UNIVERSITY OF MINNESOTA

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examining committee have been made.

Charles L. Matsch

Name of Faculty Advisor

[redacted]

Signature of Faculty Advisor

May 12, 2000

Date

GRADUATE SCHOOL

PLEISTOCENE GEOLOGY OF THE EMBARRASS AREA, ST. LOUIS COUNTY, MINNESOTA

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INTRODUCTION

Minnesota's landscapes and surficial deposits have resulted from the interaction of late Cenozoic continental glaciers upon a weathered bedrock surface. Erosion and deposition during subsequent glacial stages have largely obscured the effects of earlier glaciations. During the late Wisconsinan Stage, the Laurentide Ice Sheet advanced into Minnesota several times, becoming lobate as the ice followed preexisting regional bedrock lowlands. Each lobe of ice deposited a lithologically distinct drift, reflecting the nature of the bedrock over which it flowed. This thesis is one in a continuing sequence of detailed studies of geomorphology, sedimentology, stratigraphy, and drift lithology intended to clarify the complex glacial history of Minnesota. The Embarrass area was selected for study because of its importance to understanding the style and chronology of deglaciation in northern Minnesota.

STUDY AREA LOCATION

The study area is located in northern St. Louis County of northeastern Minnesota, approximately 75 miles north of Duluth (Fig. 1). Detailed mapping of Pleistocene deposits was carried out in the Biwabik NE and Embarrass quadrangles, and portions of the Isaac Lake, Babbitt, Babbitt NE, Biwabik, and Aurora quadrangles (U.S. Geological Survey 7.5 minute series) with the crest of the Giants Range forming the southern boundary of the map area (Fig. 1, Plate I).

A reconnaissance study of the Vermilion moraine was carried out along a line trending southeast to northwest from the west end of Birch Lake near Babbitt, along the

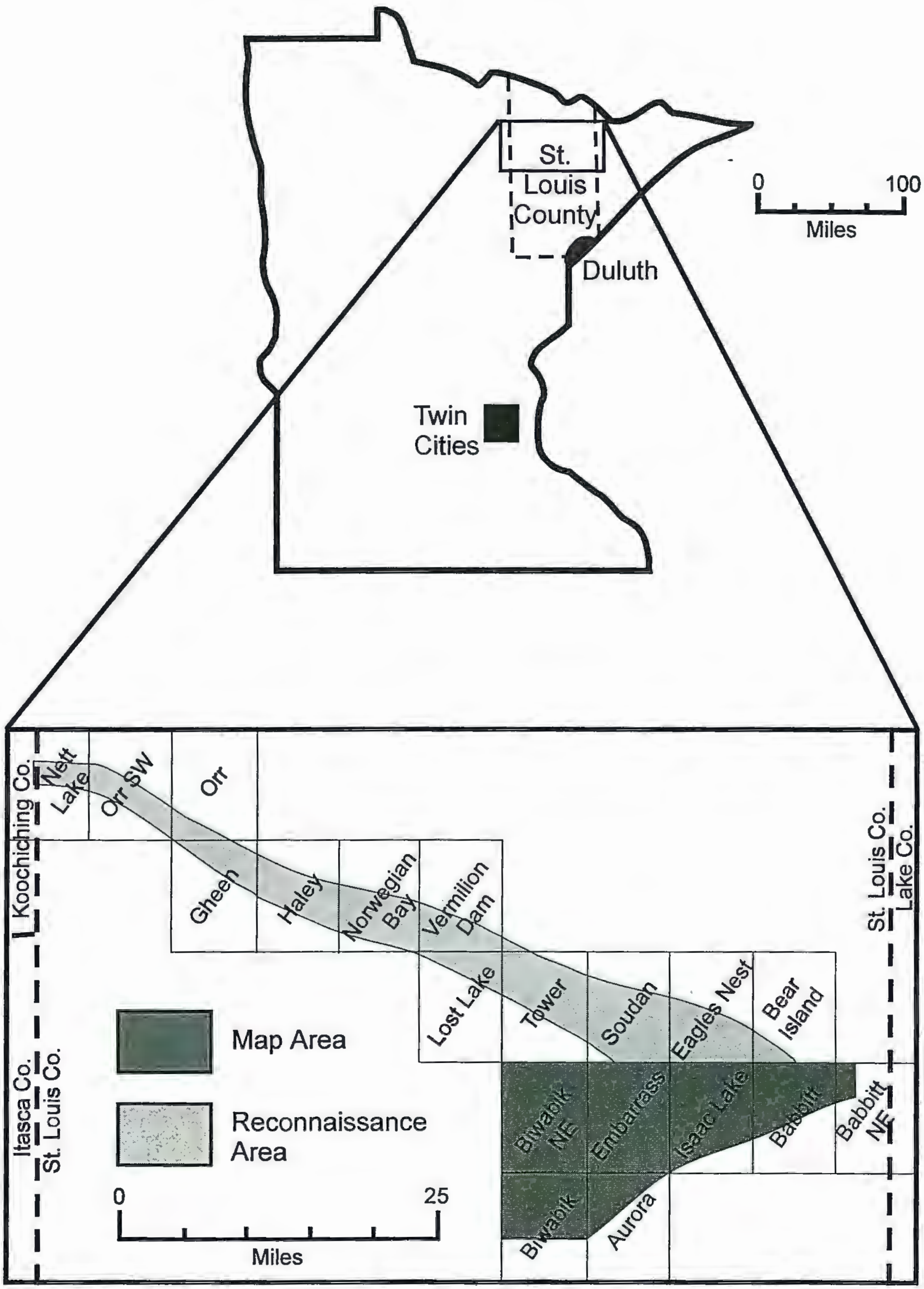


Figure 1. Study area location

south shore of Lake Vermilion to the south shore of Nett Lake approximately 14 miles northwest of Orr (Fig. 1).

A reconnaissance map of Rainy lobe landforms in the region was prepared from aerial photographs (Plate II). This map shows end moraines, lacustrine plains, drumlins, Rogen moraine, eskers and general direction of meltwater flow for an area in St. Louis County extending from south of Ely west to 8 miles east of Cook, south to Eveleth, and southeast to Brimson.

STUDY OBJECTIVES

The main objectives of this study were to: 1) map the Pleistocene deposits of the Embarrass area, 2) describe the lithologic characteristics of the sediments, 3) reconstruct the depositional environments of the various sedimentary units, 4) determine their stratigraphic relationships, 5) develop models to explain the origin of the various landforms present, and 6) correlate these results with the existing knowledge of the Quaternary history of northern Minnesota. A secondary objective was to examine the Vermilion moraine in St. Louis County and to describe its various lithofacies and genesis.

METHODS OF INVESTIGATION

Field Methods

Most of the field work for this study was done during October and November, 1984, and June through November, 1985. Additional field work was carried out during July, 1991 in preparation for the 1992 Midwest Friends of the Pleistocene Field Trip.

Exposures of Pleistocene sediments were encountered as road cuts along state highways, county roads, township roads and unimproved roads throughout the Embarrass area. Exposures were also examined in gravel pits, iron mines and railroad cuts. Within the map area, all roads that were accessible with a four-wheel-drive vehicle were driven in search of exposures. At all exposures within the map area, sediment descriptions included color, texture, sorting, and general lithology of the sediments, in addition to any sedimentary structures present. Field work was done with the aid of U.S. Geological Survey topographic maps (scale 1:24,000).

One hundred ninety-six samples were collected from a total of 220 locations for laboratory study. Orientations of the long axes of pebbles (length to width ratio of at least 2:1) were recorded at two exposures to determine preferred fabric orientation.

The locations of water wells, for which logs were supplied by the Minnesota Geological Survey, were confirmed in the field. Many of these water wells penetrate the Quaternary sediments and are finished in bedrock. This depth to bedrock data was used in conjunction with bedrock outcrop maps (Griffin, 1967; 1969; Griffin and Morey, 1969) to construct a topographic map of the bedrock surface of the Embarrass area (Plate III).

Laboratory Methods

Of the 196 samples collected, 73 were subjected to particle size analysis utilizing sieve and pipette methods described by Folk (1980) and the grain-size scale of Wentworth (1922). For practical purposes, the gravel fraction analyzed was restricted to granules and the entire pebble fraction (2 to 64 mm). If the sample contained clasts larger than 64 mm, they were discarded prior to sieve analysis. Using these results, the sediments were given descriptive textural names according to the trilinear classification schemes of Shepard (1954) and Lawson (1979) (Fig. 2).

The colors of air-dried sediment samples were classified in the laboratory under artificial light using the standard Munsell color chart (Raukas, 1982).

During the winter of 1989, the Pleistocene geology of the mapping area was re-examined by detailed stereoscopic study of high-altitude black and white aerial photographs taken in the early spring. Field observations and laboratory data supported geologic contacts that were adjusted at this time.

LOCATION NUMBERING SYSTEM

Locations referred to in this report use the U.S. Bureau of Land Management system of subdivision of public lands. The first two numbers of this system specify the township, the third number indicates the section within the township, and if letters follow the section number, they designate a portion of the section. The section is divided into quadrants and these are labeled A, B, C, and D in a counterclockwise fashion beginning in the upper right-hand quadrant (Fig. 3). For example, 60N-15W-24CCDD is the same as SE 1/4, SE 1/4, SW 1/4, SW 1/4, sec. 24, T. 60 N., R. 15 W.

PHYSIOGRAPHY

The late Cenozoic glaciation of Minnesota and the configuration of the preglacial bedrock surface have combined to produce a wide variety of landscapes, which Wright

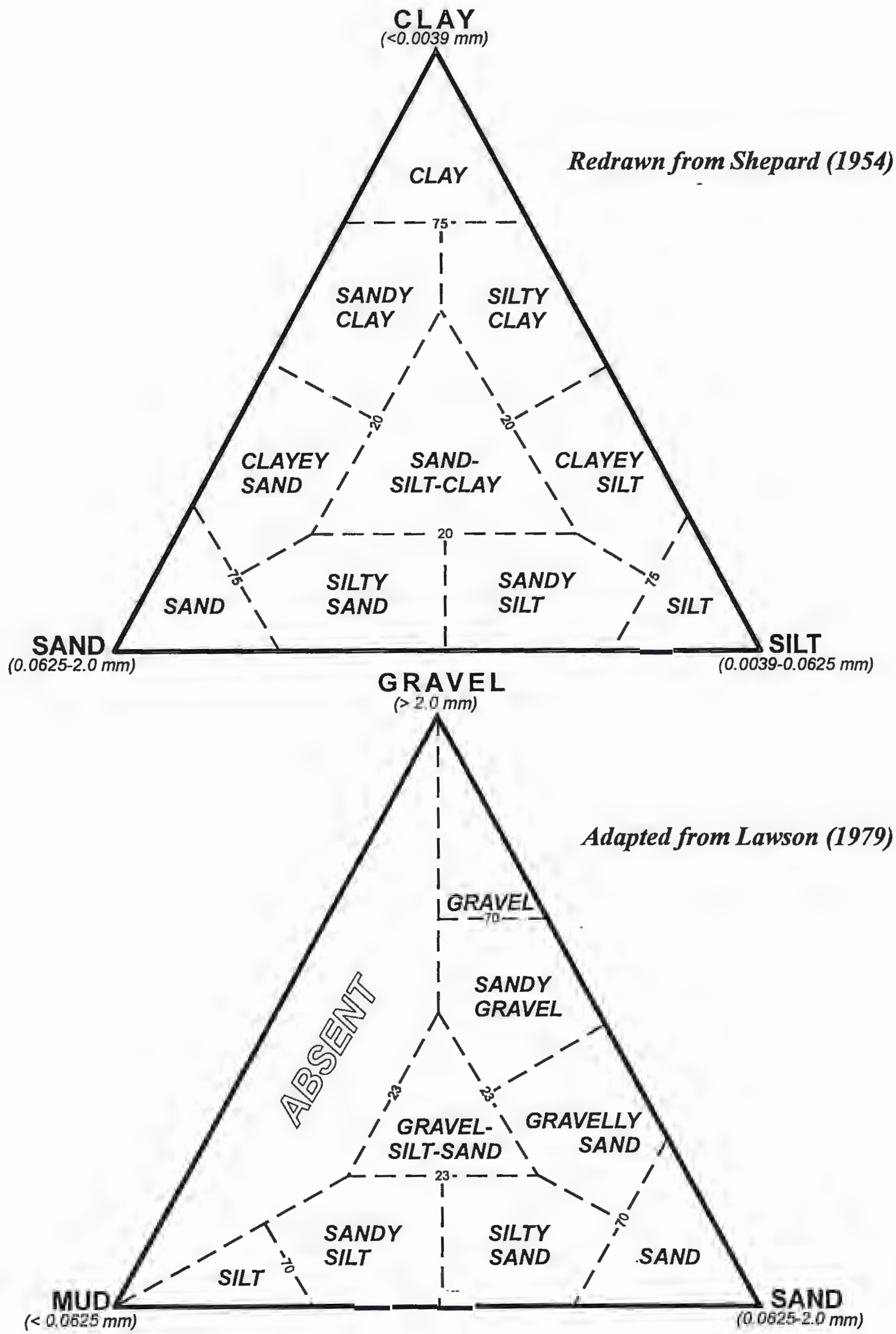


Figure 2. Sediment classification scheme

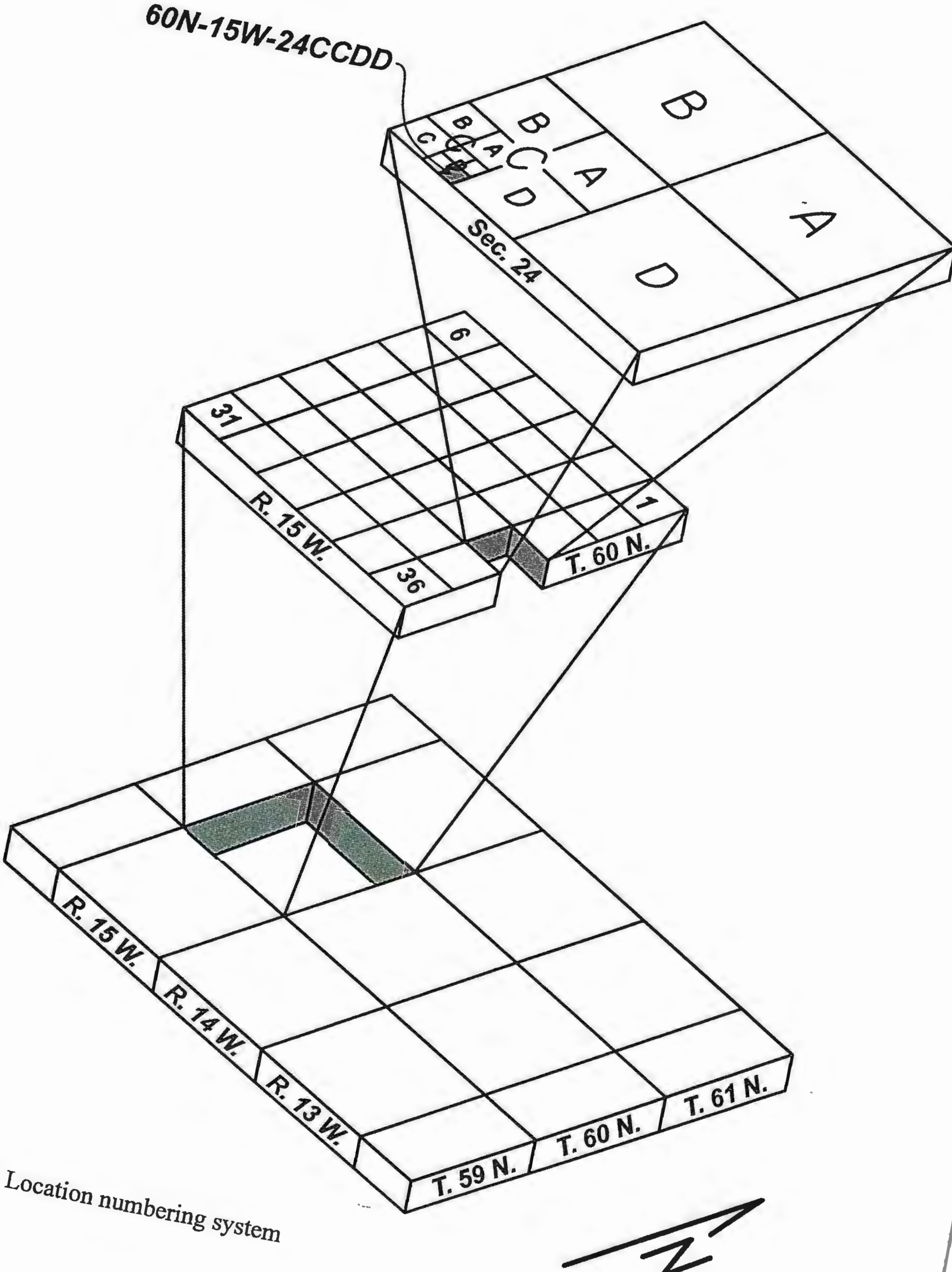


Figure 3. Location numbering system

(1972a) has classified into physiographic areas. According to his scheme, the study area occupies parts of the Chisholm-Embarrass area, the Border Lakes area and the Giants Range (Fig. 4).

The Chisholm-Embarrass area is characterized by thick accumulations of drift which fill the bedrock lows and tend to subdue the very irregular underlying bedrock surface (Cotter and others, 1965a). Most of this area is low and swampy, but it is transected by two east-west trending end moraines. Glacial diamictons, glaciofluvial and glaciolacustrine deposits, punctuated by sporadic bedrock outcrops, occur between the end moraines.

The Giants Range is a bedrock highland that rises 300 to 500 feet above the surrounding drift plains and is one of the most prominent topographic features in Minnesota. Drift cover on this bedrock-cored landscape is generally thin and patchy. The Giants Range physiographic area also includes the open-pit mines and tailings dumps of the Mesabi range immediately south of the mapping area (Wright, 1972a).

Differential glacial erosion of bedrock in the Border Lakes area has produced a pattern of lakes and ridges which reflect bedrock lithology and structure. Drift cover in this area is also thin and patchy, with bedrock outcrops common (Wright, 1972a).

The Laurentian Divide, a watershed dividing Hudson Bay drainage from Lake Superior drainage, transects the study area (Fig. 4) and significantly influenced late Wisconsinan deglaciation in the region.

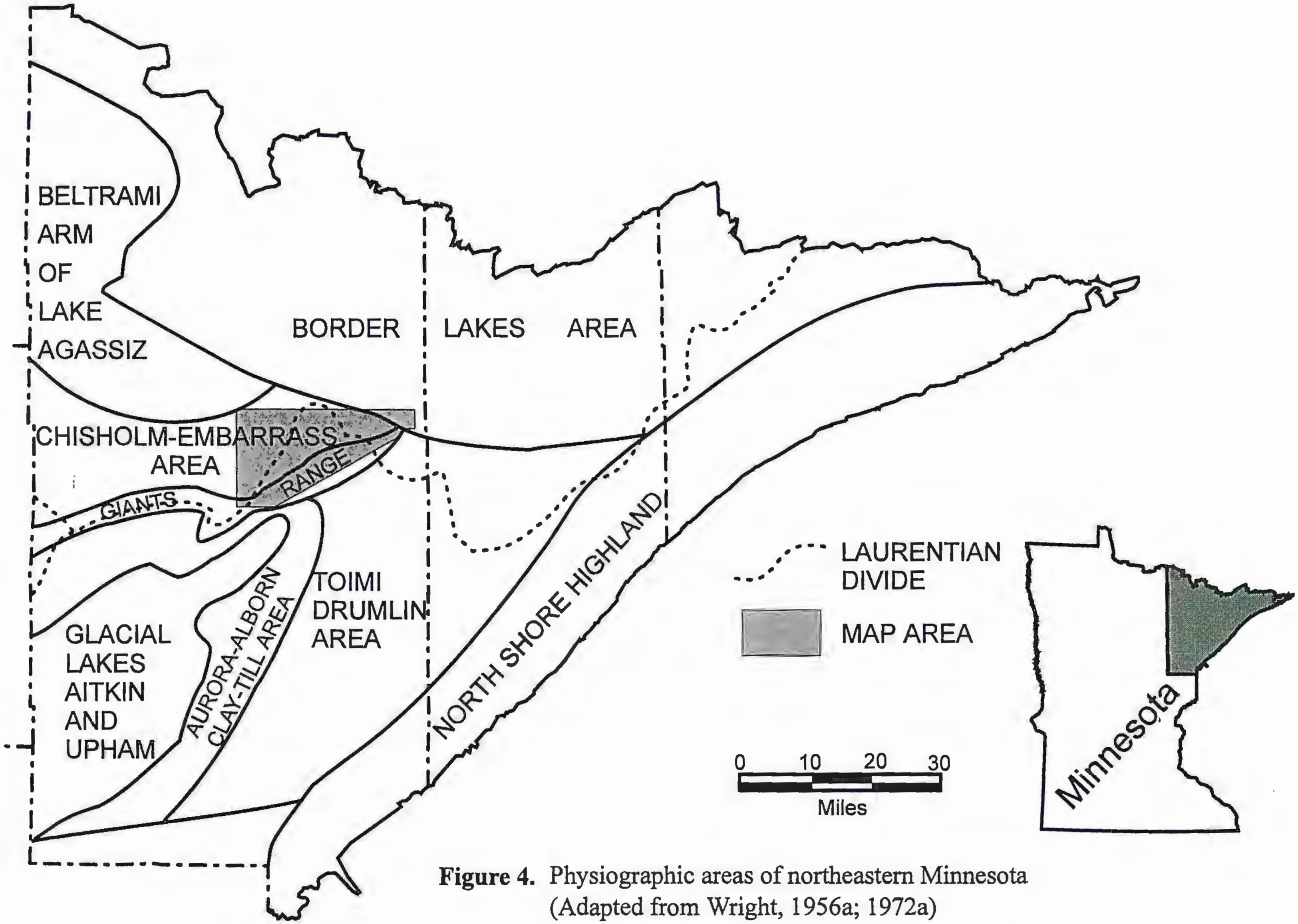


Figure 4. Physiographic areas of northeastern Minnesota
(Adapted from Wright, 1956a; 1972a)

BEDROCK GEOLOGY

The study area is located in the Vermilion district and adjacent areas. The Vermilion district is a Precambrian terrane, which appears to form a western extension of the Wawa volcanic-plutonic belt of the Superior province of the Canadian Shield (Morey, 1979). Bedrock studies have been numerous and, for the most part, the bedrock geology of the area has been mapped in detail (see Balaban, 1979 or Morey, 1980 for references).

The Vermilion district proper consists of steeply dipping subaqueous volcanic and derivative sedimentary rocks, with minor interbedded volcanic flows and pyroclastic rocks (Fig. 5). The subaqueous volcanic rocks contain interbedded banded iron formation, which has been mined in the vicinity of Tower, Soudan, and Ely (Morey, 1979).

These Archean supracrustal rocks were metamorphosed and complexly folded by the emplacement of granitic batholiths approximately 2700 million years ago (Goldich, 1972). The Vermilion fault generally separates the greenschist facies rocks of the Vermilion district from the amphibolite facies rocks of the Vermilion Granitic Complex to the north (Morey, 1979). The granitic rocks of the Giants Range batholith have normal intrusive relations with the supracrustal rocks of the Vermilion district in the study area, but are in fault contact elsewhere (Sims and others, 1970; Green, 1982) (Fig. 5).

Immediately south of the study area, the gently dipping lower Proterozoic sedimentary rocks of the Animikie Group overlie nonconformably the Giants Range batholith, and with angular unconformity undivided Archean metasedimentary and metavolcanic rocks (Sims and others, 1970; Green 1982) (Fig. 5).

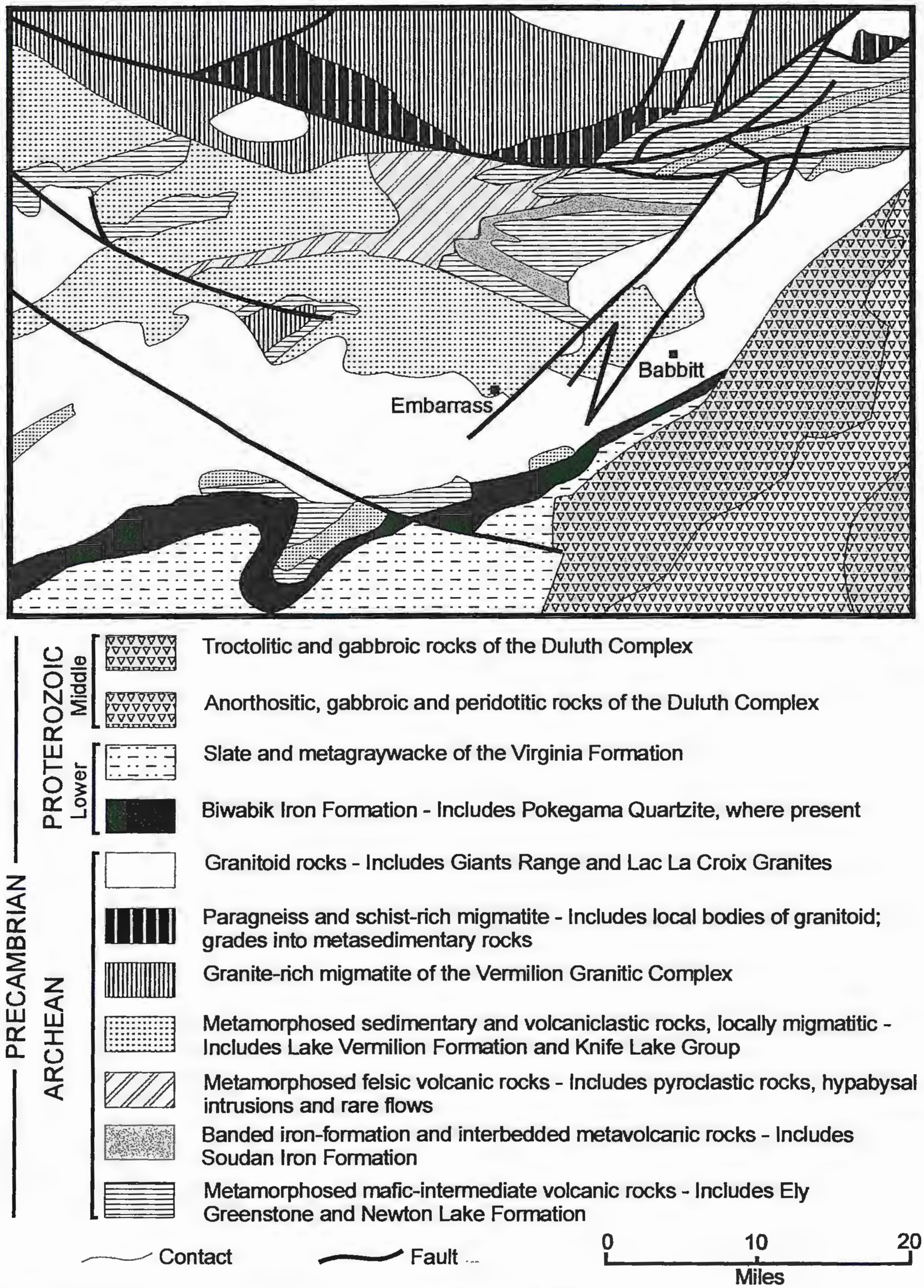


Figure 5. Bedrock geology of the Vermilion district and adjacent areas
(redrawn from Morey and others, 1982)

South and east of the study area, the Archean and lower Proterozoic rocks have been intruded by the Keweenawan (1100 ma) Duluth Complex (Green, 1982) (Fig. 5). The Duluth Complex is a layered igneous intrusion consisting primarily of troctolitic, anorthositic, and gabbroic rocks, with lesser amounts of intermediate to granitic rocks (Green, 1982).

PREVIOUS QUATERNARY INVESTIGATIONS

Upham (1894) published the first report on the glacial geology of northeastern Minnesota. He documented 12 concentric recessional moraines across the state and ascribed their formation to a single lobe of ice advancing south-southwest. In northeastern Minnesota, the southernmost moraine, named the Leaf Hills or 9th moraine, approximately parallels the highlands of Lake Superior's north shore. The Itasca or 10th moraine and the Mesabi or 11th moraine are shown as coincident along the crest of the Giants Range from Hibbing east to the Embarrass Gap north of Aurora. From this point eastward, the Itasca (10th) moraine veers southeast, then parallels the Giants Range, trending generally northeast, while the Mesabi (11th) moraine continues to follow the crest of the Giants Range to the south shore of Birch Lake. The northernmost moraine of Upham (1894) is the Vermilion or 12th moraine. It trends southwest from Ely, thence northwestward from Tower along the south shores of Lake Vermilion, Pelican Lake and Nett Lake.

Upham (1894) was the first worker to differentiate three general till types in northern Minnesota based on clast lithology. Tills in northwestern Minnesota were characterized by abundant Paleozoic limestone pebbles, till of north-central Minnesota contained abundant Precambrian crystalline lithologies, and till of northeastern Minnesota contained lithologies common to bedrock in the Lake Superior basin.

Todd (1898) disagreed with Upham's interpretation that the moraines of northeastern Minnesota resulted from the advance and retreat of a single lobe of ice. He argued that northern Minnesota was glaciated by two contemporaneous lobes of ice, the Lake Superior lobe in the east and the Red River lobe to the west. Todd (1898) postulated the Mesabi moraine of Upham (1894) to be an interlobate moraine, with further ice recession resulting in the formation of the Vermilion moraine on the southern margin of the Red River lobe and the formation of Upham's Leaf Hills moraine on the northern margin of the Lake Superior lobe.

Elftman (1898) agreed with Todd's (1898) revision of the moraines of northern Minnesota, but changed the name of the Leaf Hills moraine to the Highland moraine because of its doubtful extension to the Leaf Hills in west-central Minnesota. Elftman (1898) also interpreted the drift of northeastern Minnesota to be the result of two contemporaneous ice lobes with different provenances - the Superior lobe, which advanced southwesterly from the Lake Superior basin and the adjacent Rainy lobe, which advanced across the upland generally southward across the Canadian border.

Winchell's (1899) geologic map of northern St. Louis County shows the Vermilion moraine in a position slightly different than previous workers. He mapped it as extending northwest from the south shore to Birch Lake, along the south shore of Lake Vermilion, passing north of Pelican Lake northwestward to the southwest shore of

Kabetogama Lake. Winchell (1899) also noted another moraine, the southern or Mesabi moraine, following the Giants Range to an intersection with the Vermilion moraine near the west end of Birch Lake. He attributed the Mesabi moraine to the farthest north advance of the Lake Superior lobe. Winchell (1901) was the first worker to note the presence of lacustrine sediments in the lowlands north of the Giants Range and attributed their deposition to glacial Lake Norwood.

Leverett's (1932) regional study presented evidence for glaciation of northeastern Minnesota by three distinct lobes of ice. The oldest drift, termed Patrician, is described as being deposited by a composite Rainy-Superior lobe terminating at the St. Croix moraine in central Minnesota. Recessional moraines of this advance were reported in the Embarrass area. A later, contemporaneous advance of the Rainy and Superior lobes formed the Vermilion and Highland moraines, which intersect at a complex interlobate junction near Isabella (Leverett, 1932). Leverett was the first worker to document the deposits of the St. Louis sublobe, an offshoot of the Des Moines lobe which approached the study area from the west.

In recent years, Wright (1955, 1956a, 1956b, 1962, 1964, 1969, 1971, 1972b, 1973) and co-workers (Florin and Wright, 1969; Wright and others, 1973; Wright and others, 1970; Wright and Ruhe, 1965; Wright and Watts, 1969) have presented a refined chronology of the Wisconsinan glaciation of Minnesota. Because the surficial glacial deposits of Minnesota are the result of a series of advances and retreats of several lobes of ice originating in separate accumulation centers (Wright, 1972b), a new system of correlating events had to be established.

Wright (Wright and Ruhe, 1965) grouped contemporaneous Wisconsinan events into informal "phases" based on morphologic and stratigraphic features and gave the phases local geographic names, correlating between adjacent lobes where possible.

The St. Croix phase is defined by an advance of the confluent Rainy and Superior lobes to the St. Croix moraine in central Minnesota at least 20,500 years ago (Birks, 1976). The formation of the Toimi drumlins by the Rainy lobe is ascribed to this phase (Wright and Watts, 1969).

The next phase defined by Wright (Wright and Ruhe, 1965, Wright and Watts, 1969) was the St. Croix phase ice, when the Rainy lobe readvanced to the Vermilion moraine, truncating the Toimi drumlins on the north. At the same time, the Superior lobe advanced to the Highland-Mille Lacs moraine, which truncates the Toimi drumlins on the southeast (Wright and Watts, 1969). These advances define the Automba-Vermilion phase (Wright and Ruhe, 1965).

The Superior lobe retreated and advanced two additional times during the Split Rock and Nickerson phases. Contemporaneous with the Nickerson phase of the Superior lobe was an advance of the St. Louis sublobe. This activity defines the Nickerson-Alborn phase (Wright and Watts, 1969).

Upon the retreat of the ice at the end of the Nickerson-Alborn phase, large proglacial lakes formed in the lowlands. This interval is termed the Agassiz phase (Wright and Ruhe, 1965).

Winter (1971, 1973) and co-workers (Cotter and others, 1964; Winter and others, 1973) have presented evidence for three separate glacial advances in the Mesabi-

Vermilion iron range area, resulting from their detailed studies of the glacial drift and ground water resources of this area.

The first advance deposited what they termed the "basal till" from a lobe with a northwestern source. The matrix of the "basal till" is calcareous and the predominate clay mineral is illite (Winter and others, 1973). The pebble fraction of this unit is mostly locally derived granitic and metamorphic rocks, but it also contains a few carbonate pebbles and a few rock types derived from the Lake Superior basin. Winter (1971) explained the presence of carbonate clasts, a calcareous matrix and the northeast-southwest fabric as the result of Keewatin ice advancing northeastward south of the Giants Range, analogous to the late Wisconsinan St. Louis sublobe. This unit is present in open-pit mines of the Mesabi range and is considered to be pre-St. Croix phase (Winter, 1971).

The next advance was that of the Rainy lobe, which deposited the "bouldery till," named for its most conspicuous feature. This advance is ascribed to the St. Croix phase (Winter, 1971).

The third advance was that of the St. Louis sublobe, which itself was split by the Giants Range into two smaller ice tongues. One ice tongue advanced north of the Giants Range depositing a brown silty till, while the southern ice tongue deposited a red clayey till as a result of the incorporation of Superior lobe provenance lake sediments (Winter and others, 1973).

Most recently, Stark (1977), Friedman (1981), and Fenelon (1986) have mapped, in detail, the surficial geology of the area immediately east of the study area, extending beyond the interlobate junction of the Rainy and Superior lobes nearly to Lake Superior. Stark (1977) described two Rainy lobe end moraines south of the Vermilion moraine, which are contorted around two moraines deposited by a sublobe of the Superior lobe. He also described glacial Lake Dunka as an ice marginal lake developed as the Rainy lobe retreated north of the Laurentian divide. Friedman (1981) mapped the same moraines as Stark did to the west, but documented a readvance of the Rainy lobe to the Vermilion moraine, which buries a portion of the northernmost of the two Superior sublobe moraines. Fenelon (1986) reinterpreted the Superior sublobe end moraines of Stark (1977) and Friedman (1981) to be Rainy lobe end moraines with the Superior lobe characteristics of these sediments inherited from incorporation of proglacial Superior lobe outwash and lacustrine sediment by the Rainy lobe.

Hobbs (Hobbs and others, 1988) recently synthesized and refined the mapping of Stark (1977), Friedman (1981), and Fenelon (1986), retaining the interpretation of moraines presented by Fenelon (1986). However, he later defined the Isabella sublobe as that portion of the Laurentide Ice Sheet located between the Rainy and Superior lobes in the Isabella area which deposited the moraines previously ascribed to a sublobe of the Superior lobe (Lehr and Hobbs, 1992, p. 6).

PLEISTOCENE SEDIMENTS

INTRODUCTION

Sediment deposited from continental ice sheets in low-lying shield areas is initially derived from the substrate by erosion. It is the thermal regime of the basal portion of the ice that dictates the mode of erosion (Weertman, 1964; Boulton, 1972). Where the ice is sufficiently thick, so as to prevent dissipation of frictional and geothermal heat, a thin layer of water develops at the base of the ice (Weertman, 1961; 1964; Clayton and Moran, 1974). Under these wet-based or thawed-bed conditions, the ice advances primarily by basal sliding (Weertman, 1964) and erosion of the substrate occurs through abrasion (Clayton and Moran, 1974).

Quarrying may also occur in the basal zone if the temperature of the ice is near the pressure melting point and if the substrate is irregular. As ice encounters an obstruction to flow, a rise in pressure results in a depression of the melting point and melting occurs. This water migrates to the lee side of the obstacle, where it refreezes as a result of a decrease in pressure, resulting in the adfreezing of joint blocks and their subsequent removal. Heat of recrystallization migrates through the rock to melt more ice on the stoss side (Boulton, 1970). This melting and refreezing process is referred to as regelation and operates on small obstacles.

Quarrying, through repeated regelation, provides the tools to further erode the substrate by abrasion (Clayton and Moran, 1974). The thawed-bed zone is usually located some distance up-ice from the margin, because in the marginal areas, where the ice is thinner, the base of the ice is often frozen to the substrate owing to a cold periglacial climate (Weertman, 1961; Clayton and Moran, 1974). Under these cold-based or frozen-bed conditions, the ice advances by internal deformation and the dominant mode of erosion is quarrying (Clayton and Moran, 1974).

Repeated incorporation of eroded substrate through regelation, produces a basal stratified zone of debris-rich ice (Weertman, 1961; Boulton, 1972) containing up to 80 percent glacial debris by volume (Boulton, 1968). Glacial debris is defined as *"material being transported by a glacier in contact with glacier ice. In most cases it is disaggregated, except for clasts of various sizes, including large rafts"* (Dreimanis, 1982, p. 11).

Transportation and subsequent deposition in a purely glacial sedimentary environment result in an unsorted sediment commonly referred to as till. Deposition of till requires melting of the debris-rich ice (Sugden and John, 1976), which produces water. Also, as the debris-rich ice is melting, gravity immediately begins to act upon the sediment and it may be remobilized several times before final deposition (Lawson, 1979). The degree to which water and gravity may participate in the deposition of till has been the subject of intense debate among scholars in the field of glacial geology.

The INQUA Commission on the Genesis and Lithology of Quaternary Deposits Work Group for the Genetic Classification of Tills, which consisted of 115 members from 32 countries (Dreimanis, 1982), agreed upon the following definition of till: *"a sediment that has been transported and is subsequently deposited by or from glacier ice, with little or no sorting by water"* (Dreimanis, 1982, p. 21).

The expression *"by or from glacier ice"* implies deposition directly from glacier ice by processes such as lodgement or basal meltout, but also encompasses deposition of debris from glacier ice by processes such as mass movement, as long as little or no sorting occurs (Dreimanis, 1982). Tills deposited directly from glacier ice are termed primary or ortho-tills, while tills which have undergone redeposition in the glacial

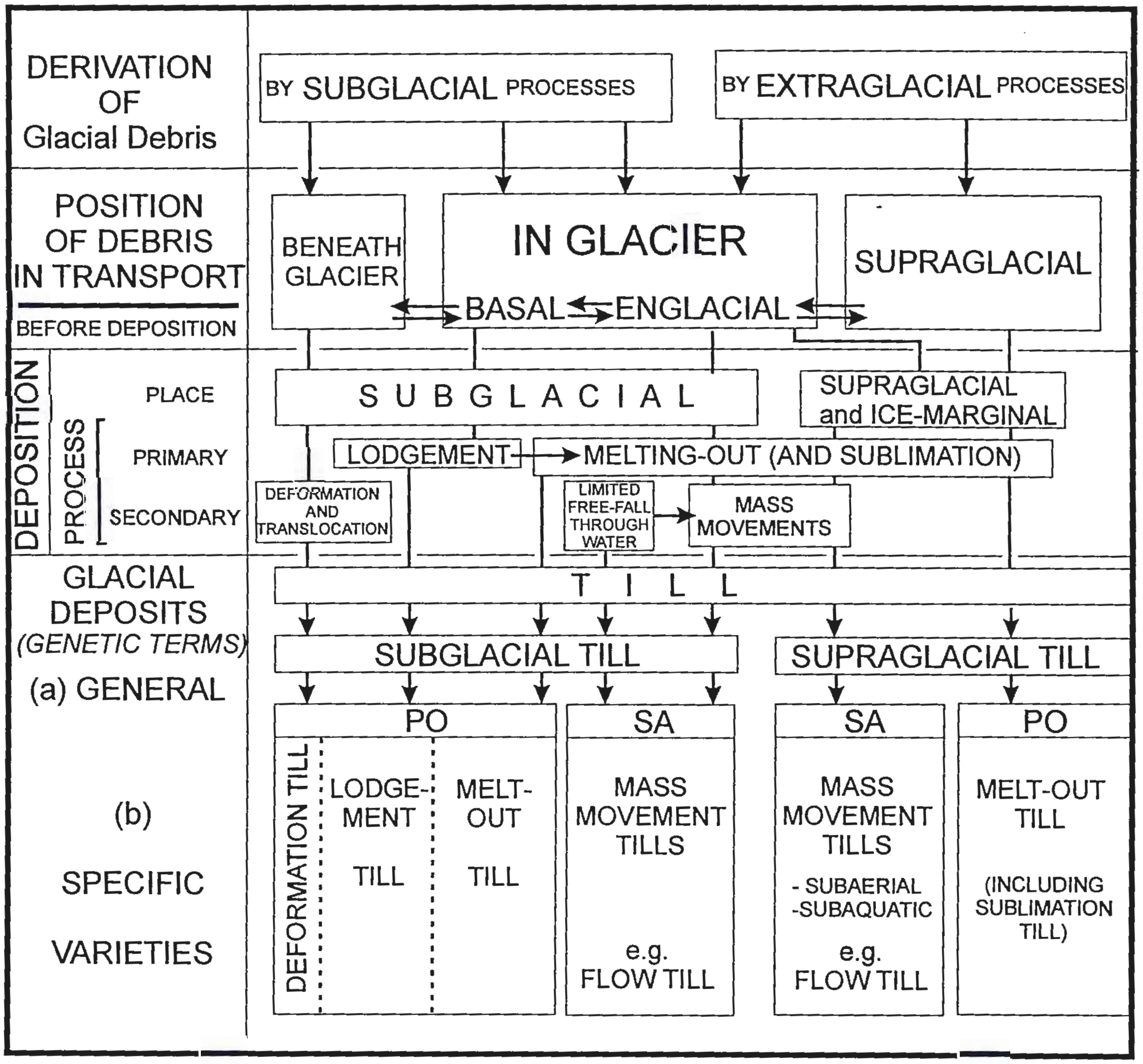


Figure 6. Factors that influnce the formation and deposition of tills (upper half), and a tentative depositonal genetic classification of tills (lower half). Abbreviations: PO - Primary or Ortho-tills, SA - Secondary or Allo-tills. (Dreimanis, 1982)

environment are secondary or allo-tills (Fig. 6) (Dreimanis, 1982).

To assign the word till to a sediment implies a mode of genesis for the sediment and *"since a scheme based upon genesis is in many cases the goal of classification"* (Aario, 1977b, p. 99); "till," as defined above, as well as other genetic terms are used throughout this report. These genetic sediment terms are used as map units on the Pleistocene geologic map of the Embarrass area (Plate I) and also provide a framework for the following section of this report. Geologic maps which indicate the environment of deposition or the origin of sediment are called lithogenetic maps (Clayton and others, 1980).

The map unit boundaries shown on Plate I are based, not only upon the sediment observed at the surface in a particular area, but also on the process which created the predominant landforms in the map unit. According to Aario (1977b) *"if a division is based on the mechanism of formation, it is the dominant process responsible for forming the landform which gives the name"* (p. 99) to the map unit.

This genetic approach to mapping not only furnishes information about the surficial sediments and landforms, but also provides a basis for making inferences as to the nature of related sediments in the subsurface (Eyles, 1983).

SUBGLACIAL TILL

A diamicton, interpreted as a primary or ortho-till of subglacial origin, occurs in limited, but widespread exposures throughout the study area (Plate I). This till has been informally referred to as the "bouldery till" by previous workers in this area (Winter, 1971; Winter and others, 1973; Stark, 1977), but has never before been distinguished based on genesis.

Sediment Description

The matrix of the subglacial till is sand to silty sand with textural analyses of seven samples (Appendix A) yielding an average composition of 76 percent sand, 22 percent silt and 2 percent clay (Fig. 7). This average plots in the sand field of Shepard's (1954) classification (Fig. 2). It is interesting to note that this average matrix texture is nearly identical to the average texture of till (75 percent sand, 20 percent silt and 5 percent clay) reported for large areas of the southwestern Canadian Shield (Dredge and Cowan, 1989). At one locality (60N-15W-03DDBD) this till exhibits thin (about 1/4-inch thick) horizontal lenses of sorted fine to medium sand. At another locality (60N-15W-15ABAA), the matrix was observed to be fissile. In general, the matrix of the subglacial till is very compact and difficult to excavate.

The matrix color of the subglacial till ranges from light brownish gray (10YR 6/2 and 2.5Y 6/2) near the land surface to light gray (10YR 7/1, 10YR 7/2 and 2.5Y 7/2) deeper in the profile (Appendix A). At one locality (60N-15W-15ABAA) the light brownish gray oxidized zone was observed to be 3-feet thick, abruptly changing to light gray colored below. The shallow depth of oxidation in this till is a reflection of its compact matrix.

Overall, the subglacial till is quite stony, with approximately 30 percent of the total sediment volume composed of angular to very angular pebbles, cobbles and boulders

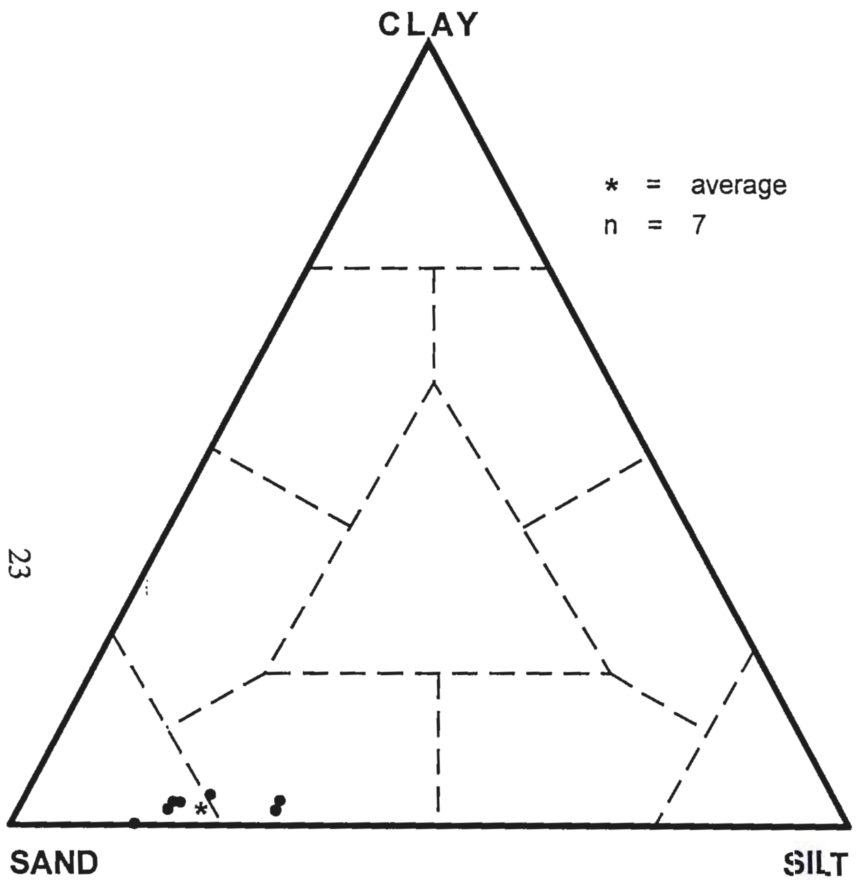


Figure 7. Matrix texture of subglacial till

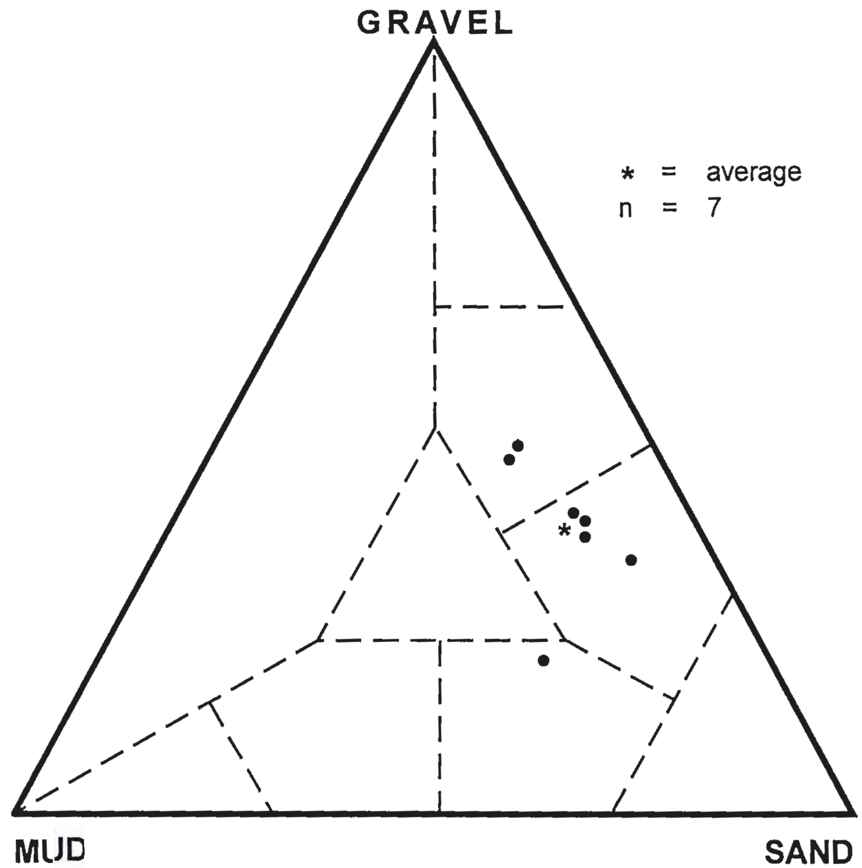


Figure 8. Overall texture of subglacial till

of apparently locally derived granite, gneiss and schist (Fig. 9). Some clasts are faceted and striated. Grain-size analyses of seven samples of subglacial till yield an average composition of 37 percent gravel, 47 percent sand and 16 percent silt + clay (Fig. 8). This average plots in the gravelly sand field of Lawson's (1979) classification (Fig. 2). Large boulders (up to 8 to 15 feet in diameter) commonly occur at the surface within this map unit.

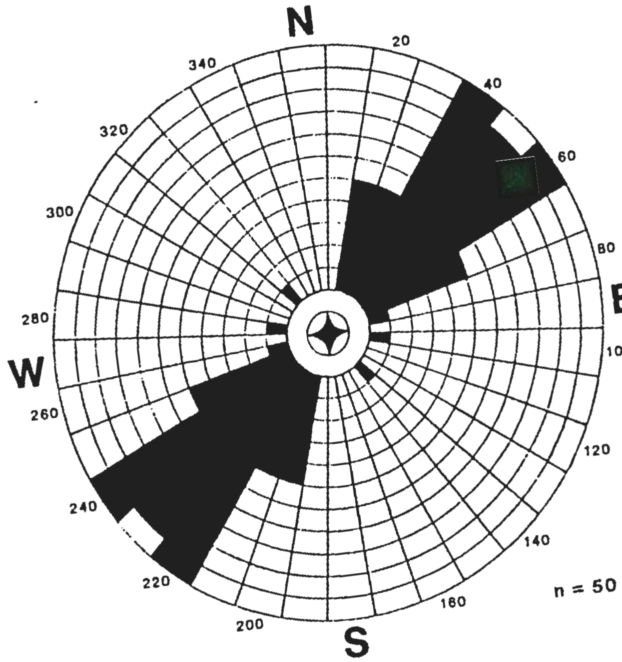
Till fabric studies have been extremely useful in determining the direction of ice movement (Harrison, 1957; Harris, 1971), as well as providing information about the genesis of the till (Goldthwait, 1971; Dreimanis, 1976; 1989). Till fabric analyses were undertaken at two exposures of subglacial till within the study area. These results show a strong fabric oriented northeast southwest (Fig. 10).

Other sediments also occur locally within this map unit as minor constituents. Silty, fine sand was observed in some low areas between ridges of subglacial till. Also, brown, sandy diamicton containing abundant angular stones was observed in some places. The stratigraphic relationship of these sediments to the subglacial till is unclear; however, they are assumed to overlie the subglacial till.

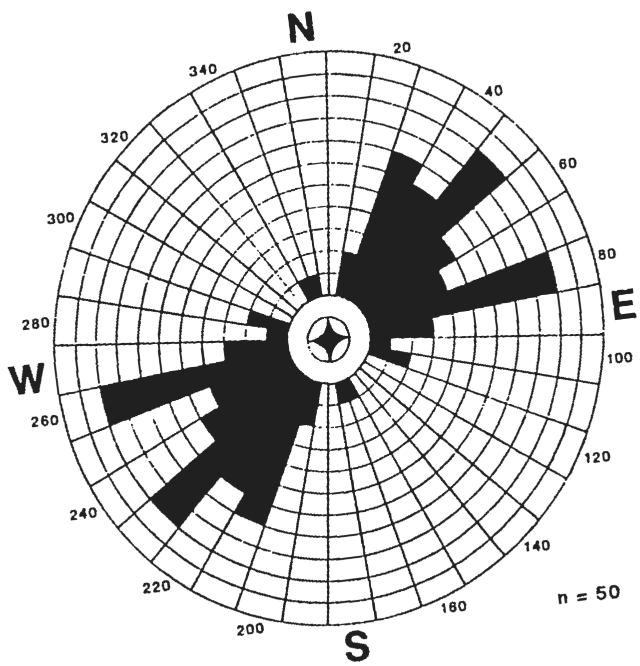
Morphology of the Transverse Ridges

The most extensive surficial occurrences of subglacial till are in the north central portions of 60N-15W and along the northern slope of the Giants Range (Plate I), where it forms a series of northwest-southeast trending ridges 20 to 50 feet high, 200 to 2200 feet long and 200 to 600 feet wide. The ridges have a regular spacing approximately equal to

Figure 9. Exposure of subglacial till in Rogen moraine (60N-15W-03DDBA)



60N-15W-03DDBA



60N-15W-15BCAB

Figure 10. Rose diagrams showing till fabric in Rogen moraine

their width and appear to be fairly symmetrical in cross-section. In some instances, larger ridges are composed of composite smaller ridges, while other ridges are linked by cross-ribs (Plate I). Transverse ridges of subglacial origin (Minell, 1977) have been referred to as Rogen moraine in Sweden (Hoppe, 1952; Lundqvist, 1969, 1981, 1983; 1989; Shaw, 1979; Markgren and Lassila, 1980), hummocky active-ice moraine by Finnish workers (Kujansuu, 1967; Aario, 1977a, 1977b, 1977c) and ribbed moraine in North America (Lee, 1959; Hughes, 1964; Cowan, 1968; Elson, 1968; Prest, 1968; Carl, 1978; Lundqvist, 1981; Goldthwait, 1985; Bouchard and others, 1989; Dyke and Dredge, 1989). Recently, some North American workers (Shilts and others, 1987; Lundqvist, 1989; Bouchard, 1989; Fisher and Shaw, 1992) have used the term Rogen moraine for this type of transverse ridge. Since this term seems to be more widely known than ribbed moraine, the term Rogen moraine will be used in this report. While an exhaustive study of the origin of Rogen moraine is beyond the scope of this report, a brief discussion is presented below.

Origin of Rogen Moraine

It appears from a review of the literature, that numerous theories regarding the origin of Rogen moraine have evolved over the years; therefore, the following discussion is based on some of the more recent literature. In spite of disagreements as to the origin of this type of transverse ridge, several authors have presented similar general observations. The ridges are composed mainly of subglacially deposited till, occur in regional depressions in the landscape, and are associated with streamlined landforms and eskers.

It is proposed that the Rogen moraine in the study area was formed as the margin of the Rainy lobe retreated past the Giants Range. The best developed Rogen moraine in the map area occurs immediately behind the Big Rice moraine (U of M Ag. Exp. Sta., 1971) (Plate II), which intersects the Giants Range near the valley at Hinsdale (hereafter referred to as the Hinsdale gap; 59N-14W-08 and 17). The hills rising to elevations of 1940 feet immediately west of the Hinsdale gap are the highest on the eastern Giants Range (Plate I) and must have been too high for the ice at the Big Rice moraine to surmount. Another factor contributing to the stabilization of the active ice margin at this position is that the crest of the Giants Range east of the Hinsdale gap is oriented approximately parallel to the inferred direction of ice flow (Plate I), providing a minimum of resistance to the flowing ice, whereas west of the Hinsdale gap, the crest of the Giants Range is oriented somewhat oblique to ice flow (Plate I).

The obstruction of ice flow caused by the Giants Range, coupled with the probable existence of a narrow marginal zone of ice which was frozen to the bed, (Clayton and Moran, 1974; Moran and others, 1980) contributed to the formation of a compressive flow regime at the ice margin. It is likely this compression would have resulted in the development of folds in the basal, debris-rich ice (Shaw, 1977); possibly involving the underlying subglacial sediments and bedrock (Moran, 1971, Christiansen and Whitaker, 1976; Minell, 1977; Shaw, 1979) (Fig. 11, Stage 1).

Continued compressive flow would have enhanced the development of larger folds which may have become superimposed through thrusting (Minell, 1977; Shaw,

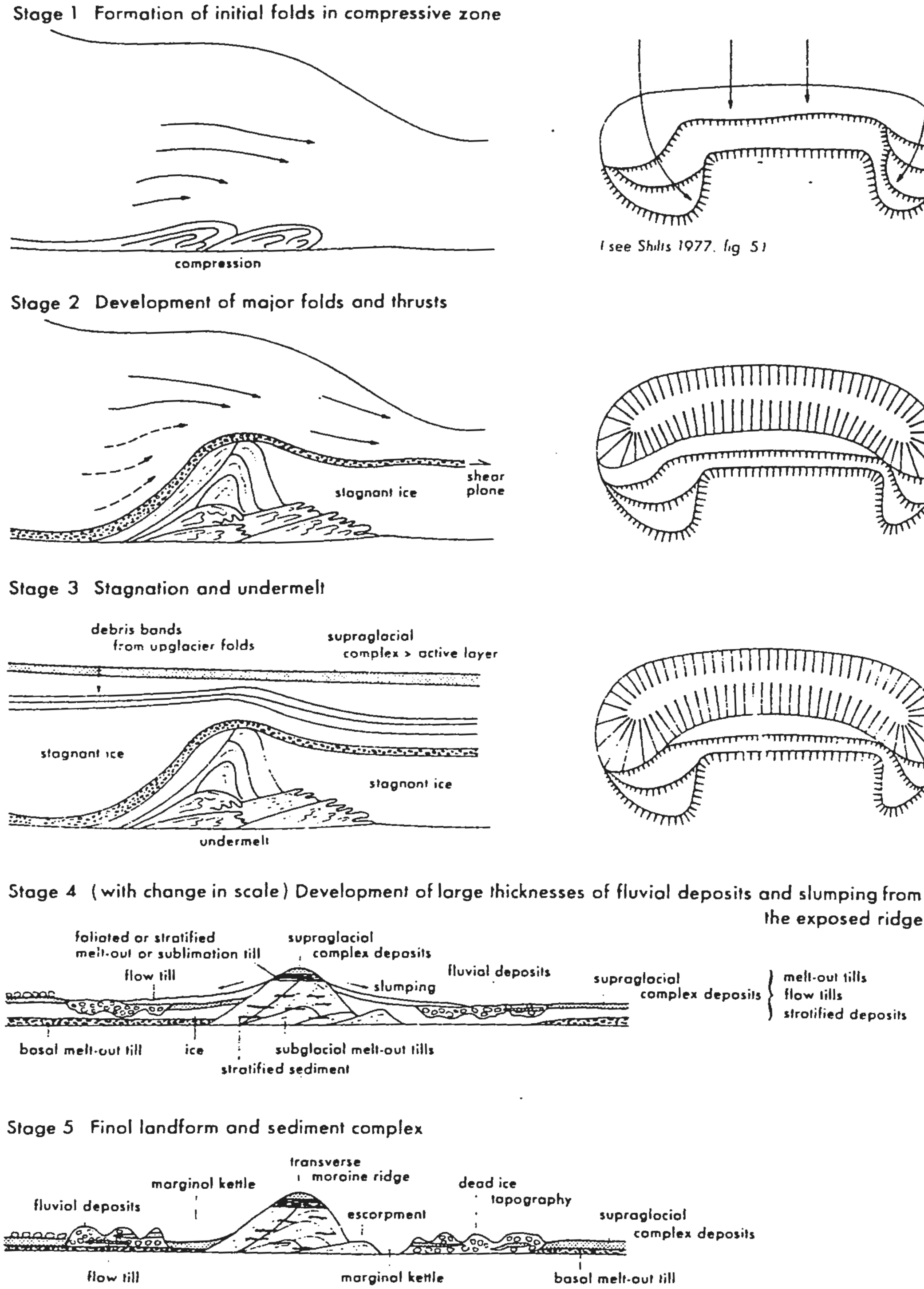


Figure 11. Shaw's (1979) model of Rogen moraine formation

1979; Bouchard, 1989) (Fig. 11, Stage 2). This thrusting within the frozen-bed zone could have plucked portions of solid bedrock from the substrate under conditions of elevated pore-water pressures (Clayton and Moran, 1974; Moran and others, 1980) and shattered bedrock may form the cores of the Rogen moraine as documented by Minell (1977) (Fig. 12).

Exposures in the Rogen moraine of the study area were not large enough to conclusively document either folds or thrust planes, which are difficult to recognize (Shaw, 1979). However, compressive flow is evidenced by the presence of abundant large, angular clasts (Minell, 1977; Shaw, 1979) of biotite gneiss (Fig. 9) derived from the underlying bedrock (Griffin, 1969). At one exposure (60N-15W-10BCACB) within the Rogen moraine, a cluster of large boulders, or possibly fractured local bedrock, was observed to underlie a thin mantle of subglacial till.

It is proposed that some of the debris raised into englacial positions was further elevated into supraglacial positions and released from the ice to become flow till (Boulton, 1968). This supraglacial sediment would have accumulated until it reached a thickness equal to the depth of summer thawing, after which time surface ablation ceased (Boulton, 1972; Shaw, 1979).

In his model, Shaw (1979) suggests stagnation could have occurred either before, or after the onset of wet-based conditions (Fig. 11, Stage 3); however, evidence from the Embarrass area suggests that ice remained active after the onset of wet-based conditions. The well-developed till fabric parallel to the proposed direction of ice flow (Fig. 10) would not be expected in an area of strongly compressive flow (Boulton, 1971; Shaw,

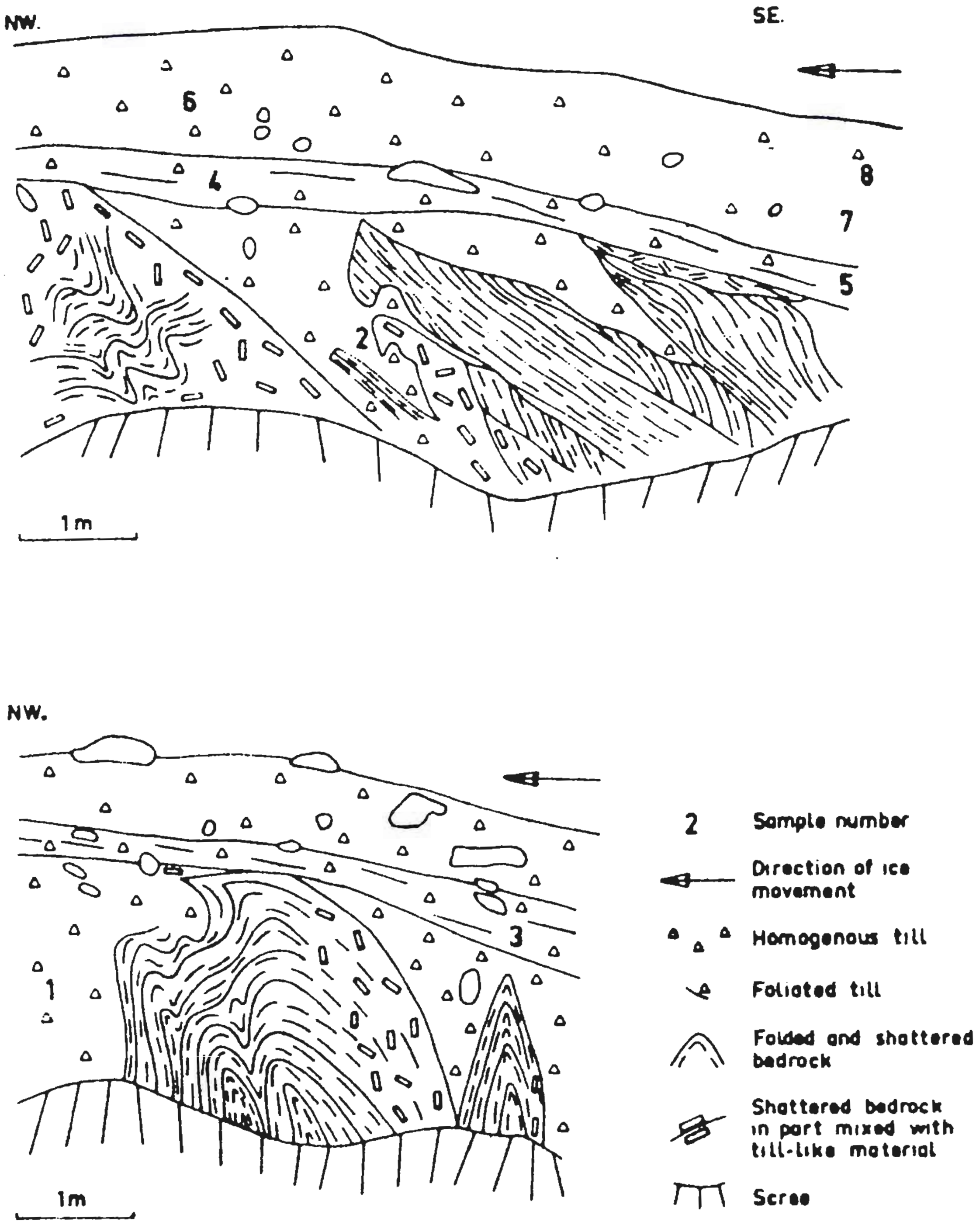


Figure 12. Cross-section of Rogen moraine from Sweden (Minell, 1977)

1979). However, this strong fabric can be explained if the ice remained active following the onset of wet-based conditions and, utilizing the lowlands immediately down-ice (Aario, 1977b), overrode the Rogen moraine, reorienting the till fabric parallel to flow (MacClintock and Dreimanis, 1964). It has been suggested that Rogen moraine does form under nearly stagnant, yet active, ice (Lundqvist, 1969; Markgren and Lassila, 1980).

According to this model, basal meltout continued (Fig. 11, Stage 4), and excess meltwater was channelized into subglacial tunnels where eskers were deposited (Shaw, 1979). This association of eskers and Rogen-type moraine is common (Hoppe, 1952; Lee, 1959; Hughes, 1964; Cowan, 1968; Elson, 1968; Lundqvist, 1969; Aario, 1977a).

As the foliated, debris-rich ice was melting, small cavities formed within the ice into which lenses of sorted sand were deposited. The sorted sand lenses present in one exposure of Rogen moraine in the Embarrass area are horizontal, which indicates deposition by meltout after deformation (Shaw, 1979).

The landforms and sediments associated with the Rogen moraine of the Embarrass area bear some resemblance to the association reported by Shaw (1979) from Sweden (Fig. 11, Stage 5) - transverse ridges composed of subglacial (meltout?) till, localized irregular dead-ice topography composed of supraglacial sediments, and eskers composed of stratified sand and gravel (Plate I). Other sediments associated with Rogen moraine of the Embarrass area are well sorted sands which occur in the swales between ridges and local occurrences of sandy, stony diamicton. The sands were probably deposited in standing water during the late stages of meltout, and correspond to Shaw's

(1979) transgressive sediments. Alternatively, they could have been deposited in standing water some time after the Rainy lobe retreated from this immediate area as proglacial lakes formed in front of the retreating ice margin. The brown, sandy, stony diamicton present locally is interpreted to be a supraglacial till deposited when the debris-covered stagnant ice finally melted (Shaw, 1979) (Fig. 11; Stage 4 and 5).

Another occurrence of subglacial till is in 60N-13W-06 and 07 and 60N-14W-12D where it also appears to form a series of northwest-southeast trending ridges that extend to the west into 60N-14W-01 and 12 (Plate I). However, upon examination of bedrock outcrop maps (Griffin and Morey, 1969), and aerial photographs, it is apparent the ridges in 60N-14W-01 and 12 are controlled by the numerous outcrops of biotite gneiss which exhibit foliation and compositional layering parallel to the ridge crests (Griffin and Morey, 1969). The till in this area is thin, but thickens to the east where the trend of the ridges is less prominent, yet appears to be Rogen moraine. Exposures in this area were small and did not provide enough detail to ascertain whether this subglacial till was deposited through lodgement or basal meltout.

Other Occurrences of Subglacial Till

The other exposures of subglacial till in the study area are in landforms interpreted to be end moraines. Light brownish gray, silty, sandy till is exposed at 61N-15W-25ACAA where erosion by meltwater subsequent to till deposition has bisected the Wahlsten moraine (U of M Ag. Exp. Sta., 1971) exposing a core of subglacial till. This till may have been deposited by lodgement (Rogerson and Batterson, 1980) before the ice stagnated and the supraglacial debris cover developed. Alternatively, this subglacial till may have been deposited by basal meltout (Rogerson and Batterson, 1980) of subglacial and englacial debris after development of a thick cover of supraglacial debris halted surface ablation. Kemmis and others (1981) suggest the development of irregular hummocky topography similar to the Wahlsten moraine to have formed time-transgressively, with the deposition of subglacial till prior to deposition of the supraglacial till which gives the landform its morphology.

The other occurrence of a silty, sandy till interpreted to be of subglacial origin is where it comprises that portion of the Vermilion moraine from the Dunka mine on the east edge of the map area (60N-12W-03 and 10) northwestward to Kramer Bay on the south shore of Birch Lake (61N-13W-36). In this area, the Vermilion moraine has a very asymmetrical cross-sectional profile with a very steep distal slope (approximately 25 degrees) and a rather gentle proximal slope. Examination of aerial photographs reveals that the proximal slope is fluted. Along part of this section of the Vermilion moraine, the lower portions of the proximal slope are overlain by Superior lobe outwash. The asymmetrical cross-sectional profile and the presence of subglacially-deposited till are in agreement with observations reported for push moraines (Andrews, 1975; Boulton, 1986) (Fig. 13).

Push moraine formation may be initiated by an ice margin advancing over a proglacial talus apron pushing this debris into the characteristic asymmetrical shape (Rogerson and Batterson, 1980) (Fig. 14). Moraine construction is continued as subglacial till is deposited by a combination of lodgement and basal meltout. Squeeze-up

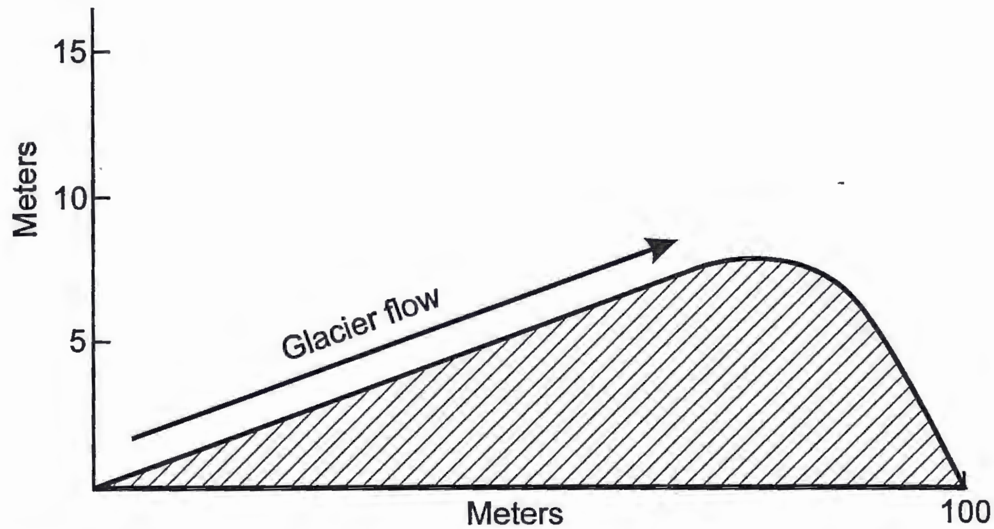
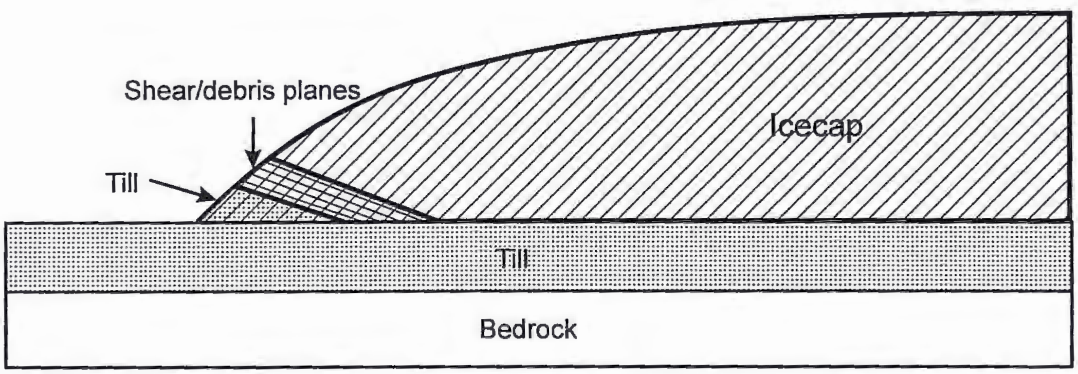


Figure 13. Cross-profile of a push moraine (data from the Outer Lewis Glacier moraine, Baffin Island) (Redrawn from Andrews, 1975)



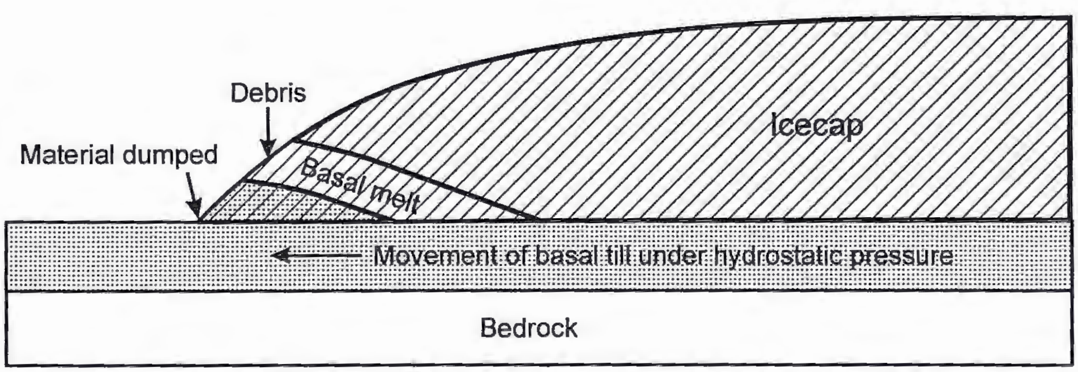


Figure 14. Model of push moraine genesis (Redrawn from Andrews, 1975)

of water-saturated subglacial sediment may also contribute to the formation of push moraines (Andrews, 1975; Rogerson and Batterson, 1980).

Although surface exposures of the subglacial till are fairly limited, it is undoubtedly present in the subsurface elsewhere in the study area. Well logs from throughout the map area commonly report 4 to 17 feet of gray hardpan, or gray hardpan with boulders overlying bedrock at depths ranging from 15 to greater than 100 feet. It is probable that these are occurrences of subglacial till.

Summary of Environment of Deposition

The well-developed fabric, the presence of faceted and striated clasts, the compact nature of this sediment, its expression as Rogen moraine, and the association with other landforms typically found in the subglacial landsystem (Eyles, 1983), support the conclusion that this till was deposited by meltout and/or lodgement in subglacial environment by ice flowing from northeast to southwest.

SUPRAGLACIAL SEDIMENT

This unit actually consists of two separate mappable units which occur in numerous widespread exposures throughout the study area. "Supraglacial sediment complex" and "supraglacial till" are discussed together here because they share several common characteristics and probably have related origins.

The chief sediments present in both units are brown sandy diamictons with loose consistencies which are interpreted as secondary or allo-tills of supraglacial origin

(Dreimanis, 1982). Within the supraglacial sediment complex map unit, these diamictons are associated with poorly-sorted sandy gravel, gravelly sand, moderately well sorted sand and silt in addition to other diamictons whose matrix is enriched in silt and clay relative to the subglacial till. The supraglacial sediment complex map unit is expressed as linear belts and irregular areas of hummocky topography with undrained depressions and short linear ridges common (Plate I). This unit contrasts with the supraglacial till unit which consists solely of brown sandy diamictons and is expressed as generally low-relief till plains, but locally exhibits higher relief where superimposed upon an irregular bedrock surface (Plate I). Both of these units correspond to what has been informally referred to as the "bouldery till" by previous workers in this area (Winter, 1971; Winter and others, 1973; Stark, 1977).

Sediment Description

Supraglacial Till

The matrix of the supraglacial till is dominated by sand. Textural analyses of seven samples (Appendix A) yield an average composition of 86 percent sand, 10 percent silt and 4 percent clay (Fig. 15) which plots in the sand field of Shepard's (1954) classification (Fig. 2). The matrix has a very loose consistency and is very easily excavated.

The color of the matrix of the supraglacial till is very pale brown (10YR 7/3) to pale brown (10YR 6/3) and light yellowish brown (10YR 6/4) to brown (10YR 5/3) (Appendix A), reflecting the oxidized nature of this permeable sediment.

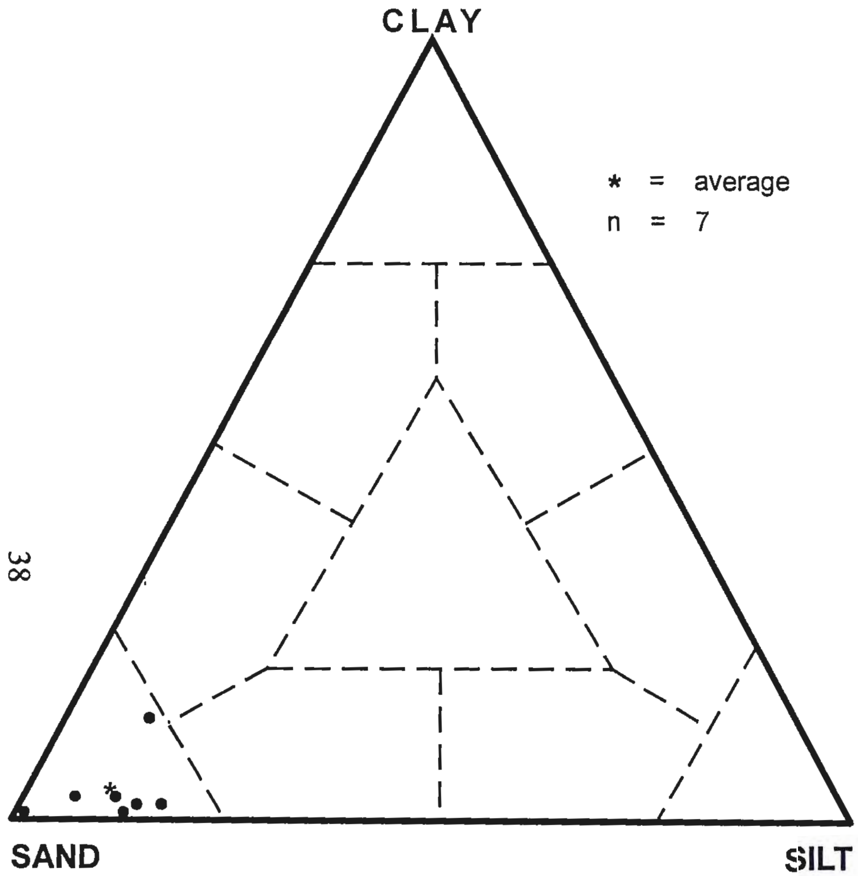


Figure 15. Matrix texture of supraglacial till

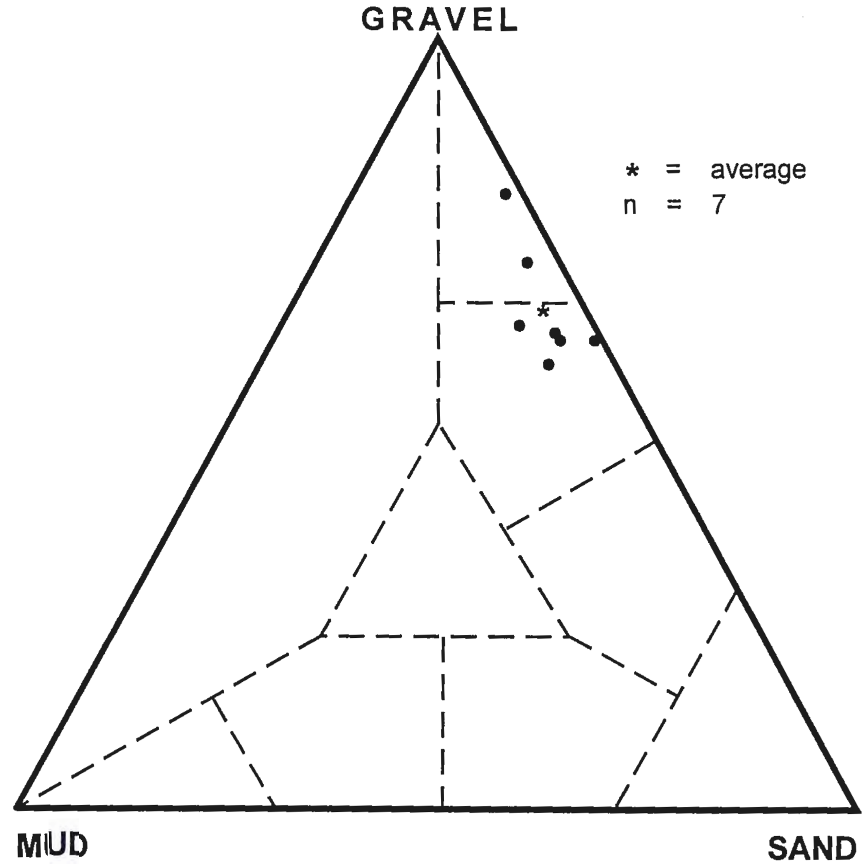


Figure 16. Overall texture of supraglacial till

Overall, the supraglacial till is very stony, with approximately 30 to 50 percent of the total sediment volume composed of very angular to rounded pebbles, cobbles and boulders of locally derived granite, gneiss, schist and greenstone. Grain-size analyses of seven samples (Appendix A) plot in the sandy gravel and gravel fields of Lawson's (1979) classification (Fig. 2). The overall average texture of these seven samples is 65 percent gravel, 30 percent sand and 5 percent silt + clay (Fig 16) which plots in the sandy gravel field of Lawson's (1979) classification.

The portion of this unit in the northwestern part of the map area which lies below 1450 feet in elevation is locally overlain by well-sorted fine sand deposited in Lake Norwood.

Supraglacial Sediment Complex

The most common sediments present in this map unit are brown sandy diamictons with very loose consistencies due to a lack of fine matrix. Textural analyses of nine samples (Appendix A) yield an average composition of 90 percent sand, 7 percent silt and 3 percent clay (Fig. 17) and plot in the sand field of Shepard's (1954) classification (Fig. 2).

The color of the matrix of these diamictons ranges from pale brown (10YR 6/3) and light yellowish brown (10YR 6/4) to yellowish brown (10YR 5/4) and brown (10YR 5/3) (Appendix A). These oxidized colors are very similar to the matrix color of supraglacial till and also reflect the coarse matrix of this till.

These diamictons are also generally very stony, with approximately 30 to 50

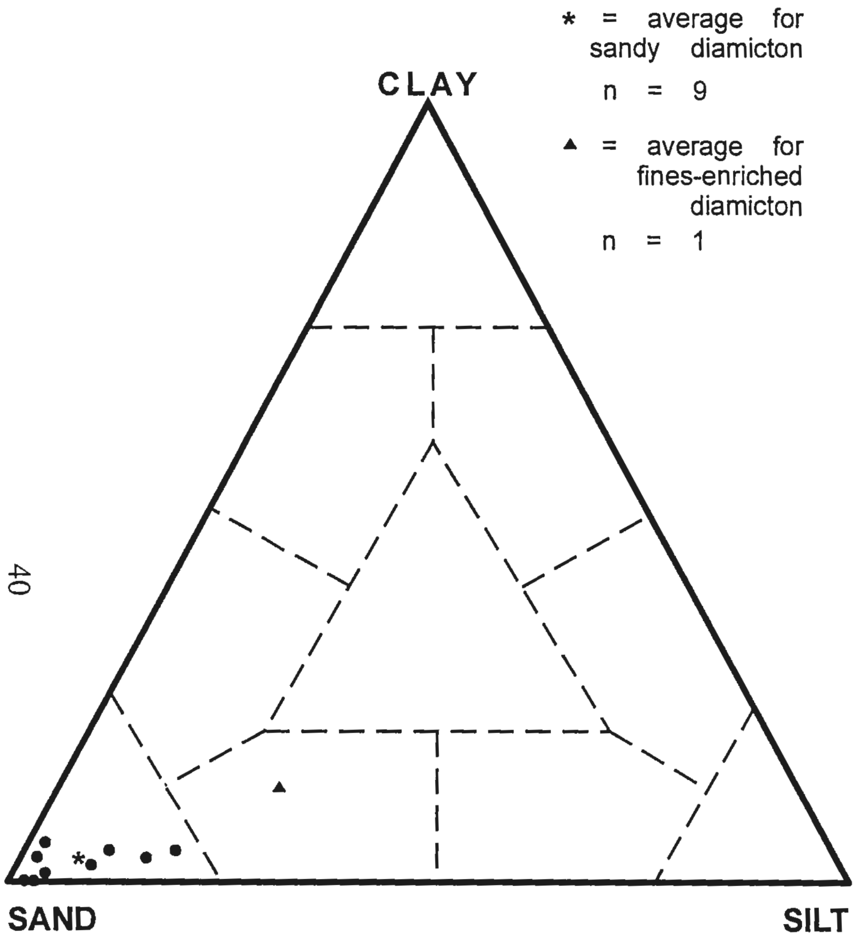


Figure 17. Matrix texture of supraglacial sediment complex diamictons

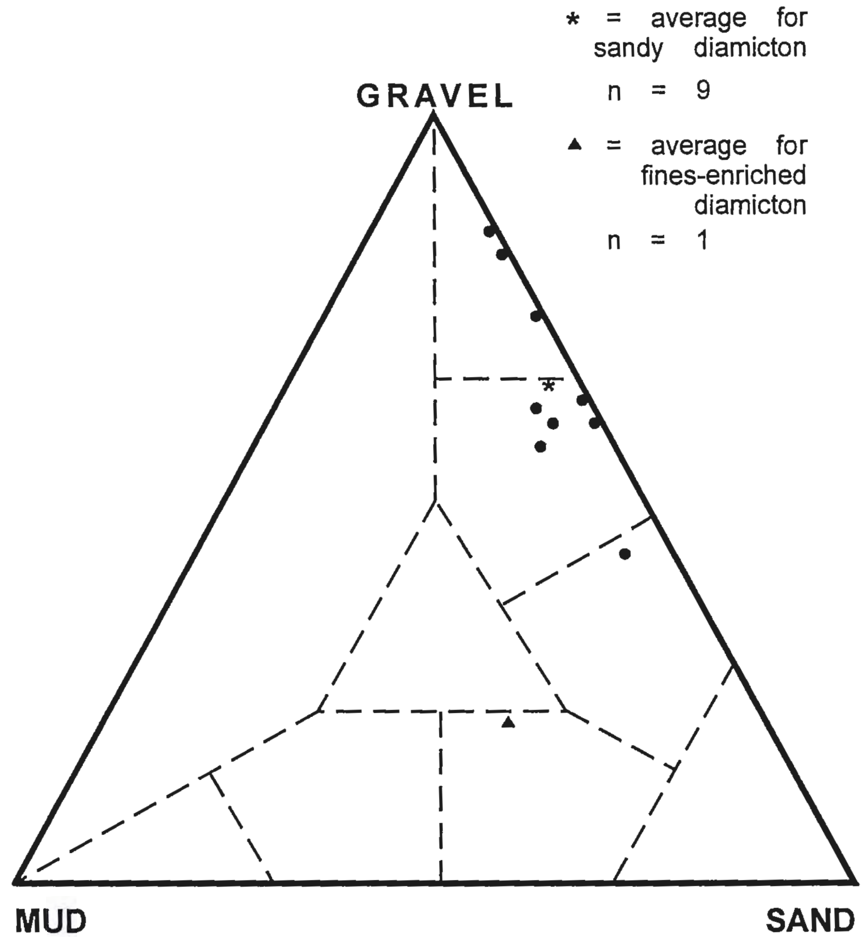


Figure 18. Overall texture of supraglacial sediment complex diamictons

percent of the total sediment volume composed of angular to rounded pebbles, cobbles and boulders of locally derived granite, gneiss, schist and greenstone (Fig. 19). Grain-size analyses of nine samples (Appendix A) plot in the sandy gravel, gravelly sand and gravel fields of Lawson's (1979) classification (Fig. 2). The overall average texture of these nine samples is 65 percent gravel, 31 percent sand and 4 percent silt + clay (Fig. 18).

This map unit also contains diamictons whose matrix is enriched in silt and clay relative to other supraglacial diamictons and the subglacial till (Fig. 19). The matrix of these diamictons is quite silty. Textural analysis of one sample (Appendix A) yielded a composition of 61 percent sand, 26 percent silt and 12 percent clay (Fig.17) which plots in the silty sand field of Shepard's (1954) classification (Fig. 2). The matrix color of this fines-enriched diamicton is dark brown (10YR 3/3) (Appendix A).

The fines-enriched diamictons are generally less stony than other diamictons in the study area. Textural analysis of one sample yielded an average composition of 21 percent gravel, 48 percent sand and 31 percent silt + clay (Fig. 18) which plots in the silty sand field of Lawson's (1979) classification (Fig. 2). The gravel fraction is composed primarily of very angular to rounded pebbles of granite, gneiss, schist and greenstone.

The fines-enriched diamictons occur as beds or lenses 1 to 5 feet thick and appear to have a limited lateral extent as they were not observed to be continuous from one exposure to another.

Occurring in association with the diamictons of the supraglacial sediment complex are a variety of sediments that have been sorted to some degree by flowing

Figure 19. Exposure of fines-enriched diamicton overlying sandy diamicton within the supraglacial sediment complex map unit. (60N-14W-29BDDBA) Sample EM-11 was taken from fines-enriched diamicton. Sample EM-12 was taken from sandy diamicton (Appendix A).

water. These sediments range from poorly-sorted sandy pebble-gravel and pebbly sand to moderately well sorted coarse to very-fine sand and silt. Some of these sorted sediments are cross-bedded, while others exhibit collapsed bedding. The individual clasts in these sediments commonly are rounded.

These sorted facies of the supraglacial sediment complex occur as beds ranging from a few inches to a few feet in thickness and are interbedded with both sandy diamictons and fines-enriched diamictons. These beds of sorted sediment appear to have restricted lateral extent since they were not traceable beyond a single exposure.

Geomorphology

Supraglacial Till

The type of feature exhibited by the supraglacial till may be referred to as a supraglacial till plain (Paul, 1983) which is characterized by rather thin accumulations of supraglacial till producing relatively low-angle slopes. Some areas of supraglacial till plain in the study area exhibit higher relief, for example, the north flank of the Giants Range. In these areas, the higher relief is apparently due to the irregular underlying bedrock surface. This map unit is comparable to what other workers in the area have called ground moraine (Winter and others, 1973; Stark, 1977; Olcott and Siegel, 1978; Friedman, 1981).

Supraglacial Sediment Complex

The supraglacial sediment complex map unit most commonly occurs as linear belts, but also as irregular areas, of hummocky topography (Plate I). Numerous undrained depressions, conical hills and short linear ridges combine to form this hummocky landscape.

Some of these linear belts of hummocky supraglacial sediment complex have previously been, and are in this study, interpreted as end moraines (Plate II). The southernmost of these moraines has been named the Big Rice moraine (U of M Ag. Exp. Sta., 1971) north of the Giants Range and the Allen moraine (U of M Ag. Exp. Sta., 1971; 1981) south of the Giants Range (Plate II). The Allen moraine has also been referred to as the first moraine (Stark, 1977) and the first pro-Vermilion moraine (Friedman, 1981). Traditionally, the Big Rice moraine and the Allen moraine were considered separate, but evidence compiled during this study suggests they represent a contemporaneous ice-marginal position and will hereafter be compositely referred to as the Big Rice moraine.

As the Rainy lobe flowed southwestward into the study area during recession, the topographic high of the Giants Range dictated the geometry of the ice margin. The obstruction to flow caused by the Giants Range resulted in thicker ice on the flanks of the Giants Range than over its crest. The thicker ice flowed at higher velocities (Sugden and John, 1976) producing a doubly lobate ice margin with the crest of the Giants Range occupying the interlobate area (Fig. 20).

Supraglacial sediment complex also forms a portion of the Wahlsten moraine (U of M Ag. Exp. Sta., 1971) which trends east-west across the northern portion of the map area (Plate I, Plate II). This moraine extends westward to beyond Pfeiffer Lake (U of M Ag. Exp. Sta., 1971; Winter and others, 1973; Hobbs and Goebel, 1982) and is truncated

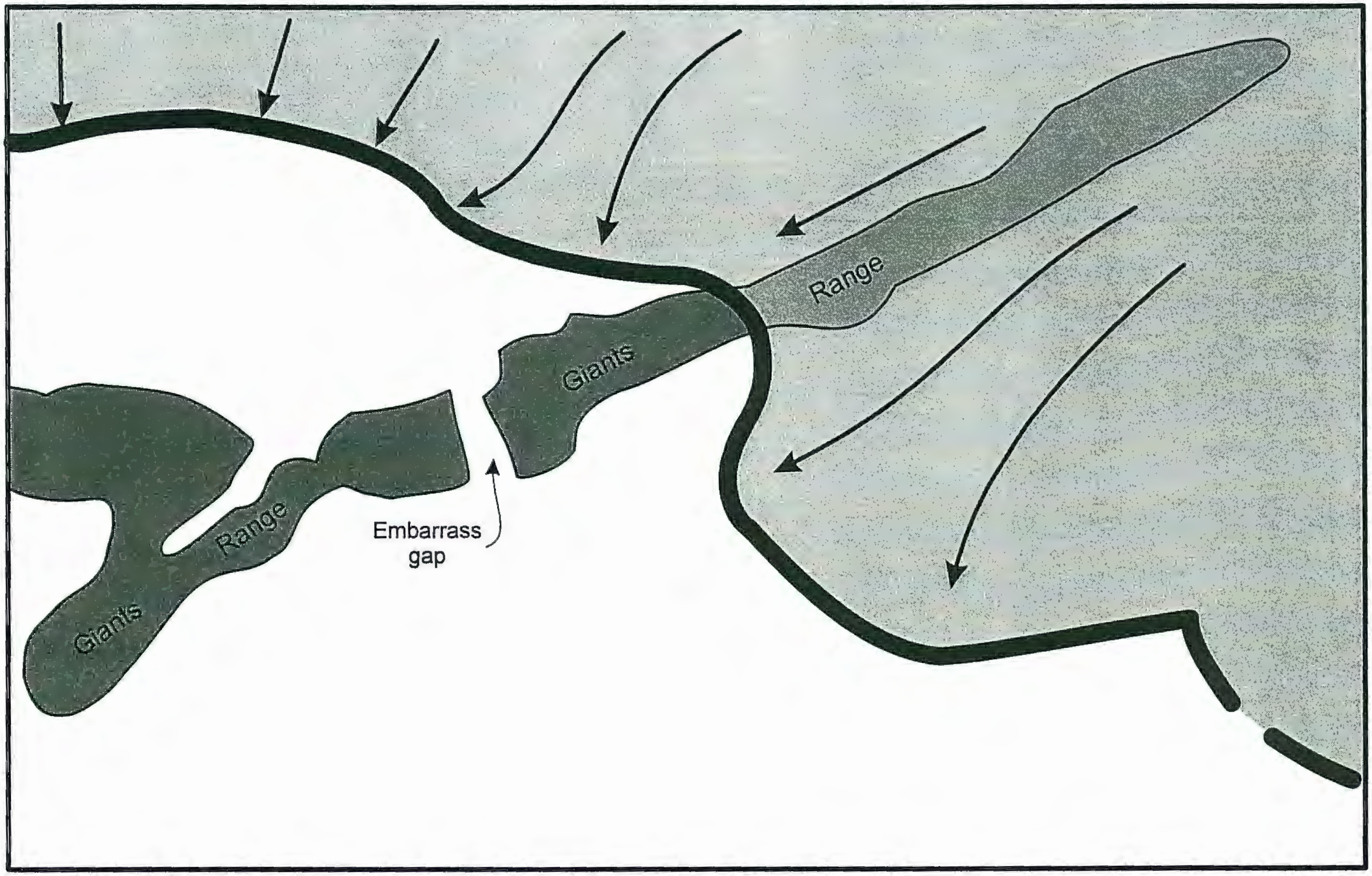


Figure 20. Rainy lobe ice margin at the Big Rice moraine

on the east by sediments related to the Vermilion moraine (Plate I, Plate II). The eastward extension and correlation of the Wahlsten moraine is uncertain; however, Stark (1977) did report another end moraine (his second moraine) north of the Big Rice moraine (his first moraine) and south of the Vermilion moraine (his third moraine). On the regional geomorphology map compiled for the 1992 Midwest Friends of the Pleistocene Field Trip (Plate I in Lehr and Hobbs, 1992), the moraine between the Allen and Vermilion moraines was named the Wampus Lake moraine. It is likely that the Wahlsten moraine correlates with the Wampus Lake moraine, but critical areas for correlation are covered by younger sediments.

Supraglacial sediment complex also forms that part of the Vermilion moraine from the southwest shore of Bear Island Lake (61N-13W-21) northwest to the southwest shore of Eagles Nest Lake No. 2 (62N-14W-33D) (Fig. 31).

Other significant occurrences of supraglacial sediment complex include a band of hummocky topography from Lempia Lakes (61N-15W-33) south to approximately 1.5 miles north of Embarrass (60N-15W-14), and a more irregular tract in the vicinity southeast of Kaunonen Lake (60N-14W-23, 24 & 25). In each of these areas, the proximal side of these tracts of hummocky topography exhibit ice-contact faces (Plate II), suggesting that they also mark recessional ice-marginal positions of short duration. The other occurrence of supraglacial sediment complex is along the west shore of Sabin Lake (59N-16W-13 and 19) (Plate I).

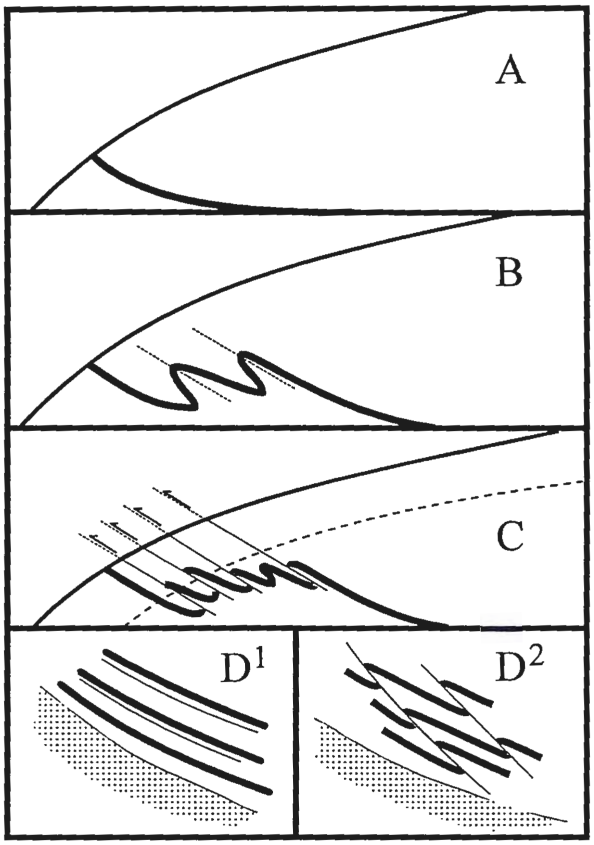
Processes of Supraglacial Deposition

As mentioned previously, "supraglacial till" and "supraglacial sediment complex" probably have related origins; therefore, their sedimentary environments will be discussed together below.

Compressional flow near the margins of ice sheets (Weertman, 1961; Clayton and Moran, 1974) deforms the subglacial and englacial debris bands into a series of isoclinal folds whose axial planes strike approximately parallel to the ice margin and dip up-ice (Paul, 1983). Continued compression results in the development of thrust faults which may lie either parallel or oblique to the axial planes of the folded debris bands (Paul, 1983) (Fig. 21).

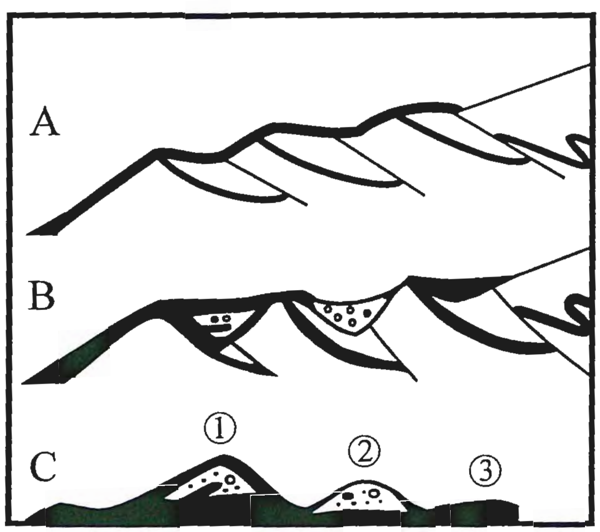
As the surface of the ice sheet in the marginal zone is lowered through ablation, the debris bands crop out as ridges (Boulton, 1972; Paul, 1983) (Fig. 21 and 22) from which debris is released as flow tills (Boulton, 1968; 1972) which are secondary or allo-tills of supraglacial origin (Dreimanis, 1982). These water-soaked tills are fluid, the degree of which depends on water content, and are susceptible to flow down even low-angle (1 to 2 degree) slopes (Lawson, 1979). Flow tills are deposited either in low areas on the ice, which may occur as troughs between debris band outcroppings (Boulton, 1968; Paul, 1983), or in the proximal proglacial zone (Boulton, 1968) (Fig. 22). During flowage, some of the fine matrix is winnowed from the sediment and deposited elsewhere, resulting in a coarse-textured supraglacial till with a loose consistency.

Sometimes, enough of the fine matrix (clay, silt and some sand) is winnowed from supraglacial sediment to produce an open-textured remnant of boulders, cobbles,



(A) Idealized path taken by the basal debris band during compressive flow. (B) Folds form on the debris band, which are drawn out by flow into tight structures whose axes lie parallel to flow lines. (C) Thrust faults dissect the folds along lines parallel to flow. Ablation of the glacier to dashed line leads to a typical structural sequence at the glacier terminus. In plan, the thrust faults may lie parallel to the strike of the debris bands (D^1) or oblique to them (D^2). Ice melt, under a cover of till derived from the debris, leads to a topography of ridges and troughs (Fig. 22).

Figure 21. Development of a multiple debris band sequence (Redrawn from Paul, 1983)



(A) shows the initial system of ridges and troughs in which outwash and flowed till accumulate (B). After deglaciation (C) the topography is "inverted", with the formation of ridges from the trough fillings. Some ridges (1) contain both outwash and flowed till; others (2) contain mostly outwash, and others (3) mostly flowed tills.

Figure 22. Development of the supraglacial landsystem by relief inverson (Redrawn from Paul, 1983)

pebbles and sand (some of the sandy gravels present within in the supraglacial sediment complex). The fine sediment accumulates in hollows on the ice and is resedimented, sometimes among preexisting boulders, cobbles and pebbles producing a diamicton which may be relatively enriched in silt and clay (Boulton, 1967). The fines-enriched diamictons in the

study area were probably deposited by processes similar to this.

Some of the fine sediment winnowed from supraglacial debris may be deposited in ephemeral supraglacial lakes formed either between debris band outcroppings or other low areas on the ice surface (Boulton, 1972; Paul, 1983). These sediments commonly exhibit collapsed bedding due to melting of the underlying ice (Clayton, 1964; Boulton, 1972). The sorted sand and silt facies of the supraglacial sediment complex, which locally exhibits collapsed bedding, was probably deposited in this type of environment.

Ephemeral supraglacial fluvial systems may also develop in marginal areas of ice sheets (Boulton, 1972) and deposit sand and gravel (Fig. 22B). Some of the sorted facies of the supraglacial sediment complex, such as poorly-sorted sandy gravel, pebbly sand, and sand may have been deposited in this type of environment.

The unequal distribution of debris on the glacier surface (Fig. 22) results in an irregular distribution of ablation rates, since only a few inches of debris cover slows ablation rates considerably (Boulton, 1972). These spatially variable ablation rates produce a very dynamic depositional environment with inversions of topography common. Resedimentation of supraglacial debris continues until all underlying ice is melted, which may require 1000 to 3000 years (Clayton, 1967; Florin and Wright, 1969).

If the topography resulting from supraglacial sedimentation reflects patterns inherited from active ice, it is classified as controlled disintegration topography (Gravenor and Kupsch, 1959). Controlled disintegration topography may form when thrust planes develop parallel rather than oblique to debris bands (Paul, 1983) (Fig. 21 D^1) leaving intact the ridges of debris and intervening troughs. When the underlying ice melts, the trough fillings will be deposited as linear ridges (Fig. 22). These linear ridges are a minor constituent of the hummocky topography of the supraglacial sediment complex.

If the topography resulting from supraglacial sedimentation is irregular and appears not to have inherited patterns from active ice, it is termed uncontrolled disintegration topography (Gravenor and Kupsch, 1959). Uncontrolled disintegration topography may result when thrust planes develop oblique to debris bands (Paul, 1983) (Fig. 21 D^2) and subsequent melting of the underlying ice produces an irregular topography. This type of supraglacial sedimentation contributed to the formation of the irregular hummocky topography of the supraglacial sediment complex map unit. An example of uncontrolled disintegration topography is the Wahlsten moraine in the vicinity of Fishing Lakes (61N-14W-20 and 29) (Plate I).

If the supraglacial sediment is not concentrated in troughs on the ice surface, in other words, is more uniformly distributed on the ice surface, the resulting landform will be a supraglacial till plain (Paul, 1983). Another mechanism by which a supraglacial till plain may form is when the sediment filling the supraglacial troughs is so fluid it does not retain its ridge-like morphology upon melting of the underlying ice, but rather flows laterally to produce low relief topography (Paul, 1983). This scenario is a less favorable explanation for the origin of supraglacial till plains in the study area. The relatively coarse nature of the diamictons present in this map unit would have enabled rapid evacuation of contained water once flow was initiated.

ICE-CONTACT DEPOSITS

This map unit is widespread in the western one third of the map area, but also occurs in the central and eastern parts of the map area. It is considered broadly gradational with the supraglacial sediment complex map unit. The supraglacial sediment complex unit consists of primarily diamictons, with lesser amounts of sorted sediment, while the ice-contact deposits unit is predominantly sorted sediment, with lesser amounts of diamicton. Furthermore, the two units exhibit contrasting geomorphology. Exposures in this unit are fairly numerous due to small-scale gravel mining of its sand and gravel facies.

Sediment Description

The sediments in this map unit range from poorly-sorted sandy pebble-gravel and pebbly sand to well-sorted, very-fine to very-coarse sand (Appendix A). Interbedded with these sorted lithologies are sandy diamictons. According to the trilinear classification of Shepard (1954), all of these sediments are classified as sand (Fig. 23). However, when the gravel fraction is considered, they range from gravel and sandy gravel to gravelly sand and sand (Fig. 24). All of these sediment types may be present at a particular exposure. For example, Table 1 shows the variability of sediment types present within a single

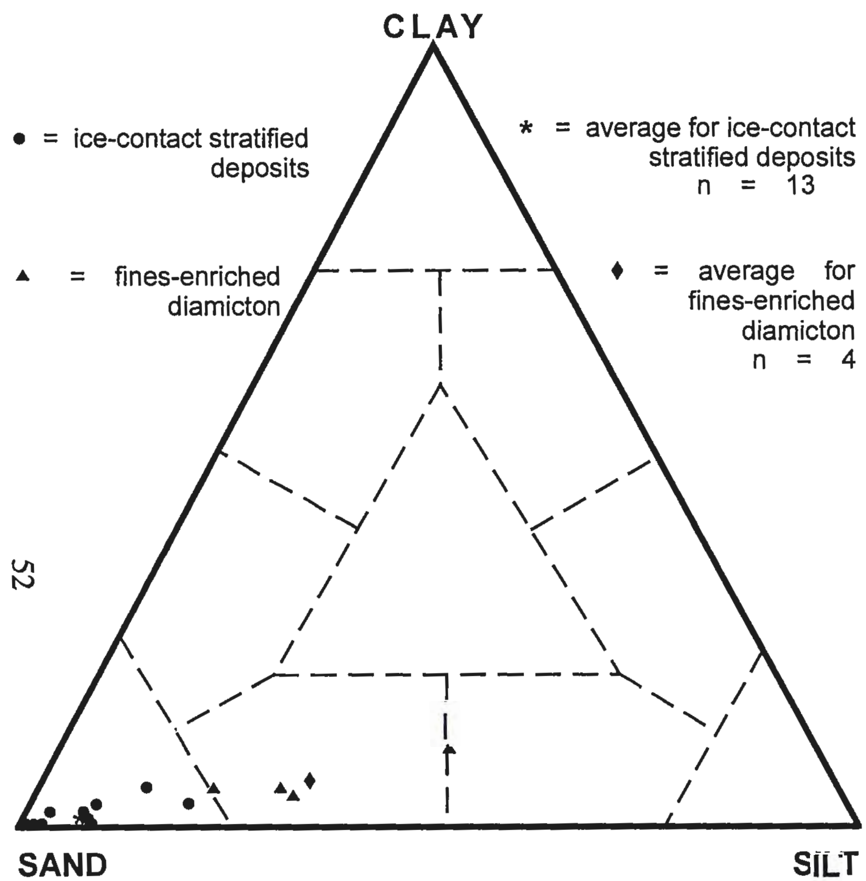


Figure 23. Texture of ice-contact deposits

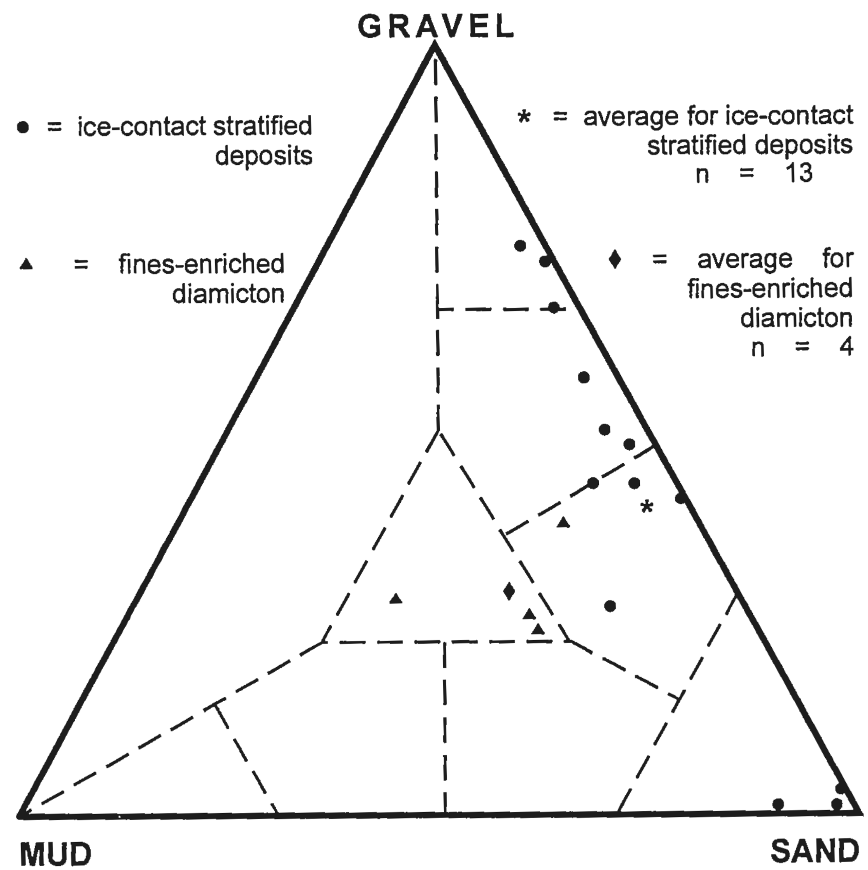


Figure 24. Overall texture of ice-contact deposits

landform (a kame at 60N-14W-20ADAD) mapped as ice-contact deposits.

TABLE 1

Sample Number	% Gravel	% Sand	% Silt	% Clay
Em-6	3	97	0	0
Em-7	41	59	0	0
Em-8	57	37	3	1

Em-6 is a weakly stratified, well-sorted sand; Em-7 is a poorly-sorted pebbly sand; and Em-8 is best described as a sandy, cobbly diamicton.

The colors of the sorted facies of ice-contact deposits range from light brownish gray (10YR 6/2) and pale brown (10YR 7/3 to 10YR 6/3) to yellowish brown (10YR 5/4 and 10YR 6/4), brown (7.5YR 4/4) and dark brown (10YR 4/3) (Appendix A). These generally brown colors reflect the oxidized nature of these permeable sediments.

The pebble, granule and very coarse sand fractions of the ice-contact sediments consist of subangular to rounded clasts of granite, gneiss, schist, greenstone and other lithologies common to the Vermilion district. The ice-contact deposits usually exhibit horizontal stratification and collapsed bedding. Some of the sands within this unit are cross-bedded.

Locally interbedded with these sandy, sorted sediments and sandy diamictons are diamictons whose matrix is relatively enriched in silt and clay (Fig. 25), ranging from silty sand to sandy silt (Shepard, 1954) (Fig. 23). The overall texture of these fines-enriched diamictons ranges from sandy gravel to gravel-silt-sand (Lawson, 1979) (Fig. 24). Textural analyses of four samples of fines-enriched diamicton from this map unit have an average matrix texture of 62 percent sand, 32 percent silt and 6 percent clay

Figure 25. Exposure of ice-contact deposits showing fines-enriched diamicton overlying sorted sand (60N-15W-24DCBB)

(sandy silt) (Fig. 23) and an overall texture of 29 percent gravel, 44 percent sand and 27 percent silt + clay (gravel-silt-sand) (Fig. 24). The matrix color of the fines-enriched diamictons range from pale brown (10YR 6/3 and 10YR 7/3) to yellowish brown (10YR 6/4) and dark grayish brown (10YR 4/2).

Geomorphology

Areas mapped as ice-contact deposits constitute several distinct landforms in the study area: eskers, solitary ice-contact fans, ice-contact complexes, and two distinct types of linear ice-marginal features.

The most prominent esker systems are located in the western portions of the map area and range from 1/2 mile to 3 1/2 miles in length (Plate I). These eskers are 30 to 60 feet high with sharp crests and fairly regular longitudinal profiles. The southwestern termini of the northeastern trending esker systems just northwest of Salo Corner broaden as they end on a lacustrine plain. These features are termed deltaic expansions and form when subglacial streams enter a standing body of water (Hughes, 1964). Other smaller eskers occur within the subglacial till unit in the eastern and southwestern parts of the map area. Eskers are also a component of the larger ice-contact complexes in the western part of the map area (Plate I).

Ice-contact fans are common within the central and western portions of the Lake Norwood basin. In most cases, these fans are characterized by their steep northeastward-facing ice-contact slopes and gently slope to the southwest. Good examples of solitary ice-contact fans are in 60N-16W-25 and 36, and in the area immediately north of Embarrass. Ice-contact complexes composed of multiple ice-contact fans and eskers are common in the western one-quarter of the map area.

Areas mapped as ice-contact deposits which comprise landforms interpreted as end moraines, exhibit two distinct morphologies. One landform is expressed as a very distinct linear ridge, typically ¼ to ½ mile wide. It has fairly straight, steep ice-contact proximal slope and a slightly more gradual distal slope which is typically scalloped. These portions have flat tops with elevations approximately 1450 to 1460 feet above sea level (Fig. 26). In a few areas, the ridge is somewhat wider and higher (approximately 1500 feet) and its surface is pitted. At all locations, this type of ridge is bordered on its distal side by lacustrine plains. These areas of ice-contact deposits comprise the western portions of the Wahlsten moraine from 61N-15W-30 to the western edge of the map area (Plate I) and the Vermilion moraine from the vicinity of Tower to the south shore of Nett Lake (Fig. 27).

Features similar to this have been referred to as a delta moraines (Sugden and John, 1976), ice-marginal deltas (Hyvarinen, 1973) or glaciofluvial marginal deltas (Gluckert, 1977; Lundqvist, 1979; Eronen and Vesajoki, 1988). End moraines with similar morphology and composed primarily of stratified sand and gravel are documented in areas north of the study area where Lake Agassiz bordered the retreating Rainy lobe (Zoltai, 1961; 1965; Dredge and Cowan, 1989; Sharpe and Cowan, 1990). This type of moraine is also similar to the Salpausselkä moraines in Finland.

The other morphology ice-marginal ice-contact deposits exhibit is a broader, less distinct ridge with a steep ice-contact proximal slope and a gradual distal slope which grades into outwash plains and lacustrine plains (Fig. 28). This feature is best described

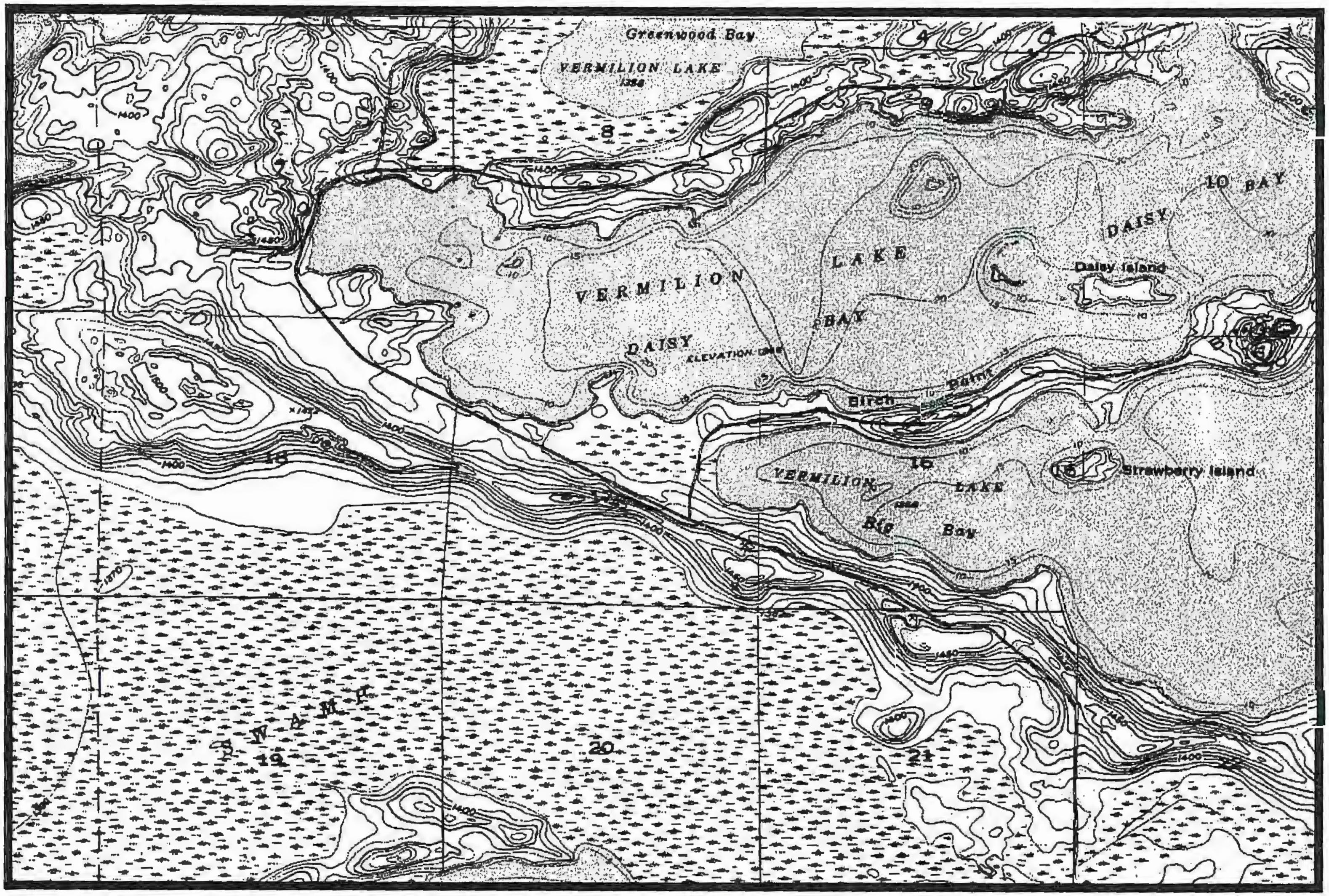


Figure 26. Morpohology of the subaqueous outwash facies of the Vermilion moraine (Portions of the Tower and Lost Lake quadranges - 62N-16W)

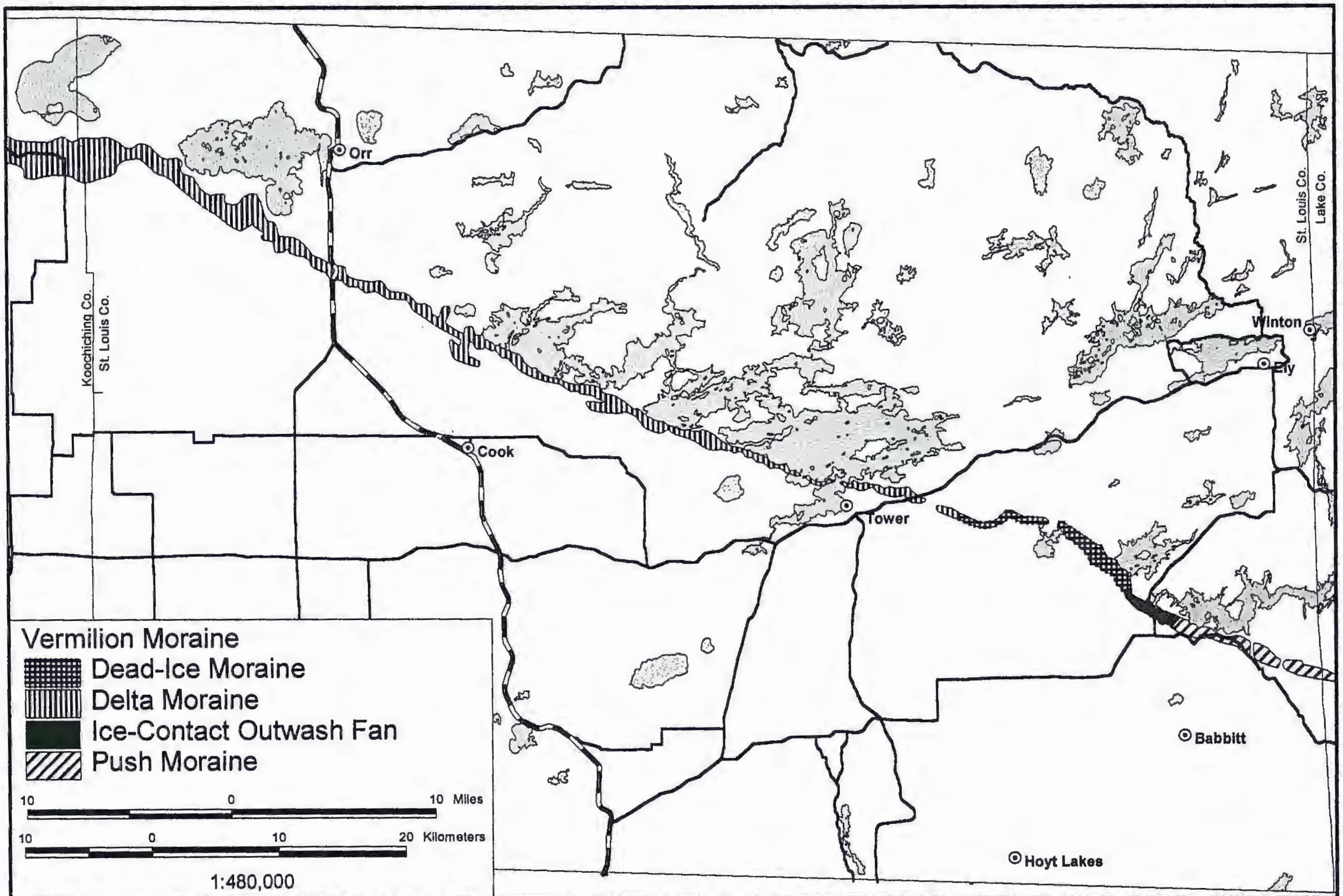


Figure 27. Summary Map of the Vermilion Moraine

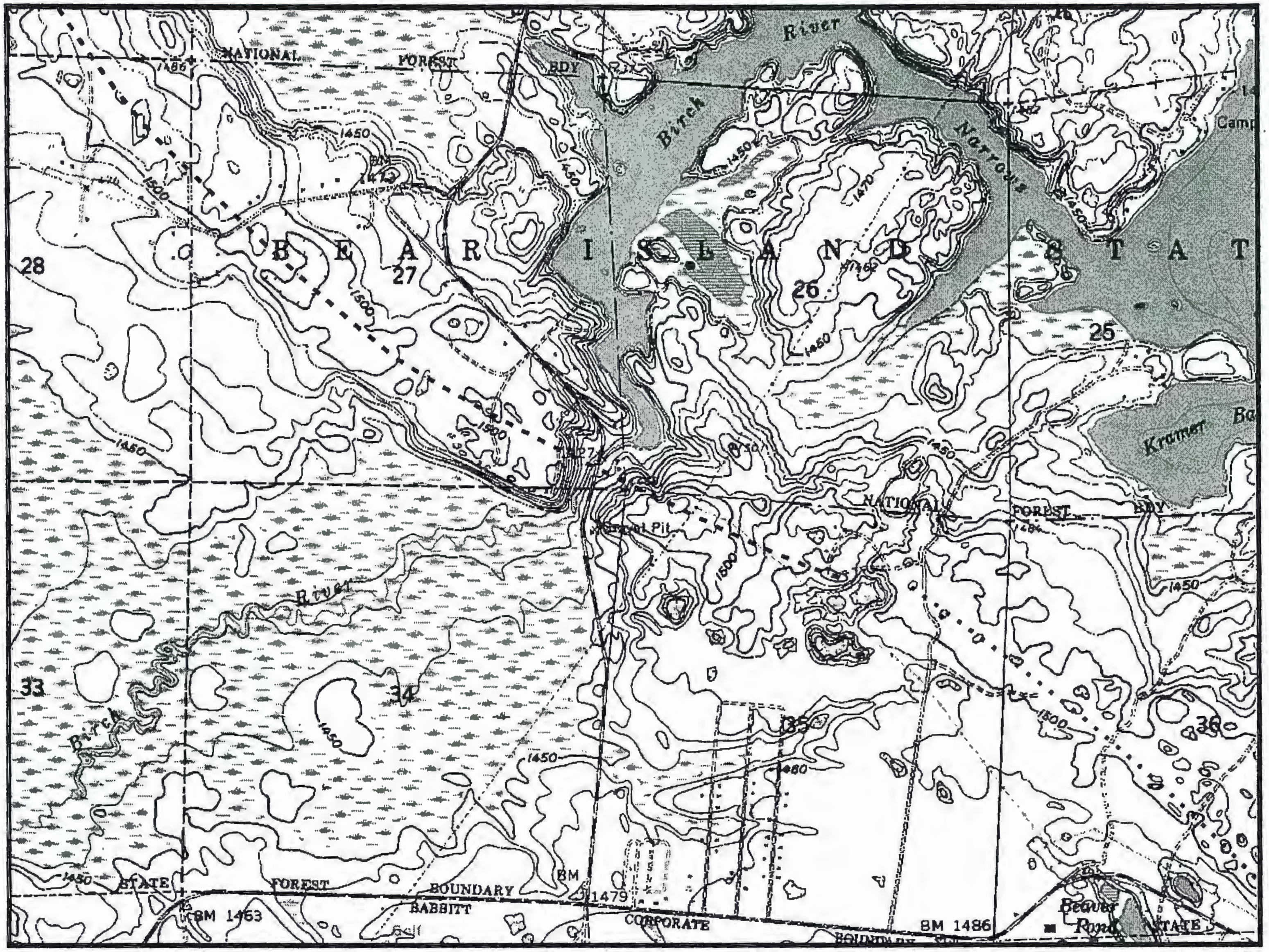


Figure 28. Morpohology of the ice-contact outwash fan portion of the Vermilion moraine (Portion of the Babbitt quadrangle - 61N-15W) - - - - - - - - Crest of ice-contact outwash fan · · · · · · · · · · Crest of push moraine

as a series of coalesced ice-contact outwash fans and constitutes a short segment of the Vermilion moraine from Kramer Bay of Birch Lake northwest to 61N-13W-21 in the eastern portions of the map area (Fig. 27; Plate I).

Sedimentary Environment

The eskers that occur in the Embarrass area were probably deposited by subglacial meltwater streams, as evidenced by the preservation of very distinct ridge crests and fairly regular longitudinal profiles (Lundqvist, 1979). The preservation of the deltaic expansions also supports a subglacial origin for these eskers.

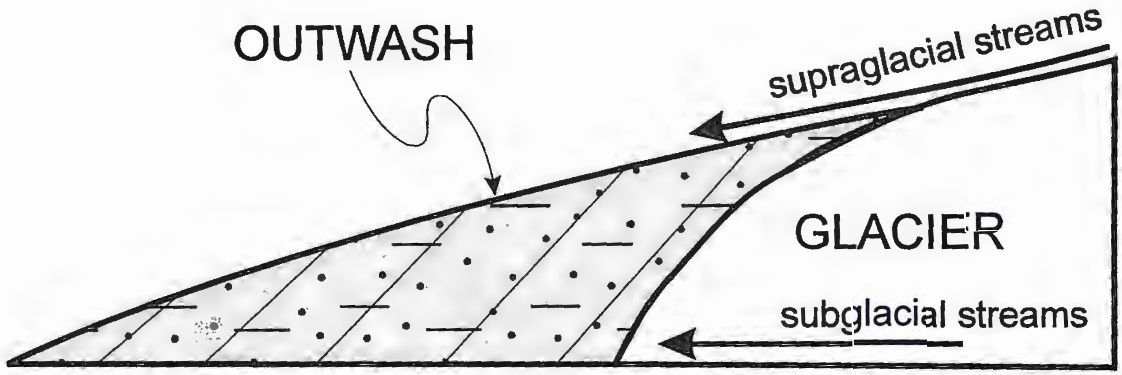
The numerous ice-contact fans and ice-contact complexes present in the areas west and southwest of Embarrass formed as a portion of the Rainy lobe ice margin retreated through Lake Norwood. Several distinct ice-contact faces oriented northwest-southeast are present in this area. Some of these ice-contact fans probably correlate with the supraglacial sediment complex facies of the Big Rice moraine in 60N-16W-15, 16, 21, 22, 23 and 60N-14W-28, 29, 30, 31, 32, 33. The environment of deposition was probably quite similar to that described in the earlier sections on processes of supraglacial sedimentation. However, the ice-contact deposits reflect a predominance of fluvial sedimentation at the immediate ice margin in a subaqueous environment, rather than supraglacial sedimentation.

The ice-contact deposits forming the Vermilion moraine along the southwest shore of Birch Lake were deposited by meltwater streams issuing from the ice margin. The exact source of these meltwater streams -supraglacial, englacial, or subglacial- is not certain. However, several eskers trend orthogonal to the Vermilion moraine immediately north of Birch Lake (Plate II), suggesting an input of subglacial meltwater to these fans. In any case, these streams would have lost energy and deposited their loads at the ice margin. Sand and gravel accumulated as outwash fans which coalesced as the meltwater streams changed course or developed multiple channels on the enlarging fan surface (Fig. 29). The ice margin was probably fairly gently sloping in this area since stagnant ice was buried by sand and gravel as evidenced by the presence of kettles in this area. These ice-contact outwash fans grade distally into outwash plains and are distinguished from them only because they were deposited in contact with glacial ice and demarcate an ice margin.

The portion of the Vermilion moraine composed of ice-contact deposits was most likely deposited as subglacial meltwaters emptied into a proglacial lake (Fig. 30), analogous to the coarse facies of turbidite successions (Rust, 1977). The type of subglacial drainage network present (Fyfe, 1990) and the depth of lake water (Rust and Romanelli, 1975; Eronen and Vesajoki, 1988; Fyfe, 1990) dictated the style of sedimentation and the morphology of resulting landform. Fluctuating lake levels may even be the mechanism which triggered moraine formation. High proglacial lake levels produced a relatively high potentiometric surface within the ice sheet. A drop in lake level increased the hydraulic gradient and subglacial drainage was initiated (Sharpe and Cowan, 1990).

The assemblage of sediments present in this section of the Vermilion moraine, bears a strong resemblance to what has been referred to as subaqueous outwash in Canada (Rust and Romanelli, 1975; Rust, 1977; Cheel and Rust, 1980). Sedimentary

A



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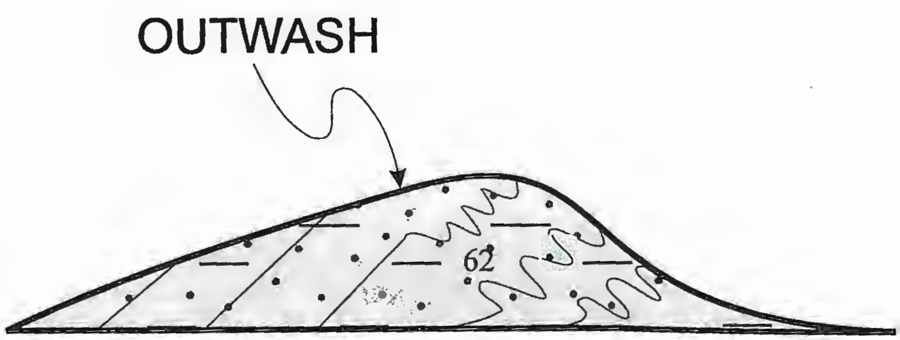

Figure 29. Depositional model of the ice-contact outwash fan portion of the Vermilion moraine

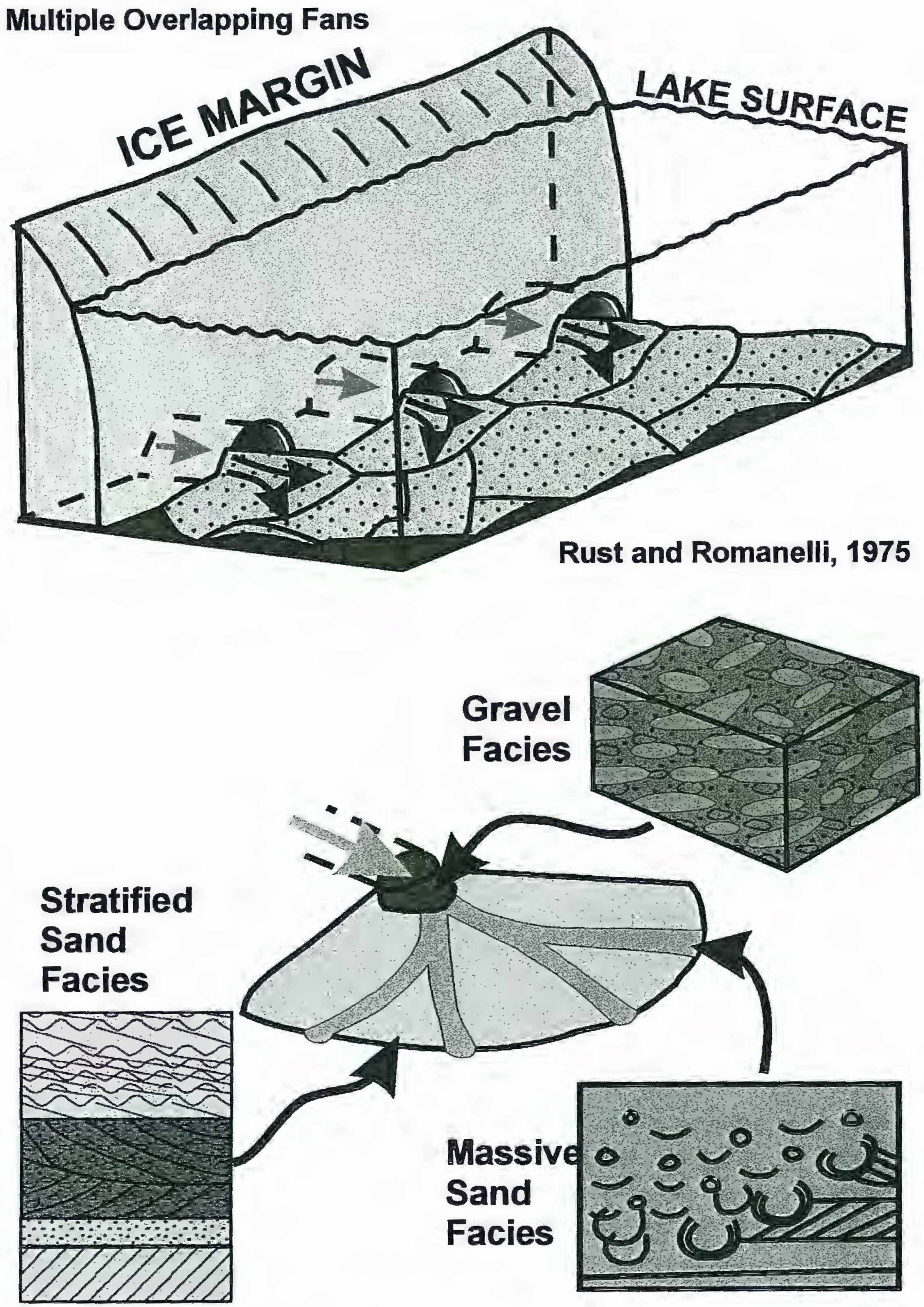

Figure 30. Depositional model of the subaqueous outwash facies of the Vermilion moraine

characteristics present which are indicative of subaqueous deposition include narrow, steep-sided channels oriented generally perpendicular to the moraine and filled with sediment which is, in most cases, coarser than the surrounding sediment (Rust and Romanelli, 1975, Rust, 1977). This type of channel is unlike those present in the subaerial outwash environment, where channels are much shallower relative to their width (Rust, 1977). These channels in the Vermilion moraine probably formed by plug-flow of highly concentrated sediment exiting the ice-margin (Rust, 1977; Postma and others, 1983). The steep walls of the channel were maintained during flow, until rapid deposition as meltwater flow diminished. The presence of normally-graded beds of gravel (Fig. 31) further support subaqueous, rather than subaerial deposition (Sharpe and Cowan, 1990). The bedding in most places in this portion of the Vermilion moraine dips at right angles away from the moraine (Fig. 32) arguing against deposition by water flowing parallel to the crest of the ridge as in esker sedimentation. It is unclear whether this dipping bedding represents delta foresets or large-scale bedforms (Sharpe and Cowan, 1990).

Deposition in contact with glacier ice is indicated by the presence of flow tills locally interbedded with (Rust and Romanelli, 1975) and overlying (Thomas, 1984; Sharpe and Cowan, 1990) the sand and gravel. Furthermore, the presence of large solitary boulders within sorted sand suggests deposition by mass-movement from the glacier (Cheel, and Rust, 1980; Thomas, 1984; Sharpe and Cowan, 1990), possibly from an overhanging ice margin (Thomas, 1984).

Well-sorted fine sand and silty sand occur interbedded with the coarser sediments

Figure 31. Normally-graded bed of gravel in Vermilion moraine. View is towards the north, generally parallel to the moraine. (62N-16W-22BCDD)

Figure 32. Large-scale dipping beds of pebbly sand in Vermilion moraine. View is towards the east, generally perpendicular to the moraine. (62N-16W-22BCDD)

of the Vermilion moraine. In the subaqueous environment, these sediments would have contained a great deal of water as they were deposited from suspension. Deformation structures present in the fine sand and silty sand units resemble ball and pillow structures (Rust, 1977; Sharpe and Cowan, 1990) and formed by dewatering in response to vertical loading as the overlying sediments were deposited (Rust, 1977).

The source of meltwater for the ice-contact deposits of this portion of the Vermilion moraine was subglacial (Fig 30). The area immediately behind the Vermilion moraine is characterized by a generally thin cover of glacial drift and numerous bedrock outcrops (Morey, 1981), which may reflect scour by subglacial meltwaters (Sharpe and Cowan, 1990). In the early stages of moraine formation, subglacial meltwater flow may have been through a linked cavity system, or closely-spaced network of minor conduits to explain the continuous nature of the ridge (Rust and Romanelli, 1975; Fyfe, 1990). During later stages of moraine formation, drainage was most likely through subglacial conduits, which is represented by the eskers trending perpendicular to the moraine immediately behind it. The only reason two stages of subglacial drainage is proposed is that subglacial drainage through major conduits usually produces a series of individual ice-marginal deltas rather than a continuous ridge (Thomas, 1984; Fyfe, 1990). An alternative explanation is that there was only one stage of subglacial drainage through a closely-spaced network of minor conduits and that the number of eskers mapped up-ice (Plate II) under represents the number of conduits supplying meltwater to the Vermilion moraine. In fact, much of the area immediately up-ice of the moraine is currently occupied by lakes, and there may be esker remnants within these lake basins which have not been recognized.

The depth of water in front of the moraine can be inferred by looking at the flat-topped segments of the moraine. The area immediately in front of the Vermilion moraine across St. Louis County is approximately 100 feet below the flat-topped portions of the moraine. The flat-topped segments of the moraine are interpreted to have formed as overlapping subaqueous fans (Thomas, 1984; Sharpe and Cowan, 1990) were built up to the water level (Fyfe, 1990) (Fig. 30). Therefore the lake bordering the Vermilion moraine was approximately 100 feet deep near the ice margin. Sharpe and Cowan (1990) suggest that a rapid drop in the level of the proglacial lake may initiate moraine formation. In their model, the rapid drop in lake level triggers catastrophic sheet-flow drainage, the bedrock and subglacial debris are scoured and deposited rapidly as the water enters the lake. If their model is correct, then the Vermilion moraine does not represent a period of ice-marginal stability as do other moraines, but rather marks the opening of a lower outlet of the ice-marginal lake (Sharpe and Cowan, 1990).

The higher, wider and hummocky segments of this portion of the Vermilion moraine were built above lake level by mass-wasting of supraglacial debris into the proglacial zone.

Summary of the Nature of the Vermilion Moraine

The term "Vermilion moraine" has previously been used in reference to a series of Rainy lobe end moraines and related deposits in the vicinity of the Mesabi and Vermilion iron ranges (Hobbs and Goebel, 1982). It is proposed that the usage of "Vermilion moraine" be restricted to the single Rainy lobe ice-marginal feature that extends from the Rainy-Superior interlobate junction near Isabella, Minnesota northwestward to the south shore of Nett Lake (Fig. 27). The easternmost portions of the Vermilion moraine are described and defined by Friedman (1981) and Stark (1977) and Hobbs (Hobbs and others, 1988). The portion of the Vermilion moraine from the vicinity of the Dunka mine (60N-12W-10BCAD) to the south shore of Nett Lake is summarized below.

Along its 70 mile course across northern St. Louis County, the Vermilion moraine is comprised of four distinct morphologic units, each with a characteristic lithofacies, a reflection of the variety of processes which contributed to its formation. From east to west, these units are interpreted to be: 1) a push moraine composed of subglacial till; 2) a series of coalesced ice-contact outwash fans; 3) dead-ice moraine composed of supraglacial sediments; and 4) a delta moraine composed of subaqueous outwash (Fig. 27).

The area mapped as jökulhlaup deposits has previously been included as part of the Vermilion moraine (Hobbs and Goebel, 1982), but is not considered part of the Vermilion moraine as defined in this report. These coarse proximal outwash (Eyles, 1979) deposits associated with the Vermilion moraine are discussed in a later section of this report.

These interpretations of the sedimentary environment of the Vermilion moraine allow some inferences to be made about the margin of the Rainy lobe while it stood at the Vermilion moraine. The push moraine segment was deposited by active ice which appears to be restricted to those areas in the vicinity of the east end of the Giants Range

(Plate I). This area of active ice may reflect an influence of the thicker, therefore more active Superior lobe ice approximately 15 miles to the east. The Giants Range may have formed a barrier to the westward propagation of this influence since it appears as though the Vermilion moraine only a few miles northwest of the end of the Giants Range was deposited from nearly stagnant ice.

The potentiometric surface immediately north of the ice-contact outwash fans must have been relatively low in order for subglacial drainage to have been concentrated there. Examination of well logs from this area suggests a buried valley trending northeast in this area (Plate II); therefore, this potentiometric low may have been bedrock controlled. Interestingly, the close association of push moraines and subaerial outwash fans has been reported from the margins of modern glaciers (Boulton, 1986).

Stagnant ice within and distal to the Vermilion moraine is suggested along the next segment of the moraine. The dead-ice moraine was, of course deposited on stagnant ice, and the jökulhlaup fan must also have been deposited on stagnant ice (discussed below in more detail).

The gap in the moraine east of Tower may be due to a series of high bedrock hills in this area. The irregular bedrock topography locally may have prevented uniform deposition along the ice margin (Persson, 1983).

The ice margin in the area fronted by subaqueous outwash also would have been a hydraulic low as well as a surficial topographic low, as suggested by the presence of subglacial streams entering a proglacial lake. Indeed, the bedrock along the Vermilion moraine from just northwest of Babbitt to Tower, consists of Archean supracrustal rocks which are topographically higher than the plutonic rocks on either side. In summary, the Rainy lobe in St. Louis County at the time of deposition of the Vermilion moraine must have been fairly thin and nearly stagnant, at least in those areas west of the eastern terminus of the Giants Range.

OUTWASH SAND AND GRAVEL

This map unit occurs primarily in the central and north-central portions of the map area, but also occurs in the Babbitt area and in the Embarrass gap. Exposures near roads are quite numerous due to mining for aggregate; however, large parts of this unit lack roads and are not easily accessible.

Sediment Description

This map unit is composed of poorly-sorted sandy pebble-gravel and pebbly sand which grades laterally and vertically into moderately well sorted very-coarse to fine sand. All of the samples analyzed plot in the sand field of Shepard's (1954) classification (Fig. 33). However, the overall textures range from sand to sandy gravel (Lawson, 1979) (Fig. 34) (Appendix A).

The outwash sand and gravel is generally brown in color, ranging from light brownish gray (10YR 6/2) to brown (10YR 5/3) and dark brown (10YR 4/3), reflecting the oxidized nature of these permeable sediments. In some instances, a particular lithology will dominate in an exposure, lending the color of the source rock to the sand and gravel. For example, the pebbly sand at 60N-14W-10BBBA is composed primarily

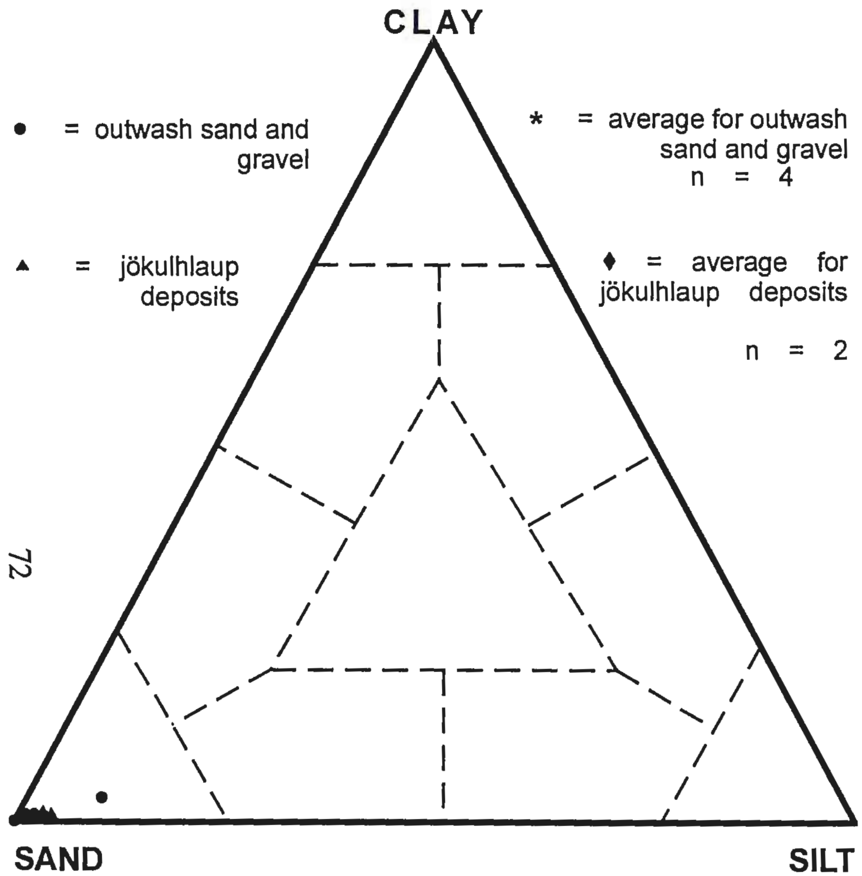

Figure 33. Texture of outwash and jökulhlaup deposits

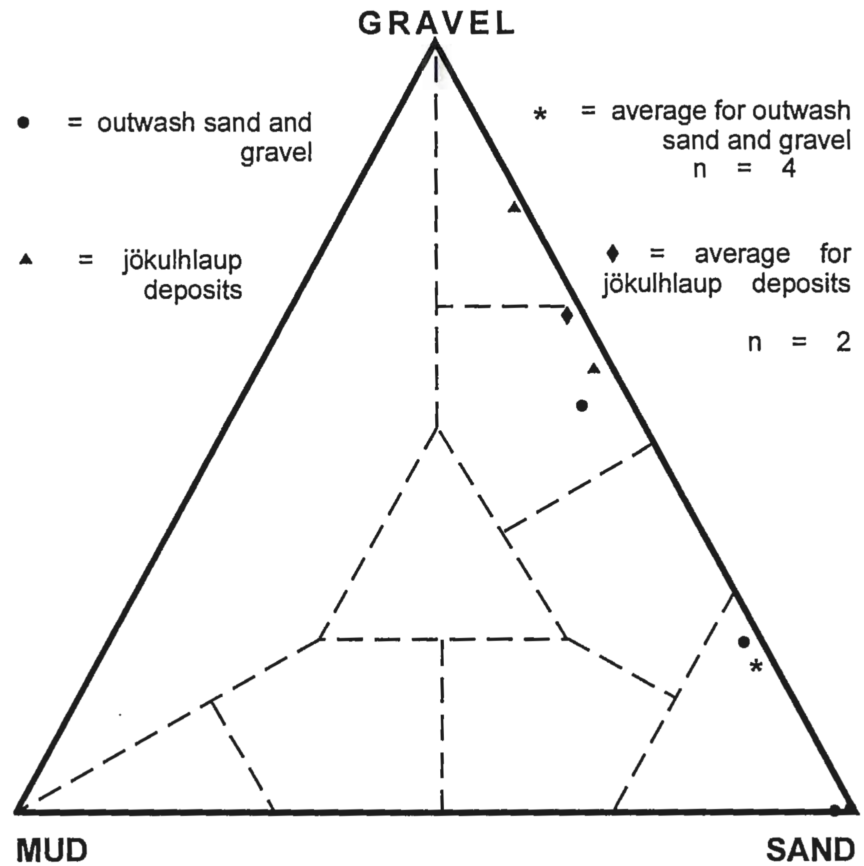

Figure 34. Overall texture of outwash and jökulhlaup deposits

of granitic clasts, which imparts a pink color to the sediment.

The pebble, granule and very-coarse sand fractions of the outwash sand and gravel in most of the study area (map unit 4b) are composed of subangular to rounded clasts of granite, gneiss, schist and other lithologies common to the Vermilion district. However, a portion of the outwash plain in the Babbitt area (map unit 6) contains subangular to rounded clasts of reddish felsite, granophyre, amygdaloidal basalt and other lithologies derived from the Lake Superior basin. Outwash sand and gravel characteristically exhibits horizontal stratification, and is less commonly cross-bedded. A vertical sequence of glaciofluvial sand and gravel often exhibits sharp transitions of particle size from one bed to another.

Geomorphology

Outwash sand and gravel occurs in the northern and eastern portions of the study area and is expressed as low-relief plains which slope gently from end moraines toward lacustrine plains. This type of geomorphic feature is referred to as an outwash plain or Icelandic term sandur. In some places (61N-13W-29, 30 and 31 and 61N-14W-29 and 30 for example) the outwash plains are pitted.

Sedimentary Environment

Glacial meltwater streams have multiple channels with low sinuosity and are considered to be braided streams (Miall, 1983). Deposition in the braided stream environment is characterized vertical and lateral accretion of longitudinal sand or gravel bars (Miall, 1983). Deposition of a clast on a bar occurs as the transporting capacity of the meltwater stream is diminished, either by a decrease in the velocity of the stream or by an increase in the sediment load of the stream (Sugden and John, 1976). Discharges of glaciofluvial streams are highly variable both seasonally and diurnally as ablation rates change in response to fluctuations in incident solar radiation (Church and Gilbert, 1975). The sediment load of glaciofluvial streams changes with changing ablation rates, but is also affected by periodic mass wasting of preexisting sediment into the meltwater stream (Church, 1972). These changes in stream velocity and sediment load are responsible for the rapid lateral and vertical variation in grain size that is present within the outwash sand and gravel map unit.

JÖKULHLAUP DEPOSITS

This unit occurs only in the north-central portion of the map area, but extends north to the Vermilion moraine. Numerous small exposures were present during the 1985 field season due to logging road construction. In July, 1992, while preparing for the Midwest Friends of the Pleistocene Field Trip, a backhoe was used at one location to create an exposure approximately 7 feet high.

Sediment Description

The northern portions of this unit, immediately north of the map area consist of poorly-sorted boulder-gravel which is primarily matrix supported, but was observed to be clast supported at some localities (Fig. 35). This coarse facies grades into poorly-sorted

Figure 35. Exposure of jökulhlaup deposits (61N-14W-26BBDC)

bouldery cobble- and pebble-gravel in the southern portions of this unit in the map area. This relatively finer grained facies of jökulhlaup deposits grades southerly into outwash plains and ultimately lacustrine plains. The two samples collected for textural analysis range from sandy gravel to gravel (Lawson, 1979) (Fig. 34).

An exercise was undertaken in an attempt to quantify this apparent southward fining of the jökulhlaup deposits. At every available exposure in this map unit, the diameters of 50 of the largest clasts present were measured. When the means of these diameter values were plotted on a map (Fig. 36), a weak trend became apparent. The highest values of 16 and 17 inches are present near the south shore of Bear Head Lake and the lowest values of 7, 8 and 9 inches are present near the southern margin of the fan. Intermediate values of 11, 12 and 13 inches are present in the central portions of the map unit. Although not conclusive, the data suggest a general trend of diminishing maximum particle-size toward the south. The maximum clast-size present at each locality was also plotted on a map resulting in no apparent trend. The individual clasts present in this sediment are subangular to well rounded, mainly granitic lithologies of local origin. The colors of these sediments are shades of pink, reflecting the pink colors of the constituent granitic clasts. The gravels present in this map unit are generally massive and appear to be structureless.

Geomorphology

The jökulhlaup deposits have an overall fan-shaped geometry, with a smaller fan superimposed upon a larger fan. The apex of the larger fan is located immediately north

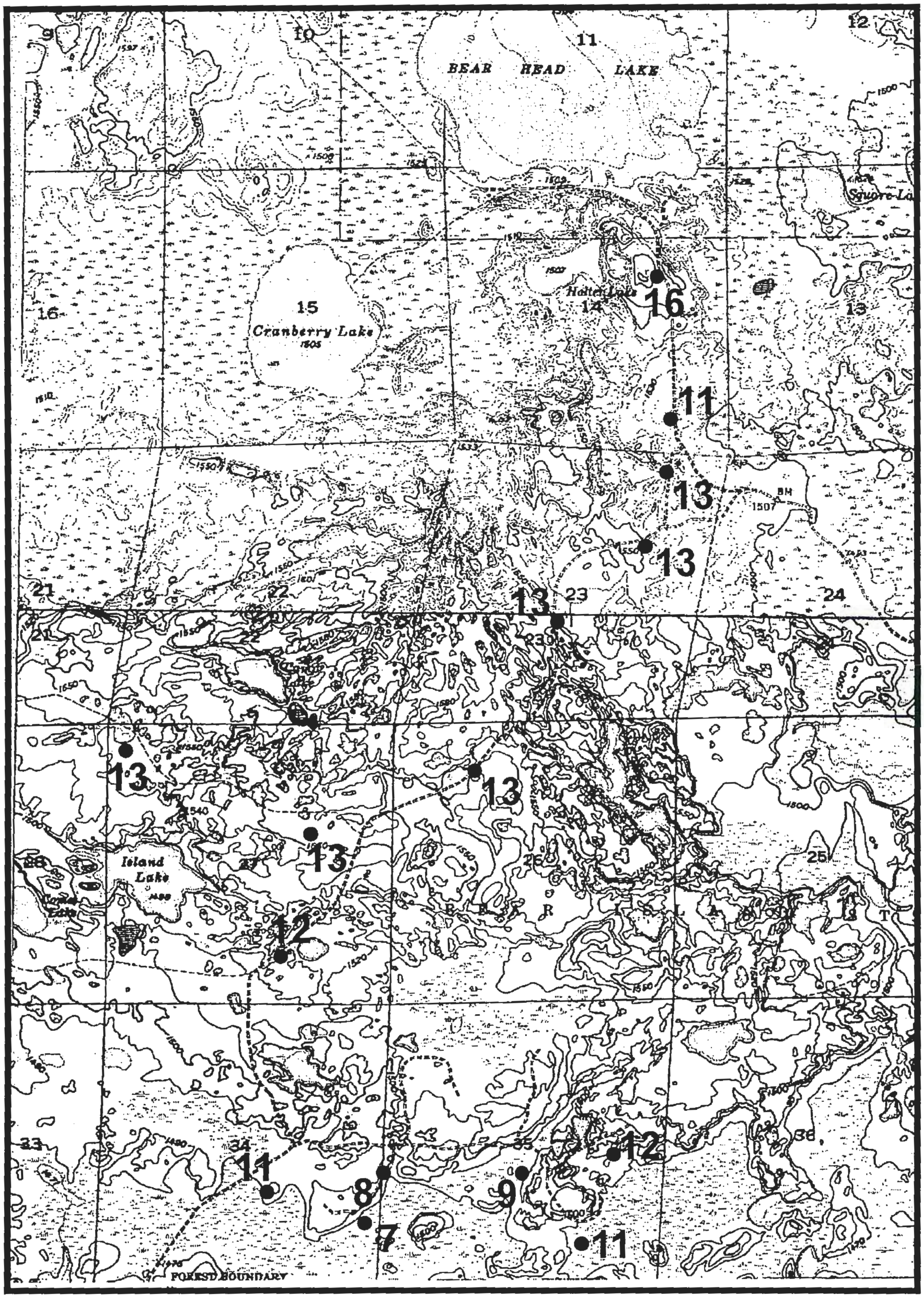

Figure 36. Jokulhlaup fan. Numbers refer to the mean diameter (inches) of the 50 largest clasts present at each locality (Portions of the Eagles Nest and Isaac Lake quadrangles)

of the map area on the south shore of Bear Head Lake (61N-14W-14) (see aerial photograph on Plate II), while the apex of the smaller fan, and overall highest part of the fan, is located just southwest at 61N-14W-22A and 61N-14W-23B. The overall fan-shaped nature of this landform is accentuated by the presence of linear ridges and kettles which radiate from the highest portion of the fan producing a very rugged topography. The relief between adjacent kettles and ridges exceeds 70 feet at some localities (i.e. 61N-14W-22BC). There is a consistent low-angle gradient from elevations of greater than 1610 feet above sea level at the highest portion of the fan to elevations of less than 1500 feet above sea level near its distal margins.

Sedimentary Environment

Jökulhlaup is "an Icelandic term for glacier outburst flood" (Bates and Jackson, 1980, p. 334), so when applying the term jökulhlaup to a sediment type, this depositional environment is inferred. Deposition of the jökulhlaup deposits of the study area, is discussed below.

The apex of the jökulhlaup fan located on the south shore of Bear Head Lake points towards a slight reentrant in the dead-ice moraine facies of the Vermilion moraine (Plate II). Subglacial meltwater became ponded behind a dam created by debris or ice (a frozen margin) or both. The jökulhlaup was initiated when the forces exerted by ponded water exceeded the strength of the dam (Clague, 1987). The ice/debris dam may have become buoyant due to hydrostatic pressure, contributing to the initiation of the jökulhlaup (Nye, 1976; Clague, 1987). Flow probably began as a trickle, but thermal erosion by the flowing water would have rapidly enlarged the outlet (Maag, 1969; Clague, 1987). Discharge rapidly increased to a maximum within a few days (Clague, 1987), during which time the flood waters were scouring the debris-rich proglacial stagnant ice into a series of long, linear channels, separated by clean-ice interfluves. At maximum discharge, the flood waters were probably transporting large boulder-sized clasts as well as large blocks of ice as bed load. A rather spectacular example of a modern jökulhlaup is given by Thorarinsson (1953) from Grímsvötn, Vatnajökull, Iceland where discharges at the peak of a jökulhlaup were 40,000 to 50,000 cubic meters per second and ice-blocks as large as three-story houses were being transported by the flood waters.

Further evidence of the high hydraulic head of the meltwaters is that the highest portion of the fan is currently approximately 100 feet higher than the small outwash plain to the north, between Bear Head Lake and the Vermilion moraine, and the apex of the fan was undoubtedly higher while stagnant ice existed in the proglacial zone. This requires that the subglacial meltwaters were exiting the ice margin as a spectacular fountain. In fact, this may explain the modern bathymetry (Gilbert, 1990) of Bear Head Lake, which is deepest in the north, nearest the source of the jökulhlaup.

Discharges of jökulhlaups rapidly decline as the reservoir drains (Whalley, 1971; Clague, 1987) as illustrated in the hydrograph of a modern jökulhlaup (Fig. 37). This rapid decrease in discharge causes nearly instantaneous deposition of the larger clasts transported by the flood waters (Maizels, 1989). The majority of the finer sediment load (sand, silt and clay) would have been transported beyond the distal margin of the jökulhlaup fan and deposited either on the outwash plains flanking the fan, or in Lake

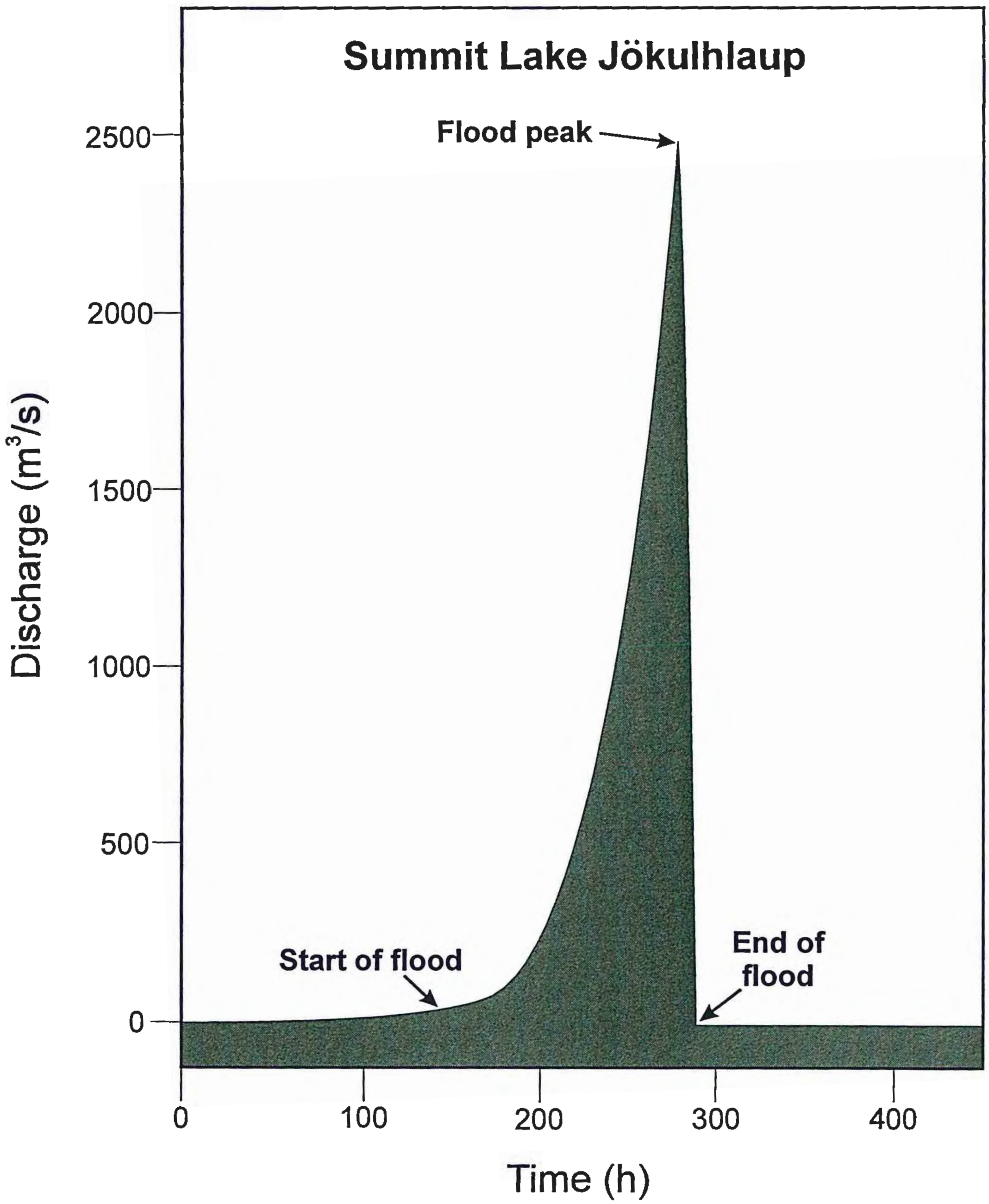

Figure 37. Hydrograph of a modern jökulhlaup (Redrawn from Clague, 1987)

Norwood to the south. This rapid deposition is evidenced by the poor sorting and lack of sedimentary structures (Maizels, 1989) in the jökulhlaup deposits in the Embarrass area.

The bedrock source of clasts in the jökulhlaup deposits is probably very local. This is evidenced by the predominance of granite clasts (Fig. 35) and the fact that granite is the bedrock lithology mapped in the area immediately up-flow from the jökulhlaup fan (Sims and others, 1970). The granitic clasts were probably eroded from the substrate primarily by glacial erosion from local sources and easily released from the ice as subglacial meltwater eroded basal ice. Bedload transportation of pebble-size and larger clasts by the flood waters would have been efficient at rounding even though the transport distance is fairly short (Elfström, 1987).

The very hummocky topography of the jökulhlaup fan suggests that the sediments were deposited on stagnant ice which subsequently collapsed as the underlying ice melted. Evidence for the presence of stagnant ice is also suggested by the occurrence of low-lying, swampy areas containing lakes in the areas surrounding the jökulhlaup fan. Some of the kettles may have resulted from meltout of large blocks of ice which were transported and buried by the jökulhlaup. The resulting topography of a rugged fan-shaped landform with radiating kettles and ridges is the result of topographic inversion, with the ridges representing former supraglacial channels and the kettles representing the clean-ice interfluves. Two separate jökulhlaups from the Vermilion moraine may have contributed to the construction of the jökulhlaup fan in the study area, as evidenced by the presence of two overlapping fans.

The jökulhlaup-deposited fan in the study area appears very similar to the tunnel-valley fans of the St. Croix moraine in east-central Minnesota (Patterson, 1994). However, one notable difference is the lack of well-developed tunnel valleys up-ice from the jökulhlaup fan in the Embarrass area. This is probably due to the fact that the tunnel valleys in east-central Minnesota are developed in a sedimentary bedrock terrane and that comparable volumes of subglacial drainage in an area underlain by more resistant crystalline bedrock would not yield well-developed tunnel valleys.

LACUSTRINE SEDIMENT

This unit is widespread in the topographically lowest portions of the study area in the areas now traversed by the Embarrass and Pike Rivers. Holocene peat deposits overlie much of this unit, especially in the lowest areas. Therefore, the exposures studied are probably biased towards near-shore facies of the lacustrine basin.

Sediment Description

The sediments which comprise this map unit range from moderately well-sorted pebbly, coarse sand in near-shore facies to well-sorted very-fine sandy silt in off-shore facies (Appendix A). Most commonly they are well-sorted medium to fine sand (Fig. 38). A total of 13 samples were subjected to particle size analysis. Ten samples plot in the sand field, one in the silty sand field, and two in the sandy silt field, with the average plotting in the sand field (Shepard, 1954) (Fig. 39). Most of these 13 samples contained no gravel. Ten samples plot in the sand field of Lawson's (1979) trilinear classification, while one sample plots in each of the following fields: silty sand, sandy silt, and silt (Fig.

Figure 38. Exposure of lacustrine sand (60N-15W-14CC)

40). The overall average texture of lacustrine sediments is sand (Fig. 39, Fig. 40).

The color of the lacustrine sediments ranges from pale brown (10YR 6/3) to dark brown (10YR 4/3) and light yellowish brown (10YR 6/4), yellowish brown (10YR 5/4) to dark yellowish brown (10 4/4), also gray (10YR 5/1), light gray (10YR 7/2) and light brownish gray (10YR 6/2) (Appendix A). The finer lacustrine sediment typically exhibits gray and brownish gray colors, while the coarser sands are shades of brown.

Sedimentary structures are generally poorly developed in this unit with most exposures exhibiting massive bedding to weakly stratified horizontal bedding. At some localities, horizontal laminae of dark mineral concentrations were observed. Cross-bedding was observed at only one exposure where the lacustrine basin becomes restricted (60N-16W-26CDCD) and currents were probably strong. Ripple-drift cross-laminations (Jopling and Walker, 1968) were observed at one exposure (60N-15W-29CAAC).

Also included in this unit are boulder-lag accumulations which are widespread at the margins of this map unit near the 1450 foot contour line in the eastern portions of the map area (Fig. 41). At one locality, a boulder-lag accumulation was noted near 1430 feet in elevation (60N-15W-23CAA and CAB). Boulder-lag accumulations are conspicuously absent from the western portions of the map area, except in the immediate vicinity of the Embarrass gap.

Geomorphology

Lacustrine sediments occur as low-relief plains in the topographically lowest portions of the study area, generally less than 1450 feet in elevation (Plate I). This low

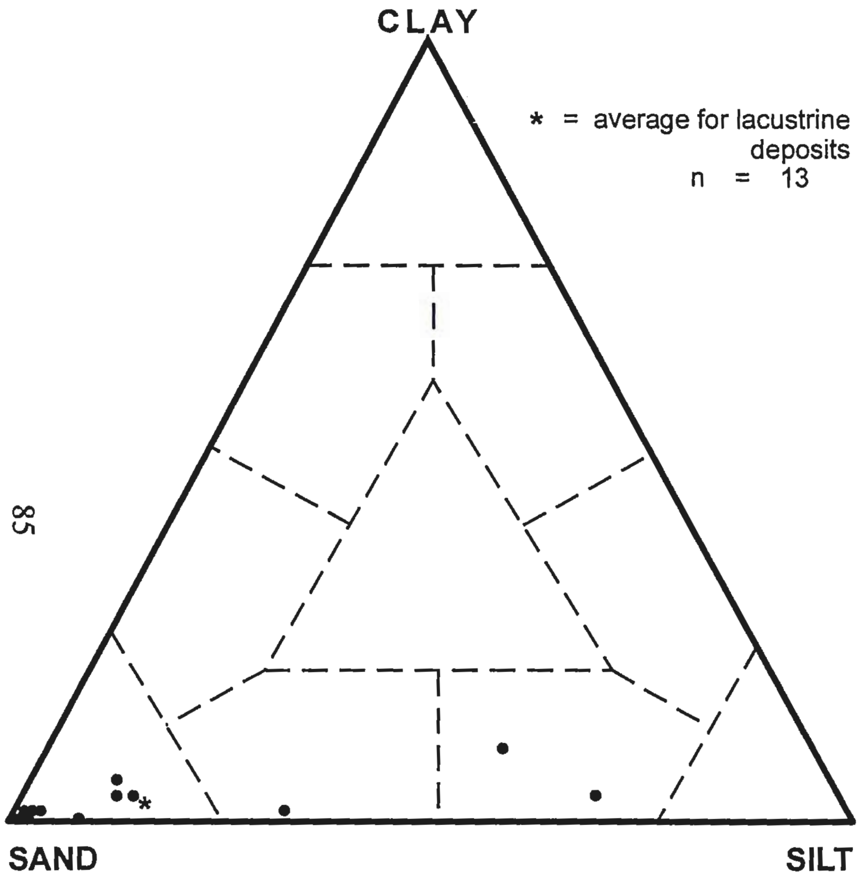

Figure 39. Texture of lacustrine sediment

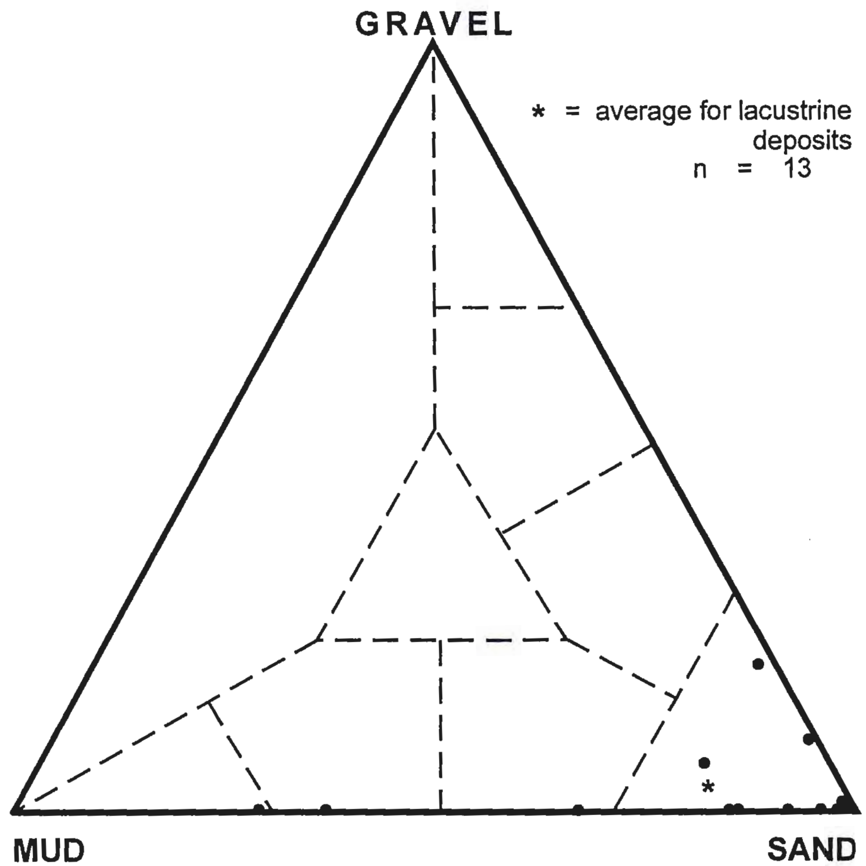

Figure 40. Overall texture of lacustrine sediment

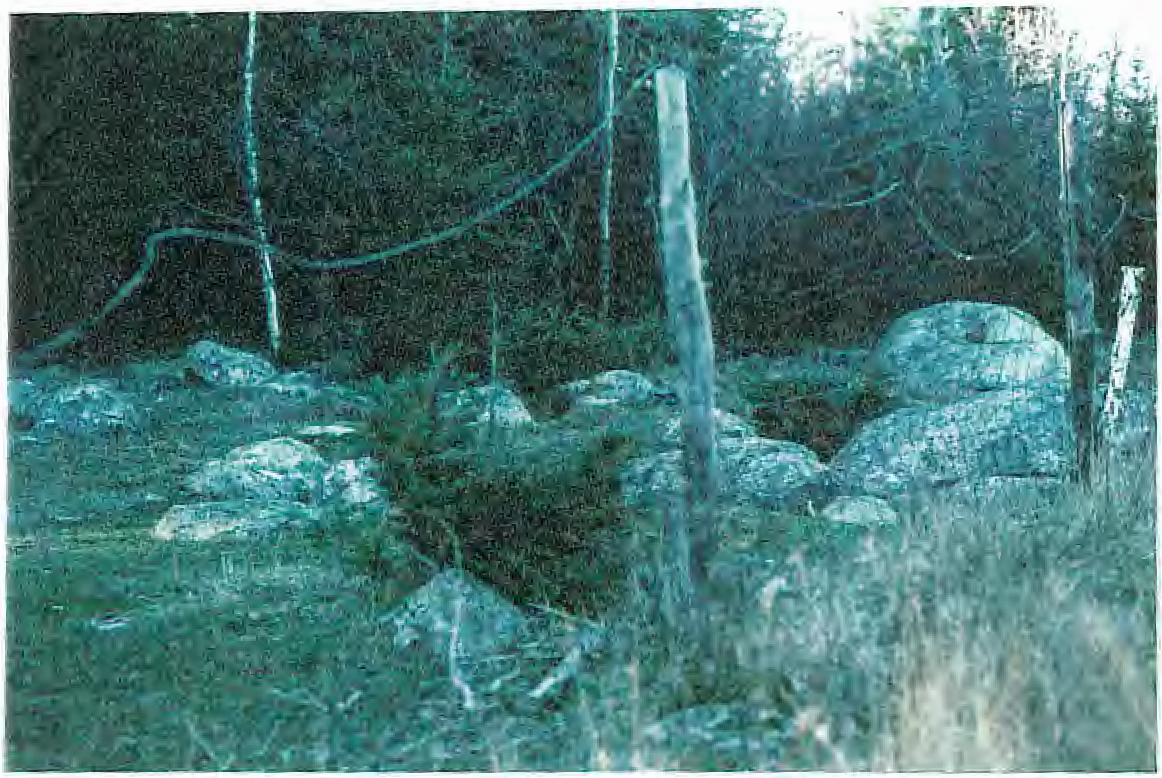

Figure 41. Lake Norwood strandline at 1450-foot level (60N-15W-01CAAD)

area generally slopes towards the low-gradient streams and bogs that occupy the axes of the lacustrine basin.

Sedimentary Environment

The sediments of this map unit were deposited in a proglacial lake which formed as the margin of the Rainy lobe retreated into the lowlands north of the Giants Range. As mentioned above, the Laurentian divide coincides with the crest of the Giants Range east of the Embarrass gap. Throughout the deglaciation of the region, drainage through the Embarrass gap was impeded by the presence of stagnant ice which altered the course of the Laurentian divide, causing lakes to form in the area currently south of the Laurentian divide.

Lake Norwood was probably in existence while the ice margin stood at the Big Rice moraine and was certainly in existence as the ice margin stood at the Wahlsten moraine as evidenced by the well developed strandline features at the 1450-foot contour line in this area. This strandline is interpreted as having formed by wave erosion of preexisting glacial sediments. The western portions of Lake Norwood must have been at least partially ice-walled and contained abundant stagnant ice as suggested by the lack of recognizable strandline features and the presence of numerous stagnant ice landforms.

The main outlet of Lake Norwood in the map area was through the Embarrass gap (Winchell, 1901; Hobbs, 1983). There are terraces present in the Embarrass gap near the 1450 foot contour line which in some locations, are boulder-armored. Stagnant ice was present in the Embarrass gap at this time, as evidenced by the presence of Sabin Lake and

Wynne Lake, and would have provided a plug which controlled the level of the lake. It is likely that at the time Lake Norwood was in existence, Lake Upham was present immediately to the south, and the Embarrass gap was not actually an outlet but a constriction in an otherwise continuous lacustrine basin. In either case, the water which flowed through the Embarrass gap would have followed the modern St Louis River drainage eventually around the Superior lobe and into the St. Croix River.

No well developed deltas were recognized either in the field or by examination of topographic maps or aerial photographs. Sediment supply was probably delivered to the basin by several small meltwater streams crossing the outwash plains to the north. A deltaic facies consisting of ripple-drift cross-laminated sand and silty fine sand was noted at 60N-15W-29CAAD. These sedimentary structures indicate a rapid decrease in the velocity of the sediment-laden water as would happen in a deltaic environment (Jopling and Walker, 1968; Gustavson and others, 1975), and record the location where a minor meltwater stream entered the lake. Masses of debris-rich ice present in, and around the basin also contributed sediment to the lacustrine basin as the buried ice melted.

The sediments of Lake Norwood are relatively coarse-grained when compared to typical proglacial lacustrine deposits. This could be attributed to one of, or a combination of two conditions present. The Lake Norwood basin is generally less than 50 feet deep, which may have been too shallow to allow the clay fraction to settle from suspension if there were appreciable currents in the lake. Secondly, the crystalline bedrock source for all of the glacial sediments in the study area yields relatively little sediment in the silt and clay size fractions.

Terraces in the Embarrass gap at 1430 and 1400 feet, as well as other geomorphic evidence from areas west of the study area, suggest glacial lakes were in existence at 1430 and 1400 feet within the study area (Hobbs, 1983). These lakes have been referred to as Lake Koochiching and are considered early stages of Lake Agassiz (Nikiforoff, 1947; Hobbs, 1983). The 1430-foot stage of Lake Koochiching would have deposited sediment with a northwestern provenance in the lower portions of the Lake Norwood basin. Although the lacustrine sediments in the lower portions of the basin are finer grained, no indicators of northwestern provenance, such as calcareous sediment, were found. However, this may be due to the lack of good exposures in these areas. The 1400-foot stage of Lake Koochiching would only have occupied the lowest portions of the Pike River valley in the northwestern part of the map area but must have extended into the low areas west of the map area. Lacustrine clay is reportedly present in the valley of the Pike River (U of M Ag. Exp. Sta., 1971). It is postulated this stage of the lake drained through the Embarrass gap crossing the present day Laurentian divide at 60N-15W-30 (Plate I) (Leverett, 1932; Hobbs, 1983). The sequence of glacial lake drainage is discussed in more detail in the section on glacial history.

EOLIAN SEDIMENT

"Eolian sediment" is not a map unit on the Pleistocene geologic map of the Embarrass area because it does not comprise any recognizable landforms and is generally quite thin throughout the study area. Eolian sediment occurs in scattered exposures throughout the study area and ranges from 12 to 24 inches in thickness (Fig. 42). Eolian

Figure 42. Exposure of eolian sediment. Upper 18 inches is artificial fill. Eolian sediment is 24 inches thick overlying outwash sand and gravel.

sediment was observed to overlie all map units except Superior lobe outwash. However, this apparent lack of eolian sediment may be related to the limited number (3) of exposures of Superior lobe outwash within the map area.

Sediment Description

The texture of the eolian sediment ranges from fine sand to silt, but is most commonly silty, fine sand or fine-sandy silt (Appendix A). Total silt content ranges from 21 percent to 68 percent and most of the sand fraction is fine and very-fine (Appendix A). When plotted on the trilinear diagrams, the results show considerable scatter, primarily within the silty sand and sandy silt fields (Shepard, 1954) (Fig. 44). All of the samples contained a variable amount of medium, coarse, and very-coarse sand, in addition to small amounts of gravel in over half of the samples (Appendix A; Fig. 43). No sedimentary structures were observed in the eolian sediment.

Colors of the eolian sediment range from very pale brown (10YR 7/4) to pale brown (10YR 6/3) and brown (10YR 5/3) and light yellowish brown (10YR 6/4) to yellowish brown (10YR 5/4) (Appendix A).

Sedimentary Environment

Loess is a sediment transported and deposited by wind that consists of predominantly silt (60 to 80 percent) (Ashley, 1985). Since the windblown sediment in the study area contains appreciable amounts of fine and very-fine sand, the more general term eolian sediment is applied.

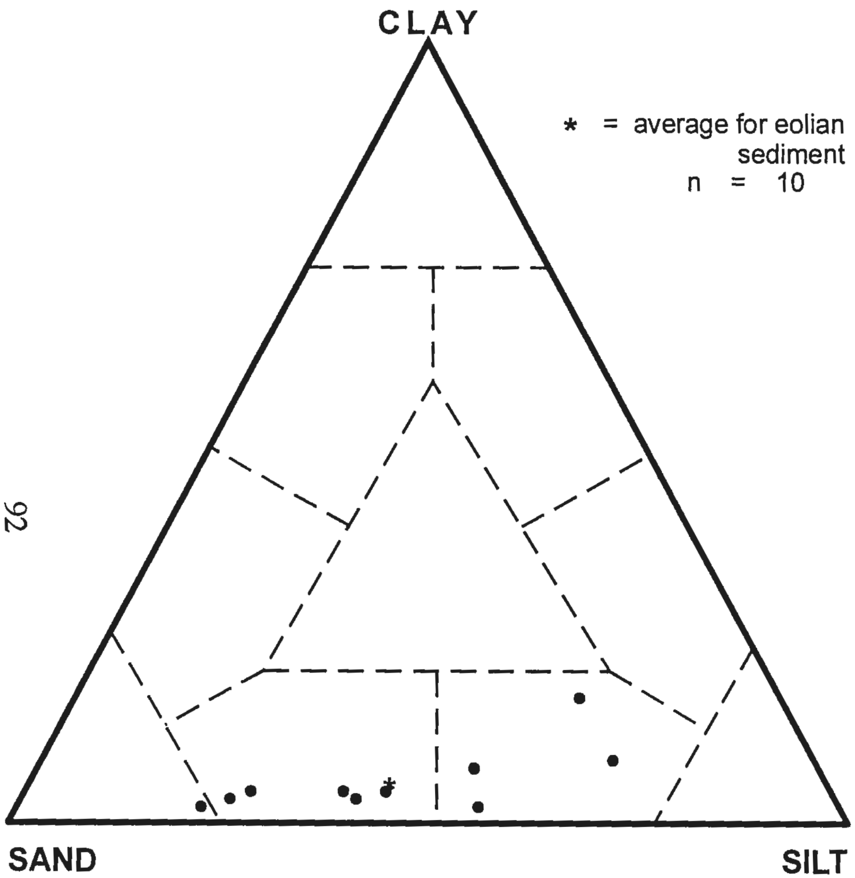

Figure 43. Texture of eolian sediment

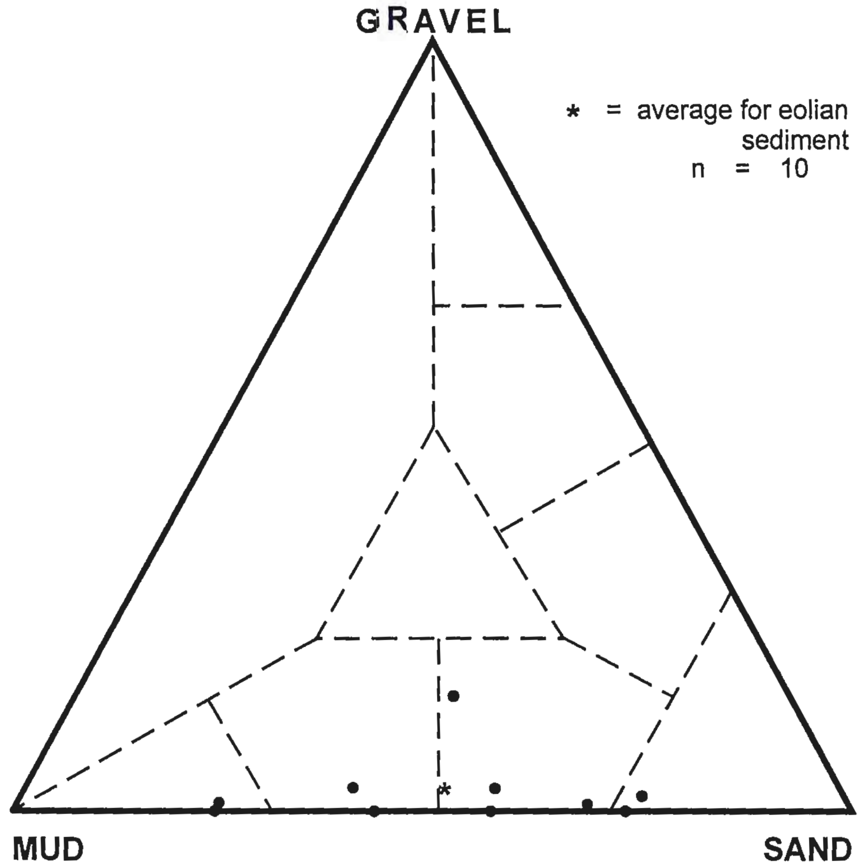

Figure 44. Overall texture of eolian sediment

The primary source of eolian sediment was probably the unvegetated, or sparsely vegetated outwash plains (Ashley, 1985). Other sources may have been lacustrine plains at times of low water (Ashley, 1985) and dry supraglacial debris (Boulton, 1972).

Examination of the particle size data for the loess (Appendix A) shows that some samples contain medium, coarse and very coarse sand and even granule size clasts in addition to the fine sand and silt. This, of course, is not common in eolian sediment, however, in all instances the coarser material can be traced to the underlying sediment. Larger clasts (coarse sand and gravel) can be elevated into the overlying fine-grained eolian sediment by a combination of bioturbation, cryoturbation, and tree-throw. The periglacial climate which was present in northern Minnesota at the end of the Pleistocene would have been effective at elevating the larger clasts through frost action. Fine-grained silt-rich sediments are particularly susceptible to cryoturbation (Eyles and Paul, 1983). Eolian sediment similar to this has been reported in areas east (Grigal, 1968; Friedman, 1981; Fenelon, 1986) and north (Zoltai, 1961) of the study area. The brown silty till reported by Winter (1971; Winter and others, 1973) to occur in thin, scattered exposures as far east as Ely may, in fact be cryoturbated eolian sediment.

PRE-LATE WISCONSINAN TILL

Sediments older than late Wisconsinan do not occur at the surface in the study area, nor are they mapped on Plate I. However, in two open-pit iron mines at the periphery of the study area, sediments interpreted to be of pre-late Wisconsinan age are exposed. Stratigraphic sections from the Dunka and Embarrass mines expose multiple tills and are described below. Assistance in describing the following stratigraphic sections was provided by Howard Hobbs during our preparation for the 1992 Midwest Friends of the Pleistocene Field Trip (Lehr and Hobbs, 1992).

Embarrass Mine Section

This section is located in the northeastern corner of the Embarrass mine (58N-15W-05BDDC) at the base of the old mine-access road that descends into the pit from the southeastern corner of the mine.

Depth (feet)	Description
0 to 16	Pebbly sand, slightly silty, slightly cobbly and bouldery, poorly to very poorly sorted. Some contorted bedding present.
16 to 66	Sandy diamicton (Fig. 45, "bouldery till"), not extremely stony, contains laminae of sorted sand, grayish brown (2.5Y 5/2), massive to faintly stratified, fairly loose consistency. This is the late Wisconsinan Rainy lobe till, the "bouldery till" of Winter (1971; Winter and others, 1973).
66 to 85	Sandy diamicton (Fig. 45, lower till), brown to dark yellowish brown (7.5YR to 10YR 4/4), leached from 66 to 70 feet, calcareous from 70 to 85 feet. Blocky structure, with individual blocks apparently cemented with iron oxide. Harder to excavate than the unit above. In general, this till is less rocky than the overlying till. Common rock types include: greenstone, granite, gneiss, Algoman iron-formation, also some limestone and graywacke.

Another section of this lower, older till occurs just west of the above-described section. The weathering profile in this section appears to be better preserved.

Height above lake (feet)	Description
24 to 20	Sandy diamicton, reddish brown (5YR 4/4), jointed, leached.
20 to 16.5	Sandy diamicton, brown to dark brown (7.5YR 4/3), jointed, with reddish material coating joint surfaces, leached.

16.5 to 10	Sandy diamicton, brown to dark brown (7.5YR 4/3), calcareous.
10 to 8	Sandy diamicton, brown to dark brown (10YR 4/3), calcareous.
8 to 0	Section mostly covered, probably same unit to water level. When this section was visited in 1985, well sorted sand was observed below the till, in places, along this portion of the section.

Winter (1971; Winter and others, 1973) also described the lower till at the Embarrass mine section where he noted that the texture, pebble lithology, clay mineralogy, and heavy mineral assemblage is very similar to the late Wisconsinan Rainy lobe till. He also measured five till fabrics in the lower till at the Embarrass mine section, which show a general northeast-southwest trend of the long axes of pebbles (Fig. 46). He interpreted the lower till at the Embarrass mine section to have been deposited by a northwestern-provenance glacier advancing northeastward south of the Giants Range, analogous to the late Wisconsinan St. Louis sublobe. However, in light of recent studies on the dispersal of Hudson Bay carbonate into Minnesota (Gowan, 1993; Mooers and Lehr, 1996; DeLong and Mooers, 1997), the simplest explanation is that the older till at the Embarrass mine section was deposited by a pre-late Wisconsinan Rainy lobe from a dispersal center northeast of the Paleozoic carbonate rocks present in the Hudson Bay lowlands.

The degree of weathering present in the lower till at the Embarrass mine section - reddish colors, development of jointing in the upper part and the depth of leaching - suggests an interglacial period of weathering, rather than interstadial. While direct chronologic control is lacking, it is likely that this weathering zone is Sangamonian. The

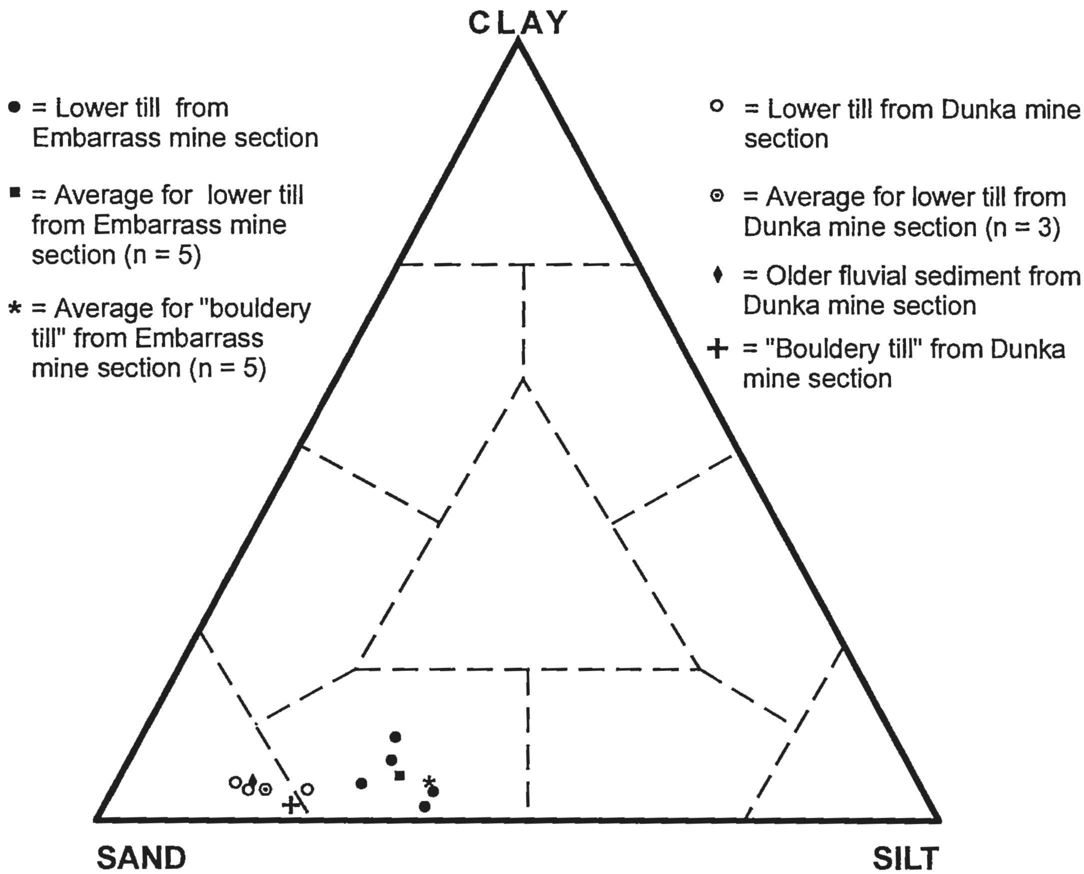

Figure 45. Matrix texture of tills from the Embarrass and Dunka mine sections

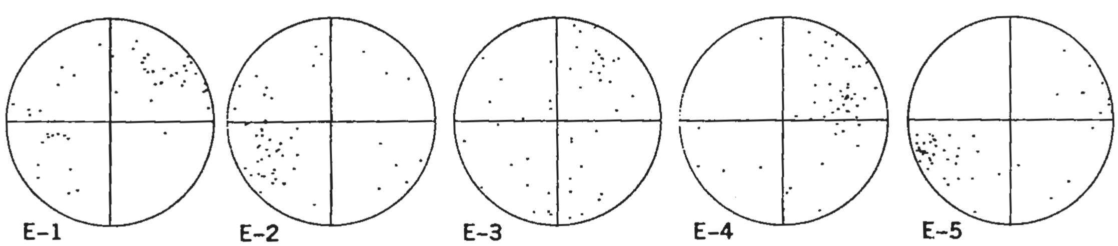

Figure 46. Fabric of "basal till" at Embarrass mine section (from Winter, 1971). Pebbles show a general northeast-southwest orientation of long axes

degree of weathering present in the lower till at the Embarrass mine section is similar to the degree of weathering in the Sangamon paleosol where exposed elsewhere in the midcontinent (Leon Follmer, personal communication, 1992). Therefore, the lower till must be at least as old as Illinoian.

Dunka Mine Section

The section described below is located in LTV's Dunka mine at approximately the midpoint along a large exposure in the southwestern portion of the pit (60N-12W-10BCBA). The section is located on the proximal slope of the Vermilion moraine. The lowest portion of the section appears to occupy a swale in the bedrock. This is the same general area from which Stark (1977) described the Dunka pit till, a dark-colored, silty, sandy, and calcareous till. This unit was not visible when this section was described (July 1991), but may be present within the lower, covered part of the section.

Interval (feet)	Description
0 to 16	Oxidized fine to medium sand containing some boulders. Bedding is subhorizontal in places, and massive in other places. Since this section is located north of the Laurentian divide, this unit is interpreted to be lacustrine sediment deposited between the moraine crest and the retreating Rainy lobe. The boulders may have been elevated into the sand from the underlying till by periglacial processes, or tree-throw.
16 to 26	Very rocky, sandy diamicton (Fig. 45; "bouldery till"), grayish brown (2.5Y 5/2). Numerous laminae of sorted sand. Few small diffuse iron stains (10YR hue). Clast types dominated by granitoid pebbles, cobbles and boulders. Also some basalt, diabase (possibly locally derived hornfels), greenstone, schist, and gneiss. No Paleozoic sandstone or carbonate clasts in this interval. Also, no Superior lobe rock types present. This unit is interpreted to be the late Wisconsinan Rainy lobe till that comprises the push moraine segment of the Vermilion moraine (Lehr and Matsch, 1987) and correlates with the "bouldery till" of Winter (1971; Winter and

others, 1973).

26 to 40	Very rocky, sandy diamicton (Fig. 45; lower till), noncalcareous, with interbedded sorted sand. The top of this interval (26 to 29 feet) is brown (10YR 5/3) with yellowish red (5YR 4/8) mottles, which are larger and more abundant at the top of this interval. A gradational color change from brown (10YR) to light olive brown (2.5Y 5/3) occurs at approximately 29 feet. Although, at the measured section this till is noncalcareous, other portions of this interval lateral to the measured section are calcareous. This unit is interpreted to be a pre-late Wisconsinan oxidized till of northeastern provenance, possibly of supraglacial origin.
40 to 44	Interbedded silty sand, silt, and pebbly sand containing occasional cobbles and boulders, calcareous, cross-bedded. Bedding is locally collapsed. Color of this interval ranges from light yellowish brown (2.5Y 6/3) to light olive brown (2.5Y 5/4). Carbonate pebbles present. This stratified sediment is texturally very similar to the tills at this section (Fig. 45). This interval is interpreted to be pre-late Wisconsinan oxidized, collapsed fluvial sediment of northeastern provenance.

The lower till at the Dunka mine section has a distinctly different texture than the lower till at the Embarrass mine section (Fig. 45). It is interesting to note that at both the Embarrass and Dunka mine sections, the older and younger tills have very similar matrix textures. The Dunka mine section is located northeast of the Giants Range, while the Embarrass mine section is located on the leeward side. This suggests that local conditions (possibly the Giants Range) were dictating the texture of till deposited.

The degree of weathering present in the lower till at the Dunka mine section is not as intense as in the lower till at the Embarrass mine section. The lesser degree of weathering in this till may have occurred under interstadial climatic conditions. Therefore, the age of the lower till at the Dunka mine section may be early Wisconsinan.

The presence of Paleozoic carbonate clasts in the lower till at the Dunka mine is in accordance with other reports of older Rainy lobe tills (Martin and others; 1988; 1989; 1991), so it is interpreted to be a pre-late Wisconsinan (early Wisconsinan?) Rainy lobe till, with the carbonate fraction derived from the Hudson Bay lowland.

BEDROCK-DRIFT COMPLEX

This unit is mapped where glacigenic sediment is thin enough to allow the topography of the underlying bedrock to be expressed. In some instances this expression is no more than a high point that is attributed to a single bedrock outcrop, or a series of outcrops. In other cases, some of the actual structures (i.e. faults and joints) and contrasting lithologies (i.e. foliation) within the bedrock have topographic expression. Areas within this map unit where bedrock does not crop out, consist of a thin mantle of undifferentiated sandy, bouldery diamicton, or sorted sediment. Large boulders are also commonly present at the surface in this map unit. The relationship between bedrock and geomorphology is discussed in the next section of this report.

THE RELATIONSHIP OF BEDROCK TO GEOMORPHOLOGY

DRIFT THICKNESS

The present day geomorphology of the study area is a close reflection of the morphology of the underlying bedrock surface, and the topographic prominence of the Giants Range is its most obvious manifestation (Plate III). However, immediately north of the Giants Range are buried bedrock valleys which are filled with 100 to possibly more than 200 feet of glacial drift, subduing the extremely irregular buried bedrock surface (Plate III). These buried valleys appear to be located beneath the major streams of the study area. Even the valleys of Bear Creek (60N-15W-12 and 13) and Camp Eight Creek (60N-14W-09) seem to occupy buried valleys.

DIFFERENTIAL GLACIAL EROSION

Origin of the Giants Range and the Embarrass Gap

The most striking example of differential glacial erosion of bedrock in the Embarrass area is exhibited by the topographic prominence of the Giants Range. A possible scenario for the development of the Giants Range is outlined below.

The Giants Range is a prominent bedrock ridge extending from the south shore of Birch Lake at the east edge of the map area, southwestward to just west of the St. Louis-Itasca County line. The Giants Range and areas immediately north are underlain by the Archean Giants Range Granite and small areas of Archean supracrustal rocks (Morey and others, 1982). South of the Giants Range are the relatively flat-lying rocks of the Animikie Group (Morey and others, 1982). The northeastern end of the Giants Range coincides with the eastern end of the Biwabik Iron Formation (Fig. 47) (Green, 1982). The topographic prominence of the Giants Range diminishes southwestward where the Biwabik Iron Formation changes from magnetite facies to carbonate facies (G.B. Morey, personal communication, 1992).

At the Embarrass mine, at the south end of the Embarrass gap, the entire thickness of Biwabik Iron Formation hosted a soft, hematite ore deposit (Fig. 48). There are

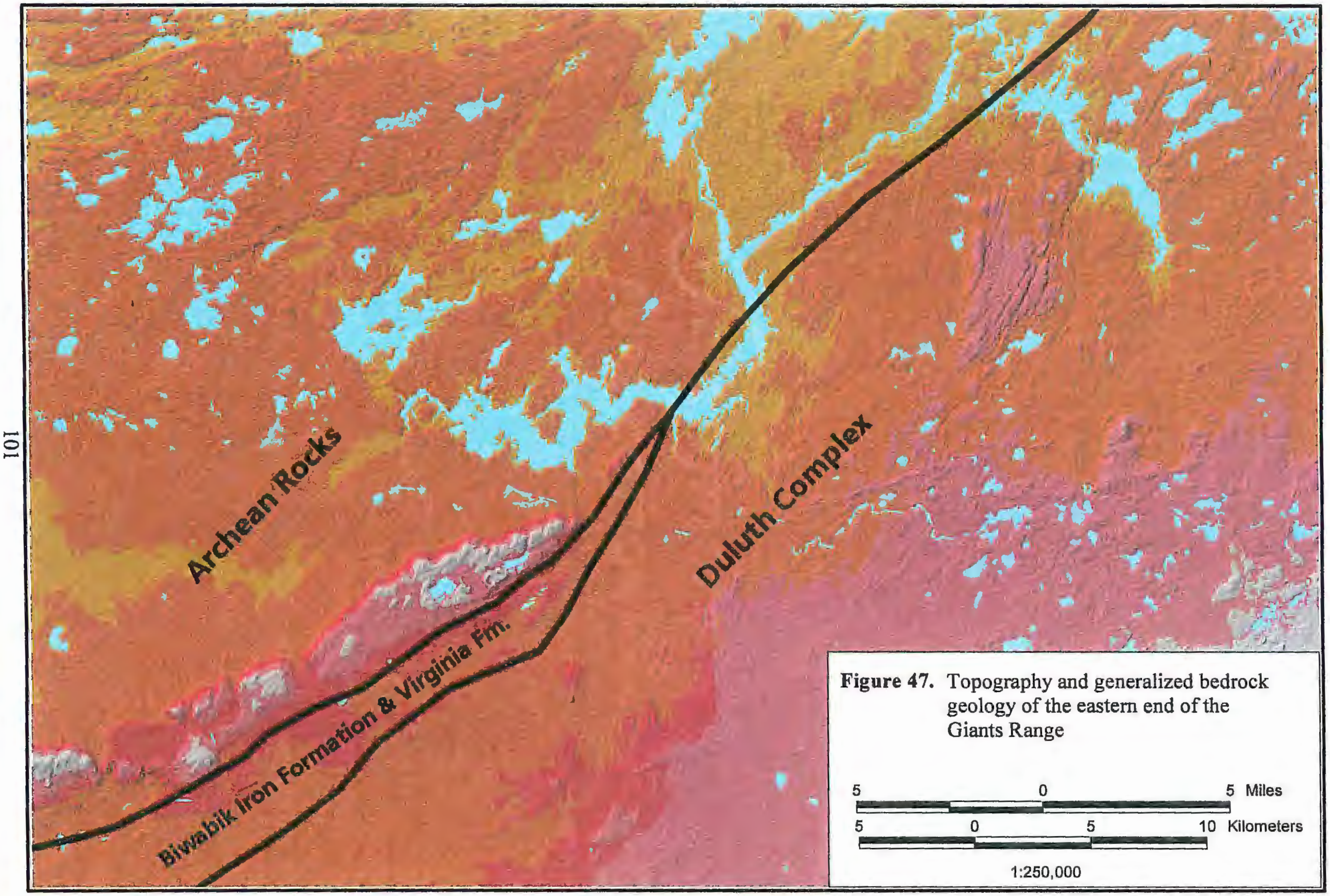

Figure 47. Topography and generalized bedrock geology of the eastern end of the Giants Range

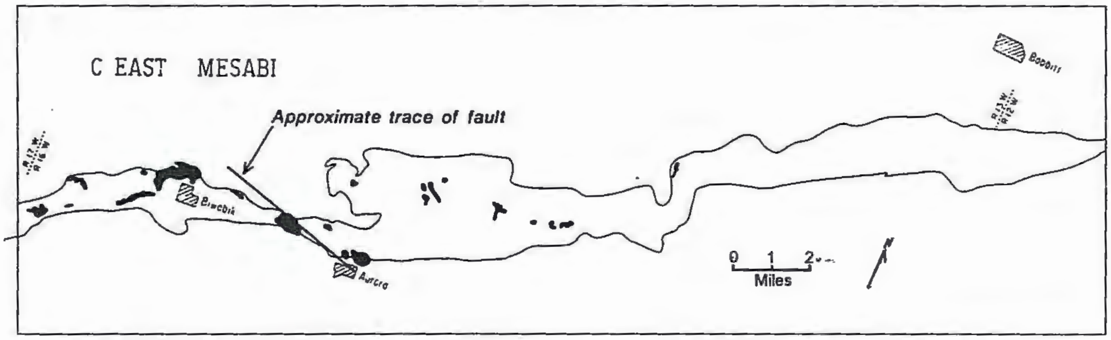

Figure 48. General map of the eastern part of the Mesabi range showing location, distribution and shape of natural ore bodies in the Biwabik Iron-Formation (modified from U.S. Steel map of 1956). From Morey (1972a)

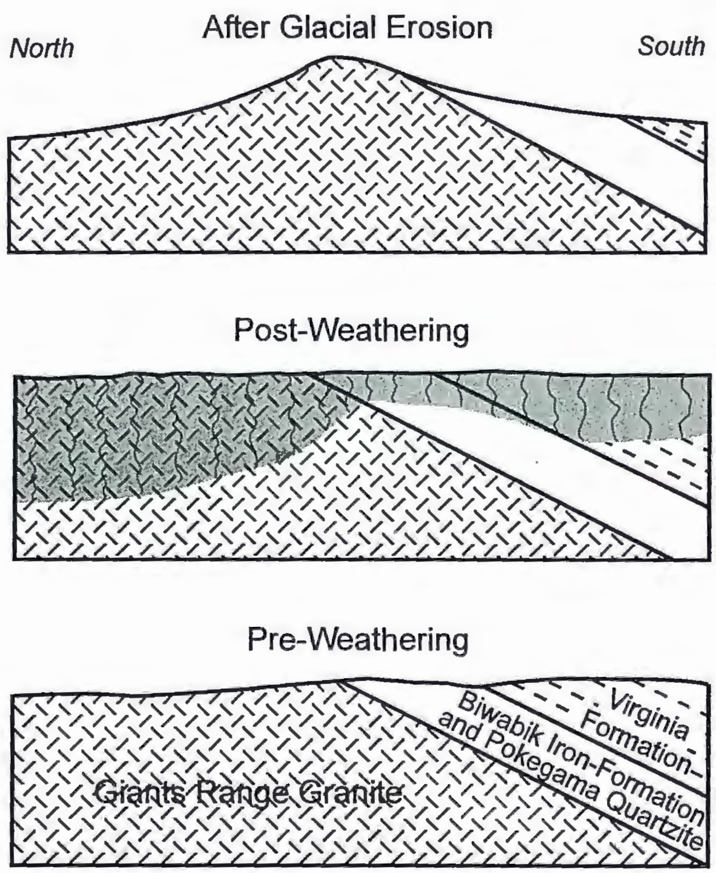

Figure 49. Proposed origin of the Giants Range

various theories for the formation of these "natural ores", but a credible theory invokes weathering of silicate iron-formation to a variety of ferric oxide minerals by ground water. At the Embarrass mine, this weathering was concentrated along a fault (Fig. 48).

The presence of the magnetite iron-formation must have played a role in the formation of the Giants Range. The Precambrian bedrock of Minnesota has been subject to at least two periods of intense chemical weathering, a pre-Late Cambrian episode (Morey, 1972b) and a late Jurassic and early Cretaceous episode (Parham, 1972). The Giants Range Granite would have been easily weathered, resulting in a thick saprolite. The silicate iron-formation would have thinned where it laps onto the Giants Range batholith (Fig. 49) and therefore may have weathered through its entire section whereas the thicker parts of this unit would have provided a barrier to deep chemical weathering (Fig. 48).

The first late Cenozoic glacial advances would have easily eroded the thick saprolite north of the Mesabi range, and some of the weathered iron-formation. The present-day Giants Range was protected from deep chemical weathering by a cap of silicate iron-formation. The Embarrass gap originated because there was no protective cap of silicate iron-formation, and it had been nearly entirely weathered to "natural ore" (Fig. 48). Therefore the Embarrass gap is probably a late Tertiary to early Pleistocene feature. In summary, the Giants Range resulted from a variable thickness of pre-glacial saprolite developed in different rock types which was then subsequently eroded (Feininger, 1971).

Other Examples of Differential Glacial Erosion

In other parts of the study area where bedrock is near the surface, the modern topography is clearly related to bedrock type and structure. For example, differential glacial erosion is expressed by the linear feature trending northeast-southwest from 60N-14W-02B to 60N-14W-09D. This feature is mapped as the Waasa fault (Griffin and Morey, 1969). The intense pre-Pleistocene chemical weathering was concentrated along fractures associated with the fault and subsequent glacial erosion selectively removed the soft saprolite. This example of differential glacial erosion is particularly pronounced because the strike of the fault lies parallel to the direction of the last ice advance.

As mentioned in the previous section on subglacial till, the northwest-southeast trending ridges present in the area of 60N-14W-01 and 12 and 60N-13W-06 can be related to the orientation of the foliation in the underlying gneisses. Quartz-rich bands within the gneiss were more resistant to chemical weathering than the amphibolite-rich bands, and so they remain as ridges after glacial erosion. These relationships of bedrock layering to differential chemical weathering and subsequent glacial erosion have been observed in other studies (Gordon, 1981).

Another example of differential glacial erosion is in the vicinity of Birch River Narrows northwest of Babbitt (61N-13W-23, 25 and 26). Here the northeast-southwest trend of the shoreline of Birch Lake near Camp Rivard is attributed a major fault (Griffin, 1967) while the northeast-southwest trending shore of Birch River Narrows is attributed to a minor fault. The northwest-southeast trending portion of Birch Lake Narrows is parallel to the foliation of the underlying gneisses (Griffin, 1967). The same processes outlined above may also be responsible for the surface expression of these bedrock features.

DEPTH OF GLACIAL EROSION

The subject of how much glacial erosion has occurred during the late Cenozoic has received a great deal of attention. Estimates of total glacial erosion vary from a few tens of feet (Flint, 1971; Gravenor, 1975; Sugden, 1976) to hundreds of feet (Bell and Laine, 1985) and even thousands of feet (White, 1972).

Kaszycki and Shilts (1979; 1980; Shilts and Kaszycki, 1986) have calculated depth of glacial erosion from near the center of the Laurentide Ice Sheet for the last glacial stage to be from six to 60 feet. If these estimates are multiplied times the 17 (Bowen, 1978) to 20 (Mix, 1987) glacial cycles for the Quaternary alone (<1.6 Ma.) (not including late Tertiary glaciations), one arrives at a conservative estimate of total glacial erosion on the order of several hundred feet.

Saprolite as much as 250 feet thick is present beneath glacial deposits in central Minnesota (Meyer, 1986). A similar cover of weathered rock undoubtedly covered northern Minnesota in pre-glacial times, but has been eroded. Based on these observations and the above discussions, glacial erosion on the order of several hundred feet in northern Minnesota seems most likely.

QUATERNARY HISTORY

Major glaciation of North America may have commenced as early as 3 ma as suggested by the oxygen isotope record (Bell and Laine, 1985) and was certainly well established prior to 2.2 ma. The presence of till in northern Nebraska which is overlain by a volcanic ash dated at 2.2 ma (Boellstorff, 1978) suggests considerable southern extent of continental glaciers by this time. Oxygen isotope records suggest that the average Pleistocene glacial cycle lasted 100,000 years (Mix, 1987); therefore, there may have been at least 25 major glacial episodes since the onset of continental glaciation 2.5 to 3 ma (Bell and Laine, 1985). Northeastern Minnesota was certainly glaciated numerous times during the late Cenozoic, but evidence exists only for the most recent events. Glacial deposits older than late Wisconsinan are present in the subsurface of northeastern Minnesota, but until recently they have received little study.

PRE LATE-WISCONSINAN

Recent rotosonic drilling by the Minnesota DNR - Division of Minerals has revealed the presence of multiple pre late-Wisconsinan tills of both northwestern and northeastern provenance in the area west and northwest of the study area (Martin and others, 1988; 1989; 1991). Tills of both northwestern and northeastern provenance contain clasts of Paleozoic carbonate and have calcareous matrixes. These data suggest that the glacial dispersal patterns of the late Wisconsinan Rainy lobe were different than during most of the late Cenozoic glaciations.

Winter and coworkers (Winter, 1971; Winter and others, 1973) reported several occurrences of a dark-colored, sandy, silty, till below late Wisconsinan Rainy lobe deposits in open pit mines on the Mesabi Range and referred to it as the "basal till. They also mentioned that this till did not look the same at all localities along the Mesabi Range (Winter and others, 1973); therefore, there may be multiple pre-Wisconsinan tills in the subsurface of the Mesabi Range area.

Spruce or tamarack wood (Preston and others, 1955) from glaciofluvial sediment overlying the "basal till" at the Duncan-Douglas mine near Hibbing yielded a radiocarbon age of >36,490 BP (Y-250) (Winter, 1971), suggesting a pre-late Wisconsinan age for the "basal till". Furthermore, at the Embarrass mine, a truncated weathered zone in the "basal till" is evidenced by a seven-foot-thick leached and oxidized zone (see description of the Embarrass mine section above). This degree of weathering must have occurred under interglacial climatic conditions, so "counting down from the top", the weathering zone is Sangamonian and the till is at least as old as Illinoian.

The extent of early Wisconsinan glaciation in the Midwest has been revised in recent years (Curry, 1989; Clark and others, 1993). It is currently thought that the early Wisconsinan glacial maximum was not as extensive as the late Wisconsinan (Richmond and Fullerton, 1987; Clark and others, 1993). Sediments of early Wisconsinan age have not been conclusively identified in northeastern Minnesota; however, the moderate degree of weathering in the older till present at the Dunka mine section suggests that the till and fluvial sediment present there may be early Wisconsinan with the weathering occurring under interstadial climate conditions of the middle Wisconsinan. If this interpretation of age is correct, the presence of Paleozoic carbonate and calcareous matrix in the sediments at the Dunka mine section indicates that glacial dispersal during early Wisconsinan was similar to that of previous glaciations, but different than the late Wisconsinan Rainy lobe.

LATE WISCONSINAN

The Hudson Bay lowland was deglaciated during the middle Wisconsinan, approximately 46-32 ka (Andrews and others, 1983; Berger and Nielsen, 1990), so this is a maximum age for the late Wisconsinan Laurentide Ice Sheet. Finite radiocarbon ages on wood from Alberta (I-4878), Saskatchewan (S-96), North Dakota (W-2450) (Clayton and Moran, 1982), and South Dakota (GX-14,675) (Gilbertson, 1990), suggest that the Laurentide Ice Sheet had considerable southwestern extent by approximately 27 to 29 ka.

Hawk Creek Phase

Little is known about the glacial history of Minnesota in the earliest part of the late Wisconsinan (35 to 20 ka). In Minnesota, the first late Wisconsinan advance was probably that of the Superior lobe, approximately 29 ka. This glacier advanced into southwestern Minnesota and to the base of the Coteau des Prairies in northeastern South Dakota, depositing the red, sandy Hawk Creek till (Matsch, 1972; Gilbertson, 1990). The Hawk Creek till may correlate with the Rainy-lobe-like Marcoux Formation in northwestern Minnesota (Moran and others, 1976), and with the "old red till" in central Minnesota (Mooers, 1988).

Assignment of a late Wisconsinan age for the Hawk Creek till is based on the observation that no interglacial weathering zone has been identified (Matsch, 1972), and that the underlying "gastropod silts" are probably Sangamonian (Stage 5) to middle Wisconsinan (Stage 3) (J.P. Gilbertson, written communication). An early or middle

Wisconsinan age for the Hawk Creek till is unlikely, since recent work suggests that in the early and middle Wisconsinan, the southwestern Laurentide ice sheet was less extensive than in the late Wisconsinan (Richmond and Fullerton, 1987).

Granite Falls Phase

Following the analogue of the latest Wisconsinan glaciation, flow probably shifted to a more northerly path across the Precambrian uplands of northern Minnesota. This glacier deposited the sandy, calcareous Granite Falls till (Matsch, 1972) in southwestern Minnesota and northeastern South Dakota (Gilbertson, 1990) soon after the Hawk Creek phase, since no buried weathering zone has been documented in the Hawk Creek till (Matsch, 1972).

The age of the Granite Falls till has received much discussion, and remains ambiguous. Two radiocarbon ages on wood from sediment beneath the Granite Falls till are >31,000 BP (W-99) and >39,000 BP (I-4932), while one is finite at 34,000 +2800 -2450 BP (GX-1309). The infinite ages can be explained as wood eroded from the widespread, wood-bearing "gastropod silts". Wood from the "gastropod silts" has yielded several infinite radiocarbon ages, the oldest being >56,000 (QL-4151 and QL-4152). But the finite age of approximately 34,000 ka from beneath the Granite Falls till represents the maximum age for the Granite Falls phase.

Toronto Phase

Flow in the Laurentide Ice Sheet continued to shift westward, because by approximately 26 to 23 ka, northwestern ice had reached southwestern Minnesota, northeastern South Dakota, and north-central Iowa depositing the Toronto till (Lehr and Gilbertson, 1988) and the Sheldon Creek Formation (Bettis, 1997). Two radiocarbon ages on wood from beneath the Toronto till and above a thick oxidized zone in northeastern South Dakota range from 22,900 $\pm$ 1000 (GX-3439) to 26,150 + 3000, -2000 (GX-2864). Several radiocarbon ages on wood and organic-rich sediment constrain the deposition of the Sheldon Creek Formation to between 40 and 24 ka (Bettis, 1997).

Alexandria Phase

Upon retreat of northwestern ice at the end of the Toronto phase, northeastern ice receded into central Minnesota and stabilized for some time at the Alexandria moraine complex. The Superior lobe at this time was probably at the St. Croix moraine in the southern part of the present Twin Cities. Both the Alexandria moraine complex and the southern part of the St. Croix moraine in the Twin Cities are massive ice-stagnation complexes, which suggest a stable ice margin for possibly thousands of years. Buried ice persisted in the Alexandria moraine complex and the southern St. Croix moraine until after the 14-ka advance of the Des Moines lobe.

Hewitt Phase

Renewed recession of the Rainy lobe resulted in formation of the Wadena drumlin field and deposition of the sandy, calcareous Hewitt till in north-central Minnesota. This chronology of an early late-Wisconsinan shift of flow direction from southwest to more southerly to southeast is the identical sequence displayed in the younger late Wisconsinan glacial deposits of Minnesota, particularly the Des Moines and Koochiching lobes.

St. Croix Phase

During the St. Croix phase maximum, the confluent Superior and Rainy lobes stood at the St. Croix moraine while the Itasca lobe was at the Itasca moraine (Mooers and Lehr, 1997). By this time, flow in the Itasca lobe had shifted to slightly west of south, while flow in the Rainy and Superior lobes was nearly straight west.

An age of greater than 20,500 $\pm$ 400 BP (I-5443) is suggested for the St. Croix phase by a radiocarbon age on basal organic sediment from a bog in the Pierz drumlin field, behind the St. Croix moraine (Wright, 1972b). Clayton and Moran (1982), on the other hand, suggest an age of 15 ka for the St. Croix phase, based on correlations with ice margins outside Minnesota. The Superior and Rainy lobes had certainly retreated from the St. Croix moraine by 14 ka, because outwash from the Des Moines lobe can be traced behind the St. Croix moraine in central Minnesota (Mooers, 1988). The St. Croix phase - from the time of the advance of the Superior lobe to the St. Croix moraine to the retreat of the Rainy lobe from the Toimi drumlin field area (Wright, 1972b), probably spans several thousand years from >20 ka to approximately 14 ka.

As the Superior lobe receded from the St. Croix moraine, it formed the Pierz drumlin field, ice-marginal fans, and a widespread network of tunnel valleys with eskers in east-central Minnesota (Mooers, 1988, 1990). The southern part of this terrane is mantled by deposits of the northwestern-provenance Granstburg sublobe of the Des

Moines lobe. As the Rainy lobe receded from the St. Croix moraine it formed the Brainerd drumlin field and hummocky end moraines (Mooers, 1988, 1990), some of which were later mantled with deposits of the St. Louis sublobe (Hobbs and Goebel, 1982). At this time, the margin of the confluent Superior and Rainy lobes was south of the Laurentian divide. In other words, its meltwaters were flowing directly into the Mississippi River watershed.

Continued recession of the Rainy lobe formed the Toimi drumlin field and deposited the Independence till (Wright and others, 1970). At most surface exposures in north-central and northeastern Minnesota, Rainy lobe tills are noncalcareous, although Björck (1990) reported some near-surface Rainy lobe tills in the study area as calcareous. Three recently described rotosonic cores from the northeastern part of the Toimi drumlin field shows that the Independence till is locally quite thick, ranging from 39 to 138 feet in thickness. Approximately the upper 30 feet of the Independence till in these cores is noncalcareous, but it is calcareous and contains a few carbonate pebbles at depth (Hobbs, 1992).

A hypothesis explaining the noncalcareous nature of near-surface Rainy lobe tills in northeastern Minnesota involves large-scale ice dynamics. While the Rainy lobe margin was in southwestern and central Minnesota, the Hudson ice divide connecting the dome in Labrador and the Keewatin dome (Dyke and Prest, 1987) was northeast of the western margin of Paleozoic rocks in the James Bay/Hudson Bay lowlands. In the earliest stages of southwestward flow (Granite Falls, Alexandria and Hewitt phases), the Laurentide Ice Sheet incorporated large amounts of carbonate from a possibly frost-shattered bedrock surface and from older carbonate-bearing tills. This debris was then elevated above the basal zone and transported into central Minnesota. The development of an ice stream in Hudson Strait (Dredge and Cowan, 1989; Dyke and others, 1989) during recession of the Rainy lobe, caused a southwestward shift in the Hudson ice divide (Dredge and Cowan, 1989). By the time the Rainy lobe receded through northeastern Minnesota, only Precambrian rocks were being eroded, thereby diluting any carbonate-bearing debris in the ice. In this hypothesis, the lower part of the Independence till was deposited earlier, when the Hudson ice divide was farther northeast, and the upper part of the Independence till, and the tills in the study area were deposited after the ice divide had been displaced to the southwest.

The precise age of the Independence till is uncertain, but an age of approximately 15 ka is suggested by two radiocarbon ages on organic sediment from bogs in low areas between Toimi drumlins. Basal radiocarbon ages of 14,690 $\pm$ 390 (W-1763) at Weber Lake (Ives and others, 1967) and 15,850 $\pm$ 240 (I-5048) at Kylen Lake (Birks, 1981) have been considered too old (Clayton and Moran, 1982), because in both cases, the dated material was lake silt. Lignite and black shale are absent from this area, but the introduction of old carbon from the calcareous Independence till (Karrow, 1992) has not previously been considered. The Independence till is older than 12.1 ka (Lu-2556), but it may be younger than 15 ka, possibly 14 ka.

Automba Phase

The Superior lobe, being thicker and more dynamic than the Rainy lobe, flowed into an area formerly occupied by the Rainy lobe in the Mille Lacs Lake area during the Automba phase (Wright, 1972b). The marginal position of the Rainy lobe at this time is uncertain, but it was probably in the area now covered by St. Louis sublobe deposits. After the Rainy lobe retreated from the Toimi drumlin field area, the Superior lobe advanced to the Highland moraine on the North Shore highland, truncating the Toimi drumlins and forming the Highland flutes (Wright and Watts, 1969).

As the Rainy lobe retreated northeast of the Aitkin area, proglacial lakes developed. Lake Aitkin I occupied the low areas northeast of Aitkin and was dammed on the south by the Superior lobe at the Highland-Mille Lacs moraine (Wright, and Watts, 1969). On the west, this lake was probably blocked, in part, by ice-cored Rainy lobe end moraines, and drained into the Mississippi River near Brainerd. Younger deposits obscure the exact path of this meltwater. Lake Upham I occupied the lowlands in the upper part of the St. Louis River watershed and was probably confluent with Lake Aitkin I (Wright, and Watts, 1969). These lakes were receiving meltwaters from both the Rainy and Superior lobes with brown-colored sediment supplied by the Rainy lobe and red-colored sediment by the Superior lobe. Little more is known about Lakes Aitkin and Upham I because the Alborn advance of the St. Louis sublobe covered the entire lake plain, incorporating reddish silt and clay.

After the Rainy lobe retreated northeast of the Hibbing area, where there is a three-way divide separating the Mississippi River, Hudson Bay and Great Lakes watersheds, proglacial lakes developed north of the Giants Range in the Hudson Bay watershed. A group of approximately 75 drumlins was formed at this time between

Keewatin and Buhl on the Mesabi Range (Wright and Watts, 1969).

The extent to which the Giants Range controlled the marginal shape of the Rainy lobe at this time is not certain, since Rainy lobe end moraines of this age are covered by St. Louis sublobe deposits (Hobbs and Goebel, 1982). It has been suggested (Mooers, personal communication) that the approximate 90-degree junction of the Rainy lobe at the St. Croix moraine and the Itasca lobe at the Itasca moraine, was caused by a division of flow lines by the Giants Range.

Vermilion Phase

Certainly by the time the margin of the Rainy lobe retreated into the study area, both the Giants Range and the Laurentian divide were influencing ice-marginal processes. From north of Hibbing northeast to the Embarrass gap, the Laurentian divide is coincident with the crest of the Giants Range. Therefore, north of the Giants Range, the Rainy lobe was generally fronted by lakes, while south of the Giants Range, at least in the areas east of the St. Louis sublobe, meltwaters flowed away from the ice margin. North of the Giants Range, the margin of the Rainy lobe was fairly straight, possibly controlled by accelerated ablation due to the presence of ice-marginal lakes. South of the Giants Range, the Rainy lobe had a lobate margin (Plate II). The Giants Range and the Embarrass gap also affected subglacial drainage, evidenced by the orientation of eskers deposited during deglaciation (Plate I; Plate II). The occurrence of Rogen moraine both north and south of the Giants Range (Plate II) indicates that the Giants Range was creating compressional flow within the Rainy lobe.

The extent and outlets of early, high-level lakes that developed between the Rainy lobe on the north, the Giants Range on the south, and the Koochiching lobe on the west are not certain. There are several places between Buhl and the Embarrass gap, where the crest of the Giants Range is lower than 1610 feet in elevation, that may have served as outlets for these local, high-level lakes.

The first ice-marginal lake to utilize the Embarrass gap outlet was at an elevation of approximately 1475 feet, and drained through a small spillway which enters the Embarrass gap near the Giants Ridge Ski Area (59N-16W-24A) (Plate I). This lake will be referred to as the upper level of Lake Norwood. The upper level of Lake Norwood was blocked on the west by high ground (greater than 1470 feet above sea level) in the area south of Sand Lake (59N-18W). The margin of the Rainy lobe at this time was probably south of the Big Rice moraine west of the Embarrass gap, possibly crossing the Giants Range at the Embarrass gap (Fig. 50). The north-south trending eskers south of the Big Rice moraine between Big Rice Lake and the Embarrass gap may have been deposited into the upper level of Lake Norwood at this time. As the margin of the Rainy lobe stood in the Embarrass gap, the resulting reentrant in the ice margin focused meltwater, burying the ice margin with sand and gravel, filling the Embarrass gap. The Koochiching lobe did not have to be blocking low western outlets at this time for there to be a lake at 1475 feet in elevation between the Giants Range and Rainy lobe south of the Big Rice moraine.

Lake Upham I must have been in existence south of the Giants Range at this time, or the ice-cored Embarrass gap would have easily been eroded to levels below 1475 feet.

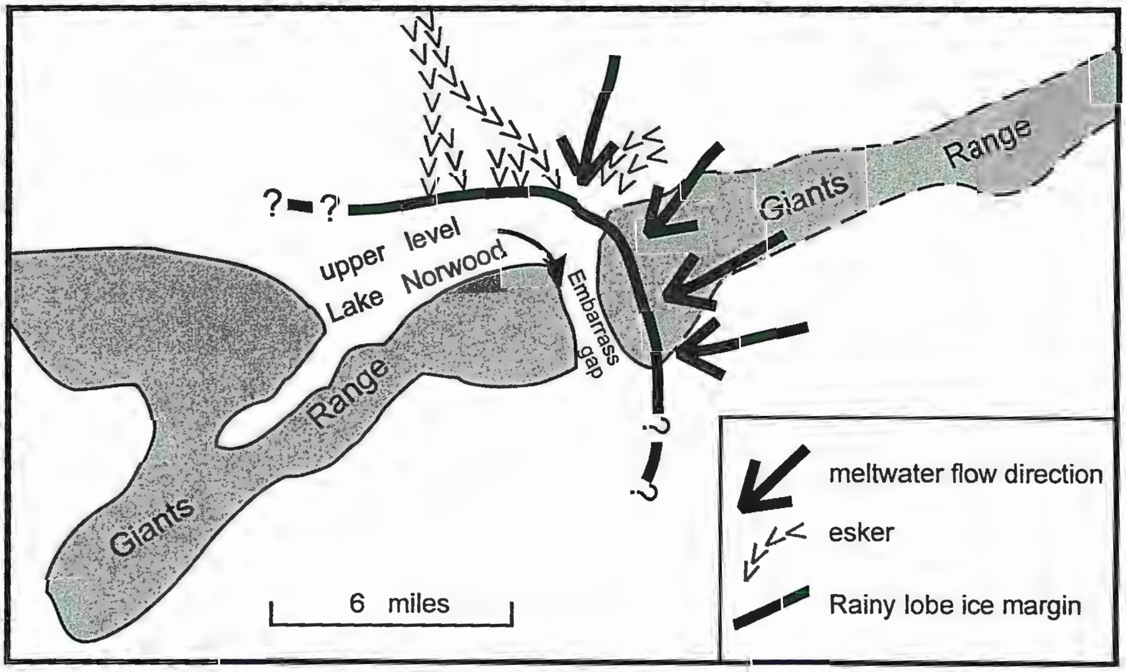

Figure 50. Rainy lobe ice margin in the Embarrass gap and development of the upper level of Lake Norwood

Another line of evidence suggests that Lake Upham I was contemporaneous with lakes north of the Giants Range. Björck (1990) reported a *"reddish-brown silty-clayey diamicton"* (p. 23) overlying glaciofluvial sediments in the Big Rice moraine five miles north of the Embarrass gap. He correlated this diamicton with the red clayey till deposited by the St. Louis sublobe, even though this is the only report of Alborn till north of the Giants Range this far east. Also, a water-well-log (Minnesota Department of Health, unique well number 117471) from one-half mile southeast of Björck's occurrence of "red till" reports "red clay" beneath 19 feet of sand and gravel. The red color of these sediments is most likely due to sediment deposited by northward overflow from Lake Upham I, which was receiving red-colored sediment-laden meltwaters from the Superior lobe (Wright and Watts, 1969).

The Rainy lobe stabilized for some time at the Big Rice moraine north of the Giants Range. This moraine is correlated with the Allen moraine south of the Giants Range. A readvance, or a reorientation of flow of the Rainy lobe, is suggested because the Allen moraine truncates Toimi drumlins on the south. A readvance (of at least minor extent) of the Rainy lobe to the Big Rice moraine helps explain the occurrences of red-colored sediment in the Big Rice moraine. In this scenario, the reddish lacustrine sediment deposited north of the Giants Range by overflow from Lake Upham I was incorporated by the Rainy lobe and redeposited locally.

The Big Rice/Allen moraine is younger than the Toimi drumlin field (approximately 14 ka) and older than 12,100 $\pm$ 150 (Lu-2556), a radiocarbon age from basal organic sediment from Heikkilla Lake (Björck, 1990), a kettle in the Big Rice moraine (60N-12W-30). The actual age of the Big Rice moraine is probably as much as a thousand years older than 12.1 ka. Tundra vegetation was present in the Heikkilla Lake area until approximately 10.5 ka (Björck, 1990; Huber, 1992); therefore, ice-block meltout, and subsequent organic deposition in Heikkilla Lake, was certainly delayed (Florin and Wright, 1969).

Tracing the Big Rice moraine through the Embarrass area to the Allen moraine south of the Giant Range is difficult. Ice-contact fans deposited into the upper level of Lake Norwood mark several ice-marginal positions. Meltwaters from the Rainy lobe at the Allen moraine flowed into the St. Louis River and then into Lake Upham I.

The Isabella sublobe may have joined the Rainy lobe at the Allen moraine near Stone and Big Lakes (58N-12W), where the Allen moraine appears to end and a few long, southwest-trending eskers terminate (Plate II). No definite marginal deposits have been identified for this inferred position of the Isabella sublobe, but a line drawn from Stone Lake to the Highland moraine separates an area of shorter drumlins to the northeast from longer ones to the southwest (Plate II).

The Superior lobe at this time stood at the Highland moraine. Its meltwaters were flowing into the lower Cloquet River and depositing the broad outwash plain in the Island Lake Reservoir area north of Duluth (Hobbs and Goebel, 1982). These meltwaters probably flowed into Lake Aitkin-Upham I.

Following the period of stability marked by the Big Rice and Allen moraines, the Rainy lobe retreated to the Wahlsten moraine north of the Giants Range and to the Wampus Lake moraine south of the Giants Range (Plate II). West of Wahlsten (61N-

15W-29), Lake Norwood fronted the Rainy lobe margin at the 1450 foot level. Here the Wahlsten moraine is a smooth ridge with a symmetrical cross-sectional profile composed of subaqueous outwash deposited into Lake Norwood. East of Wahlsten, the moraine is hummocky and diamicton dominated and fronted by outwash plains (Plate I).

A minimum age for the Wahlsten moraine is constrained by two radiocarbon ages on basal organic sediment from a kettle lake immediately in front of the moraine. Lempia Lake began receiving organic sediment between about 12,050 $\pm$ 240 (Lu-2555) and 11,500 $\pm$ 550 (Lu-2502) BP. Tundra vegetation was present in the study area from the time of deglaciation to between 10.5 ka (Björck, 1990; Huber, 1992) and 10 ka (Wright and Watts, 1969). The cold climatic conditions suggested by the pollen record may have enabled permafrost to survive in the area until after 12 ka. If permafrost was not present this late, the cold climate certainly inhibited meltout of buried ice (Florin and Wright, 1969). Therefore the actual age of the Wahlsten moraine must be older than 12.1 ka, possibly 13 ka.

Correlation of the Wahlsten moraine across the Giants Range to the Wampus Lake moraine is even more tenuous than the correlation of the Big Rice and Allen moraines. The eastern end of the Wahlsten moraine is covered by the jökulhlaup fan emanating from the Vermilion moraine, and the Wahlsten moraine ice margin was in Lake Norwood at the reentrant in the Rainy lobe north of the Giants Range (Plate I). Meltwaters from the Wampus Lake moraine, south of the Giants Range, followed the Partridge River into the St. Louis River and then into Lake Upham I (Plate II).

By the time the Rainy lobe stabilized at the Wahlsten moraine, the level of Lake

Norwood had dropped from 1475 to 1450 feet. The lake continued to use the Embarrass gap outlet, where there are terraces at approximately 1450 feet in elevation. There is also a boulder lag on the side of the Giants Range in the Embarrass gap at approximately 1450 feet. In the area of Embarrass, there are commonly boulder concentrations at approximately 1450 feet in elevation which are interpreted to be wave-washed shorelines of Lake Norwood. As mentioned above, Lake Upham I must have been in existence at approximately the same level as Lake Norwood at this time, or the ice-cored Embarrass gap would have been eroded to levels below 1450 feet.

Lake Norwood at the 1450 foot level was blocked on the south by the northern slope of the Giants Range and on the north by the Rainy lobe at the Wahlsten moraine. To the west there are presently areas lower than 1450 feet in elevation. Either Lake Norwood was held in by stagnant ice, or the Koochiching lobe was at least as far east as Swan Lake (near Pengilly, 56N-22W), blocking a gap in the Giants Range lower than 1450 feet.

As the Rainy lobe retreated through the northern part of the Toimi drumlin field, outwash from the Superior lobe at the Highland moraine was deposited along the Stony River. The Stony River at this time, probably crossed the Laurentian divide in what is now a peatland southwest of Sand Lake (59N-11W). At some point, these meltwaters became ponded in the upper Stony River basin, because the advance of the Isabella sublobe to the outer moraine incorporated a considerable amount of reddish silt, in addition to Superior lobe sand and gravel (Hobbs and others, 1988). The till of the Inner and Outer moraines of the Isabella sublobe is quite similar to till of the Superior lobe

(Friedman, 1981).

The Outer moraine may be slightly younger than the Wampus Lake moraine, because it appears to truncate the Wampus Lake moraine. If the Wampus Lake moraine correlates with the Wahlsten moraine, the Outer moraine may have been deposited approximately 13 ka. The Inner moraine appears to be truncated by the Vermilion moraine (Friedman, 1981), and is therefore older than approximately 12.5 ka.

As the Isabella sublobe retreated from the Outer to the Inner moraine, lower meltwater outlets were opened to the northwest and west. Each recessional ice position is marked by a meltwater channel or a band of outwash (Hobbs and others, 1988). One of these Superior lobe valley trains can be traced through the gap between the Giants Range and the Vermilion moraine, thence behind the moraine for a few miles, then back through the moraine into the Babbitt area (Plate I). These Superior lobe meltwaters eventually entered Lake Norwood. At the time the Isabella sublobe was at the Outer and Inner moraines, the Superior lobe was at the Highland moraine.

The Rainy lobe probably retreated somewhat before readvancing to the Vermilion moraine. A retreat and readvance is inferred only because the Vermilion moraine is oriented northwest-southeast compared to east-west for the Wahlsten and Big Rice moraines (Plate II). Alternatively, the reorientation of the ice margin at the Vermilion moraine may represent a shift to a southwesterly flow direction, possibly caused by a lowering of the glacier's profile by accelerated ablation in the areas to the west where the Rainy lobe was fronted by a lake.

From southeast of Soudan (62N-R5W-36D) northwestward to Nett Lake, a lake fronted the Rainy lobe. In this area, the Vermilion moraine represents a series of coalesced ice-contact deltas with flat-topped segments at approximately 1460 feet in elevation (Lehr and Matsch, 1987). These flat-topped segments represent the water level in Lake Norwood to which the deltas were built. Sediment was transported to the ice margin primarily by subglacial streams, as evidenced by numerous sharp-crested eskers north of the moraine. Some sediment was delivered to the ice margin by supraglacial streams and mass wasting, suggested by the higher and wider hummocky portions of the Vermilion moraine.

As the Rainy lobe retreated from the Wahlsten moraine, extensive low areas were uncovered to the west. At this time, either the Koochiching lobe was at least as far east as Swan Lake (near Pengilly, 56N-22W), or Lake Norwood drained. The early, high lake that developed in front of Keewatin ice north of the Giants Range has been referred to as Lake Koochiching, a precursor to Lake Agassiz (Nikiforoff, 1947; Hobbs, 1983). While the Rainy lobe was at the Vermilion moraine, Lake Norwood expanded westward, merging with Lake Koochiching, apparently maintaining the 1450-foot level. Field evidence suggests that Lake Norwood was not receiving meltwaters from the Koochiching lobe at this time, because in the eastern part of the Lake Norwood basin (Babbitt-Embarrass area), no Keewatin-provenance lacustrine sediments have been identified. Several exposures were examined in the Babbitt and Embarrass areas between 1430 and 1450 feet in elevation and all show only Rainy lobe lacustrine sediment. One possible explanation for the lack of Koochiching lobe lacustrine sediment in the Babbitt-Embarrass area is that currents did not distribute Koochiching lobe meltwaters into the eastern part of the Lake Norwood basin. Meltwater was also entering the eastern part of the Lake Norwood basin while the Rainy lobe was at the Vermilion moraine, both directly from the ice margin and as meltwater streams. This influx of sediment-laden meltwaters may have prevented the circulation of Koochiching lobe meltwaters into the eastern part of the Lake Norwood basin. Alternatively, Koochiching lobe lacustrine sediments may exist in the topographically lower parts of the Lake Norwood basin, but there are no exposures in these areas.

From Soudan, eastward to the junction with the Isabella moraine, the Vermilion moraine is slightly lobate, and in contact with the northeastern of the end of the Giants Range (60N-12W-09). In the area of Eagles Nest, Bear Head and Bear Island Lakes, the Vermilion moraine is hummocky and diamicton-dominated. Near the area where the Rainy lobe was in contact with the Giants Range, the Vermilion moraine has a gentle proximal slope, a steep distal slope and is composed of silty, sandy till. This segment of the Vermilion moraine is interpreted to be a push moraine.

The age of the Vermilion moraine can only be inferred from relative dating, because no radiocarbon ages are associated with this moraine. The Vermilion phase was interpreted by Wright (1972b) to be contemporaneous with the Automba phase of the Superior lobe. Because there are several Rainy lobe recessional moraines between the St. Croix and Vermilion moraines - Pleasant Lake, Stewart Lake, Outing, Sandy Lake (Mooers, 1988), Big Rice, and Wahlsten moraines - it is unlikely that the Vermilion moraine correlates with the Mille Lacs moraine of the Superior lobe (Mooers, 1988). The Vermilion moraine is younger than the Wahlsten moraine, which is interpreted to have formed approximately 13 ka. The most extensive Koochiching lobe advance overrode the western part of the Vermilion moraine, in the vicinity of Nett Lake (Martin and others, 1988). This advance is correlated with the Alborn advance the St. Louis sublobe at about 12 ka. Therefore, the age of the Vermilion moraine is between 13 ka and 12 ka, possibly 12.5 ka.

The Vermilion moraine correlates with the Isabella moraine of the Isabella sublobe (Hobbs and others, 1988). By the time the Isabella sublobe stabilized at the Isabella moraine, the till being deposited was increasingly more like the Rainy lobe till in color and stone content when compared to the tills of the Outer and Inner moraines (Hobbs and others, 1988).

As the Rainy lobe retreated from the Vermilion moraine, Superior lobe meltwaters crossed a low area in the moraine where it is bisected by the Stony River (60N-11W-08C). Superior lobe outwash can be traced both behind and in front of the Vermilion moraine into the Babbitt area, suggesting that while the Rainy lobe was retreating from the area for the last time, the Superior lobe was still thick enough for its meltwaters to flow northwest of the North Shore highland and into the area north of the Mesabi range. The Superior lobe must have retreated from the Highland moraine soon after the Rainy lobe withdrew from the Vermilion moraine, because Superior lobe outwash is found only immediately north of the Vermilion moraine in the Birch Lake area. No Superior lobe outwash has been reported from the lower areas farther to the north.

According to this chronology, the Superior lobe stood at the Highland moraine from the time the Rainy lobe vacated the Toimi drumlin field, approximately 14 ka, until after the Rainy lobe retreated from the Vermilion moraine about 12.5 ka. The combination of this long period of stability and lateral flow out of the Lake Superior basin, evidenced by the Highland flutes, produced the massive Highland moraine.

With the retreat of Rainy lobe, Superior lobe and Isabella sublobe ice approximately 13.5 to 12.5 ka, the study area was deglaciated for the first time since about 30 ka. This 17,000-year period of continuous glacial cover is at least partly responsible for the extensive areas of scoured bedrock and the general lack of glacigenic deposits older than late Wisconsinan in the region.

Alborn Phase

In the past, it has been assumed that Keewatin ice north of the Mesabi Range (Koochiching lobe) and south of the Mesabi Range (St. Louis sublobe) advanced synchronously (Winter, 1971). The Keewatin Koochiching lobe may have advanced into the low areas north of the western Mesabi Range as many as three times in latest Wisconsinan time (Martin and others, 1991), while the St. Louis sublobe probably advanced only once. The chronology of the Koochiching lobe is poorly constrained, because only one radiocarbon age (11,120 $\pm$ 250; Y-1782;) is associated with these advances. This radiocarbon age is from wood detritus at the base of a core from Myrtle Lake (Stuiver, 1969) in southeastern Koochiching County (63N-24W). Myrtle Lake probably originated as a kettle, and thus the significance of this date is uncertain. The first advance of the Koochiching lobe may correlate with the advance of the St. Louis sublobe and the advance of the Red River lobe to the Big Stone moraine.

The St. Louis sublobe of the Red River lobe advanced southeastward into the Lake Aitkin-Upham I basin south of the Giants Range, reaching a point only 22 miles from Lake Superior (Hobbs and Goebel, 1982). The St. Louis sublobe incorporated reddish-colored lacustrine sediment from Lake Aitkin-Upham I and deposited the red, clayey Alborn till (Baker, 1964). It is unclear whether Lake Aitkin-Upham I had drained by this time, or that the glacier advanced into the lake, displacing the water.

Radiocarbon ages associated with the St. Louis sublobe are in conflict. Wood fragments from peat buried in the upper part of Lake Aitkin II sediment at the main Aitkin site (Farnham and others, 1964) yielded radiocarbon ages of 11,560 $\pm$ 400 BP (W-1141) and 11,710 $\pm$ 325 BP (W-502) and provide a minimum age for Lake Aitkin-Upham II, which formed after the St. Louis sublobe advance. An age of 10,620 $\pm$ 400 (W-574) from buried peat at the west Aitkin site, two miles west, suggests the lake drained later, although this peat may not correlate with peat at the main Aitkin site (Farnham and others, 1964).

Two radiocarbon ages (11,330 $\pm$ 350 BP; W-827 and 11,100 $\pm$ 400 BP; W-1140) from wood enclosed in the Alborn till at the Mariska Mine near Gilbert (58N-17W-24DC) conflict with the dates from the Aitkin site. These ages have been rejected as too young (Wright and Watts, 1969) because the Red River lobe had retreated by this time, allowing Lake Agassiz to form in the Red River Valley (Clayton and Moran, 1982). In addition to this argument, the following explanation is offered. The geologic setting at the Mariska Mine appears, from examination of aerial photographs, to be a Rainy lobe esker complex that is, according to Farnham (Farnham and others, 1964), mantled with reddish-colored clayey till. Stagnant ice persisted in the Embarrass gap area, just 10 miles northeast of the Mariska Mine, until approximately 10 ka (Lu-2504, Lu-2506, Lu-2507). Therefore, the apparently young ages can be explained as dating the meltout of underlying Rainy lobe ice and burial by flow till of trees growing on the debris-covered ice. The older dates from the main Aitkin site (W-502 and W-1141), are probably the most accurate estimate (Clayton and Moran, 1982; Wright and Watts, 1969), of a minimum age for the Alborn advance. Therefore the St. Louis sublobe advance may have occurred about 12 ka (Wright and Watts, 1969), or a few hundred years earlier.

As the St. Louis sublobe stood at the Culver moraine, meltwater streams flowed south, merging with meltwater streams from the Superior lobe at the Nickerson-Thomson moraine. This demonstrates that the Alborn phase of the St. Louis sublobe was contemporaneous with the Nickerson phase of the Superior lobe (Wright and Watts, 1969).

The St. Louis sublobe probably stagnated soon after advancing. As the cleaner ice in the center of the lobe melted, Lakes Aitkin-Upham II formed, held in by ice-cored moraines (Hobbs, 1983). This lake drained through a variety of outlets to the south, both directly into the Mississippi River and into the St. Croix River via several channels that pass around the Nickerson-Thomson moraine (Wright and Watts, 1969; Hobbs, 1983). Lakes Aitkin-Upham II had drained by about 11.6 ka (W-502 and W-1141).

Embarrass Phase

Lake Koochiching/Norwood maintained the 1450 foot level while the

Koochiching lobe was at its maximum somewhere near the western St. Louis County line (Martin and others, 1988), and the St. Louis sublobe was at its maximum south of the Giants Range. Meltwaters flowed southward through the Embarrass gap and then around the eastern margin of the St. Louis sublobe into the Us-Kab-Wan-Ka, Cloquet and St. Louis Rivers; around the Superior lobe and into the St. Croix River (Hobbs, 1983). Upon stagnation of the St. Louis sublobe and retreat of the Koochiching lobe, the level of Lake Koochiching/Norwood dropped from 1450 to the early Mizpah level of 1430 feet (Hobbs, 1983). At Togo, in northeastern Itasca County, the highest wave-washed surfaces are at about 1430 feet (Hobbs, 1983), suggesting that this area was ice-covered when Lake Koochiching/Norwood was at the 1450 foot level. The wave-washed boulder lags in the vicinity of Embarrass at 1430 feet in elevation were formed as Lake Koochiching/Lake Norwood stood at the early Mizpah level. Drainage from the early Mizpah stage of Lake Koochiching crossed the Laurentian divide approximately six miles north of Giants Ridge (60N-15W-30, 31, 32), then flowed south through the Embarrass gap where there are terraces at approximately 1430 feet. This water must have drained into an early, high stage of Lake Upham II (Hobbs, 1983) and eventually into the Mississippi River via the St. Croix River.

With continued retreat of the Koochiching lobe and establishment of lower levels of Lake Aitkin-Upham II, the level of Lake Koochiching dropped to the later Mizpah stage of approximately 1400 feet (Hobbs, 1983). This level is marked by strandlines near Mizpah (62N-28W) at approximately 1400 feet and by terraces in the Embarrass gap at approximately 1400 feet. The drainage of the later Mizpah stage of Lake Koochiching followed the same path through the Embarrass gap as drainage in the early Mizpah stage, although the land in the area of the Laurentian divide (60N-15W-30, 31, 32) is presently approximately 1425 feet. It is proposed that at least 25 feet of peat has been deposited in this area. Thick accumulations of peat are known to form near watersheds on former lake plains in other parts of northern Minnesota (Severson and others, 1980).

Cass Phase

Continued recession of the Koochiching lobe uncovered the Prairie River outlet near Grand Rapids, lowering Lake Koochiching to the Gemmell stage of about 1350 feet (Clayton, 1983; Hobbs, 1983). While Lake Koochiching stood at the Gemmell level, water from Lake Climax, at the south end of the Red River lobe, entered the lake via the McIntosh channel and deposited a delta near Trail (Clayton, 1983; Hobbs, 1983). Therefore the River Warren outlet had not yet developed. These events help define the early Cass phase of Lake Agassiz, approximately 11.6 to 11.7 ka (Fenton, and others, 1982).

Recently, the Lake Agassiz isobases and shoreline-uplift curves have been revised (Thorleifson, 1996). The study area is located approximately at isobase 5 (Thorleifson, 1996; Figure 27). On Thorleifson's Figure 28, isobase 5 intersects the Herman beach at approximately 1450 feet above sea level, suggesting an earlier merger of Lake Agassiz with Lake Koochiching/Norwood than earlier workers (Fenton and others, 1982) had proposed, possibly during the earliest stages of Lake Agassiz. In fact, Thorleifson (1996) proposes that the earliest stages of Lake Agassiz drained through the Embarrass gap.

As Lake Koochiching dropped from the Mizpah to Gemmell level approximately 11.6 ka, significant drainage of glacial meltwaters through the Embarrass gap ceased. Final meltout of stagnant ice in the Embarrass gap is recorded by radiocarbon ages on basal organic sediment from Sabin Lake, which currently occupies the northern part of the Embarrass gap at elevations lower than any known lake drainage level. Ages of 10,230 $\pm$ 120 (Lu-2506) and 10,320 $\pm$ 170 (Lu-2507) suggest that final meltout of stagnant ice in the Embarrass gap post-dates the entire lake-drainage history of the Embarrass gap. A small lake just north of the Embarrass gap did not begin receiving organic sediment until 9.5 ka (9,510 $\pm$ 90; Lu-2504).

More rapid rates of uplift in the Lake Koochiching basin than in the Lake Climax basin resulted in a reversal of flow in the McIntosh channel, and with melting of stagnant ice in the Big Stone moraine, the River Warren outlet became established (Fenton and others, 1983; Hobbs, 1983). This lowered Lake Climax to the Herman level and Lake Koochiching to the Trail level, and the Prairie River outlet was abandoned (Hobbs, 1983). Ice recession from the area of the McIntosh channel uncovered lower ground and Lakes Climax and Koochiching merged to form Lake Agassiz at, or above the Herman level, marking the end of the Cass phase and the beginning of the Lockhart phase of Lake Agassiz (Fenton and others, 1983).

Lockhart Phase

The history of Lake Koochiching outlined above occurred between the time the Rainy lobe receded from the Vermilion moraine (approximately 12.5 ka) and approximately 11.6 ka, the beginning of the Lockhart phase. During this time, the Rainy lobe was somewhere between the International border and the Lake Nipigon area, blocking the eastern outlets of Lake Agassiz, possibly at the Eagle-Finlayson-Brule moraine (Teller and Thorleifson, 1983).

Moorhead Phase

Approximately 11 ka, the Rainy lobe retreated north of the Lake Nipigon area (Fenton and others, 1983) and the Superior lobe may have retreated north of the Lake Superior basin (Clayton, 1983). This retreat opened outlets to Lake Superior that were lower than the River Warren outlet and initiated the fall in Lake Agassiz from the Campbell level to the Ojata level (Clayton, 1983). This event marks the end of the Lockhart phase and the beginning of the Moorhead phase of Lake Agassiz.

Emerson Phase

A readvance of the Rainy lobe to the Hartman-Dog Lake moraine and of the Superior lobe to the Marks moraine approximately 9.9 ka blocked the eastern outlets and Lake Agassiz rose again to the Campbell level and drained through the River Warren outlet (Nielson and others, 1982; Fenton and others, 1983). This event marks the beginning of the Emerson phase of Lake Agassiz.

The Emerson phase lasted until approximately 9.5 ka, when the Rainy and Superior lobes retreated north of the Lake Nipigon area (Fenton and others, 1983). Soon after this retreat, the Rainy lobe readvanced to the Nipigon moraine, once again depositing till rich in Paleozoic carbonate derived from the James Bay lowlands (Zoltai, 1965; Karrow and Geddes, 1987).

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APPENDIX A: GRAIN-SIZE ANALYSIS DATA AND MUNSELL COLORS

Sample Number	Map Unit	Location (T-R-S)	Sample Wt (g)	Munsell Color	Wt % Gravel	Wt % Sand	Wt % Mud	Wt % Sand	Wt % Silt	Wt % Clay	Field Description (abbreviated)
EM-18	2	60-15-15ABBBB	667	10YR 7/2	20	53	27	67	31	2	sandy, silty till
EM-23	2	60-15-22CBCCB	438	2.5Y 7/2	36	50	14	78	19	3	sandy, silty till
EM-27	2	61-15-25ADBBC	524	10YR 6/2	46	36	18	66	31	3	silty, sandy till
EM-40	2	60-15-03DDBDD	1016	2.5Y 6/2	33	57	10	85	15	0	sandy diamicton
IL-11	2	60-13-06DBCBC	4103	10YR 7/1	48	36	16	74	22	4	sandy till
IL-16	2	60-13-07BAAAD	2352	10YR 7/2	39	47	14	80	18	2	sandy diamicton
BABNE-2	2	60-12-09ADABC	889	10YR 6/2	38	49	13	79	18	3	silty, sandy till
AVERAGE	**2**		**1427**		**37**	**47**	**16**	**76**	**22**	**2**	
BW-14	(a)	58-15-05BDDCA	47	N.D.				71	25	4	
AA-1	3a	59-15-11CADAA	640	10YR 6/4	62	33	5	86	13	1	sandy diamicton
AA-4	3a	59-15-09DCABA	630	10YR 7/3	71	25	4	84	14	2	sandy, stony diamicton
EM-3	(b)	60-14-04BABAC	1401	10YR 5/3	63	28	9	77	10	13	sandy diamicton
IL-3	(b)	60-14-10BBBBA	3317	10YR 6/3	61	38	1	98	1	1	sandy, bouldery diamicton
IL-29	(c)	60-14-35ABBBC	3256	10YR 6/3	61	34	5	86	11	3	sandy diamicton
EN-3	(d)	62-14-27CDADC	1295	10YR 5/3	80	18	2	90	6	3	sandy diamicton
SN-1	(d)	61-15-22BACCC	644	10YR 6/3	58	34	8	81	17	2	sandy diamicton
AVERAGE	**3a**		**1598**		**65**	**30**	**5**	**86**	**10**	**4**	
BNE-26	3b	60-16-21DAADC	4095	10YR 6/4	57	34	9	78	18	4	very-bouldery diamicton
EM-10	3b	60-14-29DBBAA	3928	10YR 6/4	85	14	1	95	2	3	bouldery, sandy diamicton
EM-12	3b	60-14-29BDDBA	2375	10YR 5/3	82	17	1	93	2	5	bouldery, sandy diamicton

Sample Number	Map Unit	Location (T-R-S)	Sample Wt (g)	Munsell Color	Wt % Gravel	Wt % Sand	Wt % Mud	Wt % Sand	Wt % Silt	Wt % Clay	Field Description (abbreviated)
EM-32	3b	61-15-25BBAAC	629	10YR 6/3	60	34	6	86	10	4	silty, sandy, stony diamicton
EM-34	3b	61-15-27DDCCD	640	10YR 6/3	62	31	7	82	15	3	sandy, stony diamicton
EM-36	3b	61-15-34DCBBD	651	10YR 6/4	43	51	6	89	9	2	sandy, stony diamicton
IL-18	3b	60-14-24ACDCC	3850	10YR 6/4	63	36	1	95	4	1	sandy diamicton
BW-4	3b	59-15-19CCBDC	4472	10YR 6/4	74	25	1	98	2	0	stony, sandy diamicton
EN-6	(d)	61-14-01BDCCD	1241	10YR 5/4	60	39	1	97	3	0	sandy diamicton
AVERAGE	**3b**		**2431**		**65**	**31**	**4**	**90**	**7**	**3**	
EM-11	3b*	60-14-29BDDBA	2534	10YR 3/3	21	48	31	61	26	12	silty, clayey diamicton
BNE-6	4a	60-16-36CBADD	3390	10YR 6/4	48	49	3	95	3	2	sandy, pebbly diamicton
BNE-10	4a	60-16-25DABAB	4003	10YR 6/4	72	27	1	98	2	0	pebbly, cobbly, bouldery sand
BNE-17	4a	60-16-34AADCA	3203	10YR 7/3	27	57	16	78	19	3	diamicton
BNE-50	4a	60-15-17BBBBA	608	10YR 6/4	66	31	3	91	8	1	sandy, stony diamicton
BNE-51	4a	60-15-06DABAA	351	10YR 7/3	43	47	10	82	13	5	sandy gravel & gravelly sand
BNE-56	4a	60-16-26AABBD	479	10YR 5/4	43	52	5	92	7	1	sandy diamicton, stone-poor
EM-5	4a	60-14-21BDDCD	3491	7.5YR 4/4	50	45	5	91	8	1	pebbly sand, poorly sorted
EM-6	4a	60-14-21BCBCC	57	10YR 5/4	3	97	0	100	0	0	sand, well sorted
EM-7	4a	60-14-21BCBCC	2851	10YR 4/3	41	59	0	100	0	0	pebbly sand, poorly sorted
EM-8	4a	60-14-21BCBCC	3520	10YR 4/3	57	39	4	91	7	2	cobbly diamicton
IL-15	4a	60-13-18BDADC	107	10YR 6/3	1	97	2	97	3	0	medium sand
IL-19	4a	60-13-18ABABC	110	10YR 6/2	1	90	9	91	9	0	medium sand, well sorted
BW-10	4a	59-16-12ACACA	4794	10YR 6/4	74	23	3	89	8	3	sandy diamicton

Sample Number	Map Unit	Location (T-R-S)	Sample Wt (g)	Munsell Color	Wt % Gravel	Wt % Sand	Wt % Mud	Wt % Sand	Wt % Silt	Wt % Clay	Field Description (abbreviated)
AVERAGE	**4a**		**2074**		**40**	**55**	**5**	**92**	**7**	**1**	
BNE-35	4a*	60-16-23CCBAC	3542	10YR 7/3	26	48	26	65	31	4	diamicton with fine matrix
BNE-44	4a*	60-16-26ABABB	3524	10YR 6/4	24	50	26	66	29	5	sandy diamicton, collapsed
EM-4	4a*	60-15-24DCBBD	3025	10YR 4/2	28	31	41	44	46	10	till (?)
EM-21	4a*	60-15-24CACCA	224	10YR 6/3	38	46	16	74	21	5	silty, sandy diamicton
AVERAGE	**4a***		**2579**		**29**	**44**	**27**	**62**	**32**	**6**	
EM-1	4b	60-14-06DBBAB	51	10YR 5/3	0	100	0	100	0	0	fine to med sand, well sorted
EM-2	4b	60-14-06DBBAB	3427	10YR 4/3	53	41	6	88	9	3	pebbly, cobbly sand
IL-1	4b	60-14-09ADBCB	97	10YR 5/3	22	76	2	97	2	1	coarse sand
IL-2	4b	60-14-09ADBCB	111	10YR 6/2	0	98	2	98	1	1	fine sand
AVERAGE	**4b**		**921**		**19**	**79**	**2**	**96**	**3**	**1**	
IL-5	4c	61-14-26BACBD	3141	N.D.	79	20	1	97	2	1	bouldery, pebbly sand
IL-6	4c	61-14-23DCCBB	3510	N.D.	58	40	2	95	4	1	sandy, bouldery diamicton
AVERAGE	**4c**		**3326**		**68**	**30**	**2**	**96**	**3**	**1**	
BNE-1	5	60-15-29CAADA	74	10YR 6/2	0	92	8	92	8	0	fine sand & fine sandy silt
BNE-2	5	60-15-29CAADA	139	10YR 6/2	0	98	2	98	2	0	fine sand
BNE-15	5	60-16-26CDCDD	129	10YR 6/4	0	99	1	98	1	1	fine to medium sand
BNE-16	5	60-16-34AADCA	190	10YR 6/3	1	98	1	99	1	0	medium sand, well sorted
BNE-18	5	60-16-34AADCA	55	10YR 7/2	0	29	71	29	68	3	clayey, silty, fine sand
BNE-25	5	60-16-28DACAA	84	10YR 6/3	6	79	15	84	13	3	sand, well sorted

BNE-54	5	60-16-14ADDCD	152	10YR 5/4	19	79	2	97	2	1	Medium & coarse sand
BNE-57	5	60-15-29DDCDC	67	10YR 4/4	0	86	14	86	11	3	silty, fine sand
EM-15	5	60-14-19DBBCA	118	10YR 6/2	1	98	1	99	1	0	medium sand, well sorted
EM-24	5	60-15-12DDADA	103	10YR 4/4	0	85	15	85	10	5	silty, fine sand, well sorted
IL-14	5	60-13-18ACCBC	120	10YR 5/4	0	96	4	96	3	1	medium sand, well sorted
IL-20	5	60-14-15AAADD	101	10YR 7/2	0	37	63	37	54	9	silty, fine sand
BAB-1	5	60-13-34AAADD	145	10YR 5/1	9	90	1	99	1	0	medium sand
AVERAGE	**5**		**111**		**3**	**81**	**16**	**83**	**15**	**2**	
BNE-36	(e)	60-16-23CCDDC	41	10YR 7/4	0	24	76	24	68	8	loess
BNE-39	(e)	60-16-26CBDBD	46	10YR 6/4	0	73	27	72	25	3	loess
BNE-41	(e)	60-16-26BACCC	38	10YR 5/4	2	74	24	76	22	2	loess
EM-9	(e)	60-14-21BCBCC	109	10YR 6/3	0	43	57	43	55	2	silt
EM-14	(e)	60-14-19DBBCA	37	10YR 6/3	3	56	41	58	38	4	sandy, pebbly silt
EM-16	(e)	60-14-29BBCCC	44	10YR 6/3	1	68	31	69	27	4	silt
EM-30	(e)	61-15-35BCDAC	40	10YR 6/3	3	39	58	41	52	7	clayey silt
IL-9	(e)	61-13-28BCCAA	45	10YR 6/4	15	45	40	53	43	4	silty till (?)
IL-13	(e)	60-13-17CABCC	31	10YR 5/3	0	57	43	57	40	3	silt
BW-12	(e)	59-15-09CCCBD	36	10YR 5/3	1	24	75	24	60	16	silt, well sorted
AVERAGE	(e)		**47**		**3**	**50**	**47**	**52**	**43**	**5**	

Notes: (a) "Basal till" (Winter, 1971) from Embarrass mine section
(b) supraglacial till within bedrock drift complex unit
(c) supraglacial till within subglacial till unit
N.D. not determined
d) sample from outside map area
(e) eolian sediment
(*) fines-enriched diamicton

	Particle-Size Distribution (Individual Weight Percent)													
	Gravel		Sand					Mud		Selected Statistical Parameters (Folk, 1980)				
Sample Number	Pebble	Granule	Very Coarse	Coarse	Medium	Fine	Very Fine	Silt	Clay	Graphic Mean	Median	Incl. Std. Dev.	Graphic Skewness	Graphic Kurtosis
EM-10	80.47	4.35	4.58	4.20	3.62	1.44	0.50	0.35	0.48	-3.57	-3.80	2.40	0.20	1.28
EM-11	17.96	3.26	7.72	12.00	13.27	9.58	5.79	20.58	9.83	2.03	2.20	4.51	0.01	1.12
EM-12	75.13	6.48	6.72	5.24	3.23	1.34	0.61	0.41	0.82	-3.50	-3.60	2.60	0.02	1.45
EM-4	24.98	3.34	5.53	6.01	7.06	4.79	7.86	32.85	7.58	1.66	2.50	4.88	-0.21	0.81
EM-5	36.17	13.84	17.51	16.23	7.88	2.83	0.89	4.20	0.47	-1.37	-1.00	2.62	-0.09	1.08
EM-6	0.00	2.74	10.54	39.85	41.14	4.80	0.42	0.18	0.33	0.93	1.00	0.77	-0.18	1.15
EM-7	20.91	16.56	21.61	24.00	11.05	1.55	0.36	0.26	0.20	3.20	3.40	1.73	-0.62	1.08
EM-8	51.25	6.17	7.37	10.44	8.19	6.58	6.02	2.90	1.08	-2.33	-2.20	4.47	0.51	0.85
IL-15	0.00	0.60	0.98	4.47	28.35	44.68	17.34	2.37	0.20	2.30	2.30	0.84	-0.02	1.13
IL-19	0.00	0.53	1.65	6.98	37.52	33.97	10.18	9.03	0.14	2.17	2.10	1.15	0.20	1.47
EM-1	0.00	0.37	2.24	8.22	35.94	37.81	15.40	0.00	0.00	2.08	2.10	0.85	-0.11	0.98
EM-2	46.28	6.56	11.31	12.62	12.38	4.11	0.91	4.48	1.34	-2.00	-1.50	3.90	-0.05	0.97
IL-1	0.00	21.62	21.63	21.73	21.96	8.06	2.89	1.52	0.59	1.33	0.70	1.67	0.06	1.04
IL-2	0.00	0.00	1.46	6.95	43.38	38.59	7.90	0.99	0.73	2.03	2.00	0.78	0.05	1.16
BNE-1	0.00	0.00	0.00	0.00	7.22	50.57	34.16	7.87	0.17	2.93	2.80	0.74	0.30	1.12
BNE-2	0.00	0.00	0.00	0.78	26.62	59.81	11.26	1.50	0.00	2.38	2.40	0.60	-0.03	0.99
BNE-15	0.00	0.00	0.76	10.88	73.91	11.75	1.28	0.58	0.83	1.57	1.60	0.49	-0.04	1.42
BNE-16	0.00	1.28	4.50	16.79	45.26	26.22	4.59	1.12	0.25	1.62	1.70	0.93	-0.14	1.22
BNE-18	0.00	0.00	0.00	0.35	2.39	7.71	18.18	68.46	2.92	4.95	4.90	1.55	0.04	0.99

	Particle-Size Distribution (Individual Weight Percent)													
	Gravel		Sand					Mud		Selected Statistical Parameters (Folk, 1980)				
Sample Number	Pebble	Granule	Very Coarse	Coarse	Medium	Fine	Very Fine	Silt	Clay	Graphic Mean	Median	Incl. Std. Dev.	Graphic Skewness	Graphic Kurtosis
BNE-25	0.00	5.93	8.07	13.04	32.71	19.37	5.93	12.32	2.62	1.85	1.70	2.04	0.20	1.64
BNE-53	0.00	0.00	0.00	0.00	3.25	20.40	43.33	31.57	1.45	3.90	3.60	1.32	0.38	1.35
BNE-54	7.49	11.84	20.67	26.34	19.77	9.62	1.96	1.54	0.77	0.33	0.40	1.54	-0.06	1.01
BNE-57	0.00	0.00	0.00	0.57	5.53	36.47	43.57	11.19	2.68	3.17	3.20	1.07	0.18	1.80
EM-15	0.00	0.77	1.95	3.06	12.18	51.36	29.63	0.97	0.06	2.63	2.70	0.76	-0.24	1.23
EM-24	0.00	0.00	0.00	0.00	4.98	43.34	36.54	10.42	4.70	3.13	3.30	1.26	0.47	2.32
IL-14	0.00	0.00	0.78	5.81	29.98	49.07	10.67	3.15	0.53	2.20	2.20	0.79	0.02	1.48
IL-20	0.00	0.00	0.00	0.12	1.31	6.43	29.46	54.20	8.48	5.05	4.75	1.79	0.28	0.93
BAB-1	0.00	9.08	11.54	34.75	36.61	6.69	0.82	0.52	0.00	0.73	0.80	1.02	-0.16	1.15
BNE-36	0.00	0.00	0.00	0.00	1.36	2.10	20.56	68.39	7.58	5.45	5.40	1.64	0.09	0.84
BNE-39	0.00	0.00	1.71	6.39	15.78	17.51	31.22	24.69	2.69	3.40	3.30	1.91	0.14	1.24
BNE-41	0.00	1.87	1.30	3.33	11.33	34.85	23.68	21.24	2.40	3.22	2.95	1.71	0.26	1.63
EM-9	0.00	0.00	0.04	0.20	2.82	15.20	24.55	55.11	2.08	5.27	5.30	1.32	0.03	0.92
EM-14	0.00	3.19	5.60	9.57	11.87	13.64	15.65	36.31	4.16	3.42	3.45	2.56	0.01	0.94
EM-16	0.00	1.51	2.07	4.10	10.79	21.66	29.07	27.12	3.67	3.58	3.40	2.04	0.16	1.35
EM-30	0.00	2.87	6.85	6.78	6.85	5.72	13.36	51.10	6.46	4.07	4.45	2.80	-0.17	0.97
IL-9	0.00	15.10	8.93	8.15	10.52	4.82	12.18	36.80	3.48	2.83	3.25	3.30	-0.17	0.82
IL-13	0.00	0.00	2.29	4.49	12.01	16.96	20.96	40.10	3.19	3.78	3.65	2.05	0.12	1.02
BW-12	0.00	1.11	1.50	3.69	4.71	4.24	9.48	59.84	15.42	5.58	5.65	2.52	-0.08	1.13

Limits of selected statistical parameters according to Folk (1980)

Inclusive Graphic Standard Deviation (Folk)

< 0.35	very-well sorted
0.35 - 0.50	well sorted
0.50 - 0.71	moderately well sorted
0.71 - 1.0	moderately sorted
1.0 - 2.0	poorly sorted
2.0 - 4.0	very-poorly sorted
> 4.0	extremely-poorly sorted

Inclusive Graphic Skewness (Folk)

+1.00 to + 0.30	strongly fine-skewed
+0.30 to +0.10	fine-skewed
+0.10 to -0.10	near symmetrical
-0.10 to -0.30	coarse skewed
-0.30 to -1.00	strongly coarse-skewed

Graphic Kurtosis (Folk)

< 0.67	very platykurtic
0.67 - 0.90	platykurtic
0.90 - 1.11	mesokurtic
1.11 - 1.50	leptokurtic
1.50 - 3.00	very leptokurtic
> 3.00	extremely leptokurtic