SEDIMENTOLOGY, TECTONIC CONTROL AND RESOURCE POTENTIAL OF THE UPPER DEVONIAN - LOWER CARBONIFEROUS HORTON GROUP, CAPE BRETON ISLAND, NOVA SCOTIA

By

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TITLE: Sedimentology, tectonic control and resource potential of the Upper Devonian-Lower Carboniferous Horton Group, Cape Breton Island, Nova Scotia

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ABSTRACT

The Famennian?/Tournaisian Horton Group of Cape Breton Island, Nova Scotia was deposited in two adjacent fault-bounded asymmetric sub-basins. They are the western and northern Cape Breton sub-basins, which were part of a large intracontinental distensional rift system. This system, analogous to the Fundy Basin Rift of Belt (1968), lay at approximately 10-15° S. paleolatitude in a warm, arid climate. The half-graben sub-basins had opposed asymmetry, were approximately 100 x 50 km in size, and were separated by a narrow zone of elevated basement. These features are all common to the adjacent structural segments of known rifts. In both sub-basins, but best illustrated in western Cape Breton, ten facies assemblages can be defined, arranged into four depositional systems within the Craignish, Strathlorne and Ainslie Formations of the Horton Group.

The basal Craignish Formation generally overlies Acadian basement but locally gradationally overlies Fisset Brook Formation volcanics, attributed to extrusion in an intracontinental extensional setting. The Craignish consists of three facies assemblages dominated by distal braided stream (C1, C3) and basin-central mudflat/playa (C2) deposits. These can be combined into two depositional systems attributed to 1.) a broad pre-rift sag basin with longitudinal sediment dispersal to the northeast (C3 depositional system), and 2.) the initial syn-rift formation of two half-graben sub-basins with sediment dispersal away from fault-bounded margins (C1/C2 depositional system). The intensity of fault-bounded subsidence increased and became more localized during deposition of the latter system.

The overlying Strathlorne Formation consists of four facies assemblages which include the components of lacustrine depositional systems localized in the two fault-bounded sub-basins. These parts include, from sub-basin margin to centre, fan-delta (S3), shoreline mudflat (S4), prograding delta/shoreline (S2) and open lacustrine (S1) depositional settings. The sedimentary style is characterized by vertical stacking of coarsening-upward, shallowing-upward sequences which are attributed to individual or grouped subsidence events at the controlling footwall scarps of the sub-basins. Strathlorne deposition accompanied the most active phase of fault-bounded subsidence during Horton time. Sediment dispersal was away from the margins and toward the deep axis of each sub-basin.

The upper Ainslie Formation consists of three facies assemblages which depict proximal to distal variation of alluvial fan (A1) to low sinuosity fluvial (A2) to high sinuosity

fluvial (A3) depositional settings. The alluvial fan sediments accumulated near fault scarps and are characterized by thick coarsening-upward sequences attributed to direct control by fault subsidence. In general, the Ainslie Formation is interpreted to represent a phase of waning tectonic activity and sub-basin filling by sediments derived from the high-relief fault-bounded margins. Sediment dispersal was transverse near sub-basin margins and may have been longitudinal toward the northeast in sub-basin centres. The Horton Group of the study area is overlain by the Macumber Formation of the Windsor Group, which represents a rapid marine transgression into the continental rift.

There is significant potential for metallic deposits hosted in Horton sediments. Specifically, paleoplacer gold may occur in braided stream deposits of the Craignish Formation in the Baddeck and Cheticamp areas of the western Cape Breton sub-basin, and Kupferschiefer-type copper deposits may occur at the Horton-Windsor boundary in the Baddeck and Whycocomagh areas of the western Cape Breton sub-basin. Open lacustrine mudstones of the S1 facies assemblage are thermally mature and have significant hydrocarbon (especially oil) source rock potential, especially in the Baddeck area of western Cape Breton and the exposed part of the northern Cape Breton sub-basin. Fan-delta (S3), shoreline (S2) and fluvial (A3) deposits of the Strathlorne and Ainslie Formations have significant reservoir potential, especially in the Mabou/Lake Ainslie area, and minor amounts of oil has been recovered in the western Cape Breton sub-basin during 120 years of exploration. However, past exploration efforts may not have fully evaluated the moderate hydrocarbon potential of the possible trap types.

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CHAPTER 1
INTRODUCTION AND PHILOSOPHICAL APPROACH

TECTONO-SEDIMENTARY ANALYSIS OF BASIN-FILL SEQUENCES

The conclusions I have reached in this thesis are based on the treatment of the Lower Carboniferous Horton Group of Cape Breton Island as a sedimentary succession of diachronous deposits which filled tectonically-active sub-basins. The premise of the study is that the style and spatial arrangement of facies reflect the nature and distribution of sedimentary environments, which in turn result from local and regional tectonic activity before and during deposition. These facies can be used to interpret the sedimentary and structural evolution of the sub-basins through time. In a fault-bounded basin, subject to complex later structural effects (as in this case), this approach is regarded as the most successful avenue for study.

The term "basin analysis" is commonly applied to studies that address the controls and processes involved in the structural formation of a basin. The same term has been applied by sedimentary geologists to studies of the sediments that fill the basin. The term "tectono-sedimentary analysis of basin-fill sequences" combines the two aspects of basinal study and focusses attention on the various controls which determine the sedimentary characteristics of the basin-fill deposits in a feedback process of interpretation.

This study deals with a considerable, though variable, stratigraphic thickness present over a large area where extensive outcrops are uncommon. Therefore much of the basic information is derived from sedimentological observations on isolated exposures and is synthesized into a stratigraphic framework within which further studies can be organized. Stratigraphic and sedimentological analysis at two scales is involved: a) the lithofacies observed in outcrops and drillholes which reflect local depositional environments, b) the depositional systems observed through more regional correlation of facies assemblages, which reflect interaction of fault-related subsidence and sediment supply. The result is a four-dimensional tectono-stratigraphic sedimentation model for the evolution of the basin.

Certain geological concepts, which bear on the contents of this thesis, are introduced below and discussed more fully throughout the thesis.

APPALACHIAN TECTONIC CONCEPTS

The Paleozoic history of the Appalachian Orogen involved the assembly of exotic terranes by orthogonal and oblique collisions. Deformational, intrusive and metamorphic effects, representing collisional episodes, peaked during the mid to late Ordovician Taconian Orogeny, the mid to late Devonian Acadian Orogeny, and the mid to late Carboniferous Alleghenian Orogeny (Williams, 1979; Schenk, 1978; Williams and Hatcher, 1982; Kennedy et al., 1982). The interorogenic periods, which encompass the plate motions between and leading to these collisions can be deciphered from the poorly-studied sedimentary sequences. The Horton Group was deposited during Late Devonian/Early Carboniferous time (Bell, 1960; Utting et al., in press) immediately after the Acadian Orogeny, which affected most of the Orogen, but before the more localized Alleghenian Orogeny. It is thus part of an interorogenic sequence that occurs in one part of the Orogen. Understanding the tectonic controls on deposition can provide important information regarding plate interactions between collisional episodes which may be applicable elsewhere. These concepts are discussed more fully in Chapters 2 and 5.

CONCEPTS OF BASIN ORIGIN

As discussed later, the sub-basins which received Horton Group sediments are interpreted as tectonic in origin, characterized by fault-controlled extension (Belt, 1968 and Quinlan, 1988). The primary control on such a basin is an extensional stress field and deformation is commonly manifested by vertical subsidence on fault planes. However, overprinting by the pervasive compressional regime of the Late Carboniferous Alleghenian Orogeny hinders the structural interpretation of the Horton basins. In this study I have attempted to reconstruct the tectonic evolution of the sub-basins from the results of these events in the sedimentary record. For example, the presence of coarse-grained immature facies, grain size and thickness trends, lateral facies changes, and paleocurrent data all relate to the location and trend of original basin margin faults. In addition to constraining the process of sub-basin origin on a local scale, these fundamental relationships between tectonics and sedimentation also help define the regional tectonic framework. The application of these ideas to study of the Horton Group is discussed in Chapters 4 and 5.

FACIES CONCEPTS

Viewing the Horton Group as a three-dimensional package of facies assemblages, as described in Chapter 3, allows interpretations of paleogeography and depositional environments. The facies concept, as an interpretive tool, is simply a way of organizing descriptive material so that it is more interpretable in light of modern sedimentological concepts. The functions of a facies model espoused by Walker (1984) are equally valid for a local example of basin-fill analysis or for a sedimentary ore-deposit model.

I hope to make some progress toward organizing the sedimentary characteristics of the Horton Group so that this information can be used as a) a norm for comparison with other non-marine sequences, b) a guide to future observations in the Horton and other similar sequences, c) a predictor of paleogeography and facies distribution, and therefore of potential for petroleum and mineral deposits, and d) a basis for tectono-stratigraphic interpretation of fault-bounded basins in the Appalachians. I am attempting to build, not a facies model in the conventional sense, but a tectono-sedimentary basin-fill analogue with the aim of understanding the organization and significance of the depositional units which comprise the Horton Group on Cape Breton Island. An analogue of this type forms the framework for future research because it involves a systematic approach to understanding the basic controls on the sedimentary characteristics of a unit. These ideas are more fully developed in Chapters 4, 5 and 6.

RESOURCE ASSESSMENT CONCEPTS

While description of facies and facies models aids exploration in relatively well-studied basins, a broader analysis of the depositional framework is useful in basins that are less well-known. Such an analysis can indicate the kind and quality of deposit which might be present and set the stage for more systematic exploration approaches as knowledge increases. Important questions might include whether the basin is conducive to commercial deposits, what kind of deposits might be present and what geologic factors exert control on their distribution and predictability. By identifying the characteristics of the tectono-sedimentary basin-fill model it may be possible to construct local ore deposit and hydrocarbon deposit models. The application of these concepts to the Horton Group is discussed in Chapter 6.

GENERAL PURPOSE OF THIS STUDY

The sedimentology and resource potential of non-marine sequences in fault-bounded basins have not received adequate attention in Canada. Horton Group sediments have significant and long-recognized mineral and petroleum potential both onshore and offshore, but have not been investigated in detail using sedimentological data as a predictive tool. This study was undertaken to establish the complex inter-relationships between a) tectonic setting, b) syndepositional structural movement, c) types and characteristics of facies deposited, d) spatial distribution and geometry of facies assemblages and, e) resource potential of this sequence (Fig. 1). It is an attempt to erect a predictive framework within which further studies can be organized. The combination of data from outcrops, drillholes and paleontological samples was a planned approach to maximize the information and further demonstrate the value of integrated studies of a unit from which data are sparse.

SPECIFIC OBJECTIVES OF THIS STUDY

The specific objectives of this study fall into three categories, all closely inter-related, as follows:

a) Structural/Stratigraphic - to delineate the geometry and structural style of the basins and the organization of the enclosed sediment, then decipher the geological history of Horton deposition and its relation to the regional tectonic setting.

b) Sedimentologic - to describe the facies assemblages present, the predictability of their distribution and geometry, and the effect of syndepositional tectonic movements in order to constrain the interpretation of depositional environments.

c) Resource Potential - to investigate the ways in which the structural, stratigraphic and sedimentologic characteristics of the depositional model can be used as predictive tools in exploration of this and similar units.

STUDY AREA AND DATA BASE

Field work for this project was conducted in 1987 and 1988 on selected outcrops in parts of western and northern Cape Breton Island covering an area of approximately 10 000 km^2. This area is located between latitude 45°30'N to 47°15'N and longitude 60°15'W to 61°30'W (Fig. 2). Parts of many 1:50 000 topographic sheets are included. Road access is by the Trans Canada Highway 105, Highway 19, the Cabot Trail and numerous secondary

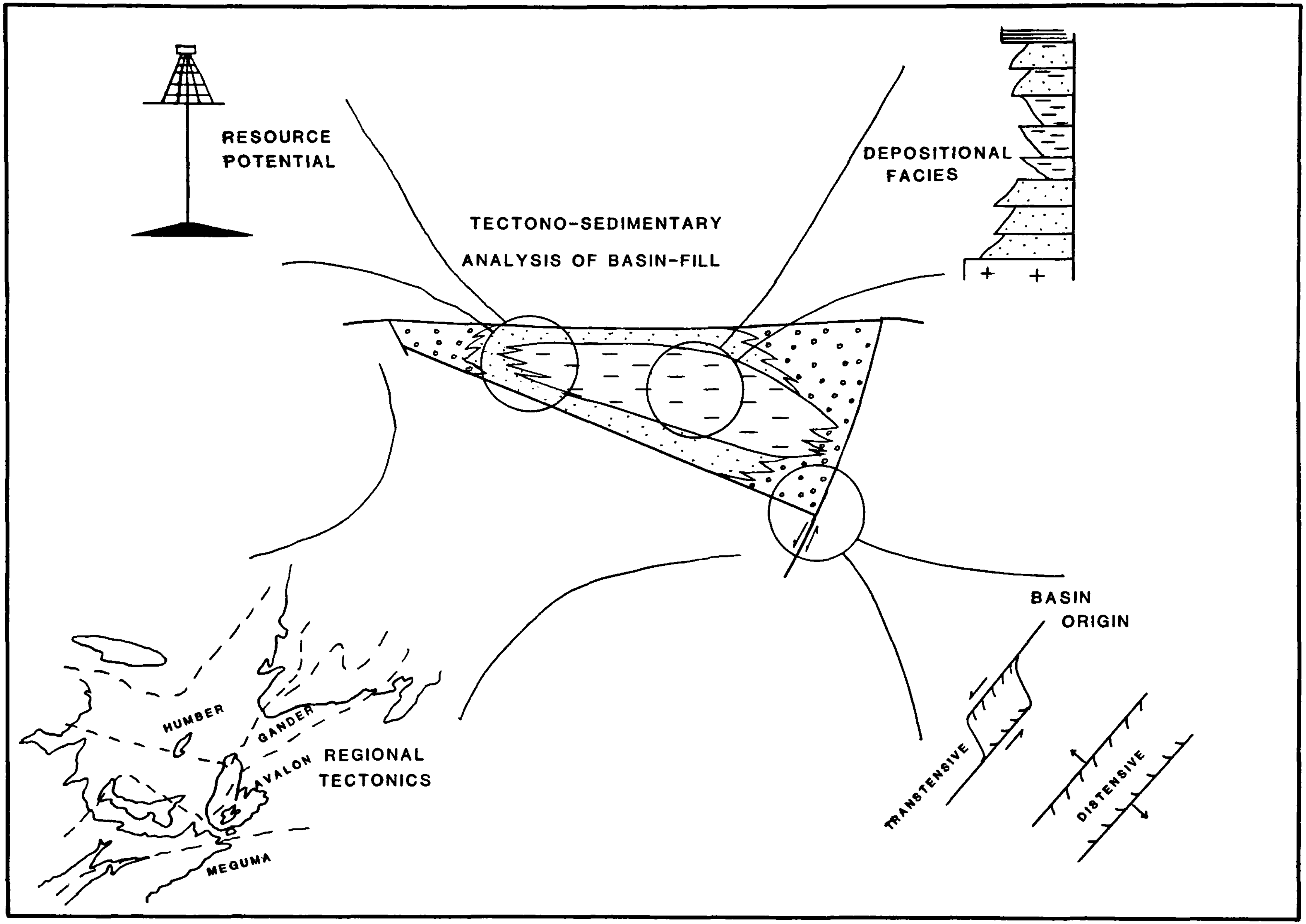

Figure 1. Relationship between various conceptual elements of tectono-sedimentary analysis of basin-fill.

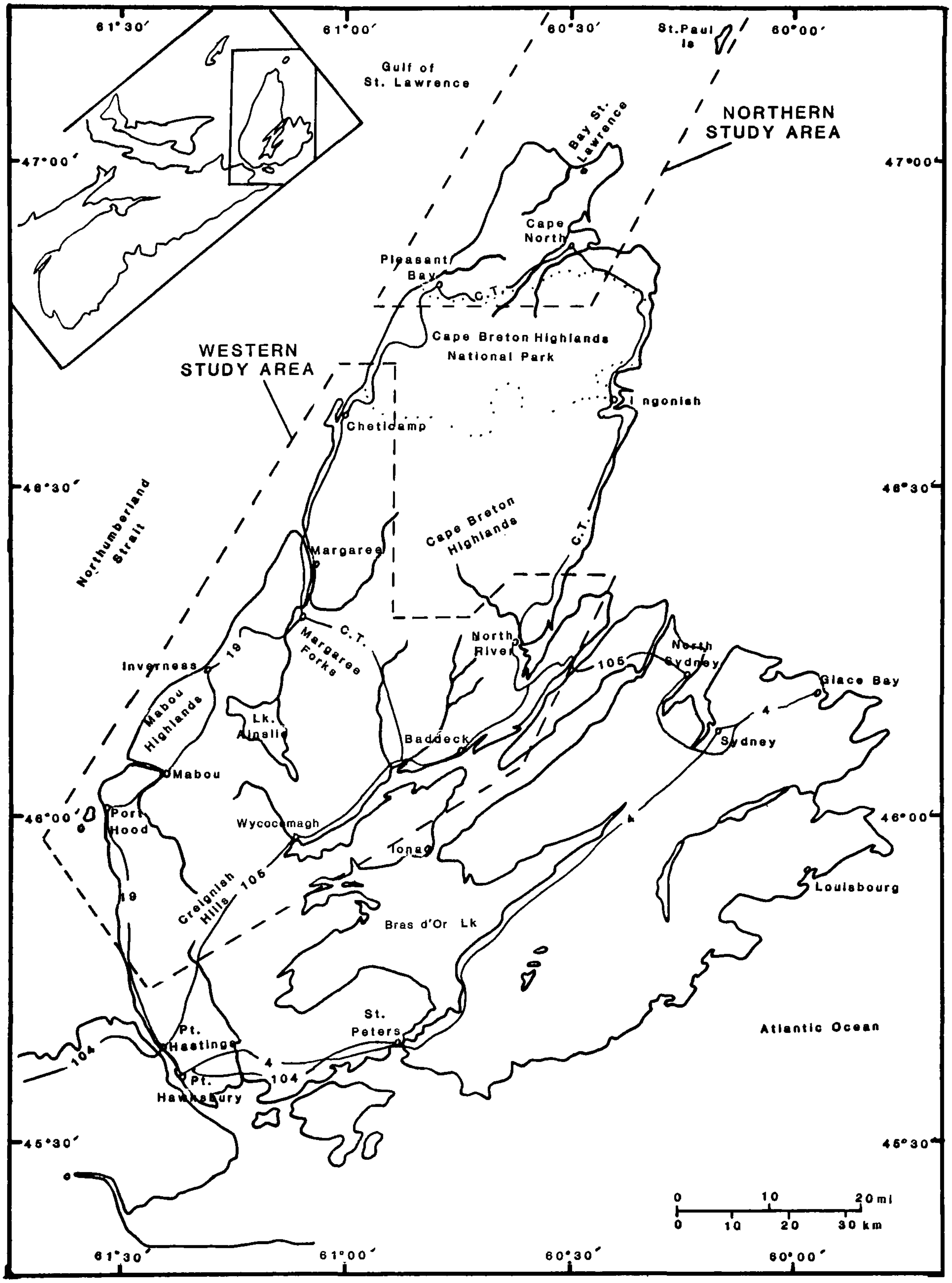

Figure 2. General geography of Cape Breton Island and location of study areas.

roads. Carboniferous strata are present along the coasts and beneath the low hills and valleys between resistant, steep-sided blocks of pre-Carboniferous basement that rise up to 500 m a.s.l.

Sixty-five outcrop sections in western and northern Cape Breton Island, an aggregate measured thickness of 21 000 m , were measured in as much sedimentological detail as time permitted (Fig. 3, App. I). Many sections in western Cape Breton are stream bottom and bank exposures, best viewed during late summer low water stages. Others, especially in northern Cape Breton, are coastal cliffs, commonly with limited access, that are subject to tidal timing. Outcrop quality is good in coastal sections, although faults are abundant, and poor to fair in stream sections where there are extensive covered intervals. Nowhere is the entire Horton Group well exposed.

Measured sections from 9 drillholes, 2000 m in aggregate, were also incorporated (Fig. 3, App. I) to enhance the three-dimensional view of the complex geology, fill in gaps between outcrops and provide some information of a more economic nature. The drillholes were selected because they had over 100 m of Horton rocks described on available lithologic logs, had some core available, or were in crucial locations. Most are shallow holes in the Mabou-Lake Ainslie area, whereas one is located offshore from Cape St. Lawrence (Fig. 3).

Basic field methods were employed throughout using only hammer, tape measure, Brunton compass and camera. Hand samples were procured from each facies assemblage and although detailed petrographic analysis was not a major objective, 75 thin sections provided significant information on composition, grain size/shape/sorting, porosity and cements (discussed in Chapter 3). In addition, since stratigraphic correlation is difficult in a nonmarine sparsely fossiliferous sequence, 53 palynological samples were obtained from appropriate Horton facies (Chapter 3, App. II). These aided in correlation, interpretation of depositional environments and evaluation of resource potential. Some trace fossil and plant fossil samples were also collected.

Nine hundred and thirty paleocurrent measurements were obtained, the first such body of data for the Horton Group of Cape Breton Island (discussed in Chapter 3). Data include directional indicators from trough cross stratification, ripple cross lamination, symmetrical ripple crests, linear sole marks, current lineation and pebble imbrication. All strata have been deformed to some extent, so all data were reoriented to their original

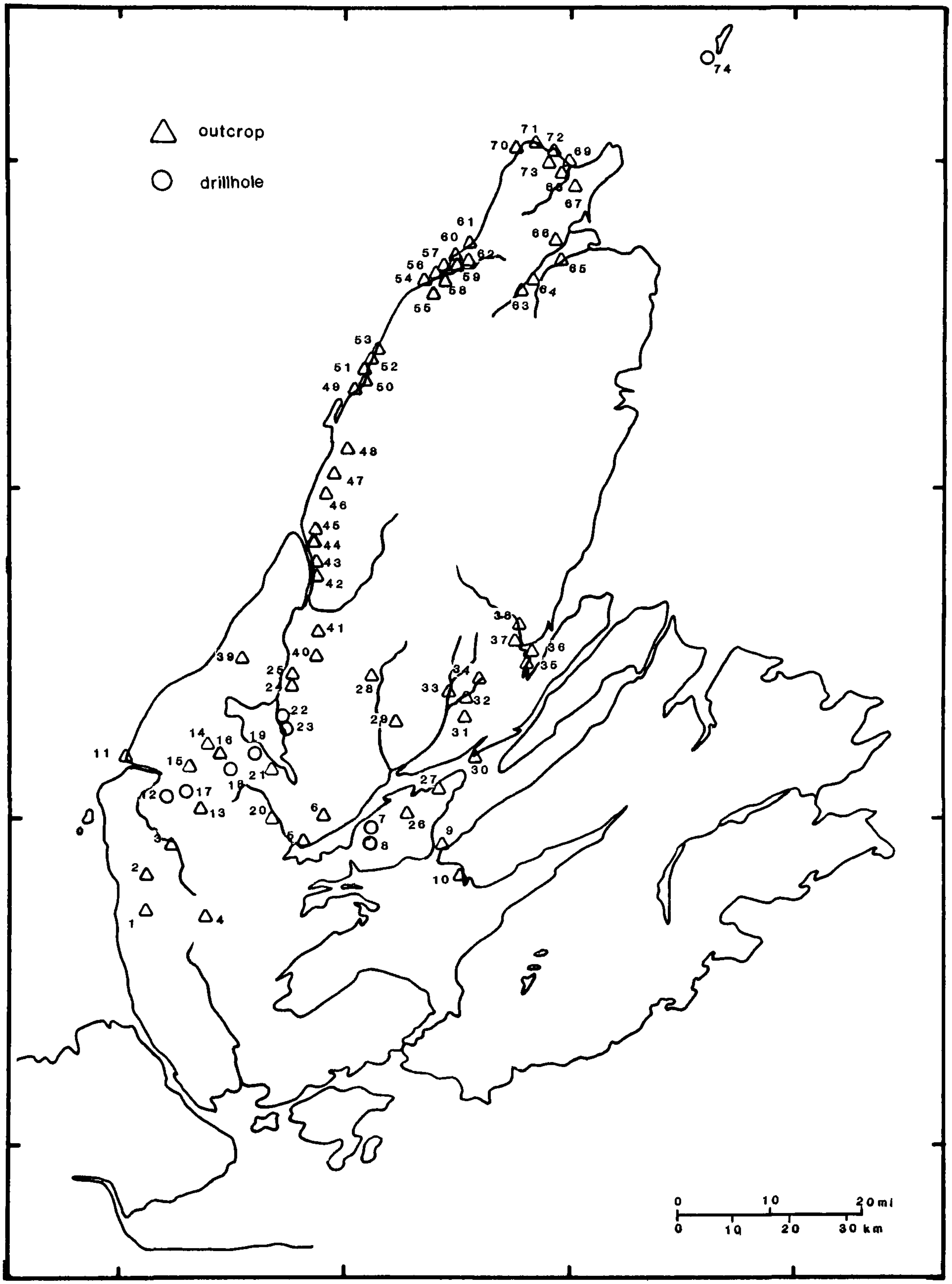

Figure 3. Locations of measured outcrop and drillhole sections. Numbers refer to section names listed in Appendix I.

horizontal position by stereonet procedures. Because of the structural complexity, strike and dip were measured for each bed from which paleocurrent data were obtained.

PREVIOUS WORK

The Lower Carboniferous rocks of Cape Breton Island have been studied intermittently since the work of Dawson (1858). Prior to 1968 the emphasis was on mapping and stratigraphy, but study of basin origins and structure became more popular after that. Descriptions of the Horton Group in Nova Scotia are scattered through the literature dating back to Dawson (1873) in the Minas Basin area of mainland Nova Scotia, and Fletcher (1884) on Cape Breton Island.

The Horton Group was formally defined by Bell (1929) with a type area designated as the southwest portion of the Minas Basin, but it can be traced throughout Nova Scotia to Cape Breton Island (see Williams et al., 1985, for historical background). The work of Norman (1935), Weeks (1954), Bell (1958), Neale and Kelley (1960), Murray (1960) and Kelley (1967) identified the basic stratigraphic framework of the Horton Group on Cape Breton Island and its correlation to the type area. These authors, particularly the latter two, also offered some interpretations of the depositional basin and environments.

Belt (1968a, 1968b), Bradley (1982) and Bradley and Bradley (1986) discussed the relation between tectonic style, basin origin and facies development in the area. Stockmal et al. (1987) summarized recent work on the tectonics of the Northern Appalachians. Ongoing studies on the pre-Carboniferous basement of Cape Breton by Barr and Raeside (1986), Raeside and Barr (1986) and on the Late Devonian Fisset Brook Formation volcanics by Blanchard et al. (1984) provide important background information to the study. Studies dealing with the resource potential of the Horton and Windsor Groups on Cape Breton Island are by Bell (1958), Kelley (1958), Côté (1959), and Kirkham (1978, 1985). Further details on these and other studies are discussed in appropriate sections throughout the thesis.

FORMAT OF THESIS

Introductory and background information on the general geological setting and stratigraphy of the study area is outlined in Chapter 2. The following two chapters present detailed definitions, descriptions, and interpretations of the facies assemblages identified.

Chapter 4 also discusses the vertical and lateral arrangement of the facies assemblages, and the trends in paleocurrent and paleontological data related to the geometry of the sediment bodies. Chapter 5 presents interpretations of basin geometry, paleogeography, sediment dispersal and tectonic controls. In Chapter 6 some comments are offered on the potential for, and possible distribution of, metallic and petroleum deposits associated with the Horton Group on Cape Breton Island. A short compendium of conclusions reached in this study is listed in Chapter 7. Several appendices include measured section locations and tabulate data relevant to the study.

CHAPTER 2
GEOLOGICAL SETTING AND STRATIGRAPHY

REGIONAL APPALACHIAN TECTONIC SETTING

The northern Appalachian Orogen is an elongate northeast-southwest belt of deformed rocks on the eastern seaboard of North America. Deep seismic studies have identified 3 lower crustal blocks in the Nova Scotia/Newfoundland area (Stockmal, 1988; Marillier, 1988), although the upper crustal rocks have been divided into 5 tectono-stratigraphic zones by Williams (1979) (Figs. 4 and 5). These zones extend the length of the Orogen and are based primarily on stratigraphic and structural characteristics of Late Precambrian to Ordovician rocks. Except for the Humber Zone (see below), these zones can be considered as "composite suspect terranes" (Williams and Hatcher, 1982; Keppie, 1985; after the definition of Coney et al., 1980), acquired by the North American craton during a long, complex tectonic history. Keppie (1985) suggested that this zonation is overly simplistic and that the Orogen is composed of many small terranes with complicated accretionary histories involving oblique subduction and large transcurrent motions (as in the Cordillera). All these terranes constitute basement to the Horton Group in some parts of the Maritime Provinces and so are relevant to this study.

Deformation, intrusion and metamorphism, though diachronous, were concentrated in the mid- to late Ordovician Taconian Orogeny, the mid- to late Devonian Acadian Orogeny, and the mid- to late Carboniferous Alleghenian Orogeny, which probably represent the main periods of terrane accretion. Certain time-stratigraphic assemblages characterize the Appalachian Orogen (Schenk, 1978; Thomas, 1977) (Fig. 6): 1) Grenvillian crystalline basement present in the Humber Zone, 2) thick Late Precambrian to Ordovician sediments and volcanics in all terranes (pre-Taconian Sauk sequence of Sloss, 1963), 3) Silurian to Middle Devonian sediments and volcanics well preserved only in the Humber zone (pre-Acadian Tippecanoe sequence of Sloss, 1963), 4) numerous Devonian and Carboniferous granitic intrusions associated with Acadian Orogeny, 5) thick Middle Devonian to middle Carboniferous sediments and volcanics which overlie all tectono-stratigraphic zones and include the Horton Group (post-Acadian Kaskaskia sequence of Sloss, 1963), 6) thick Middle Carboniferous to Permian sediments which overlie all tectono-stratigraphic zones (post-Alleghenian Absaroka sequence of Sloss, 1963).

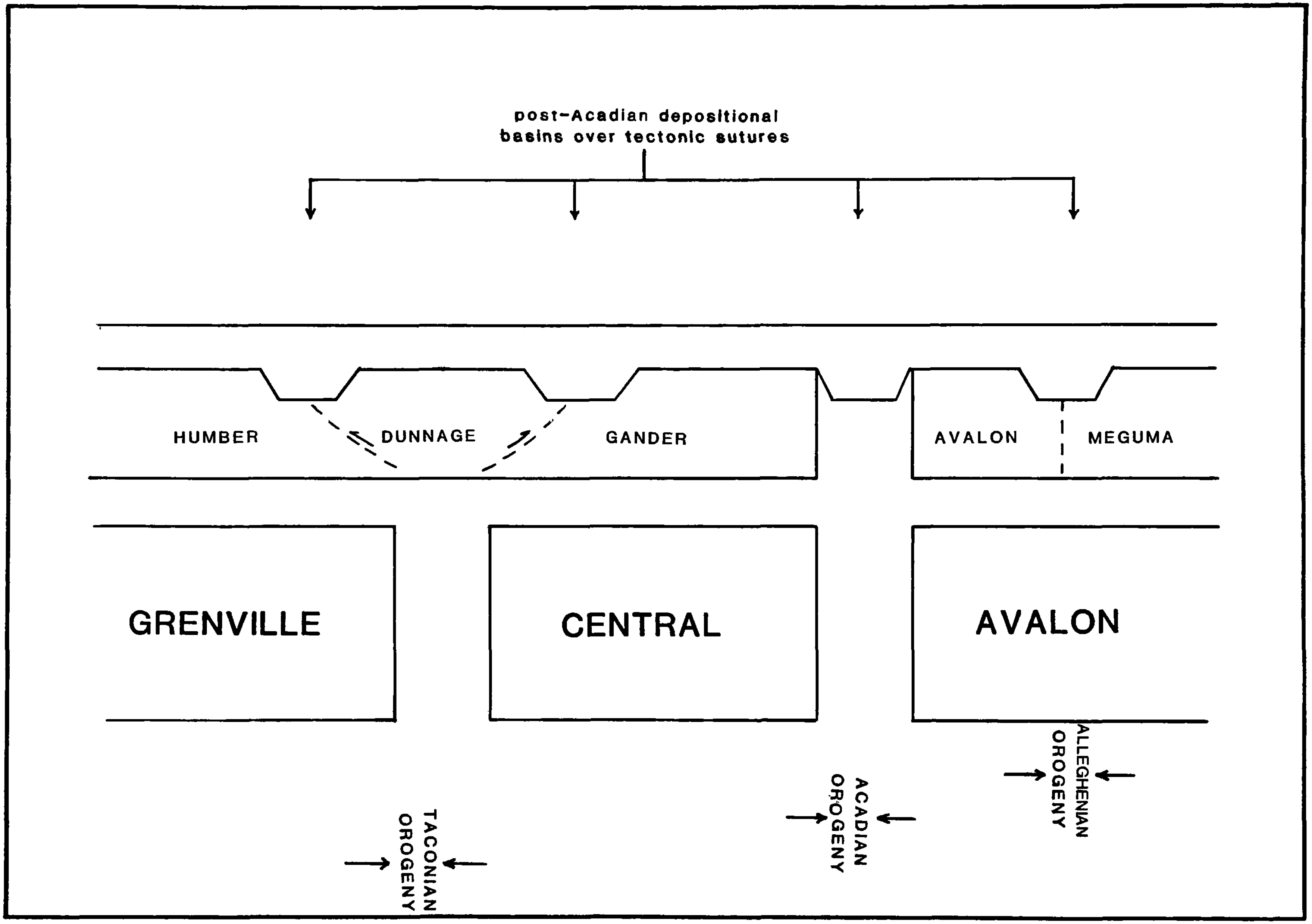

Figure 4. Relationship between three lower crustal blocks (Stockmal, 1988: Marillier, 1988), five tectonostratigraphic zones (Williams 1979) and locations of post-Acadian basins in the Canadian Appalachians.

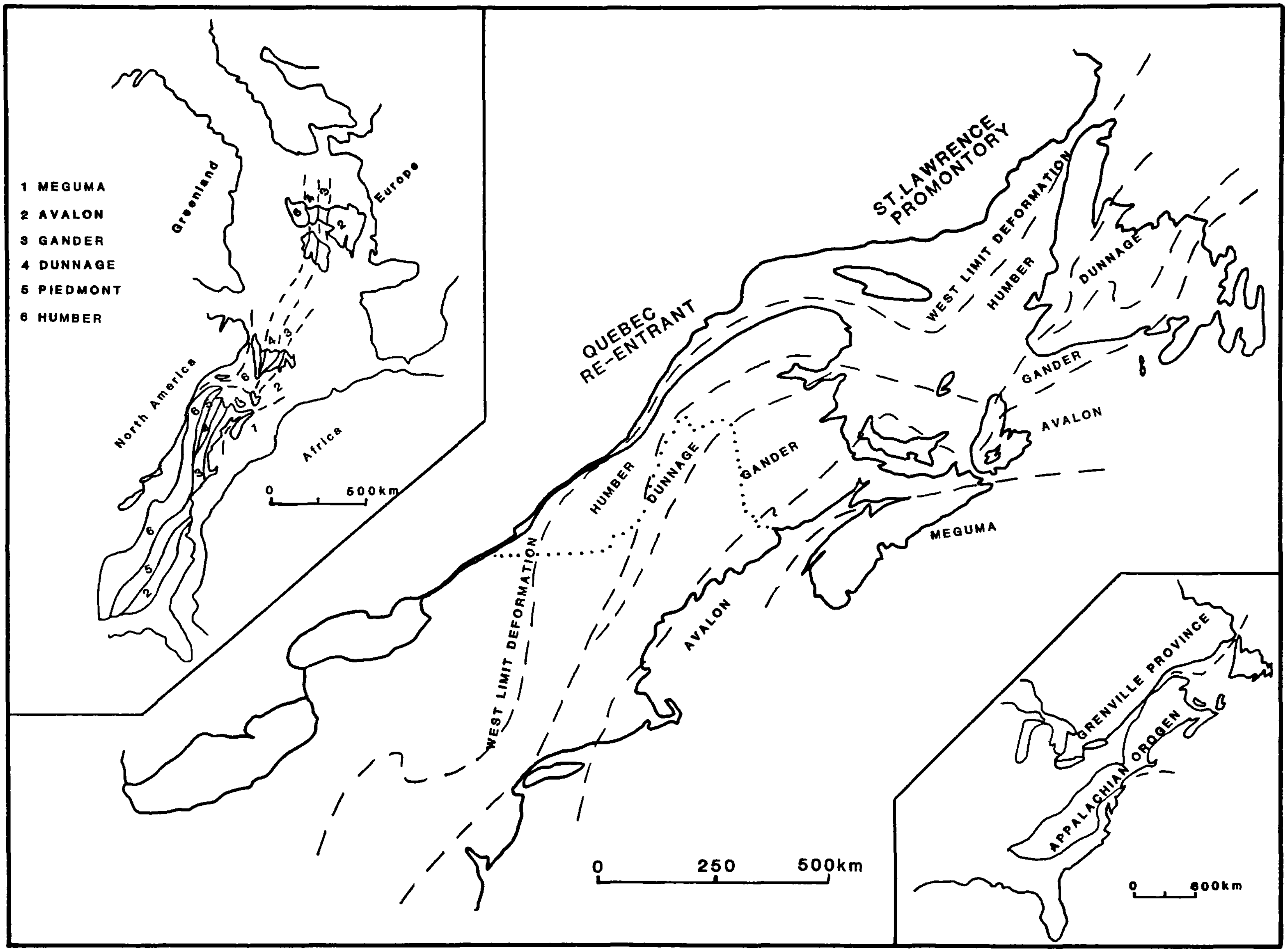

Figure 5. Tectonostratigraphic zones of the Appalachian Orogen (compiled from Williams, 1984; Williams and Hatcher, 1982; Keppie, 1985; Stockmal et al, 1987; Barr and Raeside, 1986).

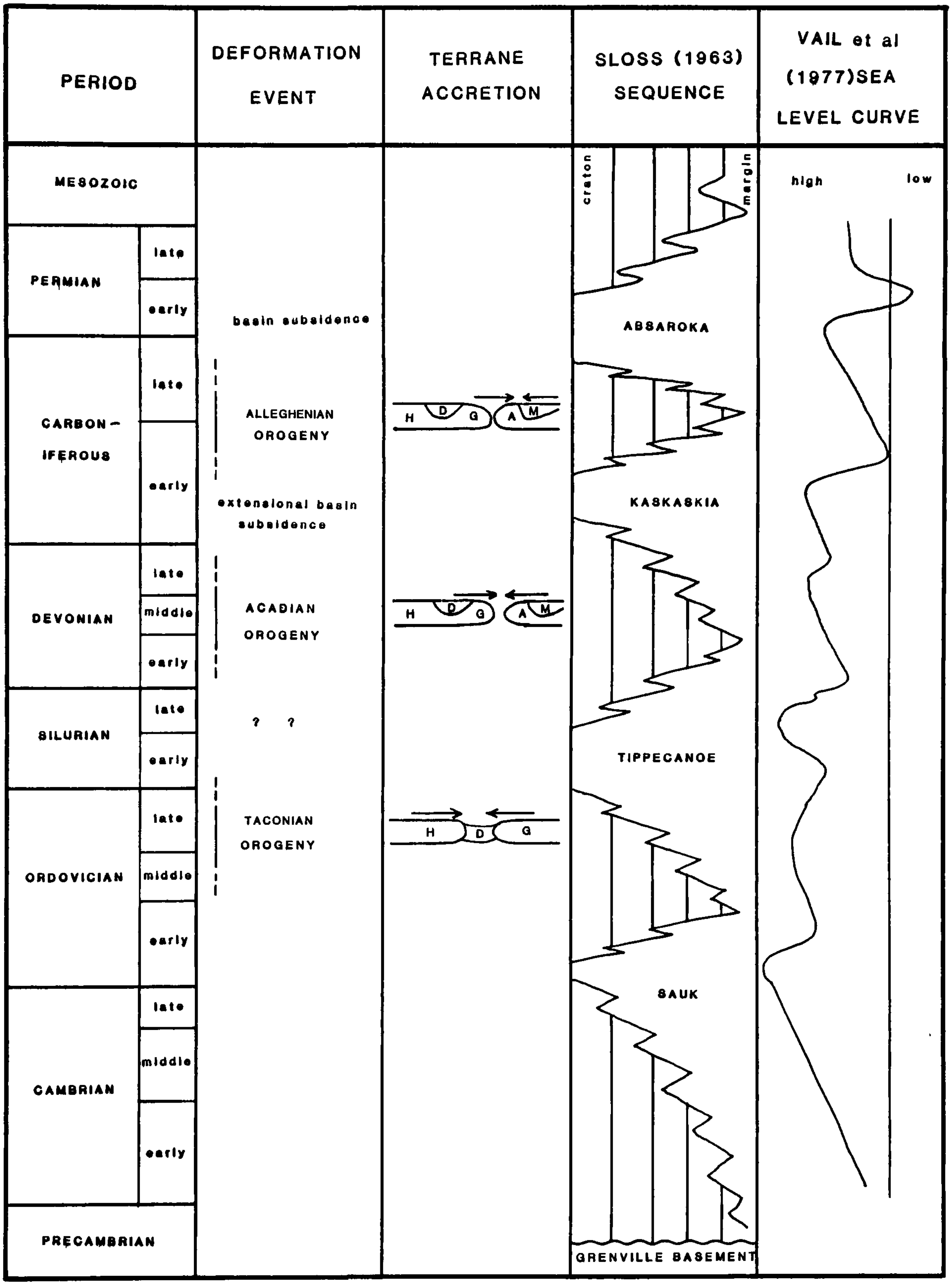

Figure 6. Terrane accretion and time-stratigraphic sequences of the northern Appalachian Orogen (compiled from Thomas, 1977; Schenk, 1978; Sloss, 1963; Vail et al, 1977).

INBOARD TERRANES (TACONIAN OROGENY)

The Humber Zone on the northwest side of the Appalachians consists of Proterozoic to Ordovician autochthonous sequences of the North American passive margin (Williams and Hatcher, 1982) overlying the Grenvillian lower crustal block (Marillier, 1988). These sequences are overlain by allochthonous sheets of oceanic crust obducted in Middle Ordovician to Silurian times (Lash, 1986). The Dunnage Terrane contains ophiolites, mélange and ultramafic plutons of ?Cambrian to Middle Ordovician age (Williams, 1979), and is generally interpreted as the vestiges of the Lower Paleozoic Iapetus Ocean (Williams, 1984). The Gander Terrane, in the middle of the Appalachian Orogen, consists of deformed Cambrian to Ordovician deepwater clastics and volcanics which overlie the Central lower crustal block (Schenk, 1978; Marillier, 1988). It may represent the southeastern margin of the Iapetus ocean basin.

These 3 zones are interpreted by Schenk (1978), Williams (1979) and Ziegler (1984) to represent Proterozoic breakup of a supercontinent, formation of the Iapetus Ocean and subsequent closure. This led to the suture of the Dunnage and Gander Terranes, carried on the Central block, to the Humber Zone, situated on the Grenville block, by orthogonal collision in the Middle to Late Ordovician Taconian Orogeny. Alternatively, Stockmal et al. (1987) view the Taconian Orogeny as the collision of the Dunnage ocean crust/volcanic arc with crustal promontories of the irregular North American margin followed by accretion of the Gander cratonic terrane in the Acadian Orogeny.

OUTBOARD TERRANES (ACADIAN/ALLEGHENIAN OROGENIES)

The Avalon and Meguma Terranes both overlie the Avalon lower crustal block of Marillier (1988) in the southeastern part of the Orogen. The Avalon Terrane is the most extensive of the composite tectono-stratigraphic zones and consists of thick Proterozoic sediments and volcanics overlain by Cambro-Ordovician shallow water clastics (Williams, 1984). There is little evidence of Taconic deformation, although the Acadian Orogeny affected the entire terrane. It has been interpreted by Williams (1979) as the margin of a thin exotic microcontinental mass which had stable platform conditions during the time when Iapetus Ocean was generated and destroyed. After Early Paleozoic motion (discussed in the next 2 sections) this terrane was accreted to the North American craton along a high angle suture which penetrated the entire crust (Keppie, 1982; Marillier, 1988) during the

Acadian Orogeny (Williams and Hatcher, 1982; Ziegler, 1984).

The Meguma Terrane is only exposed in southern Nova Scotia. It consists of a thick sequence of Cambrian to Lower Devonian marine clastics transported from the southeast (Williams, 1984), which can be correlated with rocks in North Africa (Schenk, 1978). There is no evidence of Taconic deformation, but the Acadian and Alleghenian Orogenies are recorded. The Meguma Terrane has variously been interpreted as the passive margin sequence of a continental mass, or a rift-fill sequence within the Avalon microcontinent (Williams, 1979). It is sutured to the Avalon Terrane along a high-angle fault zone. Keppie (1985) and Keppie and Dallmeyer (1987) suggest Late Silurian docking of the two terranes, followed by accretion of the resulting "Acadia composite terrane", carried on the Avalon lower crustal block, onto North America during the Acadian Orogeny. It is possible that the eastern part of the Appalachians was juxtaposed against the western part by major transcurrent displacement rather than overthrusting (Kennedy et al., 1982; Williams and Hatcher, 1982) and that the resulting deformation constitutes the Acadian Orogeny. The post-Acadian Horton Group overlies all these tectonic zones, confirming that they were assembled by Late Devonian time. The Alleghenian Orogeny, though locally intense, apparently did little to alter this situation in Nova Scotia (Stockmal et al., 1987). The main phase of Alleghenian deformation is manifested in Nova Scotia as middle to late Carboniferous dextral oblique transpression along various faults, including the Cobequid-Chedabucto fault which separates Meguma and Avalon Terranes, and associated thrusting and folding (Keppie, 1982; Nance, 1986).

PALEOMAGNETIC AND STRUCTURAL EVIDENCE

A considerable body of paleomagnetic and structural data from both sides of the Atlantic suggests that sinistral oblique convergence on major fault zones parallel to the inboard terranes was rapid and significant before the mid-Carboniferous (Keppie, 1985, 1988; Webb, 1969). Morris (1976), Kent and Opdyke (1979) and Irving (1979) documented paleomagnetic evidence suggesting 1500-2000 km (10°-15° latitude) of northward displacement of the Avalon Terrane relative to cratonic North America, probably in the early Devonian or possibly Silurian. This would have brought these 2 areas together in the Middle Devonian with Nova Scotia at about 10-15° S paleolatitude (Morel and Irving, 1978; Irving and Strong, 1984; Scotese et al., 1984; Kent, 1985; Woodrow, 1985) (Fig. 7). Several

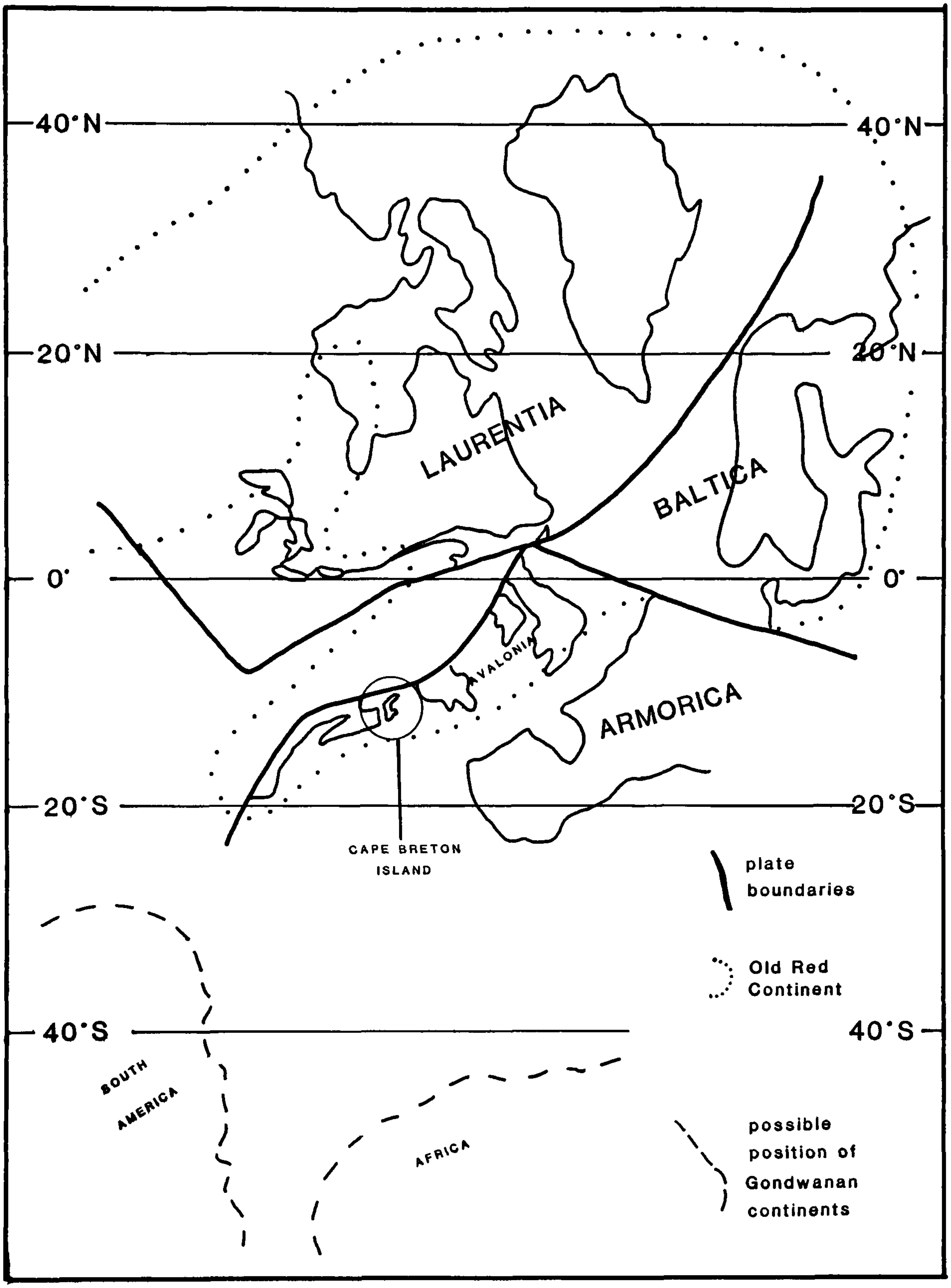

Figure 7. Paleomagnetic reconstruction of paleogeography of northern Appalachians in Middle to Late Devonian time (Morel and Irving, 1978; Kent, 1985; Woodrow, 1985). Nova Scotia was positioned at 10-15° S. paleolatitude.

studies (Van der Voo et al., 1979, 1981; Kent and Opdyke, 1979) apparently found paleomagnetic evidence for 1500-2000 km of sinistral motion of the outboard terranes in Carboniferous time. Recent re-study has shown that this conclusion was in error (Irving and Strong, 1984; Kent and Opdyke, 1985), although both later papers note that the revision does not negate the possibility of large Devonian sinistral motions.

Soper and Hutton (1984) and Hutton (1987) documented structural evidence for sinistral offset of hundreds of kilometres across northeast-striking fault zones in Britain between the mid-Silurian and mid-Devonian. Webb (1969) mentions structural evidence in New Brunswick for sinistral motions of Acadian age cut by dextral offsets, on different trends, of Alleghenian age. Based on various geological data Keppie (1988) suggests that progressive oblique sinistral convergence throughout the Ordovician, Silurian and Early Devonian led to closure of the Iapetus Ocean and culminated in transpressive emplacement of Avalonia along a deep vertical suture (as observed by Marillier, 1988).

There is ample paleomagnetic and structural evidence for oblique dextral offset of about 150-200 km and attendant folding and northwestward thrusting in the middle to Late Carboniferous Alleghenian Orogeny (Webb, 1969; Keppie, 1982; Bradley, 1982; Ziegler, 1984; Scotese et al., 1983; Piper and Waldron, 1988). However, this motion had ceased by the end of the Carboniferous (Scotese et al., 1984). If the above ideas are correct, there must have been a major re-orientation of the tectonic stress field between the Acadian and Alleghenian Orogenies, during which time the Horton, Windsor and Canso Groups (total of about 4000 m of strata, approximately equivalent to the Kaskaskia sequence of Sloss, 1963) were deposited in the area previously affected by the Acadian Orogeny. This period of subsidence and basin formation in Late Devonian to Early Carboniferous time may represent a relaxation phase (Quinlan and Beaumont, 1984) or a period of adjustment of the regional stress field to new plate configurations (Ziegler, 1984). Séguin (1986) studied dykes in the Gaspé area and postulated a period of extension after Acadian compression but before Alleghenian compression. Understanding the stress field and tectonic activity of this period is of crucial interest to this project.

TECTONO-STRATIGRAPHIC ZONES ON CAPE BRETON ISLAND

Until recently all of Cape Breton was included in the Avalon Zone. Through detailed study of pre-Carboniferous rocks of the Island Barr and Raeside (1986) and Barr et al. (1987, 1988) have identified 4 distinct tectonostratigraphic subdivisions (Fig. 8). They tentatively correlate the Southeastern Zone with the Avalon, the Aspy Zone (formerly Highlands Zone) with the Gander, and the Northwestern Zone with the Humber Zone. The Bras d'Or Zone is interpreted as a previously unrecognized terrane, apparently thrust over the Aspy Zone (Barr et al., 1988). If this zonation is correct, the Horton Group on Cape Breton Island overlies all tectonostratigraphic zones of the Appalachians (except the Meguma), indicating that the Orogen was fully assembled by the Late Devonian, with only minor alterations in this area during the Alleghenian Orogeny.

The recognition of this zonal complexity is significant because Cape Breton lies at a "strategic position in attempts to correlate geological subdivisions between Newfoundland and the mainland" (Barr and Raeside, 1986). The zonation mentioned above is supported by the deep seismic data of Stockmal et al. (1987) and Stockmal (1988) who postulated a northwest-southeast dextral shear zone in the Strait of Canso (Stockmal 1989, pers. comm.) which offset the zones by 200-250 km in pre-Carboniferous times (Figs. 5, 8). This resulted from the irregular geometry of the North American margin at the St. Lawrence Promontory/Quebec Re-entrant and its effect on incoming terranes during the Taconian and Acadian Orogenies (Stockmal et al., 1987).

STRUCTURAL GEOLOGY OF CAPE BRETON ISLAND

Western Cape Breton can be divided broadly into the Carboniferous lowlands and the pre-Carboniferous highlands, both dominated by northeasterly-oriented structural trends. All observable structural evidence relates to post-Horton (Alleghenian) compression: there is no direct evidence of Horton age deformation. In Carboniferous rocks linear folds, plunging gently to the southwest, and normal faults trend northeast and most are at least post-Windsor in age (Norman, 1935; Currie, 1977; my observations). They correlate with a period of thrusting and folding observed in New Brunswick by Webb (1969) and Ruitenberg and McCutcheon (1982). There are also subordinate northwesterly-trending reverse and normal faults (Kelley, 1967). Between areas of Carboniferous outcrop there are basement blocks, commonly bounded by high angle faults and tilted to the west (Kelley, 1967). Some of these bounding faults are high angle reverse faults which produce overturned folds in adjacent Carboniferous rocks (Kelley, 1967; Currie, 1977). Ferguson (1946) noted the presence of isolated outliers of red Horton conglomerate on top of

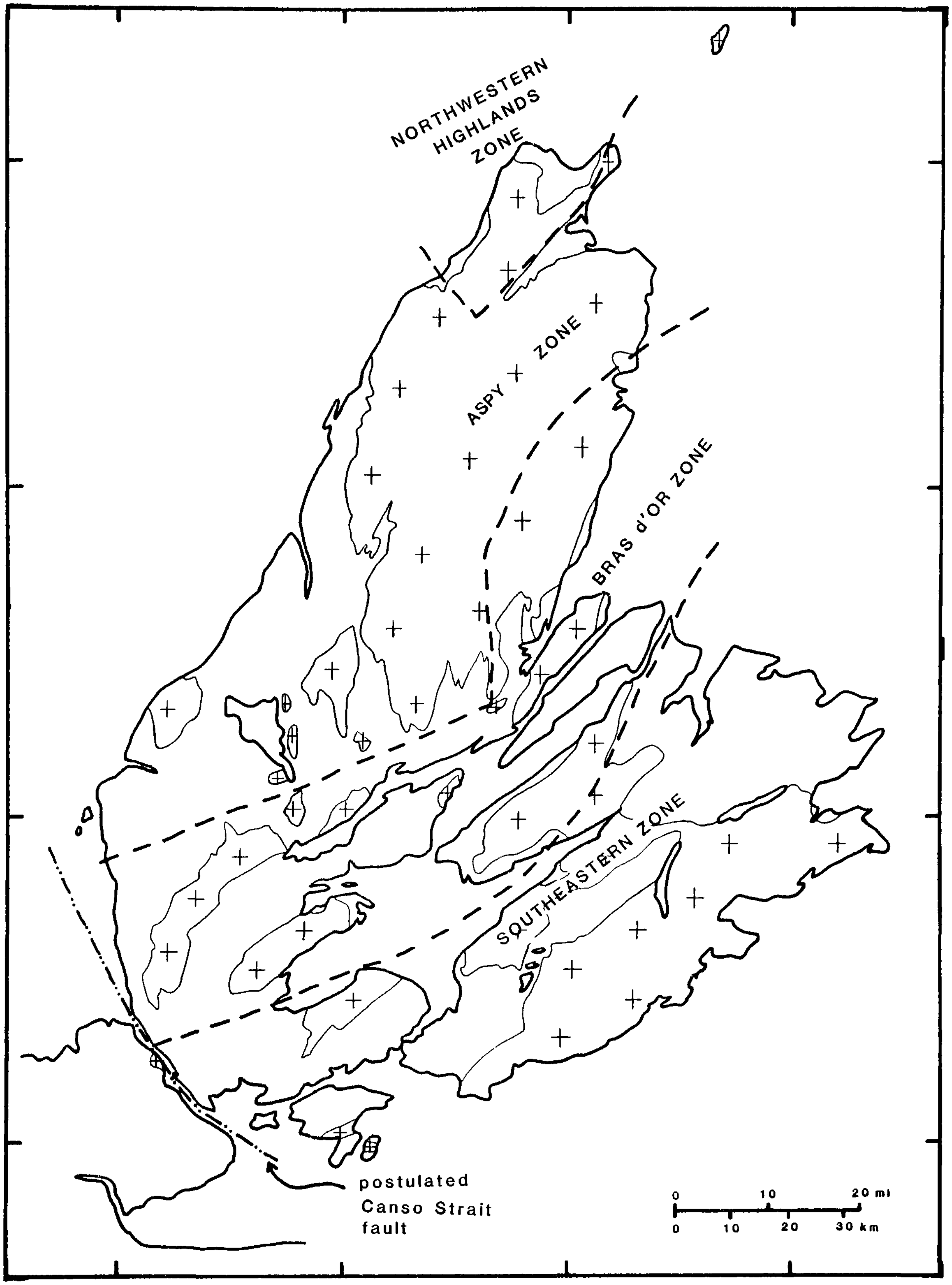

Figure 8. Four tectonostratigraphic subdivisions of pre-Carboniferous rocks of Cape Breton Island (Compiled from Barr and Raeside, 1986; Barr et al, 1987).

Creignish Hills, implying that this basement block was not a positive feature during Horton deposition but was later faulted and uplifted.

Northern Cape Breton is dominated by the Cape Breton Highland basement complex, which records multiple deformation, and is surrounded by small pockets of deformed Carboniferous rocks. In general, the Highlands rocks have been metamorphosed to a high grade and are characterized by diverse structural trends (Raeside and Barr, in press). The tectonostratigraphic zones of Barr and Raeside (1986) are separated by wide shear zones. One significant lineament, not identified as a zone boundary is the northeasterly-trending Aspy Fault, postulated to have been active during the Carboniferous (Webb, 1969; Barr et al., 1987). Carboniferous rocks preserved around the margins of the Highlands are commonly near-vertical to overturned and cut by many near-vertical faults.

Currie (1977) documented several thrust faults in northern Cape Breton and postulated others, some of which are at least post-Windsor in age (Fig. 9). These generally include Precambrian and Horton rocks thrust over Windsor Group in directions consistently away from the Cape Breton Highlands, interpreted by Fyson (1967) and Currie (1977) to represent gravity sliding away from uplifted basement blocks toward sedimentary basin areas. Neale and Kelley (1960) also noted local thrust faults in the same area. Weeks (1954) postulated a major northward-directed thrust which carried Horton and Windsor Group rocks over basement in southeastern Cape Breton (Fig. 9), although Webb (1969) suspected strike slip motion in that area. Smith and Collins (1984) found evidence in central Cape Breton for post-Windsor thrusting, and reviewed additional evidence from other sources for similar fault motion throughout Nova Scotia and New Brunswick. Boehner (1986) postulated a major thrust fault in the Windsor Group of western Cape Breton, and confirms that thrust offset in central Cape Breton is a possible explanation of extreme thickness and facies changes in Windsor Group rocks over short distances (Boehner 1988, pers. comm.).

In western Cape Breton Norman (1935), Cameron (1948) and Kelley (1967) identified northwestward-directed thrusts which cut Horton and Windsor rocks. The descriptions of Norman (1935) suggest an additional set of minor southeastward-directed (antithetic) faults (Fig. 9). In the same area, preliminary interpretation of a seismic cross section (line 62Y in McCormick, 1982) suggests a possible major thrust plate, with ramp-and-flat geometry, which carries Precambrian, Horton, Windsor and Canso rocks (Fig. 10).

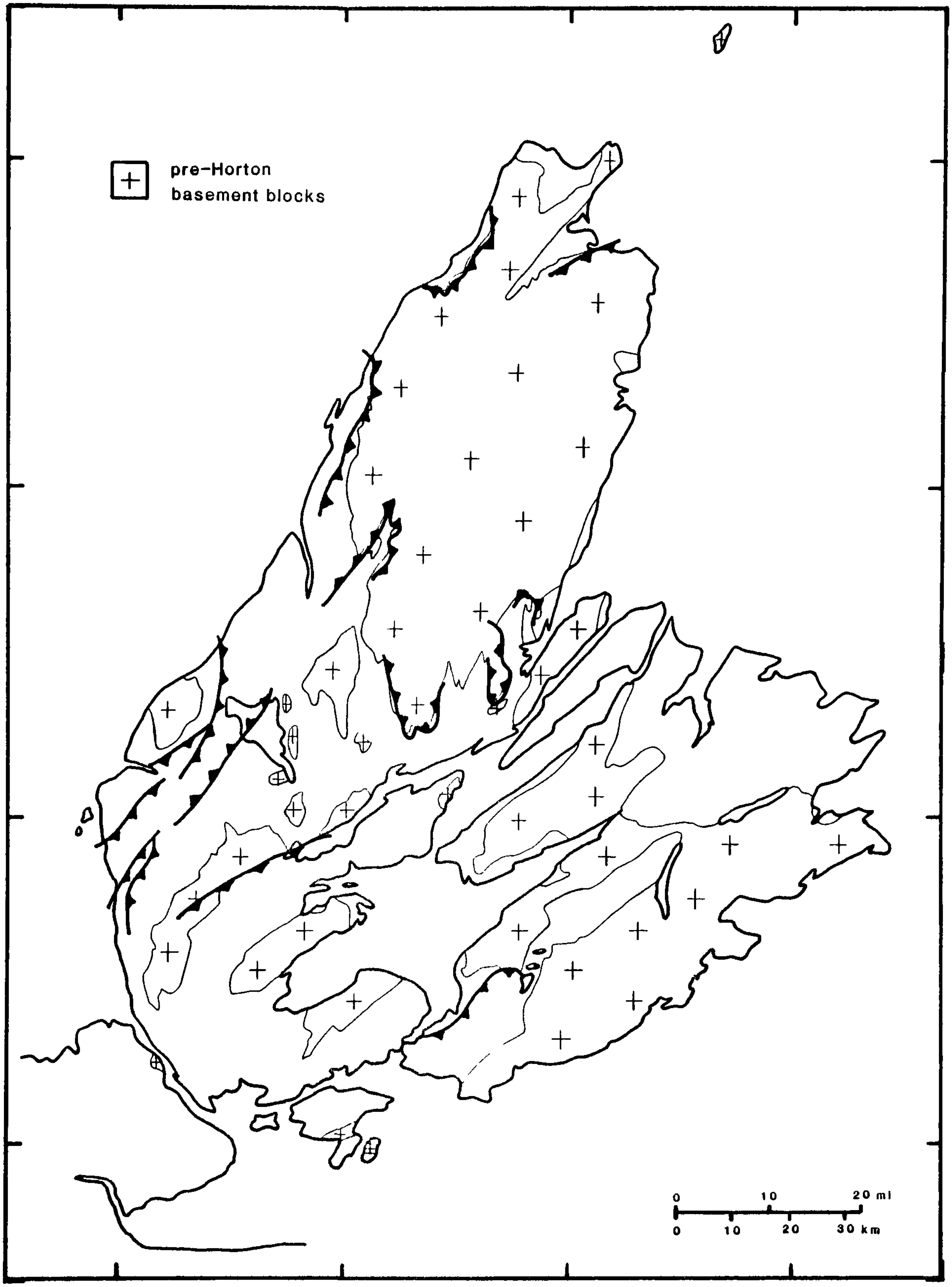

Figure 9. Postulated post-Horton thrusting on Cape Breton Island (compiled from Norman, 1935; Cameron, 1948; Weeks, 1954; Kelley, 1967; Currie, 1977; McCormick, 1982; and this study).

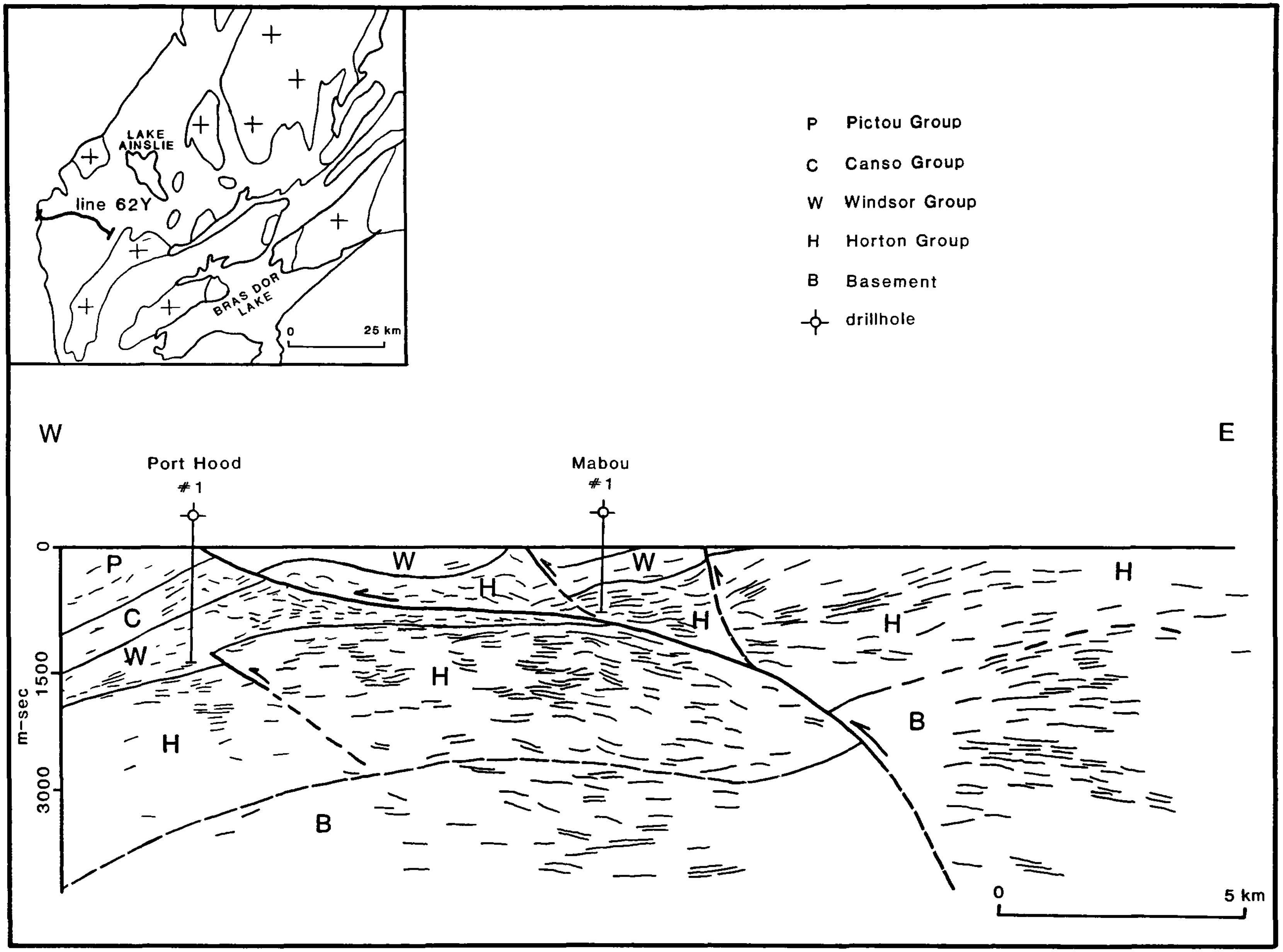

Figure 10. Portion of seismic line 62Y with preliminary interpretation of major, and associated minor, thrust faults.

Many accessory synthetic and antithetic faults with overlying folds are suggested as well as a significant zone with Horton structurally imposed over Horton. The presence of numerous subhorizontal faults and apparent repetition and thickening of section in outcrop and drillholes of western Cape Breton suggests that post-Horton thrusting has occurred in this area (Fig. 11). Confirmation of this conclusion is difficult in interbedded lithologies with abundant facies changes. However, the nearby inferred original margin of the Horton depositional basin, and associated marginal facies, are apparently offset towards the northwest by 10-15 km (Fig. 9) (further discussed in Chapter 4).

Together, these observations suggest that there is a major northwestward-directed overthrust sheet, perhaps with many accessory décollement surfaces in the fine grained facies of the middle Horton, in this area where maximum thicknesses have been attributed to the Horton Group (as by Kelley, 1967). All evidence indicates that the responsible compressive stress occurred in post-Windsor, and likely post-Canso, time (Smith and Collins, 1984). The identification of these thrusts is important in explaining facies thickness and distribution patterns, and it bears on the resource potential of the Horton Group.

THE LATE PALEOZOIC MARITIMES BASIN

The Acadian Orogeny appears to have brought eastern and western tectonostratigraphic zones together into a single intracontinental area where post-orogenic basins actively subsided. The depocentres have been referred to by the collective term "Maritimes Basin" coined by Williams (1973) (Fig. 12). This term refers to a structurally-complex, syn- and post-Acadian, intramontane or successor basin which includes platforms of thin undeformed sediments, northeast trending sub-basins with sediments up to 6 km thick, and linear northeast-trending ridges of exposed basement (Fig. 13). It is partly represented in the onshore geology of all 4 Maritime Provinces but about two-thirds of the 150 000 km^2 area of the Basin is offshore in the Gulf of St. Lawrence (Williams, 1973). It appears to have undergone a two-phase history of a) Late Devonian/Early Carboniferous fault-bounded sub-basins with thick sediments, and b) a Late Carboniferous broad sag with a thinner blanket of sediments, separated by a period of deformation representing the Alleghenian Orogeny (St. Peter, 1987).

Bell (1929, 1944, 1958) visualized the "Fundy Basin", a wide series of partly-interconnected downwarped sub-basins separated by broad uplifts of pre-Carboniferous

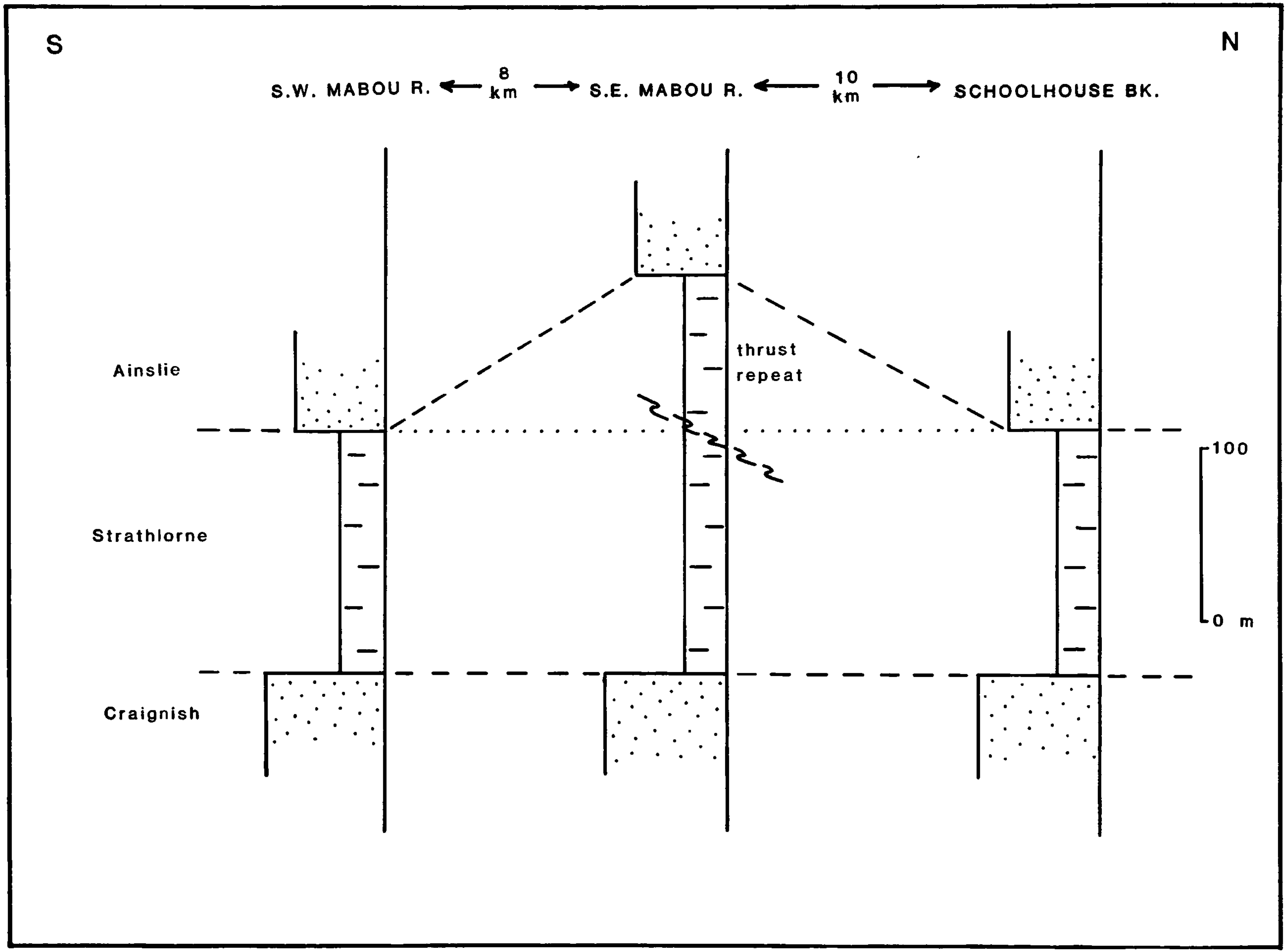

Figure 11. Observed thickening of Strathlorne section and fault zone, interpreted as thrust repetition in Mabou area.

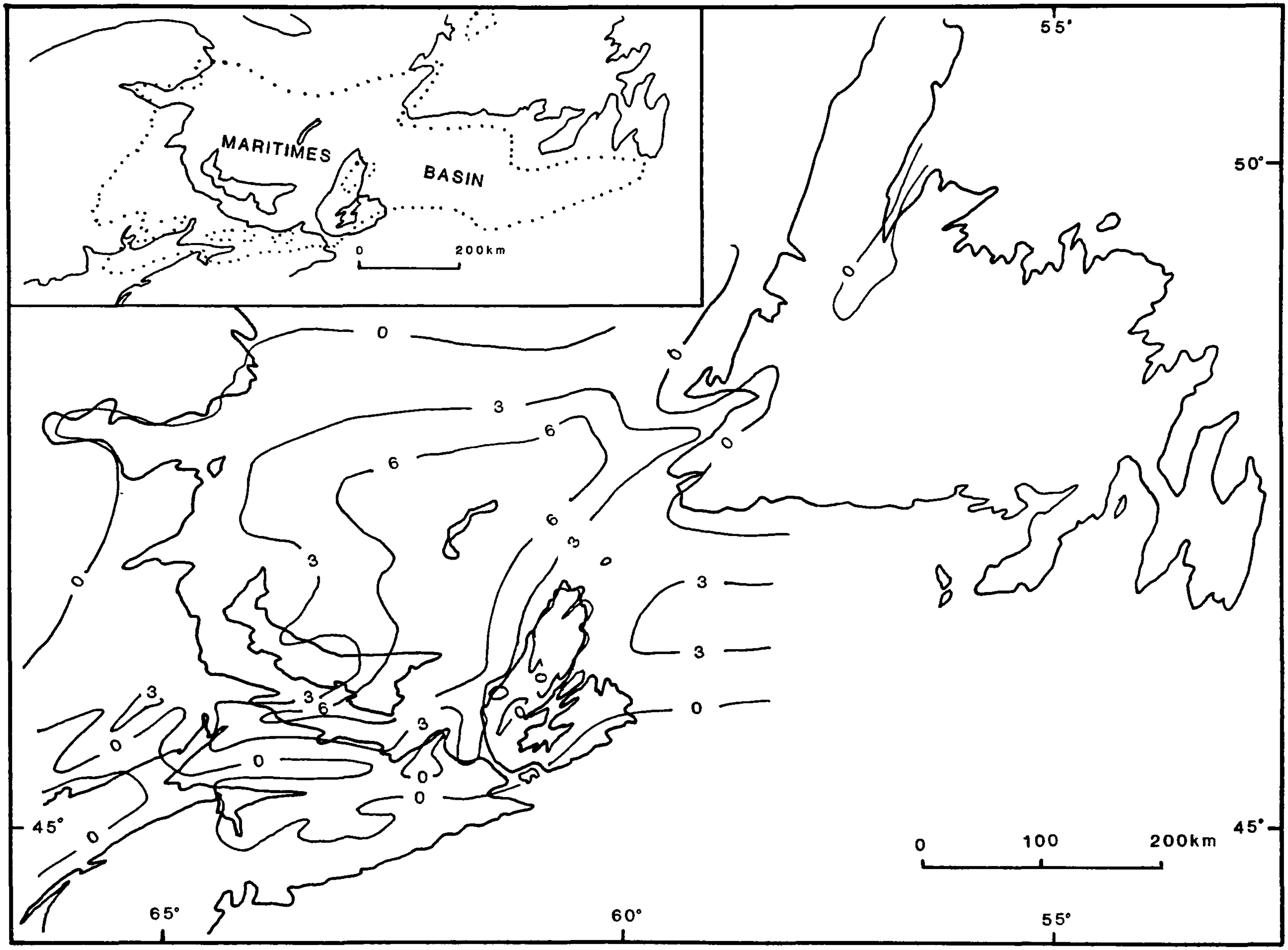

Figure 12. Thickness of Carboniferous strata in the post-Acadian Maritimes Basin (from St. Peter, 1987; Howie and Cummings, 1961; Howie and Barss, 1975). Contour values in kilometres.

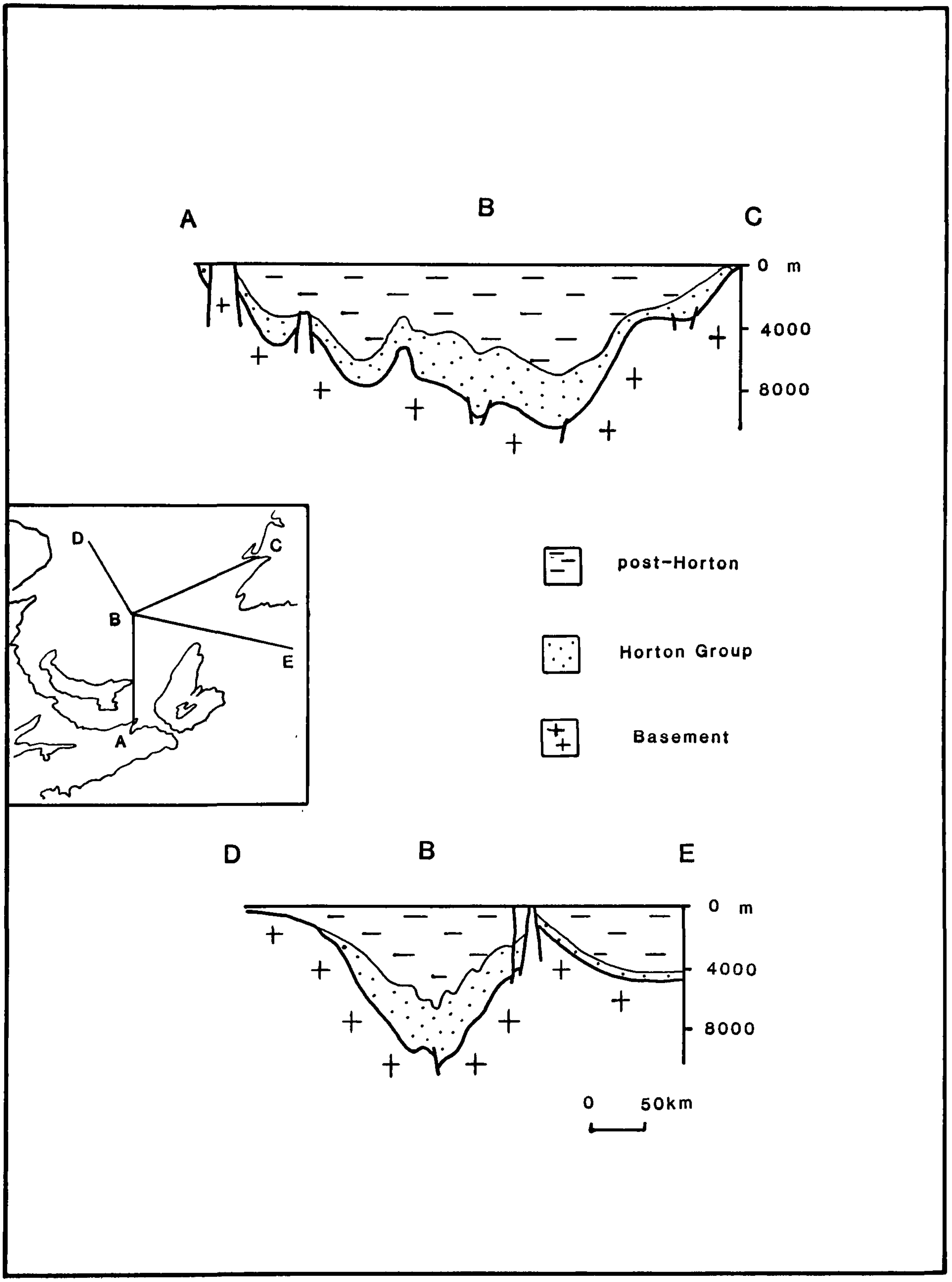

Figure 13. Generalized cross sections of the Maritimes Basin (from Howie and Barss, 1975).

basement. Poole (1967) referred to the "Fundy Geosyncline" as a composite of uplifted source areas and miniature basins, each with its own local history. Kelley (1967) had similar views and emphasized that although similar lithofacies occur in all sub-basins, they may not all correlate in age, and that deposition began in narrow basins which widened through time.

Belt (1968a, b) introduced the concept of the "Fundy Basin Rift" bounded by high angle dip-slip faults or sharp flexures which controlled facies distributions through a long history (Fig. 14). The timing, magnitude and sense of fault motion could vary along the rift, producing a different local stratigraphy in each area, although a fining of grain size toward the centre would be typical of all parts. Howie and Barss (1975) suggested the Maritimes Basin formed by a relaxation during the waning stages of the Acadian Orogeny, while Morel and Irving (1978), based on paleomagnetic data, related it to separation of Gondwana from Laurasia after the Acadian Orogeny (Fig. 15). Fyffe and Barr (1986) refer to the Basin as a failed rift, based on the chemistry of enclosed volcanic rocks.

Bradley (1982) suggested that most Carboniferous basins in Atlantic Canada are strike-slip pull-aparts developed within the very large post-Acadian dextral Magdalen Basin (Mann et al., 1983) (Fig. 16). Webb (1969) had earlier cited evidence for strike-slip motion on some faults throughout the northern Appalachians. While there is evidence for extension in the Magdalen area (Quinlan, 1988; Stockmal, 1988), the evidence for transtensive, as opposed to distensive, control is largely derived from post-Horton units. Hyde (1979) interpreted strike-slip basin formation in Viséan rocks of central Newfoundland while Bradley and Bradley (1986) interpreted a small dextral pull apart of Viséan age in southeastern Cape Breton. Although there is evidence for conflicting interpretations these have converged towards the idea that post-Acadian basins were intracontinental, fault-bounded, extensional successor basins, and that fault motions were likely episodic and multiple.

GENERAL STRATIGRAPHY OF CAPE BRETON ISLAND (refer to Table 1 and Figure 17)

PRECAMBRIAN Precambrian metamorphosed sediments and volcanics underlie most of the upland areas of Cape Breton Island and form the crystalline basement and sediment source for Late Paleozoic sediments (Fig. 18). In the past most Precambrian rocks were

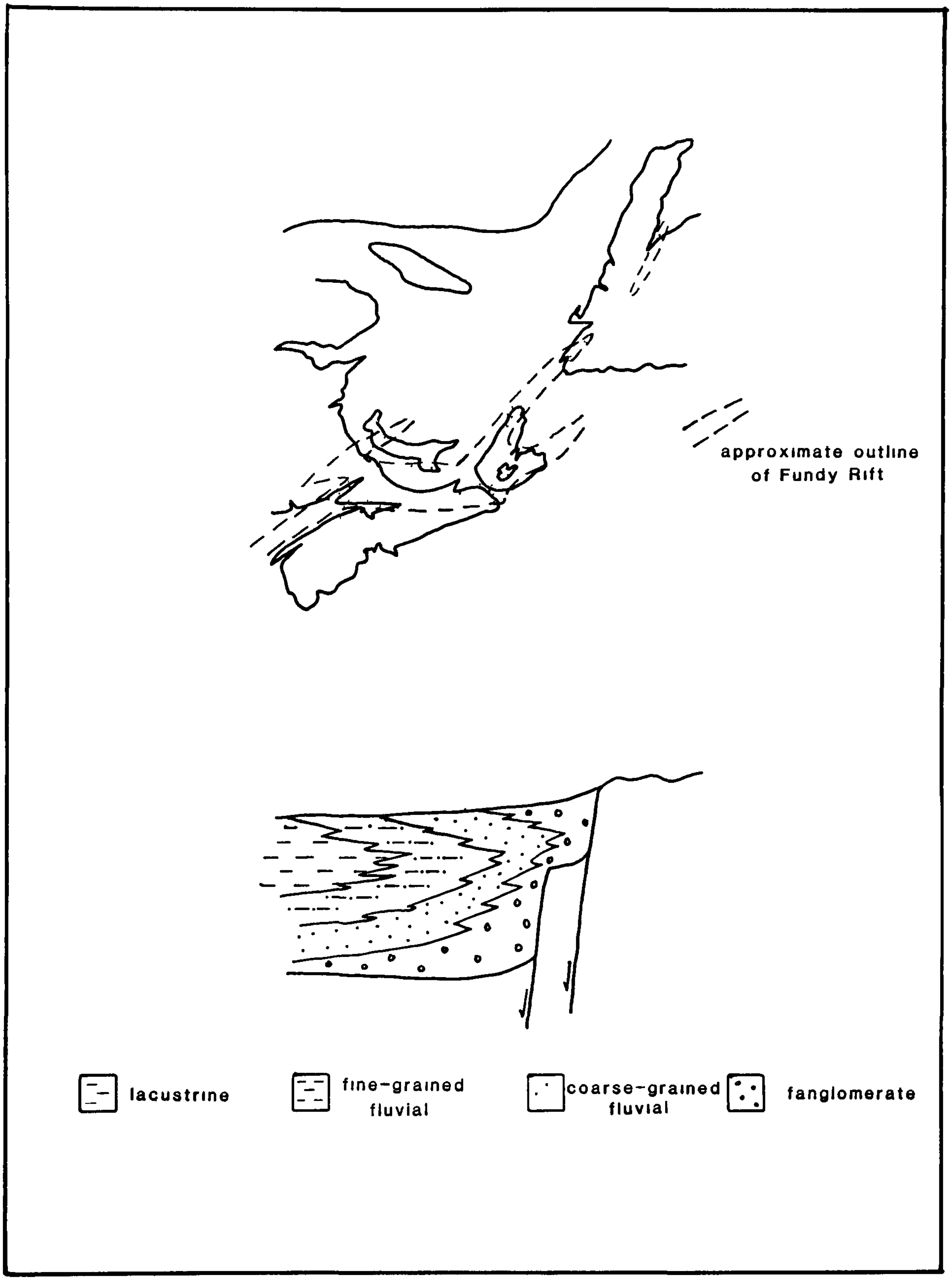

Figure 14. The concept of the Fundy Basin Rift and control of depositional facies by bounding faults (from Belt, 1968a, 1968b).

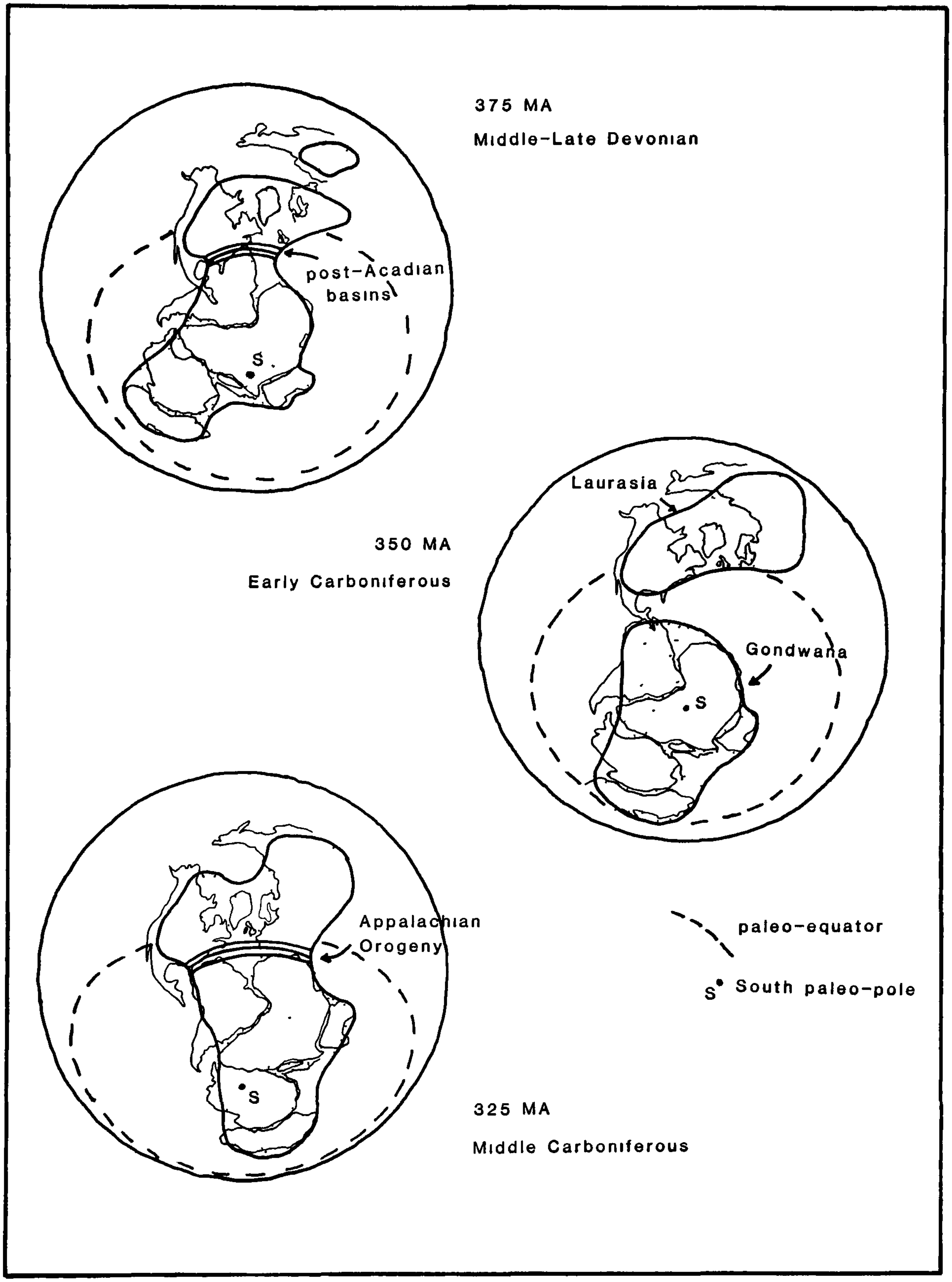

Figure 15. Interaction of Gondwanan and Laurasian cratons in Middle Devonian to Middle Carboniferous times, based on paleomagnetic data (from Morel and Irving, 1978).

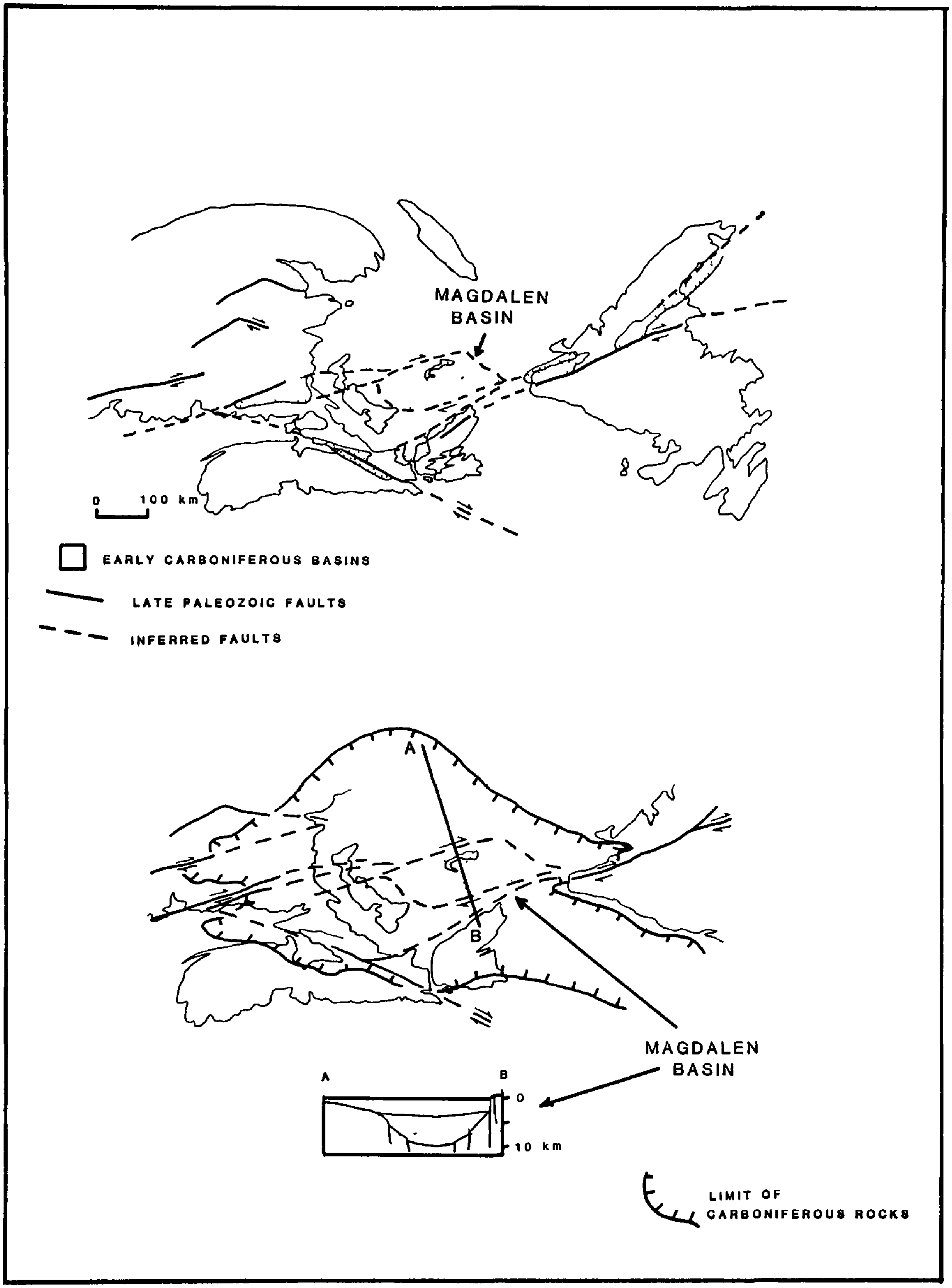

Figure 16. The position of the Magdalen Basin in the Canadian Appalachians and its interpretation as a dextral strike-slip pull-apart (from Bradley, 1982; Mann et al, 1983).

referred to the George River Group in western Cape Breton or the Fourchu Group in eastern Cape Breton. As already mentioned, recent mapping has organized the rocks into tectonostratigraphic zones, each characterized by particular rock assemblages (Barr and Raeside, 1986; Raeside and Barr, 1986).

The Southeastern Zone (outside the study area) is characterized by calc-alkaline mafic to felsic meta-volcanics and minor sediments of the Fourchu Group. The Bras d'Or Zone is underlain by a variable suite of greenschist meta-sediments of the George River Group and McMillan Flowage Formation. Grey schistose quartzite is typical, with locally prominent meta-greywacke, white marble, grey slate, gneiss (dated at 700 Ma), green meta-volcanics and some granitoid intrusions. The stratigraphy is not yet well understood. The Aspy Zone includes a) gneiss and amphibolite of the Cape North Group, b) meta-volcanics of the Money Point Group, and c) the Cheticamp Lake and Coastal Gneiss belts. The Northwestern Highlands Zone includes the Blair River Complex gneisses and amphibolites dated at 1045 Ma (Grenvillian).

<u>LOWER PALEOZOIC</u> (Fig. 18) In southeastern Cape Breton, outside the study area, slightly metamorphosed Cambro-Ordovician rocks were described by Weeks (1954). These include fossiliferous sandstone, conglomerate, shale, greywacke and bimodal volcanics of the Bourinot-Kelvin Glen Groups. Although they are not preserved elsewhere, these sediments may have originally been present throughout the Island.

Grey to pink granitic intrusives are scattered through the uplands basement rocks and may represent several generations of magmatic activity. Kelley (1967) notes granitic dykes and sills which cut Cambrian rocks but are overlain by Carboniferous sediments.

<u>UPPER PALEOZOIC</u> The Upper Paleozoic succession of Cape Breton Island rests unconformably on Acadian basement and includes most of the main Groups found throughout Nova Scotia and offshore in the Gulf of St. Lawrence (Fig. 13).

Lower Carboniferous rocks (Fig. 19) correspond to the (post-Acadian, pre-Alleghenian) Kaskaskia sequence of Sloss (1963). The Late Devonian-?Early Carboniferous Fisset Brook Formation is sporadically distributed near basement blocks and consists of mafic and felsic bimodal volcanics with some interbedded sediments. The formation is conformably overlain by Horton Group grey and red sandstone, siltstone and conglomerate, followed by Windsor Group limestone, anhydrite, salt and red siltstone, both of Early Carboniferous age. These units are discussed in more detail in the next section. Canso

AGE MA	PERIOD	STAGE	GROUP	LITHOLOGY	PLUTONICS
250	PERMIAN	late			
		early			
300	CARBONIFEROUS	late	PICTOU-MORIEN	clastics,coal	
			RIVERSDALE	clastics,coal	
		early	CANSO	clastics	
			WINDSOR	carbonates, evaporites	
350			HORTON	clastics	granitics
	DEVONIAN	late	FISSET BROOK	volcanics	
		middle	McADAM LAKE	volcanics,clastics	
		early			
400	SILURIAN	late			
		early			
450	ORDOVICIAN	late			
		middle			
		early			
500	CAMBRIAN	late	BOURINOT-KELVIN GLEN	clastics, volcanics	granitics
		middle			
550		early			
no time scale implied	HADRYNIAN		BLAIR RIVER, CHETICAMP LAKE, MONEY POINT, CAPE NORTH, McMILLAN FLOWAGE, GEORGE RIVER, FOURCHU	meta-volcanics, meta-clastics	mafics, ultramafics, granitics
	HELIKIAN		UNAMED	high grade metamorphics	

Table 1. General stratigraphy of Cape Breton Island.

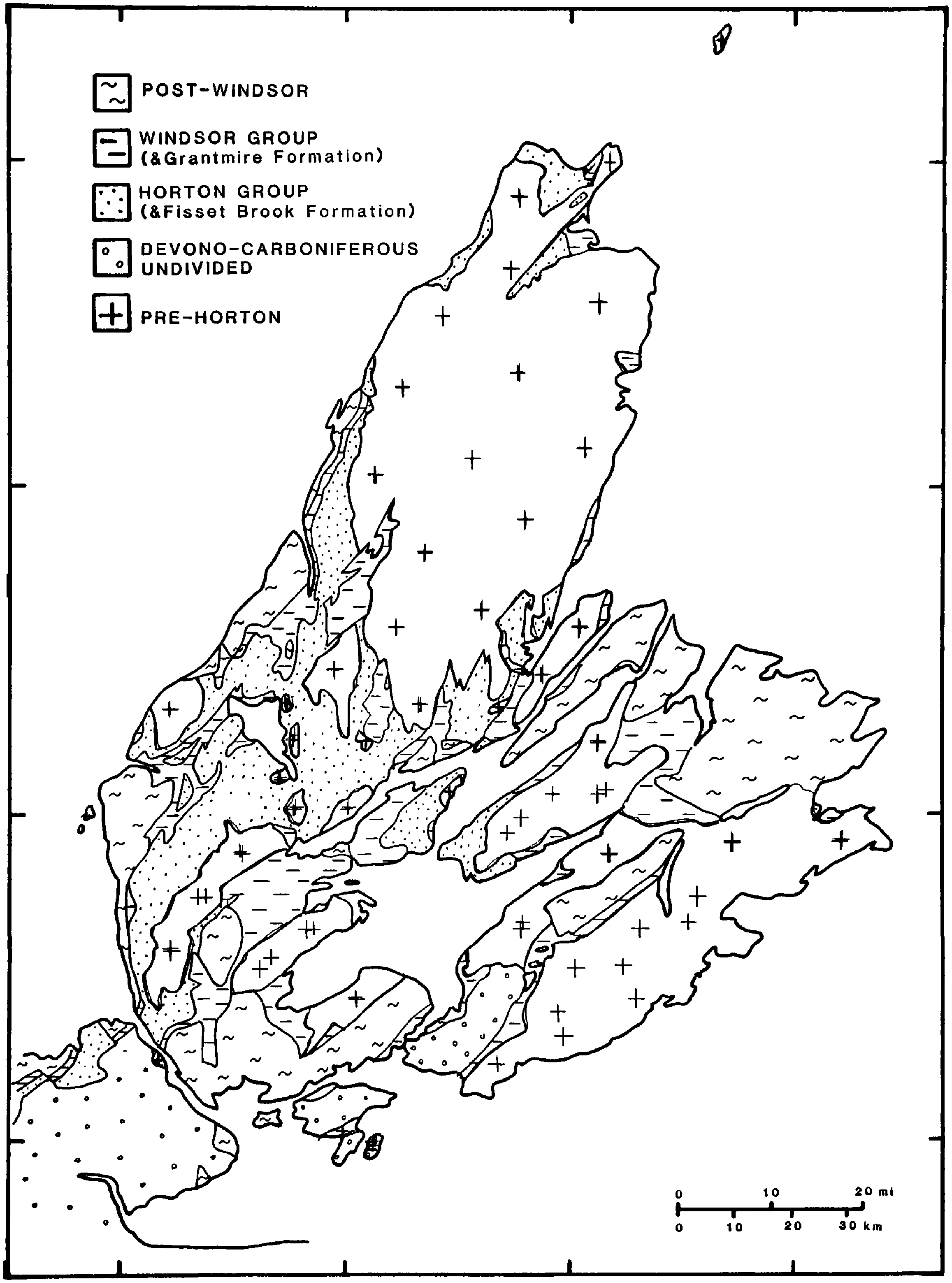

Figure 17. General geology of Cape Breton Island (from Keppie, 1979).

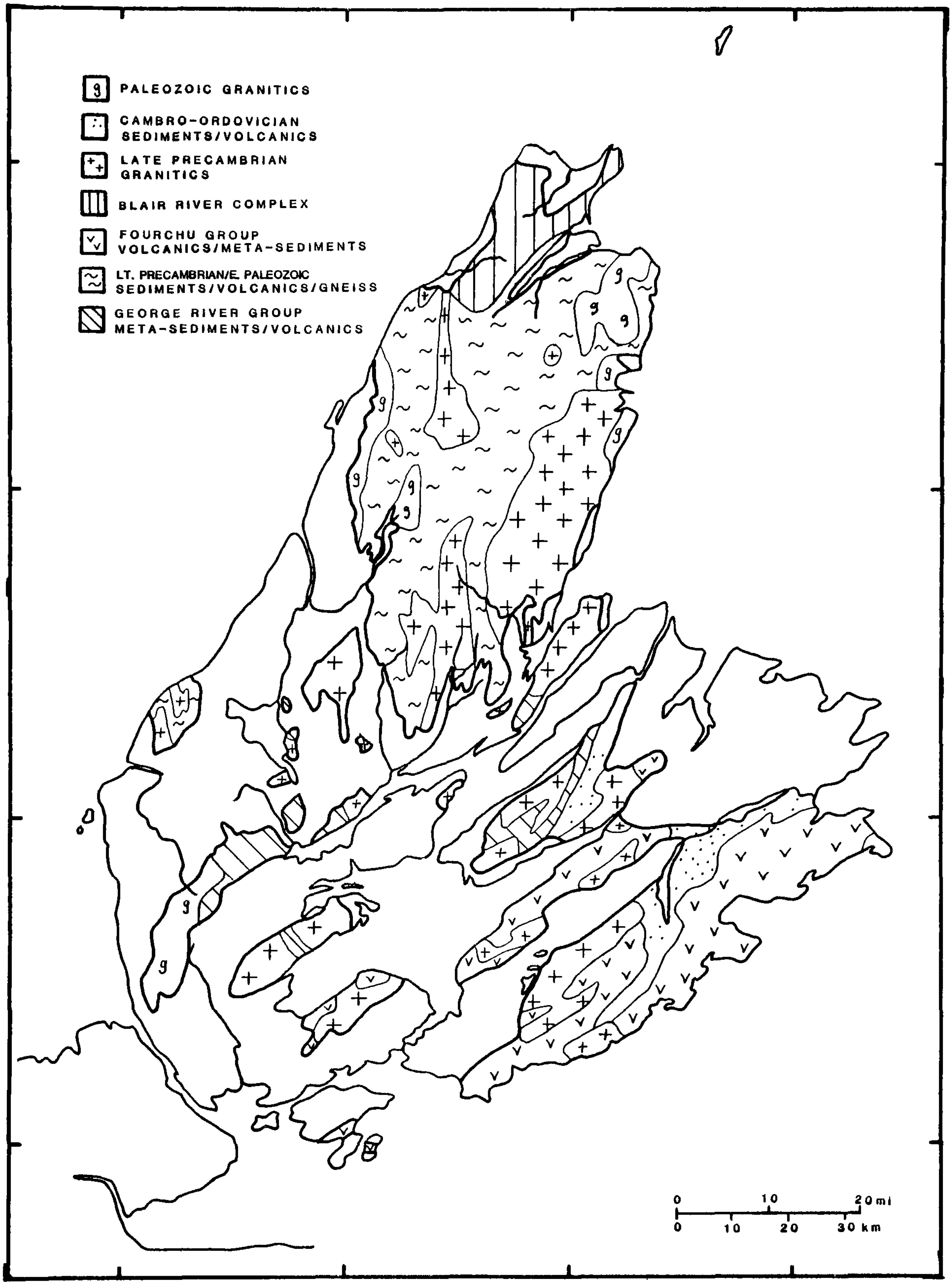

Figure 18. Precambrian and Lower Paleozoic geology of Cape Breton Island (compiled from Weeks, 1954; Keppie, 1979; Barr and Raeside, 1986; Barr et al, 1987)

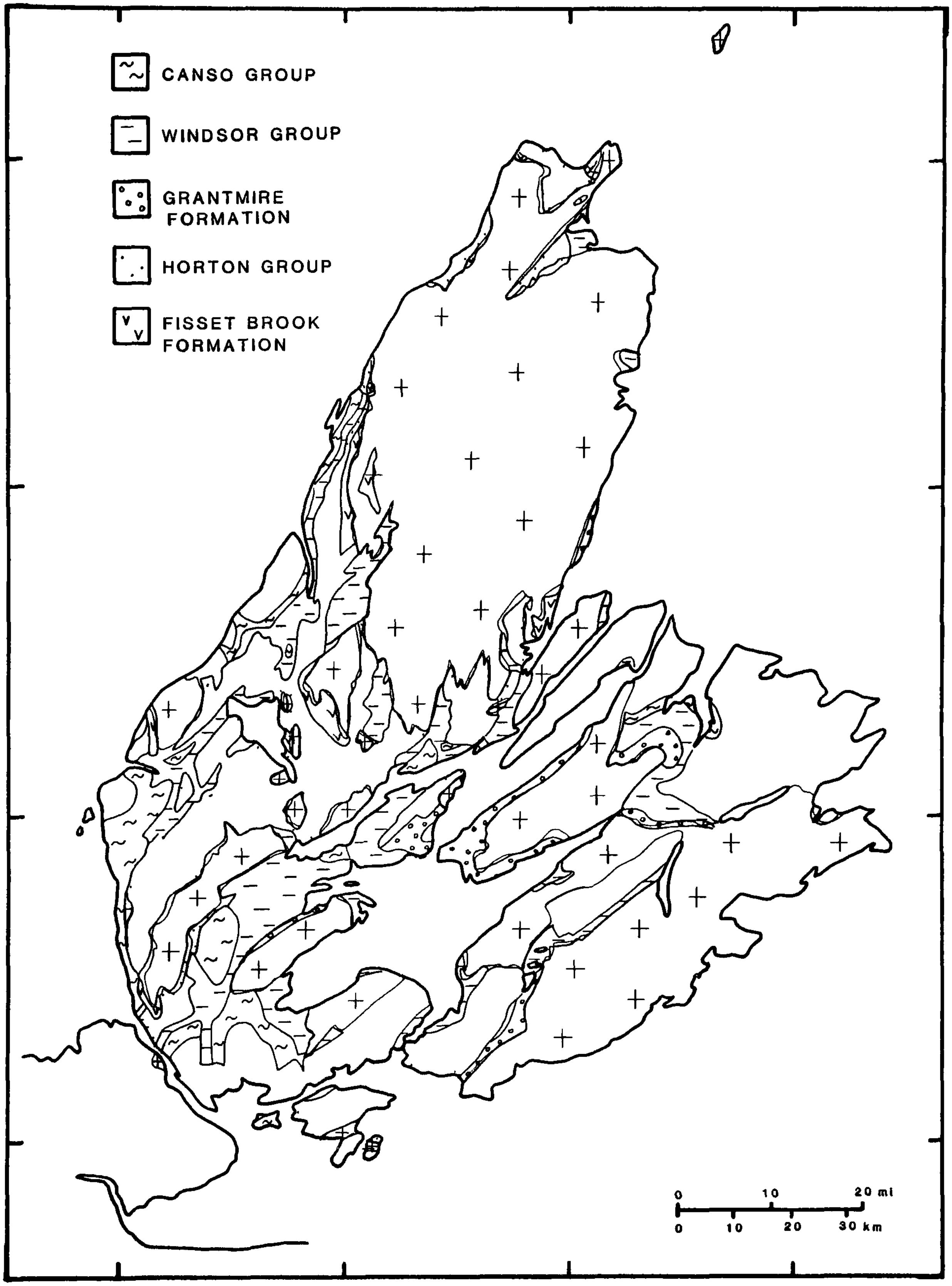

Figure 19. Lower Carboniferous geology of Cape Breton Island (from Keppie, 1979).

Group red and grey siltstone, sandstone and minor limestone conformably overlie the Windsor Group but are sporadically preserved. In outcrops in western Cape Breton there are some small dioritic dykes and sills, presumably of Carboniferous age, which cut lower Horton and Windsor strata.

Upper Carboniferous rocks (Fig. 20) correspond to part of the post-Alleghenian Absaroka sequence of Sloss (1963) and unconformably overlie the Lower Carboniferous sequence. The strata occupy most of the Sydney Basin of eastern Cape Breton Island but have only sporadic occurrence elsewhere. The Riversdale Group is present only in isolated areas around the western coastline (Gibling et al., 1987) where the Port Hood Formation consists of arkosic sandstone and shale with minor coal seams (Norman, 1935; Gersib and McCabe, 1981). The Pictou Group, an important coal-bearing unit of sandstone and shale, is well developed in the Sydney Basin as the Morien Group and in a small area on the west coast (Gibling et al., 1987). No younger bedrock units are present on Cape Breton Island.

DETAILS OF LOWER CARBONIFEROUS STRATIGRAPHY

<u>FISSET BROOK FORMATION</u> A bimodal volcanic suite with interbedded fine-grained sediments, referred to as "pre-Mississippian" or "pre-Horton" in some earlier reports (Norman, 1935; Cameron, 1948; Neale and Kelley, 1960), is exposed in widely scattered outcrops throughout western and northern Cape Breton Island (Fig. 19). It also occurs in the offshore well used in this study north of Cape Breton (St. Paul P-91). Where it occurs this sequence unconformably overlies Acadian basement, and is conformably overlain by the Horton Group. It has generally been referred to as the Fisset Brook Formation with a type section designated on Fisset Brook near Cheticamp (Kelley and Mackasey, 1965). The succession was assigned a ?Late Devonian to Early Carboniferous age based on palynological evidence (Kelley and Mackasey, 1965; Smith and Macdonald, 1982) but K-Ar radiometric dating has yielded a range of ages from 384±10 Ma (Eifelian) to 328±7 Ma (Sepukhovian) with most dates in the 370-375 Ma interval (Frasnian) (Blanchard et al., 1984). Pervasive alteration may affect these data but Blanchard et al. (1984) infer a general younging trend from east to west across Cape Breton. The Fisset Brook Formation is one of several Devono-Carboniferous tholeiitic basalt or basalt/rhyolite volcanic units in Nova Scotia and may represent the middle of 3 extrusive pulses: 1) Middle Devonian McAras Brook Formation of the Antigonish Area, 2) the Late Devonian-Early Carboniferous Fisset

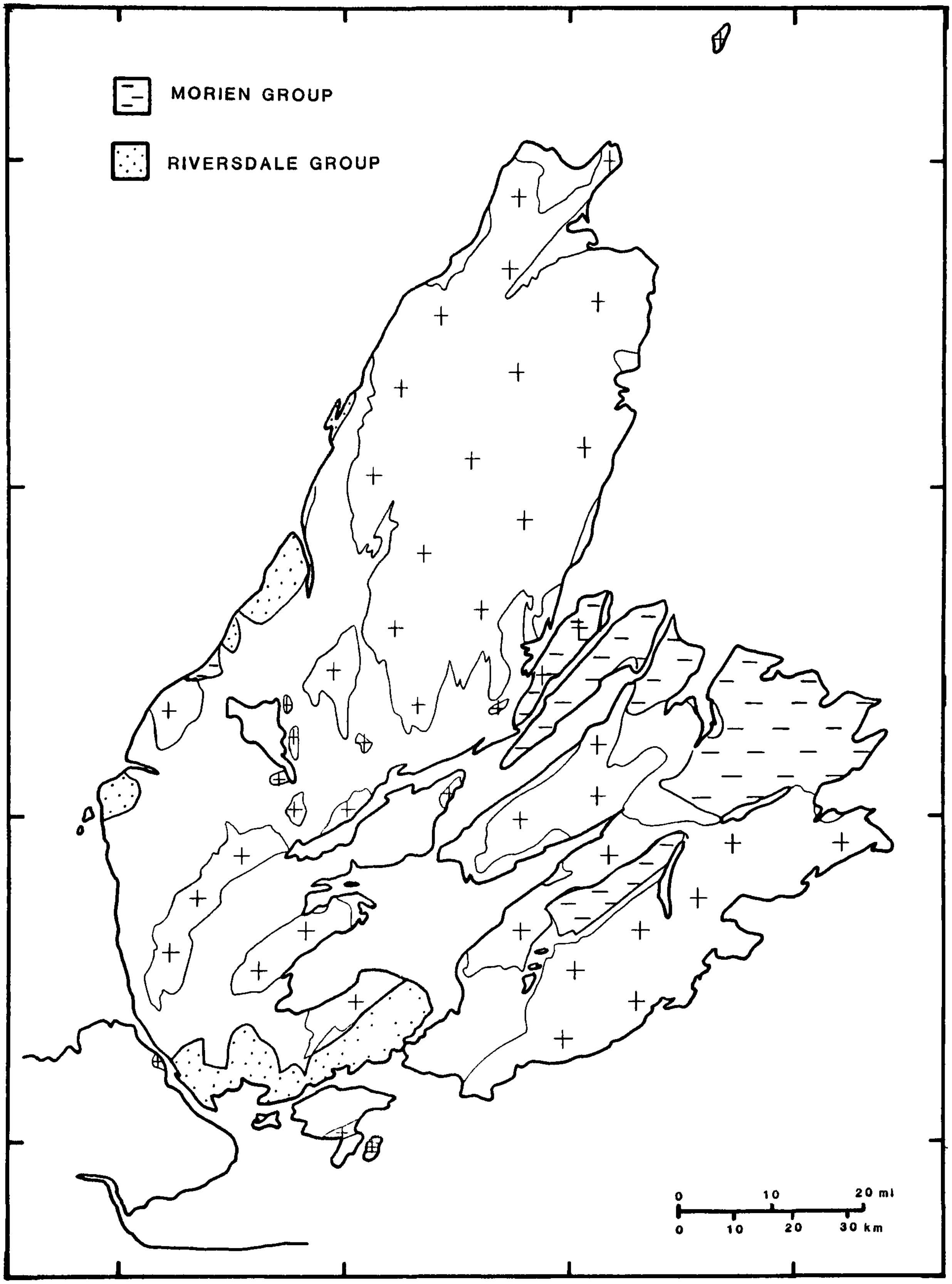

Figure 20. Upper Carboniferous geology of Cape Breton Island (from Keppie, 1979).

Brook Formation of Cape Breton, and 3) the Middle Carboniferous Fountain Lake Group of the Cobequid area (Dostal et al., 1983).

The Fisset Brook Formation outcrops in 3 major areas adjacent to basement blocks; in the East Lake Ainslie area, in a north-south belt between Cheticamp and Margaree, and at Lowland Cove at the northern tip of the Island. One section in each of these areas was reconnoitered, including the type section. The succession consists of sequences up to 500 m thick of thick mafic lava flows, thin felsic ash flows and varying proportions of interbedded sediments (Dostal et al., 1983). The Fisset Brook commonly comprises a thin lower sediment-dominated unit, a thick middle mafic lava unit, and a thick upper felsic volcanic unit (Neale and Kelley, 1960; Kelley and Mackasey, 1965; Blanchard et al., 1984).

The mafic lavas are dark green to maroon, thick bedded (up to 10 m, but typically 2-5 m) (Fig. 21), finely crystalline olivine-clinopyroxene-plagioclase basalts of tholeiitic to transitional character (Dostal et al., 1983; Blanchard et al., 1984; Murphy, 1988). They have irregular bases with cooling rinds and commonly have amygdules filled with chlorite, calcite or sericite (Dostal et al., 1983). No pillows have been reported. Several thin beds of flow breccia were observed on Cooper Brook at the tops of green basalt flows (Fig. 22). Many early reports mentioned andesitic lavas as common constituents (Norman, 1935; Kelley and Mackasey, 1965) but Blanchard et al. (1984) suggested that extensive alteration has enhanced the alkaline character of the basalts. Based on texture, chemistry (TiO_2 vs Zr) and stratigraphic relations Dostal et al. (1983), Blanchard et al. (1984) and Murphy (1988) all suggest the basalts were erupted in a terrestrial setting related to intraplate continental rifting (Fig. 23). This might have occurred in either a distensive (Blanchard et al., 1984) or transtensive (Dostal et al., 1983) stress regime. These conclusions are similar to those reached for similar Tournaisian volcanics in New Brunswick (Fyffe and Barr, 1986) and in the Magdalen Islands (Barr et al., 1985).

The felsic ash flows in the sequence are pink to orange, massive to thinly banded rhyolites, and are well exposed on Cooper Brook (Fig. 24). These are commonly porphyritic with quartz or feldspar phenocrysts set in a welded glassy matrix of quartz and feldspar (Blanchard et al., 1984). Kelley and Mackasey (1965) mention rhyolite breccias. The lack of intermediate compositions suggested to Dostal et al. (1983) that the rhyolite was not derived by differentiation of the mafic magma, but resulted from crustal anatexis during magma ascent.

Figure 21. Thick bedded green basalt flows with interbedded red siltstone, Fisset Brook Formation, Fisset Brook (section 48). Note geologist for scale

Figure 22. Flow breccia at top of basalt flow, with blocks of basalt set in a matrix of red mudstone, Fisset Brook Formation, Cooper Brook (section 24).

Figure 24. Thinly banded rhyolite ash fall, Fisset Brook Formation, Cooper Brook (section 24).

Figure 25. Red siltstone and poorly sorted conglomerate beds between basalt flows, Fisset Brook formation, Fisset Brook (section 48).

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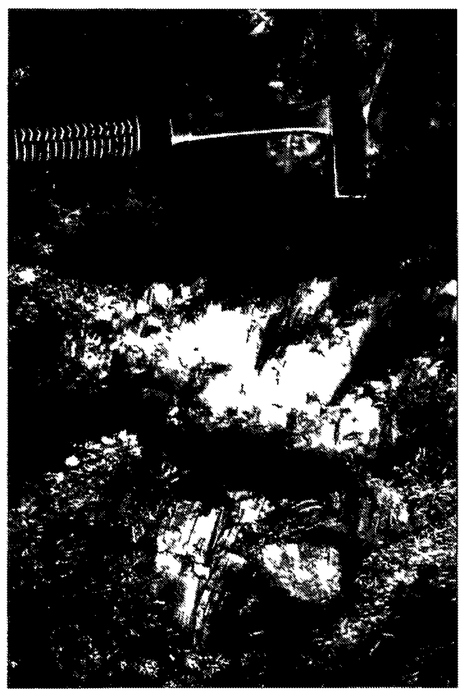

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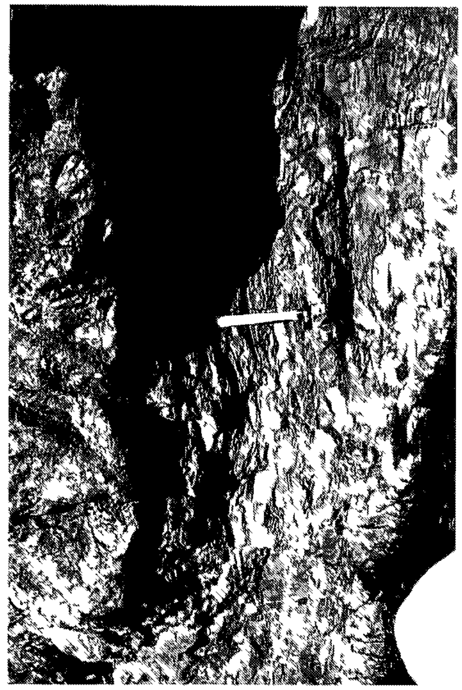

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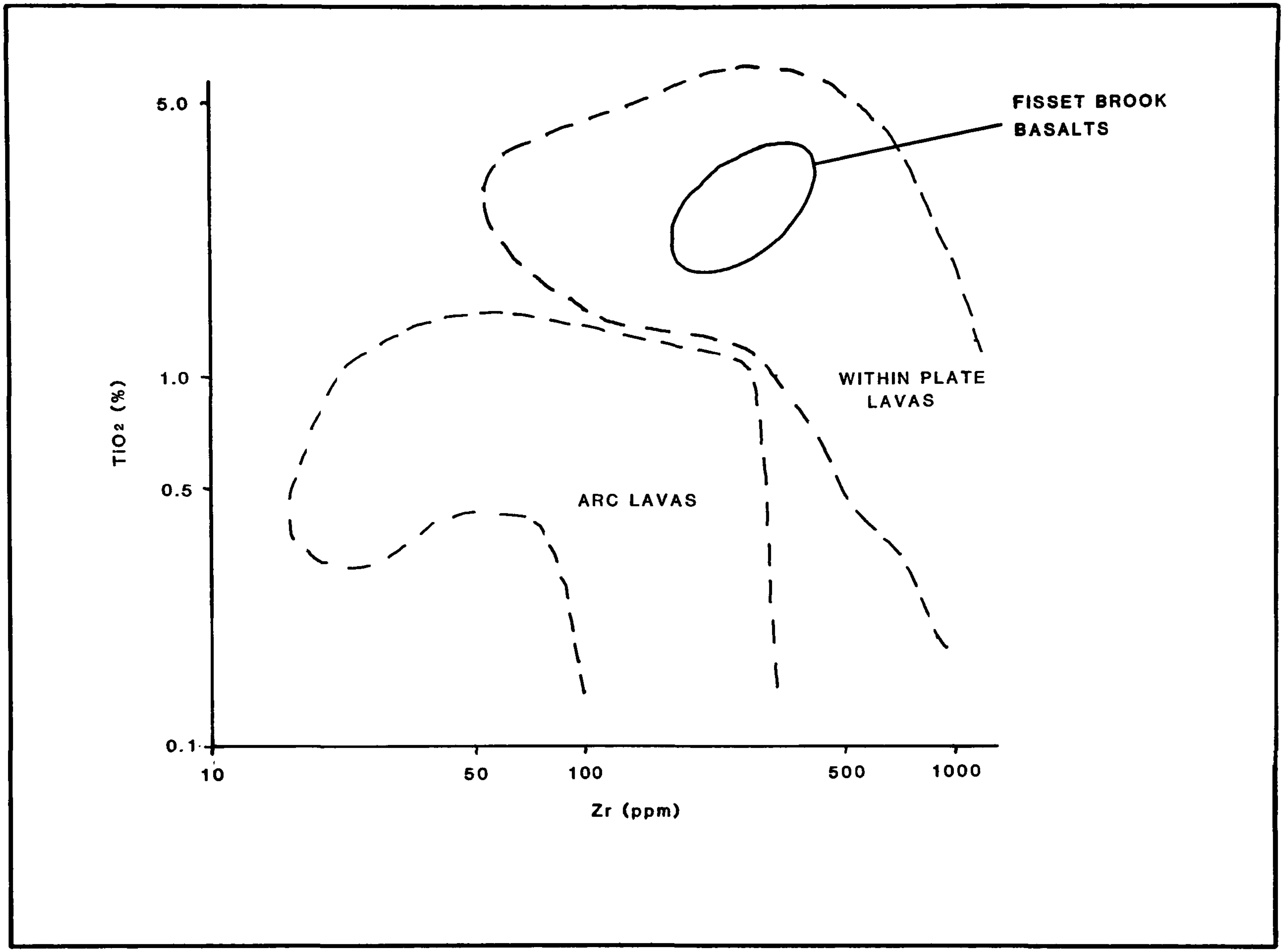

Figure 23. Chemistry and tectonic setting of basalts of the Fisset Brook Formation (modified from Dostal et al, 1983).

The interbedded sediments are typically thin red siltstone to very fine sandstone with some coarser-grained beds present near the unconformable contact with basement. On Fisset Brook red siltstone beds up to 3 m thick between basalt flows are baked to a hard brick-like consistency. A few sharp based poorly sorted conglomerate beds, up to 3 m thick, with basalt clasts set in a fine sandstone matrix also occur (Fig. 25). Plant fragments and spores are commonly present as noted by Kelley and Mackasey (1965), Smith and Macdonald (1982), and Kasper et al. (1988). The character of these sediments, similar to those of the Horton Group, is consistent with deposition in a continental setting. The sequence grades up into the lower Horton Group and the top of the Formation is generally placed at the top of the uppermost volcanic flow (Kelley, 1967).

HORTON GROUP Dawson (1873) introduced the term Horton Bluff Series for beds in the Minas Basin area, recognized as Lower Carboniferous by Logan (1845). Bell (1929) renamed these the Horton Series, of Tournaisian age based on a sparse macrofloral assemblage and designated a type area at Horton Bluff near Wolfville, Nova Scotia. Bell (1929, 1960) defined a) a lower Horton Bluff Formation of red sandstone and grey siltstone up to 1000 m thick resting unconformably on basement (of late Tn_2 to early Tn_3 age, Utting et al., in press) and b) an upper Cheverie Formation of grey sandstone and red siltstone up to 200 m thick (of Tn_3 age, Utting et al., in press), abruptly overlain by the Windsor Group of Viséan age (Table 2). Recent palynological study (Utting, 1987; Utting et al., in press, and this study) has confirmed that all Horton samples throughout the Maritimes are of Tournaisian age. These data are discussed in Chapter 3.

The Horton Group is also present in the Moncton area of New Brunswick where it has a similar stratigraphy (Gussow, 1953; Carter and Pickerill, 1985) (Table 2). At the base of the Group the Late Devonian Memramcook Formation, up to 2000 m of red sandstone and grey siltstone rests unconformably on basement. It is overlain by the Albert Formation (of late Tn_2 to early Tn_3 age, Utting, 1987), up to 1600 m of grey shale, siltstone and oil shale which thickens to the south (Foley, 1987). The succeeding Moncton Formation of red sandstone is up to 1500 m thick and is in turn overlain by the Windsor Group.

In the Antigonish area (Table 2) Murray (1960) described the Ogden Brook Formation of red sandstone and siltstone overlain by the south Lake Creek Formation of red and grey sandstone and siltstone with a distinctive black organic shale, the Big Marsh Member. Murray (1960) and Boehner and Giles (1982) recognized the upper Rights River

STAGE	MONCTON N.B.	MINAS BASIN N.S.	ANTIGONISH N.S.	GUYS BOROUGH N.S.	WESTERN CAPE BRETON N.S.	EASTERN CAPE BRETON N.S.	BAY ST. GEORGE NWFLD.	DEER LAKE NWFLD.
VIS	MACUMBER	MACUMBER	MACUMBER	WINDSOR GP	MACUMBER	MACUMBER	SHIP COVE	?
TOURNAISIAN	MONCTON ALBERT	CHEVERIE HORTON BLUFF ?	RIGHTS RIVER SOUTH LAKE CREEK	HADLEY ROAD Tracadie Road Clam Harbour Road	AINSLIE STRATHLORNE	HORTON UNDIVIDED	SPOUT FALLS FRIARS COVE SNAKES BIGHT KENNEL'S BROOK	CAPE ROUGE C3 C1 &C2
FAMMENIAN	MEMRAMCOOK ?		OGDEN BROOK ?	GLENKEEN St.Francis	CRAIGNISH	?	?	?
FRASNIAN				?	.FISSET BROOK			
GIVET					?			

Table 2. Formations of the Horton Group throughout the Maritimes Basin (compiled from Weeks, 1954; Bell, 1960; Murray, 1960; Hyde, 1978; Smith, 1981; Knight, 1983; Carter and Pickerill, 1985).

Formation red conglomerate, sandstone and siltstone (coarser-grained to the north) of late Tn_2 to early Tn_3 age (Utting et al., in press), which is overlain by the Windsor Group. In the Strait of Canso area Ferguson (1946) assigned 1100 m of metamorphosed and deformed grey conglomerate, sandstone and shale to the Horton Group. In the Guysborough area (Table 2) Smith (1981) described the intertonguing Glenkeen and St. Francis Formations of grey conglomerate, sandstone and siltstone overlain by the Clam Harbour Formation black shale, and the upper intertonguing Hadley Cove and Tracadie Road Formations of red and grey sandstone and siltstone. These are overlain by the Windsor Group.

In western Newfoundland Hyde (1978, 1979, 1983) and Knight (1983) have described the stratigraphy of the correlative Anguille Group (Baird and Côté, 1964) (Table 2). The sequence is up to 4500 m thick, rests unconformably on basement and is conformably overlain by the Codroy Group (Windsor equivalent, Utting, 1980). The Anguille of the St. Georges Bay area (100 km from Cape Breton Island) includes the Late Devonian Kennel's Brook grey sandstone and red siltstone up to 3200 m thick, overlain by the Snakes Bight Formation dark grey siltstone up to 1000 m thick, in turn overlain by the intertonguing Friars Cove and Spout Falls Formations of red and grey conglomerate, sandstone and siltstone up to 2200 m thick.

On western Cape Breton Island (Fig. 26) Norman (1935) and Cameron (1948) divided the Horton into Upper and Lower Groups by correlation with Bell's type area. Their Lower Group comprised up to 700 m of grey arkosic sandstone with pebbly conglomerate and red shale interbeds, while their Upper Group comprised 900 m of grey shale with sandstone interbeds overlain by grey or red sandstone with shale interbeds. This was sharply overlain by the basal Windsor Group limestone, generally in a conformable relation but unconformably near the basin edge.

Murray (1955, 1960) introduced the tripartite stratigraphic scheme in general use today (Table 3), designating with Southwest Mabou River as his type section. The Craignish Formation consists of a lower Skye River Member of grey sandstone and conglomerate and red siltstone 900 m thick, and upper McLeod Member of red siltstone and sandstone 650 m thick. The Strathlorne Formation comprises grey siltstone, shale, fine grained sandstone and minor limestone 300 m thick. Plant and fish fossils are present in places. The Ainslie Formation includes a lower McIsaac Point Member comprising fining-upward sequences of grey medium-grained sandstone to siltstone, and upper Glencoe

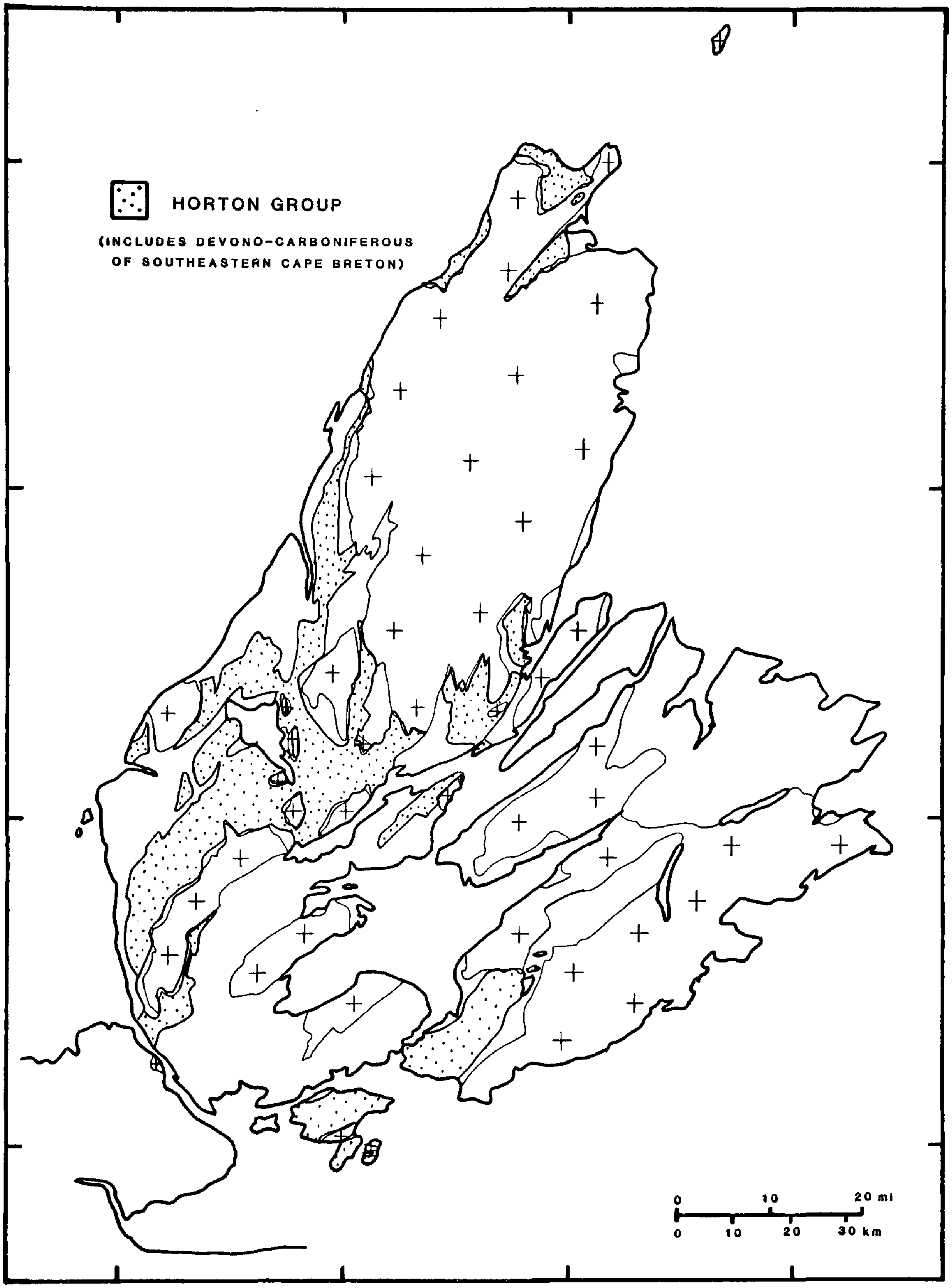

Figure 26. Geology of the Horton Group, Cape Breton Island (from Keppie, 1979).

Member of interbedded calcareous grey sandstone, red shale and limestone. The Ainslie is apparently conformably overlain by the basal Windsor limestone.

Neale and Kelley (1960) studied the Horton Group of northern Cape Breton Island (Fig. 26) and recognized the same three-part stratigraphy, sharply but conformably overlain by the basal Windsor Group. The Craignish Formation consists of up to 1600 m of red and grey conglomerate, sandstone and siltstone with a few plant fragments. The Strathlorne Formation comprises up to 500 m of grey shale, siltstone, sandstone and minor limestone and conglomerate, which coarsens to the southeast, with abundant plant fossils. The Ainslie Formation is poorly exposed but includes up to 500 m of grey and red sandstone and siltstone with minor conglomerate.

In central Cape Breton Island (Fig. 26) Kelley (1967) made several modifications to Murray's stratigraphic scheme. In the Craignish Formation he recognized an additional basal unit of limited lateral extent present only in the basin centre area (Table 3). This Graham River Member is 500 m of grey conglomerate and sandstone with some green diabase dykes and sills. For mapping purposes he used a Strathlorne-Ainslie Formation although he conceded that at any section these intertonguing facies were usually separable. He also identified the importance of red conglomerate in the Ainslie in the Bras d'Or Lakes area.

In eastern Cape Breton Island (Fig. 26) rocks assigned to the Horton Group are present in only a few areas. Weeks (1954) described a thick sequence of deformed and metamorphosed red and grey conglomerate, sandstone and siltstone beneath the Windsor Group in the Isle Madame/L'Ardoise area. From structural evidence he concluded that this represents a large overthrust sheet transported from the south to its present position. More recently Boehner and Giles (1982) reported Horton clastics beneath the Windsor Group in a few outcrops near the margins of Sydney Basin. Although eastern Cape Breton Island was apparently an area of Horton deposition, it was not included in this study.

WINDSOR GROUP The Windsor Group (Table 1), of Viséan age (Utting, 1978, 1980), is a variable sequence of fossiliferous marine carbonates, evaporites and minor red clastics distributed throughout Atlantic Canada. It conformably to unconformably overlies the Horton Group and is conformably overlain by the thick nonmarine clastics of the Canso Group. It represents the only significant marine deposits of the Upper Paleozoic of Atlantic Canada. Giles (1981) identified a basic rhythmic organization of 5 Major Cycles

AGE MA	PERIOD	EPOCH	STAGE	FORMATION (age ranges poorly defined)	MEMBER (applicable near basin centre)
350	CARBONIFEROUS	EARLY	VISEAN	MACUMBER (WINDSOR GROUP)	
			TOURNAISIAN	AINSLIE	Glencoe
					McIsaac Point
355				STRATHLORNE	
				CRAIGNISH	McLeod
360	DEVONIAN	LATE	FAMENNIAN		Skye River
365					Graham River
			FRASNIAN	FISSET BROOK	rhyolitic unit
370					basaltic unit
					basal coarse sediments
375		MIDDLE	GIVETIAN	BASEMENT	

Table 3. Stratigraphy of the Horton Group, Cape Breton Island (from Murray, 1960; Kelley, 1967).

characterized by lithological and paleontological features. Each successive cycle has a wider distribution and onlaps the preceding one. The cycles are attributed to transgressive-regressive shallowing-upward phases of sedimentation, bounded by time planes, attributed to eustatic sea level changes associated with episodes of Late Paleozoic Gondwanan glaciation (Giles, 1981).

Major Cycle 1 (A subzone of Bell, 1929) is the thickest cycle and is of significance to this study as the easily-recognized unit immediately overlying the Horton Group in all areas (Table 2). Boehner (1986) described this cycle as 300 to 600 m thick and consisting of a basal laminated dolostone 3-50 m thick (the Macumber Formation) overlain by up to 300 m of massive anhydrite and up to 300 m of stratified halite and red siltstone. Near basement blocks or basin margins a coarse red clastic facies interfingers with these lithologies.

In the study area only the basal Macumber Formation (the A1 carbonate of Bell, 1929) is well exposed and was considered as a convenient datum plane, following the conclusions of Geldsetzer (1977), Kirkham (1978) and Giles (1981). It is generally about 10 m of dark grey thinly laminated, organic-rich dolomitic marlstone. Fossils are uncommon, though Kelley (1967) mentions ostracods and gastropods, and 2 poorly preserved conodont elements were recovered from a sample in this study (a second sample was barren). These conodonts were assigned to the Diplognathus Zone at the base of the Windsor Group by Von Bitter (1988, pers. comm.). The contact with the Horton Group is always very sharp with a slightly irregular topography but nearly always appears to be conformable (Fig. 27). The lower few centimeters of the Macumber commonly have a pelleted or oölitic texture, and may contain scattered sand grains. The only outcrop location with a demonstrable angular unconformity at this level is just west of the Mabou Highlands basement block, where the Macumber overlies steeply west-dipping Horton red conglomerate (Figs. 28, 29). Palynological evidence from the Macumber (Utting, 1980; Utting et al., in press) indicates a Viséan V_2/V_3 age, suggesting a significant time gap between Horton and Windsor Groups (see Chapter 3).

Geldsetzer (1977) identified a western laminite facies (the Macumber Formation as described above) and a contemporaneous eastern fossiliferous carbonate shoal facies (Gays River Formation) separated by a north-south facies boundary just east of the inferred upper Horton basin margin of this study. The Grantmire Formation (Weeks, 1954), which

Figure 27. Sharp, but apparently conformable, contact of Horton Group conglomerate and Macumber Formation laminated marlstone, MacRae Brook (section 28).

Figure 28. Angular unconformity between Horton Group red conglomerate and Macumber Formation, Mabou Mines area.

Figure 29. Close-up of unconformity between Horton Group and Macumber Formation, Mabou Mines area.

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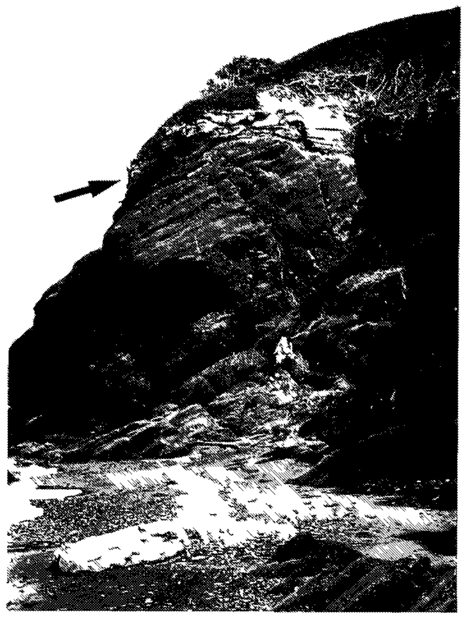

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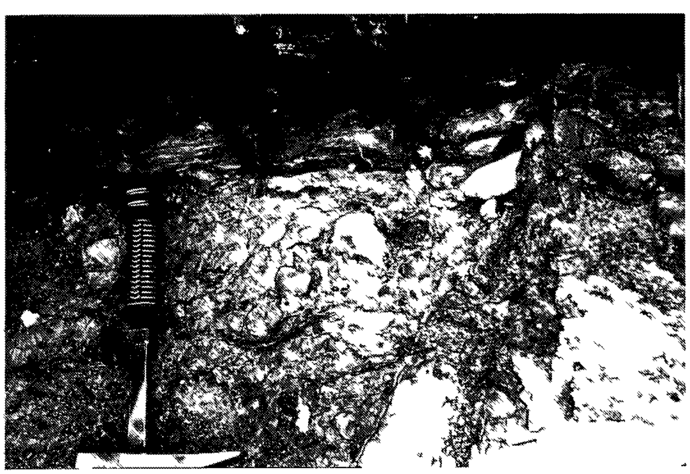

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comprises coarse red clastics, represents a time-transgressive basin margin facies present in central and eastern Cape Breton. The Grantmire was defined by Weeks (1954) as of Windsor age, but is lithologically identical to the upper Horton Group of that area, which has led to much confusion in the area close to the basin margin. There, in the absence of the Macumber Formation, upper Horton and Grantmire clastics are indistinguishable (Bell, 1958; Kelley, 1967). Kelley (1967), Geldsetzer (1977) and Boehner (1981) attribute much of the originally mapped Grantmire to the Horton Group.

The Cycle 1 sequence of the Windsor Group is interpreted as representing rapid marine transgression into a pre-existing sub-sea level continental basin, with increasing saline stratification and evaporite deposition (Geldsetzer, 1977; Kirkham, 1978; Giles, 1981). The cause of this rapid transgression may have been tectonic breaching of a presumed barrier (Kirkham, 1978) or eustatic seal level rise (Geldsetzer, 1977) due to early glacial episodes in the Carboniferous-Permian glaciation (Crowell, 1978; Giles, 1981). Von Bitter et al. (1988) interpret the Macumber as representing deposition in relatively deep water. The overlying Cycles 2 to 5 (subzones B to E of Bell, 1929) are progressively thinner but more extensive sabkha-mudflat deposits (Boehner, 1986).

CHAPTER 3
DESCRIPTIONS OF FACIES ASSEMBLAGES

INTRODUCTION

The tripartite stratigraphic scheme of Murray (1960) is obvious in the field and can be recognized at most extensive outcrops throughout Cape Breton Island. It represents the starting point for any study of the complex Horton Group. Initially an attempt was made to view the Horton Group as a collection of diachronous facies without the constraint of existing ("layer-cake") stratigraphy. However the two concepts can be related. In this study, the lithostratigraphic formations of Murray (1960) are considered as large-scale lithofacies associations which represent major tectonically-controlled three-dimensional depositional systems, with non-horizontal, probably diachronous boundaries. However, because these basic units have basinwide occurrence in an invariably consistent stratigraphic order, they cannot simply be considered as diachronous lithofacies. They also appear to have some temporal connotations and must represent general tectonic phases of basin evolution through time. In fact, correlative units can be designated in most areas of Horton deposition in Atlantic Canada, confirming the insight of previous workers and adding new significance to these divisions.

The Craignish, Strathlorne and Ainslie formations are thus considered in this study as four-dimensional units which can be interpreted in light of the spatial and temporal tectonic history of Cape Breton Island. Toward definable paleo-basin margins it appears that the Horton Group thins as lower units progressively pinch out against controlling basement blocks. A zero edge of Horton deposition is eventually reached, probably on all sides, beyond which overlying Windsor Group units onlap onto basement and similarly pinch out from the base.

On a gross scale the formations, in their characteristic stratigraphic order delineate sedimentological trends, most obvious in western Cape Breton. The Craignish Formation commonly displays an overall fining-upward trend and may represent a single depositional package. Similarly the Strathlorne and Ainslie Formations together represent an overall coarsening-upward trend and likely represent another depositional package. This is similar to the idea expressed by Kelley (1967) that the Strathlorne and Ainslie are intertonguing members of a single formation.

Within the original stratigraphic schemes of Murray (1960) and Kelley (1967) several members were designated in each formation. These are valid in local areas but the specific lithological definitions are less easily applied on a regional scale, as pointed out by Kelley (1967). For the purposes of this study I have designated several facies assemblages within each formation (Table 4). These assemblages represent specific depositional settings arranged laterally within the basin during the phase of basin evolution represented by that formation. The boundaries of facies assemblages are non-horizontal and are probably diachronous. The concept of facies assemblages is more flexible and more representative of the basinwide diversity of Horton deposits than stratigraphic members would be, and more clearly leads to interpretation of the tectonic evolution and facies distributions within depositional basins. In Table 4 and the rest of this thesis each facies assemblage is designated with a letter-number system (indicating the formation to which it belongs and its interpretive importance) and is named using a shorthand system indicating predominant colour and grain size range. Interpretations discussed in Chapters 4 and 5 are also included. Each facies assemblage comprises a series of individual lithofacies which are described in a general way in this chapter.

The same facies assemblages can be found in both the western and northern Cape Breton areas, but there are some differences. Overall, the Horton Group of the western area is more fully exposed, although individual outcrops are commonly of better quality in the northern area. As discussed more fully in Chapters 4 and 5, there appear to be differences in the distribution of facies assemblages and predominant sedimentary sources for these two areas, and they are interpreted as representing two separate sub-basins of deposition, for at least part of Horton time. The descriptions which follow indicate the characteristics of each facies assemblage and refer to the differences between the two sub-basinal areas. For the sake of clarity, the descriptions are couched in terms of the depositional position of the facies assemblages with respect to the original margins of these two sub-basins, which are defined in Chapter 4.

Petrographic description was not a major objective of this study and only a few comments on petrography are included in these descriptions (summarized in Table 5). The rocks are extremely complex and variable in thin section, including abundant multi-cycle lithic fragments and several generations of cement. The complexity is due to several factors: a) the wide, and as yet not fully studied, diversity of basement source rocks over

FORMATION	FACIES ASSEMBLAGES	POSITION
AINSLIE	A1 red/grey pebbly coarse ss-cgl (distal alluvial fan/proximal braidplain)	↑ basin margin
	A2 red fine-coarse ss & siltstn (low sinuosity fluvial)	
	A3 grey/green fine ss&siltstn (high sinuosity fluvial)	↓ basin centre
STRATHLORNE	S1 dark grey mudstone (open lacustrine)	↑ basin centre
	S2 grey/green very fine-fine ss (deltaic nearshore/shoreline)	
	S3 grey/green medium ss- boulder cgl (fan delta)	
	S4 red siltstn-fine ss (coastal mudflat)	↓ basin margin
CRAIGNISH	C3 grey/green coarse ss-granulestone (distal gravelly braidplain)	↑ basin centre
	C2 brick red siltstn-fine ss (mudflat/playa)	
	C1 red/orange coarse ss-cgl (distal-medial gravelly braidplain)	↓ basin margin

Table 4a. Formations of the Horton Group on Cape Breton Island and facies assemblages defined in this study.

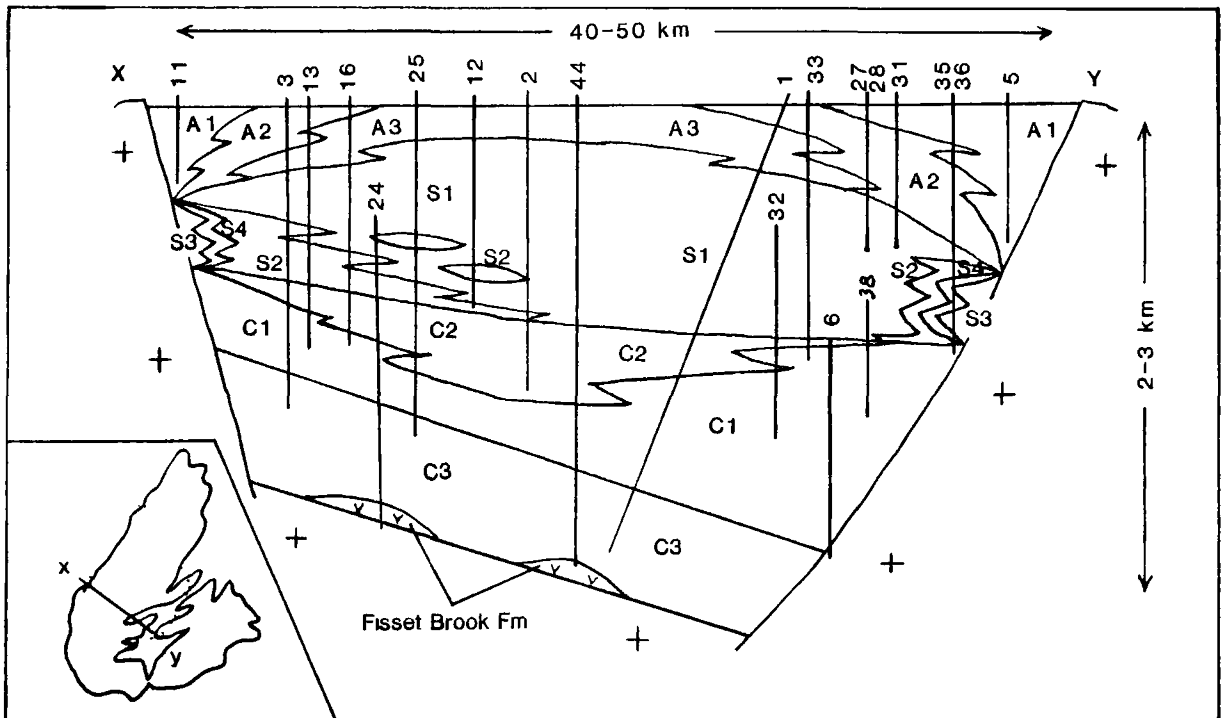

Table 4b Selected measured sections (numbered as in App I) showing control on distribution of facies assemblages and sub-basin geometry in western Cape Breton. See Chapters 4 and 5 for further discussion.

FORMATION	FACIES ASSEMBLAGE	GRAIN SIZE	QUARTZ %	FELDSPAR %	ROCK FRAGMENTS %	MATRIX/ CEMENT	CLASSIFICATION (Pettijohn, 1973)
AINSLIE	A1	css-cgl	10-60	0-10	50-90	0-20	litharenite
		f-mss	65-90	0-15	0-20	5-10	quartz arenite-sublith/subfeldsarenite
	A2	vf-fss	75-80	0-10	0-20	10-20	sublitharenite
	A3	vf-fss	85-95	0-5	0	5-10	quartz arenite
STRATHLORNE	S1	si-vfss	60-80	0	0	20-40	quartz arenite
	S2	vf-fss	80-95	0-10	0-5	5-10	quartz arenite-subfeldsarenite
	S3	f-css	80-90	10-20	0	10	quartz arenite-subfeldsarenite
	S4	si-vfss	85	0	0	15	quartz arenite
CRAIGNISH	C1	css-cgl	60-80	10-30	10-30	0	sublith/subfeldsarenite
		f-mss	40-60	10	10-50	10-20	litharenite-sublitharenite
	C2	vf-fss	70	20	0	10	feldsarenite-subfeldsarenite
	C3	m-css	70-90	5-25	5-20	5-10	quartz arenite-sublith/subfeldsarenite

Table 5. Summary of petrographic data for each facies assemblage.

a large area, b) probable common localized cannibalization of earlier Horton sediments, c) general immaturity of Horton sediments, d) distinct dependency of mineralogical content on grain size, and e) complex post-depositional history.

One thousand paleocurrent measurements were obtained during field work, comprising 225 from the Craignish, 432 from the Strathlorne, and 344 from the Ainslie Formations. The data represent all facies assemblages described and are presented as rose diagrams on basemaps in this chapter. Considering the volume of rock studied the data are sparse but are nonetheless useful in indicating basic sedimentological relations. They are important in the mapping and interpretation of facies assemblages, and hence in interpreting the separate sub-basins of deposition.

CRAIGNISH FORMATION

The Craignish Formation, up to 2000 m thick (Fig. 30) locally overlies the Fisset Brook Formation conformably, but in most areas unconformably overlies basement. The great thickness recorded in certain areas may be partly due to thrust repetition, and the maximum thickness may be less than 2000 m. For this project the formation has been studied at 37 outcrops and in 3 drillholes, although most expose only a small portion of the entire unit. In central basin positions (eg: Mabou area of western Cape Breton) there is a general trend to fining-upward and reddening-upward. Near original basin margins (eg: Baddeck area of western Cape Breton) where the Craignish progressively pinches out from the base, reddened coarse sediments are typical throughout. The three-dimensional arrangement and characteristics of facies assemblages is obscured by poor exposure in western Cape Breton and structural complications in northern Cape Breton. The Craignish Formation can be divided into 3 facies assemblages (Table 5).

C1 Red/Orange coarse ss-cgl This facies assemblage is up to 1200 m thick (Fig. 31), though 100-300 m of exposed thickness is more typical. It was described at 26 outcrop and 3 drillhole sections. It is a common facies assemblage in the upper Craignish near the margins of the western Cape Breton sub-basin, (eg: Baddeck, Little Narrows, Whycocomagh areas) and in the lower and middle Craignish near the margins of the northern Cape Breton sub-basin (eg: Pleasant Bay, Lowland Cove areas). Figure 32 presents a typical example from Baddeck River. This assemblage forms part of the McLeod Member of Murray (1960) and Kelley (1967) (Table 3) and intertongues with the red

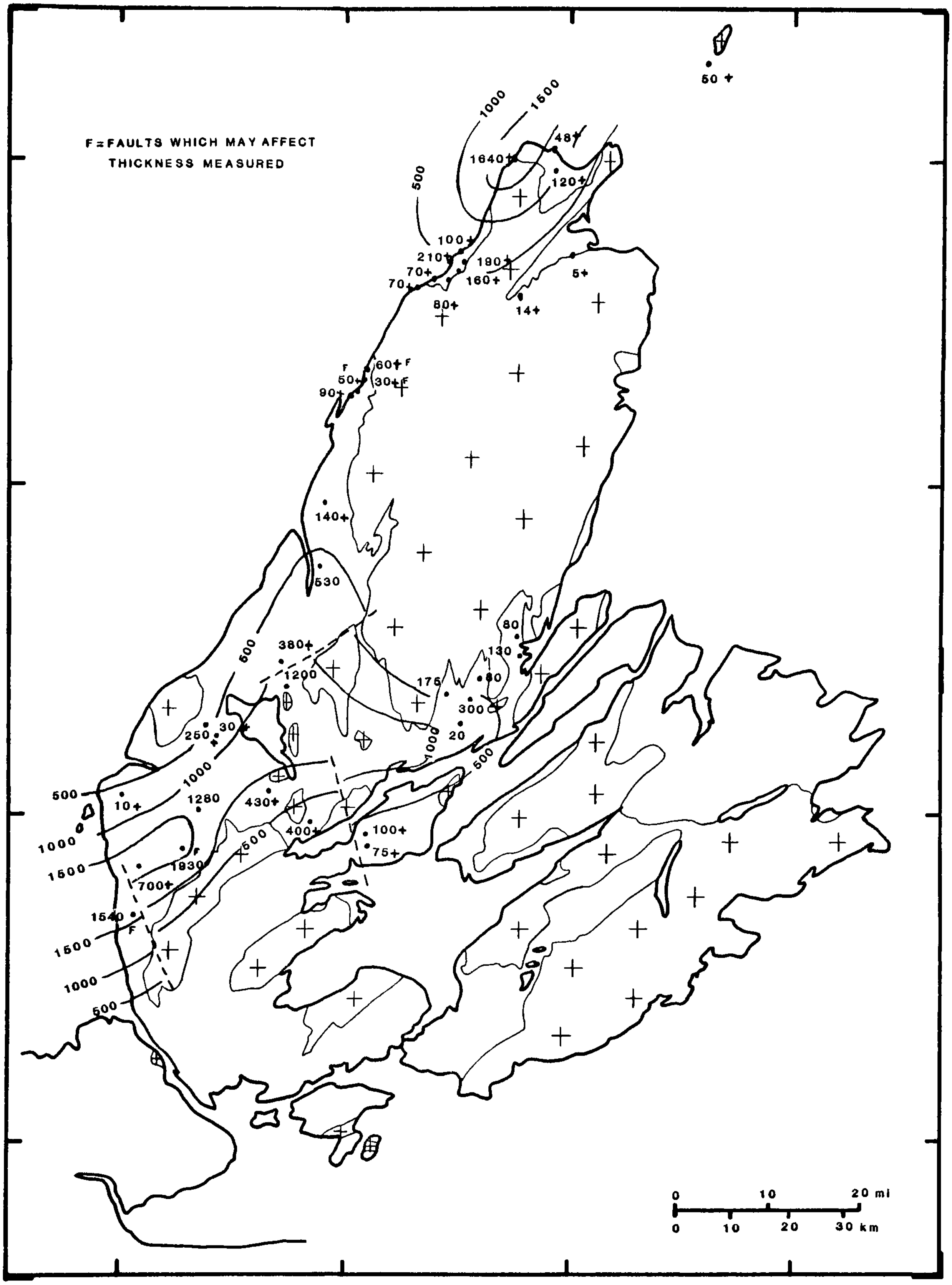

Figure 30. Isopach map, Craignish Formation. Thickness values in metres.

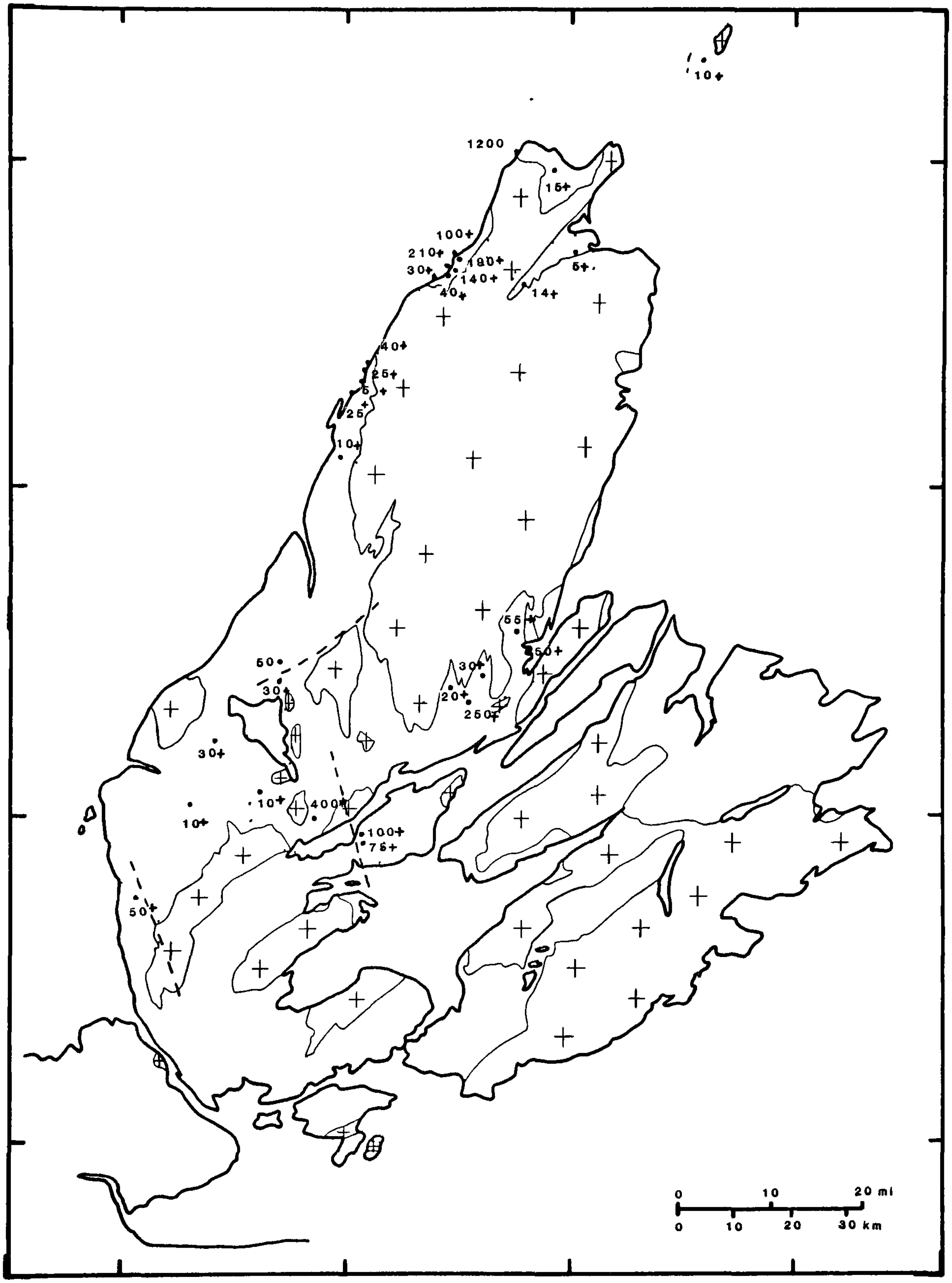

Figure 31. Isopach map, C1 facies assemblage. Thickness values in metres.

KEY TO TYPE EXAMPLE SECTIONS

txb	trough cross stratification
rxl	ripple cross lamination
ripp crest	ripple crest orientation
imb	imbrication
scour	scour orientation
tool	tool mark orientation
curr lin	current lineation

hummocky cross stratification

contorted lamination

burrows

limestone

desiccation cracks

granules

pebbles/boulders

evaporite crystal molds

calcrete

roots

rip-ups

horizontal lamination

low angle lamination

KEY TO MAPS

basement blocks

inferred faults which offset facies

85+ minimum thickness exposed

approximate isopach or zero line

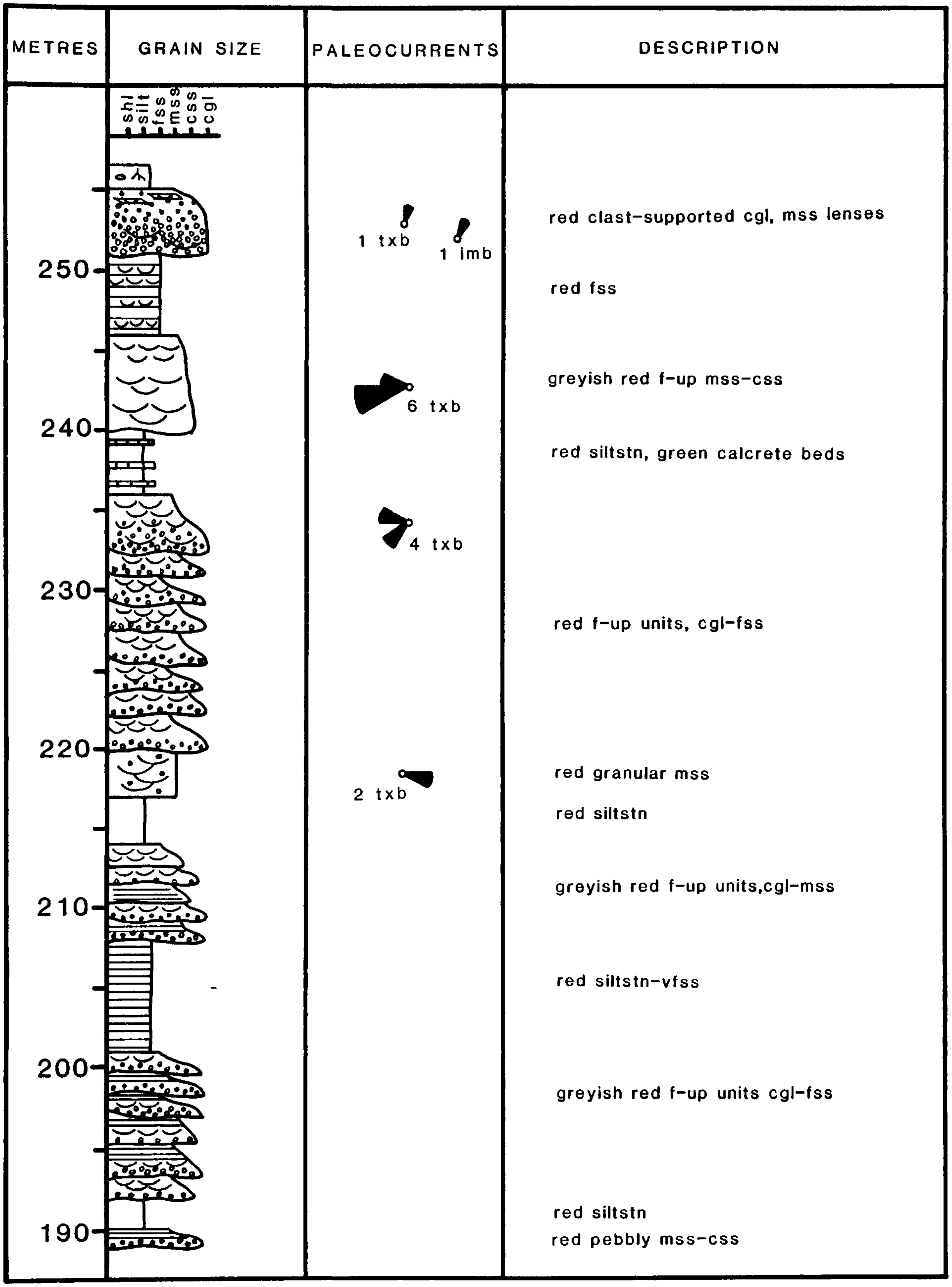

Figure 32. Type example, C1 facies assemblage, Baddeck River.

siltstone and fine sandstone of facies assemblage C2. A total of 102 paleocurrent measurements were collected from 15 sections, mostly from trough cross bedding (Fig. 33).

The predominant lithofacies is red to reddish grey or orange grey, well stratified fine- to coarse-grained sandstone, granulestone and conglomerate, most typically pebbly medium sandstone to granulestone. Beds are 1.0 to 5.0 m thick (typically 1-2 m), fine upward from sharp flat or scoured bases to fairly sharp tops and in some good exposures are seen to pinch out laterally. The beds are uniform in aspect, with internal horizontal stratification or medium-scale trough cross stratification (Fig. 34). Pebbles occur in sandstones as thin lags and large scour pockets at scoured bases (Fig. 35) or as discrete thin strata, or dispersed throughout the bed. Conglomerate beds are generally supported by a matrix of coarse sandstone to granulestone. Most are pebbly, micaceous and have poor to fair sorting. Sandstone and conglomerate commonly occur in discrete packages up to 100 m thick (generally 20-50 m) of stacked beds, separated by finer-grained red sediments (eg: Baddeck River, Fig. 32). The red colour, due to hematite rims around grains, is typical of the western area, and the orange colour, due to high feldspar content and hematite, is typical of the northern area. At the top of the Craignish Formation, immediately beneath the Strathlorne Formation, grey colours are consistently present.

In thin section these rocks have angular to subangular, equant to elongate framework grains of quartz and rock fragments (granite, quartzite, schist, marble) and are classified as litharenite/sublitharenite or subfeldsarenite. In the northern area orange, partially altered K-feldspar is very prominent, comprising up to 50% of framework grains. The rocks are relatively immature, with variable amounts of sericitized matrix or pseudomatrix, but may display bimodal size distribution. In red samples the most obvious characteristic is the ubiquitous thick hematite coating on grains, with later silica, calcareous and clay cements (Fig. 36). In orange samples abundant K-feldspar and minor hematite are characteristic. All primary porosity is occluded, but up to 3 % secondary porosity is present, due to removal of K-feldspar, especially in finer grain sizes and near the top of the Craignish.

Coarse beds are commonly separated by brick red to maroon to greyish red beds and lenses of siltstone to medium sandstone. These units range up to 10 m thick (generally 0.5-2.0 m), have sharp bases and tops, fine upward, and in some cases represent the upper parts of coarser units. They are generally well sorted, and calcareous, and have internal horizontal or trough cross stratification and ripple cross lamination. They may have very

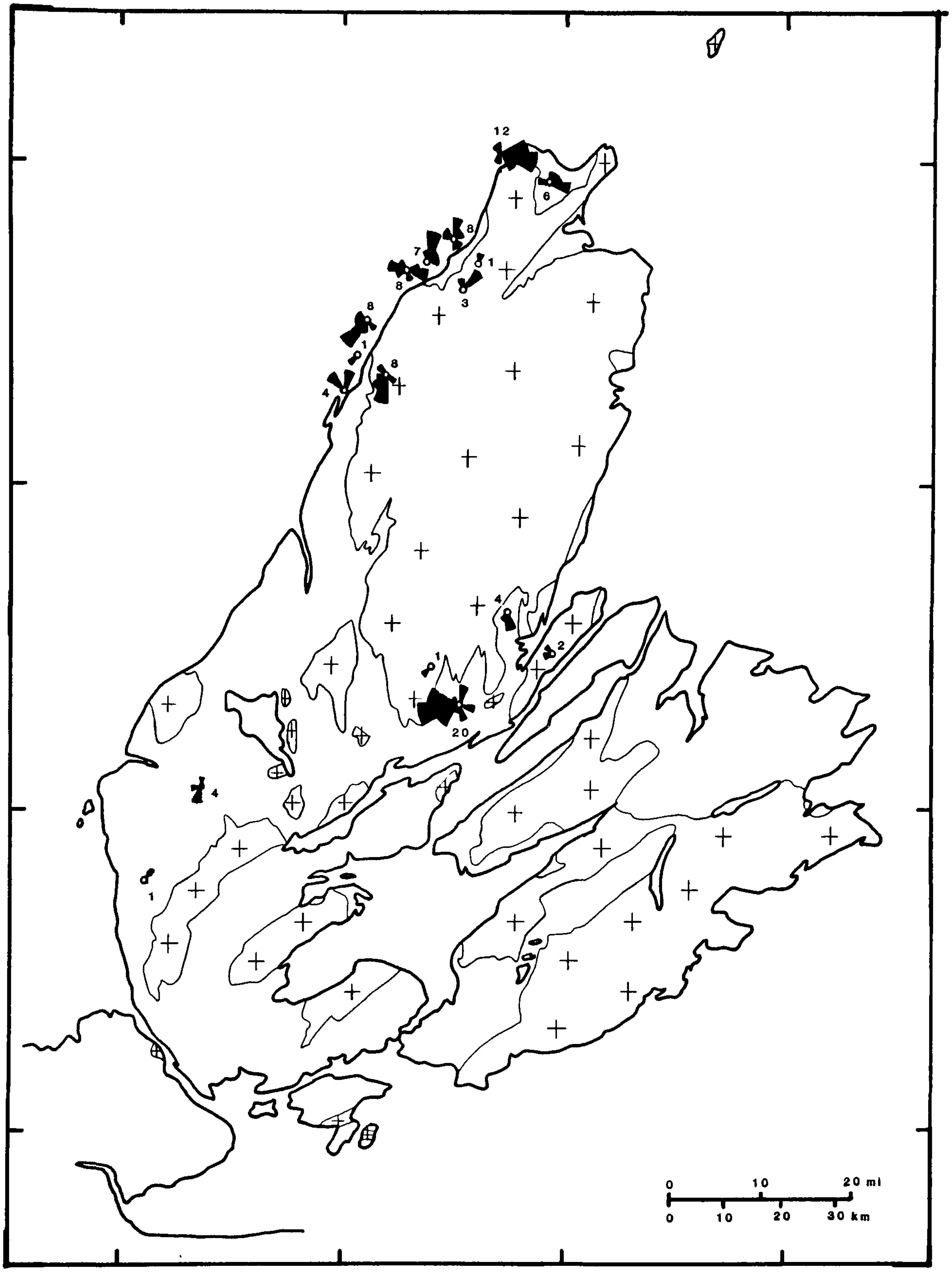

Figure 33. Paleocurrent map, C1 facies assemblage. Most measurements are from trough cross stratification.

Figure 34. Medium-scale trough cross stratification in reddish grey pebbly coarse sandstone to granulestone, C1 facies assemblage, Lowland Cove (section 70).

Figure 35. Scour pocket filled with poorly sorted conglomerate at bas of pebbly coarse sandstone, C1 facies assemblage, Lowland Cove (section 70).

Figure 36. Photomicrograph of hematite rinds surrounding grains of red pebbly coarse sandstone, C1 facies assemblage, Judique Intervale Brook (section 2). 10x magnification.

Figure 37. Green calcareous nodules in red siltstone between pebbly coarse sandstone beds, C1 facies assemblage, Corney Beach, (section 53).

Figure 38. Grey sandy limestone in red siltstone, C1 facies assemblage, North River (section 38).

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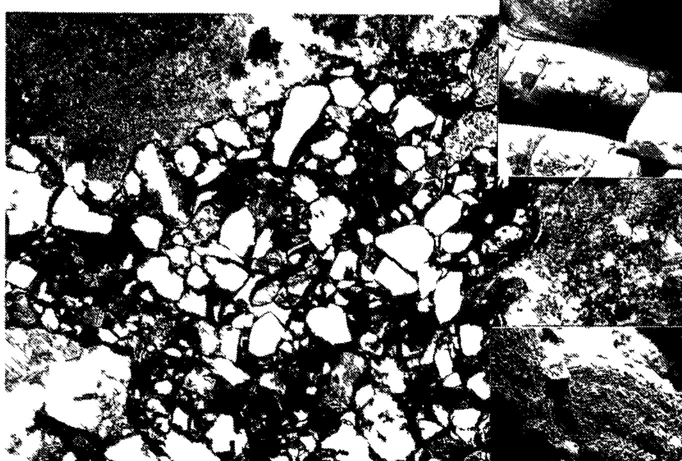

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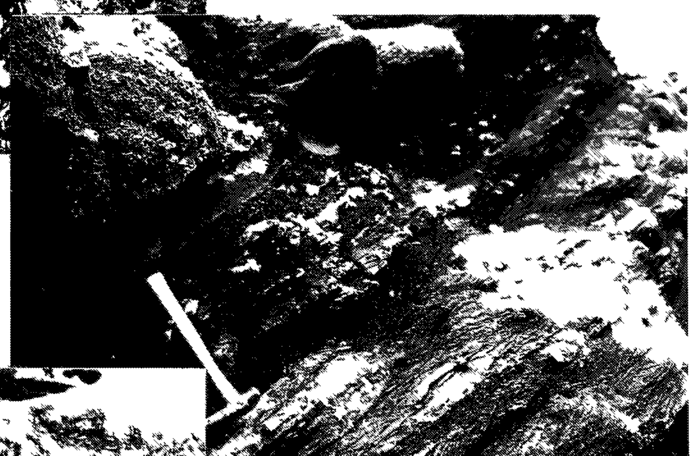

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thin streaks of granules or dispersed angular granules "floating" in a sandy siltstone matrix (eg: Argyle Brook). Thin green bands, calcareous nodules, vertical root casts and wood fragments are common (Fig. 37). On Baddeck River (Fig. 32) and North River several beds of grey sandy micritic and peloidal nodular limestone are present, with calcite-filled root casts (Fig. 38). Uncommon lined and calcite-filled (?) burrows are present in a few locations. In thin section these sediments are uniform and have angular to subangular equant quartz and rock fragments with hematite rims. They are classified as quartz arenite. Calcite cement occludes most primary porosity but up to 3% secondary porosity is present. C2 Brick red siltstn-fine ss This facies assemblage comprises up to 500 m in exposed thickness (Fig. 39), although there may be thrust repetitions and the true maximum may be 200-250 m. It was described at 17 outcrop and 2 drillhole sections with thick occurrences present only in the western Cape Breton area. A typical example from Judique Intervale Brook is presented in Figure 40. It is the most common facies assemblage of the upper Craignish in that area and occurs only as thin tongues in areas close to original basin margins. This assemblage forms part of the McLeod Member of Murray (1960) and Kelley (1967) and intertongues with facies assemblage C1. A total of 58 paleocurrent measurements were collected from 10 sections, mostly from trough cross stratification in thicker sandstones, and ripple cross lamination in thin sandstones (Fig. 41).

The predominant lithofacies is brick red to maroon to red-brown siltstone, sandy in part, in units up to 50 m thick (Fig. 40). It is typically massive and uniform, but may be thinly laminated or have an irregular rubbly (bioturbated?) texture. Abundant small green calcareous reduction spots and green or reddish silty limestone nodules in distinct horizontal beds are common (Fig. 42). On Judique Intervale Brook several 20 m thick units contain abundant evaporite crystal molds (probably gypsum rosettes, B. Rust, pers. comm.) small tracks and trails and desiccation mudcracks up to 1 cm deep (Figs. 40, 43, 44). In the Baddeck and Lowland Cove areas green calcareous streaks and nodular limestone beds with green calcareous root casts, encased in red siltstone are common (Fig. 45). On Baddeck River there is a 1 m thick greenish grey limestone with sharp base and top and an irregular fractured appearance. In thin section this bed is composed of rounded structureless peloids with external cortices set in a matrix of fine sparry calcite. There are also abundant elongate rhizoliths (?) with dark haloes and coarse sparry calcite fillings.

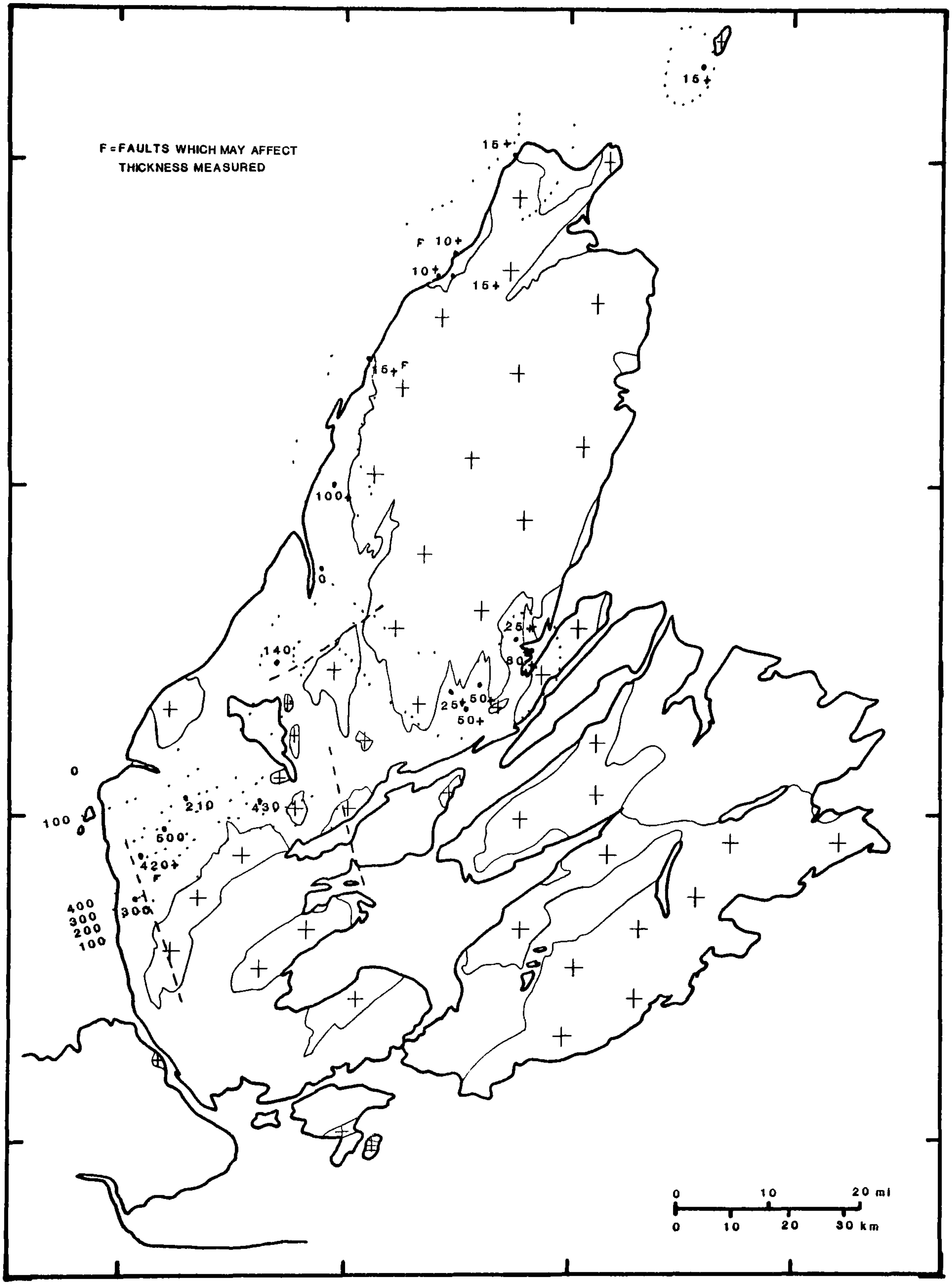

Figure 39. Isopach map, C2 facies assemblage. Thickness values in metres.

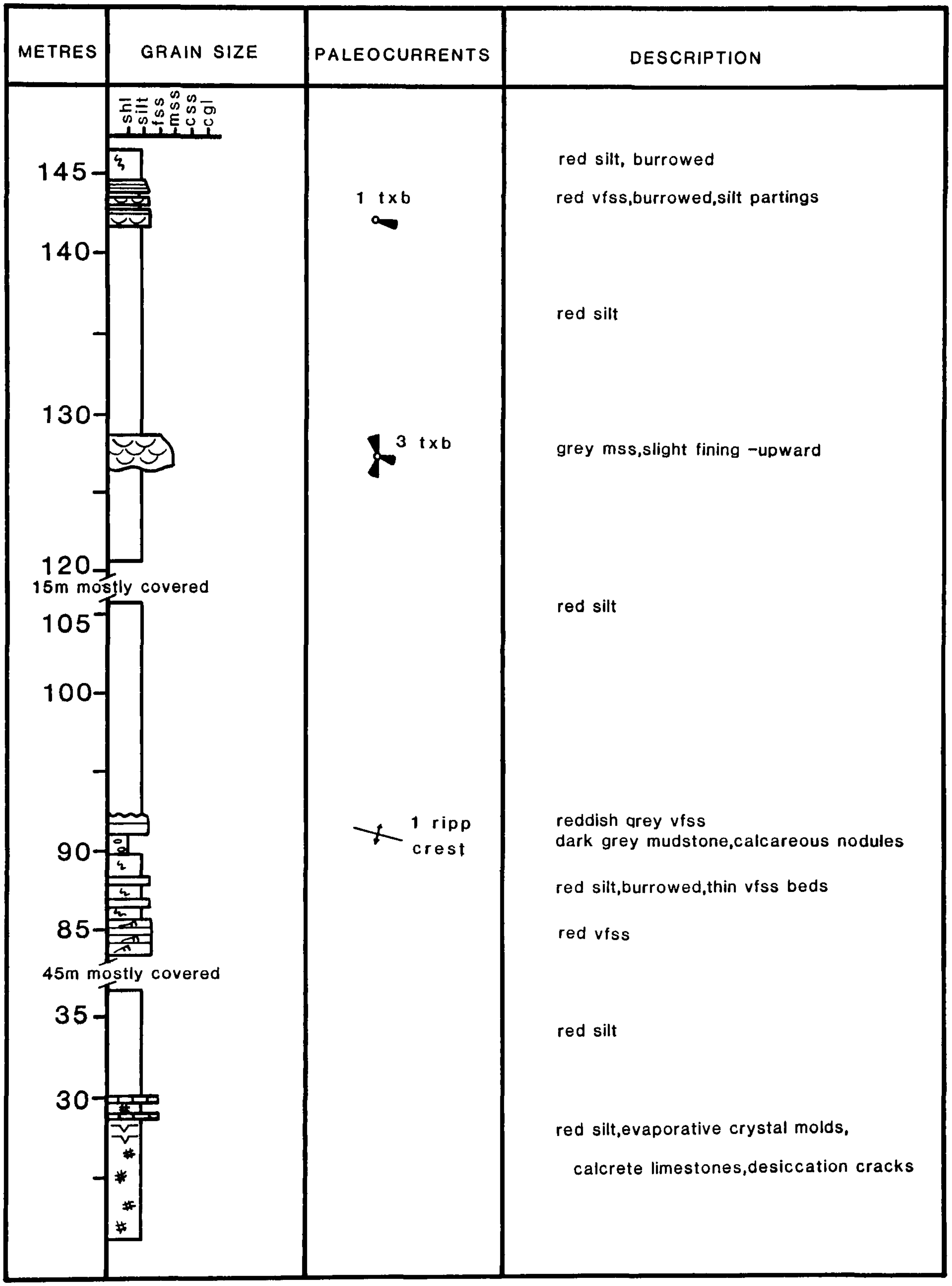

Figure 40. Type example, C2 facies assemblage, Judique Intervale Brook.

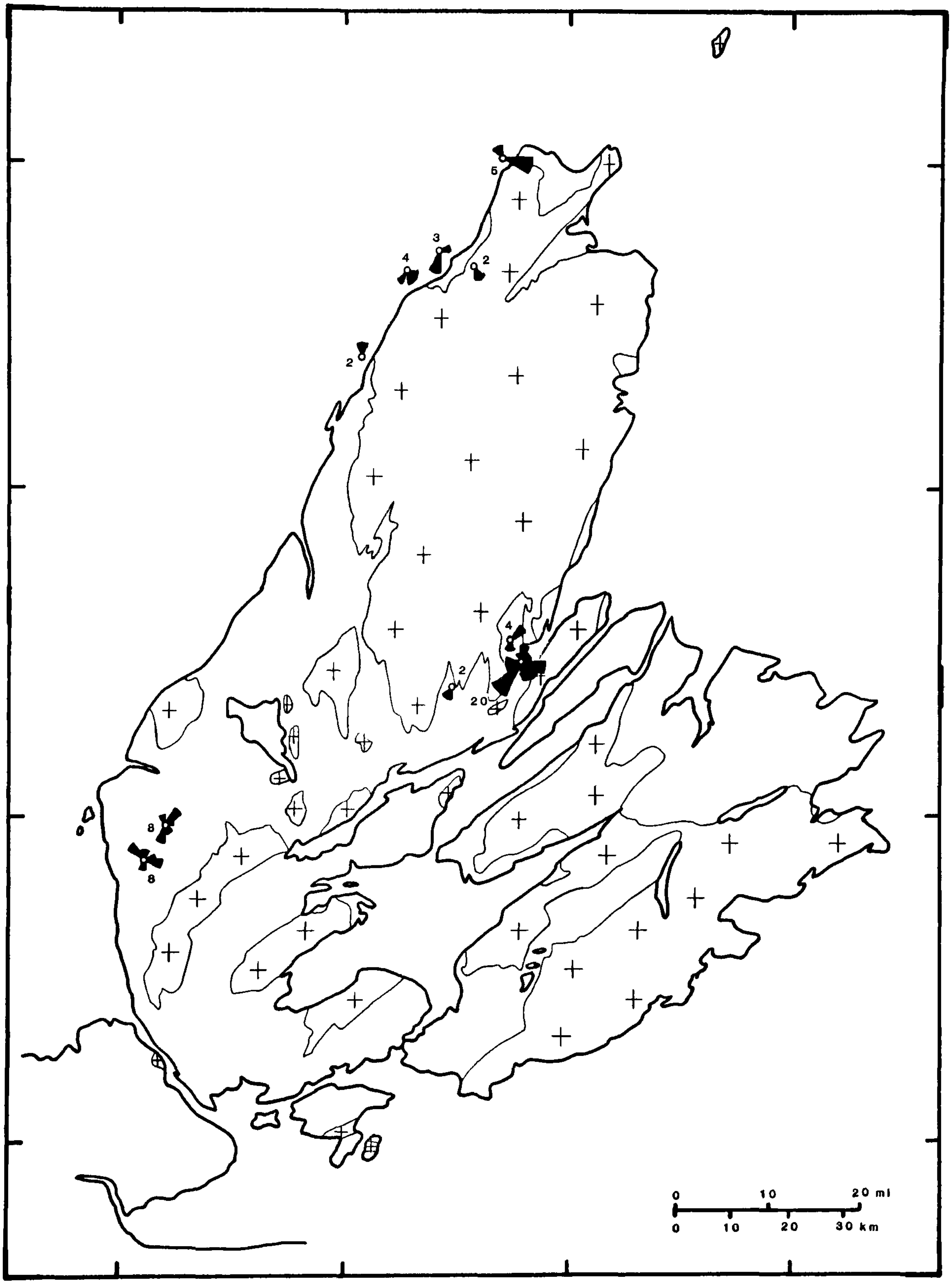

Figure 41. Paleocurrent map, C2 facies assemblage. Most measurements are from trough cross stratification and ripple cross lamination.

Figure 42.	Green reduction spots and roots in red siltstone, C2 facies assemblage, Munro Point (section 36).
Figure 43.	Evaporative crystal molds (probably gypsum rosettes) in red siltstone, C2 facies assemblage, Judique Intervale Brook (section 2).
Figure 44.	Desiccation cracks in red siltstone, C2 facies assemblage, Munro Point (section 36).
Figure 45.	Sub-vertical and horizontal green calcareous root casts in red siltstone, C2 facies assemblage, Lowland cove (section 70).
Figure 46.	Red very fine to fine sandstone beds encased in red siltstone, C2 facies assemblage, Munro Point (section 36). cliff is 7m high.

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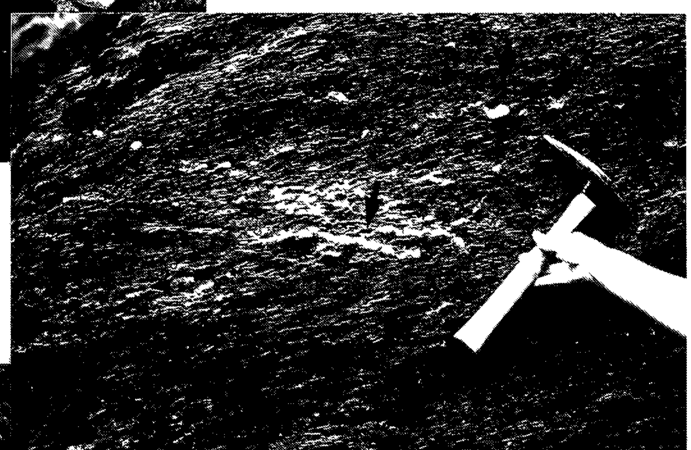

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Ubiquitous thin red or grey very fine- to fine-grained sandstone interbeds are dispersed throughout the facies assemblage (Figs. 40, 46). Most are calcareous and rippled and are less than 1 m thick, with sharp bases. In thin section they are moderately to well sorted and have angular to subangular quartz and feldspar framework grains with about 10% matrix. They are classified as feldsarenite to subfeldsarenite. There are a few possible burrows filled with calcite. Several beds 1-5 m thick, which fine upward from granulestone to fine sandstone, occur in bundles up to 20 m thick at several outcrops in western Cape Breton. These beds have sharp scoured bases, horizontal or trough cross stratification and are similar to sandstones of facies assemblage C1.

C3 Grey/green coarse ss-granulestone This facies assemblage is up to 1200 m thick (Fig. 47), although there may be thrust repetitions and the true maximum may be 800-1000 m, of which there is typically much less exposed at individual sections. It was described at 17 outcrops and 3 drillhole sections and a typical example is presented in Figure 48. It unconformably overlies basement (eg: Graham River) or conformably overlies the Fisset Brook Formation (eg: McFarlane Brook; Gallant River) and is the main facies assemblage for the lower and middle Craignish of western Cape Breton. In northern Cape Breton it occurs as relatively thin units at the top of the Craignish. On Graham River and McFarlanes Brook diabase sills up to 5 m thick intrude the lower part of this facies assemblage, causing recrystallization of the rocks. The C3 facies assemblage is approximately equivalent to the Graham River and Skye River Members of Murray (1960) and Kelley (1967). A total of 70 paleocurrent measurements were collected from 10 sections, mostly trough cross bedding (Fig. 49).

The predominant lithofacies is grey or greenish grey to brownish grey, well stratified, fine- to coarse-grained sandstone and granulestone, typically pebbly coarse sandstone. Beds are 0.5 to 5.0 m thick (typically 1-2 m) and fine upward from sharp flat or scoured bases to fairly sharp tops. The beds are uniform in aspect, may appear massive, horizontally stratified, or have medium-scale trough cross stratification (Fig. 50). Small subangular to subrounded pebbles occur as thin lags at scour bases, as discrete laminations or dispersed throughout the bed (Fig. 51). Each successive bed commonly has a different dominant grain size (Fig. 52). Beds may be stacked into packages 10 to 30 m thick with few intervening fine grained lenses (eg: Graham River). The sediments are micaceous, may be calcareous, and generally have fair to good sorting. One example of mudcracks was

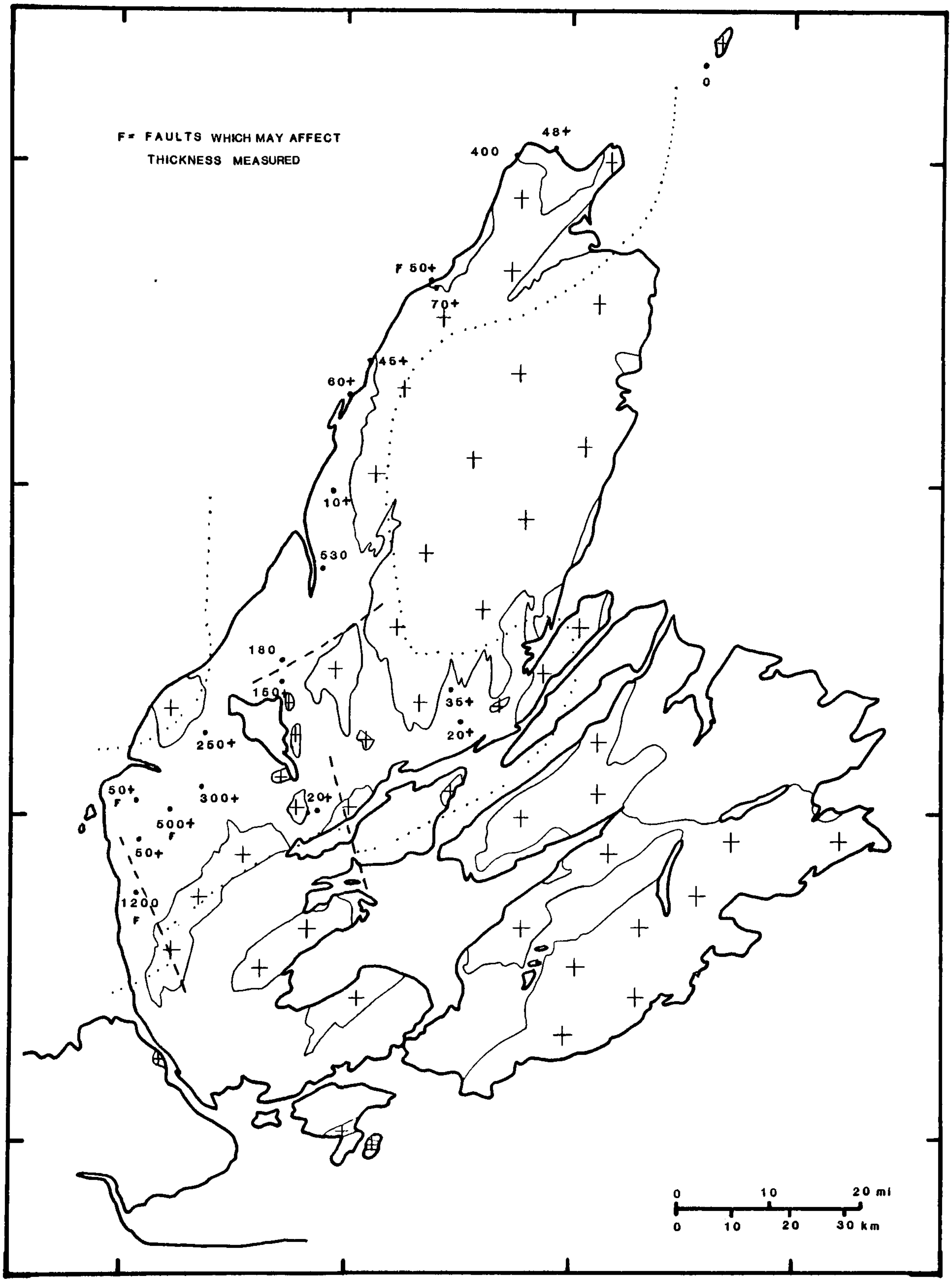

Figure 47. Isopach map, C3 facies assemblage. Thickness values in metres.

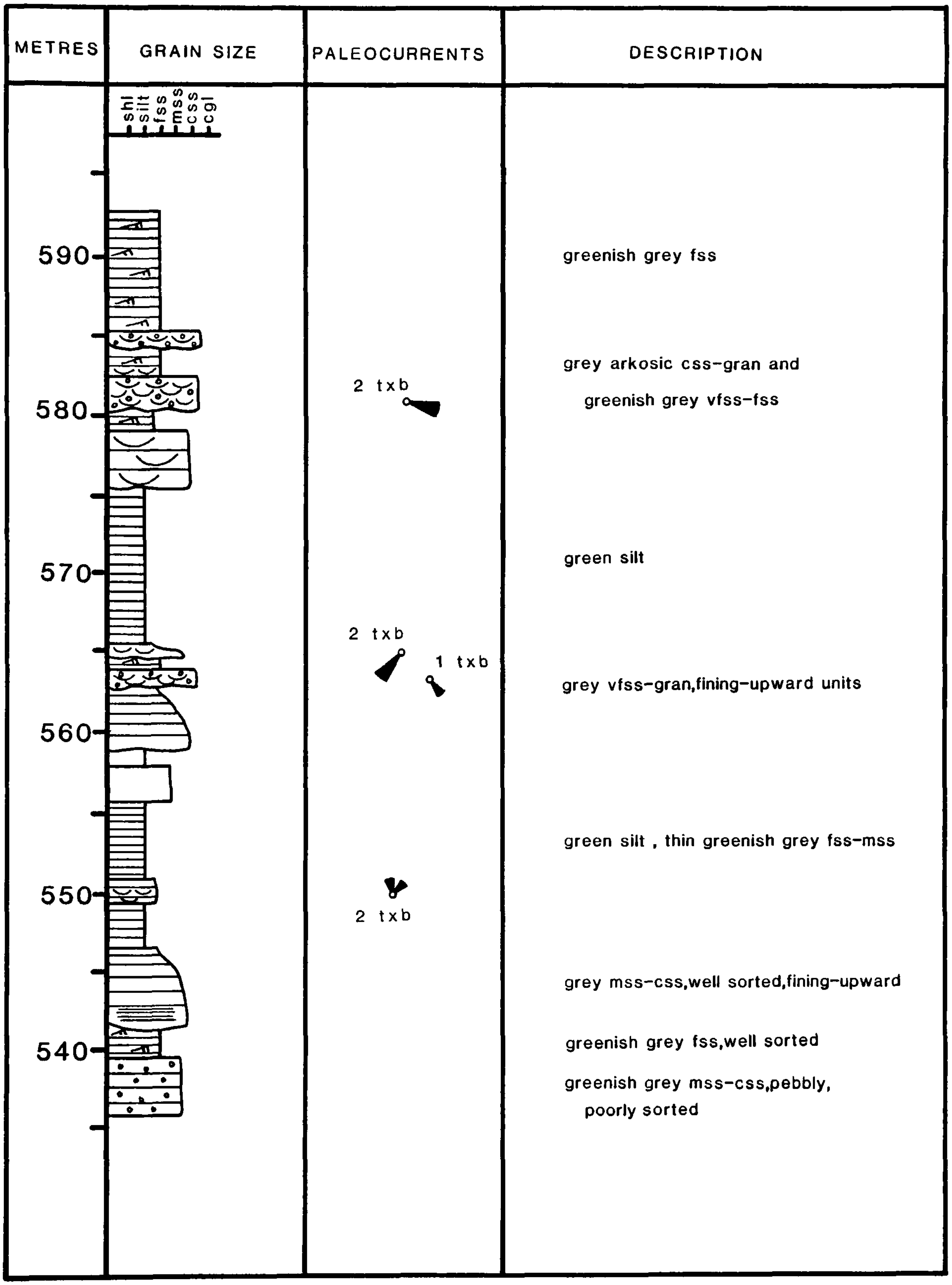

Figure 48. Type example, C3 facies assemblage, Graham River.

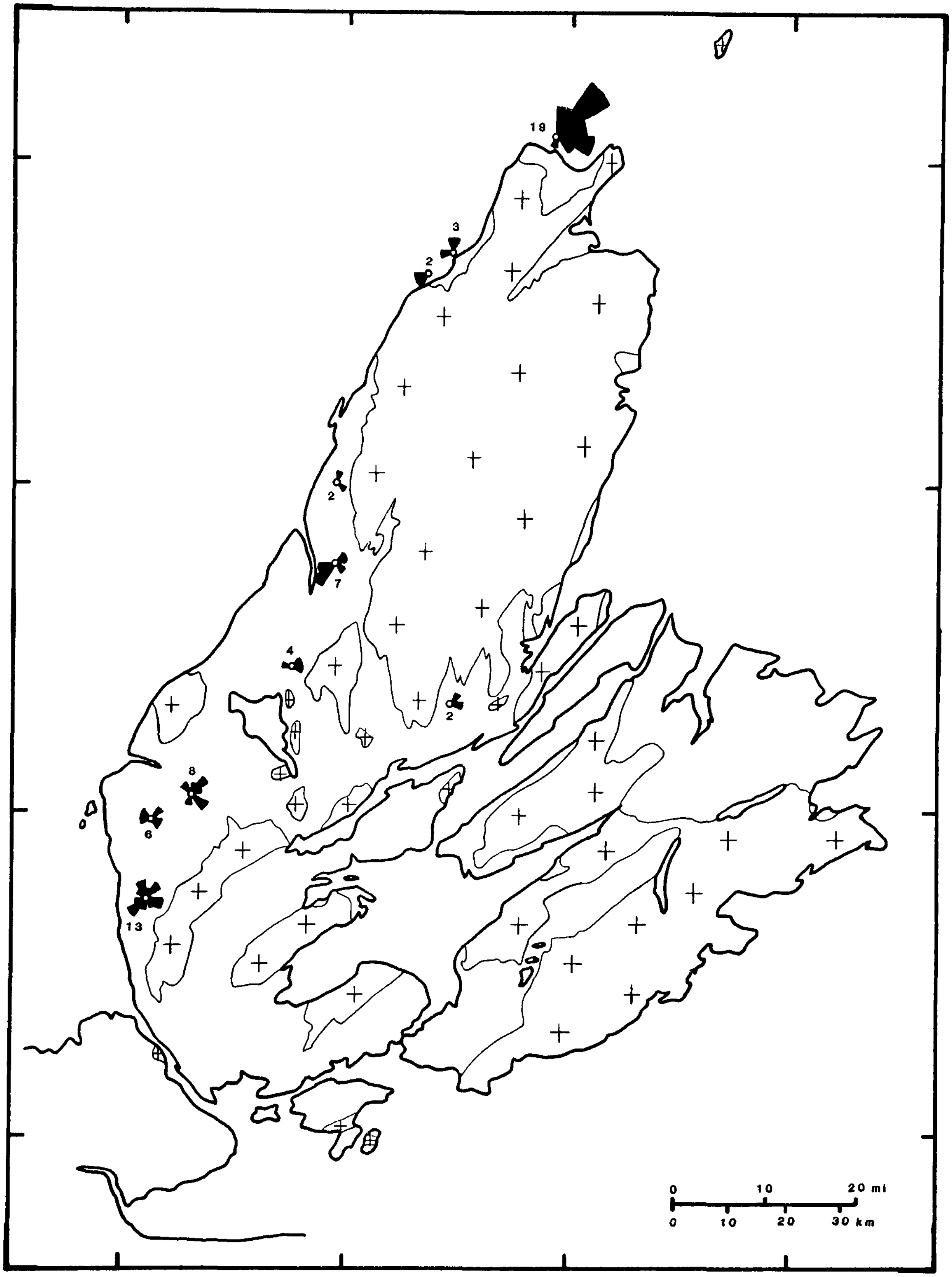

Figure 49. Paleocurrent map, C3 facies assemblage. Most measurements are from trough cross stratification.

Figure 50. Grey pebbly coarse sandstone with medium scale trough cross stratification, C3 facies assemblage, Gallant River (section 44).

Figure 51. Greenish grey pebbly coarse sandstone with medium scale trough cross stratification, C3 facies assemblage, Black Point (section 72).

Figure 52. Alternating beds of coarse sandstone and sharp-based sandy conglomerate, C3 facies assemblage, Black Point (section 72).

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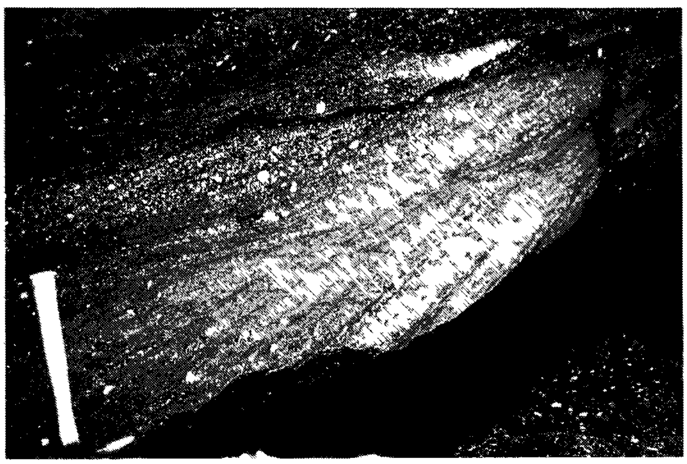

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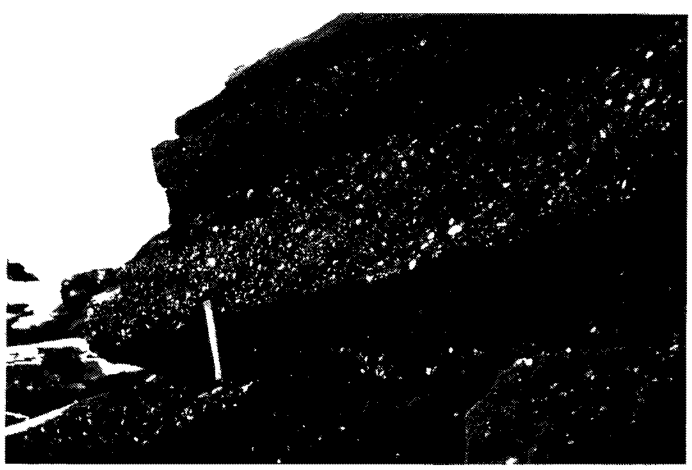

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observed at Black Point (Fig. 53). In thin section these rocks have angular to subangular equant framework grains of mono- and poly-crystalline quartz and rock fragments (granite, volcanics, quartzite, schist, marble) and are classified as quartz arenite to sublitharenite/subfeldsarenite. Extensive silica cement and sutured contacts are ubiquitous although calcareous cement is locally abundant. All primary porosity has been occluded and there is no secondary porosity.

Greenish grey fining-upward beds of sand-supported conglomerate to very pebbly coarse sandstone are locally abundant as isolated beds, or in 20-50 m coarsening-upward sequences of beds which become more massive upward. The matrix is argillaceous coarse sandstone with poor to fair sorting. Sedimentary structures include vague horizontal stratification and, in a few cases, trough cross stratification. This lithofacies is particularly common at Graham River, Gallant River, (Fig. 54) and St. Paul P-91.

Green siltstone to fine sandstone is commonly interbedded with the coarser lithofacies as thin beds or lenses less than 1 m thick. It rarely occurs as thick sequences up to 75 m thick in the middle of the Craignish at Graham River (Fig. 48), McFarlanes Brook, and Gallant River in western Cape Breton (referred to by Côté, 1959, as the "anomalous grey fine clastic unit"). The sediments are typically very uniform in aspect, well sorted, with thin horizontal lamination in siltstone and ripple cross lamination or some trough cross stratification in sandstones (Fig. 55). At Abbey Coast and Le Buttereau in northern Cape Breton units 5-20 m thick of dark grey to greenish grey siltstone with bioturbation and very fine to fine grained sandstone beds 2-3 m thick occur as tongues between units of facies assemblage C1. A single occurrence of sandy pebbly massive limestone is present on Graham River.

STRATHLORNE FORMATION

The Strathlorne Formation, up to 590 m thick (Fig. 56), appears to conformably overlie the Craignish Formation and is conformably overlain by the Ainslie Formation in an intertonguing relationship. The great thickness recorded at a few locations in western Cape Breton is apparently partly due to thrust repetition as described in Chapter 2 and a true maximum of about 350 m is more likely. For this project the Strathlorne Formation has been studied at 46 outcrops and 6 drillholes, although many expose only a portion of the entire unit. However, it is much better exposed than the Craignish Formation. It

Figure 53. Large desiccation crack molds on base of coarse sandstone, C3 facies assemblage, Black Point (section 72).

Figure 54. Grey boulder conglomerate with coarse sand to granule matrix, C3 facies assemblage, Conglomerate Creek (section 64).

Figure 55. Greenish grey very fine to fine sandstone in fining-upward units, with horizontal lamination and grey siltstone partings, C3 facies assemblage, Black Point (section 72).

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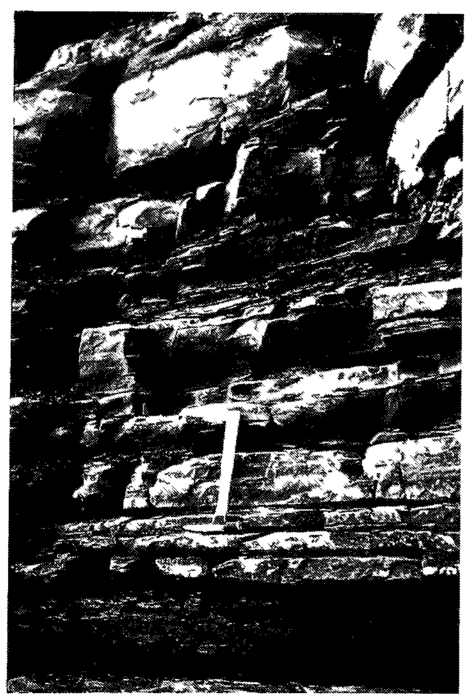

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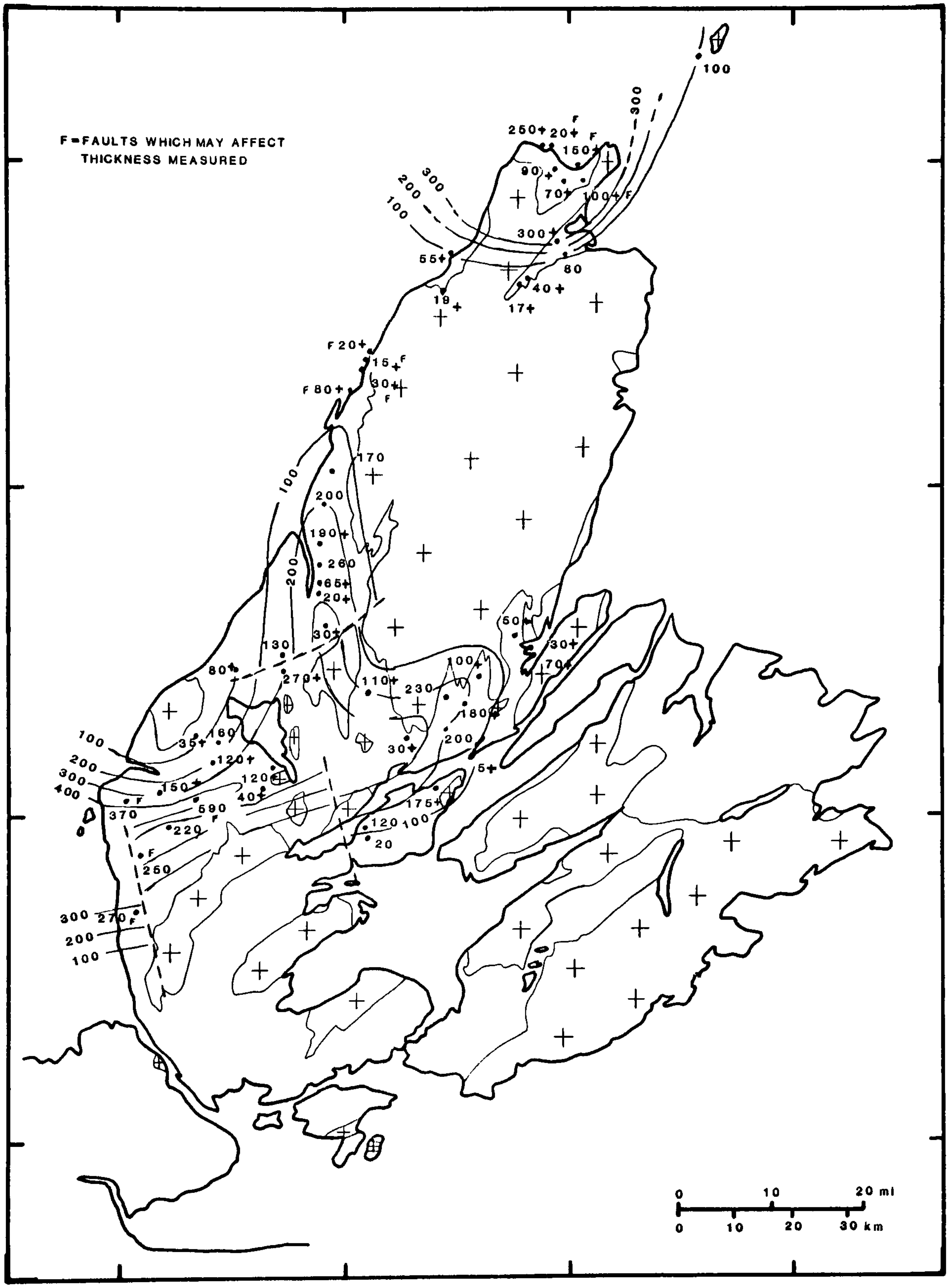

Figure 56. Isopach map, Strathlorne Formation. Thickness values in metres.

generally consists of grey mudstone with interbedded grey fine sandstone and limestone beds, in marked contrast to the redder and coarser formations above and below. It generally thickens toward the centres of depositional sub-basins. Thick, well exposed sections of the Strathlorne are common in the western Cape Breton area but uncommon in the northern area. There is a general tendency for more abundant sandstone in the lower and middle Strathlorne part of western Cape Breton and all of the exposed (southeastern) part of the northern Cape Breton area. The Strathlorne Formation is poorly preserved over much of the northern area, either through erosion or faulting. It can be divided into 4 facies assemblages (Table 4).

S1 Dark grey mudstone This facies assemblage has a cumulative thickness of up to 500 m (Fig. 57) although this is apparently due to thrust repetition and probably 300 m is the true maximum, with 150 to 250 m being a more representative thickness. It was described at 45 outcrop and 6 drillhole sections. It is well displayed in the Mabou, Baddeck and Margaree areas of western Cape Breton, and the Aspy River and Meat Cove areas of northern Cape Breton. Figure 58 presents a typical example from North Branch Baddeck River. It is the dominant facies assemblage of the Strathlorne, except in original marginal positions in the sub-basins. It overlies the upper Craignish Formation either sharply as in the Baddeck area or with a gradational transition as in the Mabou area. The S1 facies assemblage typically comprises the bulk of the middle and upper Strathlorne, and commonly occurs in units 10 to 100 m thick. A total of 185 paleocurrent measurements were collected from 22 sections, mostly ripple cross lamination, symmetrical ripple crests, some tool marks and current lineation (Fig. 59).

The predominant lithofacies is grey to dark grey siltstone, claystone and very fine sandstone or limestone thinly interbedded on a scale of 1 to 20 cm and commonly arranged in thickening- and coarsening-upward sequences 2 to 45 m thick (generally 5-20 m) (Figs. 58, 60, 61, 62). Although siltstone is the typical lithology of the bulk of the sediments, they commonly range from claystone at the base of a sequence to fine sandstone at the top. Sandstone:siltstone ratios range from 1:10 at the base to 1:1 at the tops of these sequences. These coarsening-upward sequences tend to be thinner (2-20 cm, av 6 m thick), finer-grained, more numerous and more obvious in the Baddeck area of western Cape Breton, and the Meat Cove area of northern Cape Breton (ie. toward sub-basin axes). In this setting, some stacked sequences are arranged into discrete bundles 20-50 m thick (eg.

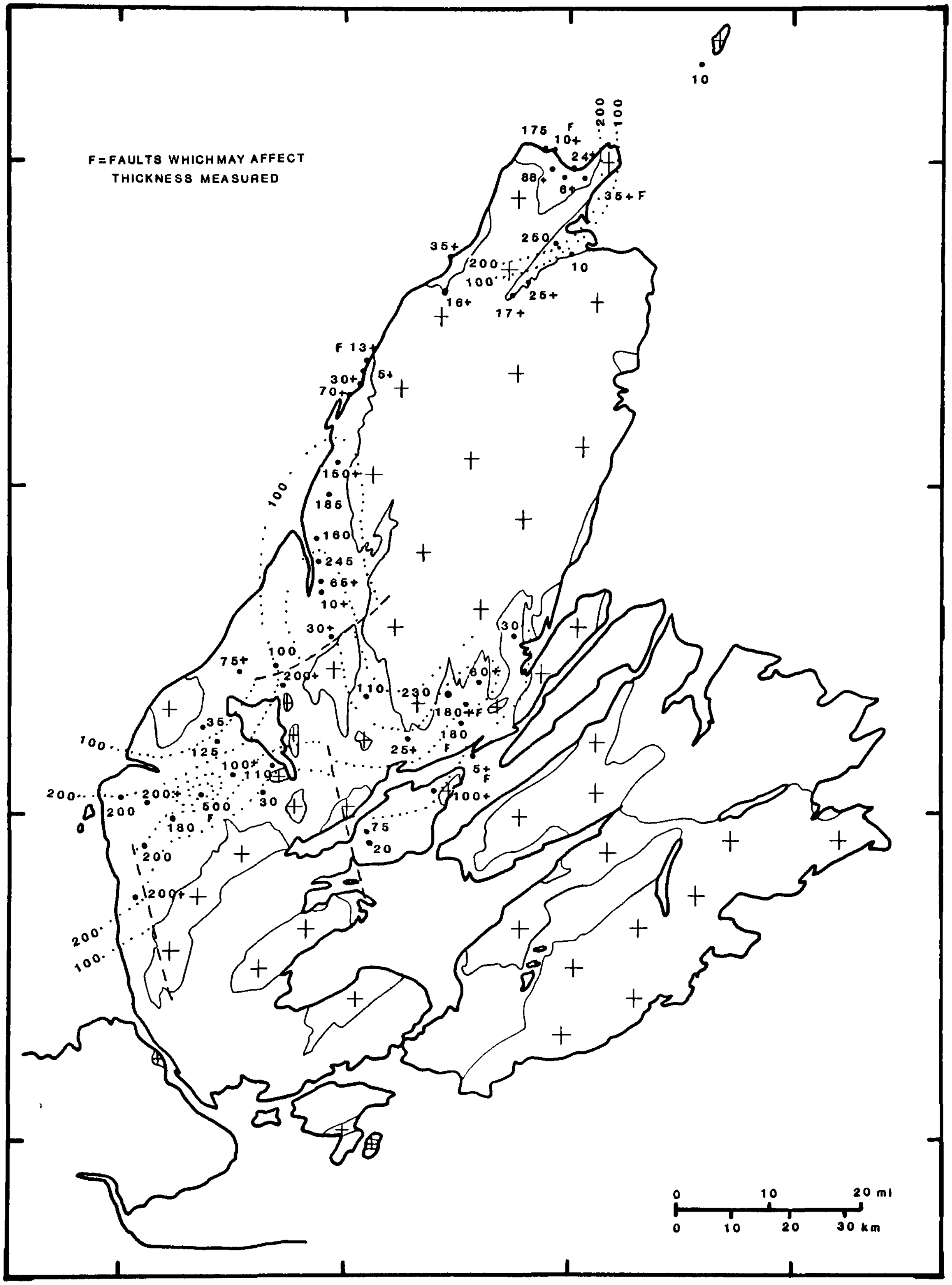

Figure 57. Isopach map, S1 facies assemblage. Thickness values in metres.

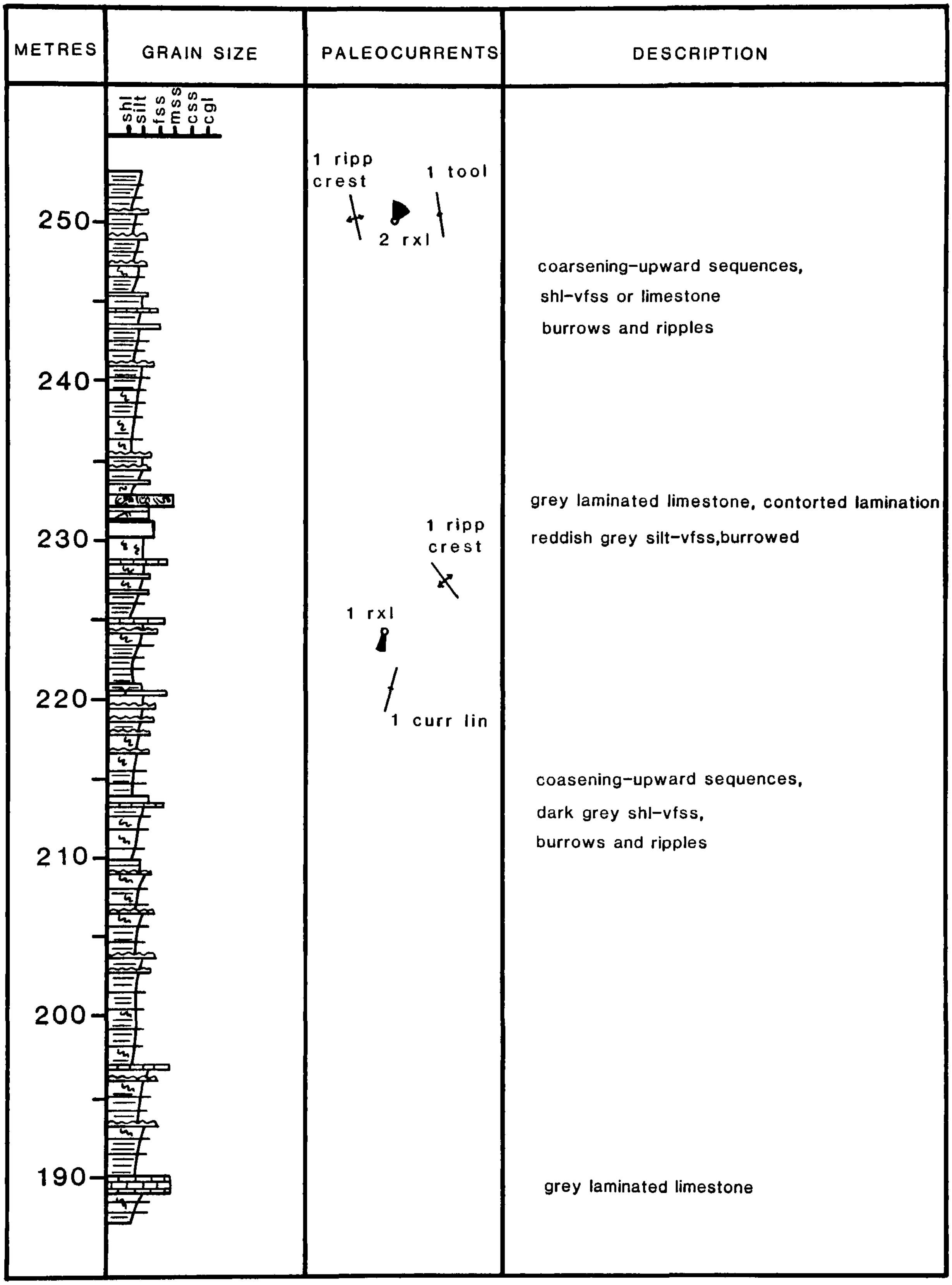

Figure 58. Type example, S1 facies assemblage, North Branch Baddeck River.

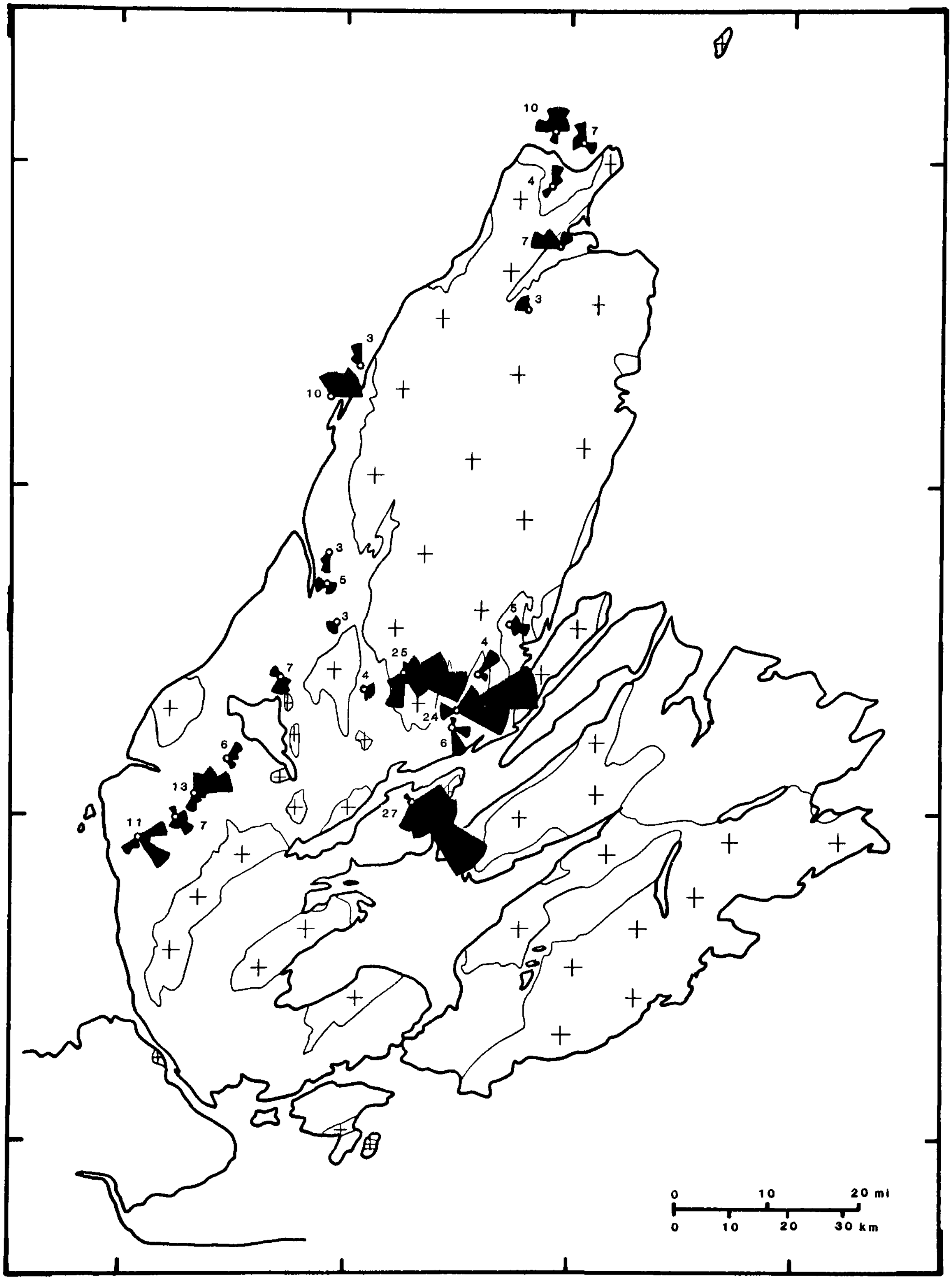

Figure 59. Paleocurrent map, S1 facies assemblage. Most measurements are from ripple cross lamination and tool marks.

Figure 60. Dark grey siltstone with thin sandy siltstone beds, S1 facies assemblage, Graham River (section 1). Scale is 15 cm.

Figure 61. Several coarsening-upward sequences 2m thick, of dark grey silty mudstone passing upward into calcareous very fine sandstone, North Branch Baddeck River (section 33). Backpack for scale.

Figure 62. Five stacked thin coarsening-upward sequences, S1 facies assemblage, Peters Brook (section 31). Backpack for scale.

Figure 63. Trace fossils (including Lockeia) on bases of very fine sandstone beds, S1 facies assemblage, McFarlanes Brook (section 24). Scale in cm.


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North Branch Baddeck River, Baddeck River; Fig. 58). Conversely, they are thicker (5-45 m, av 18 m), slightly coarser-grained, fewer in number and less obvious in the Mabou area of western Cape Breton and the Aspy area of northern Cape Breton (ie. toward sub-basin margins).

At the base of each sequence the claystone tends to be dark grey to black, uniform, organic-rich (it may have a petroliferous smell) and thinly laminated. In some cases (eg: North Branch Baddeck River) the lamination consists of alternating light and dark even laminae 1-5 mm thick, reminiscent of varves. Thin horizons of calcareous nodules up to 15 cm in diameter, rare thin limestone beds, and fish fragments are common (eg: North Branch Baddeck River, Cap Rouge, North Aspy area). Upward, grey to dark grey siltstone and sandy siltstone become dominant, with a concurrent increase of horizontal burrowing and thin calcareous very fine to fine sandstone interbeds. These interbeds range from 1-50 cm (generally 10-20 cm) with sharp bases and sharp to gradational tops. Grain size is typically uniform but a few examples show grading. The bases commonly have abundant casts of small horizontal trace fossils, small scale load structures, and tool marks such as scratches and prods (Figs. 63, 64, 65). Internal stratification comprises ripple cross lamination (Fig. 66), horizontal lamination and some contorted lamination. At the tops of beds both asymmetrical and symmetrical (usually current ripples with a secondary symmetrical shape) ripple shapes are common. These are typically straight-crested, but some bifurcate, and have somewhat variable orientation from bed to bed (Fig. 67). One bed at McRae Brook has (?)synaeresis cracks on the top (Fig. 68).

Sandstones at the tops of coarsening-upward sequences commonly display interference ripples (Fig. 69), planed-off symmetrical ripples (Fig. 70), desiccation cracks (Fig. 71) and, at Baddeck River and Salmon River Coast, raindrop impressions (Fig. 72). In thin section these sandstones have well sorted, equant, subangular framework grains of quartz and minor rock fragments (classified as quartz arenite), set in a pervasive poikilotopic sparry calcite cement which may comprise 20-40 % of the rock. The texture is very uniform, although laminae are marked by slightly greater argillaceous content. All original porosity has been occluded by calcite cement but up to 2% secondary porosity is present.

The coarsening-upward sequences may be capped by thin sandstone as above, thick sandstone of the S2 facies assemblage, or, as in the Baddeck area, thin limestone beds (Fig. 58). Many of the latter are actually very calcareous siltstones (ie. about 40-50% calcareous

Figure 64. Load structures on base of very fine sandstone bed, S1 facies assemblage, Meat Cove (section 71).

Figure 65. Casts of current crescents around mud chips on base of very fine sandstone bed, S1 facies assemblage, Murdoch Pauls Brook (section 27). Scale is 15 cm.

Figure 66. Ripple cross lamination in thin, sharp-based, very fine sandstone underlain and overlain by dark grey siltstone, S1 facies assemblage, Schoolhouse Brook (section 16).

Figure 67. Straight-crested symmetrical ripples with bifurcation at top of thin fine sandstone bed, S1 facies assemblage, southeast Mabou River (section 13). Scale is 15 cm.

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Figure 68. Synaeresis cracks filled with siltstone on top of rippled very fine sandstone bed, S1 facies assemblage, MacRae Brook (section 28).

Figure 69. Interference ripples on top of very fine sandstone bed, S1 facies assemblage, Meat cove (section 71).

Figure 70. Planed-off symmetrical ripples with straight crests at top of thin fine sandstone bed, S1 facies assemblage, Meat cove (section 71).

Figure 71. Large desiccation cracks filled with siltstone at top of grey very fine sandstone, S1 facies assemblage, Meat Cove Roadcut (section 71).

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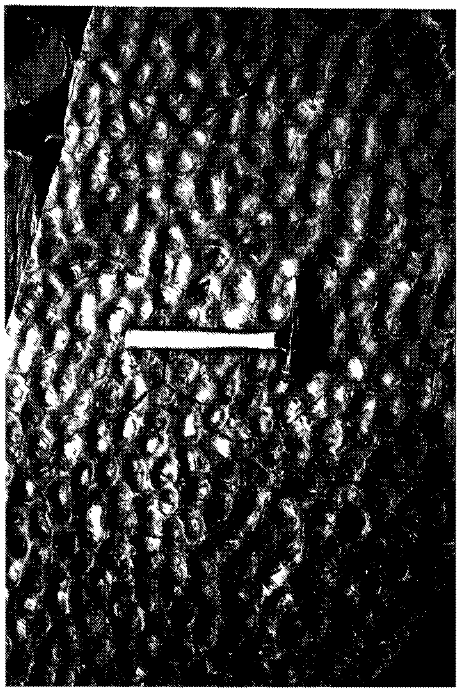

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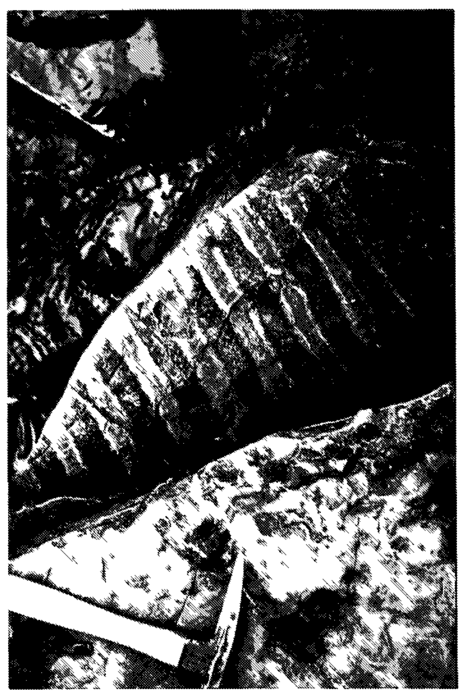

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Figure 72. Raindrop impressions on top of very fine sandstone at top of coarsening-upward sequence, S1 facies assemblage, Baddeck River (section 32). Scale is 15 cm.

Figure 73. Contorted lamination of alternating micrite and argillite within a grey limestone bed at top of coarsening-upward sequence, S1 facies assemblage, North Branch Baddeck River (section 33).

Figure 74. Stromatolitic limestone, S1 facies assemblage, North Branch Baddeck River (section 33). Scale is 15 cm.

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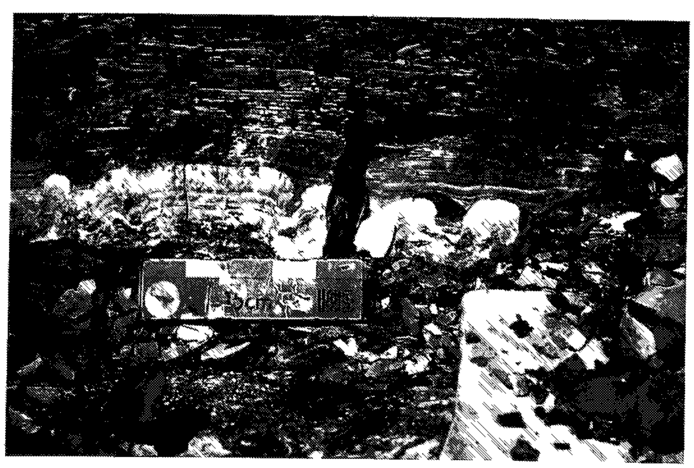

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cement) with symmetrical ripples and mudcracks. True limestones occur in at least 2 forms. Many limestones have alternating light-coloured micritic and dark-coloured argillaceous laminae, commonly with contorted lamination in the middle and a petroliferous smell (Fig. 73). In thin section, micritic laminae have abundant peloids (many partly-dissolved), spar-filled horizontal burrows, small arcuate (ostracod?) shell fragments, and possible mudcracks filled with silt grains and peloids. Argillaceous laminae have ovoid peloids, silt grains, sparry cement, iron oxide staining, and rare shell fragments. A second type of limestone, common at the tops of coarsening-upward sequences in the Baddeck area, has cyano-bacterial stromatolites and a petroliferous smell (Fig. 74). These are thin (up to 20 cm), and have convex-up laminae depicting many small high-amplitude structures at the base which become more broad and flat toward the top. In thin section the laminae are quite irregular, have birdseye structure, bitumen streaks and some incorporate silt grains. Similar stromatolitic limestones occur at the base of the Strathlorne, immediately overlying S3 sediments which overlie basement at Sugar Brook and Sutherlands Brook in northern Cape Breton. A third type of limestone was not observed directly in this facies assemblage in the study area but likely occurs. Broken fragments of oölitic limestone and individual abraded oöids occur in thin section in several Strathlorne lithofacies and at least one oölitic limestone is known from a similar facies in the Horton Bluff outcrop in mainland Nova Scotia (T. Martel, pers. comm.).

S2 Grey/green very fine- fine ss This facies assemblage has a cumulative thickness up to 150 m (Fig. 75), although this may in part be due to thrust repetition, and probably 100 m is a more likely maximum at any section. Twenty to fifty metres is a more typical cumulative thickness. It was described at 44 outcrops and 5 drillhole sections. It is well developed in the Mabou and Margaree areas of western Cape Breton and the Aspy and Meat Cove areas of northern Cape Breton. Figure 76 presents a typical example from Meat Cove. Where common, S2 is the prevalent facies assemblage of the lower and middle Strathlorne, but also occurs in the upper Strathlorne on opposite sides of the depositional sub-basins. Facies assemblage S2 typically occurs in units 2-50 m thick at the tops of S1 coarsening-upward sequences, or as a transitional facies assemblage between the Strathlorne and underlying or overlying formations. A total of 181 paleocurrent measurements were collected from 25 sections, mostly from trough cross stratification, ripple cross lamination, and current lineations (Fig. 77).

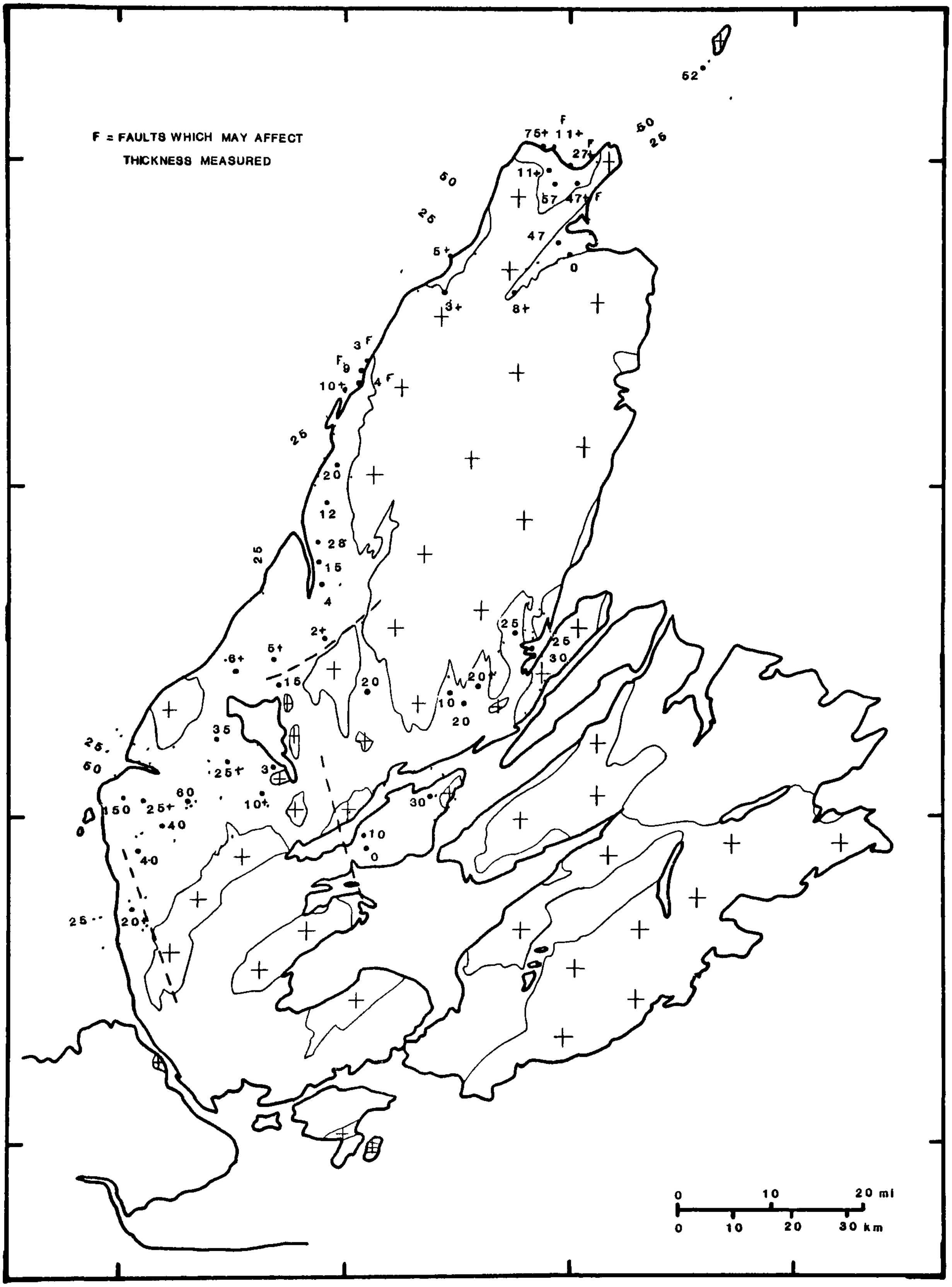

Figure 75. Isopach map, S2 facies assemblage. Thickness values in metres.

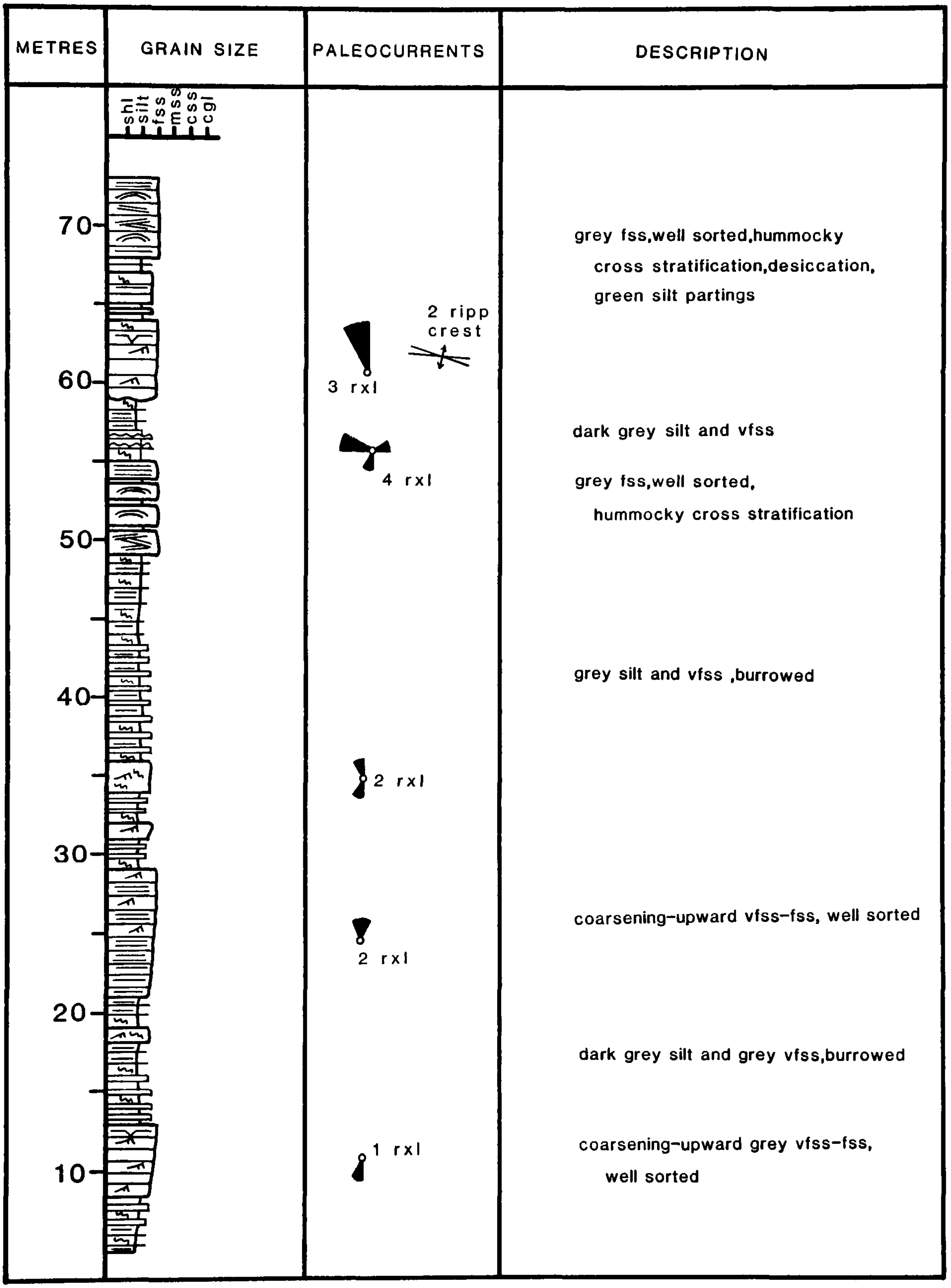

Figure 76. Type example, S2 facies assemblage, Meat Cove Roadcut.

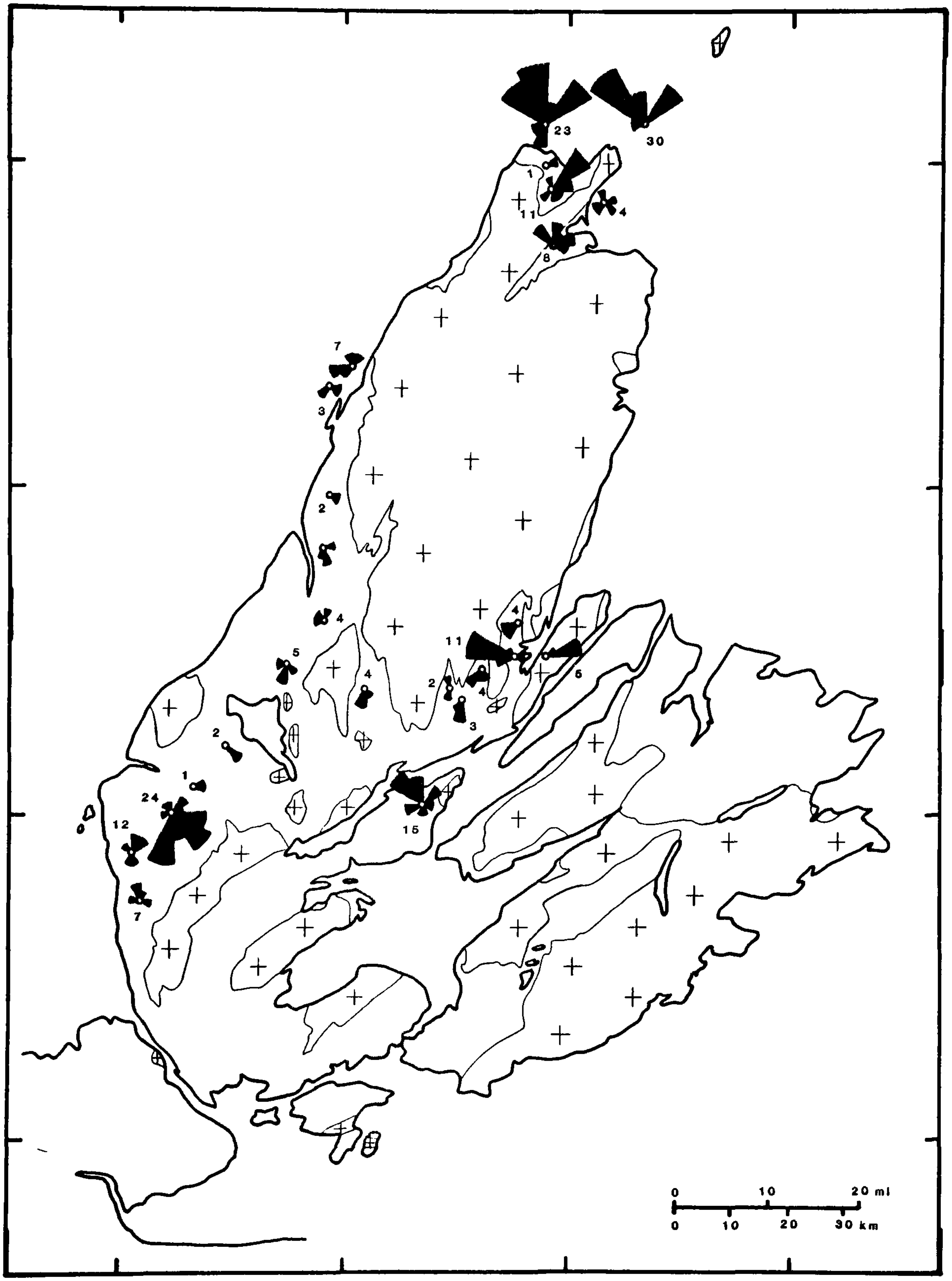

Figure 77. Paleocurrent map, S2 facies assemblage. Most measurements are from trough cross stratification, ripple cross lamination and current lineations.

The predominant lithofacies is grey, greenish grey or buff-coloured, well stratified, very fine- to fine-grained sandstone. Units are 1-10 m thick (typically 2-5 m) with sharp flat bases and bedding 10 to 50 cm thick (Fig. 78). A slight coarsening-upward trend is present in a few examples (eg: North River, Meat Cove, Fig. 76). The sandstone is moderately to well sorted and calcareous. Internal sedimentary structures include abundant horizontal lamination with current lineation, low angle lamination and medium-scale trough cross stratification, with ripple cross lamination and symmetrical ripple forms near the tops. Climbing ripples are present at several outcrops (eg: Salmon River Coast) (Fig. 79). Contorted lamination is common, typically as upright to slightly overturned sharp anticlinal and broader synclinal forms in the middle of a bed with undeformed lamination above and below (Fig. 80). Hummocky cross stratification (lambda = 50-100 cm, a= 20-40 cm) is present in a number of well-exposed sandstone bodies (eg: Baddeck River, North Branch Baddeck River, Peters Brook, Murdock Paul's Brook, Old Bridge Brook, Meat Cove) (Figs. 76, 81, 82), and may be even more common, because it is difficult to recognize in small, poorly exposed outcrops. Abundant desiccation cracks and roots were observed in a few instances (eg: Meat Cove, Salmon River Coast, Southwest Mabou River). Several sandstone bodies have sharp scoured bases with lags of wood fragments and fining-upward trends (eg: Le Buttereau, Angus Lake Brook, North Branch Baddeck River). In thin section, sandstones of S2 facies assemblages have well sorted, equant, angular to subangular, quartz framework grains with minor feldspar and rock fragments. They are classified as quartz arenite, although in northern Cape Breton plagioclase is more abundant and some samples are subfeldsarenites. The texture is very uniform with little argillaceous matrix (Fig. 83) except in possible burrow fills. Primary porosity is occluded by abundant calcite cement but 5-10% secondary porosity is present.

Greenish siltstone to sandy siltstone interbeds up to 20 cm thick are a ubiquitous accessory lithofacies (Fig. 76). They typically have horizontal burrows and may have symmetrical ripples. In a few sections (eg: Murdock Paul's Brook) rare rippled oölitic sandy limestone occurs intimately interbedded with the thicker fine sandstone. These limestones are buff-coloured, dolomitic, up to 1 m thick, laminated, and have rip-up clasts of siliceous mudstone. In thin section (Fig. 84) the limestone contains about 20% angular quartz grains of silt to fine sand size and a few fossil fragments. Quartz-rich/öoid-poor laminae alternate with quartz-poor/öoid- and peloid-rich laminae. Peloids are large (0.3-

Figure 78. Grey, well sorted, micaceous fine sandstone with low angle lamination sharply overlying dark grey mudstone (of S1 facies assemblage), S2 facies assemblage, Meat cove (section 71). Note hammer for scale.

Figure 79. Climbing ripples in greenish grey very fine sandstone with slight fining-upward trend, S2 facies assemblage, Salmon River coast (section 69).

Figure 80. Contorted bedding in sharp-based, very fine sandstone, S2 facies assemblage, Le Buttereau (section 49).

Figure 81. Hummocky cross stratification in fine sandstone at top of 7m coarsening-upward sequence overlying Craignish Formation, S2 facies assemblage, Baddeck River (section 32).

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Figure 82. Hummocky cross stratification in grey, micaceous fine sandstone, S2 facies assemblage, La Bloc (section 51).

Figure 83. Photomicrograph of well sorted quartz arenite with poikilotopic calcite cement, S2 facies assemblage, Baddeck River (section 32). 10x magnification, crossed polarizer.

Figure 84. Photomicrograph of alternating quartz-rich vs. oöid-rich lamination in rippled limestone, S2 facies assemblage, Murdoch Paul's Brook (section 27). 5x magnification.

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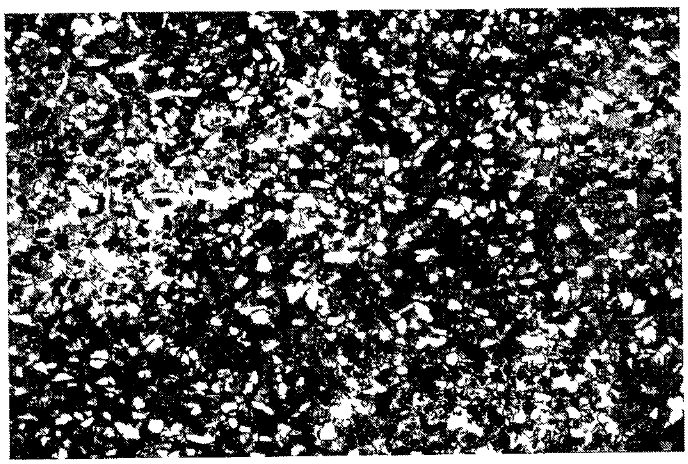

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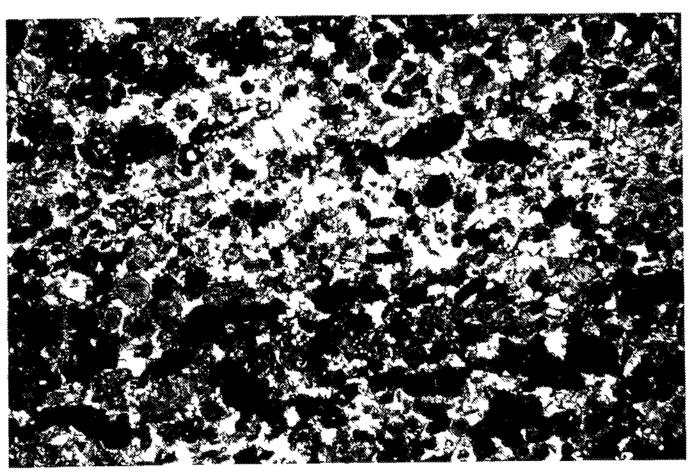

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1.0 mm), micritic and have irregular shapes, while öoids have peloidal cores, radial crystal form and distinct cortices with sutured contacts. All framework grains are set in a sparry calcite cement.

S3 Grey/green medium ss-boulder cgl This facies assemblage has a cumulative thickness up to 100 m (Fig. 85), but typically occurs as an isolated bundle of beds 2-30 m thick in sections close to depositional sub-basin margins. The bundles occur in the middle part of the Strathlorne in the Margaree area of western Cape Breton and the Aspy/Salmon River areas of northern Cape Breton. They occur in the upper Strathlorne only in the Baddeck area of western Cape Breton. This facies assemblage is best exposed in northern Cape Breton, is only abundant on Middle Aspy River, and is represented by a typical example from Conglomerate Creek (Fig. 86) where S3 is interbedded with S1. It was described from 15 outcrop and 2 drillhole sections. A total of 44 paleocurrent measurements were collected from 8 sections, mostly trough cross stratification, ripples and pebble imbrication (Fig. 87).

The predominant lithofacies is grey or greenish grey, micaceous, pebbly, medium- to coarse-grained sandstone. Beds are 2-5 m thick, have sharp scoured bases with pebble or siltstone rip-up lags, and fine-upward (Figs. 86, 88, 89). Units of this facies assemblage may scour down into S1 sediments (Gallant River, Murdock Paul's Brook, Black Cliffs) or S2 sediments (McLeod Point, Salmon River Coast) (Figs. 90, 91, 92). In thin section these sandstones have angular quartz, feldspar (plagioclase and microcline) and rock fragments, with a pseudomatrix of sericite (after feldspar) and minor calcite cement. They can be classified as quartz arenite to subfeldsarenite, and lack preserved porosity.

Greenish grey or grey, poorly sorted, matrix-supported pebble to boulder conglomerate commonly occurs with the previous lithology in massive beds 1-3 m thick (Red Point and McLeod Point in western Cape Breton, Middle Aspy River and Conglomerate Creek in northern Cape Breton, Fig. 86). The conglomerate has round to subround clasts up to 150 cm in diameter (most are 5-30 cm) set in a matrix of poorly sorted, angular, micaceous, coarse sandstone to granulestone. In some beds the large clasts project above the top of the bed and large boulders commonly have fractures filled with the surrounding matrix sandstone (eg. Middle Aspy River, Conglomerate Creek) (Fig. 93). At Conglomerate Creek, Little Narrows #1 and Little Narrows #2, units of facies assemblage S1 up to 15 m thick intertongue with these coarser sediments (Fig. 86).

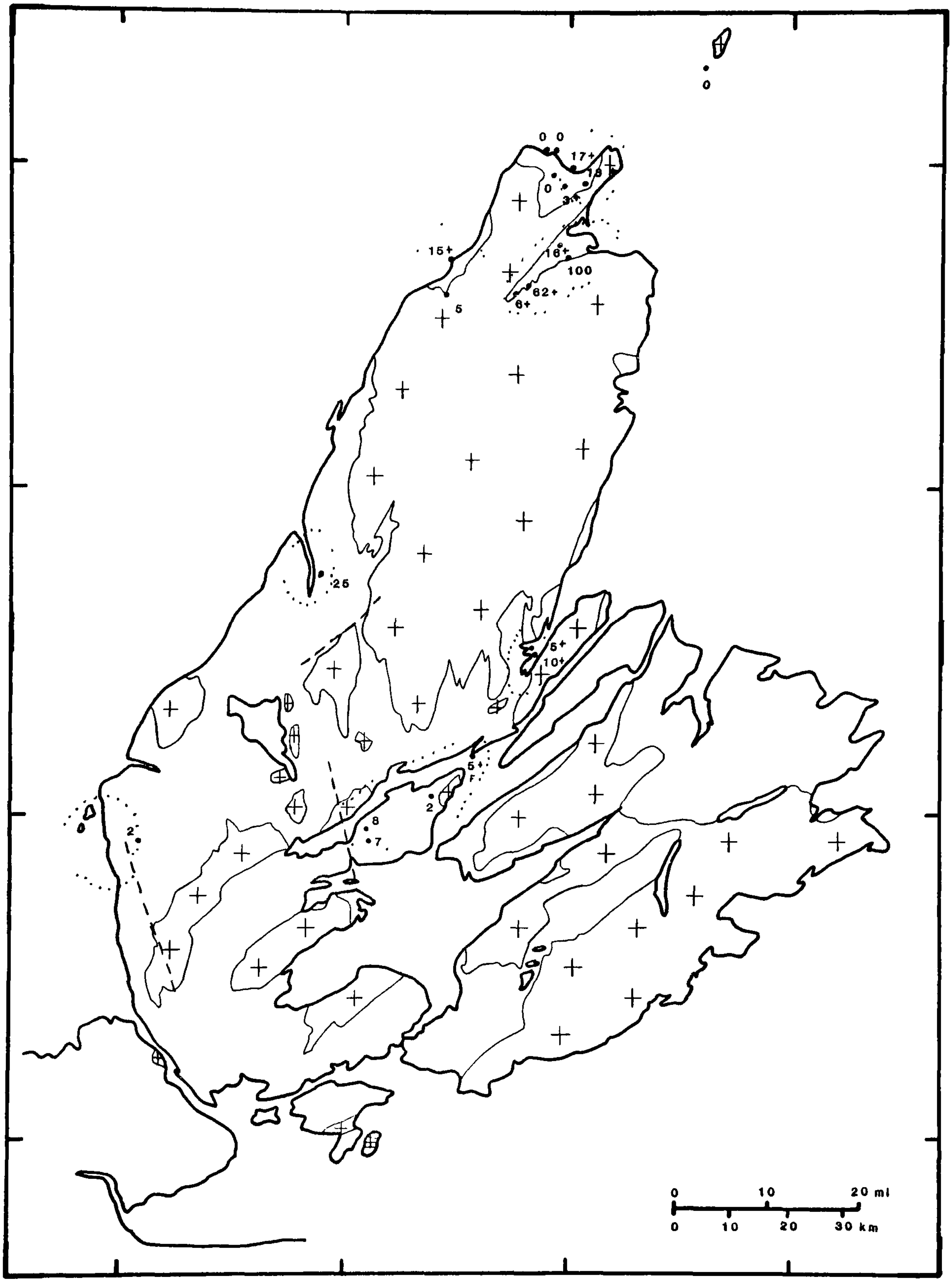


Figure 85. Isopach map, S3 facies assemblage. Thickness values in metres.

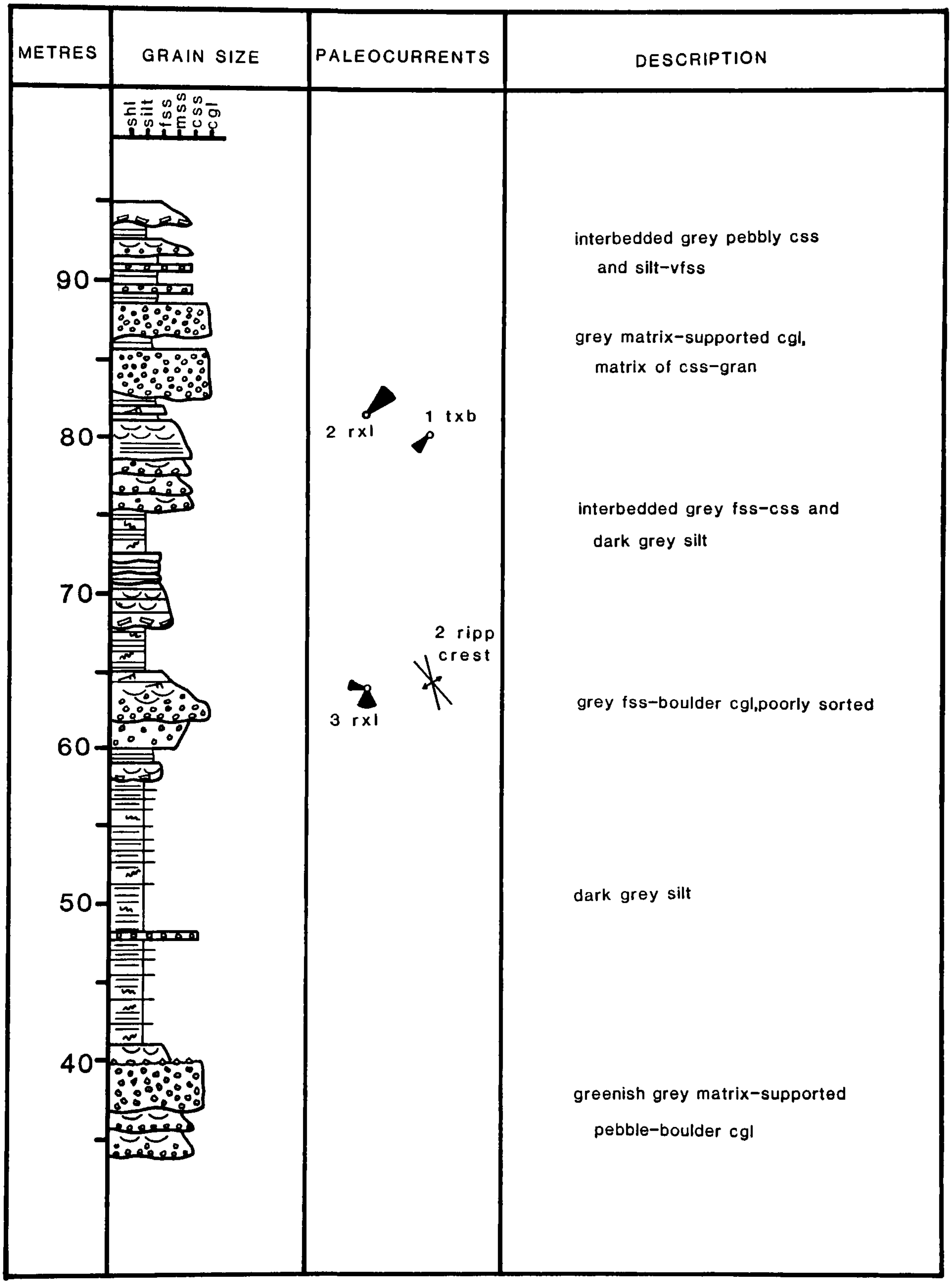


Figure 86. Type example, S3 facies assemblage, Conglomerate Creek.

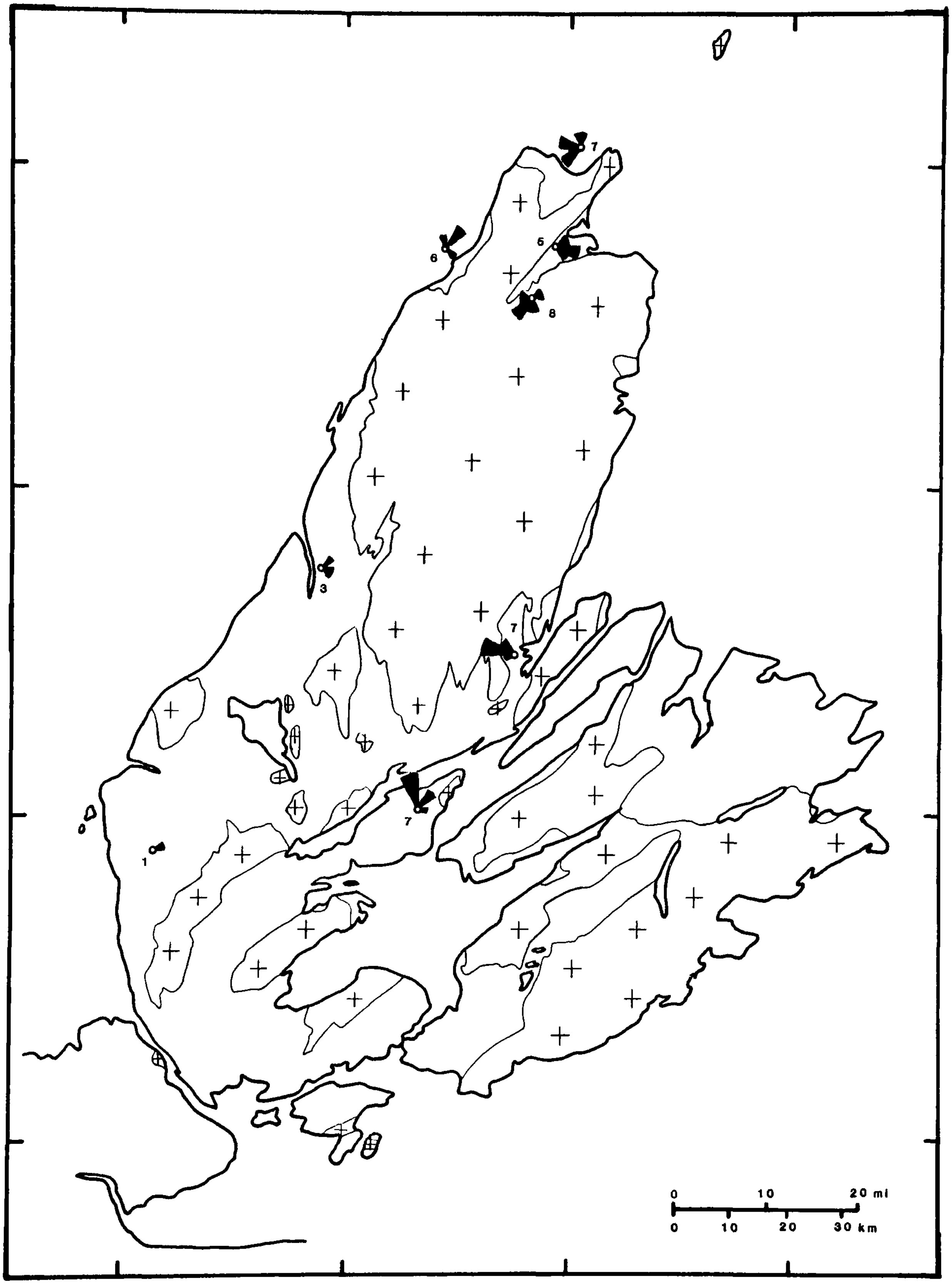


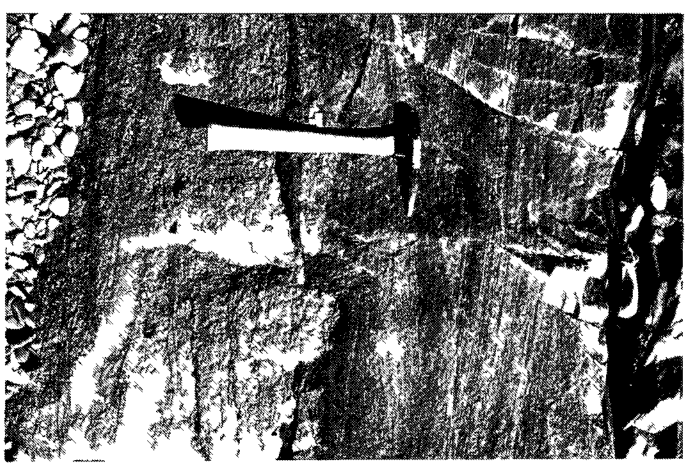
Figure 87. Paleocurrent map, S3 facies assemblage. Most measurements are from trough cross stratification, ripple cross lamination and pebble imbrication.

Figure 88. Grey medium to coarse sandstone in sharp-based bed, with rip-up clasts of mudstone, overlying grey siltstone, S3 facies assemblage, Black Cliffs (section 61).

Figure 89. Greenish grey pebbly coarse sandstone with pebble imbrication, S3 facies assemblage, McLeod Point (section 35).

Figure 90. Medium to coarse sandstone with scoured base cutting down into grey siltstone (of S1 assemblage), S3 facies assemblage, Murdoch Paul's Brook (section 27).

Figure 91. Poorly sorted pebbly coarse sandstone with scoured base cutting down into fine sandstone (of S2 assemblage), S3 facies assemblage, McLeod Point (section 35).



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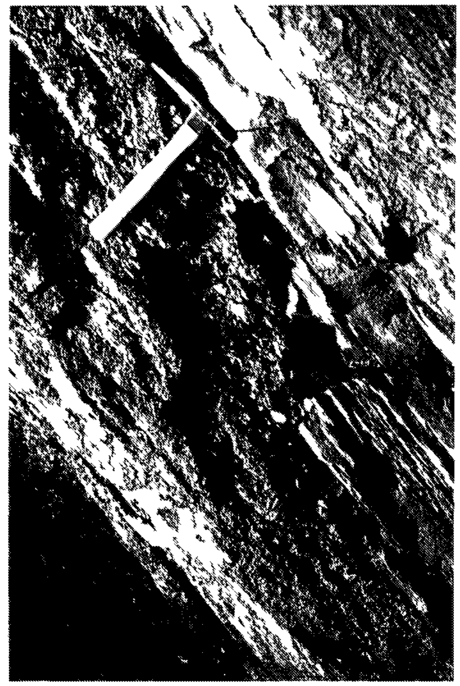

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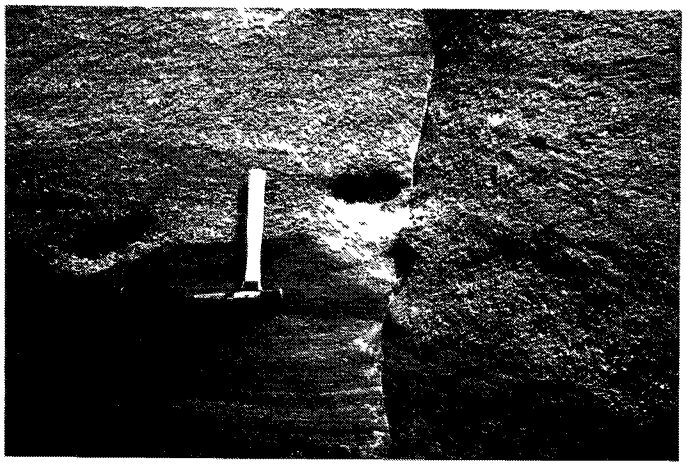
Figure 92. Grey coarse sandstone with scoured base cutting into fine sandstone, and mudstone rip-ups, S3 facies assemblage, Salmon river coast (section 69).

Figure 93. Grey matrix-supported boulder conglomerate with coarse sandstone matrix, S3 facies assemblage, Middle Aspy river (section 65). Note that matrix fills fractures in rounded granitic boulders.

Figure 94. Green clay band with volcanic shards, encased in greenish grey sandy siltstone, S3 facies assemblage, Salmon River coast (section 69). Hammer handle for scale.



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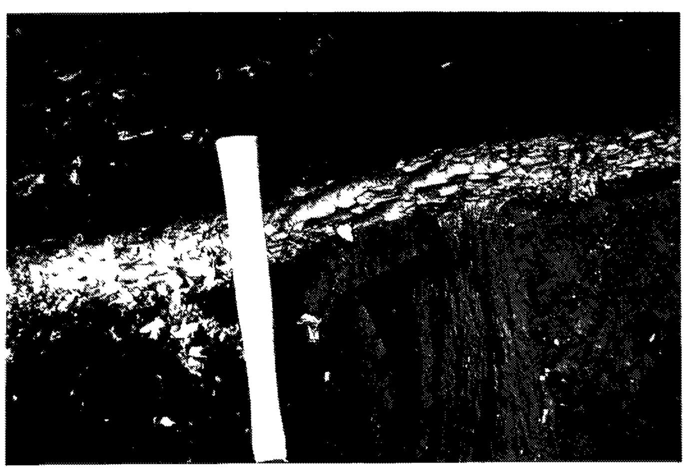

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At several sections, interbeds of greenish grey, well sorted siltstone to fine-grained sandstone are present, with trough cross stratification, ripples and root traces. At Conglomerate Creek, Sugar Brook and Sutherland's Brook S3 sediments (with associated S1) rest sharply on basement with no intervening Craignish deposits. At Salmon River Coast this lithofacies also includes two bright greenish-yellow clay bands 5 cm thick which contain a few scattered, cuspate (?) volcanic shards in a very altered clay-rich groundmass (T. Fowler, pers. comm.), and may represent ash fall beds (Fig. 95).

S4 Red siltstn-fine ss This facies assemblage has a cumulative thickness up to 45 m (Fig. 95) but typically occurs in units 2-10 m thick. A few isolated beds occur throughout the Strathlorne Formation (in the S1 facies assemblage), but thick units are present only in the middle and upper Strathlorne of the Baddeck and Mabou areas of western Cape Breton (near depositional sub-basin margins associated with S2 or S3) or at transitional units where Strathlorne sediments grade upward into the Ainslie Formation. Facies assemblage S4 is uncommon in northern Cape Breton, perhaps through lack of exposure. It was described from 16 outcrop and 2 drillhole sections, and Figure 96 presents a typical example from McLeod Point. A total of 29 paleocurrent measurements were collected from 7 sections, all in western Cape Breton, and all from ripple cross lamination (Fig. 97).

The predominant lithofacies is brick-red to reddish grey or brown, massive, micaceous, calcareous sandy siltstone (Figs. 96, 98). Bright green calcareous streaks and nodules (Fig. 99), root traces, wood fragments, desiccation cracks (Fig. 100), and small-scale loading structures are ubiquitous. Laminae, up to 5 cm thick, of red very fine sandstone with sharp bases and rippled tops are common. These ripples typically have straight crests and symmetrical shape but may contain cross lamination (Fig. 101).

Red, silty, micaceous, very fine- to fine-grained sandstone is commonly interbedded with the above lithofacies. Beds are up to 3 m thick, have sharp scoured bases (rarely with a thin pebble lag), and internal horizontal, low-angle, or contorted lamination. Trough cross stratification is uncommon but most beds have ripples at the top. In thin section, these sandstones have angular, tabular-shaped quartz grains with hematite rims, 15% clay matrix and calcareous cement, and are classified as quartz arenite to quartz wacke. In the Baddeck area greenish grey limestones up to 50 cm thick, with gradational bases and sharp tops are encased in the red siltstone (Fig. 96). Some limestones are thinly laminated while others are sandy and massive, but desiccation cracks, root traces, and irregular vugs are typical (Fig. 102).

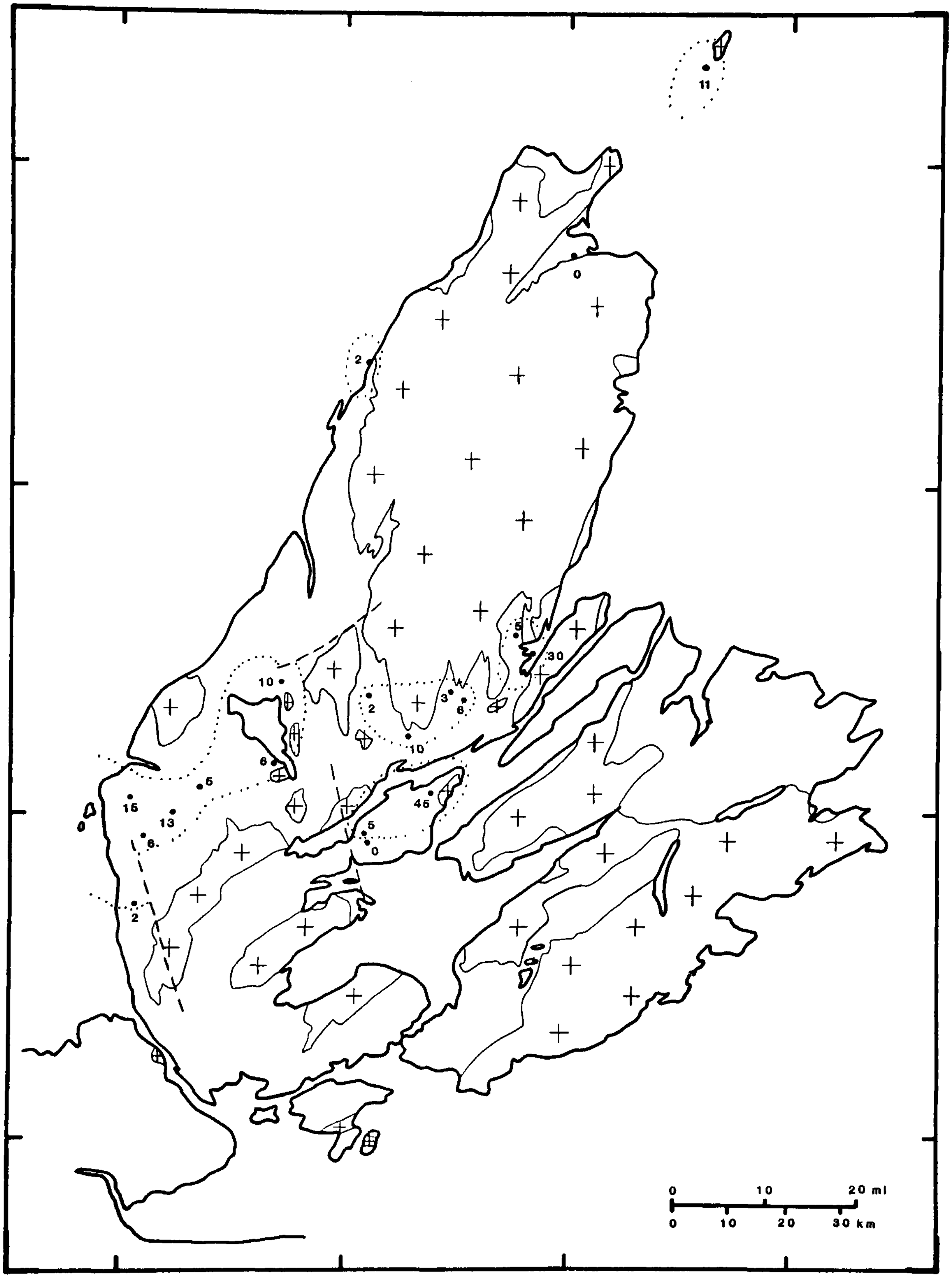


Figure 95. Isopach map, S4 facies assemblage. Thickness values in metres.

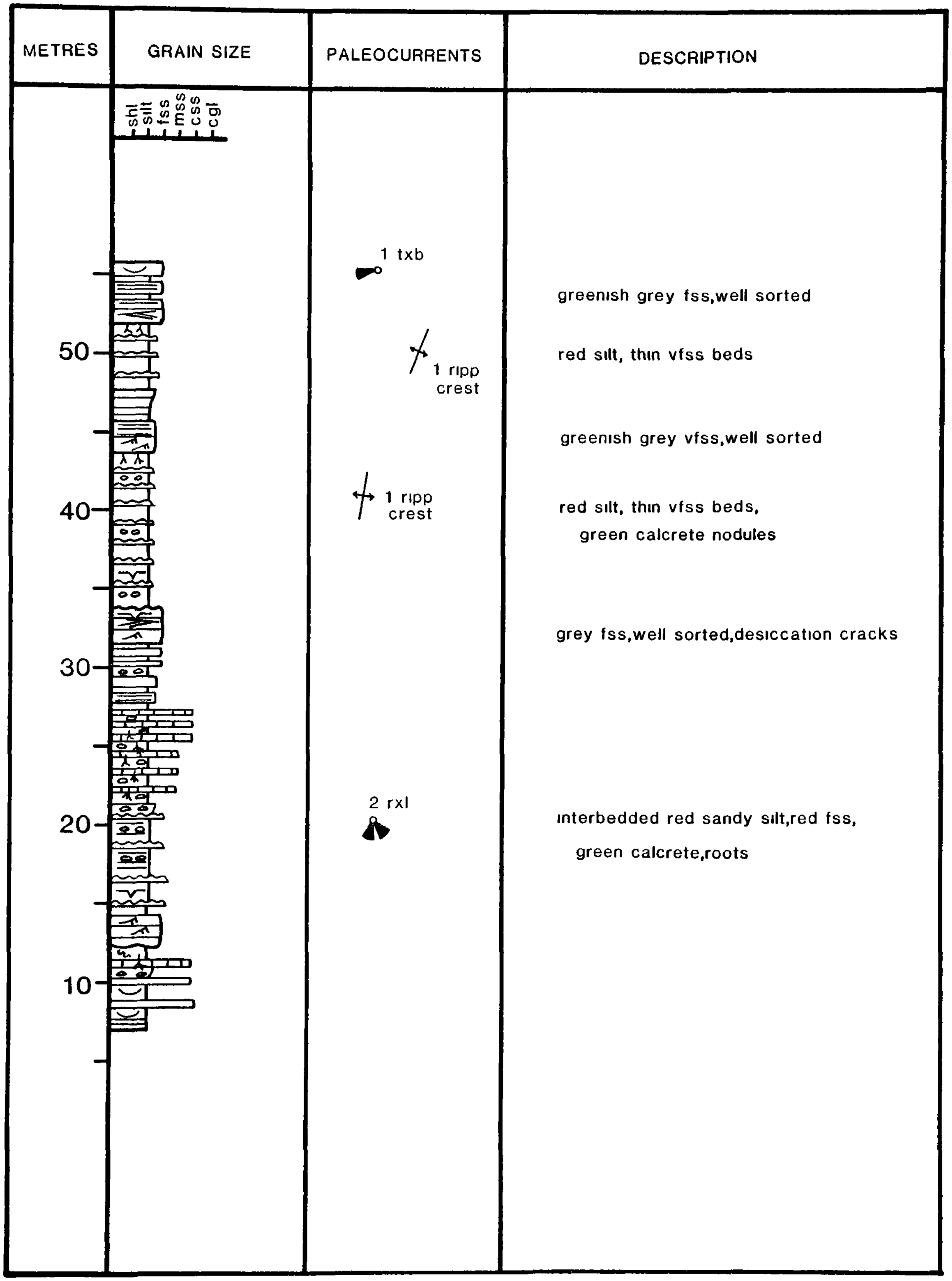


Figure 132. Type example, S4 facies assemblage, McLeod Point.

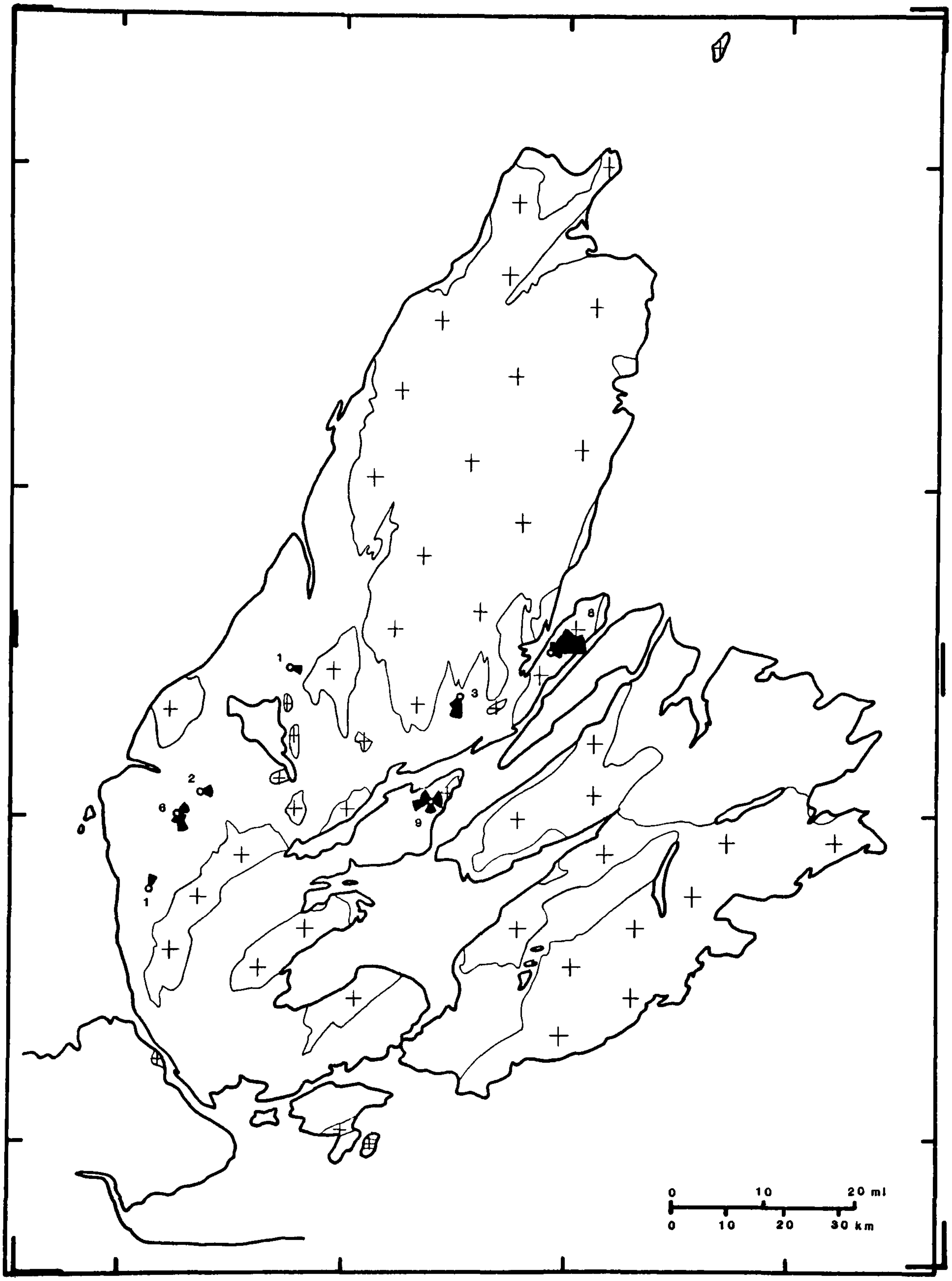


Figure 97. Paleocurrent map, S4 facies assemblage. All measurements are from ripple cross lamination.

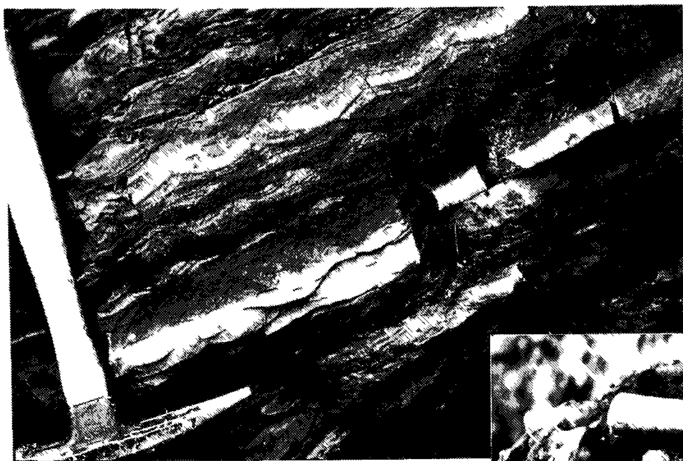
Figure 98. Red siltstone with greenish sandstone beds which have symmetrical ripples, S4 facies assemblage, McLeod Point (section 35).

Figure 99. Red sandy siltstone with green calcareous nodules and root traces, S4 facies assemblage, McLeod Point (section 35).

Figure 100. Desiccation cracks on top of grey very fine sandstone, S4 facies assemblage, McLeod Point (section 35). Hammer handle for scale.

Figure 101. Symmetrical ripple forms with straight crests and current ripple cross lamination, S4 facies assemblage, Murdoch Paul's Brook (section 27).

Figure 102. Greenish grey limestone with root traces, encased in red siltstone, S4 facies assemblage, McLeod Point (section 35).



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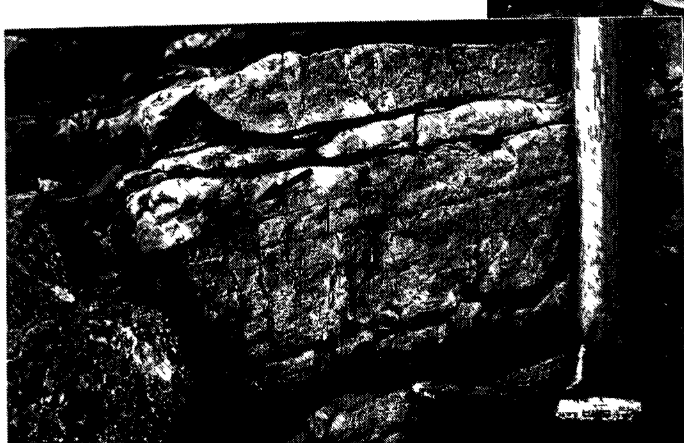

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AINSLIE FORMATION

The Ainslie Formation, up to 660 m thick (Fig. 103), conformably overlies and intertongues with the Strathlorne Formation and is sharply, but apparently conformably, overlain by the basal Macumber Formation of the Windsor Group. The great thickness recorded at some sections may be partly due to thrust repetition as described in Chapter 2 but it is apparent that there are large thicknesses of coarse-grained red sediments in the Ainslie near some original sub-basin margins. Away from those margins, thicknesses of 100-200 m are more typical. For this project the Ainslie has been studied at 38 outcrops and 8 drillhole sections. It is poorly represented in northern Cape Breton and most of the following information is from western Cape Breton where exposure is excellent. It generally consists of grey and red sandstone, siltstone and conglomerate with a distinct lateral variation from original sub-basin margins (thick, red, coarse-grained) to sub-basin centres (thin, grey, fine-grained). The Ainslie Formation can be divided into 3 facies assemblages (Table 4).

<u>A1 Red/grey pebbly coarse ss-cgl</u> This facies assemblage is up to 660 m thick (Fig. 104), although it thins to zero over a short distance toward sub-basin centres. It was described from 23 outcrops and 3 drillhole sections. It conformably overlies the Strathlorne Formation or other Ainslie facies assemblages and is sharply overlain by the Macumber Formation. It has a localized distribution, present only close to original sub-basin margins and almost exclusively in the upper Ainslie as one or several large-scale coarsening-upward sequences. A typical example from Whycocomagh Hwy 105 is presented in Figure 105. In western Cape Breton the colour is always red, while in northern Cape Breton some of the beds are grey. A total of 167 paleocurrent measurements were collected from 10 sections, most in western Cape Breton, and mostly from trough cross stratification or pebble imbrication (Fig. 106).

The predominant lithofacies is red to red-brown, thick-bedded conglomerate, granulestone and pebbly coarse sandstone (Fig. 105). The conglomerates range from sand matrix-supported to clast-supported. In northern Cape Breton the sediments may be grey and more arkosic, but otherwise are similar. Beds are distinct or indistinct, but are 1-5 m thick, with sharply scoured bases, sharp tops and a general fining-upward trend. The sediments are massive and poorly organized (eg: Donaldson Coast, Whycocomagh, Murdoch Paul's Brook) (Figs. 107, 108) or well stratified (eg: Green Point, Red Point,

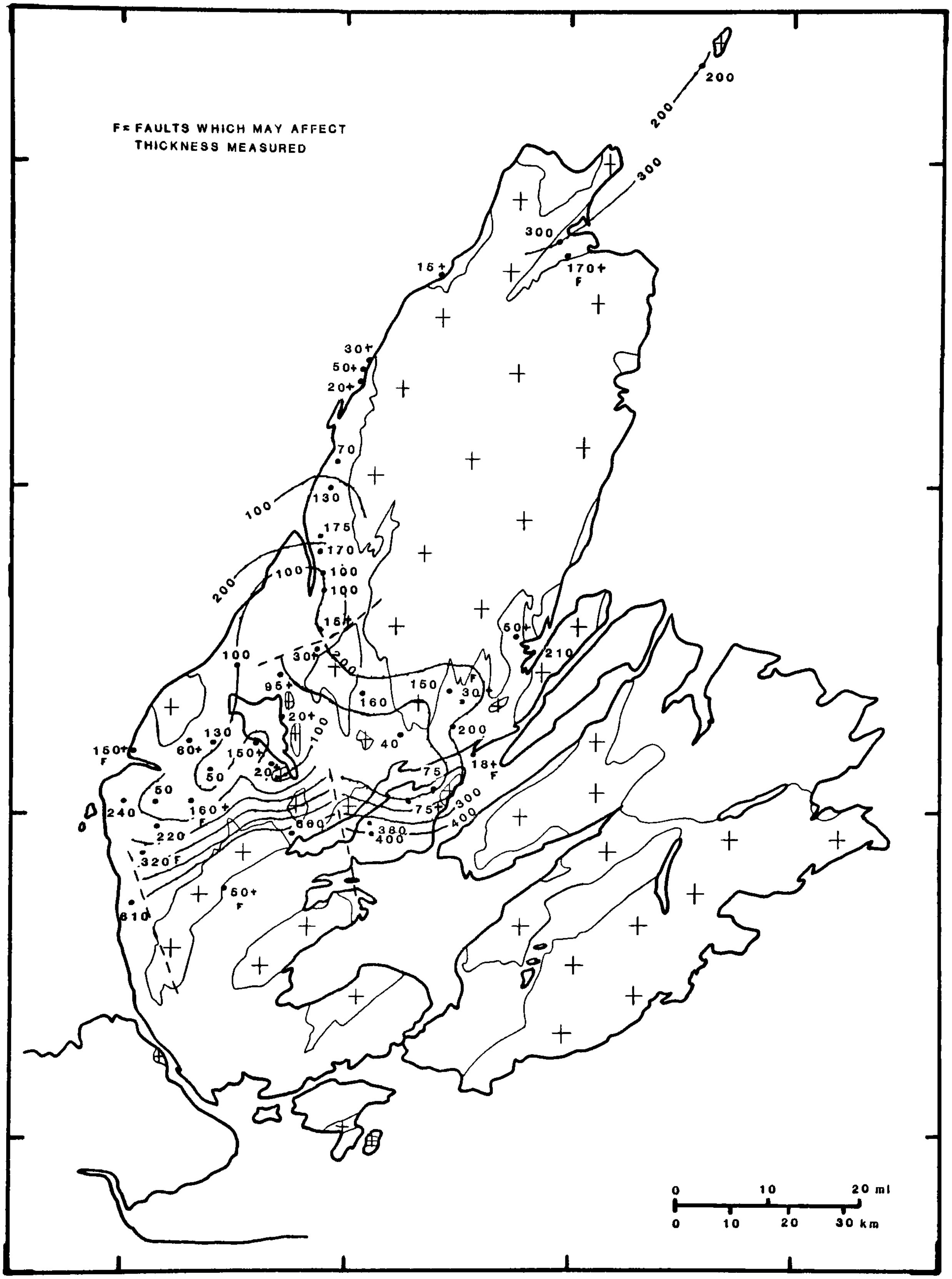

Figure 103. Isopach map, Ainslie Formation. Thickness values in metres.

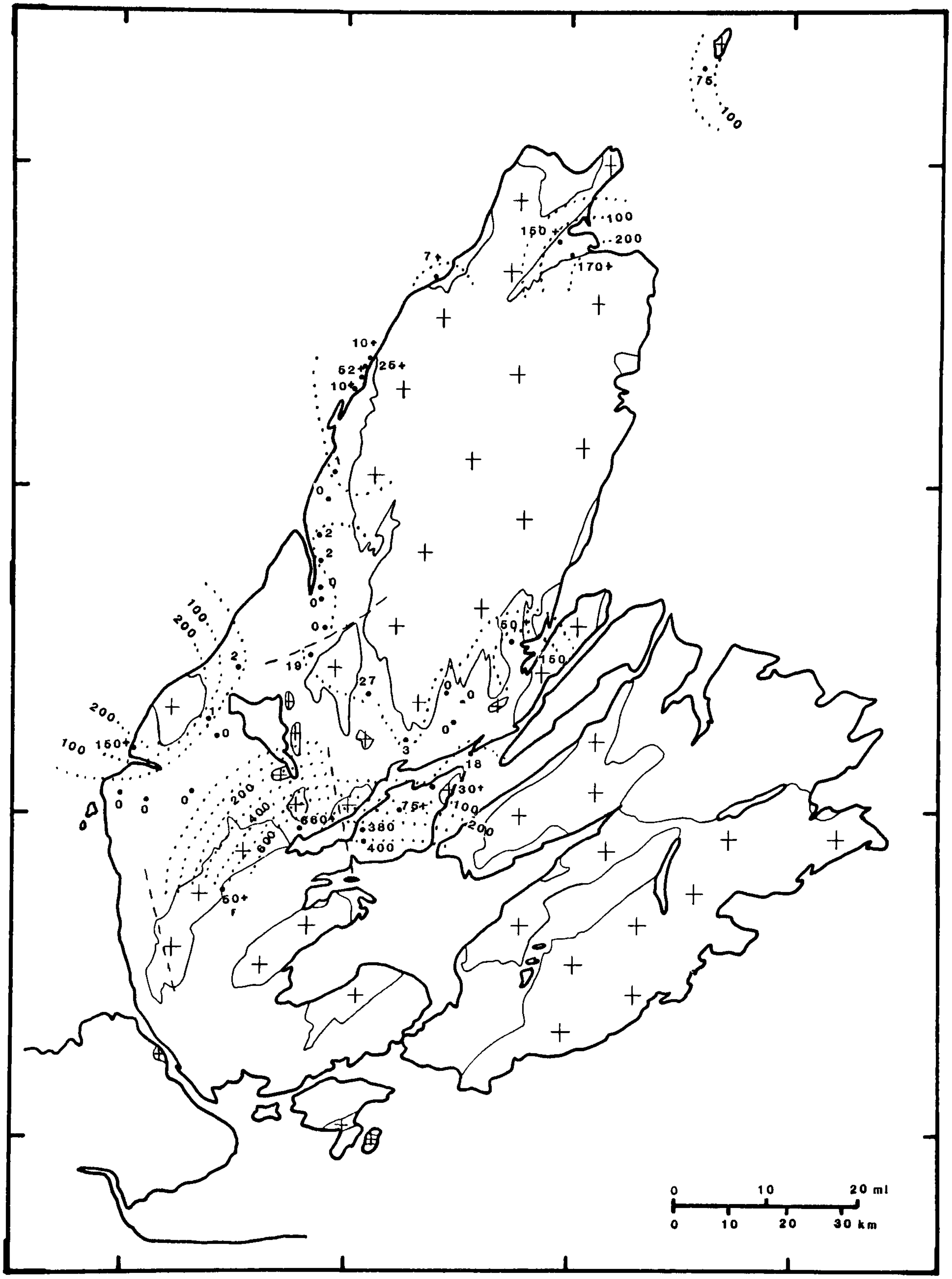

Figure 104. Isopach map, A1 facies assemblage. Thickness values in metres.

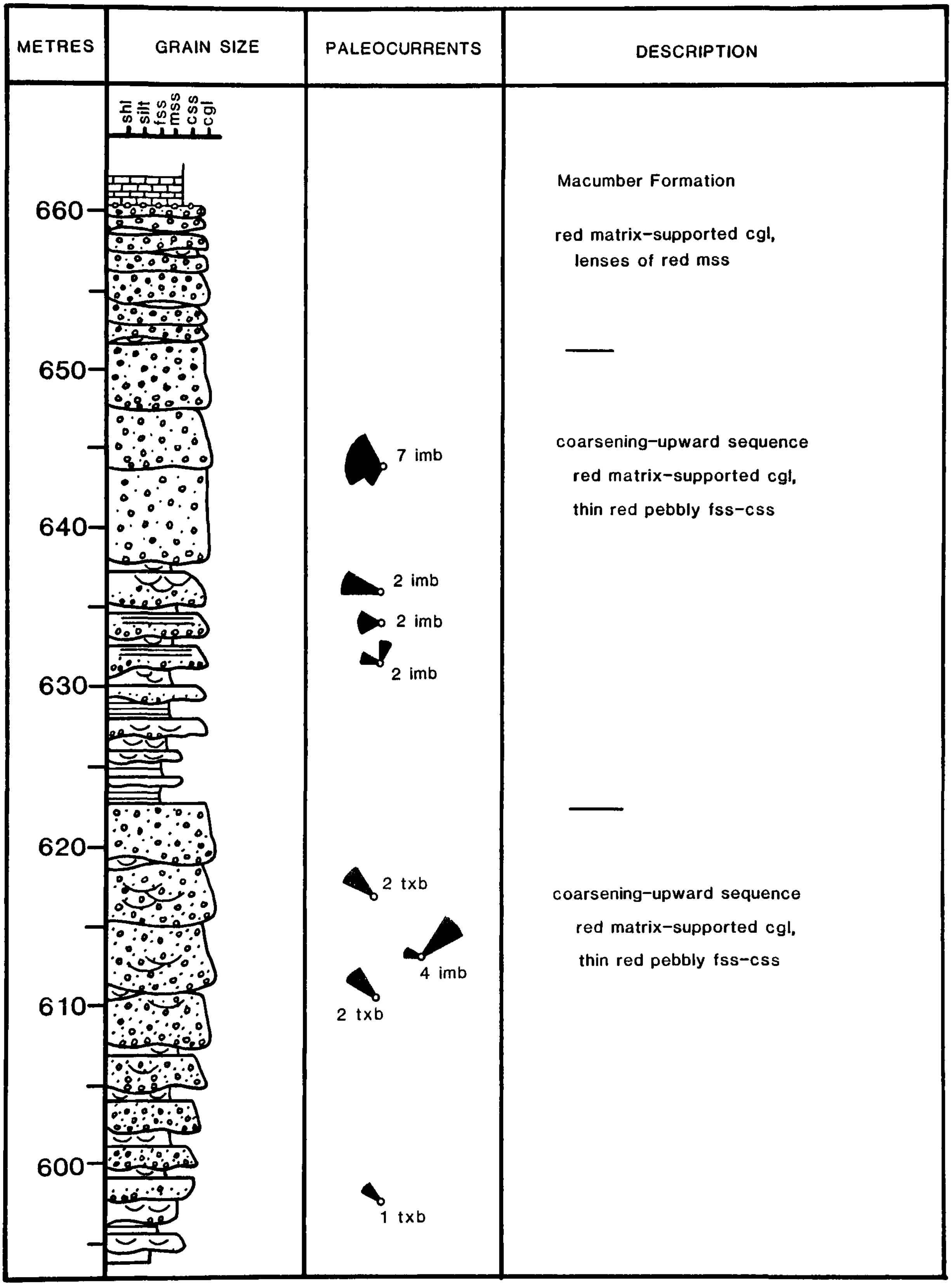

Figure 105. type example, A1 facies assemblage, Whycocomagh Hwy. 105.

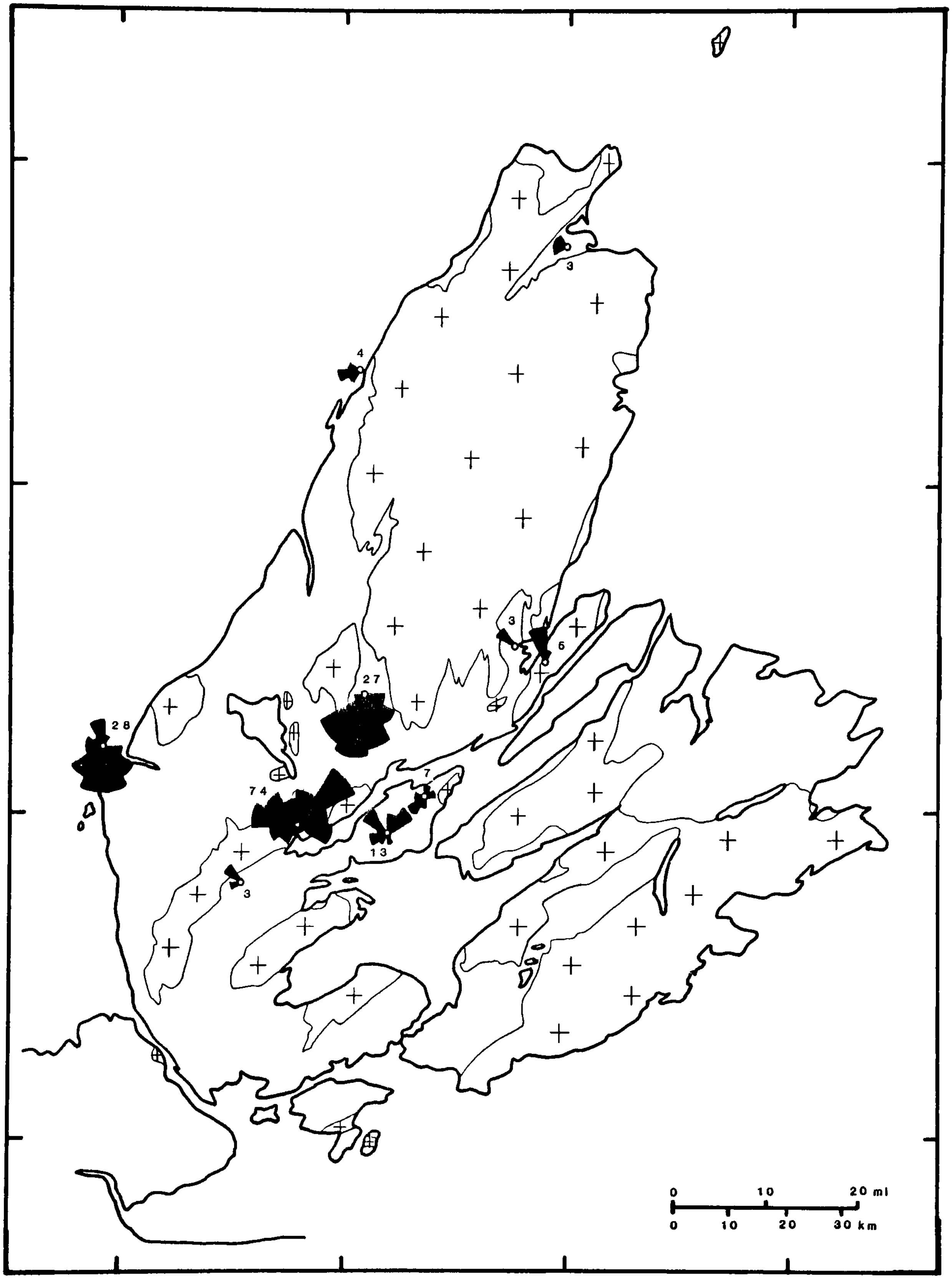

Figure 106. Paleocurrent map, A1 facies assemblage. Most measurements are from trough cross stratification and pebble imbrication.

Murdoch Paul's Brook) (Figs. 107, 108) or well stratified (eg: Green Point, Red Point, McLeod Point) (Figs. 109, 110). Beds commonly pinch out laterally over several tens of metres as broad thin lenses. Bed bases may be difficult to discern but are commonly marked by undulating scour surfaces (Fig. 111) or distinct scour pockets up to 2 m wide by 0.5 m deep where clasts are concentrated (Figs. 112, 113). When exposed in plan view, as at Green Point, these scour pockets are ovoid and several metres in long dimension (Fig. 114). Upper parts of beds commonly contain sharply defined lenses of red medium-grained sandstone with trough cross stratification (eg: Green Point, Murdock Paul's Brook) (Fig. 115), or massive silty medium-grained sandstone with "floating" pebbles (eg: Green Point). These lenses are commonly only partially preserved beneath scour surfaces and some are deformed (syn-sedimentary). The incidence of these lenses increases and their grain size decreases away from original sub-basin margins. Broad, shallow trough cross stratification is common (Fig. 116) except in the coarsest grain sizes, where poorly- to well-developed clast imbrication is predominant (Fig. 117).

At sections where large thicknesses are present (eg: Green Point, Whycocomagh) the beds are arranged into large-scale coarsening-upward sequences 35-100 m thick (Fig. 105). The coarsening-upward trend is most obvious in the variation of maximum clast size (from about 5 cm to 35 cm upward). The maximum clast size to bed thickness ratio at Whycocomagh ranges from 0.01 to 0.10. Clast lithologies are very diverse and include granitic, volcanic, meta-sediment and a few (?)Lower Paleozoic fossiliferous limestone types. The clasts have hematite rinds and are set in a matrix of moderately to well sorted coarse sandstone to granulestone which may be silty (Fig. 118). Calcareous cement has occluded all porosity. The matrix can be classified as litharenite to sublitharenite.

Red sandstone lenses in thin section have angular quartz and rock fragments surrounded by hematite rinds, with bimodal sorting, up to 10% argillaceous matrix (especially in northern Cape Breton) and calcareous cement. There is no preserved primary porosity but 3-4% secondary porosity occurs in some samples. In several sections (eg: Murdoch Paul's Brook, La Bloc, North Aspy River) units up to 5 m thick of bright red, massive, silty very fine sandstone with scattered pebbles occur in the lower parts of the facies assemblage. These units have thin irregular beds of green or cream-coloured calcareous nodules and calcified rootlets (Fig. 119). The uppermost 10 cm to 2 m of A1, directly beneath the Windsor Group, is always green-stained, calcareous, and more

Figure 107. Red, poorly-sorted, matrix-supported conglomerate, with matrix of silty coarse sandstone, A1 facies assemblage, Schoolhouse coast (section 56).

Figure 108. Red, poorly-sorted, matrix-supported conglomerate with lenses of red medium sandstone, A1 facies assemblage, Murdoch Paul's Brook (section 27).

Figure 109. Alternating coarser and finer grain sizes, A1 facies assemblage, Green Point (section 11). Note typical polymictic pebble composition.

Figure 110. Well stratified red pebbly coarse sandstone to conglomerate, A1 facies assemblage, Red Head (section 30).

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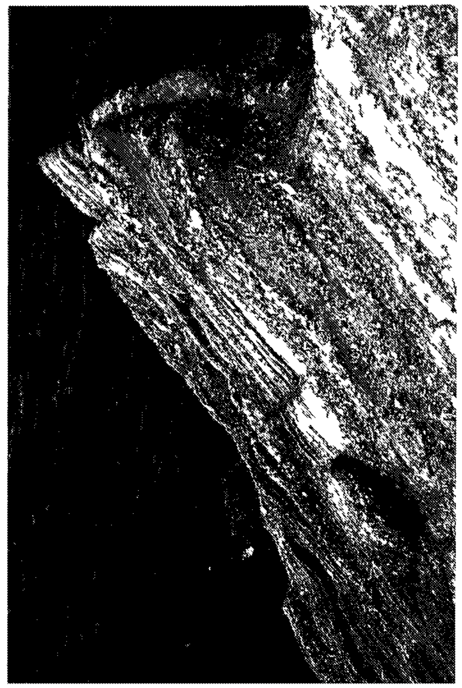

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Figure 111. Scour at base of thick cobble conglomerate which cuts down into red siltstone with calcrete nodules, A1 facies assemblage, Murdoch Paul's Brook (section 27).

Figure 112. Scour pocket marking base of one depositional unit overlying another, A1 facies assemblage, Green Point (section 11).

Figure 113. Foundered scour pocket at base of conglomerate overlying red siltstone with calcrete nodules, A1 facies assemblage, Benacadie Point (section 10).

Figure 114. Plan view of large ovoid scour pocket filled with pebbles, A1 facies assemblage, Green Point (section 11).

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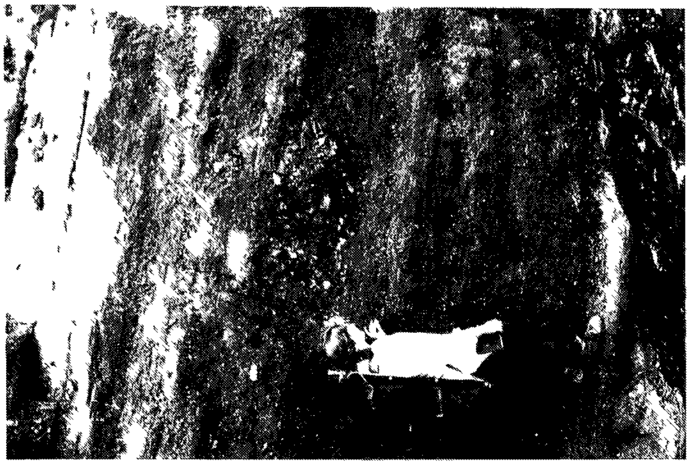

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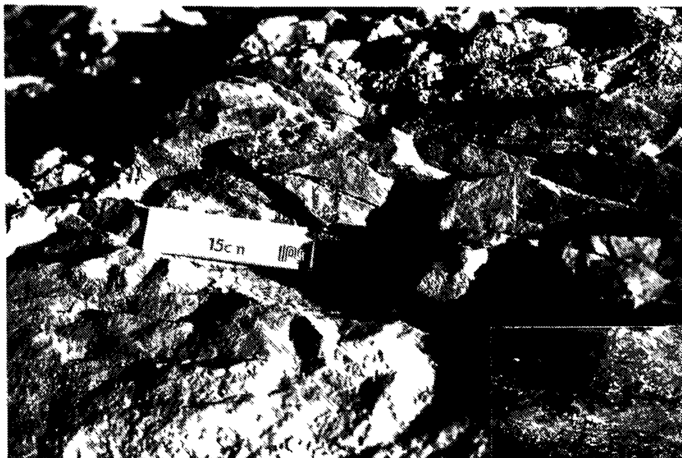
Figure 115. Discontinuous red medium sandstone lens at top of fining-upward conglomeratic unit, A1 facies assemblage, Whycocomagh (section 5).

Figure 116. Broad shallow trough cross stratification in red pebbly coarse sandstone, A1 facies assemblage, Big Brook (section 40).

Figure 117. Clast imbrication in reddish grey pebbly coarse sandstone to granulestone, A1 facies assemblage, Green Point (section 11).

Figure 118. Photomicrograph of polymictic red pebbly coarse sandstone with hematite rims, A1 facies assemblage, Murdoch Paul's Brook (section 27). 2.5x magnification.

Figure 119. Red sandy siltstone with green calcareous nodules, which occurs between coarser lithologies, A1 facies assemblage, La Bloc (section 51).



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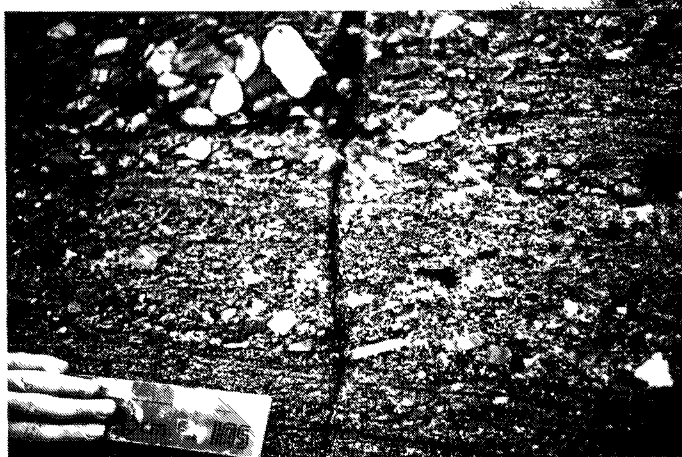

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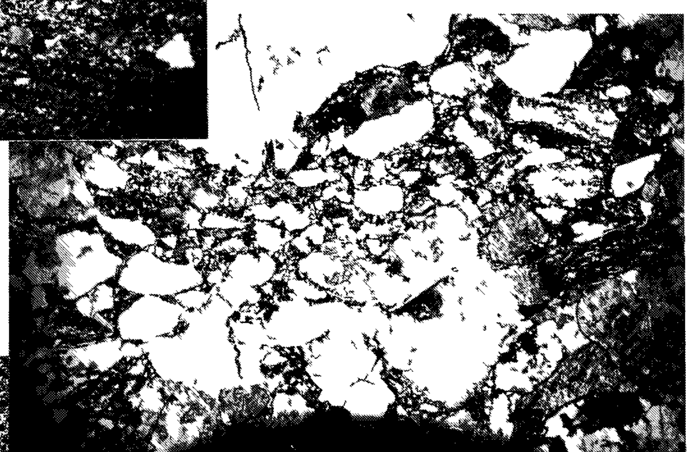

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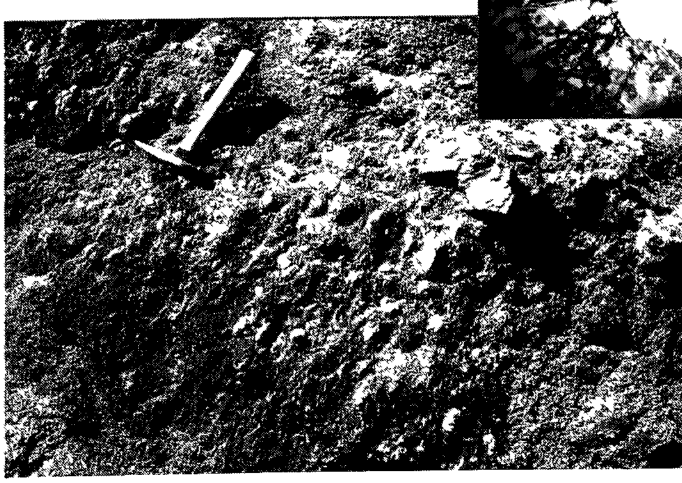

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argillaceous and poorly-sorted than the sediments described above. At Washabuck Bridge it contains small limestone rip-up clasts. This variation is important for the resource potential of the Horton Group and is discussed further in Chapter 6.

A2 Red fine-coarse ss and siltstn This facies assemblage is up to 200 m thick (Fig. 120) although it declines to zero over a short distance away from sub-basin margins. It conformably overlies the Strathlorne and is conformably overlain by facies assemblage A1 or by the Macumber Formation. Its distribution is generally in a belt between facies assemblages A1 and A3, typically in the middle and upper parts of the Ainslie Formation. It is poorly represented in northern Cape Breton. It was described from 20 outcrop and 2 drillhole sections and a type example is present on North Branch Baddeck River (Fig. 121). A total of 69 paleocurrent measurements were collected from 10 sections, almost all in western Cape Breton, and mostly from trough cross stratification and ripples (Fig. 122).

The predominant lithofacies is red to red-brown, micaceous, fine- to coarse-grained sandstone in sharp-based, fining-upward units 1-10 m thick (generally 2-5 m thick) (Figs. 121, 123). Unit bases are scoured (up to 1 m relief over 5-10 m laterally) commonly with intraformational conglomerate of imbricate red shale (Fig. 124) or distinctive green limestone clasts (Fig. 125) and twigs. Small-scale loading features are also present. There is a typical upward sequence of sedimentary structures from horizontal lamination or medium-scale trough cross stratification (Figs. 126, 127) to ripple cross lamination. Burrows, desiccation cracks (Fig. 128) and root traces may occur near the tops of units. In the Mabou area several beds have contorted lamination (eg: Southwest Mabou River, Graham River). Away from original sub-basin margins the sediments generally become finer-grained, more argillaceous and more calcareous. In thin section these sandstones are very uniform, are moderately to well sorted and are tightly packed. Angular to subangular, equant to bladed monocrystalline quartz and rock fragments (with up to 10% plagioclase and 5% K-feldspar) are surrounded by hematite rims and calcareous cement. The rocks are classified as quartz arenite to sublitharenite. All primary porosity is occluded, but some secondary porosity is present.

Red sandy siltstone in units up to 5 m thick separate the sandstone units and form the finer upper parts of the fining-upward units (Fig. 121). Streaks of very fine to fine sandstone less than 1 cm thick are ubiquitous. Less common are beds of red fine micaceous sandstone up to 1 m thick with sharp flat bases and more gradational tops.

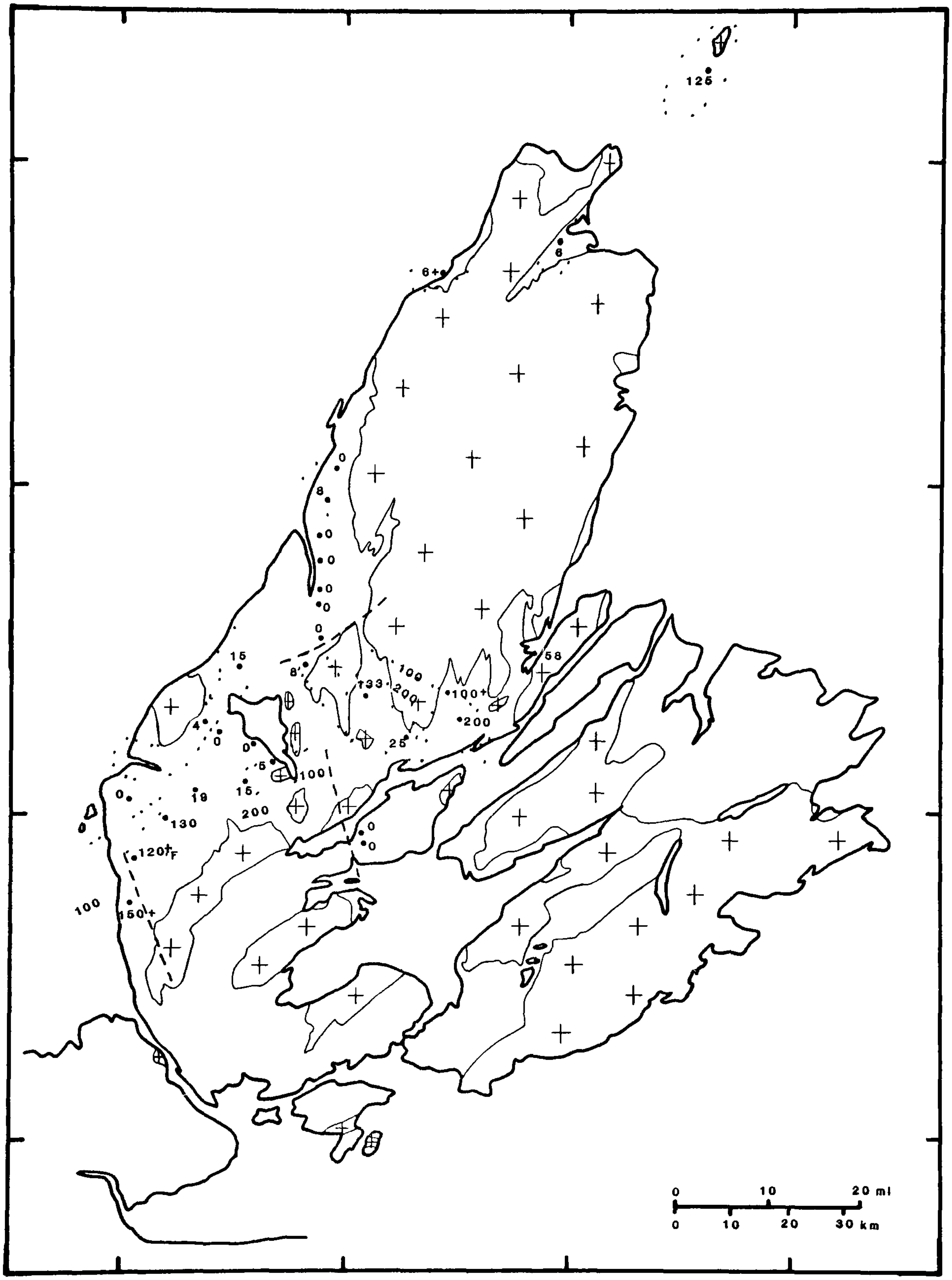

Figure 120. Isopach map, A2 facies assemblage. Thickness values in metres.

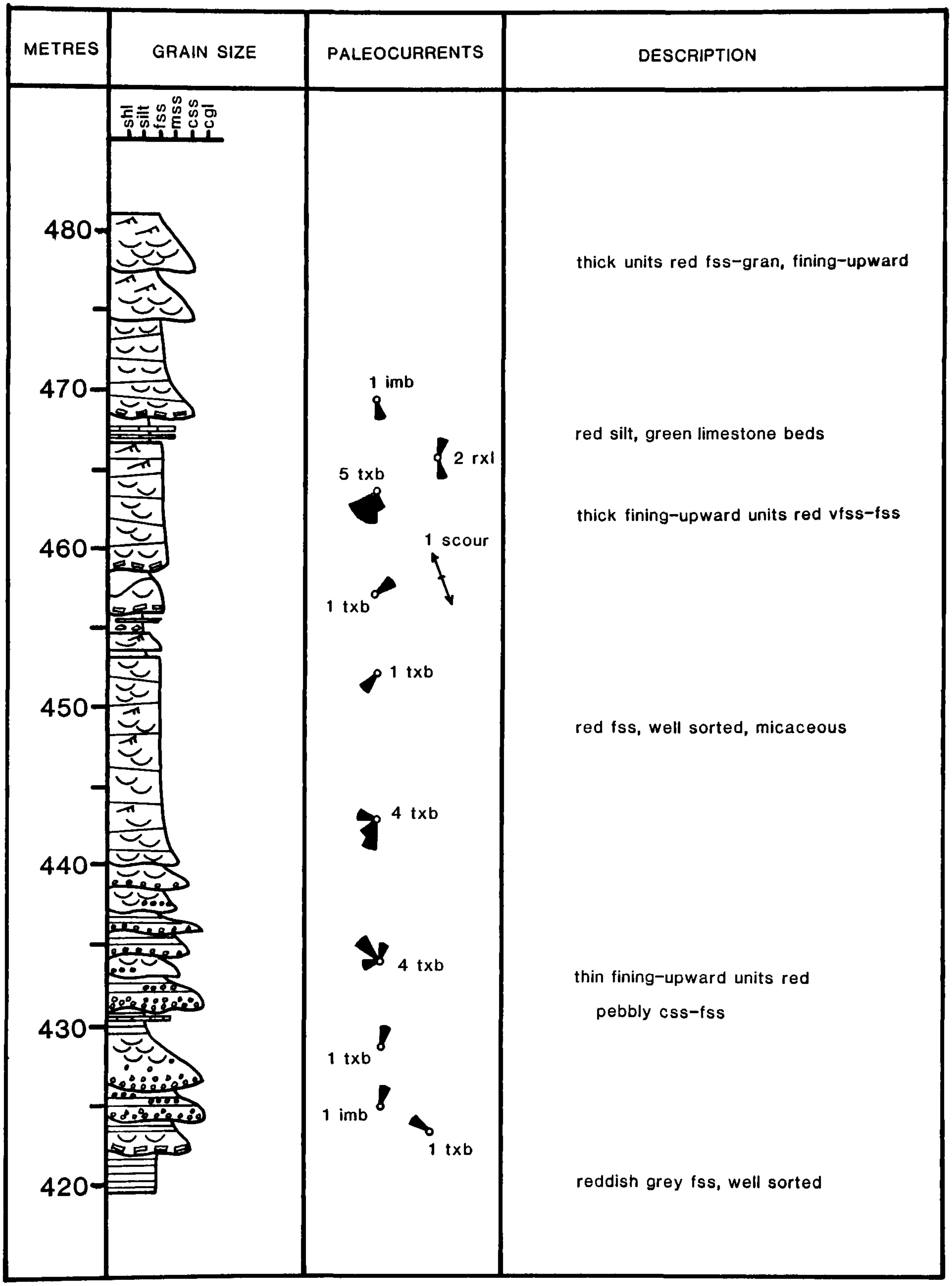

Figure 121. Type example, A2 facies assemblage, North Branch Baddeck River.

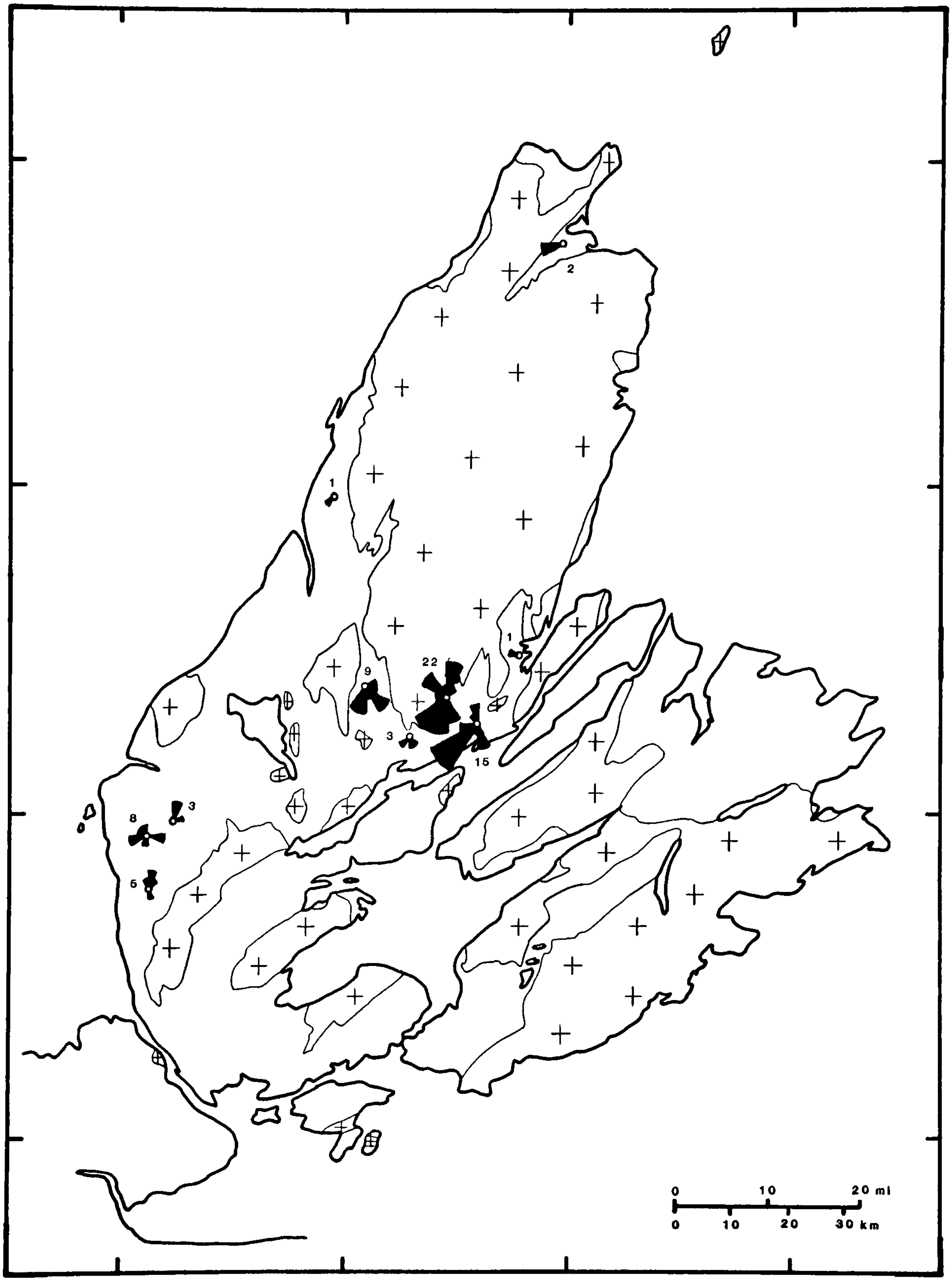

Figure 122. Paleocurrent map, A2 facies assemblage. Most measurements are from trough cross stratification and ripple cross lamination.

Figure 123. Thick red unit with scoured base which fines up from pebbly coarse sandstone to medium sandstone, A2 facies assemblage, Peter's Brook (section 31). Unit is 7m thick.

Figure 124. Large red siltstone rip-ups at sharp scoured base of thick fining-upward unit, A2 facies assemblage, North Branch Baddeck River (section 33). Scale is 15 cm.

Figure 125. Angular limestone rip-ups at base of thick red sandstone unit, A2 facies assemblage, North Branch Baddeck River (section 33). Scale is 15 cm.

Figure 126. Medium-scale trough cross stratification in red, fining-upward, medium to coarse sandstone, A2 facies assemblage, MacRae Brook (section 28).


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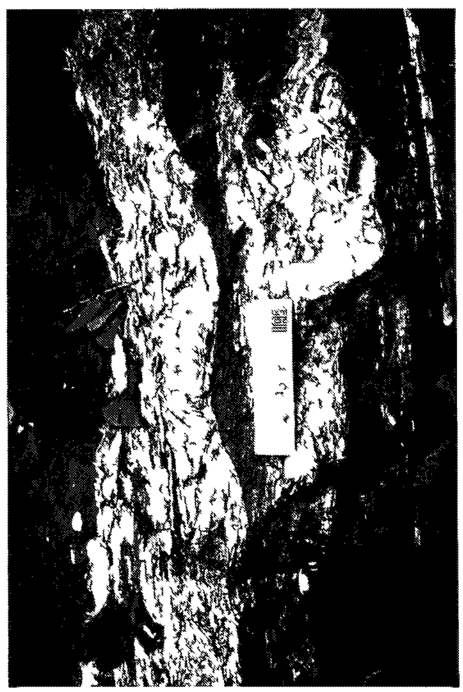
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Figure 127. Medium-scale trough cross stratification in reddish grey fine sandstone, A2 facies assemblage, Southwest Mabou River (section 3).

Figure 128. Casts of large desiccation cracks on base of red fine sandstone which overlies red siltstone, A2 facies assemblage, MacRae Brook (section 28).

Figure 129. Thin greenish limestone beds in red laminated siltstone, A2 facies assemblage, North Branch Baddeck River (section 33).

Figure 130. Photomicrograph of alternating quartz-rich vs. calcite-rich lamination in limestone, A2 facies assemblage, North Branch Baddeck River (section 33). 5x magnification.

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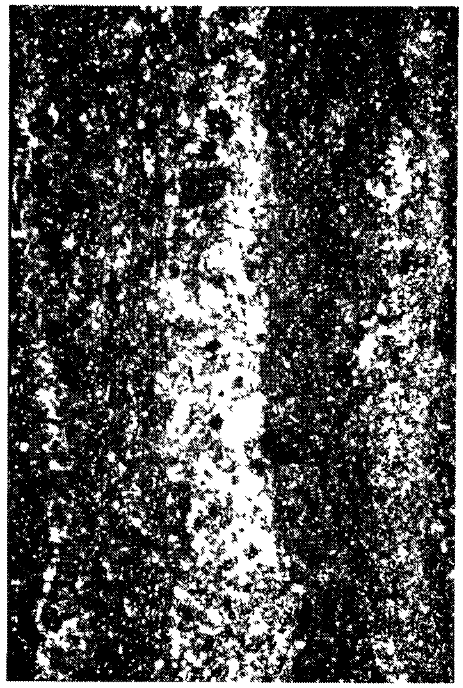

130

Internal sedimentary structures include horizontal lamination passing upward into ripple cross lamination with rooting. In some sections (eg: North Branch Baddeck River, Fig. 121) thin green- to cream-coloured limestone occurs in nodules and beds up to 10 cm thick within units of red siltstone (Fig. 129). Some are massive, with desiccation cracks, while others are laminated, and these beds are apparently the source of the distinctive limestone rip up clasts present in thick sandstone units described above. In thin section the laminated micrites have alternating calcite-rich and quartz-rich laminae (Fig. 130). Dispersed ostracod(?) shell fragments, with prismatic calcite rims, and rusty-coloured root traces(?) are common.

A3 Grey/green fine ss and siltstn This facies assemblage is up to 250 m thick (Fig. 131) but is generally 100 to 200 m thick. It was described from 22 outcrop and 4 drillhole sections. It is well developed in the Mabou, Lake Ainslie and Margaree areas of western Cape Breton (although only well exposed in a few locations), generally in basin centre positions farther from sub-basin margins than A1 or A2. Figure 132 presents a typical example from Gallant River. The A3 facies assemblage is poorly represented in northern Cape Breton. It conformably overlies the Strathlorne Formation and is apparently conformably overlain by the Macumber Formation (Fig. 132). It generally consists of interbedded fine sandstone and siltstone. A total of 60 paleocurrent measurements were collected from 10 sections, mostly from trough cross stratification and ripple cross lamination (Fig. 133).

The most conspicuous lithofacies is resistant beds of grey green very fine- to medium-grained (generally very fine- to fine-grained) sandstone in fining-upward units 3 to 15 m thick (Figs. 132, 134). These units are less common in locations away from interpreted sediment sources. The sandstone is normally well sorted, micaceous and slightly calcareous. The units have sharp flat or deeply scoured bases (Fig. 135) commonly with lags of pebbles, wood fragments or siltstone rip-ups (Fig. 136). At Angus Lake Brook and Gallant River there are angular limestone rip-up clasts (Figs. 132, 137) set in a matrix of medium to coarse grained quartz and rock fragments. These intraclasts include several types: a) pelletal micrite with ostracod fragments, b) silty öolite, c)silty micrite, d) laminated algal carbonate, e) broken öoids. Minor soft-sediment loading structures and casts of horizontal burrows are also common at unit bases. Internal horizontal lamination with current lineation (Fig. 138), medium-scale trough cross stratification and ripple cross

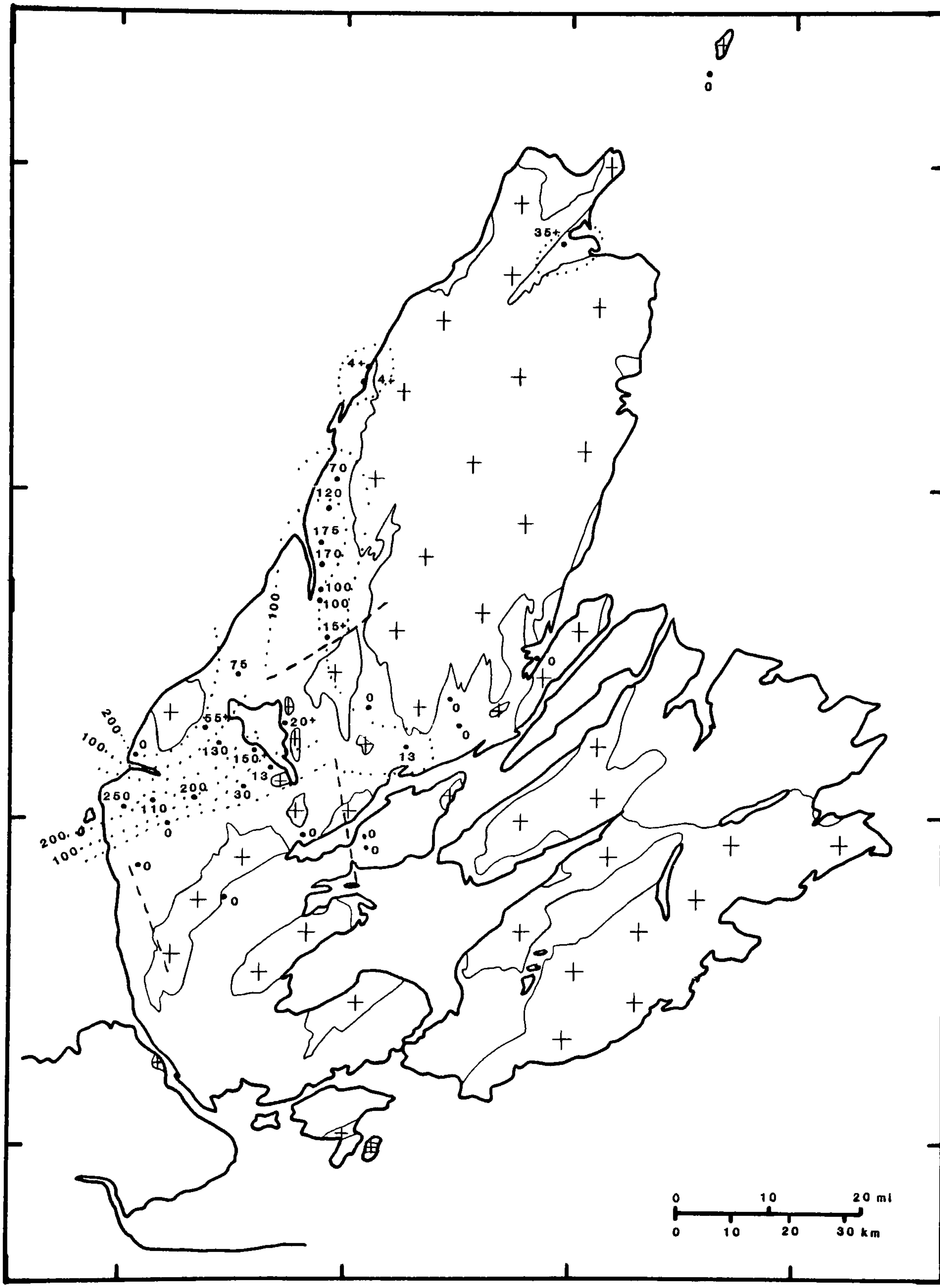

Figure 131. Isopach map, A3 facies assemblage. Thickness values in metres.

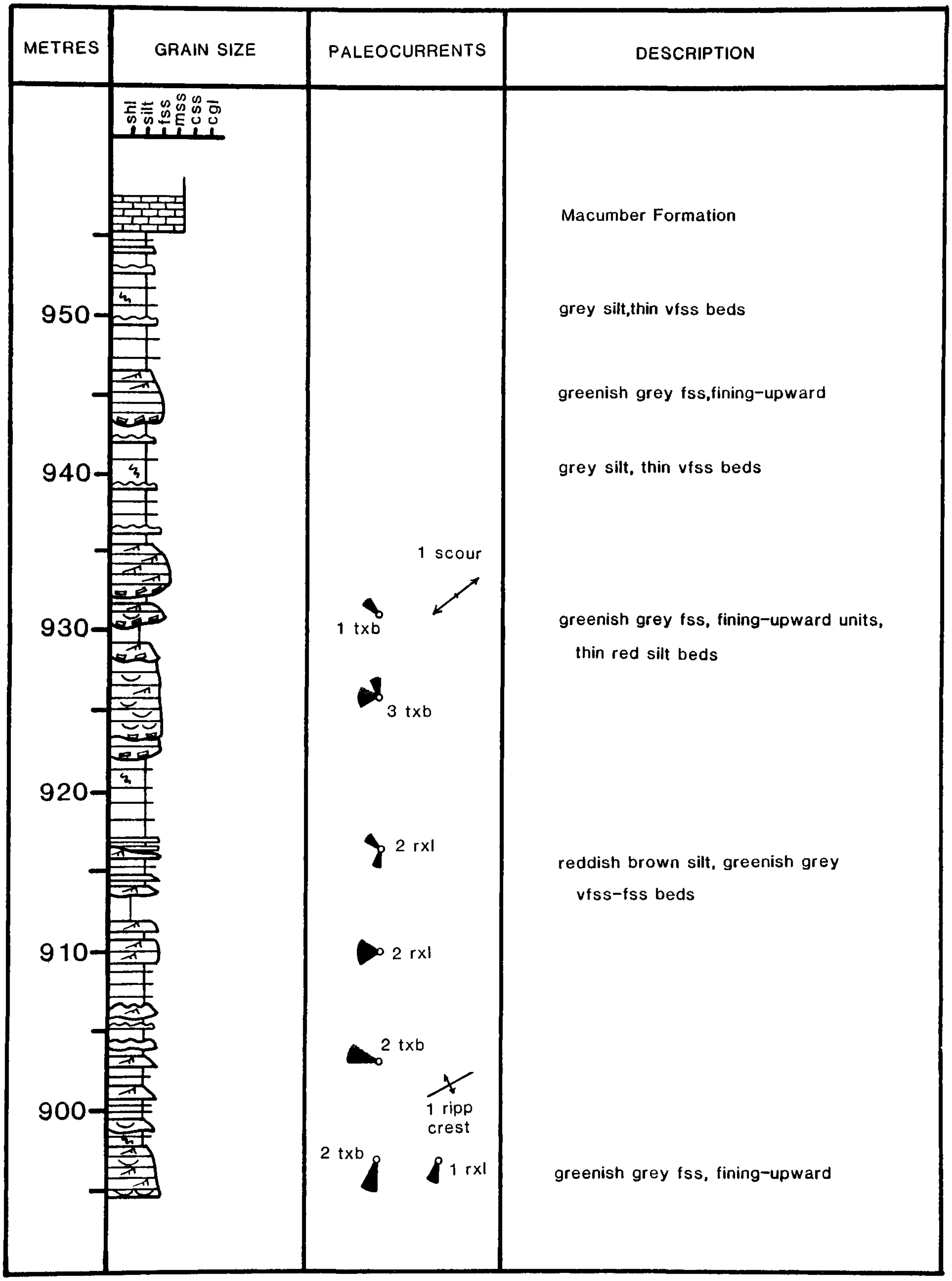

Figure 96. Type example, A3 facies assemblage, Gallant River.

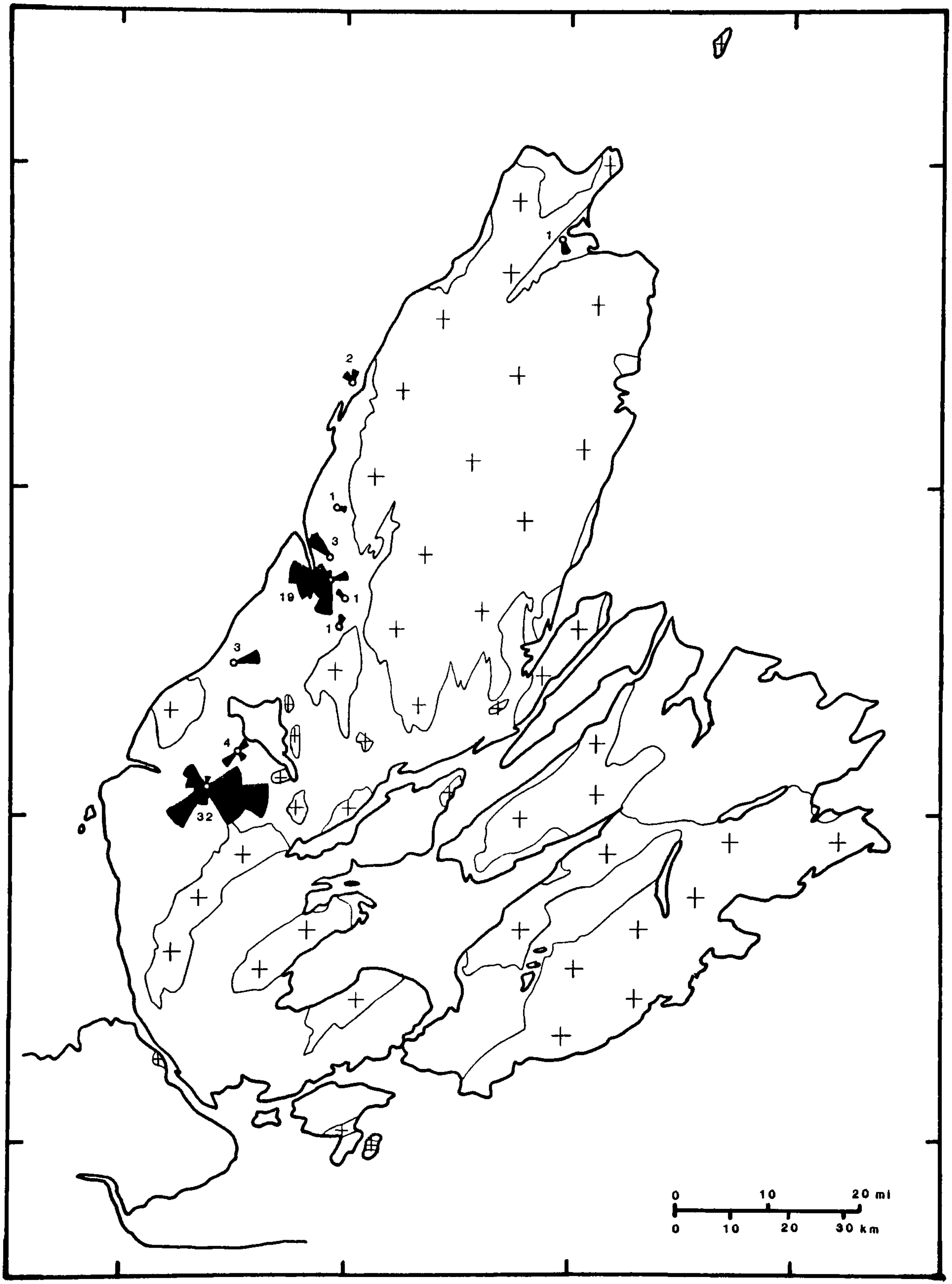

Figure 133. Paleocurrent map, A3 facies assemblage. Most measurements are from trough cross stratification and ripple cross lamination.

Figure 134. Fining-upward fine sandstone unit, 7m thick, with trough cross stratification near base and ripple cross lamination near top, A3 facies assemblage, Gallant River (section 44).

Figure 135. Deeply scoured base of thick fine sandstone unit 4m thick, which cuts down into brownish grey siltstone, A3 facies assemblage, Gallant River (section 44).

Figure 136. Lag of mudstone rip-ups and wood fragments at base of thick fining-upward fine sandstone unit, A3 facies assemblage, Southeast Mabou River (section 13). Scale is 15 cm.

Figure 137. Photomicrograph of limestone rip-ups in medium to coarse sandstone matrix, including oölitic and micritic types, A3 facies assemblage, Angus Lake Brook (section 47). 2.5x magnification.

Figure 138. Current lineation in horizontal lamination of fine sandstone, A3 facies assemblage, Southeast Mabou River (section 13). Scale is 15 cm.

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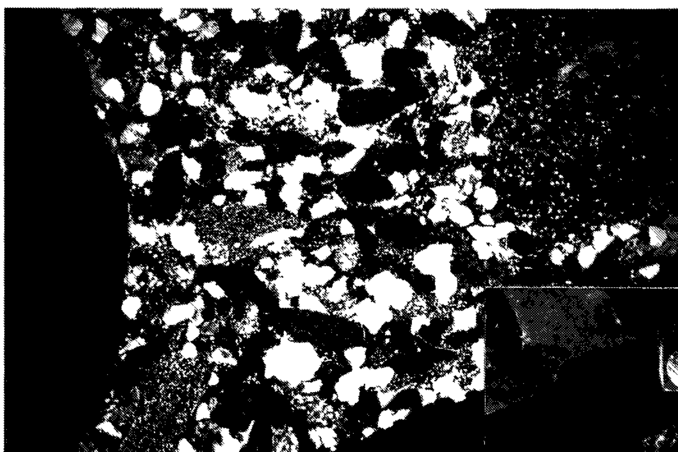

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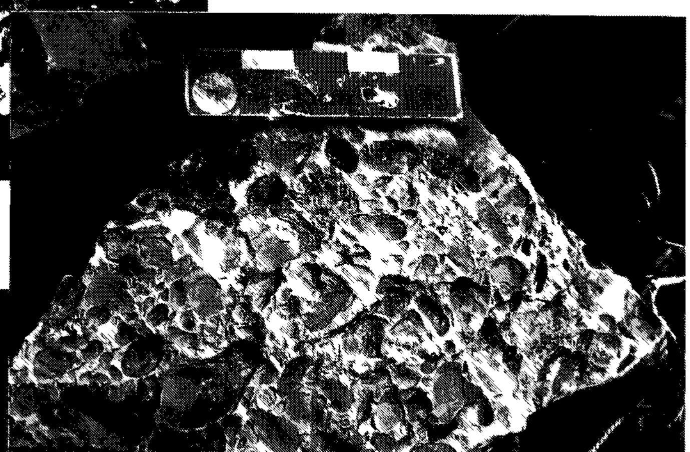

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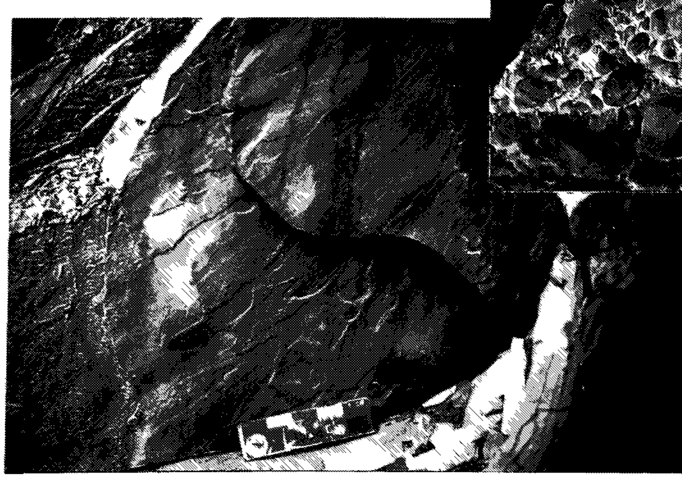

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lamination (including climbing ripples) (Fig. 139) are ubiquitous. Contorted lamination occurs at Old Bridge Brook. Many units have a slight fining-upward grain size trend, passing through a fairly gradational top (Fig. 140) with thin siltstone partings into overlying units of the siltstone lithofacies; others have no recognizable trend. These units generally occur as separate bodies but may be stacked several on top of, and cutting into, each other as evidenced by thin horizons of rip-up clasts (eg: Long Marsh Gulch, Inverness #3). Several of these units in the Lake Ainslie area have oil staining or active oil seeps. In thin section sandstones of the A3 facies assemblage have well sorted, subangular, equant framework grains of monocrystalline quartz with minor feldspar, mica and rock fragments, and are classified as quartz arenite. The texture is uniform except that 5-10% clay matrix is concentrated into alternate laminae. All primary porosity is occluded but 5-10% secondary porosity is common after calcareous cement (Fig. 141).

Separating the sandstones are units of grey, greenish grey, reddish grey or red-brown siltstone (red colours associated with numerous thick sandstone units, grey colours associated with thinner and rarer sandstone units) (Fig. 132). Reddish siltstones are sandy, bioturbated, massive and may have green calcareous nodules (eg: La Bloc), whereas grey siltstones are argillaceous, laminated and may have abundant burrowing. Thin limestone beds are present in some grey siltstones (eg: Inverness #3). Most siltstones are micaceous, slightly calcareous and have sharp-based beds of very fine- to fine-grained sandstone up to 2 m thick (Fig. 142). Tool marks, burrow casts, internal horizontal lamination and sharp rippled tops are typical of these sandstone beds. In thin section well sorted, subangular quartz grains occur in alternating clay matrix-rich and clay matrix-poor laminae. Primary porosity is occluded, but 1-2% secondary porosity is usually present.

STRUCTURAL OFFSET OF FACIES ASSEMBLAGES

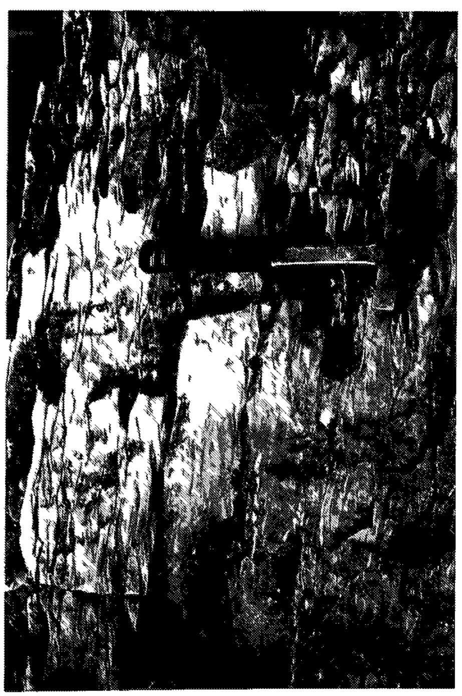
As discussed in Chapter 2, and well illustrated on Figures 57 (S1 isopach) and 104 (A1 isopach) there is some evidence to support the presence of major fault offsets in western Cape Breton. Lines of evidence include a) offset of isopach trends, b) unexpected close juxtaposition of facies assemblages, c) paleocurrent data, d) repetition of section in outcrop, e) seismic data, f) presence of isolated outliers of assemblage A1 on upfaulted basement blocks. In the Mabou area a possible explanation is the interpreted presence of a large overthrust plate which carried Horton and basement rocks northwestward by about

Figure 139. Climbing at top of very fine to fine sandstone, A3 facies assemblage, Southeast Mabou River (section 13).

Figure 140. Interbedding of siltstone and very fine sandstone at top of thick fining-upward sandstone body. A3 facies assemblage, Judique Intervale Brook (section 2).

Figure 141. Photomicrograph of well-sorted quartz arenite, A3 facies assemblage, Gallant River (section 44). 10x magnification.

Figure 142. Scour and fill in thin bioturbated fine sandstone unit encased in reddish brown siltstone, A3 facies assemblage, Gallant River (section 44).



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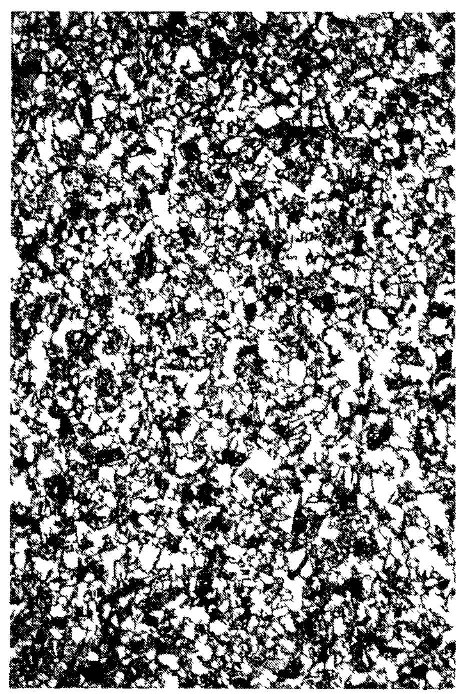

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Creignish Hills (Ferguson, 1946) and on the east side of Creignish Hills with associated paleocurrent indicators (eg: Melford, Whycocomagh) suggests this basement block was not a positive feature during Horton deposition. An apparent offset in the area north of Lake Ainslie is depicted by close juxtaposition of A1 and A3 facies assemblages in outcrops only 2 km apart along strike (Big Brook, Pat's Brook). The recognition and interpretation of these post-Horton features is vital to understanding the facies distribution, basin geometry and resource potential of the Horton Group. Therefore the tentative boundaries of proposed offset blocks are marked on each map. Fault repetition may be a factor in other areas as well.

PALYNOLOGICAL DATA

A reconnaissance collection of 53 palynological samples from 30 sections (Fig. 143 was submitted to Dr. J. Utting of the Institute of Sedimentary and Petroleum Geology. This included 37 samples from western Cape Breton (1 Craignish, 32 Strathlorne, 4 Ainslie) and 16 from northern Cape Breton (4 Craignish, 12 Strathlorne). The aim of this study was to determine the age of Horton Group Formations, but important information on organic matter type, depositional environment and thermal alteration index (TAI) is also included (App. II). Most samples contain abundant, well-preserved palynomorphs. The following summarizes Dr. Utting's findings (Paleontology Subdivision Reports 7-JU-88, 11-JU-88).

All assemblages studied are qualitatively diverse and fall into two palynological assemblage zones recognized in the Tournaisian stage of the Lower Carboniferous of Nova Scotia (Utting et al., in press). These are the 1) Emphanisporites rotatus - Hymenozonotriletes explanatus Assemblage Zone of Tn_2 age in the Craignish Formation of northern Cape Breton, and 2) the Vallatisporites vallatus Assemblage Zone of late Tn_2 to mid Tn_3 age in the Strathlorne and Ainslie Formation of both areas. The second includes a) the Umbonatisporites abstrusus - Umbonatisporites distinctus Subzone of late Tn_2 to early Tn_3 age, in a few samples from the lower Strathlorne of northern Cape Breton, and b) the Spelaeotriletes sp. A. Subzone of early to mid Tn_3 age (Utting et al., in press). All zones and subzones can be correlated with those found in Horton Group rocks elsewhere in the Maritimes and to the zonal scheme of Western Europe (Table 6) (Utting et al., in press). There is no evidence that any Horton Group sediments of Cape Breton Island are of Tn_1 or Late Devonian age, although no productive samples were obtained

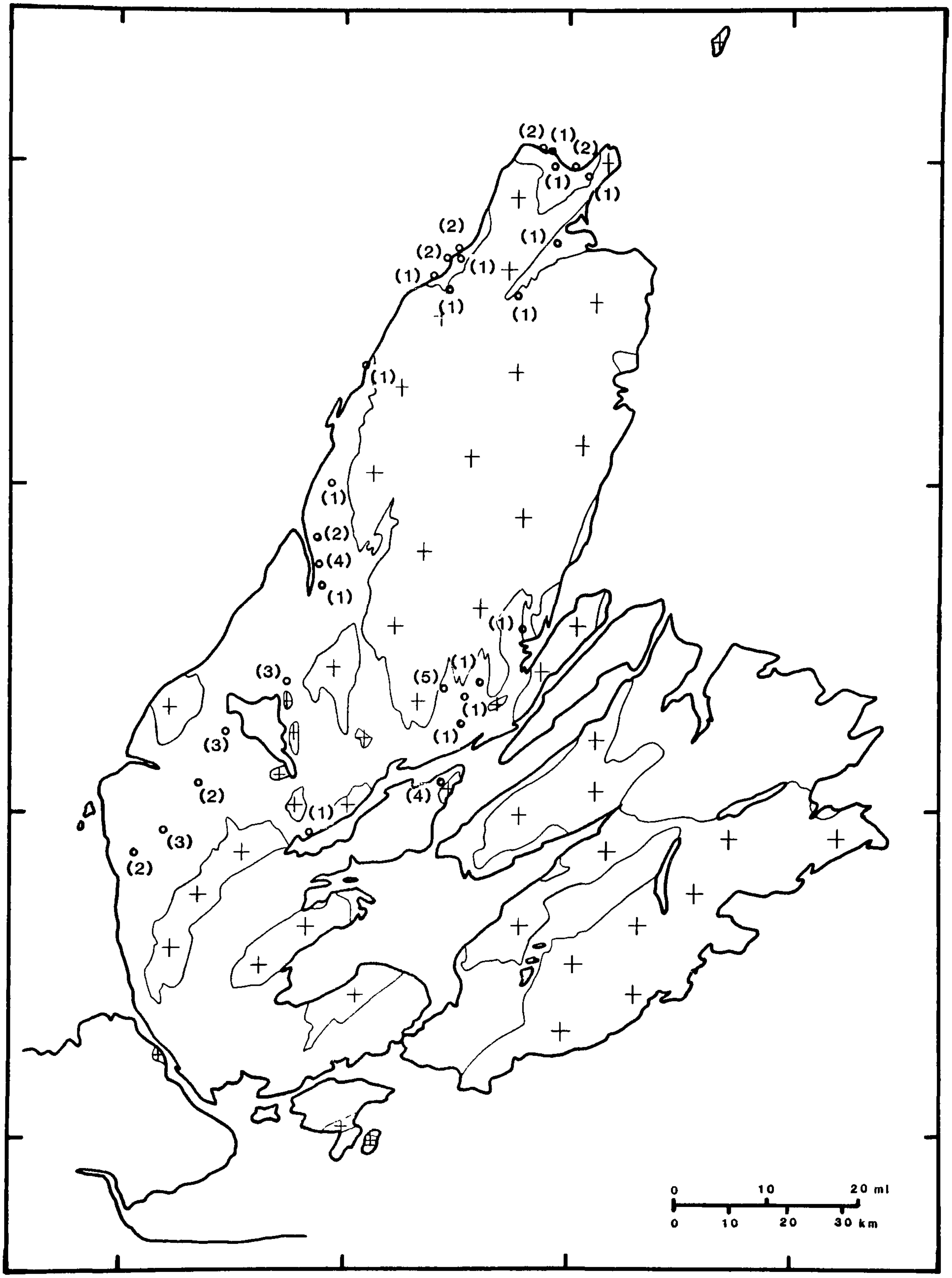

Figure 143. Map of palynological sample locations. Number in brackets refer to number of samples from each location. Refer to Appendix II for data.

SYSTEM	SERIES	SUB-SERIES	STAGES	AGE (MA)	MIOSPORE ZONES/SUBZONES OF WESTERN EUROPE	HORTON GROUP OF CAPE BRETON ISLAND	MIOSPORE ZONES/SUBZONES OF MAINLAND NOVA SCOTIA
CARBONIFEROUS	VISEAN	V3	HOLKERIAN		K. triradiatus–K. stephanephorus	Macumber Fmn	L.noctuina var.noctuina–K.stephanephorus
		V 1–V 2	ARUNDIAN		Lycospora pusilla	?	barren red beds or hiatus ?
			CHADIAN				
	TOURNAISIAN	TN 3	COURCEYAN	352	S.claviger–A.macra		C.decorus –S.claviger S.pretiosus var.pretiosus
					S.pretiosus–R.clavata	? Ainslie Fmn Strathlorne Fmn	V.vallatus: SPELAEOTRILETES SP.A; U.ABSTRUSUS –U.DISTINCTUS
		TN 2			S.baltiatus–R.polyptycha	?	?
					K.hibernicus–U.distinctus	Craignish Fmn	E.rotatus–H.explanatus
		TN 1b			V.verrucosus –R.incohatus	?	not studied
DEV.			STRUNIAN	360	S.lepidophytus –V.nitidus		

Table 6. Age ranges of Horton formations compared to miospore zones/subzones of Western Europe and mainland Nova Scotia (modified after Utting et al, in press).

Island are of Tn_1 or Late Devonian age, although no productive samples were obtained from the lower parts of the Craignish Formation.

From palynological evidence, and based on correlation with Western Europe there may be one miospore zone absent between sampled beds of the Craignish Formation (C1 facies assemblage) and those of the Strathlorne/Ainslie formations in northern Cape Breton. However undisturbed sections have an unknown (and unsampled) thickness of C3 facies assemblage between C1 and Strathlorne, which might account for this gap. There is no palynological data from the Craignish of western Cape Breton. The general consensus in the literature (Murray, 1960; Kelley, 1967; Williams et al., 1985), and in this study, is that the Craignish-Strathlorne contact is conformable. Although the actual contact is only exposed in a few localities it typically appears as a distinct change of colour and grain size with no evidence of angular discordance or erosion. In fact, at several outcrops in western Cape Breton, there appear to be relatively thick transitional sequences at that contact (eg: Southwest Mabou River, Baddeck River, Christopher McLeod Brook, North Branch Baddeck River, Munro Point). In northern Cape Breton the contact is commonly unexposed, or is a fault.

There appears to be a significant gap (of several miospore zones) between the Strathlorne and Ainslie Formations and the Macumber Formation, which has inspired some controversy. Kelley (1967), Geldsetzer (1977) and Kirkham (1978) suggest that in general this contact is conformable, with local exceptions around active basement blocks, and in field observation it is always sharp, but with no obvious evidence of erosion. Norman (1935) and Williams et al. (1985) suggest a hiatus between Horton and Windsor sedimentation. Utting et al. (in press) suggest that a hiatus may occur in some localities but not others. However, palynological data from the Macumber Formation of Cape Breton Island is virtually nonexistent.

All Horton samples studied were deposited in fluvial and/or lacustrine environments: a few marine acritarchs occur but were probably reworked from pre-Carboniferous rocks. All samples from Craignish and Ainslie Formations have few spores and are dominated by woody and coaly organic fragments (approximately 65-90% of sample, terrigenous source, commonly abraded by transport) with accessory exinous and amorphous material (approximately 10-35% of sample). The ratio of amorphous + exinous : woody + coaly material is 1:3 to 5. The Craignish and Ainslie Formations are attributed to deposition in fluvial environments; a conclusion which supports all sedimentological evidence as discussed in Chapter 4). All samples from the Strathlorne Formation (facies assemblage S1) contain significant amounts of exinous material (up to approximately 50%); spores are often abundant. Coagulated amorphous material (up to approximately 60%), includes biodegraded exinous/woody/fecal matter. At 5 different outcrops, significant amounts of Botrycoccus sp. (aquatic algae) occur and are interpreted to represent deposition in lacustrine environments. It is also apparent that samples taken from mudstones stratigraphically close to thick sandstones (interpreted as fluvial/shoreline facies assemblage S2) have different relative proportions of organic constituents (amorphous + exinous : woody + coaly ~ 1:5) compared to those samples far removed from shoreline influence (amorphous + exinous : woody + coaly ~ 2 to 1:1). In addition, there is generally more woody and coaly material in the western part of Western Cape Breton (amorphous + exinous : woody + coaly ~ 1:2) than in the eastern part of western Cape Breton (amorphous + exinous : woody + coaly ~ 1:1), suggesting an asymmetry of terrigenous clastic input. Samples from dark grey laminated mudstone with little shoreline influence commonly show pyrite degradation of miospores.

MISCELLANEOUS PALEONTOLOGICAL INFORMATION

Trace fossils are typically abundant in the Strathlorne Formation and, in some locations, are present in the Ainslie Formation. R. Pickerill and D. Fillion of University of New Brunswick and G. Narbonne of Queen's University kindly provided some comments on several of the collected samples. In the Ainslie Formation ichnofossils consist of numerous small horizontal tracks and trails (gastropods ?) in red siltstone and some isolated arthropod resting and scratch traces as casts on the bases of sandstones of the grey/green vf-f ss facies assemblage (A3). These include Rusophycus and Lockeia. No trace fossils were collected from the Craignish Formation. The Strathlorne Formation yielded many samples of well-preserved casts of small ichnofossils on the bases of siltstone to fine sandstone beds. The fauna appears to be generally of low diversity, but with large numbers of individuals of rather small size, characteristics common in lacustrine sequences due to the variable chemistry, oxygenation and sedimentation conditions of lakes (Picard and High, 1972; Hakes, 1985; R. Pickerill, 1989, pers. comm.). Most living freshwater arthropods (notably branchiopod crustaceans) are <2 cm long (R. Pickerill, 1989, pers. comm.).

The dark grey facies assemblage (S1), (sampled at Gallant River, McFarlanes brook, Baddeck River, North Branch Baddeck River, Murdock Paul's Brook), is typified by arthropod/crustacean or bivalve resting traces and simple horizontal (annelid?) feeding burrows. Abundant examples of Lockeia, "small stuffed burrows" and small Paleophycus/Planolites are characteristic. A few specimens of Rusophycus, Cochlichnus, ?Bifungites and Helminthopsis are also present. This fauna is similar to that observed in the lacustrine facies of the correlative Albert Formation of New Brunswick (Wiley, 1986).

The grey/green vf-f ss facies assemblage (S2), (sampled at Southwest Mabou River, Baddeck River, and MacRae Brook), is typified by arthropod resting and locomotion traces and simple horizontal burrows present in reddish grey siltstone and very fine sandstone. Abundant Rusophycus and Cruziana are characteristic. Large Paleophycus/Planolites, and several specimens of Monomorphichnus and Scoyenia are also present. This fauna is similar to that observed in the fluvial facies of the correlative Albert Formation of New Brunswick (Wiley, 1986).

Although the sample size is small and incomplete, there appear to be differences, in the ichnofauna and behaviours represented, between sediments with shoreline influence (S2) and those with no direct shoreline influence. As lacustrine ichnofaunas are less well known than their marine counterparts these observations may be of significance in future studies.

Fossil roots and plant fragments are abundant in a few localities although most are not well-preserved. These occur in the Fisset Brook, Craignish, Strathlorne and Ainslie Formations. At least some of these wood fragments can be identified as Lepidodendropsis a Devonian and Lower Carboniferous plant which is well known from the Horton Group throughout the Maritime Provinces (Bell, 1966). Those from the Fisset Brook Formation have been identified by Kasper et al. (1988) as the Upper Devonian progymnospermous plant Archaeopteris.

Several localities yielded rare to abundant tiny fish fragments from the S1 facies assemblage of the Strathlorne Formation. These are particularly abundant at Cap Rouge, Presqu'ile, Baddeck River and North Branch Baddeck River.

CHAPTER 4

INTERPRETATIONS OF FACIES ASSEMBLAGES AND DEPOSITIONAL SYSTEMS

The 10 facies assemblages described in Chapter 3 are each interpreted, in this Chapter, to represent a specific depositional setting prevalent in parts of the depositional sub-basins at particular periods of Horton deposition. Each facies assemblage includes several facies, which represent depositional environments within that setting. In this study, exposures provide only partial vertical sections with little lateral continuity, but the correlation of many widespread exposures indicates lateral variations on a large scale. The resulting interpretations are quite general, in keeping with the basin analysis viewpoint of the study, but serve to establish a depositional framework for the Horton Group. More detailed fieldwork would elucidate more subtle aspects of the depositional relations. As in Chapter 3, the following interpretations reflect the depositional position of the facies assemblage with respect to the original margins of the sub-basins. The interpretations of facies assemblages and paleocurrent data illustrate the vertical and lateral distribution of depositional settings and the limits of the original sub-basins of deposition. In effect, a 4-dimensional picture of several Horton depositional systems is produced.

THE CRAIGNISH FORMATION

<u>C1 Red/Orange Coarse ss-cgl.</u> The C1 facies assemblage, which intertongues with the C2 facies assemblage toward sub-basin centres, is characterized by 2 main facies: predominant reddish coarse sandstone to conglomerate in fining-upward units, and subordinate red siltstone to medium sandstone. The sediments are poorly to moderately sorted litharenite to sublitharenite, with ubiquitous hematite rims around grains. These coarse-grained, relatively immature sediments suggest proximity to a comparatively high relief source area, where mechanical weathering and rapid transport were important. Interbedded fine grained sediments have evidence of subaerial exposure, and the facies assemblage is interpreted to represent a braided fluvial setting.

Braided streams have multiple, broad, shallow, low-sinuosity channels with many braid bars, and generally carry sand- to gravel-sized bedload. Braiding is caused and controlled by a combination of several factors (Miall, 1977): a) geometry and slope of channels, b) strength and variability of flow discharge, c) amount and coarseness of bedload, and d) lack of stabilizing vegetation (in post-Silurian sediments this likely indicates an arid climate Schumm, 1968). The main processes involved are bar formation, growth and destruction during flood events, bedform migration in the channels, and falling water-stage deposition in channels, or flood deposition in overbank areas (Miall, 1977; Rust and Koster, 1984). The environment is locally rather unstable through time due to continuous active channel/bar migration and avulsion, and the resulting deposits are less well organized than in meandering systems (Collinson, 1985).

In the C1 facies assemblage the fining-upward units begin with a scoured base and pebble lag followed by conglomerate or pebbly sandstone with some internal horizontal stratification and abundant trough cross stratification (facies Gt, Gm, St, Sh of Miall, 1977), suggesting a dominance of in-channel scour-and-fill and dune migration under turbulent conditions. Discrete scour pockets filled with pebbles (facies Ss of Miall, 1977) indicate deep scour in the lee of bedforms or obstructions and emphasize the turbulent conditions of flow. Coarse beds may cut down into one another but are commonly separated by finer sediments. The coarse beds are, in places, arranged into bundles several tens of metres thick, separated by finer sediments, which may represent channel filling by multiple aggradational events and subsequent channel avulsion. The coarser lithofacies of C1 resembles the Donjek-type distal gravelly braided stream model (Miall, 1977, as distilled from Williams and Rust, 1969), characterized by relatively well-organized deposits with cyclic fining-upward channel units, avulsion packages, and overbank areas. Similarly, the abundant trough cross stratified conglomerate and sandstone with some horizontal stratification, common fining-upward units and subordinate finer sediments resemble Facies Assemblage GIII of Rust (1984), attributed to accretion in migrating gravelly channels during flood events in a relatively distal braidplain setting.

The subordinate finer grained lithofacies occurs in thin lensoid fining-upward units separating coarser beds, or in thicker units separating bundles of stacked coarse grained beds. It consists of bright red very fine to medium quartz sandstone with hematite around grains and calcite cement. Stratification includes horizontal, trough and ripple cross lamination (facies Sh, St, Sr, Fl of Miall, 1977) and common desiccation mudcracks and roots. These finer grained sediments with better textural and mineralogical maturity and smaller sedimentary structures are interpreted as in-channel falling water-stage and interchannel overbank deposits. Sandstone lenses with cross stratification are commonly deposited as wedges around gravel bar margins or as the fills of open scours on bar tops (Miall, 1977; Rust and Koster, 1984). The small overbank areas are more common in distal locations and record vertical accretion during floods and periods of exposure and non-deposition between floods or after avulsion. In the C1 facies assemblage some red siltstone units have thin, extensive beds of green, sandy, nodular limestone with root casts, wood fragments and desiccation cracks. Similar sediments were noted by Allen (1974), Steel (1974) and Collinson (1985) and are interpreted as paleosol (calcrete) horizons (Van Houten, 1982), developed during periods of lengthy interchannel stability. These are discussed more fully in the next section. Diagenetic reddening is generally considered to indicate exposure in an arid climate during deposition (Retallack, 1981; Rust and Koster, 1984). Several palynological samples from C1 mudstones yielded a flora consistent with deposition in a fluvial setting.

In summary, the C1 facies assemblage is interpreted as deposits of medial to distal gravelly braidplains in a warm arid climate. Coarse-grained deposition occurred in broad shallow channels during relatively frequent high energy flood events, separated by periods of inactivity when minor vegetation and rudimentary soil horizons were established on interchannel areas.

<u>C2 Brick red siltstone-fine ss</u> The C2 facies assemblage, which intertongues with, and overlies, the C1 facies assemblage toward sub-basin margins, is characterized by 2 main lithofacies: predominant red siltstone and sandy siltstone in thick units with minor red or grey fine sandstone. Desiccation mudcracks, root casts, green calcareous nodules and peloidal limestones are characteristic. Thick units with gypsum rosettes occur near the centre of the western Cape Breton sub-basin. These features suggest subaerial exposure of low lying areas far from significant input of coarse grained sediments, and subject to wet/dry cycles in some central areas. A mudflat setting, basinward of the coarse sediment wedge previously described, is the favoured explanation of these characteristics. The subordinate lithofacies occurs in scour-based beds of well sorted sandstone with trough and ripple cross stratification scattered throughout. These sandstones are similar to, but of more distal character than, to those of the previous facies assemblage and represent small fluvial channels extending from the edge of the adjacent C1 braidplain.

The predominance of fine grained suspension deposits with evidence of subaerial exposure suggests a mudflat/playa system (Glennie, 1972), perhaps with extensive loess deposits (Collinson, 1985), located basinward of the C1 braidplains. Relatively slow sedimentation rates dominated by fine sediment are implied. In this setting intense reddening, desiccation cracks, roots, extensive thick calcareous beds and nodules, and evaporative crystals are attributed to long periods of sediment stability and exposure in an arid climate. Mature calcrete develops just beneath the surface in tropical to subtropical semi-arid environments during relatively long periods of sediment surface stability at a rate of 1-10 cm/1000 years (Van Houten , 1982; Leeder, 1976). It forms by interaction between CO_2 in meteoric water and Ca^{++} in the soil from weathering products (Esteban and Klappa, 1983). Due to early cementation calcrete preservation is excellent and these beds are very conspicuous components of the C2 facies assemblage. Because thick calcrete horizons take thousands of years to develop they are most likely to form and be preserved in areas distal from active braided fluvial sedimentation (Leeder, 1976).

The subordinate sandstone lithofacies is interpreted as the deposits of shallow active channels which extended into the mudflat setting from the edge of the adjacent braidplain, perhaps as a result of especially extensive flood events and concurrent coarse sediment input. The thicker, coarser examples which occur in bundles might be the distal expression of fan/braidplain progradational phases. These channels may have also supplied runoff to the isolated low-lying central portions of the sub-basins to form shallow playa lakes. Playas occur in arid internal drainage sites, are fed by ephemeral streams or proximity to the water table, and commonly have saline waters due to evaporative concentration (Collinson, 1985). Drying allows desiccation cracks and evaporite crystals to form, especially where groundwater supply is dominant. A portion of the red siltstone in C2 could represent loess blown from the adjacent braidplain and alluvial fans, as suggested for the Triassic Keuper Marl by Wills (1970). More detailed sedimentological study might substantiate this possibility. Diagenetic red colours, characteristic of C2, result from conversion of yellow ferric to red ferrous minerals and are considered to be an indicator of exposure in an arid climate during deposition (Retallack, 1981; Rust and Koster, 1984).

In summary, the C2 facies assemblage is interpreted as the deposits of a distal mudflat/playa setting, beyond most coarse grained input, where deposition occurred by vertical accretion of fines from wind or flood events. Long periods of non-deposition, stability and exposure allowed thick calcrete soil horizons and minor vegetation to develop in a warm arid climate.

The C1/C2 Depositional System The red coarse grained C1 and red fine grained C2 facies assemblages intertongue and are interpreted as a single depositional system with a proximal to distal facies variation. The proximal sub-basin margin areas were dominated by bedload deposition in broad, shallow braided channels during flood stages with minor finer grained overbank flood and falling stage deposits (C1). The distal sub-basin centres were dominated by aqueous flood or aeolian vertical accretion of red fine sediments on exposed mudflats beyond the braidplain with some shallow isolated channels and playa lakes in localized depressions (C2). The linkage between these two depositional settings and their processes is suggested by their similar red colouration, textural and mineralogical immaturity and lateral intertonguing relationship. There is abundant evidence for subaerial exposure in a warm arid climate. In stratigraphic terms this depositional system is equivalent to the McLeod Member defined by Kelley (1967).

The coarse grain size and immaturity of C1 sediments imply proximity to elevated source areas where erosion was active. The braided nature and fining-upward units suggest high energy but flashy runoff. Fault-bounded margins are inferred, although the most proximal alluvial fan facies are not exposed (or preserved?). The thick bundles of beds may be due to a) autocyclic channel avulsion common in a distal braidplain setting (Miall, 1977), b) allocyclic climatic cycles related to individual flood runoff events, c) allocyclic alluvial fan progradation phases induced by tectonic events (Steel and Gloppen, 1980). Data in this study are not adequate to evaluate these explanations and it is possible that a combination of factors was present. The well developed calcretes, rooting, evaporite crystals and deficiency of coarse sediments in assemblage C2 imply low-lying distal depressions (sub-basin centres) with periodically stable sediment surfaces close to the local water table.

Paleocurrent data are limited for the C1/C2 depositional system but support the above interpretations. In western Cape Breton coarse sediments of assemblage C1 were derived from and deposited in relatively localized areas representing main input points and depositional lobes associated with the fault-bounded sub-basin margins (Figs. 31, 33). This sediment was deposited in thick wedges adjacent to several basement blocks such as the Mabou Highlands on the west, Bras d'Or Lakes and Big Bras d'Or/St. Ann's areas on the east, and the Cheticamp area on the north. The C2 fine sediments are thickest in a wide area between the marginal C1 facies belts with input from several sources approximately corresponding to those of C1 (Figs. 39, 41). These facies distributions, paleocurrents, and interpretations suggest deposition of the C1/C2 depositional system in a closed internal drainage sub-basin in western Cape Breton, with alluvial fan/braidplain wedges near inferred fault-bounded margins on all sides, and mudflat/playa deposits in the low-lying centre (Figs. 144, 145).

In northern Cape Breton the data are more limited but suggests localized C1 sediment input from basement blocks in the Pleasant Bay and south Aspy areas on the south, an unknown area offshore from Lowland Cove on the west, and possibly in the St. Paul's Island area on the east (Figs. 31, 33). Likewise data from C2 sediments (present only as tongues in C1) suggest the source area west of Lowland Cove was pre-eminent in supplying this depositional system. The western and northern Horton outcrop areas are close to each other in western Cape Breton Highlands between Cheticamp and Pleasant Bay but data from C1 facies assemblage imply a definite separation by a high standing basement block sediment source at that point, defining the two sub-basins as separate fault-bounded structural units (Fig. 145). This separation persisted through much of Horton time, as discussed later.

The C1/C2 depositional system represents a discrete depositional system present in both fault-bounded sub-basins during a particular phase of Horton deposition in the Cape Breton region. This system was characterized by high energy braidplain deposition in more proximal areas and low energy mudflat deposition in basin centre areas. In both sub-basins, but best documented in western Cape Breton there is a tendency for C1 sediments to dominate in the lower and more proximal parts and for C2 sediments to dominate in the upper and more distal parts of the depositional system. This created an overall fining-upward, increasingly distal-upward trend expressed over hundreds of metres of vertical section. This must relate to the long-term regional controls on basinal evolution, discussed in Chapter 5.

<u>C3 Grey/Green coarse ss-granulestone</u> The C3 facies assemblage, which generally does not intertongue with the C1/C2 depositional system, is characterized by 3 main lithofacies: grey sandstone and granulestone, greenish grey sandy conglomerate and subordinate green siltstone to fine sandstone. The predominant lithofacies is typically grey, coarse sandstone with pebbly laminations, which occurs in stacked fining-upward units several metres thick. The sediments are moderately to well sorted, quartz arenite to sublitharenite with abundant silica cement. These coarse-grained, moderately mature sediments suggest moderate

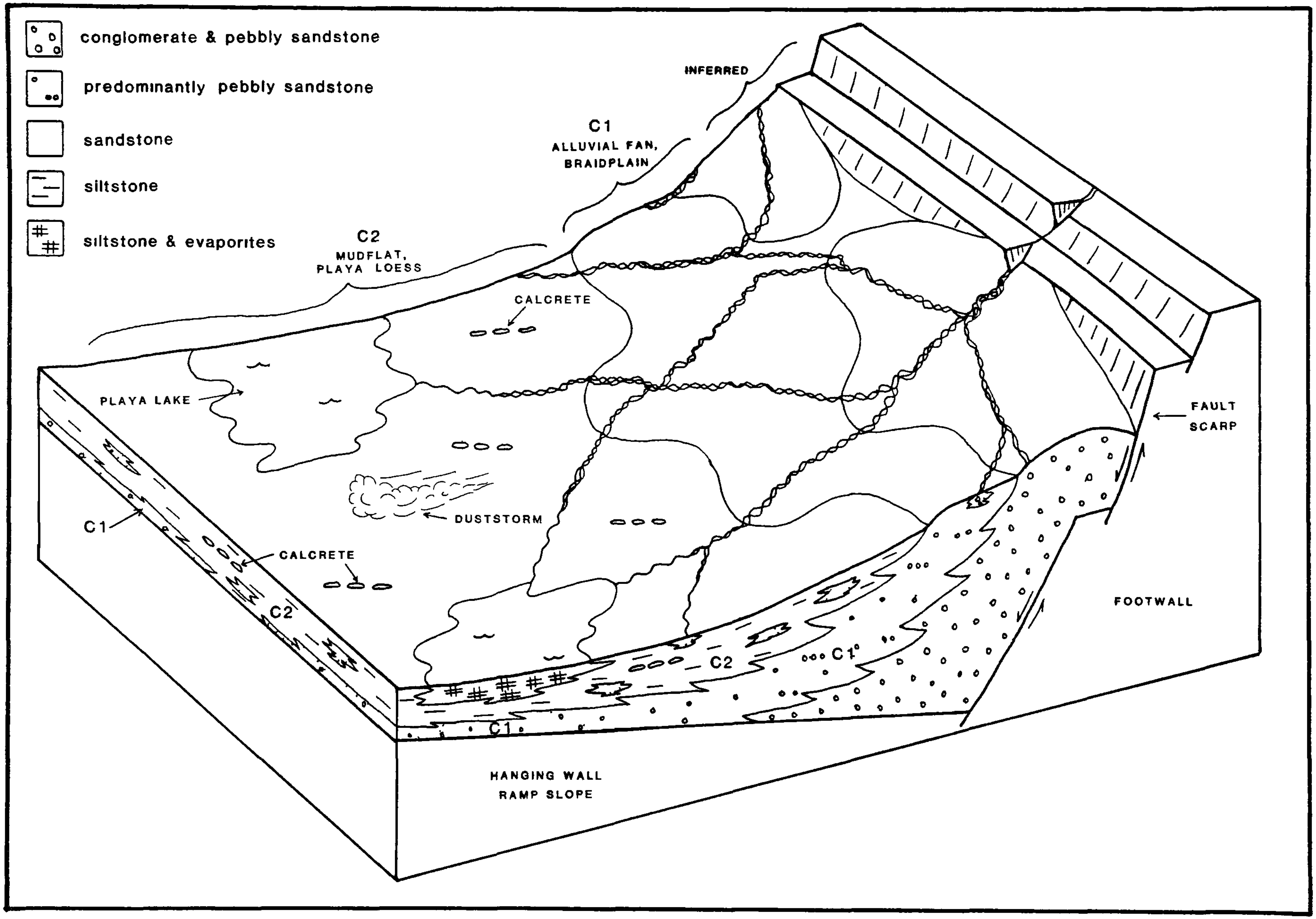

Figure 144. Block diagram illustrating depositional environments and facies distributions of C1/C2 depositional system.

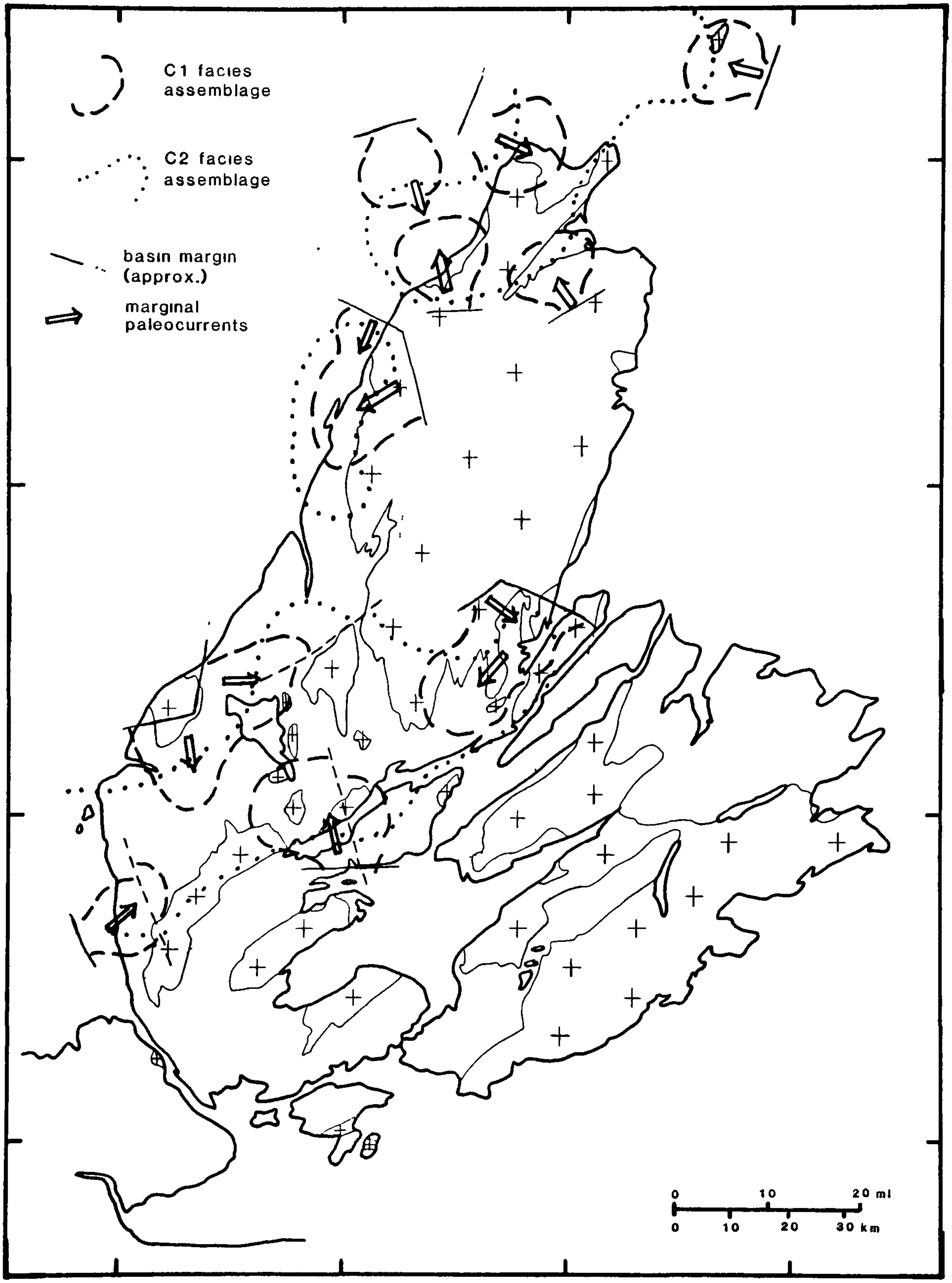

Figure 145. Map of sub-basin margins, distribution of facies assemblages and generalized sediment dispersal of C1/C2 depositional system.

proximity to an elevated source area where mechanical weathering and rapid transport were important. The deposits are interpreted in terms of a braided fluvial setting.

The fining-upward units begin with a sharp base (some with a minor pebble lag) overlain by pebbly coarse sandstone, either massive or with horizontal, trough or planar cross stratification (facies Gm, Gt, Gp, Sh, St, Sp of Miall, 1977). Successive beds have different grain sizes. In-channel scour-and-fill and dune or lateral bar migration were dominant processes under conditions of variable flow discharge and velocity. At some localities beds are conglomeratic and either massive or have vague horizontal stratification (facies Gm of Miall, 1977). Beds of both types may be stacked in bundles separated by units of finer sediment which probably represent filling of the channel system by multiple aggradation events, followed by avulsion to another area and subsequent overbank deposition. Conglomeratic versions typically coarsen upward and become more massive upward, and are reminiscent of progradational sequences on alluvial fans attributed to fault movements (Steel and Gloppen, 1980; Rust and Koster, 1984). These sequences may represent the toes of extensive alluvial fans. The coarser lithofacies of C3 resembles the Donjek-type distal gravelly braided stream model (Miall, 1977, as distilled from Williams and Rust, 1969) or the Facies GIII of Rust (1984) interpreted as a relatively distal braidplain setting.

The subordinate finer grained lithofacies occurs in thin lensoid units within and separating coarser beds, or as thick units separating bundles of coarser strata. It generally consists of green well sorted siltstone to fine sandstone with thin horizontal lamination or trough and ripple cross stratification (facies Sh, St, Sr, Fl of Miall, 1977). Thin lenses are attributed to in-channel and bar top falling-stage deposition (Miall, 1977; Rust and Koster, 1984), whereas thick units are interpreted as overbank flood deposits accumulated between periods of channel aggradation when the locus of coarse grained deposition had avulsed to another location. Evidence for subaerial exposure or aridity, so characteristic of the finer lithofacies of the C1/C2 depositional system is lacking, although in western Cape Breton the C3 facies assemblage conformably overlies the Fisset Brook Formation, regarded as having been deposited in a subaerial setting (see Chapter 2). Red colouration, desiccation cracks, roots and calcrete beds are rare or absent. Rare units of dark grey siltstone with bioturbation and one occurrence of a thick sandy limestone suggest that small short-lived ponds were present on the C3 braidplain. At several sections in the central part of the western Cape Breton sub-basin unusually thick units of green siltstone to fine sandstone occur, with minor coarser grained beds. These may represent longer shallow lacustrine phases in basin centre depressions, with periodic influxes of coarse fluvial flood sediment.

These facts may indicate that deposition of C3 occurred under climatic conditions distinctly more humid than those of the C1/C2 depositional system, although there is no evidence for the abundant vegetation that would be expected. Conversely, deposition may have occurred in a broad, shallow low-lying basin with very slow subsidence and a high water table.

The C3 Depositional System The uniform grey coarse grained deposits and subordinate green fine grained deposits of C3 are interpreted to represent a depositional system separate from, and somewhat different from, the C1/C2 depositional system. Although braided fluvial sedimentation in a relatively distal braidplain setting was dominant, the sediments remained unoxidized, with silica (rather than calcite) cement and there is no evidence of extensive exposure. Bundles of coarse sediments and conglomeratic coarsening-upward bundles represent more proximal facies. Intervening fine sediments show no evidence of abundant vegetation or calcrete soil horizons, which suggests that this braidplain was very low-lying, with ponds and long-lived lakes in central depressions. These factors may indicate cooler, more humid climatic conditions than the C1/C2 depositional system, or simply a gradually subsiding, non-faulted basin with very low relief and high water table (rain shadow effects of aridity are enhanced in fault-bounded basins, Jowett and Jarvis, 1984). However, although further study of this point is required, it is likely that C3 represents fundamentally different climatic and/or tectonic conditions from those responsible for depositional system C1/C2.

In western Cape Breton the C3 facies assemblage grades upward from the volcanics and sediments of the Fisset Brook Formation (which unconformably overlie basement) and is equivalent to the Graham River and Skye River Members of the Craignish Formation (Murray, 1960; Kelley, 1967). In fact, at one location C3 has intrusive sills in its basal part, presumably related to Fisset Brook volcanism, which is unknown in later Horton deposits. In this area the Fisset Brook and C3 together may define a single depositional system older than, and associated with different tectonic conditions than, the C1/C2 depositional system. In northern Cape Breton the C3 facies assemblage is uncommon and does not occur at the base of the Craignish between the Fisset Brook Formation and the C1/C2 depositional system. The presence in a few sections of moderate thicknesses of C3 apparently at the top of the Craignish Formation, is directly opposite to the facies distribution of western Cape Breton. However, poor exposure and structural complications obscure the significance of this arrangement, which may be an artifact of faulting.

Facies distribution, isopach and paleocurrent data are limited for the C3 facies assemblage and do not elucidate individual sediment source areas or depositional loci (Figs. 47, 49). Depositional thickness is greatest in basin central areas (although at most occurrences only a minimum thickness can be measured). There is no direct evidence of a structural saddle between Cheticamp and Pleasant Bay delineating 2 separate sub-basins (as in the C1/C2 depositional system). As a first approximation the depositional basin is considered to have been a broad linear feature with low relief margins beyond the geographic limits of this study (Figs. 146, 147). The tectonic setting of this basin may have been related to that in which the Fisset Brook volcanics were extruded, as a precursor to the narrower fault-bounded sub-basins of the C1/C2 depositional system. Although the paleocurrent data are very sparse, the majority of measurements indicate flow from southwest to northeast and may delineate the general basin paleoslope. The paucity of proximal sediments in C3 is probably because the basin margins were located in areas now beyond the erosional edges of Horton sediments, in the offshore area to the west, and where Precambrian basement outcrops to the east and north. In addition, these marginal areas would have been faulted and erosionally cannibalized as sediment sources during later Hortonian phases of fault-bounded subsidence.

THE STRATHLORNE FORMATION

Characteristics of the Strathlorne Formation suggest deposition in a lacustrine setting, as discussed below. No marine fossils are known, whereas the ichnofauna, palynoflora and miscellaneous other fossils all indicate nonmarine deposition. The principles of lacustrine sedimentation are more complex and less well known than many other depositional environments and are briefly reviewed here.

Lakes are nonmarine, relatively small, low energy, standing bodies of water, fresh to saline, enclosed in land-locked depressions. Most modern lakes occur in small intracratonic glacially-scoured or epeirogenic basins (eg: Alpine Lakes of Switzerland, Great Lakes) (Galloway and Hobday, 1983) whereas many ancient examples may represent

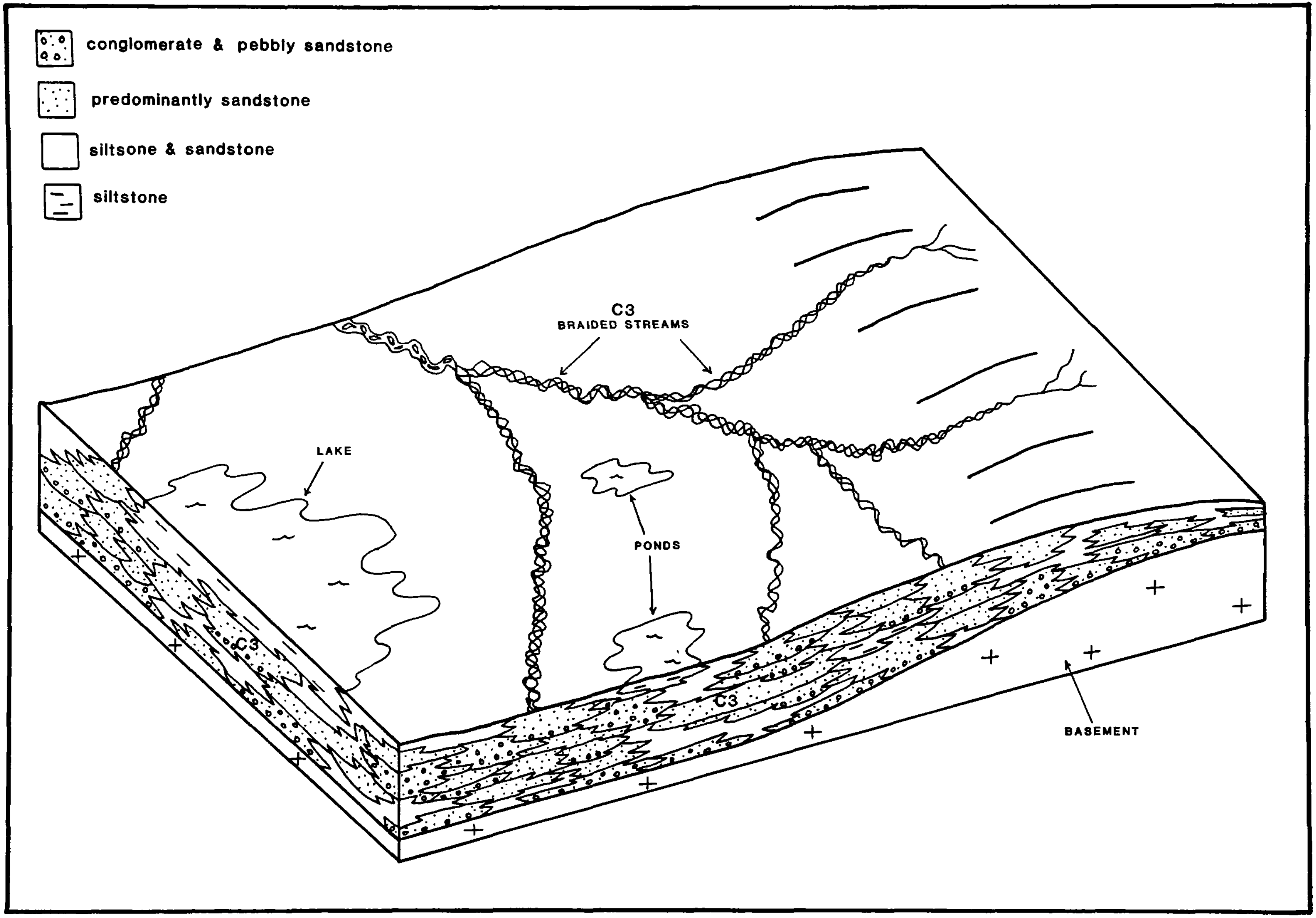

Figure 146. Block diagram illustrating depositional environments and facies distributions of C3 depositional system.

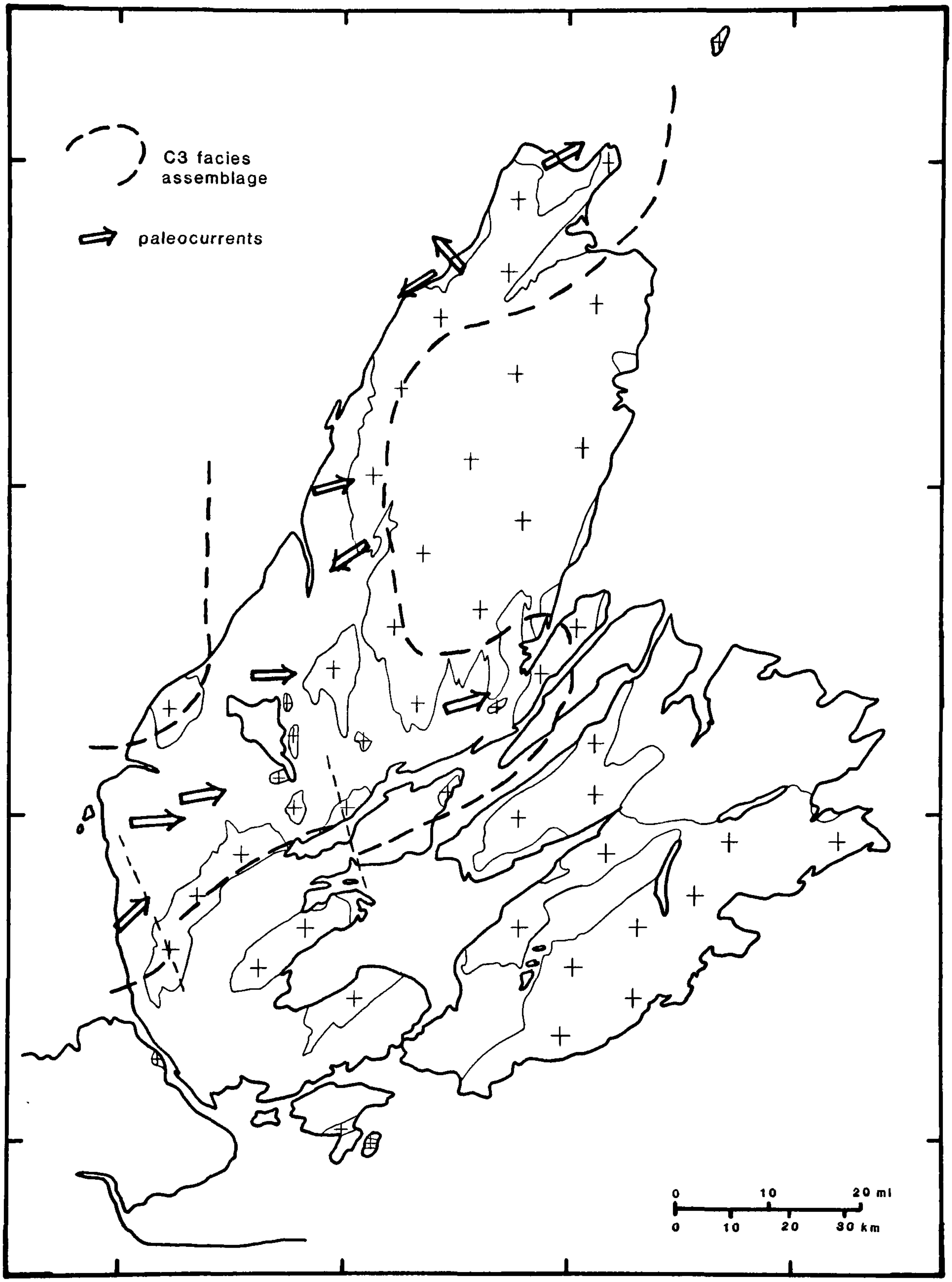

Figure 147. Map of basin margins, and generalized sediment dispersal of C3 depositional system.

large long-lived water bodies in tectonic depressions where thick sequences are preservable (eg: Green River Formation, Lockatong Formation; Davis, 1983). Lakes are very important as base level markers and as sensitive climatic indicators (Picard and High, 1981). In a closed lacustrine system, sediment supply, water chemistry, and water level all vary independently and are very sensitive to subtle fluctuations of climatic and tectonic factors (Reading, 1986). This results in complex sedimentation patterns which change through time, according to the controlling factors, producing complex multiple basin-fill sequences (Picard and High, 1981). The regional tectonic setting controls basin size, shape, depth, gradient and longevity, whereas the regional climatic regime controls precipitation, evaporation, runoff, vegetation and weathering (Selley, 1985). Warm equable climates discourage circulation and promote density/temperature/oxygen/salinity stratification of the water column, and consequent anoxia near the bottom. Climatic aridity enhances this effect by increasing surface evaporation. A closed basin traps all sediment input, allowing thick accumulations, and low energy conditions in most lakes allow good preservation and dominance of fine-grained laminated deposits.

The moderate size, small fetch and low energy characteristics of lakes create a shallow wave-mixing base. Infrequent storms may completely alter sediment patterns by redistribution of nearshore deposits and circulation patterns, but only for a relatively short time (Picard and High, 1972). Circulation below wave base is generally poor (especially in tropical climates) producing water stratification into an upper warm oxygenated epilimnion (subject to refreshment and evaporation, high organic productivity) and a lower cold anoxic hypolimnion (stagnant, good preservation of organics) (Haleanson and Jansen, 1983; Hutchinson, 1957) (Fig. 148). These are separated by a distinct thermo-/oxy-/halo-cline surface (Dean and Fouch, 1985). Sediment-laden inflow may form a suspended-load plume as an overflow at the surface or an interflow at the themocline, whereas bedload density underflows may travel downslope along the sediment surface (Sturm and Matter, 1978) (Fig. 149). The low energy regime generally results in narrow nearshore/shoreline zones of coarser clastic sediment where fairweather waves are effective. Poorly developed beaches and abrupt facies transitions are typical (Picard and High, 1972). At main input points the nearshore zones prograde rapidly into shallow low energy water as deltas (Van Houten, 1964). Finer sediment is deposited offshore below wave base and may include the greatest volume of sediment (Reineck and Singh, 1973). Density current deposits are

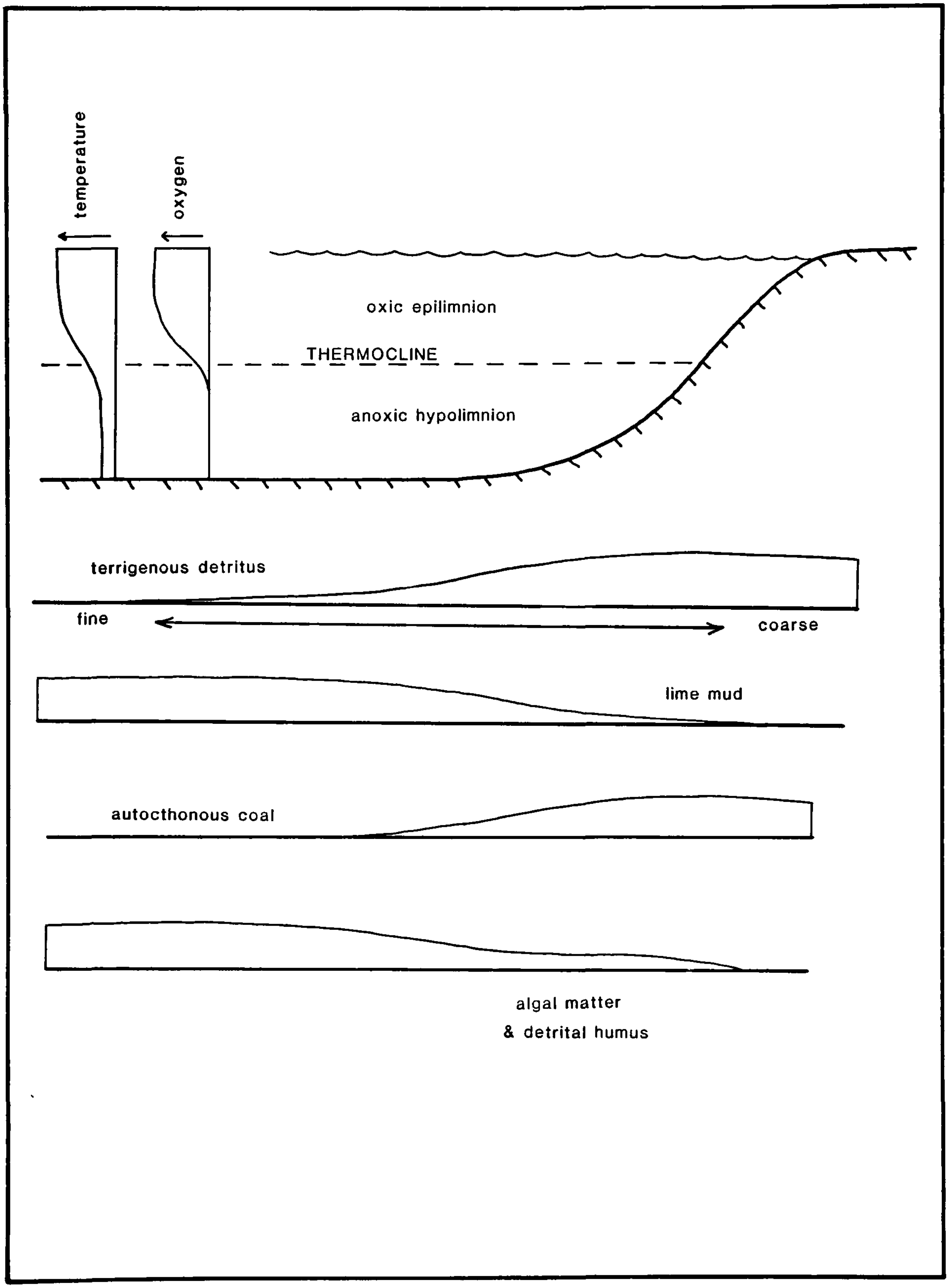

Figure 148. Circulation zones in modern lakes and trends in sediment components (from Selley, 1985).

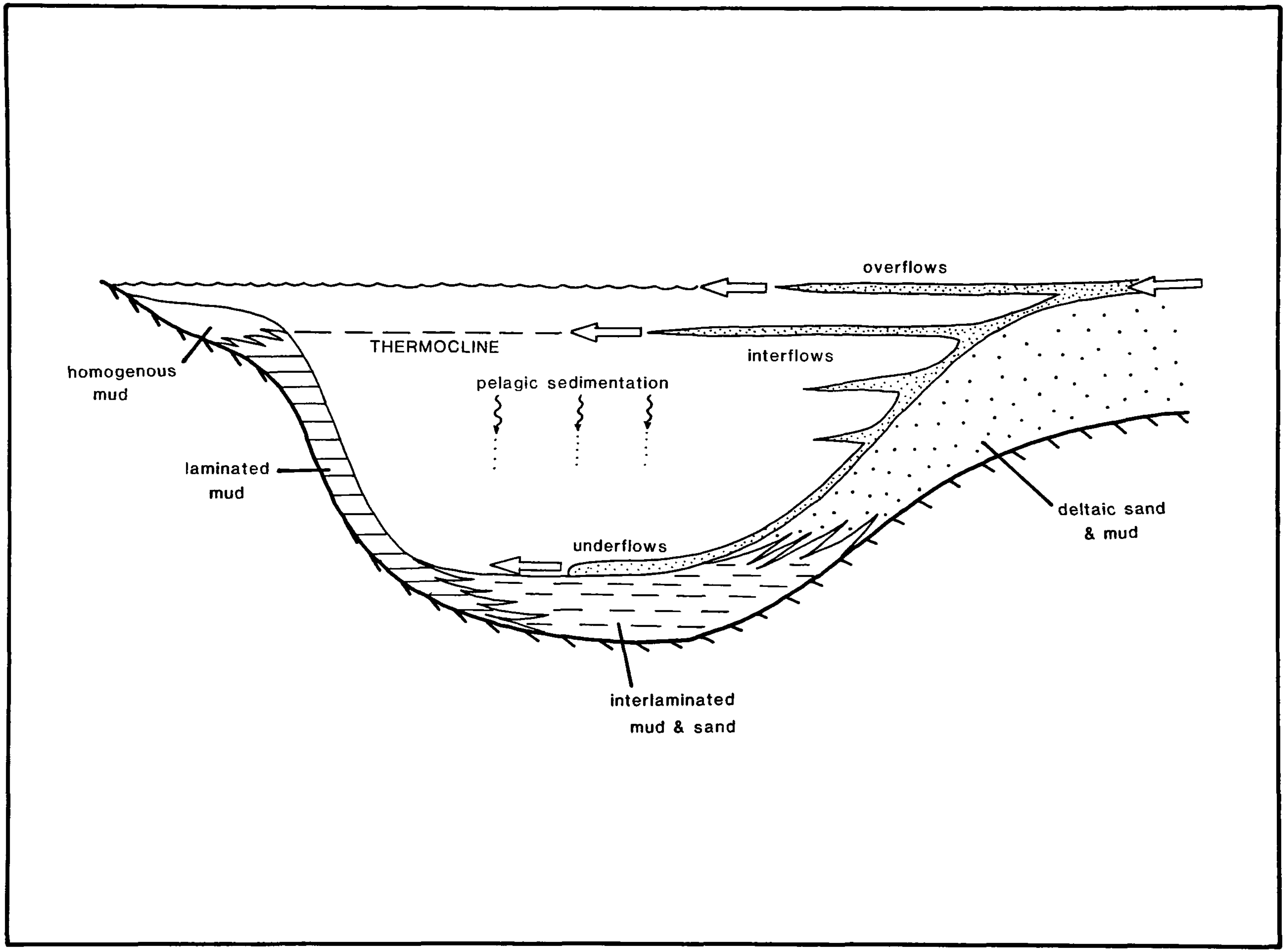

Figure 149. Relationship between water column stratification, clastic processes and clastic deposits in modern lakes (from Sturm and Matter, 1978).

common in the offshore areas where ripples (otherwise uncommon below wave base) have crestlines parallel to shoreline (Davis, 1983).

S1 Dark grey mudstone The S1 facies assemblage, which intertongues with the S2 and S3 facies assemblages toward sub-basin margins, is characterized by three thinly interbedded lithofacies: grey calcareous claystone to siltstone, grey calcareous very fine sandstone, and grey finely crystalline limestone. These lithofacies occur in specific arrangements, bioturbated siltstone being the predominant lithology. The S1 assemblage is interpreted as relatively low energy open lacustrine deposits.

Stacked thickening- and coarsening-upward sequences a few tens of metres thick are characteristic throughout the sub-basins, ranging from claystone at the base to very fine sandstone or limestone at the top. These are thicker and sandier in the Mabou and Meat Cove areas, thinner and more argillaceous in the Baddeck area. There is a concurrent upward decrease in organic matter and increase in bioturbation, sorting and current/wave indicators. Well-preserved, delicate horizontal lamination is typical of basal parts of sequences, indicating little current or wave energy. The upward increase of well sorted, graded sandstone beds is accompanied by an increase in tool marks, ripple cross lamination and straight-crested symmetrical and asymmetrical ripple forms, indicating increased current and wave energy. Sandstone beds in the upper parts of these sequences commonly display interference ripples, planed-off symmetrical ripples, desiccation cracks and raindrop impressions, all indicative of subaerial exposure. Many sandstones have contorted lamination, indicating major disturbance events, either sudden, short-lived tectonic movements or significant storms. Many sequences pass upward into thick fine sandstone bodies of S2 facies assemblage, interpreted as nearshore/shoreline deposits in the next section.

Coarsening-upward sequences in assemblage S1 are interpreted as shallowing-upward sequences which represent progradation, over a relatively broad front, of higher energy nearshore sediments over open lacustrine sediments deposited beyond the zone of shoreline influence. They therefore delineate small fill sequences of relatively shallow, large lakes, localized around clastic input points. In some cases the filling sequence was complete and subaerial exposure was achieved. Similar sequences were recognized and given the same interpretation by Picard and High (1972, 1981) and by Van Houten (1964, 1965). Grey fine-grained sediments typical of basal parts of the sequences probably do not indicate great depth, since the fill sequences are never more than a few tens of metres thick (and commonly less than 10 m). Rather, they indicate deposition beyond the range of sandy clastic input from most shoreline-related processes (in a low energy lake, this need not be far offshore, Picard and High, 1972). The dark grey, organic-rich, thinly laminated, unburrowed claystone characteristic in some areas suggests that anoxic, stagnant conditions were developed below a thermocline in at least some parts of the sub-basins. Palynological samples from these rocks typically show pyrite degradation of miospores, indicative of anoxic conditions. As progradation of a shoreline segment proceeded the shallow offshore area began to fill, manifested as an upward increase in the thickness and occurrence of extensive thin sandstone beds rapidly deposited from sediment-laden, higher energy density currents. These sandstones are separated by grey burrowed siltstone representing the slow deposition of low energy background sediments. Between periods of sandy input there was ample time for bioturbation. The pulses of more energetic deposition are interpreted as density underflows which deposited turbidite-like beds, as described from Alpine Lakes by Lambert et al. (1976) and Sturm and Matter (1978), and from Lake Mead by Grover and Howard (1938). These underflows were probably related to flood events of dense sediment-laden water (Miall, 1984), but some could also be related to storms (suggested by the presence of contorted lamination in S1 and S2, of hummocky cross stratification in S2) or to seismic events (suggested by contorted lamination in S1 and S2).

The coarsening-upward sequences must be related to progradation from major clastic input points around the lake margins, typically developed as deltas. Gravity-driven density currents are common on delta fronts (Galloway, 1975). Many of the S1 sequences did not achieve fill stage and exposure before conditions reverted to open lacustrine deposition. However, many sequences are capped by a) thin beds with indications of subaerial exposure attributed to interdeltaic deposition, b) thick S2 sandstone bodies interpreted as nearshore/shoreline deposits at main clastic input points (see below), or c) various limestone lithologies (discussed below). Proximity to main clastic input points affected the distribution of biota. Numerous palynological samples all yielded a flora indicative of a lacustrine setting but those from zones close to interpreted shoreline influence contain abundant transported woody and coaly matter, whereas those from the bases of sequences and far from interpreted shorelines contain abundant exinous and amorphous organic material. Ichnofossils record a low-diversity, dwarf fauna, typical of lacustrine sequences, which is dominated by crustacean traces in nearshore sediments (less stable substrates, variable temperature and salinity, fewer deposit feeders) and by bivalve and annelid traces in offshore sediments (more stable and muddy substrates, constant temperature and salinity, abundant deposit feeders). Fish fragments are only abundant in offshore sediments.

Limestone caprocks, although minor in volume, are important in some areas for the interpretation of the facies assemblage. Commonly, the uppermost bed of a sequence is very calcareous siltstone (40-50% poikilotopic calcite cement, quartz grains no longer in normal packing contact) which can be mistaken in the field for fine grained limestone. This texture is attributed to early cementation of a loosely packed, water-rich silt in carbonate-saturated water. $CaCo_3$ is the most abundant solute in modern fresh water (Davis, 1983; Dean and Fouch, 1985) and carbonate precipitation is very common in lakes, especially if the pH exceeds 8, as in stratified lakes and lakes with abundant algal growth (Picard and High, 1972; Reading, 1986). Other sequences are capped by thin beds with alternating micrite/argillite laminae with ostracod fragments, burrows and desiccation cracks. Both these sediment types are interpreted as low energy nearshore deposits, geographically removed from main clastic input points, where discrete pulses of fine clastic input alternated with periods of chemical sedimentation. A lateral delta margin is visualized, with minor longshore or flood currents supplying fine detritus at infrequent intervals. Small oolite shoals might also have occurred in this environment, suggested by the presence of transported oolite fragments in associated facies. Still other sequences are capped by thin cyano-bacterial stromatolite beds. These are interpreted as very shallow, low energy nearshore deposits in interdeltaic areas beyond the influence of clastic input, as interpreted in the Green River Formation (Ryder et al., 1976) and in modern lakes (Dean and Fouch, 1985).

S2 Grey/green very fine-fine ss The S2 facies assemblage, which intertongues with the S1 facies assemblage toward sub-basin centres, is characterized by grey or greenish grey very fine to fine sandstone, with minor green siltstone and oolitic limestone interbeds. The sandstone is well sorted quartz arenite, present in well stratified units several metres thick which cap S1 coarsening-upward sequences or form transitional units between the Strathlorne Formation and Craignish or Ainslie Formations. The mineralogical and textural maturity, presence of desiccation cracks and roots in some sections, and stratigraphic relations with other facies assemblages indicate that these units are lacustrine shoreline sediments found near the margins of sub-basins.

Most occurrences of S2 have a coarsening-upward or no obvious trend and include abundant horizontal and low angle stratification, trough cross stratification, hummocky cross stratification, and both symmetrical and climbing ripples. These characteristics indicate moderate to high energy processes of both wave and current activity. The S2 assemblage is interpreted as the deposits of prograding nearshore/shoreline sand bodies in which abundant horizontal and low angle stratification are characteristic (Reinson, 1984). The presence of these units requires major clastic input points such as deltas to supply large volumes of sand. A delta is a 3-dimensional mass of sediment deposited where a river enters standing water (Galloway, 1975) and the sedimentology of deltas first became well-known due to the early studies of lacustrine deltas by Gilbert (1890) and Barrell (1912). Most deltas which build into shallow, low energy water bodies are river-dominated (Galloway, 1975; Coleman and Wright, 1975) and experience rapid outward progradation. This tends to create sandstone bodies perpendicular to the shoreline, with deposition concentrated in distributary mouth bars, but these may merge laterally into sheet sandstones parallel to shore (Miall, 1984). Even with relatively low energy conditions, large lakes probably have enough fetch for wave and longshore current redistribution of sand into shoreface deposits. It appears that in S2 deposits wave and current energy was adequate to modify the delta front areas into shoreline-parallel sandstone bodies much of the time. Conversely, rivers which fed the Strathlorne lake shoreline may have been braided, and produced laterally continuous sheet-like bodies across the width of their braidplain. In this case the S2 shoreline sediments may represent "braid deltas" (McPherson et al., 1987) or "extended fan deltas" (Wescott and Ethridge, 1980). However, both terms were developed for deltaic sediments of coarse sandstone to conglomerate size, and may not be appropriate here.

The many occurrences of preserved, rather small-scale, hummocky cross stratification suggest shoreface deposition above storm wave base (which is very shallow in most lakes, Picard and High, 1972) where squalls could alter sediment distribution. Duke (1984, 1985) suggested that hummocky cross stratification could be preserved in very shallow lacustrine environments, and in fact Greenwood and Sherman (1986) have described it from the surf zone of Lake Huron (although its preservation potential may be very low). Other examples of lacustrine hummocky cross stratification are mentioned by Duke (1985) and Reading

(1986) (after Van Dijk et al., 1978). It is also present in the correlative Horton Bluff Formation of the Minas Basin area (T. Martel, 1987, pers. comm., and personal observation). In the Horton Group of Cape Breton Island contorted lamination is a conspicuous accessory structure and is attributed to sudden shock of subaqueous soft sediments either by storm wave oscillatory loading or by tectonic-seismic disturbance. Several other sandstone bodies at the tops of S1 coarsening-upward sequences have sharp scoured bases with pebble and wood fragment lags, abundant trough cross stratification and fining-upward trends. These are interpreted as the deposits of distributary channels which entered the lake, scoured into lacustrine sediments, and prograded rapidly beyond the general shoreline. These are characteristic of river domination where streams debouch into low energy standing bodies of water (Galloway, 1975; Miall, 1984).

Rippled oolitic sandy limestone, intimately interbedded with thicker fine sandstone, occurs as an accessory lithofacies in a few outcrops. Alternating quartz-rich/oöid-poor and quartz-poor/oöid- and peloid-rich laminae indicate alternating periods of siliciclastic input and chemical sedimentation in a setting where vigourous currents were present. These may represent minor interdistributary areas protected from coarse input by levées, or short time periods when channel avulsion removed clastic input to another area of the delta.

The S2 facies assemblage commonly has multiple recurrences in any vertical section of the Strathlorne Formation, suggesting that multiple shoreline progradation and lake-fill phases occurred. Deltaic cyclic sequences have been described many times since recognized by Scruton (1960) and are generally attributed to autocyclic sedimentary processes (see Miall, 1984). In the Strathlorne Formation, each cycle includes a) a prolonged phase of active progradation, manifested by a thick shallowing-upward sequence (S1) capped by one of various forms of nearshore/shoreline sediments (S1 or S2) and, b) a rapid abandonment phase, manifested by the abrupt termination of that coarsening-upward sequence as sediment input shifted to a different locale, and initiation of another sequence beginning with offshore sediments. In the Strathlorne this shift of depositional locus and shoreline position may have been related to autocyclic channel or lobe switching (common in deltas, Miall, 1984), allocyclic tectonic increase of subsidence, allocyclic tectonic backstepping of source area, or allocyclic climatic change which altered sediment supply. These recurrent controlling factors are discussed more fully below.

S3 Grey/green medium ss-boulder cgl The S3 facies assemblage, which is uncommon in the studied outcrops but which intertongues with S1 and S2 facies assemblages near several sub-basin margins, is characterized by two lithofacies; grey pebbly medium to coarse sandstone and grey sand-supported conglomerate, with minor greenish siltstone to fine sandstone interbeds. Units up to several tens of metres thick composed of stacked, scour-based, fining-upward beds, cut into S1 or S2 sediments throughout the study area, but the facies assemblage is only dominant in the Aspy area. The sediments are poorly to moderately sorted and have poor mineral and textural maturity. The coarse grain size, poor to moderate maturity, scoured bases and occurrence in isolated bundles suggest the abrupt influx of very coarse sediments into the open lacustrine and shoreline areas for short periods followed by resumption of the lower energy sedimentation. Minor interbedded fine sediments with roots suggest an association with a subaerially exposed fault margin environment.

The predominant lithofacies is micaceous pebbly medium to coarse sandstone which occurs in sharp, scour-based beds with rip-ups and pebble lags. The sandstones are thick, have fining-upward trends and occur in S1 or S2 units in widely scattered outcrops. They are interpreted as braided stream deposits. Subordinate scour-based beds of poorly sorted, micaceous, massive, sand-supported pebble to boulder conglomerate are common in the Aspy area, generally cutting into S1 sediments. They are interpeted as proximal gravelly braided or debris flow deposits (facies Gm or Gms of Miall, 1977). The characteristics of these coarse grained sediments suggest exceptionally erosive, high energy braided channels which carried large amounts of coarse sediment into distal settings normally dominated by shoreline and open lacustrine deposits. This facies assemblage is interpreted as the toes of fan-deltas, alluvial fans which prograde over a narrow coastal plain to debouch directly into a standing body of water (Ethridge and Wescott, 1984; Nemec and Steel, 1984)

Fan-deltas are common in arid, high-relief (fault-controlled) settings where coarse bedload, flashy discharge and lack of vegetation are prominent factors (Ethridge and Wescott, 1984). Narrow coastal plains allow steep fan geometry and intimate interbedding of low and high energy sediments at the distal edge. Near the controlling fault scarp these tongues of fine sediment pinch out and the bundles of coarse sediment merge into a thick wedge adjacent to the fault scarp (Ethridge and Wescott, 1984; Nemec and Steel, 1984). The latter situation is present on Middle Aspy River in northern Cape Breton where the entire exposed Strathlorne is represented by deposits of the S3 facies assemblage. In a low energy lacustrine setting the fan-delta front is generally not reworked because the progradational impingement of the fan into the lake is so rapid and far-reaching (Kleinspehn, 1984) and wave energy is low. The intertonguing of S3 with S1 and S2 sediments, with abrupt boundaries, must represent relatively isolated and short-lived progradational phases in the fan-delta evolution. These phases were most likely related to fault movements at the sub-basin margins.

S4 Red siltstone-fine ss The S4 facies assemblage, which is uncommon in the studied outcrops but is associated with S2 and S3 sediments near several sub-basin margins, is characterized by two lithofacies: brick red to brown siltstone and red silty fine sandstone. The predominant lithofacies is red massive sandy siltstone with abundant desiccation cracks, roots, wood fragments, and straight-crested symmetrical ripples. Green calcareous nodules and green limestones with sharp tops and gradational bases occur in units a few to a few tens of metres thick. As discussed for the C2 facies assemblage, these characteristics suggest subaerial exposure of low-lying areas removed from significant coarse-grained sedimentation. The calcareous nodules and limestone beds with roots are interpreted as pedogenic calcretes developed in a warm arid climate during relatively long periods of sediment surface stability and exposure (Leeder, 1976; Van Houten, 1982). Red quartzose very fine to fine sandstone in beds with sharp scoured bases, horizontal and ripple cross lamination are common and are interpreted as the deposits of small fluvial channels which cut through the area of fine-grained deposition from marginal areas.

The general aspect of assemblage S4 indicates deposition in an exposed mudflat setting associated with clastic shoreline facies in an interdeltaic area. Sedimentation was predominantly as suspension deposition from fluvial flood events or perhaps as aeolian loess deposits (Wills, 1970). The paucity of S4 may reflect a lack of wide extensive coastal plains in the Strathlorne fault-bounded sub-basins.

The S1/S2/S3/S4 Depositional System The four facies assemblages of the Strathlorne Formation are interpreted as the component parts of a single lacustrine depositional complex. All four intertongue, depending on their position relative to the original sub-basin margins, but the bulk of the exposed Strathlorne belongs to either the S1 or S2 assemblages. The proximal sub-basin margins were dominated by fan-deltas and proximal braidplains (S3), mudflats in interdeltaic areas (S4). The shoreline area was composed of sandy delta and shoreface zones at points of major clastic input, and low energy carbonate flats in adjacent interdeltaic areas with less siliciclastic input (S2). The open shallow lacustrine area, which occupied much of the depositional sub-basins for most of Strathlorne time, was dominated by slow, low energy deposition (S1). However it was subject to periods of rapid, high energy clastic input from shoreline areas, especially in front of prograding deltas at major sediment input points. A biota of limited diversity was present.

The S3 proximal braidplain/fan-delta deposits are uncommon but intertongue with S1 and S2. They represent sudden influx of unusually vigourous depositional processes into relatively quiet nearshore and open lacustrine settings. Their presence implies that, in Strathlorne time, the depositional sub-basins had narrow coastal plains adjacent to high relief fault scarps. Fault subsidence was rapid and continuous enough that coarse material only rarely built out into the lakes and did not form continuous bajada complexes along the scarps. The fans were likely steep and relatively small so the abrupt fault subsidence phases allowed open lacustrine environments to transgress close to the fault scarps. In fact, at 2 locations in northern Cape Breton, S3 sediments are directly overlain by thin stromatolitic limestone beds and then S1 dark grey siltstone. The S3 facies assemblage is well developed only near the areas of sub-basin margins already indicated by the C1/C2 depositional system, and it is suggested that during Strathlorne times fault margins followed the same trends but were displaced toward the sub-basin centres. In western Cape Breton S3 sediments are localized in the Bras d'Or Lakes/St. Ann's Bay areas, implying the eastern margin of this sub-basin was fault controlled. The western side around the Mabou area was less influenced by fault subsidence. In northern Cape Breton S3 sediments are concentrated in the Aspy area, interpreted as a sub-basin margin with prominent fault subsidence but the opposite (western) margin is not exposed.

Between the S3 fan deltas and S1/S2 sandy deltas subaerial mudflats were present (S4). Red diagenetic colours, desiccation cracks, roots and well-developed calcrete zones suggest exposure in a warm arid climate. These sediments are the only red deposits in the Strathlorne Formation and their scarcity probably indicates narrow coastal plains in sub-basins with active fault subsidence and dominantly open lacustrine conditions. In the Mabou area of western Cape Breton S4 is present as thin but extensive tongues implying wide coastal plains on the western margin of the sub-basin. Thicker, but less extensive, units in the Baddeck/St. Ann's area imply narrow but rapidly subsiding coastal plains on the eastern margin. In addition, units of these sediments are conspicuous in the uppermost Strathlorne near both sub-basin margins where there is a transition into overlying red fluvial deposits of the Ainslie Formation. Deposits of the S4 facies assemblage are rarely exposed in the studied outcrops of northern Cape Breton and no conclusions are drawn.

The S2 lacustrine shoreline deposits are present throughout the Strathlorne Formation commonly capping S1 coarsening-upward sequences or as transitional units between Strathlorne and Craignish or Ainslie deposits. They represent sandy shoreline bodies a few metres thick, deposited during recurrent phases of shoreline/delta progradation through Strathlorne time as a result of the changing balance between lake level, runoff, sediment supply and tectonic subsidence. The sediments were mineralogically and texturally mature and most probably underwent some wave modification and longshore transport forming bodies parallel to the shoreline. The presence of preserved hummocky cross stratification indicates high energy storm activity (common in tropical areas) which spread sand alongshore from distributary mouths and offshore as density underflows. Some fluvial channel deposits entered directly into the open lacustrine setting beyond the shoreline environment, indicating river-dominant processes were also common. In western Cape Breton the S2 facies assemblage is best developed in the lower and middle Strathlorne Formation of the western side of the sub-basin (Mabou-Lake Ainslie-Margaree areas). This suggests wide coastal plains with abundant sandy sediment input and prominent prograding deltas, conclusions that are also supported by data from S3 and S4 facies assemblages. On the eastern side of this sub-basin (Baddeck area) S2 is less important and only prominent in the upper Strathlorne. Narrow coastal plains with little sandy sediment input and few deltas are indicated for this side (again, supported by data from S3 and S4). It appears that in the western Cape Breton sub-basin there was a) a fundamental asymmetry of dominant sediment input and subaerial and shoreline facies distribution, and b) a distinct reversal of this asymmetry in later Strathlorne time. In northern Cape Breton, where only a portion of the sub-basin is exposed on land, the S2 facies assemblage is an important component of the lower and middle Strathlorne in the Aspy-Cape St. Lawrence-St. Paul's Island areas, located on the east side of the sub-basin (opposite to the distribution in western Cape Breton). The controls on these relations are more fully discussed below and in the next chapter.

The S1 low energy lacustrine deposits are present throughout the Strathlorne Formation, intertonguing with S2 and S3 shoreline deposits, and are dominant in areas away from original sub-basin margins. They are characterized by stacked shallowing-upward sequences, from offshore open lacustrine to nearshore or shoreline environments, and many are capped by occurrences of the S2 facies assemblage. In upward succession, slowly-deposited, bioturbated fine grained sediments are interbedded with progressively sandier sediments, deposited rapidly from underflows. These sequences record progressive shoreline progradation over offshore sediments and likely correlate with similar phases discussed for S2. In areas of little clastic input (interdeltaic) the sequences are capped by nearshore carbonate facies associated with relatively inactive muddy shorelines. These progradational phases were recurrent through Strathlorne time, each followed by an abrupt termination and return to open lacustrine fine-grained deposition.

In western Cape Breton the S1 facies assemblage, like the other facies assemblages, displays a fundamental asymmetry. On the western side of that sub-basin (Mabou-Lake Ainslie area) there are relatively few but very thick shallowing-upward sequences and many thick S2 progradational shoreline bodies (Fig. 150, left side). There is little dark grey, organic-rich mudstone at the base of sequences, many thick sandstone beds, few limestone beds and much woody/coaly organic matter. This area is interpreted as one of shallow lacustrine conditions with a wider coastal plain, abundant clastic input and thick sandy progradational cycles (especially in the lower and middle Strathlorne) indicating large deltaic complexes. There was a decreased volume of sandy input in later Strathlorne time. In contrast, on the eastern side of that sub-basin (Bras D'Or Lakes-Baddeck area) there are many thinner shallowing-upward sequences which begin with dark grey organic-rich mudstone and less woody/coaly matter, have relatively few thin sandstone beds, and are commonly capped by limestones (Fig. 150, right side). Thick shoreline sandstones of S2 are uncommon, except in latest Strathlorne time. The succession in this area is interpreted as one of somewhat deeper lacustrine conditions with a narrow coastal plain, less clastic input and thinner progradational cycles with little sandstone (especially in the lower and middle Strathlorne) indicating few large deltaic complexes. Chemical sedimentation was important even in the shoreline zone, emphasizing the paucity of sandy clastic input. Many cycles probably did not reach exposure before an abrupt shift to open lacustrine conditions recurred. There was an increased volume of sandy shoreline input in later Strathlorne time.

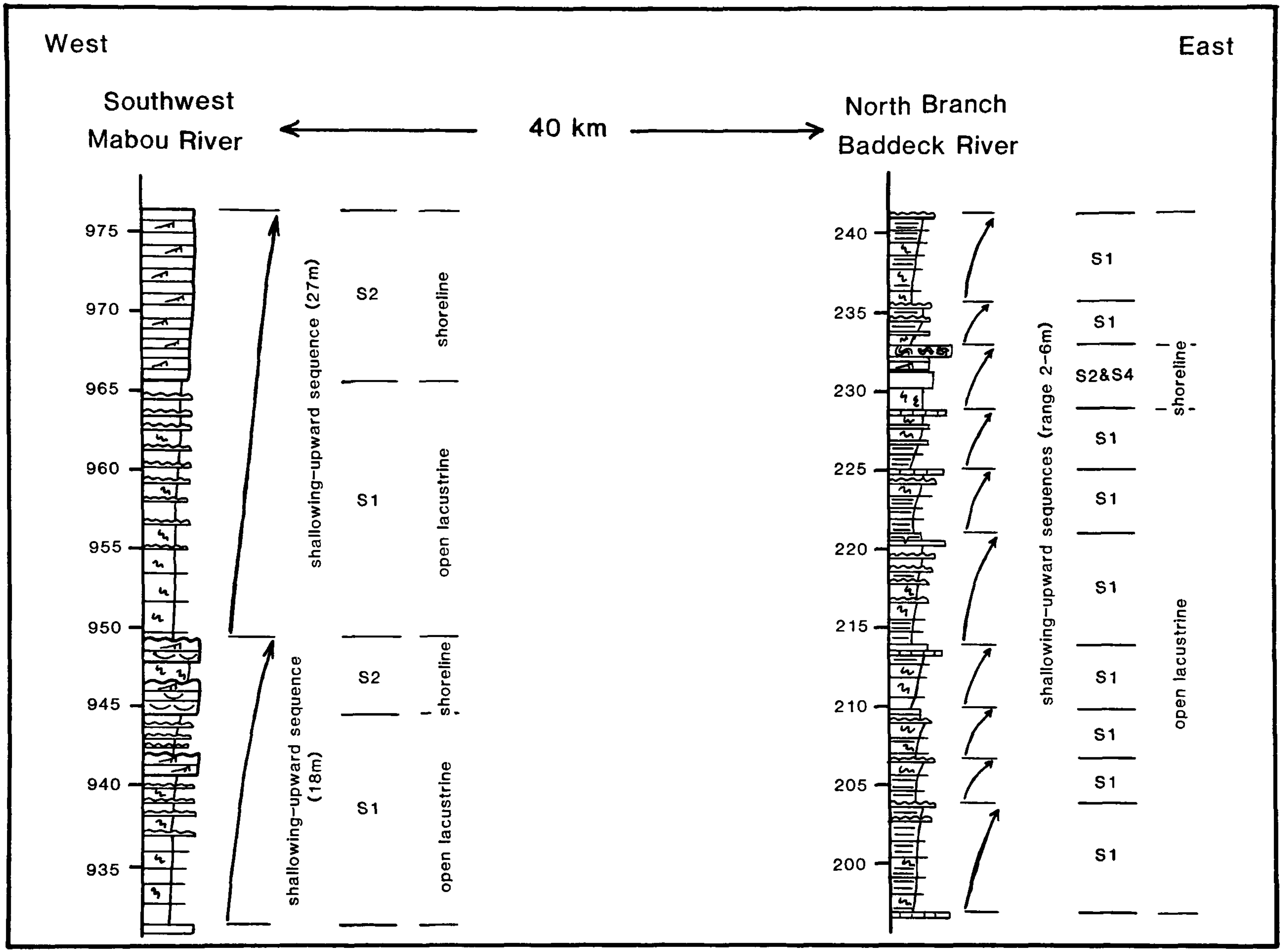

Figure 150. Comparison of shallowing-upward sequences of Strathlorne Formation, Western Cape Breton sub-basin. Note asymmetry of number, thickness and grain size of sequences.

Together this information illustrates a) a fundamental asymmetry of dominant sediment input, facies distribution and progradational style and b) a distinct reversal of dominant sediment input and facies distribution in later Strathlorne time (Fig. 151). These relations are very similar to those found in S2 and S3 facies assemblages. In northern Cape Breton, where only a portion of the sub-basin is exposed on land, a similar asymmetry is suggested. On the southeast margin (Aspy area) shallowing-upward sequences are thick with abundant occurrences of S2 and S3 facies assemblages, whereas toward the northwest (Meat Cove area) the sequences are thinner, with lesser, but still prominent S2 sediments. All data from the Strathlorne of both western and northern Cape Breton demonstrate a fundamental linkage of S1, S2, S3 and S4 facies assemblages, representing lacustrine depositional systems which display large scale inter-relationships between facies distribution, clastic input and asymmetry across entire sub-basins (Fig. 152).

The paleocurrent data of the Strathlorne Formation also indicates the linkage of these facies assemblages into a single lacustrine depositional system in each sub-basin, each with a distinctive asymmetry and history of sediment input. The total isopach map for Strathlorne sediments (Fig. 56) indicates the general limits of deposition and the likelihood of a structural high between Cheticamp and Pleasant Bay which separated the western and northern sub-basins.

In western Cape Breton, data from assemblage S3 (Figs. 85, 87) are sparse but appear to relate directly to localized coarse grained fan input at major fault-bounded margins. During middle Strathlorne time there was input from the Mabou-Margaree side, whereas in late Strathlorne time there was input from the Baddeck side. Data from assemblage S4 (Figs. 95, 97) are also sparse but generally indicate flow away from several basement blocks, such as Mabou Highlands, southern Cape Breton Highlands and the Bras d'Or Lakes area in middle and later Strathlorne times. Data from S2 (Figs. 75, 77) convincingly demonstrate that the bulk of Strathlorne shoreline/delta progradation proceeded from the Mabou-Lake Ainslie-Margaree-Cheticamp areas toward the east, particularly in early and middle Strathlorne times. In later Strathlorne time there was a significant reversal of this trend and shoreline progradation proceeded from the Bras d'Or Lakes-St. Ann's-southern Cape Breton Highland areas. Data from the S1 facies assemblage are extremely important because they represent the large volume of open lacustrine sediments and most paleocurrent measurements were taken from thin sandstone beds

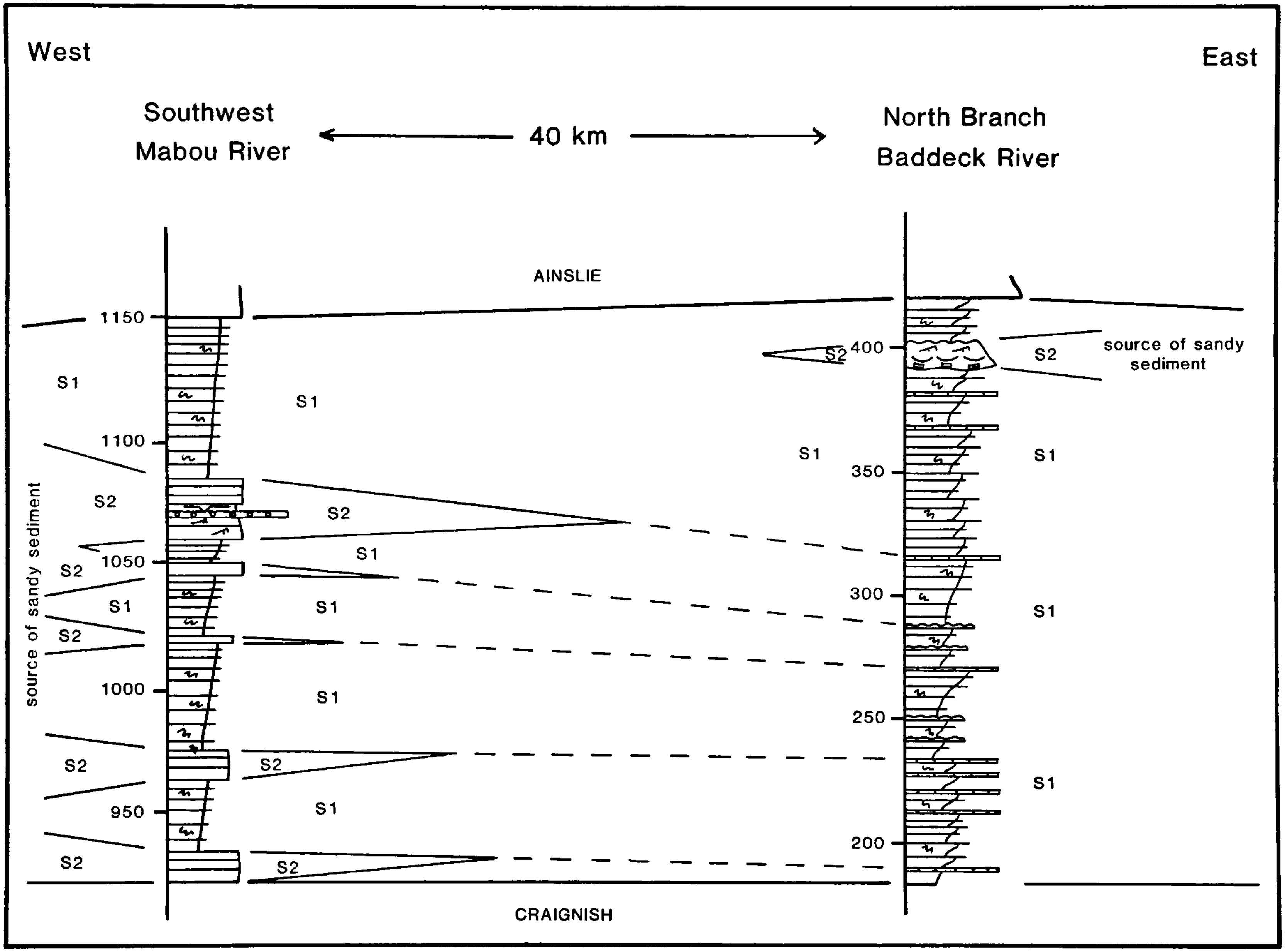

Figure 151. Cross section of Strathlorne Formation, Western Cape Breton sub-basin, illustrating fundamental asymmetry of sediment input and facies distribution, and reversal of these factors in later Strathlorne time.

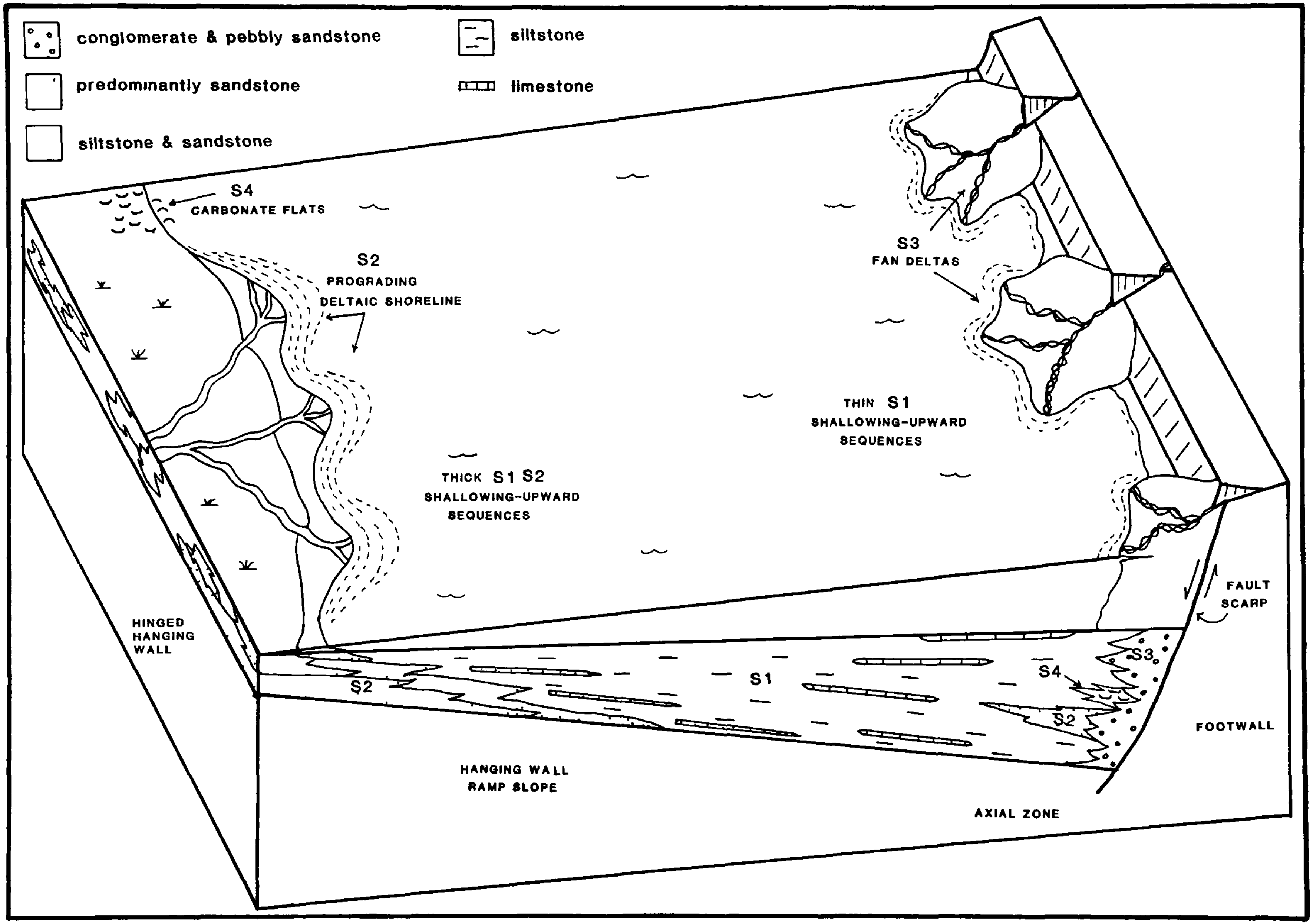

Figure 152. Block diagram illustrating depositional environments and facies distributions of S1/S2/S3/S4 depositional system.

formed by gravity-driven underflows. These data should best illustrate the regional paleoslope of the entire sub-basin, in areas generally removed from direct shoreline influence. Throughout the western Cape Breton area these data (Figs. 57, 59) indicate a single dominant sub-basin paleoslope toward the east or southeast.

In northern Cape Breton, data from S3 (Figs. 85, 87) are sparse but generally indicate flow away from several basement blocks in varying directions. This may indicate that during early and middle Strathlorne time the basement was broken by faulting into many blocks with varying tilt directions. In fact, some flow directions in the Aspy area are toward the interpreted sub-basin fault margin and suggest that the Aspy Fault was active during this time. This evidence is interpreted to show that major synthetic and antithetic faults can create isolated mini-basins within an extensional system as discussed by Gibbs (1984). Abundant data from assemblage S2 (Figs. 75, 77) convincingly demonstrate that the bulk of Strathlorne shoreline/delta progradation proceeded from the Aspy area toward the north, particularly in early and middle Strathlorne time, although only a potion of this sub-basin is exposed on land. During later Strathlorne time there appears to have been a decrease in sediment input from this margin. There is also support for the interpretation that the northern sub-basin was separate from that of western Cape Breton. Data from assemblage S1 illustrate the regional paleoslope of the sub-basin which dipped toward the north and northwest, away from the Aspy area. By analogy with western Cape Breton, that paleoslope was likely maintained in the unexposed portion of the sub-basin to the north.

The data and interpretations suggest definite fault-bounded sub-basin margins during Strathlorne time which delineate two separate sub-basins (Fig. 153). These sub-basins were likely evolutionary continuations of those defined for the C1/C2 depositional system. However, Strathlorne sub-basins are somewhat narrower, and deposition was dominantly aqueous rather than subaerial. Although there was a dominance of aqueous deposition, the presence of S3 fan deltas, indications of exposure and warm arid conditions in the S4, general lack of abundant vegetation and coal in S2 coastal sediments, and evidence of significant carbonate chemical sedimentation and desiccation phases in nearshore S1 sediments all confirm the continuation of generally warm, arid climatic conditions.

In summary, the Strathlorne Formation represents a discrete depositional system present in both sub-basins during a particular phase of Horton deposition in the Cape Breton region, with conditions somewhat different from those pertaining in Craignish time.

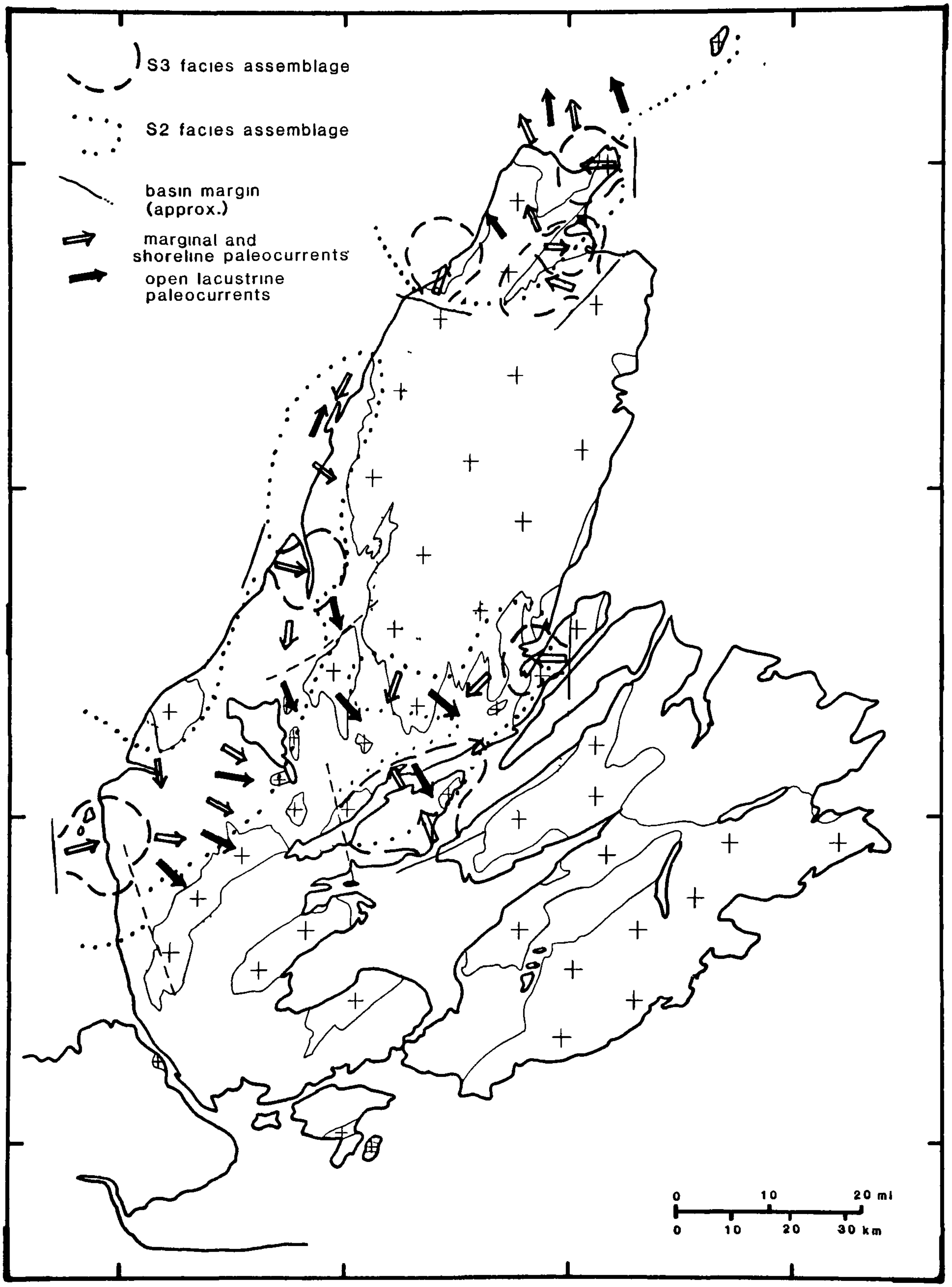

Figure 153. Map of sub-basin margins, distribution of facies assemblages and generalized sediment dispersal of S1/S2/S3/S4 depositional system.

The Strathlorne Formation was characterized by low energy open lacustrine sedimentation and moderate energy prograding shoreline sedimentation. Multiple recurrence of transgressive/regressive cycles through time and the asymmetrical distribution of facies assemblages in space (both on sub-basin-wide scales) were prominent features which must relate to the overall controls on deposition as discussed in Chapter 5.

THE AINSLIE FORMATION

A1 Red/grey pebbly coarse ss-cgl. The very thick A1 facies assemblage, which intertongues with A2 and is characteristic of the sub-basin margin areas (especially in the upper Ainslie), is characterized by three lithofacies: red, pebbly coarse sandstone to sandy conglomerate, subordinate lenses of red medium sandstone, and rare red siltstone with calcareous nodules and roots. The coarser lithofacies is predominant and occurs in thick scour-based, lens-shaped beds which fine upward. The units may be massive and disorganized or have clast imbrication and broad trough cross stratification. Diverse clast lithology, hematite rims on clasts, poor to moderate sorting and abundant rock fragments are characteristic. Partially preserved lenses of medium sandstone with trough cross stratification increase in number away from sub-basin margins. These coarse, relatively immature sediments suggest proximity to a relatively high relief source area where mechanical weathering and rapid transport were important. Rare interbedded fine grained sediments have evidence of exposure in a warm arid climate (such as pedogenic calcrete nodules and roots) and the facies assemblage is attributed to deposition in a distal alluvial fan to braidplain setting. The braidplain setting was reviewed previously but alluvial fans will be reviewed here.

Alluvial fans accumulate where drainage from a limited high-relief watershed intersects a steep highland margin. Rapid deposition of coarse bedload from flashy discharge streams occurs due to lateral flow expansion (Spearing, 1971; Miall, 1981; Rust and Koster, 1984). Aggradational accumulation of many of these individual beds results in fans which are thickest near the fault scarp and thin dramatically over a short lateral distance. Rapid lateral facies changes are characteristic. Many alluvial fans prograde outward and hence the base of definable fan deposits rises away from the fault scarp (Bull, 1972). Fan radii are generally less than 10 km and thickness less than 500 m, although in arid climates fans are typically areally smaller, with steeper gradients, and are thicker and coarser grained than humid climate fans (Spearing, 1971). Some deposition near the apex may be from debris flows which deposit mud-supported, disorganized gravels. The bulk of the sedimentation is from linear aqueous flows in complex, very shallow channel/bar systems on the constantly changing fan surface (Bull, 1972; Rust and Koster, 1984). This depositional style, comparable to that of braided streams, deposits sharp-based, broadly lens-shaped, fining-upward beds a few metres thick with pebble imbrication and/or cross stratification (Bull, 1972; Rust and Koster, 1984; Speksnijder, 1985). Short periods of dynamic deposition are separated by longer periods of relative inactivity. Lack of abundant vegetation and well developed soil horizons can be due to arid climate and/or continuous surface redistribution and aggradation. Large-scale cyclicity is common in alluvial fan sediments and is generally attributed to allocyclic (tectonic) processes because fan sedimentation is a direct response to the presence of a high-relief fault scarp (Miall, 1981; Rust and Koster, 1984). When coarsening-upward, these cycles have been attributed to episodes of basin subsidence or uplift at the basin margin (Heward, 1978). The stacking of several of these indicates periodic fault scarp rejuvenation at repeated irregular intervals (Steel and Gloppen, 1980).

In the A1 facies assemblage the lens-shaped fining-upward units typically begin with an erosive base with large scour pockets (facies Ss of Miall, 1977), followed by sandy conglomerate or pebbly sandstone, either massive, or with pebble imbrication or broad shallow trough cross stratification (facies Gm, Gt of Miall, 1977). This suggests a dominance of in-channel scour-and-fill with bedload transport as longitudinal bars and dunes migrated under high energy conditions. These units commonly cut down into one another, indicating a high energy aqueous system. Where sub-basin margin exposures are good the fining-upward units in assemblage A1 are arranged into stacked large-scale coarsening-upward sequences up to 100 m thick. The trend is expressed as an upward increase of bed thickness and maximum clast size, and a decrease in sorting and abundance of trough cross stratification. These features are attributed to rapid tectonic basin subsidence at fault-bounded margins and subsequent progradation of fan deposits over braidplain sediments during a phase of relative tectonic quiescence which would allow erosion and sedimentation rates to overtake subsidence and build a coarse grained wedge of sediment (Blair and Bilodeau, 1988). This interpretation will be discussed more fully in Chapter 5. The coarser lithofacies of A1 resembles the Scott-type proximal gravelly braided stream model (Miall, 1977, as distilled from Boothroyd and Ashley, 1975) in the dominance of pebbly deposits, textural immaturity, and paucity of sandstone and siltstone lithologies.

The subordinate red medium sandstones are better sorted and occur as thin lenses preserved at the tops of conglomeratic beds beneath scour bases. They have trough cross stratification (facies St of Miall, 1977), some syn-sedimentary deformation fabrics and their occurrence increases away from sub-basin margins, ie. in the down-fan direction. These finer grained, better sorted deposits are interpreted as in-channel, falling-water stage wedges deposited around gravel bar margins or in bar top scours (Miall, 1977; Rust and Koster, 1984). At several outcrops in the lower parts of the facies assemblage there are thicker units of red, massive, silty very fine sandstone with horizons of green calcareous nodules and roots. These are interpreted as paleosols formed on minor overbank areas at the distal toes of alluvial fans during periods of non-deposition and sediment surface stability (facies P of Rust and Koster, 1984). The red colouration, lack of abundant vegetation and minor calcretes suggest subaerial exposure in a warm arid climate.

<u>A2 Red fine-coarse ss and siltstone</u> The A2 facies assemblage is characteristic of the Ainslie Formation in a medial belt between dominantly basin margin sediments (A1) with which it intertongues and dominantly basin centre sediments (A3). It comprises three lithofacies; red fine to coarse sandstone, red sandy siltstone and minor greenish limestone beds. The predominant sandstone lithofacies is uniform, well sorted quartz arenite with hematite rims and calcareous cement. It occurs in sharp-based, fining-upward units several metres thick with lags of wood fragments and intraformational rip-ups. Horizontal or trough cross stratification passes upward into ripple cross lamination with burrows, roots and desiccation cracks near the tops. These characteristics suggest broadly scour-based depositional units, in a subaerial setting, where relatively high energy bedload transport (predominantly as dunes) waned upward. A fluvial style of deposition is indicated.

Units of assemblage A2 are interpreted as the main channel sandstones in a fluvial system which was less energetic and more organized than those described in previous facies assemblages. The characteristics are similar to the many described examples of meandering channel deposits (Allen, 1965, 1970; Walker and Cant, 1984) except that in-channel traction deposits always greatly predominate over overbank vertical accretion deposits. Channel sequences are repeatedly stacked through the thickness of the facies assemblage but are usually separated by small thicknesses of finer sediments suggesting some lateral migration of channels through time and/or multiple channel avulsion events, processes typical of meandering streams (Walker and Cant, 1984). These units record the deposits of channels several metres (up to 10 m) deep, according to the general correlation between channel sequence thickness and original channel depth (Walker and Cant, 1984). The channels probably had rather low sinuosity, underwent only moderate lateral migration, and had poorly vegetated point bars. They may depict a fluvial style partway between the braided and meandering end members of the fluvial continuum.

In-channel units are separated by red siltstone with sandy streaks and minor, thin, sharp-based beds of red fine sandstone. Horizontal and ripple cross lamination, roots and desiccation cracks occur. These are interpreted as point bar top and overbank deposits formed in close proximity to channels subject to flooding and crevasse splay generation. The siltstone is uniformly sandy and there is little claystone, except in locations farthest from paleo-basin margins which suggests that most overbank deposition occurred quite close to channels. However, the small thickness of overbank units, relative lack of splay sandstone and abundant evidence of stability may further indicate that sinuosity was rather low, most flood events were retained within the channels, and most deposition in this system was from bedload rather than suspended load. The red colour, relative paucity of vegetation and presence of desiccation cracks reflect a warm, arid climate. A distinctive component is the common presence of several types of limestones. Some are thin, massive, nodular and display desiccation cracks; they are interpreted as pedogenic calcretes. Others are thin, extensive and characterized by alternating calcite-rich and quartz siltstone-rich laminae with ?ostracod shell fragments and ?root traces. These are interpreted as the deposits of minor freshwater ponds in low-lying interchannel areas where clastic input was periodic but minimal, and vegetation and shell banks were present. The presence of both limestone types indicates relatively long periods of stability with little clastic flood input (Leeder, 1976), suggesting that channel sinuosity and the rate of meander migration were low. However, the channels certainly did migrate laterally through overbank areas since these limestones provided distinctive rip-up clasts for the basal lags of channel units (first noted as characteristic of Ainslie fluvial sediments by Murray, 1960).

<u>A3 Grey/green fine sandstone and siltstone</u> The A3 facies assemblage, characteristic of the Ainslie Formation in basin central positions comprises three lithofacies: thick very fine to medium sandstone, grey or red siltstone, and thin very fine to fine sandstone. The assemblage is approximately equivalent to the McIsaac Point and Glencoe Members of

Murray (1960). Fine sandstone occurs as fining-upward units, several metres thick, of well-sorted, micaceous, calcareous quartz arenite. These units have sharp bases (commonly deeply scoured with steep side walls) with lags of pebbles, wood fragments and intraformational rip-ups, overlain by a sequence of trough cross stratification passing upward into ripple cross lamination. These characteristics indicate erosive turbulent flow in fluvial channels with sandy bedload transport as dunes and a decreasing flow intensity upward. The mineralogical and textural maturity of the sediments is typical of a distal setting. Channel units tend to be isolated in thick finer grained units, but are stacked and cut into one another at several locations. They become less numerous, and thinner in areas farthest from interpreted sediment sources. The channels are interpreted as high sinuosity, meandering ones in which lateral migration of point bars was the predominant process, producing well developed fining-upward sequences similar to those of Allen (1965, 1970). The presence of steep-walled scours, abundant siltstone rip-ups and wood fragments at channel bases suggests strong lateral migration through competent, fine grained overbank deposits on which significant flora developed. The common presence of climbing ripples, claystone/siltstone laminae and roots near the tops of channel units suggests a large suspended sediment load available for overbank deposition, with vegetation on the point bar tops.

The sandstone units are separated by units of calcareous micaceous siltstone of equal or greater thickness, interpreted as overbank vertical accretion deposits (Walker and Cant, 1984). Thin sharp-based beds of very fine to fine sandstone with tool marks, burrows, and horizontal and ripple cross lamination are ubiquitous and are interpreted as splay sandstones representing rapid, but waning, influxes of sediment-laden flood water into the quiet overbank environment. In areas where channel sandstones are thicker, coarser and more numerous (ie. relatively proximal to sediment sources), overbank deposits are reddish, sandy, massive and have many splay sandstones and green calcareous nodules interpreted as pedogenic calcretes developed in a warm arid climate. In areas where channel sandstones are thinner, finer and less numerous (ie.relatively distal from sediment sources), overbank deposits are grey or greenish, argillaceous, laminated, bioturbated and have few splay sandstones and rare laminated limestones. These are interpreted as the deposits of ponded water. The characteristics of the ponded water deposits probably did not indicate different climatic conditions but rather represent deposition on a low-lying floodplain, close to the water table, where periods of exposure were minimal, and fine grained suspended sediment load was predominant. A few palynological samples from overbank grey siltstone of the A3 facies assemblage yielded a flora consistent with a fluvial setting and abundant transported woody/coaly material.

The A1/A2/A3 Depositional System The Ainslie Formation comprises a proximal to distal arrangement of A1, A2 and A3 facies assemblages as components of a single depositional system extending from fault-bounded sub-basin margin to sub-basin centre (Fig. 154). The Ainslie Formation is very poorly exposed and preserved in northern Cape Breton and the interpretations here are primarily demonstrable in western Cape Breton, although they probably apply in the north as well. The three facies assemblages grade into each other laterally, based on observable intertonguing and position relative to original sub-basin margins. The proximal sub-basin margin areas were dominated by high relief and vigourous coarse-grained sedimentation on alluvial fans and proximal braidplains (A1), medial zones were dominated by transitional braided/meandering fluvial channels with minor overbank areas (A2),and the basin centre area was dominated by low relief and less vigourous fine-grained sedimentation in meandering channels and extensive overbank areas (A3). Vegetation was present but limited, probably due to an arid climate, and minor palynological data support a fluvial setting. Red colours and calcretes indicate a warm arid climate.

The A1 alluvial fan/braidplain deposits represent predominance of very vigourous, steep gradient depositional processes which transported coarse grained immature sediment over rather short distances. Their presence indicates that during Ainslie time the depositional sub-basins were bounded by high relief fault scarps. This facies assemblage is vital in delineating the fault-bounded margins because alluvial fan sediments may be very thick adjacent to the fault scarp but extend only a few kilometres basinward (Spearing, 1971). Subsidence at fault margins was probably not rapid and continuous because the alluvial fans apparently coalesced laterally into extensive bajada complexes, at least in the Baddeck area, and prograded into the basin. The base of the A1 facies assemblage rises toward the basin centre where it overlies A2 deposits, and thins radically to zero. Large-scale coarsening-upward sequences indicate direct response to fault margin subsidence events, a characteristic of alluvial fans. The sequences are stacked, suggesting several phases of fault motion in Ainslie time. Thick units of A1 are present only near the Bras

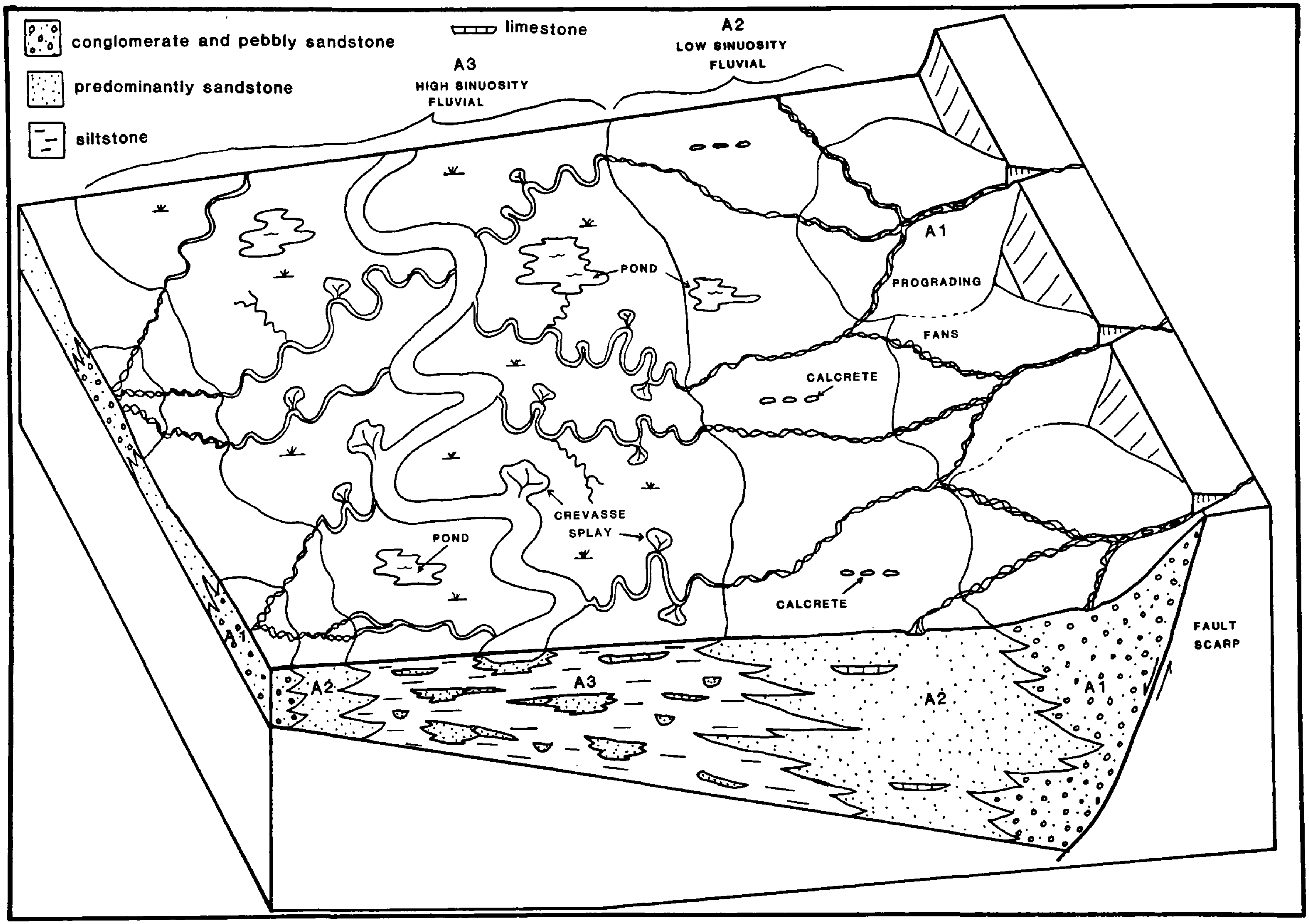

Figure 154. Block diagram illustrating depositional environments and facies distributions of A1/A2/A3 depositional system.

d'Or lakes/Baddeck/St. Ann's area and Mabou Highlands area of western Cape Breton, suggesting a closed basin. The A1 facies assemblage does not occur in basin centre positions. In northern Cape Breton, where only a portion of the sub-basin is exposed on land the A1 assemblage occurs only in the Aspy area. In western Cape Breton there is a definite asymmetry in the distribution of the A1 assemblage whereby it is much thicker and more extensive on the eastern side (implying greater sediment input and higher relief) than on the western side (Fig. 155).

The medial A2 braided/meandering fluvial deposits are dominant below and basinward of those of A1 facies assemblage and grade laterally into both A1 and A3. They represent predominantly sandy bedload transport in fluvial channels of moderate sinuosity, a fluvial style somewhere between classical braided and meandering end members. They indicate a sedimentological gradation away from alluvial fan/braidplain environments at the fault-bounded sub-basin margins into a more moderate relief setting basinward of the fan toes. There is no evidence of sequences related to fault subsidence episodes but it is common for the dynamic processes of fluvial systems to "smooth out" tectonic effects (Rust and Koster, 1984). In western Cape Breton thick developments of this facies assemblage are present, particularly in the Bras d'Or Lakes/Baddeck/Creignish Hills areas on the eastern side, but it is less well developed on the western side. This distribution is more widespread than that of the A1 facies assemblage but has a similar asymmetry (Fig. 155). The configuration implies a closed basin of deposition and a fundamental linkage of A1 and A2 deposits. The A2 facies assemblage is very poorly exposed in northern Cape Breton.

The distal A3 meandering fluvial deposits are the only Ainslie facies assemblage present in the sub-basin central area, where they overlie, and are difficult to distinguish from, Strathlorne sediments. The A3 assemblage interfingers toward sub-basin margins with the A2 facies assemblage and represents low to moderate energy transport, predominantly as abundant suspended sediment load, in a low relief setting. Deeply-cut, isolated, sinuous channels migrated laterally through extensive fine grained overbank areas where flood events frequently overtopped channel banks. Flood water was ponded in interchannel areas for periods long enough to allow freshwater limestones to form. The characteristics suggest deposition near the water table in the low-lying central portion of the sub-basin removed from the influence of coarse sediment input from fault-bounded margins. The A3 facies assemblage is well developed throughout the Mabou-Lake Ainslie-Margaree

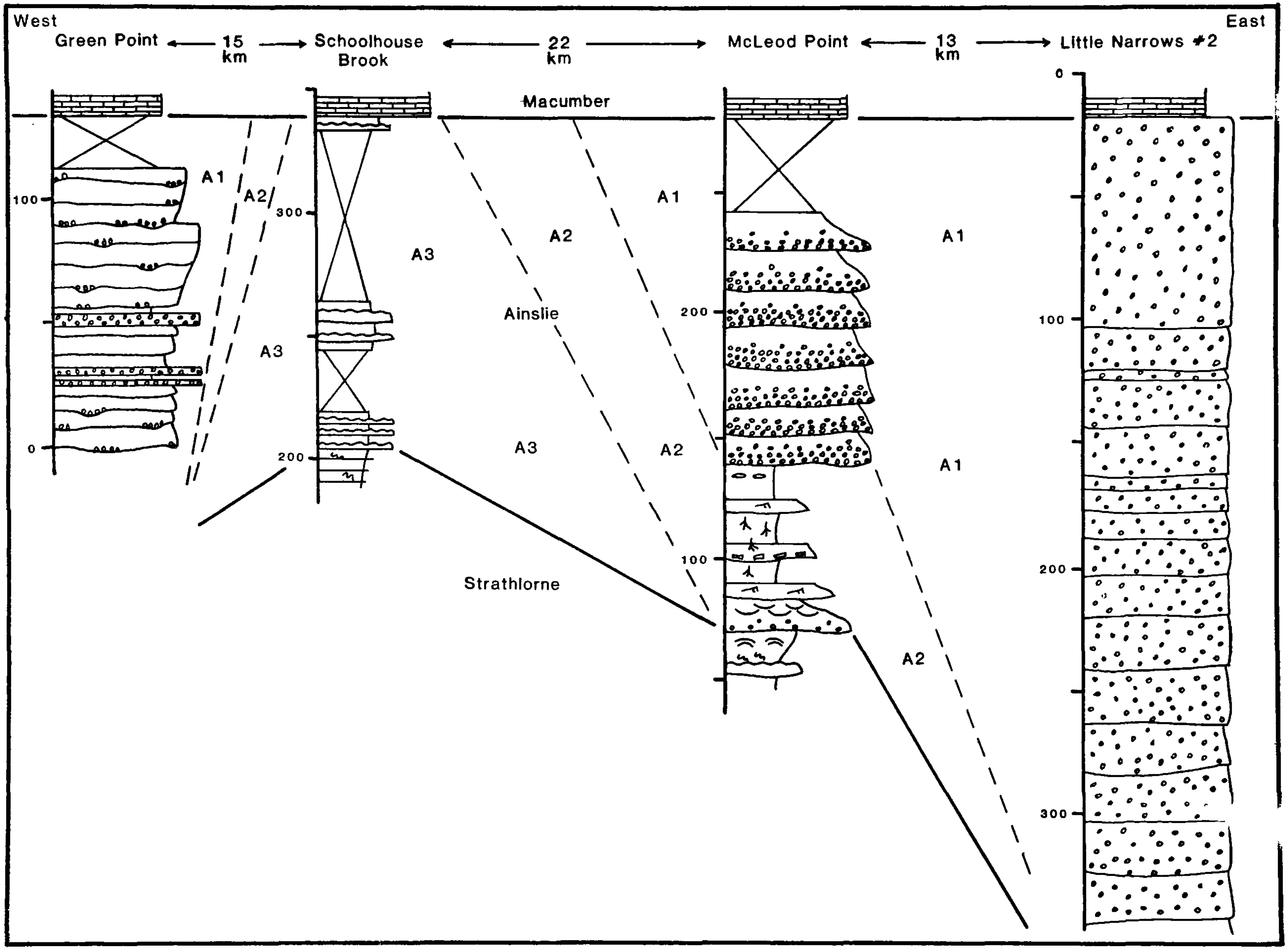

Figure 155. Cross section of Ainslie Formation, Western Cape Breton sub-basin, illustrating depositional asymmetry of sediment input and facies distribution.

area of western Cape Breton. It may represent a distal component of A1/A2 sedimentation from sub-basin margins and/or it may represent a longitudinal meandering system comprising the final fill stage of the Strathlorne lacustrine depositional system with little influence from the sub-basin marginal environments. The A3 facies assemblage is very poorly represented in northern Cape Breton.

The paleocurrent data of the Ainslie Formation also indicate the linkage of A1 and A2 facies assemblages into an alluvial/fluvial depositional system with a distinctive asymmetry and history of sediment input. The total isopach data from Ainslie sediments (Fig. 103) indicate the general limits of deposition and the likelihood of a structural high between Cheticamp and Pleasant Bay, which separated the western and northern sub-basins.

In western Cape Breton data from A1 (Figs. 104, 106) are abundant enough to indicate fault margin alluvial fans shed from Bras d'Or Lakes-St. Ann's areas on the east, Mabou Highlands on the west, and Cheticamp area in the north (Fig. 156). The latter demonstrates the structural separation of the 2 sub-basins. A distinct asymmetry is indicated, with much more sediment input from the eastern margin, indicating greater fault scarp relief and relative stability so that fan progradation was extensive. Data from A2 (Figs. 120, 122) are similar and suggest sediment input from Bras d'Or Lakes-St. Ann's area on the east, Mabou Highlands on the west, Cheticamp area in the north, and a great deal of input from the southern Cape Breton Highlands in the Baddeck area. Again, a distinct asymmetry is present with much more sediment input on the eastern side, as in A1. By analogy with other areas the thick wedge of A2 deposits in the Baddeck area implies the presence of thick A1 sediments to the north which do not occur in outcrop. It is likely that the proximal A1 deposits of the upper Ainslie in this area were either upfaulted and cannibalized during Ainslie deposition, or upfaulted and eroded at some later time. In fact, this has likely happened in many areas as minor outliers of red conglomeratic sediment are commonly preserved on tops of basement blocks throughout western and northern Cape Breton (Ferguson, 1946; S. Barr and R. Raeside, 1987, pers. comm.). Data from A3 (Figs. 131, 133) suggest rather diverse flow directions, although much of the data is from ripples which do not adequately reflect overall channel flow direction. A great deal of variation is expected in any case, because the channels are interpreted as meandering. In some cases sediment input was apparently from lateral sources related to A1/A2 deposition, whereas in other areas there is a suggestion of a general longitudinal southwest to northeast trend.

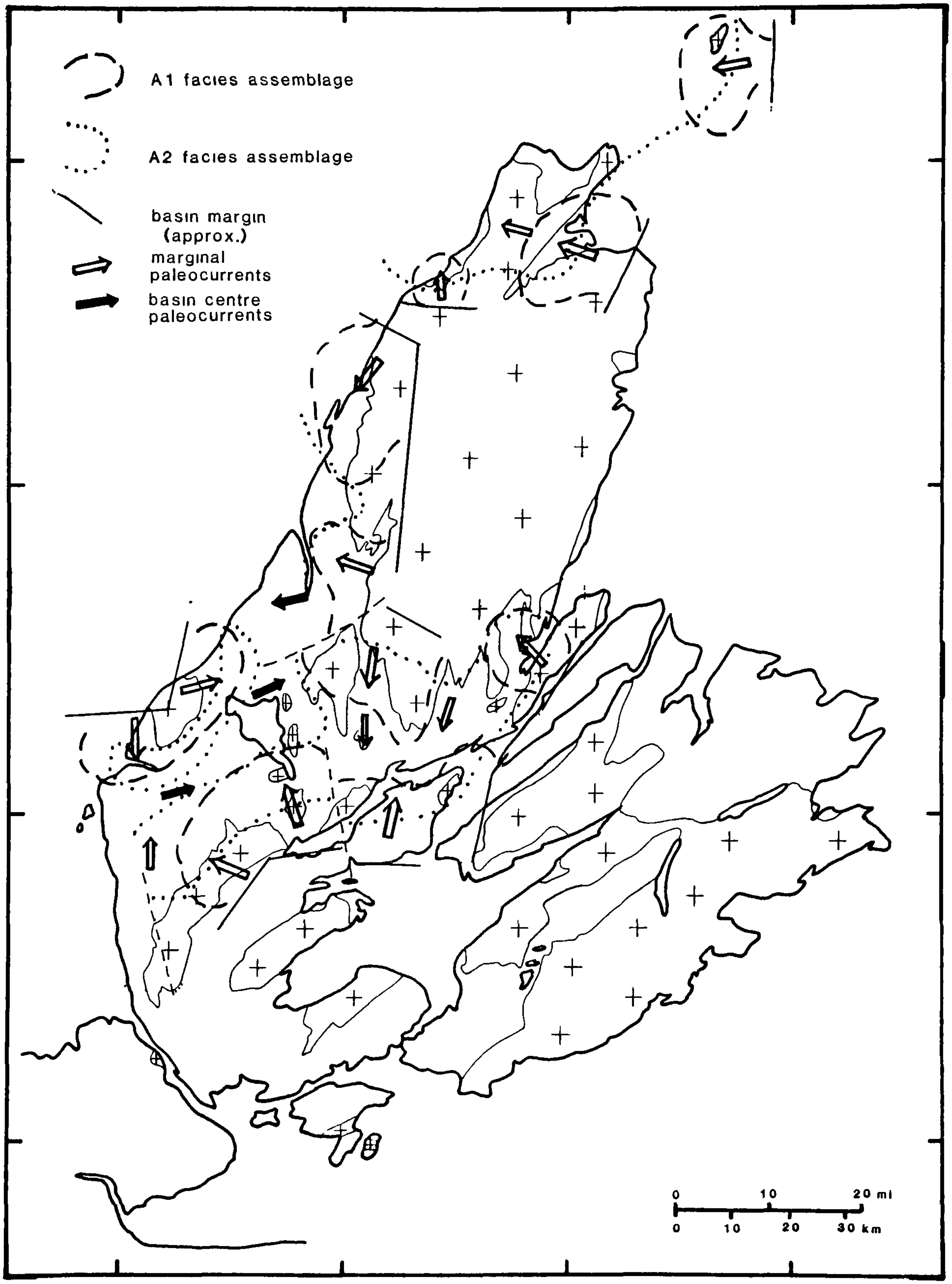

Figure 156. Map of sub-basin margins, distribution of facies assemblages and generalized sediment dispersal of A1/A2/A3 depositional system.

If this is correct it may delineate an overall basinal paleoslope dipping to the north or northeast in Ainslie time (Fig. 156).

In northern Cape Breton the Ainslie is very poorly preserved and exposed, with the bulk of information only from the Aspy and St. Paul Island areas. Minor paleocurrent data from A1 and A2 facies assemblages in the Aspy area suggest marginal sediment input toward the west from a fault-bounded margin in that area.

AFTERWORD

The facies distribution, paleocurrent and isopach data for all facies assemblages indicate the fault-bounded margins which affected and controlled deposition throughout Horton time. They also suggest that many blocks of Precambrian basement currently exposed at surface were not positive features in the Early Carboniferous. These include small basement blocks scattered through western Cape Breton as well as large blocks at the edges of both sub-basins. The presence of these, which separate Horton outcrop areas and which appear to have separated many small depositional areas, have confused interpretations in the past. The present study has illuminated this problem and concluded that, in contrast to some other interpretations, Horton deposition in the study area occurred in only 2 major sub-basins with definable and logical facies distributions. These have since been altered by much tectonic disturbance which tended to mask the overall relatively simple distribution of facies as shown in this study. In Chapter 5 the structural and sedimentary pattern of fault-bounded extensional basins are discussed. The significance of the present Horton interpretations is then placed into the context of a tectono-sedimentary analysis of the Horton basin-fill sequence.

CHAPTER 5

THE HORTON TECTONO-SEDIMENTARY BASIN-FILL SUCCESSION

INTRODUCTION

In Chapter 4 the 10 facies assemblages were given general sedimentological interpretations. Each facies assemblage is interpreted in terms of a discrete depositional setting, made up of specific depositional environments represented by the individual lithofacies. Related depositional settings are considered as components of major nonmarine depositional systems, which in turn represent a specific phase in the evolution and fill of the Horton sub-basins. All 4 systems were dominated by combinations of inter-related alluvial/fluvial/shoreline-delta/lacustrine deposits whose arrangement depicts vertical and lateral changes in environments and limits of sub-basin geometry through time. Deposition through most of Horton time was primarily influenced by motion on fault-bounded margins in extensional sub-basins.

Contrary to indications from present day outcrop patterns within the study area, the distribution, thickness and paleocurrent data of the facies assemblages and depositional systems indicate sedimentation in only 2 large fault-bounded sub-basins. Within these sub-basins three depositional systems are characterized by asymmetric facies distribution and dispersal, probably related to tectonic activity at the original fault-bounded margins. The relative simplicity of these arrangements is masked by later structural complications. The characteristics and controls of fault-bounded extensional basins are reviewed below as a preamble to synthesizing Horton interpretations.

INTRACONTINENTAL FAULT-BOUNDED EXTENSIONAL BASINS

There are 2 main end-members of fault-bounded basins in the continental setting. "Distensive" linear grabens result from pure lateral extension and are bounded by dip-slip normal faults. Basin formation and evolution are dominated by vertical subsidence of the basin floor. "Transtensive" rhombic pull-aparts result from wrench extension and are bounded by strike-slip or oblique-slip faults. Basin formation and evolution are dominated by a combination of lengthwise horizontal prolongation and vertical subsidence of the basin floor. Both types occur along linear fault zones where an extensional stress field is maintained over a relatively long period of time (Illies, 1981). Later changes in that stress field may render some structural features inactive but reactivate others with different senses of motion (Illies, 1981).

Most fault-bounded intracontinental basins are elongate parallel to structural grain, and sedimentation occurs in internal drainage systems involving both transverse and longitudinal components (Miall, 1981). A common suite of facies includes a) proximal coarse grained alluvial fan and fluvial deposits, b) medial sandy fluvial and shoreline deposits, c) distal fine grained lacustrine deposits (Miall, 1981; Reading, 1986). Asymmetry of facies distribution is common, as described by Van Houten (1965), Ryder et al. (1976) and Link and Osborne (1978).

The typically thick sedimentary section is directly linked to tectonic events in a steady state feedback process and displays geologically instantaneous reaction to those controls. Depositional environments directly "feel" the basin margin and structural events more than in most marine settings. Tectonism controls sedimentation by its influence on basin geometry and asymmetry, uplift/subsidence and erosion rate, topographic gradient, and subsequent thermal and deformational effects (Steel, 1976). Sedimentation rates are generally high due to the large topographic relief (causing rapid erosion) and internal drainage into a relatively confined depocentre (causing rapid fill). However, if subsidence is particularly rapid or continuous there could be periods of sediment starvation and slow deep water deposition (Pitman and Andrews, 1985). Because tectonic phases commonly vary in intensity and episodicity, the sedimentary record commonly displays distinct packaging and trends (Steel et al., 1978). In addition to tectonic controls, Manspeizer (1985) suggested that a large fault-bounded basin can create secondary climatic influence as rain shadow and adiabatic air flow effects, resulting in a more arid intrabasinal climate.

TRANSTENSIVE BASINS (PULL-APARTS)

Basin Characteristics Transtensive basins occur along major oblique-slip fault zones due to localized extension at a releasing curve in the throughgoing fault (Crowell, 1974; Rodgers, 1980) or overlap of en echelon fault segments (Mann et al., 1983) (Fig. 157). Typically there is no mantle upwelling, little or no volcanism and lithosphere is conserved. Zones of anastomosing subparallel wrench faults involve oblique motion, producing complex patterns of simultaneous convergence (causing push-ups) (Fig. 157) and divergence (pull-aparts) and pervasive block rotation (Reading, 1986). Transtensive basins are relatively small (generally less than 50 km long), relatively short-lived, and asymmetric. They have

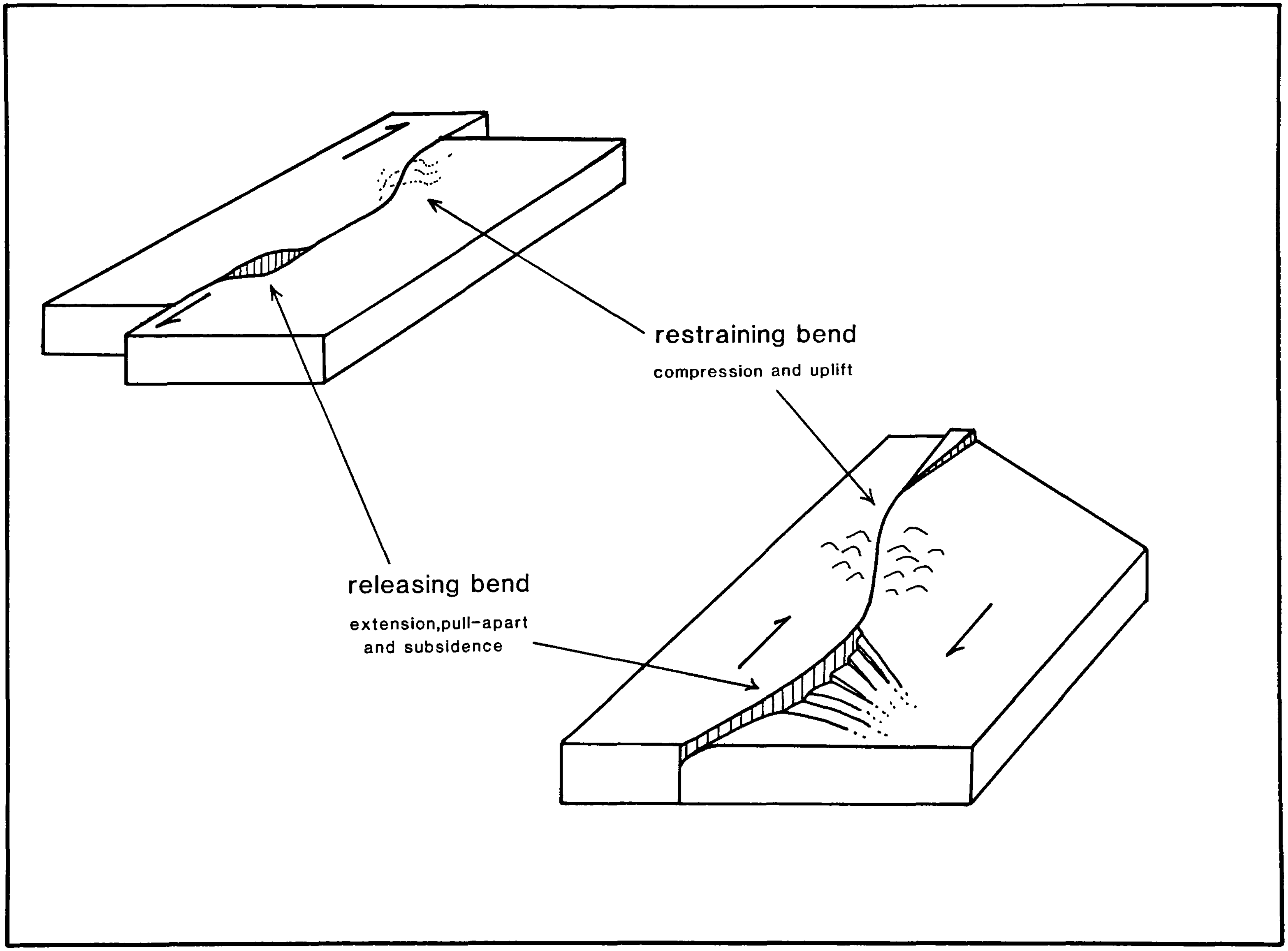

Figure 157. The formation of a transtensive pull-apart basin at the releasing bend of a strike-slip fault zone (modified from Crowell, 1974; Reading, 1985).

a rhombic shape and may occur in series along a transcurrent zone, separated by contemporaneous compressional areas (Reading, 1986). Each basin has a unique structural history and is subject to much structural alteration as regional movement continues (Reading, 1985). Extension is non-uniform and in its later phases, two deep depocentres may develop at the ends of the basin separated by a central sill raised by strike slip faults (Rodgers, 1980; Mann et al., 1983). These depocentres migrate through time as the pull-apart lengthens (Crowell, 1974). Fundamental structural patterns, including oblique faults and folds, develop in the basement (Fig. 158) and may be reflected in, or more likely concealed by, the sedimentary fill (Cloos, 1955; Freund, 1974; Christie-Blick and Biddle, 1985). Gloppen and Steel (1981) pointed out that subsidence may be very deep because fault-bounded blocks can be forced downward by adjacent ones which move and rotate independently. Most transtensive basins are short-lived because later compression and destruction are almost inevitable in an active oblique-slip system (Reading, 1986). Each basin in this dynamic setting undergoes a unique structural history (Reading, 1986) and so should have a unique stratigraphy not directly correlatable to others along the fault zone. Because of this individuality the abundant theoretical modelling in recent years has only reproduced the first-order characteristics of these basins.

Basin-fill Characteristics In transtensive basins the sediment fill is a localized complex and very thick succession (up to 10's km thick in a basin 10's km long x several km wide), resting unconformably on highly faulted basement rocks. There is a distinct lengthwise stacking of en echelon sediment bodies through time due to episodic lengthwise migration of the depocentre parallel to master faults, as one end of the basin pulls away from the other (Gloppen and Steel, 1981; Zak and Freund, 1981) (Fig. 159). This leads to lateral diachroneity and a very rapid accumulation of apparent stratigraphic thicknesses of the same order of magnitude as the size of the basin, a feature not possible in distensive basins (Speksnijder, 1985). The Devonian Hornelen basin of Norway is 70 km long but contains over 25 km of stratigraphic thickness (Steel et al., 1978); the Permian Cantabrian basin is 55 km long but accumulated 15 km of basin-fill (Speksnijder, 1985); the Quaternary pull-aparts of Turkey are up to 15 km long with 5 km of stratigraphic thickness (Sengor et al., 1985); and the Pliocene Ridge basin of California is 40 km long but contains 10 km of stratigraphic thickness (Link and Osborne, 1978). This is the most characteristic feature of deposition controlled by strike-slip motion. Petrographic trends reflect the overall migration

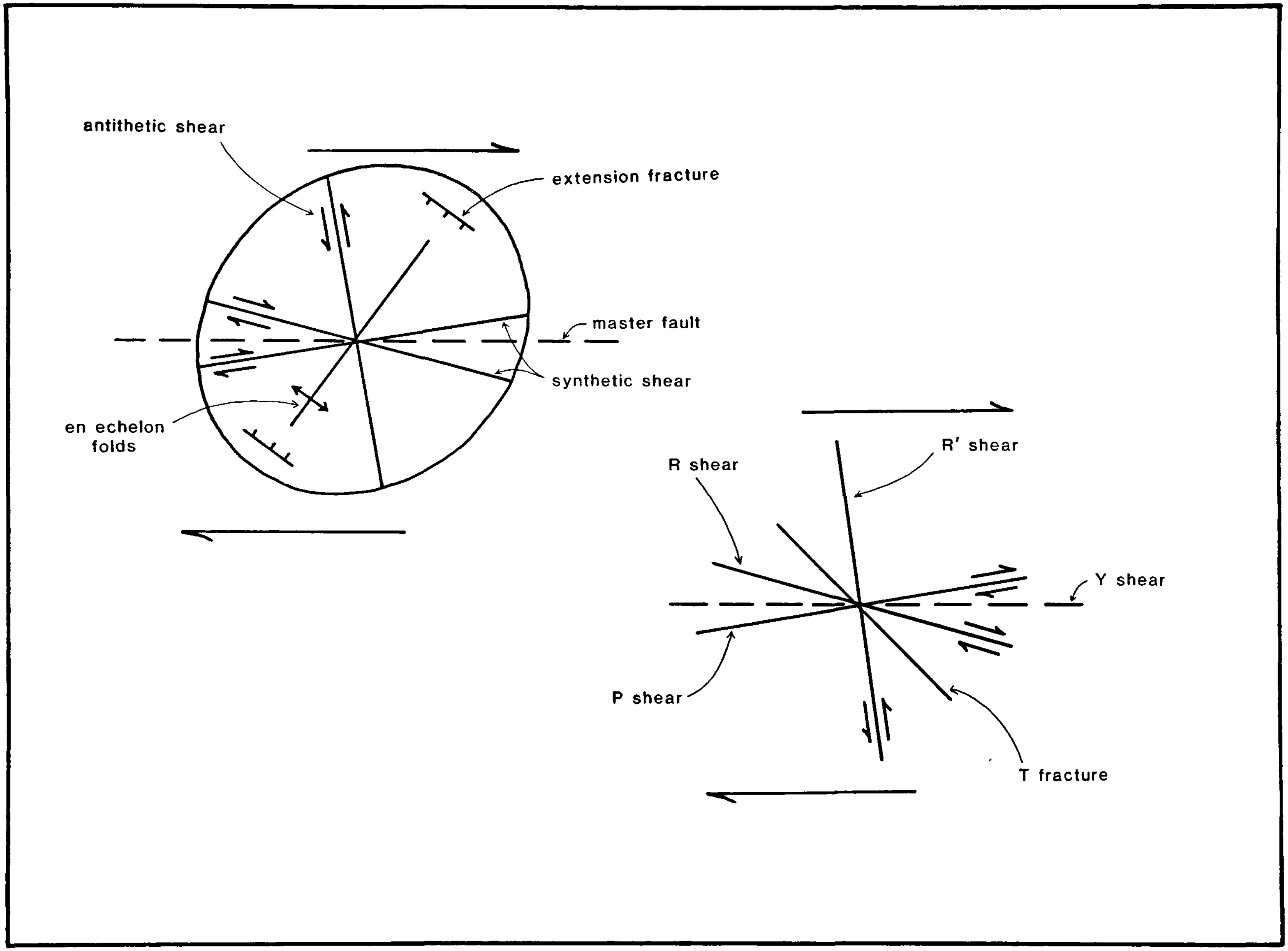

Figure 158. Strain ellipse and pattern of Riedel shear structures in basement rocks developed during strike-slip faulting (modified from Crowell, 1974).

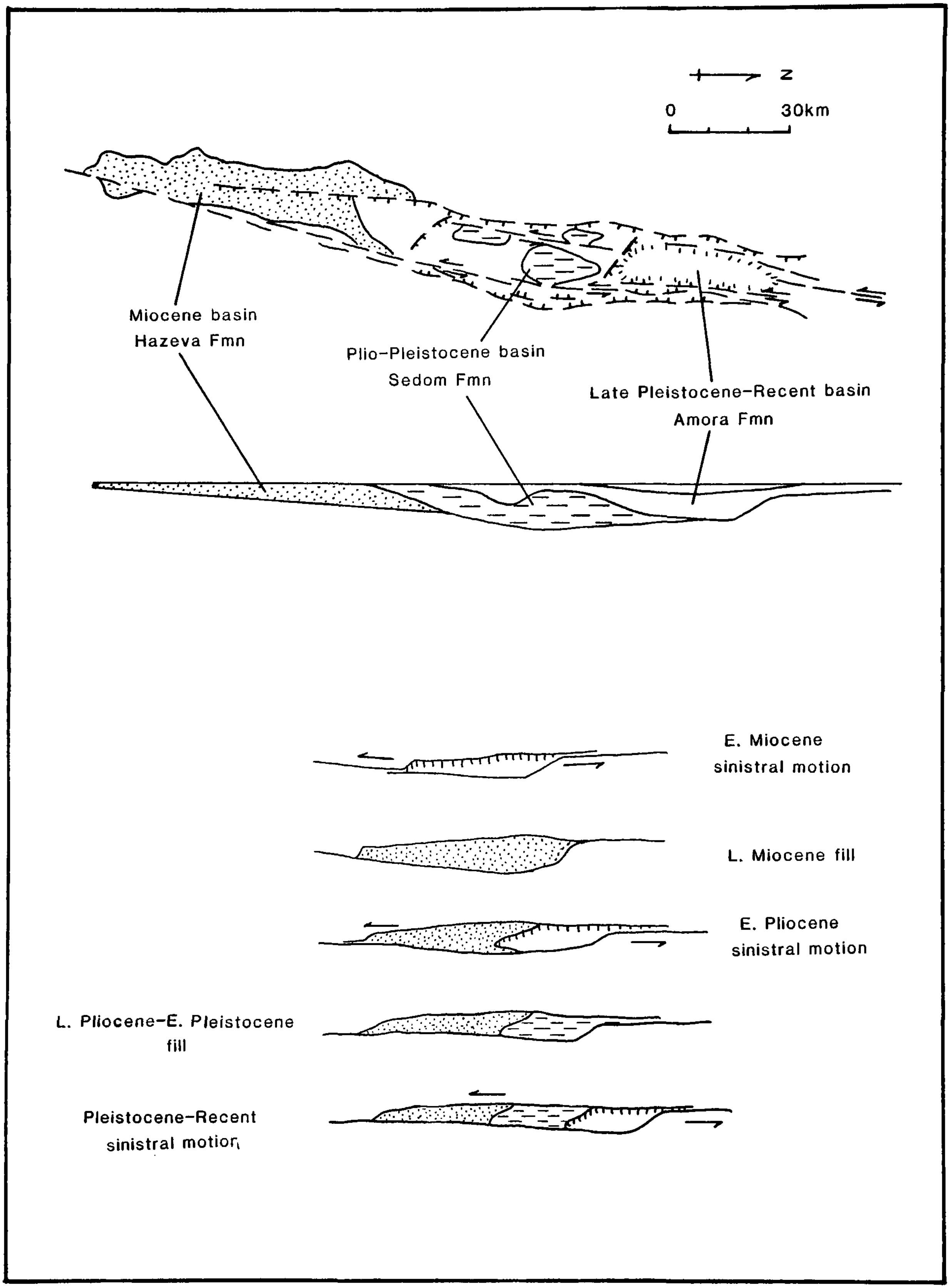

Figure 159. Progressive development of lengthwise stacking of en echelon sediment bodies as depocentre migrates during strike-slip pull-apart, Dead Sea (modified from Zak and Freund, 1981).

of the depocentre and lateral offset of the sediment source area (Ballance, 1980). Synsedimentary fault motion is continuous and soft-sediment deformation is prominent.

The sedimentation patterns are controlled by the details of strike-slip displacement, structural asymmetry, localized depocentres and uplifts. Most sediment input is from one dominant fault margin and from the basin end which pulls away and is dominated by uplift (Gloppen and Steel, 1981) (Fig. 160). The sediment patterns are very complex but Gloppen and Steel (1981) suggest that an abundance of diachronous large-scale basinwide coarsening-upward sequences is typical of transtensive basins because of the unique way in which they are filled. Because strike-slip motion is dominant the basinwide lateral offset must exceed vertical subsidence in the long term and it is recorded by the en echelon stacking of stratigraphic thicknesses 2-5 times the actual basin depth (Gloppen and Steel, 1981; Speksnijder, 1985). Narrow belts of coarse marginal facies and broad belts of sandy central facies are characteristic, whereas fine grained lacustrine sediments are less common because of the high ratio of sediment input to basin size (figure 160).

DISTENSIVE BASINS (GRABENS)

Basin Characteristics Distensive basins are typically large (100's km long x 10's km wide x several km deep) and long-lived (10's Ma). They are associated with crustal thinning, mantle upwelling, high heat flow and some volcanism (Neugebauer, 1983). Two methods of formation have been favoured by different researchers (Fig. 161). An active mantle convection mechanism would generate a crustal hotspot followed by regional doming, crustal thinning, abundant volcanism and rift subsidence (Bott, 1981; Neugebauer, 1983; Bott et al., 1983). Conversely, active tensional stress at a plate boundary would cause horizontal extension, crustal thinning and stretching, followed by passive mantle upwelling and rift subsidence with little volcanism (Turcotte and Emerman, 1983; McKenzie, 1978; Girdler, 1983). Fairhead (1986) and Dunkleman et al. (1988) noted the differences between the East African Rift (extension of a hot, thin lithosphere, regional domal uplifts, abundant volcanics, little subsidence and little sedimentation) and the West African Rift (rapid extension of a cool, thick lithosphere, no domal uplifts, few volcanics confined to the flanks, deep subsidence and abundant sedimentation). These may represent two end members of a rift continuum as espoused by Milanovsky (1972). Jarvis (1984) suggested that the difference is due to the net effect of competing crustal thinning (causing subsidence) and

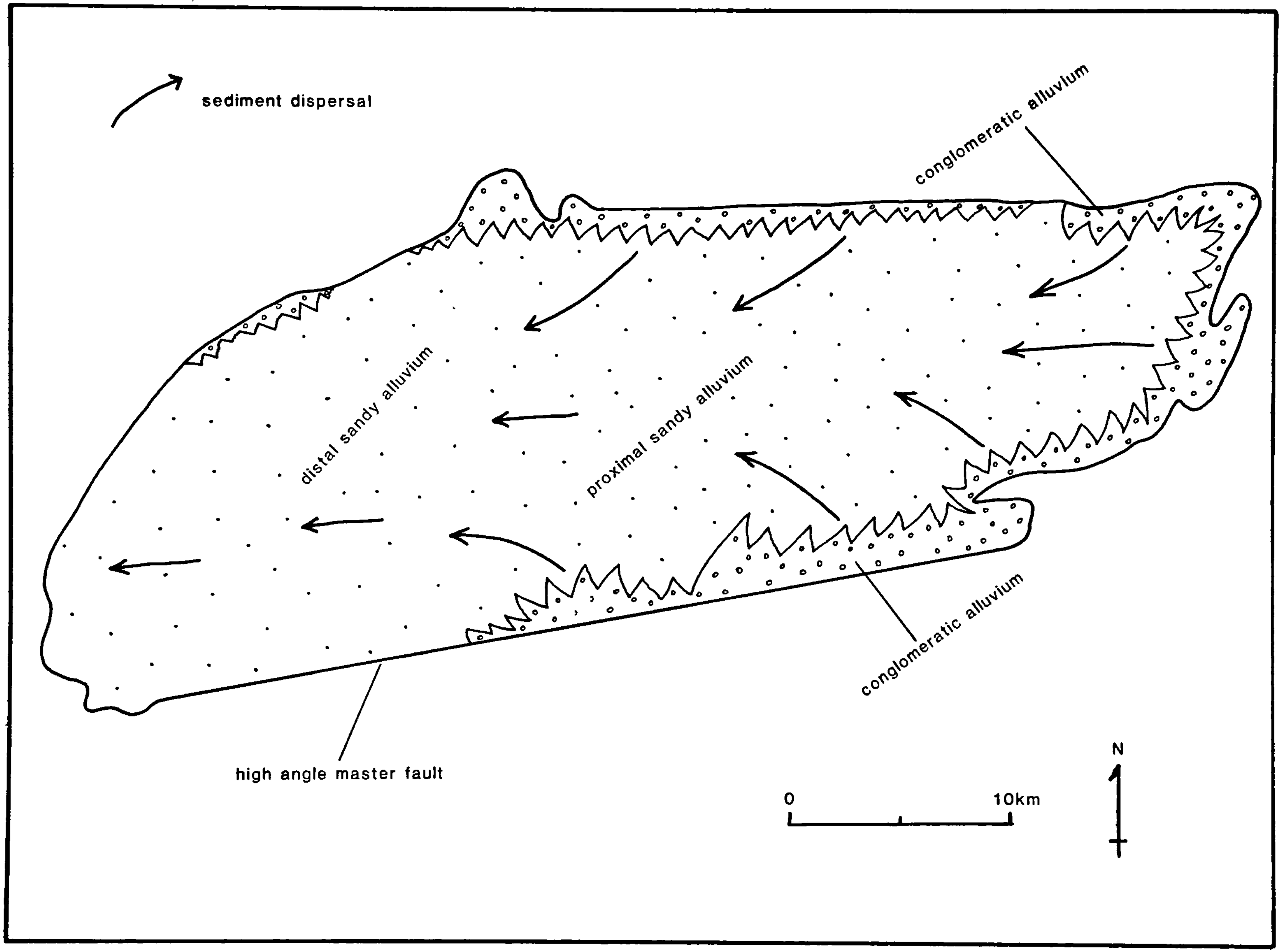

Figure 160. Predominant coarse-grained sediment input from margin which migrates away in strike-slip pull-apart basin, Hornelen Basin (modified from Gloppen and Steel, 1981).

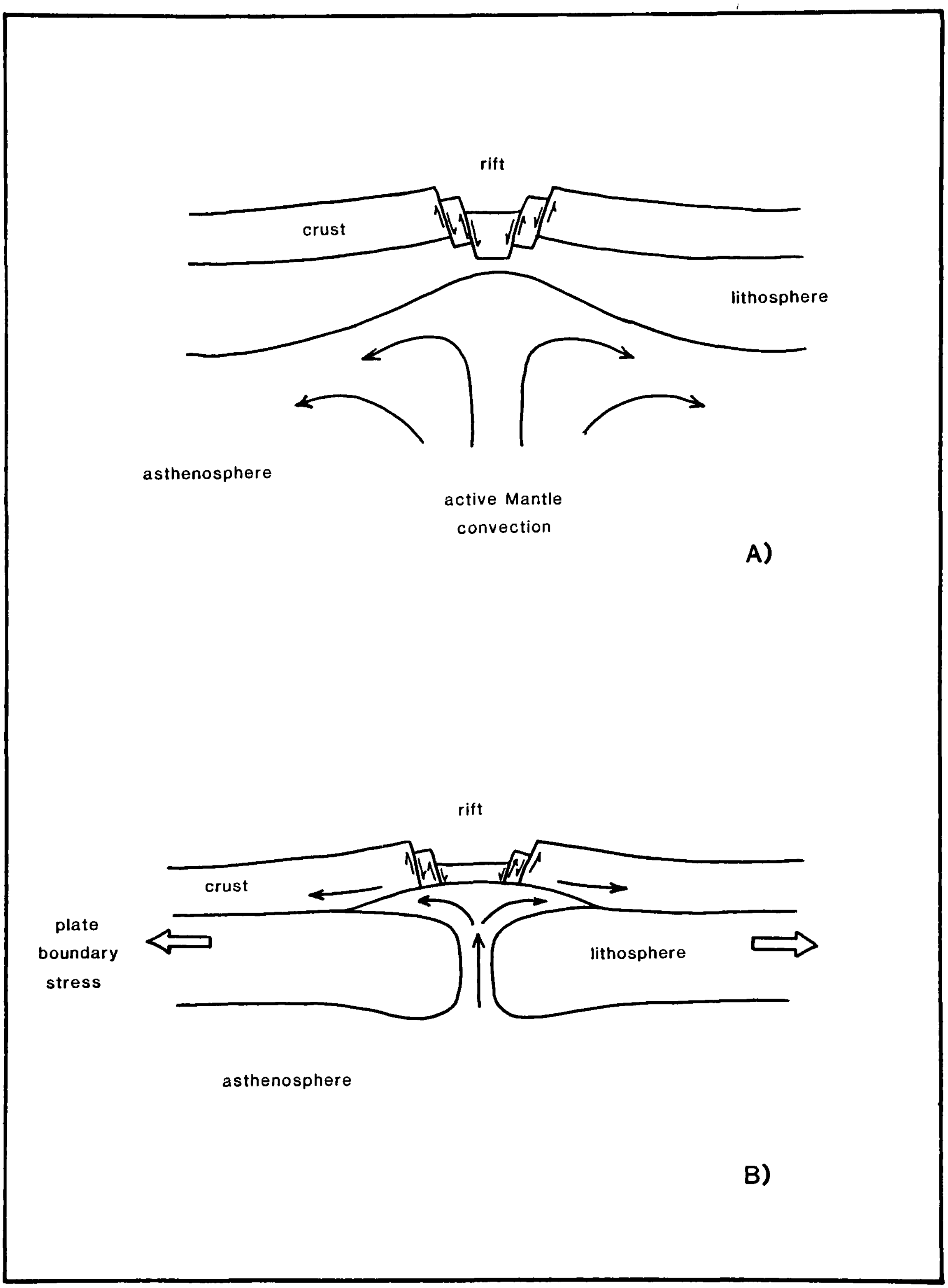

Figure 161. Two methods of formation of rifts: A) active Mantle convection and crustal doming, B) passive rifting due to plate boundary stress (from Turcotte, 1983).

crustal heating (causing domal uplift). Depending on which effect is dominant, the initial response may be broad subsidence followed by narrower rift subsidence, or may be broad doming followed by rift subsidence. It appears that the former condition is most common (Morgan and Golombek, 1984; Baker, 1986; Budnik, 1987; Watson et al., 1987; Clendennin et al., 1988).

Due to lateral heat flow, the subsidence rate is faster for narrow grabens than for wide basins, encouraging an increasingly rapid and deep subsidence, particularly in the first 10 Ma (Jarvis, 1984). Linear isostatic shoulder uplifts rim the grabens due to unloading and rotation of the footwall as the hanging wall subsides (Gawthorpe, 1987) and provide local sediment sources but also direct most of the runoff away from the graben (Frostick and Reid, 1987). Clendennin et al. (1988) found that the time lag between the onset of volcanism and the beginning of broad subsidence was about 10 Ma, and that between the onset of broad subsidence and narrow graben subsidence was about 10 Ma. Watson et al. (1987) noted a common 5-15 Ma time lag between onset of broad subsidence and the narrow graben subsidence. Post-graben thermal subsidence allows broadening of the basin again (McKenzie, 1978; Clendennin et al., 1988).

Recent seismic studies (Gibbs, 1984; Bosworth, 1987; Frostick and Reid, 1987; Rosendahl et al., 1986; Coletta et al., 1988) indicate that virtually all rifts are asymmetric in geometry (Fig. 162). Listric bounding faults merge at 12-18 km depth along a major mid-crustal decollement surface which separates upper brittle and lower ductile layers (Brun and Choukroune, 1983). In plan view the major bounding faults are curved and occur on one side only, although there may be smaller antithetic faults on the other side which give a surface form of symmetry to the basin (Gibbs, 1984; Frostick and Reid, 1987; Coletta et al., 1988; Ellis and McClay, 1988). Once initiated, block rotation subsidence on listric faults and compensating shoulder uplift offset the pre-graben shallow depression or domal uplift. Through time the displacement is localized onto fewer faults and the narrow graben development is actually a late-stage response (Mohr, 1982; Bott, 1981; Girdler, 1983; Morgan and Golombek, 1984; Clendennin et al., 1988).

Recent studies, primarily using seismic data, have also shown that most rifts are broken lengthwise into scoop-shaped segments (Fig. 163a) about 100 km long x 50 km wide (Rosendahl et al., 1986; Gibbs, 1984; Bosworth, 1987; Frostick and Reid, 1987; Leeder and Gawthorpe, 1987; Dunkleman et al., 1988; Graversen, 1988; Coletta et al., 1988).

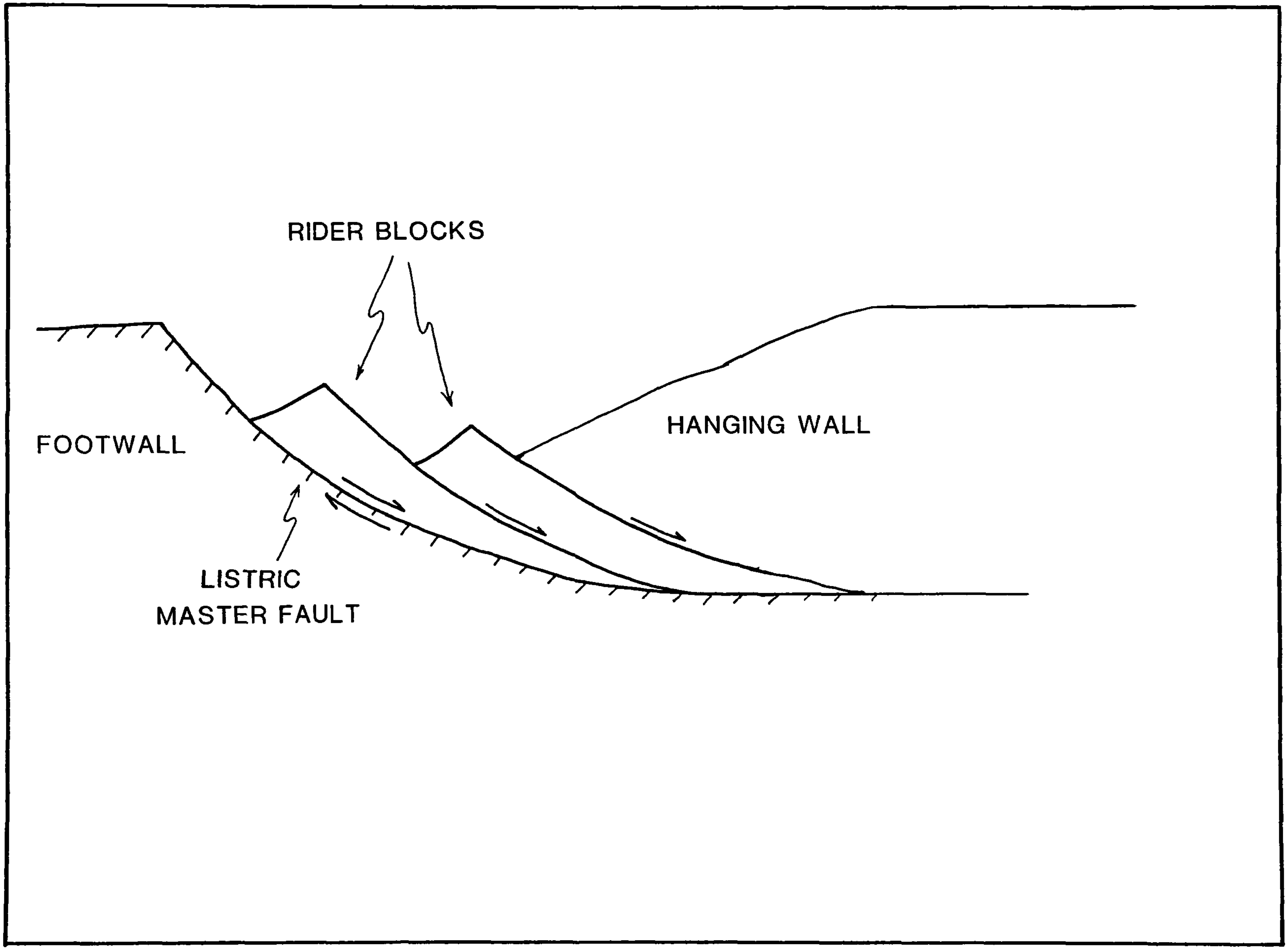

Figure 162. Structural asymmetry of extensional basins due to listric nature of main normal fault zone (modified from Gibbs, 1984).

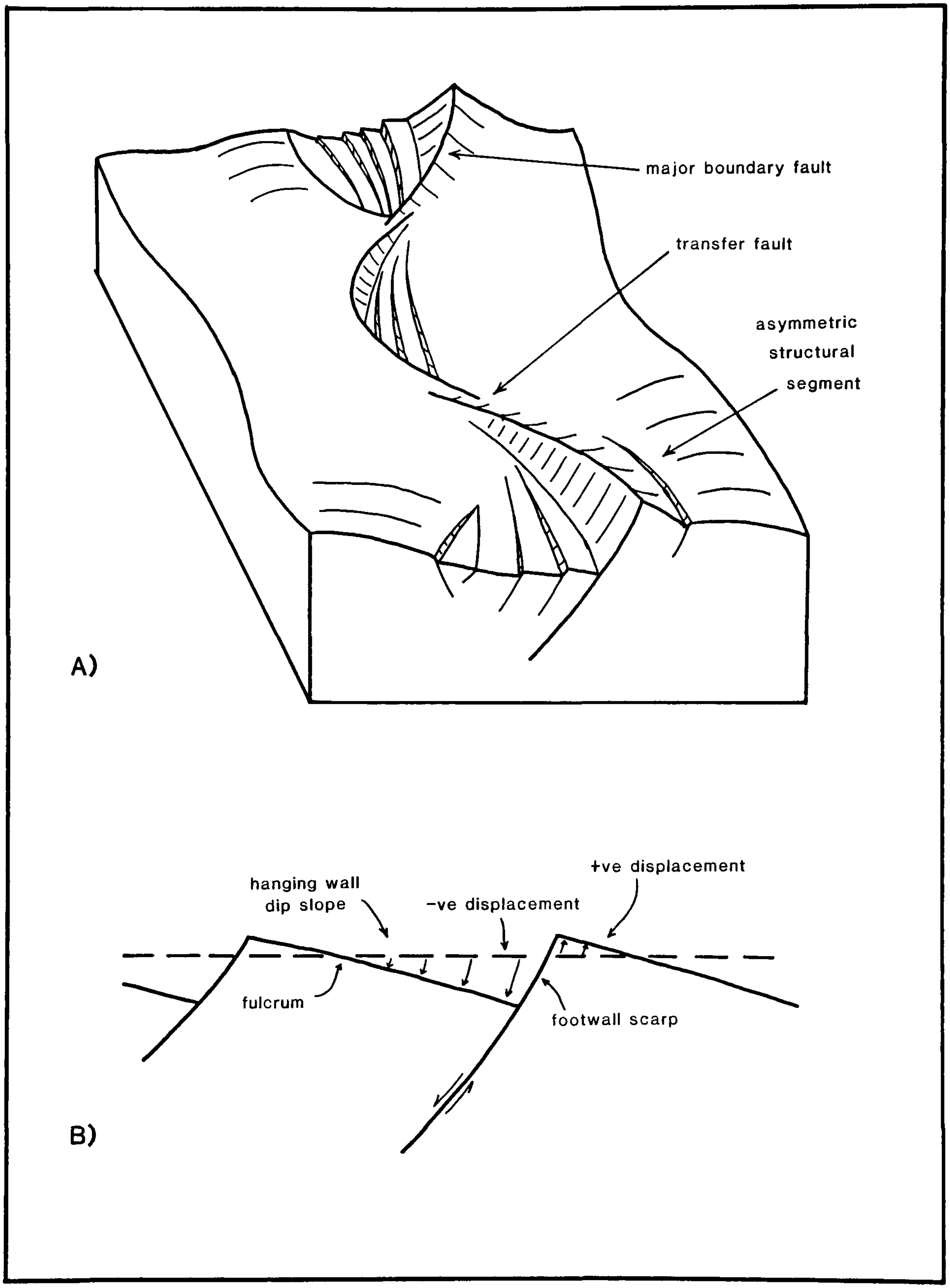

Figure 163. Terminology for asymmetric rift segments: A) asymmetric structural segments of a rift system with alternating polarity of asymmetry (from Frostick and Reid, 1987), B) tilt-block/half-graben terminology applied to rift segments (modified from Leeder and Gawthorpe, 1987).

These segments form the fundamental structural units of rifts and act as partially isolated sub-basins. Each segment comprises an asymmetric half-graben with the main arcuate listric bounding fault on one side and a hinged flexural (or faulted) hanging wall slope on the other (Rosendahl et al., 1986; Frostick and Reid, 1987) (Fig. 163b). The hinged hanging wall slope may be represented by a monoclinal ramp (tilt-block/half graben of Leeder and Gawthorpe, 1987) or by a series of stepped antithetic fault blocks (Crossley, 1984). These studies have also shown that half-graben segments typically alternate in polarity of asymmetry lengthwise along the rift. The segments are linked end-to-end across basement high accommodation zones or transfer faults which develop early in the history of subsidence (Illies, 1981; Gibbs, 1984; Rosendahl et al., 1986; Frostick and Reid, 1987; Barrett, 1988; Coletta et al., 1988) (Figs. 163a, 164). Adjacent segments may therefore have different rates and magnitudes of fault subsidence, although the entire system undergoes a generally similar structural history.

Distensive basins are also associated with post-orogenic periods (Bell et al., 1988); Jowett and Jarvis (1984) discussed the formation of these rifts after Cordilleran-type subduction orogenies. In this "foreland rift" model the extensional force under the continental foreland results from mantle material being drawn back toward the subhorizontal subducting plate by viscous drag or suction. This model has been proposed to explain the Rio Grande rift and Basin and Range extension in the western U.S. (Dickinson et al., 1988) and similar extensional basins in the Andes of Peru (Jordan and Allmendinger, 1986). Conversely, Coney (1987) suggested that post-orogenic extensional basins may simply be a result of gravitational instability and collapse of an overthickened crustal welt formed during compression.

<u>Basin-fill Characteristics</u> In distensive basins the overall basin-fill is large and linear (100's km long x 10's km wide x several km thick) and rests unconformably on basement rocks. Because virtually all tectonic motion is vertical, there is a vertical stacking of basinwide stratigraphic units which correspond to major tectonic phases (McKenzie, 1978). There should be some lengthwise consistency in the stratigraphic record, although segmentation of individual sub-basins introduces variation in detail and many sedimentary features are spatially dependant (Rosendahl et al., 1986; Leeder and Gawthrope, 1987). Intracontinental bimodal volcanics are commonly extruded in the early history, especially close to bounding faults. The overall basin-fill sequence typical of distensive basins can be summarized from

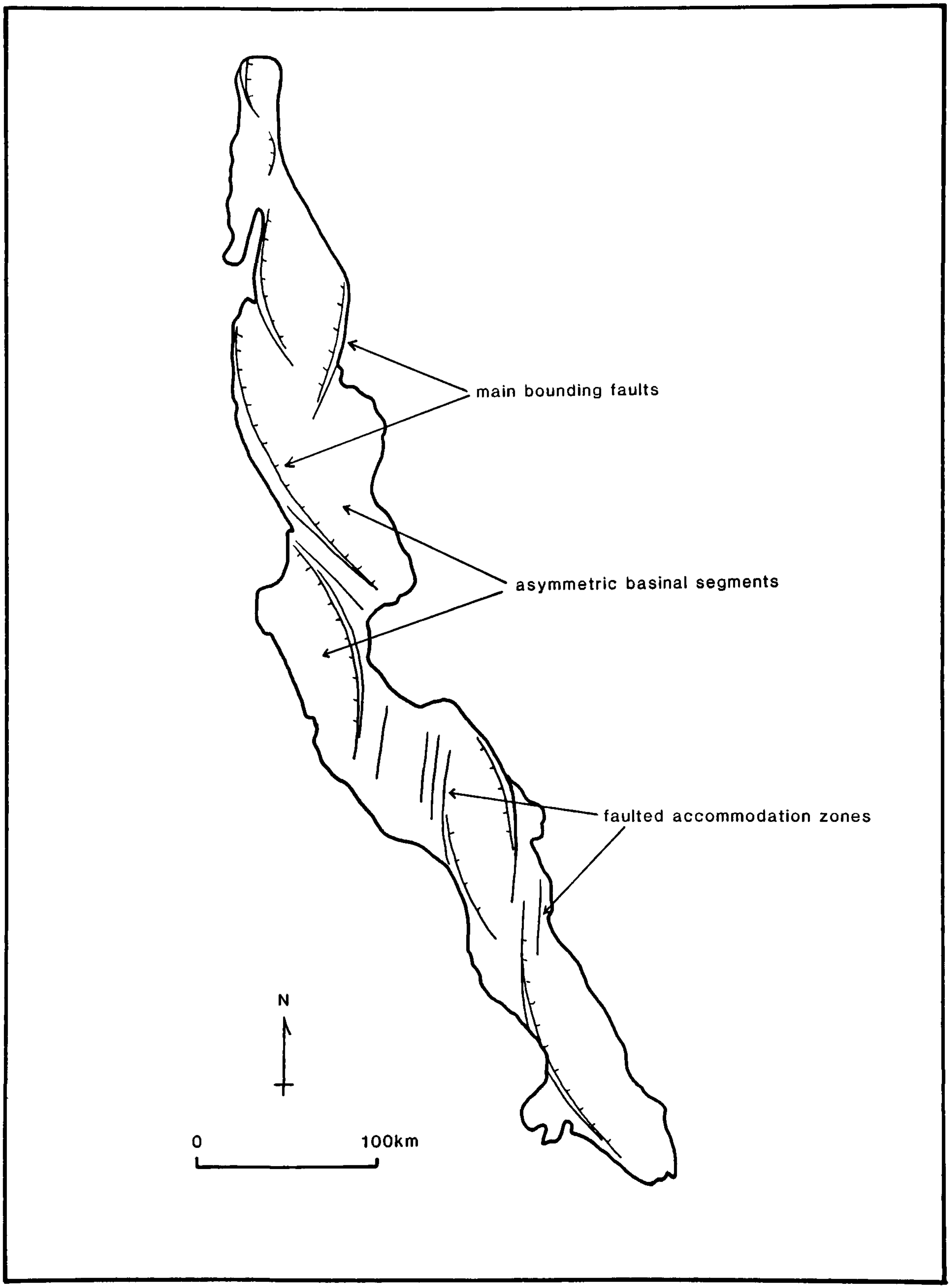

Figure 164. Asymmetric structural segments of Lake Tanganyika, African rift system (modified from Rosendahl et al, 1986).

several studies (Triassic of circum-Atlantic, Arguden and Rodolfo, 1986; Martini et al., 1986: Cretaceous/Tertiary of China, Ma et al., 1982; Watson et al., 1987) and is generally preserved in stable intracratonic areas with moderate deformation. The sequence comprises a) bimodal volcanics, b) coarse subaerial fluvial sediments in a broad sag basin, c) narrow graben (syn-rift) sequence of subaqueous followed by subaerial deposits, d) overlying (post-rift) sequence of thermal subsidence phase.

Within each segment or sub-basin the syn-rift sequence has a wedge shape, thickening toward the footwall scarp (Frostick and Reid, 1987). Each segment displays a marked basinwide asymmetry in facies and thickness (Van Houten, 1965; Ryder et al., 1976; Leeder and Gawthorpe, 1987) (Fig. 165) and the polarity of this asymmetry alternates along the length of the system (Crossley, 1984; Leeder and Gawthorpe, 1987). Depositional processes and facies arrangements in each segment are similar but may not be exactly the same or exactly synchronous (Rosendahl et al., 1986).

Lake Malawi provides a modern example of a distensive basin from the west African Rift (Crossley, 1984). Alluvial fans occur on the steep fault margin, whereas major floodplains and river deltas characterize the ramp slope margin; an asymmetry which alternates along the length of the lake in segments 100-200 km long x 50-75 km wide. The hanging wall slope is a monoclined ramp or a series of stepped fault blocks on which the dominant sediment input is spread over a wide area, and in places is trapped in small fault-bounded mini-basins. An arid seasonal climate gives flashy runoff, little vegetation, episodic high volumes of sediment input and an anoxic stratified lake. Strong southerly winds produce consistent longshore redistribution of sediment. The Triassic of the circum-Atlantic area provides an ancient example of continental sedimentation in arid distensive basins (Arguden and Rodolfo, 1986; Van Houten, 1965; Martini et al., 1986; Nadon and Middleton, 1984). A tripartite stratigraphy is present in most basins with moderate thicknesses an order of magnitude smaller than the basin size. The Newark Basin is 220 km long x 75 km wide and contains 7.5 km of sediment. The sequence is a) basal thick alluvial/fluvial conglomerate and sandstone, b) middle lacustrine shale, siltstone and sandstone, and c) upper reddened fluvial/alluvial sandstone, conglomerate and siltstone. Internal smaller-scale sequences are common and may relate to tectonic motions or climatic phases.

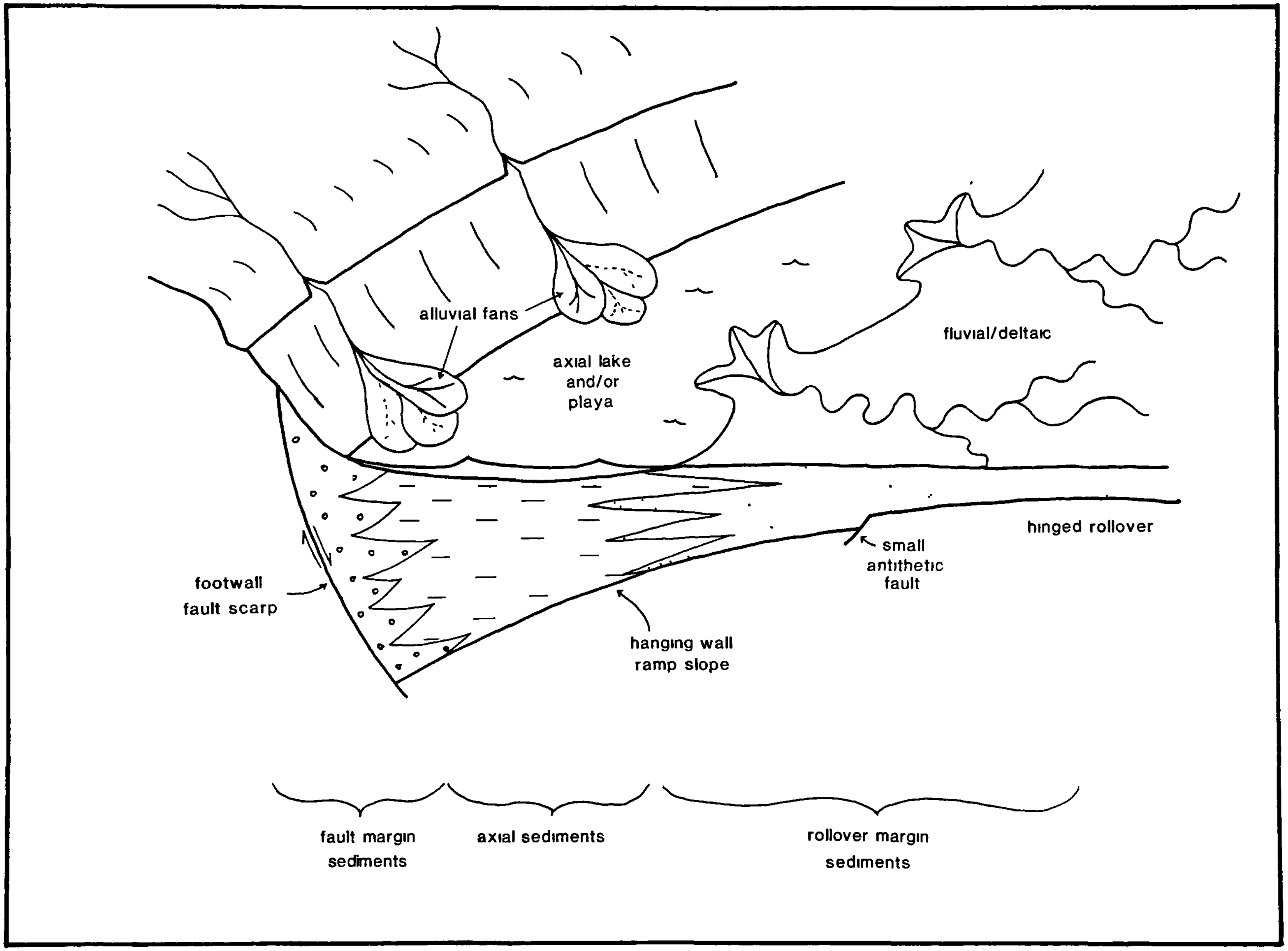

Figure 165. Asymmetry of depositional settings and facies distributions in a distensive rift segment (modified from Frostick and Reid, 1987; Leeder and Gawthorpe, 1987).

The main control on sedimentation in distensive basins is a complex interplay of sediment input, tectonic slopes and topography created by faulting and block rotation, and relative lake level affected by subsidence and climatic cycles. The fault-induced tectonic slopes and topography influence all gravity-driven sedimentary processes at basin margins and on basin floors (Leeder and Gawthorpe, 1987). After fault motion the elevated steep footwall scarp drains a very small area due to shoulder uplift and will initially produce small, thick alluvial fans with limited lateral extent (Rosendahl et al., 1986; Frostick and Reid, 1987; Leeder and Gawthorpe, 1987). That margin is characterized by narrow belts of coarse marginal facies, axial lacustrine sediments and abrupt vertical and lateral changes in facies and thickness (Gawthorpe, 1987) (Fig. 165). Over time, as erosion of that elevated scarp proceeds this side of the sub-basin provides a larger volume of sediment (Leeder and Gawthorpe, 1987; Blair and Bilodeau, 1988). The flexural, hinged hanging wall ramp drains a large area, the slope of which increases during active subsidence. This side of the sub-basin is characterized by extensive fluvial/deltaic systems which build into the shallow lacustrine setting (Rosendahl et al., 1986; Frostick and Reid, 1987), depositing broad belts of deltaic and shallow lacustrine facies with gradual changes in facies and thickness (Gawthorpe, 1987) (Fig. 165).

Each fault subsidence event has an instantaneous deepening effect (Blair, 1987) causing shoreline transgression at both the steep footwall and gentle hanging wall margins, and soft-sediment deformation (Leeder, 1987). Therefore, the lacustrine facies dominate only during periods of maximum subsidence which create topographically deep, closed sub-basins, and dominate especially near the steep footwall side (ie. axis) of the asymmetric half-graben segment (Blair, 1987; Leeder and Gawthorpe, 1987; Watson et al., 1987). As discussed by Blair and Bilodeau (1988), and contrary to some conventional wisdom, thick fine grained sediments denote the maximum phases of active subsidence in a fault-bounded basin. This concept is based on a) the disparity between rapid rate of response to tectonic subsidence and creation of topographic relief versus the slow rate of erosion and coarse sediment production (8 to 117 times slower) from high relief fault scarps, b) the instantaneous response of fine grained environments to tectonic subsidence versus the slow rate of response of coarse grained environments, and c) various observable modern examples (eg. Death Valley, Red Sea) which commonly have low gradient fluvial/lacustrine environments close to the active fault margin where subsidence is focussed, and very restricted coarse grained fans and braidplains (Blair and Bilodeau, 1988). Therefore supply of coarse sediment always lags behind fault-induced subsidence. But as subsidence decreases, scarp erosion and sediment supply catch up, coarse marginal facies prograde from the sides, and a general coarsening-upward sequence develops (Blair, 1987; Watson et al., 1987; Blair and Bilodeau, 1988).

EXTENSIONAL STYLE DURING HORTON DEPOSITION

The foregoing discussion of the two end-member styles of fault-bounded extensional basins allows comparison and interpretation of the characteristics of the Horton Group. Extensional basins in the overall compressional setting of an oblique-slip system are a special case and certain features must be demonstrated to justify use of this interpretation. Transtensive basins are generally small and poorly preserved due to later deformation which is commonly extreme. Most importantly, they are typified by great apparent stratigraphic thicknesses of the same order of magnitude as the basin size, and 2-5 times the actual basin depth. This is possible because the sediment bodies are stacked en echelon lengthwise along the basin due to the dominantly lengthwise horizontal component of structural motion during extension. As this motion represents the effective definition of transtension, the accompanying sedimentary style is the key to justifying such an interpretation. Transtensive basins have a certain set of characteristics, as reviewed above, which are not demonstrated by the Horton Group, and this tectonic style is rejected as a fruitful concept for interpretation of the Horton.

Conversely, the Horton Group of Cape Breton Island displays many of the important characteristics associated with distensive systems. The entire depositional area of the Horton is large (over 800 km long from New Brunswick to Newfoundland and 200 km wide, with minor breaks), and the sediments have suffered only moderate deformation over 350 Ma of subsequent history. There is no identifiable strike-slip component of syn-sedimentary fault motion in Horton deposits. On Cape Breton Island there is distinct structural separation into two discrete, but related, sub-basins about 100 km long and 50 km wide, for at least part of Horton time. These are linked lengthwise but the elevated basement zone between them is only 10-15 km wide, much narrower than the sub-basins themselves. Facies distribution, the presence of outliers, and paleocurrent data imply that many basement blocks now exposed at surface were not positive features in Horton time and were later elevated as fault-bounded blocks. The impression that deposition occurred in many small sub-basins is erroneous. The stratigraphic units of the Horton Group are laterally persistent, approximately synchronous and stacked vertically throughout Cape Breton Island (and, in fact, throughout the Maritimes Basin). There is no evidence of systematic lateral translation of sediment sources, and facies distributions and paleocurrent data are easily related to vertical subsidence phases and stationary sediment sources. The Horton Group of Cape Breton Island has a small total stratigraphic thickness (<3 km), an order of magnitude less than the general sub-basin dimensions. The defined sub-basins appear to have had a long-term asymmetric structural development and consequent asymmetric facies distribution consistent with the half-graben geometry typical of distensive rift segments. Within the sub-basins, the locations of steep footwall scarp and hanging wall ramp remained constant throughout the period of fault-induced subsidence.

The Horton Group of Cape Breton Island is therefore interpreted to represent deposition in part of a distensive or rift graben tectonic system. It encompasses two adjacent segments, each represented by a sub-basin with an asymmetric half-graben geometry. The interpretation of the facies assemblages and depositional systems discussed in Chapter 4 are examined in this context in more detail below.

THE OVERALL HORTON TECTONIC SUCCESSION

The overall succession characteristic of the Horton Group in the study area is best illustrated in the western Cape Breton sub-basin, where there are abundant data and fewer structural complications (Fig. 166). This succession can be related to the general structural evolution and basin-fill history typical of distensive basins as reviewed above.

Post-Acadian activity began with Late Devonian (about 370-375 Ma) extrusion of Fisset Brook bimodal volcanics and associated nonmarine sediments in isolated areas, indicating volcanic centres were not abundant. The ?Late Devonian/Early Carboniferous? C3 depositional system of unoxidized braidplain deposits with scattered lakes, low topography and a high water table was deposited in a single, extensive, unfaulted basin which slowly subsided (Fig. 166-I). It was broad and shallow with low margins beyond the present study area. The overall paleoslope of this broad sag basin may have dipped to the northeast. Basins of this nature are a common initial response to crustal thinning and passive Mantle upwelling induced by active tensional stress at remote plate boundaries, as

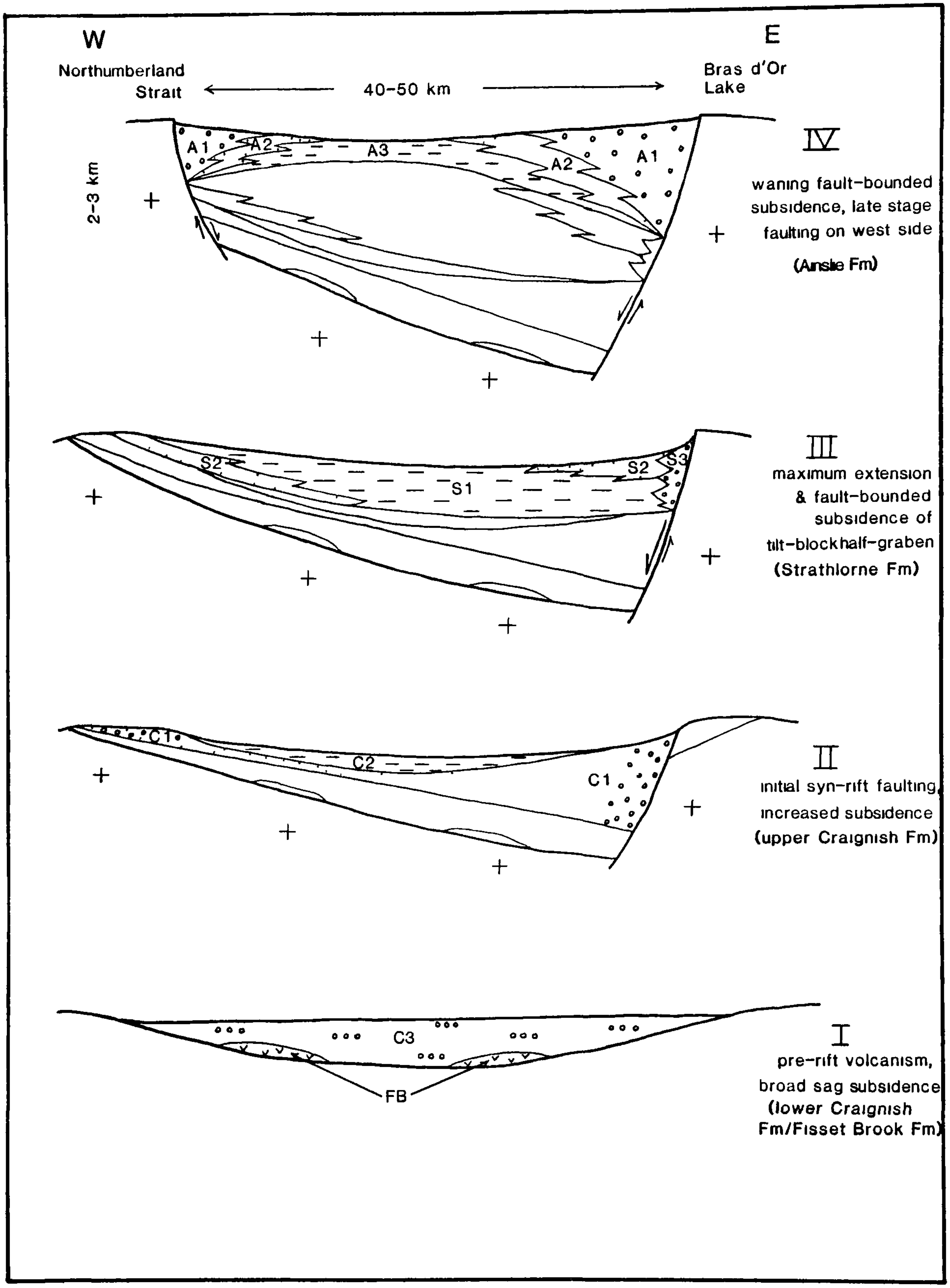

Figure 166. Four-phase tectonic evolution of Horton depositional systems, Western Cape Breton sub-basin.

reviewed above (Turcotte, 1983; Morgan and Golombek, 1984, and others). The Acadian metamorphosed basement had probably been thickened during collisional orogenesis in the mid to late Devonian and post-orogenic extension may have also been the result of mantle suction (Jarvis, 1984; Dickinson et al., 1988) or of gravitational collapse processes (Coney, 1987).

The broad shallow sag phase was succeeded by the Early Carboniferous (Tn_1?/Tn_2) C1/C2 depositional system of reddened braidplain sediments near the fault-bounded margin and mudflat/playa sediments in the sub-basin centre (Fig. 166-II). Subaerial exposure in a warm arid climate was pervasive. Facies distributions, thickness variation and paleocurrent data indicate the emergence of two closed, fault-bounded sub-basins with internal drainage, and dimensions somewhat narrower than those of the C3 depositional basin. In the western Cape Breton sub-basin there is a general fining-upward trend, indicating expansion of the mudflat setting through time, and finally a shoreline transitional sequence upward into the Strathlorne lacustrine depositional system. Following the ideas of Blair and Bilodeau (1988), the C1/C2 depositional system is interpreted as a response to the initial faulting near the southeast margin of the broad sag basin and conversion to a narrower rift system with increasingly rapid and deep fault-bounded subsidence. Over a period of time the coarse marginal facies were confined to an increasingly narrow belt adjacent to the fault scarp. The presence of transitional sequences into the Strathlorne, attributed to the most active phase of tectonic subsidence during Horton time, further supports this view. This tectonic transition, from pre-rift sag to syn-rift tilt-block/half-graben, was relatively gradual, occurring over several hundred metres of stratigraphic thickness, perhaps as fault motion became localized on fewer fault traces. This appears to follow an evolutionary path common in many grabens as noted by Morgan and Golombek (1984), Baker (1986) and Clendennin et al. (1988). The time lags between a) onset of Fisset Brook volcanism and C3 broad subsidence, and b) C3 broad subsidence and C1/C2 fault-bounded subsidence total less than 20 Ma and so are similar to those noted by Watson et al. (1987) and Clendennin et al. (1988).

The C1/C2 depositional system is succeeded, apparently conformably, by the late Tn_2-early Tn_3 S1/S2/S3/S4 (Strathlorne) lacustrine depositional system (Fig. 166-III). The proximal fault-bounded footwall margin was dominated by fan deltas and proximal braidplains with some mudflats. Lacustrine shorelines on the hanging wall ramp were extensive, sandy, prograding delta and shoreface zones, with some interdeltaic carbonate flats. The open lacustrine setting was dominated by low energy deposition with episodic density underflows. The general climatic conditions remained warm and arid. Facies distributions, thickness variation and paleocurrent data indicate that depositional environments were arranged with asymmetric distribution, suggesting two discrete fault-bounded half-graben sub-basins. These are interpreted to represent two adjacent structural segments of a rift system with nearly opposite polarities of asymmetry, discussed below. The presence of coarse grained facies indicative of fault-bounded margins, the dominance of fine grained deposits through the bulk of the Strathlorne Formation (but concentrated near the same margins), and pervasive sub-basin asymmetry are interpreted to suggest maximum fault-bounded subsidence of the distensive graben system during Strathlorne deposition. As discussed by Blair (1987) and Blair and Bilodeau (1988) the onset of fine-grained deposition in a fault-bounded tectonic basin is indicative of renewed tectonic subsidence, contrary to the misconception that coarse-grained sedimentation correlates with uplift. With increased rate of subsidence the coarse sediment supply could no longer keep pace with sub-basin floor deepening and a large standing body of water formed along the axis of subsidence. Restricted runoff at the steep footwall scarp supplied localized alluvial fan deltas but abundant sediment supply from larger watersheds on the hanging wall side built large deltas which prograded and retrograded according to the balance between sediment supply, lake level and basin subsidence. In later Strathlorne time there was mild reversal of the facies asymmetry attributed to initiation of fault subsidence at the hinged margin of the hanging wall, perhaps on smaller antithetic faults (Gibbs, 1984; Frostick and Reid, 1987), or small stepped fault blocks (Crossley, 1984).

The Strathlorne depositional sequence is succeeded conformably by the late Tn_2 - early Tn_3 A1/A2/A3 (Ainslie) depositional system (Fig. 166-IV). A range of subaerial facies includes reddened proximal alluvial fan deposits through medial braided or meandering fluvial to central unoxidized meandering fluvial and low-lying floodplain sediments. Sediment transport in the two former environments was transverse to sub-basin axes, whereas that in the latter may have been parallel to the axes. Deposition continued in two adjacent, closed fault-bounded sub-basins. The general climatic conditions remained warm and arid. Facies distributions, thickness variation and paleocurrent data indicate the depositional environments were asymmetric, suggesting two fault-bounded half-graben sub-basins, identical to those which were present in Strathlorne time. They represent adjacent structural segments of a rift system. The polarity of asymmetry was consistent throughout Horton deposition in western Cape Breton, and although the Ainslie is poorly preserved in the northern sub-basin, the same was probably true there. There is a fairly abrupt transition from Strathlorne lacustrine to Ainslie fluvial/alluvial sedimentation in an overall coarsening-upward sequence. This is interpreted to indicate that the Strathlorne period of maximum, active fault-related subsidence gradually decreased. Scarp erosion and sediment supply began to catch up with subsidence, and finally, completely overwhelmed the lacustrine conditions, as suggested by Blair (1987) and Watson et al. (1987). According to Blair and Bilodeau (1988), "only when subsidence rate greatly lessens or ceases can the rate of denudation in the source area exceed subsidence to produce a widely dispersed progradational clastic wedge". The A1/A2/A3 depositional system represents a phase of waning, but still present, fault-bounded subsidence in the distensive graben system, and essentially complete filling of the sub-basins.

There may or may not have been a significant hiatus between Ainslie deposition and that of the Macumber Formation of the Windsor Group on Cape Breton Island. Field observations in this study and known distribution indicate that the basin(s) in which the Macumber was deposited was wider than that of the Ainslie Formation, and it onlapped the basement outside the confines of the Horton depositional sub-basins. A rapid marine transgression occurred at that time, but the presence of the Grantmire facies near Windsor basin margins confirms that they were still fault-bounded. A situation like the Danikil Depression, a rift which connected to the ocean, may be indicated (Reading, 1982). Marine invasion can be instantaneous after fault-induced subsidence (Blair and Bilodeau, 1988) and this may account for the rapid basal Windsor transgression over much of the Maritimes Basin, and deposition of the dark grey, laminated, restricted marine Macumber Formation. These inferences suggest that the entire post-Acadian/pre-Alleghenian stratigraphy may represent a rift cycle 40-50 Ma long composed of pre-rift (Fisset Brook, lower Craignish), syn-rift (upper Craignish, Strathlorne, Ainslie, much of Windsor), and post-rift (upper Windsor, Canso) stages. Within that period there were several phases of very active fault-bounded subsidence indicated by thick fine grained sequences (Strathlorne, Macumber). Throughout Horton deposition the area was positioned at 10-15° S paleolatitude and climatic indicators record a warm arid climate.

SUB-BASIN SEGMENTATION AND ASYMMETRY

During all Horton Group sedimentation on Cape Breton Island, except for the C3 depositional system, there is evidence that deposition took place in two adjacent fault-bounded sub-basins (Fig. 167). These are interpreted as individual structural segments of a larger distensive rift system. The western Cape Breton sub-basin extends northeast-southwest from Srait of Canso to Cheticamp, and in width from Bras d'Or Lake to Northumberland Strait. The northern Cape Breton sub-basin, only a small portion of which is exposed on land, extends lengthwise from Pleasant Bay northward into Cabot Strait, and in width from Aspy River northwestward into Gulf of St. Lawrence. These sub-basins have the general dimensions, structural asymmetry and stratigraphic asymmetry of the fundamental structural units of all well-studied distensive rifts (Gibbs, 1984; Rosendahl et al., 1986; Frostick and Reid, 1987; Barrett, 1988). The zone of exposed Precambrian basement between Cheticamp and Pleasant Bay which separates them is attributed to the accommodation zone which typically occurs between rift segments (Illies, 1981; Gibbs, 1984; Frostick and Reid, 1987; Barrett, 1988). This area might be one of structural complexity in basement rocks (Red River mylonite zone? (Barr and Raeside, 1986). In most rifts, adjacent segments have subtle differences in structural and sedimentational history but generally record similar overall evolution. In the Horton Group the same facies assemblages and depositional systems occur in both sub-basins, in the same general sequence. Both sub-basins appear to have been asymmetric half-grabens, with consequent asymmetry of facies distribution, through much of Horton time. They record an early fluvial-dominated depositional system (moderate fault-bounded subsidence) passing upward into a middle lacustrine-dominated depositional system (maximum fault-bounded subsidence), capped by a later fluvial-dominated system (moderate fault-bounded subsidence). From palynological evidence, the timing of most of Strathlorne and Ainslie deposition was the same in both segments (early to middle Tn_3). Several late Tn_2 to early Tn_3 dates for lower Strathlorne sediments in northern Cape Breton suggest that lacustrine conditions, and therefore maximum subsidence, may have begun somewhat earlier there.

The primary difference between the two sub-basins is in the apparent polarity of asymmetry indicated by the sedimentary characteristics. Asymmetry of facies distribution, facies thickness and paleocurrent orientation suggest the polarity of asymmetry of the two half-grabens was opposed (Fig. 167).

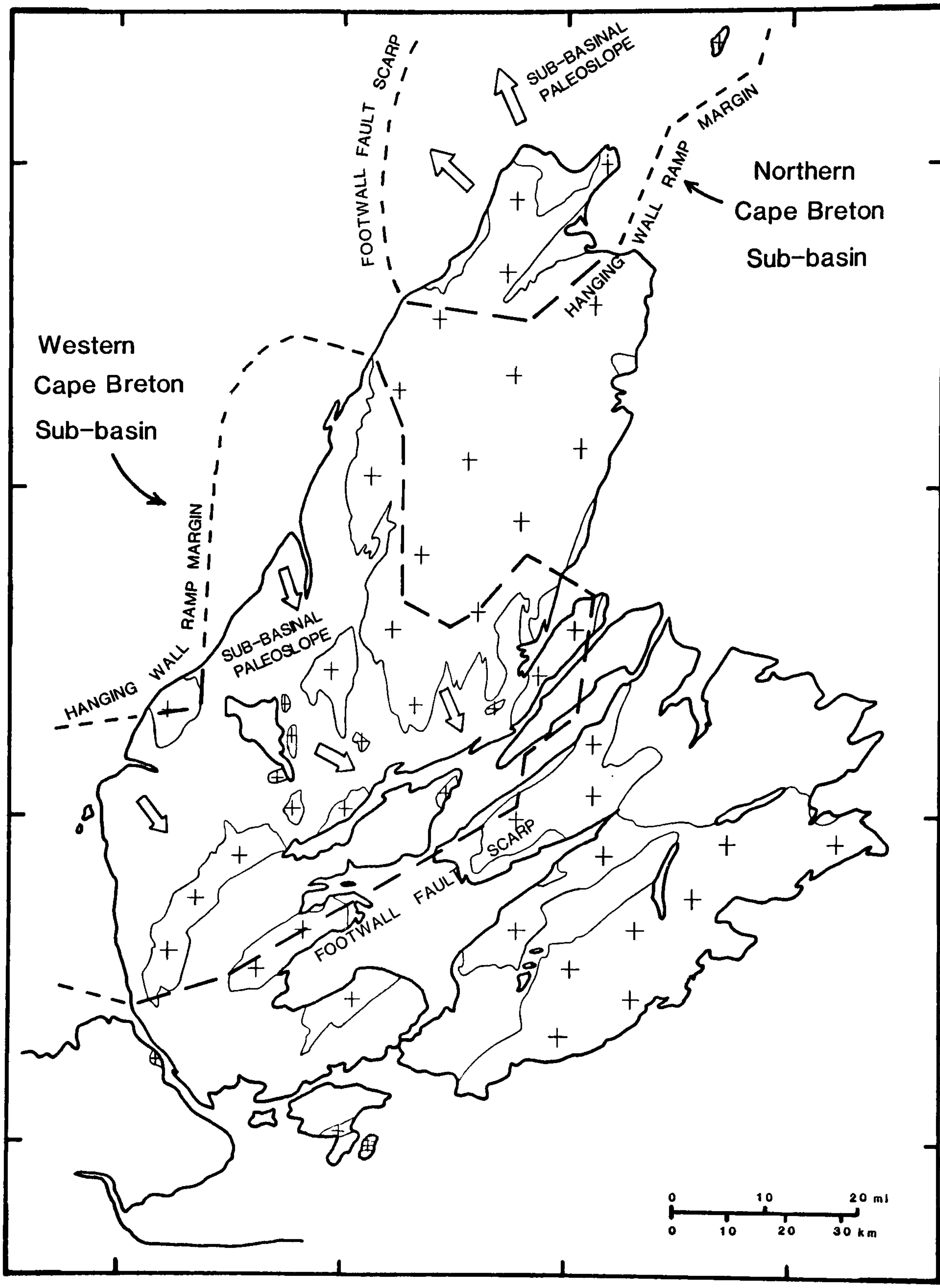

Figure 167. Limits and structural asymmetry of Western and Northern sub-basins of Horton deposition.

In the western Cape Breton segment all data from the Strathlorne and Ainslie Formations suggest that the eastern margin (Baddeck, Bras d'Or Lake, Whycocomagh areas) was the steep footwall scarp margin and the locus of sub-basin subsidence. In particular, a) C1 sediments were thickest near and mostly derived from that margin (Figs. 31, 22), b) S1 open lacustrine density underflows flowed south eastward toward a deep axis in that vicinity (Fig. 59), c) S3 fan-delta sediments were concentrated on that margin (Fig. 85) and d) A1 and A2 coarse alluvial/fluvial sediments were thickest near and derived from that margin (Figs. 104, 106, 120, 122). Conversely, all data suggest that the western margin (Mabou, Northumberland Strait areas) was the hanging wall ramp where hinging and less active subsidence took place, In particular, a) S2 delta and shoreline sediments were most extensively developed on and derived from that margin (Figs. 75, 77), and, b) A1 and A2 alluvial/fluvial sediments were thinner and less extensive on that margin (Figs. 104, 120).

In northern Cape Breton only a small portion of the southern part of the segment is exposed on land but the evidence suggests that the southeastern margin (Aspy River area) may represent the hanging wall side where hinging and subsidence occurred on a series of small stepped fault blocks. In particular, S2 delta and shoreline sediments were very extensive on and derived from that margin (Figs. 75, 77). The presence of thick S3 sediments indicates that faulting was moderately active and some paleocurrent data suggest faulting may have created a mini-basin in the Aspy area with some sediment shed back toward the sub-basin margin (as discussed by Crossley, 1984; and Ellis and McClay, 1988). Barr et al. (1987) confirm that the Aspy Fault was active during Carboniferous times, but is not a major tectonic boundary. Conversely, some data suggest the unexposed northwestern margin (offshore beneath Cabot Strait) may represent the steep footwall scarp margin, and the locus of sub-basin subsidence. In particular, a) C1 sediments were thickest on and derived from that margin (Figs. 31, 33) and b) S1 open lacustrine density underflows flowed toward a deep axis in that vicinity (Fig. 59), offshore to the north.

These lines of evidence strongly suggest that the two adjacent half-grabens had nearly opposite polarities of asymmetry. There was a basinal paleoslope dip to the east or southeast in the western Cape Breton sub-basin, and to the north or northwest in the northern Cape Breton sub-basin (Fig. 167). This arrangement resembles that recorded in all well-studied distensive rifts (Crossley, 1984; Gibbs, 1984;

Rosendahl et al., 1986; Bosworth, 1987; Leeder and Gawthorpe, 1987; Frostick and Reid, 1987; Dunkelman et al., 1988; Graversen, 1988; Barrett, 1988).

THE STRATHLORNE DEPOSITIONAL SYSTEM AND TECTONIC CONTROL

The S1/S2/S3/S4 depositional system is interpreted as the main phase of most active fault-bounded subsidence during Horton time and so deserves particular attention. Its characteristics can provide further keys to understanding the detailed interaction between controlling tectonic fault motions and the depositional response. The discussion below attempts to illustrate these relationships.

A prime characteristic of the S1 facies assemblage is the presence of stacked shallowing-upward sequences, passing from open lacustrine offshore mudstone to various nearshore or shoreline sediments over an interval of 2-45 m. These progradational phases were recurrent, each terminating with an abrupt return to open lacustrine deposition of mud. They can occur in bundles of sequences 20-50 m thick. In both sub-basins the sequences are less numerous, thicker and sandier on the hinged hanging wall ramp, and more numerous, thinner and finer grained with abundant limestone near the steep footwall scarp along the sub-basin axis (Figs. 150, 151). Most paleocurrent data indicate that sediment transport in S1 and S2 facies assemblages was toward these sub-basin axes.

The prevalence of shallowing-upward sequences in obviously fault-bounded sub-basins during a period of maximum subsidence suggests that the sequences were predominantly controlled by phases of tectonic subsidence on master faults at the footwall scarp. During a subsidence event there was instantaneous deepening of the lake floor, especially near the basin axis (Blair and Bilodeau, 1988), approximately equivalent to the throw on the controlling fault scarp. Mud began to accumulate slowly below wave base and, if tectonic conditions remained relatively stable, a shallowing-upward sequence resulted. Small throws might not have significantly shifted deposition below wave base on the hanging wall side of the basin. Larger throws or periods of several related subsidence events could collectively create bundles of sequences near the footwall scarp which would correlate with a single thick coarsening-upward sequence at the hanging wall margin. The open lacustrine (S1) part of the sequence is commonly thicker, and generally represents a longer time period, than the shoreline portion (S1 capping beds and S2). This imparts a vertical asymmetry to the sequences (Fig. 150). Contorted lamination is a common accessory sedimentary structure near the tops of sequences. These observations reinforce the interpretation that the sequences represent relatively low energy, slow fill cycles terminated by abrupt deepening, which were probably due to fault-induced subsidence.

No great depth need be implied for any Strathlorne deposits. The thickness of the original uncompacted shallowing-upward sequence could approximate the depth of water at the beginning of sequence deposition, for those sequences which filled to subaerial exposure. Not all sequences achieved exposure before subsidence but many apparently did, especially near sub-basin margins. This implies that water depth was never greater than perhaps 100 m even at sub-basin axes, and commonly was much shallower. However, even at these moderate depths in a lacustrine setting, much of the sub-basin floor could remain below the influence of waves and circulation, allowing water stratification, stagnation and anoxia. Many Strathlorne sequences, especially near sub-basin axes and footwall scarps, have laminated, organic-rich mudstone with no bioturbation near the base, attributed to deposition in the anoxic hypolimnion of a stratified lake. Near hanging wall slope margins, where deposition occurred in shallower water, these features are less common.

In the Triassic Lockatong Formation of New Jersey Van Houten (1965) described similar shallowing-upward sequences and attributed them to cyclic climatic variation and consequent expansion and contraction of the host lake. Expansion in humid climatic phases would create open circulating conditions and deposition of "detrital" mudstone-sandstone sequences averaging 5 m thick. Contraction in arid climatic phases would create closed stratified conditions and deposition of "chemical" calcareous mudstone-limestone sequences averaging 3 m thick. Counts of laminae assumed to be varves suggested an average of 21 000 years/cycle, attributed to precession of the Earth's axis. The evidence collected in this study was not geared to specifically investigating climatic variation but there is no indication of significant climatic variations through time, of either short- or long-term durations. In contrast, most sedimentary characteristics can be interpreted in terms of tectonic, rather than climatic, control, including the shallowing-upward lacustrine sequences. No attempt was made to identify true annual varves or count their occurrence as a temporal guide.

If tectonic subsidence events controlled the deposition of the shallowing-upward sequences, they should be correlative over fairly large parts of the sub-basins, although the deepening effect of any one relatively minor fault movement might not be recognizable over an extended area. Whereas sequences near the footwall scarp and sub-basinal axis probably related to individual fault-induced subsidence events, they tend to occur in discrete bundles of sequences which imply episodes of related subsidence events. In contrast, those on the hinged hanging wall ramp are thicker and fewer in number and probably record only one identifiable sequence for each bundle of sequences on the opposite side (see Fig. 151). In some sequences the S2 facies assemblage cap has a channelized fluvial aspect indicating rapid, river-dominated delta progradation and preservation by subsequent subsidence. Others are capped by shoreface facies indicating less river domination of progradation, or a lengthy period of sediment redistribution at the delta front before the next subsidence event.

Throughout the Strathlorne Formation a hierarchy of filling sequences is visualized (Fig. 168). These relate to a) water body fill after individual fault movements, creating local sequences several to several tens of metres thick, b) partial sub-basin- fill during episodes of related fault movements creating bundles of sequences tens of metres thick, and c) overall complex sub-basin-fill during and after the maximum subsidence phase, creating several hundred metres of Strathlorne lacustrine depositional sequence passing upward into the Ainslie alluvial/fluvial depositional system.

In the A1/A2/A3 depositional system the proximal A1 facies assemblage was the most influenced by fault-related motions and displays large-scale coarsening-upward progradational sequences up to 100 m thick. These are best interpeted as a direct response to fault-related relief at the footwall scarp, and subsequent fan progradation during a phase of relative tectonic quiescence. In effect, these sequences may be similar in origin to the bundles of sequences in the Strathlorne and relate to discrete episodes of tectonic motions at the fault margins during the Ainslie phase of moderate fault-related subsidence.

THE HORTON GROUP IN THE APPALACHIAN CONTEXT

The Horton Group of Cape Breton Island is part of the thick post-Acadian /pre-Alleghenian Kaskaskia sequence (Sloss, 1963) of the Appalachian Orogen. It represents deposition in several asymmetric fault-bounded half-graben segments of an overall distensive rift system positioned at 10-15° S paleolatitude in a warm arid climate. This extensional system was similar to the "Fundy Basin Rift" of Belt (1968a,b) (Fig. 169a) and the "post-orogenic relaxation phase" referred to by Howie and Barss (1975). It was certainly post-orogenic in timing and may have been related to the mega-tectonic processes espoused by

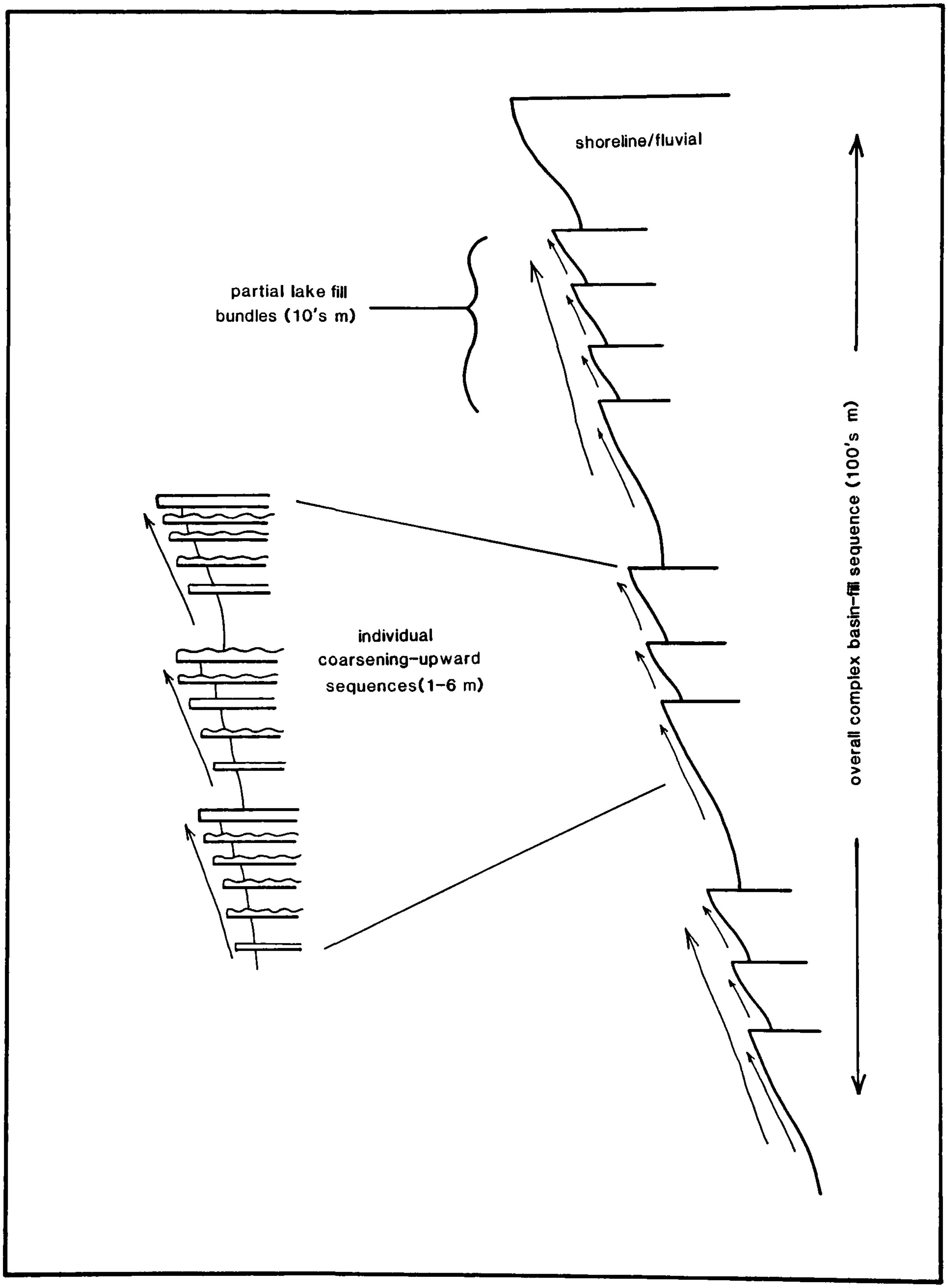

Figure 168. Hierarchy of lake-fill sequences visualized for Strathlorne Formation.

Jowett and Jarvis (1984) or Coney (1987) for "foreland rifts". This distensive period between two compressional orogenic periods must represent a relatively short-lived re-orientation of stress regime and plate motions of the type referred to by Ziegler (1984) as "a fundamental change in mega-tectonic setting due to plate reorganization". The Horton Group distensive system could be regarded as a "failed rift" (St. Peter, 1987) because it did not proceed to the stage of oceanic crust formation and was the result of plate boundary stresses and passive mantle upwelling, as described by Turcotte (1983) and others. The Horton Basin is similar to the West African Rift, characterized by extension and thinning of a thick lithosphere, dominated by subsidence and sedimentation with few volcanics (Fairhead, 1986; Dunkelman et al., 1988). It is less similar to the East African Rift, characterized by active mantle upwelling, domal uplifts, abundant volcanics and little sedimentation (Fairhead, 1986; Dunkelman et al., 1988).

Recent sedimentological and structural studies of the Horton Group (and correlatives) in other sub-basins of the Fundy Basin Rift have revealed interesting parallels. From surface and subsurface data in the Moncton sub-basin Foley and Noble (1988) have shown that lacustrine shale and siltstone of the Albert Formation (Strathlorne correlative) thicken to the south and paleocurrents indicate sediment dispersal to the south and southwest. An asymmetry of facies distribution is implied and these data may indicate a sub-basin with an axis of deepest subsidence to the south or southwest.

In the Windsor sub-basin Martel and Gibling (1988) suggest thickening of the Horton Bluff Formation (Strathlorne and Ainslie correlatives) toward an axis of greatest subsidence to the north or northeast. This polarity of asymmetry is opposed to that of the adjacent Moncton sub-basin of this study on the east. Adjacent to the Northern Cape Breton sub-basin of this study, Kilfoil (1988) has used gravity, seismic and magnetic data to interpret a half-graben structure for the Bay St. George sub-basin with the axis of deepest subsidence on the southeast side. That polarity of asymmetry is directly opposed to that defined in this study for the Northern Cape Breton sub-basin.

The foregoing discussion suggests that at least five adjacent half-graben structural segments of the linear Fundy Basin Rift can be defined, each with a polarity of asymmetry nearly opposite to those of its neighbours (Fig. 169b). The Western and Northern Cape Breton sub-basins studied in this project represent only two of these segments. In this way the Fundy Basin Rift appears to display fundamental characteristics similar to other well-

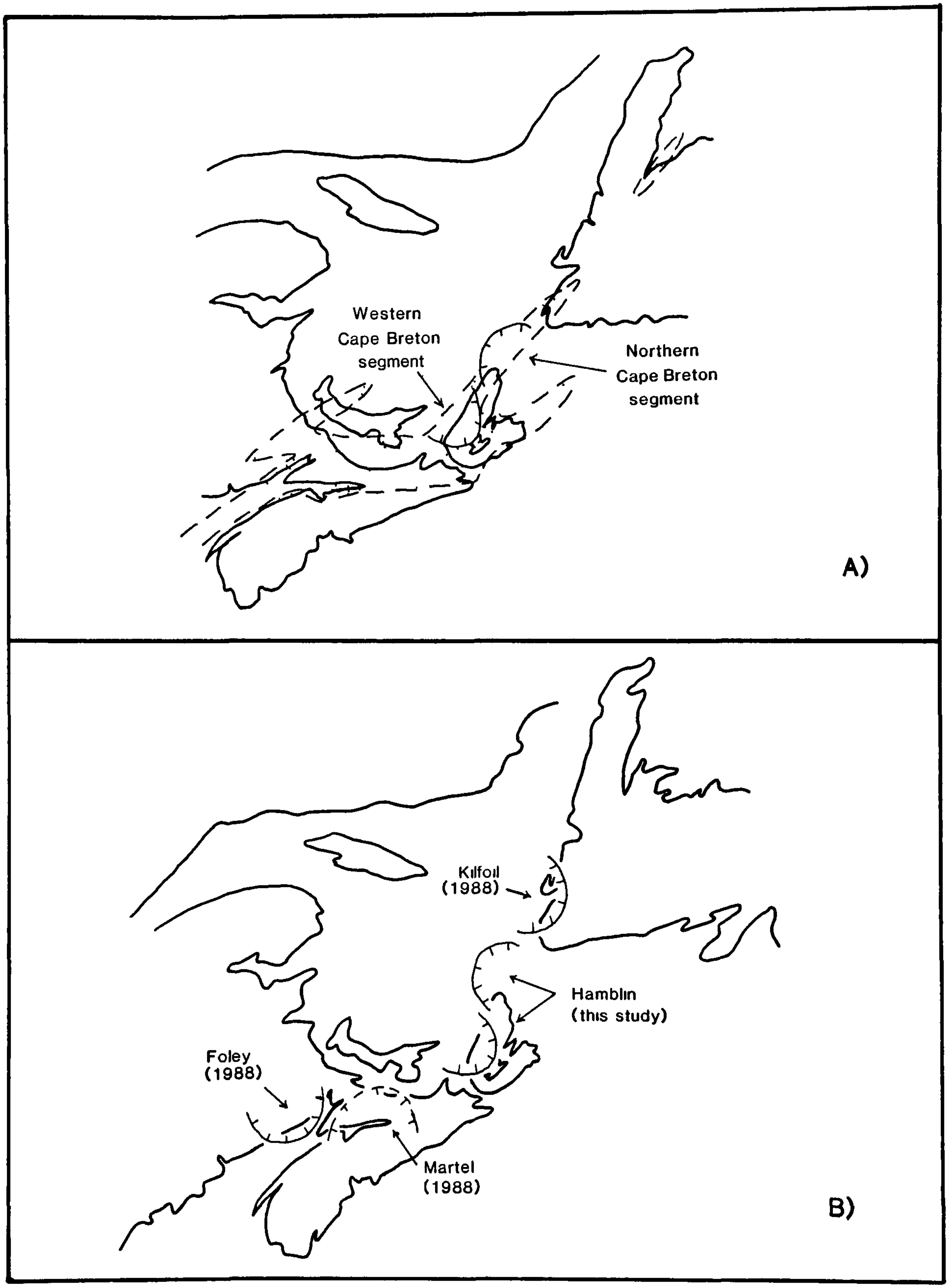

Figure 169. Sub-basins of Horton deposition in the Fundy Basin Rift (Belt, 1968a, 1968b):
A) Western and Northern Cape Breton segments of the Rift (this study),
B) five possible adjacent segments of the Rift mentioned in recent literature.

studied distensive rift systems, as discussed in this chapter.

In addition to Horton Group sediments throughout the Maritimes Basin, Early Carboniferous sediments deposited in interpreted distensive rift basins are known from other circum-Atlantic (ie. Appalachian Orogen) areas. These are in Morocco (Beauchamp and Izart, 1987) and in England (Gawthorpe, 1987; Leeder and Gawthorpe, 1987; Barrett, 1988). All display structural and stratigraphic characteristics similar to those documented in this study. Apparently the Tournaisian/Viséan was a period of regional distension and rift system formation between two compressional phases throughout the Appalachian Orogen.

Interestingly this period corresponds to a known compressional phase on the opposite side of the North American craton in the Canadian Arctic (the Famennian to Viséan Ellesmerian Orogeny) (Thorsteinsson and Tozer, 1970; Embry and Klovan, 1976). The subsequent middle to Late Carboniferous Alleghenian Orogeny in Atlantic Canada is likewise correlative with an extensional and subsidence phase of thick sedimentation in the Canadian Arctic (the Carboniferous/Permian Sverdrup Basin sequence) (Beauchamp, 1987). These mega-correlations imply that the North American craton may have shifted back and forth between confining plates on each side several times during final assembly and consolidation of the Pangean supercontinent (Fig. 170).

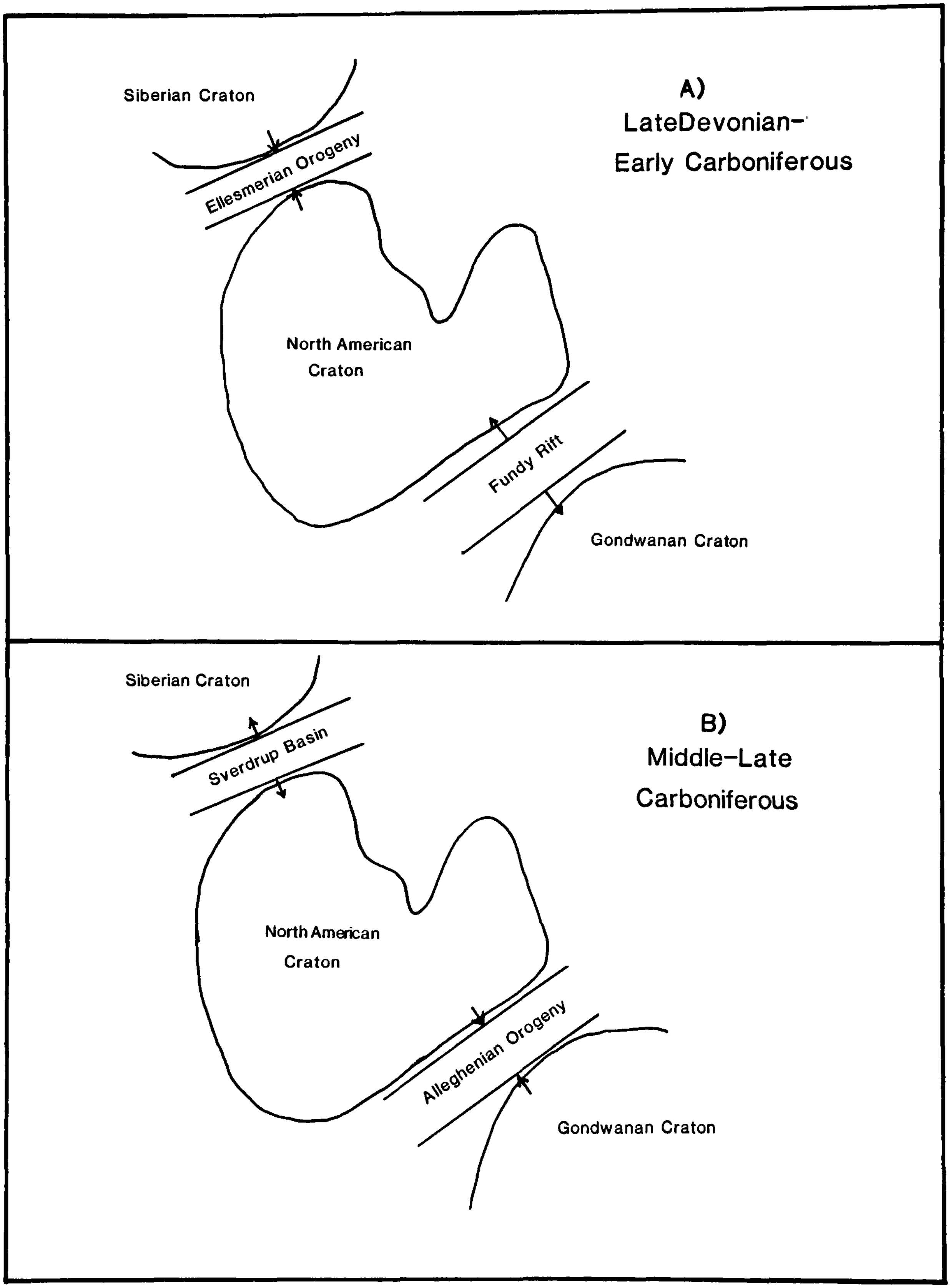

Figure 170. Speculative correlation of tectonic phases in the Canadian Appalachians and Arctic.

CHAPTER 6
RESOURCE POTENTIAL

INTRODUCTION

A practical application of basin-fill analysis is the assessment of the potential for, and predictability of petroleum and mineral resources. This involves a clear understanding of the interrelationships between tectonic controls, depositional style and facies distribution. For example, for hydrocarbon accumulations, tectonic events create the basin of deposition and some potential trap structures, but the sedimentary facies deposited within that basin form the potential source, reservoir and seal rocks. Whereas faulting and folding can be favourable features for resource potential, variable exposure and abundant post-depositional deformation make extrapolation hazardous. The details of the organization, geometry and predictability of the sedimentary facies are crucial to further evaluation of resource potential.

Distensive basins have a unique combination of characteristics which has led to a distinctive suite of natural resources worldwide (Robbins, 1983; Perrondon, 1983). The favourable factors include a) a localized basin encased in crystalline or metamorphic basement b) presence of an igneous heat source at depth and some volcanism, c) vigourous hydrothermal circulation, d) rapid tectonic subsidence in multiple phases, e) thick sedimentary sequence including porous clastics and carbonates, evaporites, and organic-rich deposits which occur in specific geometrical arrangements relative to the fault-bounded margins, f) little metamorphism and moderate structural complications (Robbins, 1983; Perrondon, 1983; Morgan and Golombek, 1984; Powell, 1986). The resulting suite of natural resources includes a) sediment-hosted metallic deposits like lead/zinc/copper, silver, gold, uranium, molybdenum, b) industrial minerals such as barite, fluorite, sodic and potassic evaporites, c) hydrocarbon resources such as coal, oil shale, oil and natural gas, and d) geothermal energy. Both structural and sedimentary style exert major controls on some of these economic deposits.

In this chapter I will review information from several types of economic deposits and compare them to the Horton Group of Cape Breton Island in an attempt to indicate its resource potential. I will focus on only a few deposit types which are most relevant to the Horton Group and which can benefit most from the sedimentological data and interpretations of this study. Deposition of the Horton Group on Cape Breton Island occurred in two large, asymmetric, fault-bounded sub-basins with definable margins, sediment input points and vertical and lateral facies distributions. Near sub-basin margins facies are coarse grained and reddened (permeable oxidized aquifers) whereas near sub-basin centres facies are fine-grained and grey (impermeable, organic-rich, reduced aquitards). Horton sediments are underlain by faulted basement rocks and generally overlain by limestone and evaporites of the Windsor Group (commonly impermeable, organic-rich, reduced aquitard).

METALLIC MINERAL POTENTIAL

Metallic sedimentary mineral deposits which could be present in the clastic units of the Horton Group include a) those hosted in relatively mature, permeable fluvial sediments, especially near sub-basin centres, and b) those hosted at redox boundaries between oxidized permeable sandstone and reduced impermeable mudstone, especially near sub-basin margins. Several of these are reviewed below.

Paleo-placer Gold Placer gold deposits are well known from several ancient sedimentary basins (Proterozoic of South Africa, Tertiary/Quaternary of Yukon) and typically occur as basal lags in coarse-grained fluvial channel deposits or in situ weathering horizons immediately overlying granitic basement (Minter, 1978). Important metallotects are a) gold mineralization in bedrock of the sediment source area, b) uplift, erosion and multiple reworking of sediments, and c) relatively mature braided fluvial deposits (B. Ryan, 1988, pers. comm., G. Burbidge, 1987, pers. comm.). Fault-bounded basins in an intramontane setting with multiple phases of subsidence can fulfill these requirements (as in the Klondike deposits of the Yukon).

Malcolm (1929) and Boyle (1979) described the Gays River paleoplacer gold deposit in the Horton Group near Windsor, N.S. in the Minas Basin area. Gold occurs as flakes and small nuggets in the basal Horton conglomerate overlying auriferous Meguma Group slate and greywacke. The deposit was discovered in 1862 and produced low grade ore (5.4 ppm Au) for about 10 years. Modern placers occur in adjacent streams. On Cape Breton Island Malcolm (1929) noted gold shows in quartz veins in schists of the pre-Carboniferous basement near Cheticamp, Middle River, Whycocomagh (all within the area of this study) and Sydney. Recently Noranda has been conducting a gold exploration program in

Creignish Hills and the southern parts of Cape Breton Highlands (McCulloch, 1988). Malcolm (1929) mentioned the presence of modern placer gold in the Margaree and Middle River drainage systems, both within the area of this study.

The foregoing discussion indicates that gold is present in significant quantities in the basement rocks of the study area and that paleoplacer gold deposits occur in the Horton Group elsewhere in Nova Scotia. Much of the sediment deposited in the Horton Group, particularly in the Craignish Formation, was originally derived from these basement rocks. The interpreted tectonic setting of a distensive rift system implies that uplift, erosion and multiple reworking of detritus was common. Sediments in both the C3 and C1/C2 depositional systems are attributed to deposition in braided fluvial environments where multiple reworking of sediment in relatively high energy conditions is probable. These Craignish braided fluvial systems were well developed all around, and were derived from, the Cape Breton Highlands basement massif where gold mineralization occurs. I conclude that all coarse clastic sediments of the Horton Group in these areas, and particularly in the Craignish C1 facies assemblage of the Cheticamp and Baddeck areas, have potential for paleoplacer gold mineralization. More detailed mapping of individual channel tracts and location of preserved paleosols on the basement would be necessary to further delineate exploration targets. Much more intensive outcrop field work, sampling, and shallow drilling would be required.

<u>Sandstone-hosted Uranium</u> Sandstone-hosted uranium deposits generally occur in post-Devonian oxidized fluvial sequences deposited in closed basins surrounded by granitic or volcanic highlands (Ruzicka and Bell, 1984). The distribution of metals is irregular and keyed directly to the sedimentological details of the fluvial channel (permeable, oxidized sediments) and interchannel (impermeable, unoxidized sediments) system (Finch, 1967). Organic matter plays a crucial role by providing point reduction sites within channel sandstone bodies and the ore is concentrated in irregular zones related to redox fronts (Ruzicka and Bell, 1984). Discordant roll-front deposits apparently result from active syndiagenetic migration of the redox front, and the accompanying geochemical cell, along a sandstone body (Rackley, 1976). The migration is arrested at some point and the deposits "frozen" due to some change in fluid circulation patterns.

Significant uranium concentrations occur in the upper Horton (Cheverie Formation) in the Minas Basin area of Nova Scotia (Charbonneau and Ford, 1978; B. Ryan, 1988, pers.

that area. The host sediments are well sorted, arkosic medium sandstone to granulestone and probably represent deposition in braided fluvial channels. These cores were cut in a Saarberg exploration program in 1981 and H. Quarch, vice-president of Interuranium Canada Ltd. kindly granted permission to view this material for comparative purposes.

On Cape Breton Island the Craignish and Ainslie Formations are dominated by fluvial deposits with detritus supplied from uplifted basement sources and older sediments. There may be some potential for sandstone uranium deposits, although carbonaceous matter, which could provide precipitation sites, is uncommon in the Horton Group due to climatic aridity. Kirkham (1978) found no high concentrations of uranium near the Horton-Windsor contact on Cape Breton Island, although his sampling was not geared directly to this metal.

Redox Boundary (Kupferschiefer) Copper/Lead/Zinc Twenty percent of world copper reserves are disseminated stratiform deposits hosted by fine-grained low energy sediments (Kirkham, 1984). Ore zones are thin but the grade can be rich (up to 5% Cu) over a large area. As discussed by Kirkham (1975, 1978,1984) and Smith and Collins (1984) the Horton-Windsor contact on Cape Breton Island is comparable to several large deposits worldwide. The known metallotects and potential in the Horton Group are discussed below.

Late Paleozoic deposits of this type are prominent around the North Atlantic in the Appalachian Orogen, approximately corresponding to a low paleolatitude belt of arid climatic conditions during and after assembly of the Pangean supercontinent (Kirkham, 1984). Host basins are linear, fault-bounded, extensional basins with internal drainage, positioned within the supercontinent and overlying thinned lithosphere with elevated heat flow (Jowett, 1986; Maiden et al., 1986; Jowett et al., 1987). A distinctive sedimentary sequence includes a) basal continental bimodal volcanics, b) a thick, commonly oxidized, succession of alluvial fan/fluvial/lacustrine deposits with a general trend of fining grain size from fault-bounded margins to basin centre, c) a rapid marine transgression which deposited a thin but extensive laminated, reducing, fine-grained deposit, and d) a capping succession of thick impermeable evaporites (Glennie, 1972; Kirkham, 1984; Haynes, 1986a,b; Jowett et al., 1987). These characteristics are illustrated by the Early Permian Zechstein Group (including Kupferschiefer Formation) of Europe, the Late Precambrian Nonesuch Formation of the U.S., the Late Precambrian Coates Lake Formation of the Northwest Territories, and the Late Precambrian Roan Group of Zaire (Kirkham, 1984). As discussed throughout this thesis, these factors are also all characteristic of the Horton Group/Windsor Group sequence of the Maritimes Basin.

In Kupferschiefer-type deposits the metallic mineralization occurs at the diagenetic redox interface between a permeable, oxidized clastic aquifer below and a less permeable, reduced, fine-grained transgressive aquitard above (Rentzsch, 1974; Brown 1978). This interface is commonly an irregular, discordant surface related to the influx of diagenetic, metal-rich fluids which were confined by the overlying impermeable evaporite seal (Brown, 1978). Mineralization generally appears only if the underlying clastics were reddened by oxidation (Haynes, 1986) and is localized near fault-bounded basement blocks near basin margins where volcanics occur (Jowett, 1986; Jowett et al., 1987). The mineralization is typically near-concordant with the fine-grained host unit in thin, extensive blankets a few metres thick. The ore zone is localized around barren diagenetic Fe-oxidized zones which may cross stratigraphic and facies boundaries at very low angles (Jowett et al., 1987) (Fig. 171). Various interpretations of the mineralization process have been considered, but most recent authors view metal emplacement as a diagenetic precipitation from metal-rich brines at the redox boundary (Kirkham, 1984; Haynes, 1986a,b; Jowett et al., 1987). Distinct metal zonation indicates active circulation of the fluids through the sedimentary sequence, which would promote migration of the redox boundary and continuous enrichment through time (Jowett et al., 1987). Metallic ions were probably stripped by these fluids from the volcanics and volcanic detritus of the basin-fill sequence (perhaps during the oxidation process, Chartrand and Brown, 1985), and ultimately from the surrounding basement (Jowett et al., 1987; Murphy, 1988; Swinden et al., 1988).

In the fault-bounded western Cape Breton sub-basin the dark grey marlstone of the Macumber Formation overlies reddened sandstone and conglomerate of the A1 facies assemblage in belts several kilometres wide near sub-basin margins. These areas, depicted in Figure 172b are considered to have significant potential for sedimentary copper/lead/zinc deposits. At numerous outcrops in the Baddeck area (McLeod Point, Murdock Paul's Brook, McRae Brook, Whycocomagh and Yankee Line Creek) the uppermost few metres of Ainslie red beds are typically bleached and green stained. This tract represents the area of thickest A1 deposits adjacent to the interpreted footwall scarp margin of the sub-basin. Kirkham (1978) conducted analyses on samples from the Horton-Windsor contact in this area and found that a) metals are zoned in a fashion similar to Kupferschiefer-type

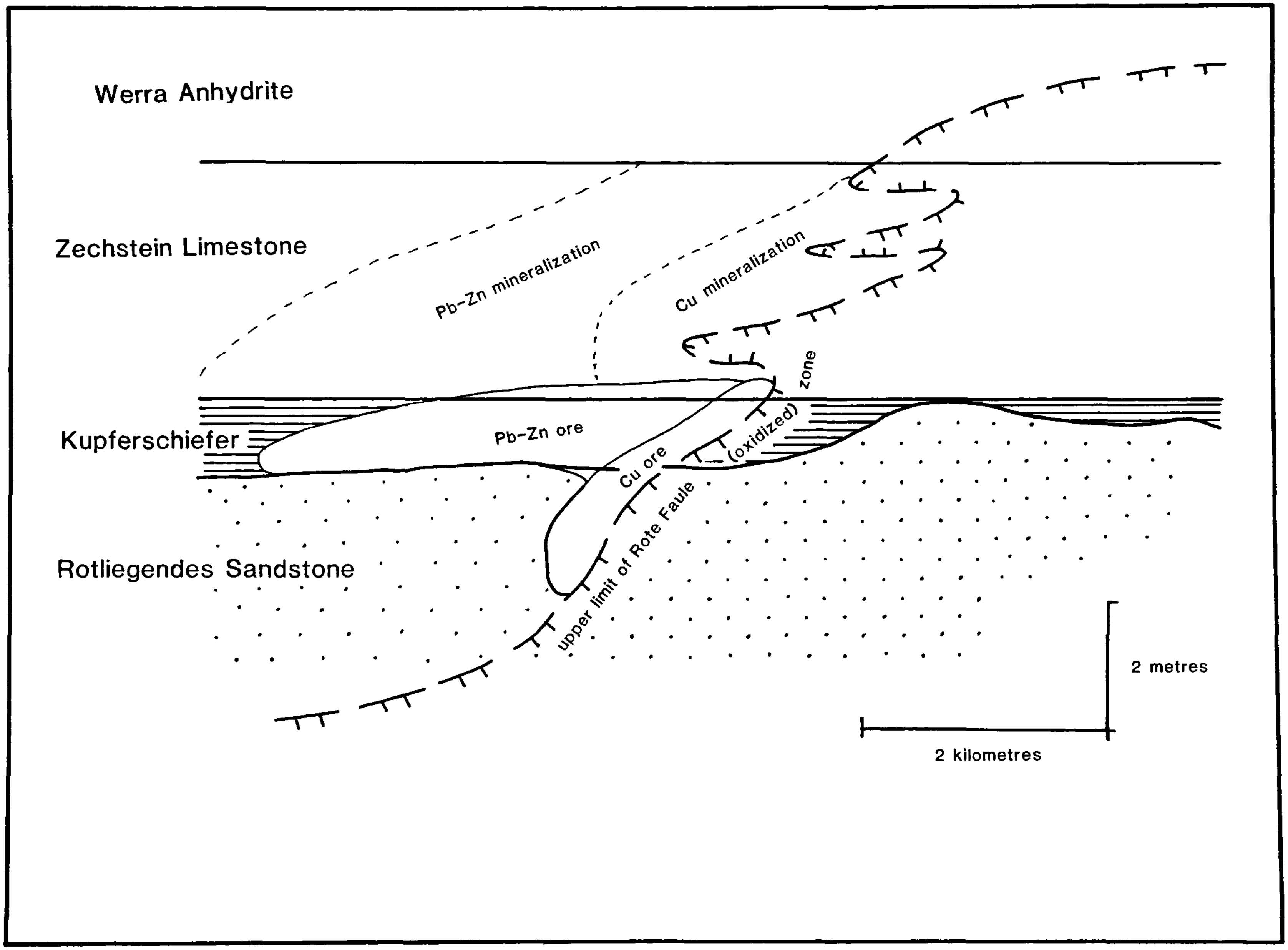

Figure 171. Ore body geometry and metal zonation of Kupferscheifer copper deposit.

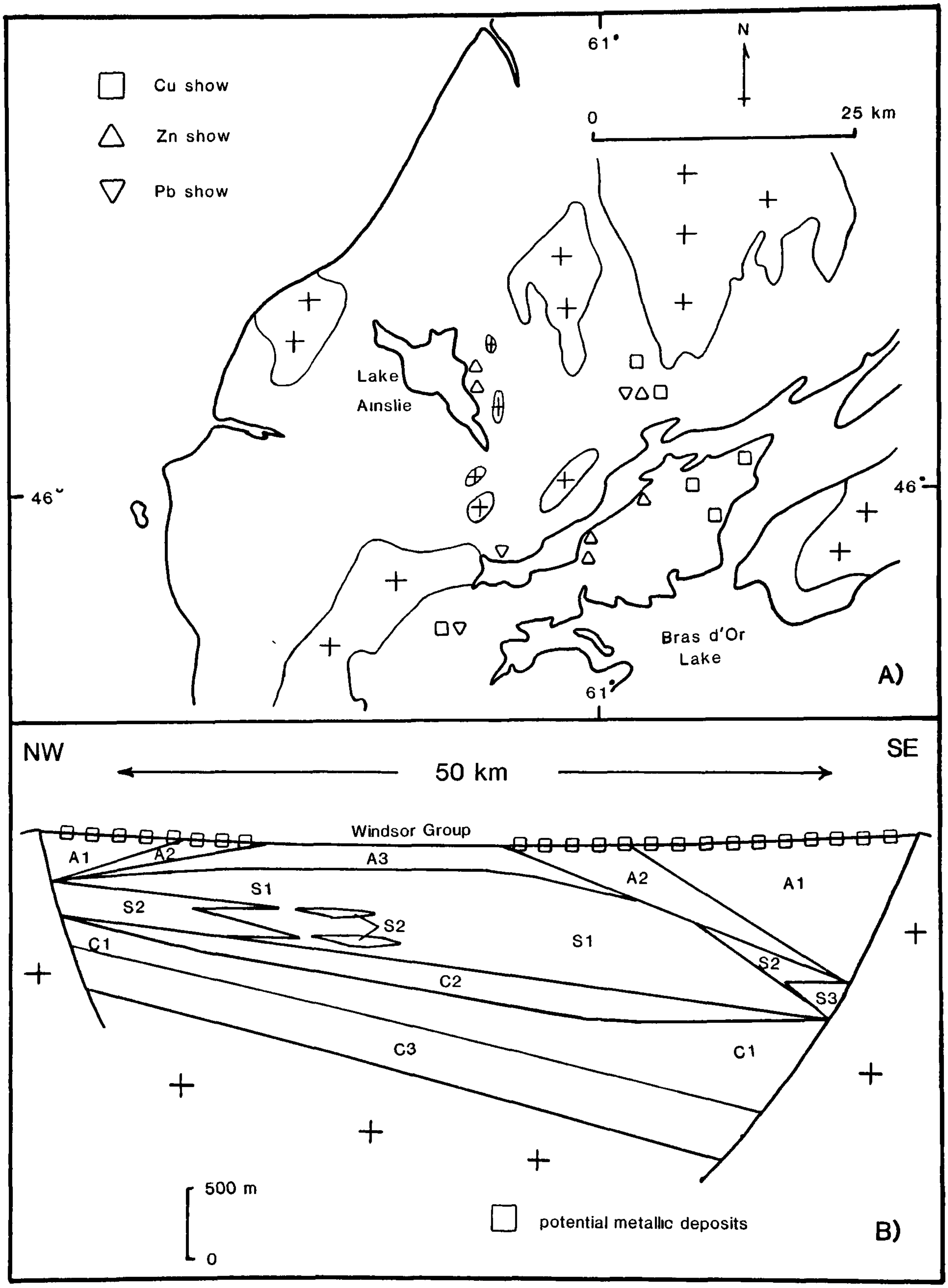

Figure 172. Base metal potential of the Horton-Windsor contact: A) elevated base metal values at Horton-Windsor contact of Western Cape Breton area (Kirkham, 1978), B) relationship of potential metallic deposits to Horton facies assemblage

area and found that a) metals are zoned in a fashion similar to Kupferscheifer-type deposits, and b) copper, lead and zinc values are erratic but generally elevated above background in both the uppermost Horton and lowermost Windsor Groups (Fig. 172a). Uranium is present, but not in significant amounts. In these circumstances the detailed distribution, extent and grade of the mineralization may relate to the contrast at the redox boundary and hence would be determined by the detailed sedimentology of the A1 assemblage sediments at any point.

By mapping the areas where Macumber laminites are underlain by reddened A1 sandstone and conglomerate a prospective area for these deposits can be delineated (Fig. 173). The coarse sediments of the A1 assemblage may have originally had fair to good porosity and permeability. Areas where Macumber overlies the A2 facies assemblage may have potential as well but in that assemblage red siltstone units are abundant between channel sandstone bodies especially near the top. Geological conditions similar to those described above obtain at the Craignish-Strathlorne contact in areas where S1 facies assemblage overlies C1 facies assemblage, and may provide an additional exploration target (Fig. 174). However, this contact is less well exposed and is commonly marked by a transitional unit of S2 sediments, decreasing the effectiveness of the redox boundary. But in several locations near the sub-basin margin (Little Narrows #1, #2) there is an abrupt contact between Craignish red conglomerate and Strathlorne dark grey shale.

Hein et al. (1988) studied the geology of the Horton-Windsor contact at the Jubilee lead-zinc deposit in the Bras d'Or area on the eastern margin of the western Cape Breton sub-basin. They concluded that growth faulting with associated fault scarp breccia on the downthrown side occurred at the time represented by the Horton-Windsor contact. They inferred that exhalative vent fluids were responsible for mineralization of the openwork breccias adjacent to faults. This was interpreted to have occurred within the confines of a small, asymmetric "mini-graben" 1-2 km across, which was positioned approximately at the master fault scarp, interpreted to have controlled deposition in the western Cape Breton sub-basin.

OIL AND NATURAL GAS POTENTIAL

Historical Background The Appalachian area was the cradle of the worldwide petroleum industry (North, 1985) and the Maritimes Basin has a long history of interest in petroleum.

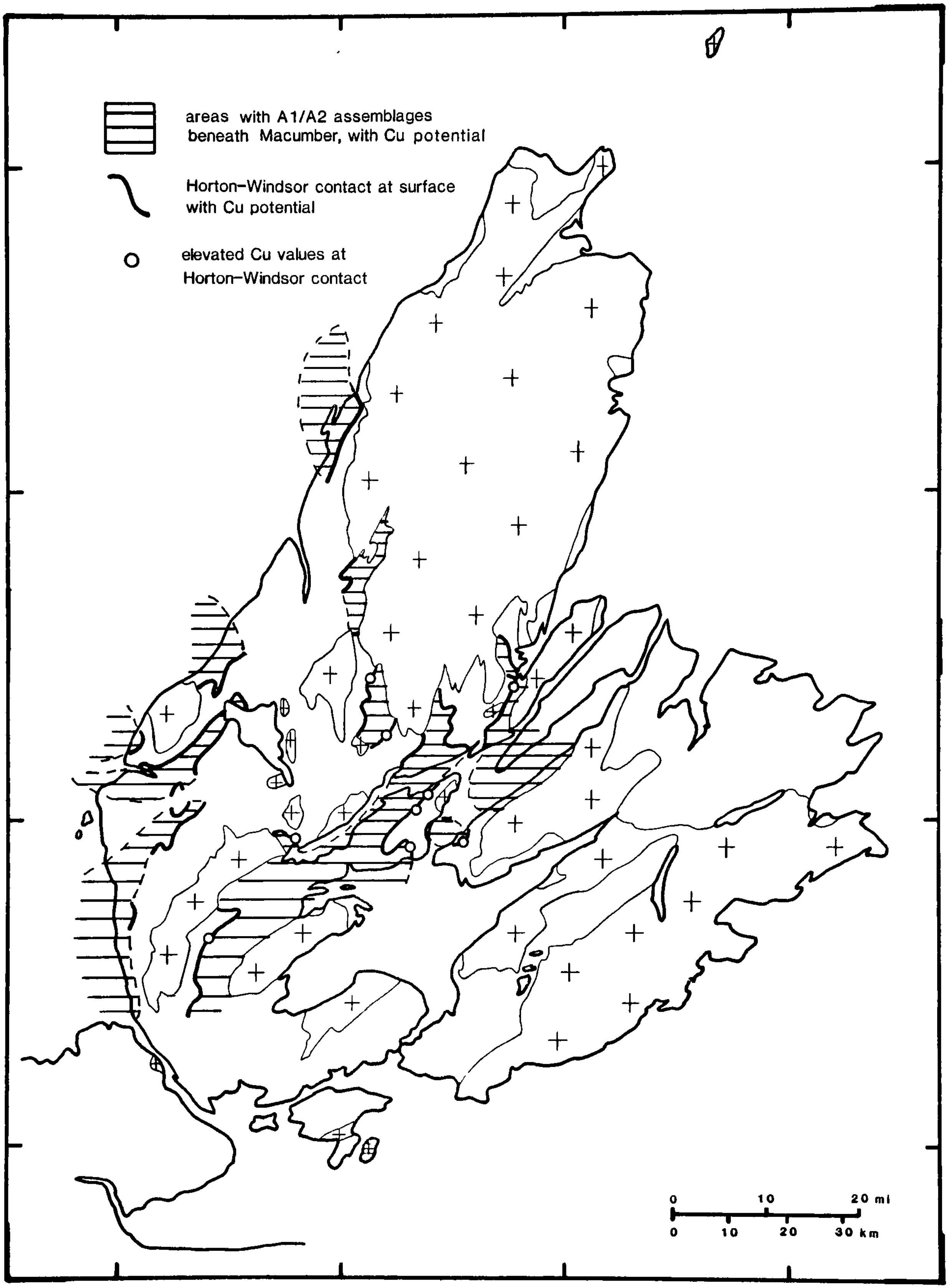

Figure 173. Map of areas where Macumber Formation overlies A1/A2 facies assemblages. These areas have potential for copper deposits.

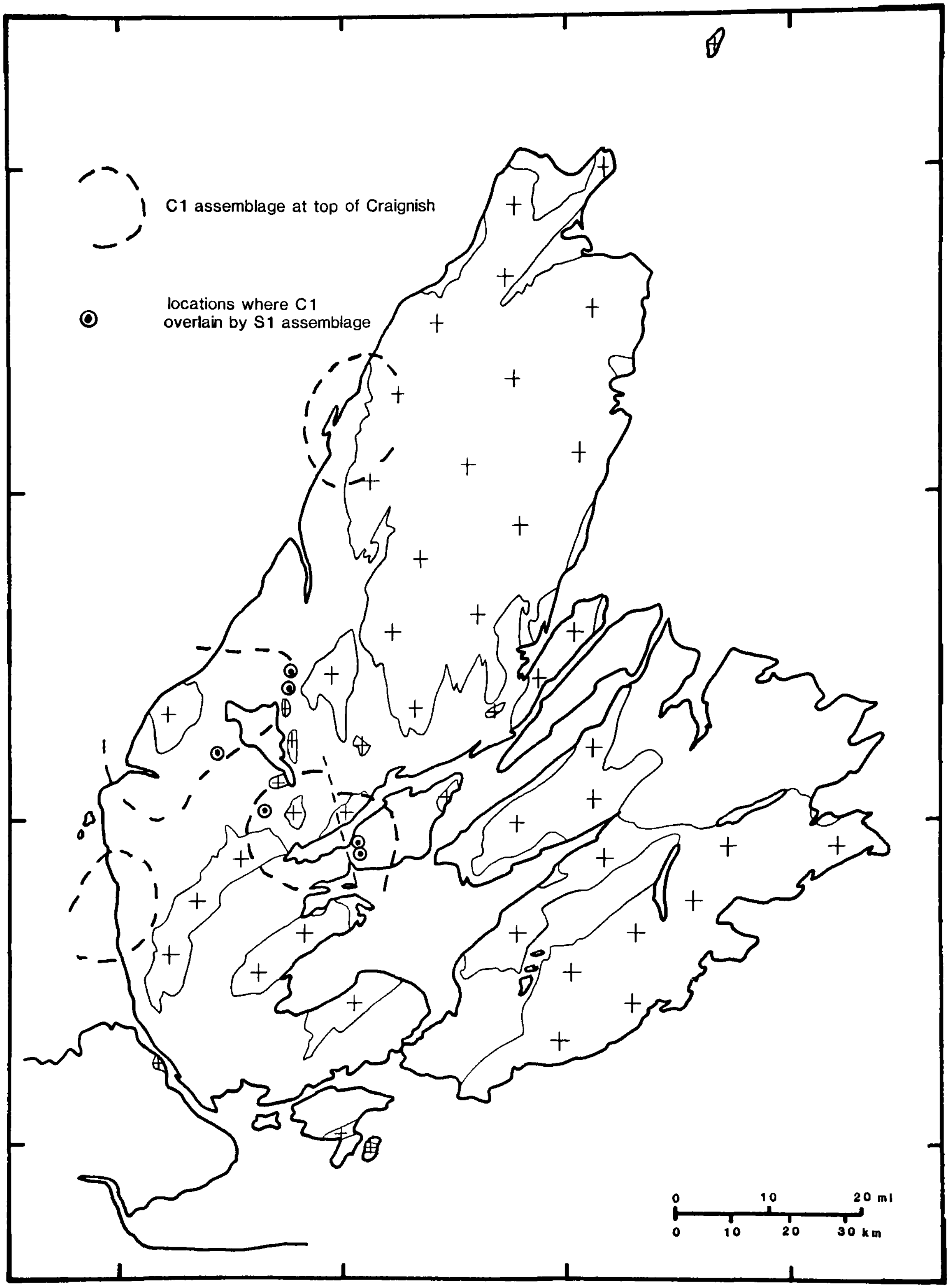

Figure 174. Map of areas where S1 facies assemblages overlies C1 facies assemblage. These areas may have potential for copper deposits.

1836 at Gaspé, Quebec (Devonian sandstone), 1849 at Moncton, New Brunswick (Horton Group) and 1864 at Lake Ainslie on Cape Breton Island (Horton Group) (Williams, 1973). Drilling began in the Moncton area in 1859 (just one year after the historic well at Oil Springs, Ontario; Hilborn, 1968) and the Stoney Creek Oil and Gas Field was discovered in 1909. Although very small, this field is significant because a) it is still the only productive oil field in the Maritimes Basin b) production is from sandstone reservoirs of the Albert Formation, correlative to the Strathlorne Formation of this study, and c) for 75 years it has represented the magnet that continues to draw some exploration to this area. The excellent reservoirs include up to 30 shoreline sandstone bodies (equivalent to S2 facies assemblage) interbedded with dark grey organic-rich source rock (equivalent to S1 facies assemblage) on the north limb of a syncline.

Meanwhile exploration efforts had spread to the oil seeps from the Horton Group (Ainslie Formation, A3 facies assemblage) at Lake Ainslie on Cape Breton Island (Fig. 175a). In 1869 the Lake Ainslie Pioneer Oil and Salt Company drilled a successful well there at the crest of the Ainslie Anticline, which may be the hole still present in the grass near the McIsaac farmhouse. Over the next 8 years, six more wells were drilled in that vicinity, all successful (McMahon et al., 1986). Between 1869 and 1905, twelve of the nineteen wells drilled in Nova Scotia were located on surface anticlines in the Mabou/Lake Ainslie area and all resulted in oil and gas shows. The discovery of the Stoney Creek Field sparked Maritime Oil and Gas Company to drill 7 more wells into the Horton on the north side of Lake Ainslie between 1910-15, again with many hydrocarbon shows (Côté, 1959; McMahon et al., 1986). Mather and Trask (1928) made a favourable report on the potential of the Horton and Windsor Groups in western Cape Breton and Eastern Gulf Oil Company proceeded with a shallow 17 well program in the Mabou/Margaree area in 1926-27. The results were disappointing and at the same time Bell (1926) downgraded the Carboniferous petroleum potential, leading to a hiatus in exploration.

In 1942-44 Lion Oil Company conducted a sophisticated program including geological mapping, gravity, seismic and two seismically-located wells in the Mabou area. They were located on salt domes and did not penetrate the Horton Group target horizon (Boehner, 1986; McMahon, 1986). During 1946-47 little Narrows Petroleum Company drilled 3 wells in the Bras d'Or area to follow up surface seeps (including Little Narrows Jubilee #1, #2) and found thin lenses of bitumen in red conglomerate (McMahon, 1986) of the A1 facies

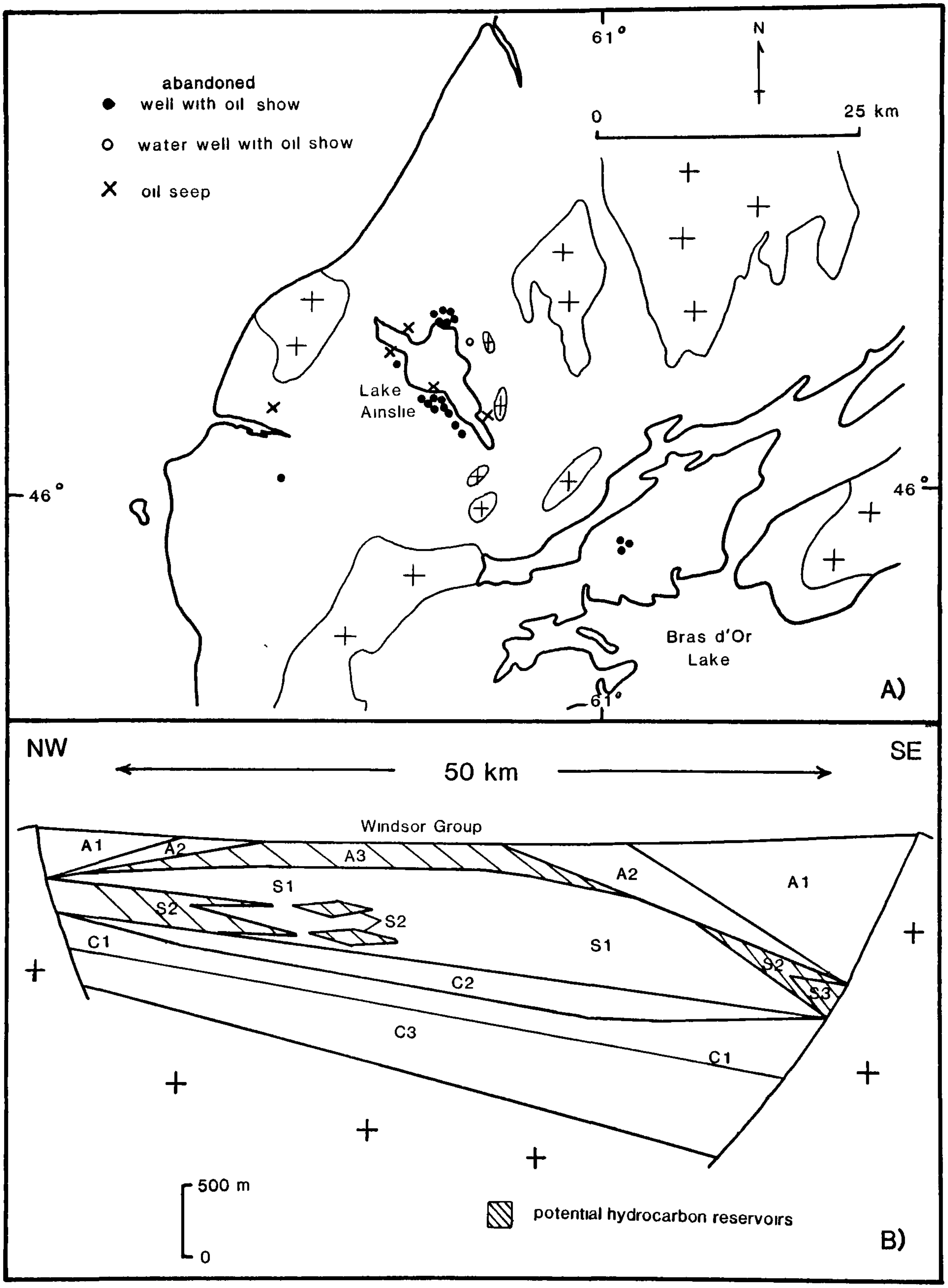

Figure 175. Hydrocarbon potential of the Horton Group: A) oil wells, seeps and shows in Western Cape Breton (McMahon, 1986), B) potential reservoir facies assemblages of the Horton Group.

assemblage, as designated in this study. A shallow hole was also drilled for structural information near Lake Ainslie (Little Narrows Hays River). In 1956-57 Imperial Oil conducted field work in western Cape Breton resulting in a favourable report on potential of the Horton Group (Côté, 1959), and proceeded with an 8 well program in 1958-60. encouraging oil shows were recorded from Strathlorne (S2 facies assemblage) and Ainslie (A3 facies assemblage) sandstone in 5 of these (including Imperial Mabou #1 and Imperial Inverness #3) (McMahon et al., 1986).

Mineral exploration by Texas Gulf, Amax and Chevron in the Bras d'Or area from 1975 to 1981 encountered many minor oil shows in Horton and Windsor Groups. The latter company followed these up in 1979 with 3 Chevron-Irving wells near Malagawatch (McMahon et al., 1986). Meanwhile, during the 1973-83 period, several wells were drilled offshore in the Gulf of St. Lawrence including Petro Canada et al., St. Paul P-91 which penetrated Horton rocks off the northern tip of Cape Breton Island. In western Cape Breton Chevron-Irving conducted seismic programs in 1982-84, completed field work in 1987, and drilled a 1500 m dry hole in 1988 through Horton rocks (Chevron-Irving Mull River #1). In 1985 a water well located at the crest of the Ainslie Anticline recovered oil from Horton sandstone (A3 facies assemblage) at a depth of 25 m. In 1988 the Nova Scotia Department of Mines and Energy twinned this with a 68 m hole and found that all sandstone beds (A3 facies assemblage), up to 20 m thick, had good porosity with bitumen and live oil shows (P. McMahon, 1988, pers. comm.).

Clearly, hydrocarbon exploration on Cape Breton Island, with the Horton Group as the primary target horizon has a 120 year history to the present day. Oil and natural gas certainly are reservoired in some sandstones of the Ainslie and Strathlorne Formations (Fig. 175), although no commercial deposit has been discovered. Data and interpretations from this study can help answer several remaining questions concerning the hydrocarbon potential of the Horton Group. These are a) are there high quality potential source rocks and potential reservoir rocks in the Horton Group?, b) are the distributions of source rock, reservoir rock and traps suitable and predictable? c) is the presence of large overthrust plates and associated structures significant?, d) is there potential offshore in adjacent areas?, and e) in general, has the exploration to date properly evaluated the hydrocarbon potential?

Hydrocarbon Potential of Fault-bounded Basins Fault-bounded basins host some of the most prolific oil fields known and include about 10% of the world's oil reserves, although they account for only 5% of the world's basins (Klemme, 1980; Barker, 1984). Productive examples include the Viking graben in the North Sea (Pegrum and Mounteney, 1978), Sirte basin in Lybia, Rhine graben in Germany, Songliao and Bohai basins in China (Li et al., 1982; Quanmao and Dickinson, 1986), Reconcavo basin in Brazil (Ghignome and Andrade, 1970), and Gabon basin in West Africa (Brink, 1974). Klemme (1980) predicted that oil reserves in fault-bounded basins would eventually reach 25% of world total. These basins are characterized by abundant potential source, potential reservoir rock, short migration paths between source and reservoir and a widespread sealing sequence (Barker, 1984; North, 1985). As discussed in Chapters 4 and 5, periods of peak subsidence generate source rock facies in the axial zone followed by periods of tectonic quiescence when there is rapid accumulation of reservoir facies near the margins (Quanmao and Dickinson, 1986). Abrupt vertical and lateral facies changes create many reservoirs (Robbins, 1983) and secondary porosity is common (Ethridge and Wescott, 1984). Fault-bounded basins are also characterized by a relatively high geothermal gradient and multiple structural trap possibilities (North, 1985).

Source Rock Potential Lacustrine sediments have been recognized as significant, oil-prone, potential source rocks only recently (Powell, 1986; Davies and Nassichuk, 1988). Rapidly subsiding, fault-bounded basins with internal drainage in a warm arid climate commonly contain large stratified lakes with abundant solute and nutrient input from intense weathering (Robbins, 1983; Lee and Kerr, 1984). This promotes abundant organic production in surface waters and excellent preservation in deeper axial zones below the thermocline (Lee and Kerr, 1984; Powell, 1986; Watson et al., 1987). Lacustrine sediments may contain up to 50% organics by volume and up to 20% total organic carbon (T.O.C.), although much of the sediment has T.O.C. < 1% (Demaison and Moore, 1980; Quanmao and Dickinson, 1986; Powell, 1986). The most important organism is the alga Botrycoccus which stores abundant amorphous lipids, the precursors to Type I kerogen, which produces light, paraffinic crude oil on thermal maturity (Robbins, 1983; Tissot and Welte, 1984; Powell, 1986). Significant quantities of terrestrial plant material (woody and coaly) may also be emplaced, the precursors to Type III kerogen which produces natural gas and minor oil when heated (Tisott and Welte, 1984; Powell, 1986). The two most important factors in establishing source rock potential are a) suitable thermal maturity, and b) sufficient organic material.

Hacquebard and Donaldson (1970) studied vitrinite reflectance (R_0) in Upper Paleozoic rocks of the Maritimes Basin, Three samples were obtained from the Horton Group of the present study area; that from Gallant River ($\%R_0$ = 2.11) is overmature with respect of oil but might generate minor gas, those from North River ($\%R_0$ = 0.83) and Yankee Line Creek ($\%R_0$ = 0.95) are within the oil window (R_0 = 0.5-1.35%, Barker, 1984). Palynological samples collected in this study were assigned a Thermal Alteration Index (T.A.I.) from 0 (thermally immature) to 5 (metamorphosed) by J. Utting (App. II, Fig. 176). These data indicate that the Strathlorne Formation lies within the oil window (T.A.I. = 2 to 3-) over much of the study area, except close to basement blocks at sub-basin margins and in the Margaree area where all samples are beyond the oil window but within the gas window (T.A.I. = 3 to 4-). Samples from S1 facies assemblage from sections with little interpreted shoreline influence generally have slightly lower T.A.I. values than those close to interpreted shoreline influence in the same measured section. As described in Chapters 3 and 5 shoreline influence is much less prevalent in the footwall scarp margin of asymmetric sub-basins. From the small amount of data available, I conclude that the Horton Group lies within the oil and/or gas windows of thermal maturity over most of the study area.

Data relevant to the amount of organic matter present in the Strathlorne Formation are presented in Appendix III and Figure 177, and were kindly provided by P. McMahon and W. Smith of Nova Scotia Department of Mines and Energy from samples collected in 1987 and 1988. Although many of the samples come from areas where there was much shoreline influence (which would dilute the organic material), there are a number which have relatively high T.O.C. In addition many samples have pyrite degradation of miospores indicating deposition in anoxic reducing conditions. In the western Cape Breton sub-basin, several samples from the Cheticamp-Margaree area have enough organic material to be considered as potential source rocks, but, as discussed above, this area appears to be thermally overmature with respect to oil. The Baddeck area is very much under-sampled and only one sample yielded values with potential source rock characteristics. However, palynology samples from this area contain abundant amorphous, exinous and Botrycoccus material. I would predict that this area, close to the controlling footwall scarp and axial

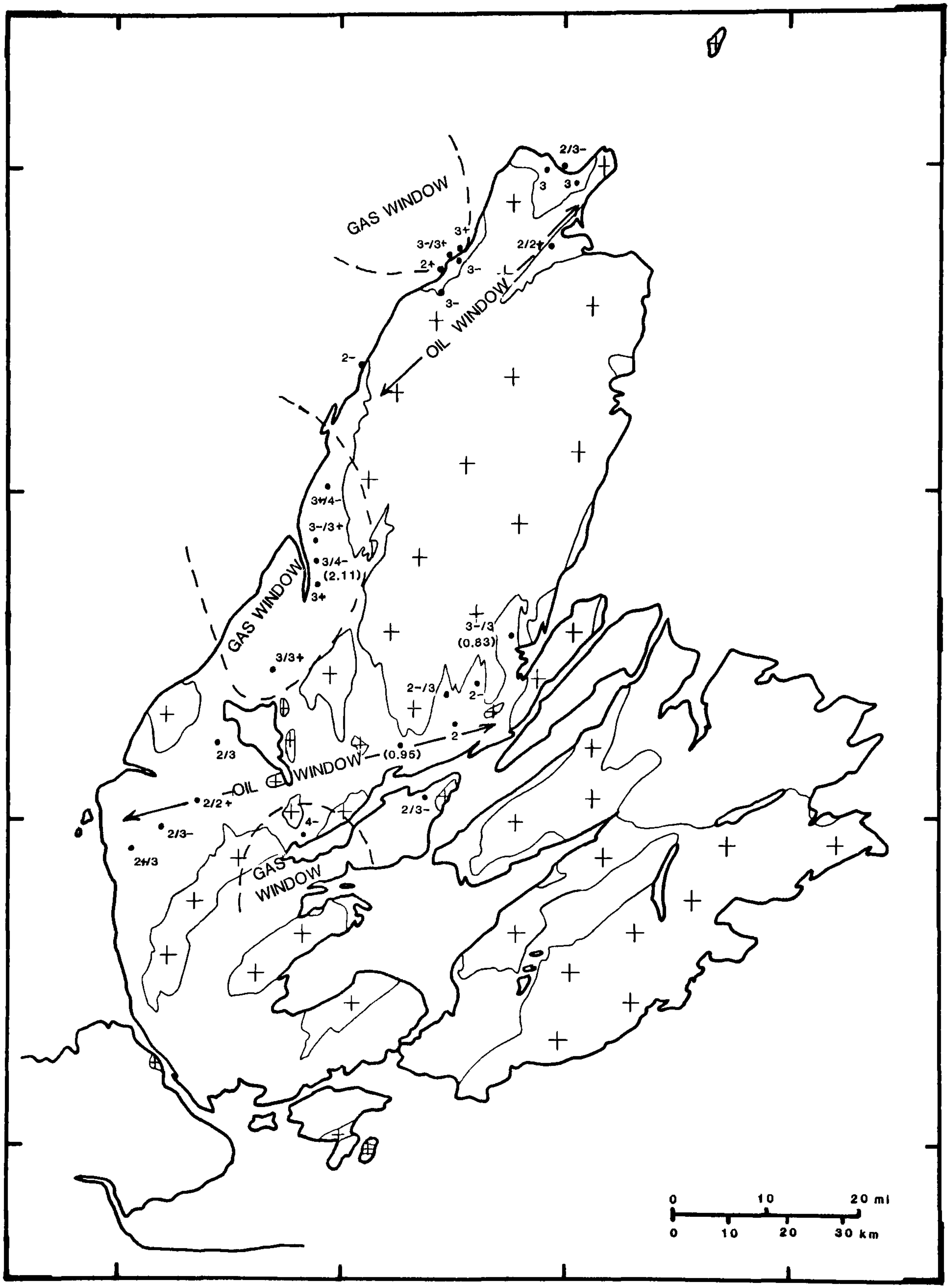

Figure 176. Map of Thermal Alteration Index data from palynological samples (see Appendix II). Several vitrinite reflectance values appear in brackets (Hacquebard and Donaldson, 1970).

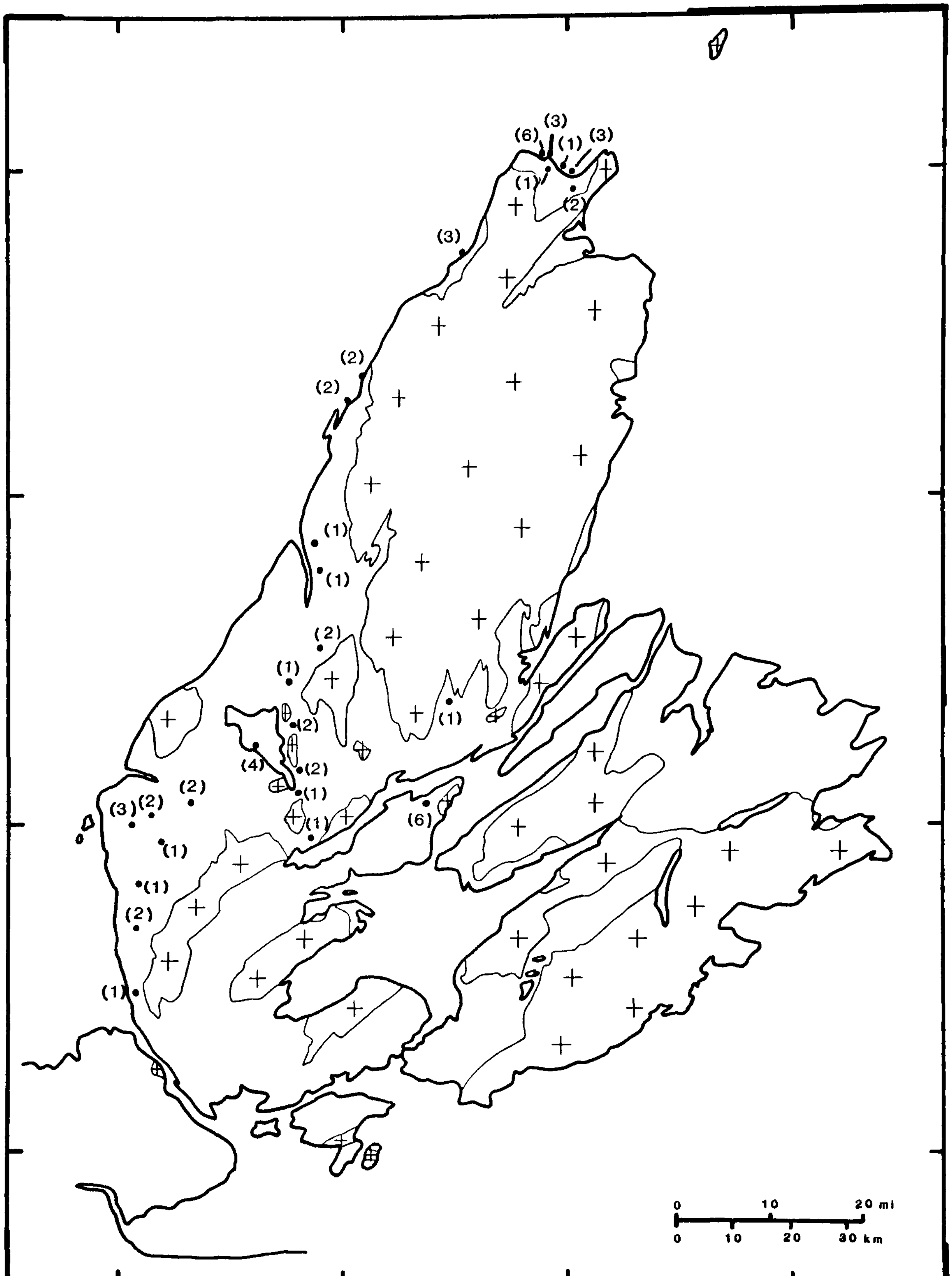

Figure 177. Locations of samples analyzed by Rock-Eval technique for hydrocarbon source rock characteristics (see Appendix III). Numbers in brackets refer to number of samples analyzed.

zone of the sub-basin, would have excellent source rock potential (especially for oil), and further sampling is vital to proper evaluation. Many samples from the exposed part of the northern Cape Breton sub-basin have excellent source rock characteristics, including abundant amorphous, exinous and Botrycoccus material. As discussed above, the Strathlorne Formation in this area is thermally mature with respect to oil, and I conclude that it is capable of generating significant quantities of hydrocarbon. A large portion of this sub-basin lies offshore in Cabot Strait and cannot be further evaluated at present. The northern sub-basin may have greater source rock potential than the western sub-basin because it appears to have been more controlled by faulting and deeper subsidence during Strathlorne time, even on the hanging wall margin of the half-graben.

I conclude that hydrocarbon (especially oil) source rocks are present in the Strathlorne Formation of the study area, particularly in the northern sub-basin and the Baddeck area of the western sub-basin. However these source rocks are not universal, but occur as extensive units of dark grey laminated mudstone several metres thick in the lower portions of coarsening-upward sequences which typify the S1 facies assemblage. The dark grey mudstone facies was deposited below the thermocline in a stratified lake and so should be best developed and most abundant near the axes of the sub-basins, where subsidence was most effective.

Reservoir Rock Potential Fault-bounded basins have thick clastic sedimentary sequences, especially near their margins and in their upper parts (Quanmao and Dickinson, 1986; Frostick and Reid, 1987). These major potential reservoir bodies occur in a confined basin, interbedded with or overlying potential source rock facies, and commonly are overlain by a regional seal of carbonates and evaporites (Fouch and Dean, 1977; Li et al., 1982; Ethridge and Westcott, 1984; Barrett, 1988). Shoreline-related facies may extend up to several tens of kilometres out from basin margins, be composed of sandstone with uniform high porosity and permeability, have excellent continuity and predictability (Candido and Wardlaw, 1985), and may be stacked in multiple sets separated by source rock facies (Quanmao and Dickinson, 1986; North, 1985). Alluvial fan sediments tend to have abundant fine grained matrix, altered feldspar and low porosity, but their distal edges may be reworked into deposits with better sorting and higher porosity and permeability. Fluvial deposits are typically well sorted and may have high porosity and permeability, but tend to be laterally discontinuous and relatively difficult to predict. Primary porosity may be preserved but as Dean and Fouch (1985) suggest, most lake water is carbonate-rich and early cementation is likely, which may inhibit excessive compaction. Therefore, preserved porosity in these sequences is commonly secondary after calcite and feldspar dissolution and can be very good but difficult to predict (Fouch and Dean, 1977; Westcott and Ethridge, 1980; Dean and Fouch, 1985; Parnell, 1985; Candido and Wardlaw, 1985). Brink (1974) found that trends of reservoir facies were localized by syn-depositional, fault-controlled "hinge lines" in the half-graben of the Gabon Basin.

The presence of oil shows and production from oil wells in the Lake Ainslie area indicates that there is suitable reservoir rock available in the Horton Group. The few thin sections examined in this study indicate the presence of secondary porosity up to 10% (visual estimate) after dissolution of calcite cement and feldspar in facies assemblages S2, S3 and A3. Data from outcrop samples, published in reports filed with Nova Scotia Department of Mines and Energy, and relevant to the reservoir rock potential, are listed in Appendix IV. These indicate that moderate to high values of porosity (ø) and permeability (k) are common in the western sub-basin, although the diagenetic history has obviously been complex and the relationship to subsurface conditions is unknown. Samples from the Craignish Formation average 12% ø, and those from the Strathlorne Formation S2 facies assemblage average 12% ø. More extensive data for the Ainslie Formation indicate that in the A2 facies assemblage there is an average ø of 14% and average k of 44 md, and in the A3 facies assemblage an average ø of 17% and average k of 79 md. Data from the Craignish are too sparse for conclusions, except that most porosity in the C3 facies assemblage is occluded by silica cement as observed in thin section. The highly oxidized and reddened A2 facies assemblage may not preserve any hydrocarbons, although it has sufficient porosity and permeability. The S2, S3 and A3 facies assemblages appear to have significant reservoir potential.

Sediments of the S2 and S3 facies assemblages represent extensive tongues of well sorted shoreline-related and deltaic deposits which are intimately interbedded with assemblage S1 source rocks. Areas where these deposits are preserved and sealed by overlying units are mapped in Figure 178. Sediments of the A3 facies assemblage represent linear well sorted fluvial channels which overlie the potential source rocks of the Strathlorne in the central portions of sub-basins. Areas where these deposits are preserved and sealed by overlying Windsor Group rocks are mapped in Figure 179. Together the areas shown

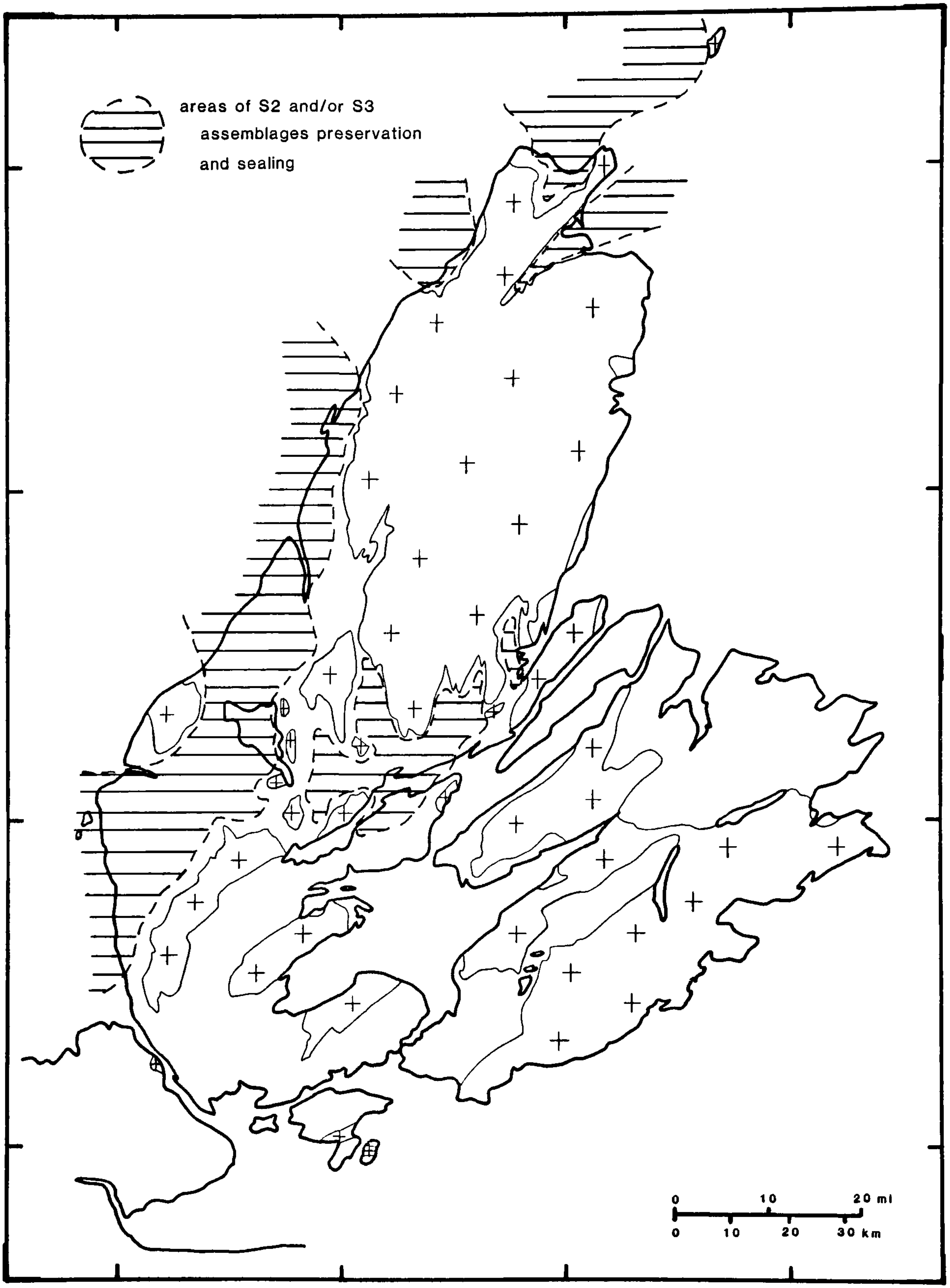

Figure 178. Map of areas where S2 and S3 facies assemblages (potential hydrocarbon reservoirs) are preserved and sealed by overlying rocks.

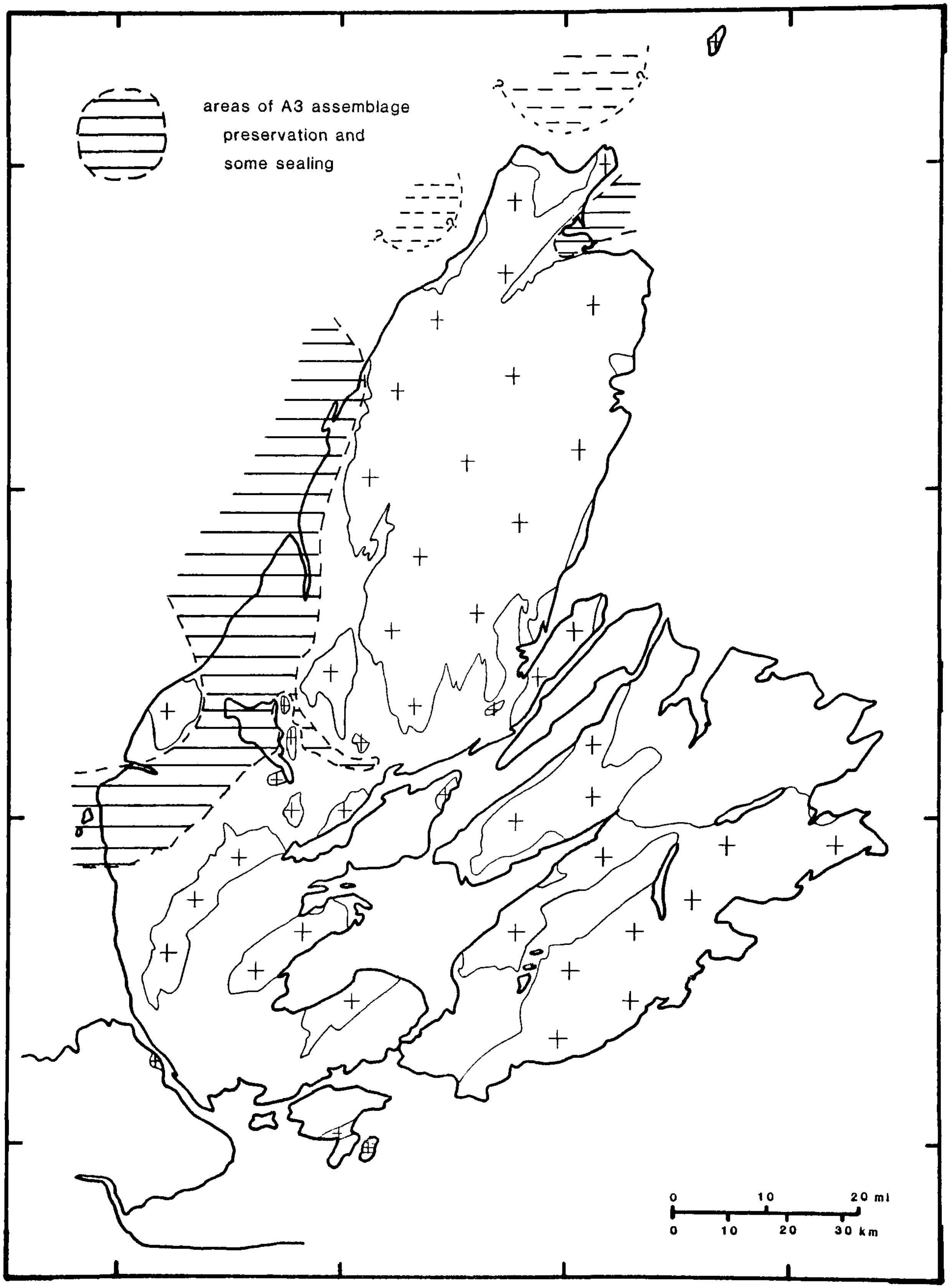

Figure 179. Map of areas where A3 facies assemblage (potential hydrocarbon reservoirs) are preserved and sealed by overlying rocks.

on these maps represent the areas of greatest exploration potential both onshore and offshore. However these are simple fairways within which more detailed sedimentological research could delineate individual reservoir bodies. Most oil seeps and shows recorded in the past occur in these same areas.

Trap Possibilities Prolific structural traps in oil and gas fields are known from fault-bounded basins such as the Los Angeles basin, Viking graben, Songliao Basin, Sirte Basin, and Gulf of Suez (Lowell, 1985). They are generally abundant but small and occur in definable fairways near basin margins, related to known structural patterns of the faults. Possible trap types include those related to rotation of tilt blocks on listric normal faults, and are related to accessory folding (Harding, 1984; Gibbs, 1984; Lowell, 1985). Prolific stratigraphic traps in oil and gas fields of fault-bounded basins are known from the North Sea, Sirte Basin, Gulf of Suez, Songliao Basin and Bohai Basin (Harding, 1984; North, 1985). They are generally of moderate size but occur in definable fairways near basin centres, related to the known facies distributions and intertonguing patterns. Possible trap types include those related to pinch-out of the coarser-grained marginal and shoreline facies, and to the coarser-grained upper basin-fill portion of the sequence (Robbins, 1983; Quanmao and Dickinson, 1986).

In the Horton Group of the study area the possible trap types can be conjectured by comparison to these known examples. Structural traps are more likely to be concentrated near the footwall scarp margin of half-graben segments. Fault-related structural traps might be present as a) high points of uptilted basement blocks rotated on listric normal faults (especially in Craignish strata), b) antithetic fault seals (especially in Strathlorne and Ainslie strata) and, c) sealing beneath overthrust plates near sub-basin margins (as discussed in Chapters 2 and 4) (Fig. 180). Fold-related structural traps might include a) rollovers on the hanging wall of listric normal faults (especially in Strathlorne and Ainslie strata), b) anticlinal drape over rotated basement blocks (especially in Strathlorne and Ainslie strata) and, c) rollovers on thrust-reactivated listric faults near sub-basin centres (as discussed in Chapters 2 and 4) (especially in Ainslie strata) (Fig. 181). Fractured reservoirs could also occur in this setting. Of the above possibilities, those related to the large overthrust plate tentatively identified in the western Cape Breton sub-basin (and perhaps present elsewhere) have the greatest potential for large hydrocarbon accumulations. There is good potential at depth, even in areas where the regional seal of

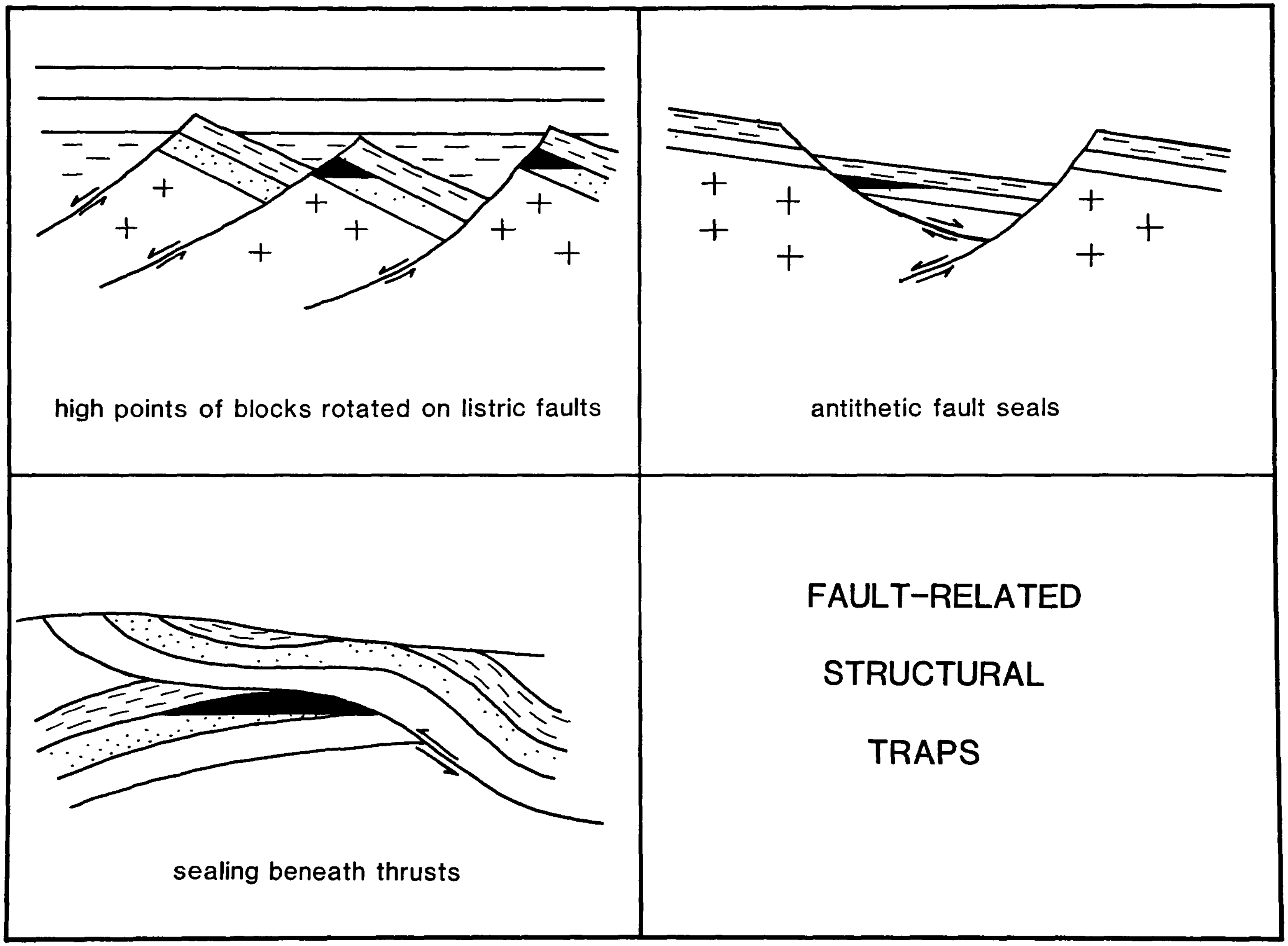

Figure 180. Fault-related structural trap types which may be present in Horton rocks of Cape Breton Island.

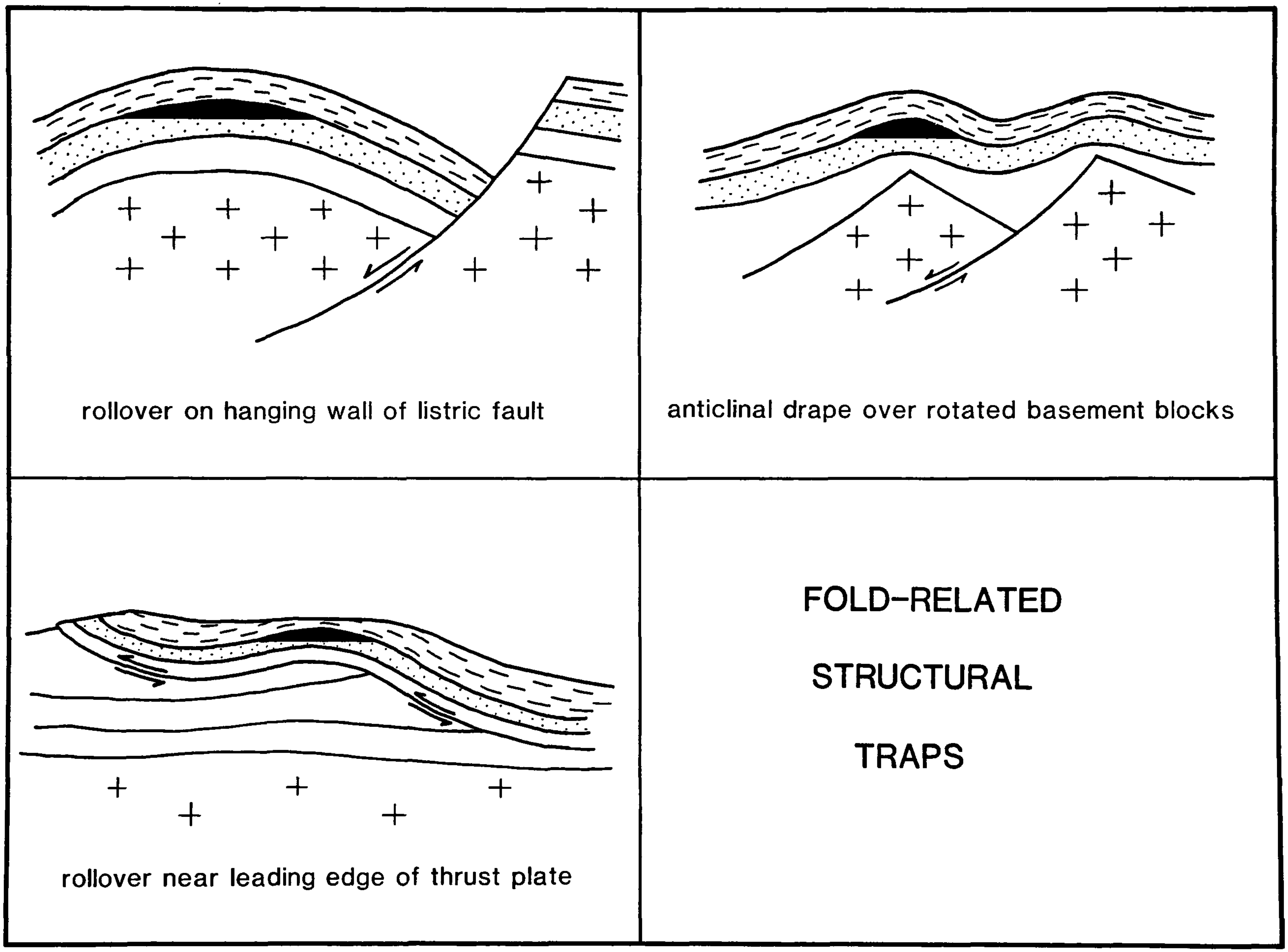

Figure 181. Fold-related structural trap types which may be present in Horton rocks of Cape Breton Island.

the Windsor Group has been eroded.

In the fault-bounded half grabens present in the study area, stratigraphic traps may occur throughout the geographic area of the sub-basins, but are ultimately closely related to the structural evolution of the sub-basins. Stratigraphic traps might be present as a) belts where rudaceous fan-delta deposits of the S3 assemblage near the footwall scarp, and arenaceous shoreline-delta deposits of the S2 assemblage near the hanging wall margins and ends of the sub-basins pinch out into assemblage S1 mudstone, b) interdeltaic shoreline-related shoal and flat deposits of various limestone types in S1 and S2 facies assemblages, c) the entire A1/A2/A3 depositional system (especially A3) which overlies the Strathlorne Formation and represents the final filling phase of the Horton sub-basins (Fig. 182). Unconformity traps near sub-basin margin faults may also occur. Of the above possibilities, trapping in A3 fluvial channels is already known, but pinch out of extensive S2 sandy shoreline tracts provides the greatest potential for large hydrocarbon accumulations. This trap type is identical to that present in the Stoney Creek oil and gas field of New Brunswick.

Summary In the Horton Group of Cape Breton Island there appear to be significant volumes of potential hydrocarbon source rock with suitable thermal maturity (S1 facies assemblage) in stratigraphic continuity with significant volumes of potential reservoir rock (S2, S3 and A3 facies assemblages). There are also abundant structural and stratigraphic trap possibilities, a regional seal sequence (Windsor Group), and many known oil shows and seeps. However the long history of exploration of the Horton Group has been sporadic, rather unsystematic, and has focussed on drilling shallow wells in the Lake Ainslie/Mabou area. Only Little Narrows Hays River, Imperial Mabou #1 and Chevron-Irving Mull River #1 in western Cape Breton have penetrated beneath the Ainslie into zones of potential S2 reservoirs and potential overthrust structures. Only one well is present in the entire northern Cape Breton sub-basin (Petro-Canada et al. St. Paul P-91). Sample and analysis density in key areas of potential source and reservoir rocks are poor to moderate. Seismic density in western Cape Breton is moderate and in northern Cape Breton is virtually nonexistent. I conclude that, whereas hydrocarbon potential in this frontier area is moderate at best, past exploration efforts have not fully evaluated all the possibilities. The observations, interpretations and basin-fill analysis offered in this study can help to suggest new exploration concepts and improve exploration efficiency.

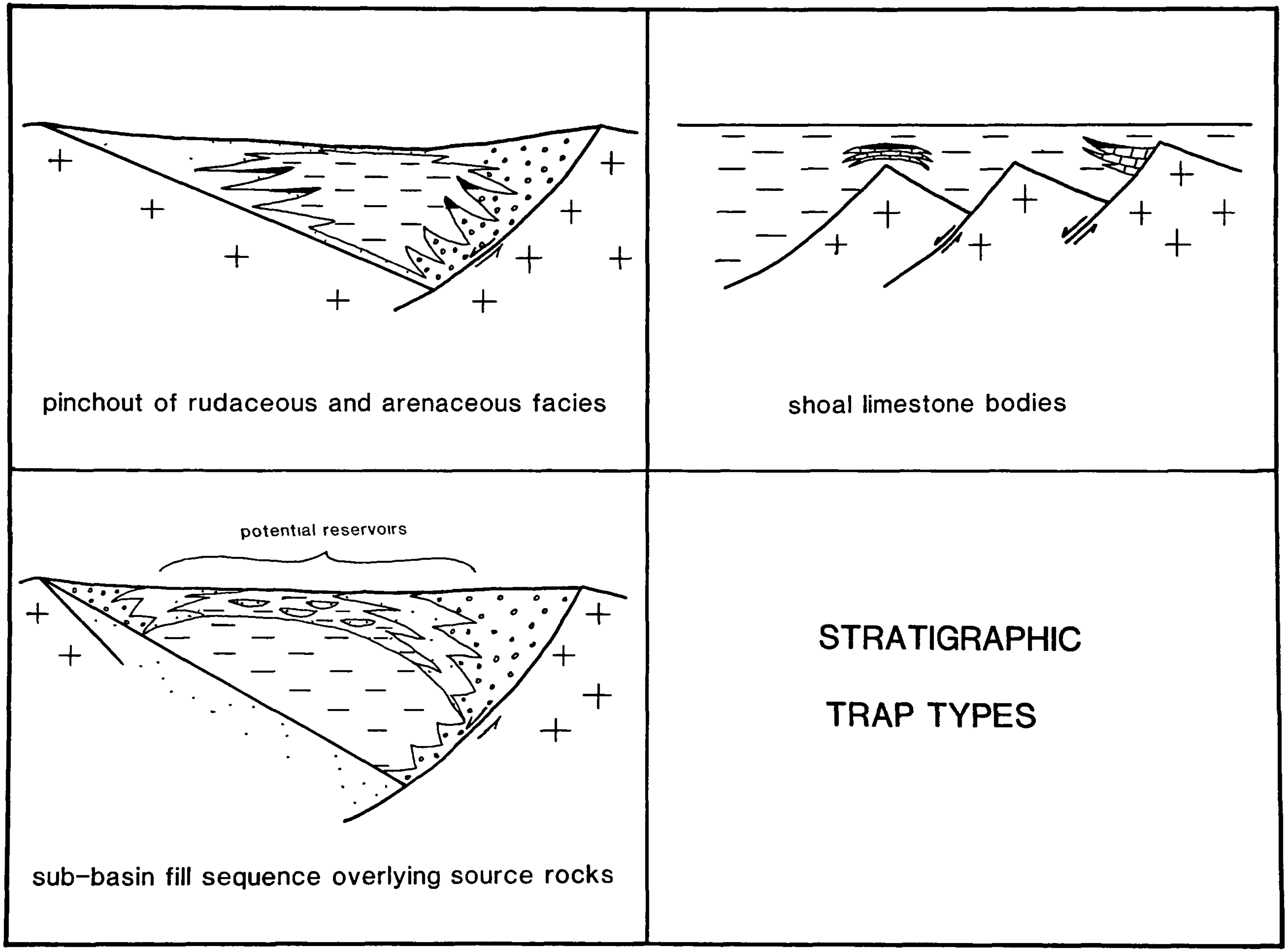

Figure 182. Stratigraphic trap types which may be present in Horton rocks of Cape Breton Island.

CHAPTER 7
CONCLUSIONS

In this study a systematic sedimentological approach is employed to establish the complex inter-relationships between tectonism, syndepositional structure, facies distribution and resource potential of the Late Devonian?/Early Carboniferous Horton Group of Cape Breton Island. Basic lithological and paleocurrent data are used to establish the basin geometries, structural style, positions of controlling margins and tectonic evolution during the various phases of Horton deposition. The style and spatial arrangement of facies reflect the nature and distribution of sedimentary environments which resulted from local and regional tectonic activity. The information is organized into a four-dimensional tectono-sedimentary analysis of the basin-fill sequence which serves as a framework for more detailed studies and comparisons.

The following conclusions summarize the important findings of this study:

1. The Late Devonian/Early Carboniferous portion of the Maritimes Basin was an intracontinental extensional basin primarily confined by fault-bounded basement blocks, which underwent multiple episodes of fault motions. This basin post-dated the Acadian Orogeny and pre-dated the Alleghenian Orogeny, both major compressional phases in the history of the Appalachian Orogen. The Cape Breton Island area was positioned at approximately 10-15° S. paleolatitude in a warm, arid climate. All observable structural evidence represents post-Horton overprinting due to Alleghenian compression, and syn-Horton deformation can only be interpreted from the sedimentary sequence.

2. Characteristics of the Horton Group basin-fill sequence are unlike those of known transtensive basin deposits, but are very like those of known fault-bounded distensive basins. During this period the Maritimes Basin is interpreted as a rift system, analogous to the "Fundy Basin Rift" of Belt (1960). It was dominated by vertical subsidence and resulting sedimentation as a result of extensional stresses at plate boundaries, rather than as a result of active mantle upwelling, doming and accompanying abundant volcanism. The Horton basin-fill sequence in this study area is characterized by a large total depositional area (10,000's km^2), near-vertical stacking of laterally continuous formations of moderate total thickness (2-3 km) in two related sub-basins, with no evidence of strike-slip control.

3. Many basement blocks presently at surface were not positive topographic features during Horton sedimentation, but were elevated along faults during Alleghanian compression. There were only two large sub-basins of deposition in the study area in Horton time, as illustrated by facies distributions and continuity, grain size and sediment dispersal trends. The sub-basins were approximately 100 x 50 km in size and were separated by a narrow zone of elevated basement between Cheticamp and Pleasant Bay. They are referred to as the western and northern Cape Breton sub-basins; the same facies and overall sequence occur in both. Facies distributions and paleocurrents indicate they were asymmetric half-grabens with opposed polarities of asymmetry. Alleghanian overthrusting altered the position of some sub-basin margins in western Cape Breton, and caused thrust repetition and thickening of the Horton sequence.

4. The Horton Group is underlain by metamorphosed basement or, locally, the Late Devonian Fisset Brook Formation volcanics and sediments of continental extensional derivation. It is overlain by the Viséan Windsor Group of marine carbonates and evaporites. In ascending order, it is composed of the Craignish, Strathlorne and Ainslie Formations. A total of ten facies assemblages can be defined which represent specific depositional settings of an alluvial/fluvial/lacustrine complex. Each includes several individual lithofacies which represent specific depositional environments. In turn, the ten assemblages combine to delineate four depositional systems which represent successive tectonic phases in the evolution of the receiving sub-basins. The distribution of these facies assemblages and depositional systems can be mapped with correlative paleocurrent data. Vertical and lateral variations in assemblages indicate the structural style and positions of controlling margins of the sub-basins through successive phases of tectonic evolution. Deposition was primarily influenced by fault-bounded margins and fault-induced subsidence within the two extensional sub-basins.

5. Horton deposition began with the C3 depositional system (lower Craignish) of distal braided stream and overbank pond deposits in a single broad basin with a general paleoslope toward the northeast. The C1/C2 depositional system (upper Craignish) comprises proximal braided stream and distal mudflat/playa sediments in an overall fining-upward sequence, and was deposited in two fault-bounded sub-basins with sediment dispersal away from fault-bounded margins. The succeeding S1/S2/S3/S4 depositional system (Strathlorne) of proximal fan-delta, medial delta/shoreline and distal open lacustrine

sediments was deposited in two fault-bounded half-graben basins with opposed polarities of asymmetry. The succession comprises multiple shallowing-upward sequences with sediment dispersal away from margins and toward sub-basin axes. The succeeding A1/A2/A3 depositional system (Ainslie Formation) of proximal alluvial fan, medial low-sinuosity fluvial and distal high-sinuosity fluvial sediments was deposited in the same asymmetric half-graben sub-basins. Multiple coarsening-upward sequences occurred near fault scarps and sediment dispersal was generally away from fault-bounded margins.

6. The overall sequence delineates the tectonic evolution of the Horton distensive rift system in the Cape Breton region. Localized bimodal volcanics were contemporaneous with and succeeded by fluvial sediments deposited in a single, broad, pre-rift sag basin (C3 depositional system). This was followed by faulting at the footwall margins to create two narrower fault-bounded sub-basins where fining-upward sequences (C1/C2 depositional system) developed as fault-bounded subsidence increased in intensity. The period of maximum subsidence and maximum extensional stress led to lacustrine conditions throughout the sub-basins (S1/S2/S3/S4 depositional system). These large tectonic lakes were deepest at the axes near the footwall scarps of the narrow, asymmetric tilt-block/half-grabens, and received abundant sediment input through delta progradation from the opposite hanging wall hinged ramp margins. Waning tectonic activity and fault-bounded subsidence allowed increased input and widespread dispersal of coarse, reddened fluvial/alluvial fan sediments (A1/A2/A3 depositional system) which overwhelmed the lakes and filled the sub-basins. Later fault-bounded subsidence probably occurred in Windsor time, but in a wider and more extensive basin.

7. The Strathlorne Formation represents the tectonic phase of most active fault-bounded subsidence and is characterized by stacked, coarsening-upward, shallowing-upward sequences attributed to tectonic control. Each coarsening-upward sequence represents instantaneous deepening due to a subsidence event and then gradual filling into shallow water as erosion and sediment input from the margins catch up. The sequences are numerous, thin, muddy and occur in bundles near the footwall scarp, but fewer, thicker and sandier on the hanging wall ramp of sub-basins. Paleocurrent data, especially from gravity-controlled underflow and delta progradation sediments, indicate paleoflow consistently toward sub-basin axes near the footwall scarps, effectively defining the sub-basin asymmetries. Bundles of thin sequences near the footwall approximately correlate to single

thick sequences on the hanging wall ramp, and are attributed to specific tectonic episodes within the major Strathlorne tectonic phase of maximum subsidence. Fault margin sediments of the Ainslie Formation (A1 alluvial fan facies assemblage) are also arranged in coarsening-upward, tectonically-controlled sequences, although this formation represents the waning phase of fault-bounded subsidence in Horton time.

8. The lower part of coarsening-upward sequences in the S1 open lacustrine facies assemblage is commonly dark grey laminated mudstone with abundant exinous and amorphous organic material, pyrite degradation of miospores, and little bioturbation. These characteristics indicate that, after a subsidence event, sediments were deposited in quiet, anoxic conditions in the hypolimnion of a stratified lake. These conditions were most common near sub-basin axes. Great depth is not necessary to develop these conditions, and the general thickness of the sequences (a few to a few tens of metres) suggests that the extensive Strathlorne lakes in the two sub-basins were relatively shallow most of the time.

9. Paleontological data support the interpretations of the Craignish and Ainslie sediments as predominantly fluvial and the Strathlorne sediments as predominantly lacustrine. The biota indicate that, during the main lacustrine phase, the lakes were stratified, with an oxidized epilimnion and a reducing hypolimnion. A low diversity biota of arthropods, bivalves, fish, and Botrycoccal algae inhabited the lakes. Two palynological zones of Tournaisian age (Tn_2, Tn_3) were identified for the upper Craignish, Strathlorne and Ainslie Formations, but no lower boundary of Horton deposition can yet be defined. A significant age gap may exist between preserved upper Ainslie deposits and the overlying Macumber Formation of the Windsor Group.

10. The Craignish Formation braided fluvial deposits, especially the C1 facies assemblage around Cape Breton Highlands near Baddeck and Cheticamp, have significant potential for paleoplacer gold deposits. A strong redox surface exists where the Macumber Formation overlies the A1 reddened alluvial fan facies assemblage and strongly resembles the classical Kupferscheifer-type copper deposits known from Europe and Africa. High copper, lead and zinc values have been recorded in the Baddeck and Whycocomagh areas, and the potential for significant deposits is excellent.

11. Lacustrine sediments of the S1 facies assemblage have significant hydrocarbon source rock potential with up to 7.63% T.O.C. in dark grey laminated mudstone at the base of coarsening-upward sequences, especially in the Baddeck area and much of the exposed

portion of the northern sub-basin. Botrycoccal algal material is present and R_0/T.A.I. values fall within the oil window over much of the study area. Sandstone and conglomerate of the S2 shoreline, S3 fan-delta, and A3 meandering fluvial facies assemblages have reservoir potential, with secondary porosity up to 39% and permeability up to 505 md (outcrop samples). These strata are interbedded with potential source rocks and overlain by the regional seal of the Windsor Group in many areas, especially the Mabou/Lake Ainslie area of the western sub-basin where oil has been recovered in the past, and the offshore portion of the northern sub-basin. Trap possibilities include structural fault and fold traps related to tilting of fault blocks and later overthrusting, and stratigraphic traps related to the pinchout of marginal facies into lacustrine mudstone and the presence of a basin-fill fluvial sequence over lacustrine mudstone. Although hydrocarbon potential is moderate, past exploration has not fully evaluated all possibilities.

APPENDIX I

Measured Section Locations
(see Figure 3 for map)

(copies of measured sections filed with
Dr. B.R. Rust, University of Ottawa, and
with author at ISPG, Calgary.)

WESTERN SUB-BASIN			
Outcrop/Drillhole Name	Latitude/ Longitude	1:50 000 NTS Sheet	Grid Location
1. Graham R.	45°51'30" / 61°26'00"		212790
2. Judique Intervale Bk.	45°55'30" / 61°26'40"		205870
3. S.W. Mabou R.	45°57'20" / 61°22'10"	11 F/14	261901
4. Melford Falls	45°52'10" / 61°16'30"		340804
5. Whycocomagh Roadcut	45°57'30" / 61°06'00"		474913
6. Argyle Bk.	45°59'30" / 61°02'30"		515945
7. Little Narrows #1	45°58'30" / 60°56'50"		591930
8. Little Narrows #2	45°59'00" / 60°56'30"	11 F/15	593940
9. Christmas Is. cemetery	45°57'00" / 60°46'00"		729921
10. Benacadie Point	45°54'20" / 60°43'50"		760855
11. Green Point	46°05'40" / 61°29'00"		171055
12. Imperial Mabou #1	46°01'20" / 61°24'20"		235975
13. S.E. Mabou R.	46°01'30" / 61°18'30"		310968
14. Saddler's Bk	46°07'30" / 61°17'50"		318089
15. Glendyer Bk trib	46°05'30" / 61°20'30"		284050
16. Schoolhouse Bk	46°07'00" / 61°15'40"	11 K/3	343079
17. Eastern Gulf #9	46°03'00" / 61°20'50"		281001
18. Little Narrows Hays R.	46°06'00" / 61°15'00"		354060
19. Imperial Inverness #3	46°06'30" / 61°11'50"		396076
20. Roseburn Bk	46°00'30" / 61°13'20"		378960
21. W. Ainslie Bk	46°05'20" / 61°10'40"		409055
22. St. Joe Lk Ainslie 137-1	46°07'50" / 61°08'40"		432103
23. St. Joe Lk Ainslie 137-2	46°08'30" / 61°08'40"		432116
24. Cooper/McFarlane Bk	46°11'20" / 61°06'30"		458167
25. Mt Pleasant Bk	46°12'50" / 61°07'00"		452194

26. Washabuck Bridge	46°00'10" / 61°52'30"		645963
27. Murdoch Paul's Bk	46°03'10" / 60°48'20"		699016
28. MacRae Bk	46°11'30" / 60°57'20"		578173
29. Yankee Line Ck	46°08'20" / 60°54'50"		612107
30. Red Head	46°06'20" / 60°43'20"	11 K/2	763072
31. Peter's Bk	46°09'00" / 60°46'00"		728135
32. Baddeck R	46°11'20" / 60°44'40"		743173
33. North Br. Baddeck R	46°11'30" / 60°46'40"		713180
34. Christopher McLeod Bk	46°13'30" / 60°42'10"		770215
35. McLeod's Point	46°14'30" / 60°36'00"		852234
36. Munro Point	46°15'20" / 60°35'55"		852253
37. Goose Cove Bk	46°15'40" / 60°38'00"	11 K/7	825259
38. North R	46°18'40" / 60°38'00"		822315
39. MacLeod's Bk	46°16'30" / 61°13'00"		374265
40. Big Bk	46°16'20" / 61°02'45"		505261
41. Pat's Bk	46°18'20" / 61°02'40"		506298
42. Long Marsh Gulch	46°22'00" / 61°03'30"	11 K/6	485364
43. Arsenault's Bk	46°22'30" / 61°03'30"		485375
44. Gallant R	46°23'00" / 61°03'20"		498385
45. Old Bridge Bk	46°25'30" / 61°04'50"		476431
46. Ruisseau des Basiles	46°29'50" / 61°03'00"		495510
47. Angus Lk Bk	46°31'30" / 61°02'00"	11 K/11	509542
48. Fisset Bk	46°36'20" / 60°58'55"		548630
49. Le Buttereau	46°40'30" / 60°57'25"		561709
50. Presqu'ile	46°41'15" / 60°57'45"	11 K/10	558722
51. La Bloc	46°41'40" / 60°57'20"		562729
52. Cap Rouge	46°42'45" / 60°56'30"		574749
53. Corney Beach	46°43'50" / 60°55'30"		585769

NORTHERN SUB-BASIN

Outcrop/Drillhole Name	Latitude/ Longitude	1:50 000 NTS Sheet	Grid Location
54. Mackenzie R. roadcut	46°49'25" / 60°49'50"		657875
55. Sugar Bk	46°48'50" / 60°47'30"		685865
56. Schoolhouse coast	46°50'30" / 60°47'40"		687894
57. Moore's Hill coast	46°51'00" / 60°46'55"		693904
58. Red R / Kerr's Pt coast	46°51'10" / 60°46'40"	11 K/15	697908
59. Eastern Bk	46°51'20" / 60°45'45"		711906
60. Gampo Abbey coast	46°52'30" / 60°45'20"		714930
61. Black Cliffs	46°52'40" / 60°44'50"		721936
62. Archies Bk	46°52'30" / 60°44'30"		725930
63. Beulach Ban Falls	46°50'00" / 60°36'20"		826885
64. Conglomerate Ck	46°49'20" / 60°37'40"		811871
65. Middle Aspy R	46°51'30" / 60°32'30"		875923
66. North Aspy R	46°53'40" / 60°32'50"		872944
67. Bay Road Ck	46°58'30" / 60°28'30"	11 K/16	923060
68. Salmon R	46°59'50" / 60°29'40		904078
69. Salmon R coast	47°00'10" / 60°29'10"	11 N/1	907082
70. Lowland Cove	47°00'50" / 60°37'20"		808095
71. Meat Cove roadcut	47°01'20" / 60°33'40"	11 N/2	855104
72. Black Point	47°02'40" / 60°32'10"		873109
73. Black Point Bk	47°01'10" / 60°31'30"		880099
74. Petro Canada et al St. Paul P-1	47°10'58" / 60°13'37"	11 N/1	099293

APPENDIX II

Palynological and Organic Matter Data

(from ISPG Reports 7-JU-88, 11-JU-88
by J. Utting)

KEY

FAC ASS:	facies assemblage sampled
LOC:	outcrop location number (See Appendix I)
ZONE:	palynological zone/subzone I = E. rotatus, II = V. vallatus, IIa = V. vallatus abstrusus, IIb = V. vallatus spelaeotriletes
TAI:	Thermal Alteration Index
ORG MATT:	approximate composition of organic matter (ratio of amorphous + exinous : woody + coaly)
DEP ENVIR:	general depositional environment indicated by organic matter.
(Botryc):	presence of Botrycoccus algal material

WESTERN SUB-BASIN					
FAC ASS	LOC	ZONE	TAI	ORG MATT	DEP ENVIR
A3	2	IIb	3-	1:2	lacust?
	16	"	3-	1:9	fluvial
	44	"	?	1:10	fluvial
	44	"	3?	1:9	fluvial
S1 (no shoreline influence)	2	IIb	3-/3	1:2	lacust
	2	"	2+	1:2	"
	3	"	3-	2:3	"
	3	"	2	2:3	"
	13	"	2	1:1	lacust?
	16	"	2	2:3 (Botryc)	lacust
	24	"	3/3+	1:2	"
	24	"	3/3+	1:2	"
	27	"	2	1:2 (Botryc)	"
	27	"	2+/3-	1:2	"
	27	"	3-	1:2 (Botryc)	"
	27	"	2+	1:2	"
	31	"	2	1:2	"
	32	"	2-	2:1	"
	33	"	2	3:2 (Botryc)	"
	33	"	2	2:3 (Botryc)	"
	33	"	2-	2:1 (Botryc)	"
	34	"	2-	4:1	"
	38	"	3-/3	1:1	"
	43	"	3+	1:2	"
	45	"	3-	1:1	"
	52	IIa	2-	3:1 (Botryc)	"
S1 (shore-line influence)	3	IIb	2	1:9	lacust
	6	II	4-	1:3	fluv/lacust
	13	IIb	2+	1:2	lacust
	16	"	3	1:9	fluv/lacust
	24	"	3-/3+	1:10	lacust
	27	"	3-	1:4	"
	33	"	2+/3-	1:9	"
	44	"	2/2+	1:4	"
	45	II	3/3+	1:3	"
	46	II	3+/4-	1:3	"
C3	44	?	3+/4-	1:3	fluvial

NORTHERN SUB-BASIN					
FAC ASS	LOC	ZONE	TAI	ORG MATT	DEP ENVIR
S1 (no shoreline influence)	55 61 63 66 67 69 71 73	IIb ? II IIb II IIa ? II	3- 3+ 3- 2/2+ 3 2 ? 3	2:3 (Botryc) 2:1 2:3 1:1 3:2 2:1 (Botryc) 3:2 1:2	lacust fluv/lacust lacust lacust anox lacust lacust " "
S1 (shore-line influence)	61 62 69 71	? II IIa ?	3+ 3- 3- ?	1:5 1:4 1:1 1:1	fluvial fluvial anox lacust lacust
C1	58 60 60 72	I I I ?	2+ 3+ 3- ?	1:2 1:4 1:4 1:2	fluvial " " "

APPENDIX III

Hydrocarbon Source Rock Characteristics

(supplied by P. McMahon and W. Smith
Nova Scotia Dept. of Mines and Energy)

KEY

FAC ASS:	facies assemblage sampled
LOC:	outcrop/drillhole location (see Appendix I for numbered locations)
TOC:	Total Organic Carbon, weight % (TOC > 0.5 = potential source rock)
S1:	indication of free hydrocarbon present in sample
S2:	indication of potential hydrocarbon which could be generated
T_{max}:	°C, temperature where S2 peak occurs ($T_{max} > 435°$ may indicate maturity)
OPI:	oil production index, $\frac{S1}{S1 + S2}$
HI:	hydrogen index, $\frac{S2}{TOC} \times 100$

WESTERN SUB-BASIN

FAC ASS	LOC	TOC	S1	S2	Tmax	OPI	HI
	Long Pond	0.77	0.00	0.01	350	0.80	1
	1	0.07	0.01	0.02	326	0.33	29
	1	0.18	0.01	0.06	436	0.14	33
	1	0.36	0.01	0.03	365	0.25	8
	2	0.34	0.00	0.14	443	0.00	41
S1	12	0.46	0.11	0.11	426	0.50	24
	12	0.33	0.04	0.00	311	1.00	0
	12	0.33	0.09	0.17	406	0.35	52
	12	0.40	0.02	0.08	443	0.20	20
	3	0.09	0.02	0.10	362	0.17	111
	13	0.11	0.00	0.00	257	0.00	0
	13	0.11	0.00	0.05	321	0.00	45
	24	0.33	0.04	0.24	451	0.14	73
	27	0.29	0.01	0.14	447	0.07	48
	27	0.32					
	27	0.13					
	27	0.27					
	27	0.41	0.03	0.13	443	0.19	32
	27	0.29					
	33	0.92	0.02	2.30	434	0.09	250
	Aberdeen Bk	0.31	0.01	0.04	284	0.20	13
	41	1.06	0.06	0.09	282	0.40	8
	41	0.64					
	45	0.05					
	47	0.22					
	49	0.33					
	49	0.32					
	52	2.23					
	52	0.43					
	Mull R	0.13	0.02	0.07	443	0.22	54
	"	0.37	0.09	0.17	436	0.35	46
	McIsaac Pt	0.06					
	"	0.27					
A3	"	0.34	0.03	0.05	436	0.38	15
	"	0.08					
	Twin Rock	0.23	0.08	0.20	439	0.29	87
	"	0.14	0.01	0.15	443	0.06	107
	Trout Bk	0.35	0.02	0.08	432	0.20	23
	"	0.16	0.02	0.07	438	0.22	44
	McKay Bk	0.09	0.01	0.10	434	0.09	111

NORTHERN SUB-BASIN							
FAC ASS	LOC	TOC	S1	S2	Tmax	OPI	HI
	61	2.15	0.44	1.03	470	0.30	47
	61	0.02					
	61	0.25					
	67	7.45	0.37	22.28	445	0.02	299
	67	5.25	0.42	22.25	439	0.02	423
	69	0.41	0.25	0.46	436	0.35	112
S1	69	7.63	1.51	37.69	448	0.04	494
	69	3.95	0.28	12.62	447	0.02	319
	Wreck Cove	2.02	0.25	0.86	458	0.23	43
	Jumping Bk	3.18	0.21	0.71	475	0.23	22
	71	1.49	0.26	0.74	440	0.26	50
	71	2.20	0.18	0.55	463	0.25	25
	71	0.99	0.06	0.04	416	0.30	4
	71	0.47					
	71	1.05					
	71	1.01					
	71	1.59	0.12	0.30	472	0.26	21
	71	2.56	0.80	1.10	467	0.42	42
	71	0.93					

APPENDIX IV

Hydrocarbon Reservoir Rock Characteristics

(supplied by P. McMahon from reports by
Berry and Wilson (1973) and Tillement (1973)
filed at Nova Scotia Dept. of Mines and Energy)

KEY

FAC ASS:	facies assemblage sampled
LOC:	outcrop location (see Appendix I for numbered locations)
POR:	porosity, %
PERM:	permeability, millidarcies

WESTERN SUB-BASIN

FAC ASS	LOC	POR	PERM
A3	McIsaac Point	20.2	38.20
	"	36.3	482.00
	"	39.0	285.00
	"	33.4	505.00
	"	14.7	0.47
	"	15.1	0.53
	North end Lake Ainslie	13.8	0.12
	"	25	89.00
	Mull River	18.1	0.37
	"	15.6	14.3
	15	10.1	0.01
	15	5.4	0.01
	15	15.0	
	12	3.0	0.01
	12	4.7	0.10
	13	6.5	0.06
	13	19.3	0.88
	13	21.7	1.00
	13	5.0	
	13	15.0	
	24	15.5	1.4
	West of McFarlane Brook	20.0	
	"	10.0	
	"	15.0	
A2	South tip of Lake Ainslie	1.1	0.01
	"	15.0	
	"	20.0	
	1	22.2	225.00
	1	19.4	4.84
	1	6.1	0.24
	1	6.0	0.24
	1	19.1	3.75
	2	20.4	63.00
	3	2.9	0.01
	3	2.3	0.01
	3	22.1	108.00
	3	21.5	202.00
	3	5.0	
	3	22.0	102.00
	3	16.2	6.60
	3	19.2	17.60
	3	16.5	2.17
	3	15.4	7.48
	3	4.9	0.76

A1	South of Graham River 5	16.8 15.0	11.20
S2	West side of Lake Ainslie " Skye River area " " " Mount Young South end of Mount Young " " " 1 13 " "	1 5 15 15 10 10 16.5 5 10 10 10 15 20 20 10	0.01
C1	Mount Young " Between SE and SW Mabou R	15 15 15	
C2	South end of Lake Ainslie West of Skye River " " " " 3 13 13	10 10 15 15 10 10 15 15 15	
C3	Between SE and SW Mabou R South of Graham River 24	10 10 5	

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