LATE QUATERNARY GLACIO-MARINE DEPOSITS OF THE STITTSVILLE AREA, NEAR OTTAWA, CANADA

by

Richard James Cheel

A thesis submitted to the School of Graduate Studies in partial fulfillment of the requirements for the degree of M.Sc. in Geology

UNIVERSITY OF OTTAWA

OTTAWA, ONTARIO, 1979

ABSTRACT

A linear topographic feature, approximately 10 km long, trends NW-SE through the town of Stittsville, Ontario. The sediments of the ridge are interpreted as subaqueous outwash deposited from a subglacial or englacial meltwater tunnel emptying into a re-entrant of the Wisconsin glacier front, which terminated in the Champlain Sea. The ridge formed by the coalescence of subaqueous outwash fans as the ice retreated to the north. Sedimentological variation along the ridge is attributed to variation in the morphology of the re-entrant. Re-entrant morphology may have been controlled by a crevasse system within the glacier or by preferential melting of one ice wall due to the accumulation of mist on the downwind side of the inlet.

Subaqueous outwash fans deposited in a narrow glacial inlet are thought to form an upward fining sequence from boulder gravel to silt and clay rhythmites. The sequence was intensely deformed by the melting of buried ice and the walls of the inlet, and the resulting landform is narrower and steeper than unconfined subaqueous outwash. However, in the Stittsville area the original form was modified by wave action during the regression of the Champlain Sea.

Subaqueous outwash deposited in asymmetric re-entrants into the ice front yield paleocurrent directions which are deflected away from the relatively straight ice wall. The

sedimentology of these deposits appears to be controlled by the degree of asymmetry and the width of the inlet. In relatively narrow asymmetric inlets the deposit is dominated by massive or laminated sand-filled channels with little interchannel cross-stratified or fine sediments preserved. In a broad asymmetric embayment the facies are asymmetrically distributed across the deposit, with thick accumulations of fine sediments preserved in the downcurrent direction. These thick accumulations gave rise to widespread soft-sediment deformation in slumps on the fan surface and above buried blocks of melting ice.

A sequence of soft-sediment deformation structures was observed in exposures in the northern portion of the ridge. In upward sequence it consists of convolute stratification, ball and pillow structures and dish structures. It is believed that the sequence formed as a result of water added to a relatively impermeable unit (silty fine sand) due to consolidation of sediment below the structures and/or from buried ice.

The stages in the development of the sequence of structures include: 1) Local fluidization of sediments with low permeability, forming convolute stratification with disrupted anticlines acting as vertical fluidization channels; 2) Upward penetration of more cohesive overlying sediments by the vertical channels, forming detached sediment masses which sank downwards as ball and pillow structures; 3) Silty fine sand which had passed upward through the vertical channels accumulated above the ball and pillow structures and may have moved down-slope as a liquefied sediment flow. Dish structures formed by relatively uniform dewatering during flow, and in some cases were compressed in the downslope direction to give oval plan forms with long axes normal to the local paleocurrent direction.

A significant feature of this structural sequence is that ball and pillow structures are exotic, so that any chronologic information which they contain is not necessarily representative of the surrounding sediments.

ACKNOWLEDGEMENTS

I thank Dr. B.R. Rust for supervising the thesis, particularly for his encouragement and time spent in the field and critically reading the manuscript. Appreciation is also expressed to Dr. H.M. French, who acted as supervisor during Dr. Rust's absence, and who, along with Dr. P. Worsley and Mr. S.H. Richard, visited the study area and provided stimulating discussion, to Mr. M. Hayward for his companionship in the field, to Maureen Czerneda for typing the thesis, and to Jack Whorwood (McMaster University) and Edward Hearn (University of Ottawa) for their assistance in preparing the photographs. Thanks also to the students, faculty and staff in the Department of Geology at the University of Ottawa for the many and varied contributions to this work and to making life in Ottawa pleasant. Finally, a special thanks to Charlotte and my parents for their support, and to Celine for being born, making the time spent writing the thesis much more enjoyable.

Financial support for 1977-78 was provided by Dr. B.R. Rust, from a National Research Council grant, and for 1978-79 by the Ontario Ministry of Colleges and Universities, in the form of an Ontario Graduate Scholarship.

TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

I. INTRODUCTION

The nature of glacial meltwater deposition in the subaqueous proglacial environment is not well understood. However, the Glacial Map of Canada (Prest et al., 1967) depicts over 25% of the total surface area as having been inundated by fresh or marine bodies of water during retreat of the Wisconsin glaciers. As well as these large bodies of water one must not forget the likelihood of relatively short-lived local ponding of meltwater in contact with the glacier front. It is surprising, then, that this environment was not given sufficient emphasis in Canada until Rust and Romanelli (1975) discussed the formation of eskers by subaqueous deposition from glacial meltwater during the Late Quaternary in the Ottawa area.

Another aspect of esker deposits which has been underemphasized is the sedimentological variation observed along the length of individual eskers. Such variations have been explained by many workers simply as a response to changing depositional controls during glacial retreat, without attempting to explain reasons for change. It is the present author's belief that without these explanations the understanding of the evolution of such landforms is far from complete.

Exposures in sand and gravel pits along the Stittsville Ridge (figure 1-1) provided an opportunity to study the sedimentology of a Late Quaternary glacio-marine deposit in relation to its geomorphic expression. A major goal of this study is to describe sediments exposed along the length of the Stittsville Ridge and to reconstruct the paleoenvironments in which they were deposited. The Ottawa area is well suited to such a study because its Late Quaternary history has been extensively studied (Johnston, 1917; Gadd, 1961, 1963, 1977; Richard, 1975, 1978; Richard et al., 1977). This particular ridge is of interest because it formed during the retreat of the Wisconsin Laurentide Ice Sheet, which in this area terminated in a standing body of water, as indicated by uplift curves (Rust and Romanelli, 1975; Rust, 1977).

A second, though equally important goal is to describe soft-sediment deformation structures exposed in excavations into the northern portion of the ridge and to offer an explanation for their formation. The structures of interest include convolute stratification, ball and pillow structures and dish structures. All three have been reported in sediments deposited in a wide range of environments, the common characteristic of which is the predominance of rapid rates of sedimentation. Of the mechanisms which are believed to be responsible for the formation of some or all of these structures cryostatic pressure, differential loading, current drag and fluid escape are the most frequently cited.

Within sediments of the Stittsville Ridge these structures have been observed in a vertical sequence consisting of (from the base) convolute stratification, ball and pillow structures and dish structures. The characteristics of the structures and the sequence in which they occur strongly supports the hypothesis that they formed in response to rapid dewatering.

THE STUDY AREA

Location

The Stittsville Ridge is located approximately 26 km to the southwest of the centre of Ottawa (figure 1-1). It is a linear feature of low relief, trending NW-SE, the town of Stittsville occupying its midpoint. The excavations which are discussed in the thesis are accessible from Carlton County Road #5, which intersects highway 7, 17 4 km northwest of Stittsville.

In all, seven sand and gravel pits were studied for the paleoenvironmental reconstruction of the ridge. The locations of these pits are shown in figure 1-1, each designated by a reference number.

Most previous workers who have studied the deposits of the ridge have referred to pits by the surnames of the owner of the excavating company working at that site. This practice has led to confusion because in some cases the properties have changed hands. Furthermore, many companies are currently excavating at more than one location in the Ottawa area. Accordingly, a completely new nomenclature is used here, while a list of the names used by other workers is included for comparison (Table 1-1).

Figure 1-1. Location map. The locations of sand and gravel pits referred to in the thesis are marked "X" on the map and are identified by their respective reference number. The dashed line is the limit of the "Fluvioglacial facies" as mapped along the Stittsville Ridge by Richard *et al.* (1977).

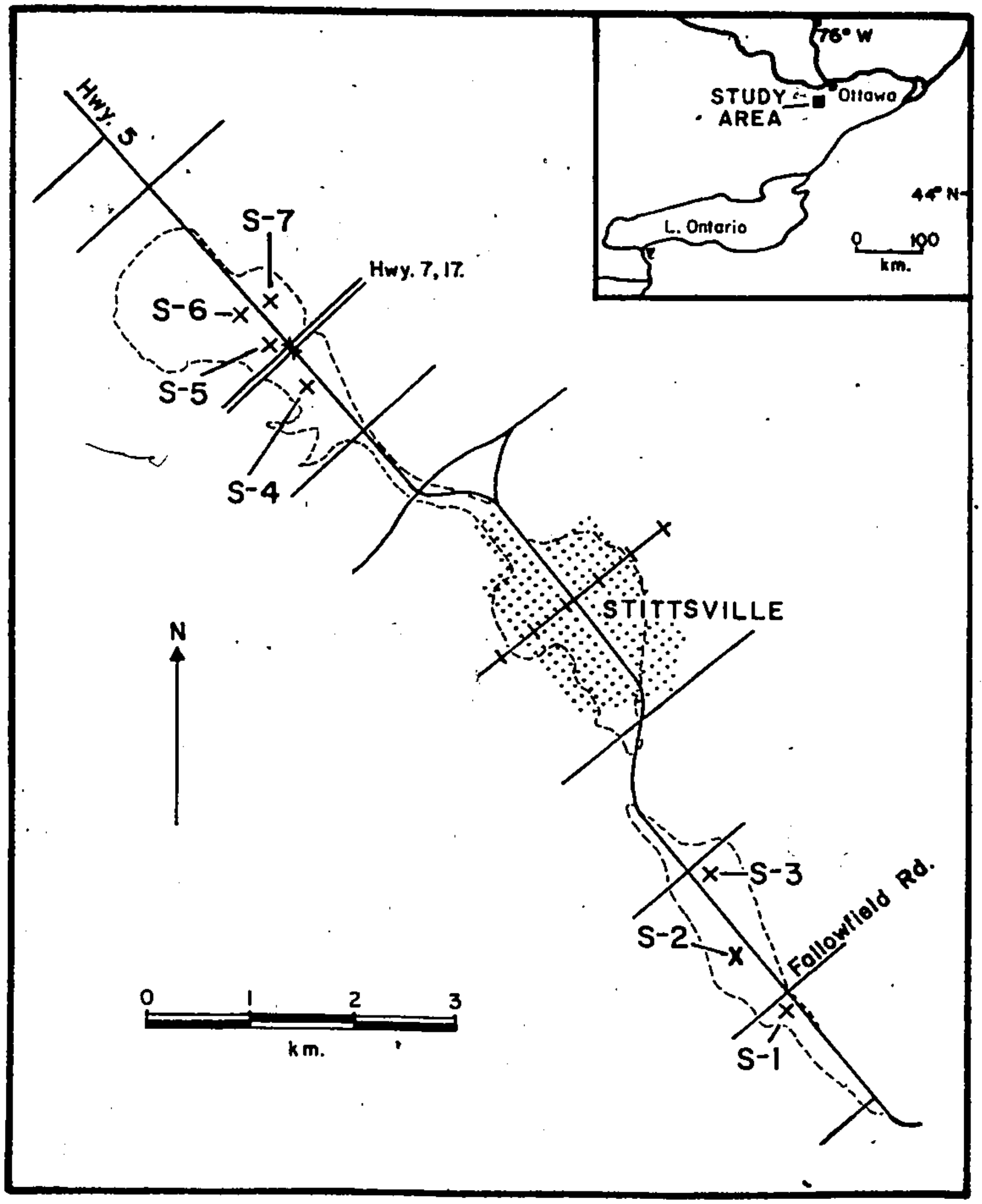

Hwy. 5
S-7
Hwy. 7, 17
S-6
S-5
S-4
STITTSVILLE
N
S-3
S-2
Fallowfield Rd.
S-1
0 1 2 3
km.
76° W
STUDY AREA
Ottawa
44° N
L. Ontario
0 100
km.

Table 1-1. Names used in previous theses and publications to identify sand and gravel pits along the Stittsville Ridge.

I.D. number used in the present thesis	Name used by earlier workers	Thesis or publication
S-1	Mulligan's Pit	Hayward, 1979
S-2	Stanley Corners Sand Pit	Thibault, 1974
S-4	Rump's Pit	Laport, 1970
S-5	Burke's Pit	Smith, 1970 Rust and Romanelli, 1975
	Bradley's Pit	Rust, 1977
S-6	Rump's Pit	Laport, 1970 Rust, 1977 Rust, 1978 Gadd, 1978
S-7	Spratt's Pit (Huntley)	Rust and Romanelli, 1975 Rust, 1977

Only three pits contained suitable exposures of the various soft-sediment deformation structures. These were sites S-4, S-5, and S-6 (figure 1-1).

Quaternary Geology

In the Ottawa area unconsolidated Quaternary sediments overlie predominantly carbonate rocks of Paleozoic age (Wilson, 1946). In general, these surficial deposits can be organized into four stratigraphic units (Rust, 1977, p. 175):

(4) Post-marine deposits: terrestrial sediments deposited after regression of the Champlain Sea.

(3) Fossiliferous sediments: includes clays, silts, sands and gravels which were deposited during the occupation of the Ottawa valley by the Champlain Sea between approximately 13,000 and 10,000 years B.P. (Richard, 1978). These are the most extensive surficial deposits in the Ottawa area.

(2) Nonfossiliferous sediments: stratified sands and gravels deposited from glacial meltwater. This unit is not uniformly distributed over the Ottawa area but is exposed in several distinct ridges (Rust and Romanelli, 1975).

(1) Till: unsorted diamict deposited directly from glacial ice. If the basal contact is observed to be overlying bedrock, this unit may be differentiated from flow tills, which may be interstratified with, or overlie, the non-fossiliferous unit. Exposures of basal till were not observed during the present study.

The bulk of the sediments making up the Stittsville

Ridge are unit 2, nonfossiliferous stratified deposits, which are overlain unconformably by a thin unit of fossiliferous sediments belonging to unit 3 (above). Johnson (1917) mapped the ridge as a marine deposit and interpreted the underlying sediments as fluvioglacial, whereas Gadd (1963, p. 2) interpreted the underlying deposit as moraine. The mapping of the ridge as a marine deposit was based on the observation of fossiliferous sediments which were exposed in relatively shallow excavations. More recent exposures have shown that these marine sediments occur extensively along the ridge as a relatively thin cover (less than 2 m), locally thickening to more than 9 m where they fill irregularities in the surface of unit 2. In light of these new exposures the ridge has been mapped most recently as a fluvioglacial landform (Richard, 1976 and Richard et al., 1977).

Rust and Romanelli (1975) and Rust (1977) interpreted the sediments within the northern portion of the Stittsville Ridge, and other ridges in the Ottawa area, as deposited from glacial meltwater on subaqueous fans at the ice front. This interpretation was based largely on regional uplift curves which showed that the ice front terminated in the Champlain Sea at this time. Additional evidence was derived from the sedimentological similarity between the sediments within the ridge and those reported in ancient submarine fan deposits. These include steep-sided channels filled with massive sand, units of climbing ripple drift and an abundance of dish structures and ball and pillow structures, which are believed to have formed by rapid dewatering of the sediment.

Gadd (1978) questioned Rust's (1977) ice-proximal interpretation of a unit of intensely deformed sediments within the ridge. Gadd's objection was based on the occurrence of marine fossils in ball and pillow structures within the unit. Furthermore, the fossils yielded a radiocarbon date indicating that the organisms lived at a time when the glacier front was well to the north of Stittsville. This led Gadd (1978) to suggest that the fossiliferous ball and pillow structures and associated sediments are strictly marine in origin, presumably deposited beyond the direct influence of glacial meltwater. Observations made by the present author suggest that the fossiliferous sands were emplaced into underlying unfossiliferous subaqueous outwash sediments at a time well after glacial retreat.

METHODS

The sediments of the Stittsville Ridge have been investigated to determine its depositional environments and the mechanisms of soft-sediment deformation which occurred abundantly in its northern portion.

The bulk of the information on the sedimentology of the Stittsville Ridge was acquired during the measurement of sections at sites S-1, S-3, S-4, S-5 and S-6. These were drawn in the field by visually correlating vertical sections measured at 10 m intervals along the exposure. Where facies changes were complex and tighter control was required this interval was reduced. A datum for each section was obtained using an abney level, for which elevations were determined by aneroid barometer.

Sediment textures were determined by visual comparison in the field. More quantitative descriptions of sediments within the Stittsville Ridge are given in Laport (1970), Smith (1970) and Thibault (1974). Primary and secondary sedimentary structures and fossil species were also identified, and their positions indicated on the sections. Paleocurrent directions were measured by Brunton compass and were analysed by the method described in Curray (1956). For each set of data the following are given: 1) $\bar{\theta}$, the resultant (mean) vector; 2) N, the number of observations; 3) L, the magnitude of the resultant vector in percent (Curray, 1956, p. 119); 4) p, the probability of obtaining a greater amplitude by pure chance combinations of random phases (Curray, 1956, p. 125).

To facilitate the description and interpretation of the Stittsville Ridge it has been divided into three areas which appear to be morphologically and sedimentologically distinct (figure 1-1). These are: 1) the southernmost part of the ridge (site S-1); 2) the area near Stanley's Corners (sites S-2 and S-3); 3) the northern part of the ridge (sites S-4, S-5, S-6 and S-7). It is important to note that these divisions are rather arbitrary, and that a gradational transition in morphology and sedimentology probably occurs between each area.

In the description of site S-1 the detailed measured section is presented within the text of the thesis to show the facies relationships at that site. The measured section at site S-3, however, was rendered of little use when new, more extensive exposures became available following the 1978 field season. The new exposures were examined in detail and the information was added to that collected earlier. Site S-2 was badly degraded during 1977 and 1978 although some information on its sedimentology was available in Thibault (1970). Within the text (Chapt. II) a generalized stratigraphic section has been constructed for ~~the~~ portion of the ridge near Stanley's Corners.

Of the sites representing the northern part of the ridge, sites S-4, S-5 and S-6 were examined in the field. While the exposures at S-7 were very badly degraded, considerable information was available for site S-7 in Rust and Romanelli (1975) and Rust (1977). This information was compiled with that for the other sites along the northern portion of the ridge to produce a generalized section for that area. Measured sections for sites S-4, S-5 and S-6 are presented in Appendix I.

II. A PALEOENVIRONMENTAL RECONSTRUCTION OF THE STITTSVILLE RIDGE

In this chapter of the thesis is presented a description of the morphology and sedimentology of the Stittsville Ridge and an interpretation of its paleoenvironmental history. The sedimentology of each part of the ridge (see page 9) is considered separately while at the end of the chapter the interpretations are summarized into a discussion of the history of the entire ridge.

To facilitate the interpretation of the Stittsville Ridge it will be useful to examine first the literature concerning the various environments in which eskers are formed and the criteria by which they may be identified.

ESKERS AND THEIR DEPOSITS - PREVIOUS WORK

Glacial meltwater deposits - general

Three types of glacial meltwater deposits are frequently described in the literature: kames, eskers, and outwash plains. Both of the terms "kame" and "esker" are defined as morphologic entities formed by glacial meltwater deposition. Neither term, by itself, bears any more specific genetic connotation. Outwash plains, however, have very specific genetic implications.

The term "kame" is most commonly used to refer to isolated mound-shaped deposits of stratified sand and gravel (Embleton and King, 1968; Price, 1973; Bloom, 1978). These are formed where glacial meltwater deposits its sediment load as flow expands on leaving the confines of a channel or tunnel. The site of deposition may be a small pond on, or cavern in, the ice or at the margins of the glacier. Because of the ice-contact environment kame sediments are characteristically intensely deformed by ice melt. Modifications of the term kame, which denote a specific genetic type, include: kame terrace, formed at ice-marginal lakes between the glacier and the walls of an adjacent valley; kame delta, formed in the subaqueous proglacial environment; moulin kame, formed in hollows on the ice (Bloom, 1978).

Eskers differ morphologically from kames in that they are narrow, often sinuous, linear ridges of stratified sand and gravel which generally trend parallel to the last direction of ice advance (Embleton and King, 1968 ; Price, 1973; Sugden and John, 1976; Bloom, 1978). They may be formed as a result of: 1) deposition within glacial meltwater conduits, their form being inherited from the shape of the conduit itself after glacial retreat; or 2) deposition at the ice margin where a conduit empties into a standing body of water, the deposit taking a linear form because the site of deposition at the mouth of the conduit retreated as the glacier retreated. In some cases this second type of esker is made up of a series of mounds and is referred to as a "beaded esker", which may be considered as an intermediate form between true kames and eskers. A more detailed discussion of the types of eskers and their sedimentology will be presented later.

The third important type of glacial meltwater deposit, the outwash plain, is formed by the coalescence of outwash fans deposited by meltwater streams in front of the terminal moraine of a glacier (after Gary et al., 1972). This type of landform may be readily distinguished from eskers and kames by its considerably lower topographic expression, greater lateral extent and lower incidence of ice-contact deformation.

From the discussion above it is apparent that the three types of glacial meltwater deposits have distinct morphologic characteristics, although a gradation probably exists between kames and eskers formed in similar environments. In the section that follows the various possible environments in which eskers form are described and the criteria for their recognition are discussed.

Esker-forming environments

Although eskers are known to have formed by deposition in a number of distinct environments no complete genetic classification is available. The simplest genetic classification is based on the recognition of eskers deposited in each of the environments described above, namely in conduits and at the ice margin in a standing body of water. Eskers formed by deposition in either environment have distinct morphological, structural and sedimentological characteristics which allow them to be differentiated in the field. It is, however, possible to subdivide these further into more specific genetic types.

Three distinct types of in-conduit environments may be identified as possible sites of deposition (Banerjee and McDonald, 1975; Price, 1973): 1) open channels, 2) tunnels flowing full, and 3) glacial lakes or ponds which occur along the conduit (which may merely be a widening of the channel.) These can in turn be classified according to the position of the conduit, whether they are on, in or under the glacier. All three of these esker-forming environments may develop in subglacial and englacial positions while open channels and glacial lakes may also develop in a supraglacial position.

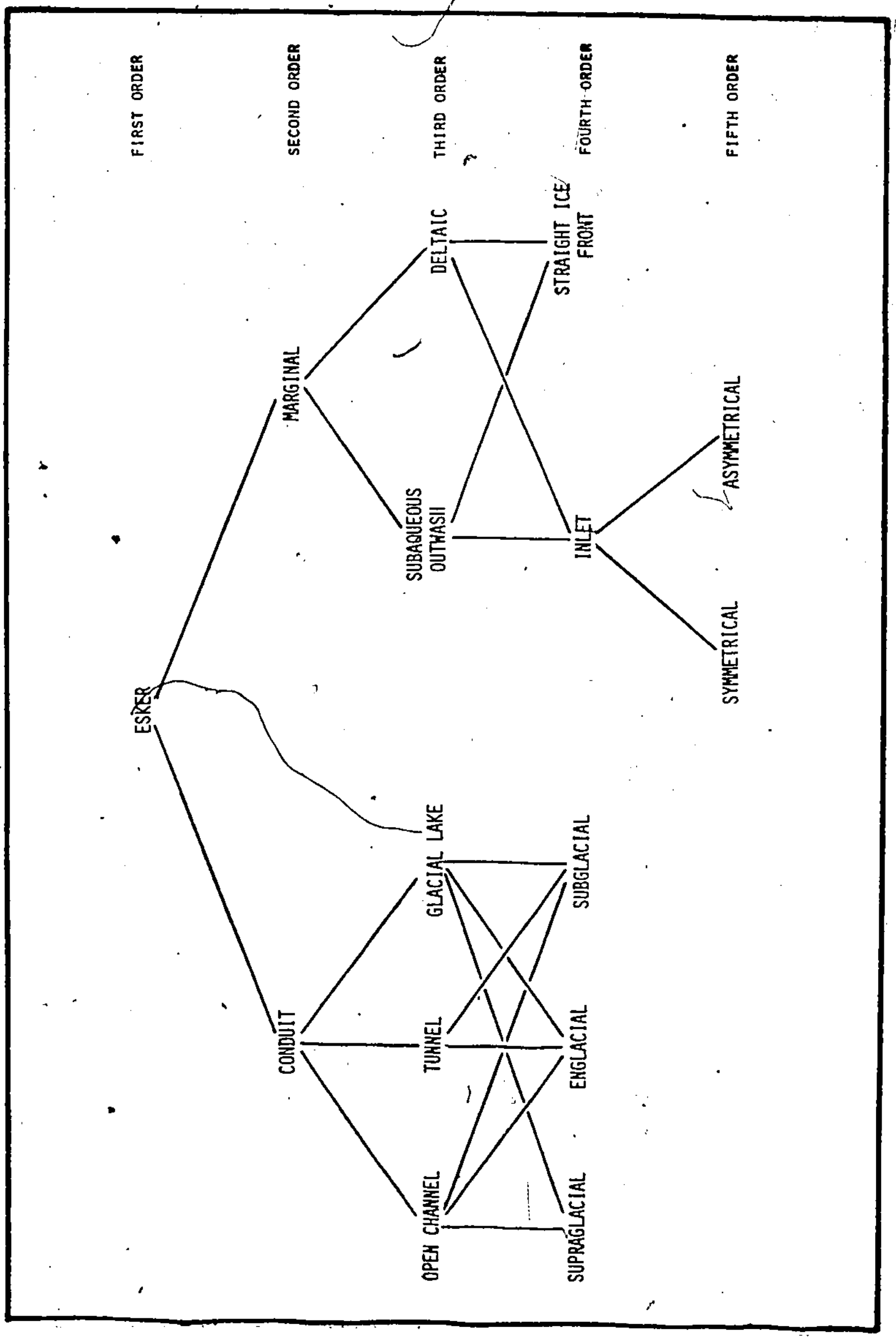
Esker formation in an ice marginal position can also take place in two distinct environments: 1) on a subaqueous outwash fan (Rust and Romanelli, 1975; Rust, 1977) at sufficient depth that the ambient water surface has no effect on the fan, and 2) on the subaerial and subaqueous portions of a delta, in shallower water (Price, 1973). The fan or delta may also vary in response to variations in the morphology of the ice front at the site of deposition. Deposition can take place at a straight ice margin or where a glacial meltwater conduit terminates at a water-filled re-entrant into the ice front. Such re-entrants may be marginal crevasses (or crevasse complexes) (Price, 1973; Rust and Romanelli, 1975) or supraglacial conduits which became flooded for some distance up-ice from the glacier terminus. Eskers formed by deposition within re-entrants may vary further depending on the form of the re-entrant, which can range from a narrow symmetrical inlet

(Price, 1973) to a broad asymmetrical embayment (Rust and Romanelli, 1975). The symmetry or asymmetry could be controlled by the orientation of crevasses or crevasse systems (Rust and Romanelli, 1975). Alternatively, Lundqvist (1955) suggested that symmetrical inlets may evolve into asymmetrical embayments in response to variations in the rate at which the ice walls melt. The process involves the concentration of fog on the down-wind side of the inlet, reducing the flux of solar radiation onto the glacial ice. As as result the ice wall on the up-wind side of the inlet will melt at a faster rate than on the down-wind side, causing the inlet to become increasingly asymmetrical over time, providing the prevailing wind direction is constant.

Figure 2-1 is an attempt to express the diversity of esker-forming environments, providing a basis for their detailed classification. An attractive aspect of this scheme is that any level of interpretation may be made, depending on the requirements of the project and the data available. For example, in geomorphic mapping it is often sufficient to make a first order interpretation of a landform simply as an esker. In a project which aims to predict the distribution of various grades of sand and gravel, however, a fifth order interpretation may be necessary.

The classification of an individual esker is complicated by the fact that many different combinations of specific environments may have been involved in its formation. For this reason it is impractical to establish a set nomenclature for

Figure 2-1. An ordered arrangement of esker-forming environments.


FIRST ORDER
SECOND ORDER
THIRD ORDER
FOURTH ORDER
FIFTH ORDER
ESKER
MARGINAL
CONDUIT
DELTAIC
SUBAQUEOUS OUTWASH
GLACIAL LAKE
TUNNEL
OPEN CHANNEL
STRAIGHT ICE FRONT
INLET
SUBGLACIAL
ENGLACIAL
SUPRAGLACIAL
ASYMMETRICAL
SYMMETRICAL

eskers. However, the hierarchy of types in figure 2-1 may be used as an outline for the desription of esker-forming environments at the various levels of interpretation.

Environmental interpretation of eskers

The environmental interpretation of eskers proposed here follows the classification scheme outlined above. Thus the first decision is whether the landform in question is an esker or not. This decision is not always straightforward, especially when the landform has been modified by wave action.

The criteria for distinguishing eskers formed by deposition in conduits and at the submerged ice margin are listed in Table 2-1. Eskers formed in conduits are generally continuous, steep-sided ridges whereas marginally deposited eskers may consist of a number of isolated mounds or "beads", and are flat topped with more gently sloping sides.

Because, by definition, glacial meltwater conduits must be in contact with ice walls, the sediments they deposit are intensely deformed by melting of the walls along the flanks of the ridge. High angle reverse faults (McDonald and Shilts, 1975) and normal faults (Boulton, 1972), both striking parallel to the trend of the ridge are common in this type of esker. Shilts and McDonald (1975) also found that sediments deposited on ice marginal fans are locally affected by large normal faults which develop as blocks of ice, buried by the fan, melt out.

Sedimentologically these two types of eskers are significantly different. With few exceptions fine sediments (silt

Table 2-1. Criteria for the differentiation between eskers formed in conduits and at the ice margin in a standing body of water. (See text for specific reference.)

Conduit	Marginal
- continuous linear steep-sided ridge	- more gently sloping sides and generally flat-topped, may consist of isolated "beads"
- extensive evidence of ice melt, especially along the flanks	- evidence of ice melt only local
- generally lacking in silt and clay	- a range of grain sizes present, from boulder gravel to silt and clay
- low to moderate variability of paleocurrent directions.	- high variability of unidirectional paleocurrent directions
- flow till uncommon	- flow till commonly preserved

and clay) are rarely deposited in a glacial meltwater conduit, because they are transported in suspension to the glacier terminus. On a marginal fan, however, a complete range of grain sizes may be deposited, from boulder gravel at the apex (Rust, 1977) to silt and clay rythmites on the most distal portion of the fan (De Geer, 1940). Banerjee and McDonald (1975) found that paleocurrent data are also useful in differentiating these two environments. Eskers believed to have been formed in conduits yield paleocurrent data which are considerably less variable that those collected from assumed fan deposits. The strength of the vector mean of paleocurrent data derived from two marginally deposited eskers were 0.49% and 0.59% while the value obtained from tunnel deposits was 0.80% (Banerjee and McDonald, 1975).

Flow tills, mixtures of till and water capable of flowing down relatively gentle slopes (Boulton, 1968), commonly occur interstratified with marginal fan deposits (Evenson et al., 1977; Hicock and Dreimanis, 1978). Boulton (1972) also reported flow tills interstratified with supraglacial open channel deposits and overlying sediments deposited in englacial and subglacial conduits. However, more common and more extensive preservation of flow tills would be expected in the ice marginal environment. There flow tills are transported across the fan, away from the mouth of the conduit where their preservation potential is low, and become interstratified with the more distal fan sediments. Flow tills which are deposited in, or near the mouth of conduits have a greater probability of being modified, or totally removed by high water discharges.

The interpretation of third and fourth order in-conduit esker-forming environments requires more information on the sedimentology and structure of the esker. Shaw (1972) suggested that an esker formed by deposition in an open channel would exhibit a fining of sediment away from the centre, where coarse channel deposits would be preserved. These flanking fines include fine sand, silt and clay deposited during overbank flooding from the main channel. In tunnels flowing full one would not expect to find this lateral fining because of the much more uniform hydraulic conditions across the conduit and the impossibility of overbank flooding. McDonald and Vincent (1972) simulated sedimentation in tunnels flowing full by passing water over a sand bed contained within a pipe. They suggested that the primary sedimentary features diagnostic of sediments deposited in natural tunnels flowing full may be: 1) dune foreset lamination which is tangential at the base and tends to become finer down the foreslope; 2) a discontinuity in the upper part of the foreslope due to slumping at the top of the dune foreset; 3) dune stoss lamination commonly preserved; 4) parallel lamination dipping gently downstream formed by the migration of low bed waves. While not all of these features have been reported in eskers they are potentially useful diagnostic criteria. McDonald and Vincent (1972) also point out that the preservation of backset bedding, formed by the migration of antidunes over an active bed will not occur in tunnel deposits because of the absence of a free water surface. Banerjee and McDonald (1975) suggested that there is less variability of paleocurrent directions in tunnel deposits than in open channel deposits where limited lateral migration of the active channel may occur between the ice walls (Shaw, 1972).

Lewis (1949) described an esker actively forming in a subglacial pond, along a meltwater channel. The ridge itself was flat-topped, like marginally deposited eskers, but had steep sides. The sediments within the esker, although poorly exposed, were reported to consist of approximately horizontal topsets overlying well-developed foreset beds. Shaw (1972, p. 39) reported similar foresets in a Quaternary deposit consisting of "planar beds of alternating coarse and fine gravel dipping at 30°". Although not yet reported in this type of deposit, one would also expect the development of topset beds. It should be noted that within the confines of a small englacial water body the deposit may not develop as extensively as fans formed by subaqueous deposition at the ice margin.

The identification of eskers formed by deposition in conduits in supraglacial, englacial and subglacial positions has been considered by Boulton (1972), although he did not distinguish between open channel and tunnel deposits. Both supraglacial and englacial conduits are, by definition, underlain by ice. Boulton (1972) observed that when an ice floor melts the bedding within the core of the resulting esker is deformed into an anticlinal structure. Price (1966) noted that a mound of ice may remain beneath the centre of a meltwater deposit after it has been lowered to the land surface. When the mound melts the centre of the ridge subsides, forming a trough along its length. Internally this trough may be reflected in the synclinal flexing of beds in the centre of the esker. Shilts and McDonald (1975) noted that large normal faults tend to develop at the site where saucer-shaped mounds of glacial ice become buried by esker sediments. Boulton (1972) found that the bedding within the core of subglacially deposited eskers remains largely undisturbed. He also noted that eskers of this type generally lack an underlying till, due to removal by erosion while the flow in the conduit was in contact with the subglacial bed. It is possible, however, that a basal till could be thick enough to resist complete removal by erosion. Till is common beneath englacially and supraglacially deposited eskers.

The distinction between eskers formed by deposition in englacial and supraglacial open channels and lakes is possibly the most difficult to make. The absence of flow tills interbedded with the meltwater deposits may be indicative of an englacial or subglacial position because the steeply sloping walls would inhibit accumulation of significant volumes of water-saturated debris. Sediments deposited in all three positions may, however, be overlain by flow tills.

The diagnostic criteria for subaqueous outwash have been described in detail by Rust and Romanelli (1975) and Rust (1977). The criteria for the recognition of deltaic eskers have not, however, been well described. The reason for the paucity of information is that the so-called deltaic eskers which have been described in the literature are not true deltaic deposits. By definition, a delta must include sediments which were deposited both subaerially and subaqueously where a river or stream enters a standing body of water (Coleman, 1976; Miall, 1976). A deltaic esker is, then, a linear ridge of sediments which were deposited both subaerially and subaqueously at the mouth of a glacial meltwater stream, during glacial retreat. The Scandinavian "esker deltas" described in great detail by De Geer (1940) are more properly classified as subaqueous outwash deposits, having been "deposited below the sea level" (De Geer, 1940, p. 46). More recently Aario (1971, 1972) has described primary sedimentary structures on a misnamed "esker delta" which was "deposited at a depth of about 100 m" (p. 151).

A major difference between true esker delta and subaqueous outwash fan deposits is the nature of the uppermost sediments. The vertical extent of the esker delta is controlled by the water level. As a result, the subaerial delta topset forms a more or less horizontal plane which later becomes the flat top of the esker. The uppermost sediments should exhibit evidence of subaerial exposure (eg. , desiccation cracks, frost polygons) and contain sedimentary facies which are common to modern deltas (Coleman, 1976). These include: distributary channels filled with a fining upwards sequence of sediments, subaerial levees, crevasse fill and argillaceous marsh deposits. One might also expect evidence of arctic flora to be preserved on the subaerial portion of the delta (Le Blanc Smith and Eriksson, 1978). Overall, a deltaic esker should exhibit a coarsening-upward sedimentary succession due to the progradation of the delta away from the mouth of the meltwater conduit.

The most commonly occurring upper unit within eskers which were deposited as subaqueous outwash fans is a wave formed gravel (De Geer, 1940; Rust and Romanelli, 1975), although it is not a facies of the fan itself. This gravel provides direct evidence that the underlying fan sediments were deposited subaqueously, having been reworked from the esker deposits when raised to wave base by isostatic uplift. Wave reworking may, however, occur along the flanks of a deltaic topset, affecting the entire upper unit if the relative water surface remains at a constant elevation for a sufficient length of time. Alternatively an overlying lag gravel may not develop on either fan type if the body of water has insufficient fetch to allow the generation of effective waves. Furthermore proglacial lakes may drain rapidly if their existence depends on the damming of drainage routes by ice. In such a situation the eskers may not be subjected to wave working for a long enough period to produce major modification.

Rust (1977) describes the following facies as diagnostic of subaqueous outwash:

1) Gravel facies: Deposited at the apex of the subaqueous outwash fan, consisting of clast-supported conglomerate which may be vaquely stratified and contains imbricate clasts with A-axis oriented parallel to the direction of flow.
2) Massive sand facies: These are channel filling sediments which include massive and faintly laminated and bedded sands, often containing dewatering structures (dish and ball and pillow structures).
3) Stratified sand facies: Deposited on the interchannel portion of the fan and characterized by the occurrence of massive and planar cross-bedded sands, trough cross-bedded pebbly sands and beds of normally graded climbing ripples.

While the foreset deposits of a deltaic fan may resemble subaqueous outwash one would expect much of the coarse sediment to be deposited on the topset. Furthermore, the channels on the subaqueous outwash fan, filled by deposition from sediment gravity flows (Middleton and Hampton, 1976) would be distinct from deltaic channels which are filled by deposition from traction currents and contain various types of cross-bedding. Not all of the facies described by Rust (1977) may be present on a subaqueous outwash fan. Aario (1971, 1972) described genetically similar deposits which lacked the gravel and massive sand facies, being dominated by stratified sands. Facies variations may be expected due to the sediment supplied to different glaciers from their catchments and due to variations in the sediment and water discharge from a given glacier at any location over time.

Deposition of deltaic or subaqueous outwash fans at a straight ice front may produce the broad eskers of relatively low relief discussed earlier as marginally-deposited eskers. However, when a glacial meltwater conduit terminates in a re-entrant of the ice front, the resulting esker will differ from the other marginal types. If the re-entrant is a symmetrical inlet the esker will be narrower and have steeper sides than those formed by deposition at a straight ice front. One would also expect that the beds on the flanks of such eskers would be faulted in a manner similar to that found in eskers deposited in conduits. Rust and Romanelli (1975) suggested that deposition in a re-entrant which formed an asymmetrical embayment in the front caused asymmetric facies distribution and paleocurrent directions oblique to the trend of the esker.

MORPHOLOGY AND ORIENTATION OF THE STITTSVILLE RIDGE

As mapped by Richard <u>et al</u>. (1977, figure 1-1), the Stittsville Ridge is approximately 10.9 km in length, although a break in the ridge occurs to the north of Stanley's Corners where the underlying bedrock crops out. The width of the ridge varies from 0.05 km to 0.3 km, generally becoming broader towards the north. The point of maximum elevation along the ridge is at site S-6 where it reaches approximately 137 m a.s.l.

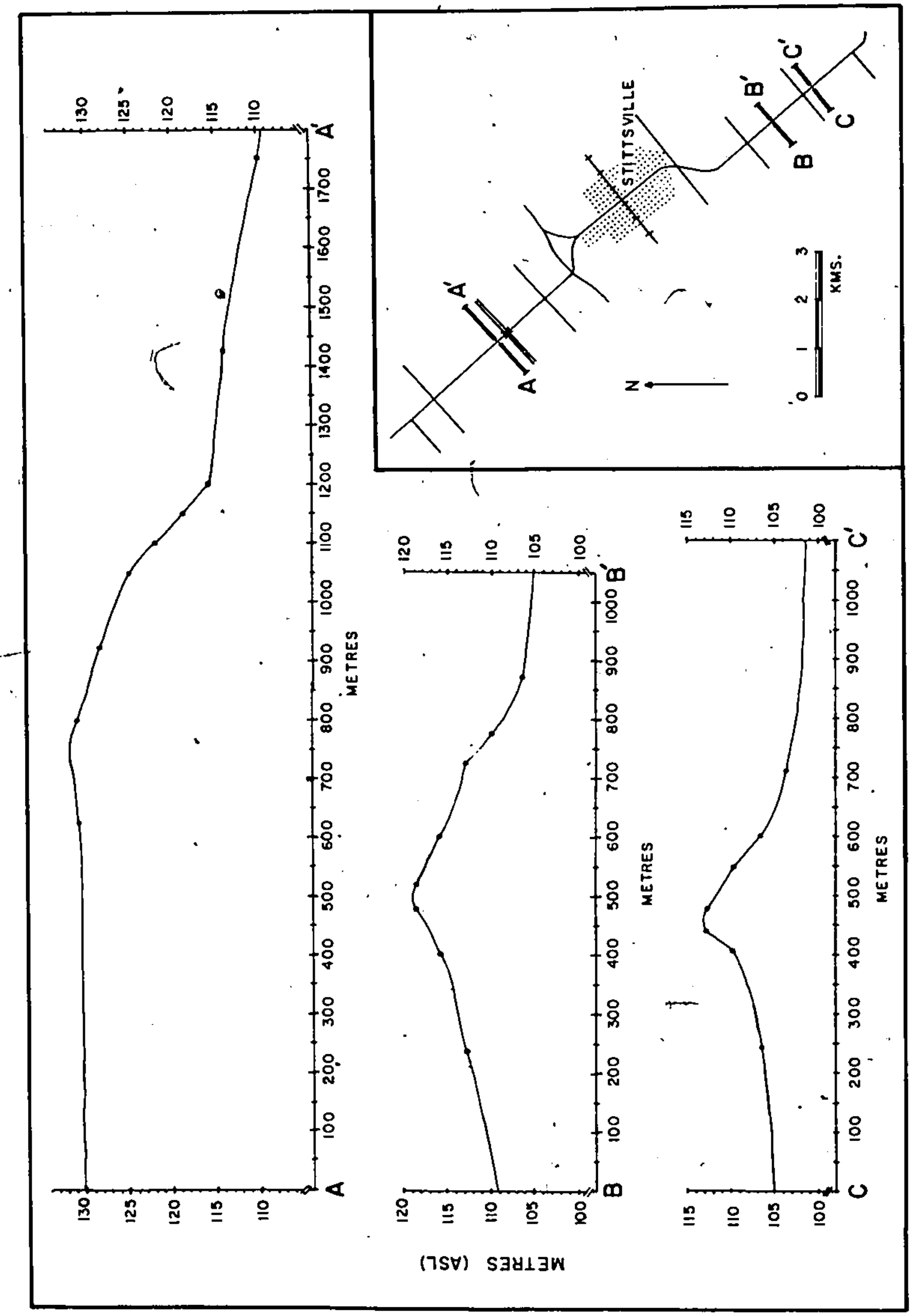
There the top of the ridge is about 30 m above the land surface to the northeast but only about 7 m above the surface to the to the southwest, where the surficial sediments become thinner. To the south, in the vicinity of site S-1, the elevation of the top of the ridge is approximately 117 m a.s.l., about 12 m above the country-side.

The morphologic expression of the ridge varies considerably along its length. To illustrate this, transverse profiles of each portion of the ridge (figure 2-2) have been derived from 10 foot contours on 1:25,000 topographic sheets for the area (Stittsville, 31G/5d, Edition 2; Munster 31G/4e, Edition 2). While fine morphologic detail cannot be determined from maps, sufficient resolution is possible to illustrate the morphologic differences between the three divisions of the ridge.

The northernmost profile across the ridge (figure 2-2A) shows a broad, asymmetric form, sloping relatively steeply to the northeast and very gently to the southwest. While the form may be partly controlled by the bedrock surface, there is a thickening of the surficial materials at the location of the ridge (Belanger and Harrison, 1976). On the northeast side of the ridge, at approximately 125 m a.s.l., a distinct break in slope occurs, changing from about 1.3°, at the top, to 3.6° below the break. This feature may be a wave-built terrace which formed during the regression of the sea and deposited along its northeastern flank.

The profile across the ridge at Stanley's Corners

Figure 2-2. Topographic profiles across the Stittsville Ridge derived from 10 foot contours on 1:25000 map sheets (Stittsville, 31G/5d, edition 2; Munster, 31G/4e, edition 2). Vertical exaggeration is X15.


STITTSVILLE
A
A'
B
B'
C
C'
N
0 1 2 3
KMS.
METRES
METRES (ASL)

(figure 2-2B) is more symmetrical than to the north of Stittsville. Again, a break in slope occurs on the northeastern side but is noticeably absent on the southwestern side. This supports the interpretation that the break is a wave-built feature. In the vicinity of Stittsville the Champlain Sea had its greatest fetch to the east, while the fetch from the west was less than 1 km when water level was at the elevation of the ridge. Furthermore, shallower depths on the western side of the ridge would have retarded wave development. As a result, wave energy was minimal on the southwest side of the ridge.

At site S-1 (figure 2-2C) the ridge is much steeper and narrower than at more northerly positions. A slight break in slope occurs on the northwestern flank, although it is not well developed.

The orientation of the ridge is also somewhat variable along its length. North of Stanley's Corners it strikes about 131° whereas south of Stanley's Corners the orientation is approximately 139°. For comparison with other oriented glacial features in the area a mean orientation weighted by the length of each segment was calculated to be 134°. Rust and Romanelli (1975) noted that the orientation of the Stittsville Ridge is approximately parallel to local glacial striae. It is also subparallel to the long axes of drumlins in the Ottawa area (Richard <u>et al.</u>, 1977) which have a mean orientation of 169° (N=138; $L = 99.01\%$; $p = 1.75 \times 10^{-59}$).

In summary, the Stittsville Ridge is a linear feature of positive relief which is oriented approximately parallel to the direction of Wisconsin glacial flow. On the basis of the morphology and orientation of the ridge, and the knowledge that it is made up of stratified sand and gravel, a first order interpretation of the ridge as an esker may be made. The presence of a terrace on the northeastern side of the ridge and the unconformable cover by fossiliferous marine gravel indicates that the esker was modified by wave action during regression of the Champlain Sea.

SEDIMENTOLOGY OF THE STITTSVILLE RIDGE

Site S-1

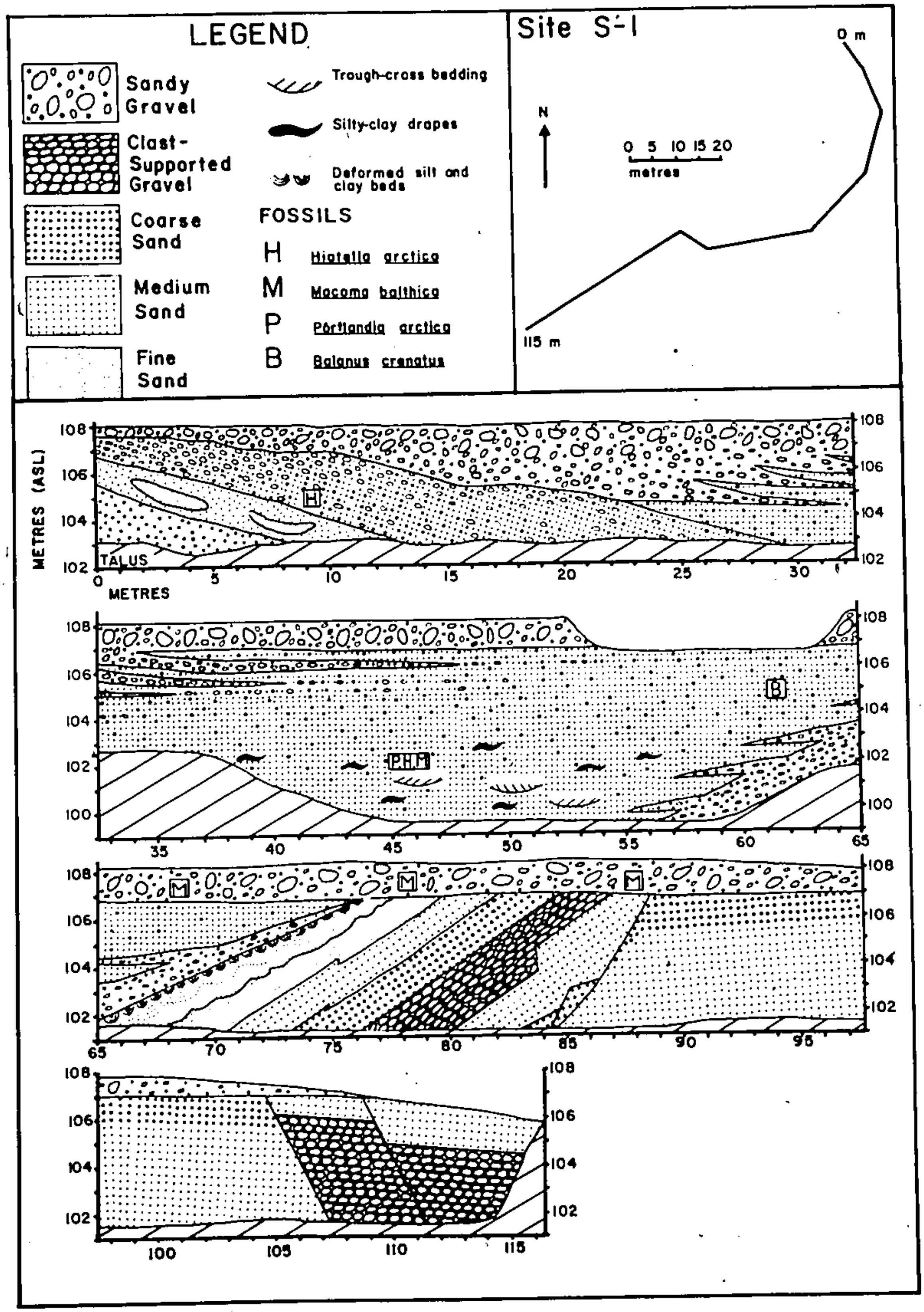
(1) Description

The sand and gravel pit at site S-1 was actively worked during the summer of 1978, providing good exposure across the central portion of the fluvioglacial unit of Richard *et al.* (1977). Figure 2-3 summarizes the stratigraphy of site S-1.

Two major stratigraphic units can be recognized in figure 2-3: 1) a deformed, unfossiliferous, stratified sand and gravel unit, unconformably overlain by 2) a sparsely fossiliferous coarsening-upwards unit.

The unfossiliferous unit is only poorly exposed on the northeast end of the pit between 0 and 13 m along the section. The sediment consists of massive granular sands at the base, fining upward to medium sand containing lenses of fine-grained

Figure 2-3. Measured stratigraphic section from site S-1.


LEGEND
Sandy Gravel
Clast-Supported Gravel
Coarse Sand
Medium Sand
Fine Sand
Trough-cross bedding
Silty-clay drapes
Deformed silt and clay beds
FOSSILS
H Hiatella arctica
M Macoma balthica
P Portlandia arctica
B Balanus crenatus
Site S-1
0 m
N
0 5 10 15 20
metres
115 m
METRES (ASL)
TALUS
METRES
PHM

sand. The upper surface of this facies dips towards the SSW at an angle of approximately 30^{o}.

A much better exposure of the unfossiliferous unit occurs between 65 and 115 m along the section. The most extensive bed consists of horizontally stratified medium sands which coarsen upward to coarse granular sand (85 to 107 m). This is gently folded into a broad anticlinal structure with its axis oriented approximately parallel to the long axis of the ridge. At 105 m this unit is terminated by a normal fault, striking NW to SE. The sediment on the western, downthrown side of the fault consists of stratified clast-supported cobble to boulder gravel with subrounded to rounded clasts in a pebbly sand matrix. This is sharply overlain by a horizontally stratified bed of medium sand. At 111 m a second normal fault displaces the gravel and medium sand by over one metre.

The eastern side of the main bed is terminated by a normal fault which appears to have experienced rotational slippage along a curved fault plane. Parallel beds on the eastern side of the fault dip 55^{o} to the east. Adjacent to the fault is a bed of medium-grained sand which is internally disturbed by high angle reverse faults which dip towards the normal fault. Immediately east of this is a less disturbed bed of stratified clast-supported gravel containing subrounded to rounded cobbles and boulders.

The sediments become finer to the east of the gravel and bedding becomes less well defined. Coarse sands adjacent to the gravel are internally faulted whereas silty fine sands exhibit soft-sediment folds which increase in amplitude towards the predeformational top of the unit where they appear as flame structures. Also within the uppermost bed are distorted alternating thin beds of silt and clay size sediments (figure 2-4, p. 45).

The easternmost beds of the fossiliferous unit are exposed between 0 and 30 m along the section. They are parallel beds of gravelly sand which dip towards the SSW. Although the dip is approximately parallel to that of the underlying unfossiliferous sediments, the upper bed shows no evidence of internal deformation. Within these beds the proportion of pebbles and cobbles increases up dip, towards the NNE. Along the base of the lowermost bed are dense accumulations of articulated valves of the marine mollusc Hiatella arctica.

Horizontally bedded gravelly sands dominate the central portion of the exposure at site S-1, overlying the previously described fossiliferous unit, from 22 to 83 m along the section. The lower beds in this unit consist of medium sand, commonly containing large scale trough cross-bedding with cross strata dipping bimodally to the NE and SW. Silt beds commonly drape over the sand beds forming lenticular and flaser bedding (Reineck and Singh, 1977). Articulated valves of Hiatella arctica, Macoma balthica and Portlandia arctica were found within some of the silt beds. These lower beds pass laterally, to the SW, into wedge-shaped pebble to cobble gravel units which coalesce along the margin of the steeply dipping unfossiliferous unit between 65 and 73 m along the section.

Above 104 m a.s.l. the silt beds are no longer present and the sediment becomes generally coarser in size. The only fossils which occur in the upper portion of this unit are <u>Balanus crenatus</u>, preserved as whole and broken shells.

A striking feature of the upper portion of this central fossiliferous unit are parallel horizontal beds which may be traced laterally across the section. Towards the NE they become increasingly coarse, passing into a clast supported gravel with pebbles and cobbles in a sandy matrix. The gravel forms long, wedge-shaped units which interfinger with the stratified sands to the SW. To the NE the largest clast size increases and the wedges coalesce to form a unit of maximum thickness 3.5 m. This same gravel unit extends across the top of the entire section, unconformably overlying both the fossiliferous and unfossiliferous stratigraphic units. Articulated valves of <u>Macoma balthica</u> were found near the base of the gravel while shell fragments were common throughout this unit.

(2) <u>Interpretation</u>

The unfossiliferous sediments at site S-1 are assigned to the fluvioglacial unit of Richard <u>et al</u>. (1977). They are overlain by fossiliferous Champlain Sea sediments. A detailed consideration of the unfossiliferous unit is required, however, to allow an interpretation of the specific depositional environment.

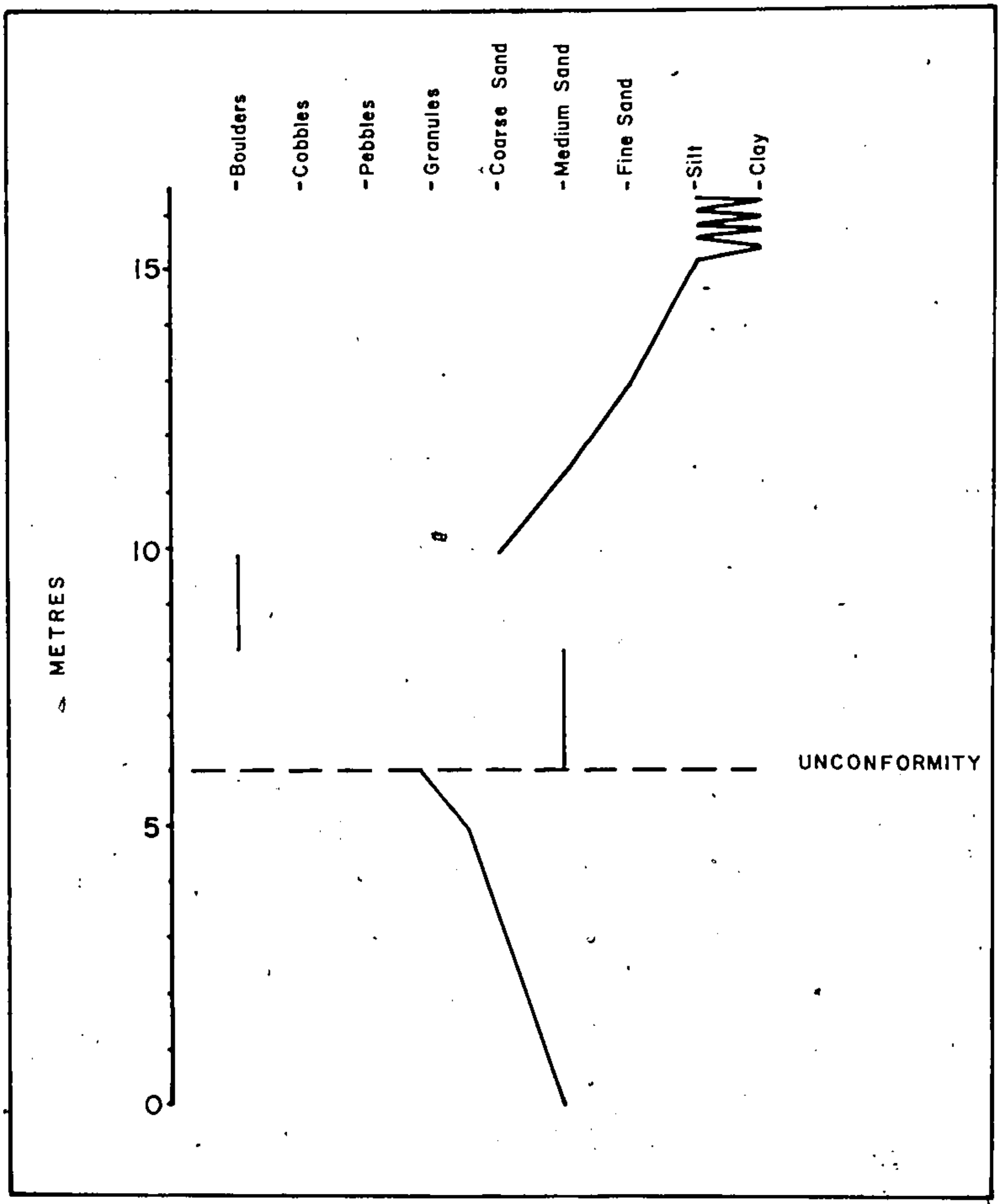
The most prominent features of the unfossiliferous sediments at site S-1 are deformation structures. Both the normal faults in sediments on the flanks of the ridge and the anticlinal flexing of beds within the ridge suggest that ice walls supported the unfossiliferous facies during deposition (Boulton, 1972). Morphologically the relatively steeply sloping sides of the ridge also suggest the former presence of ice walls (figure 2-2C).

The melting of ice may also be suggested for the deformation of sediments between 65 and 88 m along the section. Thus, as a buried saucer-shaped mass of ice melted, the beds overlying its thick core were lowered a greater vertical distance than those over its thinner flanks. At the outer margin of the ice mass a normal fault formed as the beds were displaced downward relative to adjacent beds resting on a stable base (McDonald and Shilts 1975). Small reverse faults in the coarse and medium size sand beds formed in response to minor internal adjustments during deformation. The silty fine sands, making up the outermost steeply dipping beds, were probably deformed by dewatering in response to the lateral movement of sediment as the angle of dip steepened and water was added from the underlying melting ice. The difference in the style of deformation in these beds is due to differences in permeability. The medium and coarse sands tend to have a higher permeability than the silty fine sands, allowing water to pass freely between the grains. If the grain framework was tightly packed these coarser sediments would behave as a rigid body during deformation. If loosely packed, deformation would have been accompanied by intergranular adjustments. The silty fine sands, however, have a much lower permeability, trapping water within their pore spaces. As a result, rapid dewatering occurred by liquefaction as water was added from melting ice and the grain framework was disrupted possibly by downdip flowage. The more rigid style of deformation affecting the alternating thin beds of silt and clay is due to the greater cohesion of such uniformly fine sediments.

The structural and morphological evidence indicates, therefore, that an isolated ice mass became buried by unfossiliferous sediments within the confines of ice walls. The sediments were subsequently deformed as the ice melted.

Figure 2-5 is a palinspastic reconstruction of grain-size distributions as seen in a generalized vertical section at site S-1. Of note is the upward fining of sediments from 8.2 m to 16.5 m (figure 2-5). This gradation from gravel to silt and clay at site S-1 is believed indicative of deposition on a subaqueous outwash fan. It is suggested that the clast-supported gravel was deposited near the mouth of a glacial meltwater stream while the alternating beds of silt and clay were deposited at a distal position as annual varves or from minor turbidity currents which flowed off the front of the fan.

Figure 2-5. A palinspastic reconstruction of grain size distributions at site S-1, as seen in a generalized vertical section through the unfossiliferous unit.



METRES
0
5
10
15
-Boulders
-Cobbles
-Pebbles
-Granules
-Coarse Sand
-Medium Sand
-Fine Sand
-Silt
-Clay
UNCONFORMITY

The sequence of sediments below 8.2 m (figure 2-5) does not fit this fining-upward sequence. A simple sequence would not occur, however, because the rate of ice retreat was probably sufficiently slow to allow a number of annual deposits to be superimposed at any one place on the fan. The sedimentology of each annual deposit would be strongly controlled by the glacial hydrologic regime. Furthermore, by analogy with deltas (Morgan, 1970) and submarine fans (Walker, 1978) one would expect the site of deposition on the fan to migrate with time. An active lobe would develop in one place and then be abandoned as deposition decreased slopes on that portion of the fan.

The coarsening-upward sequence of sediment from 0 to 6 m in figure 2-5 may reflect progradation of an early lobe on the subaqueous fan. The fining-upward sequence, however, is interpreted to have been deposited during the last occupation and abandonment of that portion of the fan. The continuous sequence from gravel to alternating beds of silt and clay (figure 2-5) is believed to have been deposited at the location of site S-1 in environments which were at increasing distances from the ice front as it retreated to the north. The backward migration of sediment source differentiates this environment from the submarine fan environment on continental margins. In the latter case, the source is the continental terrace which remains stationary while the fans prograde oceanward producing the characteristically coarsening-upward sequences.

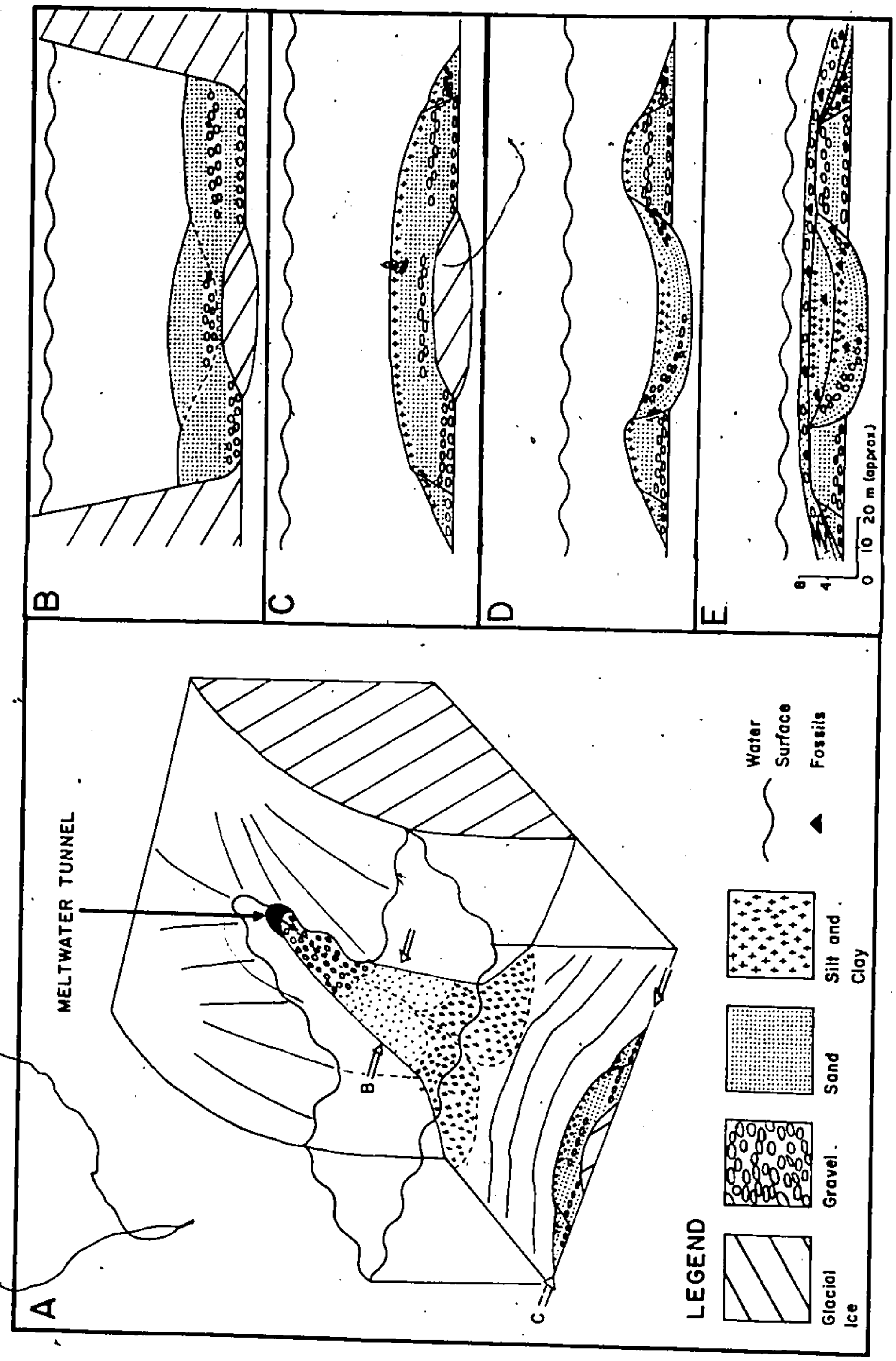
It is important, in light of discussion later in this chapter, to further consider the deformation of the alternating silt and clay beds in the unfossiliferous facies. The deformation of these distal sediments, interpreted as due to ice melt, implies that buried ice persisted for a considerable time following the retreat of the glacier front. Although actual distances cannot be determined, it is apparent that ice-melt deformation occurred after retreat of the high energy environments of the near-glacial subaqueous outwash fan. The persistence of buried ice has also been discussed by De Geer (1940) on the basis of observations of very similar deposits in Scandinavia. The absence of fossils in these distal sediments may be due to low salinities and/or low temperatures at the time of deposition.

The fossiliferous sediments exposed at site S-1 appear to be filling a depression in the underlying sediments which must have existed for some time after the buried ice melted out. The earliest sediments which were deposited within this depression, namely the wedges of gravel between 55 m and 73 m, were probably emplaced as sediment gravity flows. The coarsening-upward sequence of fossiliferous sediments within the central portion of the section was deposited during the regression of the Champlain Sea. Because of isostatic rebound, the water surface became relatively lower, and wave energy delivered at the sediment-water interface increased. The waves washed sediment into the depression and subsequently formed the lag gravel which caps all lower units exposed at site S-1. A more detailed treatment of these "hollow-filling" processes is given in Hayward (1979).

(3) Summary

The morphological, structural and sedimentological evidence suggests that the unfossiliferous sediments at site S-1 were deposited as subaqueous outwash within the confines of an ice-walled inlet at the glacier front. Figure 2-6A is a schematic diagram depicting the depositional environment and figures 2-6B and 2-6E illustrate the sequence of events which led to the development of the deposit exposed at site S-1. The inlet may have been the ice frontal expression of a crevasse system within the glacier, through which meltwater was channelled. Within the inlet the subaqueous fan was fed by sediment-laden glacial meltwater from a tunnel near the base of the glacier, as suggested by the evidence for a buried ice block rather than an extensive ice floor. The size of the sediment deposited on the fan decreased gradationally away from the tunnel mouth, from boulder gravel to sand and finally to silt and clay. Individual lobes actively prograded into the inlet, covering a block of ice, possibly derived from an iceberg which had calved off the glacier but which contained too much sediment to float. Alternatively, the buried ice may have consisted of an isolated mass of basal glacier ice which remained within the inlet during glacial retreat. The buried ice persisted long after the glacier

Figure 2-6. A depositional model for the sequence of sediments exposed at site S-1.


A
MELTWATER TUNNEL
B
C
LEGEND
Glacial Ice
Gravel
Sand
Silt and Clay
Water Surface
Fossils
B
C
D
E
8
4
0 10 20 m (approx.)

had retreated because of the insulating effect of the sediment cover. As fan lobes were abandoned they were covered by the more distal sediments of other lobes prograding from a different position in the inlet. The sequence of sediment deposited within the inlet is shown in figure 2-6B. The steep ice walls containing the inlet are believed to have inhibited the accumulation of water saturated tills on the ice, and may account for the absence of flow tills interstratified with the fan sediments. It is also possible that flow tills were not observed due to the relatively limited exposure available at site S-1.

With retreat of the ice front to the north, finer sediments were deposited at site S-1 on the distal reaches of the fan (figure 2-6C). The esker ridge developed as 1) the fans coalesced along a linear path controlled by the location of the meltwater conduit which maintained its position within the glacier during retreat, and 2) as the ice walls melted.

At some later time (figure 2-6D) the buried block of ice melted out, deforming the overlying sediments and forming a depression in excess of 8 m deep, in the submerged surface. The unstable walls of this depression became the source of sediment gravity flows which transported debris downslope.

Finally (figure 2-6E), isostatic uplift of the land caused a relative lowering of the water surface. Sediments were transported by waves into the depression, where they became populated by a marine fauna. It is this last stage which was largely responsible for masking the more hummocky terrain found on eskers in others parts of southern Ontario, which is more typical of eskers generally.

Sites S-2 and S-3

(1) Description

Figure 2-7 is a generalized stratigraphic section across the Stittsville Ridge, south of Stanley's Corners, based on data from sites S-2 and S-3.

Site S-2 is only poorly represented in figure 2-7. Thibault (1974) found that a faulted unit of unfossiliferous stratified sand and gravel dominated the exposure at site S-2 near the centre of the ridge. This is a generally fining-upward unit with gravel dominating the base and sand towards its upper limit. Unconformably overlying this unit is a clast supported gravel with a sandy matrix containing disarticulated paired valves of Macoma balthica and, less commonly, Hiatella arctica. Thibault (1974) noted that the thickness of the fossiliferous gravel increases towards the west, away from the centre of the ridge.

At site S-3 an extensive unit of unfossiliferous clast-supported gravel is exposed for approximately 60 metres along a north-south section on the eastern flank of the ridge (figure 2-8, p. 45). The clasts are subrounded to rounded and imbricate: preferred dip of ab planes normal to a-axis orientations. From north to south clast size decreases from predominantly boulders to cobbles and boulders. This unit is terminated to the west

Figure 2-7. A generalized stratigraphic section across the Stittsville Ridge at sites S-2 and S-3. Note that the section is drawn to depict the stratigraphy as it would be seen along a straight east-west exposure. In fact the pit at site S-3 is roughly circular in plan and exposures have various orientations.

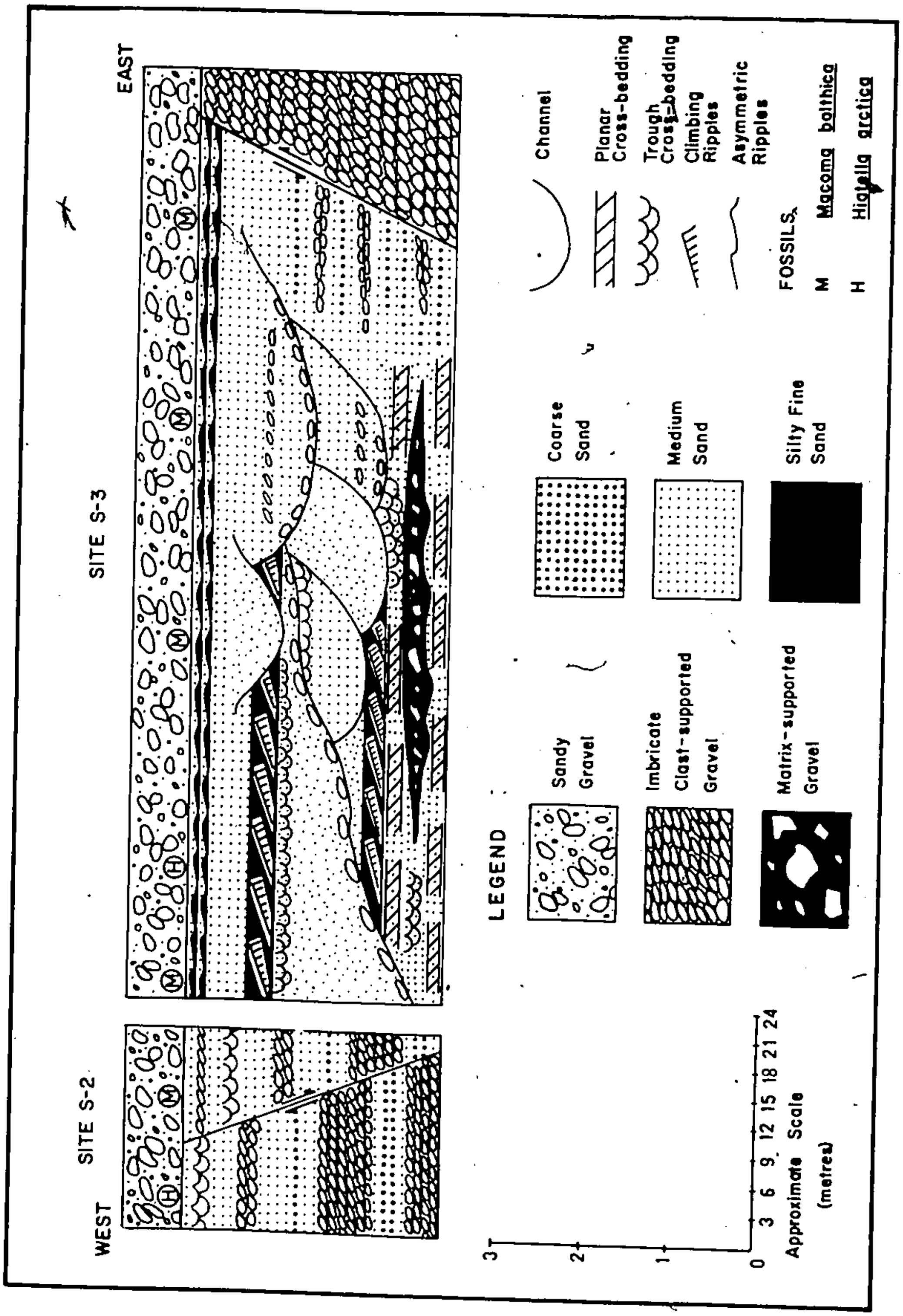

WEST
EAST
SITE S-2
SITE S-3
M
H
LEGEND
Sandy Gravel
Imbricate Clast-supported Gravel
Matrix-supported Gravel
Coarse Sand
Medium Sand
Silty Fine Sand
Channel
Planar Cross-bedding
Trough Cross-bedding
Climbing Ripples
Asymmetric Ripples
FOSSILS
M Macoma balthica
H Hiatella arctica
0
1
2
3
3 6 9 12 15 18 21 24
Approximate Scale
(metres)

Figure 2-4. Thin deformed silt and clay beds near the top of the unfossiliferous unit at site S-1. Notebook for scale is 21 cm high.

Figure 2-8. Clast-supported imbricate gravel at site S-3.

Figure 2-9. Large normal fault between clast-supported imbricate gravel and interstratified sand and gravel at site S-3. The gravel unconformably overlying the lower sands and gravels contains fossils.

Fig. 2-4

Fig. 2-8

1 m

Fig. 2-9

by a normal fault which displaced the sediment by more than 3 metres (figure 2-9, p. 45). On the opposite side of the fault there is a unit of stratified sand and pebble to cobble gravel, becoming sandier towards the top.

The lowest unit exposed on an east-west section across site S-3 is a coarse to medium sand containing planar and trough cross-bedding. The bounding surfaces of the planar cross-beds are parallel and the thickness of individual sets ranges from 10 to 25 cm. Trough cross-bed sets range from 3 to 5 cm thick and are made up of finer sands than the planar type. In figure 2-7 this unit is depicted as occurring across the section. It was only exposed, however, on the eastern side of the sand pit but is believed to underly the channelled sands described later.

Three notable features occur in the cross-stratified sand unit. The most prominent are angular to subangular boulders, in excess of 2 m in diameter (figure 2-10). The second is a lens-shaped unit of matrix supported gravel consisting of angular pebbles, cobbles and boulders in a sandy mud matrix (figure 2-11). The lower surface on the lens is almost sinusoidal in form, overlying deformed cross-stratified sand whereas the upper surface is relatively planar. The third feature is a unit of clast supported gravel which cuts into the cross-stratified sands (also shown in figure 2-11).

Stratigraphically above the cross-stratified unit, and dominating the exposure at site S-3, is a unit of channelled medium-sized sand. In this unit three types of channels were

Figure 2-10. Angular boulder within cross-stratified sands below channel-filling sediments at site S-3. Note the discontinuous gravel beds within the sands which fill the channel. (The exposure pictured is approximately parallel to the trend of the channel).

Figure 2-11. Matrix-supported gravel containing angular clasts interstratified with cross-stratified sands at site S-3. Note the occurrence of a clast-supported gravel unconformably overlying the cross-stratified sands. Stick for scale is approximately 1.5 m long.

Figure 2-12. Massive sand beds at the top of the unfossiliferous unit at site S-3. Individual beds exhibit a fining upward near their tops to silt and clay. In the uppermost beds these fines are cross-laminated and have asymmetrical ripple forms preserved. Note that the sharp, irregular lamina beneath the thin mud bed (see arrow) is a result of small tree roots which appear to have grown preferentially along a plane between the upper and lower beds, approximately 1.5 m from the surface of the ridge. Black portion of shovel handle is approximately 25 cm in length.

Fig. 2-10

Fig. 2-11

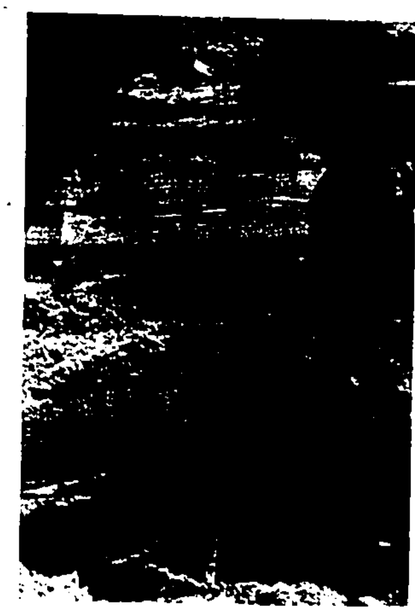


Fig. 2-12

recognized which also occur at the sites north of Stittsville (these will be described later). The three types are: Type I, channels filled with horizontally stratified medium sand with gravel along the erosional channel base and along bedding planes within the sands; Type II, channels filled with sub-horizontally laminated medium sands; and Type III, channels filled with massive medium sand.

While sections cut at right angles to the channels are rare, occasionally the channel form may be observed. They are relatively steep-sided asymmetrical erosional forms with height to width ratios of approximately 0.40. The largest channel observed was a type I channel, types II and III being generally smaller.

Type I channels contain rounded to well rounded pebbles, cobbles and boulders with <u>ab</u> planes imbricated normal to <u>a</u>-axes. An exposure parallel to the axis of a type I channel is shown in figure 2-10. Note that while gravel beds occur along particular bedding planes they are not continuous along the length of the channel. It is the discontinuous nature of the gravels beds within channels which may cause confusion in differentiating between type I channels and types II and III. Because the exposures of channels at site S-3 are only two dimensional, gravel may be present at depth into the face of the exposure. Observations at other sites along the Stittsville Ridge, and elsewhere in the Ottawa area, however, have shown that type II and type III channels do exist.

Type II and type III channels are very similar, except for the absence of internal stratification in the latter. A notable feature of both of these types is the occasional occurrence of silt or clay laminae adjacent the erosional boundary of the channels. Angular clay rip-up clasts and rounded limestone pebbles were observed within some type III channels, as well as inverse to normal grading near their bases.

At site S-3 a type I channel contains both massive and laminated sand. The transition between the two types of channel fill is not, however, continuous. The lower massive sands have been eroded at their upper limit and are overlain by trough cross-bedded coarse sands which pass upwards into type A climbing ripples (Jopling and Walker, 1968) of silty fine sand. This cross-bedded unit is restricted to the confines of the channel and is overlain by laminated medium sand. A bed of type A climbing ripples was also observed within laminated sands which probably fill a channel extending below the floor of the pit. A channel observed in a pit near Herbert's Corners, on the Gloucester Ridge, contained massive sand at the base, passing upwards into laminated sand.

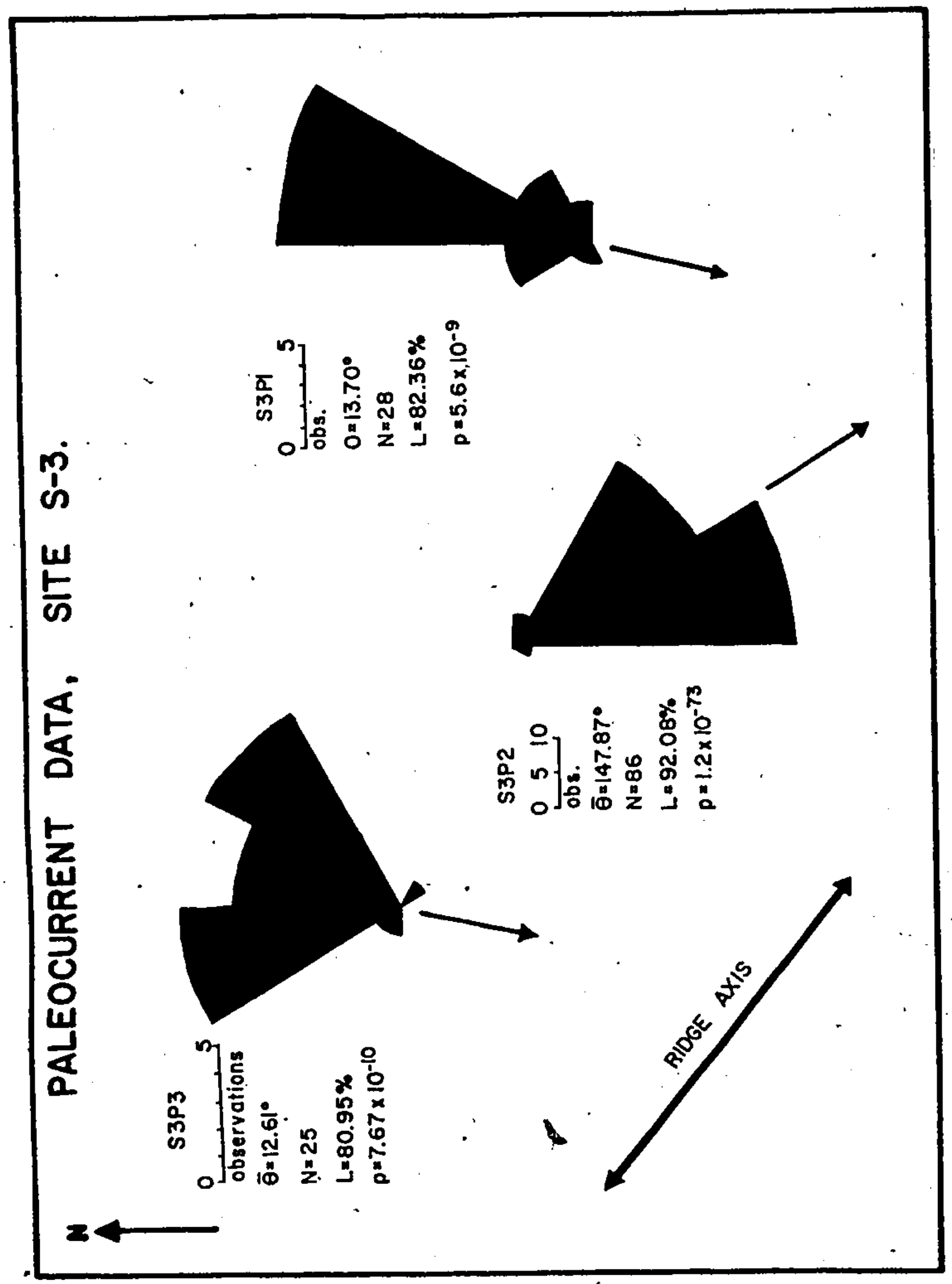
Conformably overlying the channelled unit at site S-3 is a unit of massive coarse to fine sand beds which become thinner upwards from 40 cm at the base of the unit to less than 5 cm at the top (figure 2-12). Individual beds begin abruptly and exhibit no change in grain size until within a few centimetres of their upper boundary where there is a fining of the sediment to silt and clay. These fines are cross-laminated in the upper, thinner beds.

The uppermost unit exposed at site S-3 is, again, a fossiliferous clast-supported gravel with a sandy matrix which unconformably overlies the lower unfossiliferous sediments (figure 2-9). As at site S-2, disarticulated paired valves of Macoma balthica, and rarely Hiatella arctica, are present within this unit. Note that the gravel is not present in all of the photographs because before excavating the underlying sands it is a common practice to remove this less desirable material from the surface.

Paleocurrent data was collected at site S-3 from the imbricate boulder gravel unit (S3P1), climbing ripples in a channel (S3P2) and imbricate cobbles in a type I channel (S3P3). The results of a statistical analysis of the data are presented in figure 2-13. Note that the rose diagrams for S3P1 and S3P3 represent the upstream imbrication of gravel. Paleocurrent directions are indicated by arrows.

The paleocurrents indicate a flow direction to the west of the ridge axis. Specifically, for S3P1, S3P2 and S3P3, respectively, the paleocurrent directions are 54°, 6° and 53° to the west of the ridge axis. Another important feature is the greater vector magnitude (figure 2-13) obtained for the imbrication of the boulder gravel than for the cobbles in channels. More striking is the difference in the mean deviation from the resultant vector for each set of data which

Figure 2-13. A summary of paleocurrent data collected at site S-3 from: imbricate boulder gravel (S3P1); climbing ripples (S3P2); imbricate gravel in Type I channels (S3P3). Note that each rose diagram is located to show the relative position of the specific collection site within the pit. $\bar{\theta}$=resultant vector; N=sample size; L=vector magnitude; p=probability of the data being random (Curray, 1956).


PALEOCURRENT DATA, SITE S-3.
S3P3
0 5
observations
θ̄=12.61°
N=25
L=80.95%
p=7.67 x 10-10
S3P2
0 5 10
obs.
θ̄=147.87°
N=86
L=92.08%
p=1.2 x 10-73
S3P1
0 5
obs.
O=13.70°
N=28
L=82.36%
p=5.6 x 10-9
RIDGE AXIS
N

was calculated to be 22.2 for S3P1 and 30.7 for S3P3. In both cases the gravels are comparably concentrated, although the concentrations are only local in the channels, suggesting that clast size may have controlled the degree of preferred orientation. The smaller deviation in the paleocurrent data collected from the boulder gravel is consistant with observations made by Johannson (1963) and Rust (1972) on preferred <u>a</u>-axis orientation in gravels. They found that in current transported gravels a higher degree of preferred orientation of pebble long axes is obtained from larger clasts.

<u>(2) Interpretation</u>

Because of the poor exposure of sediments at site S-2, the interpretation of this portion of the Stittsville Ridge relies heavily on the exposure at site S-3. Sedimentologically, the deposit at site S-3 contains abundant criteria for interpretation as subaqueous outwash, as described by Rust and Romanelli (1975) and Rust (1977). Some differences do exist, however, which add to the understanding of the subaqueous outwash environment.

The lowest unit at site S-3 is the clast supported boulder gravel believed to have been deposited at the apex of the subaqueous outwash fan (Rust, 1977). The reduction in clast size from north to south is consistent with the paleocurrent directions, suggesting that the sediment source, a glacial meltwater tunnel, was located to the north of site S-3. The fabric of the gravels is, however, different from that observed in coarser gravel by Rust (1977 , p. 176). He reported clasts to be imbricate parallel to their a-axes, suggesting transport by saltation, while at site S-3, the a-axes normal imbrication indicates that transport was by rolling on the bed. This difference, and the relatively smaller clast size, suggests that at site S-3 the velocity of flow near the tunnel mouth was less than that which transported the gravel observed by Rust (1977).

The sediments to the west of the boulder gravel appear to have been lowered from a stratigraphically higher position by the melting of buried ice, causing the development of a normal fault (figure 2-7). Horizontally stratified sand and gravel, adjacent the fault, is not included in Rust's (1977) model for subaqueous outwash. It is believed that this facies may be proximal sediment of limited areal extent, deposited outside a major channel containing the boulder gravel.

The cross-stratified sands which appear to dominate the lower portion of the section at site S-3 correspond to the interchannel deposits described by Rust (1977). The large angular boulders, within this unit, are best explained as dropstones deposited from glacial ice floating over the subaqueous outwash fan. The lens of matrix-supported gravel within this unit is interpreted as a subaquatic flow till. Both of these features reflect the proximity of the ice front to site S-3 during the deposition of the cross-stratified sand facies.

Of the channels observed at site S-3 both type II (filled with laminated sand) and type III (filled with massive sand) were reported by Rust (1977). Channels containing dewatering structures were notably absent, probably due to the predominance of coarser sediments at site S-3, which are less prone to liquefaction or fluidization than those described by Rust (1977).

The imbricate pebbles and boulders at the erosional base of type I channels suggest that the channels were formed by erosive currents transporting gravel as bed load. The gravel was probably locally eroded from sediments cut by the channels and transported relatively short distances to be deposited in discontinuous concentrations along the base. Sand-size sediment transported along the channels, in suspension, was deposited on more distal positions of the fan and on interchannel areas where flow expanded. Some of the time these channels may have been inactive, depositing the silt and clay laminae observed above the erosive boundary of some type II and III channels. The absence of gravel in types II and III channels probably reflects their higher stratigraphic position implying a distal position on the fan, rather than a different origin. Being stratigraphically above type I channels the only source of gravel would have been the underlying channels themselves, and as the fan evolved the possibility of channels cutting into gravel beds decreased.

The mode of channel formation outlined requires frequent pulses of dense currents over the fan. The high density of these currents, which resulted in flow across the fan rather than mixing with standing water at the ice front, may have been due to high concentrations of suspended sediment in the meltwater and/or its low temperature. The pulsating nature of the currents is a typical feature of discharge in glacial meltwater streams (Sugden and John, 1976) where, in summer, discharge commonly doubles over 24 hour periods in response to daily melting. More dramatic increases have been observed in response to precipitation on the ice. One might expect, therefore, that these high discharge events, of relatively short duration, would provide pulses of cold, dense sediment-laden water to the subaqueous outwash fan.

The sands which fill the channels were identified as mass flow deposits (syn. sediment gravity flow deposits) by Rust (1977). While four hypothetical types of mass flow deposits have been postulated (Middleton and Hampton, 1976) the criteria for their differentiation are not yet well understood. As a result, a more specific interpretation of the channel-filling sediments cannot be made at this time. However, some of the massive channel fills were definitely not transported as debris flows. Hampton (1975) found that in marine sediments a bulk clay content of 2 wt % is required to move the finest sand grains as debris flows. Four samples from a massive sand filled channel at site S-3 were analysed by dry sieving, and found to contain less than 1 wt % finer than 4ø (silt and clay). These sands could not, therefore, have been transported as debris flows. Some massive sand samples which were collected from subaqueous outwash deposits by Smith (1970) did, however, contain more than 10 wt % mud and may have been transported by non-turbulent flow.

The sands were probably introduced into the channels by slumping of the steep walls upstream of the site of deposition. The slightly coarser grain size in channels, compared to interchannel sands which are cut by the channels (reported by Rust, 1977 at other locations) is consistent with this interpretation. One would expect a fining of sediments away from the apex of the fan. As a result, sand which had slumped into channels at a given location and transported by mass flow to a more distal position, would be coarser than adjacent and underlying interchannel sediments.

Some channels appear to have been filled by a number of flows. For example type I channels with imbricate gravel along bedding planes within the sand may have been occupied by mass flow sands and later by currents transporting gravel. Repetition of this sequence of events would deposit the type of channel fill shown in figure 2-10. The occurrence of trough cross-bedded sands within the large channel in figure 2-7 indicates coarse sand transport in traction. The bed of climbing ripples, overlying the trough cross-bedded sands reflects waning flow in the channel, causing rapid fallout of fine sediment from suspension (Jopling and Walker, 1968).

The horizontal beds of massive sand above the channelled unit are probably the most distal sediments preserved at site S-3. These beds appear to have been deposited from mass flows which passed beyond channel termini or spilled over their banks. The normal grading and cross-laminated fines at the top of some beds suggests that there may have been turbulent mixing at the top of the dense flow. Clay, making up thin beds and thick laminae at the top of some massive sand units was probably deposited from suspension during the intervals of time between mass flow events on that portion of the fan.

The fossiliferous gravel, unconformably overlying the subaqueous outwash sediments, is a lag deposit formed by wave reworking of the fan surface during the regression of Champlain Sea. Reworking evidently eroded to a depth greater than 3 m, levelling the hollow which probably existed to the west of the normal fault (figure 2-9).

The morphology of the ridge suggests that deposition occurred on a subaqueous outwash fan between ice walls, although little structural evidence for such confinement was observed by the present author. Thibault (1974) reported extensive faulting of sediments at site S-2, but did not describe the deformation thoroughly. This faulting near the western flank of the ridge may have occurred during the melting of an ice wall.

At site S-3 the only major fault observed was the normal fault, between the boulder gravel and the stratified sands, which dips towards the centre of the ridge. As mentioned earlier, this fault probably formed in response to the melting of a buried block of ice beneath the central portion of the ridge. The absence of extensive faulting in the boulder gravel on the eastern flank, at site S-3, may be misleading. Firstly, the coarse grained, unstratified nature of the deposit makes it difficult to identify displacements which involved interclast movement. Secondly, faults which form in response to melting ice walls strike parallel to the length of the ridge (Boulton, 1972; McDonald and Shilts, 1975). Because the section in which the gravels are exposed trends approximately parallel to the ridge axis, such faults would be difficult to recognize.

The paleocurrent data collected at site S-3 (figure 2-13) provide evidence of an ice wall on the eastern flank. Paleoflow was predominantly to the west of the ridge axis, rather than to the east which one would expect on the eastern side of an unconfined subaqueous outwash fan.

It may be suggested, then, that the subaqueous outwash fan was deposited, as at site S-1, in an inlet into the glacier front. Furthermore, the deflection of flow to the west indicates that the inlet was probably asymmetrical in form, the meltwater tunnel being positioned on its eastern side, near the ice wall. The western side of the inlet was probably open, allowing deflection of flow towards the west.

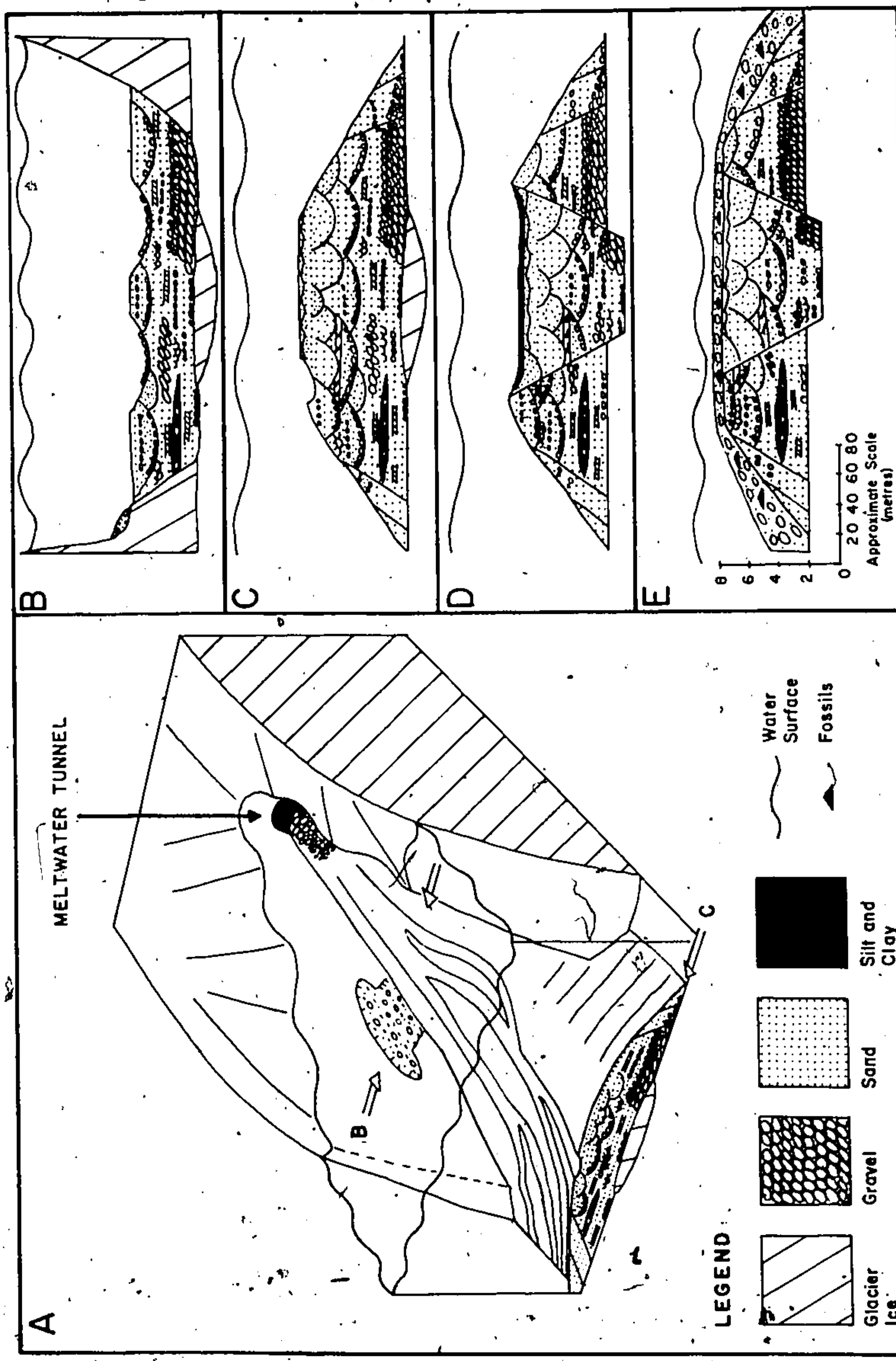
Deposition of the fan within an inlet may explain the predominance of the channelled unit at site S-3. Due to the confinement of the ice walls, the channels which eroded into the fan were concentrated within a limited area. As a result, the channels are much more prevalent than in an unconfined subaqueous outwash fan which has much of the stratified interchannel deposits.

(3) Summary

The sediments exposed at sites S-2 and S-3 are attributed to deposition on a subaqueous outwash fan within an asymmetrical inlet at the front of the retreating Wisconsin glacier (figure 2-14). These sediments differ from those deposited on the less confined subaqueous outwash fans described by Rust (1977). Most notable is the predominance of sand filled channels, and the relative absence of interchannel sediments, indicating a relatively narrow fan. Confinement also resulted in the scarcity of fine sands, which may have been transpored out of the inlet by currents deflected westward by the eastern ice wall. The interbedding of flow till with the fan sediments, probably reflects the proximity of ice walls to the fan. In figures 2-14A and 2-14B the till is shown accumulating on the ice below water level. It is equally possible, however, that accumulation was subaerial.

The sequence of sediments exposed at site S-3 generally fines upwards, because the depositional site became more distal as the ice retreated. However, the occurrence of a proximal clast-supported gravel overlying cross-stratified sand, shown in figure 2-11, disrupts this sequence. It was probably deposited as a result of channel switching near the mouth of

Figure 2-14. A depositional model for the sequence of sediments exposed at sites S-2 and S-3 (see figure 2-7 for an explanation of symbols used to illustrate sedimentary structures). Note that on the left wall of the inlet flow till is shown accumulating and that it occurs interstratified with the fan sediments.


A
MELTWATER TUNNEL
B
C
LEGEND
Glacier Ice
Gravel
Sand
Silt and Clay
Water Surface
Fossils
B
C
D
E
0 2 4 6 8
0 20 40 60 80
Approximate Scale (metres)

the meltwater tunnel or because of seasonal variations in meltwater discharge.

Figures 2-14B and 2-14E depict the sequence of events, similar to that described for site S-1, which resulted in the stratigraphy observed at sites S-2 and S-3. Figure 2-14B is a section cutting across the central portion of the inlet Proximal sands and gravels, which had buried a block of glacier ice, are cut by channels eroded into the fan surface by high energy density currents. Gravels were locally eroded from underlying and adjacent sediments and transported along the channel floors. Over time the erosive currents waned, as meltwater discharge diminished, and sand was transported through and/or deposited within the channels by mass flows. The sand was derived from slumping of the steep channel walls. Stratified sands were probably deposited on the interchannel areas but were rarely preserved due to channel migration.

Beyond the inlet (figure 2-14C) a few small channels may have been present on the western side of the ridge but most had terminated where flow was down the relatively steep slopes developed as the ice wall melted out. On top of and along the western flank of the ridge, massive sand beds were deposited from mass flows which overflowed the channel.

Over time (figure 2-14D) the buried block of glacier ice melted, lowering the central portion of the ridge and forming a hollow on the surface. Unlike the finer sediments at site S-1, the more permeable medium sands at site S-3 remained undeformed as water was released from the ice. Also at that time silt and clay were probably deposited on the ridge from suspension, but were not preserved due to wave reworking during regression of the Champlain Sea (figure 2-14E). Reworking was extensive, subduing the morphology of the ridge, and forming the fossiliferous gravel which unconformably overlies the subaqueous outwash sediments.

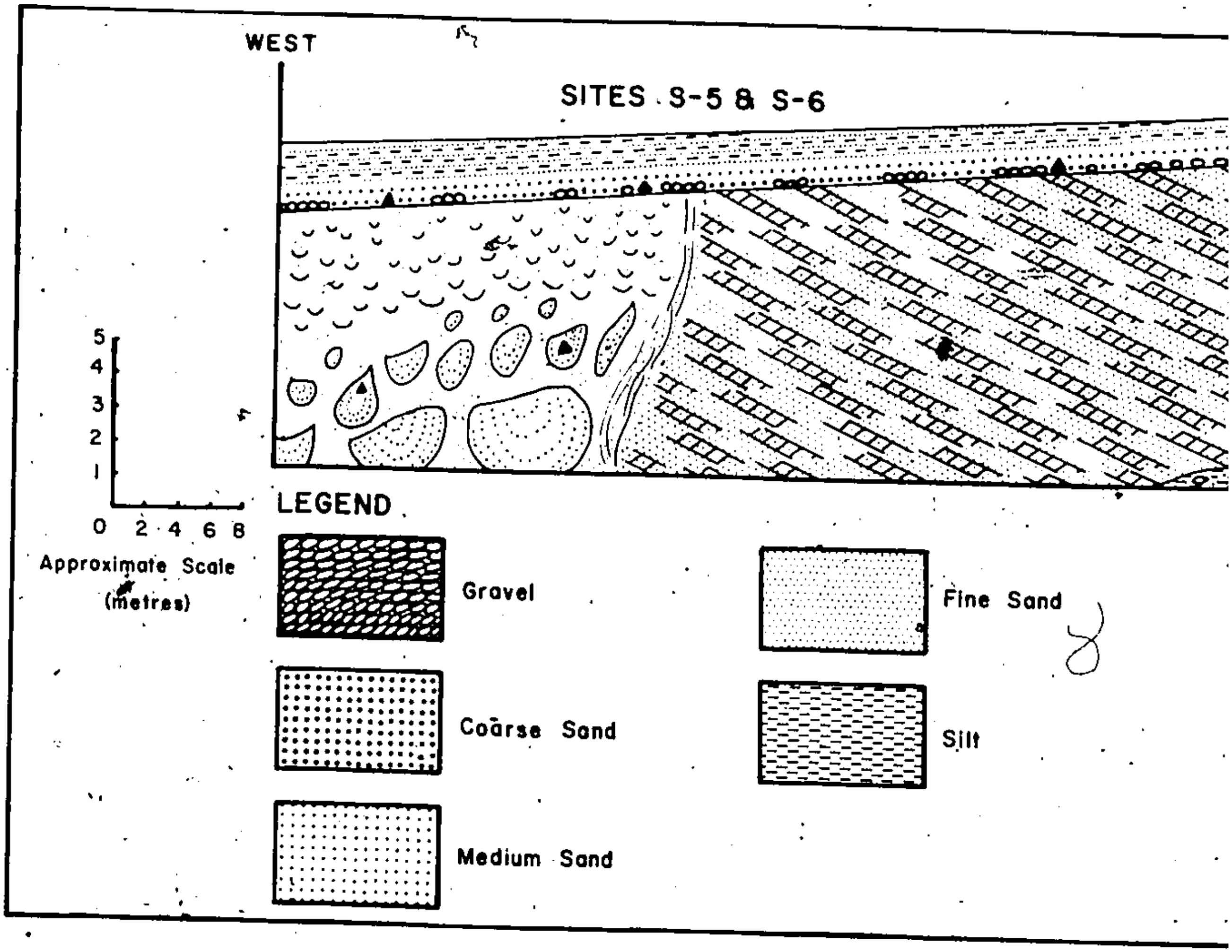
Sites S-4, S-5, S-6, and S-7

(1) Description

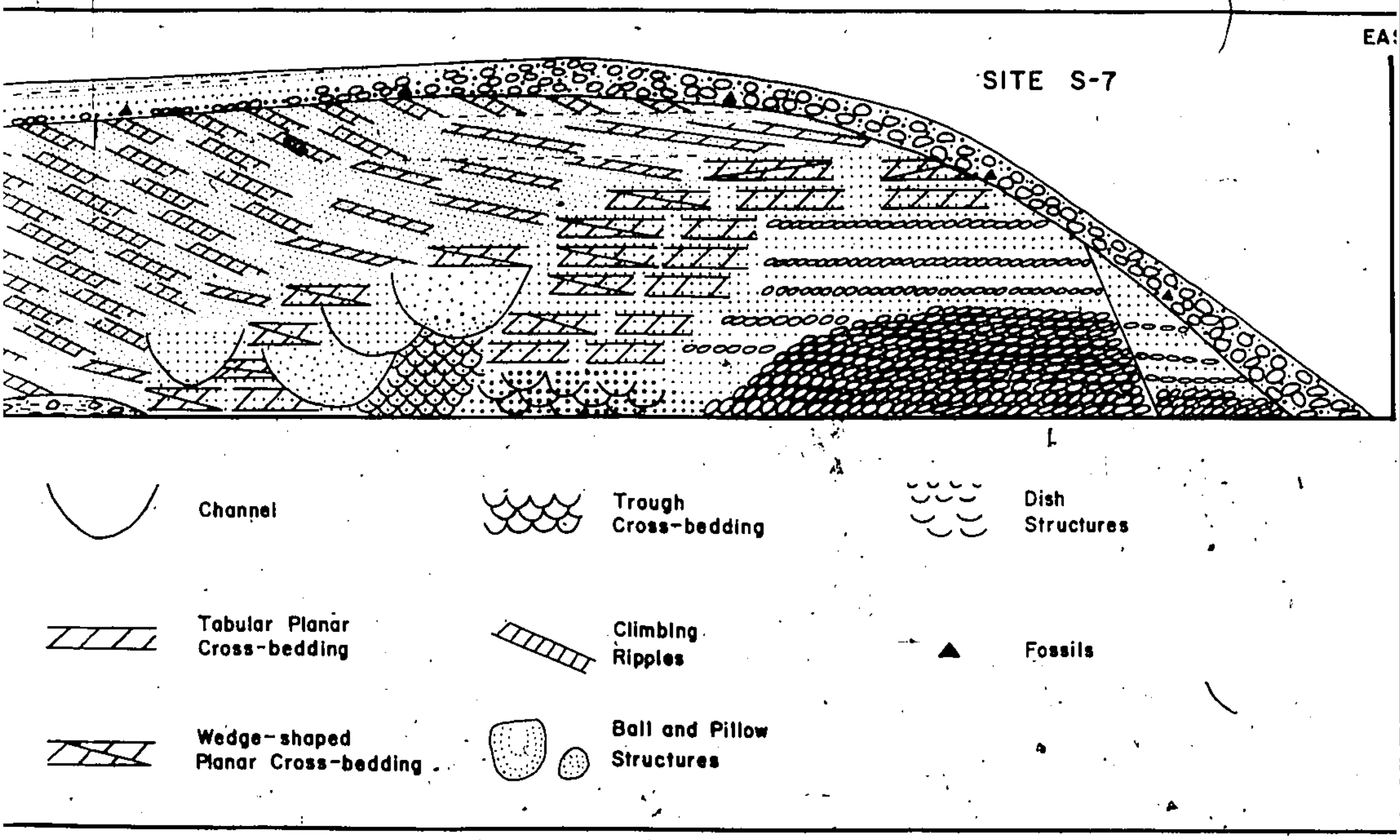
The sedimentology of the northern portion of the Stittsville Ridge has been described in detail by Rust and Romanelli (1975) and Rust (1977). However, continuing excavation in the pits allowed new data to be collected. Emphasis will be placed on new observations which add to, or alter, earlier interpretations. Figure 2-15 depicts the stratigraphy of the ridge in the immediate vicinity of sites S-5, S-6 and S-7. Exposure is not continuous across the ridge, so that figure 2-15 is an idealized section showing the major stratigraphic relationships observed at those sites.

The most obvious trend in the sedimentology of this portion of the ridge is an upward and westward fining of grain size. On the eastern flank, at site S-6, the lowest unit is a clast-supported boulder gravel in which the clasts are imbricated parallel to their a-axes (Rust and Romanelli, 1975, p. 181). Stratified sand and gravel, above the boulder gravel, is extensively disturbed by normal and reverse faults downthrown

Figure 2-15. Generalized stratigraphic section across the Stittsville Ridge at sites S-5, S-6 and S-7. Symbols denoting sediment texture are combined in some cases to denote: 1) interstratified sand and silt (the upper-most sediments on the western side); silt, sand and gravel making up a lens of matrix-supported gravel (basal unit at the centre of the ridge); and 3) sandy gravel (top unit on the eastern flank).


WEST
SITES S-5 & S-6
5
4
3
2
1
0 2 4 6 8
Approximate Scale
(metres)
LEGEND
Gravel
Coarse Sand
Medium Sand
Fine Sand
Silt

10f


EA
SITE S-7
Channel
Trough
Cross-bedding
Dish
Structures
Tabular Planar
Cross-bedding
Climbing
Ripples
Fossils
Wedge-shaped
Planar Cross-bedding
Ball and Pillow
Structures

2of2

to the east (B.R. Rust, pers. commun.)

Above and to the west of the coarse facies are sand beds containing tabular and wedge-shaped planar cross-bedding (Blatt et al., 1972), trough cross-bedding and climbing ripple drift (Jopling and Walker, 1972). While the general trend is an upward fining of grain size and a decrease in the thickness of cross-bed sets, the various cross-bed types are interstratified up the section exposed at site S-5. The boundaries between individual beds are commonly erosional and sometimes angular (figure 2-16), except at the base of climbing ripple sets.

In the lower portion of this unit, at sites S-4, S-5 and S-6, the cross-stratified sediments are cut by channels filled with massive and faintly laminated medium sands. These channels are similar to those described at site S-3 except that type I channels, with gravel beds, are rare. The channels occur in high concentrations (figure 2-16) and in isolation (figure 2-17). Note the inverse to normal grading at the base of the channel in figure 2-17.

Sets of climbing ripple drift dominate the upper portion of the cross-stratified sand facies, making up a distinct unit which thickens to the west, extending to the base of the exposure where it was locally underlain by a matrix supported gravel containing angular striated clasts. The climbing ripple unit consists predominantly of type A climbing ripples, the angle of climb increasing upwards. Type B and sinusoidal climbing ripples (Jopling and Walker, 1968, p. 973) are also

Figure 2-16. Massive and faintly laminated sand-filled channels overlain by planar and trough cross-stratified sands at site S-4. Note the angular erosional boundary between the trough cross-stratified set and overlying wedge-shaped planar cross-stratified set. The shovel is approximately 1 m in length.

Figure 2-17. Solitary massive sand-filled channel eroded into cross-stratified sands at site S-5. Note the inverse to normal grading near the base of the channel. Knife for scale is approximately 25 cm long.

Figure 2-18. Types A and B climbing ripples interbedded with massive sand at site S-4. Note the weathered cobble which was deposited as a dropstone, deforming the underlying sediment on impact. Pen for scale is approximately 14 cm long.

Fig. 2-16

Fig. 2-17

Fig. 2-18

present. All three types are commonly interbedded with massive sands (figures 2-18 and 2-19). Grain size generally decreases upward in this unit, from medium and fine sand near the base of the exposure to silty fine sand at the top where thin mud beds commonly separate climbing ripple units. Overturned climbing ripples (figure 2-20) were also observed in the uppermost portion of this unit. Large clasts, ranging from pebbles to boulders, are sparsely distributed throughout this unit (figure 2-18 and 2-21).

At sites S-5 and S-6 the climbing ripple unit ends abruptly at its western limit (figure 2-21), where individual beds are bent upwards in excess of 90°. Adjacent to this is a thick unit of ball and pillow structures which become smaller upward and pass into dish structures, all within a silty fine sand matrix. The ball and pillow structures contain massive and cross-stratified fine to medium sand. The internal stratification, where present, is generally bent upward at the edges of the structures. The dish structures are discontinuous silt laminae bent upwards at their edges. Both of these structures will be described in detail in chapter 3 of this thesis. Articulated valves of marine molluscs were observed within some ball and pillow structures. These included Macoma balthica (the most common species, which yielded a radiocarbon date of 11,300 ± 120 years B.P. (Gadd, 1978)), Hiatella arctica, Portlandia arctica and Mytilus edulis. The fossils were not distributed throughout this unit but occurred

Figure 2-19. Types A and B and sinusoidal climbing ripples interbedded with massive sand at site S-6. Shovel for scale approximately 1 m long.

Figure 2-20. Overturned climbing ripples near the top of the centre of the ridge, site S-5. 2 cm intervals on the scale.

Figure 2-21. Thick unit of climbing ripples adjacent to a unit of intensely deformed sediment at site S-6. Note the terminal edges of the climbing ripple beds bent upward, and the boulder occurring in fine sands to the left of the photograph (arrowed). Shovel for scale is approximately 1 m long.

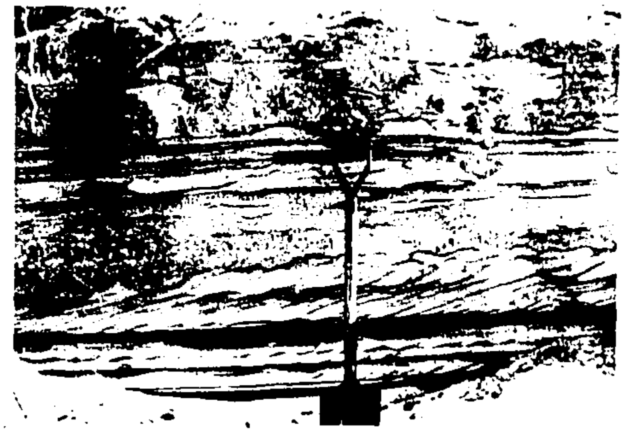


Fig. 2-19

Fig. 2-20

Fig. 2-21

near the upper limit of the ball and pillow structures. Also distributed sparsely throughout the deformed unit are large clasts, ranging from pebbles to boulders, which are typically striated.

Rust and Romanelli (1975) and Rust (1977) described this sequence of deformation structures as filling broad channels. Similar occurrences were observed at site S-4 where convoluted bedding commonly appeared at the base of the sequence of structures (figure 2-22 and figure 2-23). Figure 2-23 shows the erosional base of a silty-fine sand-filled "channel", cutting into cross-stratified sands. Note the much greater size of these channel-like structures compared to the stratigraphically lower medium sand-filled channels. At site S-4 the sediments below a channel-like structure are intruded by a silt dyke (figure A-3 at 80 m).

Paleocurrent data were collected from cross-stratified sediments at sites S-4 and S-5 and are summarized in figures 2-24 and 2-25 respectively. Also given in these figures is the position of the sampled unit within each pit (refer to figure 1) and its elevation above sea level.

All resultant vectors show deflection to the west of the ridge axis, with deflection increasing at higher stratigraphic levels. There is also an increase in the deflection from the ridge axis with increasing distance from the crest of the ridge (figures 2-24 and 2-25).

A unit of fossiliferous gravel unconformably overlies

Figure 2-22. Broad, channel-like structure, filled with deformed silty-fine and fine sand, cutting into cross-stratified sands at site S-4. Shovel for scale is approximately 1 m high.

Figure 2-23. Close-up of the base of a channel-like structure at site S-4. Note the convolute stratification and ball and pillow structures filling the channel-like structure. The black portion of the shovel handle is approximately 30 cm long.

Fig. 2-22

Fig. 2-23

Figure 2-24. Summary of paleocurrent data collected at site S-4. Note that X"s indicate the relative location of each collection site within the sand pit (see figure 1-1 and A-2) and that the elevation of each site is given. The summary includes: Θ, the resultant vector; N, the sample size; L, the vector magnitude; p, the probability of the data being randomly distributed (Curray, 1956).

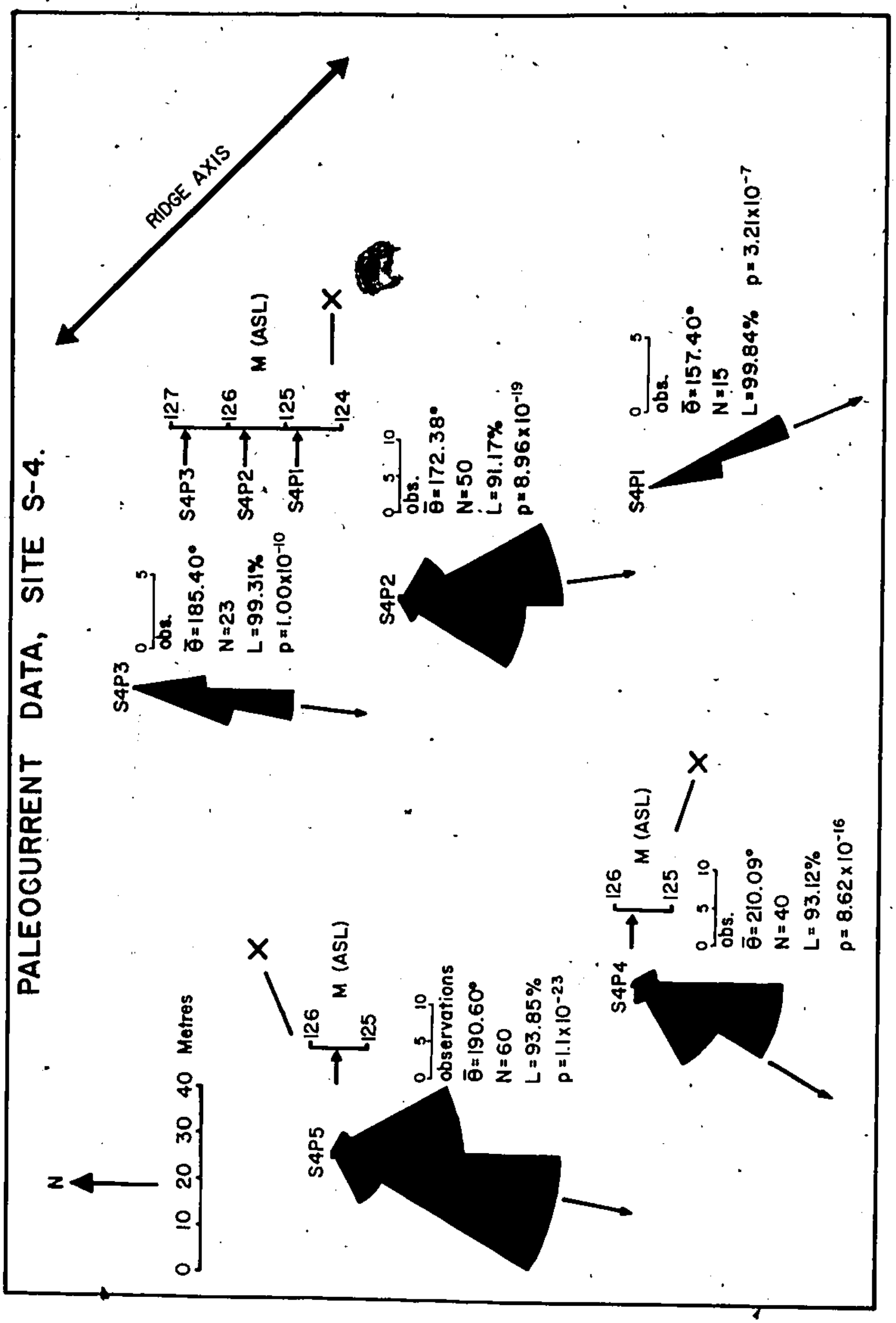
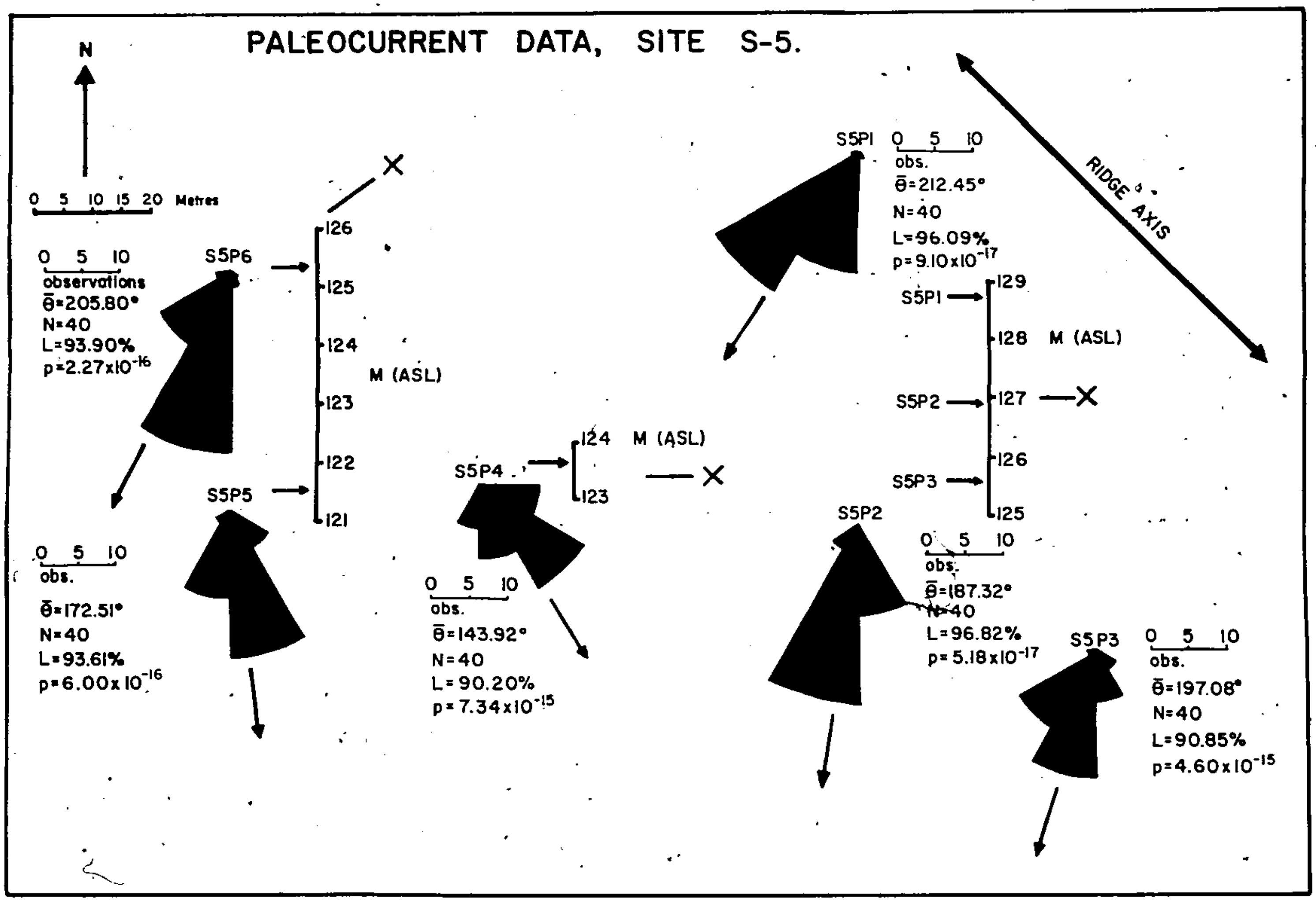

PALEOCURRENT DATA, SITE S-4.
RIDGE AXIS
N
0 10 20 30 40 Metres
S4P5
126
125
M (ASL)
0 5 10
observations
θ̄=190.60°
N=60
L=93.85%
p=1.1x10^-23
S4P4
126
125
M (ASL)
0 5 10
obs.
θ̄=210.09°
N=40
L=93.12%
p=8.62x10^-16
S4P3
0 5
obs.
θ̄=185.40°
N=23
L=99.31%
p=1.00x10^-10
S4P2
0 5 10
obs.
θ̄=172.38°
N=50
L=91.17%
p=8.96x10^-19
S4P3
S4P2
S4P1
127
126
125
124
M (ASL)
S4P1
0 5
obs.
θ̄=157.40°
N=15
L=99.84%
p=3.21x10^-7

Figure 2-25. Summary of paleocurrent data collected at site S-5. For a further explanation see figure 2-24.


PALEOCURRENT DATA, SITE S-5.
N
0 5 10 15 20 Metres
RIDGE AXIS
S5P6
0 5 10
observations
θ̄=205.80°
N=40
L=93.90%
p=2.27x10^-16
S5P5
0 5 10
obs.
θ̄=172.51°
N=40
L=93.61%
p=6.00x10^-16
126
125
124
123
122
121
M (ASL)
S5P4
124
123
M (ASL)
0 5 10
obs.
θ̄=143.92°
N=40
L=90.20%
p=7.34x10^-15
S5P1
0 5 10
obs.
θ̄=212.45°
N=40
L=96.09%
p=9.10x10^-17
S5P1
S5P2
S5P3
129
128
127
126
125
M (ASL)
S5P2
0 5 10
obs.
θ̄=187.32°
N=40
L=96.82%
p=5.18x10^-17
S5P3
0 5 10
obs.
θ̄=197.08°
N=40
L=90.85%
p=4.60x10^-15

the sediments described above. At sites S-4 and S-7 the gravel is clast-supported with a sandy matrix, identical to that at sites S-1, S-2 and S-3, and contains broken and articulated valves of Macoma balthica. At sites S-5 and S-6 this unit thickens westward and consists of a fining-upward sequence from gravel at the erosional base to sand and finally to horizontally interstratified thin beds of sand and silt (see figure 2-15).

(2) Interpretation

Rust and Romanelli (1975) interpreted the northern portion of the Stittsville Ridge as subaqueous outwash deposited in an asymmetrical embayment which opened on its western side. The form of the embayment was believed to have been controlled by a crevasse system within the glacier. The confinement of the subaqueous outwash fan to such an embayment is certainly suggested by the morphology of the ridge, the exposure of faults in fan sediments near its eastern flank, the deflection of paleocurrents to the west and, as noted by Rust and Romanelli (1975, p. 190), the asymmetrical distribution of facies within the ridge. Considering the morphology which has been postulated for the ice front at the time of deposition at sites S-1, and S-2 and S-3, however, it may alternatively be suggested that the embayment evolved by preferential melting on the upwind side of the initially narrow inlet

(Lundqvist, 1955).

The present observations of this portion of the ridge confirm those reported by Rust and Romanelli (1975) and Rust (1977) and the interpretation given here largely agrees with these authors. That is, the boulder gravel on the eastern side of the ridge was deposited near the mouth of a glacial meltwater conduit on a subaqueous fan characterized by bifurcating channels which became filled with massive and faintly laminated sands. Although channels containing gravel are rare, some of the sand-filled channels contain silt laminae marking their outer boundaries. This suggests that, as at site S-3, the channels were eroded by currents and later filled by mass flow deposits derived from slumping of the channel walls. Silt laminae were deposited from suspension during intervening periods.

The distribution of the cross-stratified sand facies in the eastern part of the exposure at sites S-5 and S-6 also supports their interpretation as sediments which were deposited on the interchannel areas of the fan (Rust, 1977). However, the thick unit of climbing ripples on the western sides of these exposures was probably deposited at a more distal position, beyond the reach of the channels. Silt beds between the uppermost beds of climbing ripples were probably deposited when meltwater and sediment discharge from the glacier was low.

The matrix-supported gravel, containing angular clasts, below the units of climbing ripples at site S-5 is probably a till. However, because the base was not exposed it is impossible to determine if it was deposited directly from the base of the glacier or as a flow till.

Observations of the intensely deformed unit within the northern portion of the ridge also require a somewhat different interpretation than that proposed by Rust (1977). The sequence of deformation structures described above, and in Rust (1977), was present in two different situations: 1) in channel-like forms, and 2) as a thick unit immediately adjacent to the unit of climbing ripples at sites S-5 and S-6.

The first situation, described above and pictured in figures 2-23 and 2-24, and figure 5 of Rust (1977, p. 179), was interpreted by Rust as channels filled with sediments which were deformed by rapid dewatering. The interpretation given here is based on extensive exposures made when highway 17 was constructed and disagrees with the channel interpretation but accepts the deformation mechanism. The stratigraphic position of these forms above the smaller massive and faintly laminated sand-filled channels and cross-stratified sands, and the finer grain size of the sediment contained within them suggests that they occupied a more distal position on the fan than the underlying facies. If they were true channels, therefore, one would expect them to be smaller than the stratigraphically lower channels due to the bifurcation of the channel system, away from the apex of the fan, as postulated by Rust (1977, p. 182).

The channel-like structures are interpreted here as the result of slumping on the subaqueous outwash fan. Such slumps are analogous to those observed on the continental slope, the surficial expression of which has been reported by Kelling and Stanley (1976, figure 27). The slumping event appears to have involved the rapid dewatering (liquefaction and fluidization) of silty fine sediments, probably beds of climbing ripples which contained sufficient mud to impede pore water escape and which had an unstable grain framework due to rapid deposition. Some mechanisms which may have caused rapid dewatering in the subaqueous outwash environment were reviewed by Rust (1977, p. 182). These include: shock from earthquakes, breaking waves, storm surges, or icebergs impacting on the bed, and rapid changes in water level and salinity as Champlain Sea invaded the Ottawa valley, previously occupied by a fresh water body. In addition to these causes other possibilities include: 1) the rapid deposition of a sediment, causing a rapid increase in the pore water pressure of the subjacent sediment; and 2) failure of silt and fine sand which had been deposited on a surface sloping at 15^{o} or more (Lowe, 1976, p. 295). Any of these may have caused slumping of sediments on the fan. An important aspect of slumping is that while and fluidization involve vertical movements of sediment and water, the material within the slumps appears to have been sufficiently mobile to flow for some distance down the gentle slope of the fan. It was during this movement that the erosional bases of the slumps were formed. The intrusion of sediments beneath the slumps probably occurred with the initial failure and the intruded bodies were subsequently eroded during the flowage of the overlying sediments.

The second situation in which the sequence of deformation structures occurs at sites S-5 and S-6 is significantly different from that described above. The uppermost ball and pillow structures contain a marine fauna, which led Gadd (1978) to question Rust's (1977) interpretation of the deposit as subaqueous outwash. Gadd (1978, p. 328) obtained a radiocarbon date of 11,300 ± 120 years B.P. from shells in the ball and pillow structures, implying that the sediments were deposited when the ice front was approximately 80 km to the north of sites S-5 and S-6 (Gadd, 1978, p. 328)

The interpretation presented here is that the fossils in the deformed unit may be explained without rejecting the subaqueous outwash interpretation. It has long been known by geologists that the formation of ball and pillow structures involves their downward displacement through mobile sediment (ie., Kuenen, 1958). Given this mode of formation and the observation that fossils occur only in the uppermost ball and pillow structures, it is reasonable to assume that the fossiliferous sediments were deposited above the subaqueous outwash but sank downwards during deformation. The question which remains is why deformation took place well after the ice had retreated.

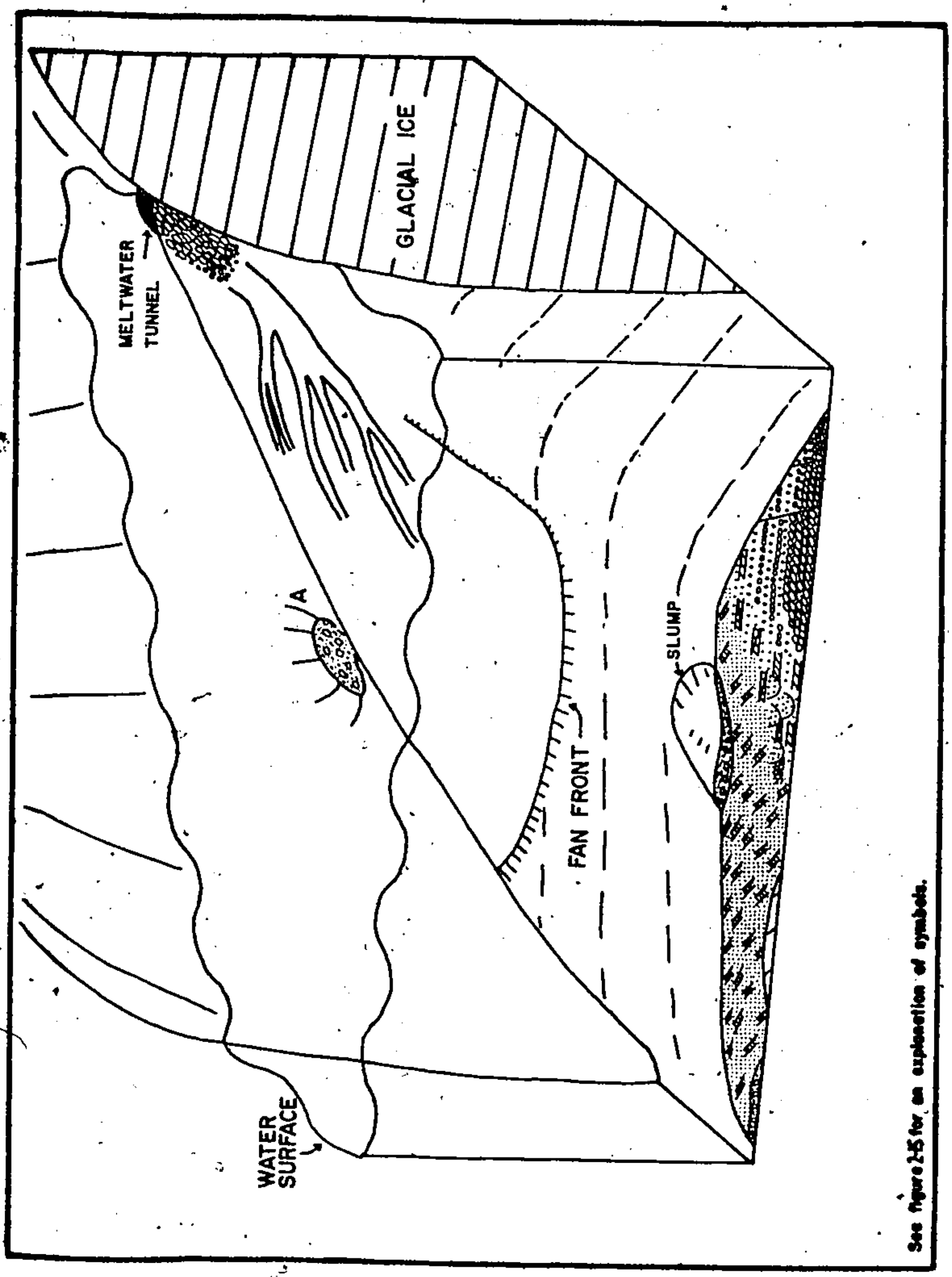
At site S-1 the sedimentological evidence indicates that deformation of the unfossiliferous sediments occurred by ice-melt after the glacier front had retreated a sufficient distance for deposition of silt and clay rhythmites. At sites S-5 and S-6 melting of buried ice may be cited as a cause of deformation while the uppermost subaqueous fan deposits were being reworked by waves and populated by a marine fauna. It is believed that at sites S-5 and S-6 the addition of meltwater to the base of the relatively impermeable climbing ripple unit caused it to become locally fluidized or liquefied. The sediment and water mixture then flowed upwards, breaking the overlying sediment into pieces which sank downwards as ball and pillow structures. Under the then lower confining pressure the uppermost sediment dewatered more slowly forming dish structures. The near vertical boundary between the climbing ripple unit and the deformed unit (figures 2-21, 2-15) probably marks the position of the eastern edge of the buried ice, the rapid vertical flow of sediment and water thrusting the edges of the beds upwards. A more detailed discussion of this mechanism of deformation will be presented in Chapter III.

The fossiliferous gravel at sites S-4 and S-7 was formed, as at other locations along the ridge, by wave reworking of the subaqueous outwash surface during the regression of the Champlain Sea. However, the fining-upward sequence of fossiliferous sediments at sites S-5 and S-6 differs from other marine deposits associated with similar ridges in the Ottawa area. The sequence is believed to reflect the breaking of waves, which approached from the east, the direction of greatest fetch. As wave base lowered, the wave energy expended on reworking the crest of the ridge increased, reducing the wave energy available to transport sediment on the western side of the ridge. As a result, the mean size of sediment decreased over time and a fining-upward sequence resulted. Horizontal thin beds of fine sand and silt interstratified within this unit are believed to have been deposited during storms and calm periods, respectively.

(3) Summary

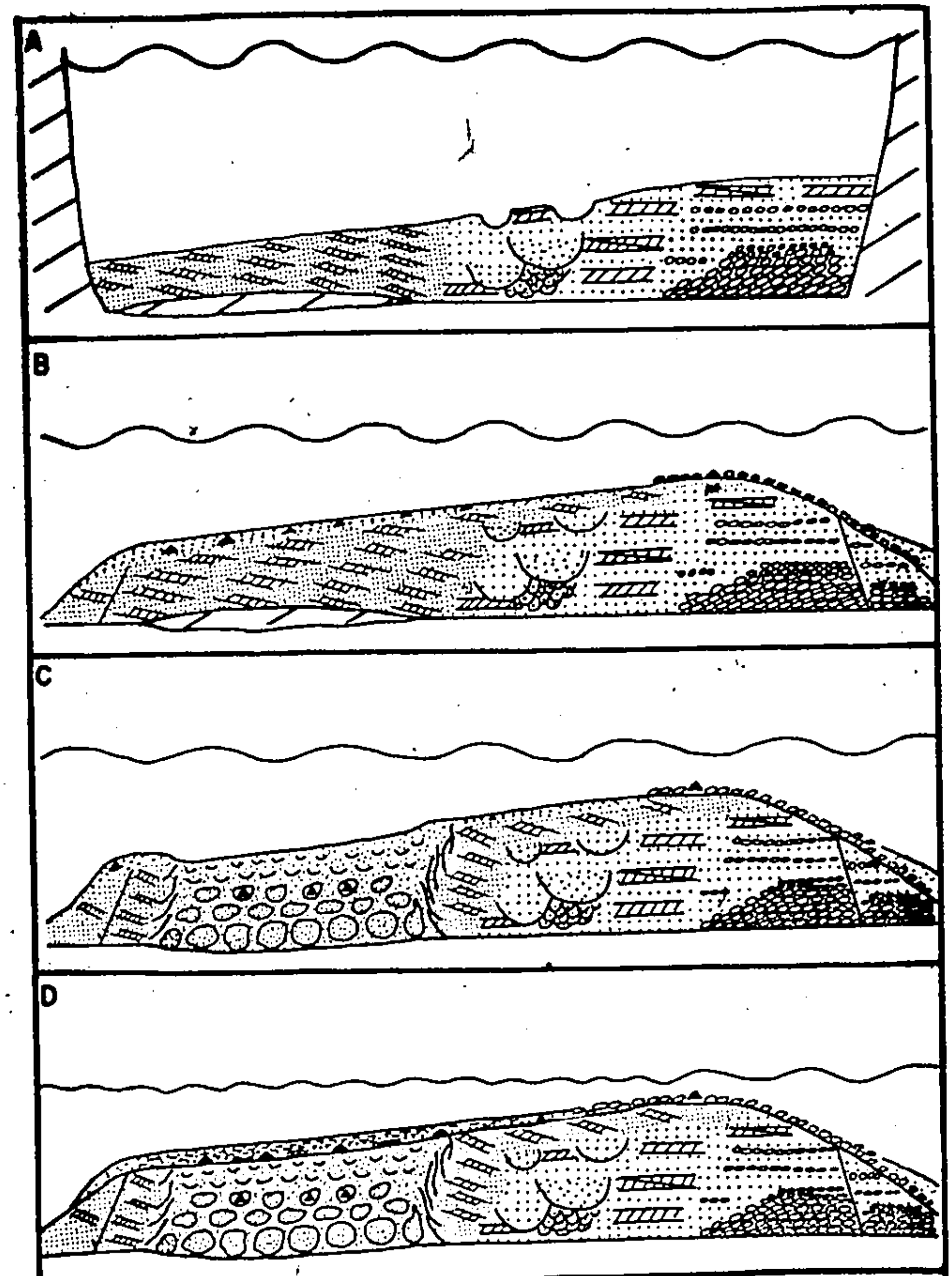
Figure 2-26 illustrates the depositional model postulated for the unfossiliferous sediments at sites S-4, S-5, S-6, and S-7. A glacial meltwater tunnel emptied into a large asymmetrical embayment at the ice front where powerful sediment laden currents deposited boulder gravel from suspension in a large channel. Outside this channel stratified sand and gravel was deposited. As flow was deflected away from the relatively straight eastern ice wall towards the more open western side of the embayment, cross-stratified sands were deposited in the areas between the channels scoured by the dense currents. As at site S-3, these channels were filled as sand slumped from the steep sides and was transported as mass flows. Because of the decrease in the current strength away from the conduit, the channels bifurcated and increasingly fine sediments were deposited. Beyond the channel termini ~~silty-fine~~ sediment was rapidly deposited from suspension as interbedded massive

Figure 2-26. A depositional model for the sequence of sediments exposed at sites S-4, S-5, S-6 and S-7. Note that the letter "A" indicates the position where flow till is accumulating.



See figure 2-5 for an explanation of symbols.

Figure 2-27. The sequence of events leading to the stratigraphy observed at sites S-5 and S-6. See text for discussion.



See figure 2-15 for an explanation of symbols.

sand and climbing ripple beds. Locally these fine sediments slumped down the fan slope, to the west, as they underwent liquefaction or fluidization. Flow tills travelled over, and became interbedded with the fan sediments, originating from areas of accumulation on adjacent glacial ice (indicated by the letter A, figure 2-26).

Figure 2-27 depicts the sequence of events which resulted in the stratigraphy observed at sites S-5 and S-6. Figure 2-27A shows the situation within the embayment. As the ice retreated sediment was deposited at increasingly distal positions on the subaqueous outwash fan. Because the meltwater tunnel was located near the eastern wall of the embayment, the fan built outward to the west where predominantly fine sediments were deposited, burying a block of glacial ice.

With further retreat the ice walls of the embayment melted and faulting occurred in sediments along the flanks of the deposit (figure 2-27B). Because exposures do not extend to the westernmost edge of the ridge we do not know whether deformation on the western side was rigid (as shown in figures 2-26 and 2-27) or plastic, but faults exposed on the eastern side indicate rigid deformation. At a later stage (figure 2-27B) the surface of the fan was reworked by waves and populated by a marine fauna.

At some later time (figure 2-27C) the amount of water derived from melting of the glacial ice beneath the fan exceeded the threshold of dewatering resistance in the overlying silty fine sand. This sediment locally fluidized, and a sediment water mixture moved rapidly upwards breaking overlying beds which sank downwards to form ball and pillow structures. In this manner the fossiliferous sediments were emplaced at a stratigraphically lower position.

With continuing regression of the Champlain Sea, wave base fell relative to the surface of the ridge. Waves breaking on the eastern side of the ridge reworked its surface, forming the fossiliferous gravel, while sand and silt were deposited on the protected western side (figure 2-27D).

DISCUSSION

The observations and interpretations presented above suggest that the Stittsville Ridge was formed by deposition in the proglacial environment on subaqueous outwash fans. Variations in the sedimentology and morphology of the ridge can be explained by changes in the morphology of the ice front as the glacier retreated to the north.

The ridge began to form when the ice front was at a position just south of site S-1. There, an englacial or subglacial meltwater tunnel terminated in standing water, depositing its sediment load within the confines of a narrow, ice-walled inlet into the glacier front. A subaqueous depositional site is suggested by the fining-upward sequence of unfossiliferous sediments formed during glacial retreat, and the coarsening-upward fossiliferous sequence deposited during marine regression, although the criteria for subaqueous outwash given by Rust (1977) are largely absent. This may be due to the narrowness of the inlet, inhibiting the development of various facies on the fan. Preservation of the sedimentary structures which are diagnostic of subaqueous outwash may also have been unlikely due to the intense deformation which took place as the ice walls and buried ice melted out. This deviation from the subaqueous outwash model is useful in that it indicates ways in which the model should be extended.

By the time the ice front had retreated to the position represented by sites S-2 and S-3, the inlet had become broader and asymmetrical in form. The subaqueous outwash fan which was deposited within this environment was similar to that described by Rust (1977), containing clast-supported gravel (deposited as bed load near the mouth of the tunnel), cross-stratified sand and sand-filled channels. Certain characteristics of the deposit, however, differ from that described by Rust (1977) and are believed to be diagnostic ~~of a more~~ confined environment. These characteristics are: 1) the predominance of sand-filled channels, and 2) the relatively steeply sloping sides of the ridge.

With continuing northward retreat of the glacier to sites S-4, S-5, S-6, and S-7, the inlet became a large, asymmetrical embayment. A possible cause of asymmetry is the accumulation of mist on its eastern, downwind side.

Within the embayment an extensive subaqueous outwash fan built to the southwest, as flow from the meltwater conduit was deflected away from the relatively straight eastern ice wall. Meltwater discharge from the conduit appears to have increased by the time the glacier had retreated to this position, depositing boulders from suspension at the apex of the fan. This increase may reflect the evolution of the conduit system, capturing a larger volume of meltwater from a larger area on the glacier.

All of the facies of the subaqueous outwash fan described by Rust (1977) are present at exposures into the northern portion of the ridge. Some changes in the model are warranted, however, on the basis of new observations. First, climbing ripple drift is believed to have been deposited mostly on distal portions of the fan, beyond the channels, as well as on interchannel areas (Rust, 1977). Second, the intensely deformed sediments are believed to have been formed by deformation in slumps on the fan and above buried blocks of melting ice, rather than in channels, as described by Rust (1977, p. 182). Futhermore, it is possible that the presence of thick units of these soft-sediment deformation structures may only form in subaqueous outwash deposited in large embayments. It is in this environment that silty-fine sand, prone to deformation by dewatering, is trapped.

III. SOFT-SEDIMENT DEFORMATION STRUCTURES OF THE STITTSVILLE RIDGE

INTRODUCTION

The sedimentological investigation of the northern portion of the Stittsville Ridge, at sites S-4, S-5 and S-6, provided an excellent opportunity to study a number of soft-sediment deformation structures. Of particular interest are convolute stratification, ball and pillow structures and dish structures, each of which has been reported in sediments deposited in a variety of depositional environments. Of the mechanisms proposed for the formation of some or all of these structures cryostatic pressure, loading and fluid escape are the most frequently cited. An important feature of deformation structures in the Stittsville Ridge is their occurrence in a sequence consisting of (from the base) convolute stratification, ball and pillow structures and dish structures.

SOFT-SEDIMENT DEFORMATION STRUCTURES - PREVIOUS WORK

Terminology

The term "soft-sediment deformation structure" will be used to refer collectively to the three structures considered here: 1) convolute stratification, 2) ball and pillow structures, and 3) dish structures. The formation of each type involves, to varying degrees, the deformation of unconsolidated sediment.

The term "convolute bedding" was first introduced by Kuenen (1953) for a type of structure observed in turbidites.

Although now accepted as being purely descriptive, Kuenen (1953, p. 1056) stated: "... it is a type of bedding produced during the deposition in the highly mobile sediment ... it is not a post-depositional deformation." The term "convolute lamination" has also been used for the same structure, but is unsuitable, because the term "lamination" refers to layers of sediment less than 1 cm thick (Blatt et al., 1972) while convoluted layers often exceed this thickness. The term "involutions" has also been applied to convolute bedding. First introduced by Sharp (1942), it has been used almost universally to refer to convolute bedding in sediments which were thought to have been deposited in the periglacial environment. Because of its genetic implications this term should be **discarded.**

Despite Kuenen's (1953) original intention, the term "convolute bedding" has been widely used in the literature in reference to structures thought to have formed by a number of mechanisms. The term "convolute stratification" will be used here because of its general application to this structure, regardless of the thickness of the deformed unit. It refers to vertically restricted beds which have been deformed into a series of laterally repetitious, broad synclinal structures alternating with narrow anticlinal or diapiric structures Other terms will be used only when referring to other published reports.

Convolute stratification appears to have little environmental significance. It has been reported from turbidites, deltaic, fluvial, glaciofluvial and nearshore marine sediments.

The term "ball and pillow structures" is a version of the terms "ball- or pillow-form structures", first proposed by Smith (1916) who briefly described the two structures to which the term intended to apply. Ball-form structures (where first recognized) "... developed on the undersides of projecting ledges of weathered sandstone as rounded lobes and curved surfaces" (Smith, 1916, p. 148). Pillow-form structures differ in that they are characterized by well-developed lobes and curved surfaces on their top sides as well. Smith (1916) realized that these structures were not merely weathering phenomena by the fact that their internal bedding conformed to the curvature of their outer margin. Potter and Pettijohn (1963) modified Smith's (1916) term to ball-and-pillow structure.

The term "ball and pillow structures" will be used here as a non-genetic name for detached bodies of sand which which have a rounded external form, and commonly contain internal stratification bent upwards at the outer edges. The term is used in the plural, the singular being "ball structure" or "pillow structure" for spheroidal and more elongate forms, respectively. The structures differ from convolute stratification in that they are detached, and structures of the same lithologic characteristics are not necessarily repeated laterally within the unit. Like convolute stratification, ball and pillow structures appear to form in any environment in which sediment is being rapidly deposited (Reineck and Singh, 1976).

"Dish structure" was first introduced by Wentworth (1967) for structures observed in turbidites. Lowe and LoPiccolo (1974, p. 487) defined dish structure as "... thin, dark-coloured, subhorizontal, flat to concave upward lamination [which ranges] from clean sharp lines less than .2 mm thick to diffuse zones up to 2 mm thick, and in width from one to 50 cm." Like the other structures discussed above, dish structures appear to form in rapidly deposited sediments in a wide range of depositional environments. Nilson et al. (1977) reported dish structures in Precambrian turbidites and deltaic sediments, Oligocene fluvial sediments, and Pliocene lacustrine deposits; Rautman and Dott (1977) reported them from Jurassic shallow marine sandstones. Middleton and Hampton (1976) believed that dish structures are diagnostic of sediments deposited from fluidized sediment flows, in which some or all of the grains are suspended by escaping water. However, Lowe (1976) has suggested that they form in liquefied sediments, in which the rate of water escape is below that required to take grains into suspension but is sufficient to reduce their resistance to shear, causing the sediment to be mobile.

Mechanisms for the formation of some soft-sediment deformation structures

(1) Cryostatic pressure

Soft-sediment deformation structures are common in Pleistocene glacial sediments and were described as early as 1859 by Sorby. Sharp (1942, p. 128) suggested that involutions formed by "... repeated differential freezing and thawing and the formation of ground ice."

Washburn (1956) defined the term "cryostatic pressure" as "...the hydrostatic pressure set up in pockets of unfrozen material trapped between the downward freezing active layer and the permafrost table where the active layer becomes irregularly anchored to the permafrost table by freezing to it some places sooner than others" (Washburn, 1973, p. 86). He attributed the formation of the structures described by Sharp (1942) and others to such freezing-induced processes.

The mechanism described above has been adopted to account for many occurrences of deformed sediments in now temperate regions of North America, Europe and Asia. The presence of such structures has been used as evidence of periglacial paleoclimate conditions (e.g., Westgate and Bayrock, 1964). Theakstone (1965) accepted the same mechanism for structures formed during the Neoglacial.

Mackay and MacKay (1976) attempted to measure cryostatic pressures in non-sorted circles at Inuvik, Northwest Territories. Their study demonstrated that variations in soil pressure occur during the convergence of the upper and lower freezing planes during freeze-back (annual freezing of the active layer). However, as freeze-back occurred, the unfrozen material was not prone to deformation. This was attributed to the formation of ice lenses, which draw water from unfrozen materials, so that they become desiccated and more cohesive, and therefore less mobile.

The main conclusion of the work of Mackay and MacKay (1976) is that the effect of cryostatic pressure is of minimal importance as a mechanism for soft-sediment deformation in periglacial or any other environments. Furthermore, the formation of ice in sediments has the effect of reducing the potential for deformation.

(2) Loading due to reverse density gradients

Kuenen (1958) described the experimental formation of ball and pillow structures by the placement of a sand onto a bed of thixotropic mud, contained within an aquarium. Deformation was initiated by a blow from a rubber hammer, simulating the shock produced by an earthquake. Kuenen (1958, p. 20) noted that "local loads sank away quickly into the mud, the edges bent upwards and saucer or kidney shapes resulted. The experimentally produced forms closely resembled ball and pillow structures in natural sediments.

The mechanism of loading due to the existance of reverse density gradients has also been widely used to explain the formation of convolute stratification. Anketell et al. (1970) presented the most complete discussion of soft-sediment deformation structures resulting from loading. This work may be considered, at least partly, as a more specific explanation of the ideas of Kuenen (1958) and earlier workers. Anketell et al. (1970) described a reverse density gradient as one in which a sediment of density d_1 rests on a sediment of density d_2, where

$d_1 > d_2$. This situation is physically unstable because there is a natural tendency for the denser sediment to attain a position below the lighter sediment. The beds may remain metastable until the lower bed undergoes a reduction of its bearing capacity (Anketell et al., 1970, p. 5), defined as the "...load per unit area which the ground can support without excessive yield" (Gary et al., 1972, p. 65).

The reasons for reverse density gradients include: 1) differences in lithology, 2) differences in the void ratio of the sediment (Anketell et al., 1970) and 3) fluidization or liquefaction of the lower bed in an initially normal density gradient system (Allen, 1977). An interesting example of how a reverse density gradient may have originated is described by Emery (1950), who envisaged the migration of a sand bar onto a base of estuarine silts (of low density). In response to gravitational instability, the sand unit was thought to have sunk into the silt, forming "contorted beds". In turbidites, deltaic and fluvial sediments, the reverse density gradient may be readily explained by the repetitious deposition of sands over muds from currents of greatly varied competence. Deformation occurs because the sands dewater rapidly, whereas muds retain their high water content for much longer periods.

Anketell et al. (1970) demonstrated that different structures are produced by similar mechanisms (e.g., loading), depending on the ratio of the kinematic viscosities of the sediments involved. Where the material composing the upper of two parallel, horizontal beds has the higher kinematic viscosity, the lower bed will penetrate the upper bed, forming upward intrusions. If, however, the material composing the lower bed has the higher kinematic viscosity, downward penetrating intrusions will form. The fact that natural sediments are never truly homogeneous accounts for the irregularity and complexity of many structures observed in the geological record.

An important aspect of sediments deformed as a direct result of loading is the presence of excess pore water in the supporting bed. This reduces the intergranular friction of the sediment, causing a reduction of its bearing capacity. This accounts, in part, for the occurrence of deformation structures in fine sands which have undergone rapid deposition. Such sediments tend to have a relatively unstable packing arrangement and commonly have significant volumes of water trapped within their pore spaces. The requirement of excess pore-water is the reason why similar structures are abundant in glacial and/or periglacial environments. Butrym et al. (1964, p. 29) outlined the conditions that favour the development of soft-sediment deformation structures in these environments. They include: 1) abundance of meltwater, 2) the presence of frozen ground (inhibiting the downward escape of water), 3) the segregation of soils into layers of differing density, and 4) the absence of vegetation (which would extract water from the sediments and tend to bind the material).

McArthur and Onesti (1970) interpreted contorted structures in Pleistocene sands overlying clays to be the result of periglacial conditions. The structures were attributed, in part, to the seasonal thawing of ground ice. The authors suggest that loading occurred as a result of liquefaction due to the contribution of water to the lower sediment by melting ice. McArthur and Onesti (1970) outlined a sequence of events thought to lead to the formation of the structures observed, which may be summarized as follows: 1) deposition of sand over a frozen clay substratum containing ground ice, and 2) thawing of the substratum and foundering of the sand into the clay.

(3) Fluid escape

Even though a reverse density gradient may exist, at least one of the beds involved in the development of soft-sediment deformation must have a high water content. Butrym et al. (1964) noted that in their experiments that liquefaction of a sediment is followed by rapid dewatering. The energy expended by the fluid was not, however, thought to be an important input for the deformation of the sediment. Allen (1977, p. 22) stated that it is "presumed that the mechanism of liquidization [syn. liquefaction] is one involving passive fluid". This statement illustrates the difference between loading and fluid escape as causes of soft-sediment deformation. Fluid escape by liquefaction and/or fluidization causes sediments to become mobile and capable of actively deforming more cohesive materials.

Some authors, who accept the explanation of deformation structures by loading, also indicate that the escaping fluid may not be passive. Butrym et al. (1964, p. 3) stated that "under certain conditions, the overlying sediment is divided into pieces by the intruding quick sand and silt" during the formation of load structures. Furthermore, "...the isolated pieces may sink down into the fluidized substratum and their edges may be bent upwards". This clearly suggests that mobile sediment-water mixtures may have a more active role in the formation of some soft-sediment deformation structures (the inference being that fluid escape may be a mechanism for the formation of ball and pillow structures). Davies (1965), who accepted loading as the mechanism for formation of convolute stratification, noted that the anticlinal portions of the structures are often ruptured. He explained these ruptures as being the result of "relatively rapid expulsion of local concentrations of pore water along an inherent line of weakness" (Davies, 1965, p. 318).

Other authors have attributed soft-sediment deformation structures directly to fluid escape. Rust (1977, p. 180), for example, implied that ball and pillow structures observed in Pleistocene deposits near Ottawa may have been formed by fluid escape. Sharp and Wood (1974) suggested that some soft-sediment deformation structures form as a result of fluid escape by the processes of fluidization, dilation, expulsion and/or elutriation. Johnston (1977, p.409) described convolute stratification in Precambrian sediments, with diapiric structures, developed at sites "where sediment and water was extruded vertically from the liquefied bed". The synclinal troughs were thought to have developed when the cohesive sediment between the diapiric structures sank downwards as underlying material was mobilized, as a result of high pore water pressures, and extruded vertically.

Lowe (1975) recognized that the anticlinal ridges of convolute stratification in Carboniferous flysch deposits functioned as escape paths for vertically flowing fluids. He also observed that the bases of the synclinal troughs were sometimes eroded. This latter feature was explained by high rates of dewatering, physically eroding the base of the deformed beds. In such cases the anticlinal ridges were extremely ruptured, forming "broad diapiric structures" (Lowe, 1975, p. 188.

Lowe (1975) asked the crucial question of whether fluid escape is the cause of convolute stratification or an effect of its formation. He answered this question by describing how fluid escape initiates deformation: "...most convolute lamination begins to form when cohesive forces and low permeability within accumulating sediments begin to inhibit the free upward flow of escaping pore fluids" (Lowe, 1975, p. 188).

Dish structures

Wentworth (1967, p. 493) attributed dish structures to scouring by currents in association with antidune bed forms in turbidity currents, with modification of the structure due to dewatering of the bed. Stauffer (1967, p. 493) also attributed their origin to largely primary depositional processes.

Lowe and LoPiccolo (1974, p. 494) suggested that dish structures "...form when consolidation occurs relatively slowly and upward-moving water becomes trapped beneath semi-permeable laminations, escaping by leakage through and flow around rather than explosive disruption of the confining laminations". The vertically flowing water reaches a lamina of fine sediment and begins to flow horizontally. Vertical flow begins again when the water breaches the lamina. The drag exerted on the lamination by the water causes an upward displacement of its edges, forming the dish structure.

SOFT-SEDIMENT DEFORMATION STRUCTURES OF THE STITTSVILLE RIDGE

Convolute stratification

Convolute stratification was observed most commonly at site S-4, while only one poor exposure containing this structure was present at site S-6. Two types of convolute stratification were recognized: 1) Those which occurred at the top of exposures, where previously overlying sediments had been removed during excavation. The synclinal portions of these structures generally extend more than 2 m across the exposures and are in excess of 1 m high. 2) Convolute stratification within the exposures, overlain by sediments containing ball and pillow structures and dish structures. The synclinal portions of this type are generally less than 1 m high and commonly extend less than 2 m across the exposures.

The larger type of convolute stratification consists of broad synclinal troughs, with relatively narrow diapiric structures intervening. At site S-6 convolute stratification was exposed in plan, appearing as ellipsoidal troughs exceeding 3 m in length. Unfortunately, only a few of the structures were exposed so that it was not possible to determine whether they have a preferred orientation.

The sediment within the synclinal portions of these large type of convolutions is predominantly massive to faintly laminated fine to medium sand, immediately underlain by a silty fine sand (figures 3-1 and 3-2). Within the diapiric structures is silty fine sand with vertically trending and swirled thin beds of fine sand. Sills and dykes of silty fine sand commonly extend from the diapiric structures, penetrating the massive and faintly laminated sands within the synclinal structures (letter "A" in figure 3-1 and figure A-3, at 15 m). An interesting occurrence of much smaller scale penetrations was observed on the outer wall of a diapiric structure which had become exposed by wind erosion of less cohesive massive sand from the synclinal trough of a convoluted unit at site S-4. They consist of small (generally less than 3 cm across), commonly elongate, bubble-like structures (figures 3-3 and 3-4). Many of these structures are open, exposing tubes which appear to be hollow to a depth of as much as 3 cm, and which have flared ends. Bubbles near the base of the diapiric structure, where the slope is about 45°, have flat tops.

Below the unit of convolute stratification at site S-4

Figure 3-1. Large scale convolute stratification at site S-4. Note the deformation of underlying sediments into convolute stratification with intact anticlines, and silty fine sand dyke penetrating the synclinal portion of the structure (arrowed). Rod 3.5 m long.

Figure 3-2. Large scale convolute stratification at site S-6. Note particularly the swirled laminae in the diapiric portion of the structure, and the downward decrease in the intensity of deformation. Pen 15 cm long.

Fig. 3-1

Fig. 3-2

Figure 3-3. The outer wall of the anticlinal portion of convolute stratification at site S-4 exposing elongate "bubble" structures. Lens cap diameter (see arrow) 5.5 cm.

Figure 3-4. Close-up of "bubble" structures in figure 3-3.

Fig. 3-3

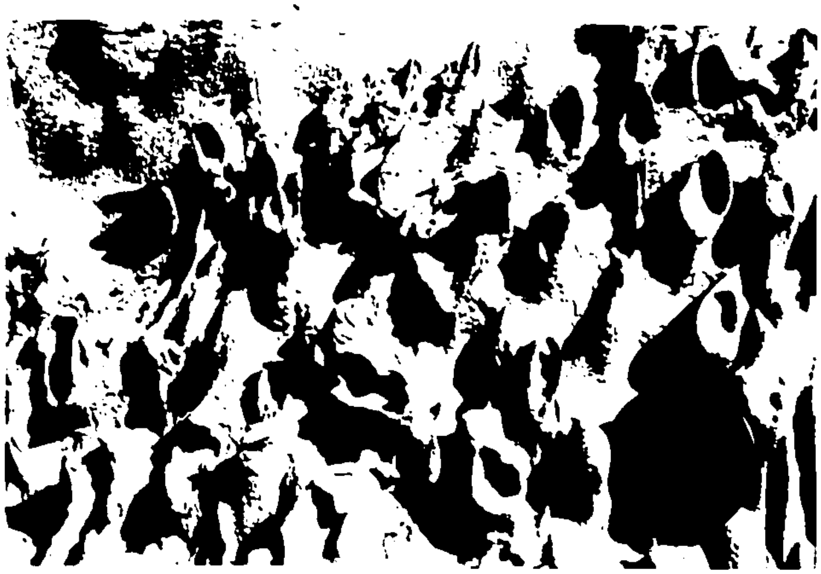


Fig. 3-4

are stratified sands deformed into convolute stratification with anticlinal structures, rather than diapirs. At site S-6 the underlying sediments are type A climbing ripples in silty fine sand and are only slightly deformed, the intensity of deformation decreasing downward from the base of the convoluted layer. The deformed unit could not be traced laterally at site S-6 due to poor exposure. At site S-4, however, the convoluted unit was observed to pass laterally into a unit of climbing ripple drift, interstratified with massive sands. The transition between the deformed and adjacent undeformed beds is gradational, the amplitude of the folds increasing towards the unit of convolute stratification.

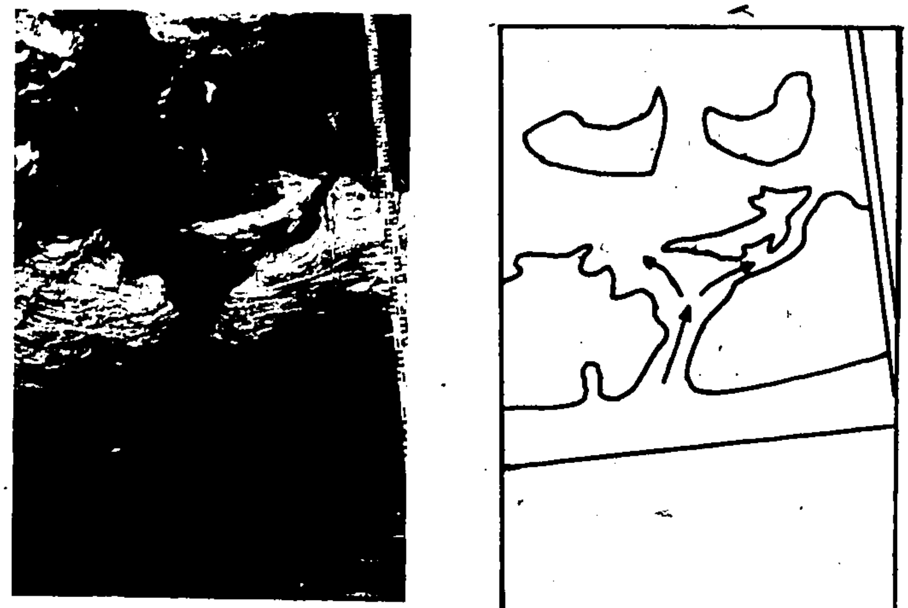
This large type of convolute stratification is similar to that described by Davies (1965) and Lowe (1975, p. 187). Both of these authors recognized the diapiric portion of the structures to be vertical channels through which fluidized sediment flowed during rapid dewatering. The mechanisms which may cause rapid dewatering of a sediment have been reviewed in Chapter II (p. 94). It is suggested that water was added from sediments below the convoluted layers, deforming them hydroplastically into synclines and anticlines. Moving upwards, this fluid encountered less permeable silty fine sands, causing flow to move laterally until sufficient volumes of water were available to locally fluidize the overlying sediment. At that point flow progressed upwards, forming the diapiric structures within the convolute stratification as vertical channels of fluid escape. Mud size sediment within the vertical channels may have been locally elutriated, forming the vertically trending fine sand beds observed within the diapirs. The synclinal troughs formed as the sediment between the vertical channels sank downwards in response to the removal of underlying material. Sills and dykes of silty fine sand probably formed as mobile sediment locally broke through the walls of the vertical channels. The smaller bubble-like structures may have developed during the final stages of deformation as silty fine sands plugged the vertical channels. The less intense fluid discharges then percolated through the relatively thin channel walls causing a mud lining to be hydroplastically deformed into the bubble-like structures. The still vertical direction of fluid escape is reflected in the vertical elongation of the structures. The flattened bubbles on the lower slopes of the vertical channel were probably further deformed by late stage adjustments of the overlying sands within the synclinal portion of the convolute stratification.

The second type of convolute stratification was observed at site S-4, and occurs near the base of the slump structures described in Chapter II, where they are overlain by ball and pillow and dish structures (figures 3-5 and 3-6). Unlike the larger type of convolute stratification, the synclinal portions of this type are filled with cross-stratified fine to medium sand, turned upwards at the edges of the structures. Like the larger type, diapiric structures filled with silty fine sand

Figure 3-5. Convolute stratification overlain by ball and pillow structures and dish structures (not visible on this photograph) at site S-4. Note also the vertically displaced portions of the convoluted unit (see arrows) and the dispersed ball and pillow structures at the top left of the photograph. The black portion of the shovel handle is approximately 30 cm long.

Figure 3-6. (A) Convolute stratification overlain by ball and pillow structures and dish structures (not visible on this photograph) at site S-4. (B) Sketch of adjacent photograph emphasizing bifurcation of the vertical channel above the convoluted unit (shown by arrows) and the partial penetration of the convoluted unit (at letter "A") by silty fine sand. Smallest interval on scale is 1 cm.

Fig. 3-5



(A) (B)

Fig. 3-6

occur between the synclinal portions of the convolute stratification.

Although this second type of convolute stratification is morphologically different from the larger type, it is believed to have formed by similar processes of rapid dewatering. Again, the process involved the injection of fluidized sediment through diapiric structures. The force of this injection is reflected in the bending of strata at the edges of the synclinal troughs where frictional drag was sufficient to deform the cross-stratified sediments. More dramatic evidence of the force exerted by vertical injection is the upward displacement of a synclinal trough to a position relatively above the general level of the convoluted unit (figure 3-5). This displacement has effectively converted the synclinal trough into a ball and pillow structure.

Note also in figure 3-5 the absence of diapiric structures in the cross-stratified sands laterally equivalent to the convoluted unit and that in figure 3-6 one of these structures (labelled "A") does not penetrate completely through the cross-stratified sands. The latter feature illustrates the relatively local nature of penetration by the mobile sediment and water mixture. Furthermore, it reflects the fact that the diapirs are not sheets which can be traced indefinitely along their length, but probably are conical or cylindrical in form. If excavation had continued further into the exposure in figure 3-3 it is probable that diapiric structures would have been observed penetrating the apparently undisturbed cross-stratified unit. Unfortunately this was not possible because of the instability of the face.

Ball and pillow structures

Ball and pillow structures were observed at site S-4, S-5 and S-6. Rust (1977, p. 178) described these structures as "...rounded, detached synclinal folds up to 2 m in width, and with limbs bent inwards towards the axial plane". He also noted (Rust, 1977, p. 179) that there was a decrease in the size of ball and pillow structures upwards, passing into dish structures.

In all but one of the exposures studied the ball and pillow structures were observed in units which extended below the floors of the sand pits. (Ball and pillow structures observed directly above convolute stratification were somewhat different and will be discussed at the end of this section.) In each of these exposures the structures were as described by Rust (1977). They occur in high concentrations within the unit, the spacing between individuals being generally less than 20 cm at their closest point, although this distance increases between larger structures. The sediment between the structures is a silty fine sand which commonly contains subvertical laminae which are similar to those reported earlier in the vertical channels associated with large scale convolutions. The material between the largest of these structures contains smaller ball and pillow structures. The largest of the ball and pillow structures observed was at site S-6 and was almost

7 m in width (figure 3-7). The smallest were observed near the base of the unit of dish structures, being generally less than 15 cm wide (figure 3-8). The smaller ball and pillow structures differ from the dish structures in their larger size, coarser grain size and the occurrence of deformed internal stratification.

The stratification within the structures is commonly bent upwards towards a subvertical axial plane, although some exceptions were observed (figure 3-9). Another exceptional case is shown in figure 3-10: a pillow with a subvertical axial plane, but with internal stratification bent downwards. Another characteristic, mentioned in Chapter II, is the occurrence of marine fossils in the smaller, uppermost structures at site S-6.

Regarding the formation of ball and pillow structures in sediments within the Stittsville Ridge, Rust (1977, p. 178-179) suggested that the process involved was one of rapid dewatering, similar to that described earlier in this chapter for the formation of convolute stratification. In the formation of ball and pillow structures, however, the unit disrupted by the rising columns of fluid is sufficiently thick and cohesive to allow the pieces to sink downwards, forming detached structures which can no longer be related to their original stratigraphic position.

Traditional models for the formation of ball and pillow structures assume that the edges of the structures became bent upwards while sinking into the underlying sediment (Kuenen,

Figure 3-7. Pillow structure almost 7 m across at site S-6. Back-pack 70 cm high.

Figure 3-8. Small ball and pillow structures at their upper limit at site S-4, overlain by dish structures. Note dish-like laminae marking the base of the ball and pillow structures. Scale in 2 cm intervals.

Figure 3-9. Large pillow structure with near-horizontal axial plane at site S-6. Back-pack approximately 70 cm high.

Fig.3-7

Fig.3-8

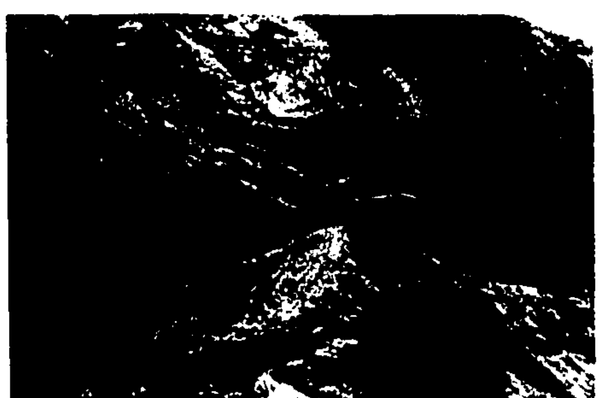


Fig. 3-9

Figure 3-10. Inverted pillow structure at site S-4. Scale in 2 cm intervals.

Figure 3-11. Small pillow structure within dish structures at site S-4. Note well-defined dish-lamina marking the lower boundary of the pillow structure. Scale in 1/10 and 1/50 foot intervals.

Figure 3-12. Sequence of dish structures overlying ball and pillow structures at site S-4. Note the general increase in dish length and vertical spacing, and the decrease in dish concavity at higher levels up the exposure. Smallest interval on scale is 1 cm. The holes are entrances to swallow's nests.


Fig. 3-10

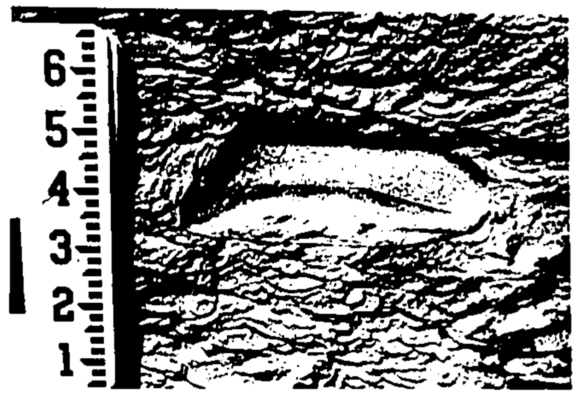


Fig. 3-11


Fig. 3-12

1958). It seems reasonable to suggest, however, that if the vertically flowing column of sediment and water was capable of turning up the edges of the convolute stratification, that sufficient force was available to deform the edges of the ball and pillow structures. The internal stratification in the structures shown in figures 3-9 and 3-10 are believed to have been deformed by the penetration of a rising column of fluid, but they appear to have rotated while sinking downwards. It is important to note that in the case of the inverted structure (figure 3-10) the frictional drag exerted on its outer edges while sinking appears to have had no effect on its internal stratification. The structure in figure 3-9, however, appears to have been further deformed during rotation, probably as the matrix became less mobile due to dewatering.

The upward decrease in size of the ball and pillow structures is believed to reflect a tendency for the vertical channels to bifurcate upwards, probably because of decreased confining pressure at shallower depths. Such bifurcation was observed at site S-4 and is shown in figure 3-6, where a single vertical channel broadens and splits into two channels after passing through the convoluted cross-stratified unit. As a result of bifurcation the active vertical channels broke up progressively smaller pieces of sediment at higher levels in the disrupted unit. The smallest ball and pillow structures were derived from the uppermost sediments in the disrupted unit, which at site S-6 contained marine mollusks at the time of deformation. Unfortunately, the bifurcating systems of vertical channels probably have a low preservation potential due to adjustments made by the detached ball and pillow structures. For this reason they are only rarely observed in the field.

Somewhat different ball and pillow structures were observed directly above convolute stratification on the edges of what are believed to be slumps which formed on the subaqueous outwash fan at site S-4 (figure 3-5). The most striking difference is that these structures occur in relatively low concentrations. They also lack internal stratification, although this is probably because they were derived from a bed of massive sand. There is, however, a well defined layer of mud at the outer margin of each ball and pillow structure. The silty fine sand between the ball and pillow structures contain dish structures.

The relatively low concentration of ball and pillow structures at this site is believed to be directly related to their position on the margin of a slump. There, the total thickness of the sediment column involved in slumping was less than towards the centre of the slump, and as a result the confining pressure exerted on the pore fluids was lower. When failure occurred, dewatering took place predominantly by uniform liquefaction on the shallow edges of the slump, rather than by fluid escape through vertical fluidization channels. As mentioned earlier, however, fluidization does appear to have occurred at the base of this portion of the slump, giving rise to convolute stratification. In the overlying sediment the widespread occurrence of dish structures, which are regarded as diagnostic of sediments which have undergone liquefaction (Lowe, 1976), supports this interpretation. The ball and pillow structures in this unit are believed to be pockets of sediments which remained unliquefied because of the textural inhomogeneity of layered sediments (Lowe, 1976, p. 294). More specifically, the presence of a cohesive bed of mud size sediment immediately below the massive sands making up the ball and pillow structures may have prevented it from being liquefied. This protective bed is now preserved as the layers of mud on the outer margins of the ball and pillow structures.

Dish structures

Rust (1977) first noted dish structures overlying ball and pillow structures in sediments within the Stittsville Ridge at site S-6. The dishes were described as concave-upward laminae which were "...sharply defined silt-rich layers about 0.2 mm thick, separated by silty sand which is generally structureless, but in some cases contains faint laminae" (Rust, 1977, p. 179). A tendency for dish concavity to decrease upwards was also reported.

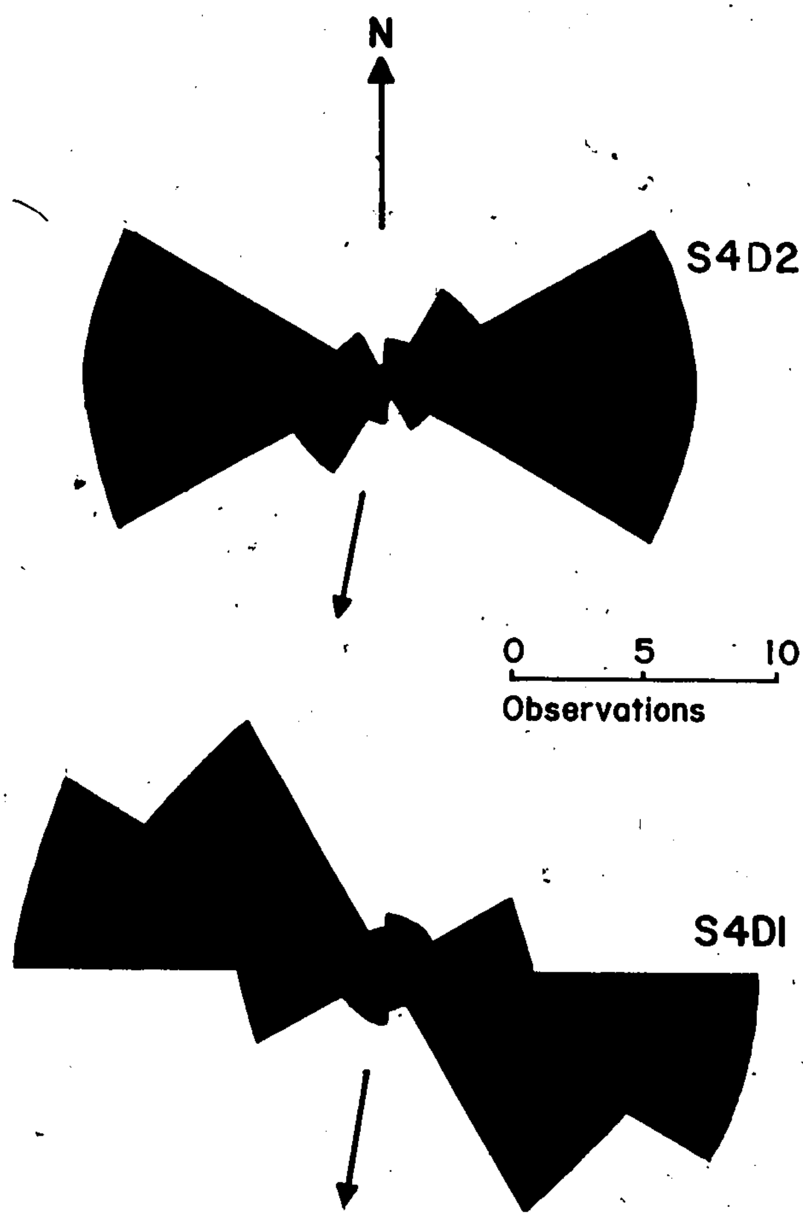
In the course of the present study dish structures were also observed primarily above ball and pillow structures in silty fine sands. The transition between the ball and pillow structures and dish structures is gradual, especially at site

S-4, occurring over a vertical distance of over 30 cm (figure 3-8). The lowermost dishes are within the vertical channels between the ball and pillow structures while the uppermost ball and pillow structures are entirely surrounded by sediment containing dish structures. A striking feature of these uppermost ball and pillow structures is that their lower margins are marked by concave-upward dish-like laminae (figures 3-8 and 3-11). They differ from true dish structures in their larger size, coarser sediment above the laminae, the greater thickness of the basal laminae and internal laminae.

As reported by Rust (1977) dish structures at site S-4 exhibit an upward decrease in concavity. There is also, at all locations included in this study, an upward increase in the apparent length of the dishes and the vertical spacing between them (figure 3-12). The horizontal spacing between the up-turned edges of adjacent structures varies from approximately 1 cm to 0, where the edge of one structure cross-cuts another.

The morphology of dish structures in plan was observed by excavating a horizontal terrace into the face of the exposure at site S-4 and by digging down from the top where the unit of dish structures is near the surface. The dish structures commonly have an elliptical form, although some are circular in plan. More complex shapes were also observed. Long axis orientation of elliptical dishes are summarized in figure 3-13, and show a preferred orientation approximately normal to the local paleocurrent direction.

Figure 3-13. A summary of orientation data derived from dish structure long-axes orientations. The arrows indicate the local paleocurrent direction based on sample S4P5, from climbing ripples, also shown in figure 2-24.

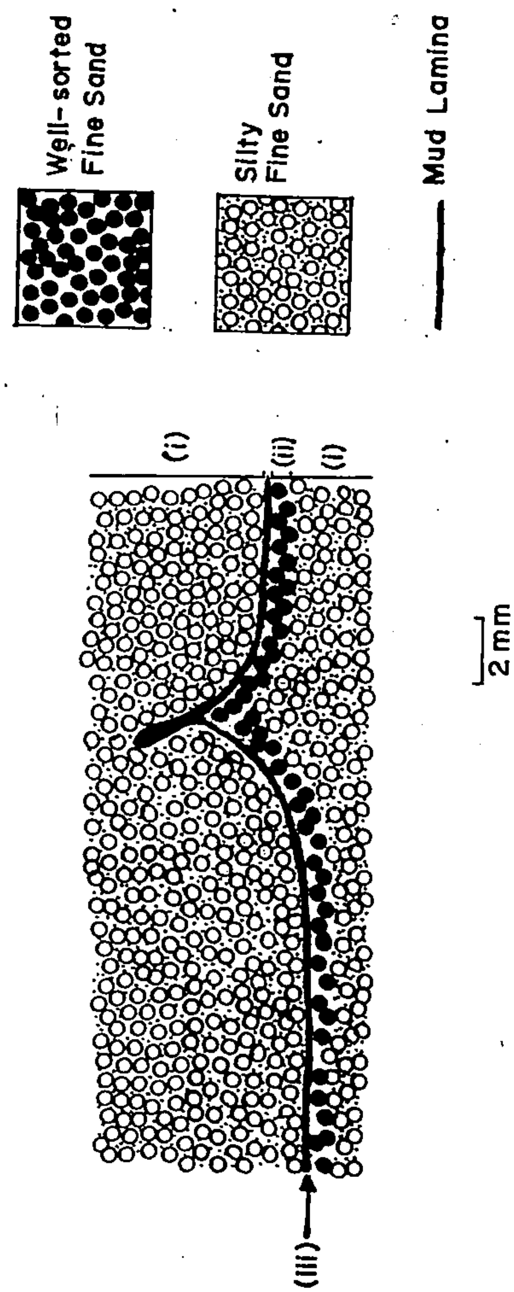

N
S4D2
0 5 10
Observations
S4DI

To provide a more detailed description of the textural characteristics of the dish structures, peels were made in the field by pouring clear shellac onto the vertical surface of the exposure. After approximately 3 hours the shellac was sufficiently dry to allow the removal of a solid block of sand. The blocks were then cut parallel to the vertical surface and examined under a binocular microscope. Figure 3-14 illustrates these textural details.

The sediment surrounding the dish laminae is a poorly sorted silty fine sand. The individual laminae, however, consist of a thin zone, approximately 1 mm thick, which contains fine sand and is largely lacking in silt and clay size sediment (figure 3-14). This zone gradationally overlies the poorly sorted sediment while its upper margin is sharply defined by a very thin lamina (less than 0.1 mm) of mud-size sediment. Immediately above this lamina the sediment is identical to that below the dish structure. The dishes themselves consist of both the zone of relatively well sorted fine sand and the overlying mud lamination. Between the up-turned edges of adjacent dish structures that do not cross-cut, the sandy laminae pass gradually into the poorly sorted sediments. Where an intersection between adjacent dishes does occur, however, the silt laminae are observed to thicken considerably, to approximately 0.3 mm (figure 3-14).

The silty fine sand in which the dish structures occur is believed to have been transported upward through vertical

Figure 3-14. Schematic illustration of grain size distributions in dish structures, showing: (i) Poorly sorted silty fine sand above and below the dish lamina; (ii) well sorted fine sand, sharply overlain by (iii) a thin lamina of mud-size sediment.


Well-sorted Fine Sand
Silty Fine Sand
Mud Lamina
(i)
(ii)
(i)
(iii)
2 mm

channels which formed the ball and pillow structures. At the upper ends of these channels the mobile sediment accumulated above the sinking ball and pillow structures. The occurrence of dish structures in these sediments indicates that following their passage from the vertical channels they were in a liquefied state (Lowe, 1976), probably due to a reduction in the rate of pore water escape.

The mechanism by which dish structures form in a liquefied sediment has been discussed at length by Lowe and LoPicollo (1974) and will be briefly reviewed here with reference to the observations described earlier. It is believed that water passing upwards through silty sand encounters an impermeable barrier (a lamination) which forces flow to move laterally, beneath the restriction. Flow progresses in this way until a weak point in the lamination is encountered or until sufficient force is available to breach the barrier. At this point the water is forced rapidly upwards, turning the edges of the lamination upward, forming the typical dish structure. It is also during this lateral flow and upward escape that mud-size sediment is elutriated from below the lamination, forming the relatively well-sorted zone of fine sand. The mud-size sediment was probably deposited along the upper margin of the zone of well sorted sand, forming the mud lamina, and where the barrier was breached (figure 3-14).

The origin of the original impermeable lamina is problematic because the sediment containing the dishes was derived from underlying materials, rather than from a laterally flowing primary current. It seems most probable that the dish structures in the Stittsville Ridge formed from laminae which developed: 1) as consolidation laminae (Lowe, 1975) by the processes of elutriation and deposition during water escape; or 2) during flowage of the liquefied sediment by some process which caused segregation of fines along horizontal planes within the flow; or 3) by some combination of these processes. The first possibility (consolidation lamination) is well accepted in the formation of dish structures which cut modified primary structures (described by Lowe and LoPiccolo, 1974). It is difficult, however, to postulate a reason for the initiation of the formation of consolidation laminae in structureless sand. The solution to this problem is made more difficult by the fact that the original laminae probably underwent some textural modification during dish formation, masking textural clues to their origin.

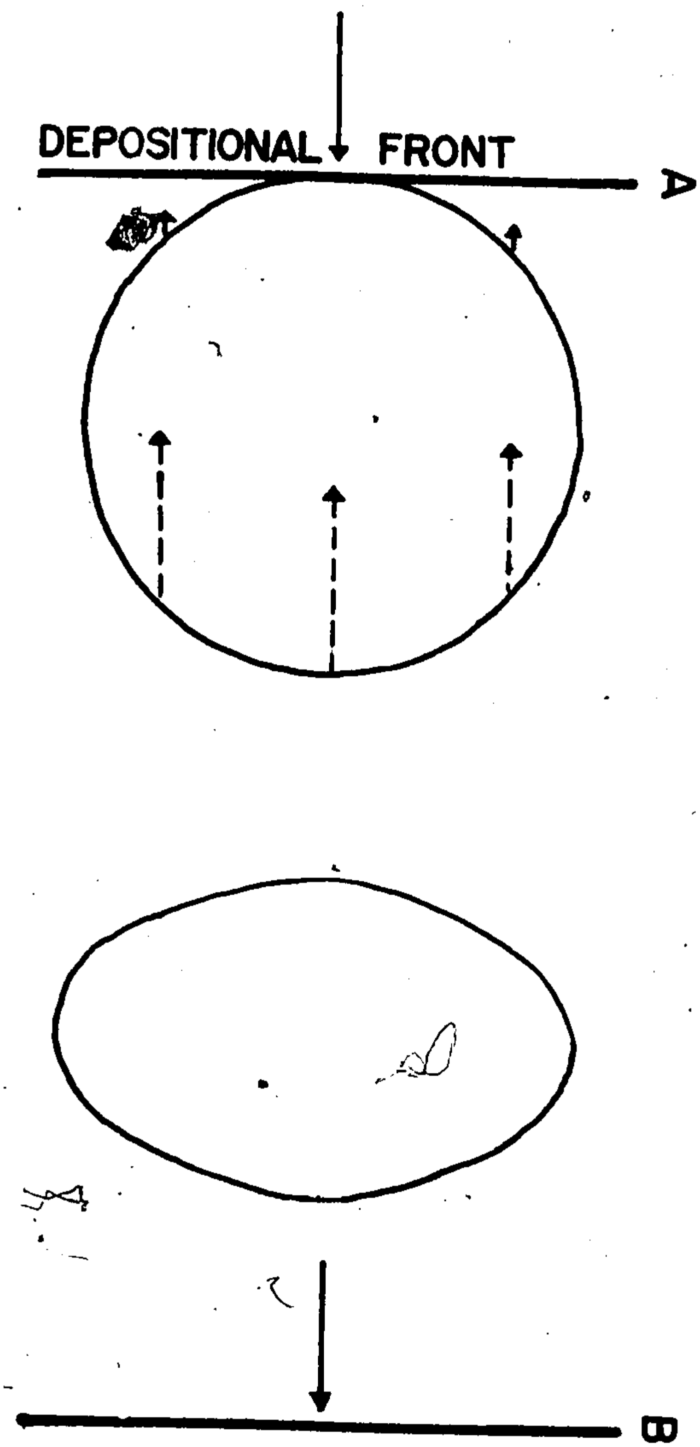
A more specific origin may be postulated for the mud laminae which define the lower margin of the smallest ball and pillow structures. It is quite possible that these are consolidation laminae which formed as water and fine sediment, passing through the silty-fine sand, encountered the more permeable sand making up the ball and pillow structures. There, flow between the pore spaces was less restricted (flow was expanded) and the fine sediment in suspension was immediately deposited, forming laminae which acted as impermeable barriers to continued vertical flow and causing the formation of the dish structure by the processes described earlier.

The preferred orientation of the dish structures at site S-4 is attributed to downslope flowage of liquefied sediments (figure 3-15). The initial plan form of the dish structure is assumed to be circular, on the assumption that their development is isotropic, i.e., the rate of escape is constant within the sediment and that the restricting lamination is of uniform composition. In figure 3-15A the dish structure is moving downslope to the left, within the flowing liquefied sediment. The heavy line in figure 3-15 is a depositional front, along which flow ceases due to rapid consolidation of the liquefied bed. As can be observed in grain flows in dry sand, the depositional front probably moved upslope, which is to the right in figure 3-15. The dashed lines with arrows indicate the horizontal distance which each respective point on the dish will move as the depositional front passes through that location. The front reaches the downslope edge of the dish before it reaches the upslope edge so that the latter will continue moving after the former has stopped. The result is a compression of the dish structure in the downslope direction, after the depositional front had passed (figure 3-15B).

DISCUSSION

While crystatic pressure and loading are widely

Figure 3-15. The modification of the plan form of dish structures by lateral compression at the cessation of downslope flowage. See text for a detailed discussion.


DEPOSITIONAL FRONT
A
B

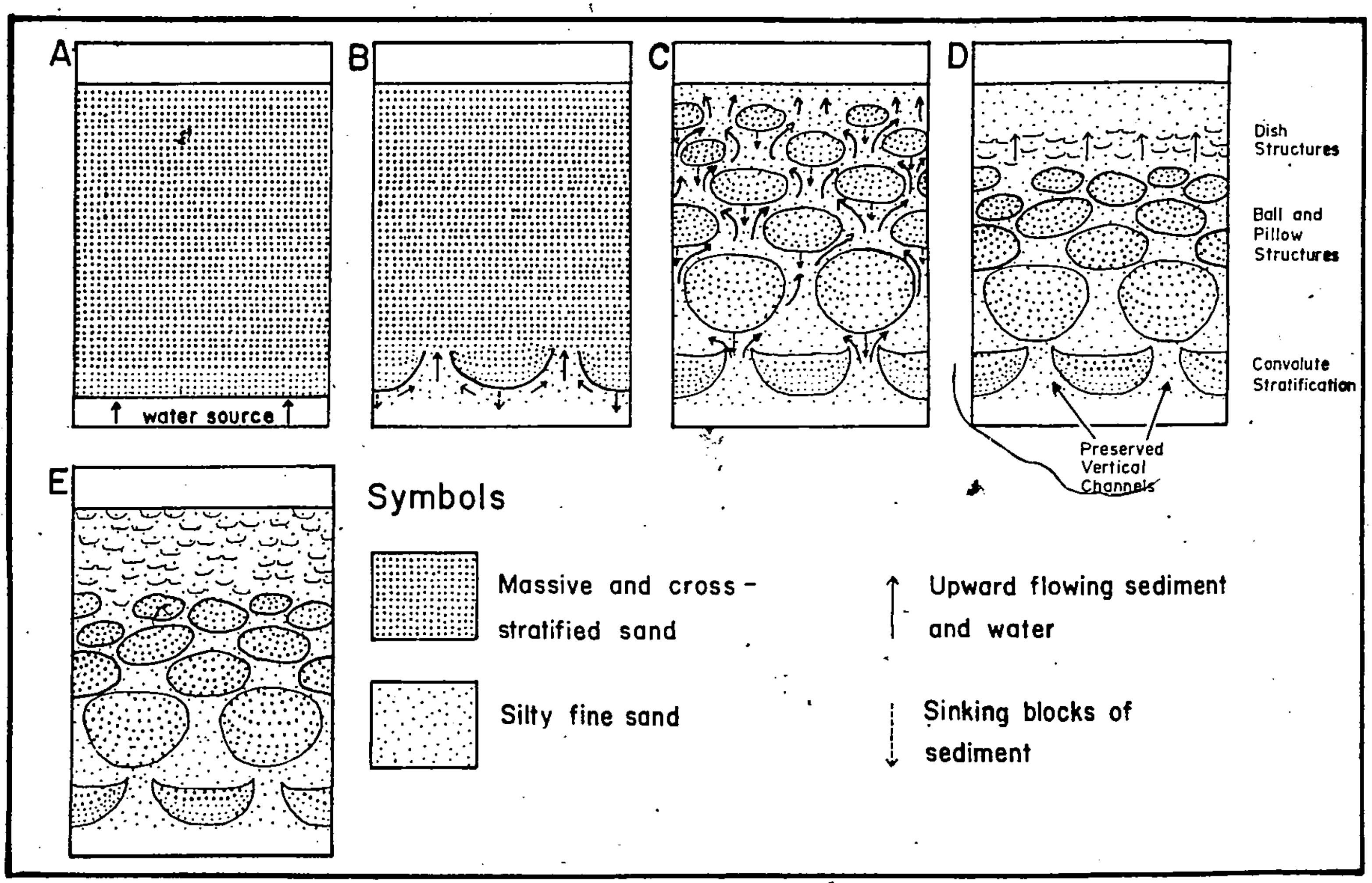
believed to be responsible for the formation of many soft-sediment deformation structures, rapid fluid escape appears to be the most important mechanism of deformation for the structures described here. Their occurrence in a vertical sequence, consisting of (from the base upwards) convolute stratification, ball and pillow structures and dish structures, provides further evidence of their common genesis.

Figure 3-16 depicts a hypothetical model which summarizes the interpretation of each structure and the probable development of the sequence. This sequence has been observed in the field, and is regarded as an ideal sequence, which may or may not occur in a complete form elsewhere.

In stage 1 of the model (figure 3-16A) the pre-deformational situation is shown. Beds of massive and cross-stratified fine sand and silty fine sand overlie a source of water. This source may be coarser sands undergoing consolidation, the excess pore water moving upward by seepage, or melting of buried ice. The fine sediments remain stable until some shock initiates rapid dewatering or until the pore water pressure is sufficient to initiate fluidization of the basal fine sediments.

In stage 2 of the model (figure 3-16B) rapid dewatering of the sediment has begun with the local fluidization of the lower beds of relatively impermeable silty fine sand. The result of this stage is the establishment of vertical fluidization channels, which give rise to a basal unit of convolute stratification.

Figure 3-16. A hypothetical model for the formation of the sequence of deformation structures consisting of (from the base) convolute stratification, ball and pillow structures and dish structures. See text for a detailed discussion.


A
B
C
D
E
water source
Dish Structures
Ball and Pillow Structures
Convolute Stratification
Preserved Vertical Channels
Symbols
Massive and cross-stratified sand
Silty fine sand
Upward flowing sediment and water
Sinking blocks of sediment

Stage 3 (figure 3-16C) includes the disruption of overlying beds by the vertical fluidization channels. Near the base, just above the convolute stratification, a few large, widely spaced channels disrupt the overlying beds into large detached blocks. The edges of these blocks are turned upward due to the frictional drag exerted by the vertical flow, forming ball and pillow structures. The detached pieces then sink downwards, in response to the removal of underlying sediment, distorting or completely destroying the vertical channels. In most cases the structures sink symmetrically, but a few develop asymmetrically or are overturned. At higher levels the vertical channels bifurcate, forming smaller, more numerous ball and pillow structures, and finally terminate at the upper limit of the stratified sediment, which is gradually moving downward. At this upper limit the fluidized sediment which was transported through the vertical channels accumulates at the top of the sediment column. Under lower confining pressures, at this level, the pore water pressure in the previously fluidized sediment decreases, allowing most of the fine sediment which had been held in suspension by the escaping fluid, to settle out. At this level the sediment is still liquefied, although no longer fluidized, because of the vertical water movement.

In stage 4 of the model (figure 3-16D) the uppermost liquefied unit has reached its maximum thickness as the addition of material from below ceases. Still mobile, due to its relatively high pore water pressure, it may flow down the gentle slope of the subaqueous outwash fan. As dewatering continues, circular dish structures form as water and clay size sediment still in suspension, encounter relatively impermeable laminae. These may be later deformed, as downslope flowage ceases, into elliptical structures with their long axes oriented normal to the downslope direction. The cessation of downslope flowage marks the completion of dewatering, the sequence of structures having been formed (figure 3-16E).

An important aspect of this hypothetical sequence of events, in terms of the paleoenvironmental reconstruction of the sediments at site S-6, is that the ball and pillow structures can sink well below their original stratigraphic position. They are therefore exotic, so that fossils occurring within them do not necessarily indicate the age or environment of the enclosing material.

IV. CONCLUSIONS

The Stittsville Ridge, along its entire length, is composed of sediments deposited on subaqueous outwash fans, within the confines of an inlet into the front of the retreating Wisconsin glacier. The inlet appears to have become larger and increasingly asymmetrical as the ice retreated to the north. The result of this change in ice front morphology was the considerable variation in the sedimentology of the ridge along its length.

The following general conclusions regarding subaqueous outwash may be drawn from the work presented in Chapter II:

1) The morphology of the ice front at the site of deposition exerts a strong control on the sedimentology of the deposit;

2) Changes in the morphology of the ice front over time cause corresponding changes in the sedimentology of the linear subaqueous outwash ridge;

3) Extensive faulting along the flanks of subaqueous outwash is believed to indicate that deposition was within the confines of an inlet at the ice front. This conclusion may not be drawn, however, without additional sedimentological evidence. The problem of interpretation is made more difficult by the fact that ice-contact deformation and subsequent wave action may destroy many features which are believed to be diagnostic of subaqueous outwash;

4) Clast supported gravel, deposited near the mouth of the

glacial meltwater conduit, may have a-axes oriented normal or parallel to the flow direction, depending on the flow activity;

5) Channels on the subaqueous outwash fan may be filled by sediment gravity flows or by primary currents depositing sediments on an active bed;

6) Subaqueous outwash fans which were deposited in relatively narrow inlets into the ice front appear to be dominated by channels which commonly contain massive or laminated sand and pebbles and boulders. These deposits preserve little inter-channel cross-stratified sediments and generally lack mud-size sediment;

7) Soft-sediment deformation structures, including convolute stratification, ball and pillow structures and dish structures, are believed to have developed in slumps on the subaqueous outwash fan and above buried masses of glacial ice.

Observations of convolute stratification, ball and pillow structures and dish structures in sediments within the Stittsville Ridge have led to the conclusion that all three are genetically related, having formed in response to rapid dewatering of the sediment. The ideal sequence of these structures is (from the base): convolute stratification, ball and pillow structures and dish structures. The sequence illustrates the relationships between the structures and the specific events which led to their formation. At the present time the origin of the laminae from which dish structures form is not well understood.

Further work on the Late Quaternary deposits in the Ottawa area should include:

1) Surveillance of active excavations into the Stittsville Ridge;

2) Detailed investigations of other sand and gravel ridges in the Ottawa area;

3) A project integrating these detailed studies into a complete history of the formation of the linear ridges in the Ottawa area and a compilation of all information regarding subaqueous outwash, with the goal of deriving a more complete depositional model;

4) The development of an experimental project on the formation of dish structures, including a detailed theoretical consideration of the mechanics of their formation.

REFERENCES CITED:

Aario, R.

1971: Associations of bed forms and paleocurrent patterns in an esker delta, Haapajarvi, Finland; Annales Academiae Scientiarum Fennicae, Series A, III. Geologica-Geographica, III, 55 p.

1972: Exposed bed forms and inferred three-dimensional flow geometry in an esker delta, Finland; 24th Int. Geol. Congr., Proc., Sect. 12, p. 149-158.

Allen, J.R.L.

1977: The possible mechanics of convolute lamination in graded sand beds; Geol. Soc. Lond. J., v. 134, p. 19-31.

Anketell, J.M., Cegla, J., and Dzulynski, St.

1970: On the deformational structures in systems with reversed density gradients; Ann. Soc. Geol., Pologne, v. 40, p. 3-30.

Banerjee, I., and McDonald, B.C.

1975: Nature of esker sedimentation; in Glaciofluvial and Glaciolacustrine Sedimentation (A.V. Jopling and B.C. McDonald, eds.), Soc. Econ. Paleontol. Mineral., Spec. Pub. No. 23, p. 132-154.

Belanger, J.R. and Harrison, J.E.

1976: Ottawa-Hull subsurface information data bank. Part I: Microfiche and Maps; Unpublished, Can., Geol. Surv., Open File 383.

Blatt, H., Middleton, G.V., and Murray, R.

1972: Origin of sedimentary rocks; Prentice-Hall Inc., Englewood Cliffs, New Jersey, 634 p.

Bloom, A.L.

1978: Geomorphology: A systematic analysis of Late Cenzoic Landforms; Prentice-Hall Inc., New Jersey, 510 p.

Boulton, G.S,

1968: Flow tills and related deposits on some Vestspitsbergen glaciers; J. Glaciol., v. 7, p. 391-412.

1972: Modern Arctic glaciers as depositional models for former ice sheets; Geol. Soc. Lond., J., v. 128, p. 361-393.

Butrym, J., Cegla, J., Dzulynski, St., and Nokonieczny, S.

1964: New interpretation of "Periglacial Structures"; Folia Quaternia, v. 17, 34 p. Krakow.

Coleman, J.M.

1976: Deltas: Processes of depositidn and models for exploration; Continuing Education Publication Company, Inc., Illinois. 102 p.

Curray, J.R.

1956: The analysis of two-dimensional orientation data; J. Geol., v. 64, p. 117-131.

Davies, H.G.

1965: Convolute lamination and other structures from the Lower Coal Measures of Yorkshire; Sedimentology, v. 5, p. 305-325.

De Geer, G.

1940: Geochronologia suecica principles'; K. svenska Vetensk-Akad. Handl., Ser. 3, 18, 6. Text and atlas.

Embleton, C., and King, C.A.M.

1968: Glacial and periglacial geomorphology; Edward Arnold (Publishers) Ltd., London. 608 p.

Emery, K.O.

1950: Contorted Pleistocene strata at Newport Beach, California; J. Sediment. Petrol., v. 20, p. 111-115.

Evenson, E.B., Dreimanis, A., and Newsome, J.W.

1977: Subaquatic flow tills: a new interpretation of the genesis of some laminated till deposits; Boreas, v. 6, p. 115-133.

Gadd, N.R.

1961: Surficial geology of the Ottawa area: Report of progress; Can., Geol. Surv., Pap. 61-19, 14 p.

1963: Surficial geology of the Ottawa map-area, Ontario and Quebec, 31G/5; Can., Geol. Surv., Pap. 62-16, 4 p.

1977: Offlap sedimentary sequence in Champlain Sea, Ontario and Quebec; Can., Geol. Surv., Pap. 77-1A, p. 379-380.

1978: Mass flow deposits in a Quaternary succession near Ottawa, Canada: diagnostic criteria for subaqueous outwash: discussion; Can. J. Earth Sci., v. 15,

p. 327-328.

Gary, M., McAfee Jr., R., and Wolf, C.C. (eds.)

1972: Glossary of geology; American Geological Institute, Washington, D.C., 805 p.

Hampton, M.A.

1975: Competance of fine-grained debris flows; J. Sediment. Petrol., v. 45, p. 834-844.

Hayward, M.

1979: Aspects of the Champlain Sea Event in the South Ottawa Region; Unpublished M.A. Thesis, University of Ottawa, 122 p.

Hicock, S.R. and Dreimanis, A.

1978: Late Wisconsinan marine subaquatic flow tills, Victoria, B.C.; Geol. Assoc. Can., Proc., v. 3, p. 421.

Johansson, C.E.

1963: Orientation of pebbles in running water. A laboratory study; Geogr. Ann., v. 45, p. 85-112.

Johnston, H.D.

1977: Sedimentation and fluid escape structures in some Late Precambrian shallow marine sandstones from Finmark, North Norway; Sedimentology, v. 24, p. 389-411.

Johnston, W.A.

1917: Pleistocene and Recent deposits in the vicinity of Ottawa, with a description of the soils; Can., Geol. Surv., Mem. 101, 69 p.

Jopling, A.V., and Walker, R.G.

1968: Morphology and origin of ripple-drift cross-lamination with examples from the Pleistocene of Massachusetts; J. Sediment. Petrol., v. 38, p. 971-984.

Kelling, G., and Stanely, D.J.

1976: Sedimentation in canyon, slope and base-of-slope environments; in Marine sediment transport and environmental management (D.J. Stanley and D.J.P. Swift), John Wiley and Sons, Inc., p. 379-435.

Kuenen, Ph., H.

1953: Features of graded bedding; Am. Assoc. Petrol. Geol., Bull., v. 37, p. 1044-1066.

1958: Experiments in geology; Geol. Soc. Glasgow, Trans., v. 23, p. 1-28.

Laport, P.

1970: Post depositional deformation structures in Pleistocene sediments, Rump's Pit, Stittsville; Unpublished B.Sc. Thesis, University of Ottawa.

LeBlanc Smith, G., and Eriksson, K.A.

1978: A fluvioglacial and glaciolacustrine deltaic depositional model for Permo-Carboniferous coals of the Northeastern Karoo Basin, South Africa; Geol. Assoc. Can., Proc., v. 3, p. 442.

Lewis, W.V.

1949: An esker in process of formation: Boverbreen, Jutunheimen, 1947; J. Glaciol., v. 1, p. 314-319.

Lowe, D.R., and LoPiccolo, R.D.

1974: The characteristics and origin of dish and pillar structures; J. Sediment Petrol., v. 44, p. 484-501.

Lowe, D.R.

1975: Water escape structures in coarse sediments; Sedimentology, v. 22, p. 157-204.

1976: Subaqueous liquefied and fluidized sediment flows and their deposits; Sedimentology, v. 23, p. 285-308.

Lundqvist, G.

1955: Raffelriktningar och rullstensasar; Geol. Foren. Forhandl., v. 77, p. 114-118.

Mackay, J.R. and MacKay, D.K.

1976: Cryostatic pressure in nonsorted circles (mud hummucks), Inuvik, Northwest Territories; Can. J. Earth Sci., v. 13, p. 889-897.

McArthur, D.S., and Onesti, L.J.

1970: Contorted structures in Pleistocene sediments near Lansing, Michigan; Geogr. Ann., Ser. A., v. 52, p. 186-193.

McDonald, B.C. and Vincent, J.A.

1972: Fluvial sedimentary structures formed experimentally in a pipe, and their implications for interpretation of subglacial sedimentary environments; Can., Geol. Surv., Pap. 72-27, 30 p.

McDonald, B.C. and Shilts, W.W.

1975: Interpretation of faults in glaciofluvial sediments; in Glaciofluvial and glaciolacustrine sedimentation (A.V. Jopling and B.C. McDonald, eds.), Soc. Econ.

Paleontol. Mineral., Spec. Pub. No. 23, p. 123-131.

Miall, A.D.

1976: Facies models -4. Deltas; Geosc. Can., v. 3, p. 215-227.

Middleton, G.V., and Hampton, M.A.

1976: Subaqueous sediment transport and deposition by sediment gravity flows; in Marine sediment transport and environmental management (D.J. Stanley and D.J.P. Swift, eds.), John Wiley and Sons, Inc., p. 197-218.

Morgan, J.P. (ed.)

1970: Deltaic sedimentation, modern and ancient; Soc. Econ. Paleontol. Mineral., Spec. Pub., No. 15, 312 p.

Nilsen, T.H., Bartow, J.A., Stump, E., and Link, M.H.

1977: New occurrences of dish structures in the stratigraphic record; J. Sediment. Petrol., v. 46, p. 1299-1304.

Potter, P.E., and Pettijohn, F.J.

1963: Paleocurrents and basin analysis; Springer-Verlag, New York, 296 p.

Prest, V.K., Grant, D.R., and Rampton, V.N.

1967: Glacial map of Canada; Can., Geol. Surv., Map 1253A.

Price, R.J.

1966: Eskers near the Casement Glacier, Alaska; Geogr.

Ann., Ser. A, v. 48, p. 111-125.

1973: Glacial and fluvioglacial landforms; Geomorphology Texts, 5, Oliver and Boyd, Edinburgh, 242 p.

Rautman, C.A., and Dott Jr., R.H.

1977: Dish structures formed by fluid escape in Jurassic shallow marine sandstone; J. Sediment. Petrol., v. 47, p. 101-106.

Reineck, H.-E., and Singh, I.B.

1975: Depositional sedimentary environments, with reference to terrigenous clastics; Springer-Verlag, New York, 439 p.

Richard, S.H.

1975: Surficial geology mapping: Ottawa Valley Lowlands (parts of 31 B, F, G); Can., Geol. Surv., pap. 75-1, Part B, p. 113-117.

1976: Surficial geology of Ottawa map area; Unpublished, Can., Geol. Surv., Open File, 366.

1978: Age of Champlain Sea and "Lampsilis Lake" episode in the Ottawa-St. Lawrence Lowlands; Can., Geol. Surv., Pap. 78-1C, p. 23-28.

Richard, S.H., Gadd, N.R., and Vincent, J.S.

1977: Surficial materials and terrain features of Ottawa-Hull area; Can., Geol. Surv., Map 1425A.

Rust, B.R.

1972: Pebble orientation in fluvial sediments; J. Sediment. Petrol., v. 42, p. 384-388.

1977: Mass flow deposits in a Quaternary succession near Ottawa, Canada: diagnostic criteria for subaqueous outwash; Can. J. Earth Sci., v. 14, p. 175-184.

1978: Mass flow deposits in a Quaternary succession near Ottawa, Canada: diagnostic criteria for subaqueous outwash; Can. J. Earth Sci., v. 14, v. page . 329-330.

Rust, B.R., ane Romanelli, R.

1975: Late Quaternary subaqueous outwash deposits near Ottawa Canada; in Glaciofluvial and glaciolacustrine sedimentation (A.V. Jopling and B.C. McDonald, eds.), Soc. Econ. Paleontol. Mineral., Spec. Pub. No. 23, p. 177-192.

Sharp, J.M., and Wood, D.S.

1974: Pore-fluid pressure release, fluidization and sediment expulsion as an explanation for many e early diagenetic features of layered sedimentary rocks; Geol. Soc. Am., Abs., v. 6, no. 7, p. 950.

Sharp, R.

1942: Periglacial involutions in north-eastern Illinois; J. of Geol., v. 50, p. 113-133.

Shaw, J.

1972: Sedimentation in the ice-contact environment, with examples from Shropshire (England); Sedimentology, v. 18, p. 23-62.

Smith, B.

1916: Ball- or pillow-form structures in sandstones; Geol. Mag., v. 3, p. 146-156.

Smith, J.

1970: A study of the primary sedimentary structures of a Pleistocene deposit, Burke's Pit, Carleton County, Ontario; Unpublished B.Sc. Thesis, University of Ottawa.

Sorby, H.C.

1859: On the contorted stratification of the drift of Yorkshire; Proceedings of the Geological and Polytechnic Society, West Riding of Yorkshire, 1849-1859, p. 220-222.

Stauffer, P.H.

1967: Grainflow deposits and their implications, Santa Ynez Mountains, California; J. Sediment. Petrol., v. 37, p. 487-508.

Sugden, D.E., and John, B.S.

1976: Glaciers and landscape; Edward Arnold (Publishers) Ltd., London, 376 p.

Theakstone, W.H.

1965: Contorted glacial lake sediments and ice blocks in outwash deposits at Østerdalsisen, Norway; Geogr. Ann., v. 47, p. 39-44.

Thibault, J.

1974: Etude d'un depot meuble stratifie de la region d'Ottawa; Unpublished B.Sc. Thesis, University of Ottawa.

Walker, R.G.

1978: Deep-water sandstone facies and ancient submarine fans: Models for exploration for stratigraphic traps; Am. Assoc. Petrol. Geol., Bull., v. 62, p. 932-966.

Washburn, A.L.

1956: Classification of patterned ground and a review of suggested origins; Beol. Soc. Am., Bull., v. 67, p. 823-865.

1973: Periglacial processes and environments; Edward Arnold (Publishers) Ltd., London, England, 320 p.

Wentworth, C.M.

1967: Dish structures, a primary sedimentary structure in coarse turbidites (abs.); Am. Assoc. Petrol. Geol., Bull., v. 51, p. 485.

Westgate, J.A., and Bayrock, L.A.

1964: Periglacial structures in the Saskatchewan gravels and sands of Central Alberta, Canada; J. Geol., v. 72, p. 641-648.

Wilson, A.E.

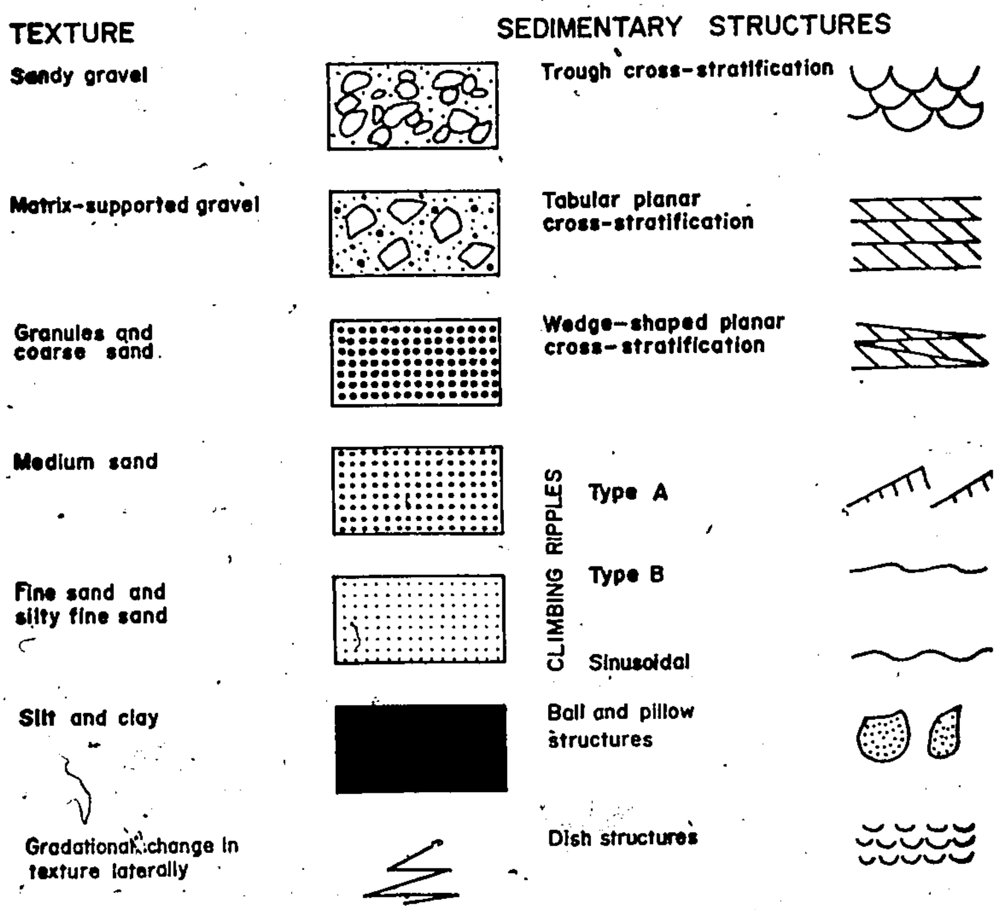
1946: Geology of the Ottawa-St. Lawrence Lowland, Ontario and Quebec; Can., Geol. Surv., Mem. 241, 66 p.

V. APPENDICES

APPENDIX I. STRATIGRAPHIC SECTIONS

Figure A-1. Key to symbols in Appendix I. Note that parallel rows of dots indicate that the sands exhibit parallel stratification. Randomly distributed dots indicate that the sands are massive, except where symbols for sedimentary structures are present, in which case the structures are present throughout the unit.

SYMBOLS



FOSSILS

<u>*Hiatella arctica*</u>	H
<u>*Macoma balthica*</u>	M
<u>*Portlandia arctica*</u>	P

Figure A-2. Maps of sand and gravel pits at sites S-4, S-5 and S-6.

MAPS OF SAND AND GRAVEL PITS AT SITES S-4, S-5 AND S-6

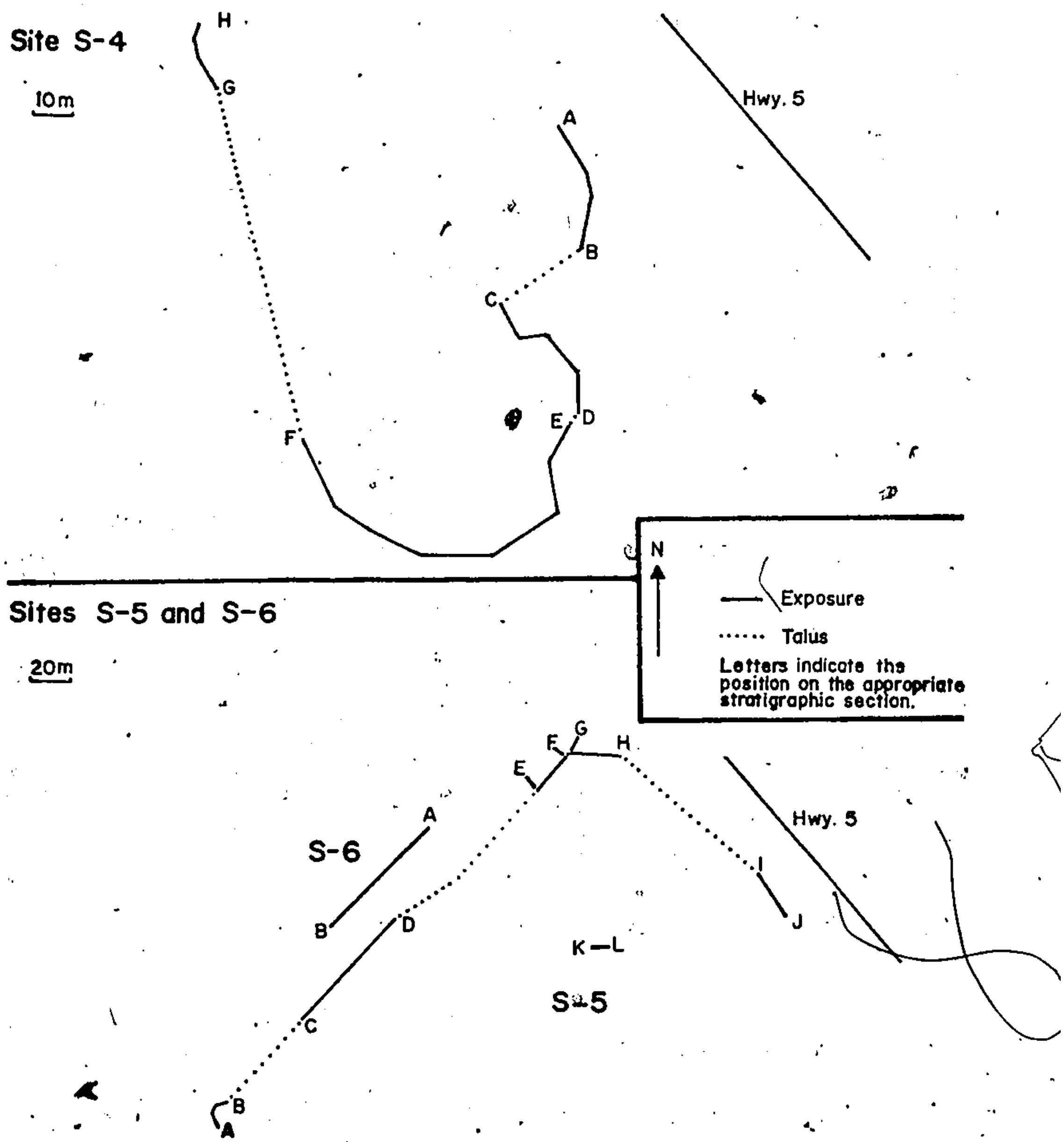
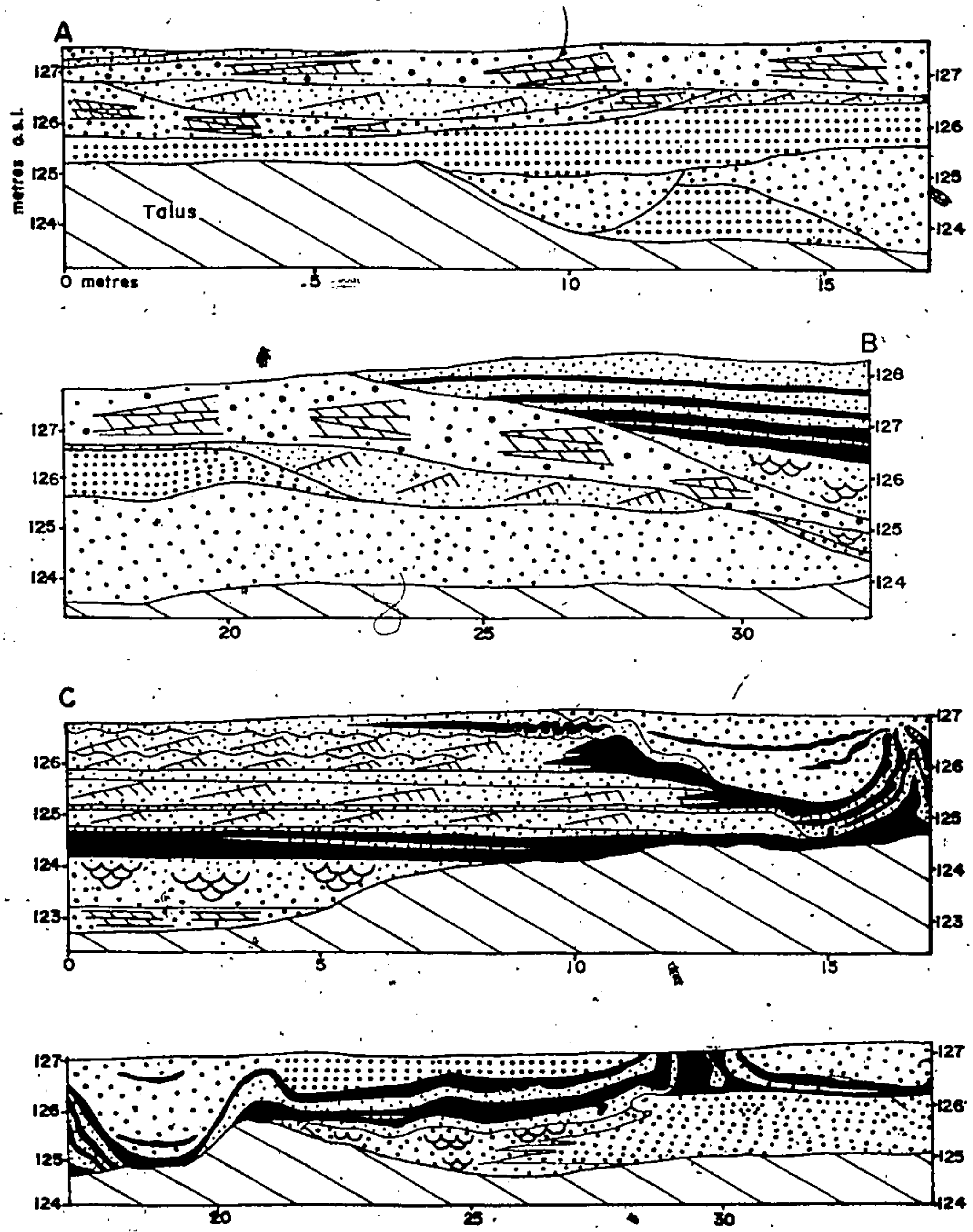
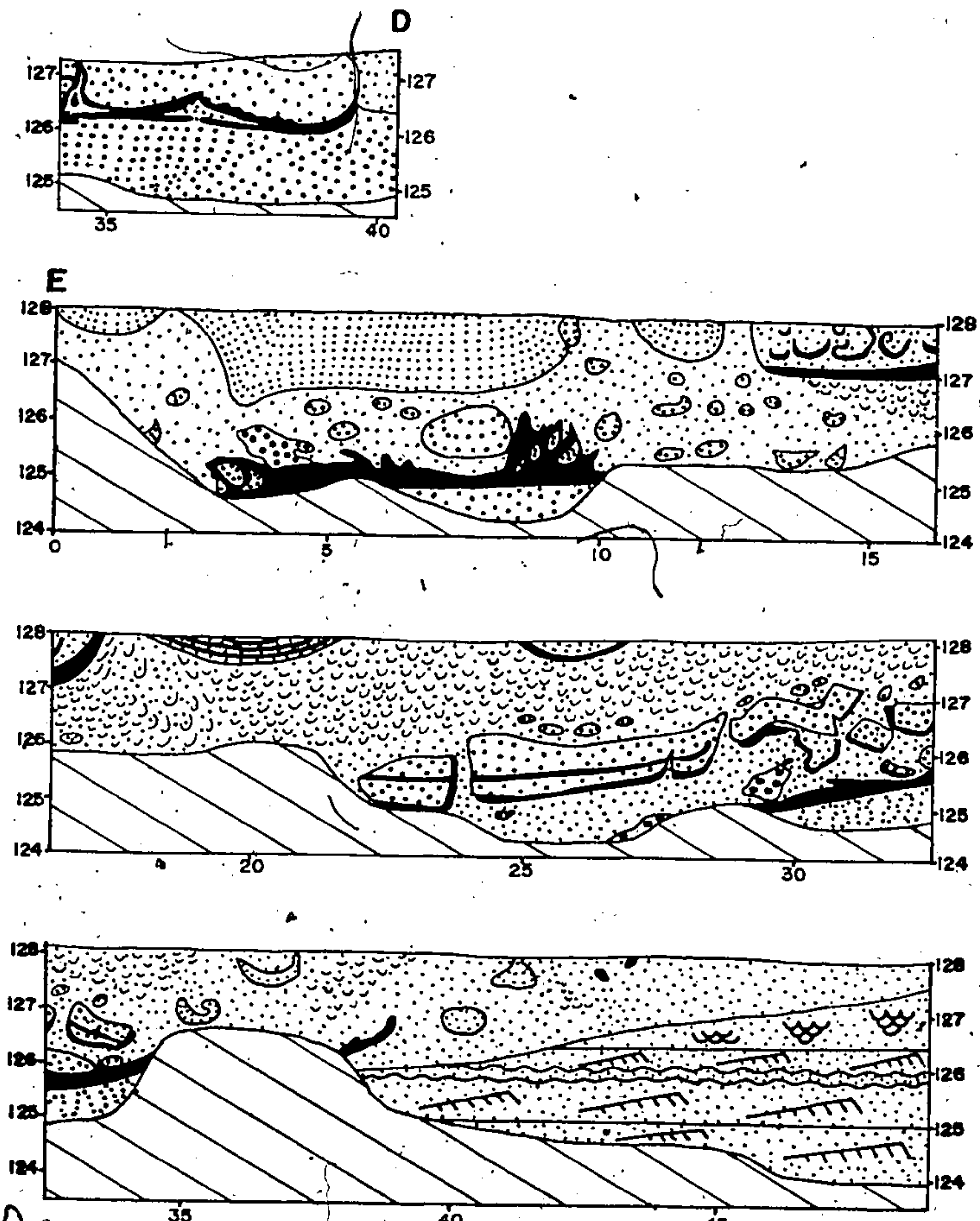
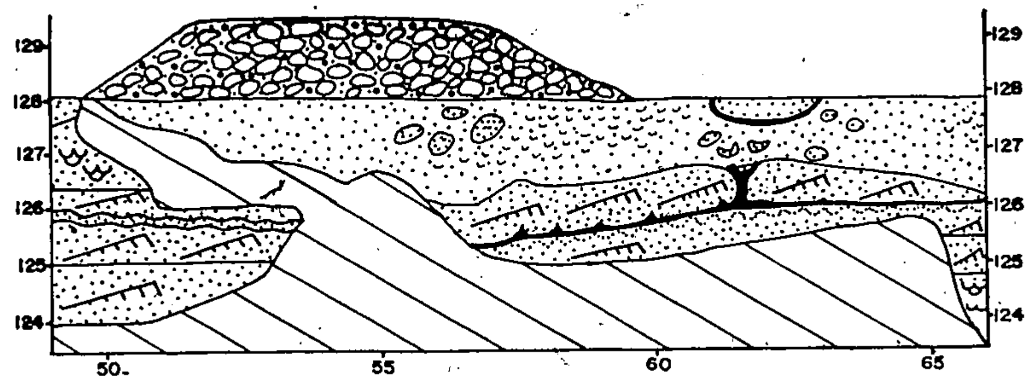


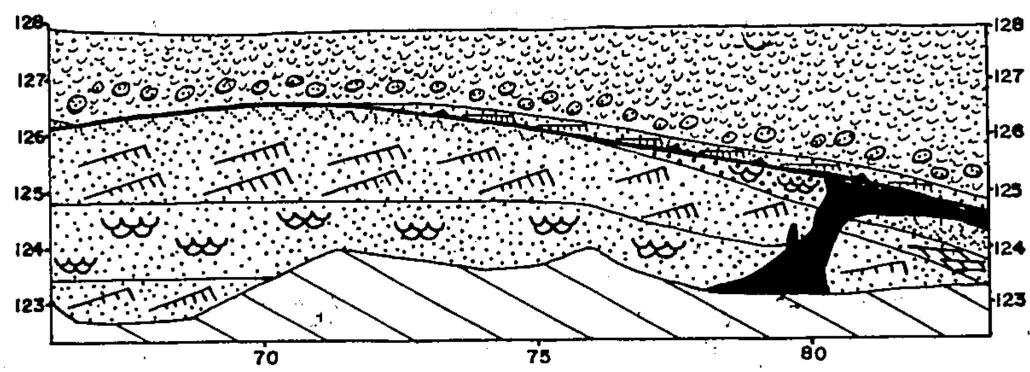
Figure A-3. Measured stratigraphic section for site S-4.

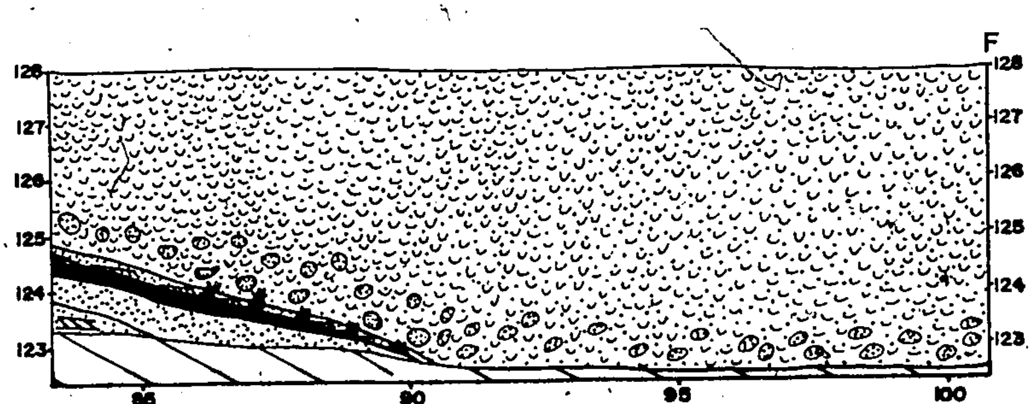
STRATIGRAPHIC SECTION —— SITE S-4




D
127
126
125
35
40
E
128
127
126
125
124
0
5
10
15
20
25
30
35
40
45


129
128
127
126
125
124
50
55
60
65


128
127
126
125
124
123
70
75
80


F
128
127
126
125
124
123
85
90
95
100

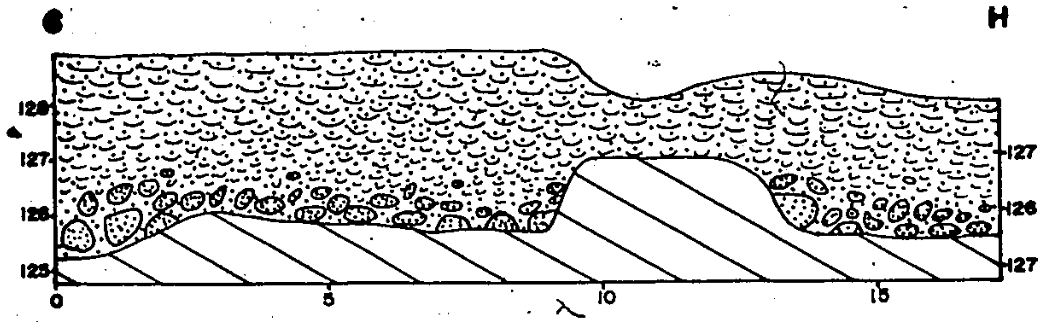
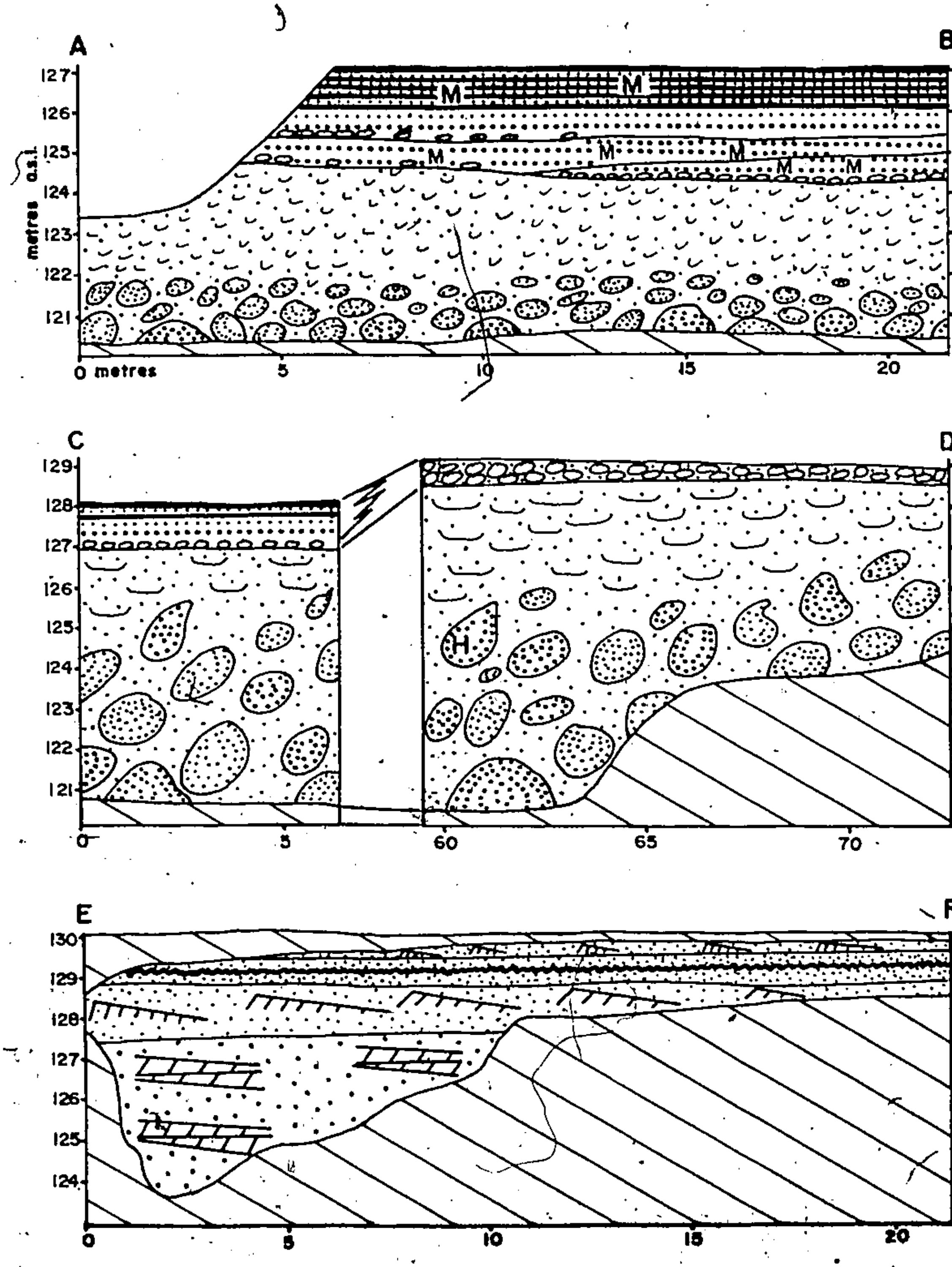
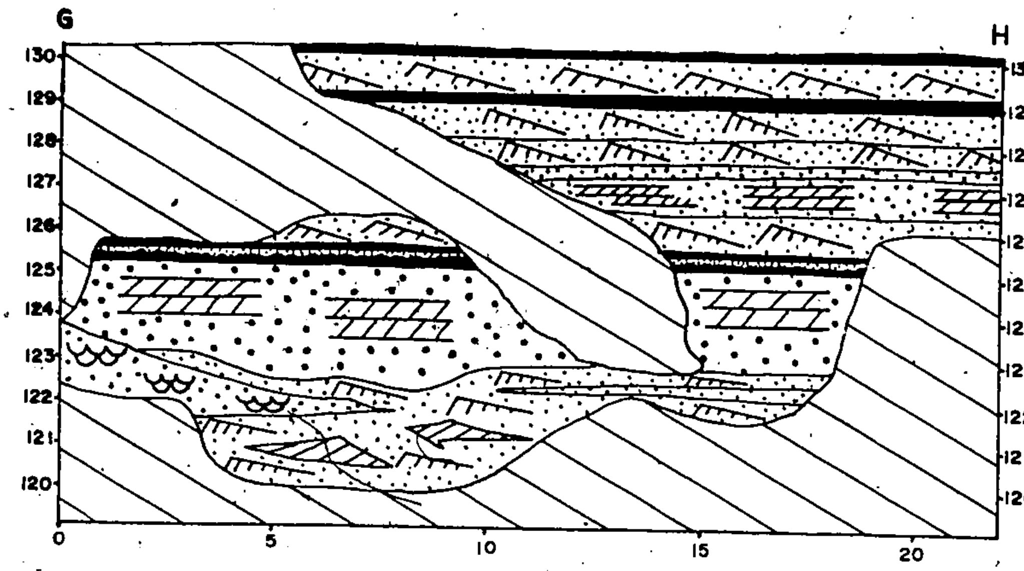
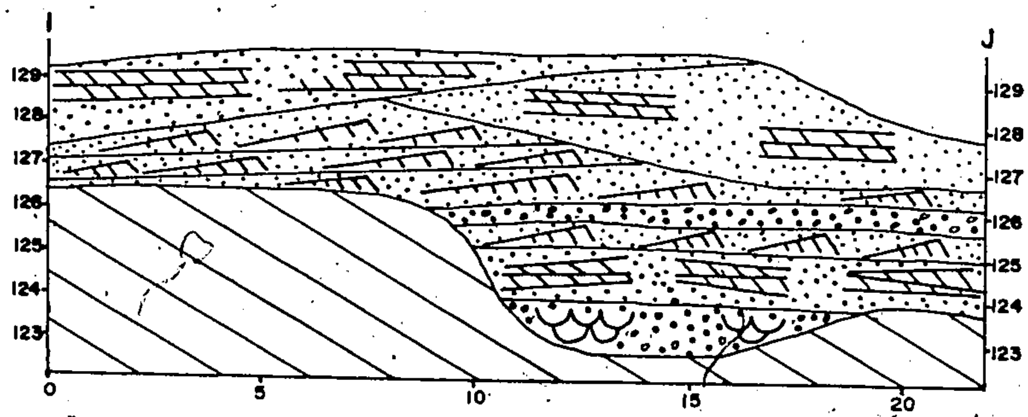

G
H
128
127
126
125
127
126
127
0
5
10
15

Figure A-4. Measured stratigraphic section for site S-5.

STRATIGRAPHIC SECTION —— SITE S-5




G
H
130
129
128
127
126
125
124
123
122
121
120
0
5
10
15
20


I
J
129
128
127
126
125
124
123
0
5
10
15
20

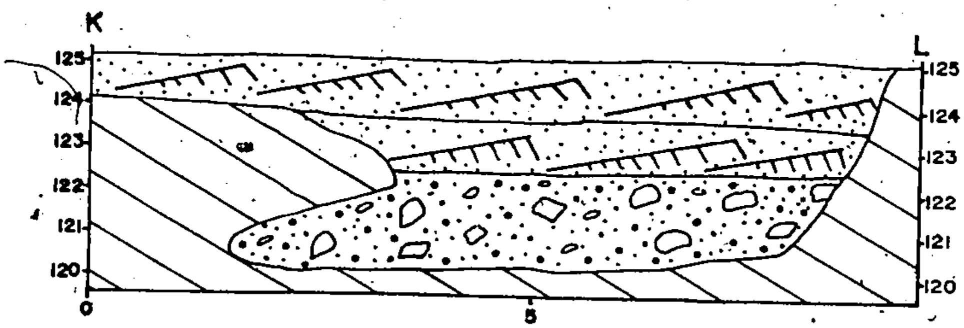
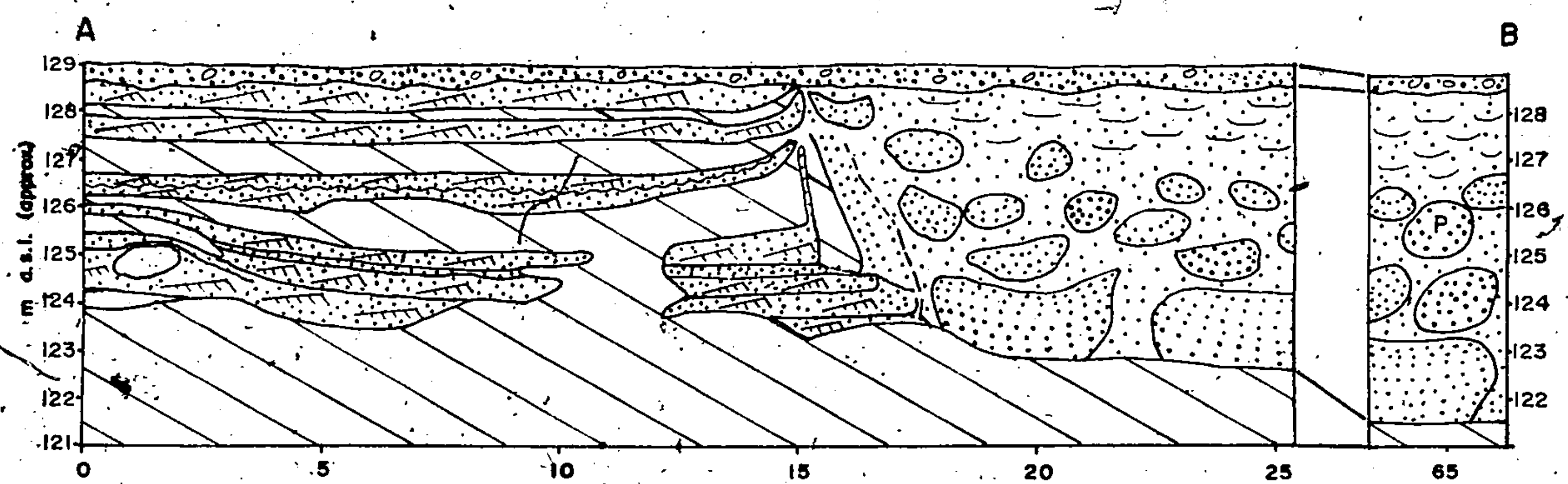

K
L
125
124
123
122
121
120
0
5

Figure A-5. Stratigraphic section for site S-6. Note that this section was partly drawn from field measurements but is largely based on information from photographs. As a result, this is not a measured section and may not be as reliable as the others presented in Appendix I.

STRATIGRAPHIC SECTION — SITE S-6



APPENDIX II. PALEOCURRENT DATA (all data in degrees)

Site S-3

Sample S3P1 - imbricate boulder gravel (direction of dip of ab plane).

295	023	003
014	020	000
357	076	036
016	086	256
020	027	016
021	054	357
022	021	011
332	041	359
318	032	
006	028	

Sample S3P2 - type A climbing ripples.

166	172	168	144	109	154	176	148	150
151	109	062	125	140	157	134	156	165
130	157	201	156	144	154	145	122	118
161	159	161	146	165	151	164	145	151
199	121	155	110	151	176	156	168	187
151	130	151	105	147	154	175	150	161
151	149	194	145	151	089	110	146	
139	154	145	112	104	141	145	125	
141	126	155	140	163	140	161	158	
130	155	217	143	154	146	149	154	

Sample S3P3 - imbricate gravel in channels (direction of dip of <u>ab</u> plane).

336	056	040
344	352	334
002	106	345
326	358	295
025	052	041
345	046	
044	018	
057	035	
020	331	
016	046	

<u>Site S-4</u>

Sample S4P1 - planar cross-beds.

155	156
158	161
161	162
158	155
150	157
153	
161	
158	
156	
160	

Sample S4P2 - trough cross-beds.

209	155	151	178	149
110	182	132	176	196
218	180	186	165	129
156	180	212	161	198
165	173	160	152	192
148	195	171	201	210
170	137	150	198	173
138	188	152	160	193
175	178	200	107	160
187	150	206	170	128

Sample S4P3 - planar cross-beds.

187	170	184
190	171	174
188	181	182
192	176	
190	179	
185	193	
196	192	
182	191	
184	186	
183	188	

Sample S4P4 - trough cross-beds.

222	195	234	196
218	205	232	209
202	201	296	229
205	202	211	191
191	179	216	196
211	184	278	238
218	177	224	221
206	194	252	241
203	190	194	255
206	221	178	202

Sample S4P5 - type A climbing ripple drift.

151	191	188	186	178	184
216	205	165	200	223	241
211	231	170	181	185	192
205	228	148	197	165	186
209	221	194	192	191	194
207	204	207	184	174	172
148	226	162	190	179	171
181	208	196	187	164	178
179	171	177	174	143	188
202	194	223	204	203	187

Sample S4D1 - long axis of dish structures.

294	327	276	272	325	351	304
305	279	308	266	270	338	260
274	252	225	266	274	287	266
280	301	200	328	218	180	226
263	322	313	327	280	301	277
322	283	252	275	263	288	292
298	214	255	304	271	292	317
299	304	298	275	298	290	285
266	313	303	293	276	289	294
304	306	323	197	267	287	317

Sample S4D2 - long axis of dish structures.

344	284	285	278	228	294	280
206	250	260	212	254	292	235
293	287	250	245	251	186	
292	264	270	265	260	247	
305	292	284	281	248	254	
283	326	290	241	270	261	
201	273	288	233	248	244	
279	255	260	294	258	237	
213	271	315	299	279	226	
316	266	256	258	238	260	

Site S-5

Sample S5P1 - planar cross-beds.

212	192	219	231
229	219	260	214
206	226	182	224
207	224	221	214
236	219	211	221
224	214	232	231
211	182	221	209
209	206	200	215
206	199	206	201
187	208	297	173

Sample S5P2 - type A climbing ripple drift.

155	206	201	176
164	194	192	180
166	181	176	185
224	186	174	211
189	197	190	183
202	192	186	177
167	215	198	185
172	168	189	202
191	173	182	201
188	203	194	179

Sample S5P3 - type A climbing ripple drift.

185	239	217	158
190	210	221	167
221	176	247	189
204	209	212	168
196	223	179	141
227	176	189	146
293	201	216	190
194	165	212	186
217	204	207	221
203	201	144	225

Sample S5P4 - type A climbing ripple drift.

114	137	154	166
116	128	135	134
123	126	168	180
139	139	149	152
191	116	184	146
196	136	128	168
107	104	112	146
101	156	191	149
156	166	109	111
181	136	159	156

Sample S5P5 - type A climbing ripple drift.

177	205	197	190
174	175	192	214
136	156	170	185
182	188	156	144
146	173	189	145
129	154	173	179
167	161	216	156
151	179	169	166
167	157	194	201
156	184	198	161

Sample S5P6 - type A climbing ripple drift.

206	223	198	214
209	209	185	207
204	172	191	239
224	222	237	204
207	199	229	219
198	190	214	200
241	192	194	210
212	201	186	206
204	203	149	197
196	217	177	243