31.5	4.09	1.33
	Glomeroporphyritic	0.00	137	0.50	0.001	47.7	4.94	
	Felsic	0.02	-34.5	0.62	0.008	75.6	3.52	
Ti-Si	Aphyric	0.00	424	0.40	0.001	37.5	3.09	1.81
	Glomeroporphyritic	0.00	455	0.24	0.001	35.5	1.93	
	Felsic	0.03	-139	0.91	0.003	30.7	10.86	
P-Si	Aphyric	0.00	28.5	0.46	0.000	3.21	3.72	1.76
	Glomeroporphyritic	0.00	32.4	0.32	0.000	6.06	2.67	
	Felsic	0.01	-27.7	0.85	0.001	11.1	7.62	
Mn-Si	Aphyric	0.00	46.2	0.88	0.000	7.84	14.65	2.18
	Glomeroporphyritic	0.00	40.6	0.82	0.000	10.6	13.67	
	Felsic	0.01	-22.2	0.92	0.001	5.68	11.25	

only significant difference in trends occurs in the plot of Mg/Zr versus Si/Zr in which the slope of the regression line for the glomeroporphyritic suite is smaller than that of the aphyric samples. The opposite trend is slightly developed in the plot of Fe^{+2*}/Zr versus Si/Zr, although the samples sets are statistically equivalent in this plot.

There are two possible explanations for the discrepancy in the individual regressions involving magnesium. First, crystallization in the aphyric sills may have involved ferromagnesian phases in which the Mg/Fe^{+2*} ratio was slightly higher than the equivalent minerals in glomeroporphyritic sills. Such a difference would presumably reflect a difference in melt geochemistry, with the aphyric initial magmas being slightly more magnesian than those of the glomeroporphyritic sills. Alternatively, it is possible that

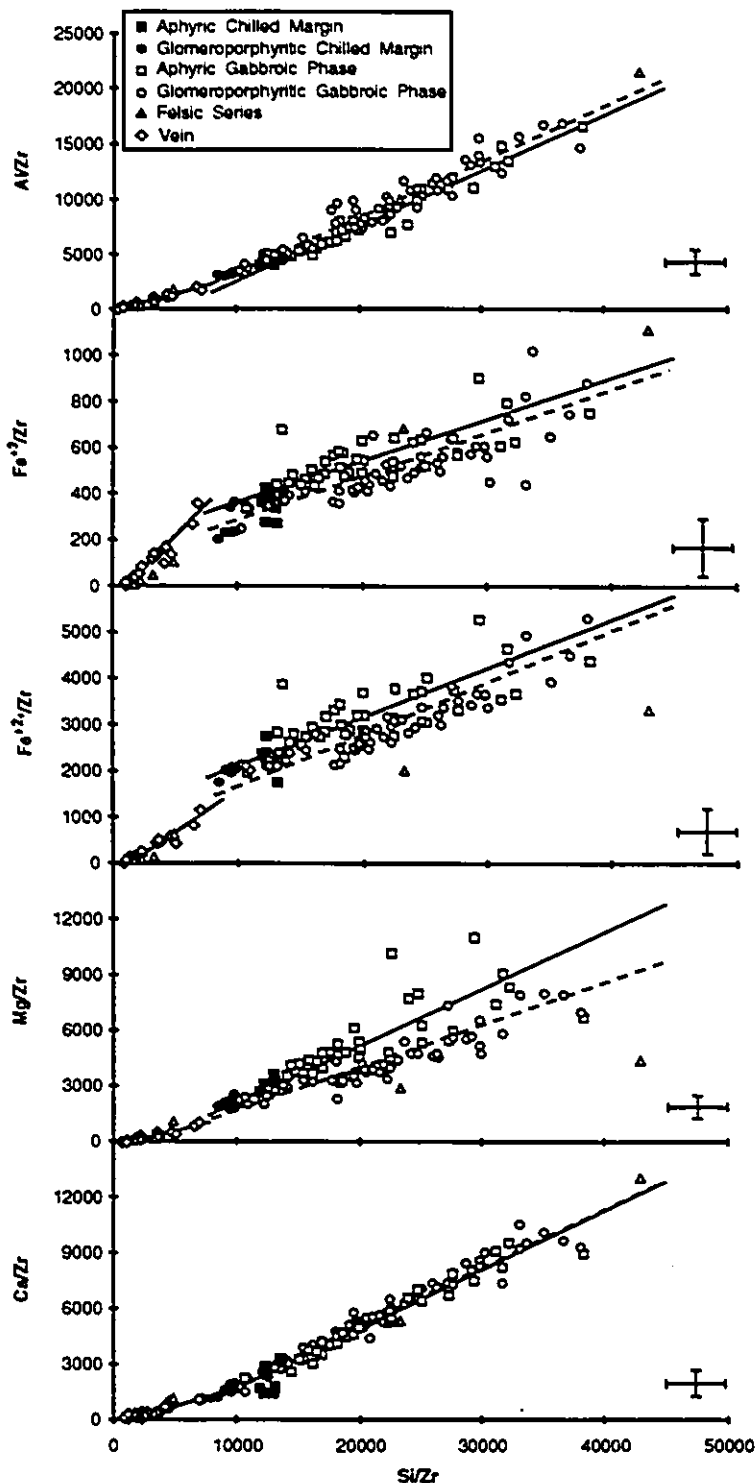


Figure 5.16. Pearce ratio diagrams of major elements versus Si using Zr as the denominator element for gabbroic phase, chilled margin and felsic series samples from aphyric and glomeroporphyritic sills. Average 2 σ analytical uncertainty is indicated by error bars. Dotted regression line = glomeroporphyritic samples, solid regression line = aphyric samples. Plot statistics are given in Table 5.10.

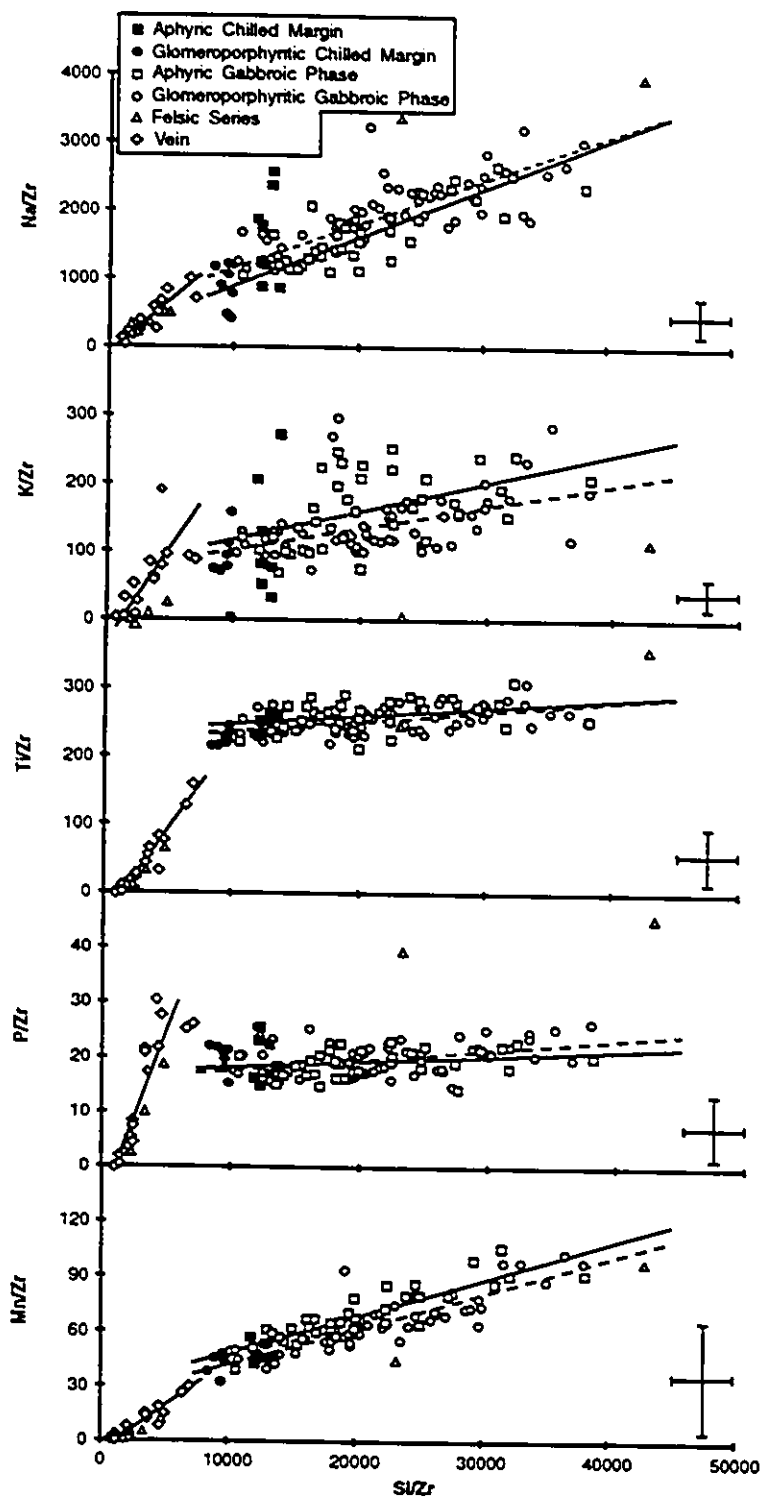


Figure 5.16 (continued). Pearce ratio diagrams of major elements versus Si using Zr as the denominator element for gabbroic phase, chilled margin and felsic series samples from aphyric and glomeroporphyritic sills. Average 2σ analytical uncertainty is indicated by error bars. Dotted regression line = glomeroporphyritic samples, solid regression line = aphyric samples. Plot statistics are given in Table 5.10.

the olivine found in sill C has influenced the slopes of the regressions in the Fe^{2+} and Mg versus silica plots. If the olivine predates sill emplacement, as was suggested in the previous section, its composition should be more primitive (i.e., richer in Mg) than pyroxene formed *in situ*. Since sill C samples comprise a large proportion of the aphyric gabbro sample suite illustrated in Figure 5.16, the more primitive character of olivine could result in the observed differences. In either case, the overall similarity of trends in Figure 5.16 is indicative of similar differentiation histories for the two types of sill, and this supports a comagmatic origin.

From the foregoing, it is apparent that samples from all sills examined, irrespective of glomerocryst content, formed by similar crystallization process acting on similar initial melts. These facts, coupled with the spatial juxtaposition of sill types and the ubiquitous nature of glomerocrysts, indicate that the initial magmas were derived from a common source. This interpretation is supported by the Sm-Nd isotopic composition of the sills. As described earlier, the initial ϵ_{Nd} ratios of a glomeroporphyritic and an aphyric sill at 1884 Ma are +3.5 and +3.2, respectively (see Table 9.2). These values indicate that the isotopic composition of the source regions to the melts of the two intrusions was essentially identical.

While the two types of intrusions have been derived from similar melts, a comparison of chilled margin samples from aphyric and glomeroporphyritic sills reveals minor geochemical differences. Specifically, the aphyric chilled margins have slightly lower concentrations of incompatible elements (P_2O_5 , TiO_2 , MnO, Zr and Y) and slightly higher concentrations of total iron ($\text{Fe}_2\text{O}_3\text{T}$), MgO, Cr, and Ni (Fig. 5.17). The differences in total iron and MgO may be sufficient to explain the differences in the trends of Fe^{3+}/Zr , Fe^{2+}/Zr and Mg/Zr versus Si/Zr for the data illustrated in Figure 5.16. The aphyric chilled margin samples may also contain slightly higher CaO and lower Na_2O , but evaluation of these differences is impaired by the loss of CaO and gain of Na_2O during albitization.

Finally, the rare-earth element (REE) patterns of aphyric and glomeroporphyritic chilled margins are similar but, with the exception of Eu, the glomeroporphyritic samples show slightly elevated REE concentrations (Fig. 5.18). The lower average Eu content (and correspondingly greater negative Eu anomaly) suggests that the two magma types may be

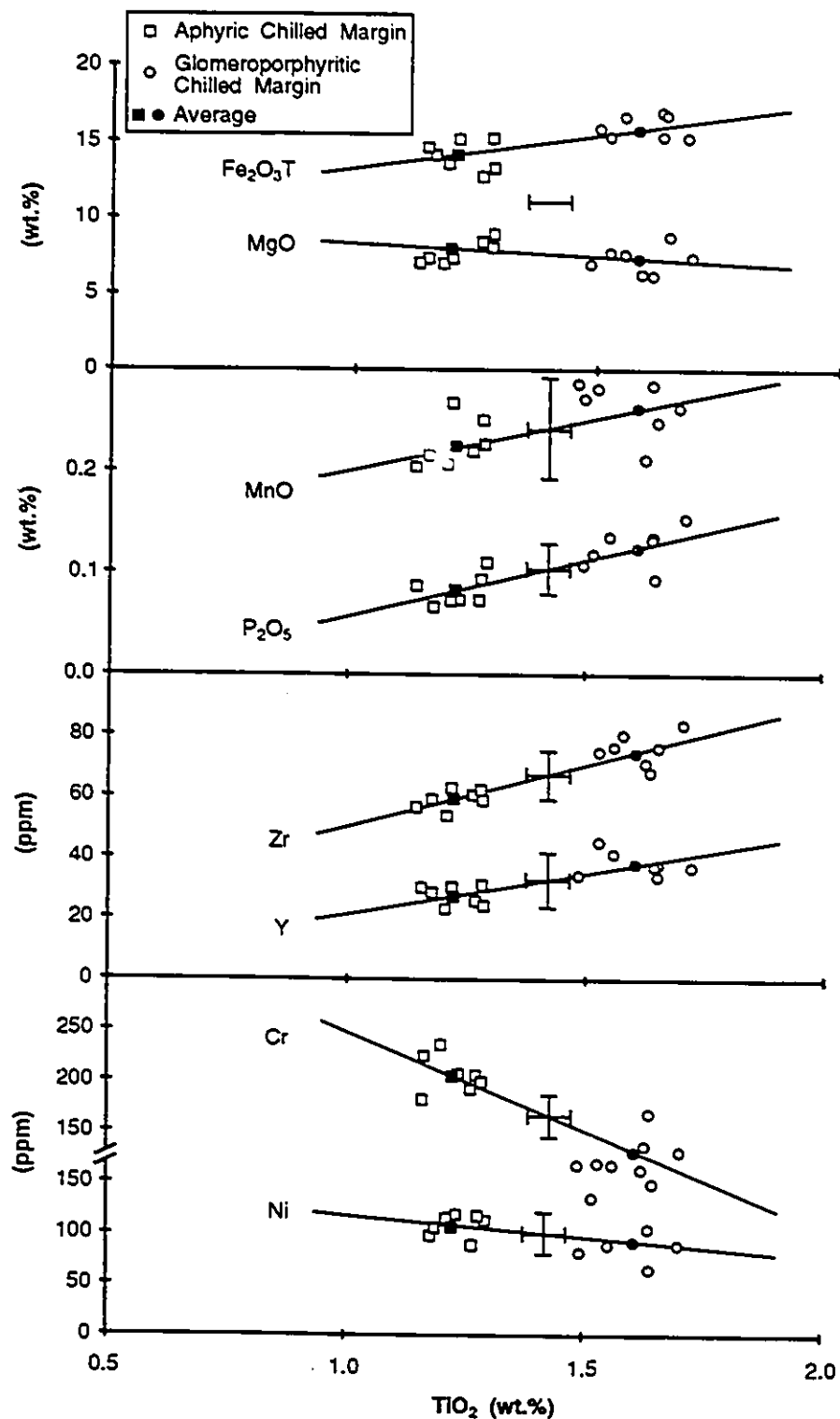


Figure 5.17. Selected elements plotted against TiO_2 for chilled margin samples from aphyric and glomeroporphyritic sills illustrating geochemical differences between the two suites. Mean values for the two groups are also illustrated. Average 2σ analytical uncertainties are indicated by error bars where greater than the size of plot symbols.

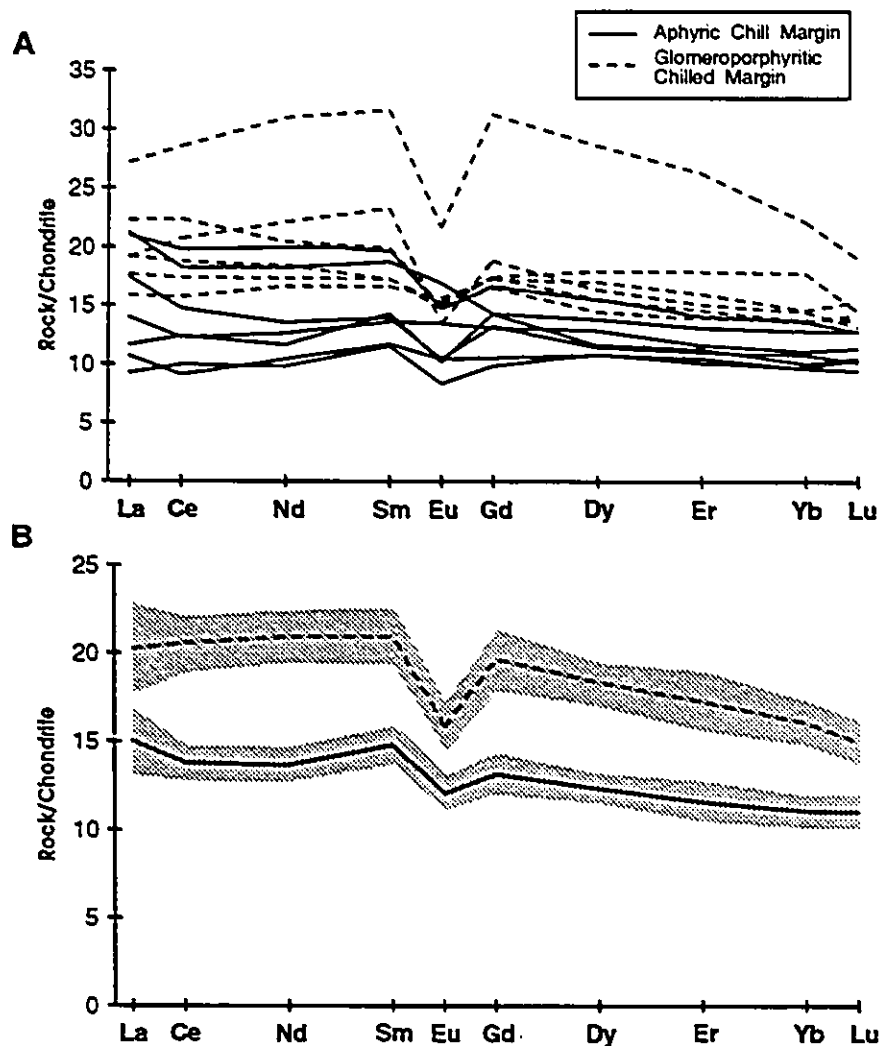


Figure 5.18. Chondrite-normalized REE plots of chilled margin samples from aphyric and glomeroporphyritic sills (A) and average values for each sill type (B). Average 2 σ analytical uncertainties are illustrated by shaded envelopes in (B). Note that the scale of the ordinate axis is arithmetic rather than logarithmic. Normalizing values are from Nakamura (1974).

related by plagioclase fractionation. The overall differences in chilled margin compositions indicate that the highly glomeroporphyritic melts were slightly more evolved than their aphyric to sparsely glomeroporphyritic counterparts. If the interpretations regarding the origins of olivine in sill C, and glomerocrysts in all of the sills, are correct, the more evolved nature of highly glomeroporphyritic melts probably reflects fractional

crystallization in the source reservoir. This possibility is examined in Chapter 8, where the relationships between Menihek gabbros and basalts are discussed.

5.7 Origin of the Felsic Rocks and Veins

In the discussions of crystallization in sills B and C, it was noted that felsic rocks ranging from quartz gabbro to granite comprise small volumes of the sills. In addition, most sills contain conformable to disconformable bands and veins of highly leucocratic material. The relationship of the felsic horizons and veins to the predominantly gabbroic host sill, however, is obscured by the absence of demonstrably conserved elements common to both felsic and gabbroic lithologies. Thus it is difficult to establish whether the felsic rocks represent the end products of fractional crystallization within the sills, if they reflect local assimilation of siliceous country rock material, or if other processes were involved.

Felsic Rocks

The aphyric and glomeroporphyritic Menihek gabbros of the Howse Lake area are interpreted to have been derived from the same source and to have undergone similar *in situ* differentiation during solidification. Based on mineralogical and geochemical similarities it is reasonable to assume that the felsic rocks which form irregular zones and lenses in the upper parts of many sills also have a common origin, and therefore can be treated as a related suite of rocks. Three lines of evidence suggest that fractional crystallization rather than assimilation was the predominant process leading to the felsic suite. First, there is no evidence in the geochemistry of the gabbroic rocks that bulk assimilation of country rock has occurred. Incompatible element ratios in chilled margins and cumulate gabbros, as well as the comagmatic basalts of the Menihek Formation (see Chapter 8), are constant. This fact precludes bulk assimilation of country rock during crystallization of the main portion of the sills.

The second line of evidence comes from the range of compositions of the felsic rocks at Howse Lake. Although there are compositional gaps between the gabbroic and felsic rocks in individual sills, this is not the case if all of the more evolved samples are

considered together. Since the different types of sills are thought to be comagmatic, with similar crystallization histories, the total compositional range of felsic rocks should approximate the compositional range which would be found in individual sills given sufficient exposure. Figure 5.19 presents plots of major element versus Zr for samples from the gabbroic phase, chilled margins, felsic series and veins of Howse Lake sills. The continuous nature of compositional variation is best illustrated by the changes in the SiO_2 , Al_2O_3 , $\text{Fe}_2\text{O}_3\text{T}$, MgO and TiO_2 content as Zr increases from chilled gabbro to granitic rocks (highest Zr). Similar, though less well defined, trends are discernible in the other elements, but these appear to have been affected by other processes. The somewhat discontinuous variation in CaO , Na_2O and K_2O , for example, probably reflects the mobility of these elements during alteration. Note that the gabbroic phase exhibits lower Zr content than the chilled margin because it is a combination of Zr-free cumulus phases and trapped liquid.

The third factor which argues against country rock assimilation derives from differences in the compositions of the felsic rocks and the compositions of rocks which could be developed by single stage mixing of gabbroic melts and Menihek sediments. The compositions of three sedimentary rocks from the Menihek Formation (from Gebert, 1988) are given in Table 5.11 and are plotted along with samples from the Menihek sills in Figure 5.20. The sedimentary samples, pyritic argillite, siltstone and arkose, represent end member compositions in the Menihek sequence, and hence should encompass the range of compositions available as contaminants. For some major elements, including SiO_2 , Al_2O_3 , $\text{Fe}_2\text{O}_3\text{T}$ and MgO (Fig. 5.20), derivation of the felsic suite through addition of sedimentary material to melts having the compositions of the chilled margins is possible. For other elements, however, concentrations in the felsic series are inconsistent with a simple mixing model. The high concentrations of P_2O_5 , Y and Zr in the felsic suite (Fig. 5.20), for example, are inconsistent with bulk melting of country rock, and instead would necessitate extensive preferential assimilation of minerals such as apatite and zircon by relatively small volumes of basaltic melt. The latter possibility is difficult to reconcile with the REE profiles of the felsic rocks (Fig. 5.21), which are relatively flat with concentrations ranging from 2 to 10 times the abundance of these elements in chilled margin samples. The REE patterns are similar to those of the chilled margins except for small light REE

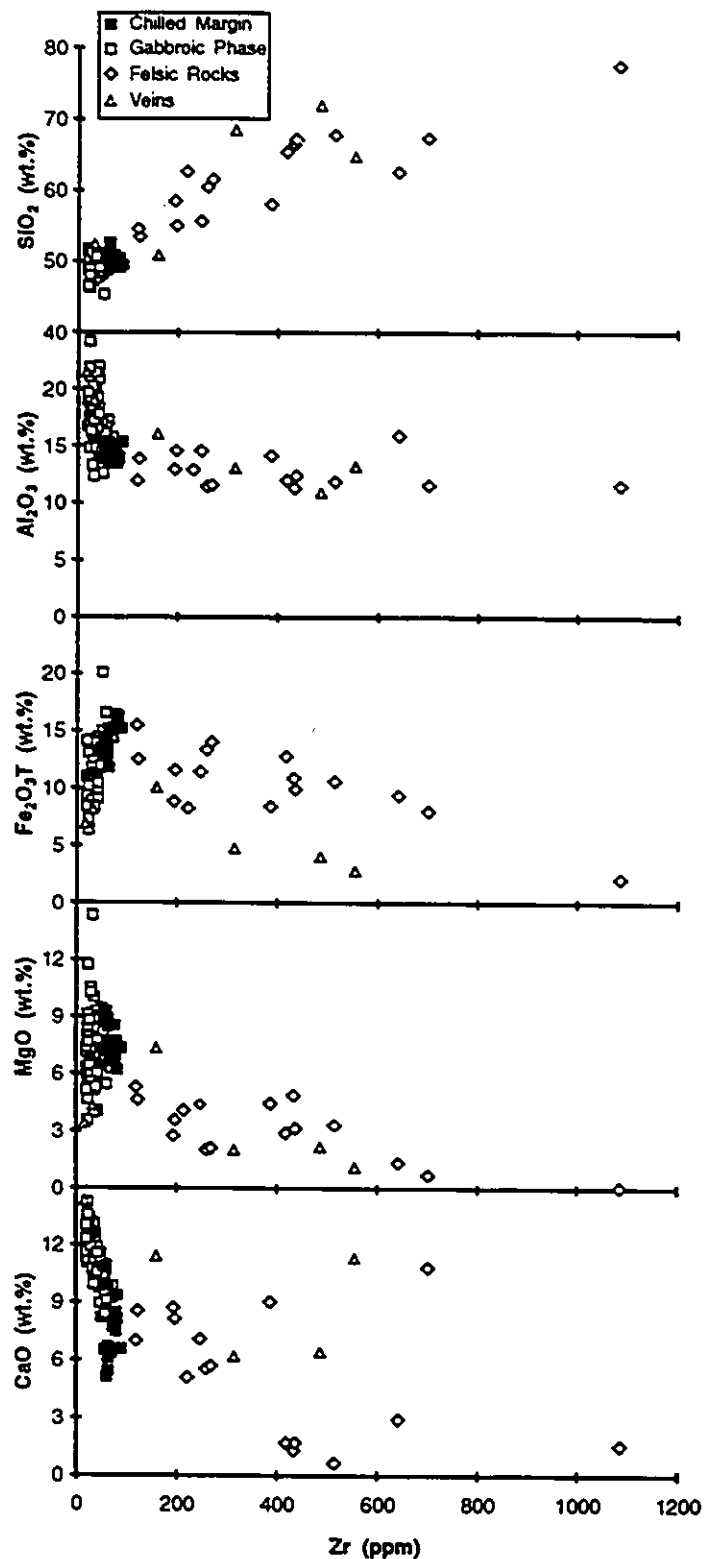


Figure 5.19. Major element compositions of the gabbroic phase, chilled margins, felsic rocks and veins from the Menihek sills plotted against Zr. For most elements there is a continuous variation in concentration from chilled margins to the most evolved compositions.

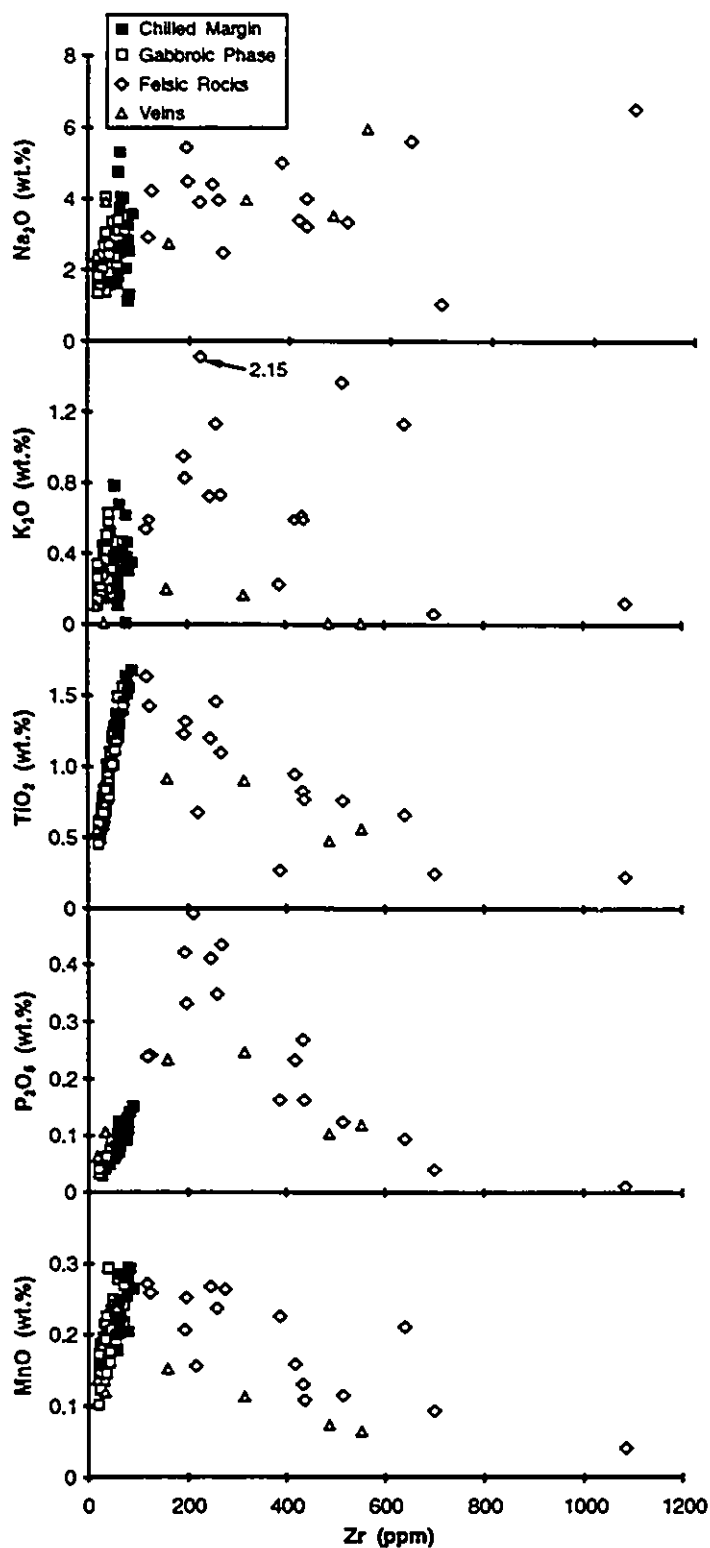


Figure 5.19 (continued). Major element compositions of the gabbroic phase, chilled margins, felsic rocks and veins from the Menihek sills plotted against Zr. For most elements there is a continuous variation in concentration from chilled margins to the most evolved compositions.

Table 5.11. Comparison of the compositions of a Howse Lake granitic differentiate with representative analyses of Menihek sedimentary rocks (from Gebert, 1987)

	F90-60C(1)	Argillite	Siltstone	Arkose
SiO ₂	77.59	43.40	59.80	74.80
Al ₂ O ₃	11.60	6.00	15.40	14.00
Fe ₂ O ₃ T	2.10	37.50	12.70	3.90
MgO	0.08	3.77	2.38	1.62
CaO	1.53	1.50	0.31	nd
Na ₂ O	6.50	0.79	2.20	0.13
K ₂ O	0.12	0.06	3.75	3.64
TiO ₂	0.23	0.57	0.37	0.40
P ₂ O ₅	0.01	0.12	0.10	0.04
MnO	0.04	0.28	0.05	nd
Ba	60	17	443	703
Cr	13	nd	nd	nd
Zr	1085	77	87	240
Sr	129	12	15	10
Rb	0	6	100	120
Y	242	17	13	13
Nb	4	13	4	nd
Zn	0	50	195	21
Ni	29	61	90	14
V	0	212	297	70

impoverishments and slightly greater negative Eu anomalies. Since apatite preferentially incorporates the light REE, preferential assimilation of apatite should produce a light REE enrichment rather than impoverishment in the felsic samples.

In light of these considerations, it is likely that the felsic suite of the Howse Lake sills originated primarily through fractional crystallization of the gabbroic melts. If the felsic rocks represent residual liquids derived by fractional crystallization, cumulate phases other than the principal silicates must have contributed to melt evolution to produce the observed trace element compositions, and to explain the lack of conservation of Ti, P, Mn, Zr and Y. In sill B it is apparent that Fe-Ti oxides joined pyroxene and plagioclase as cumulate phases in the upper part of the sill. In addition, the uppermost sample of the oxide horizon contains abundant apatite, which is also thought to be a cumulate phase. Other sills in the Howse Lake area contain cumulate oxide horizons, and it is apparent that

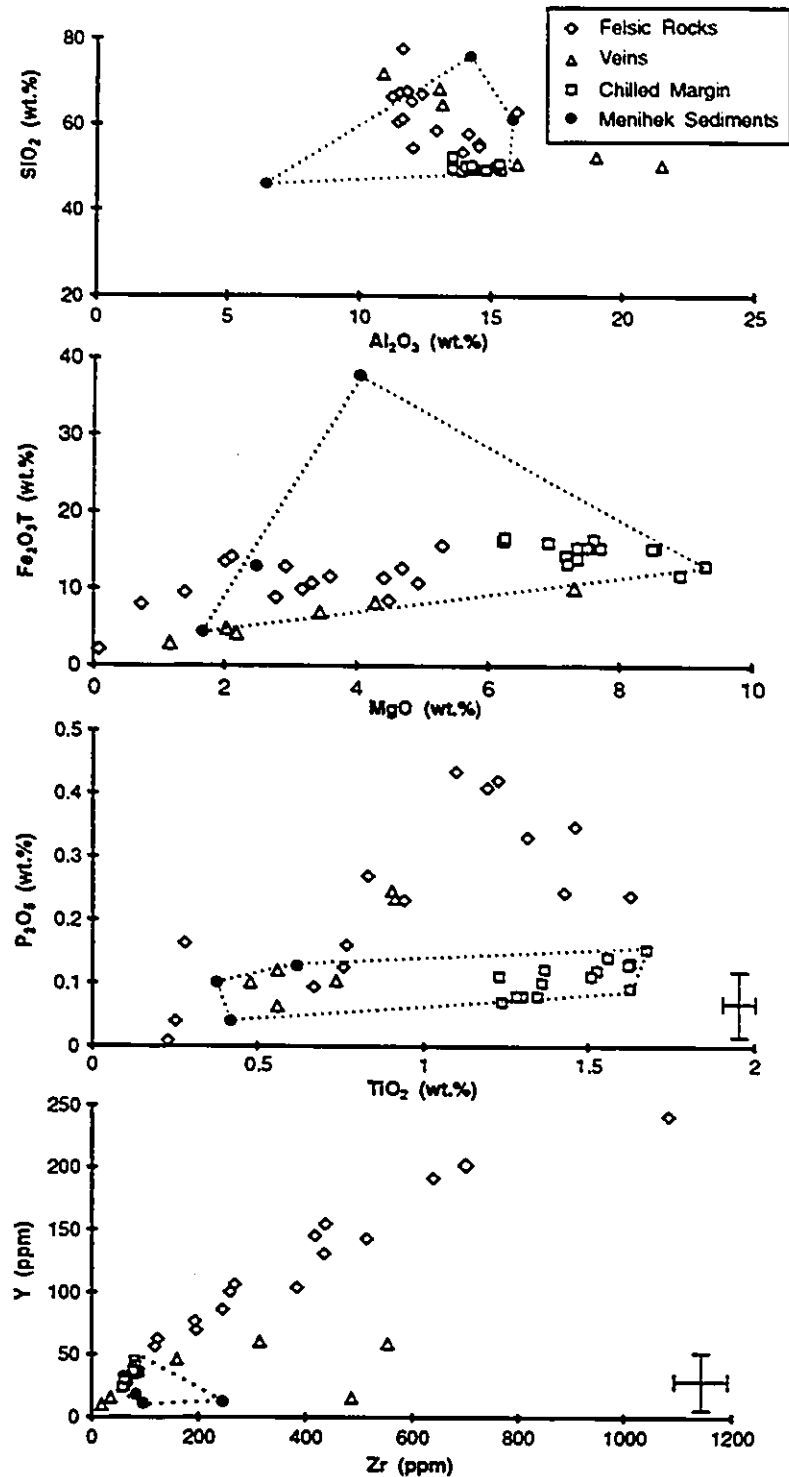


Figure 5.20. Concentrations of selected elements in chilled margins, felsic rocks, and veins from the Menihek sills, along with those of sedimentary rocks of the Menihek Formation. The dotted lines outline the compositional fields of rocks which could be derived from single stage mixing of chilled margin and sedimentary compositions. Average 2σ analytical uncertainties are indicated by error bars.

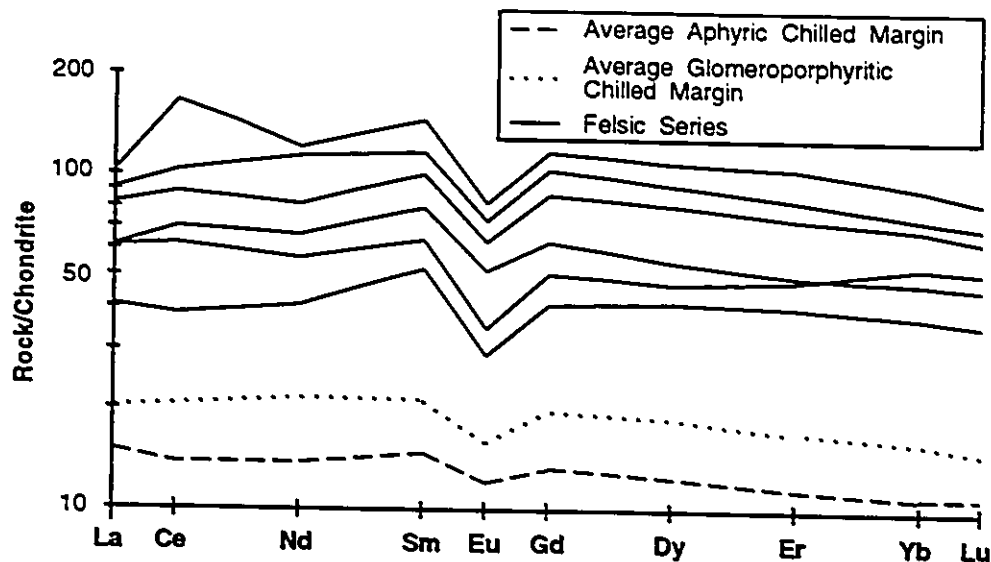


Figure 5.21. Comparison of the chondrite-normalized REE contents of the felsic series and average glomeroporphyritic and aphyric chilled margins. Normalizing values are from Nakamura (1974)

the appearance of these phases on the liquidus was a normal stage in melt evolution. Residual liquids would be depleted in the principal components of oxides and apatite, particularly Fe_2O_3 , TiO_2 , and P_2O_5 , which are relatively minor components in basaltic magmas. Trace elements with high liquid-oxide and liquid-apatite partition coefficients would also be removed (i.e., V, Y and the REE). In contrast, the silica content of residual melts would increase due to the crystallization of non-silicate phases.

The variations in the TiO_2 , P_2O_5 , Y and La contents of the felsic suite are illustrated in Figure 5.22 using Zr as a monitor of melt evolution. In Figure 5.22A, crystallization of Fe-Ti oxides impoverishes residual liquids in TiO_2 , thereby driving the melts away from the constant Ti/Zr ratio of the silicate cumulates. Because TiO_2 is a major component of the oxides, the residual concentrations decline rapidly. A similar process occurs when apatite becomes a liquidus phase (Fig. 5.22B), with residual melts successively depleted in P_2O_5 . Y and the REE are considerably more compatible in apatite than the other cumulus minerals, and thus precipitation of this phase also causes depletion of these elements relative to Zr in the residual liquids. Unlike P_2O_5 , however, Y and the REE are only minor components in apatite and therefore the decline in their concentrations in residual melts is more gradational. The departure from conservation of Y and La is depicted in Figures

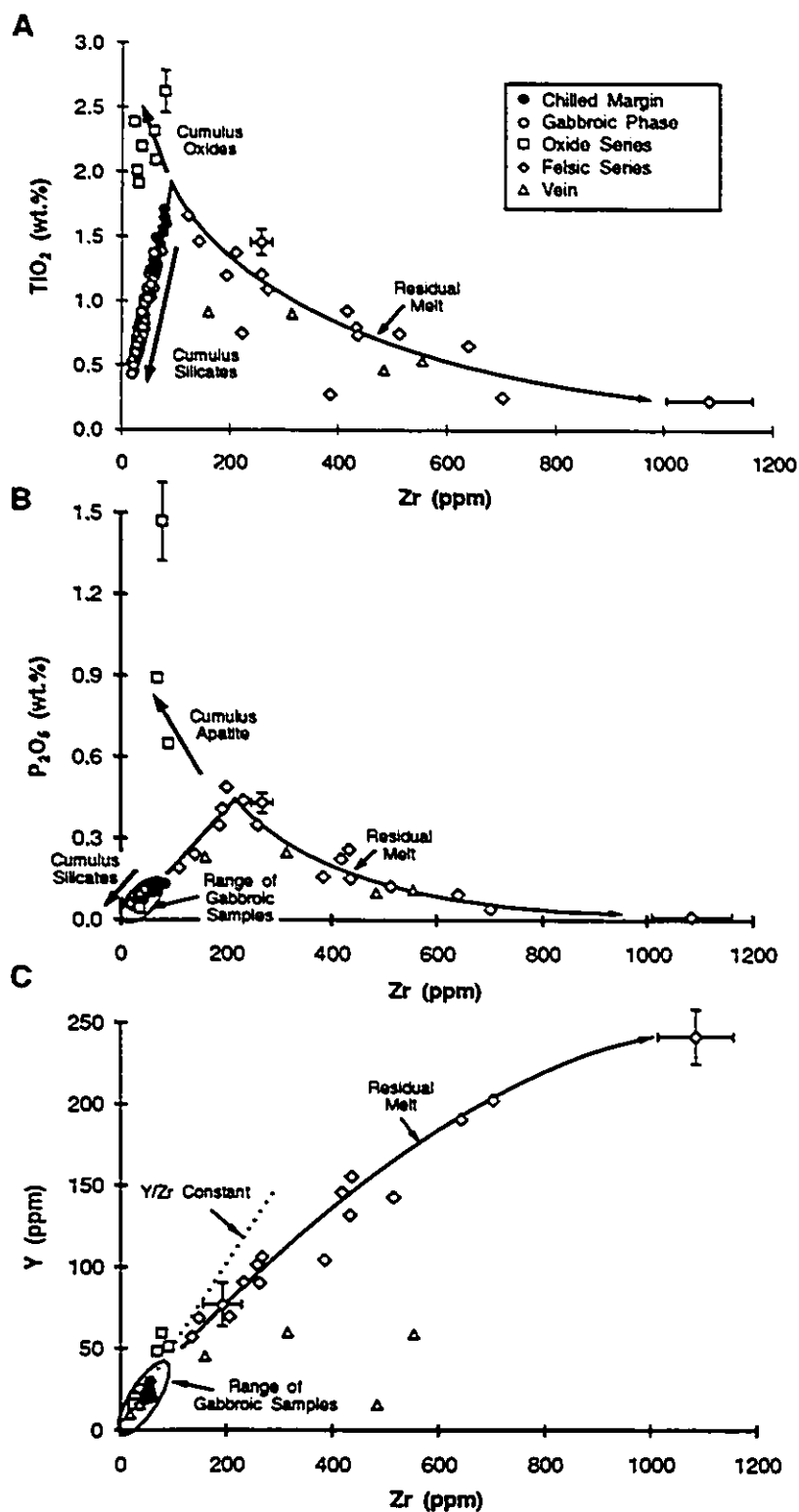


Figure 5.22. Variation in selected elements with increasing Zr in the felsic series. (A) TiO_2 versus Zr, (B) P_2O_5 versus Zr and (C) Y versus Zr. Maximum and median 2σ analytical uncertainties are indicated by error bars.

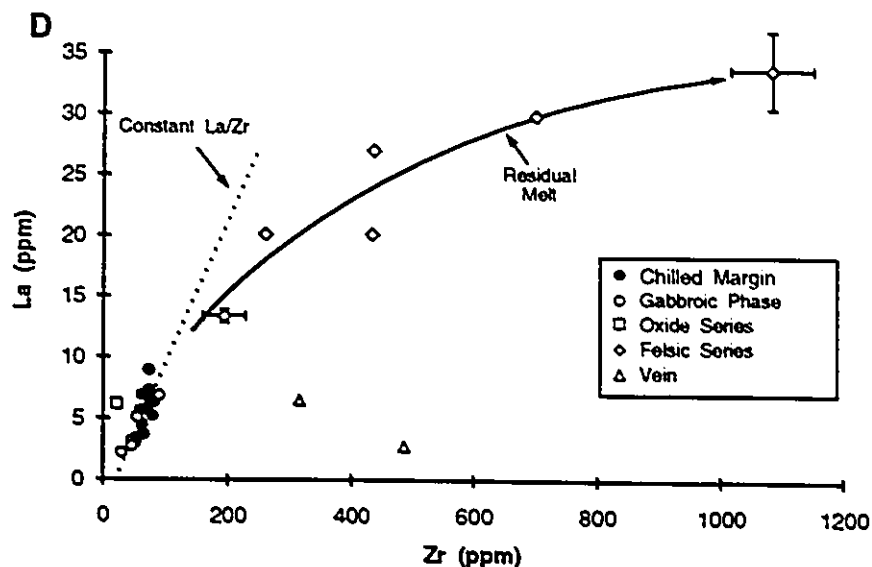


Figure 5.22 (continued). Variation in selected elements with increasing Zr in the felsic series. (D) La versus Zr. Maximum and median 2σ analytical uncertainties are indicated by error bars.

5.22C and D, and it is apparent that with increasing fractional crystallization (as measured by Zr content), the Y/Zr and La/Zr ratios of the felsic rocks gradually decline.

Figure 5.22 indicates that Zr is the only element from the group used in previous analysis which may have remained conserved during the final stages of crystallization. This element was used as the denominator in the Pearce ratio plots of Figure 5.16, in which variation in the major element content of the felsic suite is also depicted. Although the possible non-conservation of Zr and the complexity of the mineralogy of the felsic samples precludes detailed analysis of differentiation in the series, several general conclusions can be inferred from the diagrams. A comparison of the regression lines in the plots of Na/Zr and K/Zr indicates that these numerator elements were involved in the differentiation process to a greater extent in the felsic suite than in the gabbroic rocks (as indicated by the greater slopes for the felsic versus gabbroic data sets). The flatter slopes of the felsic series regressions in the Al/Zr and Ca/Zr diagrams suggest that the change in importance of Na and K was brought about by the replacement of calcic plagioclase in the cumulate by sodic plagioclase and orthoclase (both having lower Al contents), and by the crystallization of quartz. These changes are consistent with the mineralogy of the felsic samples. The trends in the Fe^{+3} , Fe^{+2*} , Ti and P contents of the felsic rocks must in part

reflect the fractionation of Fe-Ti oxides and apatite in the oxide horizons. It is clear, however, that some felsic samples contain cumulus apatite (P/Zr ratios higher than chilled margin samples) and thus this phase remained on the liquidus subsequent to formation of the ferrogabbros. In contrast, no felsic samples exhibit Fe^{3+}/Zr and Ti/Zr ratios in excess of chilled margins and this indicates that crystallization of the oxide zones sufficiently impoverished residual liquids in Fe^{3+} and Ti to remove oxide minerals from the liquidus. The steeper slopes of the regressions of these elements, however, suggests that they were incorporated into a cumulus phase in greater proportions than in the gabbroic suite. This would be consistent with the replacement of pyroxene by iron-rich amphibole in the cumulate, and a similar substitution is probably responsible for the increased slope of the felsic suite in the Mn/Zr versus Si/Zr plot. The loss of pyroxene is confirmed by the mineralogy of the suite where hornblende is the predominant mafic silicate in most felsic samples. A decline in the Mg/Fe^{2+} ratio of mafic silicates was noted in the analysis of the oxide zone samples from sill B, and the small slope of the felsic regression in the plot of Mg/Zr versus Si/Zr indicates that the mafic silicates in the felsic series were also Mg-poor.

The foregoing analysis indicates that fractional crystallization in the sills was largely responsible for the development of siliceous differentiates. While some degree of country rock assimilation is possible, the predominant process behind compositional variation was the precipitation and accumulation of plagioclase, augite and hypersthene, followed by Fe-Ti oxides and apatite. Residual liquids continued to evolved past the ferrogabbro stage, probably through precipitation of sodic plagioclase, orthoclase and amphibole.

Veins

In addition to the felsic differentiates, leucocratic rocks occur as relatively thick (up to 1 m) conformable bands and narrow (<1 cm to 8 cm) discordant veins in the Howse Lake sills. The compositions of 6 samples from these lithologies were determined and have been included in the plots of Figures 5.16, 5.19, 5.20 and 5.22. An examination of these diagrams indicates that, while four of the samples show some compositional similarities to the felsic suite and the remaining two show some similarities to gabbroic phase rocks, in detail the veins are compositionally distinct. For example, the vein and felsic series

samples appear to share the same pattern of variation in TiO_2 and P_2O_5 as Zr increases (Fig. 5.22), but the vein samples depart from the felsic series trend in the variation of Y and La.

Perhaps the most obvious compositional difference, however, lies in the REE contents of the veins (Fig. 5.23). In comparison to other lithologies, the REE patterns of the veins feature strong positive Eu anomalies and heavy REE enrichment. These characteristics contrast markedly with the relatively flat patterns of the chilled margins, and the light REE depletion and strong negative Eu anomalies of the felsic series (cf. Fig. 5.21). In terms of concentration, the REE content of the veins are closer to that of the chilled margins than the felsic suite. The Eu enriched patterns are unlikely to develop during fractional crystallization of a plagioclase-rich cumulate, but are typical of rocks in which plagioclase has broken down during rock alteration (Fowler and Doig, 1983). The compositional differences between the veins and other lithologies, coupled with the hydrous mineralogy, suggest that the veins are unrelated to the differentiation mechanism which produced the other sample series.

Given the compositional and mineralogical characteristics, the veins are thought to reflect the effects of volatile enriched fluids acting on originally gabbroic rocks. A Pb-Pb

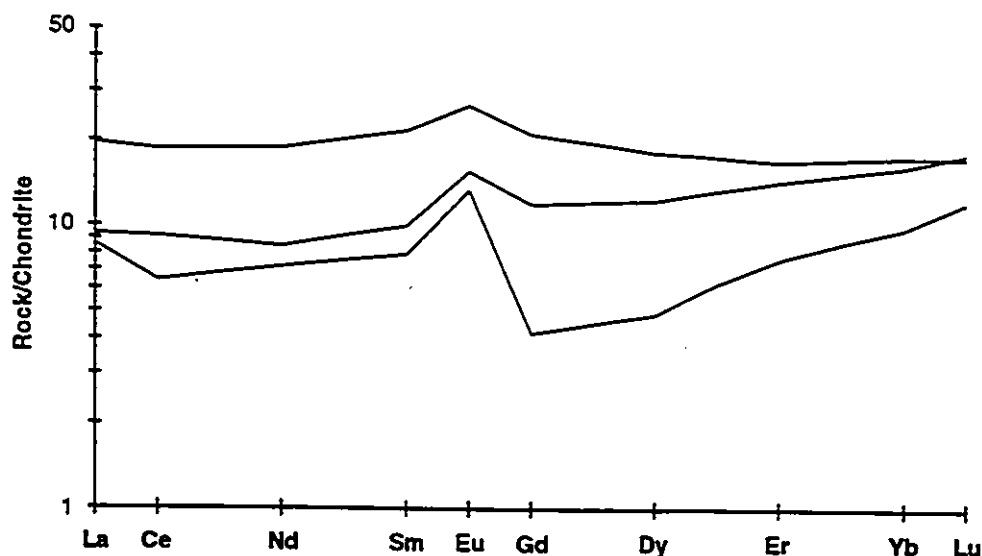


Figure 5.23. Chondrite-normalized REE contents of veins in the Menihek sills. Note the positive Eu anomalies. Normalizing values from Nakamura (1974)

isotopic age of 1793 Ma obtained from galena in a Howse Lake vein by Clark and Thorpe (1990) indicates that at least some of the veins are epigenetic. In light of this date, it is probable that some of the fluids were metamorphic in origin. Other veins, however, may be the products of late stage, hydrous magmatic liquids. The contacts between the veins and enclosing rocks are sharp but gradational. This suggests that the veins are the conduits through which magmatic and metamorphic fluids passed, altering the original gabbroic mineralogy and precipitating quartz and other phases. This interpretation is supported by the occasional presence of highly sericitized glomerocrysts in the veins. It also explains the fact that some vein samples are geochemically similar to the gabbroic rocks, while others are closer to the compositions of the felsic suite.

5.8 Summary

Aphyric and glomeroporphyritic Menihek sills at Howse Lake share geochemical and isotopic similarities which indicate that they were derived from the same magma source. Both types of sill have undergone *in situ* differentiation by fractional crystallization of plagioclase, clinopyroxene and orthopyroxene. Accumulation of fractionated phases occurred at the site of crystallization, and did not involve transport of cumulus grains. Incompatible element profiles indicate that crystallization proceeded inward from the walls of the sills (congelation crystallization). The absence of cryptic compositional variation and the apparent constancy of incompatible element concentrations throughout much of the sill stratigraphies suggest that fractional crystallization had relatively minor effect on residual melt compositions until late in their crystallization histories. The appearance of cumulus Fe-Ti oxides and apatite, and the occurrence of felsic rocks, indicate that the magmas eventually evolved to more extreme compositions. This led to precipitation of the oxides and apatite, and eventually to the formation of small volumes of dioritic to granitic rocks. The interactions of late stage magmatic fluids, as well as metamorphic fluids, produced leucocratic veins during passage through the intrusions.

Lithological variations in individual sills in part reflect *in situ* crystallization (i.e., gabbroic phase, oxide horizons, felsic rocks), and in part reflect the concentration of pre-emplacement crystallization products. The olivine-bearing rocks of sill C are thought

to reflect the concentration of pre-emplacment olivine in the centre of the sill by flow differentiation. Similarly, the anorthositic rocks which occur in many sills are thought to reflect the addition of pre-emplacment plagioclase glomerocrysts to otherwise gabbroic melts. The only lithologies that do not seem to have developed through fractional crystallization or flow differentiation are the sulphide enriched horizons. The isotopic composition of sulphur in these rocks suggests that they are the result of assimilation of country rock sulphur.

In addition to glomerocryst content, aphyric and glomeroporphyritic sills are distinguished by small geochemical differences in chilled margin compositions. Aphyric chilled margins are slightly enriched in MgO, Cr and Ni, and slightly impoverished in $\text{Fe}_2\text{O}_3\text{T}$ and incompatible trace elements. These differences are thought to reflect derivation of glomeroporphyritic melts from the source reservoir after fractional crystallization of silicate minerals. Pre-emplacment magmatic differentiation is discussed further in Chapter 8.

Chapter 6

Origin of Glomerocryst Zonation in the Menihek Gabbros

6.1 Nature of the Data Set

Glomeroporphyritic sills are conspicuous members of the stratigraphy of the Howse zone. The sills are zoned with respect to glomerocryst content, and this zonation reflects either the coincident introduction of magmas with varying glomerocryst contents, or the operation of a mechanical process acting on the glomerocryst-laden magmas during emplacement. To investigate the origin of the zonation, the glomerocryst distributions were determined in the four sills sampled in detail in the Howse Lake area. The glomerocryst distributions were obtained by image analysis of serial sections cut from samples from the four profiles. Four parameters were measured in order to define the glomerocryst distributions. These were: (1) the total area occupied by the glomerocrysts relative to groundmass, (2) the size of the individual glomerocrysts, (3) the number of glomerocrysts in unit area, and (4) the irregularity of the glomerocrysts, where irregularity is the ratio of the square of the perimeter to the area. The perimeter is squared to provide a dimensionless parameter. The irregularity of a circle is therefore 4π , while the irregularity of non-circular objects increases with increasing perimeter length relative to area.

With the possible exception of proportional area as determined by frequency measurements (cf. Chayes, 1950), there are significant difficulties involved in the transformation of two dimensional measurements into three dimensional properties. Transformative functions have been developed for uniform spheres (cf. Cashman, 1990), but since glomerocrysts in the Menihek sills are often irregular in shape the validity of these functions with respect to glomerocryst data is questionable. It is the variations in glomerocryst parameters across a given sill, however, that are of interest. Consequently, variations in three dimensional parameters are adequately approximated by variations in the two dimensional parameters as measured on the surfaces of serial sections cut from randomly oriented samples. For example, the estimated variations in mean glomerocryst

size as measured by area will reflect the true variations in mean glomerocryst size as measured by volume. This follows from the fact that the area occupied by a glomerocryst on a given plane is a function of the volume of the glomerocryst as well as the position (and angle, if non-spherical) of the plane relative to the bisectrix of the glomerocryst. The measured area cannot be interpreted in terms of volume without some type of transformational function because the probability of a plane intersecting a glomerocryst at some distance from the bisectrix is greater than the probability of symmetrical intersection. Hence maximum area will be sampled less often than smaller areas. The variations in mean area, however, will reflect variations in mean volume because the intersection probabilities will not change from one sample to the next.

6.2 Generalized Glomerocryst Distribution

The Menihek glomeroporphyritic gabbros of the Howse zone are typically zoned with respect to the quantity and size of plagioclase glomerocrysts. Because the aggregates pre-date emplacement (see Chapter 4), the systematic nature of the glomerocryst distributions and the ubiquity of the zonation pattern reflect the influence of some process other than *in situ* crystallization. The general pattern of zonation is one in which the highest concentrations of glomerocrysts occur in the central regions of the intrusions. Marginal zones contain fewer aggregates or are essentially aphyric. In addition, there is an increase in the size of glomerocrysts from sparsely glomeroporphyritic marginal zones to the more strongly glomeroporphyritic central regions. The variations in glomerocryst size are most apparent in highly glomeroporphyritic sills containing the anorthositic and core phases due to the large range in glomerocryst size in these lithologies. To some degree, there also appears to be an increase in the regularity of shape of the glomerocrysts with increasing size. Thus the aggregates of the gabbroic phase tend to be irregularly shaped, while the large glomerocrysts of the core phase appear to be relatively spherical (see Fig. 4.6). In highly glomeroporphyritic sills these size and concentration variations typically give rise to the development of interior zones composed of the anorthositic and core phase lithologies, with flanking zones are composed of the gabbroic phase. The pattern is also apparent in sparsely glomeroporphyritic intrusions, although the variation in glomerocryst

size is much less pronounced and there is no discernible variation in glomerocryst irregularity.

6.3 Potential Mechanical Differentiation Mechanisms

Systematic variations in the quantity and size of phenocrysts, glomerocrysts or xenoliths have been observed in a variety of plutonic rock types having a range of intrusive styles. Such variations may be symmetrical or asymmetrical relative to the boundaries of the plutonic body. A common pattern of asymmetric distribution involves the concentration of phenocrysts toward the upper or lower contacts of horizontal intrusions, as in the formation of dunitic cumulates through the gravitational accumulation of olivine grains on the floor of a basaltic sill. The sinking or floating of phenocrysts, however, cannot explain the distribution of such crystals in some igneous bodies. Symmetrical non-uniform phenocryst distributions in a sill, for example, cannot be produced by gravitational accumulation. Similarly, non-uniform phenocryst distribution across the width of a vertical dyke is unlikely to have developed through gravitational effects.

The Menihek glomeroporphyritic sills fall into the category of intrusive rocks in which the approximately symmetrical distribution of demonstrably pre-emplacement solids precludes an origin exclusively related to *in situ* crystallization or gravitational processes. This type of pattern has been observed in numerous intrusive bodies; examples include the Retty Peridotite of the Labrador Trough (Fahrig, 1962), leopard rock dykes of the Beartooth Mountains (Printz, 1964), the Muskox intrusion feeder dyke (Bhattacharji and Smith, 1964), and the picritic and doleritic dykes of the Isle of Skye (Drever and Johnston, 1967; Gibb, 1968; Platten and Watterson, 1987). The Menihek sills differ from most other examples in that they are glomeroporphyritic rather than porphyritic, and that the pre-emplacement phase is calcic plagioclase rather than olivine. Assuming *in situ* crystallization and gravitational processes can be excluded as the source of the glomerocryst distributions in the Menihek sills, the pattern of zonation may reflect multiple magmatic injection, wall crystallization, flow differentiation, or a combination of these mechanisms.

6.4 Multiple Magmatic Injection

Baragar (1960) suggested that some of the glomeroporphyritic sills in the Ahr Lake area are composite bodies reflecting multiple injection of contrasting magma types, and this appears to be the origin of the noritic layer which underlies the core phase in sill B at Howse Lake. The noritic layer is composed of a hypersthene-rich melanocratic rock which weathers to a chocolate coloured to reddish-brown surface in outcrop. These features, coupled with the complete absence of glomerocrysts, makes the lithology quite distinctive. When first encountered in sill B, the horizon was mapped as a pyroxenitic unit which was thought to be an intrinsic part of the sill due to the lack of chilled margins or other features indicative of an intrusive relationship. But the same distinctive lithology occurs with obvious cross-cutting relationships in other Howse Lake sills, and is clear that the noritic rocks, including the horizon in sill B, represent independent intrusions.

Although cross-cutting relationships between the norites and the glomeroporphyritic sills are more common than conformable ones, the horizon in sill B indicates that conformable emplacement of successive melts did in fact occur. It follows that conformable emplacement of aphyric or glomeroporphyritic gabbroic magmas into pre-existing sills may also have occurred. Geochemical data presented in Chapter 8, coupled with the observations of Baragar (1967) and Rohon *et al.* (1993) regarding the cyclicity of comagmatic basalt eruptions, indicate that glomeroporphyritic magmas were generated periodically over the course of cycle 2 tholeiitic magmatism. Thus it is possible that some of the structure of the Menihek sills reflects injection of glomeroporphyritic magmas into pre-existing aphyric intrusions, or vice-versa. The absence of intra-sill chilled margins might then reflect emplacement of the second injection prior to significant cooling of the first.

The process of multiple magmatic injection as described here is most consistent with two types of sill structures found in the Menihek gabbros. The first is the pattern reported by Baragar (1960) in which central aphyric gabbro zones are bounded by glomeroporphyritic lithologies. This type of structure may record the injection of aphyric magma into a pre-existing glomeroporphyritic sill. The second type of structure amenable to the multiple injection hypothesis is commonly observed in the Howse Lake intrusions and consists of marginal zones of aphyric or sparsely glomeroporphyritic gabbro enclosing

a central region of anorthositic phase gabbro. In both types of intrusion, the absence of obvious intra-sill intrusive contacts indicates that, if multiple injection was the source of the sill structure, successive magmatic injections were closely spaced in time. This is also suggested by the contrast in competency of solidified gabbroic rock and laminated turbidites; it is highly unlikely that ascending melts would preferentially invade solidified gabbro with a higher density than the melt, rather than exploiting the numerous planes of weakness in the less dense country rock. Preferential emplacement within pre-existing gabbro is even more improbable if the gabbro-turbidite assemblage of the Howse zone represents a Guaymas Basin-type sill-sediment complex, in which magmas were emplaced within unconsolidated or semi-consolidated sediments (cf. Einsele, 1986; see also Chapter 10). Consequently, the multiple injection hypothesis requires the emplacement of successive magma pulses within unsolidified or partially solidified sills. In this regard, the process is somewhat comparable to simple accretion as discussed below.

6.5 Wall Crystallization During Emplacement

Simple Accretion

A second potential process by which the glomerocryst zonation of the Menihek sills may have developed involves the solidification of magma at the walls of the conduit during passage of magmas. Harker (1904) first suggested this mechanism as the origin of variations in phenocryst content in dykes from the Isle of Skye. Delaney and Pollard (1982) have shown that accretion of magma along dyke walls will occur by crystallization in solid-liquid boundary layers during periods of flow as well as when the magma is stagnant. The thickness of the interface between accreted material and magma, and the rate of advance of this interface, are controlled by various parameters including rate of magma flow and conduit thickness (Platten and Watterson, 1987). Variations in phenocryst content could theoretically arise through accretion of magmas having variable crystal loads. For example, aphyric zones in a sill or dyke would reflect wall crystallization while the conduit was filled by aphyric magma, and porphyritic zones would reflect accretion from porphyritic melts. This process requires that the aphyric liquids were displaced from the

centre of the conduit prior to complete solidification. Although wall crystallization from variably crystal-charged magmas can produce zonation in terms of phenocryst content, the nature of the zonation pattern will be entirely controlled by the number and distribution of crystals in successive magmas. Thus a zonation pattern in which phenocryst content progressively increases or decreases across an intrusion must reflect either increasing or decreasing phenocryst content in successive melts, or the operation of another process

Differential Accretion

A second type of accretion mechanism, termed differential accretion by Platten and Watterson (1969), does not require changes in magma type to produce variations in phenocryst content. In the differential accretion model, the thermal gradient at the conduit margin is steep during periods of flow, and magma solidification is confined to a narrow boundary layer (Delaney and Pollard, 1982; Platten and Watterson, 1987). If the thickness of the solid-liquid interface is small in comparison to phenocryst dimensions, the crystals will be continually dragged along the interface, and the constant movement will prevent their incorporation into the accreted material (Fig. 6.1A). Reduction or cessation of flow leads to a reduction in the thermal gradients at the conduit margins, and a consequent expansion of the boundary layer. The combination of the broader crystallization zone and the reduction in particle movement will result in the incorporation of phenocrysts into the accreted rock in the boundary layer (Fig. 6.1B). Intermittent magma flow in the conduit could thus produce alternating aphyric and porphyritic zones. Gradational flow rate variations would produce diffuse contacts between aphyric and porphyritic horizons, while abrupt onset or cessation of flow would lead to sharp contacts.

The process of differential accretion is theoretically capable of producing three other lithological variations of interest in the present study. Since changes in flow rate will affect the thermal gradient and hence the thickness of the boundary layer, a variation in phenocryst size might accompany changes in glomerocryst content. Thus gradual flow rate reduction over a considerable time interval could give rise to a progression from aphyric rock, through sparsely porphyritic material containing small phenocrysts, to highly porphyritic lithologies with relatively large phenocrysts. The changing thermal regime would also affect the groundmass crystallization rate. High thermal gradients will promote

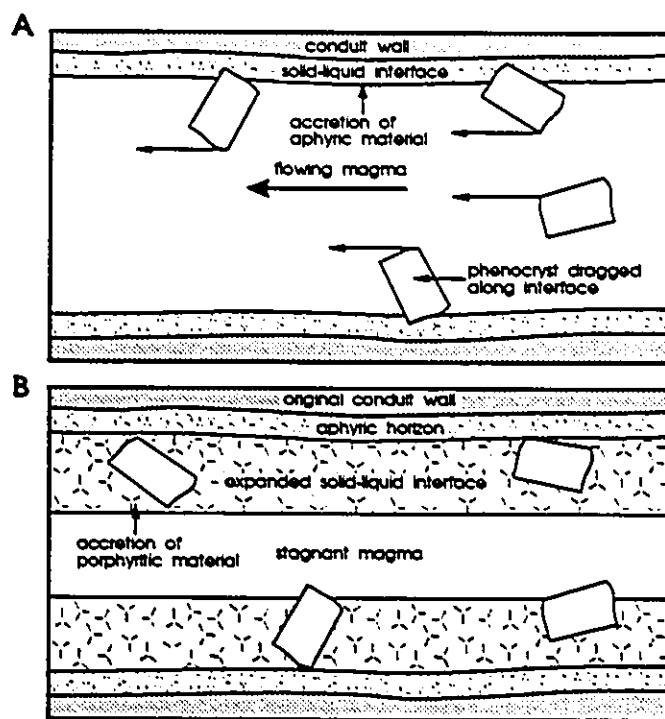


Figure 6.1. Development of contrasting lithological types by differential accretion during magma flow in a sill. (A) Relatively high flow rate leads to high thermal gradient at sill margins and development of narrow solid-liquid boundary layers. Boundary layer thickness is less than phenocryst dimensions and the crystals are dragged along the interface without being incorporated into the accreted material. (B) Reduction or cessation of flow leads to a lower thermal gradient and broader boundary layers. Phenocrysts are now incorporated into the accreted material.

rapid nucleation in the boundary layer, and limit the time available for crystal growth. This will result in fine grained accreted rock. Lower thermal gradients might reduce the degree of undercooling and nucleation rate, and increase the rate of crystallization, thereby promoting coarser grain sizes. Grain size layering and comb layering thought to have developed in this manner have been observed by Platten and Watterson (1987) in dolerite dykes from the Isle of Skye.

A third lithological variant that may be produced by differential accretion during flow consists of lenses of one lithological type within another. Since variations in conduit width will affect flow rate, a dyke or sill which pinches and swells along flow direction may develop boundary layers of varying thickness at different locations. Consequently, the

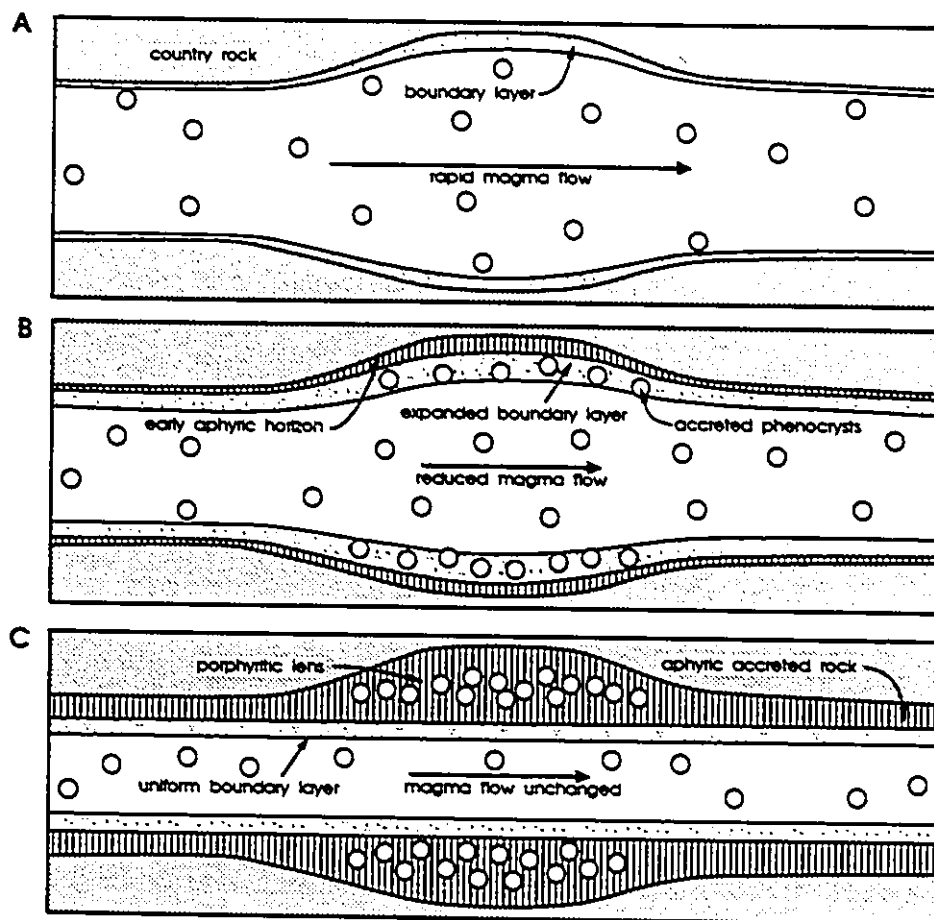


Figure 6.2. Schematic illustration (not to scale) of the development of porphyritic lenses in a matrix of aphyric rock through differential accretion in a sill of non-uniform thickness. See text for discussion.

phenocryst content of material accreted at the interface may be variable in the direction of flow. Over time, the variation in conduit thickness will tend to be damped out by more rapid advance of the solid-liquid interface in the thicker boundary layers. This may lead to isolated lenses of one lithological type within a matrix composed of second lithological type. Figure 6.2 illustrates development of this type of feature in a hypothetical sill of non-uniform thickness and equal thermal gradients at upper and lower contacts. Initially, magma flow rate is sufficiently rapid to preclude incorporation of phenocrysts even where flow rate is slightly reduced due to increased sill width (Fig. 6.2A). Consequently, a uniformly aphyric layer of rock is accreted onto the sill walls. After some time interval a reduction in flow rate occurs, and the lower thermal gradient and increased boundary layer

thickness in the wider portion of the sill allows accretion of porphyritic material in this zone (Fig. 6.2B). In the narrower portions of the sill, however, flow rate remains too high for accretion of phenocrysts. If flow rate is maintained, the width anomaly is eventually damped out by more rapid accretion in this part of the intrusion (Fig. 6.2C). The flow rate in this location thus becomes equivalent to other locations along strike, and the boundary layer is of uniform thickness. Consequently aphyric material is accreted along the entire length of the sill, thereby enclosing the porphyritic lens.

6.6 Flow Differentiation

Flow differentiation is the process by which a suspension may become heterogeneous with respect to the suspended solids due to flow in a velocity gradient. Net movement of suspended particles in a flowing fluid may occur as a result of single particle forces, forces arising from grain interactions or a combination of the two. Wall effects and Magnus-type effects belong to the first category of forces, while grain dispersive pressure results from grain interactions.

Wall Effect

The theoretical nature of single particle forces is poorly understood, but the effects have been documented experimentally. The wall effect involves the inward migration of particles in a fluid where the particles were initially close to a bounding wall and have a lateral motion approximately parallel to the wall. The inward migration may result from the same type of viscous forces in the fluid which cause mutual repulsion between suspended particles during flow (Komar, 1972a). The effect is best observed when spherical, non-neutrally buoyant particles sink or rise in a stagnant fluid along a vertical boundary. The wall effect, however, is limited to a range of only a few grain diameters from the boundary surface (Maude and Whitmore, 1956) and cannot explain particle distributions such as those found in the Menihek glomeroporphyritic sills.

Magnus Effect

The effects of Magnus-type forces and grain dispersive pressure (see below) are best illustrated in the context of a viscous fluid undergoing laminar flow in an infinitely long conduit. If the flow is considered as a series of discrete rigid layers of fluid, each molecularly bonded to adjacent layers, the behaviour of the fluid on a macroscopic scale can be examined (Fig. 6.3). The fluid layers immediately adjacent to the bounding walls will bond with the solid surface and show strong viscous resistance to movement such that the velocity (U) of these layers can be considered to be zero. Successive layers in a direction perpendicular to flow (y) slide relative to preceding layers at a rate that is dependent on the viscosity of the fluid (i.e., interlayer bonding must be overcome for sliding to occur). In effect, each layer exerts a frictional drag on the next layer. To offset frictional drag and maintain a constant rate of flow, central layers must be accelerated relative to the mean flow velocity of the fluid. This relative acceleration of central layers gives rise to a velocity gradient (dU/dy) which, given ideal Newtonian behaviour and infinite flow distance, has a parabolic profile with steep gradients at the bounding walls and zero gradient at the centre of flow (Fig. 6.3A).

Parabolic velocity gradients are replaced by plug-like gradients in non-Newtonian or pseudo-plastic fluids, such as suspensions where particle concentrations are extremely high (Leeder, 1982). The velocity distribution in such fluids is flatter in the central part of the flow, with high velocity gradients in boundary layers adjacent to conduit walls (Fig. 6.3B). The thickness of the boundary layer is proportional to the viscosity of the fluid, and in low viscosity fluids (e.g., water), which show limited viscous resistance to flow, boundary layers are small. Conversely, the significantly higher viscosities of magmas lead to larger boundary layers for equivalent flow rates. In mixtures of two fluids having different viscosities, the higher viscosity component will tend to migrate towards the central zone of the conduit such that the lower viscosity component bears the larger proportion of the velocity gradient. This is the principal underlying the use of water to reduce the pressure required to achieve a given flow rate in oil pipelines. Simultaneous flow of liquids of contrasting viscosity has also been invoked to explain the spatial relationships between some mafic and felsic volcanic eruptions (Carrigan and Eichelberger, 1990)

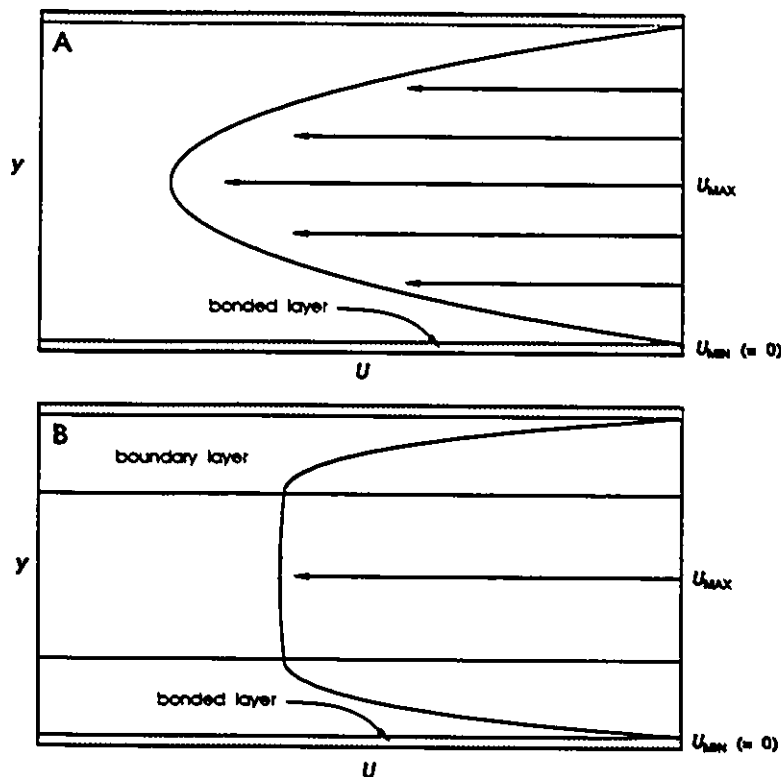


Figure 6.3. Stream line visualizations of laminar flow in (A) a Newtonian fluid giving rise to a parabolic velocity gradient, and (B) a highly non-Newtonian fluid giving rise to a boundary layer or plug flow regime, with high velocity gradients in the boundary layers and a low or zero velocity gradient in the central plug.

In a suspension in which the velocity is sufficient to maintain particle suspension, various forces will act on the solids during flow. At low particle concentrations, the large separation distance creates the situation where near approaches or collisions between particles due to velocity variations are rare. Such events will have little effect on particle distribution because forces arising from particle interactions are negligible due to the low frequency of particle approach. There will, however, be Magnus-type forces acting on the particles as a result of particle rotation imparted by the velocity gradient. Magnus-type forces are those responsible for the motion, perpendicular to flow direction, imparted to a rotating solid in a laminar fluid flow (Pao, 1961). Rotation produces differential relative velocities between the solid and fluid at the upper and lower surfaces such that a pressure imbalance develops. The effect is essentially a lift force, analogous to the pressure

imbalance resulting from differential air flow velocity along the upper and lower surfaces of an airfoil. In flowing suspensions exhibiting a velocity gradient, Magnus-type forces will act on the suspended particles since rotation is imparted to the solids by the differential velocity between adjacent fluid layers. The direction of particle rotation is such that pressure reduction occurs at the surface of the particle which is in contact with the higher velocity fluid. Because the velocity of fluid layers increases toward the centre of flow, the transverse motion of the particles will be in this direction. Magnus-type forces, then, are capable of producing particle distributions similar to those observed in some dykes and sills.

Grain Dispersive Pressure

When the concentration of a moving suspension exceeds approximately 5% (Komar, 1972b; 1976), interactions between particles will have an important effect on their distribution. This is due to the fact that particles in adjacent fluid layers will have different velocities, and consequently will approach and interact with each other. Grain interactions may be near approaches (viscous interactions) or actual collisions (inertial interactions). In viscous fluids where particle concentration is relatively low and velocity gradients small, there will be a force applied to approaching particles as a result of the incompressibility of the intergranular fluid between the two particles. This force will have a component in a direction oblique to that of flow, and the particles will tend to diverge. Increasing velocity gradients may lead to approach velocities which are sufficient to overcome this viscous repulsion and allow collisions to occur. Higher particle concentrations will have a similar effect due to the reduction in mean separation distance. Collisions will also cause divergence of particles trajectories since the approach paths can never coincide (for coincident approach paths, particle velocities are equivalent and collision will not occur). Grain interactions must then lead to a mutual repulsion between particles, or grain dispersive pressure (Bagnold, 1954), irrespective of the type of interaction. In a suspension with an initially uniform distribution of particles and a parabolic velocity profile, the greatest number of grain interactions will occur near the conduit walls where the velocity gradient is highest, and the frequency of interactions will decline to a minimum at the

centre of the conduit. Thus the grain dispersive pressure will be greatest near the boundaries of a conduit and lowest in the axial region.

Bagnold (1954) made a quantitative investigation of grain dispersive pressure by shearing viscous suspensions between concentric cylinders. No dispersive pressure was associated with the shear of a viscous fluid alone, and therefore this effect was due to the presence of particles. Bagnold (1954) found that the magnitude of the dispersive pressure was dependent on the diameter (D) of the grains, the rate of shear ($\partial U/\partial y$), the grain concentration and the viscosity of the fluid (η). He defined a dimensionless shear stress number (B) as the ratio of inertial stress to viscous stress, and used this parameter to delineate conditions under which inertial interactions or viscous interactions predominate in the development of grain dispersive pressure. The shear stress number, or Bagnold number, is related to the physical parameters of the suspension by (Bagnold, 1954):

$$B = \frac{\rho_s}{\eta} \Lambda^{\frac{1}{2}} D^2 \frac{\partial U}{\partial y}$$

where ρ_s is the grain density, and Λ is the linear concentration defined as the ratio of the mean grain diameter to mean free distance between grains. The linear concentration is related to the volume concentration (C) by:

$$\Lambda = \frac{1}{\left(\frac{C_0}{C}\right)^{\frac{1}{3}} - 1}$$

where C_0 is the maximum volume concentration and is equal to 0.74 for uniform spheres in the cubic close-packed arrangement. C_0 for natural suspensions is typically about 0.65 (Gray, 1968; Komar, 1972a)

For suspensions with $B < 40$ the viscous effects of the fluid are largely responsible for grain dispersive pressure, whereas for $B > 400$ the inertial effects of the suspended grains dominate. In the viscous region Bagnold (1954) obtained the empirical relationship for the grain dispersive pressure (P_v):

$$P_v = a_v \Lambda^{\frac{1}{2}} \eta \frac{\partial U}{\partial y}$$

where the empirically derived constant $a_v = 3.0$ for spherical grains of uniform size. Similarly, the relationship for grain dispersive pressure in the inertia regime (P_i) is given by:

$$P_i = a_i \rho_s (\Lambda D)^2 \frac{\partial U}{\partial y}$$

where the constant $a_i = 0.04$ for uniform spherical grains. Comparison of the latter two equations reveals the additional dependency of the dispersive pressure in the inertia regime on the grain size and density.

Because equilibrium flow demands constant dispersive pressure across the conduit:

$$P_v = a_v \Lambda^{\frac{1}{2}} \eta \frac{\partial U}{\partial y} = \text{constant}$$

$$P_i = a_i \rho_s (\Lambda D)^2 \frac{\partial U}{\partial y} = \text{constant}$$

Thus an increase in the velocity or shear gradient moving toward the walls of the conduit must be balanced by a reduction in particle concentration or, in the inertial regime, reductions in particle concentration, diameter or density.

As a result of the requirement for constant grain dispersive pressure, the parabolic velocity profile for a Newtonian suspension undergoing laminar flow will ideally result in a parabolic particle concentration gradient across the conduit. Porphyritic magmas which behave as Newtonian fluids and undergo laminar flow will therefore develop parabolic phenocryst distributions with the greatest concentrations in the axis of the sill or dyke. If this distribution is maintained during solidification, the resultant dyke or sill will also exhibit a parabolic phenocryst distribution (Fig. 6.5A). The exception to this generality occurs when phenocryst concentrations reach levels where they have a significant impact on viscosity. In this case, phenocryst concentration profiles may become plug-like in shape (Komar, 1976). Porphyritic magmas which behave as pseudo-plastic fluids and undergo laminar flow during emplacement will also develop plug-like phenocryst distributions as a result of the development of a boundary layer flow regime. Upon solidification, this will produce a dyke or sill with a central plug having a relatively uniform phenocryst content, and marginal zones with low concentrations of phenocrysts (Fig 6.5B).

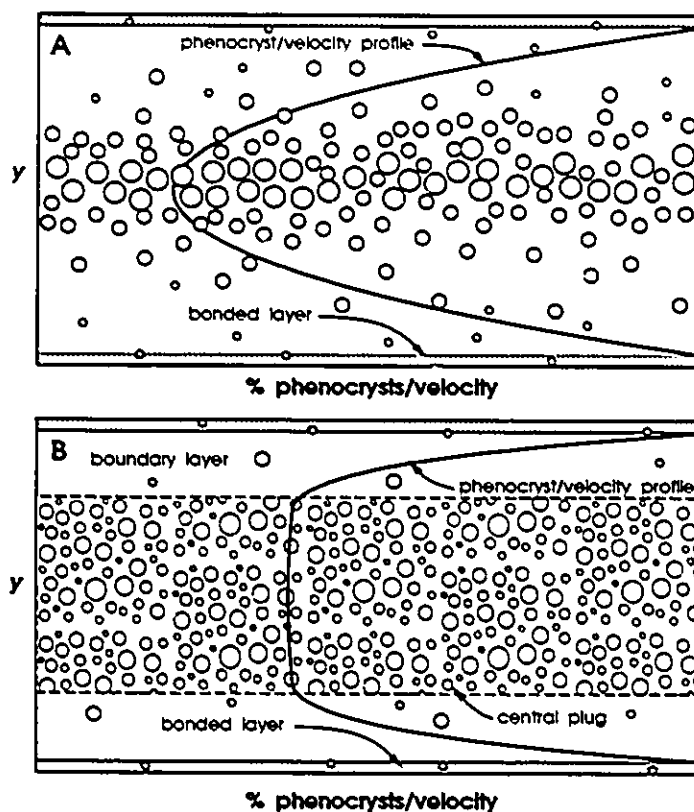


Figure 6.4. Particle distributions in a hypothetical sill having undergone flow differentiation during emplacement. (A) Parabolic increase in concentration and size of particles toward the centre of the sill as a result of a parabolic velocity gradient during flow. (B) Approximately uniform concentration of particles in a central zone with few particles in marginal zones as a result of a boundary layer or plug flow regime.

In addition to variations in phenocryst concentration, porphyritic magmas may develop a zonation with respect to phenocryst size as a result of grain dispersive pressure. In the inertial regime, the grain dispersive pressure is dependent on grain diameter, and hence constant pressure can be maintained by variations in phenocryst size as well as concentration. This could lead to more rapid axial migration of larger phenocrysts. In both the viscous and inertial regimes, the force applied to a phenocrysts is dependent on phenocryst area. At constant pressure this relationship requires that the force acting on a large phenocryst will be greater than that acting on a smaller grain. This will again lead to more rapid axial migration of larger phenocrysts. Thus a highly porphyritic magma behaving as a Newtonian fluid during emplacement may give rise to a dyke or sill which

exhibits parabolic phenocryst size and concentration gradients across its thickness, with the highest concentrations of the largest phenocrysts in the axial regions.

The effect of phenocryst concentration and size on the forces acting on a magmatic suspension during flow will also apply when the suspension behaves as a pseudo-plastic fluid. In this case, however, the velocity gradient will be plug-like, and consequently concentration and size gradients will also be plug-like. Because there will be no variation in velocity across the central plug in a perfect boundary layer or plug flow regime, the central zone will exhibit a uniform concentration and a random grain size distribution. Consequently, a porphyritic magma undergoing perfect plug flow during emplacement will form a sill or dyke with a central porphyritic zone with uniform phenocryst concentration and random grain size variations, and marginal sparsely porphyritic zones. It should be noted that, because of the equilibrium requirement of constant grain dispersive across the conduit, and because there is no grain dispersive pressure where there are no grains (Bagnold, 1954), it is theoretically impossible for completely aphyric zones in an intrusion to develop by flow differentiation. In fact, if the 5% concentration threshold suggested by Komar (1972b; 1976) for the onset of grain interactions is correct, it is theoretically impossible to develop zones containing less than 5% phenocrysts by flow differentiation.

Particle distributions in igneous dykes and sills will be influenced by the crystallization history of the magma. If flow continues for a considerable time interval subsequent to full development of equilibrium particle distribution, magma containing few particles may be continuously accreted onto the bounding surfaces. There will thus be a net decrease in the ratio of fluid to crystals. The increase in crystal content resulting from protracted flow may then lead to non-Newtonian behaviour and development of a plug flow regime in melts that were initially low in crystal content. Similarly, if initial particle content was very low (i.e., < 5%), the concentrating effect of wall crystallization may lead to the onset of grain interactions and development of grain dispersive pressure. Additionally, removal of crystal free melt at the walls during plug flow requires concomitant extraction of liquid from the central plug to preserve boundary layers which permit movement. It is hypothetically possible, given a single particle species and a complete range in particle dimensions, to produce an essentially solid, monomineralic

central plug which will stop moving only when all fluid in the boundary regions has been consumed by wall crystallization.

6.7 Glomerocryst Zonation in the Menihek Sills

Glomerocryst Content

The area occupied by glomerocrysts in a given sample was estimated by determining the average ratio of glomerocryst area to groundmass area in serial sections. Glomerocryst contents are therefore reported as percentage by area. As discussed earlier, volume concentrations will be slightly higher than these values. Average concentrations for samples from the four Howse Lake sill profiles are illustrated in Figure 6.5. From these diagrams it is apparent that there is a systematic variation in glomerocryst content with distance from sill contacts in each profile. Only in sill A, however, is the glomerocryst distribution symmetrical with respect to sill contacts. In the other profiles the zone of maximum glomerocryst content tend to be developed closer to the base of the intrusion. This asymmetry is most pronounced in sills B and C. The profile from sill A also exhibits the most smoothly varying distribution of glomerocrysts, although to some extent this probably reflects the lesser thickness of the intrusion which permitted closer sample spacing.

Sills A, B and D exhibit central zones of relatively high glomerocryst content which correspond to the anorthositic phase horizons developed in each. The anorthositic phase predominates in sill A, but comprises one third to one half of the stratigraphy of sills B and D. Although the glomerocryst concentration is somewhat variable across these horizons, the aggregates usually comprise 30-44% of the anorthositic phase. The central, well developed anorthositic phase horizon is generally flanked by less highly glomeroporphyritic anorthositic phase lithologies in which glomerocryst content declines from ~30% to 10-15%. From this rock type, the glomerocryst content declines further and is accompanied by a groundmass grain size reduction across an interval marking the transition to the marginal gabbroic phase. As was described in Chapter 4, some Howse Lake sills do not exhibit gradational transitions from anorthositic to gabbroic phase rock.

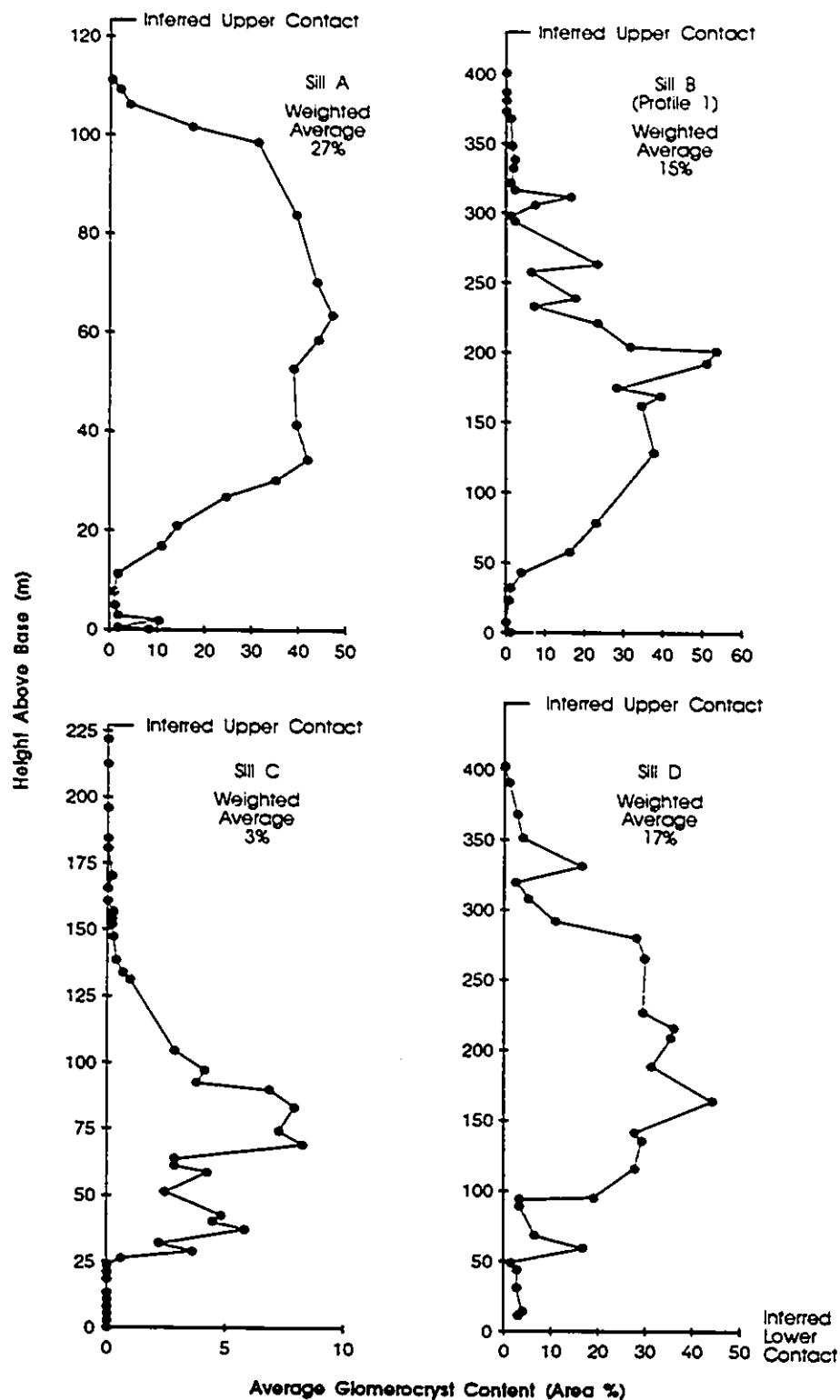


Figure 6.5. Variation in glomerocryst content with stratigraphic height in four sills from the Howse Lake area. Average 2σ analytical uncertainty is approximately twice symbol size in sills A, B and D, and is illustrated by an error bar in sill C.

This is the case at the lower contact of the anorthositic phase in sill D where there is an abrupt change from medium grained gabbroic phase rock containing a few percent glomerocrysts, to medium-coarse grained anorthositic phase gabbro containing ~20% glomerocrysts. The glomerocryst content of the marginal gabbroic phase horizons in all sills is generally less than 5%. Glomerocrysts were observed in the exposed chilled margins of sills A and B, although their distribution was highly irregular and their concentrations low.

Sills A, B and D exhibit peaks in glomerocryst content within the central anorthositic horizons which reflect the development of the core phase. In sill A the core phase concentration peak is relatively subdued because the maximum glomerocryst content (47%) is only marginally greater than the maximum (44%) found in the anorthositic phase. In sills B and D the core phase peak is more pronounced, and the glomerocryst content reaches maximums of 53% and 44%, respectively. Despite the apparent sharp increase in glomerocryst content in the core phase, this lithology is actually gradational into the enclosing anorthositic phase. Because of the large sample size required for representative measurements of core phase glomerocryst content, however, sampling was confined to a few exposures exhibiting what appeared to be maximum concentrations.

In addition to peaks in glomerocryst content in the core phases of sills B and D, smaller concentration peaks occur in the marginal gabbroic phase of these profiles. These correspond to the anorthositic phase lenses found in many of the Howse Lake glomeroporphyritic intrusions. The rather sharply defined increase in glomerocryst content in the anorthositic lenses as illustrated in Figure 6.5 is again misleading since abrupt contacts between the lenses and enclosing gabbroic phase material were not observed. Instead, the contacts were gradational, although the transition zones may be quite narrow. Again the apparent abruptness of contacts reflects sampling difficulties, in this case caused by the relatively limited thickness of the anorthositic phase lenses.

The glomerocryst profile across sill C exhibits numerous similarities to those of the other intrusions, although the glomerocryst contents are clearly much lower. The gabbroic phase occupies the entirety of sill C, but systematic variations in glomerocryst content occur within this lithology. The highest glomerocryst concentration of about 8% occurs in the centre of a 100 m interval of elevated glomerocryst content. On either side of the zone

of maximum glomerocryst content, there is a decline in the concentration of aggregates culminating with essentially aphyric marginal horizons. The distribution of glomerocrysts appears to be somewhat erratic in the lower transition zone, although this may simply reflect the elevated analytical uncertainty in glomerocryst content due to low concentrations. The upper transition zone is less well exposed and the degree of variation in glomerocryst content cannot be evaluated. No glomerocrysts were observed in the lower chilled margin of sill C, but occasional plagioclase phenocrysts are present.

Glomerocryst Size and Frequency

Field observations suggested that there was a correlation between total proportional glomerocryst content and the size and frequency of the aggregates in the Howse Lake sills. The relationships between glomerocryst content and the number and size of glomerocrysts were determined by measurements of the average size and total number of glomerocrysts in unit area in each sectioned sample. The general relationships between these three parameters are illustrated in Figure 6.6 where data from all four profiles is plotted. It is apparent that the relative importance of size and frequency in determining the proportional area occupied by glomerocrysts varies with the latter ratio. In sparsely glomeroporphyritic gabbroic phase samples, both the size and frequency of glomerocrysts determine the ratio between the area of the aggregates and that of the groundmass. The relationship between frequency and glomerocryst content, however, is better defined for low total glomerocryst contents. This indicates that frequency is more important than size in determining total glomerocryst content in sparsely glomeroporphyritic lithologies. This relationship is reversed in the anorthositic phase where frequency increases more slowly with increasing glomerocryst content. Frequency reaches a maximum at glomerocryst contents of about 30%, and samples with higher concentrations, including those obtained from the core phase, contain fewer aggregates. In contrast, average glomerocryst size increases across the entire range of total glomerocryst concentration, reaching maximum values in the core phase. Thus the variations in total glomerocryst content across a given sill reflect variations in both glomerocryst frequency and size, but the latter property exerts the predominant control on proportional glomerocryst content in highly glomeroporphyritic lithologies.

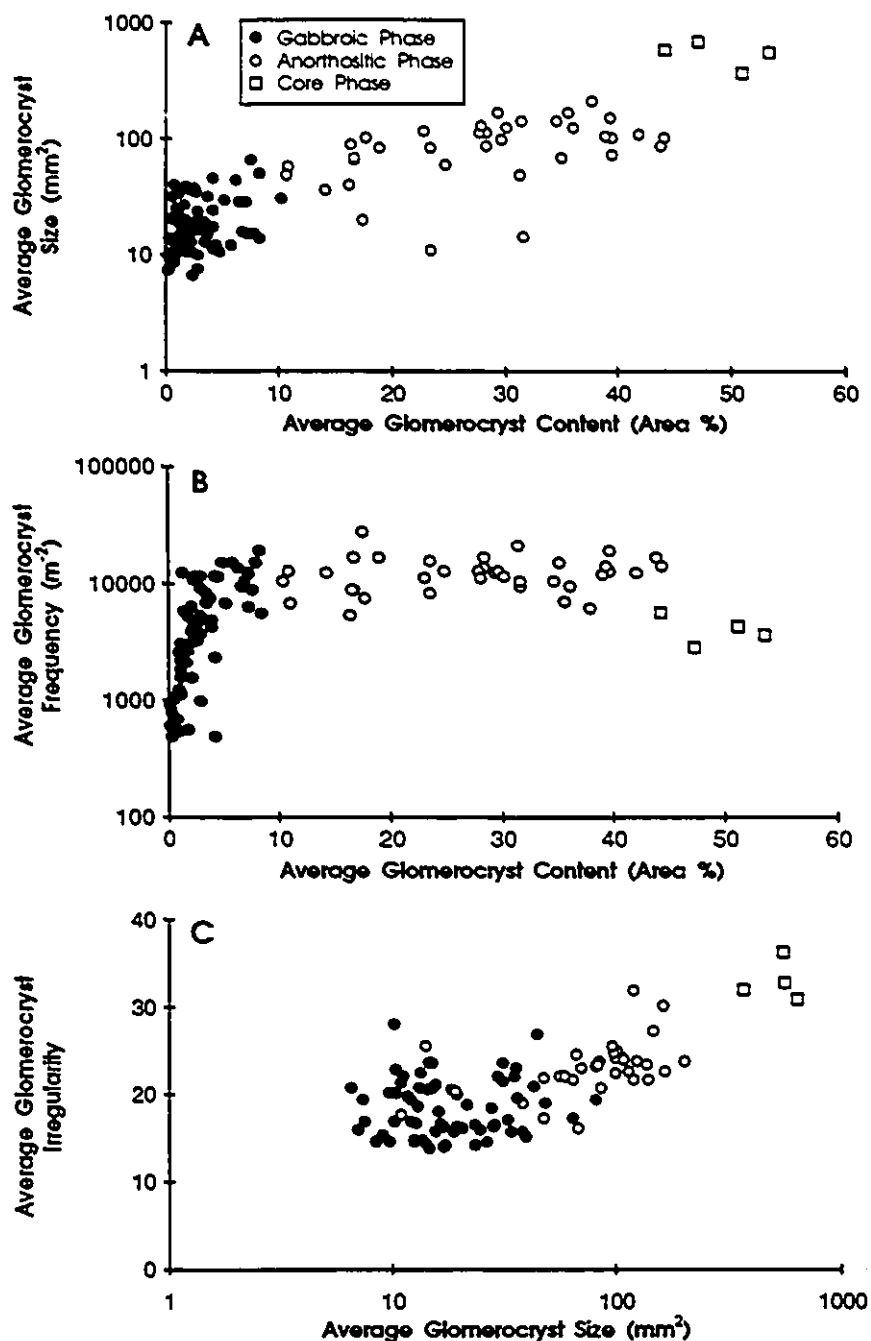


Figure 6.6. Covariation of measured glomerocryst parameters for samples from all four sill profiles examined at Howse Lake. (A) Average glomerocryst size versus total glomerocryst content, (B) average glomerocryst frequency versus total glomerocryst content and (C) average glomerocryst irregularity versus average glomerocryst size. Note that glomerocryst size and frequency axes are logarithmic due to the large variation in these parameters.

Glomerocryst Irregularity

A further feature that was noted during mapping was the apparent increase in the regularity of the shape of the glomerocrysts with increasing size. In outcrop, the core phase appears to contain relatively spherical glomerocrysts, while those of the gabbroic phase seemed more irregular in outline. As discussed earlier, a measure of the degree of irregularity in two dimensions was obtained using measurements of glomerocryst area and perimeter. The resultant ratio was thought to be a better measure of the shape of the glomerocrysts than the ratio of the two principal axes because intersection angles were unknown. The irregularity measurements are plotted against average glomerocryst size in Figure 6.6C where it is apparent that there is a positive correlation between the two parameters. This result, however, is the opposite of that expected from field observations. Thus the large, apparently spherical glomerocrysts of the core phase are in fact more irregular in outline than those of other lithologies. This is thought to reflect increasing perimeter length in larger glomerocrysts due to the increased number of marginal crystals. If this interpretation is correct, then the difference in the apparent regularity of glomerocrysts as determined by field observations and image analysis measurements is probably a scaling problem.

6.8 Origin of Glomerocryst Zonation

Multiple Magmatic Injection

The Menihek sills most likely to represent composite intrusions formed by multiple injection of dissimilar magmas are those in which there are non-gradational changes in glomerocryst content, size or frequency, possibly accompanied by changes in groundmass mineralogy or grain size. As noted earlier, there are several lines of evidence which suggest that the conformable horizon of aphyric norite in sill B represents a slightly later magma injection. Of the remaining Menihek gabbro intrusions mapped in detail at Howse Lake, however, only the sharp lower contact of the anorthositic phase in sill D is consistent with a multiple injection origin. This anorthositic horizon, however, is gradational into the overlying gabbroic phase at its upper contact. Although it is possible that the observed

upper contact is not the match to the lower contact (i.e., the intruded sill was already differentiated into gabbroic and anorthositic phases), this implies a degree of coincidence that seems improbable. The consistently gradational relationships between the various lithologies in other sills, and the general frequency of the tripartite structure among the Menihek gabbros, also indicate that multiple injection is an implausible origin for glomerocryst zonation. Thus it seems likely that the structure of the glomeroporphyritic sills at Howse Lake reflects the operation of a process acting on glomeroporphyritic magmas during emplacement of a single contiguous melt injection, which may or may not have varied in glomerocryst content with time. This does not preclude multiple injection of magma as the origin of certain types of lithological variation in some sills. For example, the central zones of aphyric gabbro enclosed by marginal horizons of anorthositic gabbro described by Baragar (1960) from Menihek intrusions in the Ahr Lake area may reflect injection of aphyric melt into a pre-existing glomeroporphyritic sill, as would seem to be the case for the noritic layer of sill B.

Wall Crystallization During Emplacement

The glomerocryst distributions observed in the Menihek sills may have been produced by wall crystallization involving simple accretion from magmas with variable crystal loads, or by differential accretion of a single magma type. In both cases, the sill cross-sections would then represent time profiles recording successive magma compositions and/or variations in flow regime. Studies of the wall crystallization process have been primarily concerned with flow in relatively narrow, sub-vertical dykes. It is therefore uncertain as to whether the types of thermal effects postulated for dykes will also apply to sub-horizontal sills of significant thickness. It is likely, for example, that the symmetrical thermal profiles formed at the walls of vertical dykes (Delaney and Pollard, 1982) will not develop at the upper and lower contacts of sills because the heat flux through the roof of the latter structures may exceed that through the floor (Irvine, 1970).

The minimum effect of differences in thermal regimes at the base and floor of a sill will be asymmetric development of solid-liquid boundary layers, and thus asymmetric development of accreted layers. As discussed by Huppert and Sparks (1989), however, the thermal flux through upper contacts in the case of flowing or convecting magmas may be

sufficient to prevent wall crystallization, and even to cause melt back of initial chilled margins and partial melting of country rock. These authors noted that emplacement of dykes or sills will be accompanied by formation of chilled margins, and subsequent wall crystallization or marginal melt back will be dependent on the relative rates of heat transfer to, and through, the contacts. Huppert and Sparks (1989) envisage three potential thermal regimes during emplacement and cooling of sills:

- (1) Heat flux from the magma never exceeds the conductive heat flux through the contact. This gives rise to formation of a chilled margin at the contact upon emplacement, followed by wall crystallization with coarsening of grain size as cooling proceeds.
- (2) Heat flux from the magma exceeds conductive heat flux through the contact for a finite time interval and is then reduced. This will cause melt back of some portion of the chilled margin formed during emplacement. If the heat flux from the magma is reduced prior to complete melt back, an internal contact with the initial chilled margin may form, and crystallization will again proceed inward from the new solid-liquid interface.
- (3) Heat flux from the magma exceeds conductive heat flux through the contact for an extensive time period. This may cause complete removal of the chilled margin and potentially partial melting of wall rock. Continued cooling will eventually lead to crystallization at the contact, but because of the increased temperatures of the wall rocks accreted material will not be chilled.

Of particular importance in the wall crystallization history of a sill is the heat flux due to magma convection, since this process will be more efficient than conductive heat transfer. Because convective heat transfer will affect only the upper contact of a sill, wall crystallization or chilled margin melt back can be asymmetric at the roof and floor. Huppert and Sparks (1989) also noted that any magma characteristics (e.g., crystal content) which give rise to high viscosity, non-Newtonian behaviour or otherwise affect magma convection will have significant effects on the heat flux to the contacts.

Where lower contacts are exposed, the Menihek sills at Howse Lake exhibit basal chilled margins. The upper contact was observed in only one sill, but this intrusion has an upper quenched margin comparable to those observed at the lower contacts of other sills. In addition, the upper gabbroic phase of sills A and C are relatively fine grained immediately below the inferred upper contacts. These observations suggest that chilled margins are present at both the upper and lower contacts of the majority of Menihek gabbros, an interpretation consistent with the observations of Baragar (1967) and Dimroth (1978). Consequently, it can be concluded that the heat flux to the solid-liquid interface was insufficient to remelt quenched material formed during emplacement of the Menihek sills. Thus these intrusions developed the first category of thermal regime described by Huppert and Sparks (1989), in which heat transfer through the contacts exceeds heat transfer to the contacts.

The presence of chilled margins and the gradational increase in groundmass grain size inward from contacts observed in most intrusions indicate that initial quenching of magmas upon emplacement was followed by some degree of wall crystallization. This is consistent with geochemical variations in sills B and C which indicate that much of the solidification histories of the intrusions were dominated by fractional crystallization which progressed inward from the contacts. Since some sills can be traced for several tens of kilometres along strike, and because the thermal flux through the contacts apparently exceeded that to the contacts, it is probable that some proportional thickness of a given intrusion formed by wall crystallization during emplacement. The proportional thickness of syn- and post-emplacement crystallization products cannot be determined, although lithological variations place constraints on the maximum thickness of horizons accreted during flow in some intrusions. In sills such as B and D, in which lithological variations are asymmetric or require closed system fractionation (e.g., oxide horizons), the maximum thickness of material accreted during flow is relatively small. In contrast, significant proportions of intrusions such as A and C, in which there is no significant lithological asymmetry, may have formed by accretion during flow. Extensive accretion is also consistent with the constancy of mineral compositions across sill C, which could reflect continual displacement of liquids residual to wall crystallization by continuous flow of magma past the profile location. The presence of felsic rocks in sill C and other intrusions,

however, indicates that at some point these systems also became closed, such that fractional crystallization led to evolution of siliceous residual melts.

Development of glomerocryst zonation in the Menihek sills by simple accretion would require continuous or intermittent displacement of unsolidified melt by increasingly glomerocryst-rich magmas during emplacement. In a typical highly glomeroporphyritic intrusion such as sill A, marginal chilled zones and adjacent fine to medium grained gabbroic phase horizons could form by initial quenching and subsequent accretion of sparsely glomeroporphyritic melts (up to 10% glomerocrysts) in the early stages of sill development. Residual sparsely glomeroporphyritic melts would then be displaced by magma containing up to 45% glomerocrysts, and accretion of the anorthositic phase would occur. The insulating effect of the marginal gabbroic phase would reduce the rate of heat loss through the solid-liquid interface, and hence promote slow crystallization and the development of the coarse groundmass characteristic of the anorthositic phase. In some sills, a third melt, containing up to 55% glomerocrysts of relatively large dimensions, would be required to produce the core phase. In sparsely glomeroporphyritic sills, such as sill C at Howse Lake, accretion from an initial aphyric magma which was subsequently displaced by sparsely glomeroporphyritic melt could produce the observed lithological variation.

Although these types of magma sequences could produce the general variations in glomerocryst concentrations apparent in the Menihek sills, there are numerous inconsistencies in such a model. This is particularly evident when glomerocryst distributions are examined in detail. A fundamental problem with the simple accretion process lies in the requirement that the glomerocryst contents of successive magmas increase. Because the glomerocrysts are thought to represent fragments of semi-consolidated plagioclase flotation cumulates, melt extraction must have occurred subsequent to, or during, an interval of plagioclase crystallization and accumulation in the crustal magma reservoir. Since it is unlikely that the formation of plagioclase cumulates would occur on a time scale large in comparison to that of melt extraction, magmas would become progressively depleted in glomerocrysts as the accumulated anorthositic material was exhausted. This situation is the reverse of that required by the accretion model. In addition, there is no obvious mechanism by which entrained glomerocrysts might become

larger as melt extraction continued. The anorthositic lenses found in many highly glomeroporphyritic sills are also inconsistent with a simple accretion model, since these would require relatively rapid fluctuations between sparsely and highly glomeroporphyritic melts. Finally, simple accretion cannot explain the asymmetric development of lithologies such as oxide or sulphide horizons without substantial variations in melt geochemistry. Since a cross-section through a sill formed by simple accretion represents a time profile, a complex and unrealistic magma succession would be required to produce the disparate rock types.

Differential accretion during emplacement offers an alternative explanation for variations in glomerocryst content and size without requiring changes in magma type. If the Menihek sills underwent differential accretion during emplacement, glomerocryst size and frequency distributions might have arisen through variations in the flow regime of a homogeneously glomeroporphyritic melt during sill development. The development of a glomerocryst zonation profile in a highly glomeroporphyritic intrusion such as sill A might occur as follows. During the initial stages of emplacement, the leading edge of the intruding magma will be quenched at the contacts with the country rock. The glomerocryst content of the quenched material will depend on the thermal contrast between country rock and magma, which will control the thickness of the chilled margin, and the size and concentrations of the aggregates in the melt. Given that the thickness of the chilled margins in the Menihek intrusions is greater than the diameters of most glomerocrysts, quenching of the magma at sill contacts should result in a glomeroporphyritic rock in which the size distribution and concentration of the aggregates is comparable to that of the magma. These parameters, however, will vary along strike since continuous accretion will produce changes in the glomerocryst content of the down-flow magma.

Continued growth of the sill will be accompanied by accretion of melt at the solid-liquid interface adjacent to the chilled margins. The size and abundance of glomerocrysts in the accreted rock will be a function of the thickness of the solid-liquid interface, and the concentration and dimensions of glomerocrysts in the magma. During periods of rapid flow, thermal gradients will be high at the solid liquid interface, and relatively narrow boundary layers will develop. If boundary layer thickness is less than the dimensions of the smallest glomerocrysts, aphyric material will be accreted. Since

glomerocryst dimensions range down to a few millimetres, however, development of sparsely glomeroporphyritic rock containing small aggregates is more likely. Relatively constant flow rates over some time interval will hence produce marginal sparsely glomeroporphyritic gabbro adjacent to the chilled margins. A gradational increase in the size and abundance of glomerocrysts in the gabbroic phase will accompany an increase in the thickness of the solid-liquid interface. The increase in boundary layer thickness will reflect a reduction of the thermal gradient due to the insulating effect of continuously accreted material, a reduction of flow rate, or both. Continued gradational changes in the thermal gradient will produce the gradational increase in glomerocryst size and concentration found in many sills, eventually leading to development of the anorthositic phase. Alternatively, an abrupt reduction in flow rate would produce a sharper transition to the anorthositic rock. Cessation of flow will allow incorporation of all glomerocrysts, irrespective of size, as the solid-liquid interface advances inward.

Changes in flow rate and thermal profiles as a glomeroporphyritic magma is emplaced could thus produce the general pattern of glomerocryst zonation observed in the Menihek sills. A few other lithological features might also be consistent with this type of process. A sustained interval of relatively rapid magma flow will promote prolonged accretion of sparsely glomeroporphyritic material with small aggregates. With time, this will increase the average size and concentration of glomerocrysts in the down-flow residual melt. The maximum glomerocryst content of an accreted horizon is therefore not restricted to the concentration in the initial magma derived from the crustal reservoir. This allows formation of horizons with concentrations of 40-50% glomerocrysts from magmas which might initially have contained only 10-20% glomerocrysts. The relatively narrow anorthositic lenses found in highly glomeroporphyritic sills might develop through variations in flow rate, possibly due to changing sill thickness as described in the previous section. Given sufficiently high flow rates, differential accretion might also produce aphyric gabbro such as occur in the marginal zones of sill C. Finally, accretion offers a potential mechanism for asymmetric development of horizons of variable glomerocryst content relative to sill margins. The zones of high glomerocryst content in sills B and C, for example, are offset toward the lower half of the intrusions rather than centred about the axes. This might reflect more rapid accretion at the roof of these sills due to more rapid

transfer of heat away from the upper contact. Such asymmetric cooling could result from convective transport of heat to the roof or increased heat flux through the roof due to pore fluid convection in the overlying country rocks.

As discussed above, accretion models are most consistent with sills which display relatively simple lithological variations. Of the four intrusions examined in detail at Howse Lake, the lithological variations in sills A and C are the most compatible with an origin by differential accretion. It seems unlikely, however, that the entire thickness of either sill at the measured profiles could form by accretion during flow. Assuming that the thermal gradient at sill contacts is compatible with wall crystallization during emplacement, the thickness of accreted rock at a given position along the strike of a sill will be a function of the crystallization and flow rates. These, in turn, will depend on various physical properties of the magmas and country rocks. An approximate estimate of crystallization times for basic sheets emplaced at upper crustal levels is given by (Jaeger, 1968):

$$t = 0.01d^2$$

where t is time in years and d is thickness in metres.

If the Menihek sills were emplaced in semi-consolidated sediments with high water contents, as is suggested in Chapter 10, cooling rates could be slightly higher due to more rapid heat removal by pore fluid convection in the unconsolidated sediments relative to indurated crustal rocks. But the relationship can be used to provide a rough evaluation of the applicability of accretion models in the Menihek sills. The relationship between crystallization time and sill width can be inverted to estimate the thickness of accreted horizons based on flow rate and sill strike length (Table 6.1). It is apparent from Table 6.1 that wall crystallization during emplacement will be limited to relatively narrow marginal horizons. Maximum thicknesses of accreted material will develop in laterally extensive sills emplaced slowly, but even with very low flow rates wall crystallization during emplacement will produce horizons only a few tens of metres thick. For the Menihek intrusions, realistic half thicknesses of accreted material at a point some distance from the sill origin are probably less than 10 m, even with more rapid crystallization due to convective removal of heat from contacts by convective circulation of water in the country rocks. Given that glomeroporphyritic sills less than approximately 100 m in thickness were

Table 6.1. Estimated thickness of rock accreted during emplacement of sills for variable flow rates and distances.

Flow Rate (m/sec)	Sill Strike Length (km)	Emplacement Time (days)	Accreted Thickness (m)
0.001	10	116	5.63
0.001	50	579	12.59
0.001	100	1157	17.81
0.01	10	12	1.8
0.01	50	58	4.0
0.01	100	116	5.6
0.10	10	1	0.6
0.10	50	6	1.3
0.10	100	12	1.8

not observed at Howse Lake, it is apparent that the glomerocryst zonation pattern could not have originated through wall crystallization during emplacement.

Flow Differentiation

Frictional drag at the bounding walls of a viscous fluid undergoing laminar flow in a conduit produces a velocity gradient across the flow. The velocity gradient gives rise to Magnus effects and grain dispersive pressure which promote migration of suspended solid particles toward the centre of flow. Particle concentration profiles are dependent on velocity and shear strain gradients, and these are influenced by physical parameters such as viscosity, flow rate and particle concentration and size. When particle concentrations exceed approximately 5% by volume, grain dispersive pressure is the predominant force acting on the particles (Komar, 1972a; 1976). At lower concentrations, the Magnus effect will predominate. The net result in both cases will be grain distribution profiles in which the highest concentrations of the largest particles will occur in the centre of the conduit, with particle size and concentration declining towards the walls (Komar, 1972a). The Magnus effect and grain dispersive pressure are thus theoretically capable of producing the type of glomerocryst zonation pattern observed in the Howse Lake sills.

The size and concentration sorting induced by flow differentiation will occur only in a laminar flow regime. The flow of the Menihek glomeroporphyritic melts must therefore have been laminar during emplacement if flow differentiation is the origin of glomerocryst zonation. The best evidence for laminar flow during intrusion of the glomeroporphyritic sills comes from the distribution of glomerocrysts in the intrusions. Since symmetrical zonation cannot be the result of gravitational effects, glomerocryst distribution must be related to fluid flow. But non-uniform glomerocryst distributions would not survive a turbulent flow regime. In addition, centimetre-scale layering, in which elongate grains are oriented parallel to sill contacts, probably reflects mechanical alignment of crystals during flow. Again this would not be possible if flow was turbulent.

Some support for laminar flow during emplacement of glomeroporphyritic magmas comes from the empirical Reynolds number (Re):

$$Re = \frac{\rho DU}{\eta}$$

where ρ is the fluid density (kg/m), D is the conduit diameter (m), U is the flow velocity (m/s), and η is the fluid viscosity (poise = 10 kg/ms). For $Re < 1000$, sheet flow between parallel walls will be in a laminar regime. The density of the Menihek magmas can be estimated from the compositions of the chilled margins of the sills, which represent quenched magma, using the method of Bottinga and Weill (1970). Because of the effects of alteration on the chilled margins, however, the compositions of the comagmatic upper Menihek and Willbob basalts (see Chapter 8) provide better estimates of melt compositions. The average density of the Menihek magmas calculated from aphyric upper Menihek and Willbob basalts is 2.62 g/cm³ at 1250°C. The presence of plagioclase glomerocrysts ($\rho \sim 2.65\text{--}2.7$ g/cm³) will increase the density of the magmatic suspensions, as will lower temperatures. But the effects on the Reynolds number of these factors will be very minor in comparison to variations in viscosity, flow rate and sill thickness. The viscosities of aphyric Menihek magmas can be estimated from the compositions of the comagmatic basalts using the method of Shaw (1972). Results range from 300 to 800 poise at 1250°C, with an average value of about 500 poise. The presence of glomerocrysts in the

melts will increase the apparent viscosity (η_a) of the suspension according to (Roscoe, 1952):

$$\eta_a = \eta(1 - \gamma C)^{-2.5}$$

where η is the viscosity of the intergranular liquid, γ is a constant ranging from 1.0 to 1.35 depending on grain size distribution, and C is the particle volume concentration. Using the calculated average melt viscosity of 500 poise and a glomerocryst concentration of 10%, the apparent viscosity will be about 700 poise.

The flow rate of magma during emplacement is difficult to estimate. Velocities on the order of 1 m/s have been attributed to relatively narrow basaltic dykes (cf. Platten and Watterson, 1987), but because of the generally much greater conduit diameter in sills, lateral flow rates will be considerably lower. Using a flow rate of 0.1 m/s, a glomeroporphyritic magma containing 10% aggregates emplaced as a 600 m thick sill, with $\rho = 2.62 \text{ g/cm}^3$ and $\eta_a = 700$ poise will have $Re = 2300$. This value exceeds that which marks the onset of turbulent flow. But since the highly glomeroporphyritic sills (A, B and D) examined in detail at Howse Lake have weighted average glomerocryst contents of 15-27% by area, and range from 120 m to 440 m in thickness, Re for these melts at the same rate of flow would be smaller. For the parental magma of sill A, with a glomerocryst concentration of 27%, $Re = 140$. Flow during emplacement of this intrusion would therefore have been laminar. Although these results are very dependant on flow rates, it seems likely that highly glomeroporphyritic magmas were generally emplaced under a laminar flow regime. For relatively narrow and sparsely glomeroporphyritic intrusions such as sill C ($Re = 980$), flow may have been laminar or turbulent. For thicker (500-600 m) aphyric sills, magma emplacement would have occurred under a turbulent flow regime unless flow rates were less than 5 cm/s or magmatic temperatures were lower. A temperature drop of 100°C , for example, would increase average melt viscosity from 500 to 1800 poise. At this viscosity, even a 600 m thick intruding magma would be in a laminar flow regime. Since there is some evidence that nucleation and growth of groundmass crystals occurred during flow, it is likely that magmatic temperatures were less than the estimated 1250°C during emplacement.

The foregoing calculations suggest that magmas containing >15% glomerocrysts emplaced as sills would undergo laminar flow. Intrusions of lower glomerocryst contents may or may not have undergone laminar flow, although the effects of heat loss and wall crystallization would tend to damp out turbulence. Consequently, flow differentiation is a potential factor in the development of glomerocryst zonation patterns in these intrusions even if the early stages of melt emplacement were turbulent. Given laminar flow, Magnus-type forces would promote migration of glomerocrysts toward the centres of the sills. Komar (1972a), however, calculated that Magnus effects will be one to two orders of magnitude less efficient in the concentration of particles than grain dispersive pressure. For grain dispersive pressure to exert a translatory force on particles in a suspension, a concentration threshold of 5% by volume is required (Komar, 1972b; 1976). The weighted average area concentrations of glomerocrysts in sills A through D are 27%, 15%, 3% and 17%, respectively, and the equivalent volume concentrations would be slightly higher. Thus glomerocryst distributions in sills A, B and D are likely to reflect the effects of grain dispersive pressure during magma emplacement. The volume concentration of glomerocrysts in the parental magma of sill C, however, was probably slightly below the threshold concentration required for development of grain dispersive pressure.

Bagnold (1954) determined that grain interactions in flowing suspensions were dominated by collisions when particle concentrations were high, velocity gradients were high and intergranular fluid viscosities were low. In comparison to Bagnold's experimental suspensions, the Menihek glomeroporphyritic magmas would be highly viscous, have relatively low particle concentrations, and would be subject to relatively low velocity gradients during emplacement. These factors give rise to Bagnold numbers $\ll 40$ and hence interactions among glomerocrysts during magma emplacement would be in the viscous domain. Komar (1976) calculated equilibrium particle distribution profiles resulting from grain interactions in the viscous regime assuming equivalent particle and liquid densities, constant intergranular liquid viscosity and Newtonian behaviour of the suspension. Profiles for wall concentrations of 10% and 40% are illustrated in Figure 6.7, along with a smoothed concentration profile for sill A. A comparison of the calculated grain concentration profiles with the measured glomerocryst distribution profiles of Figure 6.5 reveals a qualitative similarity. Although there are numerous differences in detail, sills

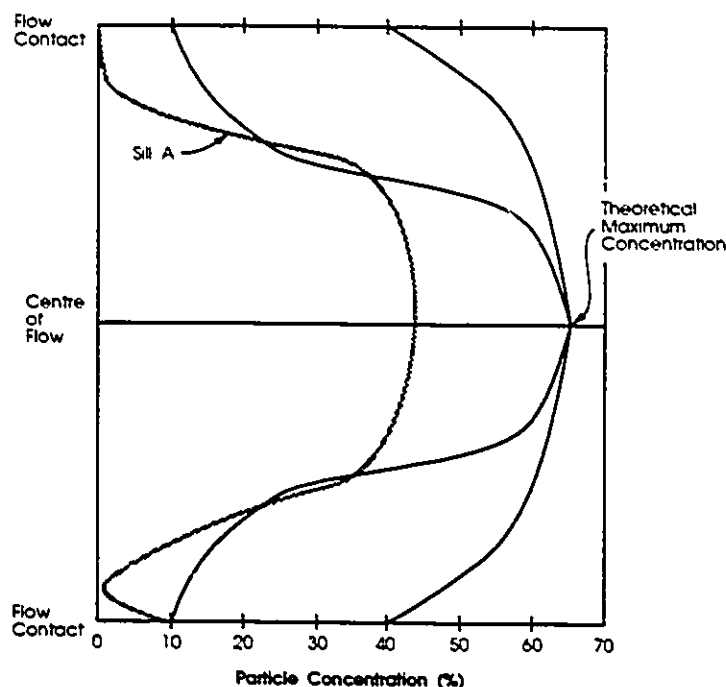


Figure 6.7. Calculated glomerocryst concentration profiles across a hypothetical sill undergoing flow differentiation during emplacement. Profiles are illustrated for wall concentrations of 10% and 40% by volume. A smoothed concentration profile from sill A is illustrated for comparison. Modified after Komar (1976).

A through D exhibit central horizons of relatively high but fairly constant glomerocryst content, and marginal zones in which glomerocryst content drops rapidly. The smoothed profile of sill A is particularly analogous to the predicted profile for low wall concentrations. In addition, the glomerocryst size variations observed in the Menihek sills are consistent with an origin related to grain dispersive pressure. Although grain size does not enter the equation for grain dispersive pressure in the viscous regime (Bagnold, 1954), size sorting will still occur because the force applied to a large particle will be greater than that applied to a small one at constant pressure. The qualitative similarity between predicted and observed particle concentrations, the attendant size variations, and the fact that the parental magmas of sills A, B and D meet the requirements for development of grain dispersive pressure, indicate that the glomerocryst distributions in these intrusions reflect flow differentiation during emplacement.

The principal differences between observed and theoretical particle concentration distributions are: (1) very low glomerocryst concentrations at sill margins, (2) low maximum glomerocryst concentrations, (3) asymmetric glomerocryst concentration gradients at upper and lower contacts of the anorthositic phase in some sills (e.g., B and D), (4) asymmetry in the thickness of gabbroic phase horizons (e.g., sill B), and (5) concentration spikes corresponding to anorthositic phase lenses in sparsely glomeroporphyritic marginal gabbro (e.g., sills B and D). With the exception of the anorthositic phase lenses within sparsely glomeroporphyritic gabbroic phase horizons, these discrepancies can be related to limited pseudo-plastic behaviour of the glomerocryst-magma suspensions, or to the effects of other forces acting on particle distributions induced by grain dispersive pressure.

The low marginal glomerocryst concentrations, which in some sills fall below the threshold for grain dispersive pressure, probably reflect the effects of Magnus-type forces and possibly wall crystallization. Komar (1972a) calculated migration velocities for particles under the influence of the Magnus effect based on the data of Jeffrey and Pearson (1965). For suspensions with physical characteristics roughly comparable to those of the Menihek glomeroporphyritic magmas, Komar (1972a) obtained migration velocities of 0.01 cm/s to 0.05 cm/s. For a glomeroporphyritic sill 20 km in length, emplaced at a rate of 10 cm/s, particle migration distances would be on the order of 20 m to 100 m based on the estimated migration rates. Numerous factors will influence particle migration rates, including velocity gradient, particle size and concentration and suspension viscosity and yield stress, and consequently estimated migration distances must be treated with a degree of caution. Nevertheless, the theoretical effect of Magnus forces would be to promote further concentration reduction in marginal zones through axial particle migration. Since low particle concentrations would reduce grain interactions that might interfere with single particle forces such as the Magnus effect, particle migration as a result of this process would be greatest where concentrations were lowest. Thus Magnus-type particle migration in marginal zones would be enhanced relative to interior regions by the marginal concentration reduction accompanying grain dispersive pressure.

This type of process might also explain the glomerocryst distribution in sill C, in which the concentration of glomerocrysts in the initial melt appears to have been below the

minimum concentration required for the onset of grain dispersive pressure. Magnus-type effects would tend to concentrate glomerocrysts towards the axis of the sill, and this migration may have been sufficient to raise the concentration to the threshold level. As discussed below, Magnus-type forces acting on glomerocrysts at the top of the sill would be augmented by gravitational settling. A further contribution to glomerocryst concentration in sill C, and to development of marginal zones of low glomerocryst content in other sills, may have been made by wall crystallization during flow. The effects of wall crystallization are impossible to evaluate in any kind of a quantitative manner, but this process may have contributed to the formation of marginal zones with few glomerocrysts through development of relatively narrow solid-liquid boundary layers. Based on earlier calculations, however, it is unlikely that effect of wall crystallization would extend farther than a few metres from sill contacts.

The maximum particle concentration in a natural suspension was estimated to be 65% by Komar (1972a). The maximum glomerocryst concentration observed in the measured profiles of the Howse Lake sills was 53% by area, which would translate into a slightly higher volume concentration. The average 2σ coefficient of variation in measured concentrations above 20% is 0.22. The maximum observed concentration ($53 \pm 12\%$) is thus within analytical uncertainty of the predicted maximum. A comparison of theoretical and measured profiles, however, suggests that concentrations in central highly glomeroporphyritic zones, which should be only slightly smaller than maximum values, are generally lower than predicted. Assuming that this discrepancy is real, there are three possible explanations for the lower glomerocryst concentrations. First, maximum observed glomerocryst concentrations in the sills underestimate true maximum values due to the vagaries of exposure. Assuming that maximum concentrations occur in the core phase, this is a reasonable interpretation since this lithology is only intermittently exposed in all sills examined. Because velocity gradients would be three dimensional in the sills, there should also be a three dimensional variation in glomerocryst content. Consequently it is possible that the core phase would not be developed as distinct horizon at a given position between upper and lower contacts, across the entire width and length of the sill. If, for example, the true three dimensional shape of a sill was cylindrical, then maximum glomerocryst content

would be observed only where the axis of the cylinder was intersected by the present erosional surface. Given that sill thickness and width are variable along strike, and that the level or position within the sill intersected by the erosional surface will change with width and thickness variations as well as with topography, it is unlikely that the horizon of maximum glomerocryst content would be everywhere exposed.

A second interpretation of the generally lower than predicted glomerocryst contents of central zones in the sills involves the increased viscosity and yield strength of magmas with high glomerocryst contents. The yield strength of intergranular fluids, and the viscous drag exerted on particles, were not considered in Komar's (1976) calculations of particle distributions. Both parameters will increase as apparent viscosity increases, and this may lead to a situation where, at high glomerocryst concentrations, the concentrating effect of grain dispersive pressure is limited by these forces. A third alternative is that maximum glomerocryst concentrations established during flow were somewhat dissipated during post-emplacement crystallization. Because of the high glomerocryst concentrations involved, however, intergranular distances would be quite small and thus very little movement would bring one given glomerocryst into contact with another. It is more probable that gravitational sinking or floating would tend to occur en masse, such that the closely packed aggregates in highly glomeroporphyritic material (e.g., the anorthositic phase) would sag into underlying sparsely glomeroporphyritic gabbro as unsolidified blocks or layers.

The asymmetric thickness of gabbroic phase horizons above and below zones of high glomerocryst content, and the asymmetric glomerocryst concentration gradients at the edges of these zones, probably reflect the effects of gravity. According to Komar (1972a), when particle and fluid densities differ, gravity will result in asymmetric concentration gradients during flow, and may modify concentration gradients after flow has ceased. In the case of particles having densities greater than that of the fluid, gravity will tend to drive particles downward in a horizontal sill such that the concentrations at the upper contact will be lower than those at the base. This produces a gradient in the grain dispersive pressure which, at equilibrium, will balance the force of gravity. As a consequence, the zone of maximum glomerocryst concentration is shifted from the sill axis towards the lower contact (Fig. 6.8). This will reduce the distance between maximum and minimum

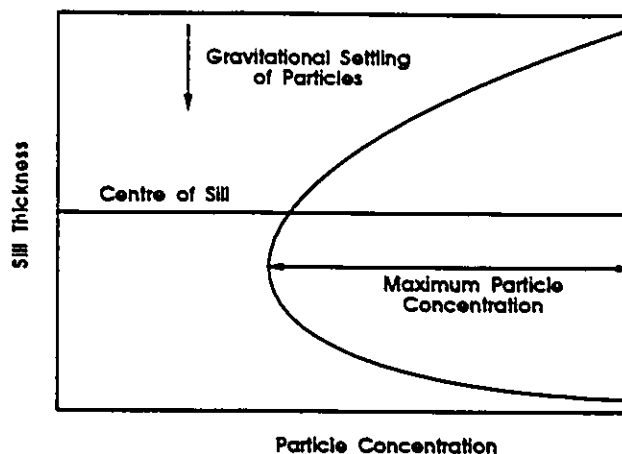


Figure 6.8. Concentration profile in a hypothetical sill in which gravity exerts a downward force on particles due to their greater density. The gravitational force results in an asymmetric concentration profile with the horizon of maximum concentration shifted toward the base of the sill. Modified after Komar (1972a).

concentrations in the lower part of the intrusion, and increase this distance in the upper part. The concentration gradient must therefore be larger in the lower part of the sill. This is precisely the type of asymmetry observed in sills B, C and D.

The development of asymmetric concentration profiles requires that particles are non-neutrally buoyant in the suspension. Since the plagioclase glomerocrysts will have densities that differ only slightly from the basaltic magma, it is worthwhile examining the rate at which the aggregates would sink or float. According to Stoke's Law, the terminal velocity (V) of a particle is given by:

$$V = \frac{2g}{9\eta} r^2 (\rho_L - \rho_S)$$

where g is the acceleration due to gravity, r is the particle radius, ρ_L and ρ_S are the densities of the liquid and the solid particles, respectively, and η is the fluid viscosity.

The average composition of glomerocryst plagioclase is An_{76} (Chapter 4), and this will have a density of approximately 2.67 g/cm^3 at 1200°C and atmospheric pressure (Kushiro, 1980). The average viscosity of aphyric Menihek magmas at 1200°C and atmospheric pressure is approximately 900 poise, and the density will differ little from the

Table 6.2. Calculated sinking velocities for plagioclase glomerocrysts (An_{70}) in basaltic magma with $\eta = 500$ poise. Distances travelled are based on a 60 hour emplacement time for a sill 20 km in length.

Glomerocryst Radius (cm)	Density Contrast (g/cm^3)	Sinking Rate (cm/hour)	Distance Travelled (m)
0.5	0.01	2.2	1.3
0.5	0.05	10.9	6.5
2.0	0.01	34.8	21
2.0	0.05	174.2	105

2.62 g/cm^3 value calculated for a temperature of 1250°C. With increasing depth of emplacement, the density of basaltic melt will increase more rapidly than that of plagioclase, and at emplacement depths of 15-20 km (~ 5 kbar) glomerocryst plagioclase would be less dense than the host magmas (Kushiro, 1980). Because the Menihek sills are thought to have been emplaced within unconsolidated or semi-consolidated sediments, however, it is unlikely that the density of the glomerocrysts was less than that of the magmas in any sill. Using density contrasts of 0.01 g/cm^3 and 0.05 g/cm^3 , sinking velocities have been calculated for two sizes of glomerocrysts in Table 6.2. Sinking rates range from 2.2 cm/hour for small glomerocrysts using the lower estimated density contrast, to 174 cm/hour for larger glomerocrysts using the higher density contrast. During emplacement of a sill 20 km in length at a rate of 0.1 m/s, glomerocrysts will travel distances of 1 to 100 m. For longer sills and larger glomerocryst dimensions these values will be higher. Consequently, it is quite likely that glomerocryst concentration gradients were influenced by gravity. This would explain the asymmetric glomerocryst profiles in the Menihek sills, as well as the asymmetric distribution of zones of high glomerocryst content relative to sill margins. This displacement may have been augmented by post-emplacement bulk settling of zones of high glomerocryst content.

The anorthositic phase lenses of elevated glomerocryst content which occur in the marginal gabbroic phase horizons of many sills are the only lithological variants which cannot be adequately interpreted in terms of flow differentiation. The origin of these lenses is not clear, but since they occur both above and below the central anorthositic phase horizons, gravitational settling is precluded. The most likely explanation for the lenses is

that they are the products of non-uniform velocity gradients developed within the sills during emplacement. Variations in the velocity gradients may have arisen through irregularity of wall rock surfaces or through non-constant flow rates. But it is uncertain as to how such variations would give rise to lenses of high glomerocryst content, and why such lenses would persist. Nevertheless, the fact that the general distributions of glomerocrysts in the Menihék sills are consistent with flow differentiation as modified by the effects of gravity suggest that the anorthositic phase lenses are also related to this process.

Three other lithological features related to flow differentiation should be mentioned here. First, the calculated concentration profiles of Komar (1976) are for the case where the suspension behaves as a Newtonian fluid. It is possible, however, that high concentrations of glomerocrysts would give rise to slightly pseudo-plastic behaviour in the melts, with the consequent development of more plug-like concentration profiles. This type of behaviour might explain the occurrence in some sills of relatively sharp contacts between the anorthositic phase (the central plug) and marginal gabbroic phase horizons (the relatively low viscosity boundary layers). The second lithological feature is the contrast between the systematic variations in glomerocryst size and frequency found in the Menihék sills, and the essentially homogeneous or random distributions in glomeroporphyritic basalts. Assuming that magmas in the feeder dykes to the basalts would be subject to the same types of forces as those flowing horizontally in sills, this contrast would seem to be a function of eruption mechanics. The most likely explanation for this difference is that flow upon eruption was turbulent, or at least convective in some manner, such that lavas were homogenized in terms of glomerocryst content. Finally, it was mentioned earlier that the distribution of olivine in sill C was similar to the distribution of plagioclase glomerocrysts. If the olivine grains represent pre-emplacement phenocrysts and micro-phenocrysts, they would be subject to the same flow differentiation effects as the glomerocrysts, and thus would exhibit a comparable distribution in the sill.

6.9 Summary

The Menihek glomeroporphyritic sills of the Howse zone exhibit systematic variations in glomerocryst size and frequency distributions which cannot have developed by *in situ* crystallization processes. Instead, this zonation must reflect some type of mechanical process acting on glomerocrysts entrained within the magmas during emplacement, or emplacement of magmas having varying glomerocryst loads. Multiple injection of magmas having variable glomerocryst contents cannot explain the variations in glomerocryst size with position in the sill, nor the gradational variations in glomerocryst concentrations across different lithologies. Wall crystallization is also unlikely to have played the predominant role in the development of glomerocryst zonation because the thicknesses of horizons accreted during emplacement will be minor in comparison to total sill thickness. The glomerocryst distributions are, however, qualitatively similar to predicted particle distributions resulting from flow differentiation during emplacement. This process will impart both concentration and size variations to entrained particles, and predicted variations in these parameters are comparable to those observed in the sills examined in detail at Howse Lake. If the modifying effects of gravity and Magnus-type forces are considered, and allowance is made for the limiting effects of increased fluid viscosity at high particle concentrations, the glomerocryst distribution profiles in the sills are remarkably similar to hypothetical profiles resulting from grain dispersive pressure. Consequently, flow differentiation during emplacement is considered the most probable source of the glomerocryst zonation patterns in the Menihek sills.

Part III
Petrology and Geochemistry of the LTBS

Chapter 7

Petrology of the LTBS

7.1 Petrology of the Intrusive Rocks

The LTBS includes three sets of gabbroic sills, the Le Fer, Menihek and Doublet gabbros, which were emplaced within volcano-sedimentary sequences that are now present in discrete, tectonically bounded, domains. The petrology of the Menihek gabbros was described in Chapter 4. The Le Fer and Doublet intrusions were sparsely sampled on a regional basis primarily to determine their relationships to spatially associated extrusive rocks. Although the sample suites are limited, field observations suggest that they are representative of the predominant lithologies in the two sets of intrusions.

Le Fer Gabbros

The Le Fer gabbros form 50-600 metre thick sills which tend to be lithologically homogeneous over much of their stratigraphy. Basal contacts are exposed in two of the sills examined, but upper contacts were not observed in this study. Where contact relations could be determined, the sills were found to intrude Le Fer sediments, but the sills also occur in the underlying arkosic rocks of the Seward subgroup (Baragar, 1967; Rohon *et al.*, 1993). Basal chilled margins are present in the two sills examined in which the lower contacts are exposed, and these consist of aphanitic to fine grained, equigranular gabbroic rock comparable to that found at the lower contacts of the Menihek sills. Grain size increases away from the basal contacts, and within a few metres the medium grained subophitic gabbroic rocks which predominate in the intrusions are present. In thin section, the chilled margins consist of a random arrangement of completely altered plagioclase microlites embedded in a fine grained dark matrix in which the only recognizable primary phase is clinopyroxene. The latter mineral is usually largely replaced by a mixture of actinolite and chlorite. In comparison with chilled material from the Menihek sills, the quenched rocks of the Le Fer gabbros exhibit a greater degree of alteration. No phenocrysts or pseudomorphic replacements of phenocrysts were observed.

Samples from the interiors of the Le Fer sills consist largely of medium grained, subophitic to ophitic gabbroic rocks. These are reddish brown to brownish grey on weathered surfaces, and greenish grey to grey on fresh surfaces. In the Tait Lake area, several sills contain 0.5-1.0 cm sized, dark green blebs which appear to be serpentized pseudomorphs of either olivine or hypersthene phenocrysts. Comparison with the Le Fer basalts of the same area, which are often olivine-phyric, suggests that the serpentine blebs were originally olivine phenocrysts. Plagioclase phenocrysts are also present, although these appear to be less common.

Groundmass textures in the medium grained gabbroic rocks from the interiors of the intrusions are typically subophitic, with clinopyroxene partially enclosing, and intergranular to, plagioclase, alkali feldspar and quartz. Less commonly, well developed ophitic textures are present. As in the chilled margins, the degree of alteration is relatively high, and modal classification is usually impossible. In general, Fe-Ti oxides and, where present, quartz and alkali feldspar, are the only phases that are commonly pristine. Alteration of plagioclase is often complete, and most grains are a mass of very fine grained, semi-opaque material which is probably largely composed of albite and clinozoisite. Some crystals, however, exhibit clear rims of sodic plagioclase. Since the central regions of these crystals are completely altered, the rims probably represent late magmatic overgrowths rather than albite formed during alteration and metamorphism. Clinopyroxene is partially to completely replaced by actinolite and chlorite, and is occasionally overgrown by hornblende. Fe-Ti oxides may be partially replaced by granular sphene at grain margins, but are usually unaltered.

Neither hypersthene nor olivine were observed in the samples examined, but intergrowths of serpentine and chlorite probably represent the alteration products of at least one of the former phases. Unlike similar occurrences in the Menihek sills, where the shapes of the pseudomorphic aggregates are often distinctive, there is no clear indication of the nature of the primary mineral in the Le Fer samples. Since groundmass augite grains tend to be partially altered to actinolite and chlorite rather than serpentine, it is likely that the latter phase replaces either olivine or hypersthene. The absence of fine grained magnetite associated with the serpentine aggregates, which is typical of altered olivine in sill C at Howse Lake, suggests that the precursor grains were hypersthene. If this is the

case, then the Le Fer sills, like the Menihek intrusions, were initially composed largely of hypersthene gabbro.

Quartz gabbro and quartz monzogabbro are common but volumetrically minor rock types in the sills, although whether the quartz is magmatic or metamorphic in origin is uncertain. These rocks contain 5-10% quartz and orthoclase as anhedral, equant grains which are easily distinguished from plagioclase because of their unaltered natures. The occurrence of granitic veins in some sills (Baragar, 1967) suggests that *in situ* differentiation produced felsic melts during crystallization of the intrusions, and this process may have also produced intermediate rock types. As described above, there is some evidence of compositional zonation in plagioclase which probably reflects the fractional crystallization of this phase. Fractional crystallization was interpreted to have led to the development of felsic rocks in the Menihek sills, and a similar evolution in melt composition is possible in the Le Fer gabbros. It is also possible that detailed examination of the Le Fer sills would reveal a somewhat more diverse range of rock types than described here.

Doublet Gabbros

The two sills examined in the Doublet terrane are relatively homogeneous bodies of metagabbro which intrude black, pyritiferous shale. Both sills exhibit narrow basaltic chilled margins at basal contacts. The chilled margins are composed of albitized plagioclase microlites enclosed by a fibrous mass of fine grained chlorite, pumpellyite, clinozoisite and sericite. Fe-Ti oxides are the only phases which appear to be primary, and these occur as fine granular crystals scattered throughout the rock. Within 1 m of the contact, the chilled margins of both sills have graded to fine to medium grained metagabbro. The upper contacts were not observed.

The two intrusions are principally composed of fine to medium grained gabbroic rock which has undergone greenschist facies metamorphism. Many samples exhibit cm-sized grains of blue metamorphic quartz, but no felsic differentiates were observed. The only significant lithological variation found was layering in a zone of leucogabbro in the lower part of one sill. In outcrop the phenomenon appears as irregular, 5-15 mm thick, light coloured laminae in slightly darker rock. In thin section, the layering is seen to be

defined by variations in grains size and modal proportions. In the light coloured layers, small plagioclase laths enclose larger equant grains of tremolite and lesser muscovite. The plagioclase content of the darker layers is greater, and the size of tremolite grains is smaller. The darker colour derives from the greater abundance of altered plagioclase, which is pseudomorphically replaced by a cryptocrystalline aggregate of semi-opaque material. The colour difference is also enhanced by the smaller amount of colourless tremolite in the darker rock. The leucogabbro was observed in a single outcrop and its relationship to surrounding metagabbro is unknown

Despite the degree of metamorphic recrystallization, the primary textures of the gabbroic rocks are generally preserved. Two textures predominate, and both are defined by the relationship between plagioclase and clinopyroxene. In fine grained metagabbro, pseudomorphically altered plagioclase laths are intergrown with granular to elongate augitic clinopyroxene grains. In medium grained rocks, the plagioclase laths are subophitically enclosed by augite. The clinopyroxene grains in both lithologies are partially to completely replaced by chlorite and actinolite. Fe-Ti oxides, partially replaced by granular sphene, occur in the interstices in both fine and medium grained rocks. Both lithologies also contain abundant fibrous to shredded aggregates of chlorite, actinolite, clinozoisite, pumpellyite and sericite or muscovite. The precursors of these secondary phases are unknown, but comparison with the Menihek gabbros suggests that hypersthene is one possibility. Quartz occurs as rounded anhedral grains and angular intergranular material in all samples examined, and in some this phase comprises up to 15% of the mode. The quartz is thought to be largely of metamorphic origin..

7.2 Petrology of the Volcanic Rocks

Le Fer Basalts

The Le Fer basalts commonly occur as relatively thin (50-100 m) horizons of essentially continuous basalt embedded within thick sequences of gabbro, which appear to represent intrusive bodies. As noted, it is generally difficult to establish contacts between basaltic and gabbroic sequences, and intervening sediments are rarely exposed. In the

Corrugated Hills, gabbroic rocks typically underlie ridges separated by narrow valleys often hosting elongate lakes. The topographic depressions probably reflect the presence of sedimentary rocks, but these are not exposed. Basalts may also underlie ridges, but are more often found on the flanks of ridges or adjacent to the shores of the intervening lakes. The extent of individual flows is generally impossible to determine within a volcanic horizon since the basalts tend to be texturally homogeneous and interflow sediments rarely outcrop. Dimroth (1978), however, reported that individual flows are 2-15 m thick and may be separated by minor thicknesses of interflow shales.

The Le Fer volcanics range from massive aphyric basalts to medium grained subophitic gabbro. The basalts often contain phenocrysts of altered olivine, plagioclase and pyroxene, and some show a mottled texture resulting from the clustering of phenocrysts and microphenocrysts. Olivine phenocrysts are more abundant in the Tait Lake basalts, while pyroxene phenocrysts predominate in the flows of the Corrugated Hills. If the interpretations of relative stratigraphic positions discussed earlier are correct, this difference may indicate a change in the predominant phenocrystic ferromagnesian mineral from olivine in the basal basalts to pyroxene in the upper volcanic sequence.

The weathered surfaces of the Le Fer basalts vary in colour from reddish brown to tan or brownish grey, and fresh surfaces are pale grey to dark greenish grey. Well developed hexagonal columnar jointing is occasionally present and pillowed flows are not uncommon, although these are less abundant than massive basalts in both areas examined. The pillows are relatively small, ranging from 20-50 cm in diameter, and white to grey weathering siliceous selvages are usually present. The Le Fer basalts tend to be non-vesicular and non-amygdaloidal. This fact, coupled with the occurrence of pillows, the paucity of pyroclastic material and the presence of interflow pelitic sediments, indicate that the basalt flows are submarine and were erupted at considerable depth.

Most of the Le Fer samples obtained in this study are aphanitic to fine grained hypocrySTALLINE rocks consisting principally of augite, plagioclase and devitrified glass. Intersertal glass is now represented by a cryptocrystalline, semi-opaque material of indeterminate mineralogy. Neither olivine nor orthopyroxene were observed in the groundmass of the basalts, although olivine may comprise up to 5% of the mode as phenocrysts. Fe-Ti oxides are ubiquitous accessory phases, and many samples contain

minor amounts of intergranular orthoclase, quartz, and sphene. With the exception of the phenocrysts the basalts are often quite fresh, with relatively minor chloritization of augite and albitization of plagioclase. In some samples, augite occasionally displays rims of a pale brown amphibole which may be a replacement feature.

Texturally, the basalts are quite variable. Most samples show a decussate framework of acicular to prismatic plagioclase crystals with interstitial granular augite, irregular Fe-Ti oxide grains and devitrified glass. The relative proportion of augite to glass varies from aphanitic intersertal basalt with little or no pyroxene, to fine or medium grained subophitic basalt with little or no glass. In some samples the subparallel arrangement of plagioclase grains gives rise to trachytic textures, while in others this phase may be intergrown with elongate, feathery augite crystals in more or less radial patterns. Less commonly, anhedral plagioclase is intergrown with equant augite grains to give a granular texture. As noted, many of the samples examined contain moderately abundant olivine, plagioclase and pyroxene phenocrysts which show strong alteration. Olivine phenocrysts are completely pseudomorphed by serpentine $\pm$ magnetite, and plagioclase phenocrysts have been completely replaced by clinozoisite and albite. The pyroxene phenocrysts were usually identified as such by the typically euhedral outlines of the pseudomorphic assemblages. In some grains, however, zones of unaltered clinopyroxene have been preserved. More commonly, the phenocrysts have been completely replaced by chlorite and actinolite, and many of the pseudomorphed grains contain bands of serpentine. This may indicate that some of the phenocrysts were orthopyroxene, or it could reflect alteration of clinopyroxene phenocrysts containing exsolved orthopyroxene.

Lower Menihek Basalts

On the basis of stratigraphic level and lithology, volcanic rocks of the Menihek Formation are subdivided into two groups, informally termed the upper and lower Menihek basalts. The more voluminous and widespread of the two is the upper horizon, which caps the formation. The upper sequence consists predominantly of basaltic flows, while the lower sequence constitutes a more limited thickness of flows, pyroclastic and volcaniclastic rocks. The upper basalts are most extensive in the northern part of the Howse zone, whereas the lower basalts were identified *in situ* only at Howse Lake to the

south. Clasts of similar lithology, however, occur in the conglomeratic horizon which underlies the upper basalts near Outlook Lake. This may indicate that similar rocks were locally erupted throughout the Howse zone, but were largely removed by uplift and erosion.

The best exposures of the lower Menihek basalts and pyroclastic rocks are found along the western and southern shores of Howse Lake where they are interbedded with volcanoclastic and sedimentary rocks. Five flows were identified in this area, and these are intercalated with approximately 100 m of turbidites, mafic agglomerate and tuff, and minor conglomerate. The pyroclastic rocks occur as thin gradational horizons showing considerable variation in lithology. A section at one location along the western shore of Howse Lake exposes mafic agglomerate containing basalt bombs occasionally exceeding 50 cm in length, overlain by lapilli tuff and culminating with a graded volcanoclastic horizon.

The conglomerates are chiefly composed of pebble to cobble sized, angular to ellipsoidal clasts of basalt and lesser red chert set in a chloritic matrix. The finer fragmental rocks consist of angular lapilli embedded in a similar chloritic matrix. The fragments consist of glassy shards and aphanitic to fine grained basaltic material showing intergranular to trachytic textures. In basaltic fragments, plagioclase forms elongate to acicular grains with interstices filled by actinolite or chlorite. A common feature of the basaltic clasts is the abundance of acicular to dendritic magnetite grains. Agglomerates are less abundant than the finer fragmental rocks, but at least two thin horizons of this lithology are exposed at Howse Lake. The bombs range from 10 cm to 50 cm in diameter and are generally ellipsoidal and smooth surfaced. Texturally these resemble the basaltic flows associated with the fragmental rocks, and initially the agglomerate horizons were thought to be pillowed basalt. The matrix to the bombs, however, is similar to that of the finer fragmental rocks.

Basaltic rocks are exposed in slight ridges or mounds bounded by swampy areas, and consequently no contacts were observed. Based on the width of exposures, it is doubtful that any flow exceeds 10 m in thickness. Despite the limited amounts of volcanic rocks, there is considerable variation in appearance. The weathered surfaces of the more massive flows are generally dark grey, while highly amygdaloidal material is a lighter

greenish grey. The flows range from dense and massive aphanitic rock with sparse, relatively large amygdales (Fig 7.1A), to medium grained subophitic material containing abundant millimetre to centimetre sized amygdales. The amygdales are filled by carbonate, chlorite, zeolites and quartz. In many exposures, amygdales are pervasively developed throughout the flows, and this texture is at least locally present in all of the basalts examined. One outcrop shows 10-20 centimetre sized pillows with cherty selvages, but these structures are generally absent. Porphyritic varieties were not observed

In addition to their amygdaloidal nature, a characteristic feature of the basalts, as well as the pyroclastic fragments, is strong magnetism. This phenomenon is the result of an abundance of groundmass magnetite, which forms squarish to irregular polygonal grains in well crystallized lithologies, and acicular to dendritic crystals in finer grained rocks. The groundmass also contains abundant plagioclase as lathy to acicular crystals which are generally fresh. The basalts are characteristically intergranular, with deeply coloured titanite occupying interstices between plagioclase laths. Sphene is a consistent accessory phase, occasionally comprising as much as 10% of the rock. Apatite and zircon also occur as abundant accessory phases. In aphanitic lithologies plagioclase and magnetite are the only primary phases, and the remainder of the rock consists of shredded chlorite and actinolite, and poorly crystallized sphene and leucosene. Coarser grained basalts, however, are quite commonly pristine.

Clasts within the conglomeratic horizon which underlies the upper Menihek basalts near Outlook Lake are dominantly amygdaloidal and strongly magnetic basalt similar to the lithologies described above. The conglomerate is coincident with a horizon identified as pyroclastic rocks on regional maps (Baragar, 1967; Dimroth, 1978), but only conglomerates were observed at Outlook Lake. Possibly both conglomerates and pyroclastic rocks are intercalated in the horizon. At Outlook Lake, the conglomerate consists predominantly of ellipsoidal cobbles of amygdaloidal to massive basalt and cherty iron formation set in a sheared chloritic matrix (Fig. 7.1B). The conglomerate probably represents the erosional products of local uplift, possibly related to lower Menihek volcanism. Similar basalt- and jasper-bearing cobble conglomerates occur in the Nimish Formation of the Schefferville zone, and these have been interpreted as alluvial fans formed around subaerially exposed volcanic centres (Evans, 1978).

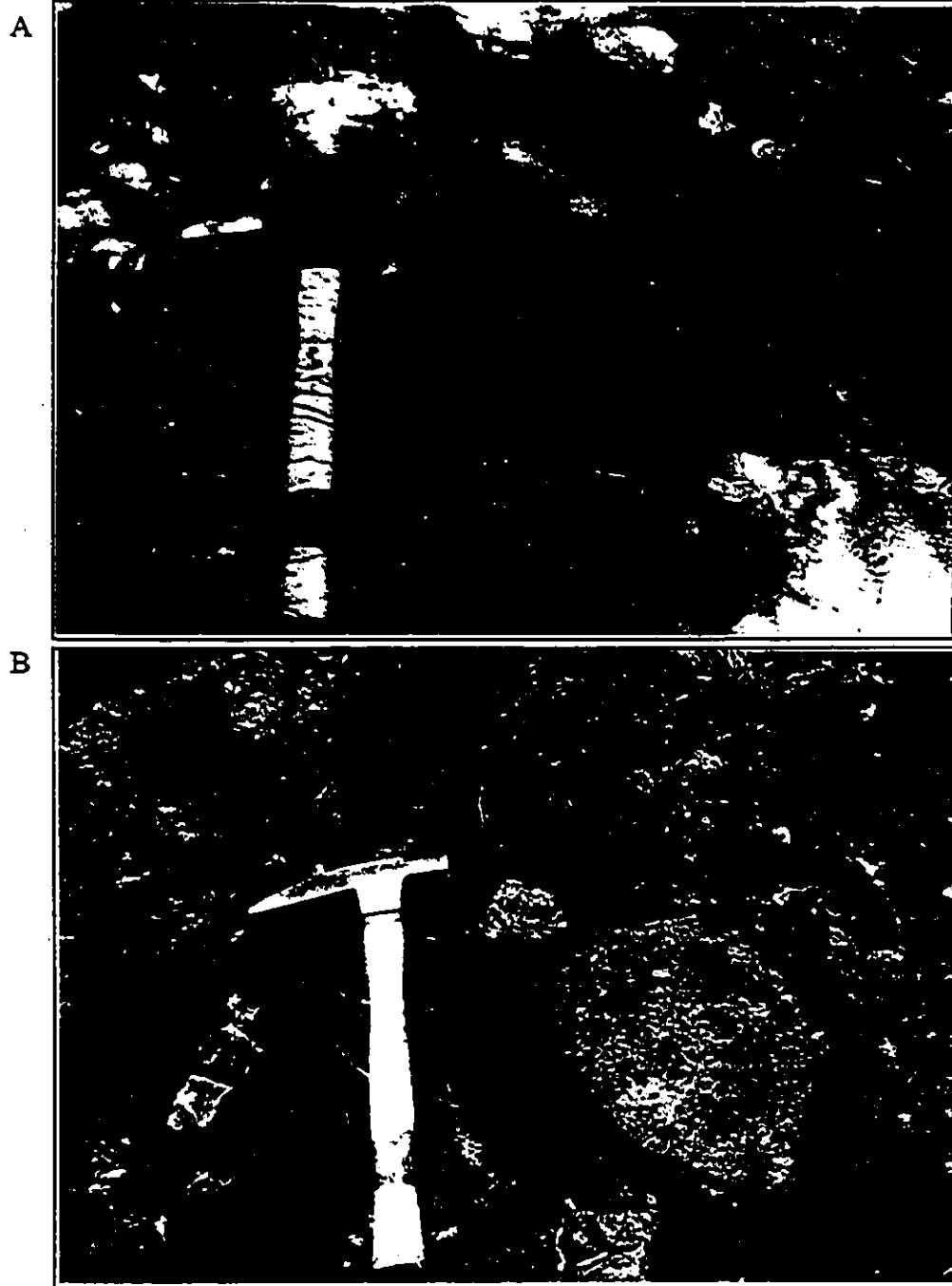


Figure 7.1. (A) Large spherical, calcite and chlorite filled amygdales in a lower Menihok basaltic flow. Hammer for scale. (B) Ellipsoidal clasts of amygdaloidal basalt in the Outlook Lake conglomerate. Hammer for scale.

The juxtaposition of pyroclastic rocks and highly amygdaloidal basalts at Howse Lake suggests that they formed either subaerially or at shallow water depths where confining pressures were insufficient to suppress gas exsolution and explosive magmatism. The presence of at least one pillowed flow, along with interflow turbiditic sediments indicates that the latter interpretation is more likely. Thus it is probable that the lower Menihek basalts were extruded under moderately shallow water conditions. This is consistent with the occurrence of the lower Menihek volcanic and pyroclastic rocks immediately above the Sokoman-Menihek transition, which records the deepening of the sedimentary basin. The occurrence in the Howse Lake and Outlook Lake areas of conglomerates containing basaltic clasts lithologically comparable to the Howse Lake flows indicates subsequent, or perhaps ongoing, uplift and erosion. Dimroth (1978) also considered deposition of the Menihek Formation to have occurred under tectonically unstable conditions, and Baragar (1967) concluded that much of the clastic sediments of the Menihek Formation were derived from a volcanic source. Local uplift and erosion of the lower Menihek sequence may have provided some of this detrital material. It is therefore probable that the lower Menihek basalts reflect magmatism associated with tectonic activity and general basin subsidence. It is interesting to note that conglomerates are also found in the Nimish and Murdoch formations, and this suggests that local uplift also occurred during the deposition of these sequences. It should also be pointed out that the Nimish volcanic sequence exhibits many lithological similarities to the lower Menihek basalts (cf. Evans, 1978). The Nimish basalts are generally massive to amygdaloidal and rarely pillowed. They are interbedded with agglomerates and tuffaceous rocks and are often strongly magnetic; some basalts even contain magnetite phenocrysts (Evans, 1978).

Upper Menihek Basalts

The upper Menihek basalts in the northern part of the Howse zone are composed of relatively homogeneous flows which are similar to those of the Le Fer Formation. Where contacts were established, individual flows were 2-10 m thick, but the occasional occurrence of medium grained subophitic texture suggests that thicker flows are present. The basalts are commonly separated by narrow intervals of dark greywacke, siltstone and shale, and the latter often contain centimetre-thick lenses and bands of massive pyrite. The

basalts are generally aphanitic to fine grained massive rocks, with grey to brownish grey weathered surfaces and grey to dark grey fresh surfaces. Pillowed flows were identified above the conglomeratic horizon at Outlook Lake, and near the top of the formation at Murdoch Lake, but appear to be less common than massive flows. As with the Le Fer basalts, the presence of pillows, the paucity of vesicles and amygdales, and the occurrence of interflow greywacke, siltstone and shale indicate that the flows were erupted in a deep water setting.

Although the upper Menihek basalts are similar to those of the Le Fer Formation, there are three general differences. The most striking of these is the occurrence of plagioclase glomeroporphyritic flows in the upper Menihek sequence. This lithology was observed at two levels in the upper part of the basalt sequence immediately west of Murdoch Lake, but such rocks do not appear to be a volumetrically significant part of the suite at this location. Dimroth (1978), however, noted the common occurrence of glomeroporphyritic basalts in the northern part of the domain (Jean Lake). The glomeroporphyritic basalts contain sparse to abundant plagioclase glomerocrysts set in an aphanitic to fine grained groundmass. The glomerocrysts range in size from 0.5 to 5 cm, and in outcrop form white to pale greenish knots in a dark groundmass. The aggregates appear to be randomly but homogeneously distributed within the flows, and the lithology is often quite similar to sparsely glomeroporphyritic marginal gabbro in the Menihek sills. Glomerocryst plagioclase is invariably pseudomorphed by clinozoisite and albite or, less commonly, fine grained, shredded aggregates of sericite. As in the Menihek gabbros, many of the glomerocrysts in the basalts show discontinuous marginal zones of unaltered sodic plagioclase. The Le Fer basalts, although not glomeroporphyritic, commonly contain abundant olivine, plagioclase and pyroxene phenocrysts. In contrast, only a few of the upper Menihek samples examined are porphyritic. Plagioclase is the most common phenocrystic mineral, although both pseudomorphed olivine and clinopyroxene phenocrysts are occasionally present. In most cases the phenocrysts are highly altered, and the secondary assemblages are similar to those found in the Le Fer basalts. The third principal difference between the basaltic rocks of the two formations lies in the generally greater degree of shearing and veining found in the upper Menihek suite.

The upper Menihek basalts are characteristically hypocrySTALLINE rocks containing subequal amounts of augite and plagioclase. As noted, olivine occurs as occasional phenocrysts in a few samples, but was not recognized within the groundmass. Hypersthene is also absent from the groundmass, although this may be an artefact of alteration. In aphanitic and finer grained samples, plagioclase forms acicular, often skeletal, grains intergrown with equant augite crystals. Interstitial areas are filled by devitrified glass and oxides. In coarser grained varieties, augite occurs as curved and branching crystals intergrown with plagioclase, or as irregular grains subophitically enclosing plagioclase. These variations appear to reflect cooling rates, and are similar to the textural range of the Le Fer basalts. Oxides are invariably present as intergranular crystals, and many samples contain accessory pyrrhotite as fine disseminations.

The primary mineralogy of the basalts has undergone variable degrees of alteration. In fine grained samples, plagioclase and augite are usually well preserved, although some albitization and chloritization have occurred. Intersertal glass is now a very fine grained aggregate of indeterminate mineralogy. In coarser varieties alteration is more pervasive, and plagioclase is largely pseudomorphed by fine grained to cryptocrystalline clinozoisite and albite. Clinopyroxene is typically partially to completely replaced by chlorite and actinolite. Several samples contain fine quartz veins, and in these the primary phases have been completely replaced and original textures destroyed in the vicinity of the veins.

Murdoch basalts

Basaltic rocks are poorly exposed and appear to be subordinate to pyroclastic rocks in the Murdoch Formation. The samples used in this study were obtained from outcrops north of Irene Lake and east of Connolly Lake. At the Irene Lake location, basalts are exposed on a low ridge presumably due to their greater resistance to erosion than surrounding schistose pyroclastic rocks. The exposure is estimated to represent 200-400 stratigraphic metres of basalt, although no flow contacts were observed and there is the potential for some thickness of interflow sedimentary rocks. With the exception of the limited occurrence of basalts at Connolly Lake, the Murdoch Formation was composed predominately of schistose pyroclastic rocks in the other exposures examined.

Basalts exposed in the Irene Lake and Connolly Lake areas consist of massive, greenish brown to greenish grey weathering rocks. Fresh surfaces reveal greenish aphanitic basalts with locally abundant quartz and chlorite filled fractures. The Irene Lake flows are generally well pillowed (Fig. 7.2A), although a hackly surface due to fracturing and cleavage locally obliterates these structures. The pillows range in from 15 cm to 2 m and are highly elongate, presumably due to the deformation which has imparted the schistosity to the Murdoch rocks in other areas. In the northern part of the Doublet terrane, Dimroth (1978) found the Murdoch Formation to consist largely of pillowed basalt similar to that described from the Irene Lake exposures, with subordinate interflow tuffs and volcanic breccias. In contrast, basalts exposed at Connolly Lake area are not pillowed, although they are otherwise similar to the Irene Lake basalts. These observations are consistent with the descriptions of Frarey (1967) who found the basalts of the Murdoch Formation in the Thompson Lake area to be massive flows with no primary volcanic structures or secondary foliation.

The igneous rocks located to the east of the Walsh Lake thrust have undergone relatively strong recrystallization during greenschist facies regional metamorphism. In comparison to other members of the LTBS, the Murdoch basalts generally exhibit more extensive alteration. In many samples little remains of primary igneous textures, and the rocks are fine grained intergrowths of actinolite, chlorite, albite, epidote, quartz and sphene. Other samples exhibit relict intergranular to subophitic textures despite complete replacement of the primary mineral assemblage. Metamorphism has resulted in the replacement of pyroxenes by actinolite and chlorite, and these two minerals are the most abundant phases in most of the rocks. Primary plagioclase has been strongly altered, resulting in either a fine grained, brownish pseudomorphic aggregate, or well crystallized clinozoisite and albite. Iron oxides have been replaced by a cloudy semi-opaque material which is probably a mixture of sphene, leucoxene and possibly hematite. Although some samples contain minor quartz which appears to be of secondary origin, there is generally little evidence of pervasive chemical reconstitution, particularly in comparison to the heavily veined and carbonated pyroclastic schists which predominate in the Murdoch Formation.

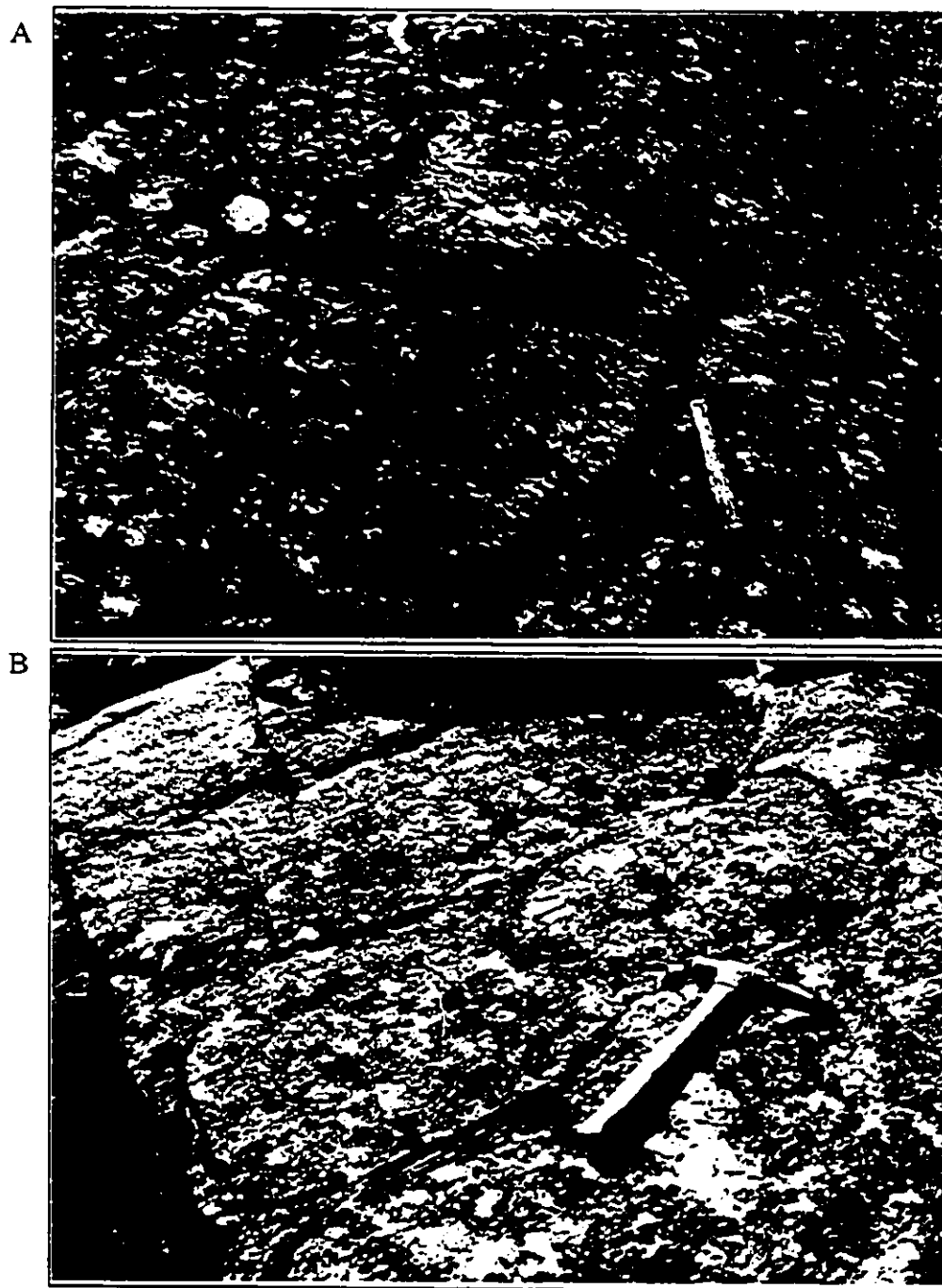


Figure 7.2. (A) Elongate pillows in Murdoch basalt forming a ridge north of Irene Lake. Dark, cherty sediment between pillows weathers recessively. Hammer for scale. (B) Large pillows in Willbob basalts from east of Retty Lake. Hammer for scale.

Willbob basalts

The Willbob basalts contrast with those of the Murdoch Formation in that they form the predominant lithology of the formation, are extremely well exposed for much of the length of the Doublet terrane, and are relatively unfoliated. They include massive and pillowed basalts (Fig. 7.2B) which appear to alternate with no regular pattern. Interflow sedimentary rocks were observed only at the bases of a few gabbroic sills within the sequence. In general, the basalts form a series of thick sequences exposed as relatively flat ridges separated by narrow valleys and lakes which probably overlie sedimentary intervals. Plagioclase glomeroporphyritic and porphyritic basalts occur in the lower part of the sequence south of Retty Lake, and these have an estimated stratigraphic thickness of 100 m. Baragar (1967) identified three such intervals within the Willbob succession in the Ahr Lake area, but these lithologies are relatively minor in comparison to massive and pillowed flows. Rohon *et al.* (1993) identified four breccia horizons in the Willbob succession and concluded that at least five volcanic cycles are preserved. Although only three glomeroporphyritic horizons have been recognized, it is possible that each volcanic cycle included eruptions of glomeroporphyritic lavas.

The basalts range from aphanitic to medium grained, and equigranular to intergranular. Both massive and pillowed flows show brownish to greyish weathered surfaces, and generally grey to greenish grey fresh surfaces. The rocks are unfoliated in the areas traversed, although fracturing and minor quartz veining is not uncommon. Many fracture surfaces show thin coatings of pyrrhotite, and this sulphide is commonly disseminated throughout the rock. Glomeroporphyritic basalts are characterized by generally sparse plagioclase aggregates which are homogeneously distributed across outcrops. These flows may exhibit gabbroic textures and grain size, although aphanitic groundmass material is also common. It is possible that some of the gabbroic lithologies represent intrusive rocks, although such relationships were not obvious in the field. It seems probable, however, that the coarser grained glomeroporphyritic samples are thick flows rather than sills since no demonstrably intrusive rocks exhibit the texture, and since the anorthositic phase which is characteristic of highly glomeroporphyritic Menihek sills was not observed.

The mineralogy of Willbob basalts is essentially identical to the secondary assemblage described for the Murdoch basalts, and there is again little petrographic evidence for pervasive compositional change. Unlike the Murdoch basalts, however, those of the Willbob Formation invariably exhibit relict igneous textures. The most prevalent primary lithology was probably fine grained intergranular basalt, with anhedral equant clinopyroxene grains interstitial to plagioclase laths. Plagioclase is now pseudomorphically replaced by clinozoisite and albite, and pyroxene has been replaced by shredded aggregates of actinolite and chlorite. Relict trachytic textures, in which pseudomorphs of plagioclase microlites exhibit subparallel orientations, are occasionally present. Interstices may have been filled by glass in some lithologies, and a mixture of glass and pyroxene in others. A few samples consist of a confused intergrowth of very fine grained material which could not be resolved except for the occasional coarser grains of actinolite, albite or epidote. These may represent originally glassy rocks. Other samples suggest that medium grained, subophitic basalts were not uncommon, particularly in porphyritic or glomeroporphyritic flows. In the latter lithology, plagioclase glomerocrysts are completely replaced by various mineral assemblages including sericite and semi-opaque, cryptocrystalline clinozoisite. The glomerocrysts range from <1 cm to about 3 cm in diameter, and locally comprise up to 20% of the rock. A few basalts contain serpentine-chlorite-magnetite intergrowths which appear to be pseudomorphic after olivine phenocrysts, but no primary olivine was observed. Polygonal pseudomorphic phenocrysts which appear to have been composed originally of pyroxene are also occasionally present. In the relative paucity of ferromagnesian phenocrysts, as well as in the occurrence of plagioclase glomeroporphyritic lithologies, the Willbob basalts are similar to those found in the upper Menihek Formation.

7.3 Summary

The Le Fer gabbros are a relatively homogeneous group of sills which intrude the sedimentary rocks of the Seward subgroup and overlying Le Fer Formation. The sills appear to be composed predominantly of gabbro and hypersthene gabbro, with minor quantities of quartz gabbro and quartz monzogabbro. The occurrence of zonation in

plagioclase grains, and the presence of granophyric rocks suggest that the sills underwent in situ fractional crystallization during solidification. The Doublet gabbros are also relatively homogeneous bodies of gabbroic rock, but these sills have undergone more intense alteration and metamorphism. The mineral assemblage consists largely of secondary actinolite, chlorite, clinozoisite, albite and sericite, although some primary augite and plagioclase is locally preserved. The Le Fer and Doublet gabbros are texturally similar to some of the Menihek intrusions, but lack the glomeroporphyritic lithologies which characterize the latter.

The Le Fer basalts include massive to pillowed flows which range from aphanitic to medium grained, and from equigranular to porphyritic. Phenocrysts and microphenocrysts consist principally of plagioclase, olivine and clinopyroxene, with olivine being more prevalent in samples from the southern part of Le Fer zone. Groundmass clinopyroxene and plagioclase are often pristine in these rocks, but unaltered olivine or hypersthene were not observed. The Le Fer basalts may be pillowed, are typically intercalated with pelitic sediments and contain relatively few amygdales. These features, coupled with the paucity of pyroclastic rocks, indicate a relatively deep water extrusive environment.

The lower Menihek volcanic sequence includes basaltic flows, pyroclastic rocks, and conglomerates in which basaltic clasts predominate. Basaltic clasts in both fragmental rocks and conglomerates are lithologically similar to the basaltic flows, and were presumably derived from a similar source. The flows themselves are rarely pillowed and often highly amygdaloidal. These features suggest that the lower Menihek basalts erupted under relatively shallow water conditions, possibly related to emergent volcanic centres. Erosion of these local topographic highs may have provided material for the interbedded conglomerates and volcanoclastic rocks. The occurrence of similar conglomerates at Outlook Lake in the northern part of the Howse zone suggests that lower Menihek magmatism was not restricted to the Howse Lake area.

The Murdoch Formation exhibits some similarity to the lower Menihek sequence in that mafic pyroclastic rocks are abundant and the intercalated sediments are turbiditic. But the majority of Murdoch basalts are well pillowed and lack the strong magnetism of the lower Menihek volcanics. The flows at Connolly Lake, however, are massive. It is

noteworthy that the Connolly Lake basalts occur in the stratigraphically lowest exposed part of the Murdoch Formation, while the Irene Lake basalts represent a higher stratigraphic level. All of the Murdoch basalts examined differ from the Le Fer, upper Menihek and Willbob flows in their strong alteration and occasionally foliated nature. In addition, pyroclastic rocks are much less abundant in the latter sequences.

The upper Menihek and Willbob basalts share many of the textural and stratigraphic characteristics of the Le Fer suite, and would also appear to represent submarine sequences formed at considerable depths. The upper Menihek and Willbob basalts, however, exhibit a greater degree of alteration and are more commonly aphyric. In porphyritic varieties, the predominance of plagioclase and rarity of olivine phenocrysts contrasts with the Le Fer basalts. The cycle 2 suites also differ from the Le Fer volcanics in the occurrence of plagioclase glomeroporphyritic flows. These extrusions are texturally similar to sparsely glomeroporphyritic marginal phase gabbro found in many of the Menihek sills.

Chapter 8

Whole Rock Geochemistry of the LTBS

8.1 Nature of the Data Set

Significant thicknesses of basalts occur in the Le Fer, Menihek, Murdoch and Willbob formations of the central Labrador Trough, while limited quantities of volcanic rocks are found in the Nimish and Chakonipau formations. In preceding sections, it was noted that these suites occur in four separate lithotectonic domains, and hence their stratigraphic relationships are obscured. It was also noted that the Le Fer, upper Menihek and Willbob basalts are lithologically similar. Geochronological work, however, has shown that there is a significant time interval between formation of cycle 1 and cycle 2 tholeiitic gabbro intrusions. If the gabbroic rocks are comagmatic with the tholeiitic basalts, then there is also a significant interval between eruption of the volcanic sequences associated with each cycle (Clark and Thorpe, 1990; Rohon *et al.*, 1993; Findlay *et al.*, 1995). Of the cycle 2 igneous suites, only the ages of the Menihek glomeroporphyritic sills (1884-1874 Ma) and the Nimish volcanic rocks (1879 Ma) are known (Machado, 1990; Findlay *et al.*, 1995). There have been no precise radiometric ages from the Doublet Group succession. Thus while it is clear that cycle 1 and cycle 2 volcanism in the Knob Lake Group represent two distinct magmatic episodes, the relationships among strata from the various tectonic domains remain poorly constrained.

This chapter tests possible relationships among the intrusive and extrusive volcanic suites through comparison of their geochemical characteristics. The source of compositional variations within each suite is also examined. The geochemical characteristics of the extrusive members of the LTBS were determined through whole rock analysis of the major and trace element compositions of 107 basalts. Whole rock analyses were also obtained for a limited number of samples from the Le Fer and Doublet gabbros. Included in some of the following discussions are the basaltic rocks of the Nimish Formation, whose compositions have been determined by Watanabe (personal

communication, 1993). The Chakonipau basalts are poorly exposed and highly altered, and consequently were not sampled in the present study

8.2 General Features

Le Fer, Upper Menihek, Willbob and Murdoch Basalts

The similarities in lithology and occurrence of the Le Fer, upper Menihek and Willbob suites are reflected in their geochemical compositions. All samples from these sequences are basic in terms of silica content, and most have SiO_2 in the range of 48-52 wt.% (Fig. 8.1). The three suites also exhibit similar ranges in the concentrations of most major and trace elements. The principal exception to this generalization lies in the slightly lower concentrations of incompatible elements (K_2O , TiO_2 , P_2O_5 , Zr and Y) in the Le Fer basalts relative to those of the upper Menihek or Willbob formations. Conversely, the Cr content of the Le Fer basalts tends to be slightly higher. The concentrations of incompatible elements and Cr in the upper Menihek and Willbob samples cover a similar range.

In the diagrams of Figure 8.1, where Zr has been used as a index of differentiation, the concentrations of several of the plotted elements define linear trends for samples from the Le Fer, upper Menihek and Willbob basalts. With increasing Zr, the Al_2O_3 , MgO (as indicated by Mg#), CaO, Cr and Ni content of the samples decline, while $\text{Fe}_2\text{O}_3\text{T}$, TiO_2 , P_2O_5 , Zr and Y increase in concentration. MnO and V tend to increase with increasing Zr in the suites, although the trends are less well defined. Na_2O increases with Zr in the Le Fer basalts, K_2O appears to decline slightly, while Ba and Sr exhibit no systematic variation. There is no consistent variation in Na_2O , K_2O , Ba and Sr content relative to Zr in the upper Menihek and Willbob basalts, and these elements show relatively large ranges in concentrations. Since Ba and Sr tend to be mobile during hydrous alteration, it is possible that both the large variation and the lack of coherent trends relative to Zr in the Le Fer, upper Menihek and Willbob suites derive from post-crystallization effects. This may also be the source of variations in Na_2O and K_2O among the upper Menihek and Willbob basalts. With exception of the variability of Na_2O , K_2O , Ba and Sr concentrations, which

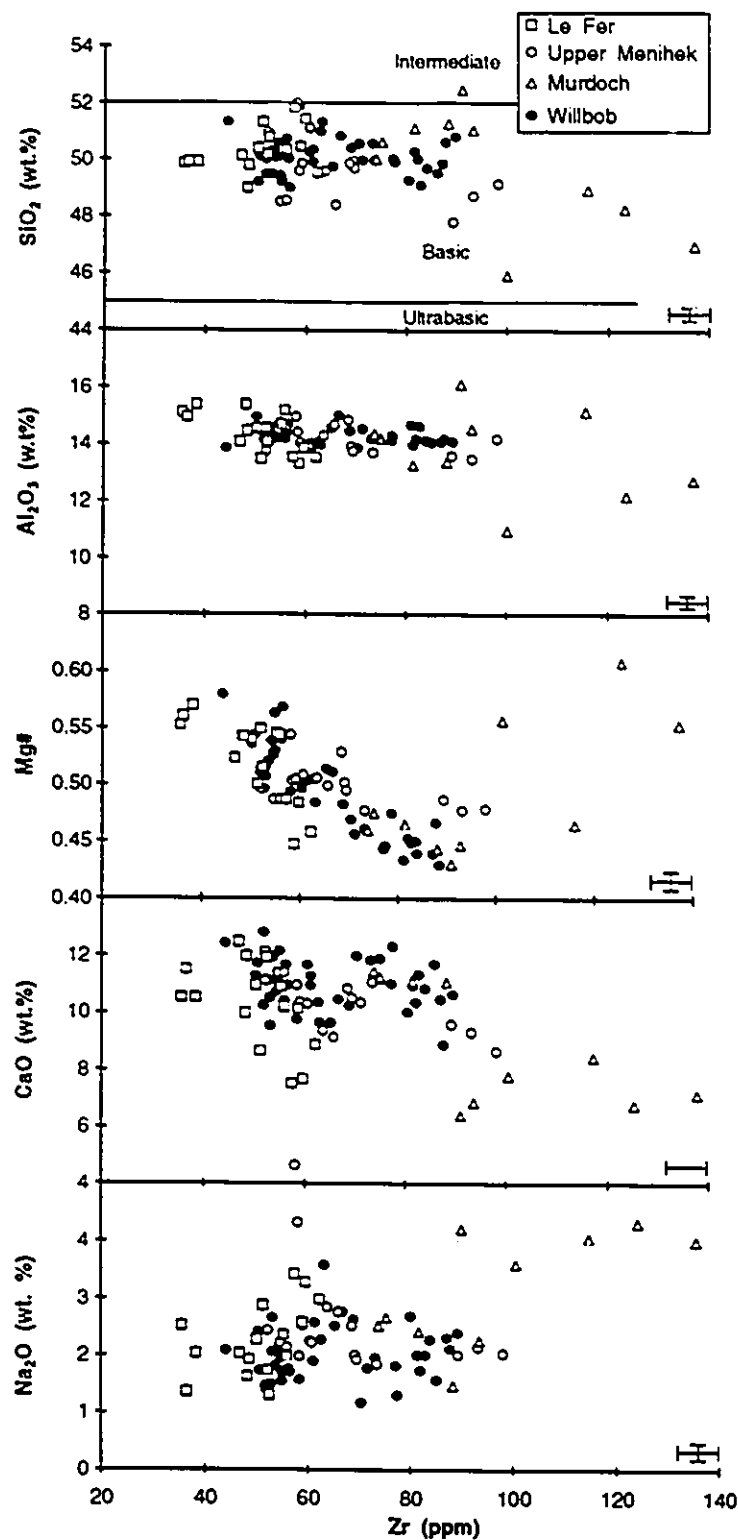


Figure 8.1. Covariation of selected major and trace elements with Zr in basalts of the Le Fer, upper Menihok, Murdoch and Willbob formations. Average 2σ analytical uncertainty is indicated by error bars except where less than symbol size.

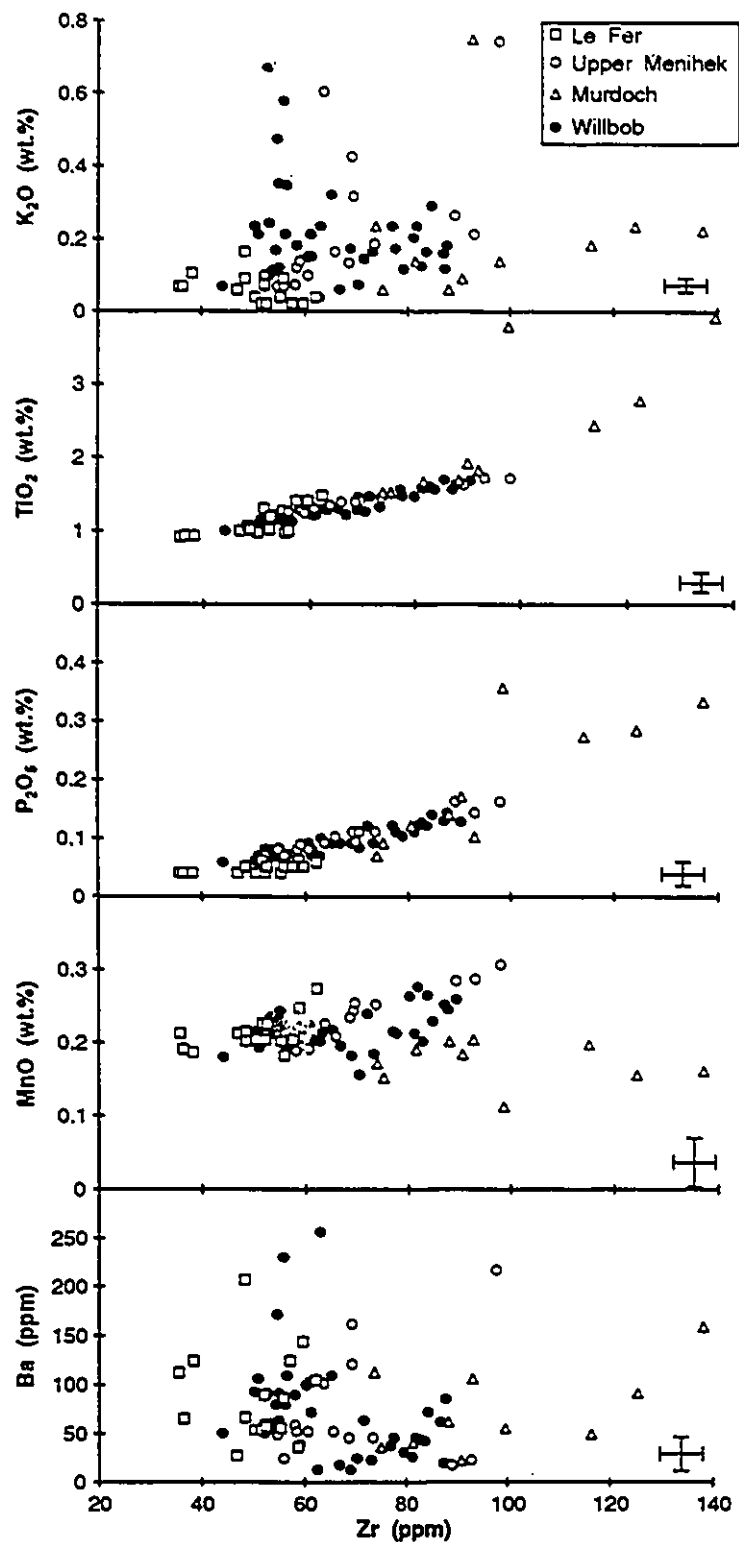


Figure 8.1 (continued). Covariation of selected major and trace elements with Zr in basalts of the Le Fer, upper Menihek, Murdoch and Willbob formations. Average 2σ analytical uncertainty is indicated by error bars except where less than symbol size.

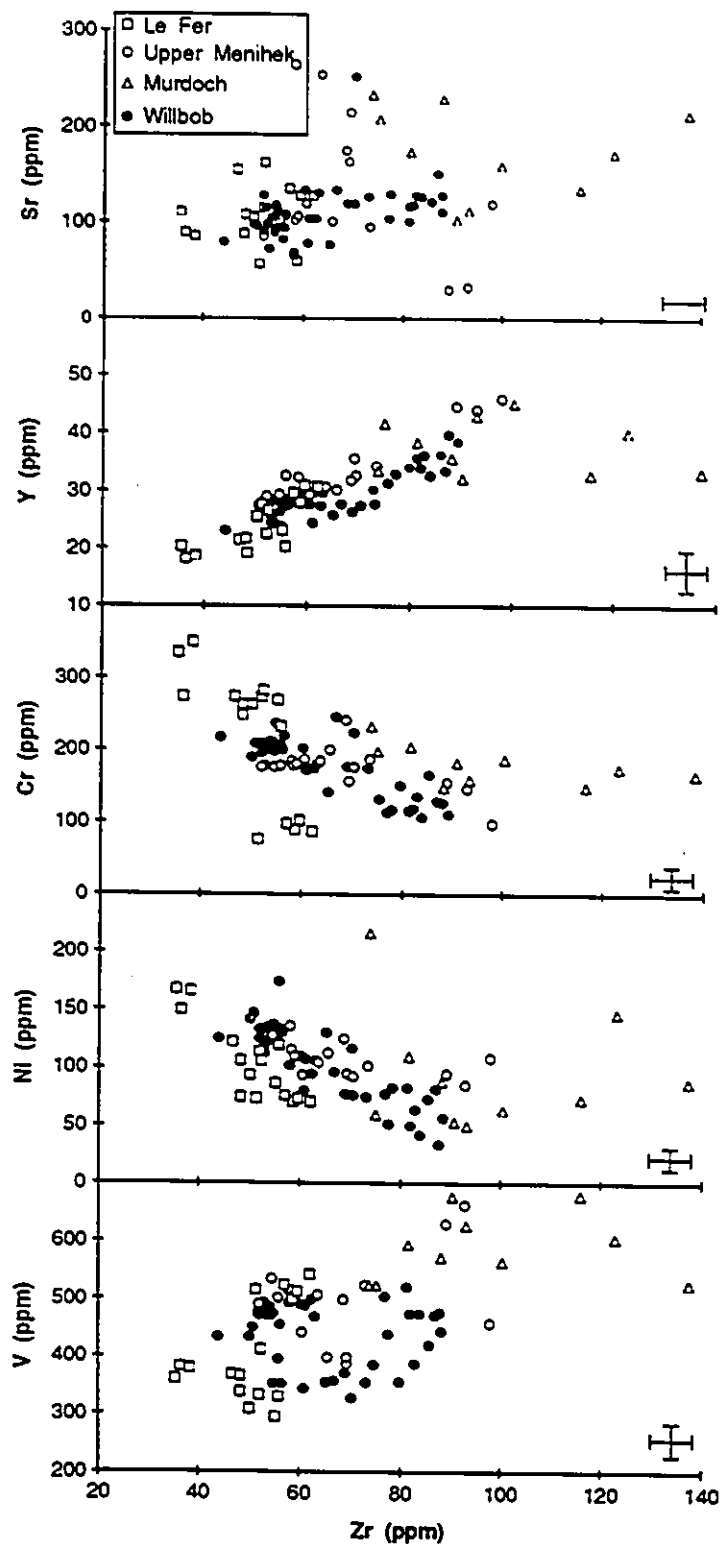


Figure 8.1 (continued). Covariation of selected major and trace elements with Zr in basalts of the Le Fer, upper Menihok, Murdoch and Willbob formations. Average 2σ analytical uncertainty is indicated by error bars except where less than symbol size.

may reflect secondary processes, the systematic geochemical variations evident in Figure 8.1 are consistent with magmatic differentiation. While it is apparent that some degree of geochemical evolution is represented by the varying compositions of samples from the three suites, it should be emphasized that the total compositional range is quite small. This homogeneity is particularly noticeable in the Willbob suite, where the data set represents a succession of basalts with a stratigraphic thickness in excess of 5 km.

The Murdoch basalts depart from the general homogeneity of the Le Fer, upper Menihek and Willbob suites in three ways. First, the ranges in concentrations of several of the elements plotted in Figure 8.1 are considerably greater. Second, the Murdoch basalts tend to have higher TiO_2 , P_2O_5 , Y and V, and lower MnO contents than the other suites. Third, covariation trends for the Murdoch samples tend to be poorly defined. The absence of coherent trends within the Murdoch suite may, in part, reflect the small sample population, but this does not explain the high degree of compositional variability in comparison to the other units. Exposures of Murdoch basalts examined in this study, however, are often intensely sheared and many samples show extensive veining. This contrasts with the usually massive nature of exposures of the other suites. In addition, the Murdoch samples exhibit the greatest degree of alteration found in the LTBS such that primary mineralogy and textures may be completely obliterated. Consequently, elemental mobility may have been enhanced in the Murdoch basalts, and such mobilization may explain some of the scatter in the concentration data.

An examination of Figure 8.1 suggests an alternative explanation for the compositional diversity of the Murdoch basalts in that the samples can be subdivided into high-Zr and low-Zr groups. The low-Zr samples have compositions which are similar to those of the Le Fer, upper Menihek and Willbob basalts, although their incompatible element contents are generally comparable to the more evolved members of the latter suites. In contrast, the high-Zr group are more geochemically variable and tend toward much higher incompatible element contents. The samples which show anomalous geochemistries were obtained from the lower part of the Murdoch Formation near Connolly Lake, while the remainder represent basalts from the middle of the formation. The two sets of samples exhibit similar degrees of alteration, and thus the compositional differences probably reflect differences in melt geochemistry rather than the effects of

alteration and metamorphism. Whichever interpretation is correct, the Murdoch basalts in general are compositionally more variable than the other suites, and tend toward higher incompatible element contents. The Le Fer, upper Menihek and Willbob basalts are compositionally similar with respect to most major and trace elements, but those of the Le Fer Formation are slightly depleted in incompatible elements and enriched in Cr. Of the four suites, the upper Menihek and Willbob basalts are the most similar, with comparable major and trace element contents except in the case of the relatively mobile components Na_2O , K_2O , Ba and Sr. This similarity is consistent with the postulated comagmatic origin of these suites.

The chondrite-normalized REE patterns of the Le Fer, upper Menihek, Willbob and low-Zr Murdoch basalts (Fig. 8.2) are all relatively flat, although there are small differences in both REE concentrations and the shapes of the patterns. The lowest REE concentrations are found in the Le Fer basalts, which have average contents of about ten times chondritic values. As with most other elements examined, the REE patterns and concentrations of the upper Menihek and Willbob basalts are essentially indistinguishable, with concentrations in the range of ten to twenty times chondritic values. The upper Menihek and Willbob suites also have similar average REE contents of approximately twelve times chondrite. The low-Zr Murdoch basalts contain slightly elevated REE concentrations, with average values of about twenty times chondritic values. This is consistent with the generally higher incompatible element content of the Murdoch suite. The two analyses from the high-Zr Murdoch group show moderate light REE enrichments in comparison to the other Murdoch samples, with La concentrations of 28 and 42 times chondritic values.

The overall shapes of the patterns also differ among the suites. The Le Fer basalts show slight light REE depletions, with chondrite-normalized La-Sm ratios ($\text{La}/\text{Sm}_{\text{cn}}$) ranging from 0.6 to 0.9, and La/Lu_{cn} values of 0.8 to 1.2 (Fig. 8.3). $\text{La}/\text{Sm}_{\text{cn}}$ in the upper Menihek and Willbob suites ranges from 0.83 to 1.0, and the slopes of the light REE patterns are correspondingly flatter. The $\text{La}/\text{Sm}_{\text{cn}}$ values of the low-Zr Murdoch samples are comparable to the upper Menihek and Willbob suites, but La/Lu_{cn} values are higher. Thus the low-Zr Murdoch basalts exhibit slight heavy REE depletions relative to the other suites. The high-Zr Murdoch samples have higher $\text{La}/\text{Sm}_{\text{cn}}$ (1.1 and 1.4) and La/Lu_{cn} (2.8

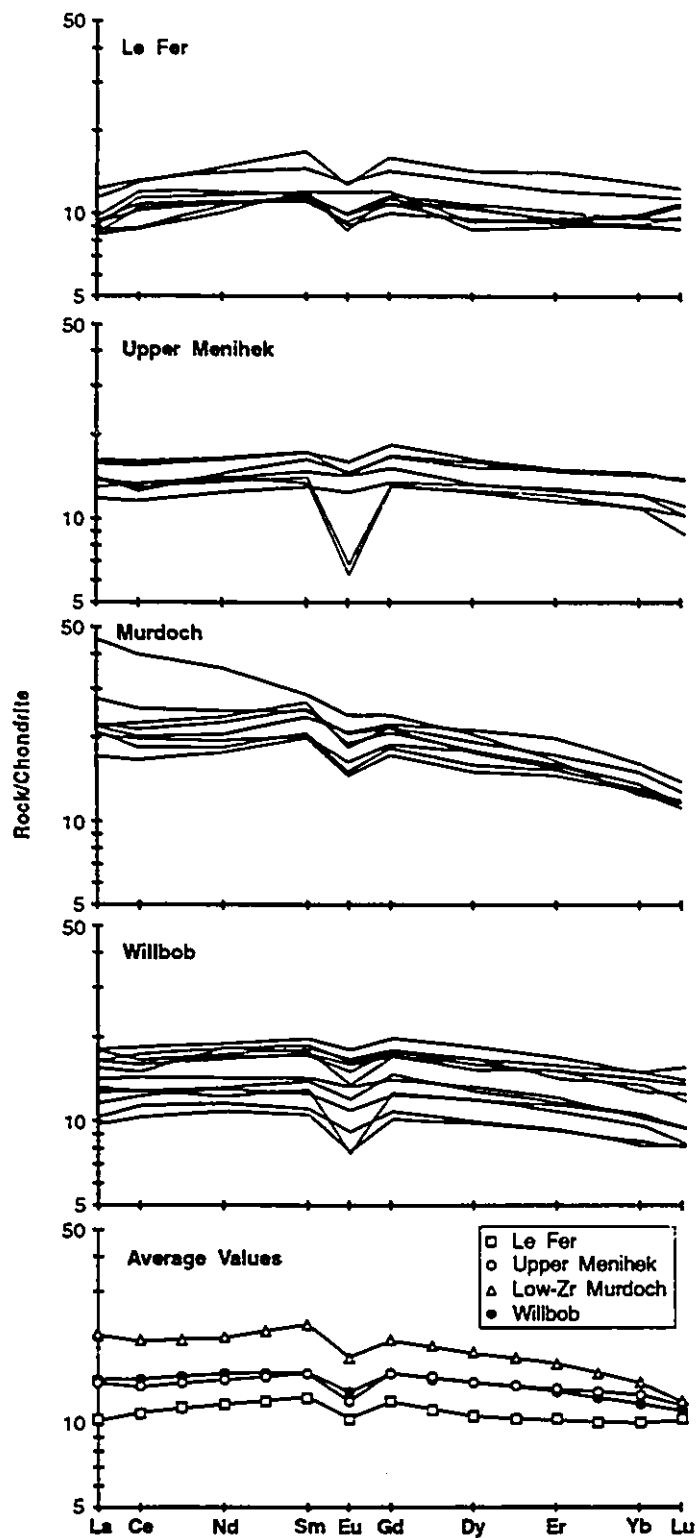


Figure 8.2. Chondrite-normalized REE patterns for basalts from the Le Fer, upper Menihék, Murdoch and Willbob formations. Normalizing values are from Nakamura (1974).

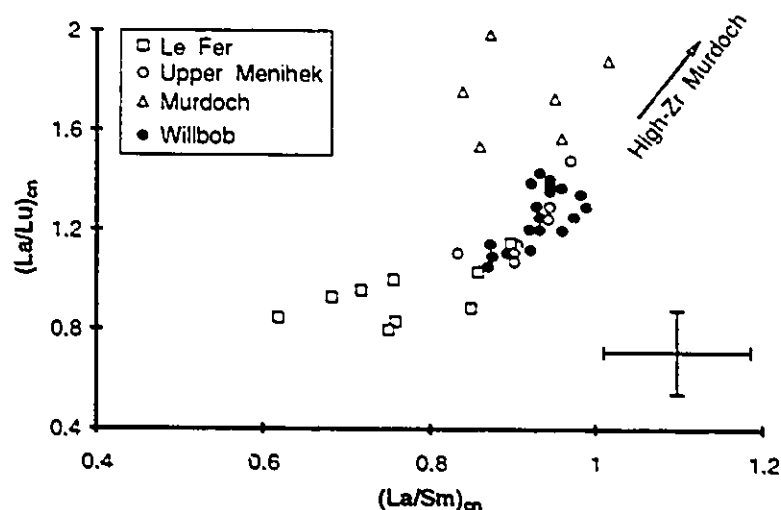


Figure 8.3. Chondrite-normalized La/Lu versus La/Sm for the Le Fer, upper Menihék, low-Zr Murdoch and Willbob basalts. The high-Zr Murdoch samples would lie to the upper right of the diagram but are not plotted to allow for a scale in which differences among the other suites are apparent. Average 2σ analytical uncertainty is illustrated by error bars. Normalizing values from Nakamura (1974).

and 4.4) values indicating more pronounced light REE enrichments. In comparison to the low-Zr samples, the high-Zr Murdoch basalts also show slight heavy REE depletions (Fig. 8.2). The La/Lu_{cn} ratios of the upper Menihék and Willbob basalts are generally intermediate between those of the Le Fer and low-Zr Murdoch suites.

The majority of samples from each of the suites show slight to moderate negative Eu anomalies (Eu/Eu^* , where Eu^* is the chondrite-normalized Eu concentration interpolated from the concentrations of Sm and Gd). There is considerable variation, however, in the values of the anomalies among samples. In each of the series one or more samples show Eu/Eu^* values close to unity, and this ratio reaches minimum values of 0.72, 0.48, 0.76 and 0.61 in the Le Fer, upper Menihék, Murdoch and Willbob suites, respectively. The presence of negative Eu anomalies indicates that plagioclase influenced the compositions of the basalts, either as a residual phase in the magma sources or as a fractionated phase during differentiation. The latter process is supported in the case of the Le Fer, upper Menihék and Willbob basalts by the plagioclase-phyric and plagioclase glomeroporphyritic nature of many of the samples.

Lower Menihek Basalts

Samples of the lower Menihek basalts fall into two categories; those obtained from *in situ* basaltic flows, and those obtained as basaltic clasts from mafic agglomerates and conglomerates. The clastic samples are typically strongly chloritized and may be schistose, but they generally exhibit the strong magnetism and amygdaloidal nature that characterize the flows of the Howse Lake area. The two groups of samples also exhibit distinctive geochemical characteristics in comparison to the basalts of the Le Fer, upper Menihek and Willbob formations (Fig. 8.4). The lower Menihek rocks range from ultrabasic to intermediate in silica content, with the low SiO_2 of the ultrabasic samples reflecting high oxide contents. The Al_2O_3 , $\text{Fe}_2\text{O}_3\text{T}$, and MgO contents are generally comparable to those of the upper Menihek basalts, but there is usually greater diversity in concentrations. In addition, the upper Menihek basalts tend to have higher CaO and lower Na_2O and K_2O , while the lower Menihek samples are distinguished by much higher TiO_2 , P_2O_5 , Zr and Nb concentrations.

Although the two groups of lower Menihek samples share distinctive geochemical characteristics, the clastic samples are enriched in Na_2O , Zr , Y and Nb and impoverished in K_2O and Ba relative to the basalts. These dissimilarities may have arisen through differences in magma compositions, or they may reflect the greater degree of alteration and deformation suffered by the clasts. An examination of normalized extended trace element patterns for clasts and flows (Fig. 8.5) suggests that the both interpretations may be correct. The patterns are similar with respect to most of the relatively immobile elements, although concentrations in the clasts are often slightly higher. Both groups exhibit elevated Ti and are depleted in Y and the heavy REE relative to normal-type mid-ocean ridge basalt (N-MORB). Both groups also exhibit strong Sr depletions. But the clastic samples exhibit consistent positive Nb anomalies which are not present in the flows, lack the moderate negative Zr anomalies found in the latter, and generally show slightly lower Ti abundances. These discrepancies are unlikely to reflect magmatic differentiation or the effects of alteration and metamorphism, and therefore may represent primary compositional differences in the parental melts. The abundances of the relatively mobile large ion lithophile (LIL) elements Rb , Ba and K are also substantially different in the two sample suites. These elements are consistently depleted in the clasts relative to the flows,

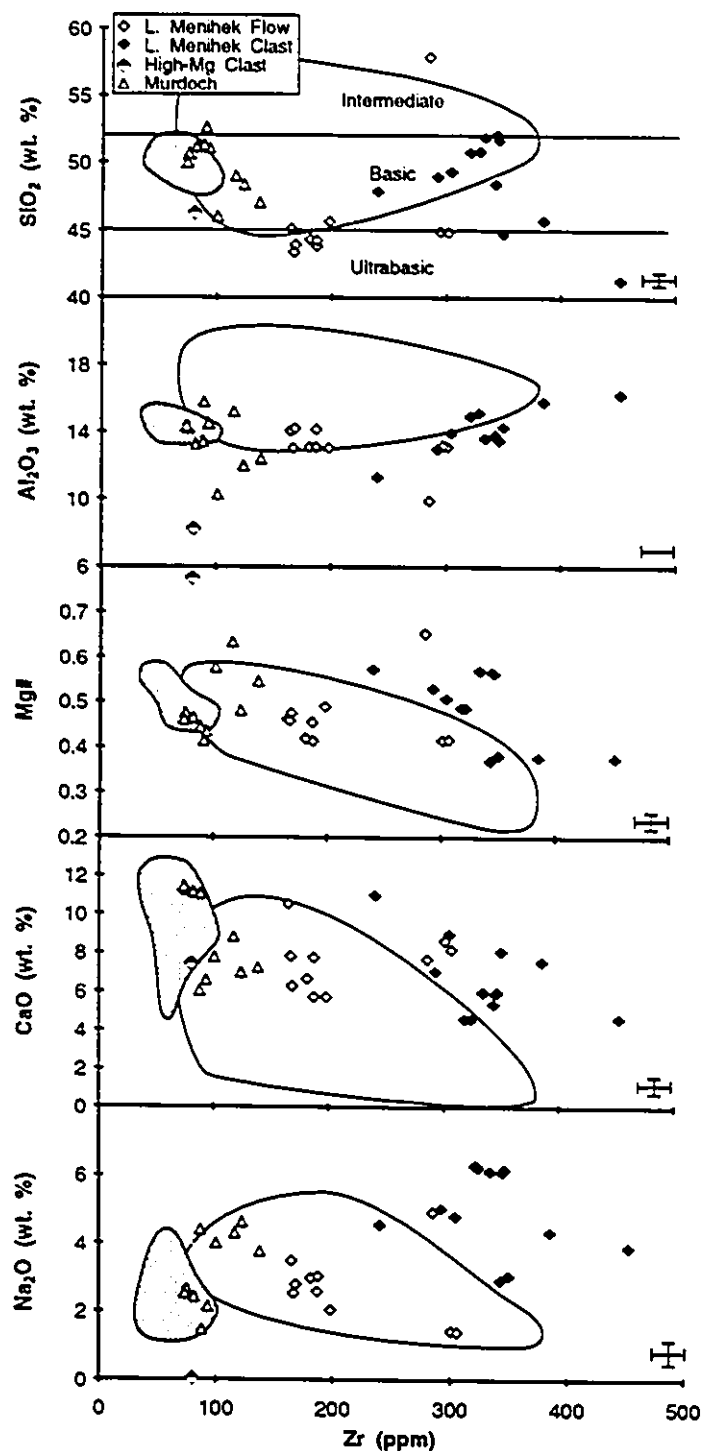


Figure 8.4. Covariation of major and selected trace elements with Zr in samples of volcanic rocks from the lower Menihek and Murdoch formations. The combined field of the Le Fer, upper Menihek and Willbobb basalts is indicated (shaded), as is the field of the Nimish volcanic rocks (unshaded; data from Watanabe, personal communication, 1993). Average 2 σ analytical uncertainty is indicated by error bars except where less than symbol size.

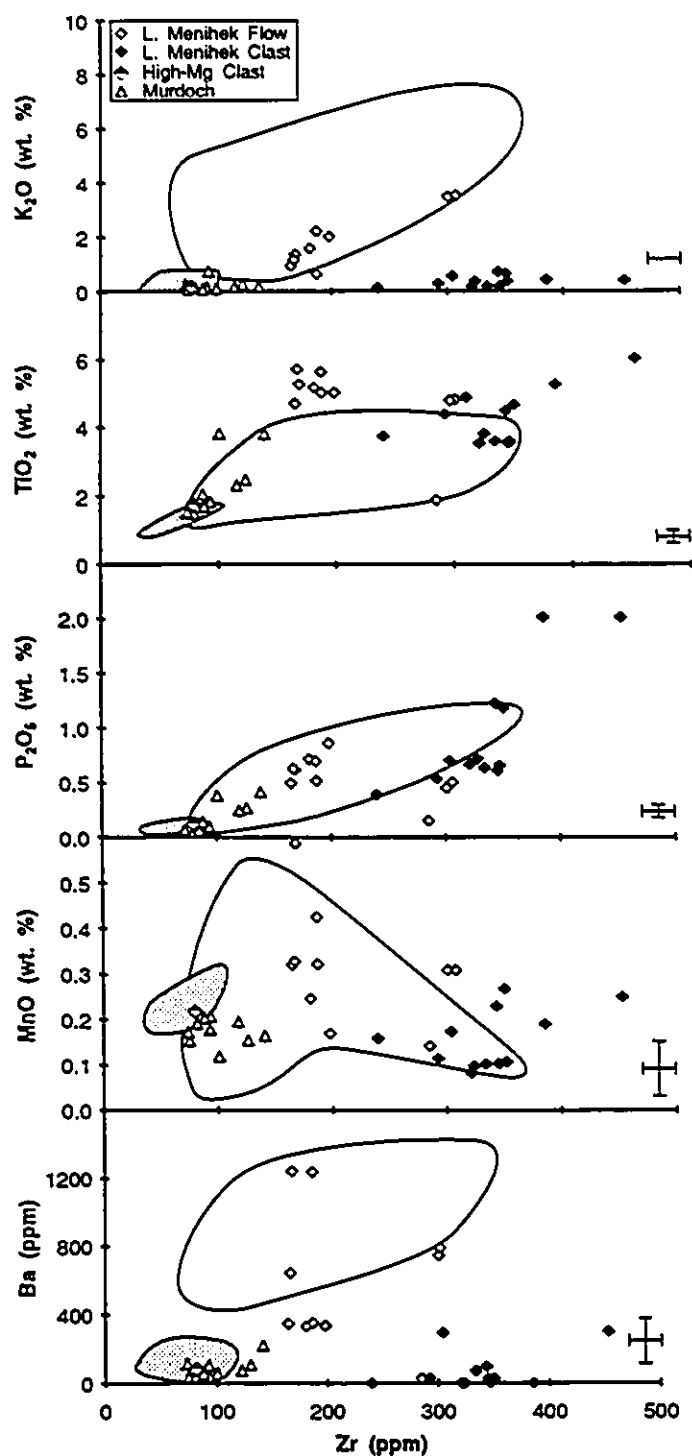


Figure 8.4 (continued). Covariation of major and selected trace elements with Zr in samples of volcanic rocks from the lower Menihek and Murdoch formations. The combined field of the Le Fer, upper Menihek and Willbob basalts is indicated (shaded), as is the field of the Nimish volcanic rocks (unshaded; data from Watanabe, personal communication, 1993). Average 2σ analytical uncertainty is indicated by error bars except where less than symbol size.

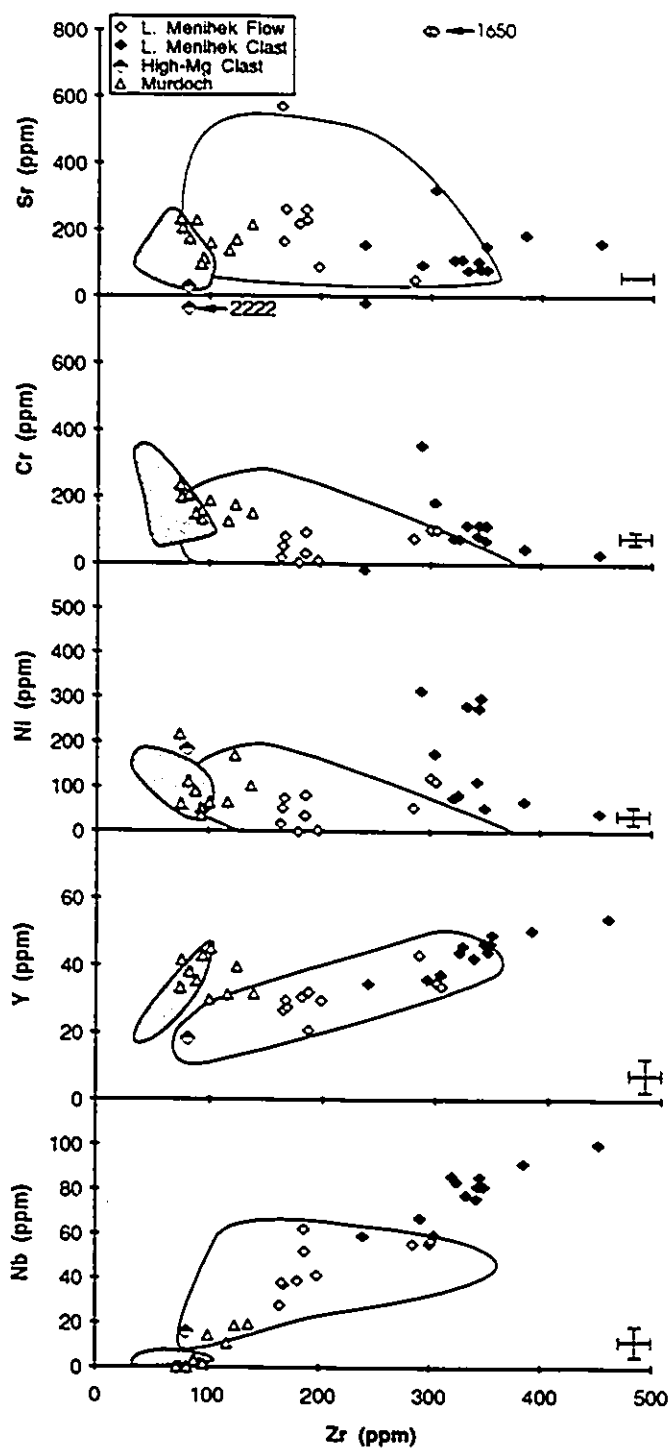


Figure 8.4 (continued). Covariation of major and selected trace elements with Zr in samples of volcanic rocks from the lower Menihok and Murdoch formations. The combined field of the Le Fer, upper Menihok and Willbob basalts is indicated (shaded), as is the field of the Nimish volcanic rocks (unshaded; data from Watanabe, personal communication, 1993). Average 2σ analytical uncertainty is indicated by error bars except where less than symbol size.

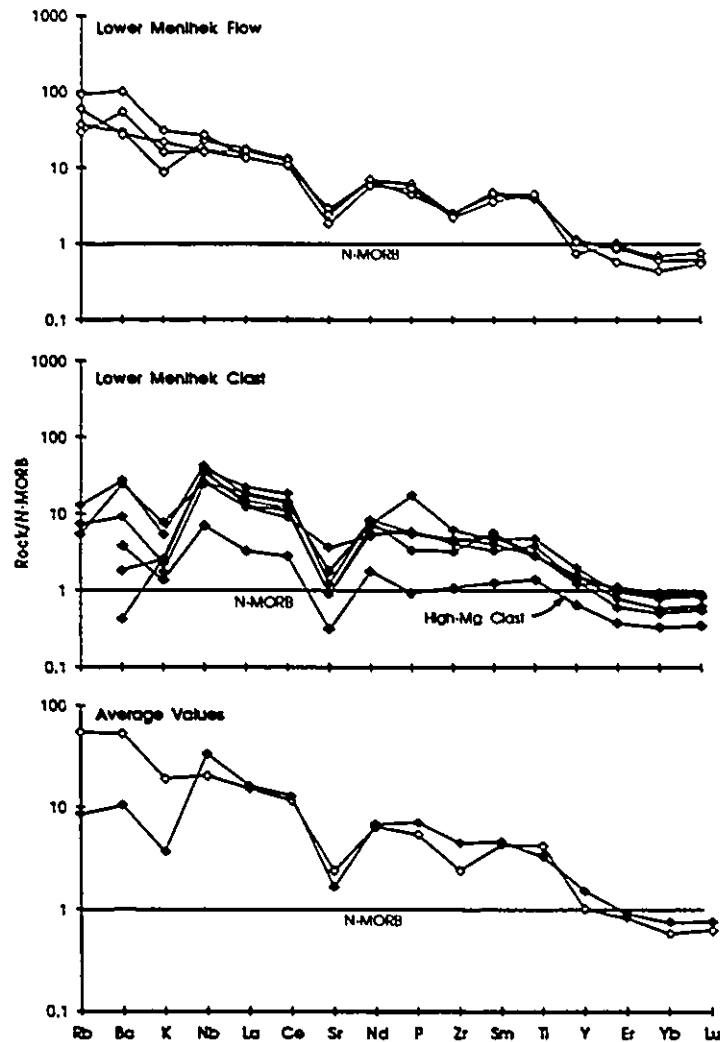


Figure 8.5. Normalized extended trace element plots for clasts and flows from the lower Menihok suite. Rb concentrations in some clastic samples are below detection limits (equivalent to a normalized concentration of approximately 4). Average abundances for the clasts do not include values from the high-Mg sample. Normalizing values are for normal-type mid-ocean ridge basalt (N-MORB) from Sun and McDonough (1989) except Ba, which is from Saunders and Tarney (1984).

and since the clasts exhibit much stronger alteration than the flows this impoverishment may reflect post-crystallization loss of these elements. Such a process might also lead to the general enrichment in immobile elements in the clasts through mass loss. The slightly lower Ti content of the clasts, however, cannot be explained in terms of mass loss since this element should be relatively immobile during alteration.

Whether the compositional differences between clasts and flows result from primary or secondary processes, it is clear that the products of lower Menihek magmatism are compositionally distinct from the Le Fer, upper Menihek and Willbob volcanic suites. The lower Menihek samples do, however, show some compositional similarities to the Nimish volcanics. This is particularly true in the case of the lower Menihek basaltic samples, which lie within the fields defined by the Nimish suite for the majority of elements plotted in Figure 8.4. The two suites share incompatible and alkali element enrichments and a compositional diversity which greatly exceeds that of the Le Fer, upper Menihek and Willbob basalts.

The incompatible element enrichment of the lower Menihek suite is also evident in the REE patterns of the samples (Fig. 8.6). Both flows and clasts exhibit strong light REE enrichment, with La/Lu_{cm} values in the range of 10.2 to 14.4, La/Sm_{cm} values of 1.8 to 2.4, and La concentrations generally exceeding 100 times chondritic values. All samples with the exception of one flow exhibit slight negative Eu anomalies, with Eu/Eu^* values ranging from 0.74 to 0.93. The light REE enrichment and REE concentrations found in the lower Menihek rocks are again comparable to those of the Nimish suite (Fig. 8.6).

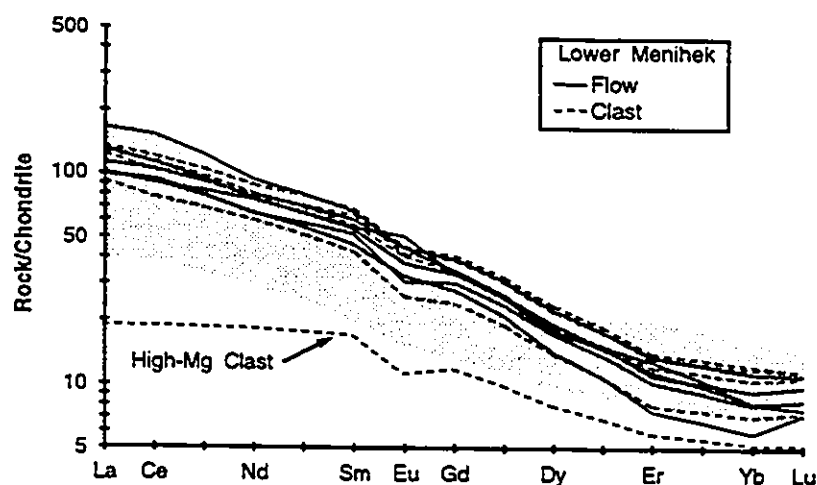


Figure 8.6. Chondrite-normalized REE abundances in clasts and flows from the lower Menihek suite. Range of values for the Nimish volcanics is shown by shaded region (data from Watanabe, personal communication, 1993). Normalizing values are from Nakamura (1974).

One lower Menihek clast differs from the others in the suite in that it lacks strong alkali enrichment and high incompatible element content. It is also highly enriched in MgO (~23 wt. %) and Cr (~2200 ppm) and impoverished in Al_2O_3 and Sr in comparison to other samples (Fig 8.4). Although the sample shares the light REE enrichment and positive Nb anomaly characteristic of the lower Menihek clastic suite (Fig. 8.5), the degree of incompatible element enrichment and the total REE concentrations are much lower (Fig. 8.6). The extended trace element pattern of the high-Mg clast suggests that it may represent an ultramafic member of the lower Menihek suite, but since the sample was obtained from conglomerate, and since its major element geochemistry is substantially different from all other samples, the relationship is uncertain. This problem is discussed further in section 8.4.

With the exception of P_2O_5 , Y and Nb, the elements plotted in Figure 8.4 exhibit little systematic covariation with Zr, despite the relatively large compositional range. The lack of coherent variation is apparent in both sample groups individually, as well as in the suite as a whole. This contrasts with relationships among elements in the upper Menihek basalts where the compositional range of most elements is smaller, but where many exhibit well defined covariation with Zr. For example, there is a fairly substantial range in SiO_2 content (42-58 wt.%) and Zr (170-460 ppm) in the lower Menihek suite, but no systematic relationship between the two is evident. Similarly, Cr and Ni show no obvious trend relative to Zr, while a negative linear correlation is apparent in the Le Fer, upper Menihek and Willbob suites. Although alteration may have contributed to the compositional variability of the clastic samples, the same cannot be said for the basaltic flows, which are relatively pristine. This suggests that compositional variation in the suite is the product of a more complex differentiation mechanism or mechanisms than that which led to variations in the Le Fer, upper Menihek and Willbob basalts.

The compositions of the Murdoch basalts are illustrated in Figure 8.4 for comparison with the lower Menihek suite. As noted earlier, the low-Zr Murdoch samples tend to fall within the compositional fields defined by the Le Fer, upper Menihek and Willbob basalts. In contrast, the high-Zr samples exhibit some compositional characteristics which are similar to those of the lower Menihek suite. This is apparent in

the low CaO and elevated Na_2O , TiO_2 , P_2O_5 , Zr and Y contents of the samples, which plot within the Nimish field in a position intermediate between the low-Zr samples and the lower Menihek suite in many of the diagrams. This is also true of the REE patterns of the high-Zr samples, which exhibit light REE enrichments more pronounced than those of the low-Zr basalts. (cf. Figs. 8.2 and 8.6). The high-Zr Murdoch sample showing the greatest light REE enrichment falls within the range of the Nimish volcanics. A comparison of the extended normalized trace element patterns of the lower Menihek and Murdoch basalts also reveals the transitional nature of the high-Zr Murdoch samples (Fig. 8.7). These exhibit, to a lesser extent, the Sr, Y and heavy REE depletions, and Ti enrichment, which characterize the alkali basalts. The high-Zr samples also lack the relative Nb and P depletions found in the other Murdoch basalts. These trace element patterns, particularly with respect to Nb, Sr and Ti, are transitional between the low-Zr Murdoch basalts and the lower Menihek alkali basalts. This suggests that the alkaline natures of the samples do not reflect post-crystallization effects, but are primary magmatic characteristics

8.3 Classification

The basaltic rocks of the Menihek Formation were previously subdivided into upper and lower sequences based on stratigraphy, lithology and petrography. From the preceding discussion it is clear that the two suites are also geochemically distinct. The discriminant diagrams of Figure 8.8 illustrate the fundamental difference between the lower Menihek volcanic rocks and the other suites in terms of geochemical affinities. In each diagram the alkaline nature of the lower Menihek basalts is evident. In contrast, the Le Fer, upper Menihek and Willbob basalts are subalkaline. The majority of the Murdoch samples are also subalkaline, but the high-Zr samples lie within the alkaline fields in Figures 8.8A and 8.8B. In the SiO_2 versus Nb/Y plot of Winchester and Floyd (1977), the same samples show elevated Nb/Y ratios, but fall within the subalkaline field (Fig. 8.8C).

In Figure 8.9 the various suites are plotted in terms of normative constituents in the basalt tetrahedron of Yoder and Tilley (1962). Most of the subalkaline samples cluster at the diopside-hypersthene join. The majority of upper Menihek and Willbob samples fall within the quartz normative field, while the Le Fer samples are split between the quartz

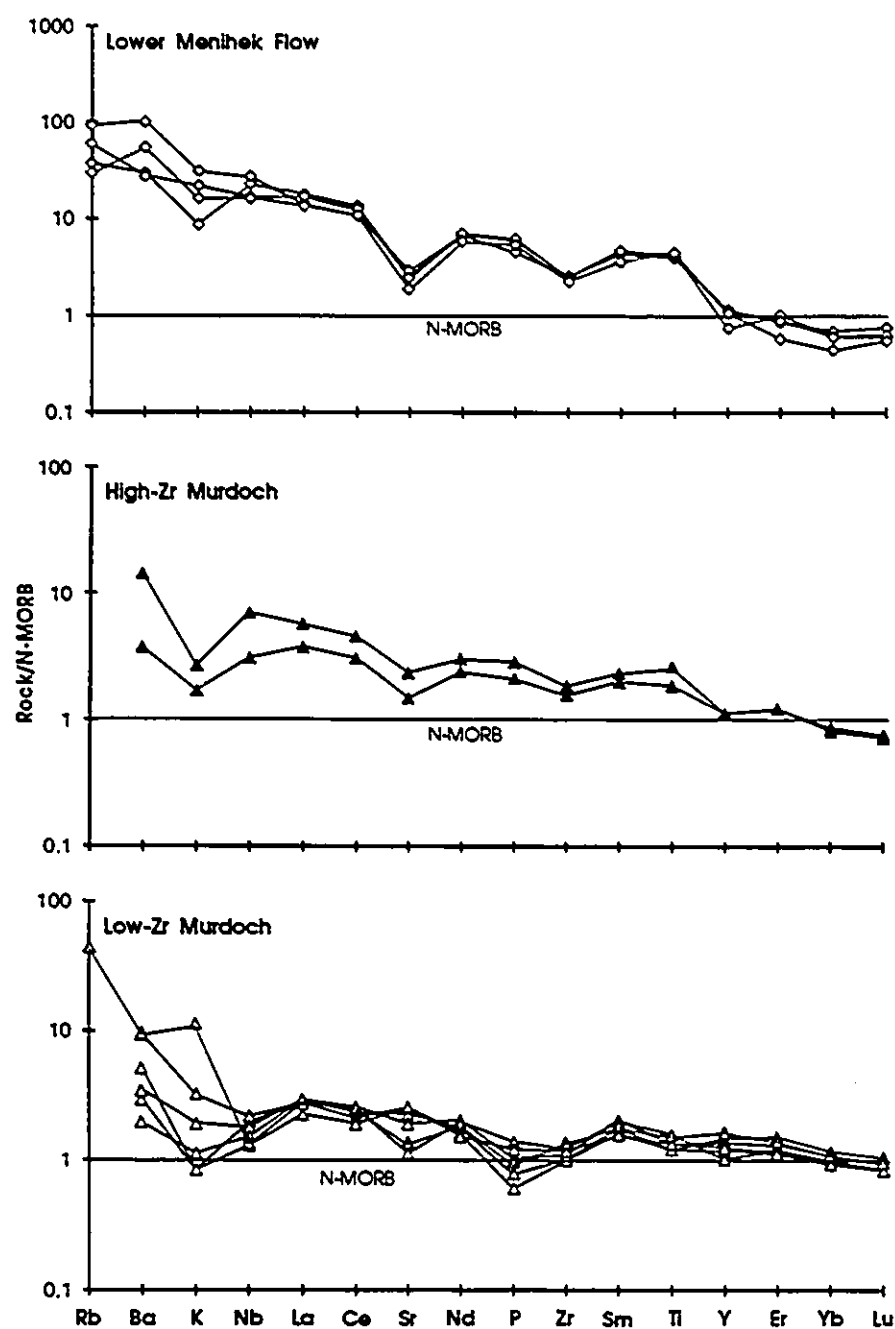


Figure 8.7. Normalized extended trace element plots for basalts of the lower Menihok and Murdoch suites. Rb concentrations for most samples of the Murdoch suite are below XRF detection limits (equivalent to a normalized concentration of approximately 4). Normalizing values are for normal-type mid-ocean ridge basalt (N-MORB) from Sun and McDonough (1989) except Ba, which is from Saunders and Tarney (1984).

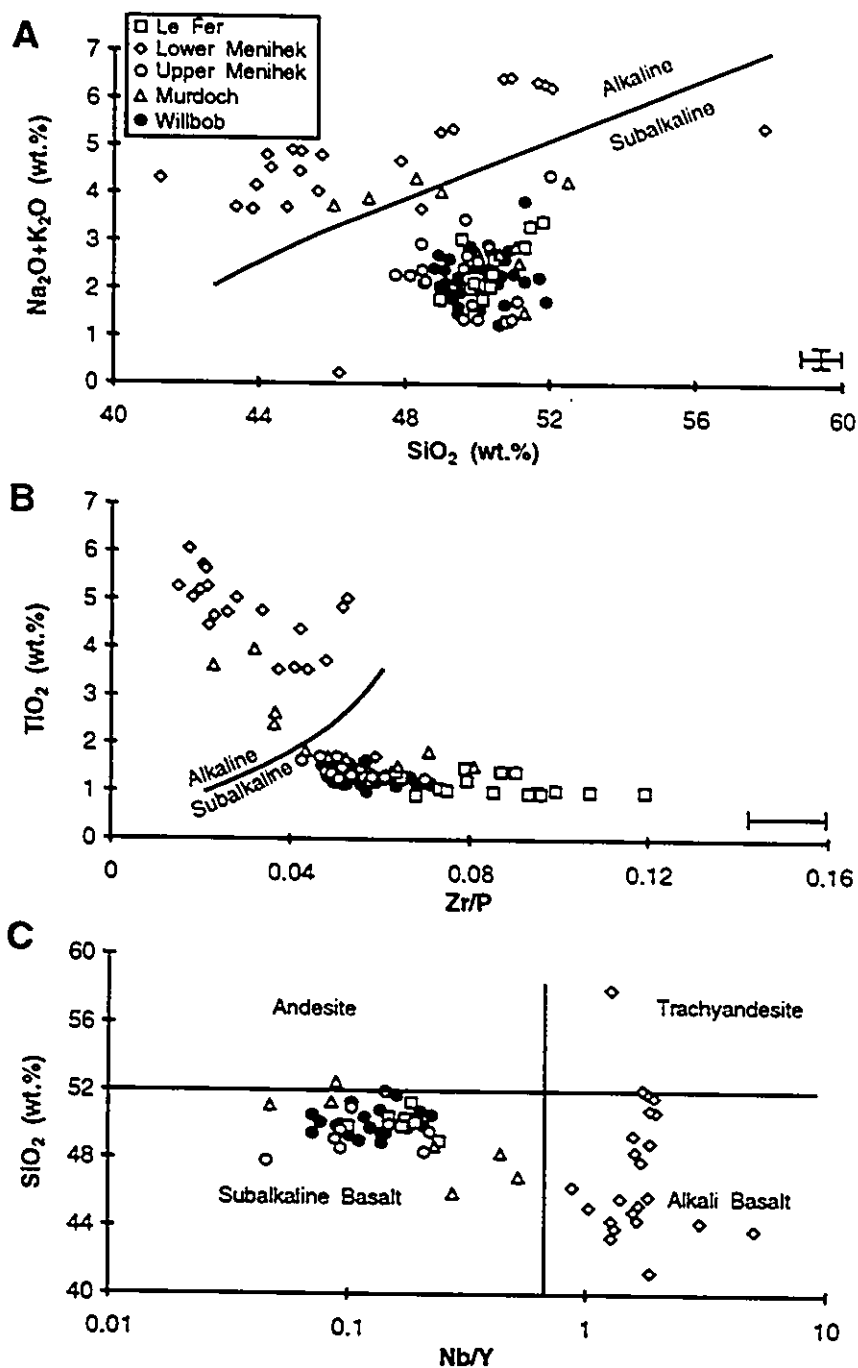


Figure 8.8. Geochemical subdivision of cycle 1 and 2 basaltic rocks. (A) Total alkalis-silica plot with dividing line from Irvine and Baragar (1971). (B) TiO_2 -Zr/P plot with dividing line from Winchester and Floyd (1976). (C) SiO_2 -Nb/Y plot with subdivision from Winchester and Floyd (1977). Average 2σ analytical uncertainty is indicated by error bars in (A), and for Zr/P in (B). Average 2σ error for TiO_2 in (B) is less than symbol size. Average 2σ error for SiO_2 in (C) is approximately equal to symbol size. Average 2σ error for Nb/Y in (C) is twice symbol size for sub-alkaline rocks and less than symbol size for alkaline rocks.

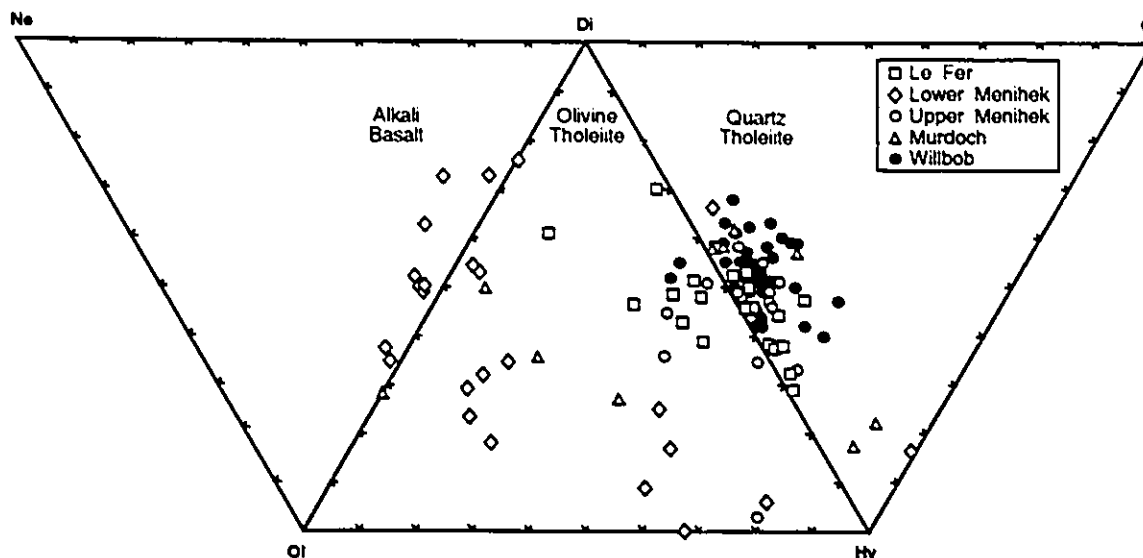


Figure 8.9. Normative classification of cycle 1 and 2 basaltic rocks using the basalt tetrahedron of Yoder and Tilley (1962). Calculation of FeO^* for use in the norm calculations is outlined in Appendix A4.

and olivine normative fields. Thus the Le Fer, upper Menihek and Willbob basalts are classified as quartz tholeiites or olivine tholeiites. The majority of Murdoch basalts also fall within the quartz normative field but, as in Figure 8.8, the high-Zr Murdoch samples show divergent compositions. These plot as olivine tholeiites closer to the diopside-olivine join, and within the compositional range of the lower Menihek suite. In terms of normative constituents, all samples of the latter are olivine normative, and approximately half contain normative nepheline. In keeping with their variable geochemical compositions, the lower Menihek volcanics exhibit considerable variation in normative compositions. The majority of samples, however, plot relatively close to the diopside-olivine join. This contrasts with the clustering of the subalkaline suites along the diopside-hypersthene join and suggests that the alkaline suite was derived from slightly silica undersaturated magmas. This interpretation is supported by the high modal oxide content of most lower Menihek samples

The tholeiitic nature of the Le Fer, upper Menihek and Willbob basalts is also evident in the AFM diagrams of Figure 8.10. Samples from each of these suites lie above the curve dividing subalkaline rocks into calc-alkaline and tholeiitic subgroups (Irvine and Baragar, 1971). The limited compositional variation of the basalts contrasts with the variation in the liquids of the Skaergaard intrusion (Wager and Brown, 1968), but is

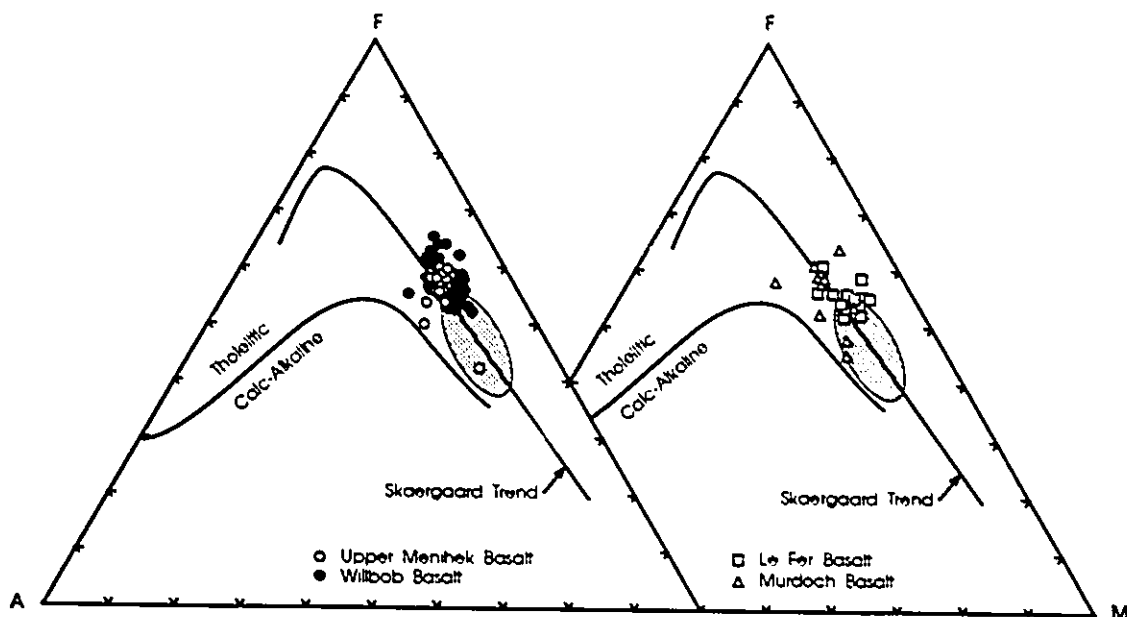


Figure 8.10. Classification of the Le Fer, upper Menihok, Murdoch and Willbob basalts according to the AFM diagram of Irvine and Baragar (1971). Skaergaard trend after Wager and Brown (1968), and abyssal tholeiite field (shaded) after Miyashiro *et. al.* (1970).

consistent with the relatively narrow compositional range of abyssal tholeiites from mid-ocean ridges (Miyashiro *et. al.*, 1970). The latter basalts, however, tend to be richer in MgO than those of the Le Fer, upper Menihok and Willbob formations. The Murdoch basalts, including the four samples showing slightly alkaline characteristics, also lie within the tholeiitic field.

The alkaline rocks of the lower Menihok Formation plot in both sodic and potassic fields of the normative Ab' (albite+5/3 nepheline)-An (anorthite)-Or (orthoclase) diagram (Fig. 8.11A) of Irvine and Baragar (1971). As discussed previously, the least altered samples consist of basaltic flows from the Howse Lake area. The clastic samples exhibit some SiO₂ and Na₂O enrichment and Rb, Ba and K₂O depletion which is thought to reflect more intensive alteration. Na-metasomatism would result in a shift toward the sodic field in Figure 8.11A, and this appears to be the case for the clastic samples. It is therefore probable that the lower Menihok rocks represent a potassic alkaline suite whose extrusive members have undergone variable Na-metasomatism subsequent to eruption. According to

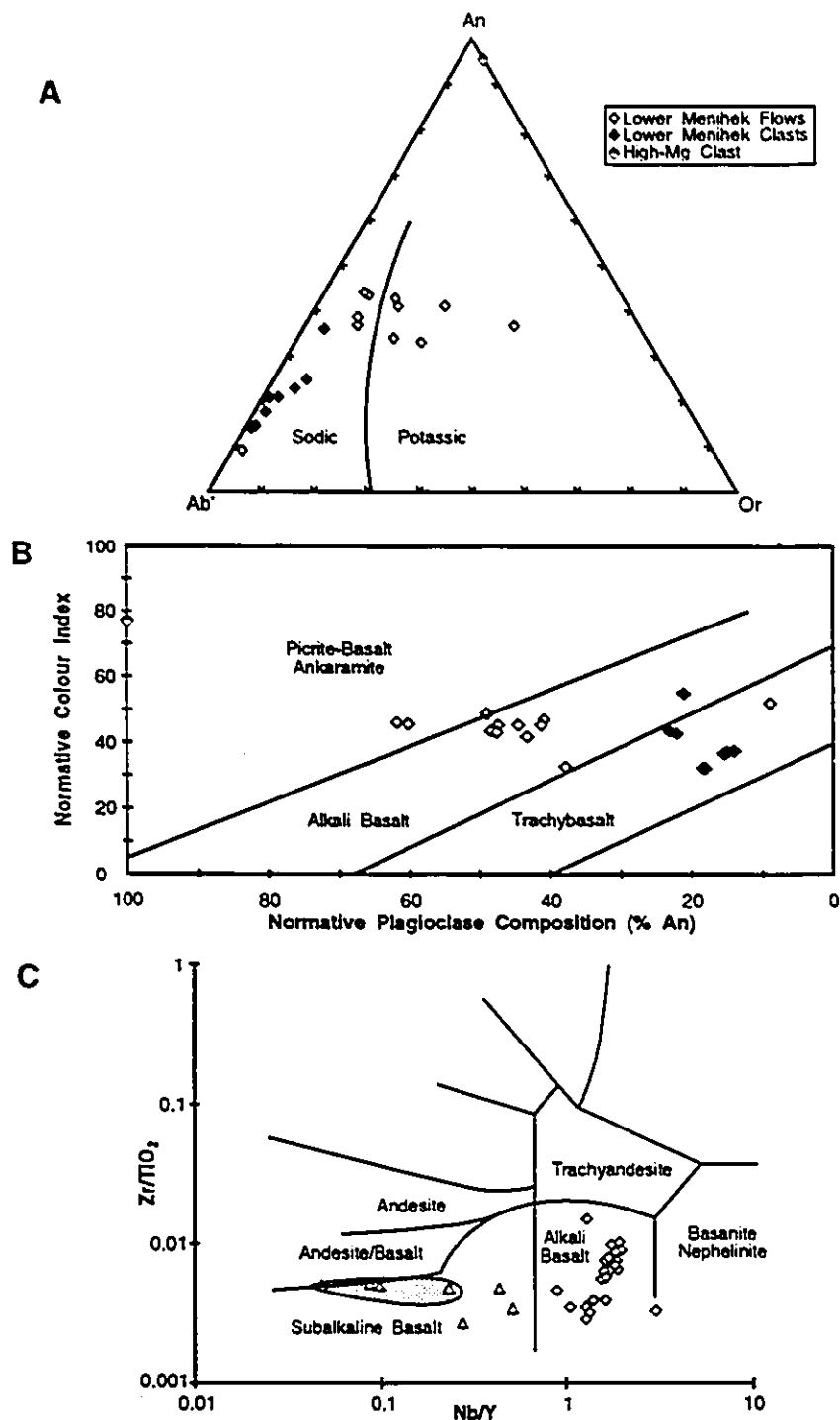


Figure 8.11. Classification of the lower Menihek alkaline rocks according to the (A) normative Ab' (albite+5/3 nepheline) - An (anorthite) - Or (orthoclase) diagram, and (B) normative colour index versus normative plagioclase composition diagram of Irvine and Baragar (1971). (C) Classification of the lower Menihek samples in the Zr/TiO₂ versus Nb/Y diagram of Winchester and Floyd (1977). The range of values for cycle 1 and 2 tholeiites is shown for comparison, as are the Murdoch basalts.,

the classification scheme of Irvine and Baragar (1971), the majority of the *in situ* lower Menihek samples are alkali basalts, while the majority of clastic samples are trachybasalts (Fig. 8.11B). Again it seems likely that Na-metasomatism is the source of this discrepancy, since such a process would result in the underestimation of the calculated An content of normative plagioclase. The fact that both clasts and flows plot in the alkali basalt field in the immobile element ratio diagrams of Winchester and Floyd (1977) supports this interpretation (Figs. 8.8C and 8.11C).

8.4 Genetic Relationships

A major difficulty in the interpretation of the tectonic setting of the Labrador Trough is the structural separation of potentially correlative sequences. With regard to the igneous rocks, the Le Fer gabbros are the presumed intrusive counterparts of the cycle 1 Le Fer basalts in the Ferrum zone. Equivalent rocks appear to be absent in the Schefferville and Howse zones. Of the cycle 2 igneous rocks, the upper Menihek basalts (Howse zone) are commonly correlated with those of the Willbob Formation (Doublet terrane), and the Menihek and Doublet sills are generally considered to be intrusive counterparts of these basalts (cf. Baragar, 1960; Dimroth, 1978; Rohon *et al.*, 1993; Skulski *et al.*, 1993). The Murdoch basalts (Doublet terrane) are considered older than the Willbob sequence on the basis of apparently conformable relationships among members of the Doublet Group (Baragar, 1967; Frarey, 1967). Given these relationships, the Murdoch Formation may be equivalent to the lower part of the Menihek Formation, and the slightly alkalic volcanics which occur in the lower part of the Murdoch sequence may be correlative with the Howse Lake alkali basalts. Alternatively, an unconformable relationship between the Murdoch and Thompson Lake formations would allow a correlation between the Murdoch basalts and the cycle 1 Le Fer basalts. In this case, the lower Murdoch alkaline rocks would be equivalent to the slightly alkaline basalts of the Seward subgroup (Chakonipau Formation), which underlies the Le Fer Formation. Alkaline volcanics comprise a significant proportion of the Nimish Formation in the Schefferville zone (Evans, 1978; Watanabe, 1989), and these are also potential correlatives to the alkaline rocks of the lower Menihek and Murdoch formations.

Upper Menihek and Willbob Basalts, Menihek and Doublet Gabbros

The geochemical and textural similarities among the Willbob and upper Menihek basalts, and Menihek sills, have often been cited as evidence of a common origin for the suites. In particular, the appearance of the striking plagioclase glomeroporphyritic texture in rocks of each suite is thought to indicate a comagmatic relationship. Identical textures, however, are found in basalts of the Sleeper and Belcher Islands (Belcher Islands Belt) and Cape Smith Belt of the Trans-Hudson Orogen (Baragar, 1974; Baragar and Scoates, 1981), and therefore textural similarity by itself does not necessarily demonstrate consanguinity. It is thus useful to compare the geochemical characteristics of the three suites.

The Willbob (Doublet terrane) and upper Menihek (Howse zone) basaltic suites are compositionally similar, and each include glomeroporphyritic flows. Of the presumed intrusive equivalents, the Menihek sills (Howse zone) also exhibit glomeroporphyritic texture, ranging from aphyric or sparsely glomeroporphyritic to highly glomeroporphyritic and anorthositic varieties. In contrast, there have been no reports of highly glomeroporphyritic/anorthositic horizons in the Doublet sills (Doublet terrane), and those examined in this study consisted of relatively homogeneous, aphyric gabbro. Highly glomeroporphyritic gabbro sills do, however, occur within the Hellancourt Formation of the northern Trough, and petrological and structural studies suggest that the Hellancourt and Willbob formations are correlative (Wares and Goutier, 1989; 1990; Rohon *et al.*, 1993; Skulski *et al.*, 1993).

It was concluded earlier that aphyric and glomeroporphyritic Menihek sills are comagmatic, but that their initial melts differed slightly in composition as well as glomerocryst content. In comparison to aphyric sills, the chilled margins of highly glomeroporphyritic sills are enriched in incompatible elements (TiO_2 , P_2O_5 , MnO, Y and Zr), impoverished in Cr and Ni, and exhibit slightly lower Mg#s. The same compositional differences are evident in the aphyric and glomeroporphyritic flows of the Willbob and upper Menihek suites (Fig. 8.12). In general, glomeroporphyritic flows of the two sequences show lower Mg#s, Cr and Ni, and higher incompatible elements than their aphyric counterparts. Figure 8.12 also illustrates the systematic compositional variations found in the Willbob and upper Menihek suites, with P_2O_5 , MnO, Y and Zr increasing sympathetically with TiO_2 , while Mg#, Cr and Ni decline. The compositional fields of the

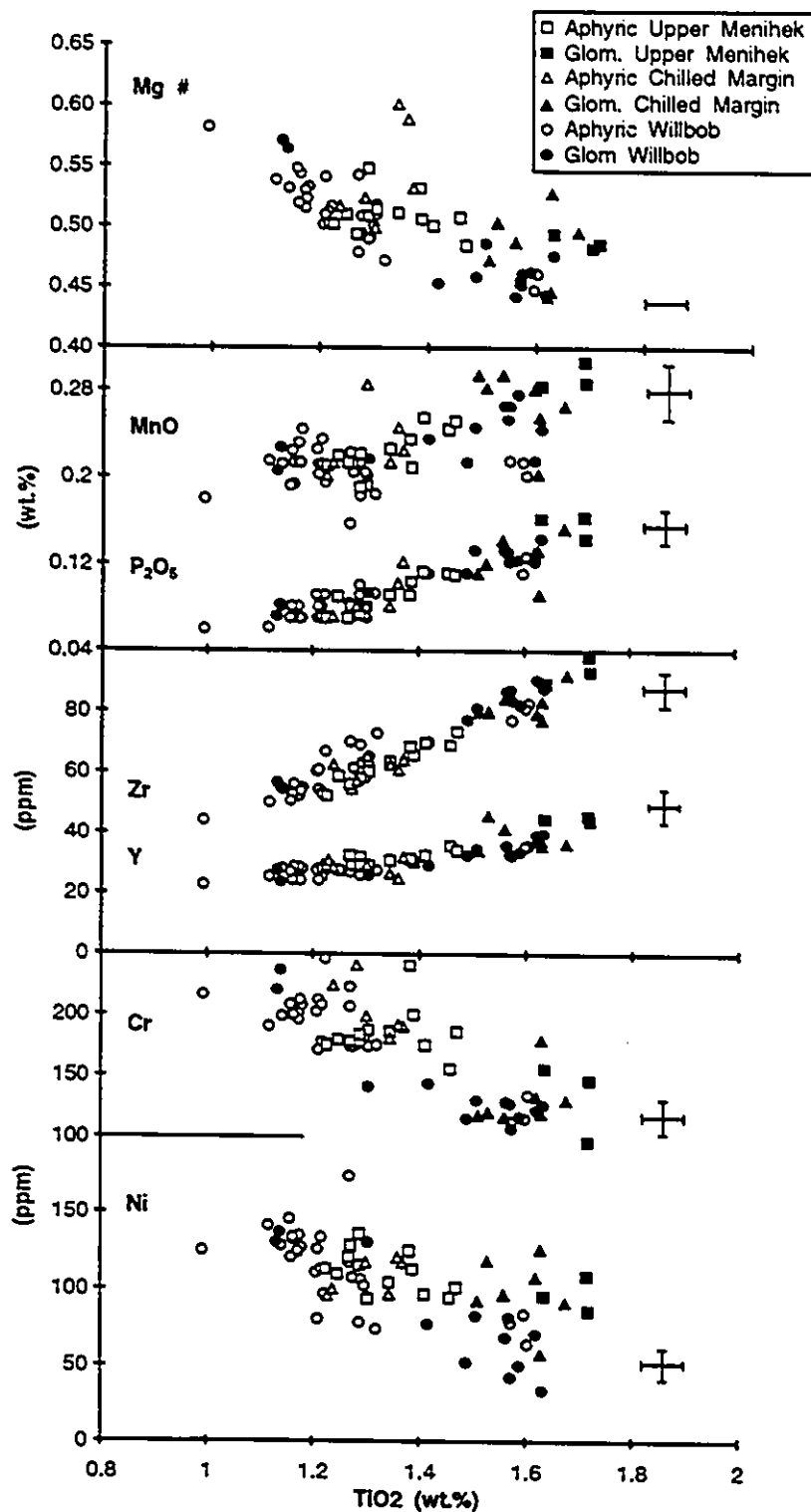


Figure 8.12. Covariation of selected elements with TiO_2 illustrating compositional differences between aphyric and glomeroporphyritic members of the upper Menihek basalts, Willbob basalts and Menihek sills (chilled margins). Average 2σ analytical uncertainty is indicated by error bars where greater than plot symbol size.

two sequences are generally similar, although the Willbob basalts show a slightly greater range in Mg# and TiO_2 content. This probably reflects the smaller sample population for the upper Menihek basalts rather than real compositional differences.

The ratios of conserved elements can be used to test the potentially comagmatic relationships among the Willbob and upper Menihek basalts, and the Doublet and Menihek sills. As discussed in Chapter 5, comagmatic rocks will have identical ratios of incompatible elements which are completely conserved during formation, and any subsequent differentiation, of the parental melts. In practical terms, the total variation in conserved element ratios among samples from two or more groups of comagmatic rocks will be less than that expected from analytical imprecision. If the suites are unrelated, or if a process in which the chosen elements participate is the source of variation, incompatible element ratios will be non-constant. The constancy of ratios can be evaluated through comparison of observed variation with that which could be expected from analytical imprecision.

The glomerocrysts found in the Menihek gabbros were interpreted as the pre-emplacement crystallization products of the parental magmas to the sills. Assuming a comagmatic relationship, the glomerocrysts found in the upper Menihek and Willbob basalts had a similar origin. If this is correct, it can be concluded that fractional crystallization of plagioclase, and possibly other phases, has played a role in the compositional variation evident in the suites. Fractionation of plagioclase during differentiation is also suggested by the range of europium anomalies found in the basalts. By analogy with the intrusive rocks, where fractional crystallization was the predominant differentiation mechanism, much of the compositional variation evident in the basalts may have also originated through crystallization processes. In this case, the same incompatible elements that were found to be conserved during differentiation of the Menihek sills might also have been conserved in the evolution of the basalts. Consequently, TiO_2 , P_2O_5 , MnO , Zr , Y and Nb were chosen as potentially conserved elements. Since REE data is available for several samples from each basaltic suite, La/Ce values have also been examined. If these elements are not conserved, or if the suites compared are not comagmatic, the ratios will diverge.

The degree of alteration and metamorphism is higher in the Doublet terrane rocks than in those from the Howse zone. The coherent trends apparent in Figure 8.12, however, suggest that secondary processes have had relatively minor effects on the concentrations of the selected elements. This is confirmed by the constancy of incompatible element ratios. Statistical data for ratios involving are given in Table 8.1, and Figure 8.13 portrays variations in selected ratios for the various cycle 2 suites, as well as 2σ error bars due to analytical uncertainty. The diagrams reveal that, given analytical uncertainty, there are no significant differences in the ratios among intrusive, extrusive, aphyric or glomeroporphyritic members of the individual suites within each tectonic domain. Neither are there significant differences in the ratios of samples from the two domains (i.e., Howse zone versus Doublet terrane). This is also evident in Table 8.1 where, for each sample grouping, the average analytical uncertainty in the ratios exceed the associated standard deviations.

Table 8.1. Average, standard deviation and average analytical error for ratios constructed from potentially conserved incompatible elements in samples from cycle 2 tholeiitic basalts and gabbros from the Howse zone and Doublet terrane. Standard deviation and average analytical error are in bold typeface for constant ratios.

Sequences Compared	Ti/Zr	Ti/Y	P/Zr	P/Y	Mn/Zr	Mn/Y	Y/Zr	Nb/Zr	Nb/Y	La/Ce
Upper Menihek Basalts and Menihek Chilled Gabbros										
Average	235.09	478.28	19.24	39.28	44.70	90.89	0.482	0.073	0.161	0.391
Standard Deviation	18.04	49.21	2.49	6.50	4.81	11.69	0.046	0.026	0.064	0.031
Average error	18.23	67.63	2.52	6.93	11.16	25.11	0.074	0.036	0.081	0.031
Willbob Basalts and Doublet Gabbros										
Average	233.54	524.01	18.51	41.94	45.75	127.14	0.461	0.056	0.125	0.379
Standard Deviation	16.81	51.03	1.77	7.41	10.33	18.36	0.052	0.020	0.054	0.022
Average error	18.11	74.09	2.42	7.48	11.42	24.61	0.071	0.027	0.063	0.029
Cycle 2 Tholeiitic Basalts and Gabbros (Upper Menihek and Willbob Basalts, Menihek and Doublet Gabbros)										
Average	234.09	507.68	18.78	40.99	45.37	98.44	0.470	0.061	0.150	0.385
Standard Deviation	17.17	54.73	2.07	7.17	6.83	17.86	0.052	0.023	0.059	0.028
Average error	18.15	71.78	2.46	7.23	11.33	27.19	0.072	0.030	0.076	0.029

Despite the coherence of incompatible element ratios among the suites, there appear to be systematic variations in Ti/Zr and Mn/Zr between glomeroporphyritic and aphyric samples, with ratios in the latter lithology generally higher. As discussed in Chapter 5, this variation is more likely to reflect slight compatibility of Mn and Ti during

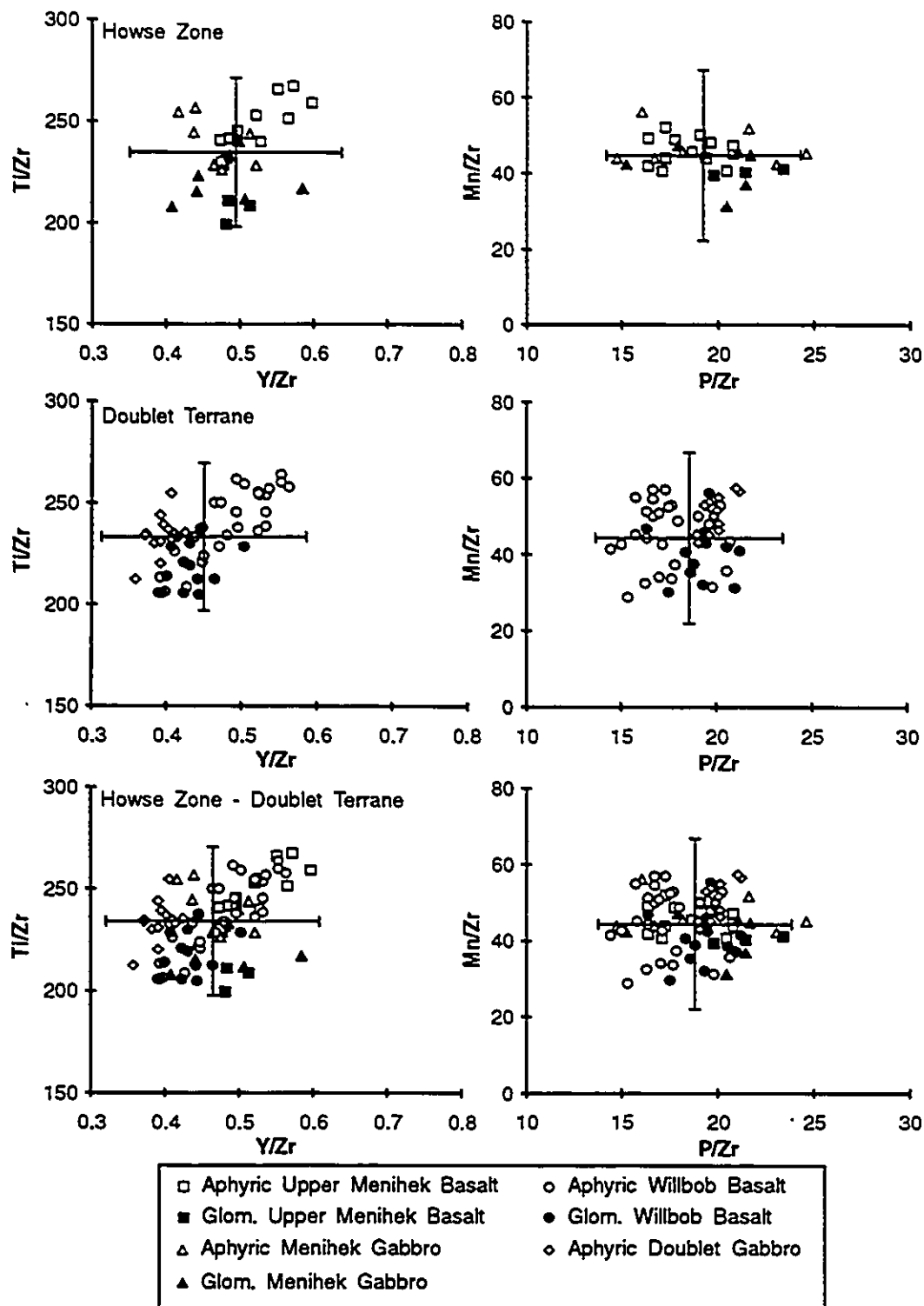


Figure 8.13. Incompatible element ratio diagrams illustrating the coherence of ratios within and among individual data groups for cycle 2 tholeiitic rocks of the Howse zone and Doublet terrane. The average 2σ analytical uncertainty is illustrated by error bars plotted at the average values of the ratios.

differentiation than as differences in the ratios of the primary melts. The lower Ti/Z and Mn/Zr of the relatively incompatible element enriched glomeroporphyritic rocks could then arise through either minor incorporation of Ti and Mn during crystal fractionation, or minor retention of these less incompatible elements during partial melting.

In light of the textural and geochemical similarities outlined earlier, and the absence of significant variation in incompatible element ratios, it can be concluded that the Willbob and upper Menihek basalts, and the Menihek and Doublet sills, are all comagmatic. If, as seems likely, the Willbob-Hellancourt correlation is correct, cycle 2 tholeiitic magmatism in the Labrador Trough produced a volcanic sequence locally in excess of 5 km thick, and extending for at least 700 km along strike, during an interval of approximately 10 Ma. From the ratio analysis it can also be concluded that P_2O_5 , Zr, Y, Nb, La and Ce were conserved during differentiation of the cycle 2 suites, while TiO_2 and MnO may have behaved as slightly compatible elements during the process. By analogy with the differentiation of the Menihek sills, the slight compatibility of these components suggests that pyroxene played a role in the geochemical evolution of the cycle 2 tholeiitic suite.

Le Fer Basalts and Gabbros

The Le Fer basalts represent the main igneous event of cycle 1 infilling identified in the central Labrador Trough. Gabbroic sills are intimately associated with the basalts, and these intrude all cycle 1 strata with the exception of the Denault dolostones. A felsic differentiate of one such intrusion in the Cramolet Lake area crystallized at 2169 Ma (Rohon *et al.*, 1993). The 2142 Ma age of the Mistamisk rhyolite (Dressler and Krogh, cited by Clark, 1984), which probably represents late cycle 1 volcanic activity (Clark and Thorpe, 1990), indicates that the Le Fer basalts and the Cramolet Lake sill are approximately coeval. This suggests that the spatially associated gabbros and basalts of the Le Fer Formation are comagmatic, and hence the intrusive rocks have been referred to as the Le Fer gabbros in the present work.

In comparison with the 1884-1874 Ma ages of glomeroporphyritic gabbros (Machado, 1990; Findlay *et al.*, 1995), which are comagmatic with cycle 2 tholeiites, Le Fer magmatism is older by 260-290 Ma. Clearly cycle 1 and 2 subalkaline basalt/gabbro

sequences represent temporarily distinct periods of igneous activity, yet the products of the two magmatic episodes are compositionally and lithologically very similar. The principal geochemical differences lie in the generally lower incompatible element contents and slightly higher MgO and Cr of the Le Fer suite. The Le Fer basalts also show slight, but consistent, light REE depletions relative to cycle 2 tholeiites.

Following the rationale of the previous section, incompatible element ratios were constructed from TiO_2 , P_2O_5 , MnO, Zr, Y, Nb, La and Ce using samples from the Le Fer basalts and gabbros to test the proposed comagmatic relationship, and to compare these values with those of the cycle 2 tholeiitic rocks. Statistical data are given in Table 8.2, and Figure 8.14 illustrates ratios involving selected elements. The Le Fer gabbros should exhibit incompatible element ratios similar to those of the Le Fer basalts if these suites are comagmatic, and if the elements were conserved during melt formation and differentiation processes. As expected, ratios in the Le Fer rocks are constant except in the case of Ti/Zr, where the standard deviation of the ratios exceeds the associated analytical uncertainty. This non-constancy does not reflect a difference between the intrusive and extrusive samples, since the ratio is also non-constant within the basaltic group alone. It is probable that the variation in Ti/Zr arose through limited compatibility of TiO_2 during differentiation, as seems to be the case in cycle 2 rocks. The apparent constancy of Ti/Y in the samples probably reflects the larger analytical uncertainty associated with measurement of Y in comparison to Zr.

Table 8.2. Average, standard deviation and average analytical error for ratios constructed from potentially conserved incompatible elements in samples of the Le Fer basalts and gabbros. Comparison of ratios from the Le Fer and cycle 2 tholeiitic suites is also shown. Standard deviation and average analytical error are in bold typeface for constant ratios.

Sequences Compared	Ti/Zr	Ti/Y	P/Zr	P/Y	Mn/Zr	Mn/Y	Y/Zr	Nb/Zr	Nb/Y	La/Ce
Le Fer Basalts and Gabbros										
Average	248.21	526.23	12.38	26.39	56.55	120.50	0.455	0.072	0.154	0.336
Standard Deviation	30.05	49.18	1.60	4.54	10.86	25.31	0.060	0.022	0.045	0.020
Average error	19.25	74.41	1.65	4.65	14.12	33.29	0.070	0.036	0.078	0.026
Le Fer Basalts and Gabbros, Cycle 2 Tholeiitic Basalts and Gabbros										
Average	238.80	500.61	16.61	35.00	47.97	100.47	0.468	0.065	0.147	0.375
Standard Deviation	23.39	42.16	3.30	7.82	9.98	20.93	0.052	0.023	0.057	0.033
Average error	18.52	70.78	2.17	6.17	11.98	27.75	0.072	0.032	0.074	0.029

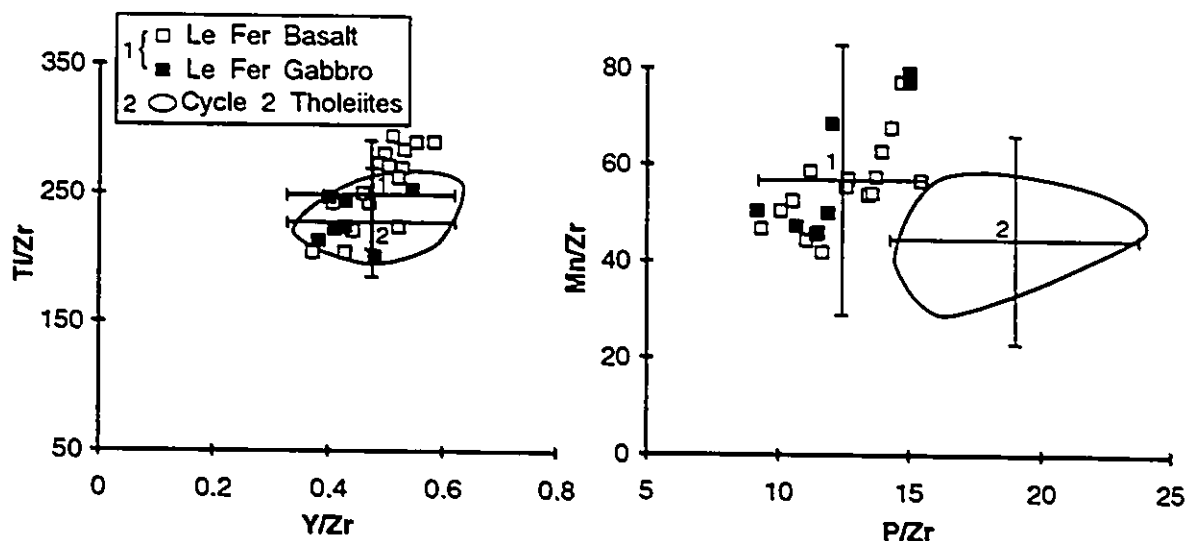


Figure 8.14. Incompatible element ratio diagrams illustrating the coherence of ratios among Le Fer gabbros and basalts. The range of values from cycle 2 tholeiitic rocks is also indicated. The average 2σ analytical uncertainty is illustrated by error bars plotted at the average values of the Le Fer ratios.

The approximate contemporaneity and intimate spatial association of the Le Fer basalts and gabbros, and the generally constant incompatible element ratios for the two groups, indicate that they are comagmatic. A comparison of the Le Fer data with cycle 2 tholeiitic basalts confirms the geochemical similarities of the two suites. With the exception of Ti/Zr, P/Zr, P/Y and La/Ce, the incompatible element ratios of cycle 1 and cycle 2 subalkaline rocks are indistinguishable within analytical error (Table 8.2). For equivalent Zr and Y concentrations, however, the Le Fer suite is slightly impoverished in P_2O_5 compared to cycle 2 tholeiites. The Le Fer suite also exhibits slightly higher MnO content (Fig. 8.14), although ratios involving this element are equivalent within error for cycle 1 and 2 rocks.

The similarity of incompatible element ratios in the cycle 1 and 2 suites, coupled with the overall geochemical similarities, suggests that the mantle source regions of the two tholeiitic suites were compositionally similar, and resultant magmas underwent similar differentiation processes. Cycle 1 basalts, however, are slightly more primitive (i.e., lower incompatible element contents, higher MgO and Cr, light REE depleted) than cycle 2 basalts. The less evolved nature of the Le Fer sequence is also suggested by the common

occurrence of olivine phenocrysts and microphenocrysts, whereas pyroxene is the predominant ferromagnesian phenocryst phase in the cycle 2 basalts. It is also apparent that the two magmatic episodes differed in the relative importance of plagioclase fractionation during differentiation, given that highly plagioclase-glomeroporphyritic basalt and gabbro are volumetrically significant members of the cycle 2 suite. Finally, incompatible elements suggest that cycle 1 magmas were slightly impoverished in P_2O_5 , and possibly slightly enriched in MnO, relative to cycle 2 melts.

Murdoch Basalts

The stratigraphic position of the Murdoch Formation relative to the Knob Lake sequence is problematical. This, in part, is due to the fact that sedimentary lithologies clearly correlative with Knob Lake strata do not occur east of the Walsh Lake Fault. The only potential exception, the Irene Lake iron formation, appears to be in tectonic contact with surrounding rocks (Clark, personal communication, 1994). The upper Menihek-Willbob correlation indicates a pre-upper Menihek age for the Murdoch Formation, while the presence of ironstone-bearing conglomerate in the lower part of the sequence (Frarey, 1952) suggests a post-Sokoman age. If the Murdoch-Thompson Lake-Willbob succession is conformable, then the Murdoch and Thompson Lake formations are the stratigraphic equivalents of the lower and middle portions of the Menihek Formation. Thus the Thompson Lake slates and middle Menihek turbidites would be correlative, while the Murdoch Formation corresponds to the turbidites, basalts and pyroclastic rocks of the lower Menihek Formation (Fig. 8.15A).

The volcanic rocks of the Murdoch Formation examined in this study range from mildly alkaline basalts which share some of the geochemical characteristics of the lower Menihek and Nimish suites, to subalkaline basalts similar to cycle 1 and 2 tholeiites. Although the database is limited, it appears that the alkalic volcanics which occur locally in the lower part of the formation are volumetrically minor in comparison to the subalkaline basalts. If the Murdoch-lower Menihek correlation is correct, then the slightly alkaline Murdoch basalts could be equivalent to the lower Menihek alkaline suite, while the more voluminous subalkaline Murdoch basalts represent an early expression of cycle 2 tholeiitic magmatism. Subalkaline basalts, however, are restricted to the upper Menihek

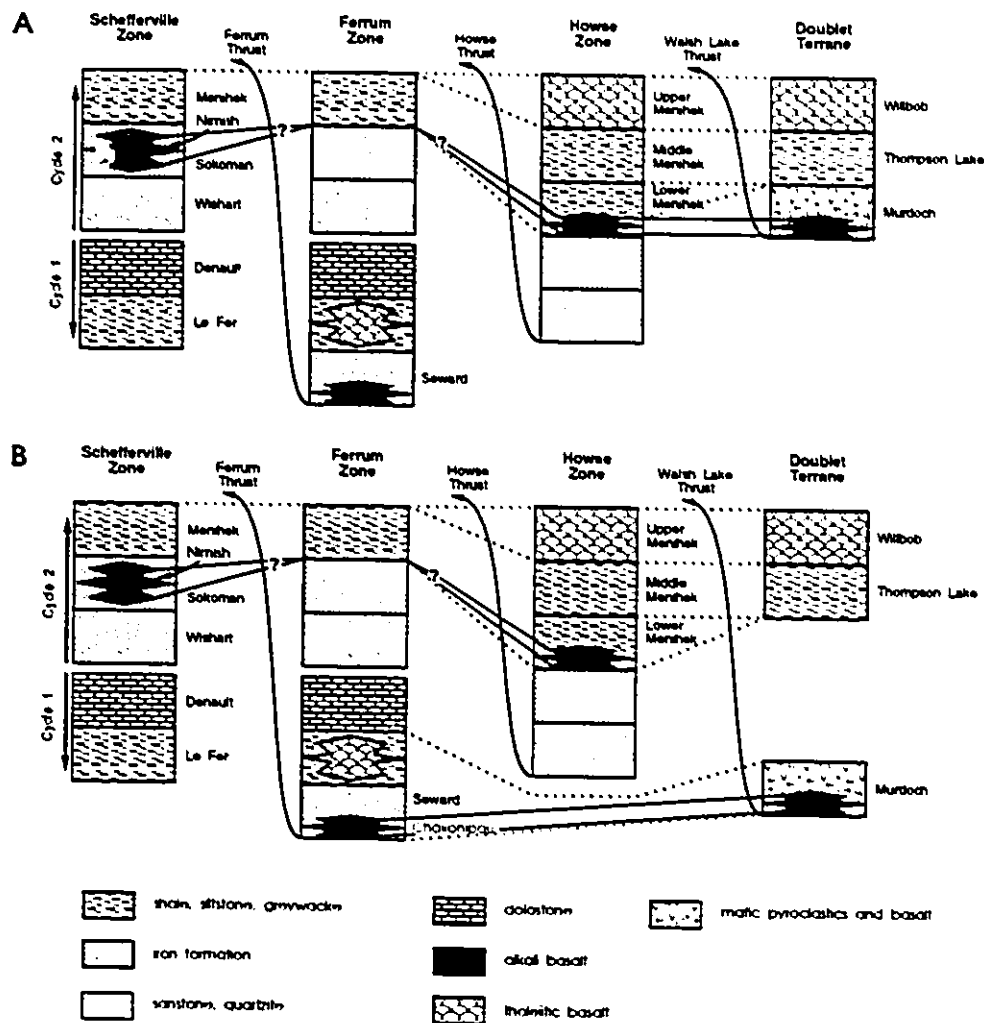


Figure 8.15. Simplified stratigraphic column of the central Labrador Trough illustrating possible correlations between the Murdoch Formation and members of the Knob Lake Group. (A) The Murdoch Formation represents early cycle 2 magmatism. (B) The Murdoch Formation is equivalent to cycle 1 Le Fer magmatism. Pertinent correlations are shown by dashed lines; solid lines illustrate possible correlations of alkaline volcanics. Unit thickness is arbitrary.

Formation in the Howse zone. This implies that early cycle 2 tholeiitic magmatism (i.e., Murdoch) was localized in the Doublet terrane.

A critical requirement of the correlations discussed above is that the Murdoch-Thompson Lake-Willbob sequence is conformable. Baragar (1967) and Frarey (1967) found no evidence of a non-conformable or tectonic contact between the Murdoch and Thompson Lake formations, but the actual contact is not exposed. This opens the

possibility that the Murdoch Formation is stratigraphically equivalent to the Lower Knob Lake Group. Thus the mildly alkaline Murdoch basalts could be correlative with the alkali basalts of the Chakonipau Formation, while the subalkaline sequence is correlative with the Le Fer basalts (Fig. 8.15B). These relationships would explain the absence of lower to middle Menihék subalkaline volcanism in the Howse zone. This interpretation, however, requires either a non-conformable or a tectonic contact between the Murdoch and Thompson Lake formations. It would also require an early (cycle 1) period of ironstone deposition to explain the ironstone cobbles in the Murdoch conglomerate.

Following the reasoning outlined earlier, incompatible element ratios in the Murdoch basalts can be compared to values found in cycle 1 and 2 samples to test possible associations. Ratio data are summarized in Table 8.3, and selected ratios are illustrated in Figure 8.16. The results of the analysis indicate that the two-fold subdivision of the Murdoch basalts suggested earlier is probably justified. When all samples are considered together, incompatible element ratios are highly variable and generally non-constant given analytical uncertainty. The subalkaline samples exhibit ratios comparable to those of cycle 1 and cycle 2 tholeiites, while the high-Zr samples tend toward values found in the lower Menihék alkaline rocks (Fig. 8.16). The low-Zr group by itself, however, exhibits relatively coherent ratios. Unfortunately, the number of samples available is insufficient to establish whether two, geochemically distinct volcanic series are present, or whether the Murdoch basalts simply represent a compositionally diverse but gradational volcanic sequence. Further sampling is required to resolve this question, but given the difference in stratigraphic position of the two groups of samples they are here treated as discrete series.

With the exceptions of P/Zr and P/Y , the extent of variations in all of the incompatible element ratios reported in Table 8.3 is less than the associated uncertainties for the low-Zr samples. The significance of the non-constancy of P/Zr and P/Y is impossible to determine given the limited database, but they most probably reflect non-conservation of P relative to Zr and Y at some stage in the history of the basalts. The alternative explanation is that the variations reflect primary differences in parental melt compositions, and therefore the samples are not comagmatic. This is unlikely given the close spatial relationships among the samples and the otherwise similar ratios. Since the Murdoch suite has undergone comparatively strong alteration, and considering the relative

Table 8.3. Average, standard deviation and average analytical error for ratios constructed from potentially conserved incompatible elements in the Murdoch basalts. Comparisons with cycle 1 and cycle 2 subalkaline rocks are also indicated. Standard deviation and average analytical error are in bold typeface for constant ratios.

Sequences Compared	Ti/Zr	Ti/Y	P/Zr	P/Y	Mn/Zr	Mn/Y	Y/Zr	Nb/Zr	Nb/Y	La/Ce
Murdoch Basalts (all samples)										
Average	259.45	688.82	23.58	64.68	24.19	60.88	0.400	0.057	0.174	0.402
Standard Deviation	63.91	299.26	9.67	36.58	5.72	12.59	0.108	0.036	0.146	0.018
Average Error	20.12	97.40	3.09	11.41	6.04	16.82	0.062	0.028	0.088	0.031
Low-Zr Murdoch Basalts										
Average	233.79	504.26	17.48	39.36	28.33	62.48	0.450	0.035	0.109	0.399
Standard Deviation	6.34	67.97	4.14	14.51	1.90	9.34	0.066	0.004	0.030	0.020
Average Error	18.13	71.30	2.29	6.94	7.07	17.26	0.069	0.017	0.055	0.030
Low-Zr Murdoch Basalts, Cycle 1 Tholeiites										
Average	244.23	521.68	13.43	29.07	50.71	108.49	0.454	0.064	0.137	0.361
Standard Deviation	27.92	52.99	3.19	9.08	15.12	33.03	0.060	0.025	0.045	0.039
Average Error	18.94	73.76	1.76	5.13	12.66	29.97	0.070	0.031	0.069	0.027
Low-Zr Murdoch Basalts, Cycle 2 Tholeiites										
Average	233.52	490.52	18.53	39.19	43.13	90.11	0.472	0.058	0.124	0.384
Standard Deviation	17.73	42.27	2.36	7.06	8.00	14.56	0.050	0.020	0.049	0.023
Average Error	18.11	69.36	2.42	6.91	10.77	24.89	0.073	0.029	0.063	0.029

mobility of P, the most reasonable interpretation is that non-conservation of P arose during post-crystallization alteration and metamorphism.

Comparisons of ratio data from the subalkaline Murdoch basalts with those from cycle 1 and 2 tholeiitic rocks are also provided in Table 8.3. It is apparent that the incompatible element ratios of the low-Zr Murdoch suite are more consistent with a genetic relationship between the Murdoch and cycle 2 tholeiitic rocks than with a Murdoch-cycle 1 association. This is also evident in Fig 8.16, where the low-Zr Murdoch samples plot close to, or within, the field of cycle 2 tholeiites in each diagram. In addition, the average ratios for the entire Murdoch suite are closer to the average values for cycle 2 rocks. It is also apparent that the high-Zr samples trend away from the tholeiitic fields in the diagrams, and typically show incompatible element ratios intermediate between those of the cycle 2 tholeiitic suite and the lower Menihek volcanics. The transitional nature of incompatible element ratios in the high-Zr Murdoch samples is in accord with the geochemical characteristics described earlier. Thus while the predominant volcanic product of Murdoch magmatism appears to have been subalkaline basalt compositionally

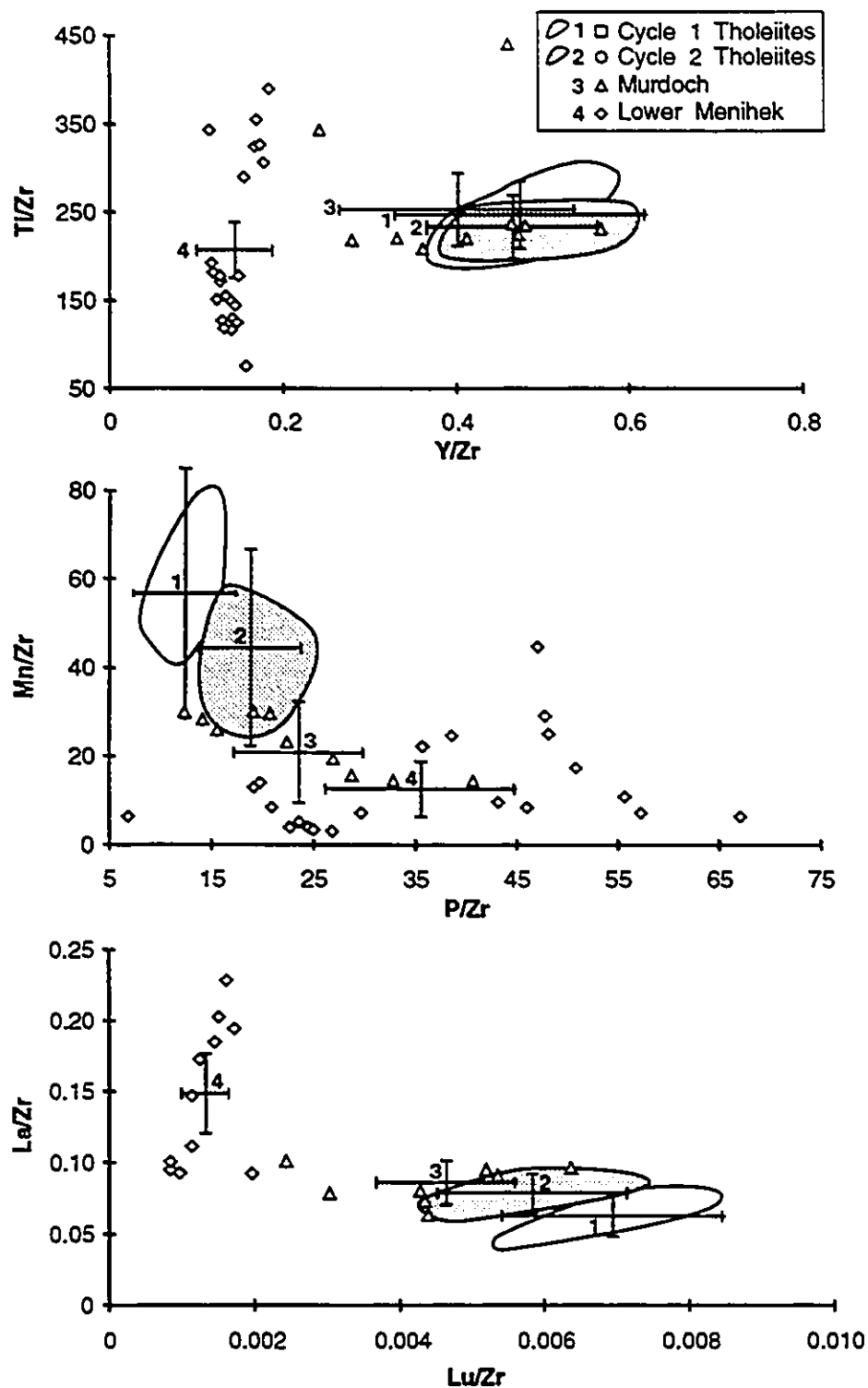


Figure 8.16. Incompatible element ratio diagrams illustrating variations among the Murdoch and lower Menihek volcanic rocks, and the cycle 1 (Le Fer) and cycle 2 (upper Menihek-Willbob) tholeiites. The average 2σ error due to analytical uncertainty is illustrated by error bars plotted at the average values of the ratios for each group of samples.

similar to the cycle 2 upper Menihek-Willbob sequence, early Murdoch magmatism seems to have produced slightly alkaline rock types.

The limited number of samples obtained from the Murdoch Formation precludes an extensive examination of compositional variation in the basaltic rocks. The available data, however, suggest that the Murdoch subalkaline basalts are related to cycle 2 tholeiitic magmatism. This indicates that the first interpretation of stratigraphic relationships discussed earlier is the more likely; i.e., the Murdoch and Thompson Lake formations are equivalent to the lower and middle portions of the Menihek Formation, respectively (Fig. 8.15A). Thus it is probable that the high-Zr Murdoch basalts which occur in the lower part of the formation are, in part, equivalent to the lower Menihek alkaline basalts. Although the high-Zr Murdoch basalts lack the strongly alkaline geochemical signatures typical of the lower Menihek suite, they may represent a transitional series between the latter-type compositions and overlying subalkaline rocks. This interpretation is consistent with the intermediate values of incompatible element ratios in the high-Zr samples, and with the transitional nature of the extended trace element patterns relative to those of the low-Zr Murdoch basalts and the lower Menihek suite.

The Murdoch-lower Menihek correlation is supported by the fact that both sequences appear to immediately predate the onset of more rapid basin subsidence. The amygdaloidal nature of the lower Menihek volcanics, and the presence of pyroclastic rocks, suggest that water levels were relatively shallow at the time of their eruption. Relatively shallow water levels are also suggested by the occurrence of interbedded jasper and shale immediately below the volcanic horizons, since these appear to record the Sokoman-Menihek transition. The turbiditic nature of the middle Menihek sediments, however, suggests that water depths had increased by this time. The Murdoch Formation includes abundant pyroclastic rocks which would also seem to indicate relatively shallow water conditions. As with the middle Menihek turbidites, however, the Thompson Lake slates indicate that water conditions had deepened by the time of their deposition. Consequently, it is probable that the lower Menihek and Murdoch sequences are stratigraphically equivalent, and both represent deposits formed prior to or during more rapid basin subsidence. The principal inconsistency of this correlation lies in the absence of subalkaline volcanic rocks in the lower to middle Menihek Formation (i.e., equivalent to

the low-Zr Murdoch basaltic suite). This may indicate that subalkaline volcanism was initially confined to areas east of the Howse zone.

Lower Menihek, Nimish, and High-Zr Murdoch Basalts

In comparison to cycle 2 tholeiites, the lower Menihek basalts are incompatible element and light REE enriched. The incompatible element ratios of the lower Menihek volcanics are also distinctly different from those of the tholeiitic rocks (Fig. 8.16). The volcanic rocks of the Nimish Formation, however, share the alkaline characteristics of the lower Menihek suite (Fig. 8.17). Both suites are distinguished by elevated alkali element concentrations and high incompatible element contents (Fig. 8.4), particularly TiO_2 , P_2O_5 , Zr, Y and Nb. Both suites also show elevated REE concentrations and light REE enriched patterns (Figs. 8.6 and 8.18). In addition, the extended trace element patterns of the Nimish and lower Menihek basalts are quite similar; both suites exhibit generally elevated high field strength (HFS) element contents, and relative Sr, Zr, Y and heavy REE depletions and Ti enrichment (Fig. 8.19). The principal difference in the patterns lies in the K and Rb depletions that are found in many of the lower Menihek samples. This may reflect loss of these elements during alteration, as appears to be the case in the lower Menihek clasts. The fact that one of the four lower Menihek patterns illustrated in Figure 8.19 does not show such depletions supports this hypothesis. It is also possible, however, that the elevated K and Ba content of the Nimish suite reflects elevated concentrations of these elements in the magma source, or the addition of these elements through crustal assimilation.

In addition to geochemical similarities, the Nimish and lower Menihek volcanics occupy similar stratigraphic positions. The Nimish extrusions are concentrated in the Sokoman Formation, but also occur in the lowest part of the Menihek Formation (Evans, 1978; Wardle, 1979) in the Schefferville zone. The lower Menihek volcanic and pyroclastic rocks occur in the basal portion of the Menihek Formation, immediately above the Sokoman-Menihek transition in the Howse zone. Both suites are overlain by the Menihek turbidites. Since the Sokoman Formation is restricted to thin, poorly exposed horizons at the western boundary of the Howse zone, it is unknown whether alkalic volcanics were erupted during ironstone deposition in this domain.

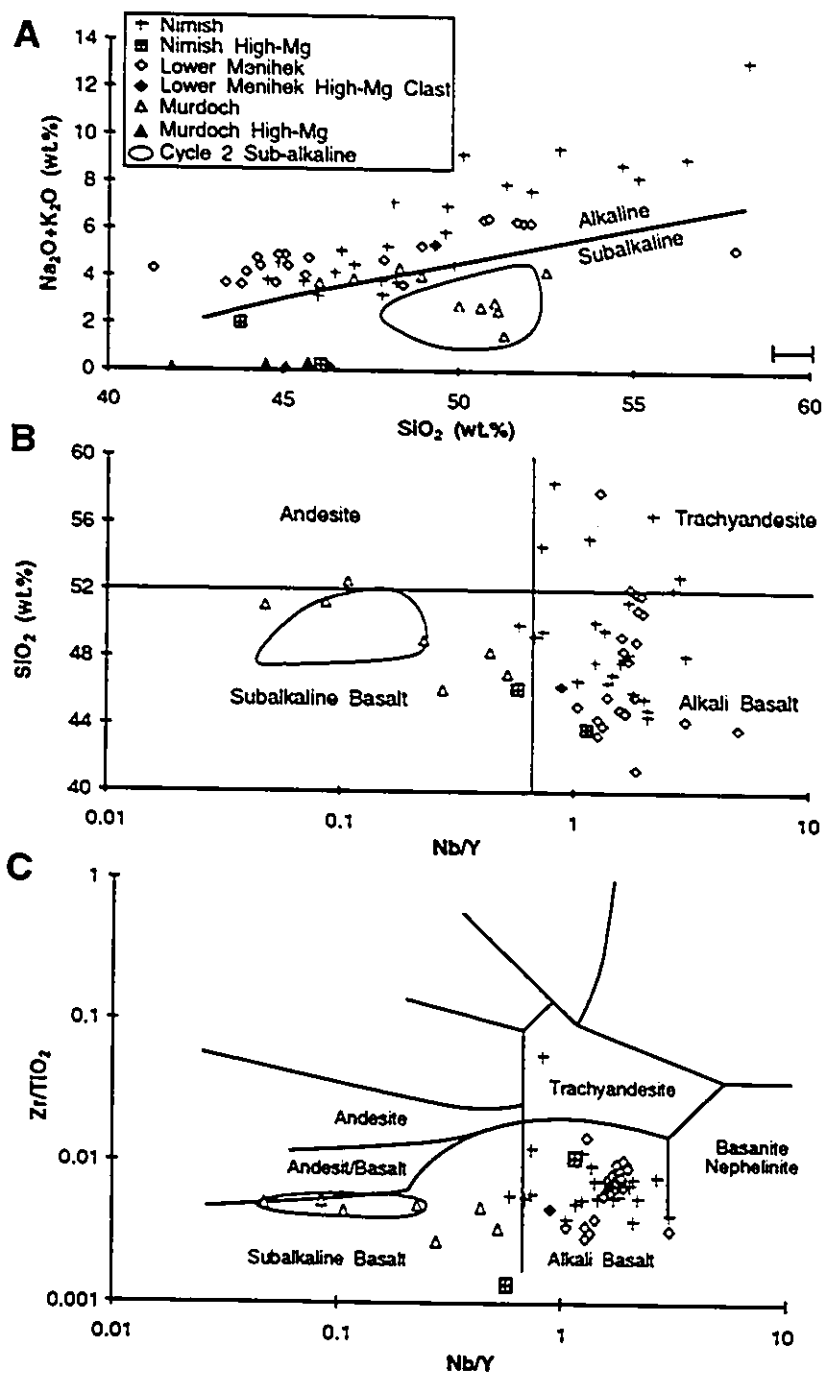


Figure 8.17. Comparison of the geochemical affinities of the Nimish, lower Menihek and Murdoch volcanic suites. (A) Total alkalis-silica plot with dividing line from Irvine and Baragar (1971). (B) SiO_2 - Nb/Y plot with subdivision from Winchester and Floyd (1977). (C) Zr/TiO_2 - Nb/Y plot with subdivision from Winchester and Floyd (1977). Nimish data is from Watanabe (personal communication, 1993), and high-Mg Murdoch volcanic samples are from Gebert (1988).

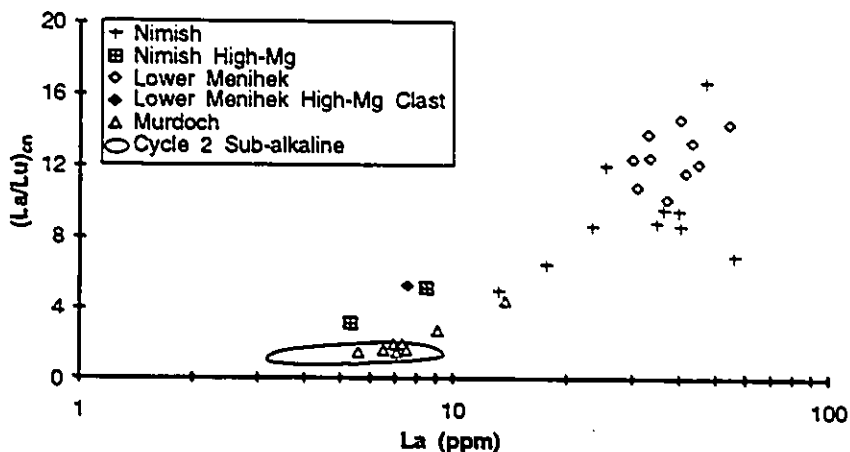


Figure 8.18. Plot of chondrite-normalized La/Lu versus La concentration illustrating the elevated REE concentrations and light REE enrichments typical of the volcanic rocks of the Nimish, lower Menihek and Murdoch formations in comparison to cycle 2 tholeiitic rocks. The high-Mg samples associated with Nimish and lower Menihek magmatism also exhibit elevated La/Lu, although absolute concentrations of La are comparable to those of the tholeiitic basalts. The Nimish data are from Watanabe (personal communication, 1993)

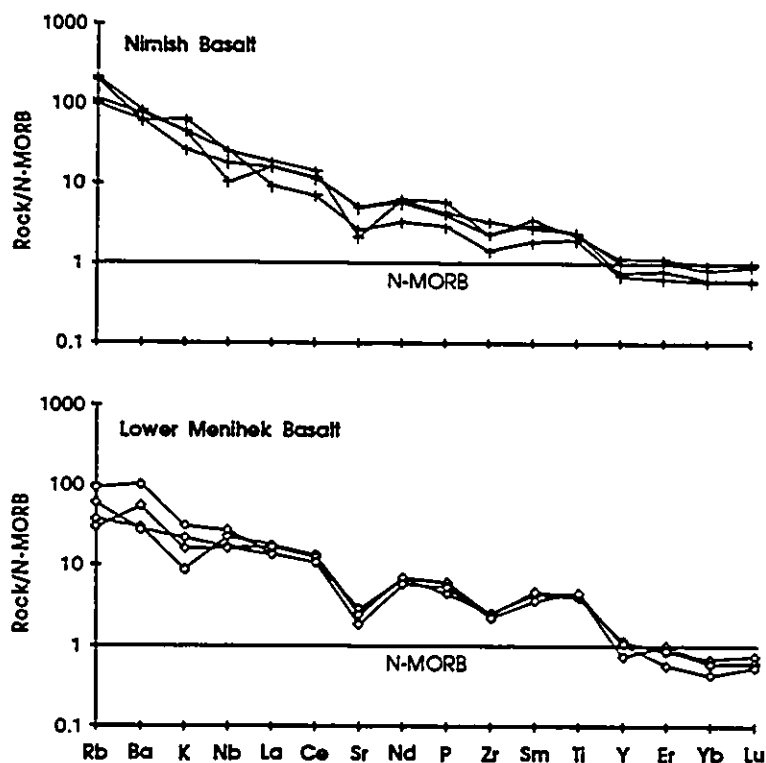


Figure 8.19. Normalized extended trace element plots for basaltic rocks from the Nimish and lower Menihek formations. Normalizing values are for normal-type mid-ocean ridge basalt (N-MORB) from Sun and McDonough (1989) except Ba, which is from Saunders and Tarney (1984).

The high-Zr Murdoch basalts exhibit some of the geochemical characteristics of the Nimish and lower Menihek suites, although their alkalinity is less pronounced. In the preceding section, the Murdoch-Thompson Lake succession was correlated with the lower to middle Menihek Formation, and the high-Zr basalts found in the lower part of the Murdoch Formation were considered the stratigraphic equivalents of the lower Menihek alkaline sequence. If this interpretation is correct, alkaline or slightly alkaline rocks occur in the Schefferville and Howse zones, and in the Doublet terrane, at approximately the same stratigraphic level. This suggests that the Nimish, lower Menihek, and lower Murdoch formations are correlative, as illustrated in Figure 8.15A.

Correlation of the three suites is further supported by the occurrence of intraformational conglomerates in each succession. Conglomerates in the Nimish Formation contain clasts of iron formation and Nimish volcanic and plutonic rocks, and these attest to at least local uplift and tectonic instability contemporaneous with alkaline magmatism. Similarly, conglomerates containing ironstone and basaltic clasts are intercalated with the alkaline volcanic and pyroclastic rocks at Howse Lake. The Outlook Lake conglomerate, which also contains clasts of jasper and alkali basalt, may be the equivalent of the Howse Lake horizon, although it appears to be located at a stratigraphically higher level. An ironstone-bearing conglomerate was also described from the lower part of the Murdoch Formation at Walsh Lake (Frarey, 1952). Although the provenance of the ironstone clasts has not been conclusively established in any of the conglomerates, the Sokoman Formation is the most likely source. This is particularly true in the case of the Nimish conglomerate which contains clasts of Nimish igneous volcanic rocks which are lithologically and geochemically similar to flows intercalated with the Sokoman Formation (Findlay *et al.*, 1995). Whatever the source of the ironstone clasts, it is apparent that local uplift and erosion was associated with periods of alkaline magmatism in each case. It is noteworthy that this tectonic instability and localized magmatic activity immediately preceded a period of relatively rapid basin subsidence.

One further feature common to the three suites is the association of high-Mg rocks with the alkaline volcanics. The Nimish Formation includes limited amounts of high-Mg intrusive rocks which were originally composed primarily of clinopyroxene and olivine (Watanabe and Fowler, 1989). The high-Mg clast found in the Outlook Lake conglomerate

is presumably related to lower Menihek magmatism since the horizon underlies the upper Menihek basalts, and since alkali basalt clasts are present in the same conglomerate. The sample has been completely recrystallized, however, and it is impossible to ascertain whether it originally formed part of an intrusive or extrusive body. The lower Murdoch Formation also includes high-Mg rocks in form of volcanics in the vicinity of Frederickson Lake (Gebert, 1988). Although the number of samples is limited, the three examples of high-Mg Murdoch rocks are geochemically similar and appear to be more closely related to the alkaline magmatic activity than to cycle 2 tholeiitic volcanism. The samples, which contain 19-23% MgO, generally fall on the extensions of covariation trends defined by the Nimish-lower Menihek volcanic rocks (Fig. 8.20). In particular, the samples tend to have higher TiO_2 (1.57-2.06 wt. %) and P_2O_5 (0.12-0.15 wt. %) concentrations than might be expected from ultramafic extrusions associated with cycle 2 tholeiitic magmatism. For example, a relatively primitive (10.7 wt. % MgO) Willbob basalt reported by Beaudoin and Laurent (1989) contains only 0.69 wt. % TiO_2 and 0.04 wt. % P_2O_5 . In addition, the Nimish and lower Menihek high-Mg samples (comparable data are not available for the Murdoch high-Mg samples) show the light REE enrichments typical of the alkaline volcanics (Figs. 8.18 and 8.21), although the degree of enrichment and the total REE concentrations are lower. The geochemical characteristics of the high-Mg samples are thus consistent with a genetic relationship with the alkaline suites.

The association of high-Mg rocks with the alkaline suite is particularly interesting in the case of the Murdoch examples, since these appear to be of extrusive origin. While the Nimish ultramafic rocks may be cumulates derived from fractional crystallization of moderately MgO-rich basaltic magmas, the Murdoch samples indicate that highly magnesian melts were produced, and erupted, at the time of formation of the alkaline sequences. In contrast, the only high-Mg rocks associated with the overlying tholeiitic suite are the ultramafic cumulates which form part of differentiated cycle 2 sills (Retty Peridotite). No high-Mg lavas have been reported from the tholeiitic succession (although Beaudoin and Laurent (1989) report one flow with ~11% MgO), and the samples obtained in this study consist entirely of relatively evolved, low-Mg basalts.

The correspondence of geochemical characteristics and stratigraphic position suggests that the Nimish, lower Menihek and lower Murdoch suites had a similar origin.

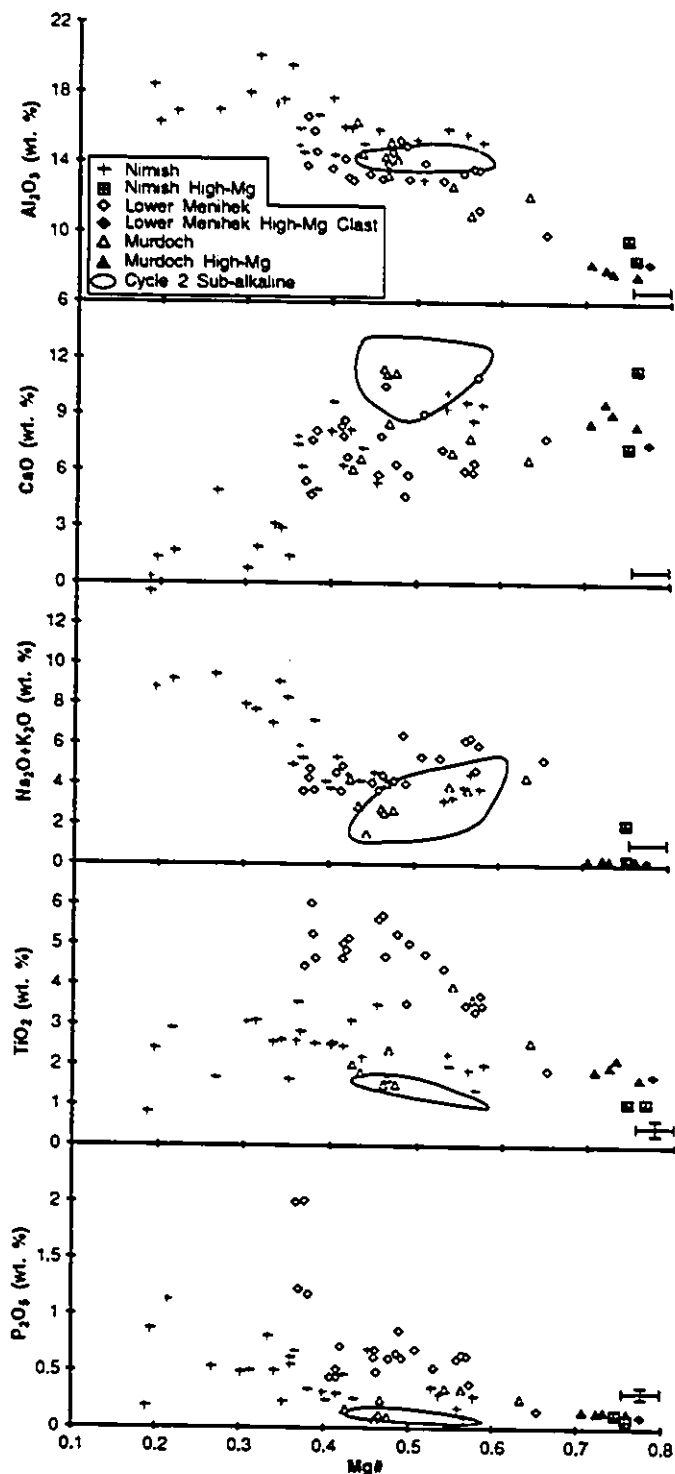


Figure 8.20. Covariation of selected major elements with Mg# in the Nimish, lower Menihok and Murdoch volcanic rocks. The limited data suggest that the high-Mg samples are members of the alkaline suites rather than part of the cycle 2 association. Nimish data from Watanabe (personal communication, 1993) and high-Mg Murdoch data from Gebert (1988). 2σ errors due to analytical uncertainty are indicated by error bars where greater than plot symbol size.

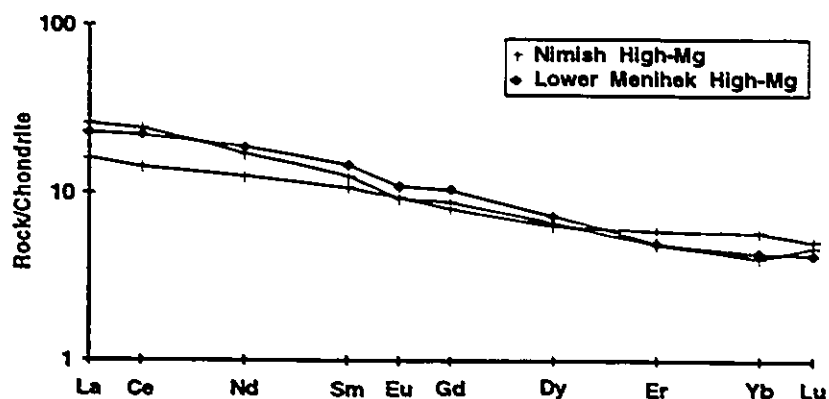


Figure 8.21. Chondrite-normalized REE patterns from the Nimish ultramafic rocks and the lower Menihek high-Mg clast. Normalizing values are from Nakamura (1978).

probably related to specific tectonic conditions. Despite these features, the limited lateral continuity and thickness of alkaline volcanism in each sequence suggest that these rocks were the results of local generation of discrete batches of magma, rather than the comagmatic products of widespread alkaline magmatism. This is confirmed in the case of the Nimish and lower Menihek volcanics by the fact that the former (Schefferville zone) post-date cycle 2 tholeiitic magmatism in the Howse zone, and are therefore younger than the lower Menihek basalts. If the alkaline rocks have a common origin, the fact that they are not coeval indicates that the necessary tectonic environment developed at different times in different parts of the basin. This is consistent with the time-transgressive nature of cycle 2 sedimentation (Findlay *et al.*, 1995).

Although the Nimish, lower Menihek and high-Zr Murdoch basalts share some geochemical characteristics, they are a compositionally diverse group of rocks. This is particularly true of the Nimish Formation which includes pyroxenitic to syenitic intrusions and basaltic to rhyolitic extrusions. In keeping with this diversity, incompatible element ratios in samples from the suites are highly variable. (Table 8.4). Of the selected ratios, only La/Ce is consistently within error of constancy in any of the sample groupings. Y/Zr values are constant given analytical uncertainty in the lower Menihek clasts, and this ratio is also constant when all lower Menihek samples are considered together. In all other cases, the variation in the ratios exceeds that expected from analytical uncertainty. The diagrams of Figure 8.22 illustrate the high degree of variability of the ratios, particularly

Table 8.4. Average, standard deviation and average analytical error for ratios constructed from potentially conserved incompatible elements in samples from the lower Menihék, Nimish and high-Zr Murdoch suites. Standard deviation and average analytical uncertainty are in bold typeface for constant ratios.

Sequences Compared	Ti/Zr	Ti/Y	P/Zr	P/Y	Mn/Zr	Mn/Y	Y/Zr	Nb/Zr	Nb/Y	La/Ce
Lower Menihék Flows										
Average	278.30	1829.86	36.91	241.73	20.74	136.84	0.153	0.218	1.477	0.434
Standard Deviation	98.72	632.25	16.42	107.58	11.13	72.44	0.026	0.048	0.522	0.015
Average error	21.58	258.73	4.83	42.63	5.18	37.80	0.024	0.019	0.214	0.033
Lower Menihék Clasts										
Average	145.47	1075.72	34.37	255.08	5.88	43.277	0.136	0.232	1.707	0.415
Standard Deviation	95.09	582.42	15.48	111.56	10.64	68.327	0.020	0.035	0.371	0.020
Average error	11.28	152.10	4.50	44.99	1.47	11.955	0.021	0.020	0.249	0.032
Lower Menihék Flows and Clasts										
Average	205.85	1418.51	35.52	249.01	12.63	85.80	0.144	0.225	1.602	0.422
Standard Deviation	95.09	582.42	15.48	111.56	10.64	68.33	0.020	0.040	0.371	0.020
Average error	15.96	200.57	4.65	43.92	3.15	23.70	0.022	0.019	0.233	0.031
Nimish Volcanics										
Average	172.21	1076.55	32.41	207.84	18.87	112.61	0.160	0.227	1.494	0.434
Standard Deviation	69.57	421.06	14.68	100.50	14.51	78.02	0.044	0.091	0.644	0.019
Average error	13.35	152.22	4.24	36.66	4.71	31.11	0.025	0.019	0.217	0.033
Lower Menihék Flows and Clasts, Nimish Volcanics, High-Zr Murdoch Basalts										
Average	197.65	1218.14	33.80	217.54	16.05	96.49	0.170	0.217	1.450	0.424
Standard Deviation	88.88	516.91	14.50	108.03	12.54	71.93	0.063	0.075	0.604	0.020
Average error	15.33	172.24	4.42	38.37	4.01	26.65	0.026	0.019	0.210	0.032

with respect to the relatively limited range of values found in all of the cycle 2 tholeiitic sequences.

The difference in age of the Nimish and lower Menihék igneous rocks indicates that they are not comagmatic. This interpretation is supported by the localized nature of the magmatic activity. The high-Zr Murdoch volcanics also appear to represent the products of localized magmatism, and hence are unlikely to be comagmatic with the other suites. The non-constancy of ratios among the three sequences is to be expected if they are the products of discrete magmas since many factors may have influenced initial ratios (i.e., source composition, melting conditions, etc.). But the extent of ratio variability within individual suites requires that either the samples are not the products of a single homogeneous partial melt, or that the selected elements were not conserved during melt generation, evolution or crystallization. In addition, there are the potential effects of

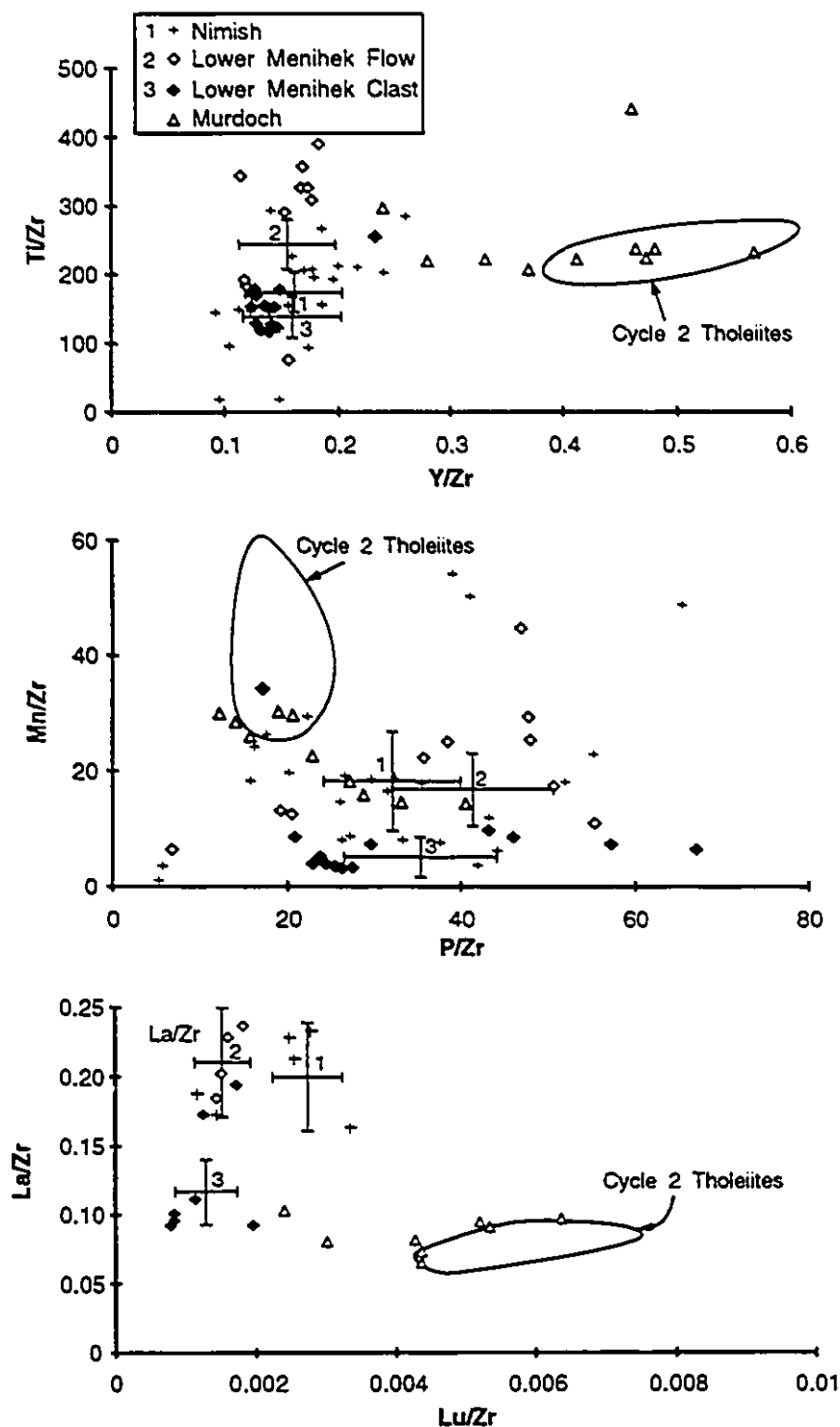


Figure 8.22. Incompatible ratio diagrams illustrating the variations in ratios among the Nimish, lower Menihék and Murdoch volcanic suites. The range of values for cycle 2 tholeiitic rocks is also shown. The average 2σ error due to analytical uncertainty is illustrated by error bars plotted at the average values of the ratios for each suite.

post-crystallization mobilization. For example, the depleted mobile LIL element contents of the lower Menihek clastic samples relative to the flows is thought to reflect the loss of these elements during alteration. The concentrations of P and Mn might also have been affected by post-crystallization processes. It seems unlikely, however, that ratios involving relatively immobile elements such as Ti, Zr, Y and Nb would be changed by alteration and metamorphism, particularly in the comparatively well preserved basalts of the Nimish and lower Menihek formations. The constancy of ratios in the cycle 2 rocks, which have been subject to similar post-crystallization effects, also argues against significant geochemical reconstitution as the source of variation in incompatible element ratios.

If post-crystallization changes are ruled out as the source of ratio variability within each sample grouping, at least in the case of the immobile elements, then the phenomenon must be due to either non-conservation of the elements during melt evolution, or to differing primary melt concentrations. The latter explanation requires that each grouping includes samples derived from discrete magma batches of varying composition. This is certainly possible in the case of the Nimish volcanics in that at least two volcanic centres, each presumably fed by a separate magma chamber, are present in the outcrop area of the suite (Evans, 1978). The two (or more) magma chambers may in turn have been fed by compositionally distinct partial melts. Compositional differences in the primary partial melts may have arisen through slight differences in source composition or melting conditions. Since the source of the lower Menihek clasts is not necessarily the volcanic horizons at Howse Lake, this explanation might also apply to compositional differences between flows and clasts. Multiple partial melts, however, seems an improbable explanation for compositional diversity in the stratigraphically, areally and volumetrically restricted lower Menihek basalts at Howse Lake.

The non-conservation of the selected elements during differentiation of the parental melts through fractional crystallization is another potential source of ratio variability. Some Nimish basalts, for example, contain Fe-Ti oxide phenocrysts (Evans, 1978). Fractional crystallization of such a phase would obviously influence the Ti content, and hence ratios involving this element, in sequentially erupted lavas. To be the source of the observed deviations, however, the involvement of phases in which all of the selected elements could be accommodated (i.e., zircon, apatite, etc.) would be necessary. The

extended trace element patterns of Nimish rhyolites reveal strong Ti and P depletions (Fig. 8.23) which indicate that minerals (e.g., Fe-Ti oxides and apatite) containing these elements as major components were fractionated prior to eruption. But comparable anomalies are not found in the trace element patterns of the basaltic rocks, and hence pre-eruption fractionation of such phases from the basic liquids is unlikely. Consequently, it is probable that ratio variability within the individual suites arose through a combination of factors, possibly including: (1) source heterogeneity, (2) non-conservation of incompatible elements through differences in melting conditions, degrees of partial melting, crystal fractionation, contamination or other mechanisms, and (3) element mobility during alteration. Such mechanisms would also have contributed to the large degree of geochemical diversity and the general lack of coherent inter-element covariation trends which characterize the alkaline suites.

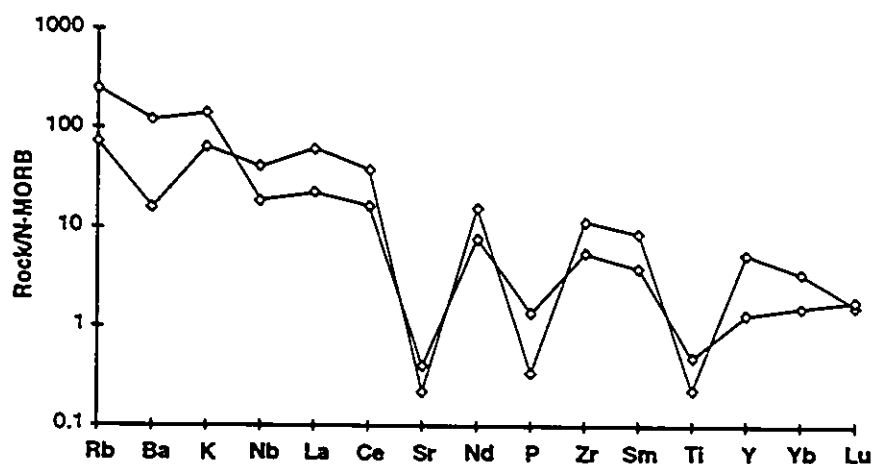


Figure 8.23. Normalized extended trace element plots for Nimish rhyolites illustrating the strong Ti and P depletions that are thought to reflect fractionation of Fe-Ti oxides and apatite. Data from Watanabe (personal communication, 1993). Normalizing values are for normal-type mid-ocean ridge basalt (N-MORB) from Sun and McDonough (1989) except Ba, which is from Saunders and Tarney (1984).

8.5 Sources of Geochemical Variation

The volcanic rocks of the LTBS exhibit compositional variations which may have arisen through five principal types of processes; partial melting, crustal contamination,

magma mixing, fractional crystallization, and alteration/metamorphism. The tholeiitic suites represent the products of large volumes of relatively homogeneous magmas. Samples from the suites exhibit generally coherent inter-element covariation trends and constant incompatible element ratios. These features suggest that either the parental magmas were derived through similar degrees of partial melting of a relatively homogeneous source, or compositionally variable parental magmas were homogenized prior to eruption. The absence of high incompatible element concentrations indicates that if crustal material was added to mantle magmas, the extent of contamination was relatively minor. If magma mixing was the source of compositional variation in the suites, the same evidence indicates that either the mixed magmas were geochemically similar, or the added volumes of compositionally distinct melts were minor. Thus the principal geochemical variations in samples from the tholeiitic suites are unlikely to be the results of fusion processes, contamination, or magma mixing. Many of the flows in these sequences, however, are porphyritic, and this suggests that fractional crystallization has contributed to compositional diversity. In addition, it is probable that the concentrations of some relatively mobile elements have been affected by post-crystallization alteration and metamorphism. This is suggested by the absence of systematic covariations between the concentrations of Na_2O , K_2O , Ba and Sr and those of less mobile elements in the cycle 2 tholeiitic basalts, and the variability of Ba and Sr in the Le Fer sequence.

Porphyritic flows are present in the Nimish and lower Menihek alkaline formations, and this suggests that fractional crystallization may also have influenced the compositions of these suites. The alkaline suites, however, are compositionally diverse, lack coherent inter-element covariations, and exhibit non-constant incompatible element ratios. As discussed previously, these features are inconsistent with fractional crystallization as the only source of variation in the suites. The differences in trace element patterns between the lower Menihek flows and clasts suggests that alteration may have influenced rock compositions, particularly with respect to the concentrations of some of the relatively mobile LIL elements. This influence, however, was probably limited to the removal or addition of Na_2O , K_2O , Ba, Rb, $\pm$ Sr in samples which underwent comparatively strong alteration. Consequently, the compositional diversity of samples from the alkaline suites must reflect processes other than, or in addition to, fractional

crystallization and alteration. Given the localized nature of alkaline magmatism, it is possible that melting of a heterogeneous source and/or variable degrees of contamination have also imparted compositional diversity to the suites

Log-Log Diagrams

The distinction between the geochemical effects of fusion and crystallization processes in a volcanic suite can be achieved by an examination of covariations in trace element concentrations. Treuil and Varet (1973) and Cocherie (1986) have shown that variable degrees of partial melting will lead to large differences in incompatible trace element contents, but relatively small variations in compatible trace element concentrations in the resultant volcanic products. In contrast, samples from a basaltic succession produced by crystal fractionation exhibit small variations in incompatible element contents, but large variations in compatible element concentrations. Thus trends in covariation diagrams using trace elements of contrasting bulk liquid-solid distribution coefficients (D_B) can be used to discriminate between partial melting and fractional crystallization as the source of variation. Log-log diagrams (log concentration element 1 vs. log concentration element 2) are often employed in this regard since the two processes produce trends of different shape. If D_B remains constant throughout differentiation, fractional crystallization gives rise to linear trends, while fusion mechanisms produce curvilinear trends (Cocherie, 1986).

The degree of curvature in the case of compositional variation due to partial melting depends on the nature of the fusion process (e.g., equilibrium melting, batch melting etc.), and the difference in D_B for the two elements plotted. For fractional crystallization, the relative D_B of the chosen elements control the slope of the trend. This arises from the Rayleigh distillation law (cf. Neumann *et al.*, 1954):

$$C_i^i = C_0^i F^{(D_B^i - 1)}$$

where:

C_i^i = concentration of element i in the derivative liquid

C_0^i = concentration of element i in the initial liquid

F = weight fraction of residual liquid

D_B^i = bulk distribution coefficient of element i

A plot of $\log C^i$ versus $\log C^j$, where C^i is the concentration of a second element, produces a straight line when D_R^i and D_R^j are constant during fractional crystallization (Allègre *et al.*, 1977). The slope of this line is given by:

$$m = \frac{(D_R^j - 1)}{(D_R^i - 1)}$$

Given the above relationship, and assuming constant D_R values, it is apparent that a suite of volcanic samples related by crystal fractionation will define trends with slopes of unity in log-log plot of two elements which are perfectly incompatible (i.e., $D_R^i = D_R^j = 0$). In the same case, a plot of a slightly compatible element ($1 > D_R^j > 0$) versus a perfectly incompatible element will have a slope between 0 and 1. The slope decreases with increasing compatibility leading to negative slopes for elements which are more compatible in the solid than in the liquid (i.e., $D_R > 1$). Slopes greater than unity indicate that crystal fractionation was not the predominant source of variation in concentration of one or both of the elements in question. For example, a log-log plot of a compatible element (e.g., Ni in olivine) versus an incompatible element (e.g., Zr) for a suite of samples related by crystal accumulation (e.g., dunitic cumulates) would exhibit a trend whose slope is greater than 1.

Sources of Geochemical Variation in the Le Fer Basalts

The Le Fer basalts are a submarine sequence of flows which were deformed and metamorphosed during Hudsonian orogenesis. It is thus possible that spillitization and metamorphism have influenced whole rock geochemistry. The effects of alteration and metamorphism on the primary mineralogy of Le Fer samples are variable. In most rocks, some degree of replacement of the primary mineralogy is apparent, and microphenocrysts are invariably pseudomorphed by secondary assemblages. Despite these effects, incompatible element ratios are constant and the samples generally define coherent inter-element covariation trends in scatterplots. These features suggest that post-crystallization processes have not significantly altered the whole rock major and trace element concentrations on a sample scale. Although this appears to be true for most elements, the absence of systematic variations in the concentrations of Na_2O , K_2O , Sr and

Ba suggests that these constituents have undergone some degree of mobilization as a result of alteration.

The Le Fer basalts now exposed are relatively homogenous, have low incompatible element contents, including the REE, and elevated Cr and Ni relative to the other suites of the LTBS. These attributes, coupled with the constancy of incompatible element ratios, argue that if assimilation of crustal material or magma mixing have influenced melt geochemistry, the effects have been relatively minor. Consequently, it is probable that the primary source of geochemical variability observed in the Le Fer basalts involves partial melting and/or fractional crystallization, with subsequent post-crystallization processes modifying the concentrations of some mobile constituents.

Previous analysis indicated that ratios involving Ti, P, Mn, Zr, Y and Nb in the Le Fer basalts and gabbros are constant within analytical uncertainty. The constant ratios, and the relatively low distribution coefficients of these elements for potential solid phases in the fusion or crystallization of mafic magmas, suggest that they were conserved during differentiation of Le Fer liquids. The REE also tend to behave incompatibly during generation and fractional crystallization of mafic melts, and hence these may also have been conserved during differentiation of the Le Fer suite. The constancy of La/Ce ratios in the Le Fer samples (Table 8.2) supports this supposition. If fractional crystallization is the principal source of variation among the Le Fer basalts, log-log plots of these elements should produce linear trends whose slopes are close to unity.

Such relationships are apparent in the log-log plots of Ti, Zr, La and Ce versus Y in Figure 8.24. The slopes are within error of unity in each case, and the good correlations between Y and the ordinate elements indicate that linear approximations of the trends are appropriate. The plots of V and P versus Y also define linear trends with slopes within error of unity, but the correlations are less strong for these elements. The linear trends coupled with slopes near 1 indicate that fractional crystallization is the most likely source of compositional variation in the Le Fer basalts, and that Ti, Zr, La and Ce were conserved during the process. V and P also appear to have been conserved, although the slightly weaker correlations with Y may indicate the influence of another differentiation process or the effects of alteration and metamorphism. Alternatively, the weak correlations may reflect changes in D_B^V and D_B^P during fractionation.

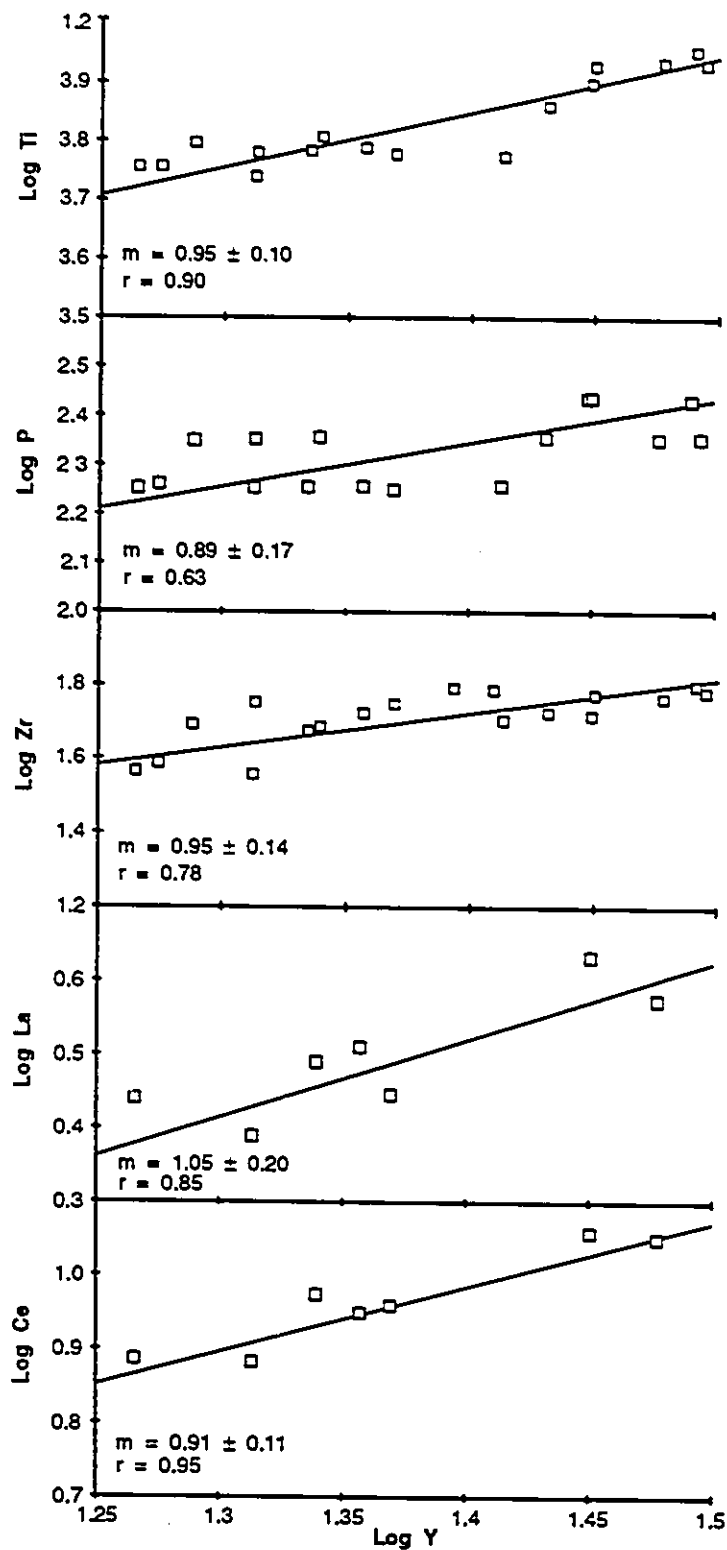


Figure 8.24. Log-log diagrams of selected elements versus Y for the Le Fer basalts. The slope (m), standard error on the slope, and correlation coefficient (r) are indicated for each regression.

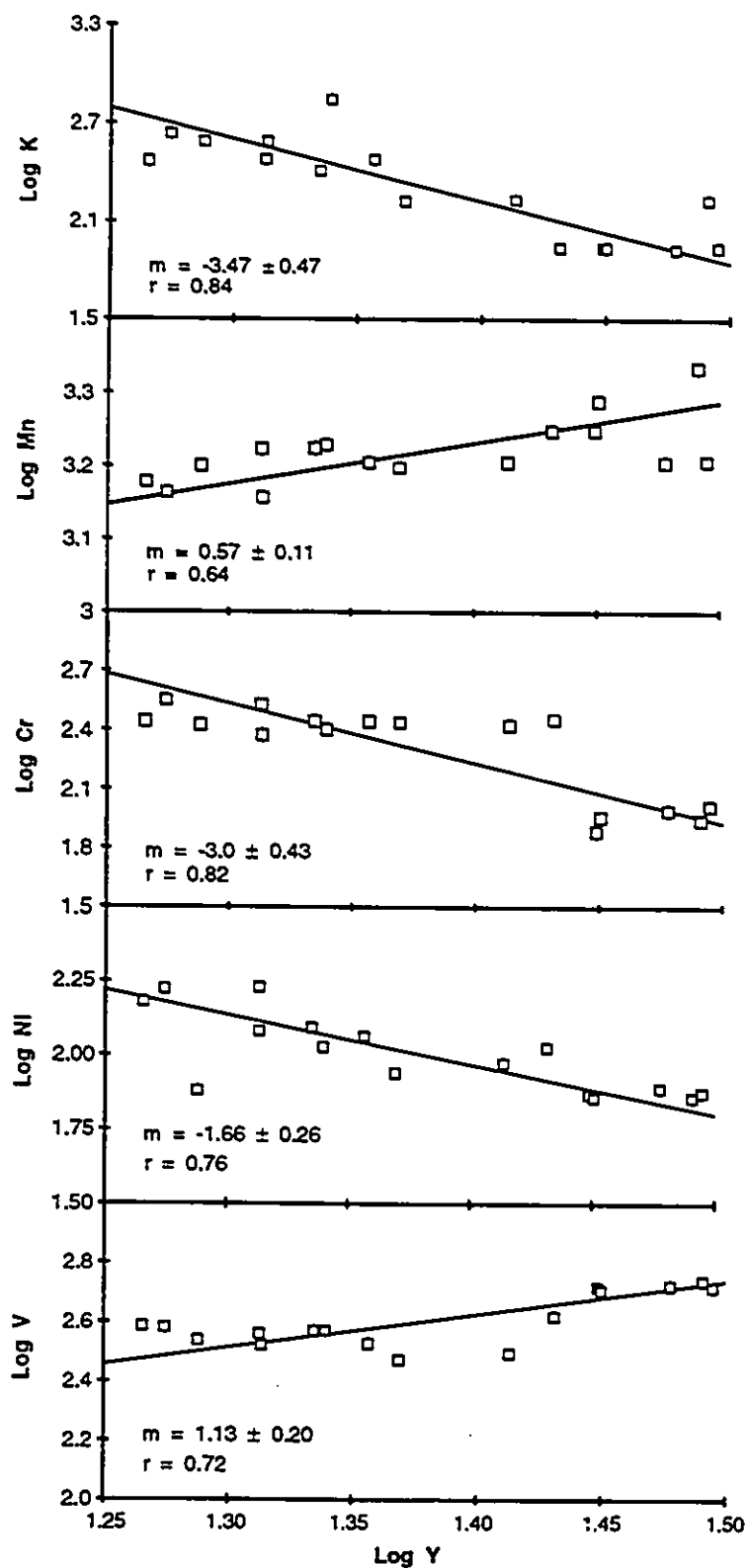


Figure 8.24 (continued). Log-log diagrams of selected elements versus Y for the Le Fer basalts. The slope (m), standard error on the slope, and correlation coefficient (r) are indicated for each regression.

Figure 8.24 also illustrates log-log diagrams in which potentially compatible elements have been used on the ordinate. K_2O , Cr and Ni exhibit well defined and negatively correlated linear trends with increasing Y, while Mn versus Y defines a positive slope between 0 and 1. The negative slopes in the plots of K_2O , Cr and Ni are consistent with fractional crystallization of phases in which these elements are relatively compatible (i.e., $D_B > 1$). The slope of 0.57 in the plot of Mn versus Y indicates that the former was not completely incompatible during crystal fractionation. Limited non-conservation of Mn was also suggested as the source of the pattern of variation found in incompatible element ratios in the compositionally similar cycle 2 subalkaline lavas (cf. Fig 8.13). In these rocks, ratios involving Mn and Ti exhibit progressive shifts with increasing degree of differentiation, and this was attributed to slight compatibility of these elements. There is no evidence of non-conservation of Ti in the Le Fer basalts, but the limited compatibility hypothesis for Mn is consistent with the intermediate slope in the log Mn versus log Y plot.

The log-log diagrams indicate that fractional crystallization was the principal source of magmatic variation in samples of the Le Fer basalts. Among the available trace elements, Ti, P, Zr, Y, V and the REE were essentially incompatible during the differentiation process, Mn was slightly compatible, and K, Cr and Ni were compatible. Because of the relationship between slope and D_B , a comparison of the slopes of Cr and Ni versus Y in the log-log plots indicates that fractional crystallization involved phases with $D_B^{Cr} > D_B^{Ni}$, and both greater than 1. The fact that the bulk distribution coefficient of Cr is almost double that of Ni (as indicated by the slopes of the plots) strongly suggests that olivine, for which $D_B^{Ni} > D_B^{Cr}$, was not the predominant ferromagnesian phase fractionated in the differentiation of the suite. The relationship between K_2O and Y indicates that a phase into which substitution of K was possible also formed part of the fractionate. Given the compositions of the basalts, this suggests that a plagioclase solid solution containing the orthoclase endmember was removed from the melts. Correlations in log-log diagrams of Ba and Sr versus an incompatible element (not illustrated) are extremely poor, and this indicates that fractional crystallization was not the principal source of variation for these elements. The lack of correlation is consistent with the postulated post-crystallization mobilization of Sr and Ba. This interpretation is supported in the case of Sr by the fact that

the slopes of trends in log-log diagrams of Sr versus an incompatible element exceed unity, which is impossible if crystal fractionation is the sole source of variation. The concentrations of Nb and Rb are above detection limits for too few samples to evaluate their behaviour during fractionation.

From the foregoing it is apparent that fractional crystallization controlled the primary compositional variations found in the Le Fer basalts, with subsequent mobilization of some elements during alteration and metamorphism. The effects of fractionation of different minerals can be examined using Pearce ratios as described in the analysis of the Menihek sills. The principal difference between the analysis of the intrusive and extrusive suites lies in the fact that the starting liquid compositions for the latter are unknown, and therefore the effects of fractionation versus accumulation cannot be resolved. The phenocryst-groundmass proportions in the basalts, however, indicate that the flows largely represent residual liquids derived from fractional crystallization and removal of the majority of cumulus crystals prior to eruption.

It has been shown that Zr and Y can be considered conserved during differentiation of the Le Fer basalts, and these elements can thus be used as denominators in the Pearce ratios. The Le Fer basalts are commonly porphyritic, containing olivine, pyroxene and, to a lesser extent, plagioclase phenocrysts and microphenocrysts. Consequently these are the most likely phases to have been fractionated from the Le Fer magmas. The fractionation of these minerals is examined in Figure 8.25 using appropriately constructed complex numerator Pearce ratio diagrams with Y as the denominator. Regression data for ratios constructed using both Y and Zr are given in Table 8.5.

Figure 8.25A illustrates the relationship between alkali feldspar and plagioclase in the suite. The average slope of the regressions in the alkali feldspar versus plagioclase diagram, calculated using Zr and Y in the denominator, is 0.026 (Table 8.5). This indicates that variations in the Al_2O_3 contents of the basalts were largely controlled by plagioclase, and only minor alkali feldspar was removed from successive melts. It was previously argued that the small amounts of cumulate alkali feldspar needed to explain the observed variations in K_2O content of the Menihek gabbros could reflect substitution of the orthoclase molecule in cumulus plagioclase. Since alkali feldspar would be an unlikely liquidus phase during fractional crystallization of the Le Fer magmas, particularly in light

Table 8.5. (A) Slope (m), standard error of the slope (SE_m), correlation coefficient (r) and t -statistic for the correlation coefficient (t) for the complex numerator Pearce ratio diagrams illustrating major element variations in the Le Fer basalts (Fig. 8.25). Critical value of t at the 95% confidence level is 2.179 ($n=14$), which is exceeded for each plot. (B) Average mineral ratios and calculated mineral proportions (mol %) in the fractionate removed from the Le Fer liquids.

(A)

Diagram	Denominator = Zr				Denominator = Y			
	m	SE_m	r	t	m	SE_m	r	t
Figure 8.22:								
(A) Kfs vs. Pl	0.02	0.004	0.72	3.88	0.03	0.003	0.90	7.84
(B) Cpx vs. Pl	0.71	0.15	0.55	2.45	0.83	0.17	0.59	2.73
(C) Ol vs. Cpx vs. Opx	0.79	0.09	0.88	6.99	0.73	0.08	0.90	7.57
(D) Ol vs. Opx	1.00	0.13	0.86	6.29	0.93	0.11	0.87	6.76
(E) Mt vs. Cpx+Opx	0.02	0.003	0.77	4.46	0.02	0.002	0.83	5.57
(F) Mg #	2.61	0.33	0.86	6.30	2.97	0.31	0.91	8.20

(B)

Mineral Ratio		SE_m	Mineral Proportion (mol. %)		2σ
Pl:Pl	1	0	Pl	37.7	5.9
Kfs:PL	0.03	0.004	Kfs	1.0	0.5
Cpx:Pl	0.77	0.16	Opx	31.3	5.6
Opx:Cpx	1.05	0.22	Cpx	29.0	5.1
Opx:Pl	0.81	0.22	Mt	1.0	0.8
Mt:Opx+Cpx	0.02	0.010	Mineral Composition		
Mt:Pl	0.03	0.01	Mg#	0.74	0.08
Mg:Fe ²⁺	2.79	0.32			

of the extremely low K_2O concentrations, such substitution is also the most probable source of the variation in K_2O . This would necessitate fractionation of plagioclase containing 2.5 ± 1.4 mol. % Or.

The source of variation in the CaO content of the basalts is examined in Figure 8.25B, which illustrates the relationship between the only probable Ca-bearing cumulus phases, clinopyroxene and plagioclase. The average slope of the regressions (0.77) is intermediate to that expected from fractionation of either plagioclase or clinopyroxene alone, and thus both phases must have influenced melt compositions. Figure 8.25C examines the possibility that clinopyroxene fractionation controlled variations in the concentrations of MgO and FeO* in the basalts. Given analytical uncertainty, the average

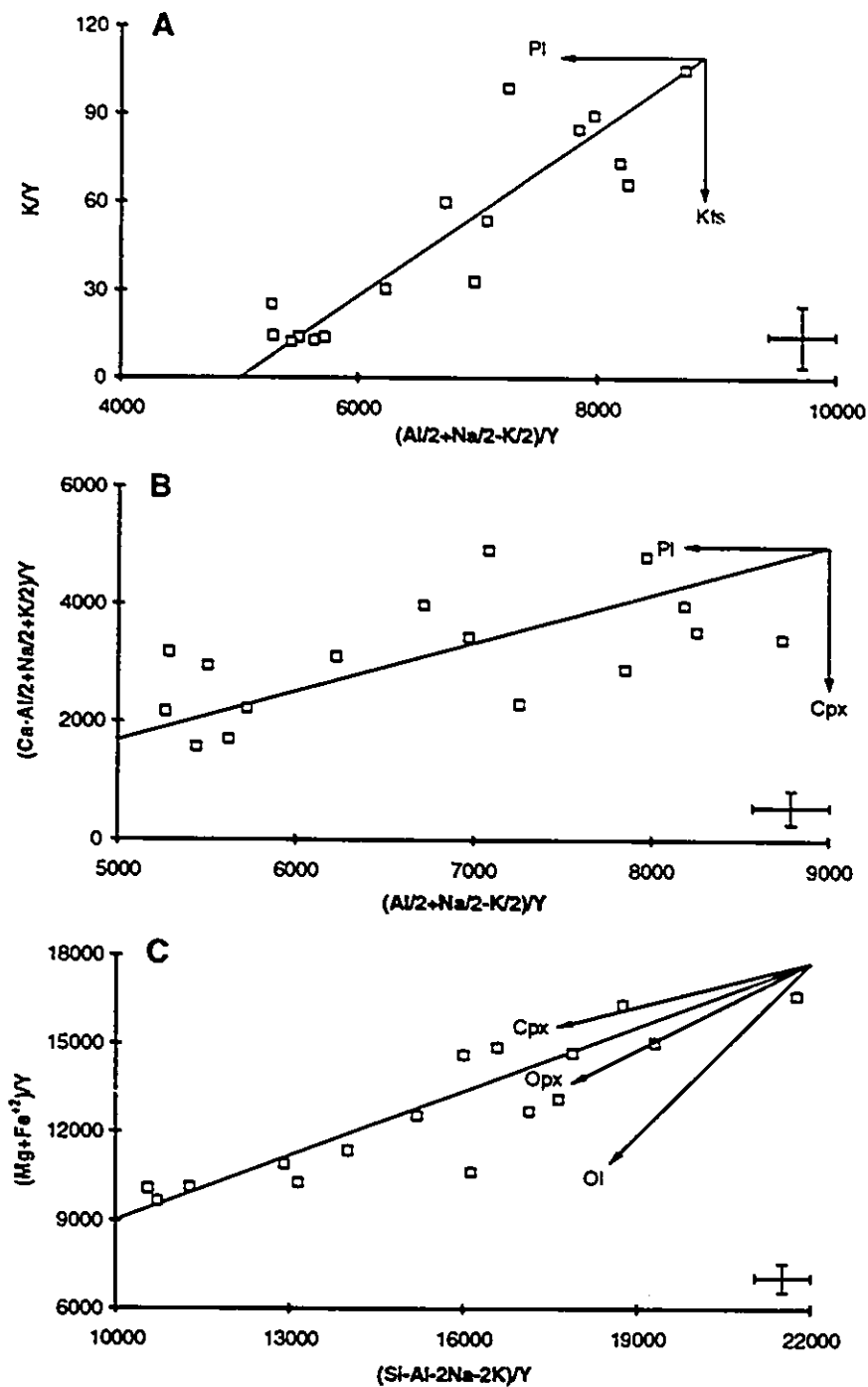


Figure 8.25. Complex numerator Pearce ratio diagrams illustrating the relationships between potential liquidus phases in the fractional crystallization of the Le Fer magmas. (A) alkali feldspar (Kfs) versus plagioclase (Pl); (B) clinopyroxene (Cpx) versus plagioclase; (C) olivine (Ol) versus orthopyroxene (Opx) versus clinopyroxene. Average 2σ analytical uncertainty is illustrated by error bars.

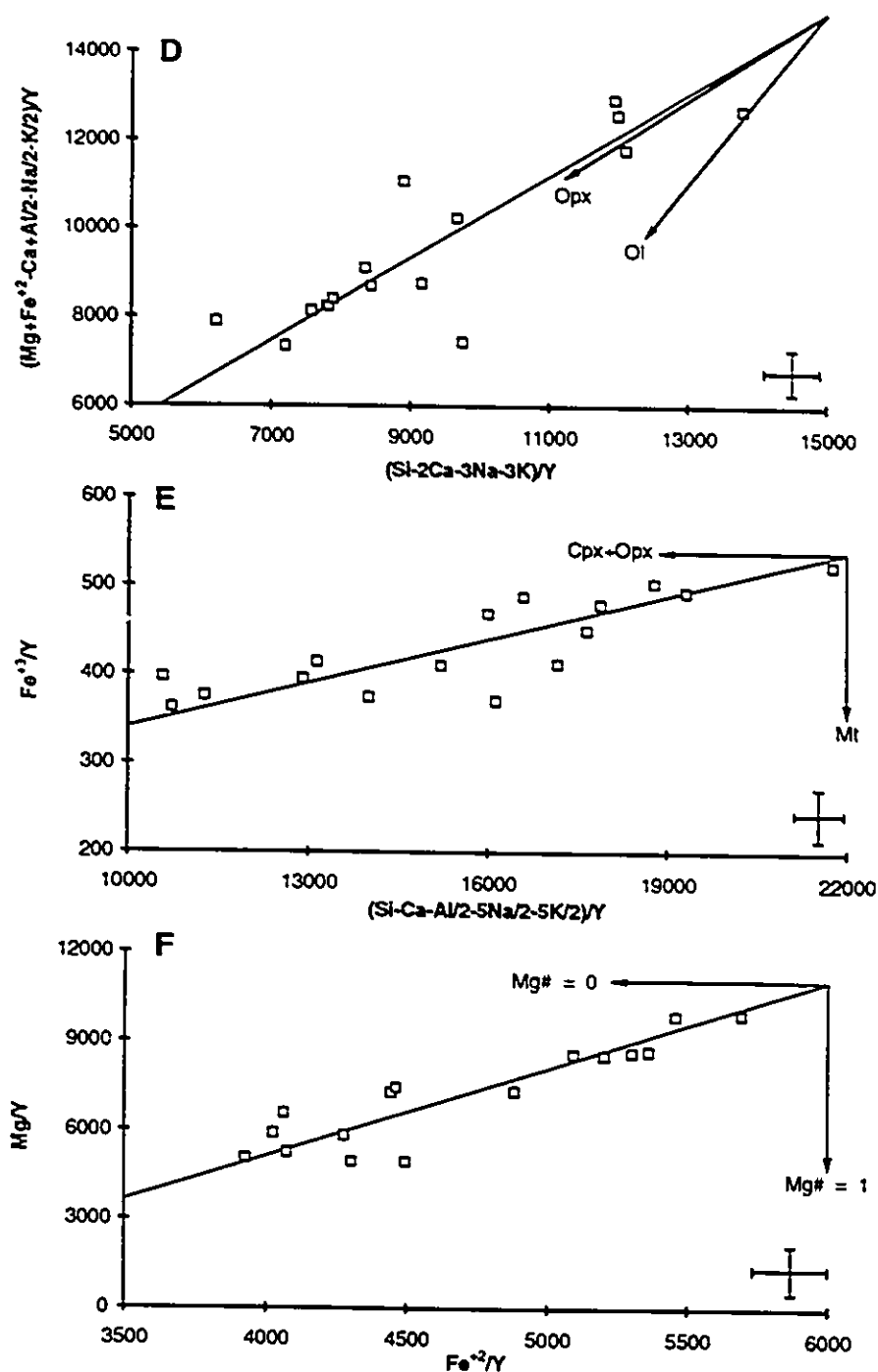


Figure 8.25 (continued). Complex numerator Pearce ratio diagrams illustrating the relationships between potential liquidus phases in the fractional crystallization of the Le Fer magmas. (D) olivine versus orthopyroxene; (E) magnetite versus clinopyroxene+orthopyroxene; (F) Mg# in clinopyroxene and orthopyroxene. Average 2σ analytical uncertainty is illustrated by error bars.

slope of the regressions does not correspond to the trends expected from removal of any one of the three probable ferromagnesian phases by itself. Consequently clinopyroxene and at least one other ferromagnesian mineral were liquidus phases at some point during differentiation of the suite. The identity of the second mineral can be resolved by removing the effect of clinopyroxene fractionation on MgO and FeO* concentrations, as illustrated in Figure 8.25D. In this diagram, the average slope of the regressions is 0.96 ± 0.12 , which is within error of the theoretical slope of 1.0 expected from fractionation of orthopyroxene. Thus crystallization of clinopyroxene and orthopyroxene satisfactorily explains the observed variations in the MgO and FeO* contents of the suite.

The only other major element which shows variations greater than expected from analytical uncertainty is Fe_2O_3 . As with K_2O , the extent of differences in the Fe_2O_3 content of the Le Fer basalts is relatively minor; the average regression in plots of magnetite versus total pyroxene is 0.02 (Fig. 8.25E). If this variation reflects precipitation of magnetite, the average fractionate would contain 1.0 mol. % of this phase. Since incompatible element ratios involving Ti are constant given analytical uncertainty, precipitation of an iron oxide, which would most likely take the form of a magnetite-ilmenite solid solution, is improbable. If the variation in Fe_2O_3 reflects substitution in the pyroxenes, the average Fe_2O_3 content of these phases was approximately 1.7 wt. %, which is slightly less than that obtained in the ratio analysis of the Menihek gabbros. Although no microprobe analyses were obtained from the basaltic rocks, this value does fall within the range of Fe_2O_3 concentrations found in the pyroxenes of the cycle 2 gabbroic rocks. It thus seems probable that variations in the Fe_2O_3 content reflect substitution of ferric iron in the pyroxenes rather than precipitation of a discrete oxide mineral.

Following the procedure used in Chapter 5, the average An content of plagioclase and the average Mg# of pyroxenes in the fractionate can be estimated. Such an analysis yields invalid results for the An content of plagioclase in that a plot of albite versus anorthite produces trends with negative slopes. Since such results are impossible if fractional crystallization was the only source of variation in CaO and Na_2O , it is apparent that another process has affected the concentration of one or both of these components. Given the mobility of Na and the fact that the Le Fer basalts represent submarine flows, the most likely source of the anomalous An-Ab ratios in plagioclase is post-crystallization

Na-metasomatism due to introduction of sea water. That such a process may have occurred is also suggested by the albitization of plagioclase in the chilled margins of the Menihok sills. The localization of albitization to the contact zones (interior samples contain unaltered plagioclase) of the sills indicates that the phenomenon was not metamorphic in origin. The contact zones, however, would be the sites of greatest exposure to sea water trapped in the pores of the intruded sediments.

The Mg# of the bulk fractionate can also be calculated using appropriately constructed ratio diagrams (Fig. 8.25F). This analysis gives an average value of 0.74 for the Mg# of the fractionate. Assuming that variations in Fe_2O_3 contents reflect substitution of ferric iron into orthopyroxene and clinopyroxene, this value (0.74) represents the average Mg# of the pyroxenes over the course of fractional crystallization. Due to the lack of microprobe data this estimate cannot be tested against measured mineral compositions, but it is in the range of values that were obtained from pyroxenes in the compositionally similar Menihok gabbros.

Using the ratios derived from Figure 8.25, the relative proportions of the principal minerals in the average fractionate of the Le Fer basalts can be estimated. Assuming variations in K_2O and Fe_2O_3 reflect substitution in plagioclase and pyroxene, the range in compositions found in the Le Fer basalts reflects fractionation of a solid having the average mineral proportions of 39 ± 6 mol. % plagioclase, 32 ± 6 mol. % orthopyroxene and 29 ± 5 mol. % clinopyroxene (Table 8.5).

The ratio analysis indicates that the compositional variation within the Le Fer suite is consistent with fractional crystallization of plagioclase, orthopyroxene and clinopyroxene prior to eruption of the residual melts. It is apparent, however, that the Le Fer basalts have been affected by post-crystallization processes which imparted further compositional variations, notably in the concentrations of Na_2O ($\pm$ CaO), Ba and Sr. In addition, the presence of olivine phenocrysts in many samples suggests that pre-eruption crystallization of this phase occurred, whereas fractionation of olivine is not consistent with the geochemical evidence. This disparity may also reflect post-crystallization processes since the distinction between fractionation of olivine and orthopyroxene was based on the changes in $\text{Mg}+\text{Fe}^{2+}$ concentrations relative to Si. Consequently,

post-crystallization modifications of the SiO_2 content of the samples offers a potential explanation for the differences in geochemical and petrographic observations.

Assuming olivine rather than hypersthene was fractionated, appropriate post-crystallization changes in SiO_2 relative to $\text{Mg}+\text{Fe}^{2+}$ could decrease the slope of the trend in Figure 8.25D from approximately 2 to approximately 1. This would lead to the conclusion that the second ferromagnesian cumulus phase was a Ca-poor pyroxene rather than olivine. It seems unlikely, however, that post-crystallization modification of SiO_2 content would lead to rock compositions that correspond exactly to those expected from orthopyroxene+clinopyroxene fractionation. A more conceivable result of SiO_2 mobility would be the development of inconclusive, inconsistent or poorly correlated trends in Pearce ratio diagrams. Such is apparently the case for the relationship between the anorthite and albite components in fractionated plagioclase. The trends defined in ratio plots of these components are inconsistent with fractional crystallization, and therefore must be interpreted to reflect a different process, such as Na-metasomatism. The greater slope of Cr relative to Ni in the log-log plots of Figure 8.23 also suggests that olivine was not part of the fractionate, since the opposite relationship would be expected if this was the case.

If post-crystallization variations in SiO_2 effects are rejected, the disparity between petrographic and geochemical evidence could be reconciled by either very minor olivine fractionation in liquids represented by the Le Fer sample suite, or by the fractionation of olivine from melts which are not represented by the suite. In the first case, the error on the calculated slope in the regressions of Opx versus Ol (Fig. 8.25D) could mask minor olivine fractionation. Assuming the maximum positive error, the resultant slope of 1.08 translates into a maximum olivine content in the fractionate of approximately 4 mol. %. Unless olivine preferentially accumulated in melts which subsequently reached the surface, this proportion is insufficient to explain the abundance of olivine phenocrysts relative to those of plagioclase and pyroxene. This is particularly true in the Tait Lake sequence where olivine is the predominant phenocryst phase.

Fractionation of olivine from more primitive magmas than those represented by the sample suite requires that either lavas with more primitive compositions were not erupted, or that the samples collected in this study are not representative of the total compositional

range of the Le Fer basalts. In either case, the fractionation of olivine from primitive magmas may have produced residual melts whose compositions were appropriate to the crystallization of plagioclase, orthopyroxene and clinopyroxene. Subsequent fractionation of the latter phases would then impart the compositional variation observed in the sample suite. Thus early olivine fractionation may have led to later precipitation of plagioclase and pyroxene, and only melts residual to the latter episode of crystallization reached the surface. The compositional changes in residual melts could have arisen from olivine fractionation alone, or from early olivine fractionation accompanied by assimilation of crustal material. Both processes would lead to increasing SiO_2 concentrations in residual melts, and silica saturation would result in the precipitation of orthopyroxene at the expense of olivine. This type of assimilation-fractional crystallization was postulated as the origin of compositional variations in the Chukotat lavas of the Cape Smith fold belt (Francis *et al.*, 1983). In either scenario, some early olivine grains could escape accumulation or resorption and be carried to the surface with the more evolved liquids. Given the low MgO contents of the Le Fer samples relative to postulated mantle partial melts (cf. Green *et al.*, 1974; Ringwood, 1975; Frey *et al.*, 1978; Jaques and Green, 1980; Sen, 1982) it is unlikely that any of the Le Fer samples obtained in this study represent undifferentiated primary magmas (see also Chapter 9). Thus olivine fractionation during evolution of primitive Le Fer magmas is the most probable source of olivine phenocrysts in the basalts.

Sources of Geochemical Variation in the Nimish, Lower Menihok and High-Zr Murdoch Basalts

The Nimish and lower Menihok alkaline volcanics exhibit similarities that suggest stratigraphic equivalence. It is apparent, however, that the suites reflect localized generation of alkaline magmas which were therefore not comagmatic. It was concluded from the non-constancy of incompatible element ratios that the compositional diversity within, and between, the two suites is unlikely to be a function of crystal fractionation alone, although this process undoubtedly contributed to compositional variability. The generally poor correlations and inconsistent trends found in log-log plots of data from the suites also suggest that crystal fractionation was not the predominant source of variation in

the suites (Fig. 8.26). Among Nimish samples, strong correlations with Zr in the log-log diagrams are found only for P, Y, La and Ce. Of these, only Ce defines a trend with a slope within error of 1, while the slopes of the trends for P and La exceed 1. Nb and K define trends with slopes within error of 1 when plotted against Zr, but the correlations among these elements are weaker. Trends among other elements relative to Zr tend to be poorly defined, with slopes ranging from 0.79 to -2.34. If Zr was conserved during evolution of the Nimish suite, log-log plots in which the slopes of the trends exceed 1 (P and La) cannot be reconciled with crystal fractionation as the only source of variation in these elements. The poor correlations between Zr and other trace elements, whether incompatible or compatible, also argues against fractionation as the only source of compositional variation. While it is possible that Zr was not conserved during melt evolution, similar highly variable and poorly correlated plots are obtained using Y and Nb as the incompatible abscissa element. This suggests that the concentrations of most, if not all, trace elements have been affected by processes other than, or in addition to, fractional crystallization. The fact that Ce and Zr define a trend whose slope is within error of one may then reflect a similar degree of non-conservation of these elements.

The log-log plots of Figure 8.26 reveal similar relationships among elements in samples from the lower Menihek suite. In these samples, P, Nb, La and Ce exhibit strong correlations with Zr, and Nb, La and Ce define trends within error of 1 in the diagrams. Other trace elements, however, tend to be very poorly correlated with Zr. While these results are similar to those obtained for the Nimish suite, interpretation of the plots for the lower Menihek suite is complicated by the fact that the samples represent two lithological types (clasts and *in situ* flows) from two geographic locations (Howse and Outlook lakes). The compositional differences between flows and clasts of the lower Menihek Formation described earlier are also evident in Figure 8.26. The most apparent differences are the higher concentrations of Zr, Y, and Nb, and lower K, Sr and Ba, in the clastic samples. These may reflect a primary compositional discrepancy, but the strong alteration of the clastic samples suggests that post-crystallization effects may have imparted some of the geochemical differences. Assuming the clastic samples were initially similar to the lower Menihek flows in composition, mass loss during alteration and metamorphism could lead to a reduction in mobile element concentrations and concomitant elevation of immobile

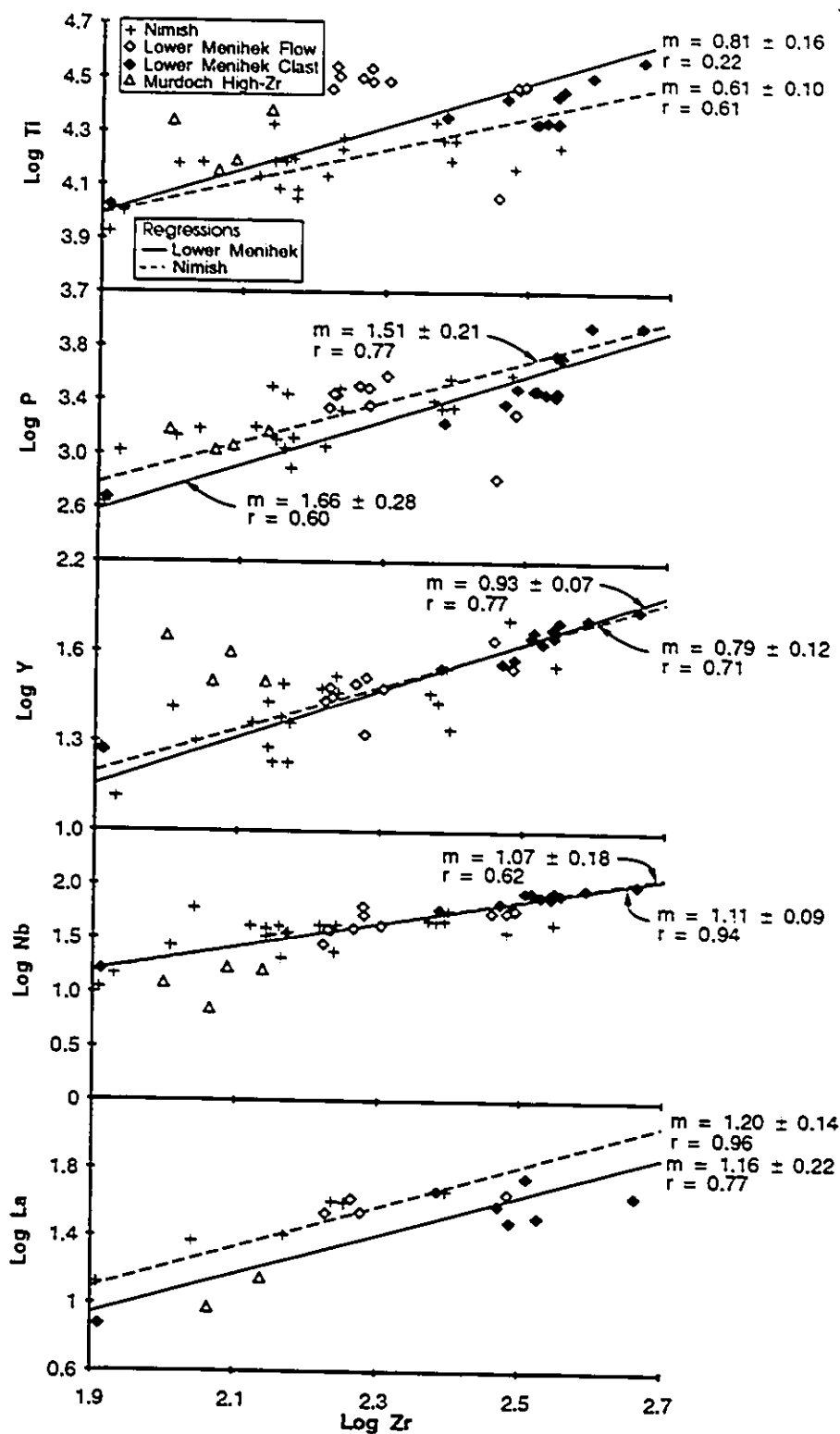


Figure 8.26. Log-log diagrams of selected trace elements versus Zr for the Nimish and lower Menihok volcanic rocks. The slope (m), standard error on the slope, and correlation coefficient (r) are indicated for each regression. The high-Zr Murdoch samples are shown for comparison.

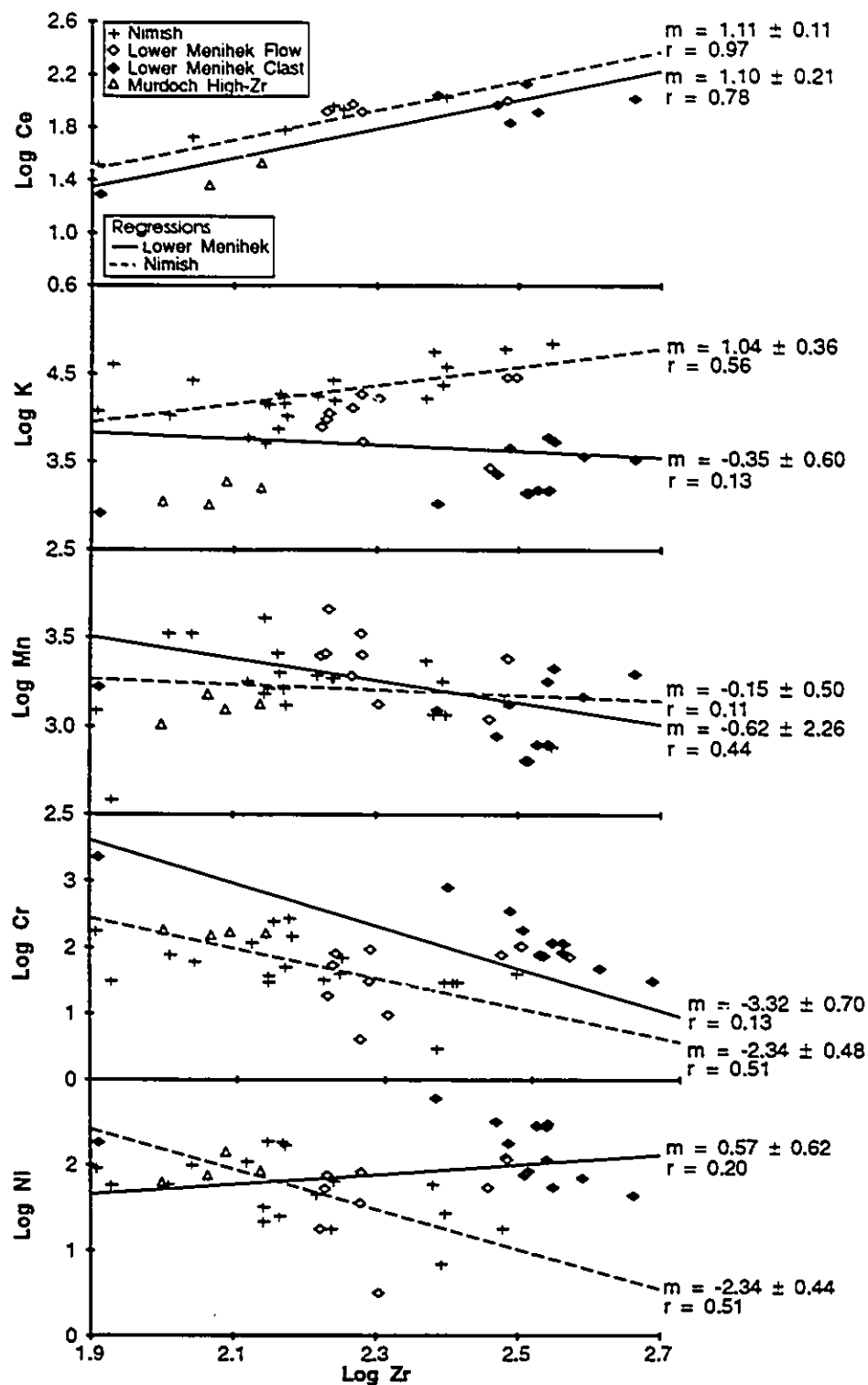


Figure 8.26 (continued). Log-log diagrams of selected trace elements versus Zr for the Nimish and lower Menihok volcanic rocks. The slope (m), standard error on the slope, and correlation coefficient (r) are indicated for each regression. The high-Zr Murdoch samples are shown for comparison.

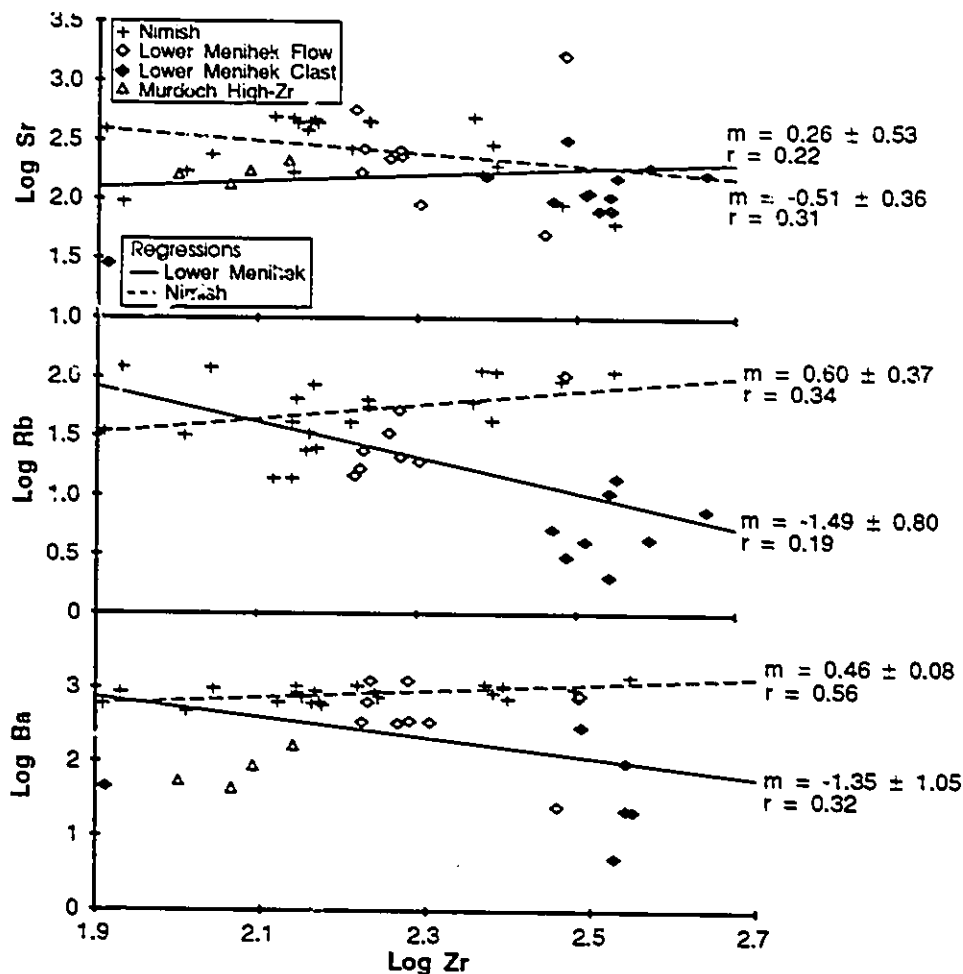


Figure 8.26 (continued). Log-log diagrams of selected trace elements versus Zr for the Nimish and lower Menihok volcanic rocks. The slope (m), standard error on the slope, and correlation coefficient (r) are indicated for each regression. The high-Zr Murdoch samples are shown for comparison.

element concentrations. Since Zr, Y and Nb are likely to be less mobile than K, Sr and Ba under such conditions, the compositional differences between the two groups are consistent with the greater degradation of the clastic samples. As noted earlier, however, the Ti depletion apparent in the clastic samples cannot be explained by such a mechanism.

While some of the compositional differences between the lower Menihok clasts and flows may reflect alteration and elemental mobility, it seems unlikely that this process would result in relative changes in the concentrations of immobile trace elements. If not, then the clasts and flows would define the same trends in log-log plots of immobile elements, but they would plot at different absolute concentrations. Although individual

regressions are not illustrated in Figure 8.26, the two groups of samples define different trends in all plots except Nb versus Zr. In all cases, including Nb, the trends are more poorly correlated than when the samples are grouped together. Thus while alteration may be the source of some of the compositional discrepancies evident between *in situ* and clastic samples of the lower Menihek suite, it does not explain the variability of trace element trends, nor the poor correlations associated with these trends.

The high-Zr Murdoch basalts, which are plotted in the diagrams of Figure 8.26 for comparison, lie on the trends defined by the other suites for some elements, and depart from these trends for others. Although the number of samples is insufficient to define trace element covariations within this group, it seems probable that the same compositional effects which have influenced the Nimish and lower Menihek rocks have affected the slightly alkaline Murdoch basalts. This is also apparent in Figure 8.27 where the low-Zr Murdoch basalts exhibit covariations comparable to those of the cycle 2 tholeiites, while the high-Zr samples depart from these trends.

In conjunction with the variability of incompatible element ratios, the lack of coherent trends in the log-log plots indicates that fractional crystallization was not the predominant source of trace element variations in the Nimish, lower Menihek and Murdoch alkaline suites. The absence of coherent curvilinear trends in the log-log diagrams would also seem to invalidate a sequential fusion mechanism as the predominant source of variation in the suites. Consequently, it is probable that processes other than variable partial melting of the source, fractional crystallization and alteration contributed to the geochemical variability of the alkaline suites.

There are two likely mechanisms by which additional geochemical variability may have been imparted to the alkaline suites; differing source compositions and crustal contamination. The fact that alkaline magmatism was followed by tholeiitic magmatism indicates that two compositionally distinct mantle reservoirs were present. Derivation of magma batches through differential sampling of the two reservoirs could thus have played a role in the compositional diversity of the alkaline suites. This process may also explain the range in geochemical character of the Nimish volcanics (subalkaline to alkaline, Fig. 8.17) and the transitional compositions of the high-Zr Murdoch basalts. Compositional differences might also reflect variable assimilation of crustal material. Thus lavas erupted

from different volcanic centres, or lavas erupted after variable residence times in a single crustal magma chamber, could display broad geochemical similarities that reflect a common origin (i.e., similar degrees of partial melting of a similar source), but variable trace element contents due to differences in the degree of contamination by crustal material.

Sources of of Geochemical Variation in the Upper Menihek, Willbob and Low-Zr Murdoch Basalts

The upper Menihek and Willbob basalts exhibit coherent inter-element relationships and constant incompatible element ratios which suggest magmatic differentiation as the principal source of geochemical variation. This possibility is examined in the log-log diagrams of Figure 8.27 in which selected trace elements have been plotted against Y. Previous ratio analysis indicated that, given analytical uncertainty, Ti, P, Mn, Zr, Y, Nb and the REE can be considered conserved during the differentiation of both the gabbroic and volcanic members of the cycle 2 tholeiitic sequence. If crystal fractionation is the principal source of variation in the volcanic rocks, these elements should exhibit linear trends with slopes of 1 in log-log diagrams. This is the case for the well correlated covariations in plots of Ti, Zr, La and Ce versus Y. Nb also defines a trend with slope within error of 1, but the correlation with Y is poor. The fact that Nb concentrations in the majority of samples are close to detection limits may contribute to the lack of correlation.

In a previous section it was noted that systematic variations in incompatible ratios involving TiO_2 and MnO probably reflect limited non-conservation of these elements. While non-conservation of Ti is not apparent in Figure 8.27, the slope in the plot of Mn versus Y is slightly less than 1, and this is consistent with slight compatibility of Mn. This result is similar to that obtained for Mn in the Le Fer basalts. As with the other incompatible elements, P displays a well defined linear relationship with Y in the log-log plot of these elements. The slope in this plot, however, exceeds the expected value of 1. The combination of a strong linear covariation and a trend whose slope exceeds 1 is interesting since these characteristics would seem to invalidate crystal fractionation, variable partial fusion and alteration as sources of variation in P. The strong correlation

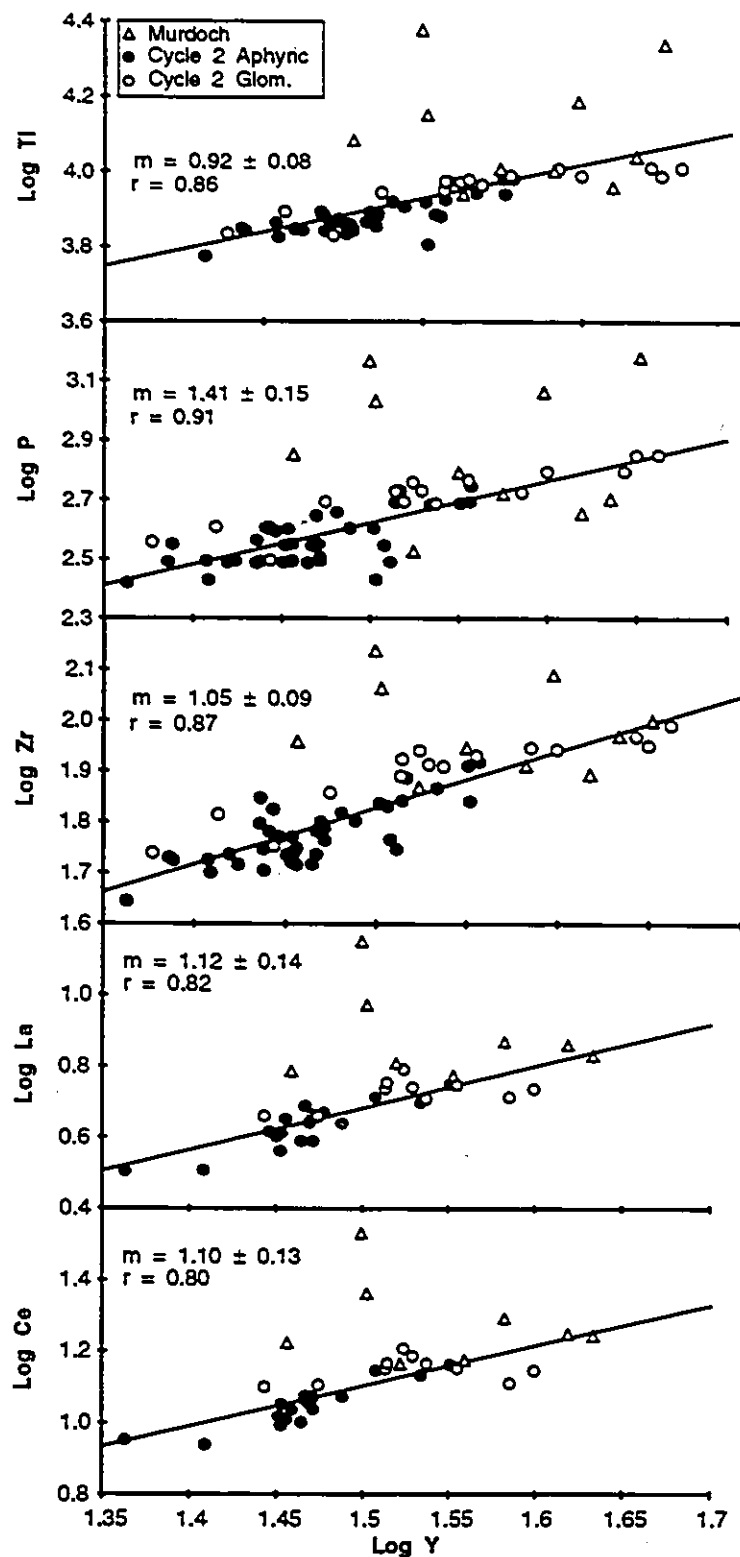


Figure 8.27. Log-log diagrams of selected elements versus Y for the upper Menihek and Willbob basalts. The slope (m), standard error on the slope, and correlation coefficient (r) are indicated. The Murdoch basalts are shown for comparison.

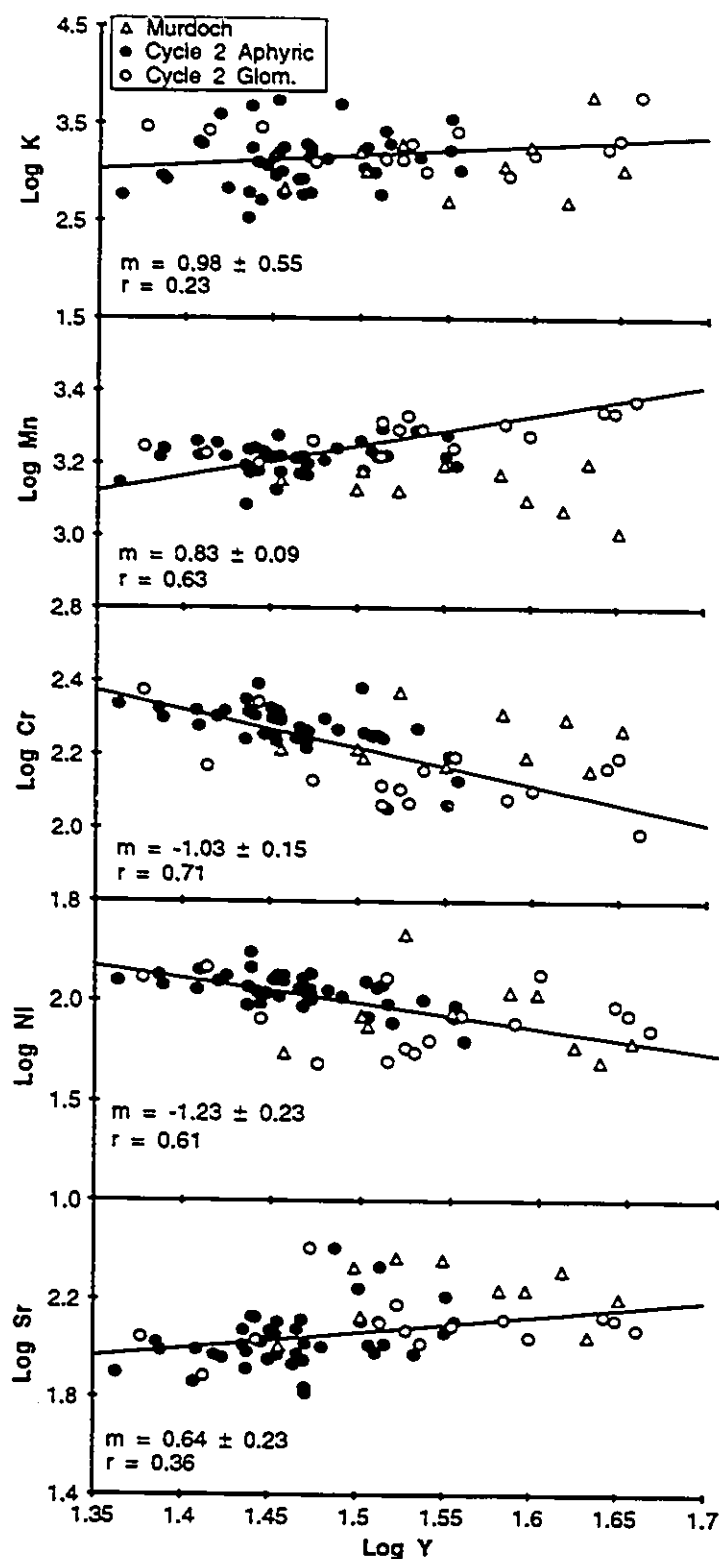


Figure 8.27 (continued). Log-log diagrams of selected elements versus Y for the upper Menihok and Willbob basalts. The slope (m), standard error on the slope, and correlation coefficient (r) are indicated. The Murdoch basalts are shown for comparison.

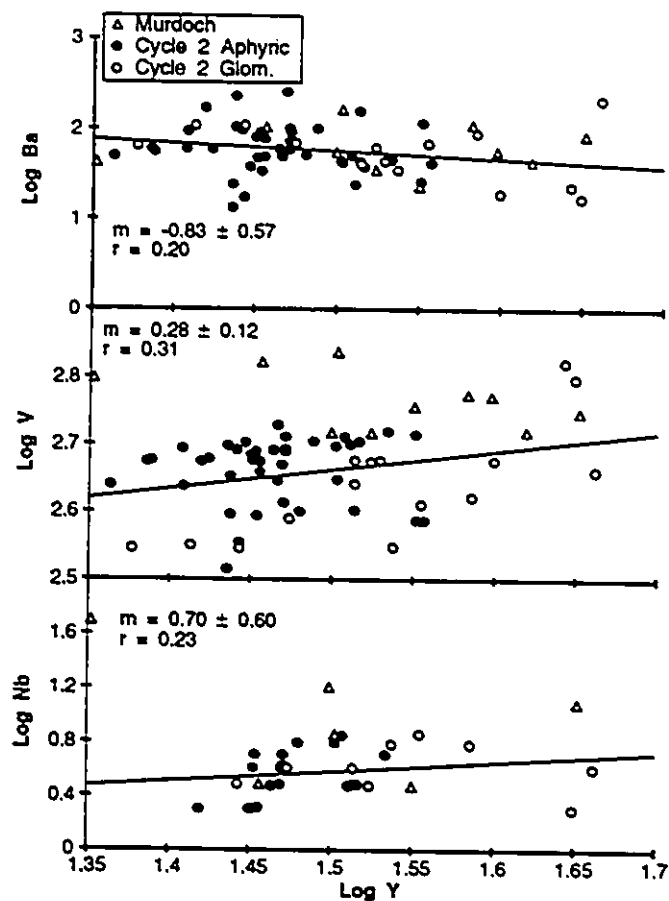


Figure 8.27 (continued). Log-log diagrams of selected elements versus Y for the upper Menihok and Willbob basalts. The slope (m), standard error on the slope, and correlation coefficient (r) are indicated. The Murdoch basalts are shown for comparison.

would also seem to be inconsistent with a change in D_B^P over the course of differentiation, since this would lead to a change in slope. That variation in P is the source of the anomalous slope is indicated by the fact that similar results are obtained using Zr and Ti as the incompatible abscissa element. Assuming the compositional variations in the cycle 2 samples generally reflect removal of a solid fractionate, the slope in the P versus Y plot indicates that P increased in concentration in residual liquids more quickly than other incompatible elements. Since P is unlikely to be more incompatible than Ti, Y, Zr and the REE in the fractionate, and if systematic analytical error is rejected, the discrepancy must be due to minor addition of P at some stage in the evolution of the samples.

Despite the anomalous covariation of P with Y in the log-log plot, the strong linear correlations among the incompatible elements indicates that crystal fractionation was the

predominant differentiation mechanism in the suite. With the exception of P, these elements also define trends in the diagrams with slopes of approximately 1, which is again consistent with differentiation through crystal fractionation. Among other available trace elements, only Cr and Ni show reasonably good correlations with Y. Both elements exhibit trends with negative slopes, as would be expected for trace elements which behave compatibly during crystal fractionation. The slopes for the two elements are identical within error, and this indicates that the two bulk distribution coefficients were similar. The equivalence of D_B for Cr and Ni contrasts with the case of the Le Fer basalts where Cr was significantly more compatible in the fractionate than Ni. Since the distribution coefficient for Cr between silicate melts and both clinopyroxene and orthopyroxene is greater than that of Ni (Henderson, 1982), both suites should show trends in which the slope of Cr was greater than Ni when plotted against an incompatible element. The explanation for this discrepancy is unknown, but it may reflect minor fractionation of chromite in the Le Fer magmas, which would substantially increase D_B^{Cr} .

The log-log plot of Ba versus Y also defines a trend with negative slope, while plots of K, V and Sr versus Y show positive slopes between 0 and 1. In each of these diagrams, however, the covariation trends are poorly correlated. Since scattered covariations also result if Zr is used in the place of Y on the abscissa, it is likely that the poor correlations result from spurious variations in Sr, Ba and V rather than Zr and Y. The absence of systematic variations in K, Ba and Sr among the cycle 2 tholeiitic samples was earlier attributed to mobility of these elements during alteration. If the total variation in K, Ba and Sr reflects the combined influence of crystal fractionation and post-crystallization processes, poor correlations with an incompatible/immobile element in the log-log diagrams would be likely.

The limited number of samples, and the presence of two geochemical types precludes an analysis of variation in the Murdoch volcanics. The Murdoch samples, however, are plotted in the diagrams of Figure 8.27 for comparison with the upper Menihok/Willbob suite. While no conclusions can be drawn from the data, it is apparent that the low-Zr group typically lie along the trends defined by the other cycle 2 suites. In contrast, the high-Zr samples depart from these trends. This generalization is most valid for the incompatible element diagrams of Figure 8.27, but is also apparent in the plots of

Cr and V versus Y. This suggests that crystal fractionation was also the predominant source of compositional variation in the subalkaline Murdoch basalts but not in the high-Zr samples.

The well defined trends among incompatible and immobile trace elements in the log-log diagrams of the cycle 2 tholeiites indicate that the concentrations of these elements are related by fractional crystallization. It is therefore probable that fractionation was also the source of variation in the major element geochemistry of the samples. This interpretation is consistent with the porphyritic and glomeroporphyritic nature of some flows. The possibility that major element variations derive from fractional crystallization can be tested using complex numerator Pearce ratio diagrams. This procedure can also be used to elucidate the composition of the fractionate. With the exception of those from glomeroporphyritic flows, geochemical data were obtained from samples of the cycle 2 volcanic rocks which were either aphyric or sparsely porphyritic. The glomeroporphyritic samples were sectioned to provide glomerocryst-free groundmass material. Thus the sample suite can be considered representative of the compositions of erupted lavas. This allows treatment of the samples as a series of residual liquids related by removal of crystalline phases. The results of the ratio analysis are summarized in Table 8.6 and illustrated in Figure 8.28

The relationship between alkali feldspar (Kfs) and plagioclase (Pl) illustrated in Figure 8.28A indicates that the former was not a major constituent of the fractionate in the cycle 2 tholeiitic melts. The average slope of the regression lines in plots of Kfs versus Pl using Zr and Y (Fig. 8.28A) as the conserved elements is 0.02 ± 0.01 (Table 8.6). The correlation coefficients in these diagrams, however, are not significant. This indicates that, while minor fractionation of Kfs is possible, another process has influenced K_2O variation. Assuming the average slope in the diagrams is representative of the true degree of Kfs fractionation, this phase would contribute 0.6 ± 1.7 mol. % to the total fractionate, where the large error derives from the poor correlation in the plots. As in previous analyses, such a quantity could easily be incorporated as orthoclase in plagioclase, in this case having a composition of roughly $(An+Ab)_{99}Or_1$.

The relationship between clinopyroxene (Cpx) and plagioclase is illustrated in Figure 8.28B, and it is apparent that both phases were fractionated from the liquids. The

Table 8.6. (A) Slope (m), standard error of the slope (SE_m), correlation coefficient (r) and t-statistic for the correlation coefficient (t) for the complex numerator Pearce ratio diagrams illustrating major element variations in the upper Menihek and Willbob basalts (Fig. 8.28). Critical value of t at the 95% confidence level is 2.179 ($n=14$) (B) Average mineral ratios and calculated mineral proportions (mol %) in the fractionate removed from the cycle 2 tholeiitic magmas.

(A)				
Diagram		Denominator = Zr		
Figure 8.22:		m	SE _m	t
(A) Kfs vs. Pl		0.02	0.014	0.15
(B) Cpx vs. Pl		0.76	0.06	0.80
(C) Cpx vs. Opx vs. Ol		0.52	0.02	0.94
(D) Fe ⁺³ vs. Cpx		0.03	0.003	0.82
(E) Ab vs. An		0.83	0.106	0.33
(F) Mg vs. Fe ^{+2*}		2.08	0.12	0.90

Denominator = Y				
m	SE _m	r	t	
0.01	0.016	0.04	0.32	
0.81	0.07	0.74	8.04	
0.50	0.03	0.91	15.72	
0.03	0.003	0.76	8.45	
0.96	0.127	0.19	1.38	
2.26	0.17	0.80	9.68	

(B)		
Mineral Ratio		SE _m
Pl:PL	1	0
Kfs:Pl	0.01	0.015
Cpx:Pl	0.79	0.07
Mt:Cpx	0.03	0.003
Mt:Pl	0.02	0.002
Ab:An	0.83	0.116
Mg:Fe ^{+2*}	2.17	0.15

Mineral Proportion (mol. %)		2σ
Pl	54.9	5.1
Kfs	0.6	0.5
Cpx	43.2	4.4
Mt	1.4	1.1

Mineral Composition		
An/(An+Ab)	0.55	0.08
Mg#	0.68	0.07

average ratio of Cpx to Pl is 0.79 ± 0.07 , indicating that plagioclase was the predominate Ca-bearing phase in the fractionate. Figure 8.28C examines the possibility that clinopyroxene was the only ferromagnesian mineral involved in the differentiation process. In the plot of $(Mg+Fe^{+2*})$ versus total Si less that in feldspar, the average slope is 0.51 ± 0.02 , which is within error of the theoretical slope of 0.5 resulting from fractionation of clinopyroxene alone. Thus it is probable that plagioclase and clinopyroxene were the principal phases involved in differentiation of the suite.

The possibility that magnetite, or a solid solution including this endmember, was also precipitated is examined in Figure 8.28D where Fe^{+3} is plotted against the Si content of clinopyroxene. The average slope of the regressions in this plot is 0.031 ± 0.003 , which would correspond to 1.4 ± 1.1 mol. % magnetite in the fractionate. The variation in ferric

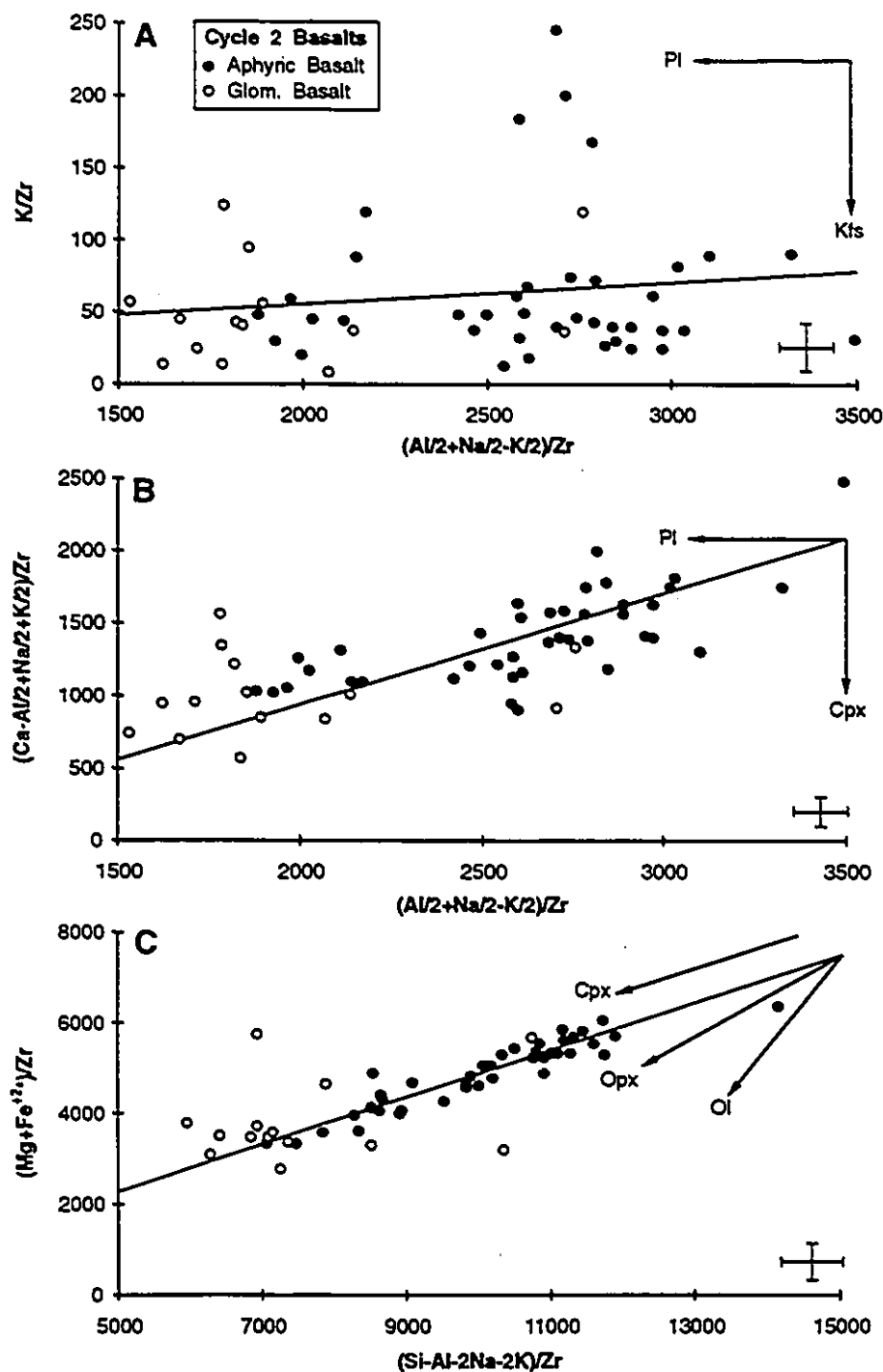


Figure 8.28. Complex numerator Pearce ratio diagrams illustrating the relationships between potential liquidus phases in the fractional crystallization of the Willbobo and upper Menihok magmas. (A) alkali feldspar (Kfs) versus plagioclase (Pl); (B) clinopyroxene (Cpx) versus plagioclase; (C) olivine (Ol) versus orthopyroxene (Opx) versus clinopyroxene. Average 2σ analytical uncertainty is indicated by error bars.

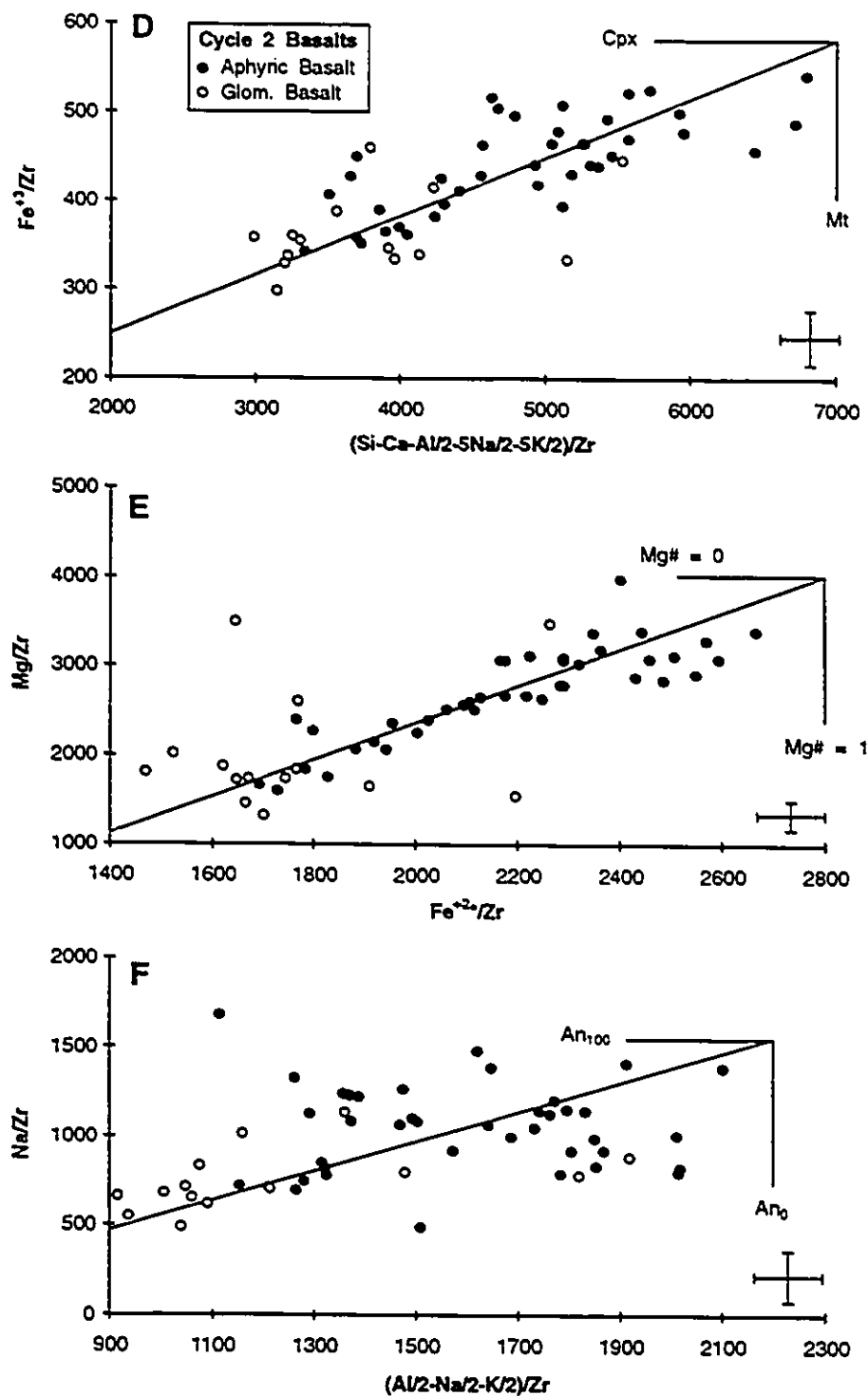


Figure 8.28 (continued). Complex numerator Pearce ratio diagrams illustrating the relationships between potential liquidus phases in the fractional crystallization of the Willbob and upper Menihek magmas. (D) magnetite (Mt) versus clinopyroxene; (E) anorthite (An) versus albite (Ab) in plagioclase (F) Mg# in clinopyroxene. Average 2σ analytical uncertainty is indicated by error bars.

iron content, however, can also be explained by fractionation of clinopyroxene containing 0.95 wt. % Fe_2O_3 . Given that the Fe_2O_3 content of augite in the comagmatic Menihek gabbros as determined by microprobe analysis is generally higher than this value, as well as the fact that TiO_2 is conserved within the suite, it is unlikely that fractionation of a discrete oxide phase contributed to the geochemical variation of the cycle 2 tholeiitic magmas.

The foregoing analysis indicates that the variations in major element concentrations in the Willbob and upper Menihek basalts as sampled in this study was principally controlled by fractionation of solid consisting of approximately 56 ± 5 mol. % plagioclase and 44 ± 4 mol. % clinopyroxene (Table 8.6B). This process has also controlled the concentrations of most incompatible and immobile elements. It is probable, however, that post-crystallization processes have influenced the concentrations of some elements in volcanic rocks. This is suggested by the poorly defined trends between relatively incompatible and immobile trace elements such as Zr, and the more mobile constituents such as Sr and Ba. This phenomenon is also revealed in the Pearce ratio analysis; for example the poor correlation found in the plots of Kfs versus Pl. In addition, attempts to evaluate the average An content of fractionated plagioclase (Fig. 8.28E), and the average Mg# in clinopyroxene (Fig. 8.28F) by ratio analysis appear to suffer from such effects. These calculations indicate that the samples are related by fractionation of plagioclase having the average composition $\text{An}_{54}\text{Ab}_{45}\text{Or}_1$, and clinopyroxene having an average Mg# of approximately 0.68. While the uncertainties associated with these values are relatively high, the compositions would seem to be inconsistent with other data. For example, the average composition of plagioclase glomerocrysts (An_{76}) in the Menihek gabbros is considerably more calcic than that calculated above. Similarly, the Mg#s of augite grains in the gabbros, while highly variable, tend to be higher than the calculated average value for clinopyroxene removed from the upper Menihek/Willbob magmas.

One probable influence on the ratio data is the mobility of Na_2O during post-crystallization processes. The albitization of chilled margins of the Menihek gabbros is clearly documented, and a similar process was suggested as the source of the highly variable Na_2O contents of the basalts. Post-crystallization introduction (or removal) of Na_2O could have an important effect on any Pearce ratios in which this component formed

part of the numerator. Na-metasomatism would certainly influence the calculated plagioclase composition in the fractionate, but would also have an effect, albeit probably less substantial, on other diagrams since sodium concentrations are involved in many of the numerators. Alteration-related variations in other elements such as K would also impact the validity of the ratio analyses.

Despite some suspect results in the analysis of fractionating phases, the strong correlations in several of the complex numerator Pearce diagrams and the evidence from log-log plots and incompatible element ratios indicate that the basic interpretation of the geochemical data is valid. As in the case of the Le Fer basalts, however, there is some discrepancy between petrographic and geochemical evidence. This arises from the fact that olivine precipitation is again suggested by the presence of this phase as phenocrysts in some flows. In addition, it is thought that olivine found in some Menihek sills (e.g., Howse Lake sill C of this study) is a product of pre-emplacment crystallization since geochemical variations in samples from the sills are inconsistent with *in situ* olivine fractionation. But olivine fractionation is inconsistent with the major and trace element variations in the volcanic data set presented here.

The discrepancy between petrographic and geochemical evidence in the case of the Le Fer basalts is thought to reflect evolution of primitive Le Fer magmas through olivine fractionation. Because the cycle 2 tholeiites examined in this study are even more evolved than the Le Fer basalts, it is also unlikely that any of these represent primary partial melts. As in the analysis of the cycle 1 samples, the standard errors on the slopes of the regressions in the ratio analysis of cycle 2 tholeiites could mask minor olivine precipitation. Because of the relatively large difference in the stoichiometric ratios of Si to (Mg+Fe²⁺*) in olivine and clinopyroxene, however, the errors associated with the regressions translate into a smaller proportion of potential olivine fractionation; in this case a maximum of 2 mol. % olivine in the fractionate. This quantity might be sufficient to explain the relatively uncommon occurrence of olivine phenocrysts in the cycle 2 tholeiites, and the minor proportions of olivine in the Menihek sills. But the co-precipitation of plagioclase, pyroxene and minor olivine cannot explain the occurrence of thick sequences of olivine cumulates in differentiated mafic/ultramafic sills (Retty Peridotite) intruding the Thompson Lake and Willbob formations. The sills have been

interpreted as comagmatic members of the cycle 2 tholeiitic suite (Beaudoin and Laurent, 1989), but they are clearly the products of magmas more primitive than those of the Menihek sills and upper Menihek/Willbob basalts. In addition, Beaudoin and Laurent (1989) reported a Willbob basalt analysis with ~ 11 wt. % MgO, and chilled margin samples from the differentiated sills with a 11-13 wt. % MgO.

Given the compositions of the Retty sills, and the occurrence of at least one relatively primitive Willbob flow, it is probable that the upper Menihek/Willbob basalts and the Menihek gabbroic sills represent the products of residual melts derived from differentiation of more primitive magmas. The thicknesses of olivine cumulates in the ultramafic portions of the sills suggest that olivine played an important role in the evolution of the primitive melts toward compositions in which plagioclase and clinopyroxene were liquidus phases. While the high-Mg basalt sampled by Beaudoin and Laurent (1989) indicates that the sample suite obtained in this study does not encompass the entire compositional range of the cycle 2 volcanic sequence, it is interesting that the vast majority of published analyses of Willbob and Hellancourt basalts have MgO contents of less than 10 wt. %, and most fall in the range 5-9 wt. % MgO. This suggests that the more primitive magmas were largely trapped within the crust.

8.6 Summary

The Le Fer volcanic rocks are a relatively homogeneous succession of olivine tholeiites and quartz tholeiites which are characterized by low incompatible element contents and low REE concentrations. The REE patterns tend to be flat, although some samples show small degrees of light REE depletions, and many exhibit negative Eu anomalies. The limited range of compositional variation evident in the suite, and the nature of this variation, is most compatible with magmatic differentiation through fractionation of plagioclase, orthopyroxene and clinopyroxene. Indistinguishable incompatible element ratios among samples of Le Fer basalts and gabbros indicate that the spatially associated intrusions and flows are comagmatic.

The upper Menihek and Willbob basalts are also relatively homogeneous quartz tholeiites or, less commonly, olivine tholeiites. In comparison to the Le Fer basalts,

however, the upper Menihek and Willbob volcanics are slightly richer in incompatible elements and slightly depleted in Cr and Ni. In addition, the REE patterns of the cycle 2 tholeiites are flat to slightly light REE enriched, and generally exhibit moderate to strong negative Eu anomalies. Incompatible element ratios among samples from the upper Menihek and Willbob basalts, and Menihek and Doublet gabbros, indicate that the four suites are comagmatic. These ratios are very similar to those of the Le Fer suites and, despite minor geochemical differences, this suggests that cycle 1 and 2 tholeiitic magmas were derived from compositionally similar sources. The limited compositional variations in the cycle 2 extrusive rocks, as well as in the chilled margins of the Menihek sills, are consistent with melt evolution through fractional crystallization of plagioclase and clinopyroxene.

The interpretations of the geochemical data from cycle 1 and 2 tholeiitic sequences are complicated by inferred compositional variations due to post-crystallization processes. Although crystal fractionation appears to be the predominant source of variation for most major and incompatible/immobile trace elements, this process is unlikely to be the source of the diverse concentrations of some constituents. The most likely explanation for this discrepancy is that certain elements (e.g. Na_2O , K_2O , Ba and Sr) have undergone changes in concentration in some samples during alteration and metamorphism. A further complication in the interpreted differentiation of cycle 1 and 2 magmas is the apparent discord between the geochemical data and petrographic features, notably the presence of olivine phenocrysts in rocks whose compositions are inconsistent with extensive fractionation of this phase. Because the two tholeiitic suites consist predominantly of relatively evolved basalts, it is considered likely that they represent residual liquids derived from primary partial melts through olivine fractionation. This may indicate that the sample suites are incomplete in terms of compositional diversity, or that more primitive magmas from which olivine was fractionated did not reach the surface.

The stratigraphic and geochemical similarities between the alkali basalts of the Nimish and lower Menihek formations indicate that they had a similar origin. The temporal and spatial relationship between these suites, however, indicates that they are not comagmatic. The high incompatible element contents, high REE concentrations, and light REE enriched patterns of the Nimish and lower Menihek volcanics contrast markedly with

the tholeiitic geochemistries of the Le Fer, upper Menihek and Willbob basalts. In addition, the alkaline rocks exhibit considerable geochemical diversity in which covariations between elements are poorly defined. As in the case of the tholeiitic suites, some of this variation appears to reflect post-crystallization effects. The considerable variations in incompatible/immobile element ratios, however, indicate that alteration and metamorphism cannot be responsible for all of the compositional diversity of the alkaline suites. Ratio analysis and log-log diagrams indicate that crystal fractionation and/or fusion mechanisms are also unlikely to have been the predominant sources of compositional variation. Consequently, it is probable that the geochemistries of the alkaline rocks were influenced by non-uniform source compositions and/or assimilation of crustal material.

The Murdoch basalts can be subdivided into high-Zr and low-Zr suites. Although the data are limited, the low-Zr type volcanics appear to predominate in the Murdoch Formation, while the high-Zr basalts appear to be confined to the lower part of the succession. In addition to the apparent stratigraphic differences, the low-Zr basalts share some of the subalkaline characteristics of the upper Menihek and Willbob tholeiites, while the high-Zr group share some of the alkaline characteristics of the Nimish and lower-Menihek suites. Despite these affinities, however, both groups appear to have transitional geochemistries. For example, while the incompatible element content of the low-Zr Murdoch basalts is less than their high-Zr counterparts, it is comparable to the more incompatible element enriched members of the cycle 2 tholeiitic suite. Similarly, the REE patterns of the low-Zr basalts are typically slightly light REE enriched and heavy REE depleted in comparison to the upper Menihek/Willbob tholeiites. A transitional character is also evident in the compositions of the high-Zr Murdoch basalts. While these are incompatible element and light REE enriched relative to the low-Zr basalts, the degree of enrichment is considerably less than that found in the Nimish and lower Menihek volcanics. Nevertheless, the stratigraphic positions, geochemical characteristics and incompatible element ratios indicate that the high-Zr Murdoch basalts are more closely related to the alkaline suites than the overlying subalkaline rocks, while the low-Zr basalts are more closely related to the cycle 2 tholeiites. This may indicate that the high-Zr suite represents the waning stages of localized alkaline magmatism, while the low-Zr suite represents the early stages of the more widespread cycle 2 tholeiitic magmatism.

Unfortunately, the limited database precludes a comprehensive examination of the range and source of compositional variations in the Murdoch basalts.

Chapter 9

LTBS Magma Types and Sources

9.1 Magma Types

Le Fer, Upper Menihék and Willbob Basalts

The Le Fer and upper Menihék/Willbob basalts examined in this study are relatively homogeneous olivine tholeiites and quartz tholeiites. In general, the Le Fer basalts exhibit slightly lower incompatible element contents and higher Cr and Ni than the majority of cycle 2 tholeiites, but the two suites are otherwise quite similar. This suggests that the magma sources for cycle 1 and 2 tholeiitic volcanism were also compositionally similar. The limited geochemical variation evident in the samples is thought to principally reflect magmatic differentiation through fractionation of plagioclase, orthopyroxene and clinopyroxene from the Le Fer magmas, and fractionation of plagioclase and clinopyroxene from the upper Menihék/Willbob melts. The relatively evolved nature of both sample suites suggests that they represent the volcanic products of residual melts derived from more primitive primary magmas through olivine fractionation. This interpretation is supported in the case of the cycle 2 tholeiites by the occurrence of thick olivine cumulate sequences and high-Mg chilled margins in the Retty mafic/ultramafic sills, and by a single example of a comparatively MgO-rich flow (Beaudoin and Laurent, 1989). No such high-Mg compositions have been reported from the Le Fer basalts, but the predominance of olivine phenocrysts in many flows supports an early stage of olivine fractionation prior to crystallization of plagioclase and pyroxene.

The Cr and Ni contents of the Le Fer and upper Menihék/Willbob basalts are plotted against MgO in Figure 9.1 for comparison with representative tholeiitic suites from potentially comparable tectonic environments. It is apparent that all of the LTBS tholeiitic samples obtained in this study have Cr, Ni and MgO contents which are comparable to the more evolved members of each suite, including the typically Mg-poor mid-ocean ridge basalts (MORB) and continental flood basalts (CFB). This is also true

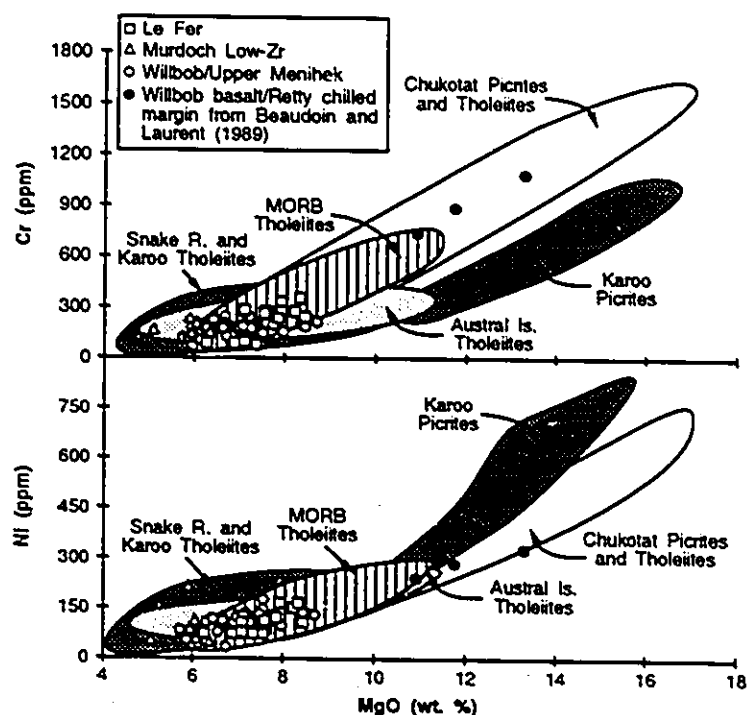


Figure 9.1. Cr and Ni versus MgO in the Le Fer, upper Menihek/Willbob and low-Zr Murdoch basalts in comparison with concentrations in selected suites from potentially comparable tectonic environments. MORB data from Sun and Sharaskin (1979), Snake River and Karoo data (CFB) from Cox (1983), Austral Islands data (ocean island tholeiites) from Dupuy *et al.* (1988), and Chukotat data from Hynes and Francis (1982) and Francis *et al.* (1983).

when the LTBS tholeiites are compared to volcanic rocks of the approximately coeval Chukotat Formation of the Cape Smith fold belt. These comparisons support the suggestion that the Le Fer and upper Menihek/Willbob basalts sampled in this study do not represent primary melts. In contrast, the high-Mg Willbob flow and Retty chilled margins tend toward the picritic fields in Figure 9.1, and are thus more likely to represent primary magmas. As indicated in the previous section, olivine fractionation in primitive cycle 1 and 2 subalkaline magmas could lead to residual liquid compositions appropriate to the fractionation of plagioclase and pyroxene. Incomplete removal of precipitated olivine grains from suspension in the residual liquids would allow eruption of olivine-bearing lavas. The difference in the relative proportion of olivine phenocrysts in the Le Fer basalts at Tait and Chassin lakes might then reflect removal of most unaccumulated olivine crystals in early eruptions (i.e., in the stratigraphically lower and olivine phenocryst

enriched Tait Lake sequence). Entrainment of unaccumulated olivine grains in residual melts tapped from magma reservoirs might also explain the occurrence of partially resorbed olivine crystals in some Menihek gabbro sills at Howse Lake (see Chapter 5). Although eruption of some high-Mg lavas cannot be excluded, as appears to be the case in the Willbob suite, the predominance of evolved flow compositions in both cycle 1 and 2 subalkaline volcanic sequences requires that primitive melts were generally trapped below the surface. This may indicate that a period of olivine fractionation in crustal or subcrustal magma chambers was a necessary prerequisite for extrusion.

Extended trace element patterns normalized to average concentrations in normal type mid-ocean ridge basalt (N-MORB) are plotted in Figure 9.2 for the cycle 1 and 2 tholeiites, excluding those of the Murdoch Formation. With the exception of a consistent trough at P and a peak at Ba, the Le Fer basalts exhibit generally flat trace element patterns with slight Lu to Nb enrichment trends and concentrations marginally greater than average N-MORB (Fig. 9.2A). The samples also show slight Zr depletions relative to Nd and Sm, and variable but consistently depleted K contents relative to Ba and Nb. Rb concentrations are generally below XRF detection limits (~ 2 ppm), but the single available analysis determined by mass spectrometry suggests that Le Fer Rb concentrations are comparable to, or slightly greater than, those of N-MORB. Because of the potential mobility of Rb, K and Ba, it is uncertain whether the abundances of these elements are primary or secondary features. But an extension of the trace element patterns from the light REE suggests that the Ba contents may reflect primary values while the Rb and K abundances have been depleted by post-crystallization processes. There is a moderate range in trace element concentrations among the Le Fer basalts, but the pattern for the most evolved sample (as determined by Mg#) is comparable to that of the most primitive sample except for the elevated overall concentrations and a marked Sr depletion (Fig 9.2B). The latter feature probably reflects fractionation of plagioclase during melt differentiation but, given the variability of Sr content among the Le Fer basalts, it is possible that depletion through crystallization was augmented by Sr loss during alteration.

The aphyric upper Menihek and Willbob basalts exhibit relatively flat extended trace element patterns which are quite similar to those of the Le Fer basalts (Fig. 9.2C), except for a slightly more pronounced Lu through La enrichment trend. Again there is a

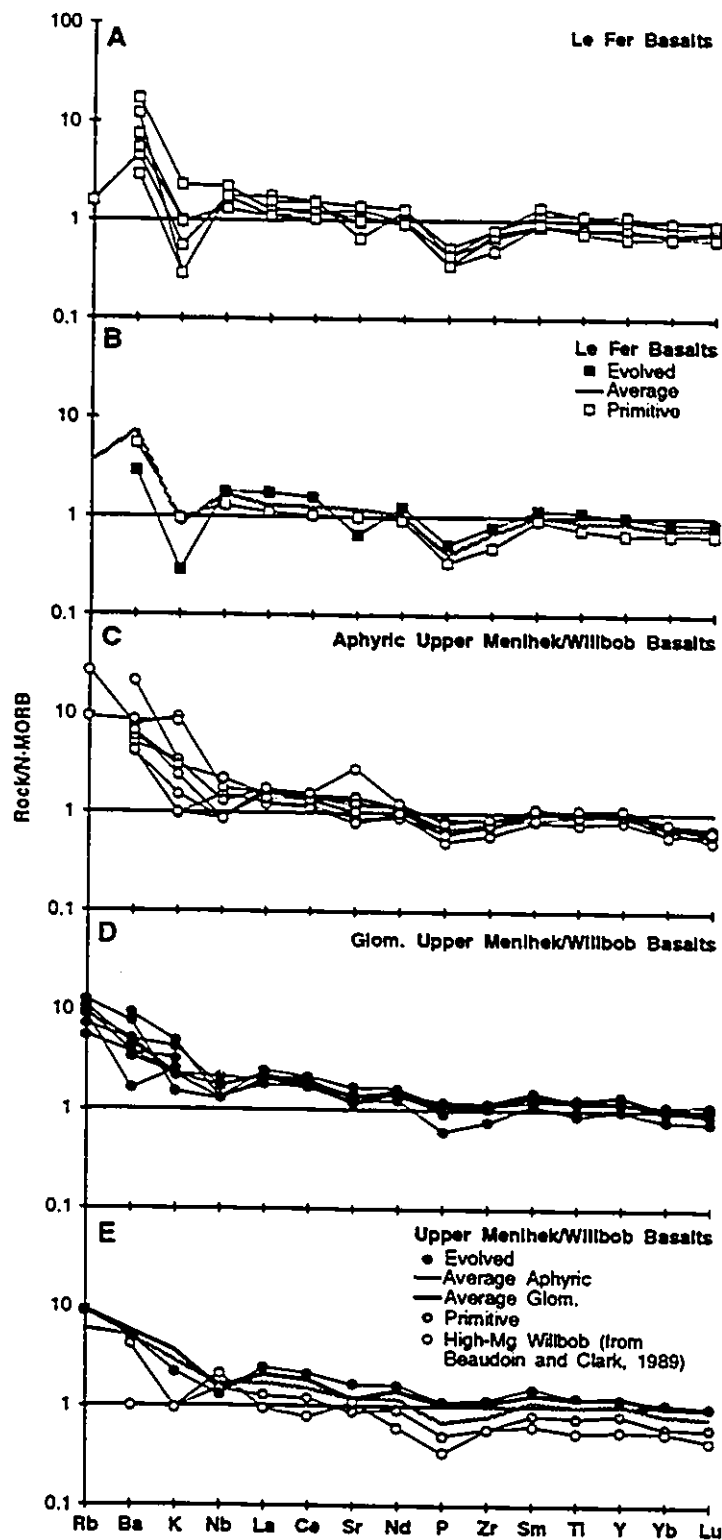


Figure 9.2. Normalized extended trace element plots for the Le Fer and upper Menihok/Willibob basalts. Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba which is from Saunders and Tarney (1984). Data for the high-Mg Willibob flow are from Beaudoin and Laurent (1989).

consistent, albeit weaker, trough at P, and a peak at Ba. Rb concentrations are again often below detection limits. The impoverishment of the Le Fer basalts in P_2O_5 relative to cycle 2 tholeiites was noted in the analysis of incompatible element ratios, and this appears to be one of the few consistent geochemical features which distinguishes cycle 1 and 2 melt compositions. Other than the discrepancy in P, the principal difference in the two sets of patterns appears to lie in the relatively low K contents of the Le Fer basalts. The large variation in K abundances in comparison to the variation in less mobile elements suggests that they do not represent magmatic values. The lack of correlation between K and Al documented in the analysis of fractional crystallization in the cycle 2 volcanics was thought to reflect non-conservation of K, and this might indicate loss of this element through post-crystallization mobility. If this interpretation is correct, the variability in the concentration of K in the cycle 2 tholeiites reflects variable loss of this element, with only the higher concentrations approximating magmatic values. The lower K content of the Le Fer basalts would presumably reflect consistently greater removal of this element. Some support for this interpretation is provided by the low Rb concentrations in the samples, since this element should be mobilized under the same conditions as K.

In comparison to the aphyric samples, the glomeroporphyritic basalts exhibit slightly higher incompatible trace element concentrations although the overall patterns are very similar (Fig 9.2D). All of the glomeroporphyritic samples, however, exhibit slight Nb and Sr depletions relative to adjacent elements, and these depletions are most pronounced when primitive and evolved samples are compared (Fig. 9.2E). Some of the aphyric basalts also display these anomalies, but in the glomeroporphyritic samples they are consistently developed. In contrast, the high-Mg Willbob basalt reported by Beaudoin and Laurent (1989) exhibit Nb and Sr enrichments. Since the glomeroporphyritic basalts are, in general, compositionally more evolved than their aphyric counterparts, the Nb and Sr depletions may be functions of the differentiation process. While this is understandable for the plagioclase-compatible element Sr, it is anomalous in the case of Nb since this element is incompatible in the phases thought to have controlled melt evolution. One potential explanation for a progressive Nb depletion could be the assimilation of sialic crust, since this material is depleted in Nb. The fact that the Nb depletions are most pronounced in samples (aphyric or glomeroporphyritic) having the highest incompatible element

concentrations is consistent with increasing degrees of crustal contamination during the evolution of the cycle 2 magmas.

Average trace element concentrations for the two tholeiitic suites have been plotted in the extended trace element diagrams of Figure 9.3 for comparison with representative examples of potentially comparable tholeiitic suites. Although there are some relative differences in trace element contents between primitive and evolved samples in the suites (Fig. 9.2), the overall patterns are similar. Consequently average abundances should provide a representative trace element pattern for comparative purposes. The averaging of trace element concentrations will also tend to reduce anomalies resulting from variable element mobility in samples that have undergone varying degrees of alteration. It should be noted that because of the low Rb contents of the majority of the tholeiitic samples, average Rb abundances have been calculated using detection limit concentrations (2 ppm) for those samples whose Rb content was below detection limits. The average values so obtained will be relatively inaccurate since true Rb concentrations will range from 0 to 4 ppm (detection limit $\pm$ 100%), but the procedure counters the effect of unreasonable weighting of the few samples whose Rb contents are perhaps anomalously high.

The cycle 1 and 2 tholeiites exhibit relatively flat trace element patterns with concentrations slightly greater than average N-MORB. The range of trace element concentrations in N-MORB, however, is considerable, as is the range of extended incompatible element patterns observed in this rock type (Fig. 9.3A). Consequently, the average trace element concentrations of the Le Fer and upper Menihék/Willbob basalts lie within the range of values found in N-MORB. The latter however, typically show overall declines in the normalized abundances of Lu through Ba, while the average cycle 1 and 2 patterns show overall increases. In this regard, the average patterns are more comparable to those of transitional MORB (T-MORB) as illustrated in Figure 9.3B. With the exception of elevated Ba contents, the average LTBS tholeiites have elemental abundances that lie within the range found in mid-Atlantic ridge T-MORB. In comparison to enriched MORB (E-MORB; Fig. 9.3C), the average trace element abundances of cycle 1 and 2 tholeiites are lower for most elements more incompatible than Zr, and the overall enrichment trend is less pronounced. In comparison to ocean island tholeiites (Fig. 9.3D), cycle 1 and 2 basalts are far less incompatible element enriched.

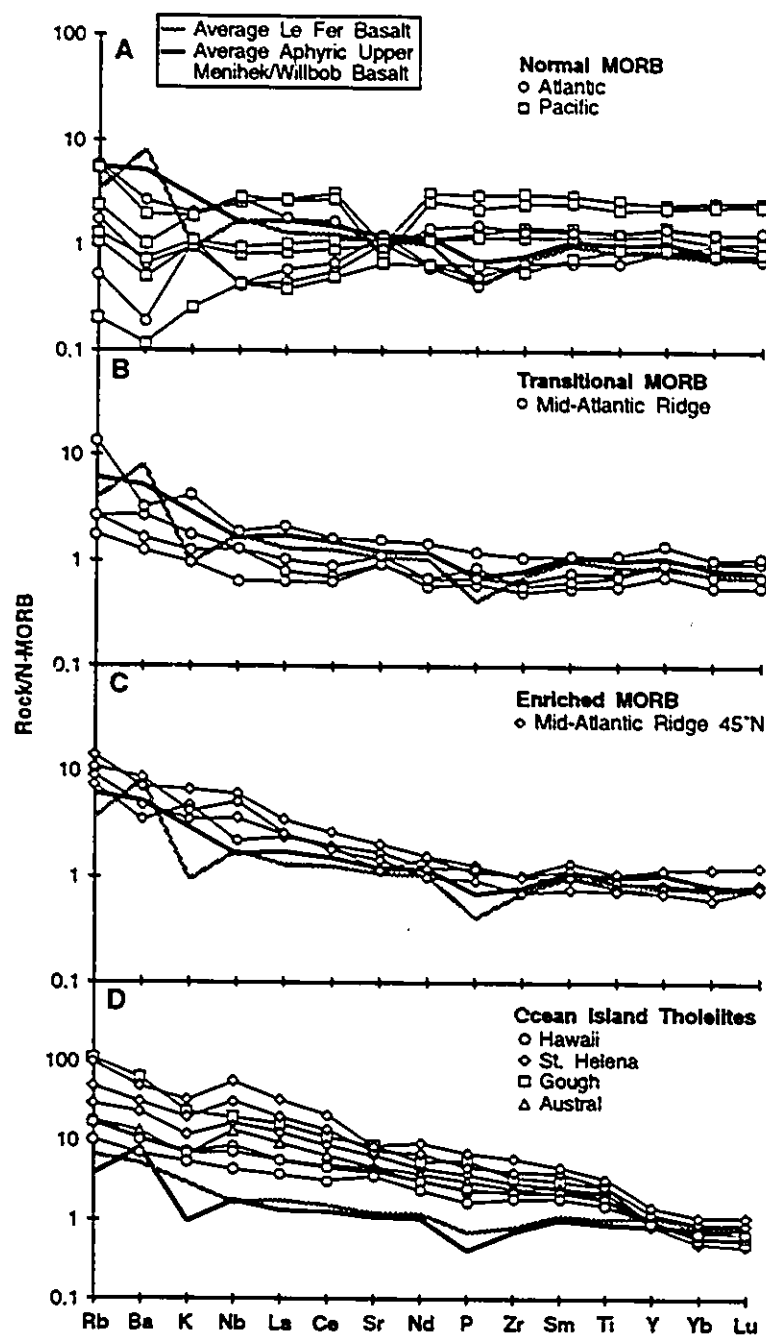


Figure 9.3. Normalized extended trace element plots comparing average abundances in the Le Fer and upper Menihek/Willbob tholeiites to representative samples of normal, transitional and enriched MORB, and ocean island tholeiites. Data from Sun and Sharaskin (1979), Newsom *et al.* (1986), Dupuy *et al.* (1988) and Sun and McDonough (1989). Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba which is from Saunders and Tarney (1984).

The small degree of enrichment in the progressively more incompatible elements suggests that the Le Fer and upper Menihek/Willbob magmas were compositionally similar to those forming modern T-MORB. It is possible, however, that some degree of incompatible element fractionation accompanied differentiation, and that the trace element abundances of the parental melts were comparable to N-MORB. The incompatible element pattern of the high-Mg Willbob flow reported by Beaudoin and Laurent (1989) suggests that this was not the case. Elemental abundances in the sample are generally lower than those of the upper Menihek/Willbob tholeiites of this study, as well as lower than average N-MORB abundances (Fig. 9.2). The low incompatible element contents are consistent with the primitive nature of the sample. The high-Mg flow does, however, exhibit the small degree of enrichment in the increasingly incompatible elements Lu to Nb that is typical of the cycle 2 tholeiites. Although this trend does not continue in the abundances of K, Ba and Rb, its presence in the relative abundances of Lu-Nb suggests that it was a characteristic of the primary melts. Although more analyses of primitive flows are required to confirm this interpretation, it seems probable that the cycle 2 tholeiites represent the products of T-MORB type melts. This suggests that the primary magmas were derived primarily from a depleted mantle reservoir with some addition of less depleted or enriched material.

Nimish and Lower Menihek Basalts

The Nimish volcanic rocks range from subalkaline to alkaline in geochemical affinity, and from basalt to rhyolite in composition. In contrast, the lower Menihek suite is strongly alkaline and consists predominantly of alkali basalt. Although the two suites share many geochemical characteristics which distinguish them from cycle 2 tholeiites, they are thought to represent discrete magmatic episodes. Compositional variations in both suites are inconsistent with simple fusion or crystallization mechanisms, and the geochemical diversity is thought to reflect additional processes, possibly including source region heterogeneity and crustal contamination. In the case of the lower Menihek clasts, it is likely that post-crystallization processes have also contributed to compositional variations.

The Cr and Ni contents of the Nimish and lower Menihek volcanics are plotted against MgO in Figure 9.4 for comparison with other alkaline suites. It is apparent that the

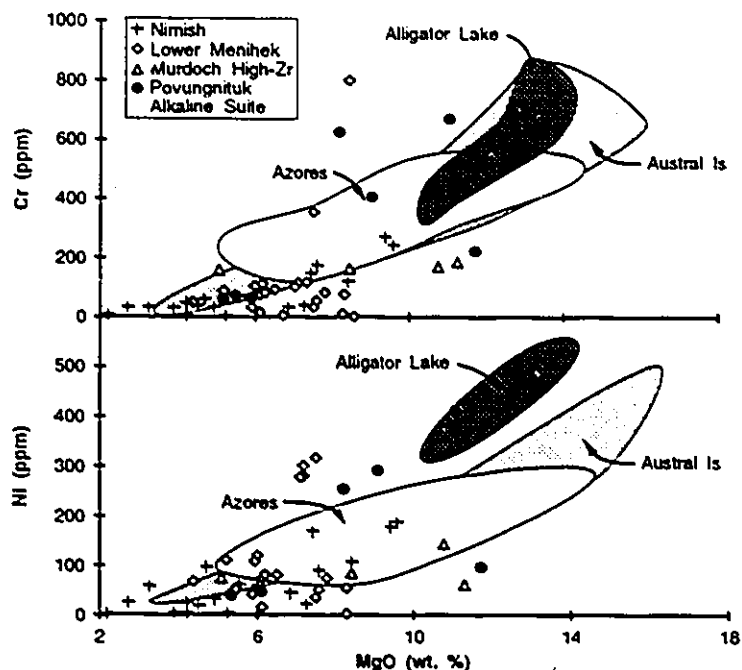


Figure 9.4. Cr and Ni versus MgO in the Nimish, lower Menihek and high-Zr Murdoch basalts in comparison with concentrations in selected suites from potentially comparable tectonic environments. Nimish data from Watanabe (personal communication, 1993), Azores data (ocean island alkali basalts) from Davies *et al.* (1989), Austral Islands (ocean island alkali basalts) data from Dupuy *et al.* (1988), Alligator Lake data (continental alkali basalt) from Eiché *et al.* (1987), and Povungnituk Alkaline Suite data from Gaonac'h *et al.* (1989;1992) and Francis *et al.* (1983).

most primitive of the LTBS alkaline rocks are impoverished in Cr, Ni and MgO relative to primitive samples from young oceanic and continental alkaline suites. This relationship also holds true in comparison to the approximately coeval Povungnituk Alkaline Suite of the Cape Smith foldbelt. This suggests that none of the Nimish or lower Menihek basalts examined in this study represent primary melts. The presence of ultramafic cumulates in some Nimish sills (Watanabe and Fowler, 1989) would also seem to require more MgO-rich magmas than those represented by the most primitive of the extrusions. In addition, the high-Mg clast from the Outlook Lake area, which is thought to be a member of the lower Menihek suite, contains ~23 wt. % MgO and ~2200 ppm. It is not known whether this sample was derived from a flow or an intrusion, but in either case it would seem to require a parental liquid richer in MgO than those of the lower Menihek volcanics.

The high-Mg Murdoch volcanics (18-21 wt. % MgO) reported by Gebert (1988) are potential representatives of such a parental liquid. As discussed in the previous chapter, these appear to be related to the mildly alkaline volcanic rocks in the lower part of the Murdoch Formation. Trace element data reported for one of these samples (#135) include Cr and Ni contents of 2300 and 1000 ppm respectively (Gebert, 1988). While it is not necessary that all of the alkaline samples were derived from parental melts as primitive as those represented by the high-Mg Murdoch volcanics, it seems likely that all were derived from magmas more primitive than those represented by observed basaltic compositions. Assuming the Nimish ultramafic sills are representative of the crystallization processes by which the primitive alkaline magmas became differentiated, it is probable that fractionation of olivine played a role in the development of the more evolved basaltic rocks. The scarcity of high-Mg alkaline volcanics is similar to the situation among the tholeiitic sequences where the predominant rock type is relatively evolved basalt. Thus it is possible that a similar period of olivine fractionation in crustal or subcrustal magma chambers was required before most alkaline melts could ascend to the surface.

The foregoing discussion suggests that the Nimish and lower Menihek volcanics represent the derivative products of more primitive liquids. The compositional diversity and the non-constancy of incompatible element ratios suggest that magmatic differentiation was not the only source of geochemical variations in the suite. Despite what appears to be a complex history of compositional evolution, the extended trace element patterns of the Nimish and lower Menihek alkali basalts exhibit many similarities (Fig. 9.5A and 9.5C). The most obvious shared characteristic is the strong Lu to Rb enrichment trend. Both suites also show Sr, Zr, Y and heavy REE depletions relative to adjacent elements. The two suites differ primarily in the extent of the latter anomalies, and in the range of trace element abundances. The most primitive and most evolved lower Menihek samples show little variation in trace element abundances (Fig. 9.5B), whereas a greater range of concentrations is apparent between the most primitive and most evolved of the Nimish basalts (Fig. 9.5D). This is consistent with the greater compositional diversity found in the latter suite. In addition, the most primitive Nimish sample displays a slight P depletion not evident in other samples, while the most evolved sample shows a small Nb depletion and a more pronounced Sr depletion than found in other samples. The variations

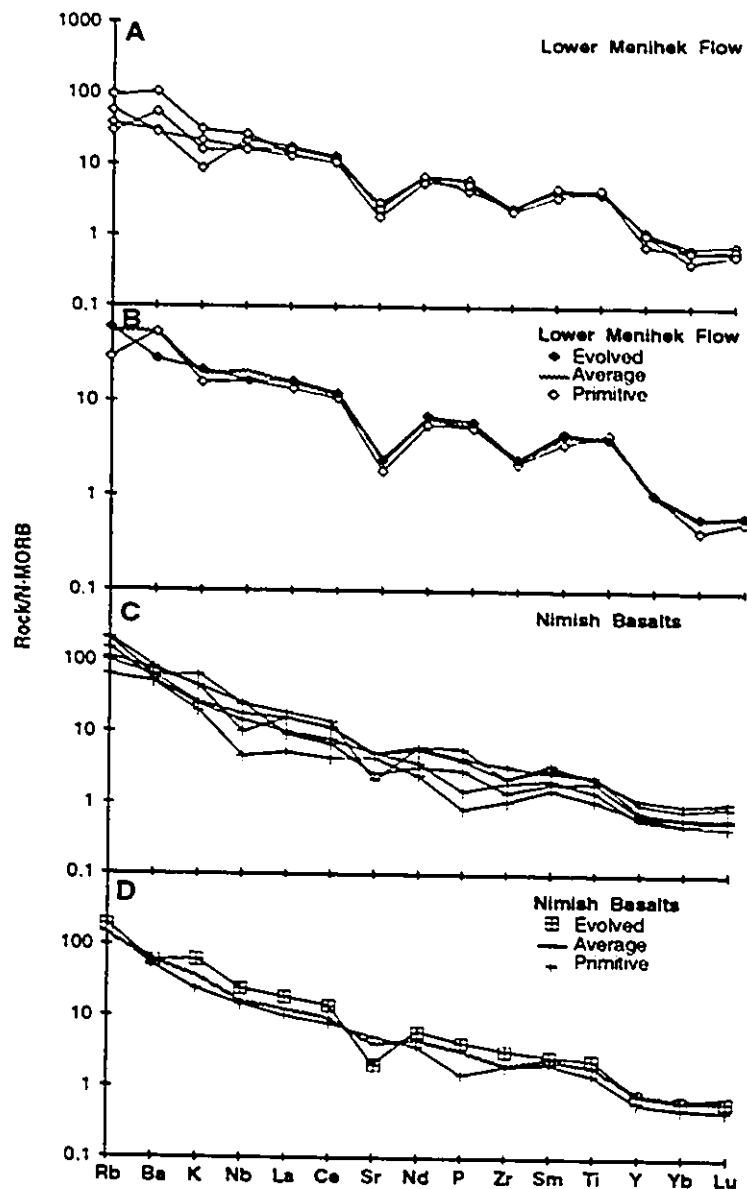


Figure 9.5. Normalized extended trace element plots for the Nimish basalts and lower Menihek flows. Nimish data are from Watanabe (personal communication, 1993). Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba which is from Saunders and Tarney (1984).

in incompatible element abundances among the Nimish and lower Menihek basalts may reflect different fractional crystallization histories, source heterogeneity, contamination or other processes, but the fact that the non-comagmatic suites share a comparably enriched trace element pattern suggests that this feature is a primary characteristic of the parental melts. This interpretation is supported by the trace element patterns of the high-Mg lower

Menihek clast and high light REE concentrations in the high-Mg Murdoch volcanics (e.g., sample #135; Gebert, 1988).

Average normalized trace element abundances for the Nimish and lower Menihek suites are plotted in Figure 9.6 for comparison with alkaline suites from continental and oceanic island settings. Representative samples of E-MORB are also illustrated, but it is apparent that the LTBS alkaline rocks are considerably enriched in Rb through Ti in comparison. The E-MORB exhibit comparable Y and heavy REE concentrations relative to the Nimish and lower Menihek basalts, but these elements are strongly depleted in the latter rocks. In contrast, the overall shape of the average trace element patterns for the LTBS alkaline rocks are quite similar to those of alkaline basalts from oceanic and continental settings. As noted by Norry and Fitton (1983) alkaline volcanics of ocean islands and continental areas are very similar in terms of both incompatible trace element abundances and isotopic composition, and both are thought to represent the products of a relatively undepleted or enriched mantle source. The similarity of trace element patterns and degree of incompatible element enrichment indicates that the parental magmas to the Nimish and lower Menihek suites were also derived from an undepleted or enriched mantle source.

Murdoch Basalts

The basaltic rocks of the Murdoch Formation can be subdivided into low-Zr quartz tholeiites which predominate in the succession, and high-Zr slightly alkaline basalts which are restricted to the lowest exposed part of the formation. The high-Zr basalts have greater incompatible trace element contents than the low Zr group, despite the fact that they are generally richer in MgO. Although the data are limited, the extended trace element patterns of the high-Zr basalts also show a greater overall Lu through Ba enrichment trend than the subalkaline suite, but are slightly depleted in K, Sr and Zr relative to adjacent elements (Fig. 9.7A). The low-Zr basalts may also show slight Sr and Zr depletions, but are characterized by small but consistent Nb and P depletions. In addition, the subalkaline rocks exhibit minor Ti depletions whereas the high-Zr samples are enriched in this element. Despite these differences, the trace element pattern of the high-Zr sample showing the smallest degree of incompatible element enrichment is not very different from

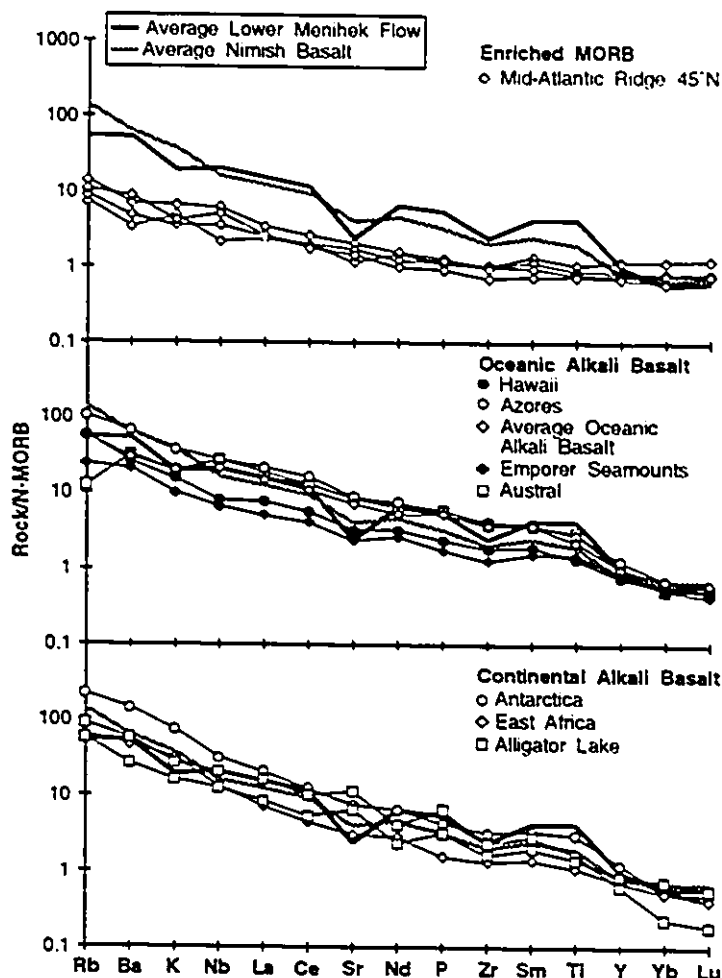


Figure 9.6. Normalized extended trace element plots comparing average abundances in the Nimish basalts and lower Menihok flows to representative samples of E-MORB, oceanic island alkali basalts and continental alkali basalts. Data from Baker *et al.* (1977), Bence *et al.* (1980), Basaltic Volcanism Study Project (1981), Kyle (1981), Newsom *et al.* (1986), Eiché *et al.* (1987), Dupuy *et al.* (1988), Davies *et al.* (1989), and Sun and McDonough (1989). Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba which is from Saunders and Tarney (1984).

the most incompatible element enriched low-Zr sample. Thus it is possible that a more detailed sampling program would reveal a complete gradation between the two groups.

In many respects, the low-Zr Murdoch basalts are geochemically comparable to the tholeiitic basalts of the cycle 2 upper Menihok and Willbob Formations. The average extended trace element pattern of the subalkaline Murdoch samples, for example, is very similar to the average patterns for the cycle 2 tholeiites, and the average abundances of the

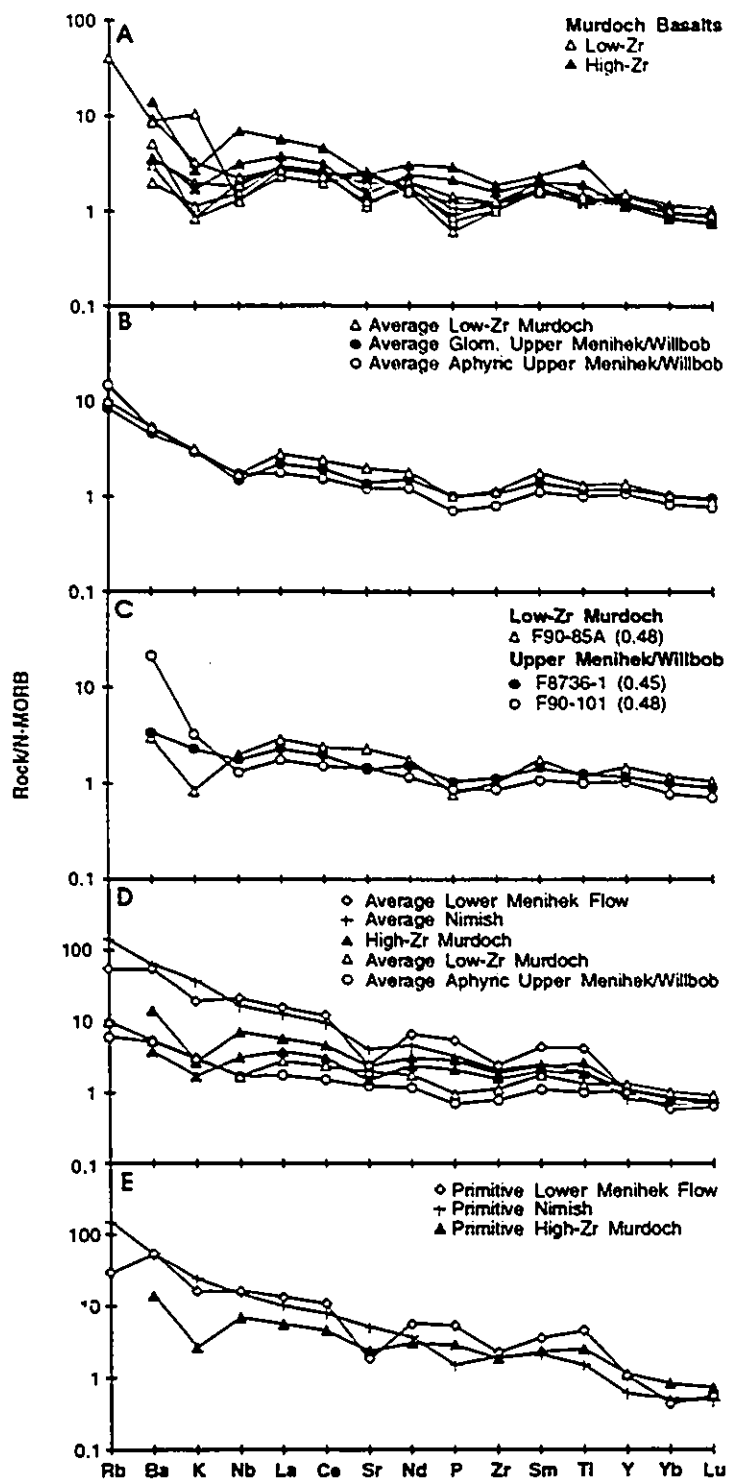


Figure 9.7. Normalized extended trace element plots for the two groups of Murdoch basalts in comparison to other cycle 2 suites. Rb concentrations are below detection limits (equivalent to a normalized abundance of approximately 4) for most Murdoch samples. Mg#s are in parentheses in (C) Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba which is from Saunders and Tarney (1984).

former are only slightly greater than those of the relatively evolved glomeroporphyritic flows (Fig. 9.7B). The low-Zr Murdoch basalts, however, tend toward higher incompatible element contents and light REE enrichments in comparison to other cycle 2 subalkaline rocks. This is apparent in Figure 9.7C where the extended trace element pattern of the most primitive low-Zr Murdoch sample is compared to that of a Willbob basalt with comparable Mg#. The Murdoch sample shows elevated incompatible element abundances relative to the comparable Willbob flow, as well as to a more evolved Willbob glomeroporphyritic basalt.

In terms of Cr, Ni and MgO contents, the low-Zr Murdoch samples are again comparable to evolved cycle 2 tholeiites (Fig. 9.1) and are consequently unlikely to represent primary melts. Although the limited database precludes an assessment of the sources of geochemical variation in the low-Zr Murdoch suite, these samples generally fall on the differentiation trends defined by the upper Menihek/Willbob suite. This suggests that the Murdoch subalkaline basalts represent the differentiation products of primary magmas whose major element compositions were generally similar to the parental melts of the other cycle 2 tholeiitic rocks. The greater incompatible element contents of the low-Zr Murdoch suite, however, indicates that either the primary partial melts were slightly incompatible element enriched relative to those of the upper Menihek and Willbob basalts, or that these magmas underwent an enrichment process subsequent to melt generation.

In previous discussions, the stratigraphic, lithological and geochemical similarities among the Nimish, lower Menihek and high-Zr Murdoch suites were cited as evidence in support of a comparable origin for these rocks. While it is clear that the slightly alkaline Murdoch samples are far less incompatible element enriched than basalts from the Nimish and lower Menihek formations, it is also clear that the former are geochemically distinct from the cycle 2 tholeiites. This is perhaps most apparent in a comparison of the extended trace element patterns of the suites (Fig. 9.7D). The slightly alkaline Murdoch basalts exhibit a greater overall enrichment trend than cycle 2 subalkaline rocks, lack the P depletion of the latter, and share the Ti enrichment and Sr, Zr, Y and heavy REE depletions of the Nimish and lower Menihek suites. This incompatible element enrichment would appear to be a characteristic of the parental melts since the most enriched high-Zr

sample is considerably more primitive (i.e., higher Mg#, Cr and Ni) than any of the low-Zr Murdoch basalts and the majority of upper Menihek/Willbob basalts.

Two of the high-Zr Murdoch basalts contain higher concentrations of MgO than any of the Nimish and lower Menihek samples, although Cr and Ni concentrations are lower than in the most primitive of the latter (Fig. 9.4). Even these samples, however, have far lower MgO, Cr and Ni contents than primitive alkaline volcanics from oceanic and continental settings. In addition, the high-Zr Murdoch basalts are the most likely suite to be related to the high-Mg Murdoch volcanics reported by Gebert (1988). It is therefore probable that, as with the Nimish and lower Menihek basalts, the high-Zr Murdoch samples do not represent primary partial melts.

A comparison of the trace element patterns of the most primitive high-Zr Murdoch, Nimish and lower Menihek basalts (Fig. 9.7E) suggests that the Murdoch flows were derived from primary melts that were only mildly incompatible element enriched relative to the highly enriched parental melts of the alkaline suite. Thus the slightly alkaline Murdoch magmas appear to have been transitional between the relatively incompatible element poor cycle 2 subalkaline melts and the incompatible element enriched cycle 2 alkaline melts. The transitional nature of Murdoch volcanics is particularly apparent when the trace element abundances of all samples from each cycle 2 sequence are considered together (Fig. 9.8). The overlapping nature of the trace element patterns from the individual suites suggests that a more or less continuous spectrum of magma compositions was produced during cycle 2 basin infilling. It is noteworthy that the volcanic activity progressed from local eruption of limited quantities of highly incompatible element enriched alkaline lavas, through more widespread eruption of compositionally transitional lavas, to widespread eruption of great thicknesses of incompatible element depleted subalkaline lavas.

9.2 Radiogenic Isotopes

Sm-Nd isotopic analyses from the Hellancourt and Baby basalts and cycle 2 gabbroic sills in the northern Labrador Trough have been reported by Smith and Ludden (1989) and Skulski *et al.* (1993). Sm-Nd and Rb-Sr isotopic analyses of Willbob basalts

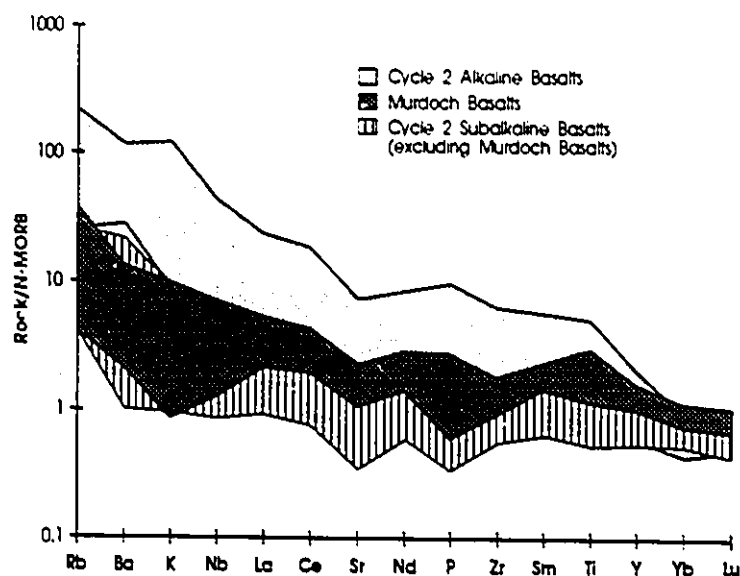


Figure 9.8. Comparison of the total range of normalized trace element abundances for the three principal compositional types of volcanic rocks forming part of cycle 2 infilling. The lower Menihek clastic samples are excluded. Note the overlap among the groups. Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba which is from Saunders and Tarney (1984).

and Montagnais (Doublet) sills of the central Trough have been reported by Rohon *et al.* (1993). In this study, Sm-Nd isotopic analyses were obtained from 17 basaltic and gabbroic samples from the Le Fer, Nimish, lower Menihek, upper Menihek, Murdoch and Willbob formations, and Rb-Sr isotopic analyses were obtained from 6 of these. The results include 2 analyses obtained from glomerocryst plagioclase separates. Analytical procedures are summarized in Appendix A7.

Rb-Sr

The results of the Rb-Sr isotopic analyses and the approximate age of the sampled units are provided in Table 9.1. For comparative purposes, the age of the Menihek gabbros and comagmatic basaltic rocks (1884 Ma) has been applied to the Murdoch and lower Menihek basalts although the latter sequences must be slightly older. The 1884 Ma date, however, is the best available estimate of the age of early cycle 2 magmatism.

Table 9.1. Results of Rb-Sr isotopic analyses for cycle 1 and 2 samples. Approximate unit ages (T) and calculated $\epsilon_{\text{Sr}}(T)$ are also given.

Sample	Unit	Rb (ppm) ¹	Sr (ppm) ¹	⁸⁷ Rb/ ⁸⁶ Sr	⁸⁷ Sr/ ⁸⁶ Sr ²	$\epsilon_{\text{Sr}}(T)$	T (Ma)
F87-35-6	Glom. Willbob Basalt	5.47	147.3	0.1073	0.70495	-3.10	1884
F87-22-15	Glom. Willbob Basalt	3.10	127.2	0.0704	0.70405	-1.65	1884
F87-33-17	Glom. Willbob Basalt	4.94	110.6	0.1291	0.70727	+21.5	1884
F20-7B	Aphyric Menihék Gabbro Chilled Margin	7.33	107.7	0.1970	0.70886	+18.0	1884
F20-9	Lower Menihék Basalt	30.89	204.5	0.4376	0.72036	+88.8	1884
B88-25	Le Fer Basalt	0.90	103.8	0.0250	0.70266	-3.92	2169

¹ Estimated 2 σ precision of concentrations = $\pm 0.5\%$
² ⁸⁷Sr/⁸⁶Sr normalized to 0.1194; estimated 2 σ precision = $\pm 0.003\%$
³ ϵ_{Sr} ratios calculated using Bulk Earth values of ⁸⁷Rb/⁸⁶Sr = 0.0827 and ⁸⁷Sr/⁸⁶Sr = 0.7045

The sample of aphyric Menihék gabbro chilled margin contains slightly higher Rb and lower Sr than those of the glomeroporphyritic Willbob basalts. This relationship is in accord with the data of Rohon *et al.* (1993) who reported average values of 3.76 ppm Rb and 132 ppm Sr, and 7.76 ppm Rb, and 120 ppm Sr, for Willbob basalts and Montagnais gabbros respectively. The elevated incompatible element contents of the lower Menihék alkaline basalts are reflected in the high Rb of sample F20-9, and this has led to an elevated ⁸⁷Sr/⁸⁶Sr ratio in comparison to the cycle 2 subalkaline rocks. The low Rb content of Le Fer basalt sample is consistent with the very low (2.6 ppm) Rb content of a granophyric differentiate in a Le Fer gabbro sill reported by Rohon *et al.* (1993). It also corresponds to the generally low incompatible element content of the Le Fer basalts.

Following DePaolo and Wasserburg (1976), $\epsilon_{\text{Sr}}(T)$ ratios at the time of formation (T) have been calculated for the samples using ⁸⁷Rb/⁸⁶Sr = 0.0827 and ⁸⁷Sr/⁸⁶Sr = 0.7045 for the bulk earth reservoir at present (McCulloch and Wasserburg, 1978; Wen *et al.*, 1987). Positive ϵ_{Sr} values have been interpreted as indicative of a source which has undergone an enrichment in incompatible elements (including Rb) relative to the bulk earth reservoir, while negative values indicate a depleted source (O'Nions *et al.*, 1979). The ϵ_{Sr} values of the cycle 2 tholeiitic rocks analysed in this study range from -3.1 to +21.5, while the data of Rohon *et al.* (1993) show an even greater range (-194 to +38). The ϵ_{Sr}

values of the lower Menihek and Le Fer samples are +88.8 and -3.92 respectively. The Le Fer granophyre analyzed by Rohon *et al.* (1993) has an ϵ_{Sr} ratio of 51.3.

The isotopic data from this study and Rohon *et al.* (1993), and trends for the bulk earth reservoir at 2169 Ma (cycle 1) and 1884 Ma (cycle 2), are illustrated in Figure 9.9A. The large variation in ϵ_{Sr} ratios are apparent in the considerable scatter in the data about the reference lines. This lack of coherence in the cycle 2 subalkaline rocks must be attributable to either contributions from source regions of dramatically different $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (e.g., depleted mantle, enriched mantle, continental crust), or to post-magmatic mobility of Rb or Sr leading to relative Rb enrichment and hence positive ϵ_{Sr} values (cf. Andrew *et al.*, 1991). The homogeneity of major and trace element compositions of these rocks despite the very large variation in ϵ_{Sr} indicates that the latter interpretation is probably correct. This is compatible with the conclusions reached earlier concerning post-crystallization mobility of Sr. A similar process may be responsible for the difference in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the Le Fer basalt and granophyre, although crustal contamination of the latter is more credible given the compositional differences between the two samples. The ϵ_{Sr} value of the lower Menihek sample is consistent with derivation of this suite from an enriched source. The incompatible element and light REE enriched nature of these rocks was also interpreted to reflect partial melting of an enriched source, but since Rb-Sr isotopic data is available for only one sample, the possible effects of Rb-Sr mobility cannot be evaluated.

Sm-Nd

The results of the Sm-Nd isotopic analysis and the approximate ages of the sampled units are provided in Table 9.2. With the exception of plagioclase glomerocrysts from the Menihek sills, the sample of Le Fer basalt contains the lowest concentrations of Sm and Nd. Again this is in accord with the low incompatible element contents of the suite. The Nimish and lower Menihek samples contain the highest concentrations of Sm and Nd, while the concentrations in cycle 2 subalkaline rocks are intermediate between the Le Fer and lower Menihek values. The slightly higher Sm and Nd contents of glomeroporphyritic cycle 2 subalkaline samples in comparison with their aphyric counterparts is consistent with the generally more evolved nature of the former. A similar

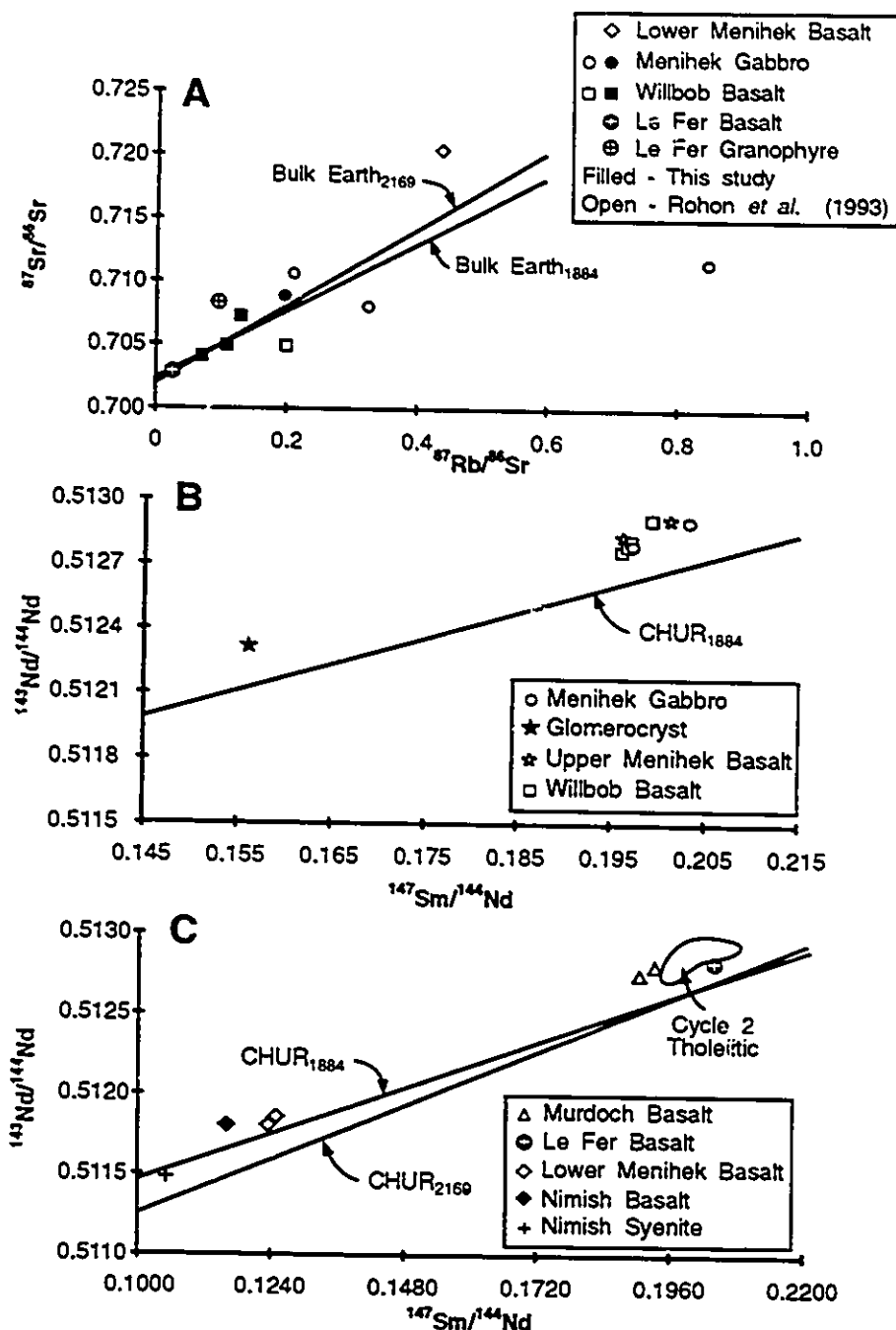


Figure 9.9. Rb-Sr and Sm-Nd isotopic data and reference lines for the bulk earth and CHUR at the appropriate ages. (A) Rb-Sr data for cycle 1 and 2 igneous rocks from this study and that of Rohon *et al.* (1993). (B) Sm-Nd data for comagmatic cycle 2 tholeiitic rocks. (C) Sm-Nd data for Le Fer, lower Menihek and low-Zr Murdoch basalts. 2σ analytical uncertainties are less than the size of plot symbols in each diagram.

Table 9.2. Results of Sm-Nd isotopic analyses for cycle 1 and 2 samples. Approximate unit ages (T) and calculated $\epsilon_{Nd}(T)$ are also given.

Sample	Unit	Sm (ppm) ³	Nd (ppm) ³	¹⁴⁷ Sm/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd ⁴	$\epsilon_{Nd}(T)$ ⁵	T (Ma)
B88-25 ¹	Le Fer Basalt	2.13	6.33	0.2038	0.512830	1.77	2169
W10-11 ²	Nimish Basalt	7.06	36.79	0.1160	0.511807	3.27	1879
B16-5a ¹	Nimish Syenite	17.74	102.20	0.1049	0.511474	-0.57	1879
F90-50 ²	Lower Menihok Basalt	10.18	49.25	0.1250	0.511855	1.95	1884
F20-9 ¹	Lower Menihok Basalt	10.49	51.39	0.1234	0.511810	1.45	1884
F90-85C ²	Low-Zr Murdoch Basalt	3.94	12.35	0.1929	0.512811	4.18	1884
F90-85G ²	Low-Zr Murdoch Basalt	3.75	11.91	0.1902	0.512748	3.61	1884
F90-87 ²	Aphyric Upper Menihok Basalt	3.27	10.08	0.1963	0.512839	4.04	1884
F90-90 ²	Aphyric Upper Menihok Basalt	2.77	8.33	0.2012	0.512919	4.42	1884
F87-35-6 ¹	Glom. Willbob Basalt	4.49	13.85	0.1961	0.512765	2.65	1884
F87-22-15 ¹	Glom. Willbob Basalt	3.38	10.39	0.1966	0.512790	3.01	1884
F87-33-17 ¹	Glom. Willbob Basalt	3.59	11.04	0.1970	0.512810	3.30	1884
F90-99 ²	Aphyric Willbob Basalt	1.67	5.07	0.1993	0.512910	4.71	1884
F20-7B ¹	Aphyric Menihok Gabbro Chilled Margin	2.83	8.69	0.1968	0.512800	3.16	1884
F15-4 ²	Menihok Glomerocryst	2.03	8.26	0.1485	0.512092	1.02	1884
F26-1A(1) ²	Menihok Glomerocryst	0.22	0.86	0.1563	0.512314	3.47	1884
F26-1A(2) ²	Menihok Glom. Gabbro Matrix	3.38	10.06	0.2033	0.512905	3.64	1884

¹ Department of Earth Sciences, Carleton University
² Geochronology Laboratory, Geological Survey of Canada
³ Estimated 2 σ analytical uncertainty $\sim \pm 0.5\%$
⁴ ¹⁴³Nd/¹⁴⁴Nd normalized to 0.7219; estimated 2 σ analytical uncertainty $\sim \pm 0.003\%$
⁵ ϵ_{Nd} ratios calculated using chondritic values of ¹⁴³Nd/¹⁴⁴Nd = 0.512638 and ¹⁴⁷Sm/¹⁴⁴Nd = 0.1967

pattern is present in the data presented by Skulski *et al.* (1993) in that glomeroporphyritic basalts and gabbros contain an average of 3.68 ppm Sm and 11.58 ppm Nd, compared to values of 3.16 ppm Sm and 9.55 ppm Nd for aphyric samples.

With one exception, the Sm and Nd concentrations of the samples are in agreement with the general geochemical characteristics of the suites, and are comparable to values reported for equivalent Labrador Trough rocks by other workers. The exception is the elevated Sm (2.03 ppm) and Nd (8.26 ppm) contents of sample F15-4, which consists of glomerocryst plagioclase separated from the gabbroic matrix. The Sm and Nd contents as determined by mass spectrometry are comparable to whole rock basaltic values, and are far in excess of the concentrations (0.14 ppm Sm and 0.50 ppm Nd; Appendix A4) determined by ICAP-AES for a split of the same sample. The latter concentrations are comparable to

those obtained by mass spectrometry for the second glomerocryst sample reported in Table 9.2, and to those obtained by ICAP-AES from other plagioclase separates. The origin of the large discrepancy in the Sm-Nd concentrations is unknown, but it is clear that the concentration data obtained from the anomalous sample are invalid and therefore the isotopic results are suspect.

The Sm-Nd isotopic data are illustrated in Figures 9.9B and C along with reference lines representing the chondritic uniform reservoir (CHUR) at 2169 Ma and 1884 Ma. All of the cycle 2 tholeiitic rocks, including the low-Zr Murdoch basalts and the valid glomerocryst plagioclase sample, lie above the CHUR_{1884} reference line. The cycle 1 Le Fer basalt lies slightly above the CHUR_{2169} reference line. Of the alkaline rocks, the Nimish basalt lies considerably above the 1884 Ma reference line, the lower Menihék basalts lie slightly above, and the Nimish syenite lies slightly below. These relationships indicate that the cycle 2 subalkaline rocks and the Nimish basalt had initial $^{143}\text{Nd}/^{144}\text{Nd}$ higher than that of CHUR at their time of formation, the Le Fer and lower Menihék basalts had initial ratios slightly higher than CHUR, and the Nimish syenite had an initial ratio slightly lower than the chondritic value. Thus the cycle 1 and 2 basaltic rocks were derived from a source with a time integrated depletion of Nd relative to Sm, while the reverse is true for the Nimish syenite.

These relationships are also apparent in the ϵ_{Nd} ratios (calculated using values for CHUR of $^{143}\text{Nd}/^{144}\text{Nd} = 0.512638$ and $^{147}\text{Sm}/^{144}\text{Nd} = 0.1967$ at present) for the samples at the time of formation (Table 9.2). The cycle 2 subalkaline basalts have $\epsilon_{\text{Nd}}(1884)$ ranging from +2.65 to +4.7, and these values are comparable to those obtained by Rohon *et al.* (1993; +3.7, +4.4) and Skulski *et al.* (1993; +2.7 to +5.2) for cycle 2 basalts in the central and northern portions of the Labrador Trough. Rohon *et al.* (1993) suggested that, among the various cycle 2 subalkaline lithologies, the ϵ_{Nd} of the Montagnais gabbroic rocks are lower than these of the Willbob basalts due to *in situ* contamination of the sills by country rock with highly negative ϵ_{Nd} ratios. The cycle 2 subalkaline data reported by Rohon *et al.* (1993) and Skulski *et al.* (1993), along with the results of this study, are plotted in Figure 9.10. The lowest ϵ_{Nd} ratio is that for an aphyric Montagnais gabbro (Rohon *et al.*, 1993), while the highest is from an aphyric Willbob basalt (Skulski *et al.*, 1993). The average

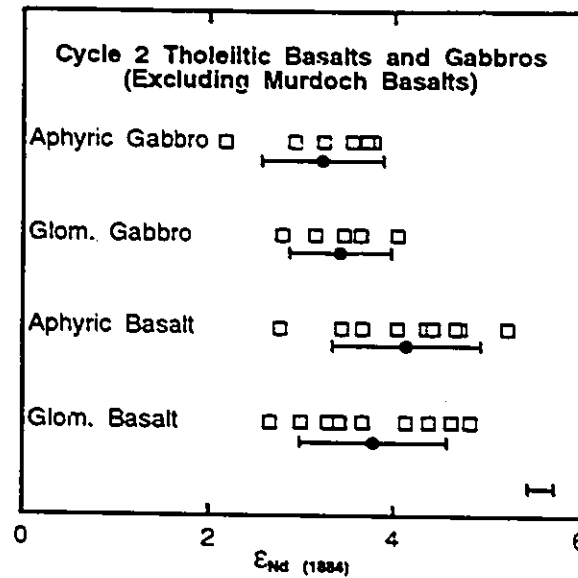


Figure 9.10. Variation in initial ϵ_{Nd} values for cycle 2 tholeiitic rocks including data from Rohon *et al.* (1993) and Skulski *et al.* (1993). Mean values and 2 standard deviations of the means are indicated. Average 2σ analytical uncertainty is indicated by error bar in the lower right corner of the figure.

ratios of the aphyric and glomeroporphyritic basalts, and glomeroporphyritic gabbros, are equivalent within two standard deviations of the means, but the average ratio in the aphyric gabbros lies outside of this bracket in comparison to the aphyric basalts. It is also apparent that the highest ϵ_{Nd} ratios of the gabbros are lower than the maximum values found in basalts. These features might be indicative of *in situ* contamination of intrusive magmas.

If the low ϵ_{Nd} ratio of the aphyric gabbro sample is disregarded, the average ratio of the aphyric samples is not significantly different than those of the other groups, and the lowest ratios from each grouping are comparable. Thus the data could also be interpreted as reflecting similar ranges in isotopic compositions among the samples, with one aphyric sill having undergone small degrees of *in situ* contamination. Whether *in situ* contamination was a widespread phenomenon in the case of the intrusions cannot be resolved without further data, but the similarity of the lowest ϵ_{Nd} ratios among sample groupings, excluding the low value from the aphyric sill, indicates that low or negative ϵ_{Nd} material has influenced the isotopic composition of both intrusive and extrusive rocks. Thus the origin of low ϵ_{Nd} ratios cannot exclusively lie in the *in situ* contamination of

intrusive magmas. This is consistent with the conclusion reached in Chapter 5 that *in situ* contamination of the Menihek sills did not greatly influence melt compositions. The fact that the ϵ_{Nd} ratios of cycle 2 basalts extend from +2.7 to +5.2 in itself suggests a mixed source for the melts, and this may reflect pre-extrusion contamination by sialic crust or source region heterogeneity.

The ϵ_{Nd} values of the low-Zr Murdoch basalts (+3.61, +4.18) fall within the range obtained from the upper Menihek/Willbob tholeiites. Equivalent data for the slightly alkaline basalts near the base of the formation are not available. The ϵ_{Nd} ratios of the lower Menihek basalts (+1.45, +1.95) fall within the range of the Nimish samples (-0.57, +3.27), and this is consistent with the proposed stratigraphic equivalency of the suites. The Le Fer basalt also has a relatively low ϵ_{Nd} ratio (+1.77) in comparison to cycle 2 tholeiites, but this value is comparable to that of +1.05 for a granophyric differentiate of a Le Fer gabbro sill reported by Rohon *et al.* (1993).

Figure 9.11 presents an isochron diagram for cycle 2 tholeiitic samples excluding the low-Zr Murdoch basalts, but including the data of Rohon *et al.* (1993) and Skulski *et al.* (1993). Although the data define a collinear trend with high correlation ($r = 0.96$), the similarity of sample compositions (except the glomerocryst plagioclase sample) produces a relatively large uncertainty in the slope of the isochron. The associated uncertainty is further increased by the lack of fit of the data in that not all points lie within error of the isochron. Such lack of fit is to be expected from the variability of ϵ_{Nd} values. The calculated age using all cycle 2 tholeiitic samples is 1971 ± 211 Ma, which is within error of both the 1884 Ma age of the Howse Lake sill, and the 2169 Ma age of the Le Fer granophyre. Similar dates, albeit with larger associated uncertainties, are obtained using only cycle 2 subalkaline basaltic samples, or using only Willbob basalts. Consequently, the Sm-Nd isotopic data fail to provide any further constraints on the age of the cycle 2 sequences.

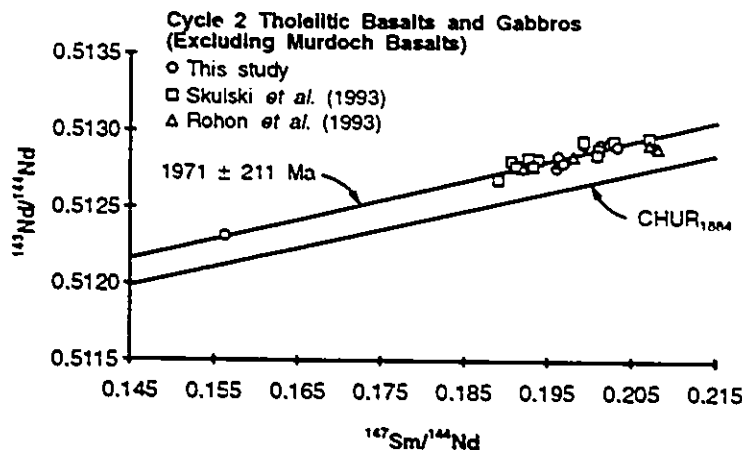


Figure 9.11. Sm-Nd isochron diagram for cycle 2 subalkaline rocks. The age defined by the isochron is 1971 ± 211 Ma. 2σ analytical uncertainties are less than the size of plot symbols.

9.3 Potential Source Reservoirs

The ϵ_{Sr} and ϵ_{Nd} of samples for which both sets of data are available are plotted in the anticorrelation diagram of Figure 9.12A. Although the data are limited, each of the samples appears to be the product of a depleted mantle source with respect to Sm-Nd and, for several of the samples, with respect to Rb-Sr as well. As noted earlier, however, there is great variability in the range of ϵ_{Sr} ratios. The ϵ_{Sr} ratios of cycle 2 tholeiitic samples, for example, range from -3 to +21.5, and if the data of Rohon *et al.* (1993) is included the range increases to -194 to +38. This variability is not reflected in the ϵ_{Nd} ratios of the suite (+2.1 to +5.2). The diversity of ϵ_{Sr} ratios indicates that the Rb-Sr isotopic system has not remained closed since crystallization, and it was suggested earlier that this may reflect post-crystallization mobility of Rb and/or Sr. Post-crystallization changes in Rb and Sr concentrations may have been related to metamorphic recrystallization, but sea-water infiltration (leading to spilitization and/or amygdale formation) has also been shown to give rise to anomalous ϵ_{Sr} values (cf. Faure, 1977; Andrew *et al.*, 1991). These processes may cause an increase in Rb/Sr, and hence $^{87}\text{Sr}/^{86}\text{Sr}$, of spilitized rocks while having little or no effect on Sm-Nd ratios. Typically, ϵ_{Sr} values are shifted to the right in anticorrelation diagrams, as illustrated in Figure 9.12A. Since all of the samples represent sequences

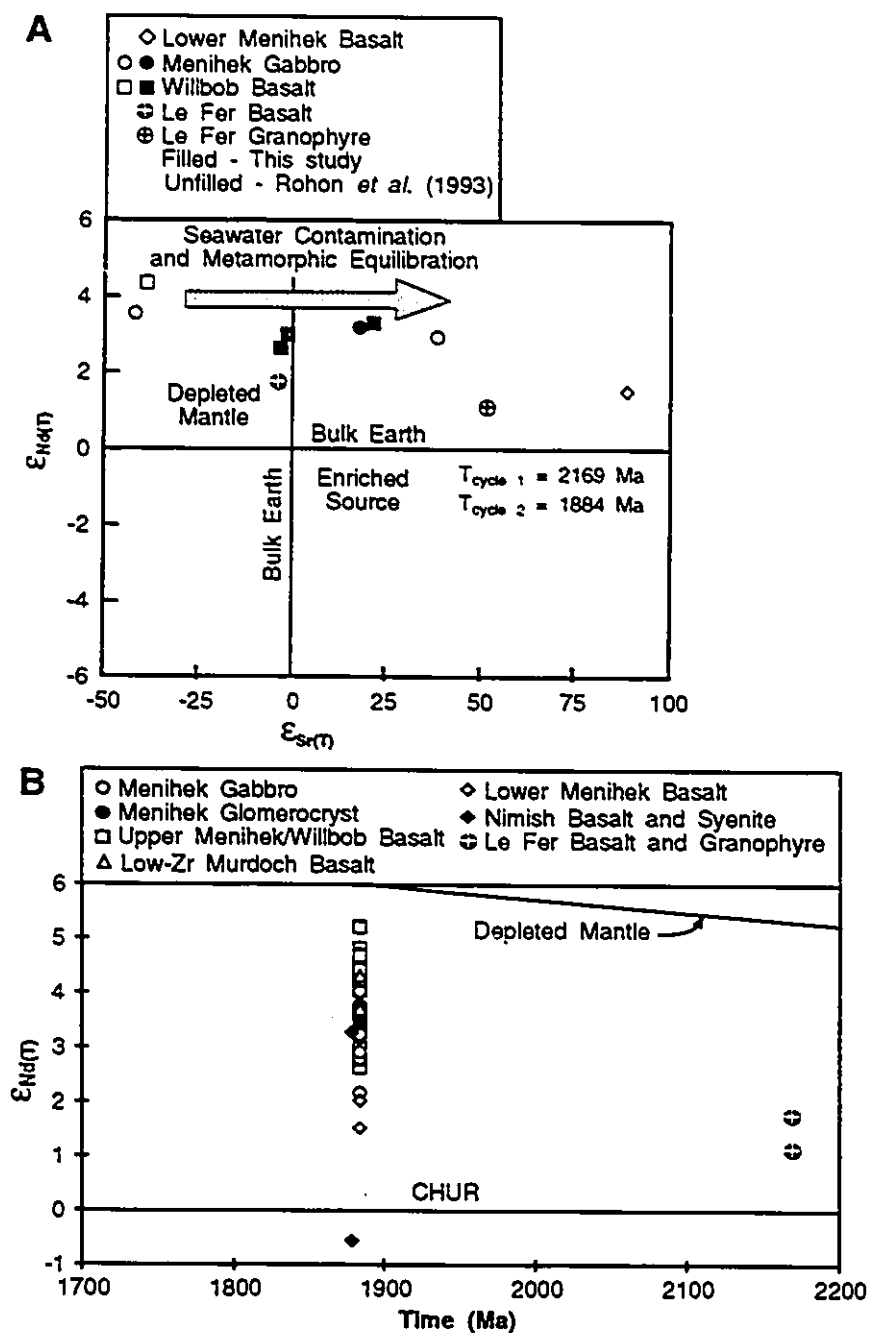


Figure 9.12. (A) Anticorrelation diagram illustrating the ϵ_{Sr} and ϵ_{Nd} isotopic compositions of samples from this study and from Rohon *et al.* (1993) at the time of formation ($T=1884\text{ Ma}$ for the cycle 2 suite and $T=2169\text{ Ma}$ for the cycle 1 Le Fer suite). Extreme ϵ_{Sr} values are excluded. (B) ϵ_{Nd} evolution diagram for samples from this study, Rohon *et al.* (1993) and Skulski *et al.* (1993). Depleted mantle curve is derived from the average isotopic composition of modern N-MORB.

formed in marine basins, and since all have been subject to alteration and metamorphism, it is probable that some degree of relative Rb enrichment has occurred. Consequently, it is doubtful that the Rb-Sr isotopic data provide valid information regarding the nature of source reservoirs for LTBS magmas.

The Sm-Nd isotopic variability of members of the LTBS relative to CHUR is apparent in the ϵ_{Nd} evolution diagram of Figure 9.12B. The majority of samples have positive $\epsilon_{Nd}(T)$ values, indicating derivation from isotopic reservoirs that were variably depleted in Nd in comparison to an undepleted CHUR source. In contrast, the negative ϵ_{Nd} value of the Nimish syenite indicates a contribution from a source enriched in Nd relative to CHUR. This indicates that sialic crust has contributed to the isotopic composition of the syenite.

On the basis of geochemical affinity, the LTBS suites can be subdivided into three principal groups: (1) cycle 1 and 2 tholeiites, (2) cycle 2 alkali basalts, and (3) cycle 2 subalkaline to mildly alkaline Murdoch basalts. The cycle 2 tholeiitic basalts exhibit trace element patterns comparable to modern T-MORB, and this suggests that they are derived primarily from a depleted asthenospheric mantle source with some contribution from less depleted or enriched material. In contrast, the cycle 2 alkaline rocks exhibit the trace element characteristics of alkali basalts derived predominantly from a less depleted or enriched source. This may represent (1) mantle material which has undergone a lesser degree of depletion than the source of the cycle 2 tholeiites, (2) primitive or primordial mantle, or (3) depleted mantle which subsequently underwent an enrichment event(s). For simplicity, this reservoir is referred to here as undepleted mantle in reference to the fact that it is less incompatible element depleted than the source of the tholeiitic rocks, and to distinguish it from incompatible element enriched continental crust. The third geochemical type (the Murdoch basalts) exhibit transitional compositions which suggest derivation from a mixed source. The incompatible element poor compositions of the Le Fer basalts indicate that a depleted mantle reservoir similar to the source of the cycle 2 tholeiites was present during cycle 1 infilling. The occurrence of alkaline volcanic rocks (Chakonipau Formation) in the lower part of the cycle 1 sequence may indicate that an undepleted mantle reservoir comparable to the source of the cycle 2 alkaline volcanics was also

present at this time. Thus there is evidence of two compositionally distinct mantle reservoirs during evolution of the Labrador Trough.

The presence of geochemically distinct mantle reservoirs during formation of the LTBS suggests that compositional differences among the igneous rocks may reflect variable sampling of the contrasting sources. In such a model, the tholeiitic rocks would represent partial melts derived predominantly from the depleted asthenospheric mantle reservoir but with minor contributions from the undepleted source. The alkaline rocks represent partial melts derived primarily from the undepleted source, but with minor contributions from the depleted reservoir. And the Murdoch basalts represent partial melts derived from variable mixtures of the two reservoirs, with undepleted material predominating in the high-Zr melts and depleted material predominating in the low-Zr melts. The mixing of heterogeneous sources may also explain the range in isotopic compositions found in the suites. The negative ϵ_{Nd} ratio of the Nimish syenite, however, requires a contribution from Archean crust, and it is possible that crustal material has also provided a component to other LTBS magmas.

A crustal contribution is also suggested by the occurrence of negative Nb anomalies in the normalized extended trace element patterns of some LTBS samples. A comparison of Nb/La in aphyric and glomeroporphyritic upper/Menihek Willbob basalts, for example, reveals a slight trend of decreasing ratios with increasing La content (Fig. 9.13). Since Nb is more incompatible than La in the minerals thought to have controlled melt evolution, the reverse would be expected if the variations were caused by crystal fractionation. But progressive crustal contamination of the melts would lead to decreasing Nb/La and result in negative Nb anomalies. The fact that the glomeroporphyritic basalts, which are generally more evolved than their aphyric counterparts, exhibit more pronounced Nb troughs might reflect longer residence time in crustal magma chambers.

9.4 Estimated Compositions of Source Reservoir Melt Components

The effect on the isotopic compositions of magmas derived from mixed source reservoirs has been evaluated following the method of Chauvel *et al.* (1987). If the melts

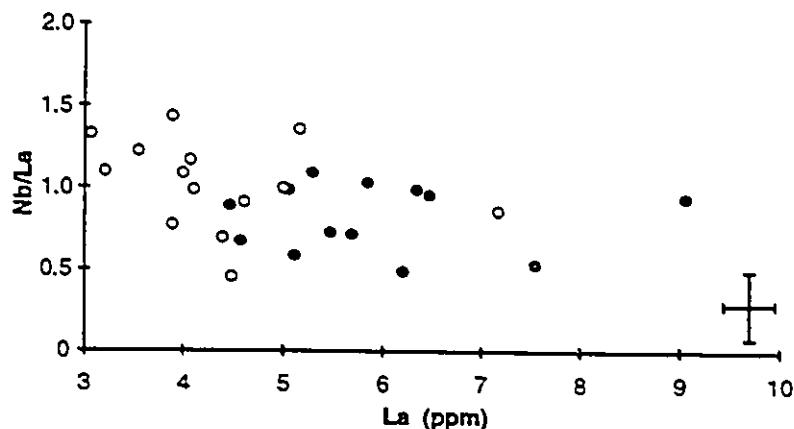


Figure 9.13. Nb/La versus La in the upper Menihek and Willbob basalts. The negative correlation suggests that continental crust having low Nb/La made increasing contributions to the melts. Average 2σ analytical uncertainty is illustrated by error bars.

are considered the products of a mantle source plus a contaminant (i.e., a second mantle source or continental crust), the two component mixing equation is given by:

$$X = \frac{C_M (\epsilon_{Nd}^S - \epsilon_{Nd}^M)}{C_C (\epsilon_{Nd}^C - \epsilon_{Nd}^M)}$$

where X is the ratio of mantle derived melt to contaminant, C_M and C_C are the Nd concentrations of mantle and contaminant endmembers, and ϵ_{Nd}^S , ϵ_{Nd}^M and ϵ_{Nd}^C are the ϵ_{Nd} ratios at the time of formation measured in the sample (S), and estimated in the mantle (M) and contaminant (C) sources. Such calculations serve only to give comparative estimates of relative contributions because of the difficulties in estimating the compositions of crustal and mantle melt components. In conjunction with geochemical data, however, the results are sufficient to identify the contributions to melt compositions made by different source reservoirs, and this provides a framework for interpretation of the petrogenetic evolution of the LTBS.

Depleted Mantle Component

The estimation of the composition of the melt component derived from the asthenospheric depleted mantle reservoir is hampered by the fact that none of the LTBS samples obtained in this study are likely to represent primary partial melts derived from

such a reservoir. The best available representative of primitive melt is the high-Mg Willbob basalt sampled by Beaudoin and Laurent (1989). As discussed earlier, even this sample is unlikely to represent a depleted mantle partial melt since it exhibits a slight Lu to Nb enrichment trend in normalized extended trace element plots. Nevertheless, it provides the only available estimate of relatively primitive melt compositions. With two exceptions, the trace element concentrations reported by Beaudoin and Laurent (1989) have been used in mixing calculations. The exceptions are Rb, which was not reported, and Nb whose reported abundance is anomalously high in comparison to adjacent elements in a normalized extended trace element plot, and in comparison to concentrations in the more primitive of the Willbob samples obtained in this study. Because the concentrations of adjacent elements (Ba, K and La) are comparable to those of average N-MORB, Rb and Nb have been adjusted to N-MORB values. The resultant extended trace element pattern is illustrated in Figure 9.14.

The isotopic composition of the depleted mantle reservoir at the time of derivation of the cycle 2 tholeiitic rocks can be estimated in two ways. First, the maximum ϵ_{Nd} ratio obtained from the cycle 2 sequence can be assumed to represent uncontaminated, depleted mantle derived melt. Samples with lower ϵ_{Nd} ratios can then be considered to represent melts derived from a depleted mantle source with a contribution from less depleted or enriched material. The second method involves the backward extrapolation of the $^{143}Nd/^{144}Nd$ ratios of modern N-MORB, which are thought to be derived from a depleted mantle reservoir.

The maximum ϵ_{Nd} ratio measured in cycle 2 tholeiitic rocks is the +5.2 value reported by Skulski *et al.* (1993) from the Hellancourt basalts of the northern Trough, while the maximum ratio obtained in this study is +4.7 for a presumably correlative Willbob basalt. The +5.2 value is comparable to the highest ratios obtained from basaltic sequences in other 1.9 Ga Trans-Hudson orogenic segments in North America. As argued by Chauvel *et al.*, (1987), the consistent occurrence of maximum ϵ_{Nd} ratios of approximately +5 in widely separated, mantle derived, basaltic sequences is best interpreted to reflect the presence of a uniformly depleted asthenospheric mantle at ~ 1.9 Ga. If the maximum ϵ_{Nd} ratio of volcanic rocks from the circum-Superior foldbelts is used

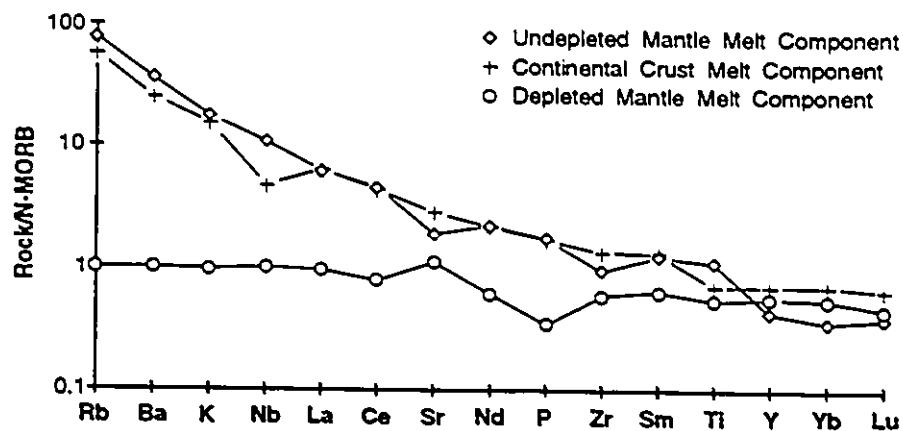


Figure 9.14. Normalized extended trace element plots for the estimated compositions of melt components derived from Archean continental crust, depleted asthenospheric mantle, and undepleted mantle. Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba which is from Saunders and Tarney (1984).

as an estimate of isotopic composition, the ϵ_{Nd} ratio of the depleted mantle reservoir was +5.3 at 1.9 Ga (as indicated by basalt from the Fox River belt; Chauvel, *et al.*, 1987).

Following the second option, the calculated ϵ_{Nd} ratio of a depleted mantle reservoir at 1884 Ma based on the average isotopic composition of modern N-MORB ($\epsilon_{Nd} = +10$, $^{147}\text{Sm}/^{144}\text{Nd} = 0.21$) is +6.4. Although the two estimates are similar, the discrepancy may indicate that none of the circum-Superior volcanic sequences are derived exclusively from a depleted mantle source. For example, the Hellancourt basalt from which the ϵ_{Nd} value of +5.2 was obtained by Skulski *et al.* (1993) is compositionally indistinguishable from the Willbob basalts sampled in this study, all of which are thought to reflect partial melting of depleted mantle with minor contributions from a less depleted or enriched source. Because of the potentially mixed origin for even the most isotopically depleted volcanic sample, and because the maximum contributions from other reservoirs are of interest, the estimate using extrapolation of modern N-MORB ratios is preferred. Consequently, an ϵ_{Nd} value of +6 was chosen for the depleted mantle component. This value is slightly lower than extrapolated average N-MORB ratio at 1884 Ma, and slightly higher than this ratio at 2169 Ma, but the resultant variations in the calculated proportions of different reservoir components are negligible in comparison to the variations introduced by uncertainty in melt component compositions. The Nd content of the estimated depleted mantle

component is taken from the concentration in the high-Mg Willbob flow (4.5 ppm; Beaudoin and Laurent, 1989).

Undepleted Mantle Component:

The nature and composition of the undepleted mantle source to the LTBS alkaline rocks is poorly constrained, and its spatial relationship to the depleted asthenospheric mantle reservoir is unknown. One possibility is that the undepleted source is subcontinental or lithospheric mantle which was isolated from asthenospheric mantle at the time of formation of the Archean continental crust (cf., Carlson, 1984; Leeman and Fitton, 1989). Various workers have suggested that compositional differences between lithospheric and asthenospheric mantle may have been enhanced by enrichment of the former through metasomatism and veining resulting from the passage of incompatible element and CO₂ enriched fluids, introduction of incompatible element rich hydrous fluids derived from subducted oceanic crust, and other mechanisms (cf., Hawkesworth *et al.*, 1984; Sun and McDonough, 1989). Bell and Blenkinsop (1987) examined the isotopic characteristics of a suite of carbonatites emplaced within the Canadian shield between 2700 Ma and 110 Ma, and found evidence for such a subcontinental mantle source below the Superior craton. Their data show that the source of the carbonatite magmas was a LIL element depleted mantle reservoir which has a mean age of approximately 2900 Ma. These authors concluded that the lithospheric mantle, although depleted relative to the bulk earth composition, is less depleted than the source of present day N-type MORB and therefore represents a distinct mantle reservoir.

A mantle plume rising below the Early Proterozoic margin of the Superior craton is a second possible configuration for the undepleted mantle source. Such plumes have been postulated as the source of the incompatible element enriched tholeiites and alkali basalts in both oceanic island and continental settings, as well as the source of the voluminous outpourings of flood basalts which often accompany continental break-up (cf., White and McKenzie, 1989). The enriched nature of postulated mantle plumes has been attributed to derivation from primordial lower mantle, derivation from delaminated lithospheric mantle carried into the asthenosphere by mantle convection, or derivation from recycled oceanic crust (cf., McKenzie and O'Nions, 1983; Creager and Jordan, 1984). A third possibility is

that the undepleted source consists of localized streaks or blobs of incompatible element enriched material in a matrix of uniformly depleted asthenospheric mantle (cf., Sleep, 1984; Fitton and Dunlop, 1985; Allègre and Turcotte, 1986).

Whatever the configuration of the undepleted mantle source to the LTBS, it would clearly have a less depleted isotopic composition than the asthenospheric mantle component. Two endmember isotopic compositions for undepleted melt components reservoirs have been used in the mixing calculations discussed below. For the first endmember, the undepleted source is assumed to have a primordial or chondritic isotopic composition (i.e., $\epsilon_{Nd} = 0$). The ϵ_{Nd} ratio of the second endmember has been estimated at +2 at 1884 Ma. This is based on the ϵ_{Nd} ratios of the lower Menihek Formation (+1.45, +1.95), and the approximately coeval Povungnituk Alkaline Suite (+1.83 to +2.40; Gaonac'h *et al.*, 1992) and Cargill carbonatite (+1.6; Bell and Blenkinsop, 1987). These ratios were obtained from highly alkaline rocks whose compositions presumably reflect a predominant contribution from the undepleted source.

Assuming the LTBS alkali basalts represent partial melts derived exclusively from an undepleted source, the composition of the undepleted mantle melt component could be estimated from primitive members of the lower Menihek or Nimish suites. As noted earlier, however, it is unlikely that any of these basalts represent primary partial melts. In addition, the most primitive samples from the two suites are Nimish basalts which exhibit the P and Zr depletions characteristic of the tholeiitic rocks, and thus probably represent melts derived from a mixed source. Consequently, the composition of the melt component of the undepleted mantle reservoir has been estimated using the average of two slightly more evolved samples, and making two assumptions about the nature of crystallization processes during melt evolution. The two samples are the most primitive (based on Mg#) of the Nimish basalts which do not show the P and Zr depletions typical of the tholeiitic rocks. These exhibit trace element patterns in which elemental abundances increase fairly consistently from Lu to Rb. The two assumptions made regarding melt evolution are: (1) the compositions of the two Nimish samples reflect 50% fractional crystallization of the primary partial melt, and (2) all of the elements plotted in the extended trace element diagram of Figure 9.14 were completely incompatible during melt evolution. Although

neither assumption is likely to be completely correct, they allow a crude estimation of the composition of the melt component from the undepleted mantle reservoir (Fig. 9.14). The estimated Nd content of the undepleted mantle component is 16 ppm.

Continental Crust Component

The estimates of the melt components contributed by the depleted and undepleted mantle reservoirs are based on inferred liquid compositions rather than on inferred mantle source compositions. In contrast, the estimates of the melt components of the crustal reservoir are based on estimated reservoir compositions (i.e., rock compositions) because the trace elements of interest will be preferentially incorporated into the liquid fraction during partial melting. In terms of these trace elements, partial melting will thus be equivalent to wholesale assimilation. Consequently, the incompatible trace element composition of the whole reservoir is a valid approximation of the melt component even if complete melting does not occur.

Two estimates of the isotopic composition of the melt component contributed by continental crust have been used in the mixing calculations. The data of Skulski *et al.* (1993) indicate an Archean model age for one sample of cycle 2 turbidites in the northern Trough. This sample has an ϵ_{Nd} value of -7.1 at 1875 Ma, and a Nd concentration of 22.4 ppm. Although the turbidites are not composed exclusively of Archean crustal detritus, the sample may offer a better approximation of the composition of local Archean basement in the Labrador Trough than broader based estimates. Consequently, the data for this sample have been used as one of the crustal component endmembers in the mixing calculations.

Chauvel *et al.* (1987) estimated the ϵ_{Nd} ratio of 1.8-1.9 Ga Archean crust at -8 to -12, with an average Nd concentration of 32 ppm. The ϵ_{Nd} ratio of -12 has been used for the second crustal endmember to establish the widest possible range of isotopic compositions. The average crustal Nd concentration of 16 ppm proposed by Taylor and McLennan (1985) is preferred to the estimate of Chauvel *et al.* (1987). This is because the smaller concentration will increase the estimated proportion of crustal contamination, and it is the maximum possible contribution that is of interest. The average trace element composition of the crustal component used in subsequent discussions of the effects of

crustal contamination on LTBS melts is that proposed by Taylor and McLennan (1985). The normalized extended trace element pattern of this estimate is illustrated in Figure 9.14.

9.5 LTBS Magma Sources

Upper Menihek and Willbob Magmas

The upper Menihek and Willbob basalts exhibit geochemical characteristics which suggest they represent partial melts derived from a depleted mantle source with some contribution from a less depleted or enriched reservoir. Table 9.3 provides results of mixing calculations for the cycle 2 tholeiites in which two models are considered. In the first, the parental melts to the tholeiitic basalts are considered to be derived from depleted asthenospheric mantle with a contribution from Archean crust (Table 9.3A). The depleted mantle component is assigned an ϵ_{Nd} ratio of +6 and a Nd concentrations of 4.5 ppm. The crustal component is represented by two endmembers having ϵ_{Nd} values of -12 and -7, and Nd concentrations of 16 and 22 ppm. Using the first crustal estimate, the ϵ_{Nd} ratios of the samples require the additions of between 2% and 6% of the crustal component to depleted mantle magmas. Similar proportions of the second endmember are required to produce the observed ϵ_{Nd} ratios, despite the less enriched isotopic composition of this crustal estimate. This is due to the counteracting effect of the higher Nd concentration in the second endmember. To produce the maximum and minimum ϵ_{Nd} ratios from comagmatic rocks reported by Skulski *et al.* (1993) and Rohon *et al.* (1993) requires between 1% and 7% of the first crustal endmember, and between 1% and 8% of the second crustal endmember. Assuming that the actual composition of the crustal component is bracketed by the two estimates, the total range of isotopic compositions of the cycle 2 tholeiitic rocks can be accommodated by the addition of 1-8% crustal material.

Calculated Nd concentrations for the depleted mantle-crust mixtures are also reported in Table 9.3A. These reflect the contributions from both depleted mantle and crustal melt components to the primary melts. In most cases, the observed sample concentrations are roughly double the calculated concentrations. This relationship is

Table 9.3. Calculated proportions of Archean crust (% AC) or undepleted mantle (% UM) components added to upper Menihok/Willbob depleted mantle magmas to give observed ϵ_{Nd} ratios. Calculated Nd concentrations in ppm (Nd^c) resulting from mixing are indicated. ϵ_{Nd} and Nd^m are the measured isotopic ratios and Nd concentrations (ppm), respectively, in the samples. Comagmatic Labrador Trough samples showing maximum¹ (Skulski *et al.*, 1993) and minimum² (Rohon *et al.*, 1993) ϵ_{Nd} ratios are included for comparison with the results of this study. The depleted mantle component is assigned an ϵ_{Nd} ratio of +6 and a Nd concentration of 4.5 ppm. All ϵ_{Nd} values are calculated as initial ratios at 1884 Ma.

(A) Depleted Mantle + Archean Crust				Archean Crust			
Sample	Description	ϵ_{Nd}	Nd^m	$\epsilon_{Nd} = -12, Nd = 16 \text{ ppm}$		$\epsilon_{Nd} = -7, Nd = 22 \text{ ppm}$	
				% AC	Nd^c	% AC	Nd^c
F8735-6	Willbob Glom. Basalt	2.7	13.85	6.0	5.2	6.6	5.7
F8722-15	Willbob Glom. Basalt	3.0	10.39	5.3	5.1	5.8	5.5
F8733-17	Willbob Glom. Basalt	3.3	11.04	4.7	5.0	5.1	5.4
F90-99	Willbob Aphyric Basalt	4.7	5.07	2.1	4.7	2.2	4.9
F90-87	Menihok Aphyric Basalt	4.0	10.08	3.3	4.9	3.5	5.1
F90-90	Menihok Aphyric Basalt	4.4	8.33	2.6	4.8	2.7	5.0
F20-7B	Menihok Aphyric Gabbro	3.2	8.69	5.0	5.1	5.4	5.4
F26-1A(2)	Menihok Glom. Gabbro Matrix	3.6	10.06	4.1	5.0	4.3	5.3
1165-86 ¹	Hellancourt Aphyric Basalt	5.2	10.30	1.3	4.6	1.3	4.7
111D ²	Montagnais (Doublet) Gabbro	2.2	3.22	7.0	5.3	7.9	5.9

(B) Depleted Mantle + Undepleted Mantle				Undepleted Mantle			
Sample	Description	ϵ_{Nd}	Nd^m	$\epsilon_{Nd} = 0, Nd = 16 \text{ ppm}$		$\epsilon_{Nd} = +2, Nd = 16 \text{ ppm}$	
				% UM	Nd^c	% UM	Nd^c
F8735-6	Willbob Glom. Basalt	2.7	13.85	26.2	7.5	59.1	11.3
F8722-15	Willbob Glom. Basalt	3.0	10.39	21.9	7.0	45.5	9.7
F8733-17	Willbob Glom. Basalt	3.3	11.04	18.7	6.7	36.9	8.7
F90-99	Willbob Aphyric Basalt	4.7	5.07	7.2	5.3	11.8	5.9
F90-87	Menihok Aphyric Basalt	4.0	10.08	12.0	5.9	21.2	6.9
F90-90	Menihok Aphyric Basalt	4.4	8.33	9.1	5.5	15.5	6.3
F20-7B	Menihok Aphyric Gabbro Chill	3.2	8.69	20.2	6.8	40.7	9.2
F26-1A(2)	Menihok Glom. Gabbro Matrix	3.6	10.06	15.4	6.3	28.9	7.8
1165-86 ¹	Hellancourt Aphyric Basalt	5.2	10.30	4.1	5.0	6.4	5.2
111D ²	Montagnais (Doublet) Gabbro	2.2	3.22	33.1	8.3	85.9	14.4

consistent with the relative trace element abundances of the high-Mg Willbob flow of Beaudoin and Laurent (1989) and the Willbob samples obtained in this study (see Fig 9.2E). For example, assuming the elevated abundances of Sm, Ti, Y, Yb and Lu (which will be least affected by addition of a crustal melt component) in the evolved Willbob

basalts primarily reflect fractional crystallization, and assuming these elements are completely incompatible, concentration differences represent 35-55% crystallization. The degree of crystallization represented by element concentrations in the aphyric basalts lies in the lower part of this range, while the glomeroporphyritic basalts generally represent larger degrees of fractionation. Crystallization of these quantities of liquid will result in calculated Nd concentrations that are consistent with observed values except in the case of the Montagnais gabbro sample analyzed by Rohon *et al.* (1993) and reported in Table 9.3. The petrographic details of this sample are not known, but the low Sm and Nd concentrations are consistent with a cumulate origin.

Using the estimated trace element abundances in the depleted mantle and crustal melt components illustrated in Figure 9.14, the trace element patterns of derivative magmas can be compared with observed profiles. The average trace element abundances of Willbob aphyric and glomeroporphyritic basalt are plotted in Figure 9.15A, along with mixing lines which have been constructed assuming complete conservation of the incompatible elements during 50% fractional crystallization of the initial magma. It is apparent that the overall trends of the mixing lines provide fairly good approximations of the average patterns of the aphyric and glomeroporphyritic basalts. The greatest degree of non-conformity is found in Sr, but this discrepancy probably reflects Sr loss in the basalts through plagioclase fractionation and/or post-crystallization mobility. It is also apparent that the 50% crystallization factor applied to the mixing lines is appropriate for the average glomeroporphyritic basalt, but overestimates the degree of evolution of the aphyric basalts. This is consistent with the generally less evolved nature of the latter.

Despite the general conformity of mixing lines and the average trace element patterns of the basalts, there are principal two inconsistencies in the comparisons. The first is that the negative Nb anomaly developed through 1-10% contamination by crustal material is less than the average deficiency in the glomeroporphyritic basalts. Because there is a fairly consistent difference in the size of the anomaly in the glomeroporphyritic and aphyric basalts, it seems unlikely that the discrepancy between measured and estimated Nb abundances is an artefact of analytical imprecision unless the measured values for all of the samples are uniformly too low. Assuming the contrast in Nb anomalies is real, the low concentrations of this element are unlikely to reflect crystal fractionation

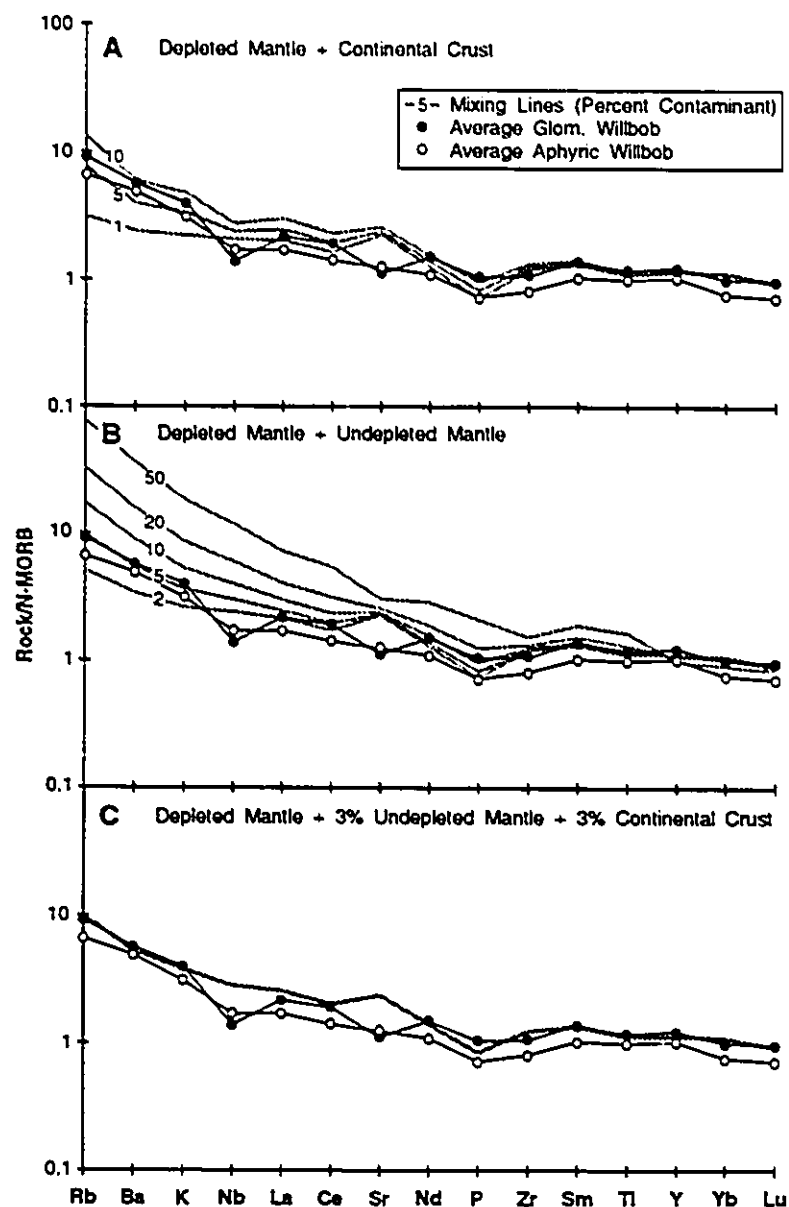


Figure 9.15. Comparison of the normalized extended trace element patterns of average cycle 2 aphyric and glomeroporphyritic basalts with patterns produced by mixing of different reservoir melt components. The mixing lines represent hybrid primary melts which have undergone 50% fractional crystallization. Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba, which is from Saunders and Tarney (1984).

because of its incompatibility. Because larger proportions of crustal assimilation are inconsistent with the isotopic data, and since the only other likely contaminant (undepleted mantle) does not exhibit a negative Nb anomaly, the discrepancy must reflect

overestimation of the Nb content of the continental crust or overestimation of the Nb content of the depleted mantle.

The second inconsistency lies in the degree of enrichment in Rb, Ba and K observed in the basalts. The abundances of the other elements can be attained by additions of 1-5% crustal material, but the concentrations of Rb, Ba and K require 5-10% contamination. The ϵ_{Nd} ratios of glomeroporphyritic basaltic samples examined in this study represent an average addition of 5.4% crustal material using the first crustal endmember ($\epsilon_{Nd} = -12$), and 5.8% using the second ($\epsilon_{Nd} = -7$). The isotopic compositions of the aphyric basaltic samples represent an average addition of 2.7% and 2.8% of the two crustal estimates. If the data of Skulski *et al.* (1993) and Rohon *et al.* (1993) are included, the average crustal contribution to glomeroporphyritic magmas drops to approximately 4%, while the average addition to the aphyric magmas remains at approximately 3%. All of these values are probably insufficient to explain the elevated Rb, Ba and K abundances in the basalts. This discrepancy may indicate that: (1) the concentrations of these elements in the basalts have increased during alteration or metamorphism, (2) the concentrations of these elements are underestimated in the crustal component, or (3) material from both continental crust and undepleted mantle was added to the depleted mantle melts.

In the second mixing model, contributions of undepleted mantle material to depleted mantle magmas is considered to have controlled the composition of the parental melts of the cycle 2 suite. The undepleted mantle melt component is assigned a Nd concentration of 16 ppm, and ϵ_{Nd} ratios of 0 and +2. It is apparent from Table 9.3B that a large melt component derived from the undepleted mantle reservoir must be added to the depleted mantle magmas to produce the observed ϵ_{Nd} ratios. For example, assuming the isotopic composition of the undepleted mantle reservoir lies between the two endmember estimates, the glomeroporphyritic basalts analyzed in this study would require additions of between 18% and 59% of the undepleted melt component to depleted mantle partial melts. The isotopically least depleted cycle 2 tholeiitic sample (111D, from Rohon *et al.*, 1993) requires even greater proportions of undepleted material. All of these estimates would increase further if the ϵ_{Nd} ratio of the undepleted source was greater than +2, as is suggested by the compositions of the Nimish alkaline rocks (see below). Clearly the

necessary proportions of the undepleted mantle component will give rise to derivative basalts far richer in incompatible elements than the upper Menihok/Willbob flows. This is apparent in the strong Lu through Rb enrichment trends of the mixing lines illustrated in Figure 9.15B. A comparison of these patterns with the trace element abundances in the average aphyric and glomeroporphyritic indicates that the enrichment trends of the latter could be produced by minor additions (~5%) of the undepleted mantle melt component. Such small quantities of the undepleted component, however, will have little effect on the isotopic compositions of resultant magmas. Consequently, the isotopic composition of the cycle 2 samples cannot be explained by fusion of a mixed mantle source.

Figure 9.15C illustrates the trace element pattern resulting from minor additions of both crustal and undepleted mantle melt components to depleted mantle primary melts. The mixing line represents a magma derived from additions of 2% of the undepleted mantle component and 4% of the crustal component to depleted mantle partial melts. The trace element abundances of the mixing line have again been adjusted to reflect 50% fractional crystallization. The agreement in trace element contents between the hypothetical hybrid magma and the average glomeroporphyritic basalt is quite good, including the relative abundances of Rb, B and K. The calculated ϵ_{Nd} ratio of the hybrid melt is +3.7, which is equivalent to the average ratio among all available analyses of cycle 2 glomeroporphyritic basalts. The calculated initial Nd content is 5.2 ppm, which is consistent with measured concentrations assuming primary melts evolved through approximately 50% crystal fractionation. The ϵ_{Nd} ratio is very susceptible to small changes in the proportion of crustal contaminant, but the effects of changes in the proportion of crustal material on incompatible trace element abundances are relatively minor, and can be offset by adjusting the contributions from the undepleted mantle source. Thus it is possible for minor variations in the proportion of crustal material to produce the observed range in isotopic composition of the basalts, while minor variations in the proportions of both crustal and undepleted mantle components control the incompatible element enrichment trends. Such variations are entirely consistent with the prolonged and widespread nature of cycle 2 magmatism. Consequently, the entire range of isotopic compositions and

incompatible trace element patterns of the cycle 2 suite can be accommodated by minor changes in the proportions of melt contributions from the three reservoirs.

The foregoing conclusion relies on the validity of compositional estimates for the various melt components, and thus it is worthwhile to consider the possible effects of errors in these estimates. Because of the strong contrasts in isotopic composition, errors in the compositional estimates for the depleted mantle and continental crust components will have the greatest impact on the interpretations. For example, if the Rb, Ba and K content of average Archean crust is greater than estimated, the addition of undepleted mantle material might not be required to explain the elevated concentrations of these elements in the cycle 2 tholeiites. There is no way to evaluate this possibility, but some aphyric basalts require such small quantities of crustal contamination to produce their isotopic compositions that it is difficult to explain their elevated Rb, Ba and K abundances without invoking undepleted mantle. Figure 9.16A presents the normalized extended trace element pattern for one such basalt (sample F90-99). The ϵ_{Nd} ratio of this sample is consistent with approximately 2% crustal contamination (Table 9.3), yet its Ba and K contents are more compatible with at least 10% contamination. In addition, the sample exhibits, to a smaller degree, the heavy REE depletions that are characteristic of the cycle 2 alkali basalts. Both of these features support a contribution from an undepleted reservoir. Further support derives from the fact that samples requiring larger crustal components do not necessarily exhibit more enriched trace element patterns than samples requiring smaller degrees of contamination. Such a relationship is illustrated in Figure 9.16B for samples F90-99 and F90-87, which are both aphyric cycle 2 basalts. Despite its less depleted ϵ_{Nd} ratio, sample F90-87 exhibits a smaller degree of Rb-Nb enrichment than F90-99. Since it is unlikely that this relationship reflects greater compatibility of these elements during the more extensive evolution of the F90-87 magma, a difference in the proportion of the undepleted mantle melt component is the probable source of the discrepancy.

The inaccurate representation of the Archean crustal component may affect the conclusions discussed above in two additional ways. Since the Nd concentration of the contaminant (crust or undepleted mantle) is large in comparison to that of the depleted melt component, errors in this estimate will have a strong influence on the calculated

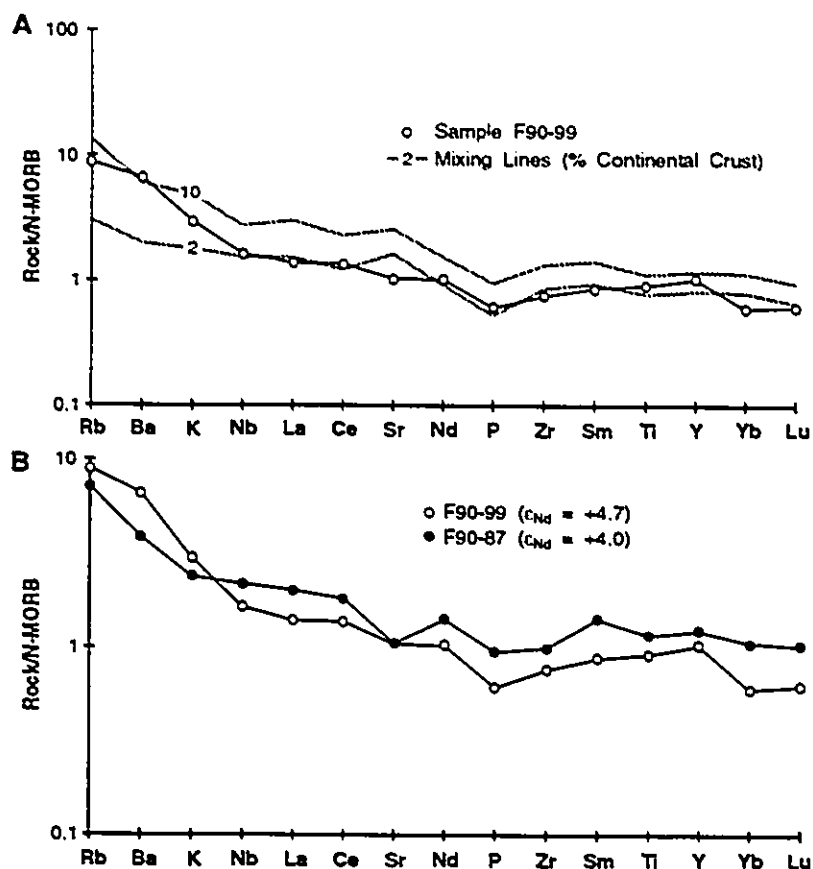


Figure 9.16. (A) Comparison of the normalized extended trace element pattern of cycle 2 aphyric basalt sample F90-99 with mixing lines representing addition of continental crust to depleted mantle partial melts. The mixing lines represent hybrid primary melts which have undergone 50% fractional crystallization. (B) Comparison of the normalized extended trace element patterns of cycle 2 aphyric basalt samples F90-99 and F90-87. Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba, which is from Saunders and Tarney (1984).

proportions of the different melt components. Because the crustal Nd content is only marginally larger than the Nd content of some of the basaltic samples, this value is more likely to have been underestimated than overestimated. If the true Nd content of the crust is higher, the degree of crustal contamination required to produce observed isotopic compositions are reduced. This strengthens the argument for a contribution from undepleted mantle. A second possible error lies in the estimate of the ϵ_{Nd} ratio of Archean crust. ϵ_{Nd} ratios of -7 to -8.7 obtained from the 1.9 Ga Eskimo Formation basalts in the Belcher Islands fold belt by Chauvel *et al.* (1987) were interpreted as indicative of crustal

contamination of depleted mantle magmas. If this interpretation is correct, the ϵ_{Nd} ratio of the Superior crust at 1.9 Ga could not have been greater than -8.7, and this is also probably true for the sialic basement to the Labrador Trough. It is possible, however, that the actual ϵ_{Nd} ratio of the crustal component was lower than -12. If this was the case, then the calculated proportions of crustal material added to the cycle 2 tholeiitic magmas are too high, and a contribution from an undepleted reservoir is again required to explain the incompatible element enrichment of the samples.

The conclusions regarding the source of the cycle 2 tholeiitic magmas may also be affected by errors in the estimated composition of the depleted mantle melt component. It is unlikely that this component was more evolved than the estimated composition since this was based on a sample which cannot be considered primitive in comparison to the picritic primary melts in other volcanic successions. If the true depleted mantle melt component was less evolved than the estimated composition, such that the concentrations of all of the incompatible trace elements were lower, the proportions of crustal or undepleted mantle material required to produce the observed isotopic composition are reduced. This follows from the fact that a less evolved depleted mantle component would have a lower initial Nd concentration, and hence the contrast in Nd content with the potential contaminants would be enhanced. If the contaminant is continental crust, the foregoing conclusions are emphasized because the resultant enrichment trends (particularly with respect to Rb, Ba and K) of the hybrid melts would be reduced, and therefore could not produce the enrichments observed in the basalts. This effect is enhanced if the depleted mantle melt component was not only less evolved but LIL element impoverished as well, as is suggested for the depleted mantle partial melts of the Le Fer basalts in the next section. If the contaminant is the undepleted mantle component, the observed enrichment trends are compatible with additions of about 5% of this material. For such small quantities of the undepleted mantle component to result in the isotopic compositions of the samples, the Nd concentration of the depleted mantle melts would have to have been about 1 ppm. But the Nd concentration in the resultant magmas would be too low (<2 ppm) to explain observed concentrations without unreasonable degrees of crystal fractionation (e.g., >80%).

A lower ϵ_{Nd} ratio in the depleted mantle component leads to the same conclusions if the contaminant is continental crust because the required proportion of the latter material is again reduced. The effect is the same if the undepleted mantle is the source of the isotopic variations, but the required proportions are still far greater than the 5% value suggested by the trace element patterns. A higher ϵ_{Nd} ratio in the depleted mantle reservoir, however, would require greater proportions of crustal material to produce observed isotopic compositions, and this might obviate the need for a contribution from the undepleted mantle source. But the degree of incompatible element enrichment of some samples with relatively high ϵ_{Nd} ratios (e.g., F90-99) requires >10% crustal contamination, and this would be possible only if the ϵ_{Nd} ratio of the depleted mantle component was very high. For the parental melt of sample F90-99, 10% crustal contamination requires that the initial ϵ_{Nd} ratio of the melt was approximately +12. Such a highly depleted source is unlikely given that extrapolation to 1.9 Ga of the ϵ_{Nd} ratios of the most depleted of modern N-MORB results in values of less than +10.

Errors in the composition of the undepleted mantle melt component will have little effect on the interpretation of the origin of the cycle 2 tholeiitic melts. This is largely due to the constraints on isotopic composition of undepleted mantle melts. The minimum contributions of this source occur when the melt component is assigned the ϵ_{Nd} ratio of primordial mantle. But even this value necessitates quite large contributions from the undepleted source to give the measured ϵ_{Nd} ratios. The observed isotopic and geochemical characteristics could only be produced by additions of an undepleted mantle component if the Nd concentration of the latter was significantly higher (to reduce the proportions necessary to produce the ϵ_{Nd} ratios of the cycle 2 tholeiites) and the enrichment trend was significantly flatter (to reduce the resultant enrichment trend in the cycle 2 tholeiites). Such a composition, however, is inconsistent with the geochemistries of the cycle 2 alkaline rocks. For example, a Nd concentration of 50 ppm in the undepleted mantle melt component is necessary to reduce the proportion of an undepleted mantle contaminant to 5-7% for the glomeroporphyritic samples. Since this concentration exceeds that of most of the alkali basalts, such a Nd content is unlikely. Consequently, it is concluded that the

isotopic and trace element characteristics of the upper Menihok/Willbob basalts and comagmatic intrusive rocks reflect the partial fusion of a depleted mantle reservoir with minor contributions from an undepleted source, and subsequent contamination of the resultant melts by continental crust. The extent of crustal contamination was also relatively minor; the maximum crustal contribution was about 8%, and the average degree of contamination was 3–4%.

Le Fer Magmas

The compositional similarity of the Le Fer basalts and cycle 2 tholeiites was interpreted to reflect derivation from a comparable mantle source. The Le Fer basalts, however, exhibit generally lower incompatible element contents and slightly flatter extended trace element patterns. Assuming the Le Fer and cycle 2 tholeiitic magmas were derived primarily from the same undepleted mantle reservoir, these features would seem to indicate that undepleted or enriched sources made smaller contributions to the former. In contrast, the single sample of Le Fer basalt for which Sm-Nd data are available exhibits an initial (2169 Ma) ϵ_{Nd} ratio of +1.8 (equivalent to +1.9 at 1884 Ma), which is lower than the values obtained from the upper Menihok/Willbob basalts (Table 9.2). A similarly low ϵ_{Nd} ratio (+1.2) was obtained by Rohon *et al.* (1993) from a granophyric differentiate of a Le Fer gabbro sill. These data suggest that the Le Fer magmas contained greater proportions of undepleted mantle or crustal melt components than those of the cycle 2 tholeiitic suites.

The proportions of the melt components of undepleted mantle and continental crustal required to produce the observed ϵ_{Nd} ratio of the Le Fer samples have been calculated using the compositional estimates outlined earlier (Table 9.4). Since the difference in age of cycle 1 and 2 magmatism is insufficient to have significant effects on the estimated Sm-Nd isotopic compositions of the source reservoirs, the same values have been used. The ϵ_{Nd} ratios of the Le Fer samples require addition of 8–9% of the first crustal component endmember ($\epsilon_{Nd} = -12$; Table 9.4A), and these values rise to 9–11% for the second crustal estimate ($\epsilon_{Nd} = -7$). If the isotopic ratios reflect additions of the undepleted mantle melt component, contributions of 40–54% are required from the primordial mantle

Table 9.4. Calculated proportions of Archean crust (% AC) or undepleted mantle (% UM) components added to Le Fer depleted mantle magmas to give observed ϵ_{Nd} ratios (Le Fer granophyre from Rohon *et al.*, 1993). Calculated Nd concentrations in ppm (Nd^c) resulting from mixing are indicated. ϵ_{Nd} and Nd^m are the measured isotopic ratios and Nd concentrations (ppm), respectively, of the samples. The depleted mantle component is assigned an ϵ_{Nd} ratio of +6 and a Nd concentration of 4.5 ppm. All ϵ_{Nd} values are calculated as initial ratios at 2169 Ma. Note that the isotopic compositions of the samples cannot be produced by additions of the second estimate ($\epsilon_{Nd} = +2$) of the undepleted mantle melt component discussed in the text

(A) Depleted Mantle + Archean Crust				Archean Crust			
Sample	Description	ϵ_{Nd}	Nd^m	$\epsilon_{Nd} = -12, Nd = 16 \text{ ppm}$		$\epsilon_{Nd} = -7, Nd = 22 \text{ ppm}$	
				% AC	Nd^c	% AC	Nd^c
B88-25	Le Fer Basalt	1.8	6.33	8.0	5.4	9.0	6.1
2038A	Le Fer Granophyre	1.2	70.70	9.4	5.6	10.8	6.4

(B) Depleted Mantle + Undepleted Mantle				Undepleted Mantle	
Sample	Description	ϵ_{Nd}	Nd^m	$\epsilon_{Nd} = 0, Nd = 16 \text{ ppm}$	
				% UM	Nd^c
B88-25	Le Fer Basalt	1.8	6.33	40.3	9.1
2038A	Le Fer Granophyre	1.2	70.70	64.0	10.7

source (Table 9.4B). The ϵ_{Nd} ratio of the second estimate of the undepleted mantle component exceeds those of the Le Fer samples, and thus cannot be the source of the observed isotopic compositions.

It is apparent from the calculated Nd concentrations reported in Table 9.4 that only the addition of the more enriched of the crustal endmember components results in a reasonable Nd content in the hybrid melts. For the Le Fer basalt, addition of the required proportion of the primordial undepleted mantle melt component results in an initial Nd concentration which is higher than that measured in the sample. Addition of the more depleted crustal endmember ($\epsilon_{Nd} = -7$) gives a resultant Nd concentration that is less than that of the sample, but which allows for only minor concentration ($\sim 4\%$) through fractional crystallization. This is substantially less than the approximately 25% crystallization required to increase the initial contents of other incompatible elements, particularly Sm through Lu, to measured concentrations in the basalt. The calculated initial Nd concentration of a depleted mantle melt mixed with the more enriched crustal

endmember allows concentration through crystallization of about 17% of the liquid which, although still low, is a better approximation of the apparent extent of melt evolution.

Mixing lines representing the additions of crustal and undepleted mantle melt components to depleted mantle magmas are presented in Figure 9.17. The mixing lines have been calculated assuming 25% fractional crystallization of the initial liquid, which is the average degree of crystallization necessary to raise the estimated initial concentrations of Sm, Ti, Y, Yb and Lu in the depleted mantle melt component to the levels measured in Le Fer basalt sample B88-25. The trace element pattern for the basalt, and the range of trace element abundances for the Le Fer basaltic suite, are also illustrated. The mixing line for addition of 40% of the undepleted mantle component exhibits a much stronger enrichment trend than the Le Fer basalts. This confirms the interpretation drawn from the Nd concentration data in that the proportion of melt component derived from the undepleted mantle necessary to produce the isotopic composition of the basaltic sample is inconsistent with the incompatible element content of the sample. Consequently, the isotopic composition of the Le Fer basalts cannot have been controlled by the undepleted mantle reservoir.

Figure 9.17B illustrates mixing curves for additions of continental crust to depleted mantle melts. It is apparent that the mixing curves are good approximations of the trace element patterns of the Le Fer basalts only for elements more compatible than Nd. This is particularly true when the pattern of sample B88-25, which requires addition of 8-9% crustal material to produce its isotopic composition, is compared to the 10% mixing line. The basalt exhibits a trace element profile with large Rb and K depletions, a smaller La depletion, and neutral Nb and Sr abundances in comparison to the mixing line. Most of these features, including the La depletion (see Figure 8.2), are characteristic of the Le Fer basalts and hence cannot be considered aberrations in the case of sample B88-25. Of these discrepancies, however, only the La depletion and the apparent lack of a Nb depletion are likely to be primary characteristics of the parental magma. The difference in Sr content, for example, could reflect plagioclase fractionation during evolution of the Le Fer magmas, Sr loss during alteration, or overestimation of the Sr content of the depleted mantle melt component. The irregular variations in the abundances of Rb through Nb are unlikely to be of magmatic origin, and it was previously suggested that this irregularity

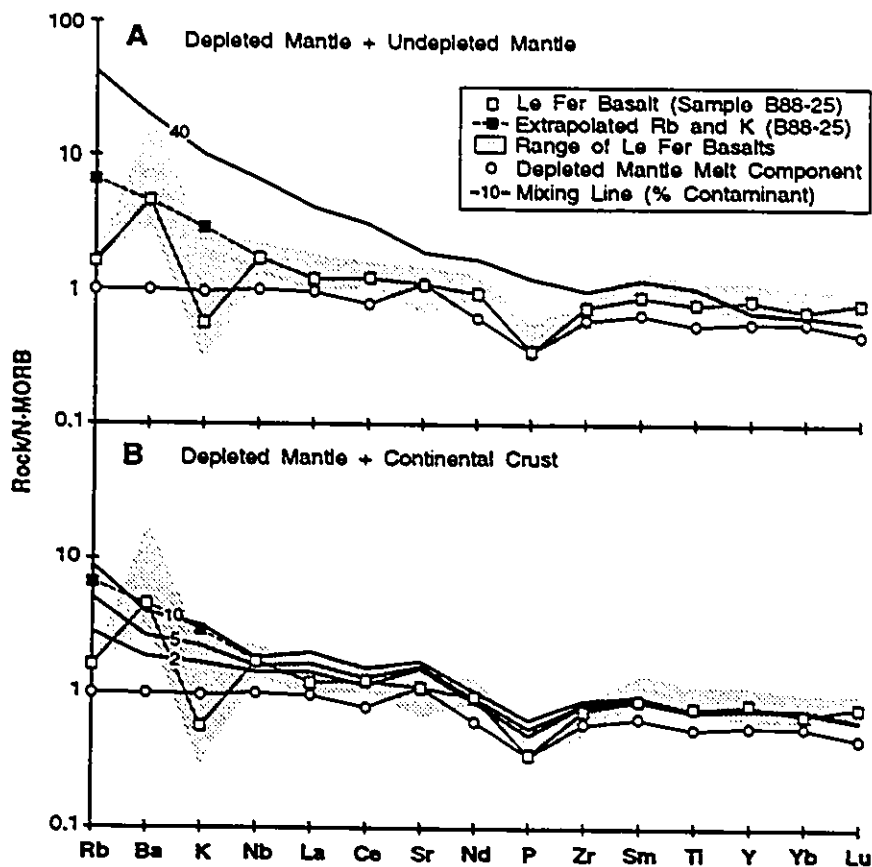


Figure 9.17. Comparison of the normalized extended trace element patterns of Le Fer basalt sample B88-25 with patterns produced by mixing of different reservoir melt components. The mixing lines represent hybrid primary melts which have undergone 25% fractional crystallization. The range in trace element abundances of the Le Fer basalts and the trace element pattern for the depleted mantle melt component are also shown. Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba, which is from Saunders and Tarney (1984).

reflects post-crystallization loss of Rb and K. If this interpretation is correct, and if extrapolated abundances of Rb and K can be considered realistic approximations of true values, the resultant profile is more or less consistent with the 10% crust mixing line (Figure 9.17B).

The absence of Nb depletions in any of the Le Fer samples obtained in this study is curious in light of the fact that many of the cycle 2 tholeiitic samples show such a depletion despite requiring smaller quantities of crustal material to produce their isotopic composition. Since Nb is incompatible in the silicate phases thought to have controlled

melt evolution in the two sequences, the origin of this discrepancy must lie in the compositions of the melt components derived from the contributing source reservoirs. If the average crustal composition was equivalent during the two magmatic episodes, this would require a larger positive Nb anomaly in the mantle partial melts parental to the Le Fer basalts. If the compositions of the crustal contributions were not equivalent, it is possible that the upper Menihok/Willbob magmas assimilated sialic material with a more pronounced Nb depletion. There is no obvious method of distinguishing between these possibilities.

The discrepancy between the La-depleted trace element profile of the Le Fer basalts and the La enriched patterns of mantle-crust mixing lines may reflect an overestimation of the La content of the crustal melt component. To produce a mixing line with a trace element profile comparable to the basalts, however, the La content of the crustal component must be reduced by about 50%. This produces a highly anomalous La depletion in the trace element pattern of the crust, and thus is unlikely to be the source of the discrepancy. An alternative explanation involves the fractionation of plagioclase from the differentiating Le Fer melts. Because the light REE are preferentially incorporated into this phase, residual melts will evolve toward light REE depleted compositions. But the total REE content of plagioclase is so low that very large quantities of this mineral must be removed before significant change in the REE pattern of residual melts is attained. In addition, the analysis of differentiation in the Le Fer suite concluded that clinopyroxene was also removed from the liquids. Since clinopyroxene has a REE pattern which is essentially the opposite of the REE pattern of plagioclase, the tendency toward light REE depleted melts will be nullified. Finally, plagioclase was an equally important part of the fractionate in the cycle 2 tholeiitic melts, the volcanic products of which do not exhibit a comparable La depletion. Consequently removal of plagioclase during melt evolution is unlikely to be the source of the La deficiency in the Le Fer basalts.

If the La depletion in the Le Fer basalts does not reflect the composition of the crust and is unrelated to differentiation processes, it must be a primary characteristic of the undepleted melt component. Such LIL element depletions are not uncommon in oceanic tholeiites (e.g., Fig. 9.3A) and these presumably reflect both the composition of the source and the partial melting conditions. The trace element pattern of two hypothetical, LIL

depleted, Le Fer primary magmas derived from a depleted mantle source are illustrated in Figure 9.18. In the first, the concentrations of Rb through Ce of the depleted mantle component have been adjusted such that the addition of 8% continental crust results in a trace element pattern comparable to that of sample B88-25 (Figure 9.18A). In comparison to the composition of the depleted mantle component estimated from the high-Mg Willbob basalt of Beaudoin and Laurent (1989), the abundances of K, La and Ce are significantly lower while Rb is significantly higher. The Nb content of the hypothetical depleted mantle melt component has been reduced slightly (2.33 to 1.7 ppm), but to a lesser extent than Rb through Ce such that a positive Nb anomaly results. This negates the Nb depletion of the crustal contaminant. The hypothetical depleted mantle melt, when contaminated with continental crust, will produce trace element profiles consistent with those of the Le Fer basalts. But it is unlikely that mantle fusion processes would give rise to melt components having such irregular abundances of Rb through Ce in comparison to average N-MORB values.

A geochemically more consistent trace element pattern for the depleted mantle melt component is illustrated in Figure 9.18B. In this example, a light REE depletion trend is continued in the abundances of K, Ba and Rb. The Nb content of the hypothetical depleted mantle component, however, has not been adjusted such that the positive Nb anomaly is retained. While such Nb anomalies are more common in strongly incompatible element enriched rocks (e.g., ocean island basalts, Fig. 9.3D), the fact that the high-Mg Willbob basalt of Beaudoin and Laurent (1989) displays a comparable Nb enrichment indicates that such patterns are possible in the LTBS tholeiites. The LIL depletion of the hypothetical depleted mantle melt component approximates the La depletion of the Le Fer basalts, but results in Rb, Ba and K abundances that are considerably lower than measured values. Minor additions of the undepleted mantle melt component could increase these concentrations to comparable values. A contaminant mixture of 8% crust and 2% undepleted mantle material, for example, would give appropriate concentrations of Rb, Ba and K while maintaining a relatively low La content (Figure 9.18B). It should be noted that a small reduction in the Nd concentration of the hypothetical depleted mantle component (not shown in Figure 9.18B) would be consistent with the LIL depleted trace

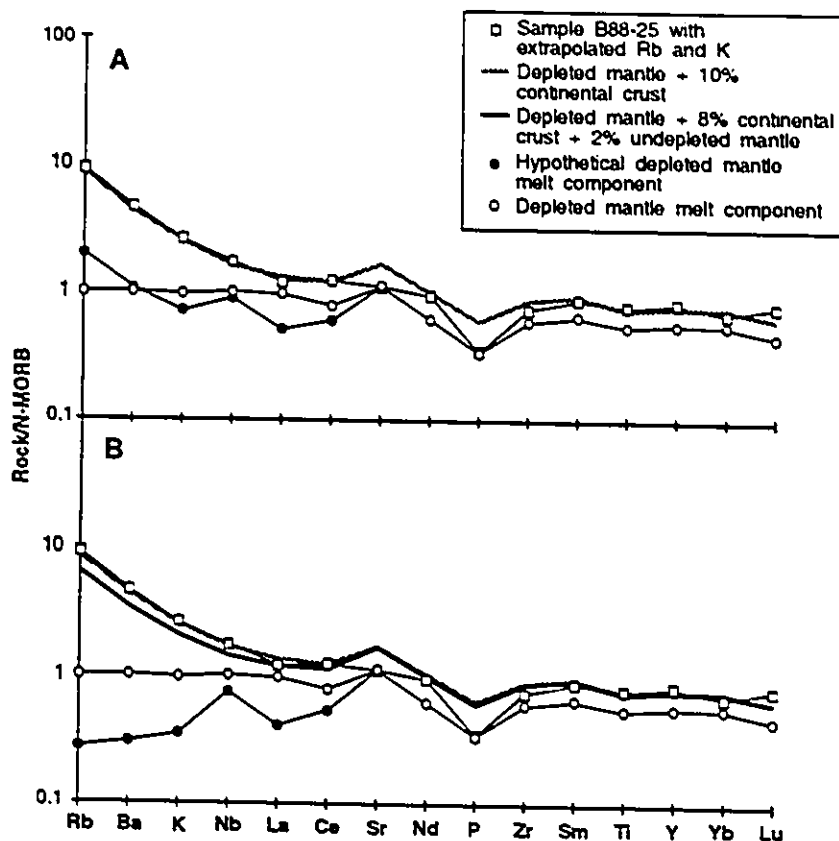


Figure 9.18. Normalized extended trace element patterns of two hypothetical depleted mantle melts, and mixing curves for additions of crustal and undepleted mantle material. (A) Depleted mantle component estimated from trace element abundances in Le Fer basalt sample B88-25 and continental crust. (B) Depleted mantle component estimated from La content of Le Fer basalt sample B88-25. LIL element abundances have been adjusted from values in (A) to provide a smooth trace element pattern. The trace element pattern of the depleted mantle component as estimated from the high-Mg Willbob basalt is shown for comparison. Mixing lines reflect 25% fractional crystallization. Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba, which is from Saunders and Tarney (1984).

element pattern, and such a reduction would decrease the proportion of crustal material required to produce the isotopic composition of the Le Fer basalt.

The absence of highly primitive Le Fer volcanic rocks, and the limited amount of isotopic data available for the suite, make the foregoing discussions highly speculative. It is clear, however, that the mildly depleted isotopic compositions of at least two Le Fer samples necessitate the assimilation of some proportion of crustal material. While some

contribution from an undepleted mantle source cannot be excluded, the incompatible element poor nature of the Le Fer basalts is inconsistent with significant proportions of such material. Because of the highly incompatible element enriched nature of the undepleted mantle component, this conclusion is insensitive to potential errors in the estimated composition of this component. Consequently, the ϵ_{Nd} ratios of the Le Fer suite were controlled by the isotopic composition of the depleted mantle and continental crust. The isotopic composition of the granophyric sample of Rohon *et al.* (1993) may reflect *in situ* assimilation of country rock at the level of emplacement, but this cannot be the case for the basalt. As with the cycle 2 tholeiites, it can thus be concluded that the contamination of the Le Fer magmas was a general phenomenon, and not restricted to melts emplaced in high level intrusions. If this interpretation is correct, then the similarity of the ϵ_{Nd} ratios of the two samples suggests that the extent of country rock assimilation in the Le Fer sills was minor. This inference is in accord with the evidence regarding *in situ* contamination of the Menihék gabbros.

While the assimilation of crustal material is required by the isotopic data, the calculated compositions of depleted mantle-crust mixtures have elevated incompatible element abundances, particularly with respect to La, in comparison to observed trace element profiles. This discrepancy may, in part, reflect minor inaccuracies in the estimated composition representation of the crustal contaminant. For example, the proportion of crustal material required to produce the isotopic composition of the Le Fer samples would be reduced to about 5% if the contaminant had an ϵ_{Nd} ratio of -14 and a Nd content of 20 ppm. But even with this reduction in the crustal contribution, the light REE abundances of the hybrid melts exceed measured values. It thus seems probable that the depleted mantle contribution to the Le Fer magmas was somewhat LIL element depleted relative to the compositional estimate derived from the high-Mg Willbob basalt of Beaudoin and Laurent (1989). Using LIL depleted compositional estimates for the depleted mantle component adequately reproduces observed La and Ce abundances, but gives rise to Rb, Ba and K depletions. If the hypothetical LIL depleted mantle component is realistic, the discrepancy between Rb, Ba and K concentrations in the depleted mantle/crustal mixtures and the Le Fer basalts might reflect minor additions of an undepleted mantle component. This might

indicate that mantle fusion processes and the physical relationship between depleted and undepleted mantle reservoirs were similar during cycle 1 and 2 magmatism.

The foregoing discussion indicates that the Le Fer magmas were probably drawn from the same general source reservoirs as the cycle 2 tholeiitic melts. Despite this common source, the two suites exhibit small difference in trace element abundances which suggest that at least one of the melt components derived from the three reservoirs differed in composition in the two magmatic episodes. The compositional differences in the two sequences are thought to be related to variations in the compositions of partial melts derived from the depleted mantle, possibly due to variations in melting depth and thickness of overlying lithosphere. But this is not a unique solution. For example, given the bulk geochemical similarities between the suites it could be concluded that the depleted mantle components were identical, but there were large compositional differences in the lesser contributions from undepleted mantle or crustal sources. A geochemical comparison of the cycle 1 and 2 alkaline rocks might provide some insight into possible heterogeneity in the undepleted mantle during the two magmatic episodes, but insufficient data are presently available from the Chakonipau lavas to undertake such a comparison. Without further information, potential differences in the origins of the two tholeiitic suites cannot be evaluated.

Nimish and Lower Menihék Magmas

The geochemical characteristics of the Nimish and lower Menihék alkaline basalts suggest that they were derived primarily from an undepleted mantle source. The isotopic compositions, however, indicate that more than one source contributed to melt compositions. This is most obvious in the case of the Nimish syenite, whose negative ϵ_{Nd} ratio necessitates a crustal contribution. Similarly, unless the ϵ_{Nd} ratio of the undepleted mantle component was at least +3.3, the isotopic composition of the Nimish basalt requires some proportion of the depleted mantle component. The lower Menihék basalts exhibit intermediate isotopic compositions which could represent undiluted melts from an undepleted mantle reservoir depending on the ϵ_{Nd} ratio of the latter source.

Mixing proportions and resultant Nd concentrations are summarized in Table 9.5A for the addition of a depleted mantle component to alkaline magmas derived primarily from an undepleted source. If the undepleted mantle component is assigned a ϵ_{Nd} ratio of 0, the contributions from the depleted mantle source required to produce observed isotopic compositions are unrealistically high. If this value is correct, for example, the Nimish basalt (sample B16-5a) would be derived primarily from the depleted mantle, and this is inconsistent with the enrichment trend apparent in the extended trace element pattern of the sample (Figure 9.19A). The proportions of the depleted mantle component required for the lower Menihek samples is also too high to be consistent with their incompatible element contents (Figure 9.19A). In addition, large degrees of crystallization of the primary hybrid melts would be required to produce the observed trace element compositions in the basalts. This is also apparent in the low calculated Nd concentrations in comparison to measured values (Table 9.5A).

The ϵ_{Nd} ratio of the undepleted mantle component has also been assigned values of +2 and +3 in Table 9.5A. Mixing of the depleted mantle component with magmas derived from an undepleted reservoir having these isotopic ratios cannot produce the observed ϵ_{Nd} values of the Nimish syenite or lower Menihek basalts, but such mixtures provide a more realistic approximation of the composition of the Nimish basalt. An undepleted mantle ϵ_{Nd} ratio of +2 reduces the required proportion of the depleted mantle component to about 60%, and a mixture such as this displays an enriched trace element profile which is similar in shape to that of the Nimish basalt (Fig. 9.19B). The elemental abundances after 50% fractional crystallization, however, are again too low. Increasing the degree of crystallization to approximately 70% results in comparable concentrations for most elements, but it is clear that the resultant mixing line is far less depleted in Y and the heavy REE than the Nimish basalt.

The proportion of the depleted mantle component required to produce the isotopic composition of the Nimish basalt is reduced to about 25% if the ϵ_{Nd} ratio of the undepleted mantle was +3. This mixture exhibits a trace element pattern which is a very good approximation of that of the basalt after 50% crystallization, except for a slightly low Nd content (Fig. 9.19C). An increase in the Nd content of the undepleted mantle component

Table 9.5. Calculated proportions of depleted mantle (% DM) or Archean crust (% AC) melt components added to Nimish and lower Menihok undepleted mantle magmas to give observed ϵ_{Nd} ratios. Calculated Nd concentrations in ppm (Nd^c) resulting from mixing are indicated. ϵ_{Nd} and Nd^m are the measured isotopic ratios and Nd concentrations (ppm), respectively, in the samples. The undepleted mantle melt component is assigned a Nd concentration of 16 ppm and variable ϵ_{Nd} ratios. All ϵ_{Nd} values are calculated as initial ratios at 1879 Ma for the Nimish samples and 1884 Ma for the lower Menihok samples. Proportions and Nd concentrations for samples which cannot be derived from the stated components are denoted n/a.

(A) Undepleted Mantle + Depleted Mantle ($\epsilon_{Nd} = +6$, $Nd = 4.5$ ppm)				Undepleted Mantle Isotopic Composition					
Sample	Description	ϵ_{Nd}	Nd^m	$\epsilon_{Nd} = 0$		$\epsilon_{Nd} = +2$		$\epsilon_{Nd} = +3$	
				% DM	Nd^c	% DM	Nd^c	% DM	Nd^c
W10-11	Nimish Basalt	3.3	36.79	81.0	6.7	62.3	8.8	25.8	13.0
B16-5a	Nimish Syenite	-0.6	102.20	n/a	n/a	n/a	n/a	n/a	n/a
F90-50	Lower Menihok Basalt	1.9	49.25	63.1	8.7	n/a	n/a	n/a	n/a
F20-9	Lower Menihok Basalt	1.5	51.39	53.2	9.9	n/a	n/a	n/a	n/a

(B) Undepleted Mantle + Archean Crust ($\epsilon_{Nd} = -12$, $Nd = 16$ ppm)				Undepleted Mantle Isotopic Composition					
Sample	Description	ϵ_{Nd}	Nd^m	$\epsilon_{Nd} = 0$		$\epsilon_{Nd} = +2$		$\epsilon_{Nd} = +3$	
				% AC	Nd^c	% AC	Nd^c	% AC	Nd^c
W10-11	Nimish Basalt	3.3	36.79	n/a	n/a	n/a	n/a	n/a	n/a
B16-5a	Nimish Syenite	-0.6	102.20	4.7	16.0	18.4	16.0	23.8	16.0
F90-50	Lower Menihok Basalt	1.9	49.25	n/a	n/a	0.4	16.0	7.0	16.0
F20-9	Lower Menihok Basalt	1.5	51.39	n/a	n/a	3.9	16.0	10.3	16.0

(C) Undepleted Mantle + Archean Crust ($\epsilon_{Nd} = -7$, $Nd = 22$ ppm)				Undepleted Mantle Isotopic Composition					
Sample	Description	ϵ_{Nd}	Nd^m	$\epsilon_{Nd} = 0$		$\epsilon_{Nd} = +2$		$\epsilon_{Nd} = +3$	
				% AC	Nd^c	% AC	Nd^c	% AC	Nd^c
W10-11	Nimish Basalt	3.3	36.79	n/a	n/a	n/a	n/a	n/a	n/a
B16-5a	Nimish Syenite	-0.6	102.20	6.0	16.4	22.5	17.4	28.8	17.7
F90-50	Lower Menihok Basalt	1.9	49.25	n/a	n/a	0.4	16.0	7.9	16.5
F20-9	Lower Menihok Basalt	1.5	51.39	n/a	n/a	4.5	16.3	11.7	16.7

from 16 to 20 ppm, however, would lead to a concentration of about 31 ppm in the hybrid melt after 50% crystallization, and this is reasonably close to the measured value of 37 ppm in the Nimish basalt. The higher Nd concentration would raise the required proportion of the depleted mantle component to about 30%, but this increase would have little effect on the resultant trace element profile. With the exception of the elevated Nd abundance, the 25% and 30% mixing lines would be almost indistinguishable in Fig. 9.19C, and

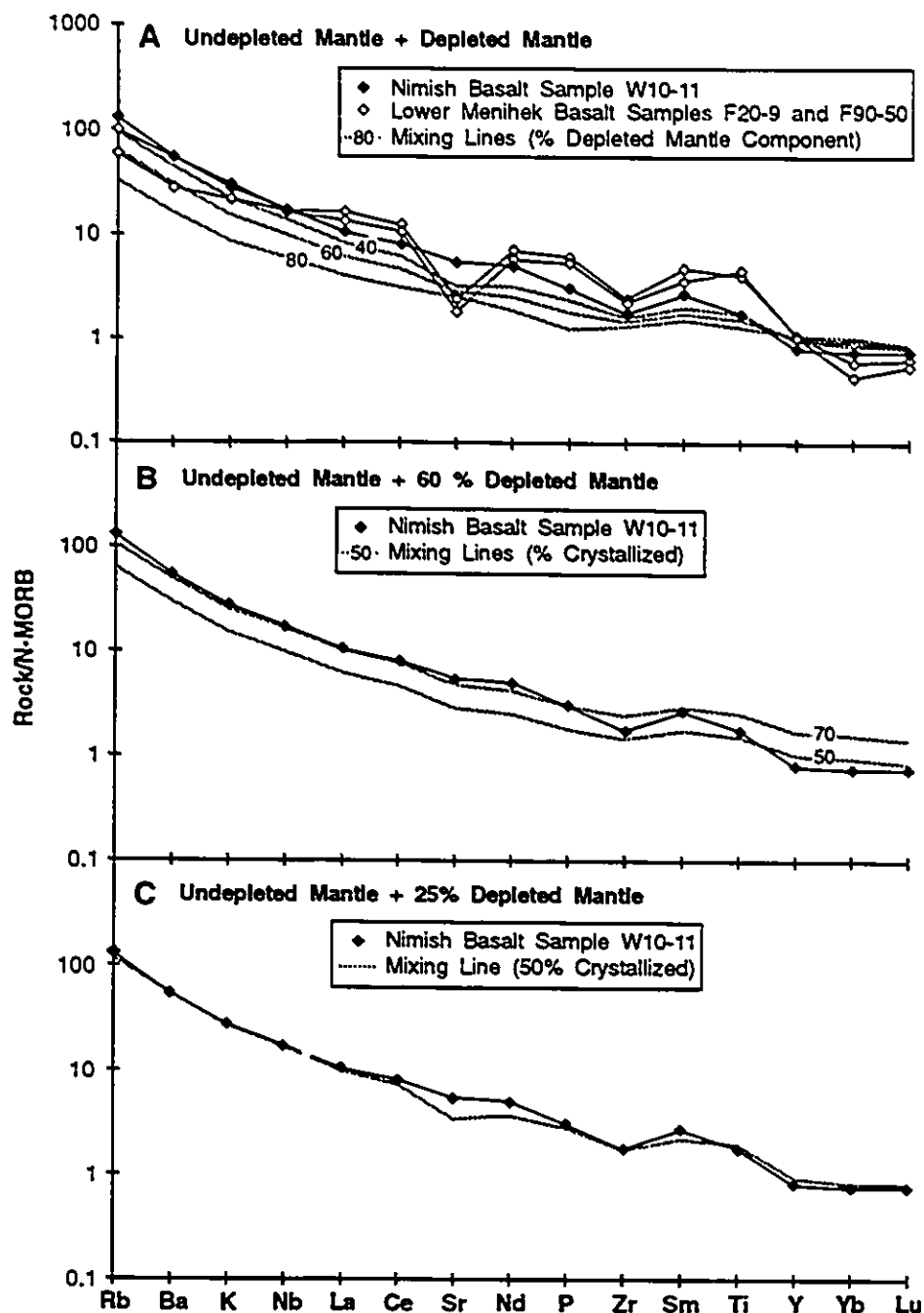


Figure 9.19. Comparison of the normalized extended trace element patterns of Nimish and lower Menihok samples whose isotopic compositions are known with patterns produced by mixing of different reservoir melt components. The mixing lines represent hybrid primary melts which have undergone 50% fractional crystallization except as indicated in (B). Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba, which is from Saunders and Tarney (1984).

therefore the latter composition has not been plotted. Thus the isotopic and incompatible element composition of the Nimish basalt is most consistent with additions of 25-30% of the depleted mantle component to magmas derived from an undepleted mantle source with an ϵ_{Nd} ratio of about +3. It should be noted, however, that a small crustal component could be accommodated by a slightly higher ϵ_{Nd} ratio in the undepleted mantle or, more realistically, by a larger contribution from the depleted mantle source. But the isotopic and trace element composition of the Nimish basalt is such that it can accommodate only very minor additions of sialic crust, and thus its ϵ_{Nd} ratio is probably close to the maximum for the suite. Given the isotopic composition of the Nimish syenite, it is probable that a more detailed isotopic study of the volcanic rocks suite would reveal a considerable range of ϵ_{Nd} ratios, extending from $\sim +3.3$ to -0.6 , and perhaps to lower values.

The isotopic and trace element compositions of the lower Menihek basalts are also more consistent with an undepleted mantle source having an ϵ_{Nd} ratio larger than that of primordial mantle. Because of the low estimated Nd content of the depleted mantle melt component, significant proportions of this material must be added to undepleted mantle magmas to produce even minor changes in the isotopic composition of the resultant liquid (Table 9.5A). For example, a 25% contribution from the depleted mantle is required to raise the ϵ_{Nd} ratio of an undepleted mantle partial melt from +3 to the measured value of +3.3 in the Nimish basalt. This effect would be more pronounced if the Nd content of the undepleted mantle component was higher than the estimated value, as is suggested by mixing calculations in the case of the Nimish basalt. But the extended trace element patterns of the lower Menihek basalts are inconsistent with large proportions of depleted mantle material. This is particularly true with respect to the highly depleted abundances of Y and the heavy REE, since the concentrations of these elements would increase rapidly with additions of depleted mantle material. Assuming the lower Menihek basalts reflect a maximum contribution of about 25% from the depleted mantle reservoir (i.e., comparable to that of the Nimish basalt), the isotopic composition of the samples requires an ϵ_{Nd} ratio greater than or equal to +1.5 in the undepleted source.

While some contribution from the depleted mantle cannot be excluded, and is in fact likely, it is clear that such contributions would not have been the controlling influence

on the isotopic and trace element compositions of the lower Menihek magmas. If the ϵ_{Nd} ratio of the undepleted mantle source was between +1.5 and +1.9, small quantities of the depleted mantle melt component would be required to produce the isotopic compositions of the samples. If the isotopic composition was greater than 1.9, contributions from continental crust are required. In this case, contributions from the depleted mantle are not necessary, but neither are they excluded. Although none of the lower Menihek samples exhibit negative Nb anomalies as might be expected from the assimilation of crustal material, a few of the Nimish basalts do show such features (cf. Figs. 8.19, 9.5). In addition, the contrast between the positive Nb anomalies of the lower Menihek clastic samples and the neutral Nb abundances of the lower Menihek flows might be indicative of differential crustal assimilation.

The proportions of crustal material added to undepleted mantle melts required to produce observed ϵ_{Nd} ratios are summarized in Tables 9.5B and 9.5C. Assuming that the ϵ_{Nd} ratio of the undepleted mantle source was +2, the maximum crustal contribution to the parental melts of the basaltic samples was less than 5% for both estimates of the crustal composition. If the ϵ_{Nd} ratio of the undepleted mantle source was +3, the maximum crustal contribution was about 12%. The equivalent values for the Nimish syenite using the two isotopic estimates are 18% and 29%. The larger crustal component required to produce the isotopic composition of the syenite probably reflects a longer residence time in a crustal magma chamber during melt differentiation. This might also be the source of the more pronounced negative Nb anomalies in Nimish rhyolites in comparison to Nimish basalts (cf. Figs. 8.19 and 8.23).

The extended trace element profiles of undepleted mantle-continental crust mixtures are illustrated in Figure 9.20, along with the patterns for the two lower Menihek samples for which isotopic data are available. The mixing lines have been calculated to reflect 50% crystallization, which gives a reasonable correspondence between calculated and measured trace element concentrations. The profiles for the hybrid melts give a good approximation of the overall trend of the trace element profiles of the lower Menihek basalts, but in detail there are substantial discrepancies. The most notable of these are the Nd, P, Sm and Ti enrichments and heavy REE depletions of the basalts. Because the

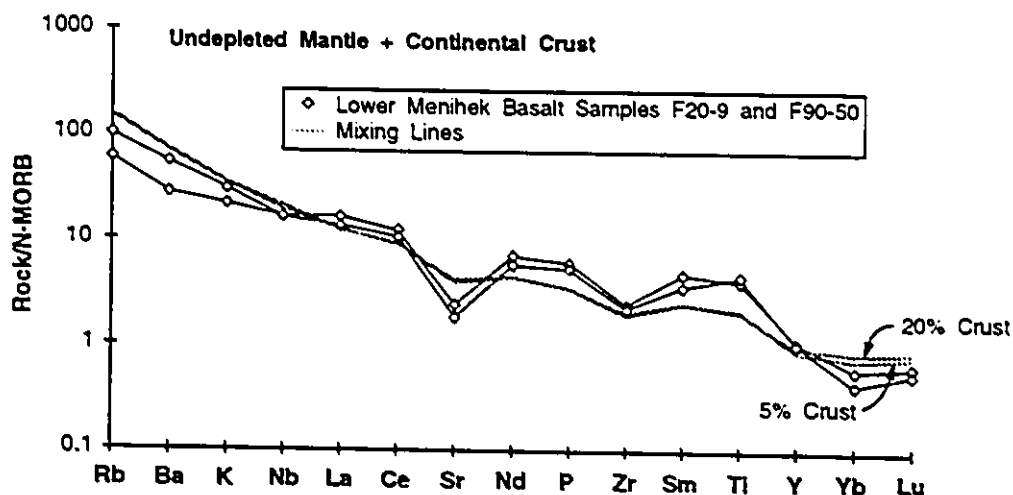


Figure 9.20. Comparison of the normalized extended trace element patterns of lower Menihiek samples whose isotopic compositions are known with patterns produced by mixing of undepleted mantle and crustal melt components. The mixing lines represent hybrid primary melts which have undergone 50% fractional crystallization. Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba, which is from Saunders and Tarney (1984).

anomalies reflect both higher and lower abundances, they cannot be functions of errors in the estimated degree of fractional crystallization. Since it is apparent that some of the alkaline magmas assimilated only very small quantities of crustal material (i.e., some Nimish basalts), or possibly none at all, these differences may derive from inaccuracies in the estimates in the calculated proportions of crustal material. But this is an unlikely explanation since the similarity of the estimated compositions of the crustal and undepleted mantle melt components gives rise to a close correspondence in mixing lines representing different crustal proportions. The mixing lines for 5% and 20% contamination illustrated in Figure 8.20, for example, are almost indistinguishable.

The differences in trace element patterns between the lower Menihiek basalts and the undepleted mantle-continental crust mixing lines may reflect contributions from a depleted mantle source. Addition of a depleted mantle component to the hybrid melts will tend to reduce the degree of LIL element enrichment of the mixing lines, thereby providing a better approximation of the lower Menihiek profiles in terms of these elements. Such additions, however, will not produce the elevated Nd, P, Sm and Ti concentrations found in the basalts, and will further increase the discrepancy in heavy REE content.

An alternative possibility is that the irregular nature of the trace element profiles of the lower Menihek flows is not a characteristic of the primary melts, but instead arose during fractional crystallization. It is difficult to imagine, however, the compositions of the liquidus phases in a mafic or ultramafic liquid which would produce a progressive Ce through Rb depletion trend, coupled with a Zr depletion and enrichments in Nd, P, Sm and Ti. In addition, although there are some pronounced differences in trace element ratios, the Nimish and lower Menihek basalts share a similar degree of incompatible element enrichment and exhibit comparable major element compositions. This suggests that the two suites would also share similar crystallization histories. The Nimish basalts would therefore exhibit equally irregular trace element profiles if fractional crystallization was the source of the anomalies. Since this is not the case, it is likely that the trace element patterns of the lower Menihek basalts derive principally from the compositions of the primary melts.

Assuming that the discrepancy between the estimated and observed trace element profiles is a function of primary melt composition, the source of the difference may lie in the use of the Nimish basalts to estimate the composition of the undepleted mantle melt component. This estimate could be inaccurate when applied to the lower Menihek suite for two possible reasons. The first is that the undepleted mantle source was compositionally heterogeneous. Thus the actual composition of the undepleted melt component may have varied among the Nimish and lower Menihek basalts, and the lower Menihek clasts. This interpretation is compatible with the inconstancy of incompatible element ratios within and among the alkaline suites.

The second potential reason for inaccurate representation of the undepleted melt component derives from the possibility that all of the available Nimish samples represent the products of a mixed source. Since the trace element patterns of all of the Nimish samples examined tend to share the profile of the Nimish basalt with the ϵ_{Nd} ratio of +3.3, it follows that all represent magmas to which the depleted mantle has made significant contributions. If the relative trace element abundances of the undepleted mantle component were similar to those of the lower Menihek basalts, addition of a depleted mantle component to hypothetical primitive lower Menihek magmas would have the effect of

smoothing out irregularities in the trace element patterns of the latter. This would lead to profiles which would resemble the more regular patterns of the Nimish basalts. While this supports the idea that all of the Nimish magmas were derived from addition of a depleted mantle component to undepleted mantle melts having the trace element characteristics of the lower Menihek suite, it does not exclude a heterogeneous undepleted source. Further isotopic data from the two alkaline sequences might help resolve the question of compositional heterogeneity in the undepleted mantle reservoir through comparison of the trace element characteristics of isotopically equivalent samples, particularly if primitive volcanic rocks could be obtained. It should be noted, however, that the compositions of the high-Zr Murdoch basalts are consistent with contributions from an undepleted mantle source having the trace element characteristics of the lower Menihek suite (see below). While this fact does not resolve the question, it does indicate that the lower Menihek clasts and flows, and the Murdoch high-Zr basalts, represent magmas derived from compositionally similar undepleted mantle sources.

The isotopic composition of the Nimish basalt (sample B16-5a) is consistent with a parental melt derived from both undepleted and depleted mantle sources. The trace element profile of the sample is compatible with an addition of 25-30% of the depleted mantle component to the estimated composition of undepleted mantle partial melts. This requires that the ϵ_{Nd} ratio of the undepleted mantle component was approximately +3. Assuming that the Nimish syenite was derived from a comparable undepleted mantle source, its isotopic composition requires the addition of close to 30% continental crust. This value would be only marginally higher if the syenite was comagmatic with the basalt (i.e., the ϵ_{Nd} ratio of the parental melt was + 3.3). The differences in isotopic and major element compositions of the two Nimish samples suggests that a longer crustal residence time, and an accordingly greater degree of crustal assimilation, was the source of the low ϵ_{Nd} ratio of the syenite.

The isotopic compositions of the lower Menihek basalts are compatible with addition of 0-12% continental crust to parental melts derived from an undepleted mantle source with an ϵ_{Nd} ratio greater than +1.5, and probably greater than +2. Since the data do not preclude the existence of a compositionally heterogeneous undepleted mantle source at

the time of cycle 2 alkaline magmatism, it is possible that this reservoir was also isotopically heterogeneous. But the isotopic data from the Nimish and lower Menihek place constraints on the isotopic composition of the undepleted mantle. These data, along with the trace element compositions of the samples, are best interpreted to reflect ϵ_{Nd} ratios of between +2 and +3 for the undepleted source. It is possible that the parental melts to the lower Menihek basalts embodied minor contributions from a depleted mantle source, although such additions are not required by the isotopic and incompatible element compositions of the basalts.

Although no isotopic data are available for the lower Menihek clastic samples, their trace element profiles are comparable in many respects to those of the lower Menihek flows, and a similar mixed source for their parental melts is postulated. The higher positive Nb anomalies of the clasts may reflect smaller degrees of crustal contamination. Alternatively, this and other small differences in the trace element patterns of the two sample suites, and the more general differences between the lower Menihek and Nimish suites, may be indicative of some degree of heterogeneity in the undepleted mantle on a scale small enough to produce compositional variations in the different alkaline magmas. Because no volcanic rocks representative of primitive melts are available, however, this possibility is difficult to evaluate.

Murdoch Magmas

The low-Zr Murdoch basalts are compositionally similar to the upper Menihek/Willbob flows, although they exhibit slightly greater incompatible element contents. This suggests that they represent magmas derived from a comparable source. This interpretation is supported by the isotopic data in that the ϵ_{Nd} ratios of the Murdoch samples (+4.2 and +3.6) fall within the range of the cycle 2 tholeiites. The isotopic composition of the cycle 2 tholeiites is thought to have been controlled by the depleted mantle and crustal melt components, rather than by the small contributions from an undepleted mantle source required to produce the observed incompatible element enrichments. This interpretation is examined in the case of the subalkaline Murdoch basalts in Table 9.6 and Figure 9.21. The trace element profiles of the Murdoch samples from

Table 9.6. Calculated proportions of undepleted mantle (% UM) or Archean crust (% AC) melt components added to low-Zr Murdoch depleted mantle magmas to give observed ϵ_{Nd} ratios. Calculated Nd concentrations in ppm (Nd^m) resulting from mixing are indicated. ϵ_{Nd} and Nd^m are the measured isotopic ratios and Nd concentrations (ppm), respectively, in the samples. The undepleted mantle melt component is assigned a Nd concentration of 16 ppm and ϵ_{Nd} ratios of 0, +2 and +3. The continental crust is assigned Nd contents of 16 and 22 ppm, and ϵ_{Nd} ratios of -12 and -7. The depleted mantle component is assigned a Nd content of 4.5 ppm and an ϵ_{Nd} ratio of +6. All ϵ_{Nd} values are calculated as initial ratios at 1884 Ma.

(A) Depleted Mantle + Undepleted Mantle				Undepleted Mantle					
Sample	Description	ϵ_{Nd}	Nd^m	$\epsilon_{Nd} = 0$		$\epsilon_{Nd} = +2$		$\epsilon_{Nd} = +3$	
				% UM	Nd^m	% UM	Nd^m	% UM	Nd^m
F90-85C	Murdoch Basalt	4.2	12.35	10.9	5.8	19.0	6.7	30.3	8.0
F90-85G	Murdoch Basalt	3.6	11.91	15.7	6.3	29.5	7.9	52.4	10.5

(B) Undepleted Mantle + Archean Crust				Archean Crust			
Sample	Description	ϵ_{Nd}	Nd^m	$\epsilon_{Nd} = -12$		$\epsilon_{Nd} = -7$	
				% AC	Nd^m	% AC	Nd^m
F90-85C	Murdoch Basalt	4.2	12.35	3.1	4.9	3.2	5.1
F90-85G	Murdoch Basalt	3.6	11.91	4.1	5.0	4.4	5.3

which isotopic data were obtained exhibit pronounced Rb and K depletions which probably reflect post-crystallization loss of these elements. Consequently, the average trace element compositions of the low-Zr suite has been used for comparison with mixing lines in Figure 9.21. Because the profiles of all of the samples are very similar except in the abundances of for Rb, K and Sr (see Figure 8.7), using average values lessens the irregularities imposed by element mobility but maintains the general pattern of the samples for which the isotopic data are available.

Assuming that the low-Zr Murdoch magmas were derived primarily from the depleted mantle reservoir, melt contributions of between 11% and 52% from an undepleted source are required to produce the observed ϵ_{Nd} ratios of the Murdoch samples. The smaller required proportions (11-16%) will result in the observed isotopic compositions if the isotopic composition of the undepleted source was comparable to a primordial reservoir (i.e., $\epsilon_{Nd} \sim 0$). The previous discussion concerning the source of the alkaline rocks concluded that the ϵ_{Nd} ratio of undepleted mantle reservoir was probably between +2 and +3. Thus the proportions calculated using a primordial ϵ_{Nd} ratio probably

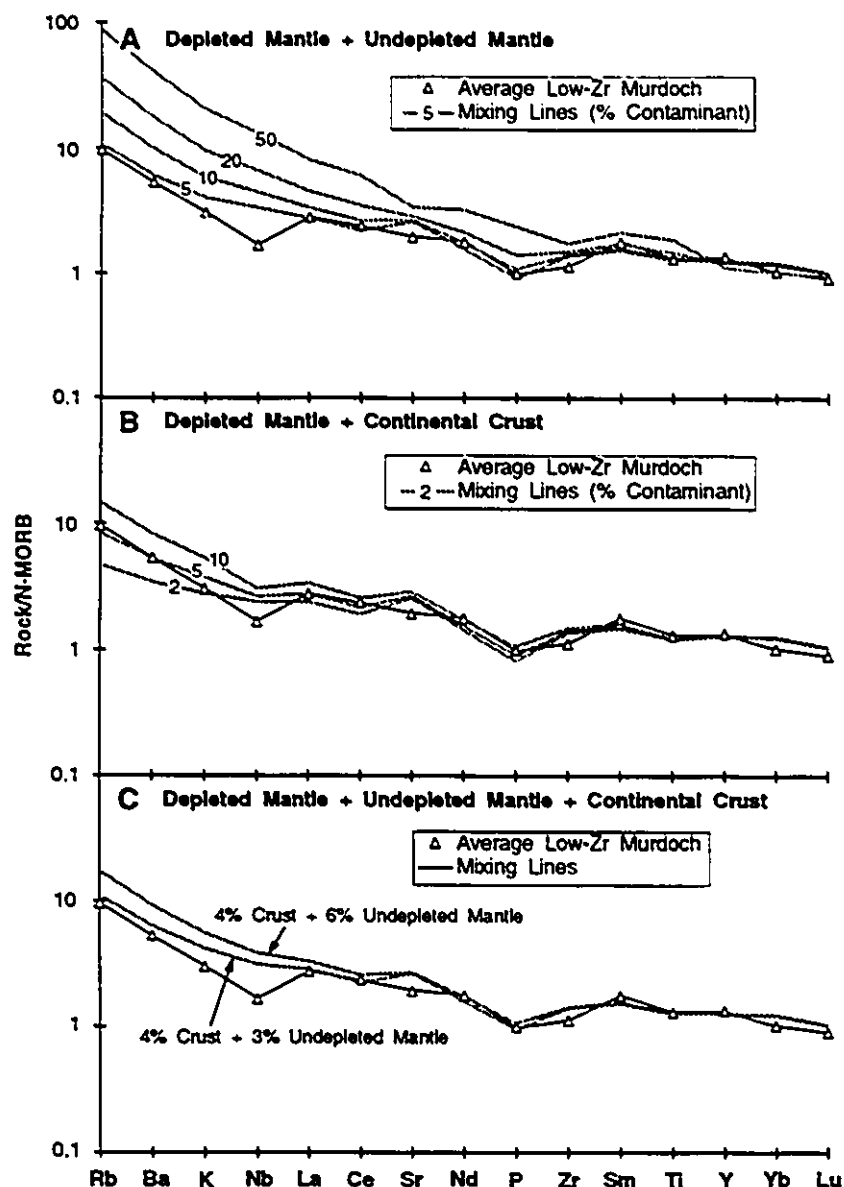


Figure 9.21. Comparison of the average normalized extended trace element patterns of the low-Zr Murdoch basalts with patterns produced by mixing of different reservoir melt components. The mixing lines represent hybrid primary melts which have undergone 55% fractional crystallization. Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba, which is from Saunders and Tarney (1984).

underestimate the contributions from an undepleted mantle reservoir required to give the isotopic composition of the samples. In any case, the addition of as little as 10-15% of the undepleted mantle component to depleted mantle partial melts gives rise to incompatible element enrichment trends after 55% crystallization which are inconsistent with those

found in the low-Zr Murdoch basalts (Fig. 9.21A). Increasing the undepleted mantle contributions to the 20-50% levels required by ϵ_{Nd} ratios of +2 to +3 in this reservoir enhances this discrepancy. Consequently, it is unlikely that the undepleted mantle source controlled the isotopic compositions of the low-Zr Murdoch suite.

The isotopic compositions of the samples can also be achieved by the additions of crustal material to depleted mantle partial melts. The required proportions of crust are less than 5% for both samples using both compositional estimates for this reservoir. The calculated Nd concentrations in the resultant melts could be increased to measured values by fractional crystallization of between 55% and 60% of the primary magmas. These values are compatible with the estimated extent of crystallization derived from the abundances of Sm, Ti, Y, Yb and Lu in the Murdoch samples. As noted previously, the concentrations of these elements will be least affected by additions of crustal material. Assuming complete incompatibility during crystal fractionation, the Sm-Lu contents of the Murdoch basalts are consistent with an average of 55% crystallization of the estimated depleted mantle melt component. The calculated extent of fractional crystallization is in accord with other geochemical parameters (e.g., Mg#) which indicate that the low-Zr Murdoch basalts are equivalent to the most evolved of the upper Menihek/Willbob volcanics.

The enrichment trends of mixing lines calculated for additions of continental crust to depleted mantle partial melts and adjusted for 55% fractional crystallization are comparable to that of the average low-Zr Murdoch basalt (Fig. 9.21B). The 5% mixing line most closely resembles the average Murdoch profile, and this is consistent with the calculated proportions of the crustal component (3.1-4.4%). As with many of the glomeroporphyritic cycle 2 basalts, however, the extent of the negative Nb anomaly of the average Murdoch profile is considerably greater than that estimated for the hybrid melts. This again may indicate that the Nb content of the continental crust or the depleted mantle melt component has been overestimated.

An additional discrepancy lies in the elevated Yb and Lu contents of the mixing lines in comparison to the average low-Zr Murdoch basalt. Because heavy REE depletions are characteristic of the alkaline rocks, this difference may reflect contributions from an undepleted mantle source. While minor contributions are possible, it is apparent that

extensive additions of an undepleted mantle melt component will give rise to enrichment trends which are incompatible with the trace element patterns of the Murdoch basalts. A 3% contribution from an undepleted source is compatible with the trace element profile of the average low-Zr Murdoch basalt (Fig. 9.21C). But increasing this proportion to 6%, which is still insufficient to produce conformity in Yb and Lu concentrations, results in a mixing line that exhibits a greater degree of Rb through Ce enrichment than the Murdoch samples. A contribution of 6% or greater might be acceptable if the LIL element content of the Murdoch basalts has been reduced by post-crystallization processes, as would seem to be the case for Rb and K. Without an unaltered example of the Murdoch basalts, however, this possibility cannot be evaluated. A larger undepleted mantle contribution would also be possible if the depleted mantle melt component was impoverished in LIL elements, as is suggested by the compositions of the Le Fer basalts.

An alternative possibility is that heavy REE content of the undepleted mantle melt component was overestimated. This would appear to be true in the case of the lower Menihek suite, since these samples exhibit stronger Y and heavy REE depletions than could be produced using the compositional estimate for the undepleted mantle component derived from the Nimish basalts. Using a lower Menihek-type undepleted mantle component would not only reduce the Yb and Lu contents of the mixing lines illustrated in Figure 9.21C, but would also reduce the degree of LIL element enrichment (Fig. 9.22). This derives from the fact that the lower Menihek basalts are less LIL element enriched than those of the Nimish Formation. This would allow slightly larger undepleted mantle melt components in the low-Zr Murdoch magmas, although contributions greater than 10% would seem unlikely given the enrichment trend in the average Murdoch sample (Fig. 9.22).

Some support for a lower Menihek-type composition for the undepleted mantle component in the Murdoch magmas comes from the trace element profiles of the high-Zr Murdoch suite. These rocks exhibit compositional characteristics which are transitional between the subalkaline Murdoch samples and the cycle 2 alkaline suites. Although no isotopic data are available for the high-Zr Murdoch basalts, it is likely that they represent hybrid melts derived from both depleted and undepleted mantle reservoirs. Since all of the LTBS magmas discussed thus far exhibit isotopic compositions consistent with crustal

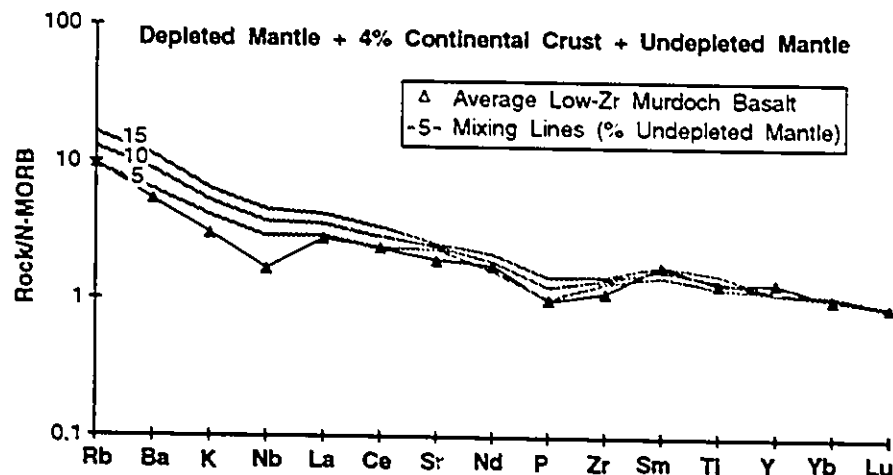


Figure 9.22. Comparison of the average normalized extended trace element patterns of the low-Zr Murdoch basalts with mixing lines for addition of a lower Menihek-type undepleted mantle component to depleted mantle partial melts. The mixing lines represent hybrid primary melts which have undergone 55% fractional crystallization. Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba, which is from Saunders and Tarney (1984).

contamination, it is probable that parental melts to the slightly alkaline also included a crustal component. But a comparison of depleted mantle-continental crust mixing lines with the trace element pattern of the most primitive high-Zr Murdoch basalt indicates that crustal contamination cannot be responsible for the enrichment trend of the latter (Fig. 9.23A). This is particularly apparent in the discrepancy between the mixing lines and the basalt in the abundances of Nd through Lu. It is also obvious that an incorrect assumption regarding the degree of melt evolution will have little impact on this conclusion.

A comparison of the trace element profile of the primitive high-Zr basalt with mixing lines representing depleted mantle melts with additions of an undepleted mantle component having the composition estimated from the Nimish basalts is presented in Figure 9.23B. Because the differences in trace element abundances between the depleted and undepleted mantle components are substantial, and because the proportions of these components are large, the effect of variations in the proportions on the resultant concentration of a given element are significant. Using Y, Yb and Lu, whose concentrations in the two components are most similar, the average degree of crystallization represented by the sample is 40%. Since the addition of the undepleted mantle component to depleted mantle partial melts will tend to decrease the concentration

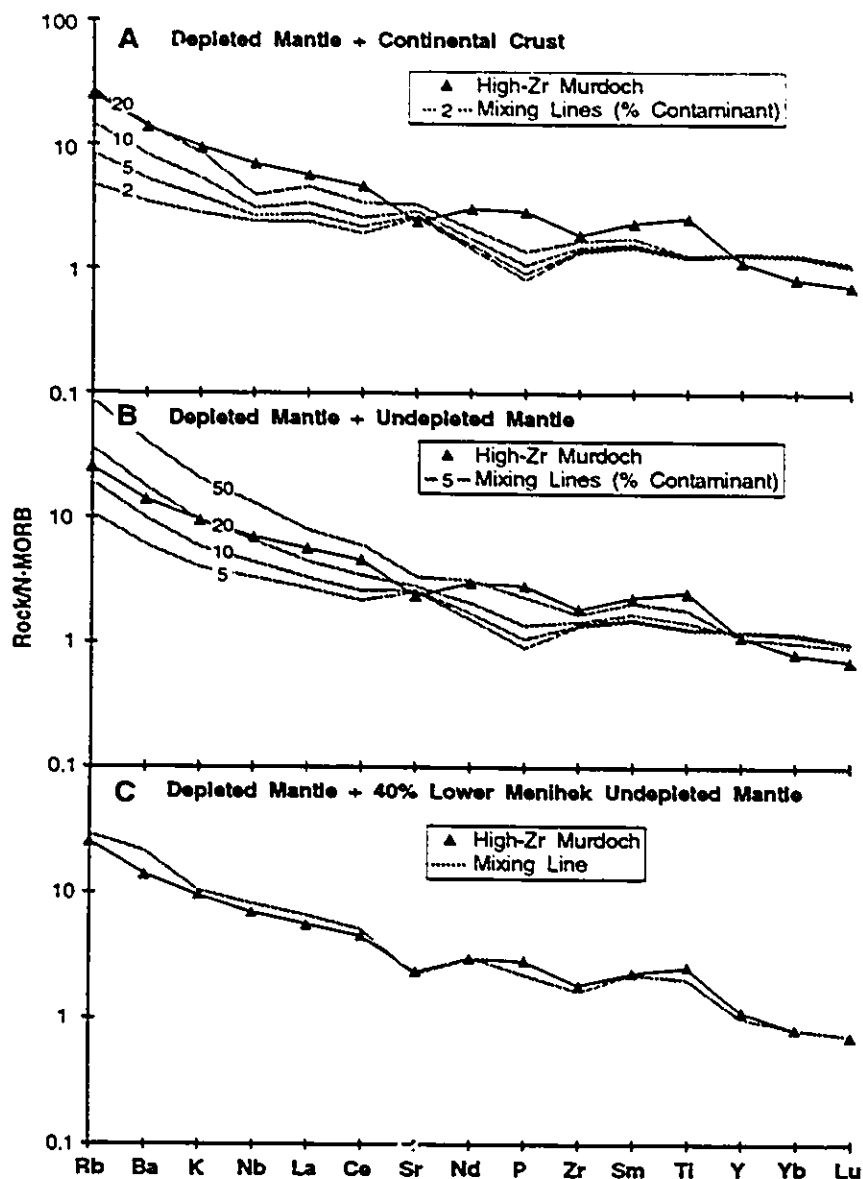


Figure 9.23. Comparison of the average normalized extended trace element patterns of the most primitive high-Zr Murdoch basalt with patterns produced by mixing of different reservoir melt components. The mixing lines represent hybrid primary melts which have undergone 45% fractional crystallization. Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba, which is from Saunders and Tarney (1984).

of these elements in the hybrid magma, this estimate is probably too low. Consequently, the mixing lines have been calculated assuming 45% crystallization of the primary magmas. This estimate is consistent with the more primitive character of the high-Zr Murdoch sample in comparison to the low-Zr suite and the alkaline rocks.

The overall enrichment trend of the primitive high-Zr sample is consistent with a 20-50% contribution from the undepleted mantle source (Fig 9.23B). But there are again substantial differences between the mixing lines and the sample in terms of the abundances of individual elements. These arise largely from the irregularity of the trace element profile of the basalt. In contrast, the use of a compositional estimate for the undepleted mantle component taken from the average composition of the lower Menihek basalts produces mixing lines which are in good agreement with the trace element pattern of the high-Zr Murdoch basalt. A 40% contribution from the undepleted reservoir, for example, gives a mixing line which is quite similar to the observed pattern after adjustment for 45% crystallization (Fig. 9.23C). Minor additions of crustal material, comparable to the proportions in the low-Zr basalts for example, would have little effect on the trace element profile of the high-Zr Murdoch basalt. Consequently, it is probable that the undepleted mantle reservoir involved in the generation of the Murdoch magmas was compositionally similar to the predominant mantle source for the lower Menihek suite. The second high-Zr sample for which REE data are available exhibits a trace element profile that is intermediate between the primitive high-Zr sample and the low-Zr suite in terms of enrichment trend. The trace element characteristics of the second sample are consistent with a 15-20% contribution from a lower Menihek-type undepleted mantle source.

The isotopic composition of the low-Zr Murdoch suite is consistent with minor additions (< 5%) of crustal material to depleted mantle partial melts. The trace element characteristics of these basalts, however, would seem to require a small contribution from an undepleted mantle source. In comparison with the cycle 2 tholeiites, the trace element profiles of the subalkaline Murdoch samples exhibit slightly greater enrichment trends, despite comparable degrees of crustal contamination. This suggests that the undepleted mantle contribution to the latter suite was greater than in the case of the upper Menihek/Willbob basalts. The trace element pattern of the most primitive of the two high-Zr Murdoch samples is consistent with a fairly large (~ 40%) contribution from an undepleted mantle component having the trace element characteristics of the lower Menihek basalts. The composition of the second high-Zr sample is consistent with smaller contributions (15-20%) from a comparable undepleted source. Crustal contamination of the resultant magmas is likely, given that such material appears to have been added to all

other LTBS melts. The trace element compositions of the Murdoch and lower Menihek samples are consistent with a lower Menihek-type undepleted mantle melt composition. The more regular trace element profiles of the Nimish alkali basalts may indicate that all of the Nimish melts contained a significant contribution from the depleted mantle source, or the Nimish undepleted mantle component was compositionally distinct from that of the lower Menihek and Murdoch suites.

The stratigraphic and compositional relationships among the cycle 2 volcanic rocks suggest that there was a decline in the influence of the undepleted mantle reservoir during progressive stages of magmatic activity. The alkaline rocks, particularly those of the lower Menihek Formation, would seem to have been derived predominantly from the undepleted source. The transitional geochemistries of the Murdoch suite may reflect the waning impact of melting in the undepleted reservoir. During early Murdoch magmatism, the undepleted mantle melt component was still a major influence on magma composition, although the proportion of melt derived from this source was significantly less than in the lower Menihek and Nimish magmas. The overlying low-Zr Murdoch basalts record a substantial reduction in the ratio of undepleted to depleted mantle melt components, and the resultant volcanic products are correspondingly impoverished in incompatible elements. The upper Menihek and Willbob basalts which cap the cycle 2 sequence, exhibit the smallest degrees of incompatible element enrichment, and hence represent magmas having the smallest undepleted mantle melt component.

9.7 Summary

With the possible exception of the high-Mg lower Menihek clast, the LTBS samples obtained in this study are unlikely to represent primary partial melts. In each sequence, there is some evidence that the volcanic rocks represent residual liquids derived through fairly extensive crystal fractionation histories. During formation of the cycle 1 and 2 tholeiitic suites, primitive magmas are thought to have undergone an early stage of olivine fractionation, which was followed by crystallization of plagioclase and pyroxene in the residual melts. The majority of tholeiitic samples represent liquids whose compositions

were controlled by the latter stage of fractional crystallization. This may indicate that a period of olivine fractionation was a necessary prerequisite to eruption.

A similar history is possible for the alkaline magmas, although there is only indirect evidence in support of such a process. The high-Mg Murdoch flows reported by Gebert (1988), the high-Mg lower Menihek clast, and the Nimish ultramafic sills (Watanabe and Fowler, 1989) indicate that some quantities of primitive melts were generated during alkaline magmatism. But the majority of volcanic products are relatively evolved basalts which cannot represent primary melts. The olivine+clinopyroxene cumulates of the Nimish ultramafic sills clearly reflect fractionation of these phases from the alkaline melts, and therefore it is thus possible that high-Mg primitive alkaline melts evolved through olivine $\pm$ pyroxene fractionation in crustal or subcrustal magma chamber and, with minor exceptions, only relatively evolved liquids were erupted.

The cycle 1 and 2 tholeiitic rocks exhibit extended trace element patterns which are comparable to those of modern T-MORB. The isotopic compositions of these suites are consistent with derivation of the parental melts primarily from a depleted mantle reservoir, with minor contributions from an undepleted mantle source and continental crust. In both sequences, it is the crustal contribution rather than the undepleted mantle component which has controlled isotopic composition. This follows from the fact that the proportion of the undepleted mantle component required to produce the measured ϵ_{Nd} ratios would result in much more incompatible element enriched compositions than are observed. The proportion of crustal material in the upper Menihek and Willbob basalts ranges from close to zero to a maximum of about 8%. The larger crustal proportions are typically observed in the more evolved samples, particularly those exhibiting plagioclase-glomeroporphyritic textures. This suggests that the plagioclase+pyroxene stage of fractional crystallization took place in crustal magma chambers where continued melt evolution was accompanied by increasing assimilation of continental crust. The more evolved samples would thus be derived from liquids with longer crustal residence times, and hence greater proportions of assimilated crustal material. This interpretation is consistent with the negative correlation between Nb/La and La content in the cycle 2 tholeiites.

The lower ϵ_{Nd} ratio of the Le Fer basalt in comparison to most cycle 2 tholeiitic samples is comparable to the relatively low value obtained for a Le Fer granophyre by Rohon *et al.* (1993). The isotopic compositions of the Le Fer samples are consistent with slightly greater (8-11%) contributions from continental crust than was the case for the upper Menihek/Willbob magmas. But the estimated proportions of crustal material would be compatible with the relatively flat extended trace element patterns of the Le Fer basalts only if the depleted mantle component was slightly impoverished in LIL elements. This suggests that either the compositional estimate for the depleted mantle component derived from the Willbob sample is in error, or the composition of the depleted mantle component was variable.

The isotopic and trace element composition of the sample of Nimish basalt is compatible with derivation of the parental magma from an undepleted mantle source with an ϵ_{Nd} ratio of $\sim +3$, coupled with a significant (25-30%) contribution from the depleted mantle. The enriched isotopic composition of the Nimish syenite probably reflects the assimilation of continental crust while the parental melt was resident in a crustal magma chamber. The isotopic composition of the lower Menihek basalts suggest that they represent the additions of minor quantities (0-12%) of crustal material to magmas derived predominantly from an undepleted mantle reservoir with an ϵ_{Nd} ratio of +2 to +3. The trace element profiles of mixtures of crustal material and the undepleted mantle component as estimated from the Nimish basalts are inconsistent with those of the lower Menihek basalts. This may indicate that the estimate was based on samples derived from a mixed source (i.e., all of the Nimish magmas included a significant depleted mantle component), or it could reflect heterogeneity in the undepleted mantle reservoir. The non-constancy of incompatible ratios supports heterogeneity of the undepleted source, while the smoother trace element patterns of the Nimish basalts are consistent with significant additions of a depleted mantle component to a lower Menihek-type undepleted mantle primary melt. Thus both interpretations may be correct. The range of ϵ_{Nd} ratios in the alkaline suites encompasses values obtained from the Povungnituk Alkaline Suite of the Cape Smith foldbelt (Gaonac'h *et al.*, 1992), and this indicates that undepleted mantle reservoirs,

comparable to those postulated as the source of modern alkaline suites, were widespread in the Early Proterozoic.

The low-Zr Murdoch basalts are isotopically and compositionally similar to the evolved members of the cycle 2 tholeiitic suite, and are thought to have been derived from similar magma sources. The slightly stronger enrichment trend in the Murdoch samples is thought to reflect slightly greater melt contributions (up to about 10%) from an undepleted source. Although no isotopic data are available for the high-Zr Murdoch basalts, the trace element characteristics of the most primitive sample is compatible with substantial additions (~ 40%) of undepleted mantle material to depleted mantle partial melts. A lesser crustal component could also be accommodated by the incompatible element abundances. As with the lower Menihek basalts, the trace element profiles of both groups of Murdoch rocks are more compatible with a lower Menihek-type undepleted mantle source.

The apparent variation in the proportions of undepleted mantle melt components contributing to the Murdoch magmas is probably the source of their transitional compositions. When all of the cycle 2 suites are compared, it is apparent that there was a progressive decline in the contributions from the undepleted mantle source to the parental magmas during evolution from the lower Menihek and Nimish alkaline suites (90-70%), through the slightly alkaline to subalkaline Murdoch basalts (40-5%), to the upper Menihek and Willbob tholeiites (<5%). It is also apparent that all of the suites, including the Le Fer basalts, require some degree of crustal contamination to produced their isotopic compositions. Thus continental crust would seem to be an important element in the tectonic setting of the LTBS.

Part IV
Evolution of the LTBS

Chapter 10

Tectonic Setting and Evolution of the LTBS

10.1 Tectonic Setting

Constraints from Geochemical Evidence

The tholeiitic basalts of the Le Fer and upper Menihek/Willbob formations are submarine flows associated with deep water marine sediments of turbidite origin. The basalts have generally been described as oceanic or transitional oceanic in geochemical affinity (Baragar, 1967; Dimroth, 1978; Baragar and Scoates, 1981; Wardle and Bailey, 1981; St. Seymour *et al.*, 1987; Wardle *et al.*, 1990a; Skulski *et al.*, 1993), and therefore compositionally consistent with magmatism in an extensional tectonic regime. This is apparent in the TiO_2 - K_2O - P_2O_5 tectonic discriminant diagram of Pearce *et al.* (1975) in which all of the cycle 1 and 2 tholeiitic samples, including the low-Zr Murdoch basalts, plot in the oceanic field (Fig. 10.1). Although there is evidence of post-crystallization loss of K_2O in many of the subalkaline basalts, the fact that none of the samples plot outside of the oceanic field suggests that their origin is compositionally more compatible with an oceanic setting than a continental setting. This is confirmed by a comparison of the extended trace element patterns of the LTBS tholeiites with those of tholeiitic volcanics from the most analogous group of continental rocks, continental flood basalts (CFB). Representative CFB from the Snake River Plateau, Deccan Traps and Parana Basin are illustrated in Figure 10.2 for comparison with the average of the Le Fer basalts and aphyric upper Menihek/Willbob basalts. The CFB have MgO contents in the range of the cycle 1 and 2 tholeiites, but are strongly enriched in most incompatible elements. As noted by Thompson *et al.* (1983) the incompatible element enrichment in CFB is comparable to that found in ocean island tholeiites rather than MORB, and this degree of enrichment is far greater than that exhibited by even the most evolved of the LTBS tholeiites.

The MORB-like geochemistry and submarine origin of the LTBS tholeiites have been cited in support of tectonic models in which the basalt-gabbro-turbidite successions

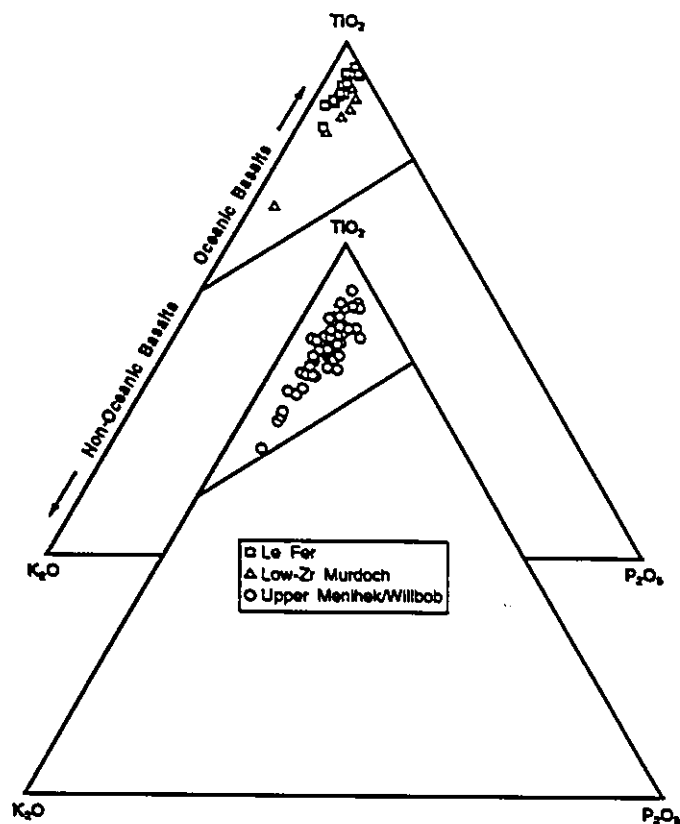


Figure 10.1. TiO_2 - K_2O - P_2O_5 discriminant diagram of Pearce *et al.* (1975) illustrating the oceanic geochemical characteristics of the Le Fer, low-Zr Murdoch and upper Menihok/Willbob subalkaline basalts.

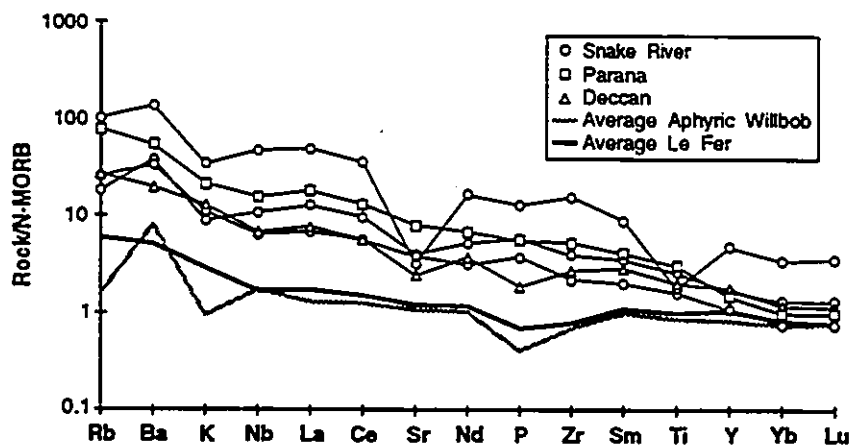


Figure 10.2. Normalized extended trace element plots comparing average abundances in the Le Fer and upper Menihok/Willbob tholeiites to representative examples of continental flood basalts. Data from Cox (1983). Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba which is from Saunders and Tarney (1984).

represent oceanic crust or embryonic oceanic crust. The isotopic data, however, indicate that Archean continental crust has made small contributions to the compositions of all members of the LTBS, with the possible exception of some Nimish melts. In addition, the basalt-gabbro-turbidite successions lack the structure of ophiolite complexes which are generally considered to be remnants of oceanic lithosphere. Finally, it is clear that the sedimentary infilling, at least in the western part of the central Trough, was deposited on continental basement. Consequently, while it is possible that the LTBS represents volcanic activity at a continent-ocean transition, it is doubtful that the cycle 1 and 2 subalkaline basalts were originally part of true oceanic crust formed at a mature ocean ridge system.

The crustal component that is necessitated by the isotopic compositions of LTBS samples may have been acquired by mantle partial melts in two principal ways. The first and most likely mechanism involves the assimilation of continental crust by LTBS magmas during passage from the mantle to the surface. This requires that the Labrador Trough was underlain by Archean basement for most of its depositional history. Although this would seem to contradict the geochemical evidence of an oceanic origin for the LTBS tholeiites, it is consistent with the sedimentary evidence from the Schefferville and Ferrum zones of the central Trough which indicates that deposition occurred on the subsiding margin of the Superior craton.

The second mechanism involves the addition of crustal material to mantle melts via subduction of continentally derived sediments. This may have been accomplished by generation of the LTBS magmas above an Early Proterozoic subduction zone, or generation of the melts from lithospheric or asthenospheric mantle into which older oceanic crust had been recycled (cf. Cohen and O'nions, 1982; Hart *et al.*, 1986; Sun and McDonough, 1989). In either model, the requirement for sialic basement rocks is removed, and this would allow much of the eastern basalt-gabbro-turbidite assemblages of the central Trough to represent some form of oceanic crust.

In the first subduction model, cycle 1 and 2 subalkaline basalts would represent the products of island arc or back-arc volcanism. To some extent, the compositions of cycle 1 and 2 subalkaline basalts are comparable to island arc tholeiites. In some tectonic discriminant diagrams, for example, the Le Fer, low-Zr Murdoch and upper Menihek/Willbob basalts plot within the field of island arc basalts or straddle the fields of

island arc and oceanic basalts (Figs. 10.3 and 10.4). On a geochemical basis alone, however, there are numerous contrasts between island arc volcanic sequences and the basaltic assemblage of the Labrador Trough. Phanerozoic island arcs, for example, are typically composed of both tholeiitic and calc-alkaline volcanics which generally include significant proportions of andesites and more siliceous rocks. Furthermore, there is often a compositional polarity to the volcanic products of island arcs, with the proportions of tholeiitic, calc-alkaline and alkaline rocks varying systematically with stratigraphic position and distance from the associated trench. In addition, the normalized extended trace element patterns of island arc tholeiites and calc-alkaline basalts are characterized by strong Lu through Rb enrichment trends and very large Nb depletions (Fig. 10.5). None of the foregoing features are consistent with the relatively incompatible element-poor tholeiitic basalts which are the overwhelmingly predominant volcanic product of LTBS magmatism.

Supra-subduction zone generation of the LTBS magmas may have occurred in a back-arc basin rather than in a magmatic arc. A back-arc setting was suggested by Rohon *et al.* (1993) who argued that the overall oceanic character of the tholeiites, coupled with the LIL element enrichment, was consistent with oceanic basalts formed at a back-arc spreading ridge. The isotopic evidence for continental basement, the preponderance of continentally derived sediments in the early stages of each depositional cycle, and the absence of true ophiolitic structure might then reflect the development of an ensialic back-arc basin on the margin of the Superior craton. Alternatively, the LTBS may represent ensimatic back-arc volcanism in close proximity to the Superior margin, possibly at the continent-ocean transition. Assuming that the upper Menihek/Willbob basalts are the expression of back-arc ridge volcanism, remnants of island arc volcanism might be expected to the east of the Howse zone or to the west of the Doublet terrane, depending on the direction of subduction. The Laporte Group, which overthrusts the Doublet terrane from the east, might represent such an arc.

Tholeiitic basalts erupted in a back-arc setting exhibit some similarities with those of island arcs. The volcanic products are often compositionally diverse, including both tholeiitic and calc-alkaline basalts and smaller quantities of andesites and more siliceous rocks. In this respect, Gill (1976) and Saunders and Tarney (1979) have suggested that

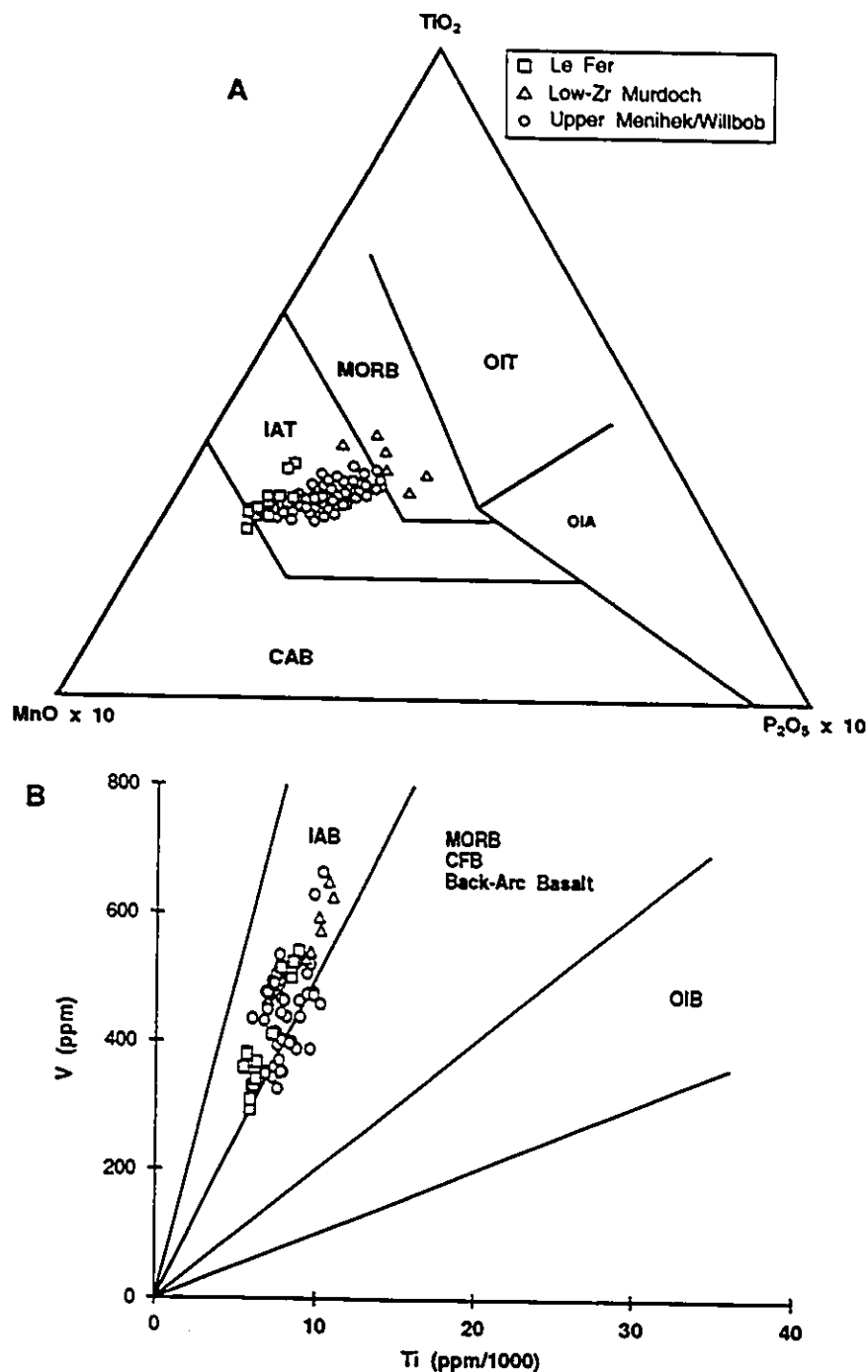


Figure 10.3. Tectonic discriminant diagrams illustrating the apparent volcanic arc affinities of the Le Fer, low-Zr Murdoch and upper Menihek/Willbob basalts. (A) TiO_2 - MnO - P_2O_5 discriminant diagram of Mullen (1982). (B) Ti-V discriminant diagram of Shervais (1982). CAB - calc-alkaline basalt, IAT - island arc tholeiite, IAB - island arc basalt, MORB - mid-ocean ridge basalt, OIT - ocean island tholeiite, OIA - ocean island alkali basalt, OIB - ocean island basalt.

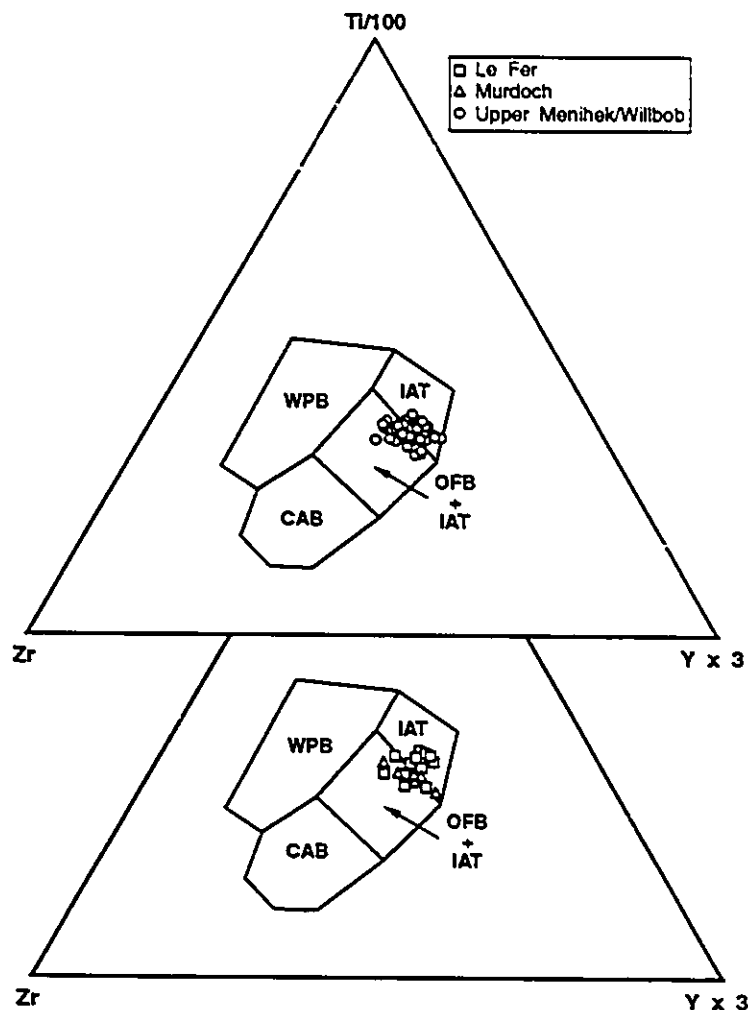


Figure. 10.4. Ti-Zr-Y tectonic discriminant diagram of Pearce and Cann (1973) illustrating the apparent volcanic arc affinities of the Le Fer, low-Zr Murdoch and upper Menihok/Willbob basalts. CAB - calc-alkaline basalt, WPB - within plate basalt, IAT - island arc tholeiite, OFB - ocean floor basalt.

back-arc volcanism is transitional between that of island arcs and normal mid-ocean ridges. Ensimatic back-arc tholeiites typically exhibit moderate LIL enrichments comparable to those found in the LTBS tholeiites, but also have, to a lesser extent, the Nb depletions characteristic of island arc basalts (Fig. 10.6A). These compositional features are generally more pronounced in ensialic back-arc volcanics (Fig. 10.6B), a feature interpreted by Saunders and Tarney (1984) to reflect increased LIL element enrichment of the mantle wedge source due to the greater longevity of continent-based arcs. In addition, back-arc basalts may show a progression from LIL enriched to LIL depleted compositions

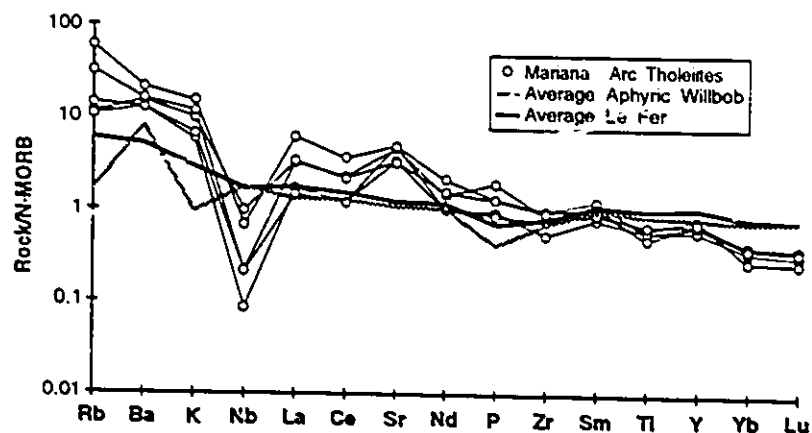


Figure 10.5. Normalized extended trace element plots comparing average abundances in the Le Fer and upper Menihok/Willbob tholeiites to representative examples of island arc tholeiites from the Marianas arc. Data from Woodhead (1988). Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba which is from Saunders and Tarney (1984).

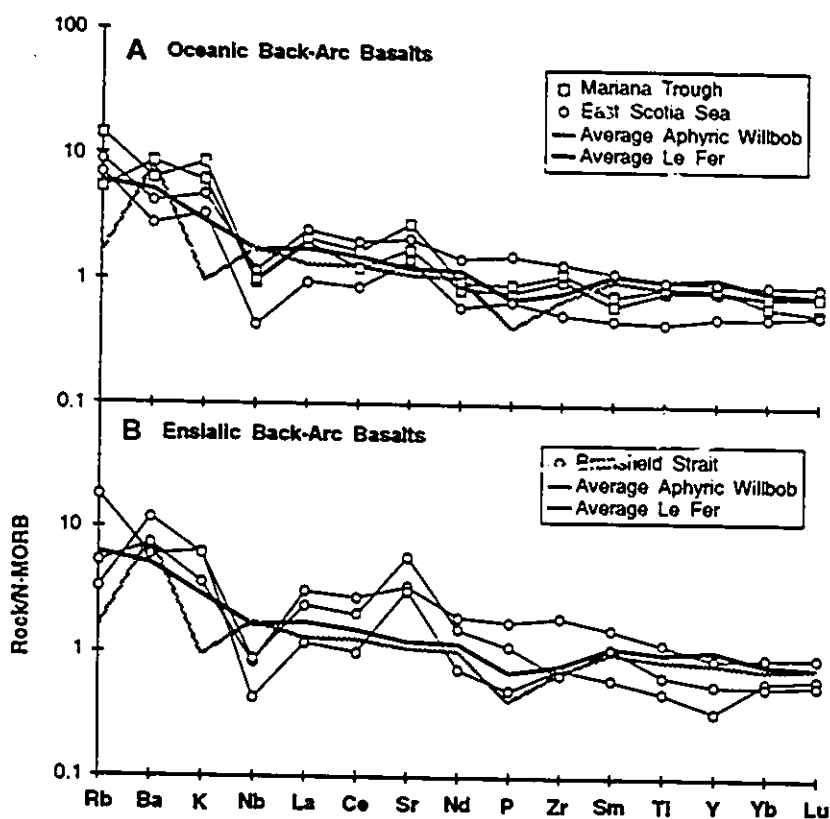


Figure 10.6. Normalized extended trace element plots comparing average abundances in the Le Fer and upper Menihok/Willbob tholeiites to representative examples of ensimatic and ensialic back-arc tholeiites. Data from Saunders and Tarney (1984). Normalizing values are for N-MORB from Sun and McDonough (1989) except for Ba which is from Saunders and Tarney (1984).

due to increasing separation of the spreading ridge and the volcanic arc (Saunders and Tarney, 1984). This arises from the fact that the mantle source to the back-arc basalts is increasingly removed from the site of LIL enrichment of the mantle wedge through dehydration of the subducted crust. In comparison, the LTBS tholeiites are compositionally homogeneous and show no evidence of progressive change in the abundances of LIL elements other than that accompanying fractional crystallization. The tholeiites also lack the strong Nb depletions characteristic of subduction related volcanism. Finally, the metabasalts and metagabbros of the Laporte Group have oceanic tholeiitic compositions (Wardle and Bailey, 1981; van der Leeden *et al.*, 1990) which are inconsistent with an origin as a volcanic arc. Thus there is no obvious candidate for an arc related volcano-sedimentary assemblage preserved in the Proterozoic sequence between the Superior Province and the remobilized Archean hinterland. Consequently, it is considered unlikely that the LTBS is a subduction related volcanic assemblage.

If the direct influence of the subduction process is ruled out, it is possible, albeit unlikely, that the recycling of oceanic crust has indirectly influenced LTBS compositions through the introduction of a sedimentary component into lithospheric or asthenospheric mantle during an earlier episode of subduction. The low ϵ_{Nd} ratios and incompatible element enriched compositions of some OIB, for example, have been attributed to melting of a mantle source which has incorporated a component of subducted lithosphere (cf. Sun and McDonough, 1989). If the LTBS was erupted in an oceanic environment, this type of process would offer an explanation for the isotopic compositions of the LTBS. While such a mechanism cannot be excluded with certainty, there are several arguments which suggest that the crustal component of the LTBS was derived through assimilation of Archean basement rather than through a tortuous routing of sediments through the mantle. The low ϵ_{Nd} ratios of the more evolved of the cycle 2 tholeiites are consistent with prolonged residence times in crustal or sub-crustal magma chambers, but there is no obvious reason why this might be so if the crustal signature was imparted during mantle fusion. The more pronounced Nb depletion of the evolved basalts offers a similar argument since larger Nb anomalies are consistent with the increasing assimilation of sialic material that might be expected to accompany prolonged crystal fractionation in crustal chambers. The

incompatible element enriched nature of the hypothetical subducted sedimentary material would presumably lead to early magmas enriched in incompatible elements and having low ϵ_{Nd} ratio. Exhaustion of the sedimentary component would give rise to incompatible element poor magmas having mantle isotopic signatures. Such progressive compositional variation is apparent in the LTBS as a whole, as in the sequence from the Nimish/lower Menihek alkaline basalts through the transitional Murdoch volcanics to the upper Menihek/Willbob tholeiites. But it is not found in the cycle 2 tholeiitic rocks which exhibit the greatest variation in isotopic composition. Thus it is difficult to reconcile isotopic variations in the tholeiitic rocks with a model in which isotopic and trace element compositions are determined by the declining proportion of a sedimentary component in the mantle source. Finally, it is not certain that the highly enriched crustal signature of recycled sediment, particularly with respect to isotopic composition, could be retained in the face of mantle convection and homogenization. Consequently, it is probable that the crustal component of the LTBS magmas was not contributed by subduction related processes. If this is the case, then the LTBS magmas were erupted through Archean crust of the Superior Province.

Lithostratigraphic Constraints

A detailed examination of lithological and stratigraphic variations in the sedimentary infilling of the Labrador Trough is beyond the scope of this study. Several attributes, however, are pertinent to discussions of tectonic setting, and these are briefly considered here. The stratigraphy of the Knob Lake Group in the central Trough is dominated by the products of two principal volcano-sedimentary cycles (Dimroth, 1978; Wardle and Bailey, 1981; Le Gallais and Lavoie, 1982). In general, both cycles record an upward and eastward progression from shallow marine or terrestrial clastic sediments largely of continental provenance (Seward subgroup and Wishart Formation), through marine clastic and chemical sediments deposited in shallow to moderate water depths (upper Seward/Pistolet/Swampy Bay subgroups and Sokoman Formation), to turbiditic greywacke-siltstone-shale sequences of mixed provenance deposited in a deep water environment (Le Fer/Bacchus and Menihek formations). The latter stage of sedimentation

in the first cycle was followed by re-establishment of shallow water conditions, at least in the western part of the basin, and widespread deposition of shallow marine chemical and clastic sediments (Denault, Fleming and Dolly formations). The equivalent of the Denault dolostone in the northern Trough (Abner Formation) marks the onset of basin enlargement to the north. The initial, shallow water stages of sedimentation in each cycle were accompanied by localized alkaline magmatism, while the terminal, deep water stages were accompanied by voluminous tholeiitic magmatism. There are numerous examples of local unconformities between the Denault/Dolly/Fleming and Wishart formations (cf. Dimroth *et al.*, 1970), but there is no evidence of a widespread deformational event. Thus the 260-290 Ma interval between tholeiitic magmatism of the two sedimentary cycles is unlikely to reflect a major orogenic episode such as closure of a 2.2-2.1 Ga oceanic basin. Instead, this interval probably records marine regression and uplift. In this regard it should be noted that the Denault-Fleming-Wishart sequence is at least locally conformable (Birkett *et al.*, 1990), and hence it is possible that the principal interval of uplift and non-deposition locally occurred prior to formation of the Denault dolostone.

The Doublet and Knob Lake groups share no obviously correlative sedimentary strata, and appear to be in tectonic contact along the Walsh Lake thrust. Consequently their stratigraphic relationship is uncertain. The geochemical data, however, suggest that the Doublet Group is stratigraphically equivalent to the upper part of the second depositional cycle in the Knob Lake Group. This follows from the fact that the upper Menihek and Willbob basalts are comagmatic, while early Murdoch magmatism exhibits some of the alkaline characteristics of the lower Menihek and Nimish volcanics. Some support for this interpretation derives from the occurrence of ironstone-bearing cobble conglomerates in association with the alkaline basalts in each of the three suites. The most likely source of the ironstone cobbles in the Nimish and lower Menihek conglomerates is the intercalated, or immediately underlying, Sokoman iron formation. While there is no evidence regarding the source of the cobbles in the Murdoch conglomerate, a comparable origin seems likely. This suggests that the Murdoch Formation post-dates the Sokoman Formation, and this interpretation is consistent with the correlation of the lower Menihek and lower Murdoch volcanics on the basis of geochemical affinity. Consequently, it is concluded that the Doublet Group is the distal equivalent of the upper Knob Lake Group.

The easternmost lithotectonic component of the central Labrador Trough is the Laporte terrane, which is generally in fault contact with the Doublet terrane to the west, and grades into the high grade gneissic hinterland to the east. The amphibolite grade metasediments, metavolcanics and metagabbros have been interpreted as the highly metamorphosed equivalents of parts of the Knob Lake and Doublet groups (cf. Wardle and Bailey, 1981). In particular, Dimroth (1978) identified several stratigraphic units of the Laporte Group to the south-east of the Wheeler dome which he considered correlative with strata of the Seward subgroup, Le Fer/Bacchus Formation and Murdoch Formation. Dimroth (1978) also noted that the proportion of metasedimentary strata derived from shallow marine sediments increased eastward. This may indicate that the Laporte Group records a transition from distal deep water pelitic sedimentation to proximal shallow water clastic and chemical sedimentation to the east, much as this transition is seen moving westward from the Doublet terrane to Schefferville zone.

The overall structure of the central Labrador Trough is that of a north-west verging fold and thrust belt consisting of an autochthonous sequence formed on the margin of the Superior craton (Schefferville zone) overthrust by north-easterly derived thrust sheets (Ferrum and Howse zones, Doublet and Laporte terranes; Wardle *et al.*, 1990a). The Laporte terrane is transitional into the high grade gneissic hinterland which is thought to represent remobilized Archean crust and infolded Proterozoic supracrustal rocks. Hoffman (1990b) suggested that the Archean component of the hinterland represents a southern splay of the Rae Province which was rotated away from the bulk of the Archean terrane in response to indentation by a Superior craton promontory. It is possible, however, that the hinterland represents, in part, remobilized crust of the Superior Province separated from the main part of the craton by continental rifting. In either case, the Labrador Trough sequence represents a basinal succession deformed and metamorphosed during convergence of the Superior craton and the Archean hinterland. In this regard, the Lac Tudor shear zone may form part of the suture zone between the two Archean blocks adjacent to the central Trough (van der Leeden *et al.*, 1990; Wardle *et al.*, 1990b).

The Seward subgroup of the first depositional cycle is largely restricted to the Ferrum zone of the central Trough, and hence was laid down in what appears to have been a relatively narrow basin developed on the margin of the Superior craton (Wardle and

Bailey, 1981). Subsidence and transgression led to more widespread deposition of succeeding strata, culminating with the Le Fer turbidites. Although more extensive than preceding sedimentation, the distribution of the Le Fer turbidites is also relatively restricted in comparison to that of the cycle 2 succession. The Le Fer Formation and preceding deposits are not exposed in the Howse zone (as defined here) or the Doublet terrane, and are also absent from the northern Labrador Trough. Skulski *et al.* (1993) have suggested that the absence of pre-Abner Formation (equivalent to Denault Formation) strata in the northern Trough reflects the existence of a crustal promontory in the north during much of cycle 1 sedimentation. The onset of carbonate deposition (Denault and Abner formations) records basin enlargement to the north, as well as the reduction of water depth in the central Trough.

The absence of cycle 1 strata in the Doublet terrane may reflect the non-deposition of these units, or this absence may simply be a function of the present erosional level. The answer to this question depends, in part, on the interpretation of the Laporte Group. If the latter sequence is essentially equivalent to the Knob Lake Group, this implies that the central Trough was a two sided marine basin with symmetrical infilling. In this case, it is probable that the stratigraphic equivalents of cycle 1 sedimentation are simply not exposed in the Howse zone and Doublet terrane. This implies that the cycle 1 basin was not as restricted as would appear from the distribution of its component sediments.

The alternative interpretation of the Laporte Group is that this sequence is not correlative with the Knob Lake and Doublet groups. These rocks, for example, may represent the remnants of an Early Proterozoic volcanic arc or back-arc basin developed above an eastward-dipping subduction zone adjacent to the Archean block which now forms the hinterland to the central Trough (Wardle *et al.*, 1990a). There are other possible interpretations of this assemblage, but it is clear that whatever the tectonic configuration, the Laporte terrane was thrust westward onto the Doublet terrane and any intervening supracrustal rocks and oceanic crust were removed. The important feature of this model is that it eliminates the requirement for symmetrical sedimentary infilling. In this case, the eastward progression towards deeper water sedimentation in both cycles may record the transition from continental margin to ocean basin. If this interpretation is correct, the absence of cycle 2 strata in the Doublet terrane may again simply be a function of

exposure. It is also possible, however, that the lower Knob Lake Group was deposited in a continental margin sub-basin that was localized in the Ferrum zone and did not extend to the Doublet terrane.

The foregoing discussions suggest two possible configurations for the depositional basins of the Labrador Trough. In the first model, the sequence records sedimentation in a continental rift basin or basins (cf. Dimroth *et al.*, 1970; Baragar and Scoates, 1981) which would probably have developed on the thinned and weakened margin of the Superior craton (Fig. 10.7A). Marginal rifting would explain the preponderance of marine sediments and submarine basalts, as well as allowing for some degree of asymmetry in the sedimentary infilling. In this model, the Laporte Group would be the eastern equivalent of the Knob Lake Group, while the Doublet Group preserves the axial, deep water igneous/sedimentary facies of the rift basin. The transgressive nature of sedimentation would reflect crustal attenuation and subsidence in the rift zone. The culminating tholeiitic magmatism may record the eventual rupture of the continental lithosphere. But the two sided nature of the sedimentary infilling implies that subsequent spreading was limited.

In the second configuration, the Labrador Trough preserves the transition from continental to oceanic lithosphere (cf. Wilson, 1968; Gibb and Walcott, 1971; Dewey and Burke, 1973). Because of the difference in ages of the two sedimentary cycles, this implies the development of two oceanic basins, presumably with an intervening period of ocean closure. Alternatively, it is possible that only the younger sequence records a continent-ocean transition, and that cycle 1 strata were deposited in a continental rift setting which failed to evolve into an oceanic basin. Some support for this hypothesis comes from the fact that late emplacement and eruption of picritic melts (e.g., the Retty Peridotite and high-Mg Willbob flows) may signify crustal rupture late in the depositional history of the second cycle. Since no picritic rocks have been reported from the first sedimentary cycle, it is possible that this sequence was underlain by continental basement throughout its depositional history. Renewed rifting, either in the same weakened zone or else further to the east, may have led to formation of a 1.9-1.8 Ga ocean basin. Subsidence of the continental margin and marine transgression would result in the overstepping of cycle 2 strata, with local deposition directly on the Superior basement (Fig. 10.7B). The Doublet Group would represent transitional oceanic crust in such a model, while the

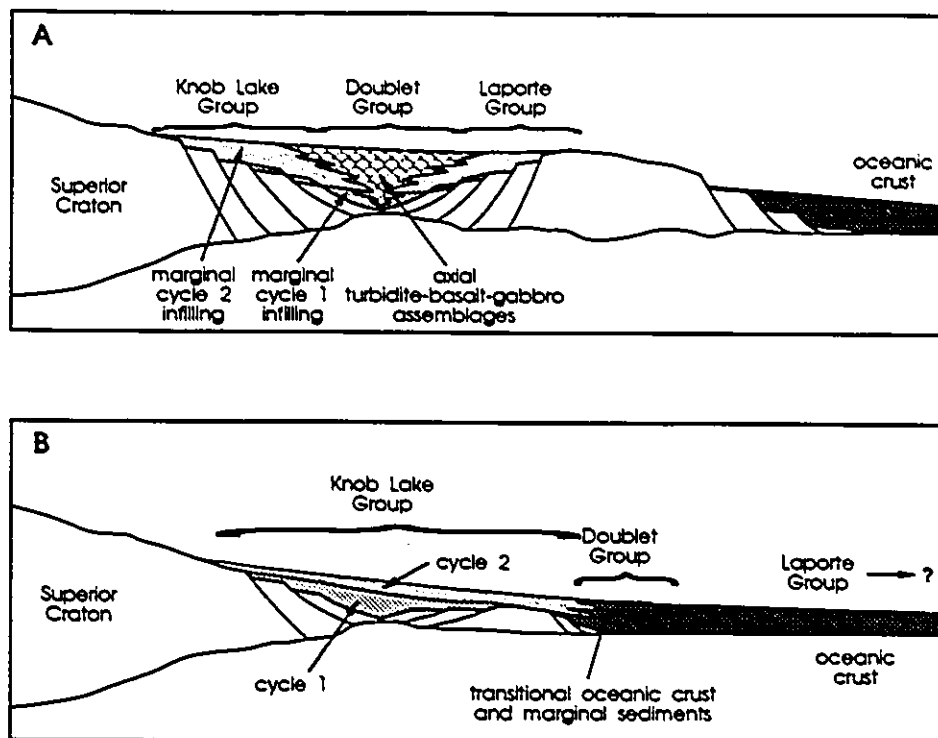


Figure 10.7. Schematic illustration (not to scale) of the development of the central Labrador Trough as (A) continental rift basins and (B) the continental margin-oceanic crust transition.

Laporte Group preserves an offshore volcano-sedimentary sequence of unknown derivation.

LTBS Formation on a Rifted Continental Margin

There are several lithostratigraphic features which suggest that Labrador Trough preserves the products of two stages of continental rifting rather than continental margin-ocean crust transitions. Perhaps the most significant is the occurrence of small bodies of Archean gneissic rocks (e.g., Wheeler dome, Snelgrove arch, and the Boulder, Renia, Moyer and Scattered gneisses) in the eastern domains of the northern and central portions of the Trough which have been interpreted as both paleotopographic basement highs and basement-involved thrust slices (Dimroth *et al.*, 1970; Dimroth, 1978; Wardle and Bailey, 1981; Le Gallais and Lavoie, 1982; Moorhead and Hynes, 1990). Irrespective of their origin, the stratigraphic importance of the Archean bodies lies in the fact that they preserve the transition from continental basement rocks to metamorphosed supracrustal

cover that has been correlated with portions of the Kaniapiskau Supergroup. The Wheeler dome, for example, is mantled by members of the Knob Lake Group to the west, and by strata of the Laporte and Doublet Groups to the east (Dimroth, 1978; Wardle and Bailey, 1981). Similar relationships are apparent in the easternmost domain (Rachel zone) of the northern Trough where the Renia gneiss is overlain by a thick supracrustal assemblage which has been correlated with Kaniapiskau strata to the west (Moorhead and Hynes, 1990). Since the Laporte Group and the equivalent metamorphosed succession in the northern Trough are gradational into the gneissic hinterland, these basement-cover occurrences suggest that much of the supracrustal assemblage preserved in the Labrador Trough was laid down on Archean crust.

The foregoing interpretation is based on the correlation between the Laporte Group (and the stratigraphic equivalent in the northern Trough) and the Knob Lake Group suggested by Dimroth (1978) and Wardle and Bailey (1981). If this correlation is correct, the lithological character of the Laporte Group offers additional support for a rifted continental margin setting. Because the Laporte Group includes metamorphosed sandstones and carbonates indicative of shallow water conditions as well as metapelites, it is unlikely that this sequence can represent offshore deposits laid down in an ocean basin. In addition, Dimroth (1978) noted that the proportion of shallow water metasediments in the Laporte sequence tends to increase to the east. If the Laporte-Knob Lake correlation is correct, the simplest interpretation of the lithological assemblage is that it is the product of deposition on the west-facing margin of the rifted Superior craton. In this case, the Archean gneisses of the hinterland represent, at least in part, remobilized Superior crust.

An additional argument for a continental rift setting comes from the occurrence of localized alkaline volcanic rocks which appear in the stratigraphic record at the transition from shallow to deep water conditions. The Chakonipau, Nimish, lower Menihek and high-Zr Murdoch basalts exhibit subaerial to shallow water lithological characteristics and are accompanied by conglomeratic sediments. This attests to at least local tectonic instability synchronous with alkaline magma activity and marine transgression. The alkaline rocks exhibit the geochemical characteristics of within plate basalts (WPB) typified by continental rift and ocean island volcanism (Fig 10.8, see also section 9.1). Given the interpretations regarding the origin of the crustal component in the LTBS

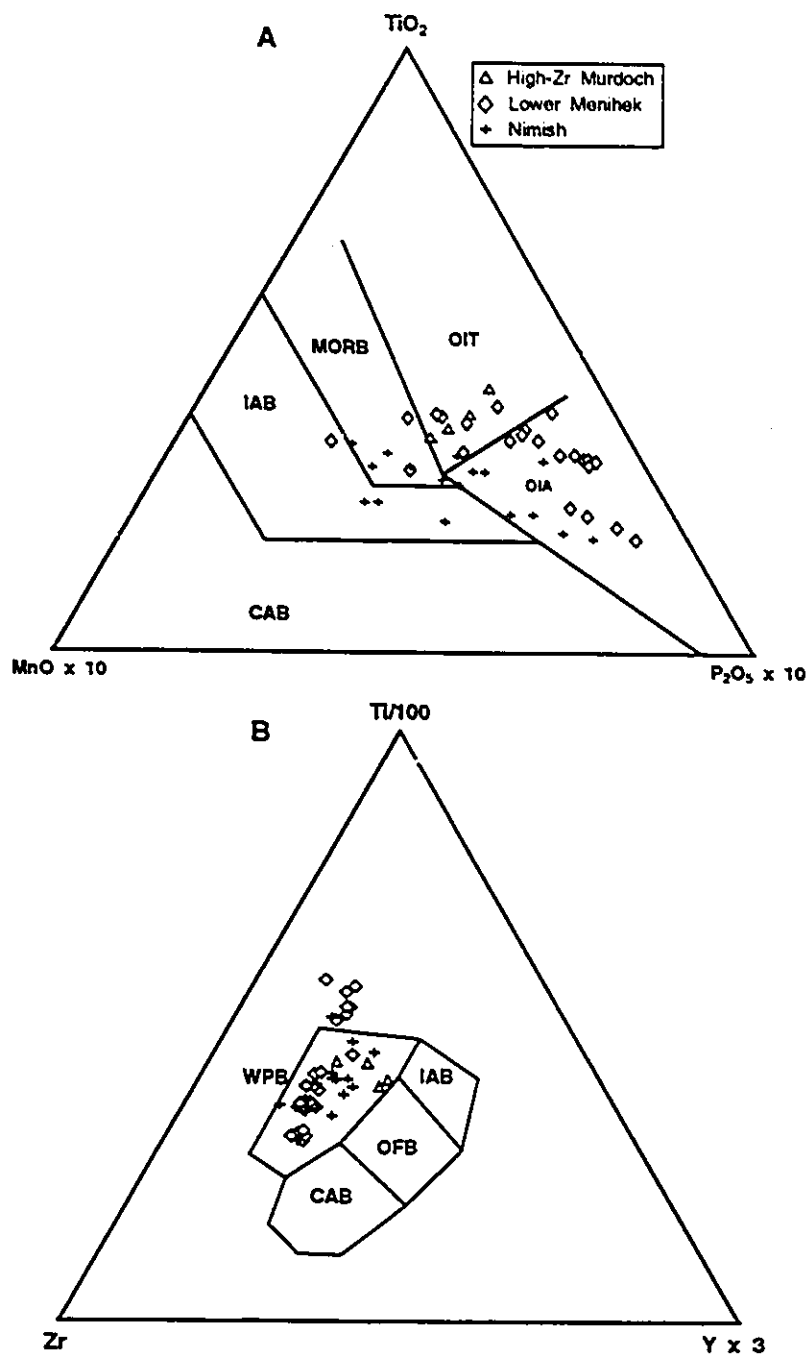


Figure. 10.8. Tectonic discriminant diagrams illustrating the within plate geochemical affinities of the high-Zr Murdoch, lower Menihek and Nimish alkali basalts. (A) TiO_2 - MnO - P_2O_5 discriminant diagram of Mullen (1982). (B) Ti - Zr - Y discriminant diagram of Pearce and Cann (1973). CAB- calc-alkaline basalt, IAT - island arc tholeiite, IAB - island arc basalt, MORB - mid-ocean ridge basalt, OIT - ocean island tholeiite, OIA - ocean island alkali basalt, WPB - within plate basalt, CFB continental flood basalt.

magmas discussed above, the alkaline rocks are unlikely to be the products of ocean island magmatism. It is thus probable that these volcanic sequences are related to continental rifting, as was proposed for the alkaline rocks of the Chakonipau (Wardle and Bailey, 1981) and Nimish (Evans, 1978) formations. The transition from alkalic to tholeiitic volcanism in the LTBS favours such a model since this type of compositional variation is common in continental rift settings (e.g., the Afar triangle; Gass, 1970). Alkalic volcanism was succeeded by turbidite deposition and voluminous and widespread eruption of tholeiitic basalts. These features suggest that tectonic instability and alkaline magmatism mark the onset of crustal attenuation and basin subsidence. Continued rifting, and possibly eventual rupture of the continental basement was accompanied by relatively rapid subsidence, turbidite deposition and production of tholeiitic melts. The oceanic geochemical character of the latter volcanism suggests that it represents the terminal stages of crustal rifting where continental lithosphere had become quite thin. Alternatively, the tholeiitic volcanism may represent the initial stages of production of oceanic lithosphere.

The nature of the compositional variations in the tholeiitic rocks is also more consistent with a continental rift setting than an oceanic environment. The geochemical diversity observed in the Le Fer, upper Menihek and Willbob basalts reflects the fractionation of pyroxene and plagioclase. The Menihek gabbros have also differentiated through fractional crystallization of pyroxene and plagioclase. The sills, however, were not the site of fractionation leading to compositional evolution in the basalts. This is indicated by the fact that the geochemical variation in the chilled margins of the sills is comparable to the basalts. This feature, coupled with the relative homogeneity of the tholeiitic basalts, suggests that fractionation of pyroxene and plagioclase occurred in magma reservoirs not presently exposed.

The cycle 1 and 2 tholeiitic rocks are low-Mg basalts which cannot be representative of a primary magma composition. It was concluded that a period of olivine fractionation was responsible for melt evolution which eventually led to cessation of olivine fractionation and onset of pyroxene + plagioclase crystallization in residual melts. The fact that few, if any, primitive rocks are preserved in the Labrador Trough suggests that there was an underlying mechanism which prevented eruption or emplacement of primitive magmas over most of the depositional history of the Trough. The most likely

mechanism is the density contrast between high-Mg picritic melts and lithospheric rocks (Stolper and Walker, 1980; Carr, 1984). Thus low density continental crust would prevent the ascent of the primitive magmas until reduced melt densities had been achieved through precipitation of olivine. The density contrast between the average Willbob basalt of this study and the high-Mg Willbob flow of Beaudoin and Laurent (1989) offers some support for this interpretation. Calculations using the method of Bottinga and Weill (1970) suggest that the high-Mg flow would have a slightly greater density (2.66 g/cm) than the average Willbob basalt (2.62 g/cm). Rupture of the crust during the late stages of cycle 2 infilling would remove the crustal density barrier and allow ascent of less evolved magmas. This type of process was suggested by Skulski *et al.* (1993) to explain compositional variations in the Hellancourt basalts of the northern Trough and the late emplacement of picritic sills.

If this interpretation is correct, it is probable that primitive melts accumulated at the mantle-crust interface or in deep crustal chambers and underwent a period of olivine fractionation (Fig. 10.9). This may have led directly to precipitation of pyroxene and plagioclase and subsequent eruption and emplacement of residual liquids. More probably, the density decrease resulting from olivine removal would allow residual melts to rise to higher crustal levels where low pressure pyroxene + plagioclase fractionation occurred. Repeated replenishment of smaller, mid-level crustal magma chambers would allow cyclic fractionation episodes and periodic eruption of plagioclase-glomeroporphyritic melts. This would explain the occurrence of several horizons of the latter lithology in the Willbob Formation. It would also explain the fact that not all glomeroporphyritic basalts are highly evolved, and therefore the suite cannot simply represent the end products of a single period of magmatic differentiation. Crustal underplating by primitive melts might also have contributed to the rift-related attenuation of the Superior margin through thermal erosion.

A comparison with the compositional evolution of volcanism in the approximately coeval Cape Smith foldbelt suggests that crustal rupture and separation in the Labrador Trough was at an incipient stage during continental collision and the onset of thrusting from the north-east. This is indicated by the fact that low-Mg basaltic volcanism of the Povungnituk Formation of the Cape Smith belt was succeeded by eruption of the voluminous picrites of the Chukotat Formation. In contrast, primitive magmas appear to have been free to ascend to the surface only in the terminal stages of infilling in the

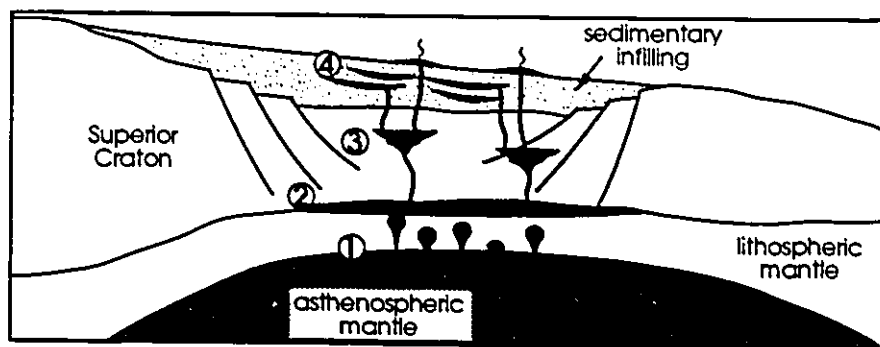


Figure 10.9. Schematic illustration (not to scale) of the hypothesized plumbing system for cycle 1 and 2 tholeiites: (1) generation of primitive magmas through partial melting of upwelling asthenosphere, (2) pooling of the ascending melts at the crustal-mantle boundary followed by olivine fractionation, (3) ascent of less dense residual melts into mid-level crustal magma chambers followed by pyroxene + plagioclase fractionation, and (4) ascent of melts residual to stage 3 into high level sills and to the surface.

Labrador Trough. As noted previously, the absence of any type of high-Mg basalts associated with cycle 1 magmatism may indicate that the continental crust was never sufficiently thinned to allow eruption of primitive lavas.

10.2 Speculation on Rifting Mechanisms and the Undepleted Mantle Source

Rifting Mechanisms

Many continental rift systems have been interpreted to reflect the presence of an underlying mantle plume or hotspot (cf. White and McKenzie, 1989), but this would not seem to have been the case in the Labrador Trough. The reasons for this are two-fold. First, continental rift zones associated with mantle plumes often exhibit a degree of overall uplift during crustal attenuation, and this leads to the subaerial volcanism typical of plateau basalt provinces (White and McKenzie, 1989). Second, tholeiitic volcanism associated with rift zones generally exhibits the geochemical characteristics of OIB (cf. Leeman and Fitton, 1989). This is also commonly true in rift settings where new oceanic crust has been produced (e.g., the Red Sea and Gulf of Aden). This geochemical affinity has been attributed to the fertile nature of the mantle plume, although other interpretations have

been advanced (cf. Leeman and Fitton, 1989). In contrast, the LTBS tholeiites are compositionally closer to MORB than OIB, and clearly represent deep water submarine extrusions. Thus it is apparent that crustal attenuation and rifting in the central Labrador Trough led to basin subsidence and marine transgression rather than uplift. It therefore seems unlikely that rifting of the Superior craton was a response to stresses imposed by a mantle plume. Instead, rifting and crustal attenuation were probably related to tectonic stresses imposed by the movements of bordering lithospheric plates. If this interpretation is correct, then the undepleted mantle source for the alkaline rocks was not a compositionally fertile mantle plume.

Because of the speculative nature of interpreted tectonic configurations in the pre-2.2 Ga Trans-Hudson Orogen, the source of lithospheric tension which led to rifting of the Superior craton is impossible to determine. Hoffman (1990b) suggested an Andaman Sea-type model in which the cycle 2 basalt-gabbro-turbidite assemblage of the Labrador Trough developed in a pull-apart basin in response to indentation of the Rae Province by the Superior craton (Fig. 10.10A). The pull-apart basin is thought to have developed along the western edge of the Rae Province, and was initially separated from the Superior Province by intervening 2.1-1.8 Ga oceanic lithosphere. The basalt-gabbro-turbidite sequence was subsequently thrust westward onto cycle 1 and 2 (foredeep ?) sediments deposited on the rifted margin of the Superior craton as east-directed subduction of the intervening oceanic crust led to oblique collision with a rotated Archean block detached from the Rae Province. By analogy with the Andaman Sea, which is situated above the subducted Indian plate (Dewey *et al.*, 1989), the tholeiitic basalt-gabbro complexes of the Labrador Trough should exhibit supra-subduction zone geochemical characteristics (Hoffman, 1990b). As discussed earlier, however, there is no evidence of supra-subduction zone compositional traits in the LTBS tholeiites.

To explain the absence of supra-subduction zone geochemical characteristics in the Hellancourt basalts of the northern Trough, Skulski *et al.* (1993) favoured a model in which oblique convergence of the Superior craton and the Rae Province triggered the formation of a pull-apart basin(s) on the Superior margin (Fig. 10.10B), rather than adjacent to the Rae Province. In this model, the cycle 1 assemblage was deposited on the rifted margin of the Superior Province and records the transition to 2.1 Ga oceanic crust.

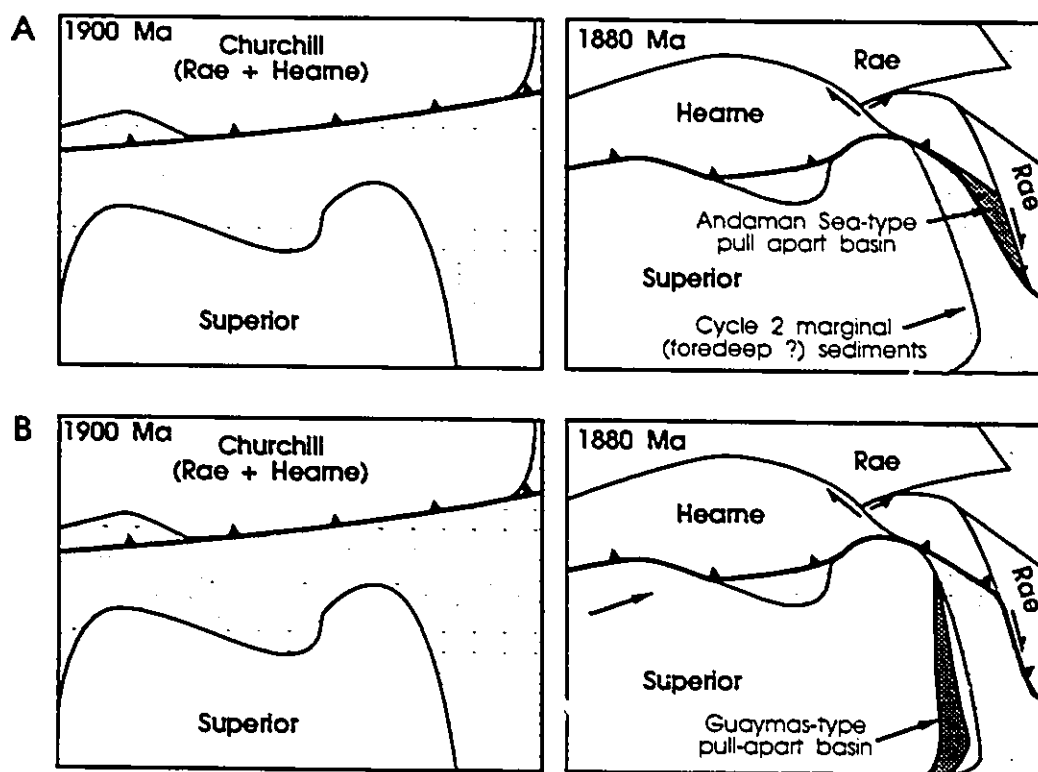


Figure 10.10. (A) Formation of the cycle 2 basalt-gabbro-turbidite assemblage in an Andaman Sea-type pull-apart basin developed along the leading edge of an Archean block of the Rae Province as proposed by Hoffman (1990b). The southern splay of the Rae Province represents an extruded block rotated clockwise in response to collisional indentation by the Superior Province. (B) Formation of the entire cycle 2 sequence in a continental margin pull-apart basin developed in response to Superior-Rae collision, as proposed by Skulski *et al.* (1993). In both models, continued rotation of the southern splay of the Rae Province leads to collision with the western edge of the Superior plate. Heavy solid line indicates subduction zone with teeth indicating polarity. Modified after Hoffman (1990b).

Skulski *et al.* (1993) suggest that localization of the younger basins on the margin of the Superior Province was the result of the inherent weakness of the thinned sialic crust relative to 2.1 Ga oceanic lithosphere. This model is somewhat analogous to the interpreted origin of the Guaymas Basin of the Gulf of California (Larson, 1972; Atwater and Molnar, 1973) except that the inferred direction of subduction is to the east, beneath the southern arm of the Rae Province (i.e., away from the marginal pull-apart basin). The Gulf of California model shares the structural style of the Andaman Sea in that spreading

is thought to occur along short volcanic ridges joined by transform faults. The principal difference between the Andaman Sea and the Guaymas Basin lies in the fact that subduction of the Pacific plate beneath Baja California ceased as much as 10 Ma prior to onset of spreading in the Guaymas Basin. (Dickinson and Snyder, 1979; Saunders and Tarney, 1984), whereas the subduction of the Indian plate beneath the Andaman Sea is synchronous with spreading. As a consequence of the arc switch-off in the Gulf of California, the volcanic products of spreading centre magmatism are more MORB-like than typical ensialic back-arc basins.

The Guaymas Basin model is attractive in that the short spreading ridges joined by transform faults, coupled with a high rate of sediment influx, have resulted in the formation of oceanic lithosphere dominated by sediment-sill complexes rather than by submarine basalts (Einsele, 1986). Einsele (1986) interpreted this situation to reflect lateral emplacement of melts within soft sediments on the basin floor when magma pressure exceeded lithostatic pressure and the tensile strength of the sediments. The emplacement of the melts leads to thermal induration of enclosing sediments, and causes subsequent magmas to rise to higher levels to again invade soft sediments. If the sedimentation rate is high relative to the periodicity of magma pulses, this gives rise to a sill-sediment complex without extrusion of lavas. Conversely, if the sedimentation rate is low, sequential magma pulses will eventually be erupted onto the floor of the basin. As spreading progresses, and the distance between the spreading zone and the rifted continental margins increases, sediment supply drops. This leads to eruption of pillowed basalts on to the floor of the basin.

The sill-sediment complex of the Guaymas Basin is analogous to the cycle 2 basalt-gabbro-turbidite assemblages in the central Trough, since it is only in the uppermost and easternmost portions of the sequence that basalt extrusion overwhelmed turbidite deposition. Thus the gabbro-turbidite sequences in the Howse zone, and possibly the Le Fer assemblage in the Ferrum zone, may reflect relatively rapid sediment accumulation which largely prevented eruption of tholeiitic basalts. Because significant thicknesses of basalts are found at three levels in the Le Fer Formation of the Ferrum zone, waning of the sediment supply may have occurred periodically in the cycle 1 basalt-gabbro-turbidite assemblage. In contrast, a reduction in the rate of sedimentary influx occurred only late in

the depositional history of the Howse zone, and this resulted in the eruption of the upper Menihek basalts. The axial nature of the Doublet Group would place these units furthest from the sediment supply. The attendant low sedimentation rate would result in the general eruption rather than emplacement of ascending magmas. A generally low rate of sedimentation in this domain is consistent with the predominantly igneous nature of the Doublet Group. The Thompson Lake shales, however, would seem to necessitate an interval in which sedimentation rates were increased, or in which the rate of magma pulse generation was reduced.

Differences in the rate of sediment supply might also explain the abundance of Menihek intrusions exhibiting anorthositic and core phase lithologies in the cycle 2 sill-sediment complex of the Howse zone. This situation contrasts with that of the Doublet terrane, where well developed anorthositic and core phase glomeroporphyritic gabbro is rare or absent. Glomerocryst-laden magmas in the Howse zone would generally form sills due to the high rate of sediment influx in this domain. As a consequence, these melts would be subject to the forces of grain dispersive pressure during intrusion, and this would lead to the observed pattern of lithological variation, particularly in the development of the anorthositic phase. A lower rate of sedimentation in the Doublet terrane would result in the extrusion of glomeroporphyritic (and aphyric) lavas onto the floor of the basin, rather than the emplacement of the melts within the sedimentary strata. The combination of the absence of an upper conduit boundary and convective or possibly turbulent effusion, would remove the potential for flow differentiation when these melts reached the surface. This would also tend to remove any flow structure developed during ascent, and homogenize the lavas in terms of the glomerocrysts. Thus the preponderance of glomeroporphyritic magmas ascending through the volcano-sedimentary infilling of the Doublet terrane would be erupted as basalt flows exhibiting fairly uniform glomerocryst distributions. In contrast, the higher sedimentation rates of the Howse zone would inhibit magmas from reaching the surface. Here the preponderance of glomeroporphyritic magmas ascending through the sedimentary infilling would be emplaced as sills, and therefore undergo flow differentiation to give the observed glomerocryst distributions.

Oblique convergence of the Superior craton with a continental land mass to the north-east offers a potential source of lithospheric tension leading to 1.9 Ga Superior

Province rifting. The result would be the formation of pull-apart basins in which spreading was dominated by short volcanic axes separated by transform faults. Initially rapid sedimentation would give rise to sediments and sediment-sill complexes. Lower sedimentation rates in axial portions of the basins, as well as across larger areas during the terminal stages of basin development, would give rise to the culminating basalt sequences. Such a model is preferable to one in which lithospheric extension is caused by a mantle plume for reasons outlined earlier. Both Hoffman (1990b) and Skulski *et al.* (1993), however, interpret the lower Knob Lake Group as a continental-oceanic transition assemblage. For reasons outlined earlier, it seems more likely that both cycle 1 and 2 strata represent the products of deposition in relatively narrow ensialic rift basins in which spreading and formation of oceanic lithosphere was limited. If this interpretation is correct, then the initial stage of rifting of the Superior craton is presumably unrelated to collisional indentation of the Rae Province because this is postulated to have occurred after 1.9 Ga (Hoffman, 1990b). One possible explanation is that initial, 2.2-2.1 Ga rifting was a response to non-uniform rates of subduction, or piecemeal subduction, of the Manikewan oceanic lithosphere (Hoffman, 1990b) beneath the Churchill Province during convergence of this domain and the Superior craton. The onset of more uniform subduction along the plate boundaries might then have removed the extensional stresses on the Superior craton and terminated cycle 1 rifting prior to development of an oceanic spreading centre. Alternatively, initial rifting may be related to opening of the Manikewan ocean basin. In either case, subsequent collisional indentation of the Churchill Province and resultant convergence of the Rae Province might have triggered renewed rifting on the Superior margin. It must again be emphasized, however, that any models regarding the source of extensional stress along the Superior margin are highly speculative.

Nature of the Undepleted Mantle Source

The foregoing discussion suggests that the predominance of T-MORB-like tholeiites in the Labrador Trough is inconsistent with a rift origin through upwelling of a fertile mantle plume. Consequently it is unlikely that such a plume represents the undepleted mantle source for the alkaline rocks of the LTBS. The fact that there is a transition from alkaline to tholeiitic magmatism in the LTBS suggests that the undepleted

mantle source was either consumed, replaced or overwhelmed during the later stages of magmatic activity. This feature is compatible with models in which the undepleted mantle source takes the form of either a fertile subcontinental mantle, or fertile zones in the otherwise depleted asthenospheric mantle. If the source of the alkaline rocks was predominantly lithospheric mantle, melting of this reservoir during the initial stages of rifting would produce the alkaline magmas. As rifting and lithospheric attenuation progressed, the site of fusion would shift to the upwelling asthenospheric mantle thereby giving rise to increasingly MORB-like melts. Alternatively, if the source of the alkaline rocks was fertile zones within an otherwise depleted asthenospheric mantle, this incompatible element enriched material would be preferentially sampled during initial magma generation due to lower solidus temperatures. As the fertile zones were consumed, or as the degree of melting increased, magma generation would progressively involve greater proportions of depleted mantle material with higher solidus temperatures. This would lead to a transition from alkaline to tholeiitic magmas as rifting continued.

Although any interpretations regarding the nature of the undepleted mantle source are highly speculative, the contrast in magmatic styles and compositions suggests that the second model may be more probable. For example, the localized nature and limited quantities of alkaline rocks, coupled with their subaerial to shallow submarine origin, would be consistent with derivation through small degrees of partial melting of a heterogeneous asthenospheric mantle source during the initial stages of rifting and crustal attenuation. In contrast, the voluminous production of homogeneous tholeiitic magma requires much larger degrees of melting, and this in turn necessitates the involvement of a much larger portion of the depleted asthenospheric mantle. This could be accomplished by upwelling of asthenospheric mantle to melting depths as the continental lithosphere was stretched and thinned, possibly to the point of rupture. An asthenospheric source for all LTBS melts is consistent with the theoretical models of McKenzie and Bickle (1988) who calculated that magma generation during lithospheric extension is confined to the upwelling asthenosphere. Nevertheless, it is possible that the subcontinental mantle contributed to the compositions of early magmas through partial melting, or by magma contamination during ascent.

10.3 Tectonic Evolution of the LTBS

Lithospheric extension within the Superior craton that resulted in the formation of the cycle 1 depositional basin commenced prior to 2169 Ma (Fig. 10.11A). The cause of the extensional regime is unknown, but it was probably related to the movements of adjacent lithospheric plates rather than to a mantle plume. The cycle 1 sedimentary basin was localized within or to the east of the Ferrum zone of the central Labrador Trough, and did not extend to the northern part of the belt. Skulski *et al.* (1993) have suggested that this localization reflects the presence of an east-facing crustal promontory in the north. It is also possible that the limited northern extent of the basin was due to greater lithospheric thickness and/or greater crustal competency in the north. The fact that all sedimentary strata deposited after the basal Seward sandstones are marine in origin suggests that rifting was also localized on the cratonic margin. Thus it is possible that a *ca.* 2.2 Ga oceanic basin was present to the east of the Superior Province prior to development of the cycle 1 pull-apart basin.

The initial sedimentary products (Seward subgroup) consisted of fluviatile arkoses and sandstones which are thought to have accumulated in a continental rift basin (Wardle and Bailey, 1981). Deposition of this sequence was preceded and accompanied by mildly alkaline volcanism (Chakonipau Formation) which appears to have been related to the onset of crustal attenuation and basin subsidence (Wardle and Bailey, 1981; Le Gallais and Lavoie, 1982). Insufficient geochemical data are available from the cycle 1 alkaline rocks for comparison with the products of later alkaline volcanism, but it seems likely that the Chakonipau basalts reflect generation of small quantities of incompatible element enriched magmas from an undepleted mantle source. By analogy with cycle 2 alkaline magmatism, the Chakonipau volcanics may represent the products of the initial stages of partial melting in the upwelling heterogeneous asthenospheric mantle.

Seward subgroup deposition was followed by one or two transgressive marine sedimentary cycles (Pistolet and Swampy Bay subgroups) formed in sub-basins within the rift zone (Dimroth, 1978; Le Gallais and Lavoie, 1982). Transgressive cycle 1 sedimentation culminated with more widespread deposition of the Le Fer turbidites, which were accompanied in the Ferrum zone by emplacement of tholeiitic gabbro sills and extrusion of comagmatic tholeiitic basalts (Fig. 10.11B). The absence of the Le Fer basalts

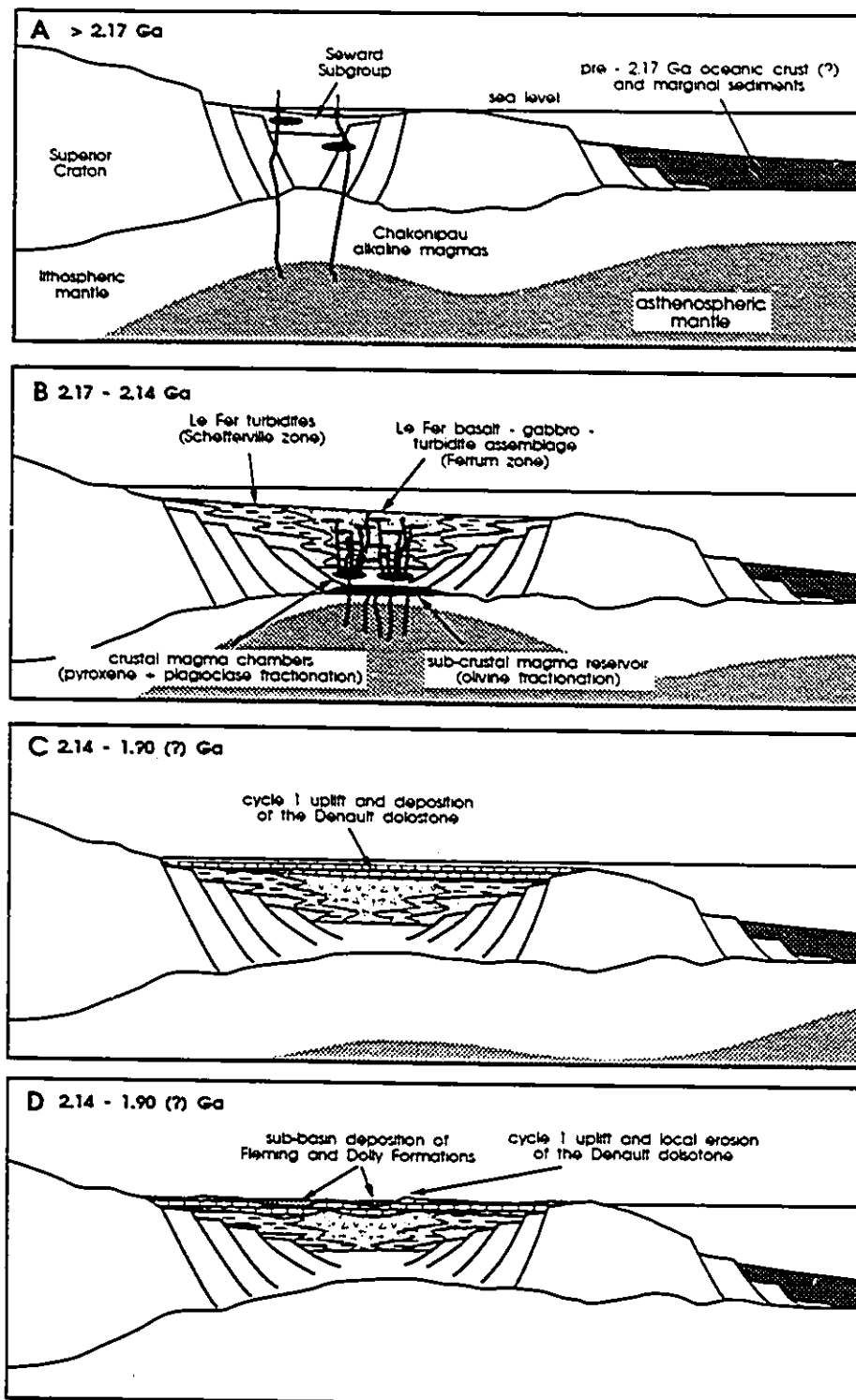


Fig. 10.11. Schematic illustration (not to scale) of the evolution of the first depositional cycle of the Labrador Trough. The approximate timing of major events are indicated, although it should be noted that these events were probably temporally transgressive across the basin, as is the case for cycle 2 strata. See text for explanation.

and gabbros in the Schefferville zone suggests that the axis of the basin at this time, as well as the locus of magma generation, was at least as far east as the Ferrum zone. Because cycle 1 strata are not exposed in the Howse zone or Doublet terrane, however, it is possible that the centre of volcanic activity was further east. If the Laporte terrane includes the metamorphosed equivalents of the lower Knob Lake Group, a more easterly centred axis is likely.

The predominance of deep water sediments and submarine basalts in the Le Fer Formation indicates that the cycle 1 sedimentary basin was undergoing relatively rapid subsidence during Le Fer deposition. This suggests that extensional stresses led to significant crustal attenuation during Le Fer deposition. The widespread introduction of tholeiitic melts as sills and flows was probably the result of upwelling and decompressive partial melting of asthenospheric mantle in response to lithospheric thinning. As discussed earlier, the absence of primitive, high-Mg basalts indicates that primary cycle 1 tholeiitic melts were trapped at depth in the crust or at the crust-mantle interface where they underwent olivine fractionation. Less dense residual melts subsequently ascended to magma chambers where plagioclase, clinopyroxene and orthopyroxene were precipitated. Because the geochemical variations in the Le Fer gabbros were not examined in detail in this study, these intrusions may be the site of the second stage of crystal fractionation. By analogy with the Menihek sills, however, it is more likely that the plagioclase + pyroxene crystallization occurred in mid-level to shallow crustal chambers. The Le Fer sills would then represent residual melts, equivalent in composition to the basalts, which were emplaced at high levels in the sedimentary pile. A more detailed geochemical study of the sills would allow evaluation of this hypothesis.

The Le Fer basalt-gabbro-turbidite assemblage of the Ferrum zone differs from its cycle 2 counterpart in the Howse zone in that basalts appear to have erupted episodically throughout Le Fer sedimentation. This contrasts with the situation in the Howse zone where tholeiitic basalts are present only in the uppermost portion of the Menihek Formation. If the interpreted origin of the sill-sediment complexes is correct, the volcanic horizons imply that the sediment supply to the Le Fer basin waned periodically, thereby allowing the tholeiitic lavas to reach the surface of the sedimentary pile. Alternatively, the rate of magma pulse generation may have increased periodically, with the same result.

Both possibilities may reflect changes in the rate of lithospheric attenuation and basin subsidence, since these factors would presumably influence basin configuration and hence sediment supply, as well as control the extent of asthenospheric upwelling and hence magma supply.

The isotopic data for the Le Fer basalt sample indicates that continental crust has contributed material to at least some of the Le Fer magmas. This is confirmed by the isotopic composition of the Le Fer granophyre reported by Rohon *et al.* (1993). It has already been suggested that the isotopic signatures of both cycle 1 and 2 tholeiitic suites reflect the fact that their respective magmas were emplaced within, and erupted through, Superior basement. The two suites show small differences in compositions, particularly in terms of incompatible trace element contents, which were interpreted to reflect slightly different melt component compositions during formation of the primary magmas. These differences are relatively minor, however, and the general range of primary magma compositions was probably quite similar. But some cycle 2 tholeiitic rocks (e.g., the Retty Peridotites) are clearly far more primitive than the most Mg-rich Le Fer sample. Although this may simply be a function of the present level of erosion (i.e., primitive Le Fer flows are currently buried below cycle 2 strata), it is also possible that crustal attenuation proceeded further during cycle 2 rifting than during Le Fer magmatism. This could be accomplished by the maintenance of the crustal density barrier throughout cycle 1 evolution, but crustal failure during the terminal stages of cycle 2 evolution. The rupture of the crust late in the depositional history of the Labrador Trough would allow ascent of more primitive melts in the axial regions of the cycle 2 basin, thereby forming the ultramafic Retty sills and the picritic flows of the Willbob Formation. The failure of cycle 1 extension to breach Archean basement might also offer an explanation for the apparent difference in isotopic composition of cycle 1 and 2 tholeiites. For example, the failure of rifting during the first depositional cycle may reflect the greater thickness of Archean basement during the early stages of development of the Trough. A greater thickness of sialic material might then result in larger degrees of contamination of the Le Fer magmas during ascent, and this would produce the lower ϵ_{Nd} ratios observed in the two samples.

Further data from the Le Fer suite, however, is necessary to confirm the apparent isotopic differences between cycle 1 and 2 tholeiites.

The Le Fer Formation is overlain by the Denault dolostone in the central Labrador Trough (Fig. 10.11C). This unit records at least local uplift of the cycle 1 sequence, marine regression, and the virtual cessation of igneous activity. The Le Fer-Denault transition would thus appear to mark the end of extensional stresses within the cycle 1 basin. The Mistamisk rhyolite appears to be correlative with late Le Fer magmatism, and therefore the 2142 Ma age of this unit (Dressler and Krogh, cited in Clark, 1984) may record the onset of uplift and regression (Fig. 10.11D). As noted earlier, there is no obvious evidence of a major period of orogenesis and deformation in the Le Fer-Denault-Fleming/Dolly-Wishart sequence of the Schefferville, Ferrum and Howse zones, although there are numerous local sub-Wishart unconformities which have traditionally been interpreted as erosional or non-depositional stratigraphic breaks. Wardle and Bailey (1981) suggested that the western part of the Trough preserves a post-Fleming/Dolly period of regional uplift, but sedimentation was uninterrupted in the Laporte terrane to the east. The latter suggestion seems unlikely given that the time interval involved in the inter-cycle hiatus in the western part of the basin may be as much as 240 Ma. It seems more probable that the entire cycle 1 sequence and underlying basement underwent a period of isostatic, essentially non-deformational uplift, presumably starting during the late stages of Le Fer deposition. This would lead to deposition of the Denault dolostone during uplift and marine regression, and local deposition of the Fleming and Dolly formations in sub-basins developed on the uplifted cycle 1 succession. The occurrence in some locations of apparently conformable relationships between the Denault, Fleming and Wishart formations may indicate that uplift was regionally diachronous. Birkett *et al.* (1990) for example describe local transitions in which the lower part of the Fleming chert breccia (adjacent to the Denault dolostone) is cemented by dolomite, while the upper part (adjacent to the Wishart quartzite) consists of chert breccia in a matrix of fine sand with chert $\pm$ carbonate cement. This may indicate that the principal interval of non-deposition occurs between the Le Fer and Denault formations in these locations.

The interval of inter-cycle uplift was terminated by marine transgression and deposition of the Wishart quartzite in shallow water conditions (Wardle and Bailey, 1981;

Fig. 10.12A). The siliciclastic sediments overstep cycle 1 strata and therefore record basin enlargement to the west (and presumably to the east). In addition, the Allison Formation in the northern Trough is the interpreted correlative of the Wishart quartzite (Wares and Goutier, 1990), and the deposition of the Abner/Allison sequence thus records basin enlargement to the north. The onset of subsidence in the north appears to have commenced at an earlier stage in the eastern part of the basin where the Abner dolostone preserves the earliest strata formed in this part of the Trough. As this sequence is correlative with the Denault dolostone, but also appears to be conformable with the overlying lower Baby Formation (Clark and Thorpe, 1990), it is probable that early subsidence in the north was coincident with the late stages of uplift in the central part of the belt.

The Wishart quartzite is conformably overlain by the Sokoman iron formation, and the latter is intercalated with the Nimish alkaline volcanics. The Wishart-Sokoman-Nimish sequence thus records renewed rifting, subsidence and alkaline magmatism in the western part of the central Trough. This renewal of crustal extension may have been related to collision between the Superior and Churchill provinces as suggested by Hoffman (1990b) and Skulski *et al.* (1993). The resultant stresses presumably led to development of pull-apart basins which were localized in the weakened zone represented by the cycle 1 volcano-sedimentary basin. The equivalent of the Wishart-Sokoman-Nimish sequence has not been recognized in the Doublet terrane, but the slightly alkaline basalts in the basal portion of the Murdoch Formation probably represent magmatic activity associated with the early stages of renewed lithospheric extension. The cycle 2 succession is demonstrably transgressive to the west; formation of the Menihek sill-sediment complex in the Howse zone at 1884 Ma predates Sokoman deposition and eruption of the Nimish alkali basalts at 1878 Ma in the Schefferville zone (Findlay *et al.*, 1995). The equivalent alkaline volcanism in the Howse zone thus developed prior to 1884 Ma, and it is reasonable to assume that the correlative lower Murdoch alkaline basalts are even older. The axis of the cycle 2 basin was thus located at least as far east as the Doublet terrane, and sedimentation and early alkaline volcanism progressed laterally and temporally away from this central zone.

The alkaline magmatism is thought to reflect partial melting of incompatible element enriched zones within the asthenospheric mantle. Assimilation of Archean crust

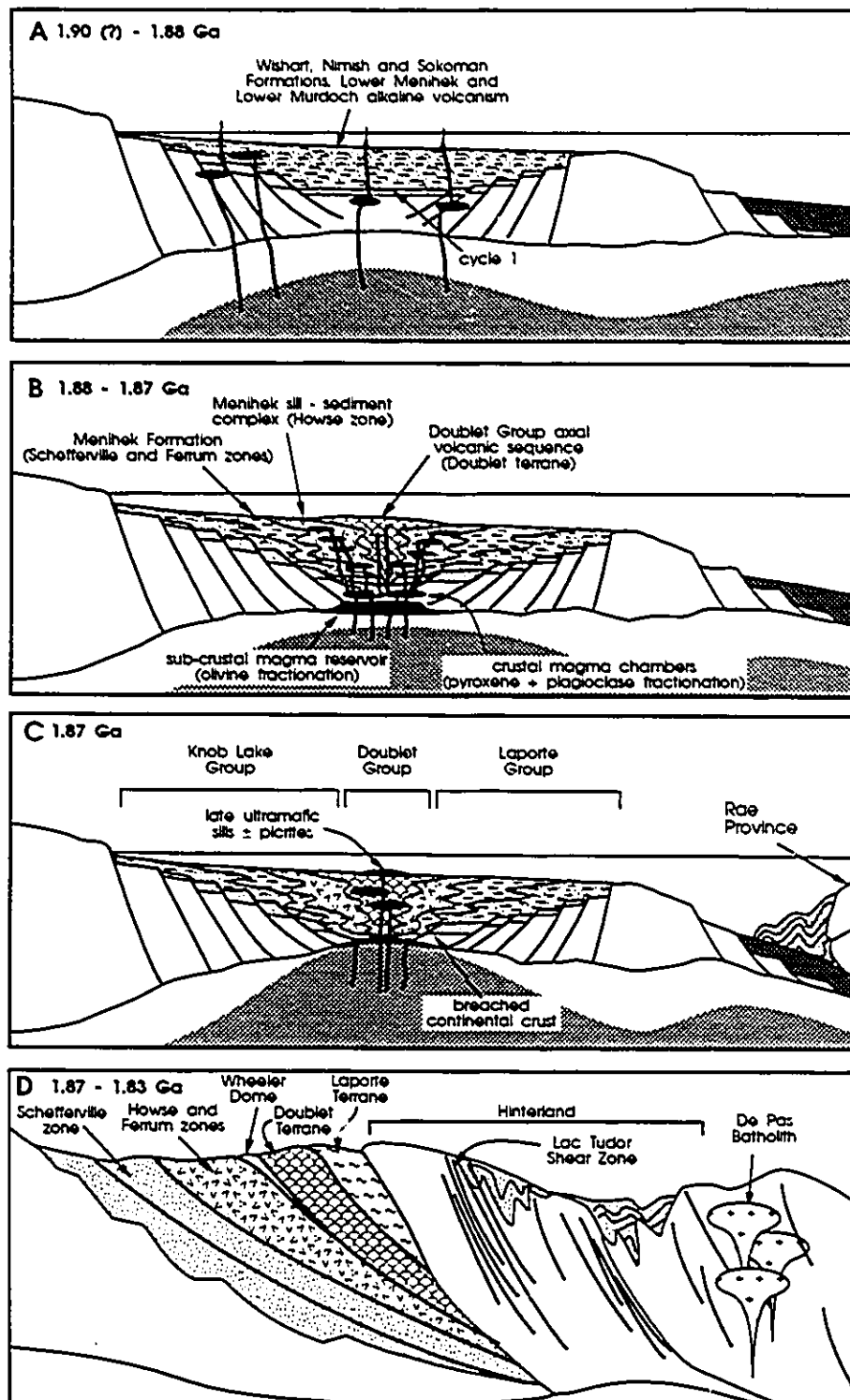


Fig. 10.12. Schematic illustration (not to scale) of the evolution of the second depositional cycle of the Labrador Trough. The approximate timing of major events are indicated, although it should be noted that these events were temporally transgressive across the basin. See text for explanation.

and melting of the depleted mantle matrix also contributed to the compositions of the resultant magmas, although the proportion of melt derived from the latter source appears to be variable among the exposed products of the three alkaline suites. It is likely that this variation reflects progressively greater degrees of partial melting in the asthenosphere during evolution of each suite such that a range of magma compositions was produced, if not preserved. This is suggested by the transitional nature of the Nimish and Murdoch volcanics, both of which range from alkaline or mildly alkaline to subalkaline in geochemical affinity.

The compositional differences among the alkaline suites can thus be interpreted as reflecting gradations in the degree of asthenospheric melting. This is consistent with the proposed rifting model in that as crustal attenuation and asthenospheric upwelling progressed, the degree of decompressive partial melting would increase. Initial melt generation would preferentially sample the undepleted zones in the asthenosphere, thus producing small quantities of highly alkaline magmas. As rifting and mantle upwelling continued, progressively larger degrees of partial melting would sample both fertile and depleted asthenospheric mantle. This would give rise to the transition from highly alkaline to mildly alkaline magmas. The transition to tholeiitic compositions could reflect: (1) exhaustion of the fertile mantle zones, or (2) sufficiently large degrees of partial melting such that the geochemical signature of the fertile zones was completely diluted, or both.

The rifting model also accommodates the transgressive nature of sedimentation and magmatism. The earliest rift-related volcanism would occur in the axial portion of the basin where crustal attenuation would be greatest, and where the upwelling asthenospheric mantle first reached melting depths (Fig. 10.13A). With continued rifting, the central portion of the upwelling mantle dome would undergo progressively larger degrees of decompressive partial melting as the flanks of the mantle dome entered the melting zone (Fig. 10.13B). Thus alkaline magmatism would migrate away from the axis of the basin, while axial alkaline rocks would be succeeded by widespread eruption of tholeiitic rocks contemporaneous with the flanking alkaline volcanism. Progressive crustal attenuation would also lead to transgression of sedimentary facies away from the rift axis as the basin subsided.

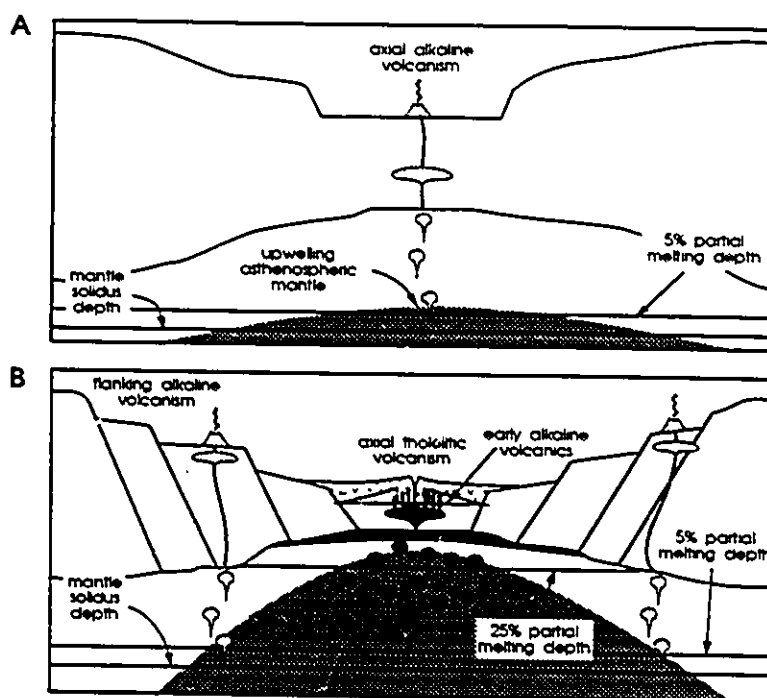


Figure 10.13. Schematic illustration (not to scale) of hypothetical magmatic evolution in a continental rift zone. (A) Initial rifting and asthenospheric upwelling cause decompressive partial melting in the mantle once the solidus depth is achieved. Small degrees of partial melting preferentially sample fertile mantle zones and give rise to localized alkalic volcanism. (B) Continued crustal attenuation and mantle upwelling leads to larger degrees of partial melting which samples both fertile mantle zones and the depleted mantle matrix. This gives rise to more widespread and voluminous tholeiitic volcanism in the axial zone of the rift. Small degrees of partial melting on the flanks of the upwelling mantle dome gives rise to contemporaneous alkaline volcanism on the rift margins. These variations may produce a transitional alkaline to tholeiitic volcanic sequence both stratigraphically and laterally.

The fact that a complete sequence from highly alkaline to tholeiitic basalts has not been identified in any single lithotectonic domain may be a function of exposure. Alternatively, the absence of a complete sequence may be related to the differentiation processes acting on the alkaline magmas. As discussed earlier, the alkaline rocks exhibit large degrees of compositionally variability which imply complex differentiation histories. With the possible exception of the high-Mg Murdoch basalts reported by Gebert (1988), there are no obvious examples of primitive alkaline volcanics. This suggests that the alkaline melts underwent some degree of fractional crystallization in sub-volcanic or crustal magma

chambers, as illustrated in Figure 10.12A. This interpretation is supported by the occurrence of ultramafic Nimish sills in the Schefferville zone. Consequently it is possible that early, highly alkaline magmas differentiating in crustal reservoirs were diluted by continuous injections of less alkaline melts, and only the resultant hybrid liquids were erupted. Mixing of alkaline magmas of variable composition might also explain the non-constancy of incompatible element ratios and the large range of element concentrations among samples from the suites. An third possible explanation may be that the small quantities of highly alkaline melts produced during the initial stages of mantle melting rarely reached the surface without solidifying during ascent through thick, relatively cool continental crust. Continued thinning and heating of the crust as rifting progressed, coupled with the more voluminous production of less highly alkaline melts, would increase the chances of some proportion of the less alkaline magmas reaching the surface.

Alkaline magmatism in the central Trough was accompanied by tectonic instability, local uplift and deposition of ironstone-bearing conglomerates. These features appear to mark the onset of more rapid crustal attenuation and basin subsidence. In the western part of the basin, the Sokoman Formation was succeeded by deposition of the Menihek turbidites (Fig. 10.12B). The Menihek Formation of the Schefferville and Ferrum zones is often poorly exposed, but regional maps suggest this sequence is devoid of gabbroic intrusions in these domains. This is consistent with the restricted occurrence of the upper Menihek tholeiitic basalts in the Howse zone. The absence of a sill-sediment complex comparable to that of the Howse zone may indicate that the lithosphere underlying the Schefferville and Ferrum zones was insufficiently attenuated to promote large scale partial melting of the asthenosphere. In contrast to the more westerly domains, the Howse zone is heavily intruded by cycle 2 tholeiitic sills. The Menihek sills are comagmatic with the upper Menihek and Willbob basalts, and were emplaced within the sedimentary pile between 1884 and 1874 Ma (Machado, 1990; Findlay *et al.*, 1995). As discussed earlier, the development of the sediment-sill complex probably attests to fairly rapid sedimentation relative to the rate of magma pulse generation. The emplacement of glomerocryst-laden melts into the sedimentary pile allowed development of the anorthositic and core phase lithologies through flow differentiation. But the occurrence of the upper Menihek basalts at

the top of the sill-sediment complex signifies a late decline in the sedimentation rate or an increase in the rate of magma production.

The sill-sediment complex that characterizes the Howse zone did not develop in the Doublet terrane, although some Doublet sills were emplaced within the strata of each of the three formations exposed in this domain. This suggests that the rate of sedimentation in the central parts of the basin was generally insufficient to prevent melt extrusion. Consequently, the alkaline rocks of the lower Murdoch Formation are overlain primarily by pillowed to massive tholeiitic basalts and pyroclastic deposits of the upper Murdoch Formation, with only minor sedimentary interbeds. Conversely, the Murdoch basalts are overlain by the Thompson Lake shales which contain only minor intercalations of basalt. This unit may thus record a relative reduction in the rate of magma pulse generation or an increase in sedimentation rate in the axis of the basin. This hiatus in magmatic activity has not been recognized in the Howse zone, although the interpreted high sedimentation rates in this zone would make such identification difficult.

In this regard, the development of the sill-sediment complex in the Howse zone offers a potential explanation for the apparent absence of basalts equivalent to the low-Zr Murdoch volcanics in the lower to middle Menihek Formations. As discussed in Chapter 8, the low-Zr Murdoch suite, which appears to represent early cycle 2 magmatism, has no apparent counterpart in the western part of the Trough. In the Schefferville and Howse zones this may reflect distance from the locus of rifting and magmatic activity. This would not seem to be the case in the Howse zone since this domain was obviously the site of both alkaline and tholeiitic magmatism. But assuming the rate of sedimentation increased rapidly across the Sokoman-Menihek transition, magmas equivalent to the low-Zr Murdoch basalts would be emplaced as sills in the Howse zone rather than erupted as flows. And since the compositions of the low-Zr Murdoch basalts are quite similar to the succeeding cycle 2 volcanics, distinction between early and late tholeiitic sills would be virtually impossible.

The Thompson Lake Formation is overlain by the thick sequence of Willbob submarine tholeiites, and this unit records the return to low sedimentary influx and rapid magma pulse generation in the Doublet terrane. The Willbob basalts were erupted essentially contemporaneously with emplacement of the Menihek gabbros in the Howse

zone, and include at least three horizons of glomeroporphyritic lithologies. The absence of well developed anorthositic phase gabbro in the Doublet terrane suggests that most or all of the glomeroporphyritic magmas were erupted onto the floor of the basin rather than intruded into the volcano-sedimentary pile. This would be expected if the interpretations regarding the rate sedimentation are correct, particularly since glomerocryst-laden melts might be slightly less dense than their aphyric counterparts. The predominance of extrusive over intrusive magmas would give rise to the horizons of homogeneously glomeroporphyritic basalts, but would preclude development of the anorthositic lithologies which are spectacular components of the Menihek sills in the Howse zone.

As with the Le Fer basalts, tholeiitic magmas generated by partial melting of the asthenosphere during evolution of the cycle 2 succession are thought to have been trapped at the base of the crust by density constraints. Precipitation of olivine in a sub-crustal reservoir allowed less dense residual melts to ascend to higher crustal levels. As discussed earlier, the similarity in the compositional ranges of the basalts and the chilled margins of the Menihek sills indicates that both intrusive and extrusive rocks were derived from the same crustal reservoirs. Thus the Menihek and Doublet sills cannot be the site of the differentiation which led to observed compositional variations in the upper Menihek and Willbob basalts. Consequently, intermediate crustal chambers are required as the site of plagioclase + pyroxene fractionation and the formation of anorthositic cumulates. An additional argument for intermediate crustal chambers at >15 km depth is the fact that these pressures would allow flotation accumulation of plagioclase, whereas this phase would sink in the high level Menihek and Doublet sills.

The final stage in the depositional history of the second cycle saw local eruption of more primitive basalts and emplacement of the Retty Peridotite (Fig. 10.12C). Although the extent of picrite eruption appears to have been relatively minor, it is possible that such lavas were initially more widespread but have since been removed by erosion. The eruption and emplacement of these primitive melts may reflect rupture of the continental crust and failure of the density barrier. This may have led to high-Mg magmas bypassing the pre-existing crustal plumbing system and ascending directly to the upper levels of the cycle 2 infilling.

Uplift and deformation of the Labrador Trough appears to have commenced during or immediately following the final episodes of cycle 2 tholeiitic volcanism. Late carbonatite emplacement and rhyodacite eruption occurred at 1870 ± 4 Ma (Machado *et al.*, 1990), and these events may have been related to early tectonic activity accompanying collision between the Superior and Rae Provinces. The encroachment of the Rae Province from the north-east, and its eventual collision with, and overriding (?) of, the Superior margin, triggered south-west directed thrusting of the outer lithotectonic domains onto the autochthonous Schefferville zone (Fig. 10.12D). This was accompanied by deformation, metamorphism and considerable lateral shortening of the cycle 1 and 2 supracrustal assemblage. The suture zone between the Rae and Superior Provinces may, in part, lie along the Lac Tudor shear zone. If the two-sided rift model is correct, then the gneissic rocks occupying the region between the Laporte terrane and the Lac Tudor shear zone would represent the eastern remnant of the rifted Superior margin. In this case, the subduction of the leading edge of the Superior plate beneath the Rae Province may have triggered the continental arc magmatism of the De Pas and Kuujjuaq Batholiths in the hinterland (van der Leeden *et al.*, 1990). Alternatively, the suture may be farther east such that the eastern remnant of the Superior margin comprises much of the hinterland. In this case, the De Pas and Kuujjuaq Batholiths might reflect subduction of the Rae Province beneath the leading edge of the Superior margin (Perreault and Hynes, 1990). Igneous zircons in the Kuujjuaq Batholith, which post-dates peak metamorphism, indicate that granulite facies metamorphic conditions in the hinterland were achieved between 1845 and 1833 Ma (Perreault and Hynes, 1990). Deformation and metamorphism of the Labrador Trough were thus accomplished by this time.

10.4 Summary

The LTBS tholeiites exhibit some textural, compositional and stratigraphic features which have been invoked as arguments in support of an oceanic basin setting for the Labrador Trough. An oceanic setting, however, is unlikely given the lack of ophiolite structure and the isotopic requirement for melt contamination by Archean continental crust. A subduction related setting is inconsistent with the limited compositional diversity

and lack of supra-subduction zone geochemical signatures in the cycle 1 and 2 subalkaline basalts. It thus seems likely that the Labrador Trough represents a continental rift zone developed on the margin of the Superior Province. The cycle 1 and 2 sedimentary basins probably developed in response to stresses imposed by adjoining lithospheric plates rather than as a response to an underlying mantle plume. It is likely that the two episodes of rifting were related to formation of pull-apart basins on the continental margin in a manner analogous to the formation of the Guaymas basin. The two volcano-sedimentary cycles were separated by a significant time interval reflecting the cessation of crustal attenuation during the first cycle, and regional uplift. Renewed attenuation in response to plate movements was probably localized in the weakened crustal zone marked by the lower Knob Lake Group.

Magmatic activity was focused in the axes of these basins where sedimentation rates were low. This produced a predominantly volcanic axial succession now represented only by the cycle 2 Doublet Group. Further from the axes, more rapid sedimentation gave rise to the sill-sediment complexes of the Howse and Ferrum zones. Adjacent to basin margins, magmatic activity was minimal, and predominantly sedimentary successions accumulated (Schefferville zone). The Schefferville zone - Howse zone - Doublet terrane lithostratigraphic succession is probably reflected in the Laporte Group (cf. Dimroth, 1978), although further work is necessary to confirm this interpretation.

Lithospheric attenuation across the axes of the pull-apart basins would allow upwelling and partial melting of underlying asthenospheric mantle. Initial magmas would be alkalic, derived from small degrees of partial melting of fertile zones in the asthenosphere. Subsequent tholeiitic magmatism would reflect larger degrees of partial melting as crustal attenuation and asthenospheric upwelling continued. Crustal attenuation did not proceed to the point of rupture in the first depositional cycle, but separation of continental crust during cycle 2 tholeiitic magmatism is suggested by the occurrence of high-MgO lavas and the Retty Peridotites in the latter sequence. The Superior crust thus acted as a density barrier to primitive magmas until late in the depositional history of the Labrador Trough. The presence of Archean basement rocks throughout most of the evolution of the central Trough is consistent with the isotopic compositions of the various members of the LTBS. It would also explain the occurrence of basement domes mantled

by Kaniapiskau supracrustal assemblages in the eastern lithotectonic domains. Finally, crustal magma chambers would serve as the site of plagioclase + pyroxene fractionation from the tholeiitic melts and, in the case of the cycle 2 tholeiites, as the site of glomerocryst formation through flotation accumulation of plagioclase.

The youngest igneous event in the infilling of the cycle 2 basin appears to have been the formation of minor quantities of acid rocks immediately preceding or during uplift and deformation accompanying Superior-Rae collision. The Labrador Trough was compressed and imbricated, with eastern segments thrust westward over the allochthonous Schefferville zone. Deformation and metamorphism of the Trough was largely completed by the time of emplacement (1845-1833 Ma; Perreault and Hynes, 1990) of the Kuujjuaq and De Pas Batholiths in the eastern hinterland.

Chapter 11

Principal Conclusions

11.1 The Menihek Gabbros

The Menihek gabbros occur in the Howse zone of the central Labrador Trough, where they form a distinctive sill-sediment complex along with turbidites of the Menihek Formation. The intrusions range from less than 100 m to greater than 500 m in thickness, and have strike lengths of several tens of kilometres. The sills were emplaced in the interval 1884–1874 Ma, and are comagmatic with a sequence of tholeiitic basalts which caps the Menihek Formation in the Howse zone. The gabbros and basalts of this domain are considered comagmatic with comparable tholeiitic rocks in the Doublet terrane (Willbob basalts and Doublet gabbros).

The Menihek gabbros may be aphyric, but more often the sills exhibit some degree of development of plagioclase glomeroporphyritic textures. Glomerocrysts consist of aggregates of numerous plagioclase grains which are compositionally and texturally distinct from groundmass plagioclase. Glomerocryst plagioclase grains are typically large and equant to slightly elongate, while groundmass grains are generally lath-like and an order of magnitude smaller. Maximum anorthite content reaches An_{87} in the glomerocrysts, and average compositions as determined from mineral separates are about An_{76} . Although the cores of some groundmass chadacrysts are as calcic as the glomerocryst grains, average compositions are considerably more sodic (An_{65}). Glomerocrysts may show resorption textures, highly sodic peripheral rims and augen-like relationships with groundmass grains. All of these features indicate that groundmass and glomerocryst feldspar represent two distinct generations of crystallization. The similar initial ϵ_{Nd} ratios of glomerocrysts and groundmass, the adcumulate textures of the glomerocrysts, and the density contrast between calcic plagioclase and basaltic liquid at moderate to high pressures, suggest that the aggregates originated as flotation cumulates at the roofs of mid-level crustal magma chambers in which plagioclase and clinopyroxene were fractionated. Disruption of the partially to completely consolidated anorthositic cumulates, and entrainment of the

glomerocrysts is thought to have occurred during periods of magma extraction from the chambers.

Plagioclase glomeroporphyritic rocks also occur in the Doublet terrane, but the highly glomeroporphyritic lithologies grading to anorthositic gabbro are confined to the Howse zone. Glomeroporphyritic intrusions in this domain exhibit systematic variations in glomerocryst size and frequency with stratigraphic position. The greatest number and largest aggregates occur in the central regions of the sills, and both parameters decline toward the contacts. This pattern of zonation is consistent with flow differentiation during emplacement due to the establishment of velocity gradients in the flowing glomeroporphyritic magmas. The ideally symmetrical distribution of glomerocrysts resulting from the actions of grain dispersive pressure during magma emplacement was modified by gravity such that the maximum glomerocryst content occurs in a horizon displaced towards the floor of the sill. This displacement, as well as the asymmetry of concentration gradients, reflects the negative buoyancy of the glomerocrysts in the basaltic magmas at shallow emplacement depths. The glomerocryst concentration profiles developed in response to grain dispersive pressure were further modified by the effects of limited pseudo-plastic behaviour of the magmatic suspension, Magnus-type forces, and possibly wall crystallization during flow. These processes resulted in marginal sparsely glomeroporphyritic zones in which the glomerocryst contents are lower than those expected from the operation of grain dispersive pressure alone. Magnus-type effects, wall crystallization and gravitational settling are thought to have led to pre-concentration of glomerocrysts, such that grain dispersive pressure developed in sills where initial glomerocryst content was below the threshold level of 5 volume %.

The restriction of highly glomeroporphyritic/anorthositic intrusions to the Howse zone is thought to reflect the relative rates of magma pulse generation and sedimentation in the various lithotectonic domains. In the Howse zone, magma pulses were emplaced within the sedimentary pile rather than erupted on the surface. The formation of the sill-sediment complex reflects the rapid sediment influx attendant on the relative proximity of detrital sources. The emplacement process allowed concentration of glomerocrysts by flow differentiation. In the Doublet zone, which represents the axial region of the basin, sedimentation rates were lower. As a consequence, magma pulses were generally erupted

as basalts. The turbulent or convective flow of these lavas during eruption prevented development of grain dispersive pressure and caused homogenization of the extrusions with respect to glomerocryst content.

The Menihek gabbros exhibit geochemical variations which are consistent with the fractional crystallization of plagioclase, hypersthene and augite during solidification. These phases exhibit no features indicative of accumulation through transport, and it is probable that fractional crystallization was an *in situ* process. Wall or congelation crystallization in boundary layers at sill contacts is consistent with the variations in incompatible element concentrations across the intrusions. Concentrations are highest in the chilled margins where crystallization was rapid and the proportion of trapped pore liquid high. Concentrations decline towards the interiors, and this variation is accompanied by increasing groundmass grain size. Together, these features indicate that as the solid-liquid interface advanced away from the chilled margins, the thermal gradient declined such that more time was available for diffusional exchange between intergranular spaces and the main body of melt. This allowed more extensive fractional crystallization, and growth of larger crystals, prior to isolation of intergranular spaces. The development of oxide horizons and lenses of felsic rocks in the upper levels of the sills suggests that solidification progressed most rapidly from the floors and walls of the intrusions.

The lack of cryptic mineral layering in sill C may reflect displacement of residual liquids through continued magma flow over some interval of solidification. But the occurrence of felsic differentiates in this intrusion, and the development of other lithological variants in other sills, particularly cumulate oxide horizons, indicate that the intrusions became closed to fresh magma influx at some stage in their solidification histories. A comparison of flow and crystallization rates suggests that wall crystallization during emplacement would result in the accretion of only a few meters of rock at sill contacts. Once the sills became closed to magma influx, continued crystal fractionation led to differentiated melts which eventually become saturated in magnetite-ilmenite and apatite such that these phases were precipitated to form the oxide horizons. Further evolution led to the formation of siliceous rock types in many sills. The coarse grain size, hydrous nature and incompatible element enriched geochemistries of the felsic rocks suggest that they represent the crystallization products of late stage, volatile-rich magmatic

fluids which accumulated in irregular pods and lenses near the upper contacts of the sills. Although it is possible that some degree of local crustal assimilation contributed to the formation of the felsic rocks, such a process had little effect on the overall compositions of the sills.

11.2 LTBS Volcanic Rocks

The volcanic rocks of the LTBS can be subdivided according to age and composition. In the first depositional cycle, tholeiitic basalts occur in the Le Fer Formation of the Ferrum zone where they are spatially associated with the voluminous Le Fer gabbro sills. The basalts and gabbros are geochemically similar and are interpreted as comagmatic products of cycle 1 tholeiitic magmatism. The U-Pb zircon date of 2169 Ma obtained from a Le Fer gabbro by Rohon *et al.* (1993) provides a minimum age for cycle 1 tholeiitic magmatism and sedimentation. The U-Pb date of 2142 Ma obtained by Dressler and Krogh (cited by Clark, 1984) from the Mistamisk rhyolite indicates that cycle 1 igneous activity continued for at least 27 Ma. This range is a minimum, however, since localized alkaline volcanism of the Chakonipau Formation was coeval with early cycle 1 sedimentation and clearly predates the Le Fer Formation.

Limited quantities of alkaline or mildly alkaline volcanic rocks also occur in the Nimish, lower Menihek and Murdoch formations of the second depositional cycle. These rocks are stratigraphically equivalent, but since the cycle 2 succession is transgressive westward, the alkaline volcanics are neither coeval nor comagmatic. The Nimish basalts, for example, predate the Menihek turbidites and were erupted at 1879 Ma in the Schefferville zone. The lower Menihek basalts are the equivalent sequence in the Howse zone, but these rocks formed prior to 1884 Ma, which is the age of the Menihek gabbros in this domain. The localized alkaline magmatism was succeeded late in the depositional history of the central Labrador Trough by widespread and voluminous tholeiitic magmatism in the Howse zone and Doublet terrane. The earliest products of this igneous event appear to be the subalkaline Murdoch basalts. Later activity produced the comagmatic Menihek gabbros, Doublet gabbros, upper Menihek basalts and Willbob basalts. The glomeroporphyritic intrusive members of this suite were emplaced between

1884 and 1874 Ma (Machado, 1990; Findlay *et al.*, 1995), and these dates provide a minimum estimate of the longevity of cycle 2 tholeiitic magmatism. These dates also indicate that the principal igneous events of the first and second depositional cycles were separated by an interval of ~ 260 Ma.

The tholeiitic basalts of the Le Fer and upper Menihok/Willbob successions exhibit many textural and geochemical similarities despite their difference in age. Both sequences are composed predominantly of massive to pillowed submarine basalts with only relatively minor quantities of pyroclastic material. Both represent deep water flows erupted during turbidite deposition. Both have generally incompatible element-poor tholeiitic geochemistries. Despite these similarities, the Le Fer basalts tend to be slightly impoverished in incompatible elements and enriched in Cr and Ni relative to the cycle 2 tholeiites. In addition, the Le Fer basalts appear to have lower initial ϵ_{Nd} ratios and flatter extended trace element patterns. Finally, the geochemical variation in the Le Fer sequence is consistent with fractional crystallization of plagioclase, clinopyroxene and hypersthene, whereas variation in the cycle 2 tholeiites is consistent with fractional crystallization of plagioclase and clinopyroxene.

The cycle 1 and 2 tholeiites are almost exclusively low-Mg basalts which cannot represent primitive partial melts. As noted, the compositional variation in these suites is consistent with fractionation of plagioclase + pyroxene. There are several lines of evidence, however, which indicate that primitive partial melts differentiated by olivine fractionation to relatively SiO_2 , CaO and Al_2O_3 enriched, and MgO impoverished, residual melts. These melts subsequently evolved through crystallization of pyroxene and feldspar to produce the liquid compositions represented by the basalts. It is thought that Archean continental crust provided a density barrier to primitive, high MgO magmas such that initial olivine crystallization occurred in deep crustal or sub-crustal magma chambers. Less dense residual liquids then migrated to mid-level crustal magma chambers where plagioclase and pyroxene were fractionated. It is in these chambers that flotation accumulation of plagioclase is thought to have formed the anorthositic cumulates which were subsequently entrained as glomerocrysts in the cycle 2 glomeroporphyritic magmas during melt extraction. The absence of equivalent textures in cycle 1 tholeiites may be

related to the depths of the crustal magma chambers. At depths of less than 15 km, for example, calcic plagioclase would be neutrally or negatively buoyant. Thus the formation of plagioclase flotation cumulates in the Le Fer magma chambers might have been precluded by shallower depths of emplacement relative to the cycle 2 reservoirs.

Despite their minor compositional differences, the cycle 1 and 2 tholeiites both exhibit extended trace element patterns comparable to those of transitional MORB. This indicates that both sequences were the products of partial melting in a depleted mantle source. The incompatible element enrichment trends in both suites, however, indicate that undepleted mantle made minor contributions to the primary magmas. In addition, isotopic data indicate that the parental magmas assimilated minor quantities of Archean sialic crust. This is consistent with the hypothesized crustal plumbing system for the two successions. The minor compositional differences between the cycle 1 and 2 tholeiites appear to reflect small differences in the composition of the depleted mantle melt component. This may have arisen through small variations in mantle composition or, more probably, through differences in melting depths.

The alkaline rocks of the Nimish, lower Menihek and lower Murdoch formations include pyroclastic rocks and basaltic flows which are associated with shallow to deep water chemical and clastic sediments, including ironstone-bearing conglomerates. The alkaline igneous activity appears to have been subaerial to shallow submarine in origin, and the sequences are thought to mark the onset of relatively rapid basin subsidence. The Nimish and lower Menihek suites exhibit compositional diversity far in excess of that observed in the tholeiitic rocks. Elemental concentrations and incompatible element ratios are highly variable within and between the two suites. In some samples it is apparent that post-crystallization processes have contributed to geochemical heterogeneity. But this cannot explain all of the compositional variation. The general absence of coherent inter-element trends and the non-constancy of incompatible element ratios indicate that fractional crystallization or progressive changes in the degree of partial melting are also unlikely to be the sources of all of the geochemical diversity. Thus it is probable that variations in mantle source composition and/or melting depths, as well as crustal assimilation, influenced melt geochemistry. As with the tholeiitic suites, the alkaline basalts are unlikely to represent primary partial melts. This is confirmed by the occurrence

of high-MgO volcanic rocks in the lower Murdoch formation. Assuming that the low density continental crust would also impede the ascent of the Nimish and lower Menihek primary melts, it is possible that a period of olivine $\pm$ pyroxene fractionation was a prerequisite for eruption of most alkaline lavas. The compositions of the alkali basalts might then be controlled by a complex series of factors including source composition, depth and degree of partial melting, fractional crystallization, crustal assimilation and post-crystallization alteration and metamorphism.

The Nimish and lower Menihek basalts exhibit extended trace element patterns which are comparable to those of Phanerozoic continental and oceanic alkali basalts. This suggests that their primary melts were derived primarily from a relatively undepleted mantle reservoir. The low ϵ_{Nd} ratios of the lower Menihek basalts confirms this interpretation, but the +3.3 value obtained from the Nimish basalt indicates a mixed source. Thus it is probable that the lower Menihek suite formed by partial melting of an undepleted mantle source, with only minor contributions from the depleted mantle, while the Nimish basalts reflect larger contributions from the depleted reservoir. Both sequences may also have assimilated geochemically and isotopically enriched crustal material, as is indicated by the negative ϵ_{Nd} ratio of the Nimish syenite. Differences in the extended trace element patterns of the Nimish and lower Menihek basalts may reflect the differences in contributions from the depleted mantle source, or they may indicate that the composition of the undepleted mantle was heterogeneous.

The Murdoch basalts are transitional between the alkaline compositions of the Nimish and Menihek suites, and the tholeiitic compositions of the upper Menihek and Willbob formations. The Murdoch volcanics can be subdivided into an areally restricted lower sequence of mildly alkaline basalts, and a more widespread and voluminous overlying succession of subalkaline basalts. The basal high-Zr sequence exhibits elevated incompatible element concentrations which are transitional to the alkaline compositions of the Nimish and lower Menihek suites. In contrast, the low-Zr subalkaline Murdoch basalts are compositionally closer to the cycle 2 tholeiites. Insufficient data are available to examine the range and source of geochemical variations in the Murdoch succession, but again the basalts can be subdivided according to similarities with other sequences. The

low-Zr basalts, for example, exhibit incompatible element ratios comparable to those of the cycle 2 tholeiites, and sample compositions which generally fall on the trends defined by the latter in covariations diagrams. The high-Zr basalts, however, exhibit non-constant incompatible element ratios and considerable compositional diversity, despite the fact that only four samples of this suite were obtained.

The alkaline compositions of the lower Murdoch volcanics, and their apparent shallow water eruption during a period of tectonic instability, indicate that they are stratigraphically equivalent to the alkaline suites of the more westerly domains. Although no isotopic data are available for the high-Zr basalts, the extended trace element pattern of the most primitive sample is consistent with a relatively large melt contribution (~40%) from an undepleted mantle source. The trace element pattern of a second, more evolved, sample is consistent with a smaller (15-20%) undepleted mantle melt contribution. The compositions of the high-Zr Murdoch basalts may also reflect crustal assimilation, but this possibility cannot be evaluated without isotopic data. Such data, however, is available for the subalkaline Murdoch basalts, and ϵ_{Nd} ratios indicate that up to 5% crustal material was added to these melts. The undepleted mantle melt component in the primary low-Zr magmas was probably less than 10%.

The range in compositions of the members of the LTBS requires that both depleted and undepleted sources were present in the Early Proterozoic mantle. All samples for which isotopic data are available, with the exception of the Nimish basalt, require crustal assimilation to produce observed isotopic and trace element compositions. Cycle 2 magmatism progressed from alkaline, through transitional alkaline/subalkaline, to tholeiitic compositions. This may also have been the case during cycle 1 magmatism, but there is currently no evidence of transitional compositions. Early cycle 2 magmatic activity (lower Menihek basalts) largely reflected partial melting of the undepleted mantle source. Subsequent melts were increasingly derived from the depleted mantle reservoir. Thus the melt contribution from the undepleted mantle source predominated in the highly alkaline lower Menihek basalts, dropped to 40-5% in the Murdoch basalts, and declined still further (<5%) in the upper Menihek and Willbob basalts.

11.3 Evolution of the LTBS

The LTBS tholeiites are geochemically comparable to transitional MORB, but have isotopic compositions which require assimilation of Archean continental crust. In addition, part of the cycle 1 and 2 sedimentary successions were deposited on the margin of the Superior craton. The volcano-sedimentary sequences lack the ophiolitic structure of mature oceanic crust, and the LTBS does not include the picritic compositions found in oceanic volcanics (Chukotat Formation) of the approximately coeval Cape Smith belt. These features suggest that the LTBS tholeiites represent incipient oceanic crust developed in a continental rift environment. This would explain the continental margin aspects of sedimentation, as well as the geochemical and lithological characteristics of LTBS magmatism. The absence of supra-subduction zone compositional traits indicates that the LTBS magmatism was not related to Early Proterozoic subduction processes, and therefore the succession cannot have formed in a back-arc basin. The LTBS tholeiites also lack the geochemical and lithological characteristics of continental flood basalts, and this indicates that continental rifting was not a response to an underlying mantle plume.

The best interpretation of the origin of the LTBS involves magmatism associated with crustal extension in response to tensional stresses resulting from plate movements. The LTBS, and the sedimentary infilling of the Labrador Trough, are thus thought to have formed in pull-apart basins on the thinned margin of the Superior craton. The nature of the plate interactions which led to crustal attenuation and rifting are poorly constrained. Cycle 2 extension may have been related to 1.9-1.88 Ga collisional indentation of the Churchill Province by the Superior craton, and oblique convergence of a rotated arm of the Rae Province and the Superior plate (Hoffman, 1990a, 1990b, Skulski *et al.*, 1993). The potential source of extension leading to the first volcano-sedimentary cycle is obscured by the lack of knowledge regarding the configuration of pre-1.9 Ga tectonic plates. One possibility is that cycle 1 extension was related to the early stages of subduction of the 2.2(?) - 1.9 Ga Manikewan ocean basin to the north of the Cape Smith belt. These interpretations, however, are highly speculative.

Crustal attenuation and formation of the cycle 1 pull-apart basin had commenced prior to 2169 Ma. Initial sedimentation was fluvial, and associated with alkaline volcanism of the Chakonipau Formation. Subsequent subsidence and marine transgression

led to deposition of shallow to deep water sediments, culminating with the Le Fer turbidites. In the axial portions of the basin, where sedimentation rates were low, turbidite deposition was interspersed with eruption of the Le Fer submarine tholeiitic basalts. The turbidite-basalt assemblage was heavily intruded by the Le Fer gabbros. Magmatic activity and sedimentation continued until at least 2142 Ma, at which point crustal extension and basin subsidence ceased. This led to infilling of the basin and marine regression, thereby producing the shallow water conditions required for deposition of the Denault dolostone. Intra-cycle sedimentation was interrupted, and in places pre-existing strata were eroded, during an interval of uplift between 2140 Ma and 1900 Ma. There is conflicting evidence regarding the extent of uplift and non-deposition/erosion, but it is clear that the interval did not see a major orogenic event.

Renewed extension led to the formation of the cycle 2 pull-apart basins. These were localized in the same general area as the first cycle basins, presumably as a result of crustal weakness related to cycle 1 extension and attenuation. The cycle 2 basin, however, extended farther to the west and north. Crustal attenuation and basin subsidence again led to a transgressive marine sequence. This commenced with deposition of continentally-derived clastic sediments of the Wishart and Allison formations, which overstepped cycle 1 dolostones and were locally deposited directly on Superior basement rocks. Subsidence and transgression led to deposition of the Sokoman iron formation and the Menihek turbidites. Alkaline volcanism of the Nimish and lower Menihek formations accompanied ironstone deposition and the early stages of Menihek sedimentation, and hence records the onset of relatively rapid basin subsidence. This stage is marked by the slightly alkaline volcanics of the Murdoch Formation in the axial portion of the basin. Subsequent turbidite deposition was accompanied by tholeiitic volcanism in the axial regions of the basin (Doublet terrane), and by emplacement of numerous gabbro sills in flanking regions (Howse zone). The variation in magmatic style is thought to reflect variations in sedimentation rates. Cycle 2 tholeiitic magmatism and turbidite deposition spanned an interval of at least 10 Ma. The eruption of the upper Menihek basalts in the Howse zone, and high-Mg basalts in the Doublet terrane, coupled with the emplacement of high-Mg magmas as sills in the latter domain, may record the rupture of continental crust. The youngest cycle 2 magmatic activity appears to have produced rhyolitic lavas. This

may have been associated with the early stages of uplift accompanying collision with the Rae (or Nain ?) Province. The Labrador Trough was subsequently deformed and metamorphosed during Hudsonian orogenesis.

The Superior Province is likely to have formed the basement to the Labrador Trough succession throughout most of its evolution. Thus continental crust acted as a density barrier to the primitive partial melts which evolved to produce both the alkaline and subalkaline volcanic rocks. Initial magmatic activity during both depositional cycles was of alkaline affinity. This is thought to have resulted from small degrees of decompressive partial melting in the upwelling asthenospheric mantle during the early stages of lithospheric attenuation. The initially small degrees of melting would preferentially sample relatively undepleted components with lower solidus temperatures in the heterogeneous asthenospheric mantle. This gave rise to the alkaline magmas of the LTBS. Continued crustal attenuation and mantle upwelling led to progressively greater degrees of partial melting. Partial melts were then derived from both undepleted and depleted components, thereby producing transitional alkaline to subalkaline magmas. These are represented in the cycle 2 sequence by the Murdoch Formation. With increased partial melting, the compositional effects of the undepleted mantle component were largely subdued, and the voluminous, transitional MORB - like tholeiitic magmas were formed.

The formation of large quantities of primitive, and relatively dense, tholeiitic magma led to crustal underplating or formation of deep crustal magma chambers in which the melts pooled. Because of the evolved nature of most of the basalts sampled in this study, it is thought that the crust acted as a density barrier to primitive melts. Thus a period of olivine fractionation was necessary before the tholeiitic magmas could rise to higher crustal levels. Coupled assimilation-fractional crystallization may have resulted in relatively SiO_2 - Al_2O_3 - CaO enriched melts whose density permitted further ascent. Assimilation of crustal material would tend to promote olivine crystallization (cf. Francis *et al.*, 1983), as well as imparting the mixed isotopic signatures to the magmas.

Further melt differentiation took place in crustal chambers at higher levels prior to eruption of the basalts and emplacement of the sills. Geochemical constraints indicate that the sills were not the site of the second stage of fractionation, at least during cycle 2 tholeiitic magmatism. This is also suggested by the fact that for the glomerocrysts to have

formed by flotation accumulation of fractionated plagioclase, crustal depths of at least 15-20 km are required. Thus it is likely that mid-level crustal chambers were the site of the plagioclase+pyroxene fractionation that led to the magma compositions represented by the cycle 1 and cycle 2 tholeiitic basalts and gabbros. These may also have been the site of some crustal assimilation. Late failure of the density barrier due to crustal rupture would allow more primitive melts to ascend directly to the surface, thereby forming the Retty sills and possibly some quantity of high-Mg flows.

The generally evolved nature of the alkaline volcanics suggests that a density controlled plumbing system may also have been involved in the differentiation of their parental magmas. While a two stage fractionation history is not required, it is clear that the alkaline melts evolved to relatively MgO-poor compositions prior to eruption. Given the presence of alkaline mafic/ultramafic sills associated with the Nimish Formation, it is possible that magmatic differentiation occurred in relatively high level chambers. The contrast in isotopic compositions of the Nimish basalt and syenite, however, suggest that crustal contamination of alkaline melts locally occurred in such high level sub-volcanic sills. This may indicate that less contaminated, but still relatively evolved alkaline melts were drawn from deeper reservoirs, and were locally erupted without further differentiation. The occurrence of relatively primitive alkaline volcanics in the Murdoch Formation (Gebert, 1988) suggests that undifferentiated melts occasionally reached the surface in the early stages of crustal attenuation. Since primitive tholeiitic melts reached the surface only when crustal rupture had occurred, the presence of the high-Mg Murdoch basalts may reflect lower melt densities due to higher volatile content.

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Appendix A1

Sample Locations

Sample locations are indicated on the simplified geological maps of the Chassin Lake, Tait Lake, Howse Lake and Retty Lake areas provided in Figures A1.1-A1.4 (see Fig. 3.1-3.3 for general locations of map areas). The locations of the traverses conducted across the four Howse Lake sills examined in detail are indicated in the Howse Lake map. Individual sample locations in the Howse Lake profiles are identified by the position above the lower contacts as indicated in Appendices A2, A3, A4 and A6. The locations of other gabbroic samples are indicated on the appropriate maps below. The geology of the sampled areas other than Howse Lake is largely based on the regional maps of Baragar (1967), Frarey (1967) and Wardle (1982).

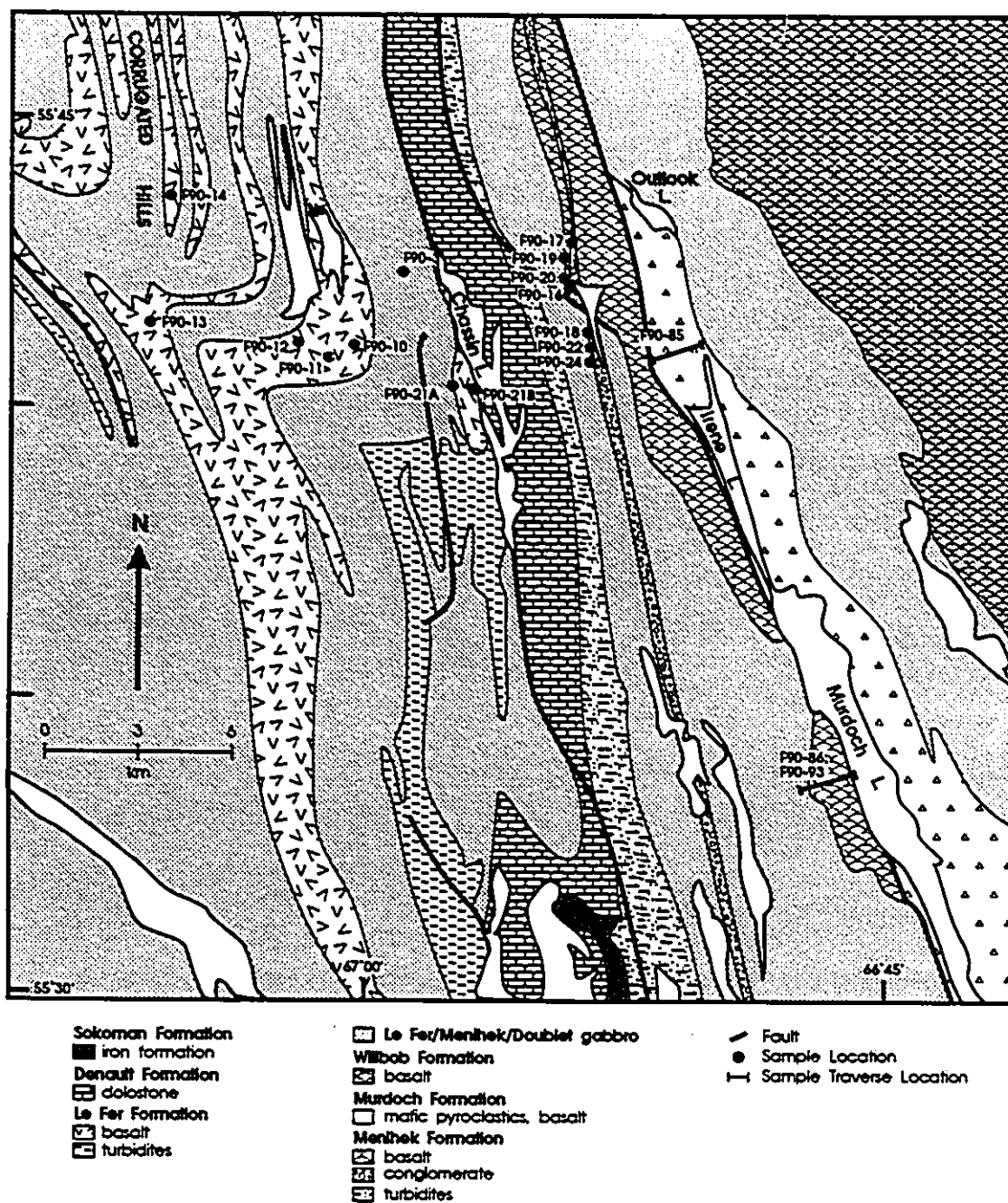


Figure A1.1. Simplified geological map of the Chassin Lake area showing sample locations.

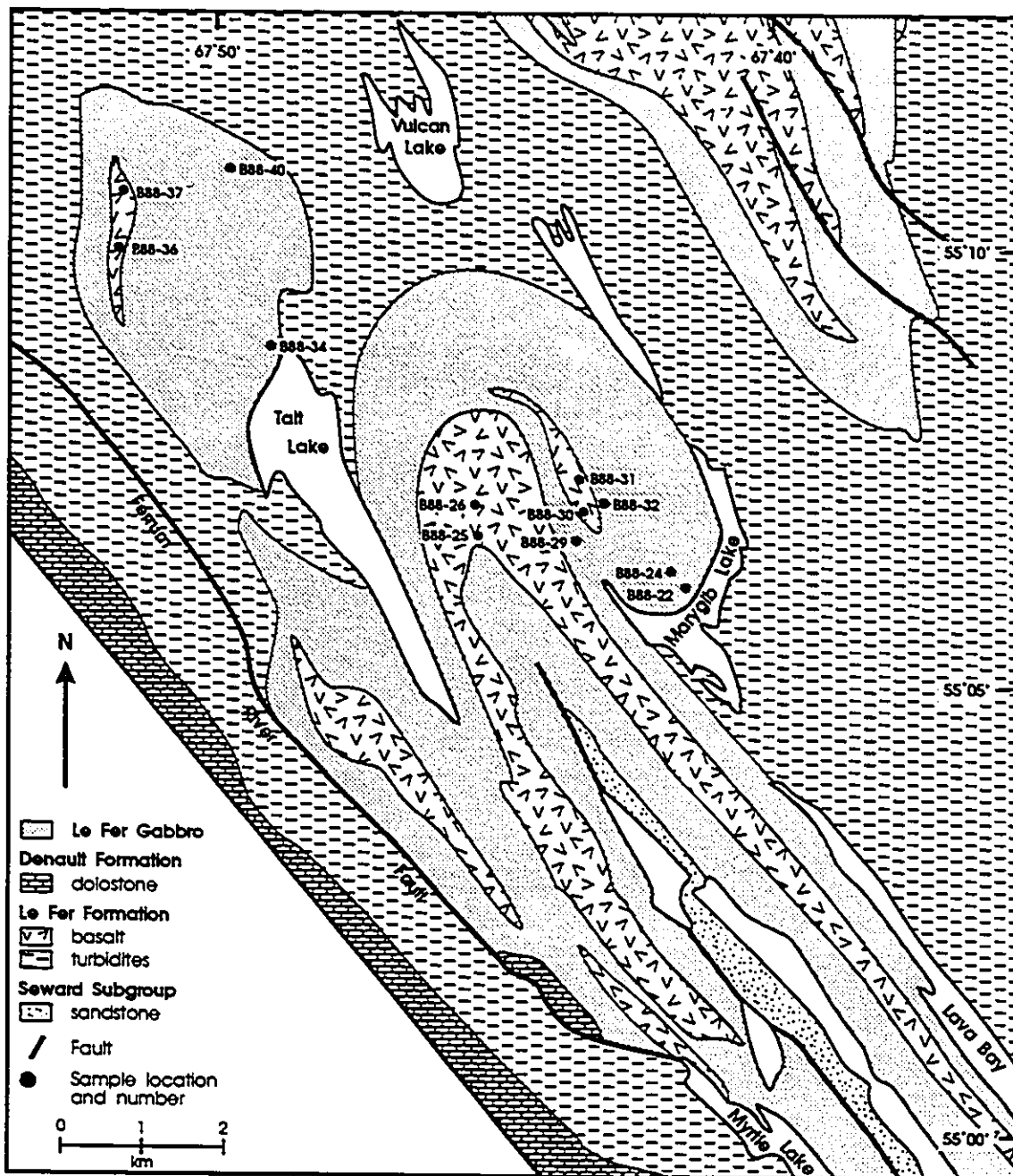


Figure A1.2. Simplified geological map of the Tait Lake area showing sample locations.

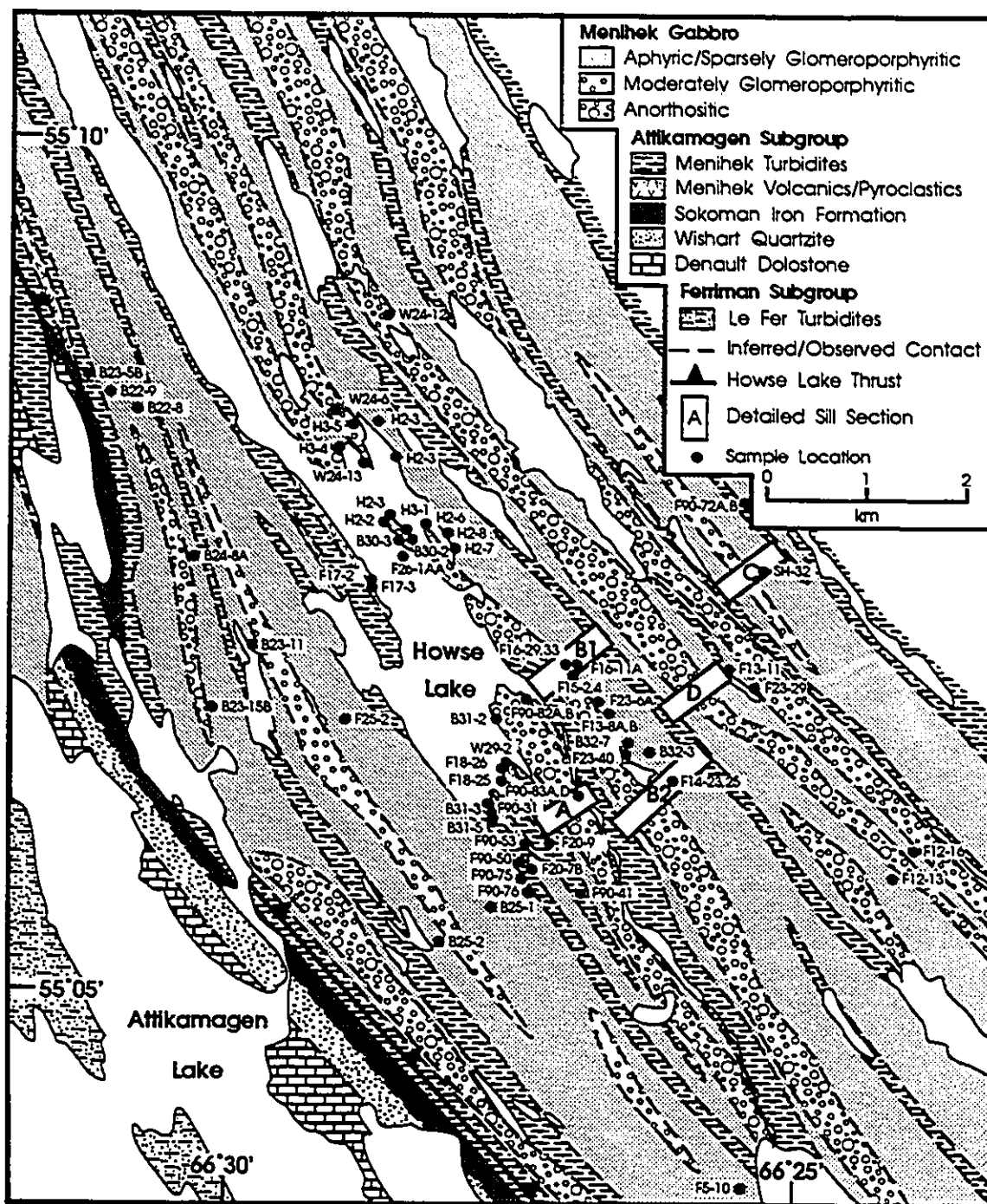


Figure A1.3. Simplified geological map of the Howse Lake area showing sample locations.

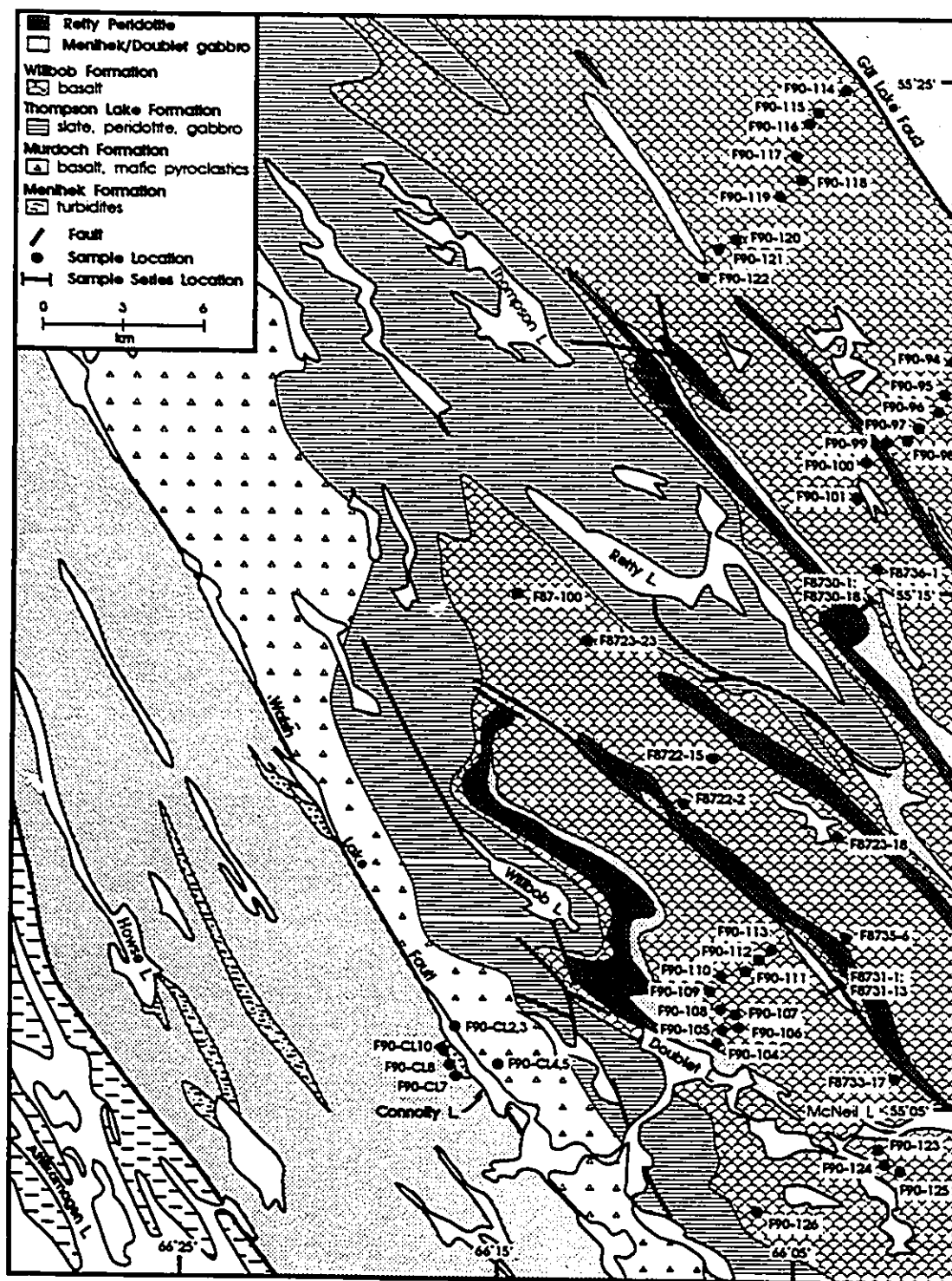


Figure A1.4. Simplified geological map of the Retty Lake area showing sample locations.

Appendix A2

Modal Analysis

A2.1 Analytical Procedure

The modal compositions of 64 samples, including 30 gabbroic and 9 anorthositic phase samples from sill B, 15 gabbroic phase samples from sill C, and 10 felsic samples from sills A, B and C, were determined by point counting. The modes are based on approximately 500 points from thin sections whose dimensions are 2.5 x 4.5 cm for the gabbroic and felsic samples, and 3.5 x 7.5 cm for the anorthositic samples. Accessory minerals (principally apatite and zircon) are not included in the analyses, and secondary phases were assigned to the primary minerals where possible. Where this determination could not be made, the points were excluded from the total. For the felsic samples, ferromagnesian minerals were assigned to clinopyroxene (where present) and an undivided group which is labelled hornblende in Table A2.2, but which also includes biotite, and where clinopyroxene was not observed, actinolite and chlorite.

A2.2 Analytical Uncertainty

Analytical uncertainty was evaluated by replicate measurements on three thin sections from sample F90-80, and three from sample F874-17. The poor precision of replicate measurements for sample F874-17 derives from the coarse grained nature of the anorthositic phase. Following Kretz (1985), the proportional uncertainty in modal abundance is expressed as the coefficient of variation (standard deviation/average value). In general, the smaller the modal proportion of a mineral, the larger the associated coefficient of variation. Estimated average uncertainties are provided in Table A2.1 and results are given in Table A2.2. Samples are listed by individual sills (Sill), lithological type (Lithology) and height above the lower contact (Height). The lithological abbreviations used are:

Ol-free = olivine free

Ol-bearing = olivine bearing

Gabbroic = gabbroic phase

Anorth. = anorthositic phase

Oxide = cumulate oxide series

Felsic = felsic series

Table A2.1. Precision of modal analyses determined by replicate measurements on three thin sections of each sample

F90-80				F874-17			
Mineral	Average	Standard Deviation	Coefficient of Variation	Mineral	Average	Standard Deviation	Coefficient of Variation
Plagioclase	37.5	1.2	0.03	Plagioclase	76.9	6.1	0.08
Clinopyroxene	32.2	1.2	0.04	Clinopyroxene	10.8	1.9	0.18
Orthopyroxene	10.2	0.5	0.05	Orthopyroxene	8.1	1.3	0.16
Olivine	15.6	0.8	0.05	Olivine	0.0		
Oxide	3.2	0.2	0.07	Oxide	2.1	0.5	0.24
Orthoclase	0.6	0.1	0.12	Orthoclase	0.9	0.2	0.22
Quartz	0.7	0.1	0.12	Quartz	1.2	0.3	0.25

Table A2.2. Modal analyses.

Sample	Sill	Lithology	Height	Plagioclase	Clinopyroxene	Orthopyroxene	Olivine	Oxide	Orthoclase	Quartz
78C	Sill C	Ol-Free	3.0	48.4	30.8	9.2	0.0	4.7	4.8	2.1
78D	Sill C	Ol-Free	6.8	48.0	33.0	13.1	0.0	3.7	1.7	0.5
SH-4	Sill C	Ol-Free	7.9	53.1	33.5	3.9	0.0	4.9	3.4	1.3
SH-8	Sill C	Ol-Bearing	18.5	47.2	31.3	12.7	2.6	4.4	1.3	0.5
SH-14	Sill C	Ol-Bearing	34.6	43.8	31.4	17.1	5.7	1.2	0.7	0.0
SH-16	Sill C	Ol-Bearing	56.2	46.0	29.8	14.4	8.1	1.3	0.5	0.0
F90-80	Sill C	Ol-Bearing	104.5	37.5	32.2	10.2	15.6	3.2	0.6	0.7
SH1-38	Sill C	Ol-Bearing	118.2	42.2	34.2	14.8	5.7	1.9	1.3	0.0
SH1-24	Sill C	Ol-Bearing	131.4	42.6	27.3	18.4	9.2	2.1	0.0	0.4
SH1-35	Sill C	Ol-Bearing	156.5	51.3	24.7	18.9	3.0	1.2	0.5	0.4
F876-1	Sill C	Ol-Free	176.8	46.9	26.8	20.0	0.0	3.3	1.9	1.1
F876-5	Sill C	Ol-Free	186.0	51.3	24.8	17.3	0.0	2.5	3.0	1.0
F876-9	Sill C	Ol-Free	199.0	44.8	22.8	20.1	0.0	3.2	5.0	4.1
F876-13	Sill C	Ol-Free	209.0	46.0	28.9	14.2	0.0	4.2	4.0	2.7
F876-22	Sill C	Ol-Free	219.1	41.7	29.7	13.4	0.0	4.1	5.1	6.1

Table A2.2 (continued). Modal analyses.

Sample	Sill	Lithology	Height	Plagioclase	Clinopyroxene	Orthopyroxene	Olivine	Oxide	Orthoclase	Quartz
F871-6	Sill B	Gabbroic	31.7	50.5	32.2	8.4	0.0	3.6	3.4	2.0

F16-21	Sill B	Gabbroic	42.4	51.4	27.2	16.4	0.0	3.3	1.3	0.5
F871-5	Sill B	Anorth.	58.0	64.3	18.1	16.4	0.0	0.8	0.0	0.3
F871-4	Sill B	Anorth.	77.9	65.6	15.2	16.9	0.0	1.3	0.7	0.3
F871-3	Sill B	Anorth.	161.8	64.4	17.9	12.6	0.0	2.4	1.2	1.6
F16-18	Sill B	Anorth.	204.3	72.6	15.7	9.2	0.0	1.1	0.6	0.8
F16-17	Sill B	Anorth.	220.4	64.5	14.5	16.7	0.0	2.2	1.2	0.9
F16-14	Sill B	Gabbroic	242.9	51.1	30.2	15.1	0.0	2.1	1.1	0.4
F16-11B	Sill B	Gabbroic	288.3	50.8	24.9	19.7	0.0	4.2	0.4	0.0
F16-9	Sill B	Gabbroic	297.7	51.0	24.5	20.5	0.0	2.4	0.7	0.8
F16-6	Sill B	Gabbroic	321.3	48.2	28.9	12.3	0.0	7.1	0.6	2.9
F16-5	Sill B	Gabbroic	328.5	45.9	33.0	11.2	0.0	3.8	5.3	0.9
F16-4	Sill B	Gabbroic	332.1	51.1	33.9	8.4	0.0	4.7	1.2	0.6
F16-2	Sill B	Gabbroic	338.5	37.6	37.7	9.4	0.0	12.6	1.1	1.6
F16-1	Sill B	Gabbroic	343.9	42.7	38.4	7.9	0.0	7.6	2.1	1.3
F90-67	Sill B	Oxide	351.5	41.1	23.4	13.0	0.0	18.4	2.3	1.8
F90-64	Sill B	Oxide	373.3	42.4	19.2	16.3	0.0	16.2	3.5	2.5
F90-61	Sill B	Oxide	387.0	47.6	22.7	10.6	0.0	7.3	7.3	4.6
F874-9	Sill B	Anorth.	115.2	62.4	19.7	13.9	0.0	2.3	1.2	0.4
F874-12	Sill B	Anorth.	137.0	68.8	13.7	15.2	0.0	1.8	0.6	0.0
F874-15	Sill B	Anorth.	160.0	68.0	20.3	4.8	0.0	3.4	1.6	1.9
F874-17	Sill B	Anorth.	200.0	76.9	10.8	8.1	0.0	2.1	0.9	1.2
F14-36A	Sill B	Gabbroic	302.1	46.9	25.7	23.8	0.0	1.7	0.5	1.4
F14-35	Sill B	Gabbroic	304.0	54.7	21.4	19.8	0.0	2.2	1.2	0.7
F14-38	Sill B	Gabbroic	306.0	51.5	21.4	21.7	0.0	3.5	1.2	0.7
F14-31	Sill B	Gabbroic	307.6	61.7	14.6	14.7	4.1	2.9	0.0	2.1
F14-30	Sill B	Gabbroic	310.5	56.6	22.8	16.2	0.0	2.5	1.1	0.7
F14-29	Sill B	Gabbroic	312.4	54.7	5.7	35.8	0.0	3.3	0.0	0.5
F14-28	Sill B	Gabbroic	313.7	51.6	28.6	15.2	0.0	2.8	1.2	0.6
F14-26	Sill B	Gabbroic	324.5	54.9	26.9	11.0	0.0	2.3	3.7	1.1
F14-22	Sill B	Gabbroic	334.8	57.0	17.7	20.0	0.0	1.6	2.8	0.9
F14-21	Sill B	Gabbroic	352.6	53.9	16.0	26.6	0.0	1.6	1.3	0.6
F14-16	Sill B	Gabbroic	360.2	61.2	24.6	11.1	0.0	1.2	1.4	0.4
F14-18	Sill B	Gabbroic	372.0	61.7	16.4	15.6	0.0	1.4	3.3	1.5
F14-12	Sill B	Gabbroic	379.7	56.1	32.0	7.9	0.0	1.8	0.8	1.3
F14-8	Sill B	Gabbroic	389.2	64.1	24.6	6.9	0.0	1.6	2.1	0.6
F14-6	Sill B	Gabbroic	402.0	50.8	35.1	8.3	0.0	3.0	1.9	0.9
F14-5	Sill B	Gabbroic	422.0	51.8	31.9	9.1	0.0	2.1	4.2	0.9
F14-3	Sill B	Gabbroic	444.5	47.0	39.1	5.7	0.0	3.0	4.0	1.3

Table A2.2 (continued). Modal Analyses

Sample	Sill	Lithology	Height	Plagioclase	Orthoclase	Quartz	Hornblende	Clinopyroxene	Oxide
F876-14	Sill C	Felsic	211.7	16.0	26.4	34.4	19.9	0.0	3.3
F876-15	Sill C	Felsic	212.1	30.0	16.4	25.6	23.2	0.0	4.8
F876-16	Sill C	Felsic	214.3	44.8	9.3	7.8	13.1	21.1	3.9
F876-17	Sill C	Felsic	215.1	21.9	28.9	32.1	15.1	0.0	2.1
F90-60B	Sill B	Felsic	400.0	22.2	23.0	16.8	13.2	19.9	4.8
F90-60A	Sill B	Felsic	404.0	27.5	21.2	22.4	7.3	18.3	3.2
F90-60C1	Sill B	Felsic	406.0	30.8	19.9	31.2	10.7	5.1	2.3
F90-60C2	Sill B	Felsic	408.0	10.2	31.6	36.5	11.2	7.4	3.0
F90-83A	Sill A	Felsic		36.4	15.3	11.7	14.1	17.0	5.6
F90-83B	Sill A	Felsic		43.2	11.4	7.0	15.4	19.4	3.6

Appendix A3

Electron Microprobe Mineral Analyses

A3.1 Analytical Procedure

Mineral compositions were determined by quantitative energy dispersive analysis of individual mineral grains in carbon coated polished thin sections. The instrument used was a Materials Analysis Company (MAC) electron microprobe equipped with a Kevex Corporation detector and housed in the Mineralogy Division of the Geological Survey of Canada. The instrument was operated with a 20 kv accelerating voltage, 10 nanoampere sample current, 10 ev/ch spectrometer range and 60-80 second acquisition time. Data reduction was by means of in-house software (system EDDI).

Analysis of mineral chemistry was undertaken to provide information on: (1) the nature of crystallization processes in the Menihek sills, and (2) the relationship between groundmass and glomerocryst plagioclase in the sills. Analytical work was thus directed at establishing the extent of compositional zonation in individual grains, examining variations in mineral chemistry with stratigraphic position, and comparing groundmass and glomerocryst plagioclase compositions. Sill C at Howse Lake was chosen for evaluation of cryptic variation in mineral chemistry because it represents a relatively homogeneous intrusion across which suitable samples could be obtained. Because olivine and hypersthene are subject to strong alteration in the sill, analytical work was concentrated on clinopyroxene and plagioclase. A limited number of analyses were obtained from sill B for inter-sill comparison of mineral compositions. The number of individual glomerocrysts from which analytical data was obtained was also limited by the effects of alteration. Sufficient analyses were obtained, however, to allow compositional comparisons of the two sets of plagioclase crystals.

A3.2 Analytical Uncertainty

Quoted precision in the analyses is $\pm 2\%$ of the concentration for concentrations greater than 5%. The principal exception to this precision is Na, for which the estimated

precision is $\pm 5\%$ for concentrations greater than 5%. To better estimate analytical precision during periods of operation, multiple analyses of the mineral standards labradorite and kaersutite were obtained. Average, standard deviation and coefficient of variation were calculated from 16 analyses of the kaersutite standard and 21 analyses of the labradorite standard (Tables A3.1 and A3.2). The coefficient of variation (= standard deviation/average) so determined was used to estimate analytical uncertainty in unknown samples.

Table A3.1. Average, standard deviation and coefficient of variation calculated from replicate analyses of the kaersutite standard. Coefficient of error = (standard deviation /average)

Oxide	Average	Standard Deviation	Coefficient of Error
SiO ₂	40.92	0.43	0.01
Al ₂ O ₃	14.06	0.18	0.01
FeO _T	11.11	0.26	0.02
TiO ₂	4.87	0.13	0.03
Cr ₂ O ₃	0.04	0.01	0.33
MgO	13.18	0.18	0.01
MnO	0.06	0.02	0.29
CaO	10.50	0.25	0.02
Na ₂ O	3.19	0.45	0.14
K ₂ O	2.07	0.06	0.03

Table A3.2 Average, standard deviation and coefficient of variation calculated from replicate analyses of the labradorite standard. Coefficient of variation = (standard deviation /average)

Oxide	Average	Standard Deviation	Coefficient of Error
SiO ₂	52.07	0.25	0.005
Al ₂ O ₃	30.57	0.15	0.005
CaO	13.1	0.12	0.009
Na ₂ O	3.68	0.28	0.075
K ₂ O	0.18	0.04	0.246
FeO _T	0.41	0.08	0.198

A3.3 Explanation Of Table A3.3

Microprobe analyses of mineral compositions are listed in able A3.3. The analyses are tabulated according to the mineral identity, the source of the sample (sill B or C), and

the stratigraphic position of the sample, denoted as height above base (Height), in the sill. The pyroxene analyses are classified as interior or rim depending on their positions relative to grain boundaries. Rim analyses are located near the boundaries with other grains, while interior analyses were obtained from the central regions of grains. Groundmass plagioclase analyses are classified as core, intermediate (abbreviated intermed. in Table A2.3) and rim. Rim analyses were obtained from zones adjacent to grain boundaries, while core analyses represent identifiable crystal cores. Intermediate analyses are those obtained from intermediate areas, and from the central regions of crystals where there was no identifiable core. Glomerocryst analyses are classified as central (within the central part of primocrysts), intergranular (intergranular to primocrysts, abbreviated Intergr.) and rim, which represent the peripheral sodic rims. These classifications are subjective and serve only to establish general patterns of zonations in individual mineral grains and within glomerocrysts. The groundmass plagioclase analyses are also classified according to the type of grain; enclosed (E) grains are those which appear to be completely enveloped in pyroxene oikocrysts, the unenclosed (U) category includes all other grains.

The Label column gives the sample and analysis number i.e., the label SH-8-12 represents the twelfth analysis obtained from sample SH-8. Because of poor totals for some analyses, the labels are not necessarily consecutive. Analyses of individual grains in a sample are grouped together, with spaces separating groups of analyses. The analyses include the concentrations of major elements in the mineral, and calculated end member compositions. For the pyroxenes, the endmembers are wollastonite (Wo), enstatite (En) and ferrosilite (Fs). FeO, Fe₂O₃, and the Mg#s ($100 \times \text{Mg}/(\text{Mg} + \text{Fe}^{+2})$) of the pyroxenes are also given. For plagioclase, the endmembers are anorthite (An), albite (Ab) and orthoclase (Or).

Table A3.3. Microprobe Analyses.

Sill C: Orthopyroxene

Label	SH-11-1	SH-11-2	SH-11-3	SH-11-4	SH-11-5	SH-11-6	SH-11-7	SH-11-8
Height	40.1	40.1	40.1	40.1	40.1	40.1	40.1	40.1
Position	Interior	Interior	Interior	Rim	Rim	Rim	Interior	Interior
SiO ₂	51.54	51.51	51.15	50.64	50.31	51.12	51.13	51.51
Al ₂ O ₃	0.23	0.52	0.48	0.00	0.00	0.27	0.40	0.13
FeO _T	26.34	25.42	24.02	29.11	30.07	26.24	26.28	24.44
MgO	19.55	20.31	21.36	18.39	17.86	19.04	20.34	19.87
CaO	1.94	1.91	2.01	1.24	1.25	2.28	1.94	2.83
K ₂ O	0.00	0.00	0.05	0.04	0.00	0.05	0.00	0.00
TiO ₂	0.19	0.31	0.38	0.39	0.31	0.40	0.32	0.36
MnO	0.45	0.54	0.57	0.57	0.51	0.52	0.50	0.40
Cr ₂ O ₃	0.00	0.03	0.06	0.01	0.01	0.02	0.02	0.00
Total	100.24	100.58	100.08	100.40	100.32	99.94	100.93	99.54
Fe ₂ O ₃	2.58	3.00	4.32	3.63	3.98	2.54	4.55	2.18
FeO	24.02	22.72	20.14	25.84	26.49	23.95	22.19	22.48
Wo	0.04	0.04	0.04	0.03	0.03	0.05	0.04	0.06
En	0.55	0.57	0.60	0.52	0.51	0.54	0.57	0.56
Fs	0.41	0.39	0.36	0.45	0.46	0.41	0.39	0.38
Mg#	56.9	58.7	61.3	53.0	51.4	56.4	58.0	59.2

Sill C: Orthopyroxene

Label	SH-11-16	SH-11-27	SH-11-28	SH-11-29	SH-24-1	SH-24-2	SH-24-3
Height	40.1	40.1	40.1	40.1	131.4	131.4	131.4
Position	Interior	Rim	Interior	Rim	Rim	Interior	Interior
SiO ₂	52.02	50.68	51.69	51.02	51.54	52.48	52.82
Al ₂ O ₃	0.26	0.40	0.28	1.36	0.23	0.38	0.57
FeO _T	19.44	26.75	22.07	24.06	21.04	16.51	16.24
MgO	24.18	19.56	21.86	19.86	23.65	26.79	27.04
CaO	3.16	1.97	2.64	2.73	1.94	2.24	2.48
K ₂ O	0.00	0.05	0.00	0.05	0.00	0.04	0.05
TiO ₂	0.21	0.32	0.28	0.32	0.26	0.23	0.17
MnO	0.43	0.54	0.45	0.37	0.45	0.40	0.31
Cr ₂ O ₃	0.03	0.03	0.00	0.03	0.00	0.03	0.04
Total	99.74	100.30	99.27	99.80	99.11	99.10	99.72
Fe ₂ O ₃	5.14	4.38	3.12	2.43	4.74	5.21	5.29
FeO	14.82	22.81	19.26	21.88	16.77	11.82	11.48
Wo	0.06	0.04	0.05	0.06	0.04	0.04	0.05
En	0.67	0.56	0.61	0.57	0.66	0.73	0.74
Fs	0.27	0.40	0.33	0.38	0.30	0.22	0.22
Mg#	68.9	56.6	63.8	59.5	66.7	74.3	74.8

Table A3.3 (continued). Microprobe Analyses.

Sill C: Orthopyroxene

Label	SH-24-7	SH-24-8	SH-24-10	SH-28-63	SH-28-64	SH-28-65
Height	131.4	131.4	131.4	140.6	140.6	140.6
Position	Interior	Interior	Interior	Interior	Interior	Rim
SiO ₂	52.43	52.80	51.35	52.91	52.92	52.16
Al ₂ O ₃	0.40	0.46	0.32	0.81	0.74	0.40
FeO _T	19.28	21.22	24.44	13.56	15.44	18.47
MgO	25.04	23.58	19.31	28.46	27.08	24.56
CaO	1.94	1.95	2.83	2.33	2.46	3.25
K ₂ O	0.00	0.00	0.00	0.05	0.00	0.00
TiO ₂	0.24	0.21	0.38	0.26	0.04	0.33
MnO	0.50	0.56	0.40	0.19	0.32	0.50
Cr ₂ O ₃	0.04	0.04	0.03	0.03	0.06	0.04
Total	99.87	100.82	99.06	98.60	99.06	99.71
Fe ₂ O ₃	4.43	3.31	1.27	4.57	4.29	4.71
FeC	15.29	18.24	23.30	9.45	11.58	14.23
Wo	0.04	0.04	0.06	0.05	0.05	0.06
En	0.69	0.65	0.55	0.78	0.74	0.67
Fs	0.27	0.31	0.39	0.18	0.21	0.26
Mg#	69.8	66.4	58.5	78.9	75.8	70.3

Sill B: Orthopyroxene

Label	F14-28-1	F14-28-2	F14-28-3	F16-14-1	F16-9-3	F16-9-4	F16-9-5
Height	297.7	297.7	297.7	287.9	243.7	243.7	243.7
Position	Rim	Interior	Interior	Interior	Interior	Interior	Interior
SiO ₂	51.36	51.46	51.79	52.43	50.31	50.62	51.68
Al ₂ O ₃	0.71	0.72	0.00	0.25	0.09	0.40	0.26
FeO _T	24.47	22.79	20.82	18.13	29.15	25.21	24.26
MgO	21.34	23.51	24.15	26.01	18.31	20.74	19.94
CaO	1.71	0.40	1.47	1.65	2.12	1.49	3.20
K ₂ O	0.00	0.00	0.00	0.04	0.00	0.00	0.00
TiO ₂	0.36	0.32	0.26	0.20	0.35	0.25	0.34
MnO	0.49	0.37	0.39	0.40	0.60	0.46	0.30
Cr ₂ O ₃	0.02	0.03	0.02	0.03	0.03	0.02	0.04
Total	100.46	99.60	98.89	99.14	100.96	99.19	100.02
Fe ₂ O ₃	3.83	4.17	4.43	4.73	5.15	4.21	2.33
FeO	21.02	19.03	16.84	13.88	24.51	21.42	22.16
Wo	0.03	0.01	0.03	0.03	0.04	0.03	0.06
En	0.60	0.66	0.67	0.71	0.52	0.59	0.56
Fs	0.37	0.34	0.30	0.25	0.44	0.38	0.37
Mg#	60.8	64.8	67.4	71.9	52.8	59.4	59.4

Table A3.3 (continued). Microprobe Analyses.

Sill C: Clinopyroxene

Label	SH-2-1	SH-2-2	SH-2-3	SH-2-4	SH-2-5	SH-3-32	SH-3-35	SH-6-1
Height	2.6	2.6	2.6	2.6	2.6	5.3	5.3	13.2
Position	Rim	Rim	Rim	Rim	Interior	Rim	Rim	Rim
SiO ₂	51.40	50.69	51.56	52.31	53.05	52.96	50.42	50.90
Al ₂ O ₃	1.17	0.00	0.42	0.00	0.00	1.01	2.46	1.92
FeO _T	13.38	17.14	16.01	12.13	9.53	10.29	10.17	9.47
MgO	11.45	9.83	10.85	12.84	14.20	17.36	16.25	17.16
CaO	22.32	20.77	21.14	21.80	23.01	18.32	18.67	18.57
K ₂ O	0.08	0.13	0.11	0.08	0.14	0.03	0.11	0.02
TiO ₂	0.79	0.13	0.21	0.45	0.13	0.39	0.63	0.46
MnO	0.45	0.34	0.26	0.21	0.08	0.26	0.20	0.20
Cr ₂ O ₃	6.19	0.11	0.13	0.21	0.19	0.14	0.28	0.45
Total	101.23	99.14	100.69	100.03	100.33	100.76	99.19	99.15
Fe ₂ O ₃	1.01	1.42	1.32	0.50	1.42	1.53	3.04	3.14
FeO	12.47	15.86	14.82	11.68	8.26	8.91	7.44	6.64
Wo	0.45	0.44	0.43	0.44	0.46	0.36	0.38	0.38
En	0.32	0.29	0.31	0.36	0.40	0.48	0.46	0.48
Fs	0.22	0.28	0.26	0.20	0.14	0.16	0.15	0.14
Mg#	60.4	50.5	54.7	65.4	72.6	75.0	74.0	76.4

Sill C: Clinopyroxene

Label	SH-6-2	SH-6-3	SH-6-4	SH-6-5	SH-6-7	SH-6-8	SH-6-9	SH-6-10
Height	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2
Position	Interior	Rim	Rim	Interior	Rim	Interior	Rim	Interior
SiO ₂	50.72	50.29	50.49	50.85	50.27	50.16	50.14	53.59
Al ₂ O ₃	1.55	1.66	1.75	1.83	1.49	0.93	1.19	1.39
FeO _T	9.41	10.73	9.14	9.15	13.10	13.78	13.49	8.39
MgO	16.79	16.30	16.43	17.30	12.04	11.97	12.18	18.19
CaO	19.04	18.58	19.65	19.48	21.27	20.73	21.53	18.79
K ₂ O	0.06	0.03	0.02	0.07	0.01	0.07	0.12	0.04
TiO ₂	0.45	0.53	0.46	0.43	0.73	0.66	0.64	0.38
MnO	0.15	0.26	0.26	0.21	0.51	0.23	0.22	0.09
Cr ₂ O ₃	0.33	0.17	0.33	0.54	0.11	0.11	0.12	0.23
Total	98.50	98.55	98.53	99.86	99.53	98.64	99.63	101.09
Fe ₂ O ₃	3.34	3.70	3.49	4.64	1.77	1.72	3.16	0.75
FeO	6.40	7.40	6.01	4.98	11.51	12.23	10.65	7.72
Wo	0.39	0.38	0.40	0.39	0.44	0.43	0.44	0.37
En	0.48	0.46	0.47	0.48	0.34	0.35	0.35	0.50
Fs	0.14	0.16	0.13	0.12	0.22	0.22	0.21	0.13
Mg#	76.1	73.0	76.2	77.1	62.1	60.8	61.7	79.4

Table A3.3 (continued). Microprobe Analyses.

Sill C: Clinopyroxene

Label	SH-6-47	SH-6-50	SH-6-51	SH-11-9	SH-11-11	SH-11-12	SH-11-13
Height	13.2	13.2	13.2	26.3	26.3	26.3	26.3
Position	Rim	Interior	Interior	Interior	Interior	Interior	Interior
SiO ₂	52.35	51.86	52.79	52.56	52.11	51.54	52.05
Al ₂ O ₃	0.00	0.31	1.06	1.47	1.36	1.71	1.58
FeO _T	10.90	13.80	11.91	9.16	8.75	9.81	9.33
MgO	13.32	11.61	11.61	18.62	18.10	16.87	18.40
CaO	22.64	21.80	22.50	18.05	18.55	19.01	17.64
K ₂ O	0.05	0.09	0.07	0.07	0.04	0.10	0.08
TiO ₂	0.10	0.32	0.14	0.35	0.34	0.43	0.37
MnO	0.21	0.30	0.43	0.16	0.24	0.30	0.18
Cr ₂ O ₃	0.11	0.06	0.19	0.30	0.33	0.33	0.24
Total	99.68	100.15	100.70	100.74	99.82	100.10	99.87
Fe ₂ O ₃	1.47	0.78	0.87	2.98	2.80	3.14	2.86
FeO	9.58	13.10	11.12	6.48	6.23	6.99	6.75
Wo	0.46	0.45	0.45	0.36	0.37	0.38	0.35
En	0.37	0.33	0.36	0.51	0.50	0.47	0.51
Fs	0.17	0.22	0.19	0.13	0.13	0.15	0.13
Mg#	68.5	60.0	63.5	78.4	78.7	75.4	77.8

Sill C: Clinopyroxene

Label	SH-11-15	SH-11-17	SH-11-18	SH-11-19	SH-11-20	SH-11-21	SH-11-22	SH-11-23	SH-11-24
Height	26.3	26.3	26.3	26.3	26.3	26.3	26.3	26.3	26.3
Position	Interior	Rim	Rim	Rim	Interior	Rim	Rim	Rim	Interior
SiO ₂	51.99	51.71	50.84	50.55	50.95	50.99	50.82	50.75	50.74
Al ₂ O ₃	1.34	1.45	1.42	1.41	1.20	1.72	1.31	1.82	1.57
FeO _T	10.47	11.04	13.30	12.51	13.46	10.76	12.92	12.85	12.78
MgO	17.13	17.05	15.21	15.33	14.43	16.49	15.13	14.06	14.97
CaO	17.90	18.17	18.44	18.59	19.09	18.66	18.72	19.14	18.05
K ₂ O	0.01	0.02	0.11	0.09	0.04	0.11	0.03	0.08	0.04
TiO ₂	0.46	0.53	0.69	0.71	0.78	0.65	0.67	0.68	0.65
MnO	0.15	0.26	0.37	0.32	0.37	0.17	0.32	0.34	0.18
Cr ₂ O ₃	0.20	0.10	0.07	0.07	0.12	0.11	0.11	0.14	0.12
Total	99.65	100.33	100.45	99.58	100.44	99.66	100.03	99.86	99.10
Fe ₂ O ₃	1.70	3.02	3.70	3.52	2.79	3.34	3.24	1.91	1.81
FeO	8.94	8.32	9.97	9.35	10.95	7.75	10.01	11.13	11.15
Wo	0.36	0.36	0.37	0.38	0.39	0.38	0.38	0.39	0.37
En	0.48	0.47	0.43	0.43	0.41	0.46	0.43	0.40	0.43
Fs	0.16	0.16	0.20	0.19	0.21	0.16	0.19	0.21	0.20
Mg#	74.5	73.3	67.1	68.6	65.6	73.2	67.6	66.1	67.6

Table A3.3 (continued). Microprobe Analyses.

Sill C: Clinopyroxene

Label	SH-11-25	SH-16-12	SH-16-13	SH-16-14	SH-16-16	SH-16-17	SH-16-18	SH-16-19
Height	26.3	40.1	40.1	40.1	40.1	40.1	40.1	40.1
Position	Rim	Rim	Interior	Rim	Rim	Interior	Interior	Rim
SiO ₂	51.28	50.55	50.88	50.52	51.05	51.74	51.21	50.95
Al ₂ O ₃	1.26	1.28	1.39	1.46	1.44	1.71	1.46	1.72
FeO _T	12.42	12.82	13.59	14.40	10.20	11.24	11.57	12.10
MgO	15.76	14.88	15.91	15.10	17.10	16.97	16.37	15.79
CaO	17.66	18.55	16.74	16.99	18.34	17.84	18.22	18.05
K ₂ O	0.10	0.06	0.08	0.11	0.09	0.12	0.03	0.08
TiO ₂	0.67	0.70	0.70	0.70	0.42	0.49	0.57	0.63
MnO	0.35	0.21	0.44	0.37	0.21	0.30	0.30	0.16
Cr ₂ O ₃	0.05	0.10	0.14	0.09	0.08	0.03	0.11	0.12
Total	99.55	99.15	99.87	99.74	98.93	100.44	99.84	99.60
Fe ₂ O ₃	2.08	2.70	2.90	3.06	3.60	3.00	3.04	2.54
FeO	10.55	10.39	10.98	11.65	6.96	8.54	8.83	9.82
Wo	0.36	0.38	0.34	0.35	0.37	0.36	0.37	0.37
En	0.45	0.42	0.45	0.43	0.48	0.48	0.46	0.45
Fs	0.20	0.20	0.21	0.22	0.14	0.17	0.17	0.19
Mg#	69.3	67.4	67.6	65.1	74.9	72.9	71.6	69.9

Sill C: Clinopyroxene

Label	SH-16-20	SH-22-1	SH-22-2	SH-22-3	SH-22-6	SH-22-7	SH-22-8	SH-22-9
Height	40.1	61.1	61.1	61.1	61.1	61.1	61.1	61.1
Position	Rim	Interior	Rim	Interior	Rim	Interior	Interior	Interior
SiO ₂	50.56	51.54	50.01	50.91	51.42	51.29	52.80	51.13
Al ₂ O ₃	0.87	1.81	1.38	1.53	2.28	1.69	1.49	1.60
FeO _T	14.18	11.80	12.37	10.56	9.79	9.66	8.96	7.58
MgO	13.09	16.06	15.15	15.84	16.15	14.48	17.94	16.02
CaO	19.01	18.15	18.88	19.19	19.99	21.17	19.14	21.16
K ₂ O	0.05	0.06	0.07	0.04	0.02	0.07	0.01	0.10
TiO ₂	0.20	0.49	0.82	0.53	0.61	0.60	0.42	0.45
MnO	0.45	0.36	0.27	0.25	0.28	0.49	0.10	0.27
Cr ₂ O ₃	0.00	0.17	0.09	0.16	0.23	0.30	0.38	0.53
Total	98.41	100.44	99.04	99.01	100.77	99.75	101.24	98.84
Fe ₂ O ₃	2.05	2.39	3.90	2.67	2.77	1.59	2.30	2.53
FeO	12.34	9.65	8.86	8.16	7.30	8.23	6.89	5.30
Wo	0.40	0.37	0.39	0.39	0.40	0.43	0.38	0.43
En	0.38	0.45	0.43	0.45	0.45	0.41	0.49	0.45
Fs	0.22	0.18	0.18	0.16	0.15	0.16	0.13	0.12
Mg#	62.2	70.8	68.6	72.8	74.6	72.8	78.1	79.0

Table A3.3 (continued). Microprobe Analyses.

Sill C: Clinopyroxene

Label	SH-22-10	F90-80-12	F90-80-13	F90-80-14	F90-80-15	F90-80-16	F90-80-17	F90-80-18
Height	61.1	104.5	104.5	104.5	104.5	104.5	104.5	104.5
Position	Rim	Rim	Interior	Rim	Interior	Interior	Interior	Rim
SiO ₂	50.86	50.27	50.66	51.49	51.53	51.17	51.52	51.44
Al ₂ O ₃	1.55	1.40	1.19	1.74	1.68	1.69	1.47	1.74
FeO _T	9.86	12.44	13.38	10.86	8.44	11.12	11.64	12.22
MgO	16.70	15.24	14.35	16.65	16.94	16.78	16.47	15.94
CaO	19.09	18.49	18.98	18.84	20.27	17.64	18.33	18.22
K ₂ O	0.15	0.09	0.04	0.11	0.08	0.12	0.03	0.08
TiO ₂	0.43	0.71	0.78	0.66	0.48	0.48	0.57	0.64
MnO	0.17	0.32	0.37	0.17	0.41	0.30	0.30	0.16
Cr ₂ O ₃	0.22	0.07	0.12	0.11	0.14	0.03	0.11	0.12
Total	99.03	99.02	99.87	100.63	99.97	99.34	100.45	100.56
Fe ₂ O ₃	3.90	3.50	2.77	3.37	3.31	2.97	3.06	2.56
FeO	6.35	9.29	10.89	7.83	5.46	8.45	8.89	9.91
Wo	0.39	0.38	0.39	0.38	0.40	0.36	0.37	0.37
En	0.47	0.43	0.41	0.46	0.46	0.48	0.46	0.45
Fs	0.14	0.19	0.21	0.16	0.13	0.17	0.17	0.19
Mg#	75.1	68.6	65.6	73.2	78.1	72.9	71.6	69.9

Sill C: Clinopyroxene

Label	F90-80-19	F90-80-20	SH-38-29	SH-38-30	SH-38-31	SH-38-32	SH-38-33	SH-38-34
Height	104.5	104.5	118.2	118.2	118.2	118.2	118.2	118.2
Position	Rim	Rim	Rim	Interior	Rim	Rim	Rim	Rim
SiO ₂	51.32	50.40	51.46	51.64	50.42	50.46	51.09	51.37
Al ₂ O ₃	0.88	1.55	0.22	1.47	1.12	0.00	0.27	0.39
FeO _T	14.40	11.05	14.63	11.67	13.07	15.68	15.81	16.23
MgO	13.29	14.11	12.48	16.51	13.89	10.57	10.94	11.03
CaO	19.29	21.71	20.81	18.37	19.47	22.32	20.83	20.38
K ₂ O	0.05	0.11	0.11	0.03	0.05	0.04	0.13	0.08
TiO ₂	0.79	0.85	0.31	0.57	0.62	0.02	0.28	0.15
MnO	0.46	0.32	0.21	0.30	0.37	0.21	0.40	0.39
Cr ₂ O ₃	0.00	0.21	0.12	0.11	0.13	0.05	0.15	0.08
Total	100.48	100.32	100.37	100.68	99.13	99.36	99.91	100.12
Fe ₂ O ₃	1.49	4.05	2.53	3.07	2.71	3.45	1.62	1.21
FeO	13.06	7.40	12.36	8.91	10.63	12.57	14.36	15.15
Wo	0.39	0.44	0.42	0.37	0.40	0.46	0.43	0.42
En	0.37	0.40	0.35	0.46	0.40	0.31	0.31	0.32
Fs	0.23	0.16	0.22	0.17	0.20	0.23	0.26	0.26
Mg#	62.2	69.5	60.3	71.6	65.4	54.6	55.2	54.8

Table A3.3 (continued). Microprobe Analyses.

Sill C: Clinopyroxene

Label	SH-38-35	SH-24-11	SH-24-12	SH-24-14	SH-24-15	SH-24-16	SH-24-18
Height	118.2	131.4	131.4	131.4	131.4	131.4	131.4
Position	Interior	Rim	Rim	Rim	Interior	Rim	Rim
SiO ₂	51.60	50.62	51.57	50.89	50.92	51.60	50.39
Al ₂ O ₃	1.51	1.52	1.67	1.65	1.60	1.44	1.70
FeO _T	8.61	10.67	8.44	9.26	8.44	12.12	10.55
MgO	16.97	16.53	17.03	16.36	16.74	14.96	14.01
CaO	20.17	18.62	20.23	20.12	20.02	19.06	21.97
K ₂ O	0.09	0.07	0.08	0.07	0.08	0.06	0.05
TiO ₂	0.42	0.40	0.47	0.45	0.44	0.78	0.48
MnO	0.43	0.21	0.11	0.32	0.29	0.25	0.26
Cr ₂ O ₃	0.12	0.24	0.44	0.32	0.27	0.14	0.24
Total	99.92	98.88	100.04	99.44	98.80	100.41	99.65
Fe ₂ O ₃	3.43	3.91	3.31	3.86	3.53	1.38	3.77
FeO	5.52	7.15	5.46	5.79	5.26	10.89	7.15
Wo	0.40	0.38	0.40	0.41	0.41	0.38	0.45
En	0.46	0.47	0.47	0.46	0.47	0.42	0.40
Fs	0.13	0.15	0.12	0.13	0.12	0.20	0.15
Mg#	77.8	73.4	78.2	75.9	78.0	68.7	70.3

Sill C: Orthopyroxene

Label	SH-24-19	SH-24-20	SH-28-38	SH-28-39	SH-28-40	SH-28-41	SH-28-42	SH-28-43
Height	131.4	131.4	140.6	140.6	140.6	140.6	140.6	140.6
Position	Interior	Rim	Rim	Rim	Interior	Rim	Rim	Interior
SiO ₂	52.27	51.30	50.73	51.36	51.46	51.35	51.51	51.53
Al ₂ O ₃	1.54	1.47	0.98	1.28	1.79	0.64	1.80	1.80
FeO _T	9.59	11.20	13.70	7.71	6.82	13.31	7.48	7.49
MgO	16.83	14.79	13.67	14.81	15.64	12.89	15.11	15.72
CaO	19.97	20.98	19.34	22.68	22.58	21.07	22.76	21.84
K ₂ O	0.02	0.00	0.08	0.05	0.10	0.12	0.06	0.12
TiO ₂	0.50	0.53	0.60	0.33	0.50	0.40	0.37	0.25
MnO	0.23	0.30	0.40	0.18	0.13	0.45	0.11	0.21
Cr ₂ O ₃	0.27	0.33	0.19	0.46	0.51	0.19	0.33	0.50
Total	101.22	100.90	99.69	98.86	99.53	100.42	99.53	99.46
Fe ₂ O ₃	2.79	3.25	2.54	1.98	2.31	2.60	2.15	2.47
FeO	7.08	8.28	11.41	5.92	4.74	10.97	5.54	5.27
Wo	0.40	0.42	0.40	0.46	0.46	0.43	0.46	0.44
En	0.46	0.41	0.39	0.42	0.44	0.36	0.43	0.44
Fs	0.14	0.17	0.22	0.12	0.11	0.21	0.11	0.11
Mg#	75.8	70.2	64.0	77.4	80.3	63.3	78.3	78.9

Table A3.3 (continued). Microprobe Analyses.

Sill C: Clinopyroxene

Label	SH-28-44	SH-28-45	SH-28-46	SH-28-47	SH-28-48	SH-28-49	SH-28-50
Height	140.6	140.6	140.6	140.6	140.6	140.6	140.6
Position	Interior	Rim	Rim	Interior	Interior	Interior	Interior
SiO ₂	51.64	50.47	51.47	50.56	51.24	51.38	50.90
Al ₂ O ₃	1.88	2.00	1.27	1.84	2.12	1.47	1.50
FeO _T	7.35	10.38	9.82	7.63	6.86	6.52	7.19
MgO	16.44	14.68	14.54	15.40	15.93	15.73	15.70
CaO	21.05	20.54	21.58	22.18	21.75	22.61	22.23
K ₂ O	0.06	0.06	0.06	0.08	0.13	0.04	0.06
TiO ₂	0.36	0.66	0.66	0.53	0.52	0.49	0.41
MnO	0.06	0.32	0.41	0.16	0.12	0.19	0.21
Cr ₂ O ₃	0.51	0.17	0.14	0.56	0.54	0.51	0.48
Total	99.35	99.28	99.95	98.94	99.21	98.94	98.68
Fe ₂ O ₃	1.99	2.69	2.05	3.29	2.11	2.18	3.13
FeO	5.56	7.96	7.98	4.67	4.96	4.56	4.38
Wo	0.42	0.42	0.44	0.45	0.44	0.46	0.45
En	0.46	0.42	0.41	0.44	0.45	0.44	0.44
Fs	0.11	0.16	0.16	0.11	0.11	0.10	0.11
Mg#	79.9	71.6	72.5	78.2	80.5	81.1	79.6

Sill C: Clinopyroxene

Label	SH-28-51	SH-28-52	SH-28-53	SH-28-54	SH-28-55	SH-28-56	SH-28-57
Height	140.6	140.6	140.6	140.6	140.6	140.6	140.6
Position	Interior	Interior	Interior	Interior	Interior	Interior	Rim
SiO ₂	51.67	50.79	50.25	50.92	50.61	49.95	51.90
Al ₂ O ₃	1.63	1.89	1.74	1.89	1.94	1.54	1.54
FeO _T	9.14	9.12	10.60	7.84	7.88	10.95	10.95
MgO	15.49	14.04	13.63	15.52	15.89	13.98	13.98
CaO	20.95	22.43	21.49	21.68	21.18	21.52	21.52
K ₂ O	0.01	0.08	0.06	0.05	0.07	0.11	0.11
TiO ₂	0.39	0.46	0.78	0.46	0.28	0.84	0.84
MnO	0.21	0.20	0.36	0.18	0.35	0.32	0.32
Cr ₂ O ₃	0.20	0.59	0.23	0.54	0.53	0.21	0.21
Total	99.69	99.60	99.14	99.08	98.73	99.42	101.37
Fe ₂ O ₃	1.88	2.42	2.43	2.57	3.48	4.01	1.42
FeO	7.45	6.94	8.42	5.53	4.75	7.34	9.67
Wo	0.42	0.46	0.44	0.44	0.43	0.44	0.43
En	0.44	0.40	0.39	0.44	0.45	0.40	0.39
Fs	0.14	0.14	0.17	0.12	0.12	0.16	0.18
Mg#	75.1	73.3	69.6	77.9	78.2	69.5	69.5

Table A3.3 (continued). Microprobe Analyses.

Sill C: Clinopyroxene

Label	SH-28-58	SH-28-59	SH-28-61	SH-28-60	SH-28-62	SH-35-1	SH-35-2
Height	131.4	140.6	140.6	140.6	140.6	156.5	156.5
Position	Rim	Rim	Rim	Rim	Rim	Rim	Interior
SiO ₂	50.89	50.82	50.16	50.63	50.86	52.43	50.85
Al ₂ O ₃	0.22	1.13	0.00	0.27	0.39	1.51	1.85
FeO _T	14.47	13.17	15.56	15.67	16.07	8.57	8.36
MgO	12.34	14.00	10.49	10.84	10.92	17.28	15.31
CaO	20.58	19.62	22.15	20.64	20.18	19.71	21.71
K ₂ O	0.11	0.05	0.04	0.13	0.08	0.08	0.04
TiO ₂	0.31	0.62	0.02	0.28	0.15	0.39	0.44
MnO	0.41	0.37	0.45	0.36	0.55	0.15	0.37
Cr ₂ O ₃	0.12	0.13	0.05	0.15	0.08	0.24	0.42
Total	99.45	99.91	98.92	98.97	99.28	100.36	99.35
Fe ₂ O ₃	2.72	2.73	3.57	1.55	1.38	2.19	3.06
FeO	12.02	10.71	12.35	14.27	14.83	6.60	5.60
Wo	0.42	0.40	0.46	0.43	0.42	0.39	0.44
En	0.35	0.40	0.30	0.31	0.32	0.48	0.43
Fs	0.22	0.20	0.23	0.25	0.26	0.13	0.13
Mg#	60.3	65.4	54.6	55.2	54.8	78.2	76.5

Sill C: Clinopyroxene

Label	SH-35-3	SH-35-5	SH-35-4	SH-35-6	SH-35-7	SH-35-8	SH-35-9	SH-35-11
Height	156.5	156.5	156.5	156.5	156.5	156.5	156.5	156.5
Position	Rim	Interior	Rim	Rim	Interior	Interior	Rim	Rim
SiO ₂	51.70	52.17	50.65	52.08	52.36	51.96	51.81	51.19
Al ₂ O ₃	1.33	1.25	1.72	1.58	1.60	1.60	1.36	1.59
FeO _T	10.93	7.85	10.72	9.53	8.22	8.11	10.21	10.67
MgO	15.88	16.43	13.96	16.22	16.83	16.64	16.43	16.06
CaO	19.25	20.18	21.36	19.80	20.31	20.37	19.41	19.08
K ₂ O	0.04	0.08	0.05	0.08	0.10	0.04	0.02	0.10
TiO ₂	0.42	0.45	0.80	0.51	0.51	0.48	0.53	0.62
MnO	0.24	0.23	0.31	0.23	0.21	0.17	0.23	0.32
Cr ₂ O ₃	0.16	0.36	0.16	0.34	0.30	0.40	0.25	0.46
Total	99.95	99.00	99.73	100.37	100.44	99.77	100.25	100.09
Fe ₂ O ₃	2.29	0.75	2.39	1.72	1.87	1.77	2.47	2.89
FeO	8.87	7.17	8.57	7.98	6.54	6.52	7.99	8.07
Wo	0.39	0.41	0.43	0.40	0.40	0.41	0.39	0.38
En	0.45	0.46	0.40	0.45	0.47	0.46	0.46	0.45
Fs	0.17	0.13	0.17	0.15	0.13	0.13	0.15	0.17
Mg#	72.1	78.9	69.9	75.2	78.5	78.5	74.1	72.8

Table A3.3 (continued). Microprobe Analyses.

Sill C: Clinopyroxene

Label	SH-35-10	SH-35-12	SH-35-13	SH-35-14	SH-35-15	SH-35-16	SH-35-17
Height	156.5	156.5	156.5	156.5	156.5	156.5	156.5
Position	Rim	Interior	Rim	Interior	Interior	Interior	Interior
SiO ₂	50.64	51.77	51.44	51.61	51.72	51.24	52.32
Al ₂ O ₃	2.37	1.73	1.56	1.64	1.52	2.11	1.80
FeO _T	8.39	8.18	9.82	9.29	9.17	9.38	8.16
MgO	16.51	17.13	15.94	14.77	15.82	15.92	16.44
CaO	19.87	20.63	19.41	21.20	20.59	20.20	20.65
K ₂ O	0.12	0.07	0.00	0.08	0.10	0.06	0.04
TiO ₂	0.39	0.40	0.37	0.45	0.48	0.39	0.45
MnO	0.18	0.26	0.24	0.48	0.24	0.19	0.05
Cr ₂ O ₃	0.37	0.44	0.33	0.34	0.22	0.28	0.29
Total	98.84	100.61	99.11	99.86	99.86	99.77	100.20
Fe ₂ O ₃	3.24	3.73	1.66	1.54	2.24	2.65	1.25
FeO	5.47	4.83	8.32	7.90	7.15	6.99	7.04
Wo	0.41	0.41	0.39	0.43	0.42	0.41	0.41
En	0.47	0.47	0.45	0.42	0.44	0.45	0.46
Fs	0.12	0.11	0.16	0.15	0.14	0.14	0.13
Mg#	77.8	78.9	74.3	73.9	75.5	75.2	78.2

Sill C: Clinopyroxene

Label	SH-35-18	SH-35-19	SH-35-20	SH-41-19	SH-41-20	SH-41-21	SH-41-22
Height	156.5	156.5	156.5	170.1	170.1	170.1	170.1
Position	Interior	Interior	Interior	Rim	Rim	Interior	Interior
SiO ₂	51.74	51.87	51.54	52.18	51.31	52.97	52.12
Al ₂ O ₃	1.75	1.59	1.32	1.44	1.78	1.48	1.80
FeO _T	7.96	7.35	7.13	8.97	11.24	9.10	8.26
MgO	16.15	16.13	15.78	15.21	13.80	16.27	15.64
CaO	21.37	21.92	22.16	21.35	21.25	19.83	21.16
K ₂ O	0.11	0.08	0.08	0.08	0.11	0.06	0.15
TiO ₂	0.47	0.43	0.46	0.57	0.68	0.56	0.41
MnO	0.23	0.17	0.21	0.39	0.35	0.28	0.18
Cr ₂ O ₃	0.45	0.35	0.48	0.15	0.10	0.20	0.23
Total	100.23	99.89	99.16	100.34	100.62	100.75	99.95
Fe ₂ O ₃	2.67	2.43	2.29	1.29	1.98	0.14	1.32
FeO	5.56	5.16	5.07	7.81	9.46	8.97	7.07
Wo	0.43	0.44	0.45	0.43	0.43	0.39	0.43
En	0.45	0.45	0.44	0.42	0.39	0.45	0.44
Fs	0.12	0.11	0.11	0.15	0.18	0.16	0.13
Mg#	78.3	79.6	79.8	75.1	68.6	76.1	77.1

Table A3.3 (continued). Microprobe Analyses.

Sill C: Clinopyroxene

Label	SH-41-23	SH-41-24	SH-41-25	SH-41-26	SH-41-27	SH-41-28	F876-3-11
Height	170.1	170.1	170.1	170.1	170.1	170.1	183.2
Position	Interior	Interior	Interior	Interior	Interior	Interior	Rim
SiO ₂	51.50	50.87	51.73	51.25	52.33	52.63	51.86
Al ₂ O ₃	1.55	2.17	1.90	2.85	1.73	1.55	1.49
FeO _T	11.96	9.52	9.70	12.50	8.85	7.90	8.48
MgO	13.98	15.86	15.35	14.34	15.77	15.43	17.09
CaO	20.01	19.62	20.62	18.99	21.01	21.61	19.49
K ₂ O	0.02	0.04	0.07	0.08	0.07	0.01	0.08
TiO ₂	0.69	0.38	0.38	0.54	0.41	0.47	0.39
MnO	0.21	0.27	0.17	0.20	0.30	0.33	0.15
Cr ₂ O ₃	0.14	0.25	0.13	0.10	0.32	0.20	0.24
Total	100.06	98.98	100.05	100.85	100.79	100.13	99.26
Fe ₂ O ₃	0.65	2.39	1.85	1.18	1.61	0.11	2.17
FeO	11.38	7.37	8.04	11.44	7.40	7.80	6.53
Wo	0.41	0.40	0.42	0.39	0.42	0.43	0.39
En	0.39	0.45	0.43	0.41	0.44	0.43	0.48
Fs	0.20	0.15	0.15	0.20	0.14	0.14	0.13
Mg#	67.6	74.8	73.8	67.2	76.1	77.7	78.2

Sill C: Clinopyroxene

Label	F876-3-12	F876-3-13	F876-3-14	F876-3-14	F876-3-15	F876-9-32	F876-9-33	F876-9-34
Height	183.2	183.2	183.2	183.2	183.2	199.0	199.0	199.0
Position	Interior	Rim	Interior	Interior	Rim	Rim	Interior	Interior
SiO ₂	51.27	51.32	52.90	50.85	51.15	50.94	50.43	50.96
Al ₂ O ₃	1.87	1.32	1.27	1.45	1.14	1.42	1.19	2.10
FeO _T	8.43	10.85	7.96	11.49	13.26	12.61	13.32	9.33
MgO	15.44	15.76	16.66	16.26	14.09	15.45	14.28	15.83
CaO	21.89	19.11	20.46	18.09	19.75	18.73	18.89	20.09
K ₂ O	0.04	0.04	0.08	0.03	0.05	0.09	0.04	0.06
TiO ₂	0.44	0.42	0.46	0.57	0.62	0.72	0.77	0.39
MnO	0.37	0.24	0.23	0.30	0.37	0.32	0.37	0.19
Cr ₂ O ₃	0.42	0.16	0.37	0.11	0.13	0.07	0.12	0.28
Total	100.17	99.21	100.39	99.14	100.56	100.34	99.41	99.23
Fe ₂ O ₃	3.09	2.27	0.76	3.02	2.75	3.54	2.76	2.64
FeO	5.65	8.80	7.28	8.77	10.78	9.42	10.84	6.96
Wo	0.44	0.39	0.41	0.37	0.40	0.38	0.39	0.41
En	0.43	0.45	0.46	0.46	0.40	0.43	0.41	0.45
Fs	0.13	0.17	0.13	0.17	0.20	0.19	0.21	0.14
Mg#	76.5	72.1	78.9	71.6	65.4	68.6	65.6	75.2

Table A3.3 (continued). Microprobe Analyses.

Sill C: Clinopyroxene

Label	F876-9-36	F876-9-39	F876-9-40	F876-9-41	F876-9-42	F876-9-43	F876-23-3
Height	199.0	199.0	199.0	199.0	199.0	199.0	221.7
Position	Interior	Interior	Interior	Interior	Rim	Rim	Rim
SiO ₂	52.01	51.32	52.20	52.02	51.62	50.58	50.96
Al ₂ O ₃	1.79	1.74	1.60	1.33	1.42	1.75	0.42
FeO _T	8.11	7.89	7.40	7.20	8.87	11.08	15.82
MgO	16.34	16.02	16.23	15.93	15.05	13.60	10.72
CaO	20.53	21.20	22.06	22.37	21.12	20.95	20.89
K ₂ O	0.04	0.11	0.08	0.08	0.08	0.11	0.11
TiO ₂	0.45	0.47	0.43	0.46	0.56	0.67	0.21
MnO	0.05	0.23	0.17	0.21	0.39	0.35	0.33
Cr ₂ O ₃	0.29	0.45	0.35	0.48	0.15	0.10	0.13
Total	99.61	99.41	100.53	100.08	99.26	99.19	99.59
Fe ₂ O ₃	1.24	2.64	2.44	2.31	1.28	1.95	1.38
FeO	7.00	5.52	5.20	5.12	7.72	9.32	14.58
Wo	0.41	0.43	0.44	0.45	0.43	0.43	0.43
En	0.46	0.45	0.45	0.44	0.42	0.39	0.31
Fs	0.13	0.12	0.11	0.11	0.15	0.18	0.26
Mg#	78.2	78.3	79.6	79.8	75.1	68.6	54.7

Sill C: Clinopyroxene

Label	F876-23-6	F876-23-7
Height	221.7	221.7
Position	Rim	Interior
SiO ₂	51.49	51.98
Al ₂ O ₃	1.36	0.00
FeO _T	12.26	9.46
MgO	12.97	14.10
CaO	22.03	22.84
K ₂ O	0.08	0.14
TiO ₂	0.57	0.13
MnO	0.21	0.08
Cr ₂ O ₃	0.21	0.19
Total	101.19	98.92
Fe ₂ O ₃	2.20	2.32
FeO	10.28	7.37
Wo	0.44	0.46
En	0.36	0.40
Fs	0.19	0.14
Mg#	65.4	72.6

Sill B: Clinopyroxene

Label	F16-6-11	F16-6-12	F16-6-13
Height	321.3	321.3	321.3
Position	Interior	Rim	Rim
SiO ₂	51.67	52.17	51.72
Al ₂ O ₃	1.59	1.37	1.61
FeO _T	8.06	10.28	10.78
MgO	16.55	16.54	16.23
CaO	20.26	19.55	19.28
K ₂ O	0.04	0.02	0.10
TiO ₂	0.48	0.53	0.63
MnO	0.17	0.23	0.32
Cr ₂ O ₃	0.40	0.25	0.46
Total	99.21	100.95	101.12
Fe ₂ O ₃	1.76	2.49	2.92
FeO	6.48	8.04	8.16
Wo	0.41	0.39	0.38
En	0.46	0.46	0.45
Fs	0.13	0.15	0.17
Mg#	78.5	74.1	72.8

Table A3.3 (continued). Microprobe Analyses.

Sill B: Clinopyroxene

Label	F16-6-15	F16-6-16	F16-6-19	F16-9-1	F16-9-2	F16-9-12	F16-9-15
Height	321.3	321.3	321.3	297.7	297.7	297.7	297.7
Position	Rim	Interior	Rim	Interior	Interior	Interior	Rim
SiO ₂	50.96	51.02	52.22	52.09	51.44	50.94	51.25
Al ₂ O ₃	2.39	1.70	1.58	1.66	1.51	2.10	0.00
FeO _T	8.44	8.06	9.97	9.38	9.12	9.32	15.72
MgO	16.62	16.88	16.18	14.91	15.73	15.83	10.60
CaO	20.00	20.33	19.71	21.40	20.48	20.08	22.37
K ₂ O	0.12	0.07	0.00	0.08	0.10	0.06	0.04
TiO ₂	0.39	0.39	0.38	0.45	0.48	0.39	0.02
MnO	0.18	0.26	0.24	0.48	0.24	0.19	0.47
Cr ₂ O ₃	0.37	0.43	0.34	0.34	0.22	0.28	0.05
Total	99.47	99.15	100.62	100.79	99.31	99.18	100.52
Fe ₂ O ₃	3.26	3.67	1.69	1.56	2.23	2.64	2.85
FeO	5.51	4.76	8.45	7.98	7.11	6.95	13.15
Wo	0.41	0.41	0.39	0.43	0.42	0.41	0.46
En	0.47	0.47	0.45	0.42	0.44	0.45	0.30
Fs	0.12	0.11	0.16	0.15	0.14	0.14	0.24
Mg#	77.8	78.9	74.3	73.9	75.5	75.2	54.6

Sill B: Clinopyroxene

Label	F16-9-16	F16-14-5	F16-14-6	F16-14-7	F16-14-13	F16-14-14	F14-16-2
Height	297.7	242.9	242.9	242.9	242.9	242.9	285.2
Position	Interior	Interior	Interior	Interior	Rim	Rim	Rim
SiO ₂	51.68	51.40	51.46	52.09	51.70	50.58	51.53
Al ₂ O ₃	1.78	1.74	1.58	1.33	1.43	1.75	0.40
FeO _T	8.06	7.91	7.29	7.21	8.89	11.08	16.28
MgO	16.24	16.05	16.00	15.95	15.07	13.60	11.06
CaO	20.40	21.23	21.75	22.39	21.15	20.95	20.45
K ₂ O	0.04	0.11	0.08	0.08	0.08	0.11	0.08
TiO ₂	0.44	0.47	0.43	0.46	0.56	0.67	0.15
MnO	0.05	0.23	0.17	0.21	0.39	0.35	0.51
Cr ₂ O ₃	0.29	0.45	0.35	0.49	0.15	0.10	0.08
Total	98.97	99.58	99.11	100.21	99.42	99.19	100.54
Fe ₂ O ₃	1.23	2.65	2.41	2.31	1.28	1.95	1.34
FeO	6.95	5.53	5.12	5.13	7.74	9.32	15.08
Wo	0.41	0.43	0.44	0.45	0.43	0.43	0.42
En	0.46	0.45	0.45	0.44	0.42	0.39	0.32
Fs	0.13	0.12	0.11	0.11	0.15	0.18	0.26
Mg#	78.2	78.3	79.6	79.8	75.1	68.6	54.8

Table A3.3 (continued). Microprobe Analyses.

Sill B: Clinopyroxene								
Label	F14-16-5	F14-16-6	F14-16-17	F14-16-18	F14-16-19	F14-16-31	F14-16-32	F14-16-33
Height	285.2	285.2	285.2	285.2	285.2	285.2	285.2	285.2
Position	Interior	Interior	Interior	Rim	Interior	Rim	Rim	Interior
SiO ₂	51.32	50.29	51.46	50.49	52.20	52.18	51.09	51.66
Al ₂ O ₃	1.88	2.87	1.70	1.12	1.54	1.50	0.27	1.88
FeO _T	9.62	12.61	8.70	13.08	7.84	8.53	15.81	8.49
MgO	15.23	14.46	15.51	13.91	15.30	17.20	10.94	15.55
CaO	20.46	19.15	20.66	19.49	21.43	19.62	20.83	22.06
K ₂ O	0.07	0.08	0.07	0.05	0.01	0.08	0.13	0.04
TiO ₂	0.38	0.54	0.40	0.62	0.47	0.39	0.28	0.45
MnO	0.17	0.20	0.30	0.37	0.33	0.15	0.46	0.38
Cr ₂ O ₃	0.13	0.10	0.31	0.13	0.20	0.24	0.15	0.43
Total	99.25	100.31	99.11	99.26	99.31	99.89	99.96	100.93
Fe ₂ O ₃	1.83	3.06	1.58	2.72	0.10	2.18	1.68	3.11
FeO	7.97	9.86	7.28	10.64	7.74	6.57	14.30	5.69
Wo	0.42	0.39	0.42	0.40	0.43	0.39	0.43	0.44
En	0.43	0.41	0.44	0.40	0.43	0.48	0.31	0.43
Fs	0.15	0.19	0.14	0.20	0.14	0.13	0.26	0.13
Mg#	73.8	67.2	76.1	65.4	77.7	78.2	55.2	76.5

Sill B: Clinopyroxene								
Label	F14-16-37	F14-16-38	F14-28-7	F14-28-8	F14-28-19	F14-28-23	F14-28-24	F14-28-25
Height	285.2	285.2	243.7	243.7	243.7	243.7	243.7	243.7
Position	Rim	Interior	Interior	Interior	Interior	Interior	Rim	Rim
SiO ₂	52.05	52.24	52.31	52.40	51.88	51.51	50.52	51.44
Al ₂ O ₃	1.34	1.25	1.46	1.81	1.56	2.20	1.56	0.22
FeO _T	11.00	7.86	8.99	8.30	12.05	9.64	11.08	14.63
MgO	15.99	16.45	16.07	15.72	14.08	16.06	14.14	12.47
CaO	19.38	20.21	19.58	21.27	20.16	19.87	21.77	20.80
K ₂ O	0.04	0.08	0.06	0.15	0.02	0.04	0.11	0.11
TiO ₂	0.42	0.45	0.55	0.41	0.70	0.38	0.85	0.31
MnO	0.24	0.23	0.28	0.18	0.21	0.27	0.32	0.53
Cr ₂ O ₃	0.16	0.36	0.20	0.23	0.14	0.25	0.21	0.12
Total	100.62	99.13	99.50	100.48	100.79	100.23	100.56	100.64
Fe ₂ O ₃	2.31	0.75	0.14	1.33	0.65	2.42	4.06	2.88
FeO	8.93	7.18	8.86	7.11	11.46	7.47	7.42	12.03
Wo	0.39	0.41	0.39	0.43	0.41	0.40	0.44	0.42
En	0.45	0.46	0.45	0.44	0.39	0.45	0.40	0.35
Fs	0.17	0.13	0.16	0.13	0.20	0.15	0.16	0.22
Mg#	72.1	78.9	76.1	77.1	67.6	74.8	69.5	60.3

Table A3.3 (continued). Microprobe Analyses.

Sill C: Groundmass Plagioclase

Label	SH-2-7	SH-2-8	SH-2-9		SH-2-10		SH-2-11	SH-2-12
Height	2.6	2.6	2.6		2.6		2.6	2.6
Type	U	U	U		U		U	U
Position	Intermed.	Core	Core		Core		Core	Core
SiO ₂	64.37	64.89	65.61		63.92		65.76	62.94
Al ₂ O ₃	19.75	21.91	19.87		20.93		20.33	22.64
FeO _T	0.28	0.17	0.13		0.36		0.22	0.44
CaO	0.70	1.42	0.39		1.03		0.63	1.41
Na ₂ O	11.81	11.13	12.28		11.67		11.96	11.67
K ₂ O	0.21	0.29	0.31		0.19		0.27	0.15
Total	97.12	99.81	98.59		98.10		99.17	99.25
An	0.03	0.06	0.02		0.05		0.03	0.06
Ab	0.96	0.92	0.97		0.94		0.96	0.93
Or	0.01	0.02	0.02		0.01		0.01	0.01

Label	SH-3-1	SH-3-2	SH-3-3	SH-3-4	SH-3-5	SH-3-6	SH-3-7	SH-3-8	SH-3-9	SH-3-10
Height	5.3	5.3	5.3	5.3	5.3	5.3	5.3	5.3	5.3	5.3
Type	U	U	U	U	U	U	U	U	U	U
Position	Intermed.	Core	Core	Intermed.	Rim	Rim	Intermed.	Core	Intermed.	Core
SiO ₂	55.80	54.09	53.65	55.00	61.44	58.40	55.98	55.97	55.98	55.07
Al ₂ O ₃	27.58	28.68	27.98	29.18	24.98	26.06	28.46	28.46	28.04	28.34
FeO _T	0.82	0.96	0.89	0.92	0.60	0.69	0.74	0.83	0.75	0.85
CaO	10.34	11.78	11.00	11.66	5.87	7.85	10.68	10.64	10.58	10.92
Na ₂ O	5.43	4.78	4.94	4.95	8.41	7.35	5.84	5.82	5.42	5.64
K ₂ O	0.24	0.17	0.12	0.10	0.39	0.20	0.08	0.12	0.15	0.19
Total	100.21	100.46	98.58	101.81	101.69	100.55	101.78	101.84	100.92	101.01
An	0.51	0.57	0.55	0.56	0.27	0.37	0.50	0.50	0.51	0.51
Ab	0.48	0.42	0.45	0.43	0.71	0.62	0.50	0.49	0.48	0.48
Or	0.01	0.01	0.01	0.01	0.02	0.01	0	0.01	0.01	0.01

Label	SH-3-11	SH-3-12	SH-3-18	SH-3-19	SH-3-17	SH-3-20	SH-3-21	SH-3-22
Height	5.3	5.3	5.3	5.3	5.3	5.3	5.3	5.3
Type	U	U	U	U	U	U	U	U
Position	Intermed.	Rim	Rim	Rim	Rim	Intermed.	Core	Intermed.
SiO ₂	55.43	62.05	61.97	61.37	58.29	64.21	54.66	55.47
Al ₂ O ₃	28.31	24.16	24.12	23.91	26.59	22.92	27.90	27.89
FeO _T	0.97	0.51	0.48	0.61	0.62	0.16	0.75	0.67
CaO	10.82	4.97	5.21	5.30	8.48	3.93	10.30	10.64
Na ₂ O	5.63	9.44	9.35	8.86	7.29	10.37	5.67	5.85
K ₂ O	0.18	0.18	0.14	0.09	0.21	0.36	0.07	0.18
Total	101.34	101.31	101.27	100.14	101.48	101.95	99.35	100.70
An	0.51	0.22	0.23	0.25	0.39	0.17	0.50	0.50
Ab	0.48	0.77	0.76	0.75	0.60	0.81	0.50	0.49
Or	0.01	0.01	0.01	0.01	0.01	0.02	0.00	0.01

Table A3.3 (continued). Microprobe Analyses.

Sill C: Groundmass Plagioclase

Label	SH-3-23	SH-3-26	SH-3-27	SH-3-31		SH-6-13	SH-6-14	SH-6-15	
Height	5.3	5.3	5.3	5.3		13.2	13.2	13.2	
Type	U	U	U	U		U	U	U	
Position	Intermed.	Intermed.	Core	Intermed.		Rim	Intermed.	Intermed.	
SiO ₂	56.68	55.46	55.37	54.65		63.65	61.03	58.88	
Al ₂ O ₃	26.57	28.08	28.10	27.72		23.71	24.37	26.26	
FeO _T	0.70	0.76	0.64	0.69		0.24	0.56	0.58	
CaO	9.12	10.33	10.36	10.86		4.53	5.85	7.83	
Na ₂ O	6.49	5.77	6.01	5.58		9.18	8.21	6.96	
K ₂ O	0.14	0.24	0.13	0.11		0.25	0.19	0.17	
Total	99.70	100.64	100.61	99.61		101.56	100.21	100.68	
An	0.43	0.49	0.48	0.52		0.21	0.28	0.38	
Ab	0.56	0.50	0.51	0.48		0.77	0.71	0.61	
Or	0.01	0.01	0.01	0.01		0.01	0.01	0.01	
Label	SH-6-16	SH-6-17	SH-6-18	SH-6-19	SH-6-20	SH-6-21	SH-6-22	SH-6-23	SH-6-24
Height	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2
Type	U	U	U	U	U	U	U	U	U
Position	Rim	Intermed.	Intermed.	Core	Core	Intermed.	Rim	Rim	Intermed.
SiO ₂	59.05	62.74	62.31	57.43	55.93	58.58	66.96	53.90	54.58
Al ₂ O ₃	26.18	23.96	23.98	26.83	27.37	27.03	20.18	28.20	27.86
FeO _T	0.72	0.50	0.44	0.40	0.52	0.62	0.48	0.77	0.77
CaO	7.73	5.05	4.98	8.94	9.61	8.20	1.15	10.64	10.74
Na ₂ O	7.36	8.93	8.96	6.44	6.13	6.67	10.83	5.53	5.30
K ₂ O	0.15	0.39	0.44	0.21	0.21	0.20	0.61	0.23	0.24
Total	101.19	101.57	101.11	100.25	99.77	101.30	100.21	99.27	99.49
An	0.36	0.23	0.23	0.43	0.46	0.40	0.05	0.51	0.52
Ab	0.63	0.75	0.75	0.56	0.53	0.59	0.91	0.48	0.47
Or	0.01	0.02	0.02	0.01	0.01	0.01	0.03	0.01	0.01
Label	SH-6-25	SH-6-26	SH-6-27	SH-6-28	SH-6-34	SH-6-29	SH-6-30	SH-6-31	
Height	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	
Type	U	U	U	U	U	U	U	U	
Position	Rim	Intermed.	Core	Intermed.	Rim	Rim	Core	Core	
SiO ₂	57.99	59.13	59.63	60.11	60.33	59.64	57.75	57.00	
Al ₂ O ₃	24.87	26.20	25.90	25.33	25.04	25.41	26.09	26.71	
FeO _T	1.60	0.47	0.48	0.28	0.33	0.49	0.46	0.68	
CaO	7.32	8.13	7.70	7.38	6.41	7.12	8.13	8.79	
Na ₂ O	6.92	6.91	7.32	7.30	7.80	7.50	6.60	6.28	
K ₂ O	0.17	0.35	0.40	0.34	0.23	0.16	0.27	0.25	
Total	98.87	101.19	101.43	100.74	100.14	100.32	99.30	99.71	
An	0.37	0.39	0.36	0.35	0.31	0.34	0.40	0.43	
Ab	0.62	0.59	0.62	0.63	0.68	0.65	0.59	0.56	
Or	0.01	0.02	0.02	0.02	0.01	0.01	0.02	0.01	

Table A3.3 (continued). Microprobe Analyses.

Sill C: Groundmass Plagioclase

Label	SH-6-32	SH-6-33	SH-6-36	SH-6-35	SH-6-37	SH-6-38	SH-6-39	SH-6-40	SH-6-41
Height	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2	13.2
Type	U	U	U	U	U	U	U	U	U
Position	Intermed.	Rim	Intermed.	Rim	Core	Core	Intermed.	Intermed.	Rim
SiO ₂	50.92	57.42	50.88	62.54	50.71	51.53	52.76	52.14	55.83
Al ₂ O ₃	29.69	25.93	29.50	22.25	30.70	31.03	30.18	29.93	27.64
FeO _T	0.94	0.53	0.91	0.52	0.72	0.58	0.76	1.00	0.65
CaO	12.95	7.97	13.06	4.08	13.26	12.82	12.50	13.35	9.88
Na ₂ O	3.98	7.02	3.80	9.08	3.31	2.97	3.89	2.81	5.64
K ₂ O	0.14	0.11	0.16	0.30	0.26	0.50	0.20	0.09	0.20
Total	98.62	98.98	98.31	98.77	98.96	99.43	100.29	99.32	99.84
An	0.64	0.38	0.65	0.20	0.68	0.68	0.63	0.72	0.49
Ab	0.35	0.61	0.34	0.79	0.31	0.29	0.36	0.27	0.50
Or	0.01	0.01	0.01	0.02	0.02	0.03	0.01	0.01	0.01
Label	SH-6-42	SH-6-43	SH-6-44	SH-6-45	SH-6-46	SH-11-35	SH-11-36	SH-11-37	
Height	13.2	13.2	13.2	13.2	13.2	26.3	26.3	26.3	
Type	U	U	U	U	U	E	E	E	
Position	Rim	Intermed.	Core	Intermed.	Intermed.	Core	Intermed.	Intermed.	
SiO ₂	60.34	58.34	58.71	58.61	63.50	51.26	55.08	48.26	
Al ₂ O ₃	24.59	26.09	26.33	26.72	23.93	31.21	28.60	32.63	
FeO _T	0.34	0.55	0.51	0.75	0.40	1.16	0.88	0.77	
CaO	6.02	7.77	8.11	8.48	4.82	13.76	10.68	15.13	
Na ₂ O	7.73	6.63	6.29	6.46	8.03	3.40	5.60	2.31	
K ₂ O	0.21	0.26	0.27	0.32	0.28	0.12	0.29	0.13	
Total	99.23	99.64	100.22	101.34	100.96	100.91	101.13	99.23	
An	0.30	0.39	0.41	0.41	0.24	0.69	0.50	0.78	
Ab	0.69	0.60	0.57	0.57	0.74	0.31	0.48	0.21	
Or	0.01	0.02	0.02	0.02	0.02	0.01	0.02	0.01	
Label	SH-11-38	SH-11-39	SH-11-40	SH-11-43	SH-11-44	SH-11-45	SH-11-46		
Height	26.3	26.3	26.3	26.3	26.3	26.3	26.3		
Type	E	E	E	U	U	U	U		
Position	Rim	Intermed.	Rim	Intermed.	Intermed.	Intermed.	Intermed.		
SiO ₂	55.69	52.72	54.04	51.79	53.02	53.37	53.55		
Al ₂ O ₃	27.03	29.30	29.33	30.11	28.99	29.10	29.51		
FeO _T	0.65	0.63	0.67	0.62	0.71	0.57	0.77		
CaO	9.18	10.83	12.03	13.12	12.16	12.04	11.32		
Na ₂ O	7.12	5.58	4.60	3.68	4.63	4.46	4.70		
K ₂ O	0.30	0.25	0.22	0.16	0.21	0.21	0.22		
Total	99.97	99.31	100.89	99.48	99.72	99.75	100.07		
An	0.41	0.51	0.58	0.66	0.58	0.59	0.56		
Ab	0.57	0.48	0.40	0.33	0.40	0.40	0.42		
Or	0.02	0.01	0.01	0.01	0.01	0.01	0.01		

Table A3.3 (continued). Microprobe Analyses.

Sill C: Groundmass Plagioclase

Label	SH-11-47		SH-11-54		SH-11-55		SH-11-56		SH-11-57		SH-11-58	
Height	26.3		26.3		26.3		26.3		26.3		26.3	
Type	U		E		E		E		U		U	
Position	Intermed.		Intermed.		Core		Core		Rim		Intermed.	
SiO ₂	52.59		48.24		49.29		48.60		54.06		53.86	
Al ₂ O ₃	29.14		33.38		32.59		32.46		29.60		29.92	
FeO _T	0.96		0.71		0.48		0.50		0.73		0.58	
CaO	11.59		16.71		15.75		15.96		12.00		12.52	
Na ₂ O	4.75		1.67		2.07		2.18		4.74		3.88	
K ₂ O	0.19		0.07		0.10		0.14		0.18		0.17	
Total	99.22		100.78		100.28		99.84		101.31		100.93	
An	0.57		0.84		0.80		0.80		0.58		0.63	
Ab	0.42		0.15		0.19		0.20		0.41		0.36	
Or	0.01		0.00		0.01		0.01		0.01		0.01	

Label	SH-11-59		SH-11-60		SH-11-62		SH-11-64		SH-11-63		SH-11-65	
Height	26.3		26.3		26.3		26.3		26.3		26.3	
Type	U		U		U		U		U		E	
Position	Core		Rim		Intermed.		Core		Intermed.		Intermed.	
SiO ₂	53.96		53.95		54.10		49.46		55.13		50.17	
Al ₂ O ₃	29.28		29.21		29.39		32.71		28.90		32.35	
FeO _T	0.93		0.65		0.40		0.55		0.70		0.56	
CaO	12.70		11.84		11.66		15.45		11.23		14.68	
Na ₂ O	4.24		4.49		4.60		1.99		4.92		2.58	
K ₂ O	0.24		0.17		0.22		0.15		0.27		0.18	
Total	101.35		100.31		100.37		100.31		101.15		100.52	
An	0.61		0.59		0.58		0.80		0.55		0.75	
Ab	0.37		0.40		0.41		0.19		0.44		0.24	
Or	0.01		0.01		0.01		0.01		0.02		0.01	

Label	SH-11-67		SH-11-68		SH-11-69		SH-11-70		SH-11-71		SH-11-72		SH-11-76		SH-11-78	
Height	26.3		26.3		26.3		26.3		26.3		26.3		26.3		26.3	
Type	E		E		E		E		E		E		U		U	
Position	Intermed.		Rim		Core		Intermed.		Intermed.		Rim		Rim		Intermed.	
SiO ₂	52.02		52.34		50.49		48.23		51.48		53.29		52.16		52.99	
Al ₂ O ₃	30.16		29.80		31.23		32.96		31.36		30.02		29.25		29.37	
FeO _T	0.65		0.67		0.71		0.62		0.76		0.99		0.84		0.64	
CaO	12.98		12.49		14.00		16.07		14.03		12.57		12.13		12.09	
Na ₂ O	3.91		3.82		3.41		1.97		3.36		3.91		4.26		4.68	
K ₂ O	0.21		0.15		0.13		0.06		0.15		0.17		0.23		0.28	
Total	99.93		99.27		99.97		99.91		101.14		100.95		98.87		100.05	
An	0.64		0.64		0.69		0.82		0.69		0.63		0.60		0.58	
Ab	0.35		0.35		0.30		0.18		0.30		0.36		0.38		0.41	
Or	0.01		0.01		0.01		0.00		0.01		0.01		0.01		0.02	

Table A3.3 (continued). Microprobe Analyses.

Sill C: Groundmass Plagioclase

Label	SH-11-79		SH-11-81	SH-11-85	SH-11-82	SH-11-83	SH-11-84	SH-16-2
Height	26.3		26.3	26.3	26.3	26.3	26.3	40.1
Type	U		U	U	U	U	U	U
Position	Intermed.		Core	Intermed.	Intermed.	Intermed.	Core	Intermed.
SiO ₂	53.74		47.67	53.75	52.50	54.00	51.80	52.53
Al ₂ O ₃	30.15		32.61	29.32	28.90	29.99	29.39	30.76
FeO _T	0.47		1.59	0.62	0.68	0.53	0.49	1.14
CaO	12.70		15.69	12.20	12.40	12.37	12.80	13.88
Na ₂ O	4.09		1.88	4.48	4.26	4.39	3.96	3.41
K ₂ O	0.19		0.09	0.16	0.19	0.13	0.20	0.16
Total	101.34		99.53	100.53	98.93	101.41	98.64	101.88
An	0.62		0.82	0.60	0.61	0.60	0.63	0.69
Ab	0.36		0.18	0.40	0.38	0.39	0.35	0.30
Or	0.01		0.01	0.01	0.01	0.01	0.01	0.01
Label	SH-16-3	SH-16-4	SH-16-6	SH-16-8	SH-16-9	SH-16-10	SH-16-11	
Height	40.1	40.1	40.1	40.1	40.1	40.1	40.1	
Type	U	U	U	E	E	U	U	
Position	Intermed.	Intermed.	Rim	Core	Core	Intermed.	Intermed.	
SiO ₂	52.15	51.96	51.81	51.72	49.03	47.53	52.94	
Al ₂ O ₃	29.53	30.13	30.20	30.56	32.14	32.50	29.19	
FeO _T	0.96	1.03	1.08	0.76	1.06	0.74	0.83	
CaO	13.03	13.01	13.23	13.39	15.70	15.99	12.35	
Na ₂ O	3.56	3.77	3.71	3.47	2.47	1.95	4.17	
K ₂ O	0.11	0.05	0.15	0.13	0.14	0.11	0.22	
Total	99.34	99.95	100.18	100.03	100.54	98.82	99.70	
An	0.66	0.65	0.66	0.68	0.77	0.81	0.61	
Ab	0.33	0.34	0.33	0.32	0.22	0.18	0.37	
Or	0.01	0.00	0.01	0.01	0.01	0.01	0.01	
Label	SH-16-22	SH-16-23	SH-22-12	SH-22-13	SH-22-14	SH-22-16	SH-22-18	
Height	40.1	40.1	61.1	61.1	61.1	61.1	61.1	
Type	U	U	U	U	U	U	E	
Position	Intermed.	Intermed.	Rim	Rim	Intermed.	Core	Core	
SiO ₂	52.13	51.25	53.30	57.35	53.88	47.57	51.23	
Al ₂ O ₃	29.23	31.50	29.12	26.53	28.79	31.92	31.73	
FeO _T	0.89	0.49	0.77	0.59	0.72	0.88	0.88	
CaO	12.48	14.26	11.26	8.95	11.39	15.59	14.63	
Na ₂ O	4.19	3.17	4.67	6.37	4.80	2.11	2.93	
K ₂ O	0.20	0.14	0.11	0.19	0.20	0.12	0.09	
Total	99.12	100.81	99.23	99.98	99.78	98.19	101.49	
An	0.61	0.71	0.57	0.43	0.56	0.80	0.73	
Ab	0.37	0.28	0.43	0.56	0.43	0.20	0.26	
Or	0.01	0.01	0.01	0.01	0.01	0.01	0.01	

Table A3.3 (continued). Microprobe Analyses.

Sill C: Groundmass Plagioclase

Label	SH-22-19	SH-22-20	SH-22-22	SH-22-23	SH-22-24	SH-22-25	F90-80-1
Height	61.1	61.1	61.1	61.1	61.1	61.1	104.5
Type	U	U	U	U	U	U	U
Position	Intermed.	Intermed.	Rim	Intermed.	Intermed.	Intermed.	Rim
SiO ₂	47.68	48.62	53.60	48.06	48.22	49.89	48.56
Al ₂ O ₃	32.69	32.79	28.87	32.50	32.46	32.13	32.77
FeO _T	0.80	0.88	0.66	0.73	0.86	0.93	0.75
CaO	15.91	15.75	11.64	15.80	15.60	14.98	15.87
Na ₂ O	2.09	2.19	4.32	1.85	1.78	2.48	1.80
K ₂ O	0.10	0.17	0.30	0.07	0.09	0.10	0.09
Total	99.27	100.40	99.39	99.01	99.01	100.51	99.84
An	0.80	0.79	0.59	0.82	0.82	0.76	0.83
Ab	0.19	0.20	0.39	0.17	0.17	0.23	0.17
Or	0.01	0.01	0.02	0.00	0.01	0.01	0.01

Label	F90-80-2	F90-80-3	F90-80-4	F90-80-5	F90-80-6	F90-80-7	F90-80-8	F90-80-10
Height	104.5	104.5	104.5	104.5	104.5	104.5	104.5	104.5
Type	U	U	U	U	U	U	U	U
Position	Intermed.	Core	Intermed.	Core	Rim	Core	Intermed.	Intermed.
SiO ₂	48.19	48.27	48.24	48.93	52.84	47.95	50.48	49.83
Al ₂ O ₃	32.97	33.13	32.97	32.83	27.88	32.89	31.89	32.17
FeO _T	0.86	0.77	0.81	0.70	2.91	0.83	0.93	0.84
CaO	16.00	16.20	16.15	15.94	12.79	15.75	14.76	15.42
Na ₂ O	1.55	1.62	1.82	1.75	3.39	1.73	2.59	2.31
K ₂ O	0.10	0.08	0.07	0.07	0.17	0.27	0.12	0.10
Total	99.67	100.07	100.06	100.22	99.98	99.42	100.77	100.67
An	0.85	0.84	0.83	0.83	0.67	0.82	0.75	0.78
Ab	0.15	0.15	0.17	0.17	0.32	0.16	0.24	0.21
Or	0.01	0.00	0.00	0.00	0.01	0.02	0.01	0.01

Label	F90-80-11	F90-80-21	F90-80-22	SH-38-1	SH-38-2	SH-38-3	SH-38-4	SH-38-7
Height	104.5	104.5	104.5	118.2	118.2	118.2	118.2	118.2
Type	U	U	U	U	U	U	U	U
Position	Rim	Intermed.	Rim	Intermed.	Intermed.	Intermed.	Intermed.	Rim
SiO ₂	49.53	49.93	54.28	52.76	53.46	52.67	51.75	55.90
Al ₂ O ₃	31.64	32.16	28.83	29.70	29.99	30.11	30.80	28.70
FeO _T	0.97	0.83	0.64	0.70	0.60	0.87	0.95	0.50
CaO	14.87	15.37	11.56	13.14	12.24	12.85	13.45	10.78
Na ₂ O	2.43	2.21	4.16	3.49	3.96	3.61	3.26	4.81
K ₂ O	0.07	0.05	0.26	0.10	0.30	0.18	0.21	0.30
Total	99.51	100.55	99.73	99.89	100.55	100.29	100.42	100.99
An	0.77	0.79	0.60	0.67	0.62	0.66	0.69	0.54
Ab	0.23	0.21	0.39	0.32	0.36	0.33	0.30	0.44
Or	0.00	0.00	0.02	0.01	0.02	0.01	0.01	0.02

Table A3.3 (continued). Microprobe Analyses.

Sill C: Groundmass Plagioclase

Label	SH-38-8	SH-38-9	SH-38-10	SH-38-11	SH-24-41	SH-24-50	SH-24-51	SH-24-52
Height	118.2	118.2	118.2	118.2	131.4	131.4	131.4	131.4
Type	U	U	U	U	U	U	U	E
Position	Core	Rim	Intermed.	Rim	Core	Rim	Rim	Rim
SiO ₂	48.60	51.82	52.86	52.55	51.09	50.21	52.06	48.89
Al ₂ O ₃	32.05	30.13	29.96	30.08	31.22	32.27	30.68	32.19
FeO _T	0.83	0.85	0.96	0.86	0.86	0.75	0.76	1.80
CaO	15.66	13.00	13.03	12.62	13.99	15.70	13.62	15.37
Na ₂ O	1.99	3.97	3.70	3.76	2.99	1.95	2.73	2.25
K ₂ O	0.13	0.23	0.18	0.16	0.08	0.08	0.28	0.11
Total	99.26	100.00	100.69	100.03	100.23	100.96	100.13	100.61
An	0.81	0.64	0.65	0.64	0.72	0.81	0.72	0.79
Ab	0.19	0.35	0.34	0.35	0.28	0.18	0.26	0.21
Or	0.01	0.01	0.01	0.01	0.00	0.00	0.02	0.01
Label	SH-24-54	SH-24-55	SH-24-56	SH-24-57	SH-24-58	SH-24-59	SH-28-5	
Height	131.4	131.4	131.4	131.4	131.4	131.4	140.6	
Type	E	E	E	U	U	U	U	
Position	Intermed.	Core	Rim	Rim	Core	Rim	Rim	
SiO ₂	48.38	48.17	48.44	48.00	48.52	51.64	53.14	
Al ₂ O ₃	32.96	32.13	32.75	32.48	32.26	30.65	29.86	
FeO _T	0.86	1.08	1.08	0.88	0.97	0.85	0.85	
CaO	16.11	16.12	16.13	16.02	15.91	13.65	13.08	
Na ₂ O	2.07	1.97	1.93	1.82	2.05	3.17	3.46	
K ₂ O	0.13	0.13	0.19	0.14	0.11	0.14	0.20	
Total	100.51	99.60	100.52	99.34	99.82	100.10	100.59	
An	0.81	0.81	0.81	0.82	0.81	0.70	0.67	
Ab	0.19	0.18	0.18	0.17	0.19	0.29	0.32	
Or	0.01	0.01	0.01	0.01	0.01	0.01	0.01	
Label	SH-28-6	SH-28-8	SH-28-9	SH-28-10	SH-28-11	SH-28-12	SH-28-13	
Height	140.6	140.6	140.6	140.6	140.6	140.6	140.6	
Type	U	U	U	U	U	U	U	
Position	Core	Core	Intermed.	Rim	Rim	Core	Core	
SiO ₂	53.29	49.31	52.87	53.46	52.67	52.23	52.78	
Al ₂ O ₃	29.52	31.37	29.71	29.40	30.00	29.48	30.25	
FeO _T	1.06	1.32	0.68	0.60	0.91	0.76	0.75	
CaO	11.89	15.74	12.93	12.47	13.04	13.07	12.64	
Na ₂ O	3.60	2.28	3.82	3.72	3.48	3.88	3.99	
K ₂ O	0.69	0.22	0.23	0.21	0.18	0.13	0.19	
Total	100.05	100.24	100.24	99.86	100.28	99.55	100.60	
An	0.62	0.78	0.64	0.64	0.67	0.65	0.63	
Ab	0.34	0.20	0.34	0.35	0.32	0.35	0.36	
Or	0.04	0.01	0.01	0.01	0.01	0.01	0.01	

Table A3.3 (continued). Microprobe Analyses.

Sill C: Groundmass Plagioclase

Label	SH-28-14	SH-28-15	SH-28-16	SH-28-17	SH-28-18	SH-28-19	SH-28-20	SH-28-21
Height	140.6	140.6	140.6	140.6	140.6	140.6	140.6	140.6
Type	U	U	U	U	U	U	U	U
Position	Rim	Intermed.	Rim	Core	Rim	Rim	Core	Core
SiO ₂	52.40	51.98	52.48	51.66	52.36	52.52	52.49	53.17
Al ₂ O ₃	30.02	29.70	30.26	30.40	29.60	29.74	29.66	30.21
FeO _T	0.72	0.92	1.00	0.87	0.96	0.82	0.78	0.93
CaO	13.24	12.85	12.95	13.24	12.69	12.51	12.79	11.46
Na ₂ O	3.72	3.60	3.96	3.19	3.74	3.71	3.88	3.80
K ₂ O	0.14	0.14	0.21	0.26	0.21	0.24	0.19	0.40
Total	100.24	99.19	100.86	99.62	99.56	99.54	99.79	99.97
An	0.66	0.66	0.64	0.69	0.64	0.64	0.64	0.61
Ab	0.33	0.33	0.35	0.30	0.34	0.34	0.35	0.37
Or	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.03

Label	SH-28-22	SH-28-23	SH-28-24	SH-28-25	SH-28-26	SH-28-27	SH-28-28	SH-28-29	SH-28-30
Height	140.6	140.6	140.6	140.6	140.6	140.6	140.6	140.6	140.6
Type	U	U	U	U	U	U	U	U	U
Position	Intermed.	Core	Intermed.	Intermed.	Rim	Core	Rim	Intermed.	Core
SiO ₂	52.44	49.77	52.45	51.74	52.27	49.24	55.96	53.67	52.78
Al ₂ O ₃	29.47	31.60	29.65	30.00	29.46	32.30	27.95	29.44	29.42
FeO _T	0.76	0.85	0.64	0.70	0.67	0.74	0.61	0.69	0.75
CaO	12.54	14.84	12.79	12.57	12.67	13.66	10.49	12.12	12.59
Na ₂ O	3.79	2.66	4.10	3.70	3.87	2.26	5.29	4.05	3.82
K ₂ O	0.20	0.10	0.14	0.47	0.20	1.02	0.43	0.23	0.13
Total	99.20	99.82	99.77	99.18	99.14	99.22	100.75	100.20	99.49
An	0.64	0.75	0.63	0.63	0.64	0.72	0.51	0.61	0.64
Ab	0.35	0.24	0.36	0.34	0.35	0.22	0.47	0.37	0.35
Or	0.01	0.01	0.01	0.03	0.01	0.06	0.02	0.01	0.01

Label	SH-28-31	SH-28-32	SH-28-33	SH-28-34	SH-28-35	SH-28-36	SH-28-37	SH-35-21
Height	140.6	140.6	140.6	140.6	140.6	140.6	140.6	156.5
Type	U	U	U	U	U	U	U	E
Position	Core	Rim	Core	Rim	Intermed.	Intermed.	Rim	Core
SiO ₂	52.82	59.76	53.44	56.23	53.08	55.12	55.78	49.69
Al ₂ O ₃	29.54	25.15	29.32	27.86	29.88	28.49	27.63	32.79
FeO _T	0.68	0.38	0.58	1.01	0.77	0.56	1.08	0.73
CaO	12.65	7.32	12.27	9.96	12.84	11.39	10.05	15.38
Na ₂ O	4.07	6.38	4.31	5.19	3.94	4.55	5.35	2.11
K ₂ O	0.14	0.88	0.27	0.28	0.24	0.32	0.33	0.14
Total	99.90	99.87	100.19	100.53	100.75	100.43	100.22	100.84
An	0.63	0.37	0.60	0.51	0.63	0.57	0.50	0.79
Ab	0.36	0.58	0.38	0.48	0.35	0.41	0.48	0.20
Or	0.01	0.05	0.02	0.02	0.01	0.02	0.02	0.01

Table A3.3 (continued). Microprobe Analyses.

Sill C: Groundmass Plagioclase

Label	SH-41-9		SH-41-11		SH-41-12		SH-41-13		SH-41-14		SH-41-15		F876-3-1			
Height	170.1		170.1		170.1		170.1		170.1		170.1		183.2			
Type	E		E		U		U		U		U		U			
Position	Core		Core		Intermed.		Intermed.		Rim		Core		Rim			
SiO ₂	47.84		48.51		52.35		48.42		53.03		47.74		53.53			
Al ₂ O ₃	32.02		32.65		29.72		32.13		29.83		32.59		29.30			
FeO _T	1.07		1.05		0.85		0.90		0.83		0.73		0.43			
CaO	15.86		16.09		12.60		15.42		12.45		15.83		11.40			
Na ₂ O	2.35		2.10		3.79		2.47		4.15		2.05		4.46			
K ₂ O	0.09		0.06		0.14		0.05		0.19		0.17		0.22			
Total	99.23		100.46		99.45		99.39		100.48		99.11		99.34			
An	0.78		0.81		0.64		0.77		0.62		0.80		0.58			
Ab	0.21		0.19		0.35		0.22		0.37		0.19		0.41			
Or	0.01		0.00		0.01		0.00		0.01		0.01		0.01			
Label	F876-3-2		F876-3-3		F876-3-4		F876-3-5		F876-3-6		F876-3-7		F876-9-4		F876-9-5	
Height	183.2		183.2		183.2		183.2		183.2		183.2		199.0		199.0	
Type	U		U		U		U		U		U		U		U	
Position	Core		Intermed.		Rim		Intermed.		Core		Intermed.		Core		Rim	
SiO ₂	53.59		52.43		53.56		48.31		48.70		49.57		52.39		54.20	
Al ₂ O ₃	29.73		30.43		29.62		32.64		31.66		31.57		29.87		29.43	
FeO _T	0.63		0.71		0.76		0.77		0.68		0.77		0.85		0.70	
CaO	12.22		13.35		12.11		16.03		15.72		14.68		13.03		11.91	
Na ₂ O	4.35		3.94		4.37		1.97		2.43		2.91		4.33		4.01	
K ₂ O	0.19		0.16		0.26		0.03		0.17		0.12		0.24		0.17	
Total	100.71		101.02		100.68		99.75		99.36		99.62		100.71		100.42	
An	0.60		0.65		0.60		0.82		0.77		0.73		0.62		0.51	
Ab	0.39		0.34		0.39		0.18		0.22		0.26		0.37		0.37	
Or	0.01		0.01		0.02		0.00		0.01		0.01		0.01		0.01	
Label	F876-9-6		F876-9-7		F876-9-11		F876-9-12		F876-9-13		F876-23-17					
Height	199.0		199.0		199.0		199.0		199.0		221.7					
Type	U		U		U		U		U		U					
Position	Intermed.		Intermed.		Core		Intermed.		Intermed.		Rim					
SiO ₂	52.78		52.57		48.20		52.09		48.77		63.51					
Al ₂ O ₃	29.93		29.48		32.70		29.69		32.03		22.93					
FeO _T	0.79		0.83		0.70		0.65		0.60		0.35					
CaO	12.94		12.78		16.03		12.76		15.48		3.78					
Na ₂ O	3.80		4.00		1.98		4.54		2.57		10.21					
K ₂ O	0.16		0.14		0.10		0.10		0.11		0.22					
Total	100.40		99.80		99.71		99.83		99.56		101.00					
An	0.65		0.63		0.81		0.60		0.76		0.17					
Ab	0.34		0.36		0.18		0.39		0.23		0.82					
Or	0.01		0.01		0.01		0.01		0.01		0.01					

Table A3.3 (continued). Microprobe Analyses.

Sill C: Groundmass Plagioclase					Sill B: Groundmass Plagioclase		
Label	F876-23-18	F876-23-20	F876-23-22	F876-23-25	F16-6-1	F16-6-3	F16-6-4
Height	221.7	221.7	221.7	221.7	321.3	321.3	321.3
Type	U	U	U	U	U	U	U
Position	Intermed.	Intermed.	Core	Intermed.	Core	Intermed.	Intermed.
SiO ₂	64.78	61.51	61.91	64.48	48.66	53.04	50.41
Al ₂ O ₃	22.17	24.73	24.24	22.83	32.91	27.99	30.80
FeO _T	0.52	0.52	0.26	0.15	0.74	2.92	0.85
CaO	2.77	5.96	5.61	3.69	16.00	12.84	13.80
Na ₂ O	10.41	8.78	9.09	10.50	1.87	3.40	2.95
K ₂ O	0.38	0.28	0.34	0.32	0.07	0.17	0.08
Total	101.03	101.78	101.45	101.97	100.25	100.36	98.89
An	0.13	0.27	0.25	0.16	0.82	0.67	0.72
Ab	0.85	0.72	0.73	0.82	0.17	0.32	0.28
Or	0.02	0.02	0.02	0.02	0.00	0.01	0.00

Label	F16-6-5	F16-6-6	F16-6-7	F16-6-7	F16-6-8	F16-9-18	F16-9-19
Height	321.3	321.3	321.3	321.3	321.3	297.7	297.7
Type	U	U	U	U	U	U	U
Position	Rim	Intermed.	Intermed.	Intermed.	Core	Intermed.	Core
SiO ₂	54.68	54.47	54.95	52.29	48.35	57.36	51.66
Al ₂ O ₃	29.04	28.50	28.05	30.82	31.83	26.88	30.12
FeO _T	0.64	0.78	0.78	0.76	1.78	0.68	0.95
CaO	11.65	10.75	10.81	13.68	15.20	8.85	13.14
Na ₂ O	4.19	5.59	5.34	2.74	2.22	6.32	4.04
K ₂ O	0.26	0.23	0.24	0.28	0.11	0.25	0.14
Total	100.47	100.32	100.17	100.58	99.49	100.34	100.06
An	0.60	0.51	0.52	0.72	0.79	0.43	0.64
Ab	0.39	0.48	0.47	0.26	0.21	0.56	0.35
Or	0.02	0.01	0.01	0.02	0.01	0.01	0.01

Label	F16-9-20	F16-9-21	F16-9-23	F16-9-24	F16-9-25	F16-14-9	F16-14-10
Height	297.7	297.7	297.7	297.7	297.7	242.9	242.9
Type	U	U	U	E	E	U	U
Position	Rim	Rim	Rim	Core	Core	Core	Core
SiO ₂	56.83	55.57	57.64	48.88	49.11	48.46	48.58
Al ₂ O ₃	26.55	27.19	26.60	32.91	31.63	32.51	33.32
FeO _T	0.40	0.52	0.61	0.87	0.92	0.69	0.84
CaO	8.85	9.55	8.07	15.81	14.75	15.79	15.96
Na ₂ O	6.37	6.09	6.56	1.80	2.44	1.73	1.75
K ₂ O	0.21	0.21	0.20	0.09	0.10	0.07	0.27
Total	99.21	99.13	99.67	100.37	98.94	99.25	100.72
An	0.43	0.46	0.40	0.82	0.76	0.83	0.82
Ab	0.56	0.53	0.59	0.17	0.23	0.17	0.16
Or	0.01	0.01	0.01	0.01	0.01	0.00	0.02

Table A3.3 (continued). Microprobe Analyses.

Sill B: Groundmass Plagioclase

Label	F16-14-11	F16-14-12	F16-14-15	F16-14-16	F16-14-18	F16-14-19	F16-14-20	F14-16-7
Height	242.9	242.9	242.9	242.9	242.9	242.9	242.9	285.2
Type	U	U	U	U	U	U	U	E
Position	Intermed.	Rim	Core	Core	Rim	Rim	Rim	Core
SiO ₂	49.81	52.17	52.17	52.60	59.11	58.56	58.49	47.88
Al ₂ O ₃	31.47	30.25	30.47	30.54	25.35	25.95	25.40	32.72
FeO _T	0.92	0.93	0.45	0.43	1.63	0.47	0.47	0.80
CaO	14.56	13.39	13.12	13.20	7.46	8.05	7.55	16.03
Na ₂ O	2.56	3.90	3.16	3.41	7.05	6.84	7.18	1.81
K ₂ O	0.12	0.16	0.15	0.08	0.17	0.35	0.39	0.07
Total	99.43	100.81	99.52	100.26	100.78	100.21	99.49	99.31
An	0.75	0.65	0.69	0.68	0.37	0.39	0.36	0.83
Ab	0.24	0.34	0.30	0.32	0.62	0.59	0.62	0.17
Or	0.01	0.01	0.01	0.00	0.01	0.02	0.02	0.00

Label	F14-16-9	F14-16-12	F14-16-13	F14-16-20	F14-16-22	F14-16-23	F14-16-24	F14-16-25
Height	285.2	285.2	285.2	285.2	285.2	285.2	285.2	285.2
Type	E	U	U	U	U	U	U	U
Position	Core	Core	Core	Core	Core	Rim	Rim	Rim
SiO ₂	49.29	47.77	49.91	52.41	52.48	56.21	59.57	59.87
Al ₂ O ₃	31.82	32.79	31.88	29.98	30.13	27.83	25.10	24.85
FeO _T	0.83	0.76	0.98	0.75	1.01	0.65	0.28	0.33
CaO	15.25	16.03	14.99	12.42	13.44	9.95	7.31	6.36
Na ₂ O	2.28	1.60	2.45	3.86	2.83	5.68	7.23	7.74
K ₂ O	0.10	0.08	0.07	0.20	0.09	0.20	0.34	0.23
Total	99.58	99.03	100.28	99.62	99.97	100.52	99.83	99.37
An	0.78	0.84	0.77	0.63	0.72	0.49	0.35	0.31
Ab	0.21	0.15	0.23	0.36	0.27	0.50	0.63	0.68
Or	0.01	0.00	0.00	0.01	0.01	0.01	0.02	0.01

Label	F14-16-25	F14-16-26	F14-16-27	F14-16-29	F14-16-30	F14-28-2	F14-28-3
Height	285.2	285.2	285.2	285.2	285.2	243.7	243.7
Type	U	U	U	U	U	U	U
Position	Rim	Rim	Rim	Core	Intermed.	Core	Core
SiO ₂	59.16	58.54	58.14	58.09	57.49	51.58	51.90
Al ₂ O ₃	25.21	26.44	26.25	26.05	26.21	31.23	31.25
FeO _T	0.49	0.47	0.54	0.50	0.74	0.73	0.58
CaO	7.06	8.24	8.07	8.02	8.32	13.49	12.91
Na ₂ O	7.44	6.69	7.11	6.22	6.34	3.37	2.99
K ₂ O	0.16	0.27	0.11	0.27	0.31	0.26	0.50
Total	99.52	100.65	100.22	99.16	99.41	100.66	100.15
An	0.34	0.40	0.38	0.41	0.41	0.68	0.68
Ab	0.65	0.59	0.61	0.57	0.57	0.31	0.29
Or	0.01	0.02	0.01	0.02	0.02	0.02	0.03

Table A3.3 (continued). Microprobe Analyses.

Sill B: Glomerocryst Plagioclase

Label	SH-16-25	SH-16-27	SH-16-28	SH-16-29	SH-16-30	SH-16-31	SH-24-25	SH-24-26	SH-24-27
Height	40.1	40.1	40.1	40.1	40.1	40.1	131.4	131.4	131.4
Position	Central	Central	Central	Central	Central	Intergr.	Central	Central	Central
SiO ₂	48.37	48.25	49.12	52.49	48.50	57.52	49.57	52.50	47.40
Al ₂ O ₃	33.20	32.98	32.95	28.69	32.90	26.87	31.05	30.40	32.67
FeO _T	0.77	0.81	0.70	0.89	0.83	0.76	0.93	0.87	1.00
CaO	16.23	16.15	16.00	12.70	16.21	8.95	14.74	13.26	16.08
Na ₂ O	1.35	1.82	2.06	3.37	2.22	5.24	3.08	3.82	2.16
K ₂ O	0.08	0.07	0.07	0.17	0.11	0.21	0.15	0.12	0.17
Total	100.01	100.09	100.90	98.31	100.77	99.56	99.52	100.97	99.48
An	0.86	0.83	0.81	0.67	0.80	0.48	0.72	0.65	0.80
Ab	0.13	0.17	0.19	0.32	0.20	0.51	0.27	0.34	0.19
Or	0.01	0.00	0.00	0.01	0.01	0.01	0.01	0.01	0.01

Label	SH-24-28	SH-24-29	SH-24-30	SH-24-31	SH-24-32	SH-24-33	SH-24-34	SH-24-35	SH-24-36
Height	131.4	131.4	131.4	131.4	131.4	131.4	131.4	131.4	131.4
Position	Central	Central	Central	Rim	Central	Central	Central	Central	Central
SiO ₂	47.62	48.34	47.90	62.21	49.34	48.83	48.14	48.32	49.70
Al ₂ O ₃	32.56	32.15	32.13	23.76	31.61	32.37	32.55	32.53	32.01
FeO _T	0.79	0.80	0.78	0.50	0.76	0.84	0.73	0.86	0.93
CaO	16.23	15.58	15.73	5.01	15.11	15.31	15.83	15.63	14.92
Na ₂ O	1.52	2.26	2.09	8.86	2.42	2.40	1.85	1.78	2.47
K ₂ O	0.10	0.10	0.11	0.39	0.11	0.14	0.07	0.09	0.10
Total	98.82	99.23	98.74	100.72	99.35	99.89	99.17	99.22	100.13
An	0.85	0.79	0.80	0.23	0.77	0.77	0.82	0.82	0.76
Ab	0.14	0.21	0.19	0.75	0.22	0.22	0.17	0.17	0.23
Or	0.01	0.01	0.01	0.02	0.01	0.01	0.00	0.01	0.01

Label	SH-24-37	SH-24-38	SH-24-39	F90-79-1	F90-79-2	F90-79-3	F15-4-1	F15-4-2
Height	131.4	131.4	131.4	69.1	69.1	69.1	220.0	220.0
Position	Central	Central	Rim	Central	Central	Rim	Central	Central
SiO ₂	48.40	48.01	61.55	51.61	51.52	64.77	48.92	48.72
Al ₂ O ₃	32.66	32.84	22.90	31.18	30.51	21.33	33.13	32.67
FeO _T	0.75	0.86	0.44	0.84	0.91	0.52	0.82	1.08
CaO	15.82	15.94	4.95	14.70	14.76	3.59	15.45	15.81
Na ₂ O	1.79	1.54	8.58	2.37	1.91	9.61	1.90	1.27
K ₂ O	0.09	0.10	0.44	0.20	0.14	0.30	0.30	0.12
Total	99.52	99.29	98.85	100.91	99.75	100.13	100.53	99.66
An	0.83	0.85	0.24	0.76	0.80	0.17	0.80	0.87
Ab	0.17	0.15	0.74	0.22	0.19	0.81	0.18	0.13
Or	0.01	0.01	0.02	0.01	0.01	0.02	0.02	0.01

Table A3.3 (continued). Microprobe Analyses.

Sill B: Glomerocryst Plagioclase

Label	F15-4-3	F15-4-4	F15-4-5	F15-4-6	F15-4-7	F16-14-21	F16-14-21	F16-14-22	F16-14-26
Height	220.0	220.0	220.0	220.0	220.0	242.9	242.9	242.9	242.9
Position	Central	Central	Central	Central	Central	Central	Central	Central	Central
SiO ₂	47.91	48.09	51.32	51.53	48.37	49.36	52.00	49.43	49.11
Al ₂ O ₃	33.19	32.67	30.43	30.36	32.19	32.59	29.79	31.71	32.41
FeO _T	1.01	0.84	0.80	0.54	0.83	0.76	0.91	0.82	0.91
CaO	15.96	15.91	12.89	12.90	15.28	15.45	12.68	15.46	15.71
Na ₂ O	2.10	2.05	3.72	3.81	2.53	2.27	4.01	2.19	1.91
K ₂ O	0.16	0.11	0.17	0.18	0.13	0.11	0.12	0.12	0.07
Total	100.34	99.68	99.32	99.31	99.33	100.54	99.51	99.73	100.13
An	0.80	0.81	0.65	0.64	0.76	0.78	0.63	0.79	0.82
Ab	0.19	0.19	0.34	0.34	0.23	0.21	0.36	0.20	0.18
Or	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.00

Label	F16-14-27	F16-14-29	F16-14-30	F16-14-31	F16-6-21	F16-6-25	F16-6-26	F16-6-27	F16-6-28
Height	242.9	242.9	242.9	242.9	321.3	321.3	321.3	321.3	321.3
Position	Central	Central	Intergr.	Intergr.	Central	Central	Central	Central	Rim
SiO ₂	48.37	49.40	53.07	56.69	48.61	51.89	51.68	51.38	58.22
Al ₂ O ₃	32.37	32.56	30.13	27.74	33.10	29.72	30.64	30.28	26.86
FeO _T	1.08	0.77	0.86	0.53	0.87	0.91	0.65	0.70	0.62
CaO	16.04	15.30	12.77	9.74	16.22	12.65	12.97	12.86	8.15
Na ₂ O	2.02	2.73	3.84	6.21	1.96	4.00	4.20	3.80	6.63
K ₂ O	0.09	0.00	0.14	0.21	0.09	0.22	0.17	0.18	0.20
Total	99.97	100.76	100.82	101.13	100.85	99.39	100.31	99.20	100.67
An	0.81	0.76	0.64	0.46	0.82	0.63	0.62	0.64	0.40
Ab	0.18	0.24	0.35	0.53	0.18	0.36	0.37	0.34	0.59
Or	0.01	0.00	0.01	0.01	0.01	0.01	0.01	0.01	0.01

Label	F16-9-27	F16-9-28	F16-9-29	F16-9-30	F16-9-31
Height	297.7	297.7	297.7	297.7	297.7
Position	Central	Central	Central	Central	Intergr.
SiO ₂	47.94	48.25	48.56	47.74	54.37
Al ₂ O ₃	33.21	32.78	32.32	32.50	27.75
FeO _T	1.01	0.85	0.83	0.85	0.77
CaO	15.97	15.97	15.34	15.93	10.70
Na ₂ O	2.11	2.06	2.36	2.40	5.28
K ₂ O	0.16	0.11	0.13	0.09	0.24
Total	100.40	100.01	99.54	99.51	99.11
An	0.80	0.81	0.78	0.78	0.52
Ab	0.19	0.19	0.22	0.21	0.47
Or	0.01	0.01	0.01	0.01	0.01

Appendix A4

Major And Trace Element Analyses

A4.1 Analytical Procedures

The major and trace element compositions of 282 whole rock samples and 13 plagioclase mineral separates were determined by X-ray fluorescence (XRF) analysis in the Department of Geology, University of Ottawa. The REE contents of 79 of the whole rock samples, and all of the mineral separates, were determined by inductively coupled argon plasma-atomic emission spectrometry (ICAP-AES), also in the Department of Geology, University of Ottawa. The REE were also determined in 14 duplicate samples by instrumental neutron activation analysis (INAA) at the Institut de génie énergétique, École Polytechnique, l'Université de Montréal. The results of INAA were equivalent within error to those obtained by ICP-AES, and only the latter data are used in the preceding discussions. The ferrous iron content of 18 of the whole rock samples was determined by titration. Weight loss on ignition (LOI) was determined for all samples which were not heated to expel volatile elements prior to fusion. A few samples with high oxide or sulphide contents were preheated to facilitate complete fusion, and LOI values are reported as zero for these. Analytical data are provided in Table A4.3.

Preparation of Sample Powders

Whole rock samples were prepared by removing weathered surfaces and crushing to approximately 1 cm sized fragments in a steel jaw crusher. Sparsely glomeroporphyritic samples were serially sectioned prior to crushing to allow selection of groundmass material only. Samples were repeatedly split until approximately 150-200 g of material was obtained, and this was crushed in its entirety in a tungsten-carbide shatterbox. The glomerocrysts in anorthositic phase samples could not be separated from groundmass plagioclase, and therefore these analyses are representative of the groundmass+glomerocrysts. Because of the extreme size of glomerocrysts in the core phase, no attempt was made to obtain geochemical data for these samples.

Glomerocryst plagioclase was obtained by serial sectioning of samples into plates between 3 mm and 25 mm thick, the thickness depending on the average dimensions of glomerocrysts. The glomerocrysts were cut from matrix material, crushed in a ceramic plate crusher, and passed through a Frantz magnetic separator to remove non-plagioclase constituents. Groundmass plagioclase was obtained by crushing a portion of the splits from whole rock samples in a ceramic plate crusher, and magnetically separating plagioclase from other phases. Both glomerocryst and groundmass plagioclase separates were subsequently hand picked under a low power binocular microscope to remove remaining impurities. Crystals from the two sample groups were then crushed to powder in an agate ball mill.

XRF Procedure

Powdered samples were prepared as fused discs using 1.300 g sample, 3.900 g $\text{Li}_2\text{B}_4\text{O}_7$, and 0.433 g of Li_2CO_3 . Analyses were performed on a Phillips PW1410/20 AHP instrument using a rhodium excitation tube. The rhodium tube was energized at 60 kv and 40 mA. Analytical uncertainty was estimated by replicate measurements on samples of international standards (Table A4.1A). Estimates of analytical uncertainties for gabbroic samples are based on 61 replicate analyses of 4 samples of gabbro standard MRG-1 and 58 analyses of 3 samples of diorite standard DR-N. Generally, the higher calculated coefficient of variation (i.e., MRG-1 vs. DR-N) was used in subsequent error calculations. Where the higher value reflects extremely small concentrations, however, the alternative coefficient of variation was used. Analytical uncertainties for felsic samples (>55 wt. % SiO_2) are based on 8 replicate analyses of two samples of syenite standard SY-2. Analytical uncertainties for plagioclase separates were estimated by 5 replicate measurements on a single sample of feldspar standard 70A.

FeO Titration Procedure

The FeO content of selected samples was determined by titration following the method of Wilson (1955) with modifications after Whipp (1974). Sample solutions were prepared by digestion of approximately 0.2 g of sample in 10 ml of HF (conc.) and 5 ml of 0.1N ammonium metavanadate solution. The samples were left to digest on a shaker

Table A4.1(A). Mean (M), standard deviation (SD) and coefficient of variation (CV) for replicate XRF analyses of MRG-1, DR-N, SY-2 and 70A. The CV values calculated from these standards are used in error analysis involving the sample types indicated. Coefficient of variations used for gabbro and basalt samples are printed in bold typeface.

Sample Element	Gabbro/Basalt/Alkali Basalt						Felsic/Vein			Plagioclase		
	MRG-1			DR-N			SY-2			70A		
	M	SD	CV	M	SD	CV	M	SD	CV	M	SD	CV
SiO ₂	39.28	0.22	0.006	53.12	0.19	0.004	59.82	0.23	0.004	66.41	0.37	0.006
Al ₂ O ₃	8.46	0.06	0.007	17.59	0.06	0.004	11.96	0.04	0.003	18.04	0.03	0.002
Fe ₂ O ₃ T	17.65	0.15	0.008	9.69	0.08	0.008	6.24	0.03	0.005	0.08	0.01	0.061
MgO	13.56	0.14	0.011	4.34	0.09	0.020	2.65	0.04	0.015	0.00	0.01	2.000
CaO	14.77	0.08	0.006	7.03	0.03	0.004	7.93	0.02	0.002	0.22	0.01	0.023
Na ₂ O	0.70	0.34	0.488	2.94	0.12	0.040	4.38	0.13	0.029	2.37	0.23	0.097
K ₂ O	0.18	0.00	0.028	1.70	0.02	0.009	4.44	0.01	0.003	11.63	0.05	0.004
TiO ₂	3.76	0.13	0.034	1.05	0.02	0.022	0.13	0.00	0.000	0.00	0.01	2.000
P ₂ O ₅	0.02	0.02	1.223	0.22	0.02	0.111	0.42	0.02	0.050	0.02	0.01	0.286
MnO	0.16	0.04	0.240	0.21	0.05	0.218	0.24	0.14	0.597	0.02	0.01	0.547
S	0.04	0.04	1.026	0.05	0.05	0.959	0.11	0.14	1.254	0.02	0.01	0.385
Ba	9	16.6	1.866	397	22.0	0.055	468	11.3	0.024	166	35.3	0.213
Cr	445	20.7	0.046	44	5.5	0.123	7	6.6	0.965	20	2.2	0.108
Zr	91	6.3	0.070	129	3.8	0.029	275	11.3	0.041			
Sr	264	3.9	0.015	399	4.9	0.012	268	2.0	0.007	66	0.8	0.012
Rb	2	2.7	1.080	67	3.0	0.044	221	3.6	0.016	529	1.6	0.003
Y	16	2.2	0.137	29	1.7	0.059	130	1.2	0.009	3	0.8	0.272
Nb	19	7.7	0.402	4	1.9	0.486	24	1.1	0.047			
Zn	183	19.1	0.104	144	14.8	0.103	256	8.4	0.033	17	4.3	0.260
Ni	185	22.5	0.121	20	10.0	0.493	22	19.4	0.872	16	4.9	0.306
V	512	46.7	0.091	223	14.2	0.064	56	8.2	0.146			

Table A4.1(B). Mean (M), standard deviation (SD) and coefficient of variation (CV) for replicate REE (ICAP-AES) analyses of MRG-1 and BAS-1. CV values calculated are used as above.

Element	Gabbro/Basalt /Plagioclase			Felsic/Vein/Alkali Basalt		
	BAS-1			MRG-1		
	M	SD	CV	M	SD	CV
La	2.56	0.16	0.062	9.23	0.64	0.070
Ce	5.97	0.26	0.044	25.84	1.14	0.044
Nd	4.24	0.22	0.051	17.27	0.96	0.056
Sm	1.34	0.05	0.040	4.83	0.23	0.048
Eu	0.59	0.03	0.050	1.44	0.05	0.032
Gd	1.99	0.12	0.058	4.36	0.25	0.057
Dy	2.41	0.10	0.042	2.90	0.15	0.050
Er	1.42	0.12	0.082	1.08	0.11	0.099
Yb	1.37	0.06	0.046	0.82	0.04	0.047
Lu	0.21	0.01	0.045	0.13	0.00	0.037

table for 24 hours, and then mixed with 10 ml of a phosphoric-sulphuric acid solution (1:2:2), 75 ml of boric acid (conc.), 10 ml 0.05N ferrous ammonium sulphate and approximately three drops of barium diphenylamine sulphonate indicator solution. The samples were then titrated to their end points with 0.125 N potassium dichromate.

The Fe_2O_3 content of samples was calculated from the difference in the FeO and total iron ($\text{Fe}_2\text{O}_3\text{T}$) contents as determined by titration and XRF, respectively. The average FeO/ Fe_2O_3 ratio of titrated samples within a petrological suite (i.e., the Willbob basalts) was used to estimate FeO content (reported as FeO^*) for all samples in the suite, and this value was used in all calculations. Analytical uncertainty in FeO content was estimated by replicate analyses of international standard SY-3 (Table A4.2). Analytical uncertainty in calculated FeO^* and Fe_2O_3 was estimated by comparing FeO^* and FeO for all samples for which titration data was obtained. The large uncertainties obtained from these comparisons probably reflects relatively large variations in the FeO/ Fe_2O_3 ratios within a given suite, which is thought to arise from post-crystallization oxidation.

Table A4.2. Average (M), standard deviation (SD) and coefficient of variation (CV) derived from multiple determinations of ferrous and ferric iron contents in standard and unknown samples. Uncertainty associated with FeO is calculated from replicate measurements of the FeO content of standard SY-3. Uncertainty associated with FeO^* (see text for definition) and Fe_2O_3 is calculated from comparison of FeO and FeO^* for numerous samples, following the procedure of Kretz (1985) for duplicate measurements.

Element	M	SD	CV
FeO	3.88	0.31	0.095
Fe_2O_3	2.62	0.28	0.108
FeO^*	8.60	0.82	0.010

LOI Procedure

The volatile content of whole rock samples was estimated by determination of weight loss on ignition (LOI). Samples were weighed before and after heating to approximately 1050°C in ceramic crucibles, with the percentage weight loss reported as LOI. Duplicate measurements were identical to the third decimal place of reported LOI values.

ICAP-AES Procedure

The REE were determined by ICAP-AES analysis of digested sample powders following the procedure of Walsh *et al.* (1981). Approximately 2 g of sample was digested for 24 hours using 10 ml of HF (conc.) in a teflon beaker. 3 ml of perchloric acid (conc.) were added to the samples which were then placed on a hotplate at approximately 140°C for 24 hours. When dried, 3 ml of HCl (conc.) and 10 ml of distilled water were added to the samples to dissolve the dried residue. Once dissolved, the samples were taken to a volume of 50 ml in graduated nalgene polytubes using distilled water. This provided a sample solution in approximately 1 N HCl.

REE extraction was accomplished in cation exchange columns using BIO-RAD AG-50W-X8 (100-200) mesh cation exchange resin. The resin height was approximately 20 cm in columns of 1 cm diameter. Resin height and elution volumes were calibrated using prepared REE solutions with major and trace elements added to duplicate rock sample compositions. The columns were equilibrated with 50 ml of 1N HCl prior to addition of samples. Major and trace elements other than the REE and Y were eluted using 100 ml of 2N HCl. This solution was discarded. The REE fraction were eluted using 300 ml of 6N HCl which was collected in 400 ml beakers. The columns were backwashed with approximately 200 ml of 6N HCl and allowed to resettle. The backwash solution was discarded. The resin was subsequently re-equilibrated with approximately 100 ml of 1N HCl. The REE fraction collected in 6N HCl was evaporated to dryness in covered 400 ml beakers on a hotplate at approximately 90°C. The residue was redissolved with 15 ml of 1N nitric acid in preparation for analysis.

The instrument used in the analysis of the REE was a Thermo Jarrell Ash Atomscan 25 Inductively Coupled Argon Plasma Atomic Emission Sequential Spectrometer. Operating conditions were:

RF power: 1350

nebulizer pressure: 37 psi

observation height: 17 mm

torch gas flow: 1 l/m

Sm and Gd concentrations were corrected for interferences from Gd and Ce, respectively, using $Sm = Sm - 0.11Gd$ and $Gd = Gd - 0.007 Ce$. Analytical precision and uncertainty were

monitored using blanks and international standard samples MRG-1 and SY-2. Analytical uncertainty was estimated by replicate analyses of in house standard BAS-1 (8 replicate measurements) and international standard MRG-1 (5 replicate measurements). Mean values, standard deviations and analytical uncertainty expressed as coefficient of variation for these standards are provided in Table A4.1B.

Explanation of Table A4.3

Analytical data are reported in Table A4.3. Values are reported as zero for undetected elements. Elements which were not analyzed in a given sample are left blank. Results are tabulated alphabetically according to sample number within individual petrological suites (note that numbers are treated as characters and therefore samples may not be in numeric order, i.e., sample F90-100 will occur before sample F90-99 in the table). Sill type (Glom.=highly glomeroporphyritic sill, Aphyric=sparsely glomeroporphyritic or aphyric sill) and stratigraphic height above base (Height) are indicated where applicable. Samples are identified according to broad lithological type using the following abbreviations: Chill=aphanitic chilled margin, Fine=fine grained chilled margin, Gabbroic=gabbroic phase, Anorth.=anorthositic phase, Oxide=oxide cumulate series, Felsic=felsic series, Ol.-Bearing=olivine bearing series, Ol.-Free=olivine free series, Vein=leucocratic veins, Sulphide=high sulphide series, Clast=clast from conglomerate or agglomerate, Basalt=basalt, Trachy.=trachybasalt, Gl.Basalt=glomeroporphyritic basalt, Cryst=glomerocryst plagioclase, Matrix=groundmass plagioclase.

Table A4.3. Whole rock major (wt. %) and trace (ppm) element analyses**Menthek Gabbros****Sill B: Profile 1**

Sample	F90-70A	F90-70B	F90-70C	F13-19B	F13-19A	F871-7	F871-6	F16-21	F871-5	F871-4
Sill Type	Glom.	Glom.	Glom.	Glom.	Glom.	Glom.	Glom.	Glom.	Glom.	Glom.
Lithology	Chill	Chill	Chill	Fine	Fine	Gabbroic	Gabbroic	Gabbroic	Anorth.	Anorth.
Height	0.0	0.3	0.5	1.0	7.5	23.0	31.7	42.4	58.0	77.9
SiO₂	49.46	48.40	48.56	48.36	49.32	49.51	49.85	49.19	48.10	46.94
Al₂O₃	13.45	15.02	13.69	14.13	13.99	14.33	14.61	15.17	21.40	21.36
Fe₂O₃T	15.30	14.92	16.13	15.10	15.72	15.26	14.33	13.57	8.47	8.06
Fe₂O₃					2.22	0.65	0.59	1.35	1.26	1.16
FeO					12.15	13.15	12.36	11.00	6.49	6.21
MgO	7.67	7.21	7.52	8.42	6.84	7.20	6.23	6.65	4.48	4.92
CaO	8.45	6.43	8.05	7.68	7.38	7.73	9.12	10.73	12.99	12.83
Na₂O	2.83	3.47	2.49	2.00	3.22	3.10	2.69	2.52	2.46	2.31
K₂O	0.38	0.34	0.30	0.61	0.45	0.37	0.39	0.36	0.28	0.31
TiO₂	1.52	1.76	1.54	1.60	1.54	1.45	1.54	1.23	0.65	0.67
P₂O₅	0.12	0.15	0.14	0.09	0.11	0.10	0.11	0.08	0.05	0.05
MnO	0.28	0.26	0.29	0.25	0.29	0.28	0.24	0.22	0.12	0.11
S	0.00	0.00	0.01	0.07	0.04	0.04	0.02	0.04	0.01	0.03
Ba	76	59	56	103	177	72	56	123	72	85
Cr	119	127	116	177	117	102	84	199	145	125
Zr	79	89	83	76	79	74	70	60	28	29
Sr	112	126	104	82	132	141	139	143	166	162
Rb	9	0	4	22	30	10	9	7	7	7
Y	45	35	41	37	34	33	32	27	14	13
Nb	10	11	7	0	6	0	0	0	0	0
Zn	57	68	44	49	215	151	237	59	78	35
Ni	117	88	96	125	92	81	79	76	92	87
V	387	438	446	407	454	445	485	412	259	206
LOI	1.82	3.45	2.13	2.72	1.83	1.24	2.23	1.60	2.16	2.10
Total	101.28	101.41	100.85	101.03	100.73	100.61	101.36	101.36	101.17	99.69
Fe₂O₃	1.57	1.53	1.66	1.55	1.61	1.57	2.03	1.93	1.11	1.06
FeO*	12.35	12.05	13.02	12.19	12.69	12.32	11.06	10.48	6.62	6.30
La	8.9	7.3	6.3		5.8	6.8				
Ce	24.7	19.3	17.9		15.0	15.6				
Nd	19.5	12.9	13.9		10.9	11.1				
Sm	6.4	4.0	4.7		3.5	3.7				
Eu	1.7	1.1	1.0		1.1	1.2				
Gd	8.6	4.6	5.2		4.6	4.4				
Dy	9.8	5.0	5.6		5.3	4.9				
Er	5.9	3.1	3.4		3.3	3.2				
Yb	4.9	3.0	3.2		3.0	3.2				
Lu	0.65	0.46	0.44		0.48	0.54				

Table A4.3 (continued). Whole rock major (wt. %) and trace (ppm) element analyses.

Menihok Gabbros (continued)

Sill B: Profile 1

Sample	60A	60C(1)	60C(2)
Sill Type	Glom.	Glom.	Glom.
Lithology	Felsic	Felsic	Felsic
Height	402.0	403.0	404.0
SiO ₂	62	77	66
Al ₂ O ₃	12	12	11
Fe ₂ O ₃ T	13	2.10	8
Fe ₂ O ₃			
FeO			
MgO	1.01	0.08	0.68
CaO	5.57	1.53	10.36
Na ₂ O	4.00	6.48	1.01
K ₂ O	1.14	0.12	0.06
TiO ₂	1.47	0.23	0.24
P ₂ O ₅	0.35	0.49	0.61
MnO	0.24	0.04	0.09
S	0.00	0.00	0.00
Ba	265	60	61
Cr	26	13	23
Zr	262	1082	683
Sr	171	129	80
Rb	34	0	0
Y	102	241	197
Nb	14	4	5
Zn	104	0	21
Ni	16	29	25
V	0	0	0
LOI	0.90	0.59	2.43
Total	101.71	100.64	100.31
Fe ₂ O ₃	3.38	0.53	1.91
FeO*	9.08	1.42	5.13
La	20.1	33.6	29.9
Ce	60.5	143.7	89.1
Nd	41.8	76.2	71.8
Sm	16.2	29.6	23.8
Eu	4.0	6.4	5.6
Gd	17.5	32.5	28.7
Dy	19.0	37.6	32.2
Er	11.3	23.8	19.1
Yb	10.6	20.4	16.5
Lu	1.58	2.86	2.41

Sill B: Profile 2

Sample	F873-1	F873-4	F873-9	F873-14	F873-17
Sill Type	Glom.	Glom.	Glom.	Glom.	Glom.
Lithology	Gabbroic	Gabbroic	Gabbroic	Gabbroic	Gabbroic
Height	6.5	12.5	25.0	49.0	59.0
SiO ₂	49.51	49.86	49.49	49.58	49.14
Al ₂ O ₃	15.26	15.07	15.92	15.82	15.44
Fe ₂ O ₃ T	14.63	13.03	12.60	12.36	12.88
Fe ₂ O ₃	1.56				
FeO	11.76				
MgO	7.02	6.63	6.80	6.88	6.94
CaO	6.14	8.90	9.47	9.83	10.31
Na ₂ O	3.85	2.44	2.52	2.32	2.30
K ₂ O	0.45	0.30	0.37	0.38	0.30
TiO ₂	1.38	1.18	1.15	1.17	1.09
P ₂ O ₅	0.09	0.08	0.07	0.07	0.07
MnO	0.26	0.22	0.19	0.18	0.20
S	0.07	0.09	0.09	0.09	0.09
Ba	348	119	66	79	49
Cr	184	216	194	171	243
Zr	70	62	56	58	54
Sr	132	111	139	154	163
Rb	9	0	8	0	0
Y	31	24	22	24	22
Nb	2	0	0	0	0
Zn	148	101	123	72	79
Ni	93	117	106	117	103
V	370	353	378	360	330
LOI	3.09	1.54	1.29	1.42	1.17
Total	99.75	99.31	99.97	100.10	99.93
Fe ₂ O ₃	2.08	1.85	1.79	1.75	1.83
FeO*	11.30	10.06	9.73	9.54	9.94
La	4.4				
Ce	11.2				
Nd	7.8				
Sm	2.5				
Eu	1.2				
Gd	3.6				
Dy	4.2				
Er	2.7				
Yb	2.3				
Lu	0.36				

Table A4.3 (continued). Whole rock major (wt. %) and trace (ppm) element analyses.

Menibek Gabbros (continued)

Sill B: Profile 2

Sample	F14-6	F14-5	F14-3	F14-1	F875-1	F875-2	F875-4	F875-5
Sill Type	Glom.	Glom.	Glom.	Glom.	Glom.	Glom.	Glom.	Glom.
Lithology	Gabbroic	Gabbroic	Gabbroic	Gabbroic	Gabbroic	Gabbroic	Gabbroic	Oxide
Height	402.0	422.0	444.5	451.7	465.0	478.0	489.4	495.0
SiO ₂	49.03	50.20	49.53	50.70	47.03	49.06	49.70	44.90
Al ₂ O ₃	16.07	16.13	15.69	17.66	18.82	18.76	16.94	12.73
Fe ₂ O ₃ T	10.37	10.80	10.49	10.01	8.72	8.69	10.39	19.41
Fe ₂ O ₃	1.06	2.38	0.76					
FeO	8.38	7.58	8.76					
MgO	6.80	6.21	6.35	5.84	6.05	5.91	6.38	6.54
CaO	11.79	12.31	12.30	12.73	12.92	12.41	10.54	11.23
Na ₂ O	1.92	2.58	2.35	2.22	2.04	2.28	2.58	2.06
K ₂ O	0.19	0.22	0.16	0.23	0.20	0.26	0.28	0.24
TiO ₂	0.79	0.81	0.72	0.73	0.55	0.69	0.80	1.91
P ₂ O ₅	0.06	0.05	0.05	0.04	0.04	0.04	0.06	0.04
MnO	0.17	0.18	0.19	0.16	0.14	0.16	0.19	0.21
S	0.06	0.04	0.05	0.02	0.02	0.08	0.06	0.02
Ba	79	47	64	54	81	93	237	5
Cr	124	72	43	79	257	178	60	0
Zr	38	36	31	29	26	29	33	30
Sr	112	105	111	112	166	150	194	87
Rb	0	3	0	5	3	0	13	0
Y	21	20	20	16	11	15	21	17
Nb	0	0	0	0	0	0	0	0
Zn	47	57	55	75	74	53	110	108
Ni	50	76	43	60	128	115	77	79
V	288	335	387	286	198	283	356	1642
LOI	1.70	1.22	1.37	0.72	2.43	2.02	1.98	1.04
Total	98.95	100.75	99.25	101.06	98.96	100.36	99.90	100.33
Fe ₂ O ₃	1.47	1.53	1.49	1.42	1.24	1.23	1.48	5.39
FeO*	8.01	8.34	8.10	7.73	6.73	6.71	8.02	12.62
La								6.0
Ce								17.6
Nd								13.5
Sm								3.5
Eu								1.1
Gd								2.6
Dy								1.9
Er								0.7
Yb								0.5
Lu								0.08

Table A4.3 (continued). Whole rock major (wt. %) and trace (ppm) element analyses.**Menihok Gabbros (continued)****Sill C**

Sample	F90-78A	F90-78B1	F90-78B2	SH-2	F90-78C	SH-3	F90-78D	SH-4	SH-6	SH-8
Lithology	Chill	Chill	Chill	Fine	Ol-Free	Ol-Free	Ol-Free	Ol-Free	Ol-Free	Ol-Bearing
Sill Type	Aphyric	Aphyric	Aphyric	Aphyric	Aphyric	Aphyric	Aphyric	Aphyric	Aphyric	Aphyric
Height	0.0	0.2	0.5	2.6	3.0	5.3	6.8	7.9	13.2	18.5
SiO ₂	48.45	51.09	48.93	49.37	49.03	48.93	48.90	49.08	47.91	48.07
Al ₂ O ₃	14.70	13.13	14.80	13.78	13.89	13.42	14.38	14.64	14.12	14.11
Fe ₂ O ₃ T	14.59	11.47	12.50	15.00	14.81	16.15	14.33	13.48	12.72	13.91
Fe ₂ O ₃								1.02	1.84	
FeO								11.21	9.79	
MgO	8.23	8.68	8.99	7.40	9.33	7.44	8.90	7.77	8.88	9.07
CaO	5.32	5.96	4.99	6.55	8.12	6.33	10.37	9.88	9.53	11.08
Na ₂ O	3.62	5.15	4.59	3.99	1.98	3.12	1.55	1.94	1.87	2.12
K ₂ O	0.16	0.10	0.23	0.67	0.26	0.35	0.29	0.25	0.33	0.23
TiO ₂	1.33	1.31	1.32	1.28	1.23	1.17	0.98	1.06	1.01	0.98
P ₂ O ₅	0.12	0.08	0.10	0.08	0.07	0.07	0.07	0.07	0.07	0.05
MnO	0.22	0.21	0.24	0.28	0.22	0.27	0.20	0.20	0.23	0.20
S	0.01	0.01	0.01	0.00	0.04	0.04	0.03	0.06	0.05	0.02
Ba	136	50	86	155	56	197	125	90	115	43
Cr	184	176	186	196	172	202	169	183	179	161
Zr	62	61	59	64	52	58	42	49	50	43
Sr	85	80	104	118	81	94	89	96	106	83
Rb	0	0	0	11	6	4	8	8	16	4
Y	31	26	24	29	28	26	24	25	23	21
Nb	0	6	0	0	0	2	0	0	3	0
Zn	53	60	41	57	108	69	68	85	64	92
Ni	114	94	117	116	129	91	73	88	80	332
V	502	558	498	470	425	451	402	403	361	353
LOI	4.06	3.60	4.09	2.39	1.27	2.81	1.71	1.31	2.32	1.03
Total	100.81	100.79	100.78	100.79	100.24	100.11	101.71	99.74	99.05	100.87
Fe ₂ O ₃	2.21	1.39	1.66	2.00	2.15	2.16	2.08	1.96	1.85	2.02
FeO*	12.94	8.17	9.75	11.70	11.39	12.60	11.02	10.37	9.78	10.70
La	6.9	7.0	5.7	3.8						
Ce	17.1	15.8	12.7	10.7						
Nd	12.6	11.5	8.5	7.3						
Sm	4.0	3.8	2.8	2.9						
Eu	1.1	1.3	0.8	0.8						
Gd	4.8	3.9	3.6	3.9						
Dy	5.3	4.0	3.9	4.7						
Er	3.2	2.5	2.5	2.9						
Yb	3.0	2.2	2.4	2.8						
Lu	0.43	0.35	0.34	0.43						

Table A4.3 (continued). Whole rock major (wt. %) and trace (ppm) element analyses.

Menibek Gabbros (continued)

Sill C

Sample	SH-41	F876-1	F876-3	F876-5	F876-7	F876-9	F876-12	F876-13	F876-14	F876-15
Sill Type	Aphyric	Aphyric	Aphyric	Aphyric	Aphyric	Aphyric	Aphyric	Aphyric	Aphyric	Aphyric
Lithology	Ol-Bearing	Ol-Free	Ol-Free	Ol-Free	Ol-Free	Ol-Free	Ol-Free	Ol-Free	Felsic	Felsic
Height	170.1	176.8	183.2	186.0	194.0	199.0	205.0	209.0	211.7	212.1
SiO ₂	48.48	49.06	49.49	49.16	48.79	47.86	48.96	49.11	63.75	66.47
Al ₂ O ₃	17.72	17.26	16.48	18.13	13.80	15.03	14.56	15.23	10.76	11.51
Fe ₂ O ₃ T	9.06	9.40	10.81	10.18	12.62	11.87	11.63	11.03	10.34	10.38
Fe ₂ O ₃	2.09	1.82	0.75		0.64		1.91	1.45		
FeO	6.27	6.82	9.05		10.78		8.75	8.62		
MgO	7.08	6.65	7.17	5.72	8.57	8.08	8.37	7.52	4.72	3.24
CaO	12.89	11.85	10.88	11.40	10.05	11.36	10.82	11.97	1.22	0.67
Na ₂ O	2.23	1.93	2.17	2.48	2.15	2.07	2.33	1.96	3.09	3.26
K ₂ O	0.32	0.38	0.42	0.48	0.33	0.43	0.36	0.40	0.59	1.34
TiO ₂	0.65	0.73	0.82	0.88	1.05	0.85	1.00	0.84	0.79	0.74
P ₂ O ₅	0.03	0.05	0.06	0.06	0.06	0.06	0.05	0.06	0.26	0.12
MnO	0.17	0.17	0.19	0.19	0.22	0.19	0.20	0.18	0.12	0.11
S	0.03	0.03	0.09	0.04	0.04	0.07	0.09	0.05	0.05	0.00
Ba	77	125	165	198	205	119	66	76	109	308
Cr	208	175	141	84	321	312	307	226	22	41
Zr	27	34	34	41	46	37	40	38	435	514
Sr	154	162	181	221	158	185	122	104	15	41
Rb	0	5	9	16	21	12	6	6	9	72
Y	15	19	17	22	24	20	21	21	132	143
Nb	3	0	0	0	3	0	0	0	25	41
Zn	45	47	63	125	156	151	78	29	96	85
Ni	136	39	46	37	70	69	69	83	11	2
V	263	280	281	290	347	317	316	324	0	0
LOI	1.84	1.85	2.03	2.07	1.99	2.44	1.55	1.90	3.91	2.62
Total	100.49	99.36	100.61	100.79	99.67	100.30	99.92	100.25	99.60	100.47
Fe ₂ O ₃	1.32	1.37	1.57	2.03	1.84	1.73	1.69	1.60	2.60	2.61
FeO*	6.97	7.23	8.31	7.33	9.70	9.13	8.94	8.48	6.97	7.00
La					2.8					27.0
Ce					6.5					76.8
Nd					6.0					51.7
Sm					2.1					20.4
Eu					0.8					4.9
Gd					2.8					24.2
Dy					3.5					28.1
Er					2.2					16.8
Yb					2.1					15.4
Lu					0.30					2.2

Table A4.3 (continued). Whole rock major (wt. %) and trace (ppm) element analyses.**Menthek Gabbros (continued)**

Sill C					Additional Gabbroic Samples				
Sample	F876-16	F876-17	F876-22	F876-23	B22-9	F23-40	H2-2	H3-5	W24-13
Sill Type	Aphyric	Aphyric	Aphyric	Aphyric	Aphyric	Glom.	Glom.	Glom.	Glom.
Lithology	Felsic	Felsic	Ol-Free	Fine	Anorth.	Anorth.	Anorth.	Anorth.	Anorth.
Height	214.3	215.1	219.1	221.7					
SiO ₂	64.35	63.98	49.34	49.28	46.20	47.40	48.50	44.30	48.80
Al ₂ O ₃	11.87	11.68	14.81	12.56	19.90	20.70	20.60	18.73	21.80
Fe ₂ O ₃ T	9.53	12.45	12.95	13.13	8.66	7.25	8.03	13.49	6.53
Fe ₂ O ₃									
FeO									
MgO	3.03	2.87	7.75	7.51	6.15	4.76	3.82	4.87	3.78
CaO	1.68	1.68	9.64	8.68	10.10	11.60	9.56	11.68	12.40
Na ₂ O	3.84	3.33	1.99	3.24	2.55	3.24	3.89	1.27	2.97
K ₂ O	0.57	0.58	0.51	0.39	1.32	0.48	0.35	0.13	0.73
TiO ₂	0.74	0.92	0.98	1.16	0.76	0.61	0.80	0.43	0.57
P ₂ O ₅	0.16	0.23	0.07	0.07	0.08	0.05	0.06	0.03	0.05
MnO	0.10	0.16	0.23	0.24	0.17	0.12	0.14	0.16	0.12
S	0.03	0.01	0.08	0.03	0.00	0.00	0.00	0.00	0.00
Ba	119	185	223	150	240	80	135	105	80
Cr	23	19	85	174	360	90	135	63	40
Zr	436	418	45	48	40	27	37	21	20
Sr	20	60	141	80	200	190	188	157	230
Rb	7	9	12	3	24	10	10	0	31
Y	155	145	24	23	0	20	21	11	0
Nb	23	14	0	0	20	20	21	21	10
Zn	65	73	76	73	48	77	1146	86	41
Ni	12	53	89	87	110	34	104	1573	25
V	0	0	376	477					
LOI	3.39	3.04	2.31	3.14	3.54	3.54	3.39	3.62	2.00
Total	99.28	100.92	100.66	99.43	99.43	99.75	99.14	98.71	99.75
Fe ₂ O ₃	2.39	3.13	1.88	1.75	1.15	0.97	2.02	1.78	0.87
FeO*	6.42	8.39	9.96	10.24	6.75	5.66	5.41	10.54	5.09
La	20.1								
Ce	54.1								
Nd	35.7								
Sm	13.0								
Eu	2.7								
Gd	14.0								
Dy	16.3								
Er	11.0								
Yb	11.9								
Lu	1.77								

Table A4.3 (continued). Whole rock major (wt. %) and trace (ppm) element analyses.**Menthek Gabbros (continued)****Additional Gabbroic Samples**

Sample	W24-6	F20-7B	F90-75	F90-76	F90-72A	F90-72B	F12-16	F25-2	F90-83A2	B22-8
Sill Type	Glom.	Aphyric	Aphyric	Aphyric	Glom.	Glom.	Aphyric	Aphyric	Glom.	Glom.
Lithology	Anorth.	Chill	Chill	Chill	Chill	Chill	Felsic	Felsic	Felsic	Felsic
Height										
SiO ₂	45.90	49.24	48.87	49.68	49.63	49.20	61.14	52.90	56.98	53.39
Al ₂ O ₃	21.10	14.79	14.06	14.05	14.03	13.69	15.59	13.72	13.91	11.73
Fe ₂ O ₃ T	8.47	14.30	13.68	12.98	16.26	15.81	9.20	12.38	8.27	15.11
Fe ₂ O ₃		1.03					4.20			
FeO		11.94					4.50			
MgO	7.56	7.19	7.24	7.10	6.16	6.07	1.35	4.61	4.39	5.18
CaO	12.10	9.86	10.77	10.69	9.22	9.17	2.88	8.49	8.87	6.88
Na ₂ O	1.75	2.61	1.82	1.60	1.11	1.29	5.47	4.17	4.91	2.84
K ₂ O	0.23	0.26	0.40	0.77	0.01	0.32	1.11	0.58	0.22	0.53
TiO ₂	0.48	1.23	1.22	1.26	1.59	1.59	0.65	1.41	0.27	1.59
P ₂ O ₅	0.05	0.11	0.07	0.08	0.13	0.13	0.09	0.24	0.16	0.23
MnO	0.11	0.20	0.21	0.20	0.28	0.20	0.21	0.26	0.22	0.27
S	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.05
Ba	70	118	130	57	23	9	1270	425	45	522
Cr	320	174	221	237	131	116	41	117	30	145
Zr	23	61	61	56	78	81	640	124	386	119
Sr	130	103	156	140	137	138	217	72	10	98
Rb	10	1	10	21	5	11	83	4	0	18
Y	0	31	28	24	37	35	191	63	104	57
Nb	20	0	4	4	6	6	114	5	0	3
Zn	50	150	0	0	0	0	103	38	34	50
Ni	220	95	98	113	106	56	7	105	165	83
V	262	378	309	291	368	355	0	23	314	488
LOI	1.70	1.04	2.07	2.15	2.45	3.19	2.62	1.43	2.11	2.08
Total	99.45	100.83	100.42	100.56	100.88	100.65	100.32	100.19	100.31	99.88
Fe ₂ O ₃	1.13	1.47	2.04	1.95	2.44	2.36	2.31	3.11	2.08	3.80
FeO*	6.61	11.54	10.47	9.93	12.43	12.10	6.20	8.34	5.57	10.19
La		4.6	3.0	3.5	6.3	5.2		6.1		
Ce		10.6	8.6	7.9	16.2	13.6		20.3		
Nd		7.9	6.2	6.5	11.6	10.5		14.9		
Sm		2.7	2.3	2.4	3.5	3.4		4.9		
Eu		1.0	0.6	0.8	1.2	1.2		1.6		
Gd		3.6	2.7	2.9	4.8	4.7		6.9		
Dy		4.4	3.7	3.7	6.1	5.8		8.3		
Er		2.6	2.3	2.4	4.0	3.6		5.1		
Yb		2.4	2.1	2.1	3.9	3.2		4.8		
Lu		0.38	0.35	0.32	0.49	0.52		0.71		

Table A4.3 (continued). Whole rock major (wt. %) and trace (ppm) element analyses.**Menthek Gabbros (continued)****Additional Gabbroic Samples**

Sample	F90-82A	F90-82B	F90-83A1	F90-83D	B23-11	B25-1	B25-2	B31-3	B31-5	F5-10
Sill Type	Glom.	Glom.	Glom.	Glom.	Aphyric	Aphyric	Aphyric	Aphyric	Aphyric	Aphyric
Lithology	Felsic	Felsic	Felsic	Felsic	Gabbroic	Gabbroic	Gabbroic	Gabbroic	Gabbroic	Gabbroic
Height										
SiO ₂	55.60	54.60	56.96	60.70	47.75	49.40	48.90	48.30	47.25	48.22
Al ₂ O ₃	14.56	14.42	12.56	12.70	14.04	13.03	16.20	14.00	14.39	13.85
Fe ₂ O ₃ T	11.36	11.52	8.63	8.00	14.04	14.03	11.40	14.20	13.59	13.91
Fe ₂ O ₃										2.52
FeO										10.25
MgO	4.40	3.57	2.70	3.60	8.24	6.55	5.86	6.24	8.61	8.04
CaO	7.13	8.09	8.49	5.25	10.03	9.62	11.50	9.35	10.99	9.53
Na ₂ O	4.38	4.46	5.29	3.83	1.93	2.45	1.96	2.49	1.37	2.19
K ₂ O	0.72	0.82	0.92	2.10	0.41	0.44	0.25	0.32	0.14	0.27
TiO ₂	1.19	1.31	1.19	0.65	0.94	1.35	0.96	1.44	0.67	1.10
P ₂ O ₅	0.41	0.33	0.41	0.50	0.06	0.11	0.07	0.12	0.05	0.08
MnO	0.27	0.25	0.20	0.14	0.20	0.21	0.18	0.24	0.22	0.23
S	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.05
Ba	283	225	16	896	235	113	72	103	82	207
Cr	80	64	64	88	184	206	92	124	544	181
Zr	247	196	193	219	41	72	41	63	37	53
Sr	76	102	57	33	82	113	154	155	72	113
Rb	5	8	0	103	20	15	31	0	0	5
Y	86	69	77	84	31	41	21	31	17	29
Nb	3	14	0	9	20	10	0	10	10	3
Zn	38	52	19		72	60	66	82	92	76
Ni	92	55	79	85	184	70	56	100	380	92
V	69	41	64	94						409
LOI	1.19	1.31	3.22	1.90	2.39	2.31	2.34	2.47	2.00	2.20
Total	101.21	100.68	100.58	99.37	100.03	99.50	99.63	99.17	99.27	99.66
Fe ₂ O ₃	2.85	2.89	2.17	1.88	2.04	2.04	1.66	2.02	1.98	2.02
FeO*	7.66	7.76	5.82	5.51	10.80	10.79	8.77	10.96	10.45	10.69
La			13.4							3.1
Ce			33.2							7.4
Nd			25.7							5.9
Sm			10.6							2.2
Eu			2.2							0.8
Gd			11.3							2.8
Dy			14.3							3.6
Er			9.1							2.3
Yb			8.4							2.1
Lu			1.22							0.31

Table A4.3 (continued). Whole rock major (wt. %) and trace (ppm) element analyses.

Menthek Gabbros (continued)

Additional Gabbroic Samples

Sample	B32-3	B32-7	F26-1AA	F13-6A	H2-3	H2-5	H2-6	H3-1	H3-4	W24-12
SW Type	Glom.	Glom.	Glom.	Glom.	Glom.	Glom.	Glom.	Glom.	Glom.	Glom.
Lithology	Gabbroic	Gabbroic	Gabbroic	Gabbroic	Gabbroic	Gabbroic	Gabbroic	Gabbroic	Gabbroic	Gabbroic
Height										
SiO ₂	46.65	47.60	48.73	47.06	47.60	48.18	45.81	50.16	46.55	49.98
Al ₂ O ₃	19.58	19.25	14.13	20.77	21.44	18.75	18.38	16.35	16.98	16.52
Fe ₂ O ₃ T	8.63	8.20	13.69	8.93	8.81	9.15	9.71	10.80	10.19	10.64
Fe ₂ O ₃				1.74			1.48			
FeO				6.47			7.40			
MgO	7.11	7.25	7.28	6.99	3.99	7.02	7.33	6.18	8.46	6.17
CaO	11.49	12.78	10.90	12.26	11.22	11.87	11.88	11.41	11.79	10.85
Na ₂ O	1.98	1.78	2.22	1.62	2.35	1.82	1.40	2.06	1.57	2.11
K ₂ O	0.27	0.30	0.15	0.25	0.61	0.16	0.25	0.19	0.15	0.22
TiO ₂	0.65	0.48	1.28	0.54	0.82	0.47	0.51	0.44	0.55	0.59
P ₂ O ₅	0.05	0.04	0.09	0.04	0.06	0.03	0.04	0.04	0.03	0.04
MnO	0.13	0.16	0.23	0.12	0.17	0.16	0.16	0.15	0.14	0.18
S	0.00	0.00	0.04	0.06	0.00	0.00	0.00	0.00	0.00	0.00
Ba	62	51	58	32	340	72	115	102	72	144
Cr	331	357	282	368	51	215	105	82	456	62
Zr	31	21	53	24	41	20	22	20	27	25
Sr	166	133	103	127	268	154	178	143	145	144
Rb	0	10	0	1	31	6	10	0	0	31
Y	14	10	33	15	23	10	9	12	10	10
Nb	21	20	3	3	10	10	2	31	0	31
Zn	67	55	69	55	81	67	83	69	76	99
Ni	207	194	78	202	85	348	669	81	352	56
V			393	180						
LOI	3.08	1.93	1.83	1.52	2.77	1.85	3.62	2.16	2.70	2.47
Total	99.63	99.78	100.57	100.15	99.84	99.46	99.09	99.94	99.11	99.78
Fe ₂ O ₃	1.23	1.16	1.83	1.27	1.25	1.30	1.50	1.53	1.45	1.51
FeO*	6.66	6.33	10.68	6.89	6.80	7.07	8.16	8.34	7.87	8.22
La			5.1							
Ce			13.7							
Nd			10.5							
Sm			3.5							
Eu			1.3							
Gd			4.9							
Dy			5.8							
Er			3.6							
Yb			3.5							
Lu			0.50							

Table A5.3 (continued). Whole rock major (wt.%) and trace (ppm) element analyses

Menthek Gabbros (continued)			Le Fer Gabbros						
Additional Gabbroic Samples									
Sample	F15-2	F16-11A	B88-22	B88-24	B88-29	B88-32	B88-34	B88-40	B88-40A
Sill Type	Glom.	Glom.							
Lithology	Vein	Vein	Chill	Chill	Gabbro	Gabbro	Gabbro	Gabbro	Gabbro
Height									
SiO ₂	62.04	67.83	48.27	47.78	50.13	48.23	48.23	50.55	50.69
Al ₂ O ₃	12.62	12.88	14.62	15.52	12.50	13.86	14.49	14.29	14.30
Fe ₂ O ₃ T	2.79	4.71	10.48	8.85	10.05	13.49	12.83	12.97	13.31
Fe ₂ O ₃			0.92	1.44		1.35	1.15	2.23	2.57
FeO			8.60	6.67	8.26	10.92	10.51	9.66	9.66
MgO	1.13	2.00	8.90	8.54	8.26	7.04	7.42	5.55	5.43
CaO	10.85	6.09	12.21	12.56	13.46	10.11	10.13	8.79	8.80
Na ₂ O	5.72	3.93	1.42	2.28	2.26	1.99	2.66	2.86	2.80
K ₂ O	0.01	0.16	0.53	0.19	0.07	0.16	0.22	1.52	1.53
TiO ₂	0.54	0.89	0.71	0.79	1.05	1.28	1.17	1.17	1.18
P ₂ O ₅	0.12	0.24	0.03	0.04	0.04	0.05	0.05	0.07	0.07
MnO	0.06	0.11	0.17	0.16	0.22	0.22	0.21	0.37	0.36
S	0.01	0.04	0.09	0.01	0.04	0.14	0.05	0.06	0.11
Ba	16	48	98	95	72	54	131	1003	994
Cr	27	25	783	595	339	212	259	18	15
Zr	554	315	32	45	56	60	54	60	60
Sr	13	79	171	117	114	105	158	246	245
Rb	0	0	7	0	0	0	0	27	26
Y	59	61	17	21	21	25	21	24	25
Nb	18	15	2	0	2	0	0	3	0
Zn	25	165	214	51	137	78	141	488	492
Ni	40	16	154	187	108	69	76	27	31
V	16	47	233	198	331	367	375	339	333
LOI	4.27	1.83	1.83	3.01	2.32	2.74	2.94	1.49	1.43
Total	100.17	100.73	99.26	99.73	100.40	99.31	100.40	99.69	100.01
Fe ₂ O ₃	0.70	1.18	1.41	1.19	1.35	1.81	1.72	1.74	1.79
FeO*	1.88	3.17	8.16	6.89	7.83	10.51	9.99	10.10	10.37
La	3.1	6.4		3.2			3.0		
Ce	8.0	16.2		10.1			9.5		
Nd	5.4	11.9		7.5			7.3		
Sm	2.0	4.4		2.6			2.4		
Eu	1.2	2.1		0.7			0.9		
Gd	3.3	5.9		3.2			3.3		
Dy	4.3	6.3		3.0			3.2		
Er	3.3	3.9		2.0			2.2		
Yb	3.6	3.9		2.0			2.2		
Lu	0.62	0.60		0.33			0.37		

Table A5.3 (continued). Whole rock major (wt.%) and trace (ppm) element analyses

Doublet Gabbros					Le Fer Basalts				
Sample	F8731-2	F8731-3	F8731-4	F8731-5	B88-25	B88-25A	B88-26	B88-30	B88-31
Lithology	Gabbro	Gabbro	Gabbro	Gabbro	Basalt	Basalt	Basalt	Basalt	Basalt
SiO ₂	48.29	49.23	44.04	48.88	50.05	49.18	49.25	47.65	49.50
Al ₂ O ₃	14.81	14.67	15.87	14.78	14.57	14.25	13.83	14.98	14.91
Fe ₂ O ₃ T	12.90	13.45	13.04	12.62	12.57	12.48	12.60	13.67	12.78
Fe ₂ O ₃							1.98	1.61	1.67
FeO							9.56	10.85	10.00
MgO	8.42	7.88	6.55	8.08	7.60	7.42	7.01	8.21	7.72
CaO	11.36	9.83	17.88	11.85	10.85	10.70	12.27	9.73	10.04
Na ₂ O	1.83	2.87	0.15	1.52	2.35	2.22	2.00	1.59	1.94
K ₂ O	0.07	0.17	0.04	0.08	0.04	0.04	0.06	0.16	0.09
TiO ₂	0.96	1.17	1.18	1.06	0.98	0.96	0.98	1.03	0.98
P ₂ O ₅	0.07	0.08	0.09	0.08	0.04	0.04	0.04	0.05	0.05
MnO	0.19	0.20	0.19	0.20	0.20	0.20	0.21	0.21	0.18
S	0.01	0.04	0.00	0.09	0.08	0.09	0.01	0.01	0.12
Ba	60	66	31	47	55	52	28	201	85
Cr	227	222	223	194	269	256	270	242	229
Zr	47	57	58	51	55	49	46	47	55
Sr	123	79	112	123	99	102	152	85	104
Rb	2	5	1	2	0	0	0	0	0
Y	17	23	23	20	23	25	21	21	20
Nb	2	6	5	4	4	0	4	5	0
Zn	75	83	60	85	72	95	88	85	92
Ni	168	115	41	163	86	91	120	103	118
V	295	353	368	325	292	301	361	357	325
LOI	2.18	2.27	2.20	2.13	1.15	1.35	1.98	2.98	2.28
Total	99.97	101.16	100.25	100.42	100.48	98.93	100.24	100.27	100.59
Fe ₂ O ₃	1.06	1.61	1.22	1.27	1.69	1.68	1.70	1.84	1.72
FeO*	9.65	10.06	9.75	9.44	9.79	9.72	9.81	10.64	9.95
La					2.8			3.1	2.4
Ce					9.1			9.4	7.6
Nd					6.8			7.0	6.4
Sm					2.3			2.2	2.4
Eu					0.8			0.7	0.7
Gd					3.2			3.0	2.8
Dy					3.7			3.2	3.3
Er					2.3			2.2	2.1
Yb					2.1			2.2	2.0
Lu					0.36			0.36	0.30

Table A5.3 (continued). Whole rock major (wt.%) and trace (ppm) element analyses

Le Fer Basalts										
Sample	B88-36	B88-37	F90-10	F90-11	F90-12	F90-13	F90-14	F90-21A1	F90-21A2	F90-21B
Lithology	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt
SiO ₂	49.04	49.29	49.07	49.29	48.31	49.95	48.97	50.07	50.68	48.67
Al ₂ O ₃	14.26	14.34	14.90	14.76	14.89	13.12	12.91	13.48	13.23	13.32
Fe ₂ O ₃ T	12.07	12.51	12.52	12.59	12.07	14.20	14.91	14.48	14.53	15.88
Fe ₂ O ₃	1.46			1.74		2.04	2.01			1.83
FeO	9.55			9.76		10.94	11.61			12.64
MgO	7.44	7.50	7.79	8.12	8.08	7.17	6.09	6.86	6.96	6.77
CaO	11.81	11.89	10.37	11.38	10.21	8.46	9.86	7.48	7.34	8.75
Na ₂ O	1.69	1.90	2.50	1.35	1.98	2.81	2.51	3.20	3.34	2.95
K ₂ O	0.07	0.09	0.07	0.07	0.10	0.02	0.02	0.02	0.02	0.04
TiO ₂	0.99	1.02	0.89	0.93	0.91	1.27	1.36	1.37	1.39	1.45
P ₂ O ₅	0.04	0.05	0.04	0.04	0.04	0.06	0.06	0.05	0.05	0.06
MnO	0.20	0.20	0.21	0.19	0.18	0.22	0.24	0.20	0.20	0.27
S	0.13	0.11	0.03	0.02	0.01	0.01	0.06	0.06	0.02	0.03
Ba	88	66	110	65	120	52	34	139	121	103
Cr	268	260	329	271	339	74	87	100	95	85
Zr	51	48	35	36	37	50	57	58	56	61
Sr	113	107	109	89	82	55	58	124	132	126
Rb	0	0	0	0	0	0	0	0	0	0
Y	22	19	20	18	18	27	27	30	29	30
Nb	0	0	2	3	0	5	4	0	0	0
Zn	93	94	65	59	50	61	96	68	78	93
Ni	112	74	165	148	161	71	69	72	75	70
V	325	336	354	377	367	501	485	499	511	532
LOI	1.51	1.53	2.68	2.35	3.24	3.37	2.88	3.22	2.95	2.65
Total	99.25	100.43	101.07	101.09	100.02	100.66	99.87	100.49	100.71	100.84
Fe ₂ O ₃	1.63	1.69	1.69	1.70	1.63	1.91	2.01	1.95	1.96	2.14
FeO*	9.40	9.74	9.75	9.80	9.40	11.06	11.61	11.27	11.31	12.36
La	3.2			2.8			4.3	3.8		
Ce	8.9			7.7			11.4	11.2		
Nd	6.9			6.7			8.9	9.3		
Sm	2.3			2.4			3.0	3.4		
Eu	0.8			0.8			1.0	1.0		
Gd	3.0			3.1			3.9	4.4		
Dy	3.6			3.6			4.5	4.9		
Er	2.1			2.1			2.7	3.2		
Yb	2.1			2.0			2.6	2.8		
Lu	0.32			0.30			0.38	0.42		

Table A5.3 (continued). Whole rock major (wt.%) and trace (ppm) element analyses

Le Fer Basalts		Lower Menihok Basalts							
Sample	F90-3	B23-5B	F18-25	F18-26	F20-9	F90-31	F90-50	F90-53	F90-53A
Lithology	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt
SiO ₂	49.45	43.05	42.10	43.73	43.32	42.72	42.32	43.63	43.85
Al ₂ O ₃	13.70	12.31	13.60	13.62	12.79	13.82	12.72	12.75	12.88
Fe ₂ O ₃ T	12.87	16.12	17.50	13.68	18.02	16.56	17.33	16.26	16.34
Fe ₂ O ₃		2.68	2.81	2.18	2.64	2.33	2.96	2.29	2.37
FeO		12.09	13.22	10.35	13.84	12.80	12.93	12.57	12.57
MgO	6.92	7.80	6.24	5.93	6.57	7.58	7.40	5.83	5.88
CaO	11.62	5.41	7.48	10.21	6.53	6.14	7.65	8.41	8.42
Na ₂ O	1.26	1.93	2.92	3.40	2.91	2.72	2.49	1.39	1.38
K ₂ O	0.02	1.89	0.61	0.91	1.53	1.31	1.12	3.38	3.40
TiO ₂	1.17	4.75	4.84	4.56	5.06	5.12	5.58	4.68	4.76
P ₂ O ₅	0.05	0.81	0.50	0.48	0.70	0.60	0.61	0.44	0.44
MnO	0.22	0.16	0.31	0.31	0.24	0.57	0.32	0.30	0.30
S	0.06	0.10	0.09	0.08	0.02	0.09	0.05	0.05	0.07
Ba	56	317	340	336	325	1205	628	724	772
Cr	275	9	90	18	4	79	52	102	102
Zr	51	187	180	160	177	164	163	292	295
Sr	157	83	220	553	213	255	162	1606	1611
Rb	0	18	20	14	32	23	16	100	100
Y	26	28	31	26	30	27	29	34	34
Nb	0	39	50	27	38	36	37	54	56
Zn	90	293	150	300	89	87	92	131	125
Ni	103	3	77	17	0	73	51	117	111
V	401	190	0	356	299	55	0	68	54
LOI	2.11	5.52	3.16	2.91	1.90	2.99	2.37	2.12	2.20
Total	99.45	99.85	99.26	99.82	99.59	100.22	99.96	99.24	99.92
Fe ₂ O ₃	1.73	2.47	2.69	2.10	2.77	2.54	2.66	2.50	2.51
FeO*	10.02	12.28	13.33	10.42	13.73	12.61	13.20	12.39	12.45
La					40.4		33.0	43.1	
Ce					89.9		79.0	97.2	
Nd					50.2		41.1	47.7	
Sm					12.3		9.4	11.1	
Eu					3.4		2.5	2.9	
Gd					9.6		7.7	9.2	
Dy					6.2		4.9	6.6	
Er					2.6		1.7	2.5	
Yb					1.8		1.3	2.1	
Lu					0.28		0.25	0.33	

Table A5.3 (continued). Whole rock major (wt.%) and trace (ppm) element analyses

Lower Menihok Basalts										
Sample	W29-2	F17-2/WR	F17-3/FC	F17-3/MC	F17-3/WR	F90-16A	F90-16B	F90-16C	F90-16D	F90-16E
Lithology	Basalt	Clast	Clast	Clast	Clast	Clast	Clast	Clast	Clast	Clast
SiO ₂	42.60	46.56	39.57	43.41	41.83	45.45	51.44	51.24	51.34	47.41
Al ₂ O ₃	12.60	13.28	15.56	15.00	13.36	10.73	13.48	13.40	13.39	12.57
Fe ₂ O ₃ T	17.20	16.93	18.51	13.56	16.38	11.83	10.65	10.59	10.91	12.73
Fe ₂ O ₃	2.59									5.35
FeO	13.15									6.64
MgO	7.23	5.01	5.64	4.15	5.11	8.01	7.12	6.97	7.10	7.26
CaO	5.53	5.20	4.46	7.18	7.57	10.44	5.94	5.82	5.95	6.86
Na ₂ O	2.49	2.81	3.73	4.13	2.85	4.32	6.07	5.99	6.15	4.87
K ₂ O	2.13	0.68	0.38	0.41	0.60	0.12	0.18	0.17	0.18	0.26
TiO ₂	5.43	4.30	5.79	5.00	4.35	3.55	3.56	3.49	3.55	4.25
P ₂ O ₅	0.67	1.18	1.93	1.91	1.10	0.37	0.63	0.60	0.65	0.52
MnO	0.41	0.22	0.24	0.18	0.25	0.15	0.10	0.10	0.10	0.11
S	0.24	0.02	0.04	0.01	0.06	0.01	0.01	0.01	0.08	0.01
Ba	1190	92	0	0	20	0	5	22	0	0
Cr	30	82	30	45	67	757	115	114	114	344
Zr	180	330	434	366	327	228	331	339	344	283
Sr	250	100	154	176	142	148	79	80	80	93
Rb	50	10	7	4	13	0	0	2	0	5
Y	20	45	52	48	46	33	42	46	44	35
Nb	60	73	96	87	76	56	77	80	85	65
Zn	120	120	127	141	182	86	76	71	72	79
Ni	34	108	41	65	50	569	279	272	297	305
V		44	0	34	50	177	110	97	89	105
LOI	3.16	4.11	4.86	4.66	6.14	5.51	2.67	2.61	2.61	3.03
Total	99.45	100.30	100.71	99.60	99.60	100.49	101.85	100.99	101.93	99.88
Fe ₂ O ₃	2.64	2.60	2.84	2.08	2.51	1.82	1.63	1.62	1.67	1.95
FeO*	13.10	12.90	14.10	10.33	12.48	9.01	8.11	8.07	8.31	9.70
La	33.2		41.5	9.0	8.1	44.3	37.0		32.3	
Ce	78.2		98.8	44.3	37.0	104.1	90.8		81.6	
Nd	47.4		50.7	104.1	90.8	55.8	48.9		40.6	
Sm	11.3		11.6	55.8	48.9	13.9	12.9		10.5	
Eu	3.9		3.1	13.9	12.9	3.5	3.2		2.4	
Gd	9.3		9.5	3.5	3.2	11.3	10.7		8.4	
Dy	6.0		6.4	11.3	10.7	8.1	7.6		5.9	
Er	2.9		2.7	8.1	7.6	3.1	3.0		2.3	
Yb	1.8		2.3	3.1	3.0	2.7	2.5		1.8	
Lu	0.26		0.37	2.70	2.49	0.39	0.38		0.29	

Table A5.3 (continued). Whole rock major (wt.%) and trace (ppm) element analyses

Lower Menihok Basalts						Murdoch Basalts				
Sample	F90-17B	F90-17D	F90-18	F90-24	F90-41	F90-CL2	F90-CL3	F90-CL4	F90-CL5	F90-85A
Lithology	Clast	Clast	Clast	Clast	Trachy.	Basalt	Basalt	Basalt	Basalt	Basalt
SiO ₂	49.48	49.78	43.04	48.64	57.48	47.56	45.16	46.11	47.68	49.89
Al ₂ O ₃	14.51	14.58	7.67	13.74	9.85	11.98	10.79	12.43	14.78	13.98
Fe ₂ O ₃ T	12.52	12.65	12.19	10.86	8.67	13.49	16.23	15.37	13.85	13.13
Fe ₂ O ₃	5.79	5.82								4.23
FeO	6.06	6.06								8.01
MgO	5.88	6.01	21.15	5.66	8.23	11.72	10.59	9.13	6.12	5.99
CaO	4.37	4.51	6.93	8.86	7.65	6.47	7.61	6.82	8.31	11.08
Na ₂ O	6.13	6.22	0.00	4.74	4.89	4.06	3.54	3.66	3.79	2.63
K ₂ O	0.16	0.11	0.09	0.54	0.32	0.22	0.13	0.19	0.12	0.06
TiO ₂	3.45	3.40	1.62	4.70	1.87	2.54	3.58	3.91	2.31	1.50
P ₂ O ₅	0.65	0.61	0.10	0.69	0.15	0.26	0.34	0.33	0.24	0.09
MnO	0.08	0.11	0.20	0.17	0.14	0.16	0.13	0.17	0.19	0.15
S	0.01	0.01	0.00	0.02	0.01	0.00	0.00	0.00	0.01	0.01
Ba	0	31	42	289	25	87	54	165	43	35
Cr	76	71	2067	181	77	167	183	161	151	196
Zr	313	327	75	300	283	121	98	135	113	74
Sr	107	119	26	316	50	169	156	209	131	204
Rb	4	0	0	3	0	0	0	0	0	0
Y	43	48	17	37	43	39	44	31	31	41
Nb	84	84	15	59	55	17	12	16	7	0
Zn	75	68	106	123	17	54	66	31	67	45
Ni	73	81	169	173	54	141	62	83	73	59
V	78	79	169	234	107	581	549	511	671	517
LOI	2.36	2.34	6.93	2.68	1.06	2.48	2.42	2.60	3.28	2.04
Total	99.88	100.46	99.92	101.30	100.32	100.94	100.52	100.72	100.68	100.55
Fe ₂ O ₃	1.92	1.99	1.87	1.67	1.33	2.40	2.89	2.73	2.46	2.22
FeO*	9.54	9.63	9.29	8.27	6.60	9.98	12.01	11.37	10.25	9.82
La	54.2		6.9	30.2				13.9	9.2	7.2
Ce	132.0		18.1	66.8				33.5	22.5	17.5
Nd	59.1		12.7	37.8				21.7	17.0	12.9
Sm	13.4		3.8	8.6				6.0	5.2	4.6
Eu	3.4		0.9	2.0				2.0	1.7	1.6
Gd	10.9		3.3	6.7				7.0	6.3	6.2
Dy	7.7		2.5	4.8				7.1	6.2	7.3
Er	3.0		1.1	1.8				3.6	3.6	4.5
Yb	2.6		1.0	1.6				2.5	2.6	3.5
Lu	0.39		0.15	0.25				0.33	0.34	0.47

Table A5.3 (continued). Whole rock major (wt.%) and trace (ppm) element analyses

Murdoch Basalts						Upper Menihok Basalts			
Sample	F90-85B	F90-85C	F90-85D	F90-85F	F90-85G	F90-127	F90-128	F90-129	F90-19
Lithology	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt
SiO ₂	49.41	50.80	49.83	51.39	50.60	49.10	48.70	46.40	49.22
Al ₂ O ₃	14.19	13.16	14.17	15.78	13.20	13.70	13.50	14.10	14.17
Fe ₂ O ₃ T	13.56	13.71	16.01	13.33	14.50	14.00	14.50	14.70	13.23
Fe ₂ O ₃	3.71	3.23	2.96	3.52	3.44				1.62
FeO	8.86	9.43	11.74	8.83	9.95				10.45
MgO	5.84	6.01	6.21	4.97	5.85	7.12	7.19	7.42	7.99
CaO	11.29	11.07	6.45	5.92	10.92	10.10	10.10	8.78	4.38
Na ₂ O	2.49	2.40	2.09	4.09	1.45	1.98	1.90	2.65	4.08
K ₂ O	0.23	0.14	0.73	0.08	0.06	0.42	0.31	0.16	0.07
TiO ₂	1.51	1.67	1.79	1.98	1.68	1.43	1.38	1.33	1.22
P ₂ O ₅	0.07	0.12	0.10	0.16	0.14	0.11	0.11	0.10	0.07
MnO	0.17	0.19	0.20	0.18	0.20	0.24	0.25	0.20	0.18
S	0.04	0.04	0.03	0.01	0.04	0.00	0.00	0.00	0.04
Ba	111	42	103	23	61	118	158	50	56
Cr	231	204	154	160	146	154	172	191	174
Zr	73	81	91	103	87	68	68	63	55
Sr	230	172	109	100	226	160	210	97	65
Rb	0	0	22	0	0	16	11	5	0
Y	33	38	42	28	35	35	32	29	28
Nb	0	0	2	3	3	0	3	6	4
Zn	44	56	71	70	70	93	76	83	76
Ni	214	110	49	54	87	93	95	108	129
V	516	589	612	651	564	380	391	381	467
LOI	2.01	1.79	3.14	3.28	2.42	2.60	2.72	3.31	4.65
Total	100.81	101.10	100.75	101.17	101.06	100.80	100.66	99.15	99.30
Fe ₂ O ₃	2.29	2.32	2.70	2.25	2.45	2.36	2.45	2.48	2.23
FeO*	10.14	10.25	11.97	9.97	10.84	10.47	10.84	10.99	9.89
La	6.9	7.4	6.6	7.3	5.6				4.4
Ce	15.9	19.5	17.1	18.6	14.4				11.2
Nd	11.6	14.9	12.3	14.3	11.2				8.6
Sm	4.2	5.4	4.3	5.2	4.0				2.8
Eu	1.2	1.4	1.3	1.5	1.1				0.5
Gd	5.1	6.0	5.2	5.7	4.8				3.6
Dy	5.5	6.6	6.1	6.2	5.2				4.2
Er	3.4	3.9	3.6	3.5	3.3				2.6
Yb	2.9	3.3	3.0	3.0	2.8				2.4
Lu	0.38	0.43	0.40	0.38	0.38				0.35

Table A5.3 (continued). Whole rock major (wt.%) and trace (ppm) element analyses**Upper Menihok Basalts**

Sample	F90-20	F90-87	F90-88	F90-89	F90-89A	F90-90	F90-91	F90-92	F90-93	F90-86A
Lithology	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Gl. Basalt
SiO₂	48.36	49.73	48.66	48.00	47.90	49.28	49.92	50.73	50.61	47.65
Al₂O₃	13.97	13.66	14.51	14.39	14.38	14.30	14.05	13.71	13.81	13.22
Fe₂O₃T	13.74	14.46	12.46	14.49	14.39	13.99	14.18	13.31	13.43	15.96
Fe₂O₃	2.58	2.46		2.74	2.64	2.12	2.07	2.26	2.67	
FeO	10.04	10.80		10.57	10.57	10.68	10.90	9.94	9.68	
MgO	7.11	6.68	7.07	6.94	6.91	7.16	7.29	6.62	7.03	7.38
CaO	9.12	11.01	10.61	11.29	11.23	10.88	10.39	11.07	10.22	9.13
Na₂O	2.79	1.85	2.48	2.21	2.11	1.97	2.54	2.45	2.21	2.11
K₂O	0.59	0.17	0.13	0.07	0.07	0.12	0.14	0.10	0.10	0.21
TiO₂	1.31	1.46	1.35	1.26	1.25	1.28	1.25	1.22	1.29	1.68
P₂O₅	0.09	0.11	0.09	0.08	0.07	0.08	0.09	0.07	0.08	0.14
MnO	0.22	0.25	0.23	0.21	0.21	0.22	0.22	0.21	0.19	0.28
S	0.16	0.08	0.01	0.05	0.05	0.07	0.07	0.06	0.05	0.05
Ba	99	46	44	49	24	51	39	59	52	23
Cr	181	186	235	174	176	177	180	175	186	143
Zr	62	73	67	54	55	58	59	52	60	91
Sr	248	94	171	93	94	102	106	85	118	32
Rb	5	0	0	0	0	0	0	0	0	0
Y	30	34	31	29	32	32	28	29	29	43
Nb	0	5	0	0	3	7	0	3	0	0
Zn	83	84	63	78	70	70	65	76	60	54
Ni	102	101	122	127	119	115	110	113	93	84
V	494	521	488	529	494	511	504	488	438	649
LOI	2.42	1.36	3.42	2.12	1.95	1.36	1.67	1.37	1.75	3.21
Total	99.88	100.82	101.02	101.11	100.52	100.71	101.11	100.92	100.77	101.02
Fe₂O₃	2.32	2.44	2.10	2.45	2.43	2.36	2.39	2.25	2.27	2.69
FeO*	10.28	10.81	9.32	10.84	10.76	10.46	10.61	9.95	10.04	11.94
La	4.3	5.0		4.8		5.1		3.9	4.6	
Ce	11.5	13.5		11.8		13.9		10.0	11.5	
Nd	8.9	10.2		9.1		10.4		7.8	8.8	
Sm	2.7	3.7		3.3		3.5		2.6	3.0	
Eu	0.5	1.2		1.1		1.1		1.0	1.1	
Gd	3.6	5.1		4.6		4.6		3.7	4.2	
Dy	4.3	5.6		5.2		5.4		4.5	4.6	
Er	2.7	3.4		3.3		3.3		2.9	2.8	
Yb	2.4	3.2		3.2		3.1		2.7	2.7	
Lu	0.30	0.46		0.46		0.47		0.35	0.38	

Table A5.3 (continued). Whole rock major (wt.%) and trace (ppm) element analyses

Upper Menhek Basalts			Willibob Basalts						
Sample	F90-86B	F90-86C	F90-100	F90-101	F90-104	F90-105	F90-106	F90-107	F90-108
Lithology	Gl. Basalt	Gl. Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt
SiO ₂	47.12	48.20	51.21	50.51	49.50	46.30	48.20	48.45	50.18
Al ₂ O ₃	13.42	13.90	13.84	13.78	14.60	13.80	13.20	14.18	14.50
Fe ₂ O ₃ T	16.36	15.40	11.12	13.08	12.30	13.90	13.70	13.01	12.64
Fe ₂ O ₃									
FeO									
MgO	7.85	7.13	7.75	6.19	6.51	5.73	6.13	7.41	7.44
CaO	9.46	8.50	12.39	9.49	10.20	10.70	11.40	11.98	10.64
Na ₂ O	1.98	2.00	2.08	3.54	2.70	1.89	1.11	1.69	2.12
K ₂ O	0.26	0.73	0.07	0.23	0.06	0.12	0.07	0.12	0.17
TiO ₂	1.61	1.68	0.99	1.27	1.19	1.51	1.21	1.16	1.20
P ₂ O ₅	0.16	0.16	0.06	0.10	0.09	0.12	0.08	0.07	0.07
MnO	0.28	0.30	0.18	0.21	0.19	0.19	0.15	0.24	0.21
S	0.09	0.24	0.05	0.02	0.20	0.24	0.24	0.04	0.04
Ba	17	213	50	252	17	41	23	90	80
Cr	154	95	216	177	240	127	213	204	210
Zr	88	96	44	62	65	78	67	54	54
Sr	29	117	79	128	130	121	113	112	89
Rb	0	0	0	0	0	0	0	0	0
Y	44	45	23	29	27	34	26	28	28
Nb	2	4	0	3	0	0	0	0	2
Zn	53	132	71	102	84	56	76	80	89
Ni	94	107	125	105	94	60	112	125	125
V	620	450	434	461	348	365	312	466	481
LOI	2.04	1.41	1.74	2.51	2.08	4.91	4.38	1.84	1.71
Total	100.63	99.41	101.48	100.93	99.42	99.17	99.63	100.19	100.92
Fe ₂ O ₃	2.76	2.60	1.88	2.21	2.08	2.35	2.31	2.20	2.13
FeO*	12.24	11.52	8.32	9.78	9.20	10.40	10.25	9.73	9.45
La			3.2	4.3					4.0
Ce			8.9	11.1					9.8
Nd			6.8	8.3					7.7
Sm			2.1	2.8					2.6
Eu			0.6	0.9					0.8
Gd			2.8	4.1					3.6
Dy			3.4	4.4					4.1
Er			2.1	2.6					2.5
Yb			1.9	2.3					2.3
Lu			0.27	0.32					0.32

Table A5.3 (continued). Whole rock major (wt.%) and trace (ppm) element analyses**Willibob Basalts**

Sample	F90-109	F90-110	F90-111	F90-112	F90-113	F90-114	F90-115	F90-116	F90-117	F90-118
Lithology	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt
SiO ₂	49.64	50.02	48.71	49.12	49.23	47.99	49.05	48.96	48.51	49.40
Al ₂ O ₃	14.14	14.00	13.79	13.65	13.98	14.59	13.95	14.10	13.85	14.01
Fe ₂ O ₃ T	13.02	12.59	14.44	13.13	13.73	12.61	14.31	13.43	13.20	12.94
Fe ₂ O ₃			4.05	3.79	2.97	3.55				
FeO			9.35	8.40	9.68	8.15				
MgO	7.32	7.47	5.81	6.54	7.16	7.35	7.67	7.60	6.96	7.09
CaO	10.95	10.28	10.71	11.38	9.34	11.01	10.05	11.19	12.57	11.80
Na ₂ O	1.93	2.00	1.78	2.21	2.62	2.35	1.44	2.11	1.39	1.79
K ₂ O	0.47	0.57	0.23	0.15	0.24	0.23	0.08	0.11	0.07	0.11
TiO ₂	1.16	1.25	1.53	1.18	1.19	1.09	1.14	1.13	1.15	1.16
P ₂ O ₅	0.07	0.07	0.12	0.09	0.07	0.06	0.07	0.08	0.08	0.07
MnO	0.23	0.22	0.21	0.22	0.23	0.21	0.21	0.21	0.21	0.21
S	0.07	0.05	0.03	0.05	0.03	0.03	0.02	0.06	0.02	0.07
Ba	170	226	37	97	59	91	59	48	49	59
Cr	199	204	110	198	204	185	204	197	192	209
Zr	54	55	75	59	52	49	51	54	51	53
Sr	93	81	101	131	71	96	89	116	125	104
Rb	9	9	3	0	0	0	0	0	0	0
Y	26	27	32	27	25	25	26	28	28	24
Nb	2	0	3	0	0	0	0	2	2	0
Zn	92	117	100	76	84	83	81	66	82	72
Ni	123	172	76	108	111	138	130	127	123	133
V	468	389	492	480	482	422	467	470	462	466
LOI	1.55	1.85	2.16	2.33	1.99	2.03	2.55	2.05	2.05	1.82
Total	100.55	100.37	99.52	100.05	99.81	99.55	100.54	101.03	100.06	100.47
Fe ₂ O ₃	2.20	2.13	2.44	2.22	2.32	2.13	2.42	2.27	2.23	2.19
FeO*	9.74	9.42	10.80	9.82	10.27	9.43	10.70	10.04	9.87	9.68
La					3.0			4.0	4.4	
Ce					8.1			10.3	10.3	
Nd					6.5			7.8	8.2	
Sm					2.6			2.7	2.9	
Eu					0.8			0.5	0.9	
Gd					3.4			3.7	3.6	
Dy					4.0			4.3	4.0	
Er					2.3			2.6	2.3	
Yb					2.1			2.4	2.1	
Lu					0.23			0.32	0.29	

Table A5.3 (continued). Whole rock major (wt.%) and trace (ppm) element analyses

Willibob Basalts

Sample	F90-119	F90-120	F90-121	F90-122	F90-94	F90-95	F90-96	F90-97	F90-98	F90-99
Lithology	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt	Basalt
SiO ₂	48.56	49.70	49.10	49.10	50.50	50.81	49.35	48.84	49.51	49.06
Al ₂ O ₃	14.07	14.30	13.80	13.80	13.95	14.09	13.57	13.91	13.77	14.15
Fe ₂ O ₃ T	14.07	13.74	13.74	13.33	13.46	13.48	13.23	14.20	14.56	12.55
Fe ₂ O ₃		2.07	2.07	2.03	3.28	2.73	3.44			3.60
FeO		10.50	10.50	10.17	9.16	9.67	8.81			8.05
MgO	7.55	6.47	7.08	5.77	6.91	6.66	6.79	7.09	5.65	7.46
CaO	10.33	10.10	11.10	11.50	10.25	9.57	10.76	10.96	10.78	11.45
Na ₂ O	2.02	2.61	1.88	1.90	2.26	1.54	2.52	1.48	1.97	1.72
K ₂ O	0.10	0.17	0.15	0.16	0.04	0.18	0.21	0.66	0.20	0.21
TiO ₂	1.14	1.27	1.19	1.28	1.29	1.27	1.25	1.20	1.57	1.14
P ₂ O ₅	0.08	0.09	0.08	0.09	0.07	0.07	0.07	0.08	0.11	0.07
MnO	0.22	0.18	0.20	0.18	0.20	0.20	0.20	0.21	0.21	0.19
S	0.01	0.24	0.21	0.21	0.07	0.03	0.03	0.06	0.06	0.03
Ba	55	13	101	22	13	88	70	90	26	78
Cr	196	174	168	170	173	179	171	176	114	196
Zr	52	68	60	71	62	57	60	52	80	55
Sr	96	118	77	123	102	64	102	113	114	92
Rb	0	5	5	5	0	0	0	15	0	0
Y	24	26	24	27	27	29	29	28	35	28
Nb	0	5	5	6	0	0	5	4	0	0
Zn	80	0	0	0	103	100	89	93	95	66
Ni	118	77	79	72	94	100	105	132	82	131
V	466	366	339	344	493	503	479	482	512	447
LOI	2.28	2.17	2.09	1.91	2.32	3.18	2.43	1.95	1.77	2.13
Total	100.43	100.80	100.41	99.02	101.32	101.08	100.41	100.64	100.16	100.16
Fe ₂ O ₃	2.38	2.32	2.32	2.25	2.27	2.28	2.23	2.40	2.46	2.12
FeO*	10.52	10.28	10.28	9.97	10.07	10.08	9.89	10.62	10.89	9.39
La							3.8	4.1	5.5	3.4
Ce							10.7	11.1	14.3	10.0
Nd							8.2	7.9	10.7	7.3
Sm							2.5	2.5	3.5	2.2
Eu							0.8	0.6	1.2	0.7
Gd							3.4	3.5	4.9	3.0
Dy							4.1	4.1	5.5	3.4
Er							2.4	2.6	3.4	2.1
Yb							2.1	2.3	2.8	1.8
Lu							0.28	0.32	0.42	0.28

Table A5.3 (continued). Whole rock major (wt.%) and trace (ppm) element analyses

Willibob Basalts

Sample	F90-99A	F87-100	F8722-15	F8722-2	F8723-18	F8723-23	F8733-17	F8735-6	F8736-1	F90-123
Lithology	Basalt	Gl. Basalt	Gl. Basalt	Gl. Basalt	Gl. Basalt	Gl. Basalt	Gl. Basalt	Gl. Basalt	Gl. Basalt	Gl. Basalt
SiO ₂	49.38	48.06	48.95	49.27	47.66	47.52	49.66	48.81	48.68	48.57
Al ₂ O ₃	14.29	14.10	13.83	13.90	14.28	14.22	13.93	13.87	13.83	14.07
Fe ₂ O ₃ T	12.36	13.55	14.16	14.92	12.72	12.63	14.89	14.70	14.44	13.87
Fe ₂ O ₃	3.41	2.16	2.84	3.63	2.56			2.96	3.22	2.09
FeO	8.05	9.80	10.19	10.16	9.14			10.56	10.10	10.62
MgO	7.47	7.25	5.76	5.92	8.47	8.22	6.59	6.14	5.98	5.53
CaO	11.56	9.34	12.06	10.34	10.68	10.65	8.71	10.12	10.61	12.91
Na ₂ O	1.71	2.43	1.27	2.28	1.66	1.51	2.07	1.69	2.23	1.54
K ₂ O	0.21	0.31	0.17	0.16	0.34	0.34	0.18	0.23	0.16	0.05
TiO ₂	1.14	1.26	1.46	1.55	1.10	1.10	1.60	1.55	1.54	1.39
P ₂ O ₅	0.07	0.09	0.11	0.13	0.07	0.08	0.14	0.12	0.12	0.11
MnO	0.19	0.21	0.21	0.25	0.20	0.22	0.24	0.27	0.26	0.23
S	0.06	0.01	0.09	0.03	0.10	0.08	0.02	0.08	0.06	0.03
Ba	105	106	44	61	107	62	19	44	40	14
Cr	205	137	113	126	214	228	124	114	105	142
Zr	50	63	76	86	55	53	86	80	82	69
Sr	95	74	126	149	105	107	108	115	124	248
Rb	0	0	0	0	0	0	3	3	0	6
Y	27	25	32	33	27	23	39	33	32	29
Nb	0	0	0	3	3	0	0	0	4	4
Zn	64	121	80	117	83	69	80	76	93	0
Ni	144	126	51	81	127	132	33	49	41	76
V	443	342	430	467	342	339	466	463	464	371
LOI	1.59	2.80	2.21	1.95	2.47	2.79	2.38	2.12	1.78	2.13
Total	100.03	99.41	100.27	100.70	99.75	99.36	100.41	99.70	99.69	100.40
Fe ₂ O ₃	2.09	2.29	2.39	2.52	2.15	2.13	2.51	2.48	2.44	2.34
FeO*	9.24	10.13	10.59	11.16	9.51	9.45	11.14	10.99	10.80	10.37
La			5.4	6.1	4.4		5.1	5.4	5.6	
Ce			13.8	15.9	12.2		13.0	15.0	14.3	
Nd			10.5	11.9	9.0		10.6	11.5	11.0	
Sm			3.6	4.0	2.9		3.6	3.8	3.7	
Eu			1.2	1.4	1.0		1.0	1.3	1.3	
Gd			4.7	5.5	3.9		4.7	4.8	5.0	
Dy			5.2	6.3	4.5		5.5	5.7	5.7	
Er			3.4	3.8	2.7		3.4	3.6	3.2	
Yb			3.1	3.3	2.3		3.1	3.3	3.0	
Lu			0.46	0.46	0.34		0.46	0.51	0.40	

Table A5.3 (continued). Whole rock major (wt.%) and trace (ppm) element analyses

Willibob Basalts				Glomerocryst Plagioclase					
Sample	F90-124	F90-125	F90-126	F14-18G	F871-4G	F15-4G	F26-1AAG	F90-74AG	F874-11G
Lithology	Gl. Basalt	Gl. Basalt	Gl. Basalt	Cryst.	Cryst.	Cryst.	Cryst.	Cryst.	Cryst.
SiO ₂	50.49	48.60	49.70	45.97	46.23	47.56	47.78	47.87	48.00
Al ₂ O ₃	13.89	13.86	13.71	30.97	30.24	31.01	29.43	30.77	29.84
Fe ₂ O ₃ T	13.61	14.15	14.52	1.45	1.70	1.51	2.83	1.16	1.91
Fe ₂ O ₃	2.05	2.15	2.18						
FeO	10.39	10.85	11.07						
MgO	5.18	6.54	5.51	0.35	0.32	0.22	0.67	0.12	0.48
CaO	10.29	9.76	11.00	16.72	16.48	15.54	14.61	15.35	14.94
Na ₂ O	2.15	2.74	1.59	1.70	2.02	2.46	2.06	2.34	2.03
K ₂ O	0.04	0.10	0.06	1.44	0.77	0.38	0.56	0.67	0.89
TiO ₂	1.58	1.47	1.53	0.04	0.03	0.04	0.04	0.03	0.03
P ₂ O ₅	0.12	0.13	0.13	0.00	0.00	0.00	0.00	0.00	0.00
MnO	0.21	0.24	0.26	0.01	0.00	0.01	0.01	0.01	0.01
S	0.06	0.31	0.08	0.07	0.06	0.04	0.07	0.03	0.04
Ba	9	29	10	204	169	140	219	134	183
Cr	120	128	126	21	20	21	19	23	23
Zr	88	79	85	0	0	0	0	0	0
Sr	126	108	119	332	261	249	369	266	347
Rb	5	5	5	54	22	9	47	15	30
Y	38	34	35	0	4	3	3	3	4
Nb	6	6	7	0	0	2	0	2	0
Zn	0	0	0	22	16	30	27	17	17
Ni	69	81	67	11	5	9	11	13	6
V	399	396	395	0	0	0	4	4	18
LOI	2.40	2.68	2.15						
Total	99.96	100.27	100.16	98.72	97.85	98.77	98.06	98.35	98.17
Fe ₂ O ₃	2.30	2.39	2.45						
FeO*	10.18	10.58	10.86						
La				1.52	0.61	0.55	1.03	0.97	1.24
Ce				2.81	1.13	0.80	1.66	2.00	2.41
Nd				1.81	0.76	0.50	0.78	1.27	1.46
Sm				0.52	0.23	0.14	0.19	0.40	0.41
Eu				0.43	0.34	0.37	0.62	0.39	0.32
Gd				0.26	0.07	0.17	0.37	0.15	0.35
Dy				0.29	0.11	0.21	0.43	0.17	0.42
Er				0.13	0.00	0.09	0.19	0.06	0.25
Yb				0.15	0.05	0.05	0.18	0.10	0.23
Lu				0.03	0.00	0.01	0.03	0.00	0.05
			An	78.2	78.7	76.4	78.0	75.5	76.7
			Ab	14.0	17.0	21.4	18.7	20.6	18.1
			Or	7.8	4.3	2.2	3.3	3.9	5.2

Table A5.3 (continued). Whole rock major (wt.%) and trace (ppm) element analyses

Glomerocryst Plagioclase		Groundmass Plagioclase					
Sample	F14-32G	F15-3M	F14-16M	F14-18M	F26-1AAM	F15-4M	F875-5M
Lithology	Cryst.	Matrix	Matrix	Matrix	Matrix	Matrix	Matrix
SiO ₂	48.56	49.10	52.06	52.08	55.01	56.22	56.76
Al ₂ O ₃	29.68	29.98	29.77	28.56	25.63	25.57	26.08
Fe ₂ O ₃ T	1.72	1.88	1.32	1.43	2.70	2.17	1.51
Fe ₂ O ₃							
FeO							
MgO	0.52	0.64	0.21	0.26	1.00	0.70	0.17
CaO	14.54	14.56	13.72	12.83	9.90	11.21	10.60
Na ₂ O	2.97	2.94	3.38	2.08	4.65	3.42	4.17
K ₂ O	0.57	0.17	0.12	0.42	0.38	0.08	0.34
TiO ₂	0.04	0.05	0.05	0.05	0.06	0.07	0.06
P ₂ O ₅	0.00	0.00	0.00	0.00	0.11	0.05	0.00
MnO	0.01	0.01	0.04	0.05	0.02	0.02	0.02
S	0.06	0.03	0.02	0.02	0.03	0.03	0.02
Ba	126	92	46	85	90	51	92
Cr	21	25	24	27	26	25	26
Zr	0	0	0	13	135	26	0
Sr	287	205	194	232	196	176	198
Rb	18	0	0	8	0	0	5
Y	0	3	4	5	13	7	3
Nb	0	0	0	0	0	5	9
Zn	21	22	36	24	29	23	22
Ni	10	20	21	15	30	21	18
V	4	0	0	0	26	0	2
LOI							
Total	98.67	99.36	100.69	98.78	99.49	99.54	99.73
Fe ₂ O ₃							
FeO*							
La	0.60	2.60	1.46	2.31	4.11	2.28	2.91
Ce	1.10	4.34	3.06	4.20	8.88	4.72	4.41
Nd	0.70	1.98	2.10	2.20	5.47	2.87	2.36
Sm	0.20	0.46	0.65	0.62	1.54	0.81	0.69
Eu	0.51	0.68	0.63	0.72	1.01	0.74	0.77
Gd	0.23	0.58	0.37	0.74	1.69	0.84	0.84
Dy	0.19	0.54	0.31	0.58	1.49	0.87	0.92
Er	0.13	0.29	0.16	0.37	0.77	0.47	0.48
Yb	0.13	0.18	0.15	0.29	0.67	0.45	0.44
Lu	0.02	0.02	0.02	0.05	0.10	0.07	0.06
An	71.6	73.7	69.2	68.5	56.0	66.0	57.7
Ab	25.2	25.4	30.1	28.9	41.7	33.5	40.1
Or	3.2	1.0	0.7	2.6	2.2	0.5	2.2

Appendix A5

Regression Analysis And Pearce Ratio Diagrams

A5.1 Regression Analysis And Error Propagation

Regression Analysis

Regression analysis utilizes the least squares reduced major axis technique as outlined by Davis (1986). The parameters of the regression line ($Y=mX+b$) are :

$$m = \frac{S_Y}{S_X}$$

$$b = \bar{Y} - m\bar{X}$$

where:

m = slope

b = y-intercept.

S_Y = standard deviation of Y values

S_X = standard deviation of X values

$\bar{Y}$ = mean of Y values

$\bar{X}$ = mean of X values

The standard errors of the slope and intercept are:

$$SE_m = m \sqrt{\left(\frac{1-r^2}{n} \right)}$$

$$SE_b = S_Y \sqrt{\left(\frac{1-r^2}{n} \right) \left(1 + \frac{\bar{X}^2}{S_X^2} \right)}$$

where:

SE_m = standard error of the slope

SE_b = standard error of the y-intercept.

r = correlation coefficient

n = number of values

Error propagation

Error propagation for uncertainty associated with geochemical ratios excluding Pearce ratios follows Kretz (1985) using:

$$v=x+y \quad \sigma_v = \sqrt{\sigma_x^2 + \sigma_y^2}$$

$$v=\frac{x}{y} \quad \sigma_v = v \sqrt{(\sigma_x^2 + \sigma_y^2)}$$

$$v=\frac{x}{(x+y)} \quad \sigma_v = \frac{(y^2\sigma_x^2 + x^2\sigma_y^2)}{(x+y)^2}$$

$$v=\frac{x}{(x+y+z)} \quad \sigma_v = \sqrt{\frac{((y+z)^2\sigma_x^2 + x^2(\sigma_y^2 + \sigma_z^2))}{(x+y+z)^2}}$$

where:

v = value obtained from combination of operands x , y and z

σ_v = error associated with v due to analytical uncertainty σ_x , σ_y and

σ_z in x , y and z

Error propagation for uncertainty associated with Pearce ratios follows Stanley and Russel (1988) using:

$$\sigma_R = R \sqrt{\left(\frac{1}{K^2}\right) \left(\sum_{j=1}^n \frac{(c_j e_j \sigma_j)^2}{W_j^2}\right) + \left(\frac{\sigma_z^2}{W_z^2}\right)}$$

$$K = \sum_{j=1}^n e_j c_j$$

where:

σ_R = propagated uncertainty associated with the Pearce ratio

R = the Pearce ratio

n = number of elements in the numerator

c_j , e_j and W_j = the coefficient, concentration in cation proportion, and concentration in wt. % of element j in the numerator

σ_x and W_x = the analytical error and concentration in wt. % of the denominator element x

A5.2 Derivation Of Complex Numerator Pearce Ratios

Complex numerator Pearce ratios are based on the ideal stoichiometric compositions of minerals or solid solution series which are thought to have been involved in magmatic differentiation. The minerals employed in the analysis discussed in the text, the abbreviations used for these minerals, and their stoichiometric compositions, are given in Table A5.1. The most useful numerators are those in which variation can be attributed to a single phase, and in which the element or elements used in the numerator are proportioned such that they reflect the amount of that element in one mole of the phase. To examine forsterite (Mg_2SiO_4) fractionation for example, the appropriate numerator would be $\text{Mg}/2$ since there are two moles of magnesium in one mole of forsterite.

Table A5.1. Minerals and mineral compositions used in complex numerator Pearce ratios.

Mineral	Representative Mineral or Solid Solution	Abbreviation Used In Text	Stoichiometric Composition
Plagioclase	Anorthite	An	$\text{CaAl}_2\text{Si}_2\text{O}_8$
	Albite	Ab	$\text{NaAlSi}_3\text{O}_8$
Alkali feldspar	Alkali feldspar	Kfs and Or	KAlSi_3O_8
Augite	Diopside	Cpx	$\text{CaMgSi}_2\text{O}_6$
	Hedenbergite	Cpx	$\text{CaFeSi}_2\text{O}_6$
Hypersthene	Enstatite	Opx	MgSiO_3
	Ferroenstatite	Opx	FeSiO_3
Olivine	Forsterite	Ol	Mg_2SiO_4
	Fayalite	Ol	Fe_2SiO_4
Fe-Ti Oxides	Magnetite	Mt	FeFe_2O_3
	Ilmenite	Ilm	FeTiO_2

Often several elements could be used to represent a given mineral (e.g., Fe^{+2} or Fe^{+3} for magnetite), and in such cases the chemistry of other potentially fractionating phases should be considered. If magnetite alone is involved in the process, either Fe^{+2} or Fe^{+3} could be used as numerator elements. If pyroxene is also involved, Fe^{+3} is a better choice as numerator element for magnetite since this obviates the need to apportion total Fe^{+2}

between the two phases. In the following derivations, the choice of elements used in Pearce ratio numerators is influenced by the stoichiometry of other phases potentially involved in the differentiation process (i.e., those listed in Table A5.1).

Plagioclase and Alkali Feldspar

Alkali feldspar is the only K-bearing phase likely to have influenced the concentrations of this element in samples used in this study. Consequently, changes in alkali feldspar content can be represented by changes in K content. For similar reasons, Na can be used to represent the albite component of plagioclase. Ca, however, occurs in both clinopyroxene and plagioclase. Consequently, Al is a better choice to represent anorthite even though this requires partitioning of Al between the Al-bearing phases. The total change in Al content reflects the change in the proportions of Al-bearing minerals; in the case of basaltic or gabbroic rocks these are anorthite, albite and alkali feldspar. Consequently:

$$\Delta Al_T = \Delta Al_{An} + \Delta Al_{Ab} + \Delta Al_{Kfs}$$

Since:

$$\Delta Ab = \Delta Al_{Ab} = \Delta Na_{Ab} = \Delta Na_T$$

and:

$$\Delta Kfs = \Delta Al_{Kfs} = \Delta K_{Kfs} = \Delta K_T$$

then:

$$\Delta Al_T = \Delta Al_{An} + \Delta Na_T + \Delta K_T$$

and:

$$\Delta Al_{An} = \Delta Al_T - \Delta Na_T - \Delta K_T$$

But since there are two moles of Al in one mole of anorthite:

$$\Delta An = \Delta Al_{An}/2 = \Delta Al_T/2 - \Delta Na_T/2 - \Delta K_T/2$$

Changes in plagioclase content can then be represented by:

$$\Delta Pl = \Delta Al_{An} + \Delta Al_{Ab} = \Delta Al_T/2 + \Delta Na_T/2 - \Delta K_T/2$$

And changes in the total feldspar content can be represented by:

$$\Delta Feldspar = \Delta Pl + \Delta Kfs = \Delta Al_T/2 + \Delta Na_T/2 + \Delta K_T/2$$

Since one mole of plagioclase consists of x moles of anorthite and $(1-x)$ moles of albite, the equation above is independent of plagioclase composition. This relationship is true for all of the numerators representing solid solutions determined in subsequent derivations. A plot of Ca_{An} versus Na_{Ab} can be used to obtain the relative proportions of albite and anorthite in plagioclase using as numerators:

$$\Delta Ca_{An} = \Delta Al_{An}/2 = \Delta Al_T/2 - \Delta Na_T/2 - \Delta K_T/2$$

$$\Delta Na_{Ab} = \Delta Na_T$$

Clinopyroxene

For the purposes of Pearce ratio analysis, augite can be considered a member of the diopside-hedenbergite solid solution series. Since both endmembers contain stoichiometrically equivalent amounts of Ca, changes in clinopyroxene content can be represented by the change in Ca content if the effect of other Ca-bearing phases are removed. Since plagioclase is the only other Ca-bearing phase likely to have fractionated this gives:

$$\Delta Cpx = \Delta Ca_{Cpx} - \Delta Ca_{An}$$

From the preceding section:

$$\Delta Ca_{An} = \Delta Al_T/2 - \Delta Na_T/2 - \Delta K_T/2$$

and therefore:

$$\Delta Cpx = \Delta Ca_T - \Delta Al_T/2 + \Delta Na_T/2 + \Delta K_T/2$$

Olivine and Orthopyroxene

Olivine and orthopyroxene found in the rocks of the LTBS form parts of the forsterite-fayalite and enstatite-ferrosilite solid solutions respectively. If either phase is involved in the differentiation process to the exclusion of the other, changes in the olivine or orthopyroxene content can be monitored by changes in Mg and Fe^{+2} content. This requires partitioning of these elements between olivine or orthopyroxene and any other

Mg-Fe⁺² bearing minerals which may be participating. Thus if orthopyroxene is the only Fe-Mg phase involved it can be represented by:

$$\Delta\text{Opx} = \Delta(\text{Mg} + \text{Fe}^{+2})_{\text{Opx}} = \Delta(\text{Mg} + \text{Fe}^{+2})_{\text{T}}$$

while olivine, which contains two moles of (Mg+Fe⁺²), can be represented by:

$$\Delta\text{Ol} = \Delta(\text{Mg} + \text{Fe}^{+2})_{\text{Ol}}/2 = \Delta(\text{Mg} + \text{Fe}^{+2})_{\text{T}}/2$$

The effect of clinopyroxene can be removed by using Ca to represent the magnesium and ferrous iron in this phase:

$$\Delta(\text{Mg} + \text{Fe}^{+2})_{\text{Cpx}} = \Delta\text{Ca}_{\text{Cpx}} = \Delta\text{Ca}_{\text{T}} - \Delta\text{Al}_{\text{T}}/2 + \Delta\text{Na}_{\text{T}}/2 + \Delta\text{K}_{\text{T}}/2$$

Therefore:

$$\begin{aligned}\Delta\text{Opx} &= \Delta(\text{Mg} + \text{Fe}^{+2})_{\text{Opx}} = \Delta(\text{Mg} + \text{Fe}^{+2})_{\text{T}} - \Delta(\text{Mg} + \text{Fe}^{+2})_{\text{Cpx}} \\ &= \Delta(\text{Mg} + \text{Fe}^{+2})_{\text{T}} - \Delta\text{Ca}_{\text{T}} + \Delta\text{Al}_{\text{T}}/2 - \Delta\text{Na}_{\text{T}}/2 - \Delta\text{K}_{\text{T}}/2\end{aligned}$$

and for olivine:

$$\Delta\text{Ol} = \Delta(\text{Mg} + \text{Fe}^{+2})_{\text{T}}/2 - \Delta\text{Ca}_{\text{T}} + \Delta\text{Al}_{\text{T}}/2 - \Delta\text{Na}_{\text{T}}/2 - \Delta\text{K}_{\text{T}}/2$$

Similar compensations can be made for other phases such as magnetite or ilmenite, which contain ferrous iron. If olivine and orthopyroxene occur together, their individual contributions to compositional change can be evaluated by examining the change in Si content relative to the change in (Mg+Fe⁺²) content. The change in (Mg+Fe⁺²) due to changes in olivine and orthopyroxene when clinopyroxene is also contributing to compositional variation is:

$$\Delta(\text{Mg} + \text{Fe}^{+2})_{\text{Ol+Opx}} = \Delta(\text{Mg} + \text{Fe}^{+2})_{\text{T}} - \Delta(\text{Mg} + \text{Fe}^{+2})_{\text{Cpx}}$$

and again further corrections can be applied as necessary.

The change in Si content due to change in the olivine+orthopyroxene content when clinopyroxene and feldspar are also involved is:

$$\Delta\text{Si}_{\text{Ol+Opx}} = \Delta\text{Si}_{\text{T}} - \Delta\text{Si}_{\text{Cpx}} - \Delta\text{Si}_{\text{Pl}} - \Delta\text{Si}_{\text{Kfs}}$$

Using the representations of the various minerals derived earlier, and making the appropriate corrections for the number of moles of Si in each gives:

$$\Delta Si_{Kfs} = 3\Delta K_T$$

$$\Delta Si_{Pl} = \Delta Al_T + 2\Delta Na_T - \Delta K_T$$

$$\Delta Si_{Cpx} = 2\Delta Ca_T - \Delta Al_T + \Delta Na_T + \Delta K_T$$

The slope in a plot of $(Mg+Fe^{+2})_{Ol+Opx}$ versus Si_{Ol+Opx} is then:

$$\frac{\Delta(Mg+Fe^{+2})_{Ol+Opx}}{\Delta Si_{Ol+Opx}}$$

Because of the difference in Si content of olivine and orthopyroxene, the slope will vary between 2 (100% olivine) and 1 (100% orthopyroxene) and can thus be used to establish the relative proportions of the two phases.

The change in Si among samples can also be used to monitor changes in clinopyroxene+orthopyroxene+olivine when all are involved in the process simultaneously:

$$\Delta(Cpx+Opx+Ol) = \Delta Cpx + \Delta Opx + \Delta Ol$$

Orthopyroxene and olivine can be expressed in terms of Si as:

$$\Delta Opx = \Delta Si_{Opx}$$

$$\Delta Ol = \Delta Si_{Ol}$$

Because clinopyroxene contains two moles of Si:

$$\Delta Cpx = \Delta Si_{Cpx}/2$$

But:

$$\Delta Cpx = \Delta Ca_{Cpx}$$

and therefore:

$$\Delta Cpx = \Delta Si_{Cpx} - \Delta Ca_{Cpx}$$

Then:

$$\Delta(Cpx+Opx+Ol) = \Delta Si_{Cpx} - \Delta Ca_{Cpx} + \Delta Si_{Opx} - \Delta Si_{Ol}$$

If these are the only phase involved:

$$\Delta Si_T = \Delta Si_{Cpx} + \Delta Si_{Opx} + \Delta Si_{Ol}$$

$$\Delta Ca_{Cpx} = \Delta Ca_T$$

and:

$$\Delta(Cpx + Opx + Ol) = \Delta Si_T - \Delta Ca_T$$

If feldspar is also involved:

$$\Delta(Cpx + Opx + Ol) = \Delta Si_T - \Delta Ca_{Cpx} - \Delta Si_{Pl} - \Delta Si_{Kfs}$$

where:

$$\Delta Ca_{Cpx} = \Delta Ca_T - \Delta Al_T/2 + \Delta Na_T/2 + \Delta K_T/2$$

$$\Delta Si_{Pl} = \Delta Al_T + 2\Delta Na_T - \Delta K_T$$

$$\Delta Si_{Kfs} = 3\Delta K_T$$

Fe-Ti Oxides

Magnetite and ilmenite are both endmembers in solid solution series, both of which involve changes in the proportions of Fe^{+2} , Ti and Fe^{+3} . If these phases were treated as solid solutions and represented by Ti and Fe^{+3} , there would be no way to partition these elements between the two. This problem is further complicated by exsolution phenomenon. Therefore they are treated as separate pure minerals in the Pearce ratio analysis discussed in the text. Changes in the amount of magnetite and ilmenite can thus be represented by the changes in Fe^{+3} and Ti contents respectively:

$$\Delta Mt = \Delta Fe^{+3}/2$$

and:

$$\Delta Ilm = \Delta Ti$$

Appendix A6

Image Analysis

A6.1 Image Analysis Procedure

Image data was acquired from slabs cut from samples obtained from the four sills examined in detail at Howse Lake. Slab thickness was determined by the approximate diameter of the predominant glomerocryst size in the sample, except for the core phase in which slabs were cut at intervals of approximately 3 cm. For most samples, this produced 5-8 slabs which ranged in area from 50 cm² to 250 cm². The slabs were lightly oiled to enhance contrast, and digital images were then acquired using a Panasonic video camera. The images were processed using Bioscan Corporation's Optimus image analysis software. The contrast between the white to pale green glomerocrysts and the dark groundmass was used to isolate the aggregates. Contrast thresholds and scale were adjusted for each sample, and a minimum area requirement was set at 5 mm² to remove any groundmass material which was light coloured enough to exceed the contrast threshold. This principally applied to altered orthopyroxene grains which occasionally produced light coloured areas in the images. Glomerocryst size and irregularity were obtained directly from the image data using algorithms included in the software. Glomerocrysts which were intersected by slab edges were excluded from these measurements. Frequency and proportional area were calculated from the number and size of glomerocrysts within a given region of interest (ROI) on the slabs. The ROI was the largest possible rectangular area within the borders of the slab.

A6.2 Analytical Uncertainty

There were several potential sources of uncertainty in the measurements obtained from the slabs. Because threshold limits were determined by visual inspection, this parameter varied from slab to slab. Glomerocryst alteration affected colour, and intense

alteration occasionally led to dark green glomerocrysts which could not be resolved from the groundmass. This usually occurred in aphanitic and fine grained samples from the chilled margins. Because the number of glomerocrysts in these samples were small, the use of tracings of the slabs, rather than the slabs themselves, was viable. This ensured relatively accurate representation of the glomerocrysts. In other samples intense alteration was confined to small areas within a given glomerocryst. Where these areas occurred within the glomerocryst, they could be excluded from the measurements by manual de-selection. Where altered areas were in contact with groundmass material, however, they could not be removed. This was due to the lack of contrast, such that in the digital images the altered zones were continuous with the groundmass in terms of brightness. Consequently, alteration may have introduced small errors to the measurements. Other small errors may have arisen through the occasional exclusion of very small glomerocrysts as a result of the size threshold.

The sources of potential error described above are negligible in comparison to variations in the measured parameters among slabs. Since the data were averaged among slabs to provide sample values, the variations in slab measurements provide the best estimates of analytical uncertainty. Thus variations among slab measurements were used to estimate the coefficients of variation for the glomerocryst parameters (Table A6.1). To provide usable estimates of analytical uncertainty, samples were subdivided according to glomerocryst content into three groups. Thus the coefficients of variation are reported for samples with <5 area % glomerocrysts, 5-20 area %, and >20 area %. Coefficients of variation for all samples combined are also given.

A6.3. Explanation of Table A6.2.

Glomerocryst data are provided in Table A6.2. Average values derived from slab measurements are reported for each sample. The locations of the four sills are indicated in Figure A1.3. Sample position is denoted by height in metres above the lower contact of the sill.

Table A6.1. Estimated analytical uncertainty for glomerocryst measurements, as determined from the average, standard deviation and coefficient of variation for individual slabs as determined from sill A samples. Samples were grouped according to glomerocryst content (by area) as indicated.

Sample Grouping	Average	Standard Deviation	Coefficient of Variation
Proportional Area			
<5%	1.02	0.67	0.66
5-20%	10.52	3.24	0.31
>20%	37.95	4.23	0.11
All	15.47	2.76	0.18
Glomerocryst Size			
<5%	21.27	11.27	0.53
5-20%	37.13	11.51	0.31
>20%	134.61	29.61	0.22
All	71.56	17.89	0.25
Glomerocryst Frequency			
<5%	717.42	344.36	0.48
5-20%	4184.06	920.49	0.22
>20%	4577.78	595.11	0.13
All	3048.12	487.70	0.16
Glomerocryst Irregularity			
<5%	19.76	7.31	0.37
5-20%	21.41	4.50	0.21
>20%	24.48	3.67	0.15
All	22.07	4.41	0.20

Table A6.2. Image analysis data.

Sample	Height	% Area Glomerocrysts	Average Glomerocryst Size (mm ²)	Average Glomerocryst Frequency (m ²)	Average Glomerocryst Irregularity
Sill A					
F90-S6	0.0	1.16	21.3	532	18.9
F90-S5	0.2	8.40	48.1	1830	19.1
F90-A	0.4	1.81	37.9	187	15.7
F90-B	1.9	10.35	29.5	3531	22.2
F90-C	2.8	1.71	19.1	893	20.0
F90-D	4.7	1.13	18.3	620	20.5
F90-F	7.7	0.90	11.1	855	22.1
F90-H	11.3	1.79	10.2	1737	22.9
F90-I	16.9	10.93	55.5	2277	22.1

Sample	Height	% Area Glomerocrysts	Average Glomerocryst Size (mm ²)	Average Glomerocryst Frequency (m ⁻²)	Average Glomerocryst Irregularity
F90-L	20.9	14.24	35.6	4113	23.2
F90-N	26.9	24.81	58.4	4281	22.2
F90-P	30.1	35.11	66.2	5021	24.6
F90-R	34.3	42.02	103.4	4087	24.5
F90-S	41.3	39.61	99.3	4211	24.2
F90-57	52.6	39.02	100.0	3945	25.0
F90-58	58.6	44.26	97.7	4657	24.7
F90-58B	63.5	47.23	643.6	824	30.9
F90-59	70.1	43.90	83.6	5507	23.8
F90-59A	83.9	39.60	69.7	6317	23.0
F90-82D	98.5	31.42	46.8	6927	22.0
F90-83B	101.7	17.40	19.1	9169	20.5
F90-83C	106.2	4.22	43.8	772	26.9
F90-82B	109.1	2.15	12.4	524	14.6
F90-82A	111.1	0.44	20.1	338	16.2
F90-82E	117.5				
Sill B					
F90-70A	0.0	1.00	12.0	181	17.0
F13-19A	7.3	0.00	0.0	0	
F871-7	22.7	0.78	8.4	228	14.6
F871-6	31.7	1.17	13.0	379	20.7
F16-21	42.4	4.13	23.1	164	16.5
F871-5	58.0	16.30	38.1	1804	19.0
F871-4	77.9	22.90	114.0	3766	22.8
F16-20	128.7	37.80	201.0	2049	23.9
F871-3	161.8	34.60	138.6	3484	21.7
F871-2	169.0	39.40	146.1	4776	27.3
F871-2A	174.7	28.30	107.1	4559	24.0
F90-81B	185.1	0.00	0.0	0	
F15-3	187.4	0.00	0.0	0	
F15-4	192.3	51.01	365.4	1730	31.7
F90-81A	201.4	53.41	538.5	1410	36.5
F16-18	204.1	31.60	13.8	3130	25.6
F16-17	220.7	23.40	10.7	2740	17.7
F16-16	233.2	7.20	27.3	2130	18.5
F16-15	238.6	17.70	99.3	2537	22.6
F16-13	258.0	6.30	42.3	4640	21.0
F16-11C	263.1	23.40	81.3	5233	23.3
F16-10	293.7	2.19	34.8	3660	22.2
F16-9	297.7	1.23	32.4	4101	17.1
F16-8	306.2	7.61	64.1	2913	17.3
F90-70	311.8	16.70	64.2	2990	21.8

Sample	Height	% Area Glomerocrysts	Average Glomerocryst Size (mm ²)	Average Glomerocryst Frequency (m ⁻³)	Average Glomerocryst Irregularity
F16-7	316.7	2.19	16.1	1440	18.1
F16-6	321.3	1.07	12.7	738	16.7
F16-4	332.1	2.03	14.3	1297	13.8
F16-2	338.5	2.13	11.9	1051	19.5
F90-67	348.0	1.48	16.4	1898	16.8
F90-65	367.8	1.17	14.0	1037	14.4
F90-64	373.3	0.00	0.0	0	
F90-63	381.0	0.00	0.0	0	
F90-61	387.0	0.00	0.0	0	
F90-60B	400.0	0.00	0.0	0	
SH C					
F90-78A	0.0	0.00	0.0	0	
SH-2	2.6	0.00	0.0	0	
SH-3	5.3	0.00	0.0	0	
SH-4	7.9	0.00	0.0	0	
SH-5	10.6	0.00	0.0	0	
SH-6	13.2	0.00	0.0	0	
SH-8	18.5	0.00	0.0	0	
SH-9	21.0	0.00	0.0	0	
SH-10	23.7	0.00	0.0	0	
SH-11	26.3	0.61	30.7	230	24.0
SH-12	29.1	3.60	15.0	2391	21.0
SH-13	31.8	2.17	10.2	2149	28.0
SH-15	37.3	5.81	11.6	5040	20.0
SH-16	40.1	4.46	11.7	3834	20.0
SH-17	42.7	4.83	10.4	5033	20.0
SH-19	51.4	2.43	8.5	3838	21.0
SH-21	58.7	4.23	10.8	3924	21.0
SH-22	61.1	2.84	7.3	3853	19.0
SH-23	63.6	2.84	9.6	3066	20.0
90-79	69.1	8.28	13.2	6312	23.0
90-79A	74.2	7.26	14.5	4166	24.0
90-80D	83.4	7.94	14.9	5066	24.0
90-80C	89.7	6.89	15.3	3622	21.0
90-80B	92.4	3.78	14.2	2530	21.0
90-80A	97.1	4.12	16.9	3854	16.0
90-80	104.5	2.83	15.6	328	16.0
SH-24	131.4	0.94	24.2	409	16.0
SH-25	133.7	0.66	39.0	227	15.0
SH-27	138.2	0.39	12.7	351	19.0
SH-31	147.4	0.25	9.0	260	15.0
SH-33	152.0	0.22	7.4	299	17.0

Sample	Height	% Area Glomerocrysts	Average Glomerocryst Size (mm ²)	Average Glomerocryst Frequency (m ⁻²)	Average Glomerocryst Irregularity
SH-34	154.2	0.21	10.0	203	17.0
SH-35	156.5	0.23	13.5	166	15.0
SH-37	161.1	0.00	0.0	0	
SH-39	165.6	0.00	0.0	0	
SH-41	170.1	0.19	7.0	314	16.0
876-1	180.8	0.00	0.0	0	
876-4	184.2	0.00	0.0	0	
876-8	196.0	0.00	0.0	0	
876-15	212.4	0.00	0.0	0	
876-23	222.1	0.00	0.0	0	
Sill D					
F27-1	11.6	2.89	23.1	1252	14.3
F27-2	13.9	3.90	17.1	1401	14.2
F27-4	31.2	2.60	33.6	1127	15.7
F27-6	43.5	2.85	19.3	1774	16.4
F27-9	49.1	1.61	26.2	698	14.6
F27-10	59.1	16.70	66.9	5509	16.2
F27-11	68.0	6.63	27.9	3207	16.3
F27-12	88.7	3.38	12.4	2281	14.7
F27-14	93.7	3.35	18.5	2744	15.7
F27-15	95.3	19.00	81.3	5574	19.5
F27-16	116.0	27.83	107.7	4215	24.1
F27-17	135.8	29.36	163.2	4143	22.8
F27-19	141.3	27.91	123.9	3680	23.9
F27-20	163.7	44.20	557.1	1810	32.9
F27-21	188.9	31.50	136.2	3467	23.4
F27-22	209.0	35.63	160.8	2353	30.1
F27-23	216.1	36.09	118.6	3164	31.9
F27-24	227.0	29.65	95.4	4223	25.6
F27-27	265.9	30.09	119.6	3813	21.8
F27-29	280.0	28.34	82.8	5487	23.5
F27-30	291.9	10.78	46.9	4327	17.4
F27-31	308.2	5.17	28.5	2241	16.6
F27-32	319.6	2.55	35.7	1104	19.6
F27-34	331.2	16.47	85.6	3008	20.8
F27-36	351.2	3.79	30.6	1642	21.6
F27-37	368.1	2.67	16.9	1580	14.1
F27-38	390.7	0.95	9.6	410	14.7
F27-39	402.3	0.00	0.0	0	

Appendix A7

Rb-Sr And Sm-Nd Isotopic Analysis

Isotopic analyses were performed by Dr. John Blenkinsop, Department of Earth Sciences, Carleton University, and by Reginald Thériault, Geochronology Laboratory, Geological Survey of Canada. Analytical procedures in the two laboratories, as described by Morisset (1992) and Thériault (1990) are summarized below. Analytical results are presented in Chapter 9 (Tables 9.1 and 9.2).

A7.1 Analytical Procedure at Carleton University

Approximately 0.1 g of whole rock powder was accurately weighed then dissolved with concentrated HF. The solution was evaporated to dryness on a 110°C hotplate, then taken up in approximately 1 ml of 2.5 M HCl. The sample was then quantitatively transferred with ~1 ml of 2.5 M HCl to a centrifuge tube. After centrifugation, the sample solution was transferred to a weighed Teflon bottle, which was then re-weighed. Approximately half of the solution was transferred to a clean centrifuge tube. The Teflon bottle was again weighed, thus providing the weights of both sample fractions. One fraction was spiked with Sm, Nd, and Sr spike solutions. The spiked fraction was evaporated to dryness on a hotplate, then taken up in ~ 1 ml 2.5 M HCl. This promoted isotopic equilibration between the sample and spike solutions. Both sample fractions were again centrifuged prior to transfer to ion exchange columns.

Isotopic analysis was performed on a Finnigan-MAT 261 multi-collector solid source mass spectrometer. The machine is a 90° deflection magnetic sector instrument with a 23 cm curvature radius, but extended geometry results in an effective curvature radius of 46 cm. A Hewlett-Packard 9845 computer controls the operation of the instrument. All elements were loaded with 1µl of nanopure water onto the evaporating filament of a Re double filament assembly. The filament was then heated to about 1.0 A in air until dry. Sr

and Nd samples were taken to glowing (~ 2 A) for a fraction of a second to ensure complete dryness without causing oxidation.

Preheating and running conditions are given by Morriset (1992). 2σ Precision of the $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios is about $\pm 0.003\%$ and 0.004% , respectively, based on replicate analyses of rock and chemical standards. The precision in concentration determinations is about 0.5% for Sr, Sm and Nd, and about 1% for Rb. The NBS 987 Sr standard gives a ratio of 0.71025 ± 0.00002 ; the La Jolla Nd standard gives a $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.51186 ± 0.00002 . Concentrations of Sm and Nd in standard BCR-1 were 6.55 ppm and 27.72 ppm, respectively, and $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios obtained for this standard were 0.70501 ± 0.00003 and 0.51266 ± 0.00002 , respectively.

A7.1 Analytical Procedure at the Geological Survey of Canada

Approximately 0.2 g of whole rock powder was accurately weighed in Teflon vessels and the sample was moistened with ~ 0.5 ml of 16 M HNO_3 . Samples were then spiked with a solution enriched in ^{149}Sm and ^{148}Nd and with a known isotopic composition. This allowed determination of isotopic composition and concentration from the same sample solution. Samples were dissolved in approximately 3 ml of concentrated HF in Teflon bombs, and heating them on a hotplate for 12 hours at 150°C . The dried residues were taken up in ~ 3 ml Hf and ~ 0.5 ml HNO_3 , and the bombs were then placed in pressure resistant steel jackets. The samples were then decomposed under pressure at $\sim 200^\circ\text{C}$ for one week. Samples were dried and taken up in 1 ml of 2.5 M HCl, then centrifuged. The resultant solutions were transferred to ion exchange columns and the REE, and then Sm and Nd, were eluted.

Isotopic analyses were performed on a Finnigan MAT 261 solid source thermal ionization mass spectrometer equipped with 6 variable Faraday cups and one fixed collector. Samples were loaded in 2 μl of H_3PO_4 on double Re filament assemblies. Operating conditions are given by Thériault (1990). Isotopic ratios were corrected for mass fractionation by normalization to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. Precision of $^{143}\text{Nd}/^{144}\text{Nd}$ ratios is

estimated at better than 2.5×10^{-5} . Analyses of the La Jolla Nd standard yielded a mean $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.511832 ± 0.000028 .