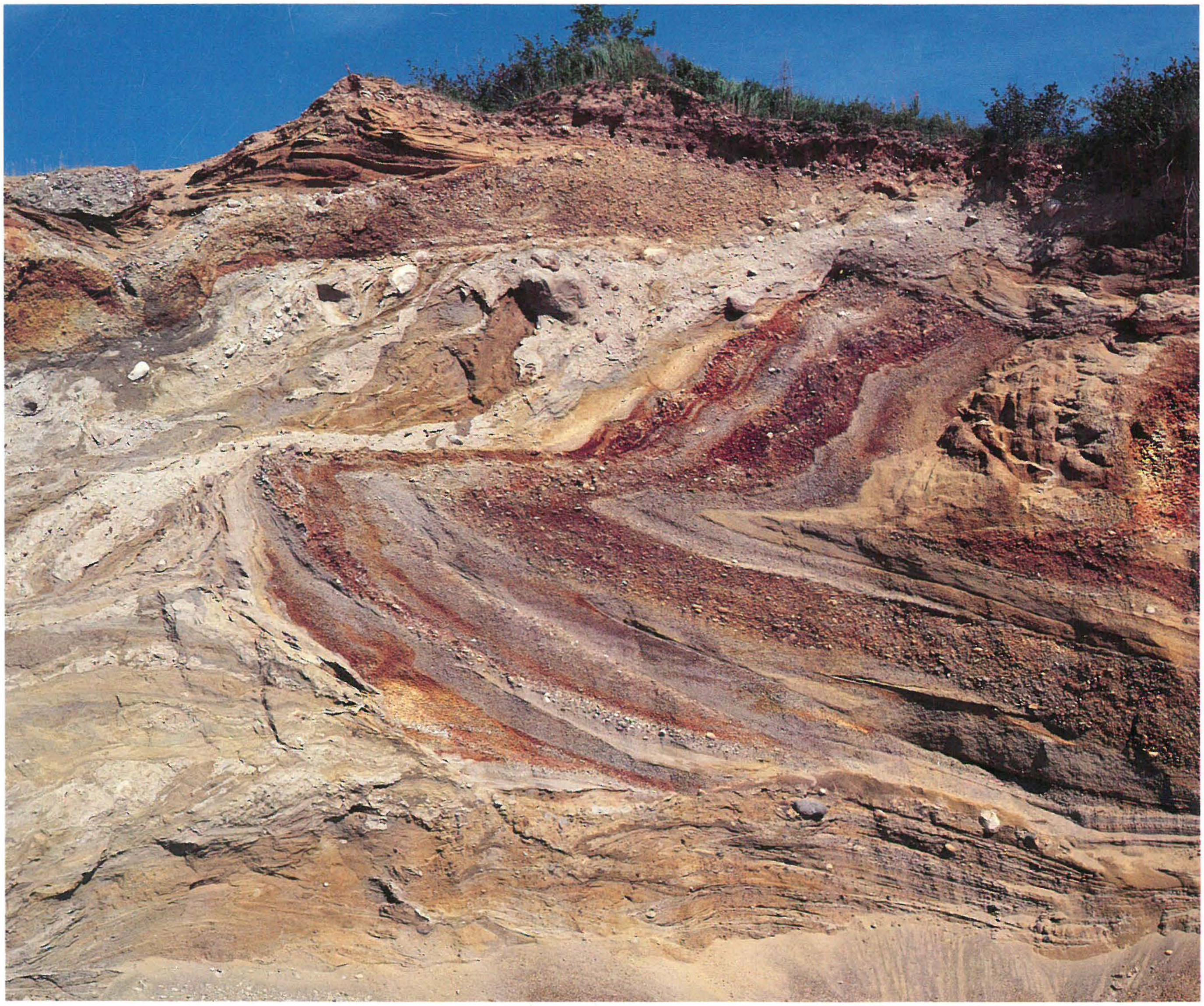
Studies in Maine Geology



Volume 6: Quaternary Geology

Edited by Robert D. Tucker and Robert G. Marvinney

Walter A. Anderson, State Geologist
Maine Geological Survey
DEPARTMENT OF CONSERVATION

Studies in Maine Geology

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1989

Cover photograph:

Ice-shove structures in an end moraine complex comprising diamicton, and subaqueous outwash deposits, Tracy Corner, Addison, Maine. Pit face is approximately 15 m high. Photograph by D. Tepper.

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Studies in Maine Geology

Papers to commemorate the 150th anniversary of
C. T. Jackson's reports on the geology of Maine

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C. T. Jackson's reports on the geology of Maine

Charles Thomas Jackson (1805-1880)

Foreword

During the period 1836-1839, Charles Thomas Jackson conducted the first comprehensive, government-funded survey of the geology of the state of Maine. As part of our commemoration of the 150th anniversary of that monumental undertaking, the Maine Geological Survey is publishing *Studies in Maine Geology*, a series of 6 volumes which cover a broad spectrum of geological investigations. The response to our initial call for papers was overwhelming not only in number, but also in the quality of contributions, which would have been well received by professional journals. I appreciate the efforts of each contributor in ensuring the success of these volumes.

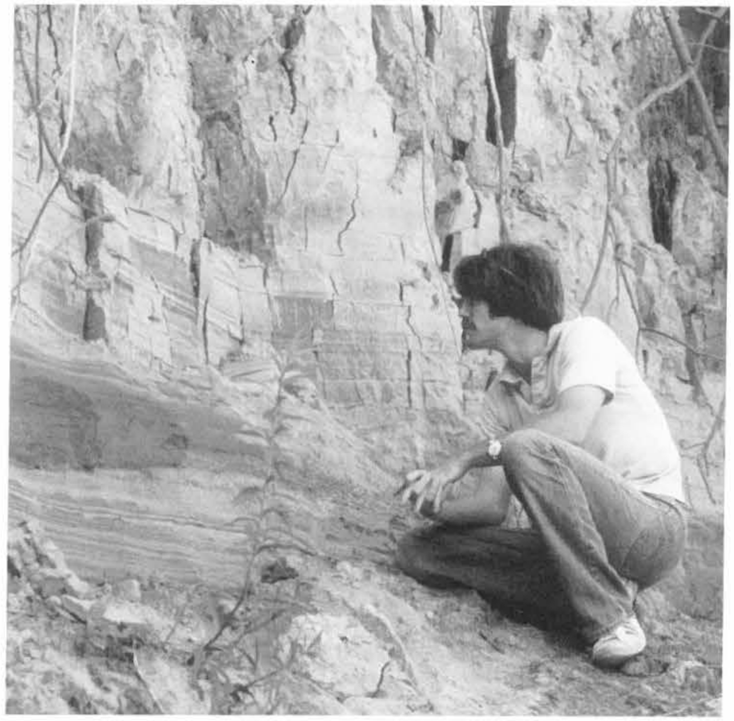
Many issues have challenged the geologic community since the time of that first survey. Jackson was charged with the seemingly insurmountable task of surveying the geology of the entire state over a period of 3 field seasons without the aid of accurate maps or modern equipment. One of the main objectives of his work was to assess the potential for mineral deposits, coal, and building materials in this unexplored territory. In the latter part of the 19th century the geological issues of the state concerned sources of dimension stone and pegmatite gemstones. During this period, the gold and silver rushes accelerated the pace of geologic exploration in Maine, causing a short-lived metal-mining boom. In the first half of the 20th century and with the advent of two world wars, the need for "strategic minerals" once again stimulated geologic investigation. The economic downturn in the domestic minerals industry of the past few decades has reversed and interest in exploration and extractive commodities is once again rapidly gaining ground in the state.

Today, a growing environmental awareness has led to a concern for protecting our resources, both now and in the future. Geologic information provides the basis for a multitude of decisions aimed at confronting the complex problems of modern society. Primary issues are those of ground water resources and contamination, coastal development and shoreline protection, and the disposal of nuclear wastes in geological repositories.

To meet the changing and expanding needs of the state over the past century and a half, the Maine Geological Survey has been under the auspices of many different government departments, culminating in its current position within the Department of Conservation. Since Jackson's time, the survey has expanded from essentially a one-man agency to include full bedrock and surficial geology, marine geology, hydrogeology, and cartographic divisions. In spite of the changes in issues and manpower, one of the primary objectives of the Maine Geological Survey has remained unchanged since Jackson's time: to provide the public with the highest quality information about the geology of the state of Maine. These volumes meet the challenge of that objective.

Walter A. Anderson

Walter A. Anderson
Director and State Geologist
Maine Geological Survey
DEPARTMENT OF CONSERVATION
Augusta, Maine



This volume is dedicated to

William R. Holland
(February 8, 1950 - April 3, 1989)

This volume is dedicated to the memory of William R. Holland, whose death in a mountain climbing accident in the Canadian Rockies was a great loss to his friends and colleagues in the Maine geological community.

Bill graduated from Colby College in 1972 with a B.A. degree in sociology. He then worked as a cartographer at the Maine Geological Survey in the mid-1970's, during which time he developed an interest in geology. Bill returned to school for training in this field, earning a B.S. degree in geology from Eastern Washington State College in 1978, and a M.A. degree, also in geology, from Boston University in 1980. While in graduate school, he gained experience in Quaternary geology through a mapping project with the U.S. Geological Survey in eastern Massachusetts.

During parts of 1980-1984, Bill carried out surficial quadrangle and aquifer mapping for the Maine Geological Survey. He made a major contribution to the 1985 Surficial Geologic Map of Maine through compilation of his work. The superior quality and detail of Bill's mapping reflect his long days in the field and devotion to understanding the glacial history of Maine. He made considerable progress in documenting the complex glacial deposits in the vicinity of the inland marine limit in eastern Maine, as well as the surficial geology of the Brownfield-Cornish area. Bill was a certified Maine geologist, and was active in the Geological Society of Maine and other professional organizations. He worked for E. C. Jordan Company, and since 1983 was employed by Robert G. Gerber, Inc., of Freeport. Bill's enthusiasm for ground-water studies and computer applications in geology were valuable assets in his consulting work. In spite of being a very observant and productive geologist, he was remarkably modest about his accomplishments.

This short summary of Bill Holland's professional career cannot do justice to his other talents, which ranged from music to mountaineering. Bill was a skilled ice and rock climber with much experience in mountain ranges across the United States and Canada. This hobby complemented his interests in glacial geology. Bill's inquisitive character and enjoyment of the outdoors gave him a well-rounded outlook on life and won many friends. Those of us who knew him will long remember this friendship.

Woodrow B. Thompson
Maine Geological Survey

Acknowledgments

Studies in Maine Geology began as a project to commemorate 150 years of geologic investigation in Maine. The original intent was to assemble a volume of papers which summarized our current understanding of all aspects of the geology of Maine. As word of the project spread, however, the volume grew into a six-volume set centered around three broad subject categories: structure and stratigraphy, igneous and metamorphic geology, and Quaternary geology.

The publication of a project of this size involves the time and goodwill of a great many people. The editors thank Marc Loiselle who initiated the project and contacted the authors. The illustrations and manuscript assembly were done by the Cartography and Information Services Division of the Maine Geological Survey. Additional editorial assistance was provided by Steve Dickson, Woody Thompson, Tom Weddle, and John Williams. Particular thanks are given to Ben Wilson, John Poisson, and Steve Jacques who drafted many of the illustrations and to Cathy Stultz and Cheryl Fiore who typed and corrected the manuscripts. We also wish to thank the geologists who willingly undertook the task of reviewing the manuscripts.

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Changing Perspectives of the Quaternary Surficial Geology of Maine

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ABSTRACT

The major historic publications reporting research on the surficial geology of Maine are reviewed to illustrate significant changes in scientific thinking that have led to the present understanding of Quaternary events and history.

The earliest research attributed the surficial geology of Maine to the biblical flood. By 1862, although the biblical flood concept still held scientific sway in Maine, glaciation started to gain importance as a factor; and by 1899 glaciation as the cause of the surficial geology of Maine had entirely replaced the flood theory in the scientific community. During the twentieth century, the collection and interpretation of geological data has rapidly expanded in Maine and adjacent parts of New England, Quebec, and New Brunswick. This progress in turn has allowed researchers to refine our understanding of the geological processes and history related to the surficial geology of Maine. However, although we know a great deal about the geological events of the late Pleistocene Epoch, almost nothing is known of early Pleistocene or Holocene events.

THE ICE-AGE THEORY

Louis Agassiz, a young Swiss naturalist from the University of Neuchatel, Switzerland, in his presidential address before the Swiss Society of Natural Sciences in 1837, revealed his revolutionary new theory that advocated the existence of a large glacier of continental proportions that had covered northern Europe and the Alps in the recent past. This glacier was said to be responsible for most of the erosional and depositional features previously attributed to the biblical flood, also referred to as the great deluge. Subsequently in 1840, in his monumental publication "Studies on the Glaciers," Agassiz provided the evidence for his conclusions. This was largely derived from his own observations in the Alps, integrated with and building upon the observations and conclusions of earlier and contemporary researchers in Europe such as Bernhardi in 1832, and Charpentier in 1834. Not only did Agassiz visualize an ice sheet covering northern Europe and the Alps, he also realized that the growth and demise of such an ice sheet would be directly related in a cause-and-effect manner to global changes of major magnitudes, including climatic cooling and warming, shifts in the distribution of plants and animals, and major sea-level changes. In addition, he predicted contemporary ice sheets on other continents. Hence, in total, the holistic concept of an Ice Age was born. Subsequently, Agassiz devoted his career to both researching his speciality, (modern and paleo ichthyology) and to compiling field data to support the concept of a global ice age. In 1846 Agassiz joined the faculty of Harvard College in Cambridge, Massachusetts with the charge to develop programs in the natural sciences. During the next 30 years, until he died in 1873, he did just this. He became a major influence on American science, and in his own right continued to be a leader in the field of ichthyology and an advocate of the Ice Age theory. In addition to his extensive research travels around North America and abroad, he came to Maine on several occasions, where he documented a variety of Ice Age phenomena (Agassiz, 1867, 1876). Although

Agassiz arrived in the United States in 1846, T. A. Conrad (1839) was the first American to accept the glacial theory, having been stimulated by Agassiz's 1837 paper.

THE BIBLICAL FLOOD VS. ICE AGE THEORY IN MAINE

Quaternary researchers in Maine (which was part of Massachusetts until 1821) prior to the proclamation of the Ice Age theory generally considered the unconsolidated glacial and glacial-marine sediments and bedrock erosional features to be the products of the biblical flood. The flood theory also included the concept of drifting icebergs as active participants in the processes of deposition and erosion. Sediments deposited by the flood were collectively called "diluvium." The persistence of the biblical flood/drifting iceberg origin for surficial deposits continued in Maine long after Agassiz's announcement of the global Ice Age theory in 1837. However, a growing number of workers presented an expanding body of data and conclusions which attributed these phenomena to the work of glaciers.

In the First and Second Reports on the Geology of Maine, Dr. Charles T. Jackson, the first State Geologist, continued to adhere to the popular biblical flood origin for the surficial sediments and modified bedrock surfaces that we currently attribute to glaciation (Jackson, 1837, 1838). Jackson synthesized his rather uncritical observations concerning the origin and distribution of various sediments and erosional features, including the emerged fossiliferous marine sediments of coastal and central Maine, the northwest-southeast trending eskers, the northwest-to-southeast transport of erratics, the trend of striae on bedrock surfaces, and the elevational distribution of both striae and erratics. He assigned the origin of these phenomena to the biblical flood coupled with the concept of iceberg drift. In doing so, Jackson concluded:

(1)A cold sea had completely covered the state to a minimum depth of 5,000 feet, at least equal to the height of the highest point in Maine, Mt. Katahdin.

(2)A great oceanic current flowed over Maine from northwest to southeast, causing icebergs to drag on and striate bedrock surfaces and to distribute erratic boulders to all elevations on the landscape. Within this sea the eskers were formed by the current as longitudinal sand bars, and the sizes of boulders that they contain reflect the high velocity attained by this ocean current.

It should be noted that the Jackson reports predate the wide dissemination of Agassiz's glacial theory, and Jackson's conclusions reflect the generally held geological views of his day.

During one of Agassiz's early forays into Maine in search of glacial features, he observed a well-developed, extensively smoothed and striated roche moutonnée, which is still observable adjacent to US Route 1A in Ellsworth Falls (Agassiz, 1876). This observation led him to proclaim for the first time that this region had, like northern Europe, been covered by a great ice sheet in the recent past. This may be the first such proclamation in the United States, but in any event provides one of the benchmarks in Maine science.

It is interesting to note that the general view, held by theologians and other scholars of the time, held that the "great deluge" was God's punishment of mankind for its sins. However, Jackson questioned this theology in part, by noting that the flood had in fact caused changes that were often beneficial rather than detrimental to mankind, at least in Maine.

> Although we are informed in the scriptures, that the deluge was ordained for the punishment of wicked men, it is certain, that there was mercy mingled with this dispensation, for the soils were comminuted, transported, and mixed in such a manner, that their qualities were improved, and rendered more suitable for the growth of plants, so that new and more fertile soils were prepared for coming generations, who literally reap advantage from the deluge. -- C. T. Jackson (1838).

Charles H. Hitchcock's subsequent reports on the geology of Maine (1861, 1862) are more scientifically sophisticated than the reports of C. T. Jackson. The geology is more thoroughly described and is systematically organized into a simple stratigraphic framework. In his classification scheme Hitchcock placed all (superficial) surficial deposits under the inclusive heading of "alluvium." He indicated that they constituted a geological formation and that the study of these deposits should be called surface geology. In addition, he reduced the number of terms for the superficial deposits from a confusing plethora of terms such as drift, diluvium, Pleistocene, Post-Pleistocene, and alluvium, used by previous authors, to the terms alluvium and drift. Hitchcock used "alluvium" as an inclusive term equivalent to drift and relegated drift to subdivision status. However, other workers used alluvium not only to denote the deposits as a whole, but also to denote the time period when the original deposition took place. Their "drift" comprised the deposits resulting from the modification of alluvium as well as the period when this modification was accomplished. Faced with this confusion Hitchcock divided drift as shown in Table 1. This classification simply appears to have added more confusion to an already confusing array of terms and usages.

Hitchcock's theory of the origin of surficial geologic features was still fundamentally based upon the concept of the biblical flood. However, relative to the conclusions of C. T. Jackson (1837, 1838), he added substantially more critical observations to the regional data base and was therefore able to reach more valid conclusions than Jackson concerning regional

TABLE 1. HITCHCOCK'S CLASSIFICATION OF QUATERNARY SEDIMENTS.

I. DRIFT	II. MODIFIED DRIFT
1. The Drift Period	2. The Beach and Sea Bottom Period 3. The Terrace Period 4. The Historic Period

events. In this regard he visualized that all of Maine and possibly the rest of northern North America were submerged beneath the sea, with currents flowing across the area from the north and northwest during the Alluvial period. Hitchcock further believed that the now-emergent "marine clay" of coastal Maine was deposited in the sea, and that most of the bedrock striae, including those on high mountains, were formed by drifting icebergs. He concluded, as had C. T. Jackson, that the sea was much colder than it is presently along the coast, judging from the assemblage of cold-water fossils contained in the emerged clay. Furthermore, he believed that icebergs were primarily responsible for the sea's coldness.

Hitchcock's definitions, coupled with the basic characteristics of the alluvial deposits, allowed him to partition the drift into the two general categories and four subdivisions mentioned above. This classification reflects the differences he visualized between alluvial sediments deposited at the stage of maximum submergence and those resulting from their reworking, first by marine action, subsequently by estuarine action, and finally by fluvial processes as the land emerged from the sea. In this scheme he visualized that icebergs, both large and small, became grounded throughout the region. Melting of the icebergs was responsible for the formation and deformation of sediments deposited against them, and ultimately for depressions left by their final melting.

Although Hitchcock clearly felt that the biblical flood was ultimately responsible for the surface geology of Maine, he conceded that small glaciers may have been present before the maximum submergence, as well as during the emergence, and coexisted with the icebergs. He reasoned, probably based upon his image of valley glaciers following Agassiz's research in the Alps, that the glaciers on our landscape must also have been confined to valleys. He further reasoned that numerous striae would be found confined to valleys and would indicate flow through the valleys. Hitchcock also concluded that markings left by valley glaciers would be similar to those left by drifting icebergs, but felt that they would be more consistent in their implications of flow directions than would those striae caused by randomly drifting icebergs. While glacier striae would be restricted to valleys and essentially parallel to their axes, striae produced by drifting icebergs would be found everywhere, including the tops of formerly submerged mountains in Maine.

As an example of glacial action, Hitchcock (1861) cited the great numbers of striae and their parallelism within the upper portion of the St. John River valley as the only evidence found to date. However, he predicted that similar evidence would be found in the mountains of Oxford and Franklin Counties, and indeed he presented such field evidence in the following year (Hitchcock, 1862).

In summary, it should be noted that Hitchcock (1861) reported the first evidence in any major geological report on Maine to support the concept of glaciers on the landscape. It is also clear in his report that the great deluge explanation for the origin of surficial deposits was still preeminent in his mind as well as in the scientific community. However, here for the first time in Maine, some credence was given to the possibility that glaciers, albeit valley glaciers, could possibly have played a role in forming the surface geology of Maine. This apparently was in deference to the growing acceptance of the Ice Age theory.

In retrospect, the evidence Hitchcock presented does indeed indicate glaciation, but of a style and magnitude beyond his experience and perhaps comprehension. Features of the surface geology of Maine were being increasingly attributed to glacial action, and in general Agassiz's Ice Age theory was being tested and accepted by the broad scientific community.

Dr. John De Laski, a physician and naturalist, in a letter (1854) to C. H. Hitchcock, reported on the distribution of striae in central coastal Maine (Hitchcock, 1862). He concluded that the striae were produced by glacier flow towards the south, which was inconsistent with the generally held view of iceberg-produced striae indicating flow towards the southeast over the entire state. In 1862 De Laski sent a solicited letter to Mr. George L. Goodale which was subsequently published in the Report of the Scientific Survey of the State of Maine (Hitchcock, 1862). In his letter De Laski carefully pointed out and documented the "gigantic system of sculpturing" reflected in the streamlined, elongated hills of the central coastal region. He observed that these hills are smooth on the sides and tops, but have bold, rough southerly ends, and noted that these characteristics are present at all elevations. These, along with many other critical observations of geological features, led De Laski to question their great deluge origin and to suggest alternatively that glacial action was responsible. Expanding on his analysis, he visualized that the bedrock sculpturing could not have been done by a small glacier confined to a valley, but rather "a glacier that filled the basin between the Camden Hills on the west to those of Mt. Desert on the east, was 40 miles wide, extended to a great distance north, involving several hills besides those mentioned, of a thousand feet high and certainly was not less than three thousand feet thick." This benchmark presentation of excellent observations and conclusions, although regionally limited, provided critical evidence for the major shift from the great deluge to the glacier theory as being responsible for the surficial geology of Maine.

In a later publication (De Laski, 1864) on glacial action in Penobscot Bay, in which he presented considerable field evidence, De Laski concluded that the glacier was not limited to this "great fjord of Maine." He alternatively reasoned that it must have extended far to the east and west and indeed was probably part of the universal glacier that covered the northern part of the continent in areas wherever striae had been observed (De Laski, 1864). In summary, Dr. John De Laski was the first worker in the region to conclude that a continental ice sheet had not only covered Maine, but areas far beyond.

THE NEW GLACIAL CONCEPT

In addition to De Laski, a growing number of contemporary researchers of Maine geology invoked glacial action as the cause

for the features previously attributed to the flood. For example, A. S. Packard, Jr. (1866), in studying drift phenomena from Labrador southward along the Atlantic coast, including all of northern New England, clearly believed in their glacial origin. The first modern comprehensive work devoted to the glacial geology of Maine was that of G. H. Stone. He did most of the field work for his monumental pioneering publication, The Glacial Gravels of Maine (1899), in the 1870's and 1880's, only a decade after the publication of C. H. Hitchcock's Geology of Maine (1862). We see in Stone's and Packard's publications a complete change from the biblical flood to nearly unconditional acceptance of widespread glaciation to account for the surficial geology of Maine.

Significantly, Stone (1899) was the first to produce a geographic map of ice-margin retreat in Maine. His map (Fig. 1) displays retreatal positions of the ice margin progressing from southeast towards the northwest. It is based upon his very careful and insightful observation and correlation of ice-marginal deposits and directional indicators of ice flow. Subsequent research has expanded his area of interpretation, but has only slightly modified his conclusions over much of the state.

It is worth noting that Stone recognized in his introduction to The Glacial Gravels of Maine (1899) the need to study the activities of modern glaciers in order to more adequately understand the results of past glaciers.

> The investigation made slow progress, not only because there were several thousand miles to be carefully explored, but especially because the nature of the subject renders such an investigation exceedingly difficult. The scout of the Western frontier who undertakes to guide a body of troops in pursuit of hostile Indians--to follow the trail, and, from the traces left behind, to give a history of the enemy's performances from day to day--has a difficult task before him: but in thus reconstructing history he has the advantage of knowing, from direct observation, the habits of the Indians. In his study of glacial deposits the glacialist labors under the disadvantage of not knowing, by observation, the exact nature of the geological work going on beneath and with an ice-sheet. It is comparatively easy to theorize regarding the probable behavior of such a body of ice, and, if properly held in check, imagination is of the greatest use in such an investigation, but the chances for error are very great. The method here adopted has been to collect as large a body of facts as possible, and then carefully to test various hypotheses by the facts, rejecting or holding in abeyance all theories not supported by positive field evidence. Glacialists are exploring a comparatively untrodden field, and it behooves them to proceed cautiously and to avoid dogmatism and denunciation. -- G. H. Stone (1899)

By 1935, when Leavitt and Perkins published their Glacial Geology of Maine, the first major study after that of Stone (1899), there was complete acceptance of the former continental glaciation of Maine within the scientific community. The major purpose of their study was to locate and characterize the gravels of Maine as potential road-building materials. In addition, they also produced a surficial geologic history. They reported a variety of equivocal evidence for periods of glaciation earlier than the last episode which we now assign to Late Wisconsin time (Leavitt and Perkins, 1935). It should be noted that considerable differences in time and effort went into the coverage of different areas of the state, which resulted in variations in the quality of results from various areas. Stone's coverage of Maine appears to have been more uniform by comparison.

Leavitt and Perkins proposed two provisional models for dissipation of the last ice sheet. In the first model, a generally east-west trending ice margin receded northward with a continuous but irregular front, progressively uncovering the state like a window shade going up, and continuing into the St. Lawrence Lowland. Final dissipation of the glacier occurred in central Canada. This view was generally held by workers in New England up until about 15 years ago, when a new model started to evolve.

Leavitt and Perkins' alternative model called for simultaneous widespread stagnation of the ice followed by melting downward and inward until the ice sheet separated into irregular masses and ice-tongues in the lowland basins and valleys. Their less-than-critical evidence documents both marginal recession in some areas and separation and stagnation in others. This dichotomous conclusion reflected a major controversy raging in New England at the time concerning which of these methods, marginal retreat or downwasting, was the major mode of glacier dissipation. The same argument persists today, but in a far less polarized and rigid manner.

Regardless of this controversy, Leavitt and Perkins did report observations from which a sequence of geological events was established. In turn, this allowed them to correlate glacial events in Maine with events in the Great Lakes region. The latter area served for many years as a "type section" for the glacial history of North America primarily because of the early and intensive research done in the Midwest by pioneers in the field of glacial geology in the United States.

The interpretations and correlations of Leavitt and Perkins were based upon relative chronology and therefore suffered from the lack of absolute dates. However, they assumed that the sequence of Late Wisconsin deglacial events established in the mid-continent would be the same all along the southern margin of the Laurentide ice sheet in terms of sequence and timing. This was a generally held view at the time. Hence they correlated, without much success in retrospect, the sequence of events as they saw them in Maine with events of the mid-continent. This task has subsequently proven to be much more complex than was imagined in 1935. In addition, they visualized that Maine was totally deglaciated for the last time by 25,000 years ago. Again, this conclusion was based upon correlation with interpretations from the mid-continent, but with no absolute chronological data available from either region.

Following the state-wide study of Leavitt and Perkins, there were a number of studies on the surficial deposits of Maine which provided considerable data on a variety of local topics. Some of these studies contributed to the analysis of major glacial events. For example, beginning with C. T. Jackson (1837), there are scattered discussions in the literature of the glacial-marine

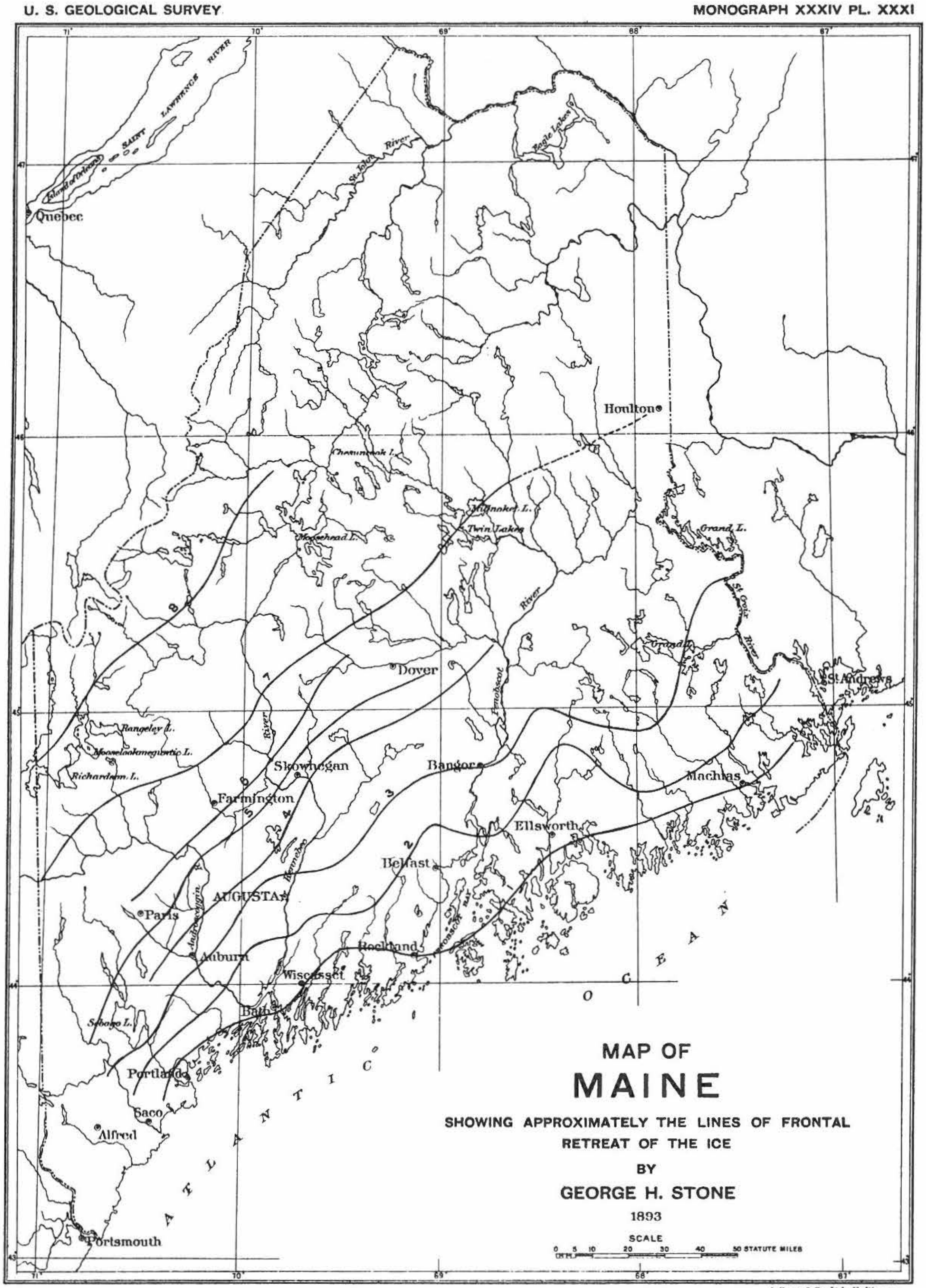

Figure 1. A map of Maine showing approximately the lines of frontal retreat of the Late Wisconsin ice sheet (from Stone, 1899).

sediments that are found throughout the lowlands of coastal and central Maine. Contemporary scientific questions focused on the origin of these sediments, first in the context of the great deluge, and later their relationship to glaciation and related land and sea-level changes. In addition, many published papers reported the nature of the marine macro-fossil assemblage within these sediments, and not only described the taxonomy of the fossils, but reported on their paleoenvironmental significance. Throughout all of these studies there was a thread of general concern about the age of the deposits and their relationship to the margin of the receding glacier and isostatic crustal and sea-level adjustments. In this regard Leavitt and Perkins (1935) were first to compile a map contouring the elevations of emerged glacial-marine features (Fig. 2).

The first significant studies of the emerged fine-grained marine sediments were published in the Report of the State Geologist for 1943-1944 (Trefethen, 1945; Trefethen et al., 1947). This was followed by a series of "clay reports" (e.g. Goldthwait, 1949) for several years, which addressed the geographic distribution as well as mineralogical and grain-size characteristics of the "clay" in regard to possible industrial uses such as lightweight aggregate. Modern studies of the late-glacial to early postglacial submergence and emergence of central and coastal Maine began with Bloom's work in southwestern Maine (1960). In studying the glacial geology, he defined the "clay" unit and named it the Presumpscot Formation. He was the first to critically discuss its relationship to land and sea-level changes and to the receding Late Wisconsin glacier margin. Stuiver and Borns (1975) expanded this style of modern research to include the entire area of late-glacial submergence in Maine. They reported that the Presumpscot Formation was entirely deposited between about 13,500 and 12,300 yr B.P. and applied their chronological data to an existing crustal rebound model. This study produced approximately 40 radiocarbon dates on fossil marine shells and on plant remains from bog bottoms, all of which were analyzed with state-of-the-art methods. Until that time there were fewer than five randomly distributed and poorly controlled radiocarbon dates, all from fossil marine shells collected at scattered locations in central and coastal Maine.

These two studies set the stage for expanded research. H. W. Borns, Jr., in a series of papers, reported and defined an extensive complex of marine grounding line moraines and ice-marginal marine deltas and fans. Borns' work focused on the central and eastern coastal zone (e.g. Borns, 1967, 1973; Borns and Hughes 1977; Borns et al., 1983). This style of work was expanded into the western coastal zone by G. W. Smith and W. B. Thompson, who examined the distribution and chronology of grounding line deposits and their implications for styles of deglaciation (e.g. Smith, 1981, 1985; Smith et al., 1982; Thompson, 1979, 1982). These and other field investigations were carried out as part of the surficial geologic mapping programs of the Maine Geological Survey. W. B. Thompson and co-workers further expanded the research on the marine transgression with studies of ice-marginal deltas throughout the entire

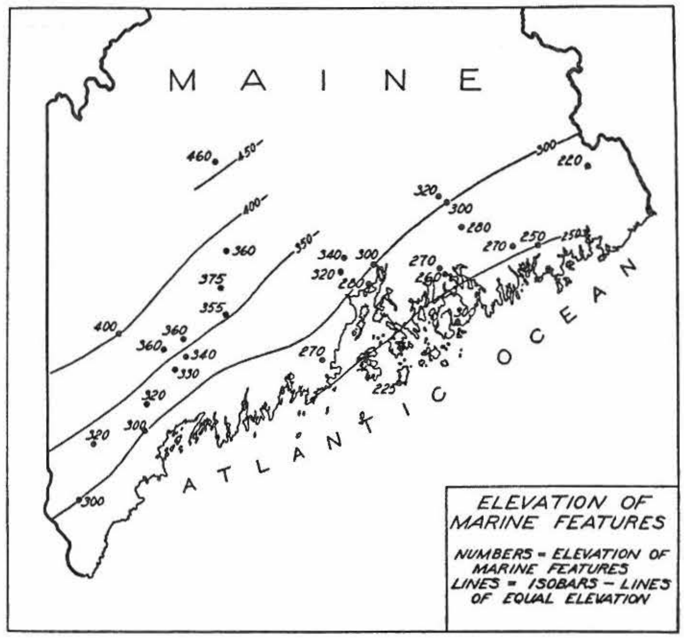

Figure 2. A map of Maine showing contoured elevations of emerged late-glacial deltas and other indicators of the upper marine limit (from Leavitt and Perkins, 1935).

area of submergence and used these data to contour the postglacial rebound pattern (Thompson et al., in press).

In addition to many topical and geographically local studies, increasing attention has been given to defining the former extent of the Laurentide ice sheet in the Gulf of Maine. Clearly, most researchers long ago agreed that the last ice sheet had flowed generally towards the southeast, across the present coastal zone of Maine, and terminated to the southeast at some unknown position on the continental shelf. Shepard et al. (1934) were the first to quantitatively address the question of the extent of the glacier on the continental shelf. In their paper on the geology of Georges Bank they determined that the last ice sheet reached that position after expanding across the entire Gulf of Maine.

Since the work by Shepard et al. (1934), a number of Canadian and American studies have been done in the different areas of the Gulf of Maine. These studies have yielded two opposing ideas on the former extent of the glacier (e.g., Borns et al., 1983; Dyke and Prest, 1987; Mayewski et al., 1981; Stone and Borns, 1986). In summary, there is little doubt that ice extended to the edge of the shelf; the question is during which ice advance. Presently, the evidence is ambivalent as to whether the Laurentide ice sheet terminated just off the Maine coast or on the edge of the shelf. However, accumulating evidence favors the latter hypothesis (Stone and Borns, 1986).

MULTIPLE GLACIATION

Leavitt and Perkins (1935) were the first to present data in an organized manner that suggested the possibility of multiple

phases of glaciation in Maine. These data consist of regional variations in glacial ice-flow directions determined by striae, coupled with localized observation of crossed sets of striae, and stratigraphic interpretations of readvances at various locations in Maine. All of these interpretations suffer from a general lack of rigorous understanding of glacier dynamics, a lack of absolute chronology, and limited stratigraphic exposures. However, Leavitt and Perkins' field observations were good, worth consideration, and certainly suggestive.

The first stratigraphic evidence for multiple glaciations in Maine, in contrast to Late Wisconsin glacier fluctuations, was reported from New Sharon in central Maine by Caldwell (1959). At this location he reported two tills separated by a forest litter zone. This was subsequently examined by Borns and Calkin (1977) who reported a radiocarbon date from wood collected at New Sharon at greater than 52,000 yr B.P. (Y-2683). Borns and Calkin also reported other sites displaying undated multiple-till sections. However, recent detailed work by Weddle (this volume) on the exposures at New Sharon shows no evidence of pre-Late Wisconsinan glaciation.

Schafer and Hartshorn (1965), in their synthesis of Quaternary geology of New England, indicated that there had been at least two extensive glaciations of the region based upon their interpretation and that of others of two tills exposed at several locations, largely in southern New England. Koteff and Pessl (1985) extended this till stratigraphy to northern New Hampshire; two equivalent till units representing separate glaciations are believed to occur in southwestern Maine (Thompson and Borns, 1985b). The observations reported in these papers document multiple tills and strongly suggest that there were at least two major episodes of extensive glaciation, the last being of Late Wisconsin age, but a clearer definition of these and possibly other glacial events awaits future research.

LATE WISCONSIN DEGLACIATION

Earlier workers beginning with Stone (1899), and continuing with Leavitt and Perkins (1935), visualized that Maine was last deglaciated from south to north by progressive retreat of the glacier margin. Shafer and Hartshorn (1965) likewise concluded, based upon older as well as some new evidence accumulated since the work of Leavitt and Perkins, that the margin of the last glacier retreated progressively to the north and northwest across New England and into Canada. In addition, they felt, along with Koteff and Pessl (1981), that this was accomplished by "stagnation-zone retreat," a process in which they visualized the progressive development of a stagnation zone of varying widths at the margin of the retreating glacier. Their model was based upon interpretations of the local geology, largely from southern New England, and studies of modern glaciers in Alaska, in particular the Malaspina Glacier. However, there was little new pertinent data available for Maine beyond that presented in 1935 by Leavitt and Perkins and, hence, the marginal retreat model originally suggested by Stone (1899). The regional model of stagnation-zone retreat presented by Shafer and Hartshorn (1965) was indirectly addressed by Borns (1973), who added new field evidence from Maine and re-evaluated the stratigraphic significance of existing radiocarbon dates from the northeast. His interpretation supported the view of a progressive retreat of the glacier margin towards the north and northwest, at least through coastal and central Maine and into Canada. However, the validity of a marginal stagnation zone was still in question. In his conclusion, Borns (1973) cited modern glaciological thinking and pointed out the inadequacy of attempting to correlate glacier margin positions over great distances along the southern and eastern margins of the Laurentide ice sheet.

The long-held "rising window shade" concept of marginal retreat implying progressive retreat of the active ice margin to the north and northwest through New England, the St. Lawrence Lowland, and into central Canada, was generally advocated into modern times (Schafer and Hartshorn, 1965; Borns, 1973). However, an alternative and very different view was expressed by Chalmers (1890, 1899), based upon his work in adjacent southeastern Quebec. He correctly observed evidence that led him to conclude that during the last glaciation, ice flowed from Maine into southeastern Quebec. This indicated for the first time that a local ice sheet occupied the highlands of southeastern Quebec, Maine, and probably New Brunswick. Chalmers named this ice cap the Appalachian Glacier. Unfortunately, his conclusion was not widely accepted and became all but lost with time.

G. H. Stone, working in Maine, certainly had the opportunity to discuss with R. Chalmers the evidence for late ice flowing from Maine into adjacent southeastern Quebec. Apparently with this knowledge Stone (1890) expanded on these ideas, along with his own evidence, and presented his hypothesis for the late-glacial history of the region as follows:

> The above stated hypotheses are consistent with the opinion of Mr. R. Chalmers of the Geological Survey of Canada that from the highlands south of the St. Lawrence River in Quebec the ice flowed north and eastward. This hypothesis would make the valley of the St. John River in Maine the area of accumulation from whence glaciers radiated north, east and south. In a paper on Glacial Erosion in Maine (published in the Proceedings of the Portland Society of Natural History in 1882), I dwelt at some length on the fact that the glaciation of Maine is less intensive in the northern part of the State. This indicates neve-like conditions prevailing over northern Maine for a large part of the glacial period. This conclusion would be consistent with the hypothesis that the radiating flow discovered by Mr. Chalmers continued throughout the whole of the glacial age, or with the hypothesis that it was only a feature of the last days of the ice-sheet. For even if we suppose with Prof. Dana that the highlands near Hudson's Bay were the radiating area during the time of maximum glaciation, it is as yet permissible to suppose that in late glacial time the rising Champlain sea melted its way up the valley of the St. Lawrence, thus isolating the portion of the ice-sheet lying south of that valley. If so, the ice would for a time flow northward from the water-shed of the St. John and from the Notre Dame hills. In other words, late in the ice age northern Maine and the adjacent

territory would for a time be the area of accumulation from whence the ice-flow radiated, no matter what may have been the earlier history of the region. -- G. H. Stone, 1890

The origin of this residual ice cap hypothesis in terms of the contributions from both researchers is not clear. However, it is obvious that their combined observations, conclusions, and experience produced the concept presented in the above quote as early as 1890. Unfortunately, the observations of Chalmers and the hypothesis advocated by Stone for the region were not widely accepted. This concept of a late-glacial residual ice cap over Maine was revived by H. W. Borns (1963) in a paper in which he stated that radiocarbon dates on ice-marginal features in coastal Maine were essentially the same as those for the Champlain Sea. This fact led him to conclude that a remnant ice cap probably existed between the St. Lawrence Lowland and the coast of Maine approximately 13-12,000 years ago. At that time Borns had not discovered the earlier works of Chalmers and Stone on this matter.

Subsequently, workers in Canada (e.g. Chauvin et al., 1985; David and Lebuis, 1985) and in Maine (Kite et al., 1982; Borns, 1985; Lowell, 1985; Newman et al., 1985) have focused upon this "new" ice cap concept, and their work has led to the current understanding of the framework of deglaciation. The greatly increased activity in surficial geological mapping since 1963 and a large body of radiocarbon dates now available from both Maine and Quebec clearly demonstrate that at the time the late-glacial marine invasion had reached the Ottawa River valley, perhaps as early as 12,600-12,800 years ago (Richard, 1978), a residual ice cap in central and northern Maine coexisted with the receding ice-sheet margin lying along the north slope of the St. Lawrence Lowland (Borns, 1985). Thus it is clear that the older "rising window shade" concept for deglaciation at least cannot be applied to the entire northern Appalachian region.

The current model representing the deglaciation of Maine, for example Figure 3, has been presented in publications by Mayewski et al. (1981); Hughes et al. (1985); Jacobson and Davis (1984); and Thompson and Borns (1985a).

ALPINE GLACIATION

Integral with the history of the ice-sheet glaciation of Maine is the associated history of alpine glaciation. As early as 1900 Tarr reported glacial cirques on Mt. Katahdin. He and subsequent workers concluded that these basins were most probably occupied by glaciers during or after the dissipation of the last ice sheet, and since then there has been general acceptance that the basins of Mt. Katahdin are indeed cirques. However, the time of their formation and occupation by alpine glaciers is in question. Caldwell (1966) reported that these cirques held alpine glaciers following retreat of continental ice from the mountain. Davis (1976), the first researcher to produce a surficial geologic map of the mountain, argued that there is no evidence to firmly support Caldwell's conclusions. In 1951, based upon a model of late-glacial climate in the northeast, Flint reported that the highlands of western Maine, including the Boundary Mountains, and also the White Mountains of New Hampshire were probably late-glacial centers of radial ice flow.

However, the work of Borns and Calkin (1977) has clearly demonstrated that the highlands of western Maine did not support either an ice cap or cirque glaciers in late-glacial time, based upon the lack of evidence for post-ice-sheet occupation of the cirques. These authors showed that the cirques were overrun by the last ice sheet, and the mountains emerged first from the ice as deglaciated nunataks. Similar conclusions were reached concerning the White Mountains (Goldthwait, 1970). Clearly, cirques exist in Maine on Mt. Katahdin, Sugarloaf Mountain, and possibly on other mountains (Borns and Calkin, 1977) above the elevations of approximately 800 m. However, the accumulating evidence indicates that these cirques were developed prior to overriding by the Late Wisconsin Laurentide ice sheet and not reoccupied during or after dissipation of the ice sheet. This suggests that by the time these mountains emerged as nunataks the regional snowline had risen to elevations of about 1,500 m or more. This concept has been expanded on a regional basis by Mayewski et al. (1981).

SURFICIAL GEOLOGIC MAPS

The first comprehensive, albeit very limited, surficial geologic map of Maine was compiled by Leavitt and Perkins (1935). Since that time there has been an accelerated program of surficial geologic quadrangle mapping by the Maine Geological Survey, especially during the last two decades. This activity has resulted in the Survey's publication of a new, second-generation Surficial Geologic Map of Maine, co-edited by Thompson and Borns (1985a) and based upon the work of many researchers.

CONCLUSION

The distribution and history of the surficial geology of Maine evolved from a model totally embracing the biblical flood origin, as in C. T. Jackson's First Report on the Geology of Maine in 1837, to a hint of the possibility of some glaciation associated with the flood, in C. H. Hitchcock's report on the Geology of Maine in 1862. However, Dr. John De Laski (1864) was the first researcher who clearly saw and documented that glaciation, and not the great deluge, was responsible for the surficial geology of Maine. Subsequently, origins based upon the flood rapidly disappeared from the literature, and research concentrated upon the regional nature and distribution of glacial processes involved in both sedimentary and erosion features. These studies have led us into modern times, when similar work has continued and provides the data base for modern interpretations of the glacial history and processes that have been active in Maine. Although much is known, there is much yet to learn about the pre-Late Wisconsin history, the maximum extent of ice sheets, the geography and chronology of the last deglaciation, and the nature and

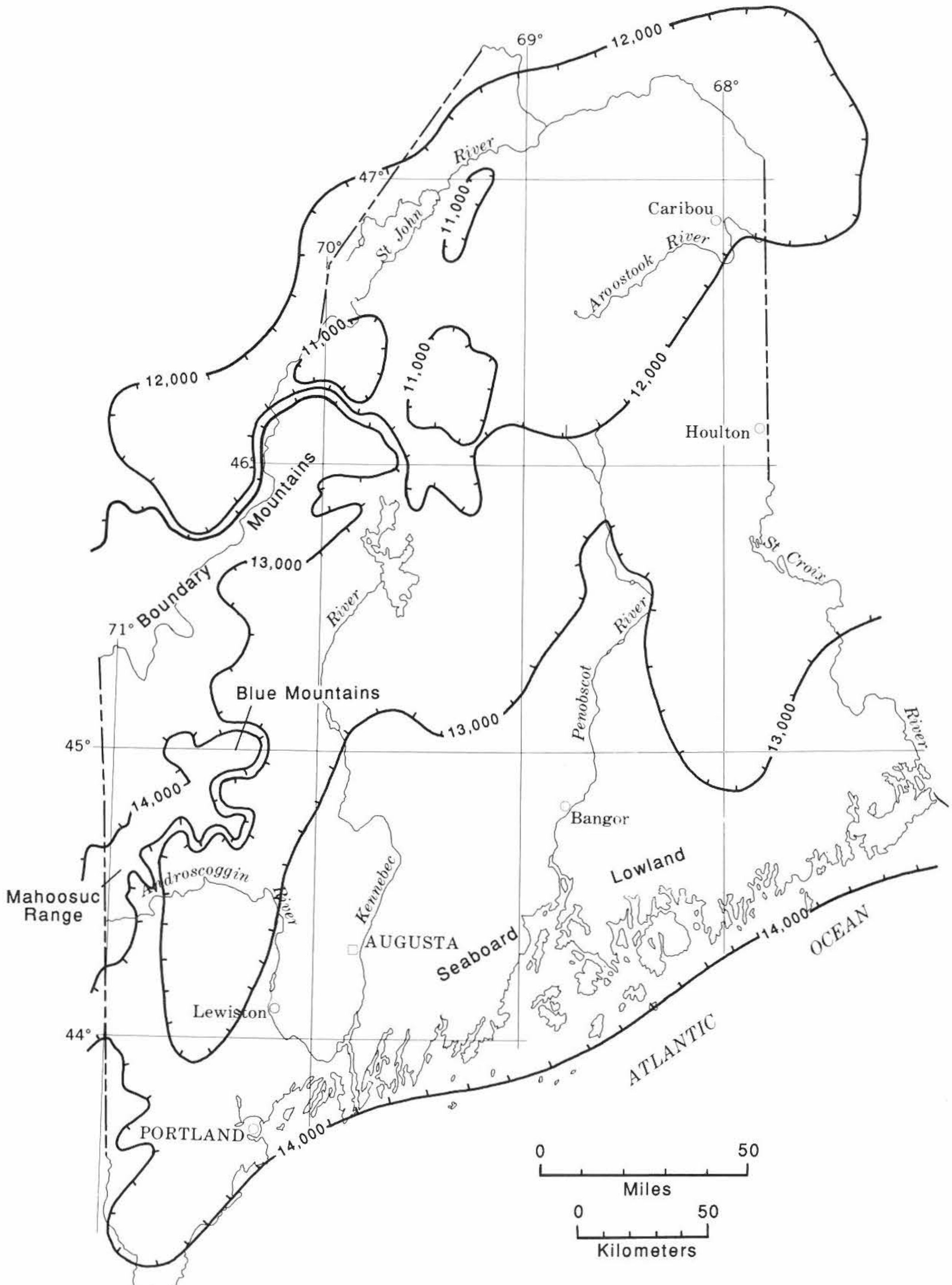

Figure 3. Ice-margin positions during the last deglaciation (from Thompson and Borns, 1985a).

distribution of glacial and glacial-marine sediments. New data, coupled with reconstructions drawn from other Quaternary sciences, will eventually allow the development of a much clearer history of changing late-glacial and postglacial environments which eventually allowed the human occupation of Maine.

To extend our current understanding of the history of the Quaternary Period in Maine, attention should be focused upon expanding the basic knowledge of the sedimentary and stratigraphic record, especially that of pre-Late Wisconsin time, integrating modern glaciological theory into understanding glacial-geological processes in Late Wisconsin time, refining the prevailing but largely generalized deglacial model, and beginning in a concerted way to identify and map the Holocene sediments with the goal of integrating the glacial and non-glacial surficial geology record to understand the entire Quaternary history of Maine.

In total, this data base is becoming an increasingly valuable and necessary tool in making proper environmental decisions in Maine and will provide input into the international effort to predict future global environmental change associated with our current Ice Age.

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Late Wisconsinan Deglaciation of Coastal Maine

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ABSTRACT

Withdrawal of the Late Wisconsinan Laurentide ice sheet from coastal Maine is recorded by a complex assemblage of glacial and glacial-marine sediments and morphologic features (end moraines, deltas, subaqueous fans). This paper summarizes the work of many people who, for the past decade, have studied the glacial geology of the Maine coastal region and have contributed to the development of a general picture of final deglaciation of the State of Maine and adjacent areas.

Detailed examination of stratigraphic sections throughout the Maine coastal zone has led to the development of a working model for glacial-marine sedimentation during Late Wisconsinan deglaciation of the state. Four principal glacial-marine lithofacies types and five major lithofacies associations have been defined to characterize the glacial-marine sedimentary succession in the coastal zone, and to provide the basis for the glacial-marine depositional model. In general terms, this model depicts final deglaciation of the coastal zone as follows: (a) establishment of marine-based ice conditions during deglaciation; (b) deposition of sediments by both ice-dominated and water-dominated processes; (c) rapid sedimentation in an environment characterized predominantly by subaqueous fan formation and remobilization of sediments by gravity flow mechanisms; (d) stillstands of the retreating ice front to produce partial- and fully-developed deltas; (e) periodic fluctuations of the ice margin (grounding line) to produce end moraines; and (f) gradual transition from marine-based to terrestrially-based deposition of glaciogenic sediments.

End moraines of a variety of forms outline the general pattern of ice retreat from the coastal zone. The most important of these moraines are DeGeer moraines. Both large stratified moraines and DeGeer moraines record minor fluctuations (or stillstands) of the ice margin and either depositional or glacial-tectonic thickening of the "normal" sedimentary succession.

The chronology of deglaciation of the coastal zone is currently an issue of some debate. Several chronologies and their implications are discussed. Although there is presently no clear consensus on the timing of the events of final deglaciation, it is now generally accepted that ice retreat from the coastal zone was not interrupted by significant glacial readvance.

INTRODUCTION

Over the past decade, field study of the surficial geology of coastal Maine has provided an extensive body of information on the nature of Late Pleistocene (in particular, Late Wisconsinan) glaciation and deglaciation of the state. This work has led to the

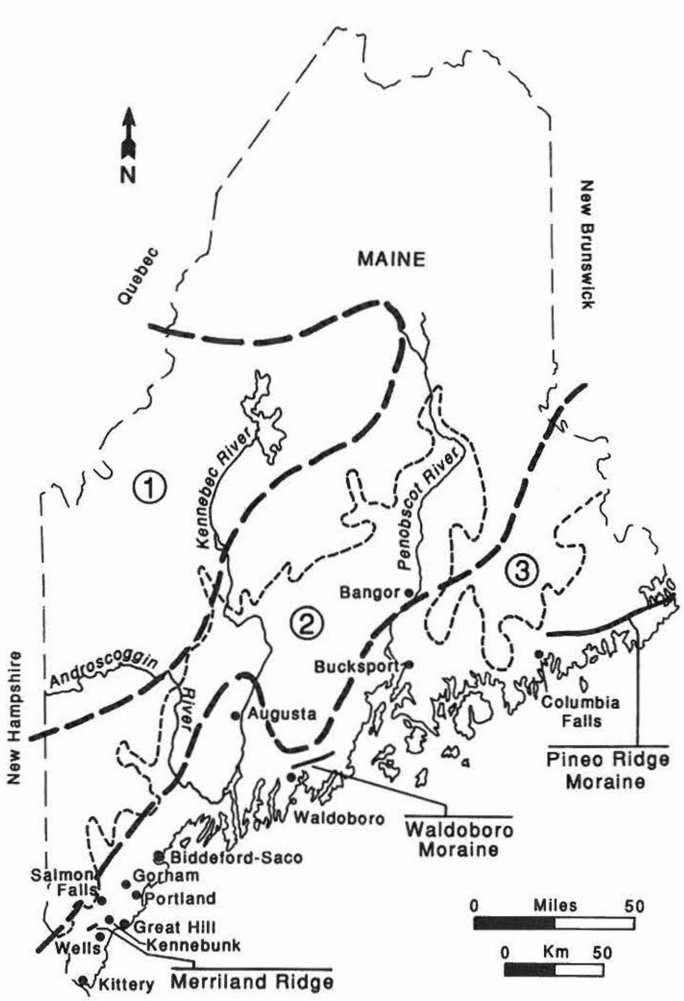

Figure 1. Location map of the Maine coastal zone. Major physiographic sections (dashed lines) include: (1) White Mountains, (2) New England Upland, (3) Seaboard Lowland. Short dashed line is inland limit of late-glacial marine submergence.

development of the first clear picture of the glacial history of a large portion of northern New England and adjacent parts of Canada (Larson and Stone, 1982; Borns et al., 1985), and has provided important information regarding the processes of glacial-marine sedimentation, a topic of substantial current interest.

The objective of this paper is to provide a synthesis of the work of many individuals who have been involved with the glacial geology of coastal Maine, in large part under the auspices of the Maine Geological Survey.

GEOLOGIC SETTING OF COASTAL MAINE

Coastal Maine is situated in the New England Province of the Appalachian Highlands (Fig. 1). The inland portion of the coastal region lies within the New England Upland Section, while the seaward portion occupies the Seaboard Lowland Section (Thornbury, 1965). The Seaboard Lowland is generally coincident with the area of late-glacial marine submergence (Fig. 1). As a result, this section displays less topographic relief than does the New England Upland. Present elevations within the coastal region are generally below 90 m, but rise inland to altitudes in excess of 400 m.

The entire coastal region is underlain by a variety of intrusive igneous rocks and complexly deformed metamorphic rocks that dip steeply and strike in a general northeast-southwest direction (Osberg et al., 1985). The dominant structural grain is clearly reflected both in the present topography of the upland areas and in the courses of many streams that drain the coastal lowland. This structural control is also evident in the orientation of the many embayments along the present coastline in the central coastal region.

During the Late Wisconsinan episode of glaciation, ice advanced from the northwest across Maine to a terminal position on the continental shelf (Borns, 1973). Glacial erosion produced a distinct northwest-southeast lineation, superimposed upon the northeast-southwest structural grain (Thompson and Borns, 1985a). Streamlined erosional forms are common, and many valleys paralleling the direction of ice movement display the effects of erosional deepening and steepening of valley sides. Glacial deposition resulted in a general reduction of preglacial relief by preferential infilling of valleys. This effect is most pronounced in that portion of the coastal zone below the limit of late-glacial marine submergence.

Withdrawal of Late Wisconsinan ice from its terminal position was underway between 17,000 and 15,000 yr B.P., and the ice had retreated across the Gulf of Maine to a position roughly parallel to, but some distance offshore of, the present coastline by 14,000 yr B.P. (Fastook and Hughes, 1982; Smith, 1985; see Fig. 25). Ice retreat, accompanied by marine submergence, progressed rapidly across the coastal region of Maine. End moraines, in a variety of forms, were produced at or near the ice front during the period of retreat, and outline in detail the pattern of ice withdrawal from the coastal zone (Smith, 1981, 1982; see Fig. 22).

GLACIAL STRATIGRAPHY OF COASTAL MAINE

The glacial deposits of coastal Maine have been described in terms of a generalized stratigraphic succession (Fig. 2) developed for the purpose of reconnaissance field mapping. This stratigraphy is discussed by Smith (1982, 1985) and has been employed by other workers in the coastal region of Maine (e.g.,Thompson, 1979). In very general terms, this stratigraphic succession includes, in ascending stratigraphic order, till, ice-contact stratified drift, subaqueous outwash, silt and clay of the marine Presumpscot Formation, and subaerial outwash (delta).

This original stratigraphy was closely tied to the conventional subdivision of glaciogenic sediments into direct glacial, ice-contact (ice-marginal), and proglacial deposits and processes. Within this stratigraphic framework, most till deposits are described as lodgement till and are considered to have been

deposited subglacially during ice advance and retreat. For the most part, then, these materials represent the oldest (lowest) unit within the coastal stratigraphic succession. Locally, lodgement till is found as a carapace over the proximal slopes of large moraines and overlying ice-contact sediments and subaqueous outwash. This relationship implies local readvance of the retreating ice margin. In addition to lodgement till, other genetic till types (flow till, melt-out till) are recognized within the coastal sedimentary succession.

Ice-contact stratified drift includes stratified deposits of sand and gravel that display collapse deformation. In general, these materials, where they occur below the marine limit, comprise eskers and morainal sediments that overlie till and underlie silt and sand of the Presumpscot Formation. Ice-contact sediments are recognized to be much more extensive above the marine limit where they occur as valley fill sequences and associated large esker systems. Distal portions of these ice-contact deposits form the heads of subaerial outwash plains that are transitional to deltas and fans constructed into the late-glacial sea. The subaerial outwash (delta) units are considered to be the youngest (highest) materials in the coastal stratigraphy, both overlying and intertonguing with sediments of the Presumpscot Formation.

The term "subaqueous outwash" (Rust and Romanelli, 1975) has been applied (Smith et al., 1979) to deposits of sand and gravel generally beneath, but often intertonguing with, sediments of the Presumpscot Formation. Common throughout the coastal region, this unit is thickest and coarsest in proximity to eskers. It is considered to have been deposited by meltwater discharge from the grounding line of the retreating ice sheet. Very often, the subaqueous outwash deposits comprise the cores of small moraines and are severely deformed by glacial-tectonic shearing and folding.

Fine sand, silt, and clay of the Presumpscot Formation of Bloom (1960, 1963) is recognized as the most extensive surface material in the area of coastal Maine below the limit of marine

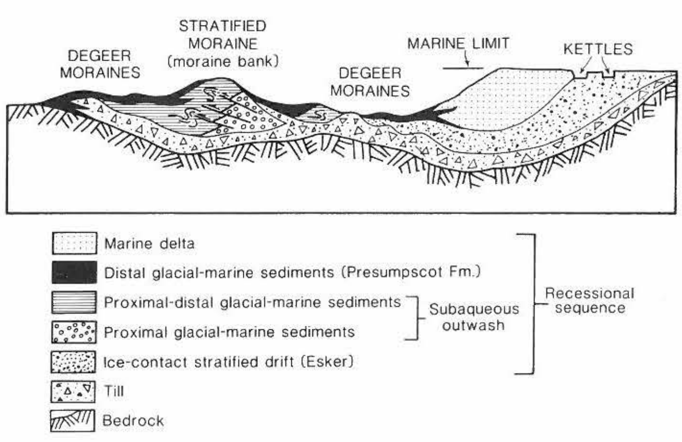

Figure 2. Generalized stratigraphy of Late Wisconsinan glacial deposits of coastal Maine (after Smith, 1985).

submergence. These sediments are considered to be quiet water marine deposits, the most distal of the glaciogenic sediments. The faunal assemblage within the Presumpscot sediments indicates deposition in initial cool (subarctic), deep marine waters with gradual shoaling and the accumulation of organic tidal flat sediments. Throughout much of coastal Maine, marine silt and clay of the Presumpscot Formation grades upward to fine sand that is locally reworked into eolian dunes (e.g., the Desert of Maine). The sand may record gradual shoaling during coastal emergence, or it may be the distal facies of outwash deposits, or it may be both.

More recent and detailed study of the glacial sediments in the Maine coastal zone (Smith, 1984a, 1984b, 1988; Hunter, in prep.; Hunter and Smith, 1988; Retelle and Konecki, 1986; Retelle and Bither, this volume) indicates that the glacial sedimentary succession is more complex than was originally thought. This is particularly the case for the widespread glacial-marine succession in the coastal region. A provisional sedimentary facies model has been developed to better accomodate our current understanding of the glacial-marine sediments. Pertinent aspects of this model will be dealt with in the subsequent discussion of glacial-marine lithofacies and lithofacies associations.

DEGLACIATION OF COASTAL MAINE

General Processes of Glacial-Marine Deposition

Within the glacial-marine setting that persisted as ice withdrew across the coastal zone, sediments were deposited by a variety of complexly interrelated processes (Smith, 1984a, 1984b). The general nature of these processes can be inferred from the sediments themselves and by analogy to processes described by other workers in other areas (e.g., Powell, 1980, 1981, 1983, 1984; Domack, 1983; Molnia, 1983; Mackiewicz et al., 1984).

During ice retreat and coastal submergence, sediments accumulated under two general depositional regimes: (1) ice-dominated and (2) water-dominated. Interestingly, the overwhelming volume of sediments in coastal Maine appears to have been deposited by (melt)water-dominated processes. Under both regimes, deposition of sediment was influenced by the juxtaposition of glacial ice and meltwater (an active depositional subsetting) and standing marine water (a passive depositional subsetting). The general array of depositional processes is illustrated in Figure 3.

In the ice-dominated environment, general depositional processes included: (a) subglacial lodgement, (b) subglacial melt-out, (c) brash deposition at the ice front, (d) debris flow deposition, and (e) brash deposition from rafted bergs. Several factors influenced the relative contributions of each of these processes. Among these factors were: (a) bathymetry, (b) configuration of the ice front, (c) rate and volume of sediment influx,

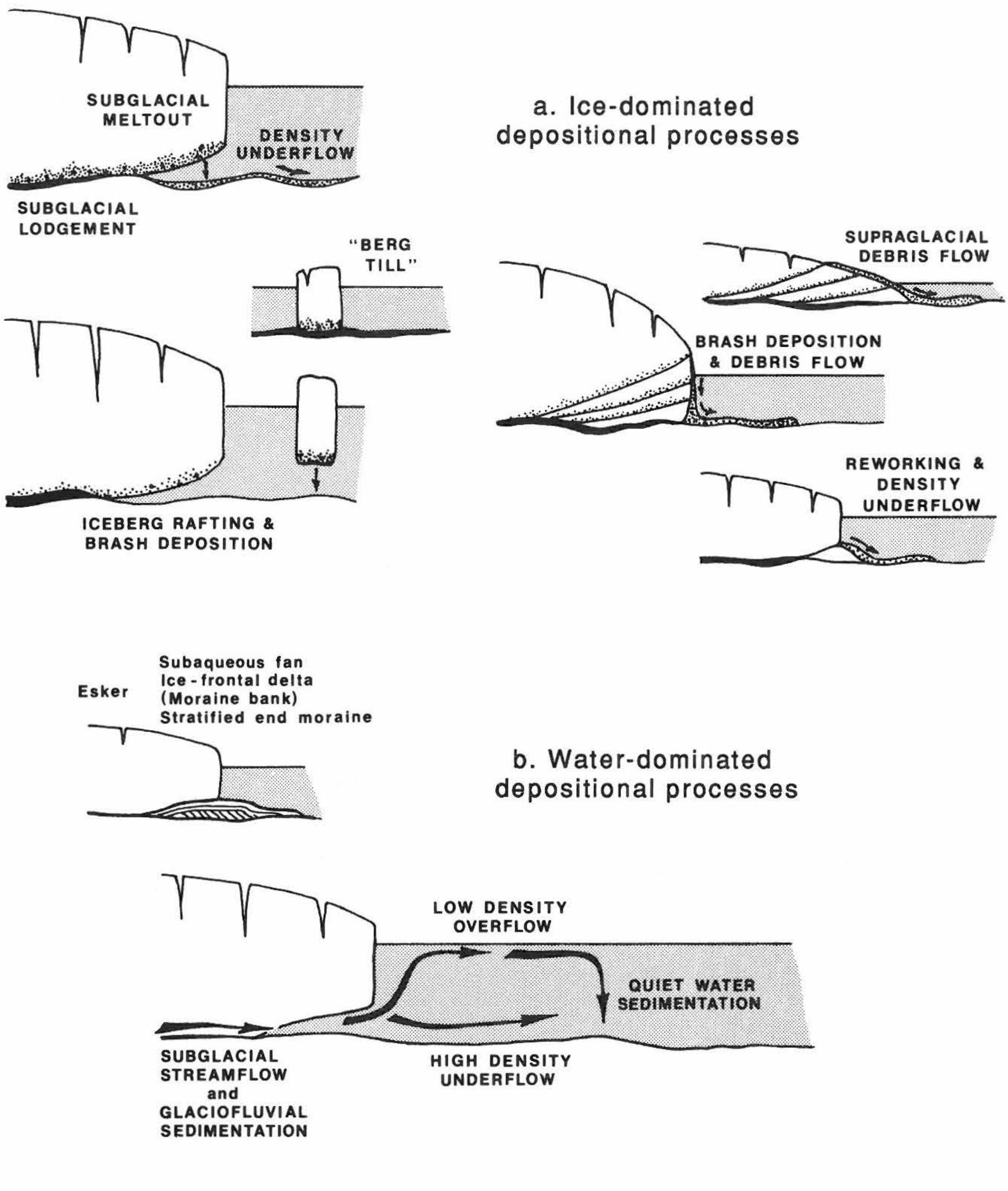

Figure 3. Processes of glacial-marine sediment deposition (from Smith, 1984b). (a) Ice-dominated depositional processes. (b) Water-dominated depositional processes.

(d) rate and nature of ice retreat, and (e) occurrence and distribution of ice-frontal features.

In the water-dominated environment, depositional processes included: (a) subglacial streamflow and glaciofluvial sedimentation, (b) ice-frontal overflow and interflow and quiet water deposition, and (c) ice-frontal density underflow. The factors influencing the relative importance of each process were much the same as those for the ice-dominated environment. In addition, the processes operating in each environment overlapped, and the relative roles of each varied in accord with these same factors.

In both depositional environments, but particularly in the ice-dominated environment, reworking of sediments would very likely have occurred as a result of the initiation of sediment gravity flows in saturated, quasi-stable deposits.

In the late stages of marine submergence and marine regression, previously deposited sediments were modified by beach and coastal processes (not illustrated in Fig. 3).

Glacial-Marine Lithofacies

On the basis of detailed examination of exposures throughout the coastal zone, several lithofacies types have been defined to describe the complex assemblage of glacial-marine sediments. A summary of these lithofacies types is provided in Table 1. This listing should not be considered exhaustive. It is, however, representative of the more common lithofacies types in the coastal zone. The description of lithofacies types presented here employs a classification scheme modified from Miall (1978) and Eyles et al. (1983). Within this scheme, we describe four principal lithofacies types: diamict (D), gravel (G), sand (S), and fine-grained sediment (F). For each of the principal lithofacies types, there are several subtypes, defined on the basis of the nature of sediment support and the internal structure or bedding characteristics of the sediment. The physical attributes of the principal lithofacies types are outlined below. Descriptions of subtypes and a more detailed discussion of the basis for genetic interpretations of sediments are presented in Smith and Socci (in prep.).

Diamict Lithofacies. Diamicts within the coastal zone are highly variable, both texturally and compositionally. Commonly, the diamicts consist of admixtures of silt or sand matrix and cobble to boulder clasts that are predominately subangular to subround. Clast composition is dependent upon underlying bedrock, and varies from scattered, small platy clasts of argillite to numerous large boulders of granite. Clast shapes include equant, faceted, and elongate ("bullets"). Clasts are also commonly striated. In general, diamicts display a well-developed fabric of elongate clasts that parallels the direction of ice movement inferred from striations on adjacent bedrock outcrops. Where diamicts have a silty matrix, they generally display a well-developed subhorizontal fissility. Diamicts, when they are exposed in end moraines, may be moderately to intensely sheared.

TABLE 1. LITHOFACIES TYPES OF GLACIAL-MARINE DEPOSITS IN COASTAL MAINE

Diamicts:	Dmm	matrix-supported, massive
	Dms	matrix-supported, stratified
	Dcm	clast-supported, massive
	Dcs	clast-supported, stratified
	Dcg	clast-supported, graded
Gravels:	Gmm	matrix-supported, massive
	Gms	matrix-supported, stratified
	Gmg	matrix-supported, graded
	Gmt	matrix-supported, trough cross-stratified
	Gcm	clast-supported, massive
	Gcs	clast-supported, stratified
	Gcg	clast-supported, graded
	Gct	clast-supported, trough cross-stratified
Sands:	Sm	massive
	Ss	stratified
	St	trough cross-stratified
	Sr	ripple laminated
	Sg	graded
Fine-grained Sediments:	Fm	massive
	Fl	laminated
Genetic Interpretations:	--d	soft sediment deformation
	--(d)	dropstones
	--(s)	sheared

Figure 4. Dmm (massive, matrix-supported diamict) lithofacies, vicinity of Salmon Falls, York County. Diamict of diamicton/proximal fan association truncates stratified sand (Ss) lithofacies of subaqueous fan association. Erosional contact between diamict and underlying sand is characteristic of the diamict lithofacies. Shovel for scale.

For the most part, diamicts are matrix-supported (Dm; Fig. 4), though rare clast-supported diamicts (Dc) have been observed in local exposures in end moraines. In addition, most diamicts are massive (D_m). However, both stratified (D_s; Fig. 5) and crudely graded (D_g) diamicts do occur within the glacial-marine succession.

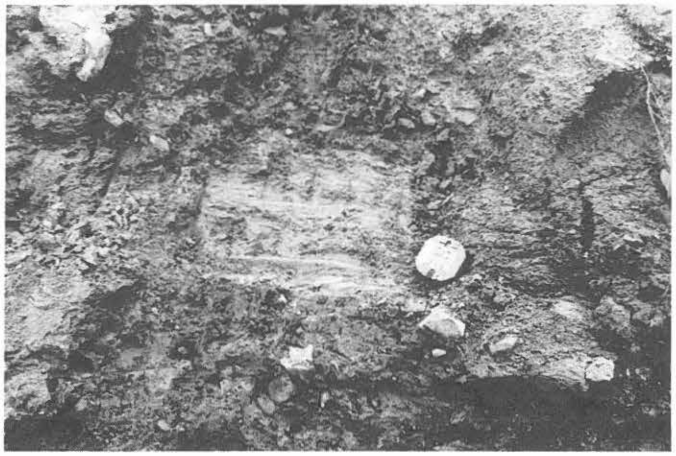

Figure 5. Dms (stratified matrix-supported diamict) lithofacies, Great Hill, York County. Compass for scale.

Figure 7. Gcm (massive clast-supported gravel) lithofacies, vicinity of Kennebunk, York County. Trowel for scale.

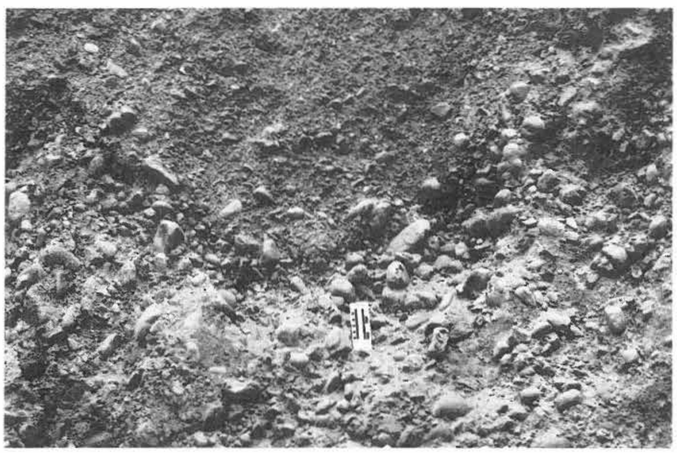

Figure 6. Gmm (massive, matrix-supported gravel) lithofacies, vicinity of Salmon Falls, York County. Upper pebble and cobble gravel in erosional contact with lower cobble gravel. Both gravel units are representative of the proximal subaqueous fan association.

Contact relationships between diamicts and other lithofacies types are virtually always sharp and well-defined. Basal contacts are most commonly erosional, sometimes showing relief of a meter or more. Upper contacts are generally abrupt except in instances where clast-rich diamicts are overlain by poorly-sorted or massive cobble or boulder gravels. In this latter situation, sediments of the diamicton/proximal fan association (see subsequent discussion) record deposition by a complex array of ice-frontal processes, including subglacial lodgement, dumping, melt-out, flowage, and ice-proximal fluvial sedimentation. Diamicts deposited in this setting are clearly gradational to sediments of the subaqueous fan association, and it is often difficult to distinguish between the two depositional regimes.

Considerations of texture, internal structure (including fabric), and contact relationships indicate that most diamicts are subglacial lodgement tills, deposited either during the last (Late Wisconsinan) glacial advance or during minor readvances of ice during general glacial retreat. Stratified diamicts are more reasonably interpreted as subglacial melt-out tills or remobilized tills. These latter materials may include flow tills.

Gravel Lithofacies. Gravels within the coastal glacial-marine succession range in texture from pebbles to boulders with fine sand to pebble matrix. Sorting is moderate to very good. Internal structures and bedding characteristics range from massive to distinctly stratified, and include cross-stratification and both normal and reverse grading.

Most gravel units are matrix-supported (Gm), and include both massive (Gmm; Fig. 6) and stratified (Gms) sediments. Locally, clast-supported gravels (Gc; Fig. 7) occur in proximal portions of moraines, deltas, and subaqueous fans. These gravels are well sorted cobble units that are typically normally graded (Gcg), though they may also display reverse grading.

Geometry of bedding in the gravel units is commonly diagnostic of their sedimentary origin. Within morainal successions, gravels are commonly coarse, massive, and intermixed with till and sand. The gravels in these situations generally have no clearly definable geometry, although in some cases they tend toward lenticularity and commonly have erosional basal contacts. These materials are interpreted as proximal fan or debris flow sediments.

The gravels described above commonly grade laterally (down-ice ?) to better sorted units that are interstratified with well-sorted sands. These lenticular or single-bed gravels are well-stratified, may display crude cross-stratification, and typically dip away from inferred ice-frontal positions at angles of 5-20 degrees (Fig. 8). These sediments are considered to be mid-fan to distal-fan deposits.

Figure 8. Subaqueous (mid-)fan lithofacies, vicinity of West Gorham, Cumberland County. Massive and stratified gravels (G_m, G_s) interbedded with stratified and cross-stratified sands (Ss, St). Shovel for scale.

Figure 10. Chevron folding produced by glacial-tectonic shearing of sand lithofacies (Ss), vicinity of Salmon Falls, York County. Sediments of subaqueous fan association have been deformed by local glacier overriding.

Figure 9. Convolution of massive and stratified sand (Sm, Ss) lithofacies produced by rapid sediment dewatering, vicinity of Kittery, York County.

Locally, gravel "mounds" occur within exposed sequences of interbedded gravel and sand. Gravels in these features display sedimentary features (cross-stratification, imbrication, cut-and-fill) that indicate deposition by fluvial processes. Gravels that occur in these situations are considered to have formed in positions of subglacial tunnel openings and subaqueous fan cores.

Massive to crudely cross-stratified pebble and cobble gravels overlie well-defined foreset beds in sediments that are considered (by virtue of internal sedimentary features and surface morphology) to be deltaic deposits. These gravels tend to concentrate along the western and northern margins of late-glacial marine submergence, and are considered to be topset beds of delta sediments that accumulated at or near the ice margin when ice was at (or close to) the late-glacial marine limit.

Coarse gravel lithofacies, particularly those that occur in association with diamicts, typically have erosional basal contacts, though they may locally display gradational contact with underlying clast-rich diamicts. These materials are considered to be ice-proximal debris flow sediments that include both proximal remobilized tills and proximal subaqueous fan deposits. In many instances, these gravels grade laterally (in an inferred down-ice direction) to well-sorted gravels that are interbedded with sand lithofacies. Both coarse and fine gravels not associated with diamicts tend to have planar or gradational contacts with adjacent sediments that are generally sand. These are sediments of the mid-fan and distal-fan associations (see subsequent discussion).

Sand Lithofacies. Sand lithofacies range from fine grained to very coarse grained. Lithofacies subtypes include massive (Sm), stratified (Ss), trough cross-stratified (St) and ripple laminated (Sr). Soft sediment deformation (S_d), representing both downslope movement of saturated material and rapid sediment dewatering (Fig. 9), occurs commonly in sand lithofacies. Where sand lithofacies are incorporated into moraines, sediments often display significant glacial-tectonic deformation, both low-angle shearing (often manifested as chevron folds) and folding (Fig. 10).

The majority of sand units are either massive (Sm) or planar stratified (Ss). Sands are most generally interstratified with either gravel lithofacies (G) or with fine-grained lithofacies (F). Sand lithofacies tend to thicken away from inferred sources of sediment supply (ice-frontal positions), and in distal occurrences may be several meters thick. Thickest sand units are typically massive and interbedded with silt and clay of the fine-grained lithofacies (Presumpscot Formation).

Horizontally stratified sand facies (Ss) consist of thin bedded (or laminated) to thick-bedded units that are either mas-

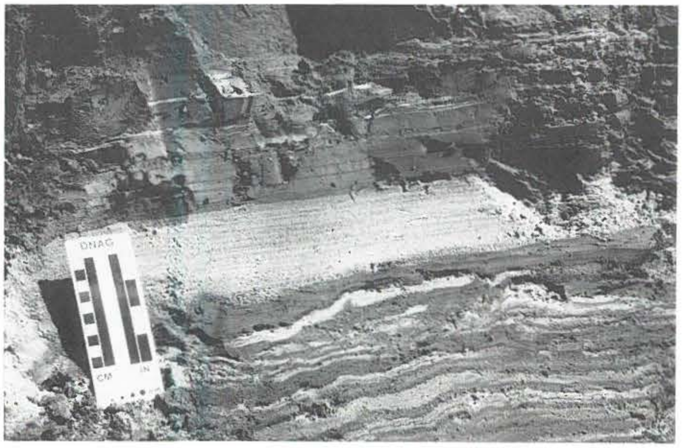

Figure 11. Interbedded fine sand (Ss) and silt (Fl) of the distal subaqueous fan/marine (glacial-marine) mud associations, vicinity of Gorham, Cumberland County. Sand and silt lithofacies overlain by laminated silt (Presumpscot Formation) that grades upward to massive silt (Fm) lithofacies.

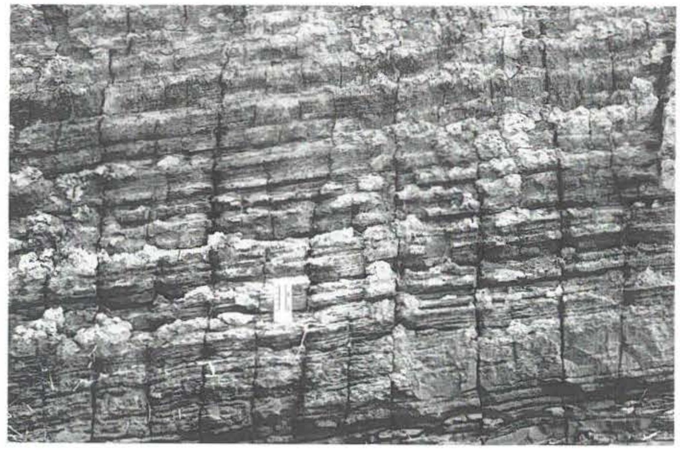

Figure 12. Laminated silt (Fl) of the marine (glacial-marine) mud lithofacies association, vicinity of Portland, Cumberland County.

sive or, more commonly, normally graded within strata. This facies is commonly interbedded with silt and clay of the fine-grained lithofacies (Presumpscot Formation; Fig. 11).

Contacts between sands and other lithofacies are generally sharp, and either planar or erosional. Not uncommonly, basal contacts between coarse sands and gravels are gradational. Also, gravel units commonly grade laterally (both normal to and parallel to inferred ice-frontal positions) to coarse and fine sands. Contacts within multi-storied sand sequences are planar, erosional, or gradational. Planar and erosional contacts within sand sequences are commonly defined by pebble horizons. Glacial-tectonic deformation of sand lithofacies is common (Fig. 10).

Based upon analysis of sediment texture, internal sedimentary structures, bedding contact relationships, and bed geometry, most sediments of the sand lithofacies are considered to be mid-fan to distal fan deposits. Some sand units, particularly those that are interstratified with fine-grained lithofacies, are considered to be more distal fan and subaqueous plain deposits.

Fine-Grained Lithofacies. Sediments of the fine-grained lithofacies include fine sand, silt, and clay. Virtually all of the units included in this lithofacies can be ascribed to the Presumpscot Formation, as originally defined by Bloom (1960, 1963). Assignment of these materials to the Presumpscot Formation is based upon lateral continuity between sand, silt, and clay units and sediments found in the type localities used by Bloom to define the Presumpscot Formation.

Fine-grained lithofacies are either massive (Fm) or laminated (Fl; Fig. 12). In general, laminated silt and sand grade laterally (in a distal sense) and vertically upward to more massive units. Also, massive silt and silty clay often grade upward to massive or bedded sand. Laminated fine-grained sediments are typically normally graded within laminae.

TABLE 2. COMMON LITHOFACIES ASSOCIATIONS OF COASTAL MAINE

DPF	Diamict/Proximal Fan	
SF	Subaqueous Fan:	(p) proximal fan
		(m) mid-fan
		(d) distal fan
MM	Marine (Glacial-Marine) Mud	
SEM	Subaqueous End Moraine	
GMD	Glacial-Marine Delta	

Contacts between fine-grained lithofacies and coarser lithofacies (diamict and gravel) are abrupt. Contacts with sand lithofacies are commonly gradational, though they may be abrupt and truncate bedding within sand units. Soft sediment deformation and dropstones occur within the fine-grained lithofacies. Locally, fine-grained sediments are glacial-tectonically deformed.

Glacial-Marine Lithofacies Associations

For purposes of discussion and interpretation of the sedimentary units (lithofacies) mapped in the Maine coastal zone, lithofacies types have been grouped into five broad categories, or associations (Table 2). Lithofacies within any association commonly occur together in exposure and are considered to have genetic affinities. That is to say, that each lithofacies association (and its lithofacies types) records sediment deposition under a unique or diagnostic sedimentary regime. The general lithofacies associations are summarized above (Table 2).

Diamict/Proximal Fan Association. This association consists predominantly of massive matrix-supported diamicts (Dmm) with subordinate amounts of stratified diamicts (Dms) and both massive and stratified clast-supported gravels (Gcm,

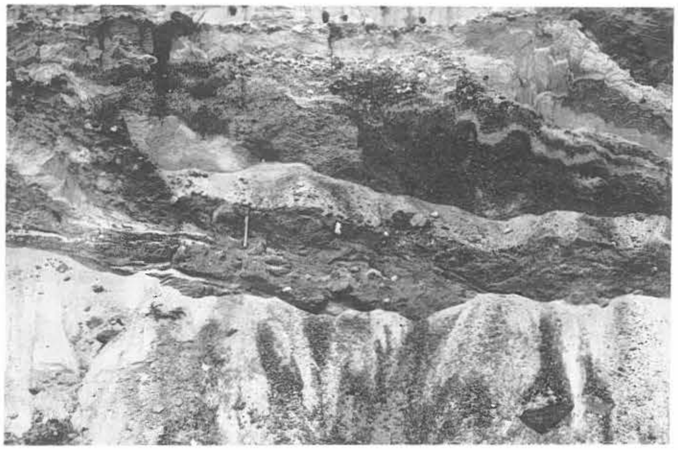

Figure 13. Diamict (Dmm) of the diamict/proximal fan association interbedded with gravel (Gcm), and sand (Ss, Sm) of the proximal subaqueous fan association, vicinity of Salmon Falls, York County. Shovel for scale.

Figure 14. Deformed gravel (Gcs) lithofacies surrounded by sheared diamict (Dmm) lithofacies, vicinity of Columbia Falls, Washington County. Sediments exposed in core of large end moraine include lithofacies of both the diamicton/proximal fan association and the proximal subaqueous fan association.

Gcs). Matrix-supported diamicts are silt/sand-rich, and contain subround to subangular clasts that range in size from pebbles to boulders and that show a large range of lithologic variability. Basal contacts of these units are typically erosional (Fig. 4), and the sediments commonly display internal shearing (D__(s)). Locally, stratified diamicts (Dms) and clast-supported gravels (Gc) overlie massive diamicts with gradational contact.

Within the coastal zone, sediments of the diamict/proximal fan association are commonly exposed in end moraines, and most generally occur on proximal slopes of these moraines. The matrix-supported diamicts of this association are interpreted to be subglacial lodgement tills, deposited during minor readvance of the terminus (or grounding line) of the retreating ice sheet. Stratified diamicts and clast-supported gravels are thought to record subglacial melt-out and resedimentation of previously deposited materials, as well as fluvial deposition of proximal fan sediments.

Subaqueous Fan Association. The subaqueous fan association consists of three sub-associations that record a continuum of sediment deposition within the range of proximal to distal fan environments. The proximal fan sub-association is dominated by lithotypes that reflect deposition in proximity to ice under very high energy fluvial regimes. Coarse matrix- and clast-supported gravels (Gm, Gc), predominantly massive (G_m) or crudely stratified (G_s; Fig. 13), are commonly interbedded with facies of the diamict/proximal fan association (Fig. 14), as well as with remobilized sediments (sediment gravity flows) of both the diamict/proximal fan association and the subaqueous fan association. Basal contacts of sedimentary units within this association are typically erosional.

Proximal fan sediments grade distally (away from inferred ice-frontal positions) and laterally (parallel to ice-frontal positions) to better sorted and finer grained sediments. The mid-fan

Figure 15. Inclined beds of pebble and cobble gravel (Gcm, Gcs) of proximal subaqueous fan association, vicinity of Salmon Falls, York County. Cobble gravels display crude normal grading. Shovel for scale.

association is characterized by sediments that range in size from pebble and cobble gravel to fine to coarse sand. Sedimentary units may be massive (G_m, Sm), but are more typically stratified (G_s, Ss; Fig. 15). Most units are clast-supported (Gc, S_). Basal contacts of units in this sub-association are planar or gradational. Gravel units of the mid-fan association may display

Figure 16. Ripple laminated sand (Sr) lithofacies of the mid to distal subaqueous fan association, vicinity of Kittery, York County.

crude cross-stratification (Gct). Sand units include plane bedded (Ss), cross-stratified (St), and ripple laminated sediments (Sr; Fig. 16). Sand units are often massive (Sm) in both upper and distal portions of the mid-fan sub-association. Sedimentary units may be horizontal, but more commonly dip at angles of 5 to 20 degrees away from the inferred position of the ice front.

Distal fan sediments are essentially transitional between mid-fan sediments and deposits of the marine (glacial-marine) mud association. They consist of coarse to fine sand and silt, and are typically interbedded with fine sand, silt, and clay (Fig. 11). Massive sands (Sm) that commonly display dewatering deformation grade upward (or distally) to thinly bedded sands that are planar stratified (Ss) or ripple laminated (Sr). These sediments, in turn, grade upward (or distally) to massive (Fm) or laminated (Fl) silt and clay. Graded units are common in the transition between the distal fan sub-association and the glacial (glacial-marine) mud association. Dropstones are likewise found (occasionally) in sediments of this sub-association.

Marine (Glacial-Marine) Mud Association. The marine (glacial-marine) mud association is, in general, equivalent to the Presumpscot Formation as defined by Bloom (1960, 1963). Lithotypes comprising this association include massive (Fm; Fig. 11) and laminated (Fl; Fig. 12) silt and clay and massive (Sm) and stratified (Ss) fine sand. Where the contact between the distal fan sediments and the marine mud sediments is gradational, sedimentary units are commonly graded, contacts are planar, and the marine muds are intimately interstratified with distal sands of the subaqueous fan association. Where the contact is abrupt between these associations, the marine mud is generally massive, and it truncates stratification within the distal-fan sands (Fig. 17).

Outsized clasts, many of which can be identified as dropstones by virtue of disrupted bedding relationships, occur within the marine (glacial-marine) mud association. Where sediments of this association are incorporated into the end moraine association, they will often display glacial-tectonic shearing and folding.

Subaqueous End Moraine Association. This association can, and most often does, include sediments of each of the preceding associations. It should be recognized that this association is less of a sedimentary association than it is a glacial-tectonic association. The sediments that comprise this association consist of whatever material is available to be incorporated into an end moraine during a period of glacial readvance. General models of the subaqueous end moraine association have been proposed by Smith (1981, 1982; Smith et al., 1979), Thompson and Borns (1985b), Retelle and Konecki (1986), and Retelle and Bither (this volume). In general, cores of end moraines consist of gravel and sand of the subaqueous fan association that have been glacial-tectonically deformed (both by shearing and by folding; Fig. 18). These sediments are overlain, on proximal slopes of moraines, by sediments of the diamict/proximal fan association, and grade distally, or are overlain on distal slopes, by deposits of the marine (glacial-marine) mud association. Often, remobilization of sediments of all associations complicates the local stratigraphy of this association.

Glacial-Marine Delta Association. Along the inland limit of marine submergence, subaqueous fan associations give way to glacial-marine delta associations. By-and-large the lithotypes involved in the two associations are very much the same. And, in fact, there is a regular gradation from fans through partially-developed deltas (Gluckert, 1975) to fully-developed deltas in the transition from marine to terrestrial depositional settings (Fig. 19).

Sediments of the glacial-marine delta association typically consist of gravel, sand, and silt in topset, foreset, and bottomset

Figure 17. Laminated silt (Fl) of the marine (glacial-marine) mud association in sharp contact with sand (Ss, Sm) of the distal subaqueous fan association, vicinity of Kittery, York County. The sand has been deformed by rapid dewatering. Between the deformed sand and the laminated silt is a zone of brecciated fine sand and silt. Brecciation of this unit was probably also the result of rapid dewatering.

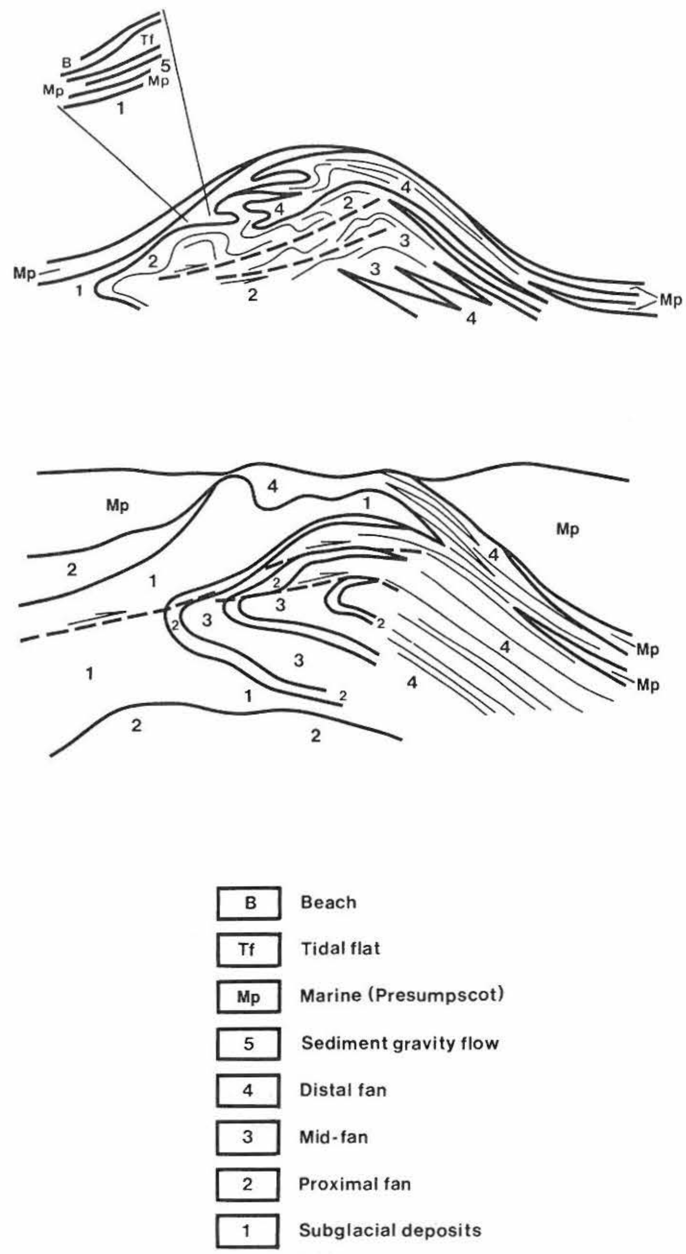

Figure 18. Schematic cross-sections of two coastal end moraines illustrating general stratigaphic relationships and nature of glacial-tectonic deformation. Dashed lines represent major shear planes. Proximal slopes of both moraines are to the left.

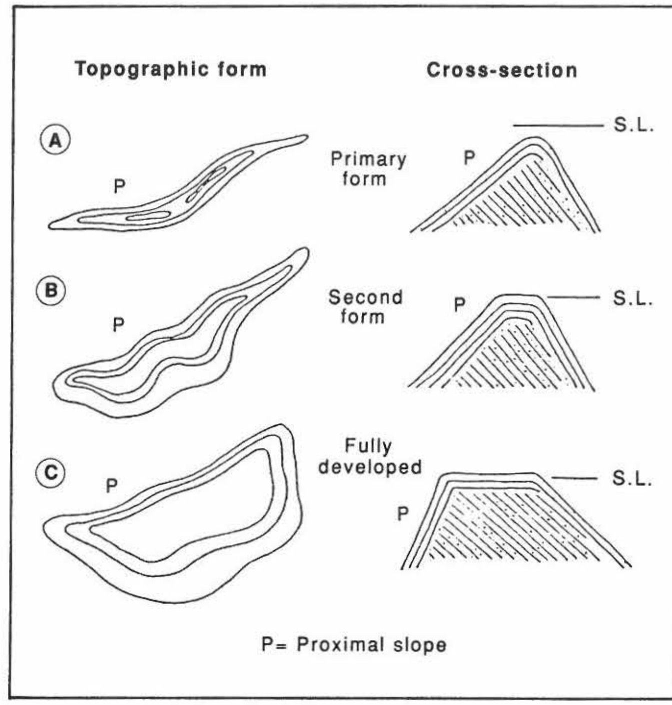

Figure 19. Varieties of ice-frontal delta. (a) Primary form: sharp-crested ridge, not constructed to sea level. (b) Second form: narrow flat-topped ridge, constructed to sea level. (c) Fully-developed delta: broad, flat-topped surface, constructed to sea level (from Gluckert, 1975).

successions. Dominant lithotypes in most exposed delta sections include fine to coarse sand that may be massive (Sm), horizontally stratified (Ss), trough cross-stratified (St), or rippled (Sr). These sediments generally dip at angles of 15 to 30 degrees away from the delta source. Thin gravel beds that may be either massive (G_m) or graded (G_g) often occur within foreset successions.

Foreset beds within deltas are overlain by pebble to cobble gravel topsets that are generally matrix-supported massive (Gmm) or trough cross-stratified (Gmt) units. In most exposed sections, topset sediments are in erosional contact with underlying foreset sediments.

Delta bottomset sediments consist of fine sand and silt that is massive (Sm, Fm) or laminated (Ss, Fl). These sediments grade laterally from coarser foreset units to deposits of the glacial (glacial-marine) mud association.

A Working Model for Glacial-Marine Sedimentation

Detailed measurement and description of stratigraphic sections throughout the Maine coastal zone has led to the recognition of the lithofacies types and lithofacies associations described above. A working model for Late Wisconsinan glacial-marine sedimentation in coastal Maine has evolved from evaluation of these associations at a regional scale. The essential elements of this model are illustrated in Figures 20 and 21 and are outlined below.

At some point (15,000-14,000 yr B.P.) as ice withdrew across the isostatically depressed Gulf of Maine, eustatic sea-level rise brought marine waters against the ice front. The terminus of the ice sheet became marine-based (grounded below prevailing sea level), and sedimentation was characterized by

deposition from ice and glacial meltwater into a marine setting. Sediment accumulated over till that was deposited subglacially (lodgement and melt-out tills) by brash deposition from ice or by remobilization and flowage of sediment at the ice margin.

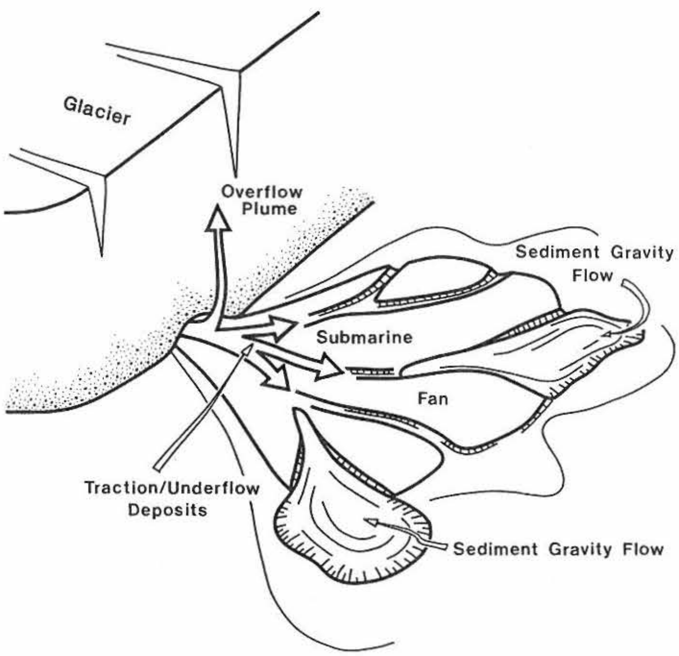

Figure 20. General model for subaqueous fan deposition. Sediment discharge at tunnel mouth consists of traction/underflow deposits and overflow (and interflow) sediments. The former are deposited as proximal to distal subaqueous fan lithofacies. The latter are deposited primarily as distal fan and submarine plain lithofacies. Sediment gravity flows within the fan complex may redistribute fan deposits.

Subglacial streams constructed eskers (tunnel/channelized flow) and spread aprons of subaqueous outwash along the grounding line, both adjacent to eskers and as unconfined subglacial sheet-wash away from esker sources.

Sediments deposited by a variety of glaciogenic processes (subglacial lodgement, subglacial melt-out, brash deposition, supraglacial flowage, flowage off ice-frontal constructional features, etc.) gave rise to debris flow deposits and high density underflows that carried sediments a few meters to several hundred meters away from the ice front (Fig. 3). At the same time, subglacial streamflow and unconfined subglacial sheetflow were transporting sediment to the ice front (grounding line). A great deal of this sediment accumulated as subaqueous fans (Fig. 20) or, to a lesser degree, partially-developed deltas against the ice front. Much of the sediment was carried into deeper water as density overflows and interflows to be deposited ultimately by quiet water sedimentation (rain-out).

Regular fluctuations of the ice margin (grounding line), induced by tidal fluctuations or changes in glacier regimen, produced a variety of types of end moraines, the most important of which were DeGeer moraines (Fig. 22). The moraines consist predominantly of subglacial/ice-frontal deposits and proximal fan sediments that have been glacial-tectonically deformed and thickened (Figs. 18 and 21).

End Moraines and the Pattern of Last Ice Retreat

End moraines and related ice-frontal features (Fig. 21) have been mapped throughout the Maine coastal region from the present coastline to the inland limit of late-glacial marine submergence (Smith, 1980, 1981, 1982; Thompson, 1982;

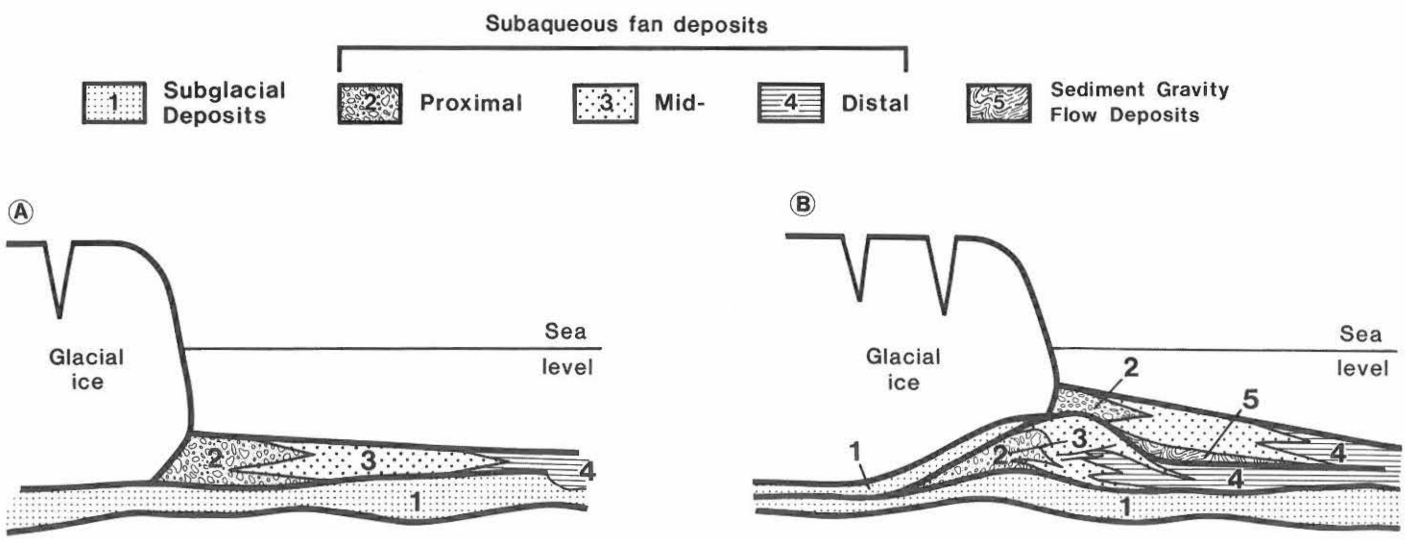

Figure 21. Stratigraphic relationships of subaqueous fan association (a) and subaqueous end moraine association (b). The simple stratigraphic situation of the subaqueous fan association is complicated during minor glacial readvance by glacial-tectonic deformation and thickening of the sedimentary package to produce an end moraine.

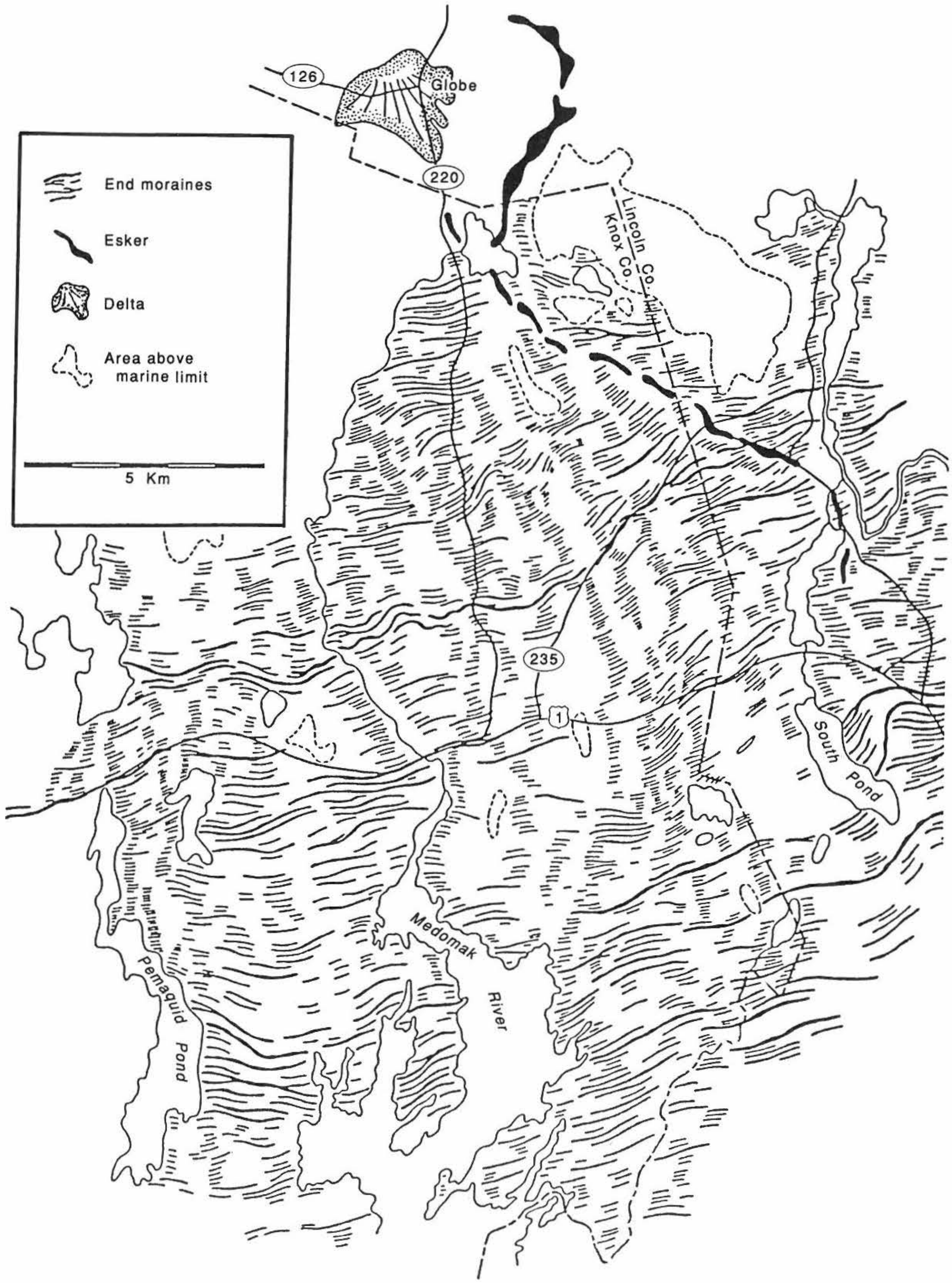

Figure 22. End moraines and glaciofluvial deposits of portions of Lincoln and Knox Counties, Maine.

Figure 23. DeGeer moraines. (a) Morphology of DeGeer moraine in the vicinity of Bucksport, Hancock County. Proximal side of moraine is to the right. (b) Gravel core of DeGeer moraine, vicinity of Kennebunk, York County. Moraine is overlain by sand of the subaqueous fan association. Shovel for scale.

Thompson and Borns, 1985b; see also Thompson and Borns, 1985a). Work by King et al. (1972), Fader et al. (1977), and Oldale (1985) indicates that similar features occur within the Gulf of Maine and offshore Massachusetts. There is, in addition, evidence that larger moraines can be traced inland above the marine limit.

The most abundant of the ice-frontal features are DeGeer moraines (Fig. 23) that occur as regularly spaced linear ridges overlain by, and often interbedded with, glacial-marine sediments. Interspersed between groups of DeGeer moraines are larger stratified end moraines and ice-frontal deltas ("moraine banks"). These features are composed predominantly of sand and gravel of the subaqueous fan association that intertongues with sand, silt, and clay of the marine (glacial-marine) mud association.

Pertinent aspects of the distribution of the moraines can be summarized as follows (Smith, 1982): (a) DeGeer moraines occur exclusively below the marine limit, and larger moraines are optimally developed below the marine limit; (b) the axes of moraine ridges are aligned perpendicular to the direction of last ice movement as recorded by glacial striations on local bedrock outcrops; (c) the moraines define a broadly lobate pattern, generally concave downvalley in topographic lows and convex downvalley over topographic highs (Fig. 22); (d) axes of large and small moraines are generally parallel to one another, although local crosscutting relations are common.

The moraines, both large and small, were formed at or near the retreating ice front during retreat of the marine-based Late Wisconsinan ice sheet. They record deposition of both subglacial and proximal glacial-marine sediments from warm-based ice, and deformation of these sediments by periodic fluctuations of the retreating ice margin (grounding line) or squeezing of the sediments into subglacial crevasses.

Associated with the end moraines, and sometimes a part of them, are a variety of delta forms (partially or fully developed) and subaqueous fans (Fig. 22). The delta forms have been described by Leavitt and Perkins (1935) as moraine banks, a term recently employed by Powell (1980) to describe a more general assemblage of ice-frontal ridges. The deltas include: (a) partially developed forms that are linear ridges with topset and foreset bedding constructed to prevailing sea level, but without the classic Gilbert-type delta form (Merriland Ridge, Fig. 24), (b) classic Gilbert-type deltas with ice-contact proximal slopes (L-Pond delta, Fig. 24), and (c) composite deltas that were constructed in segments to prevailing sea level (Pineo Ridge, Fig. 1). The subaqueous fans have much the same morphology as subaerial alluvial fans, though all available stratigraphic evidence indicates that they were formed below sea level.

Several of the most prominent end moraines (Pineo Ridge, Merriland Ridge, Waldoboro Moraine; Fig. 1) actually consist of till segments, composite (till and stratified drift) segments, and deltaic segments. This suggests that during any interval of stillstand or readvance, moraine construction involved the entire range of sedimentologic processes that existed along the ice front (grounding line) at the time. Thus, some segments were formed by subglacial processes, while others were formed by subglacial streamflow and construction of deltaic features or subaqueous fans.

DeGeer moraines display the same range of composition (Smith, 1982; Bingham, 1981; Bingham and Powell, 1982). In the case of these features, the moraine form is simply a function of either ice shove at the ice front (or grounding line) or squeezing of material into subglacial crevasses. In either case, the composition of the moraine is purely a function of the material available to the ice for modification by shoving or squeezing.

The full assemblage of ice-frontal features (end moraines, and ice-frontal deltas and fans) records, in detail, the configuration of the retreating ice margin and can be used to reconstruct the pattern of Late Wisconsinan ice retreat below the limit of marine submergence. Reconstruction of the pattern of Late Wisconsinan ice retreat based upon the distribution of these features (Fig. 25) indicates that initial ice withdrawal from

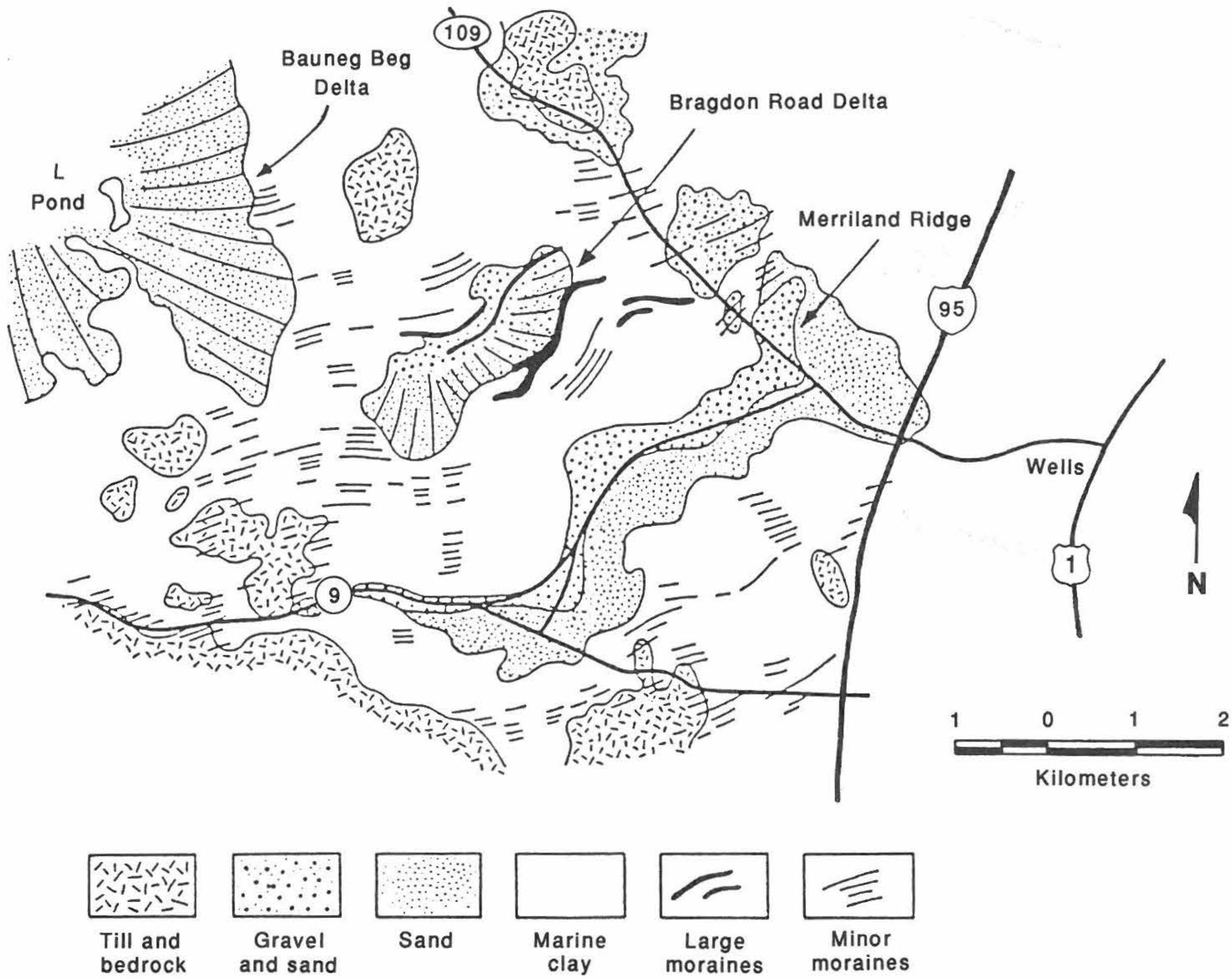

Figure 24. End moraines and glaciofluvial deposits in the vicinity of Wells, Maine (from Smith, 1982).

coastal Maine was generally parallel to the present coastline. In eastern and southwestern Maine, water depth was shallow (probably less than 30 m.), so that ice retreat was generally controlled by existing topography. In central coastal Maine (e.g., Penobscot River valley), water was deeper, and the rate of ice retreat was greater as a function of more rapid calving along the ice front.

While in eastern Maine ice retreat continued inland generally parallel to the present coastline, in central Maine the calving embayment became more pronounced (Smith, 1985). In southwestern Maine, the pattern of ice retreat was controlled in large part by emergence of the White Mountains from beneath the ice sheet (Davis et al., 1980; Spear, 1981). As a result, the direction of ice retreat in this region became (at least locally) more northerly or northwesterly, almost normal to the present coastline.

Chronology of Deglaciation

The chronology of deglaciation of coastal Maine has not yet been satisfactorily resolved, despite an abundance of radiocarbon dates that are tied either directly to ice retreat or to coastal submergence and emergence. Bloom (1963) published the first dates pertinent to deglaciation of coastal Maine. On the basis of three radiocarbon dates from the Presumpscot Formation or overlying bog-bottom sediments, Bloom suggested the following sequence of late-glacial events. Submergence of the coastal region of southwestern Maine was initiated by 12,100 yr B.P.,

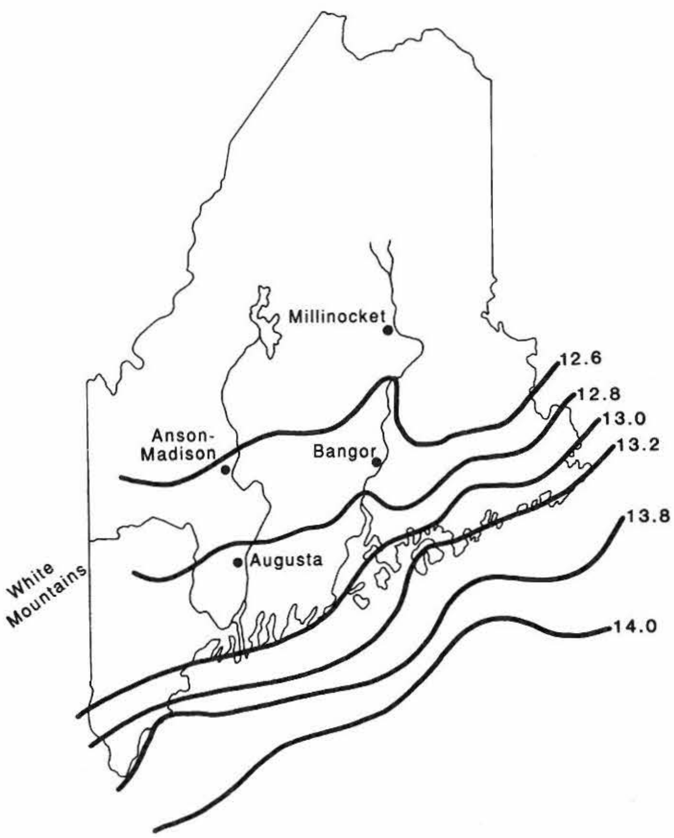

Figure 25. Generalized model of ice retreat from coastal Maine (from Smith, 1985). Location and form of ice-frontal positions based on radiocarbon-dated glacial-marine deposits and distribution of coastal end moraines. Ice-frontal position at 14,000 yr B.P. after Fastook and Hughes (1982). Ages in thousands of years B.P.

and maximum submergence was attained by 11,800 yr B.P. Emergence of southwestern coastal Maine was accomplished between 8,000 and 7,000 yr B.P. In addition, Bloom suggested that ice retreat was interrupted by a significant readvance (Kennebunk readvance), which he correlated tentatively with the "Valderan" stadial of the midcontinent.

Stuiver and Borns (1975) collected radiocarbon dates from shells and seaweed at 23 localities throughout coastal Maine. On the basis of these dates, the authors developed the following sequence of events: (a) retreating Late Wisconsinan ice was at or slightly inland of the present coastline by 13,200 yr B.P.; (b) ice had withdrawn from central Maine by 12,700 yr B.P.; and (c) coastal emergence was complete by 12,100 yr B.P. In addition, Borns (1967, 1973; Borns and Hughes, 1977) suggested that Late Wisconsinan ice readvanced to the position of Pineo Ridge (Pineo Ridge readvance) at approximately 12,700 yr B.P. (Port Huron readvance of the mid-continent).

Smith (1985), employing the dates of Bloom (1963) and Stuiver and Borns (1975), as well as nineteen additional dates (a total of forty two dates), proposed a somewhat different chronology of deglaciation as follows (Fig. 25). By at least 13,800 yr B.P., the retreating Late Wisconsinan ice margin was at the position of the present coastline in southwestern Maine, but was well seaward of this position in central and eastern Maine. Ice had withdrawn to the position of the present coastline in eastern Maine by 13,200 yr B.P. At this time, ice was still seaward of the central coast, but had begun to withdraw inland from the coast in southwestern Maine. Late-glacial submergence of coastal Maine was at its maximum between 12,600 and 12,400 yr B.P., at which time the ice margin had retreated to a position above the marine limit along its entire extent. Emergence of coastal Maine, resulting from isostatic recovery, was complete in eastern Maine by 12,000 yr B.P., and in southwestern Maine by 11,500 yr B.P.

Time-distance curves based on the available data indicate that the rate of ice retreat from coastal Maine was between 0.20 and 0.25 km/yr. While ice retreat was characterized by frequent and regular fluctuations of the ice margin, there is no evidence for significant readvance of the retreating ice sheet.

The Kennebunk readvance was originally described by Bloom (1960, 1963) to explain the occurrence of deformed glacial-marine sediments in the vicinity of Kennebunk. Bloom tentatively extended the line of the proposed readvance from Biddeford-Saco to the Maine-New Hampshire border, and incorporated elements of the Newington Moraine (Katz and Keith, 1917) in that reconstruction. On the basis of very limited chronologic data, Bloom suggested that the Kennebunk readvance was of climatic significance and could be provisionally correlated with the "Valderan" stadial of midcontinental United States.

Smith (1981) reinterpreted the Kennebunk readvance in terms of minor fluctuations of the retreating margin of the Late Wisconsinan ice sheet. The presence of DeGeer moraines (Fig. 23) throughout the area of the proposed readvance suggests an alternative explanation for deformation of the glacial-marine sediments. Furthermore, the absence of other supporting evidence in the form of changes in the direction of ice flow detracts from the idea of significant glacial readvance in this part of the coast. The Newington Moraine, employed by Bloom in reconstruction of the limit of readvance, consists of an unrelated assemblage of ice-frontal and ice-marginal deposits constructed by the retreating ice at different times and at very different positions of the ice margin.

Definition of the Pineo Ridge readvance in eastern coastal Maine (Borns, 1967, 1973; Borns and Hughes, 1977) has been based largely on the occurrence of a large delta-moraine complex (Pineo Ridge) and its relationship to other ice-marginal features. Pineo Ridge is a significant coastal geomorphic feature that sharply crosscuts a series of DeGeer moraines, suggesting by virtue of the crosscutting relationships that there was a major (40 km) readvance of the retreating Late Wisconsinan ice margin. The Pineo Ridge readvance was dated at about 12,700 yr B.P. (Borns, 1973) and tentatively correlated with the midcontinent Port Huron readvance.

Subsequent work by Miller (1986), Miller and Borns (1987), and Holland (1983) suggests that Pineo Ridge does not, in fact, record a major glacial readvance, but was instead constructed as an ice-frontal feature during Late Wisconsinan ice retreat (without readvance). A similar view has been advocated by others (Gaddis and Smith, 1983). The entire assemblage of deposits and features (moraines, deltas) that comprise the Pineo Ridge complex can be readily explained in terms of a composite delta sequence constructed within a minor calving embayment. In broad terms, the retreat of ice and the deposition of glaciogenic sediments in eastern Maine was very much the same as it was in southwestern Maine, a function of the fact that the topographic and bathymetric controls of ice retreat were similar in both areas.

Alternative models for the chronology of Late Wisconsinan deglaciation of coastal Maine have been proposed by Davis and Jacobson (1985) and, more recently, by Stone and Borns (1986). The chronology of Davis and Jacobson is based almost exclusively on dates from lake bottom and bog bottom sediments from the inner coastal zone and interior Maine. Virtually none of the previously recorded dates from the marine succession of the outer coastal zone were used by these authors in their reconstruction of ice retreat. At issue is not the validity of the dates from bog or lake bottom sediments, but the exclusion of the bulk of available radiometric data from the coastal glacial-marine succession.

In developing their reconstruction of ice-marginal positions, Davis and Jacobson made questionable assumptions regarding the role of topography in controlling the pattern of ice retreat, particularly in coastal Maine (e.g., "the thinning ice sheet would have survived longest in valleys..." - this is clearly a condition that would not have existed while the ice sheet was marine-based). Especially troublesome is Davis and Jacobson's proposed configuration of the ice margin at 13,000 yr B.P. (Fig. 26). Not only is the reconstruction inconsistent with available data from the coastal zone, it is also drawn to accomodate the Pineo Ridge readvance (even though nearby pollen data does not record the readvance), an event that most probably did not occur. Inasmuch as the Davis and Jacobson chronology has been adopted for the recently published Surficial Geologic Map of Maine (Thompson and Borns, 1985a), potential problems inherent in the chronology should be recognized.

A still different scenario for the chronology of Late Wisconsinan deglaciation has been proposed by Stone and Borns (1986). This chronology is more regional in scope, providing, as it does, a chronologic model for the entire northern Appalachian region. There are, however, internal problems with the chronology. The textual discussion indicates a chronology generally consistent with that proposed by Stuiver and Borns (1975) and by Smith (1985). The time-distance presentation of chronologic events (Stone and Borns, 1986, Chart 1), on the other hand, suggests retreat of ice from coastal Maine at least 500 to 1000 years earlier than the time proposed by either Stuiver and Borns or Smith (Fig. 27). The implications of this scenario are significant in that Stone and Borns propose that Late Wisconsinan ice had withdrawn well into central Maine by at least 14,000 yr B.P. Virtually all previous reconstructions (Stuiver and Borns, 1975; Fastook and Hughes, 1982; Smith, 1985; Davis and Jacobson, 1985) place the ice margin at or near the present coastline at 14,000 yr B.P. The reasons for this disparity are not altogether

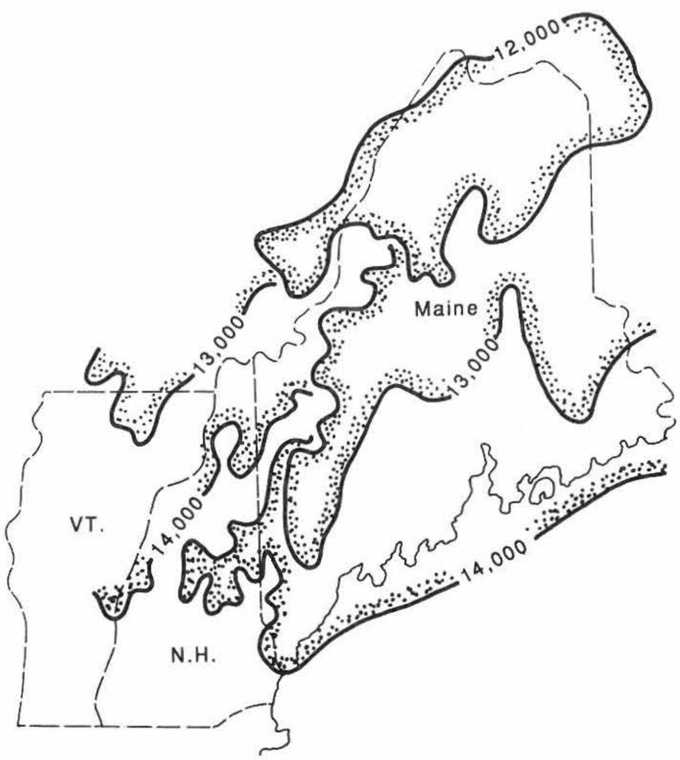

Figure 26. Generalized model of deglaciation of coastal Maine, according to Davis and Jacobson (1985). Ice-frontal positions compiled from Figures 9, 10, 11 of Davis and Jacobson (1985).

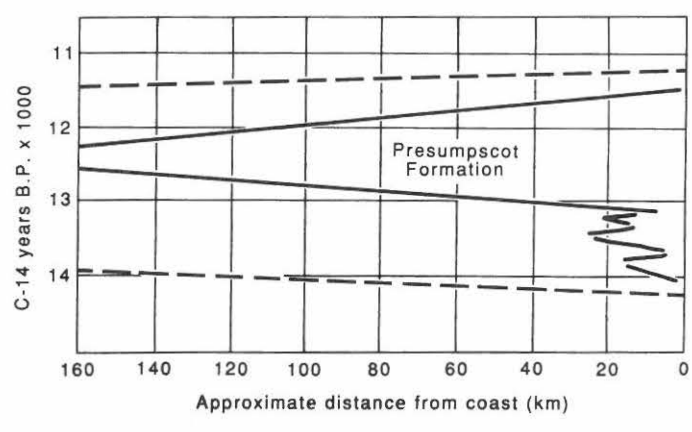

Figure 27. Time-distance curves depicting ice-retreat from coastal Maine. Solid line depicts: ice retreat and coastal submergence (lower line), and isostatic recovery and coastal emergence (upper line) according to Smith (1985). Dashed line depicts: ice retreat and coastal submergence (lower line), and isostatic recovery and coastal emergence (upper line) according to Stone and Borns (1986).

clear. It would appear that Stone and Borns have not included evidence from coastal Maine in an effort to accomodate dates from areas outside of Maine. The premise that ice had withdrawn from coastal Maine prior to 14,000 yr B.P. is not only inconsistent with available chronologic data from that region, it also ignores abundant sedimentological and morphologic information from the coastal zone.

The Regional Context of Deglaciation of Coastal Maine

Several models have been proposed for the nature and timing of Late Wisconsinan retreat of the southeastern portion of the Laurentide ice sheet. In recent years, these models have focused on the development of marine calving embayments in the Gulf of Maine and the Gulf of St. Lawrence early in the history of deglaciation (Borns and Hughes, 1977; Chauvin et al., 1985; Denton and Hughes, 1981; Gadd, 1980; Hughes et al., 1985; Thomas, 1977, among others). Calving embayments, fed by large ice streams, led to rapid drawdown of the Laurentide ice sheet, and ultimately to separation of an independent ice mass over the northern Appalachians. The extent of this independent ice mass was further constrained by emergence of portions of the White, Longfellow, and Boundary Mountains from beneath the retreating ice sheet.

Borns (1985) has summarized the salient aspects of existing data that bear on a workable model for deglaciation of Maine, in general, and coastal Maine, in particular. Separation of northern Appalachian ice from Laurentide ice produced a thin (700 m or less) residual ice sheet that extended from northern and eastern Maine to the present Maine coast (Borns, 1985). There is abundant field evidence to indicate that, although thin, this ice mass was internally active and capable of at least minor marginal readvances.

In general, the timing of these events clusters between 13,000 and 14,000 yr B.P. Emergence of the White Mountains appears to have occurred by at least 13,000 yr B.P., and possibly as early as 14,000 yr B.P. (Davis et al., 1980; Spear, 1981). Marine transgression, accompanied by ice retreat, in the St. Lawrence Valley had progressed to at least the position of Quebec City by 12,400 yr B.P. (LaSalle, 1972), though there is some suggestion that further ice retreat to the position of Ottawa had been accomplished by 12,800 yr B.P. (Richard, 1975). At the same time, there is clear evidence that the southern margin of ice in coastal Maine was at or near the position of the present coastline between 13,800 and 13,500 yr B.P. (Borns, 1973; Smith, 1985), and had withdrawn into central Maine by 13,000 yr B.P. (Smith, 1985; Thompson and Borns, 1985b).

If one accepts the general model of regional deglaciation and the general timing of development of an independent northern Appalachian ice mass at between 13,000 and 14,000 yr B.P., all of which is well documented by field evidence and substantiated by conceptual data, it is difficult to accept the chronologic model of Stone and Borns (1986) that requires ice retreat from coastal Maine at least 1,000 years earlier. While there may be disagreement with sedimentological interpretations of some dated successions, both the dates and the direct tie of those dates to the presence of ice in coastal exposures is clear and is based on sound field data and reliable sedimentologic and glaciologic interpretation.

RECOMMENDATIONS FOR FUTURE STUDY

There remain in the study of the Late Wisconsinan glacial history of coastal Maine a variety of fundamentally important problems that are substantially unresolved. The general model of deglaciation that involves partitioning of a local northern Appalachian ice mass early in the history of deglaciation is difficult to refute. It appears to be reasonably substantiated on both evidence from the field and on conceptual grounds. Questions, however, do arise when dealing with both the details of glacial and glacial-marine sedimentology and the chronology of deglaciation of coastal Maine.

Considerable work has been done in coastal Maine to delineate the areal distribution of glaciogenic materials and glacial morphologic features. On the other hand, with few exceptions (Hunter, in prep.; Hunter and Smith, 1988; Smith, 1984a, 1984b; Retelle and Konecki, 1986; Retelle and Bither, this volume), the sedimentology and detailed stratigraphic relationships of glaciogenic units have been documented and studied in only the most rudimentary fashion. Because these aspects of the glacial geology of the coastal zone are crucial to a complete understanding of the processes involved in deglaciation, it is of fundamental importance that more attention be given to them in the course of current and future detailed quadrangle mapping and topical studies.

The problems related to the chronology of deglaciation in coastal Maine have several sources. There are very few dates that can be tied directly to ice retreat. The depositional settings of those that do exist are subject to a range of interpretations. Clearly, the development of a detailed lithofacies model for the glacial and glacial-marine sediments of the coastal zone will limit the range of interpretations of the significance of these dates. Until such a model is developed and more such dates become available, real caution must be exercised in temporal reconstructions of coastal deglaciation.

If we can develop a reasonable model for glacial-marine sedimentation and a mutually acceptable model for chronology of deglaciation, then we will have a sound basis for relating the deglaciation of coastal Maine to similar events elsewhere. We face a real challenge in that regard, and should bend every effort to accomplish that goal.

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The authors wish to acknowledge the important contribution made by Walter Anderson, W. Bradford Caswell, and Harold Borns, Jr. in providing the impetus for the successful development of a program of mapping and study of the surficial geology of the State of Maine. The importance of their efforts in this regard cannot be overstated. The work of many field scientists in the early stages of the Maine Geological Survey's surficial geology mapping program has provided the basis for much of what is discussed in this paper and the great bulk of what we know of the glacial history of the state today. Included in this group of people should be the names of: Harold Borns, Andrew N. Genes, William Newman, Thomas Brewer, Dabney Caldwell, Lindley Hanson, Kristine Crossen, William Holland, Carolyn Lepage, Thomas Lowell, Steven Kite, Woodrow Thompson, and many student assistants. The work begun by these people continues today under the able direction of Woodrow Thompson.

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Late Wisconsinan Glacial and Glaciomarine Sedimentary Facies in the Lower Androscoggin Valley, Topsham, Maine

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ABSTRACT

A facies model is proposed for late Wisconsinan glaciomarine deposits in the lower Androscoggin Valley that includes four lithofacies assemblages that are defined by morphology and detailed stratigraphic and sedimentologic analysis. The end moraine facies assemblage includes subglacial and resedimented diamicton, and interbedded and locally deformed sand and gravel beds. The sediments form linear ridges which are former grounding line positions of the tidewater glacier margin. The submarine outwash fan facies assemblage commonly drapes or flanks the end moraine assemblage. In proximal regions of the fan, gravel, bedded sand, and diamicton lithofacies predominate and represent rapid deposition at the mouth of the meltwater tunnel by fluvial and mass flow processes. Distal and lateral to the ice margin, fan sediments consist of graded and cross-laminated sands deposited by underflow currents and slump-generated turbidites interbedded with rhythmically bedded silt attributed to suspension deposition from overflow or interflow plumes. The submarine plain facies assemblage is laterally and distally transitional with the submarine fan assemblage and consists of apparently massive, structureless fine-grained sediments deposited from suspension. The shallow marine facies assemblage consists of well-sorted tidal to subtidal sand lithofacies, poorly sorted gravelly and bouldery lag deposits on moraine crests, and lagoonal muds. Collectively, these lithofacies were deposited as a result of reworking previously deposited sediments during isostatic emergence.

INTRODUCTION

Retreat of the late Wisconsinan Laurentide ice sheet through the coastal zone of Maine is documented by a complex stratigraphic sequence of glacial and glaciomarine sediments. The purpose of this paper is to report the results of detailed stratigraphic logging and sedimentology of such deposits in two adjacent borrow pits in south-coastal Maine. The data include lithofacies distribution, thickness, and texture, and are provided to lead to accurate modeling and detailed paleoenvironmental reconstruction of the sedimentary environments associated with retreat of a tidewater glacier margin.

Study Area

The study area is located in Topsham, Maine, in the northwestern one-ninth of the Brunswick 7.5 minute quadrangle (Bath 15 minute quadrangle). Two gravel pits, the Webber pit and the Bisson pit are located along the west side of Meadow Road approximately one-half mile west of the Cathance River (Fig. 1). The gravel pits are both located on the west side of a bedrock structural and topographic high that reaches a maximum elevation of 198 feet in the Webber Pit.

The larger of the pits, the Webber Pit (Fig. 1), has exposures in a moraine ridge complex. The land surface drops off to the

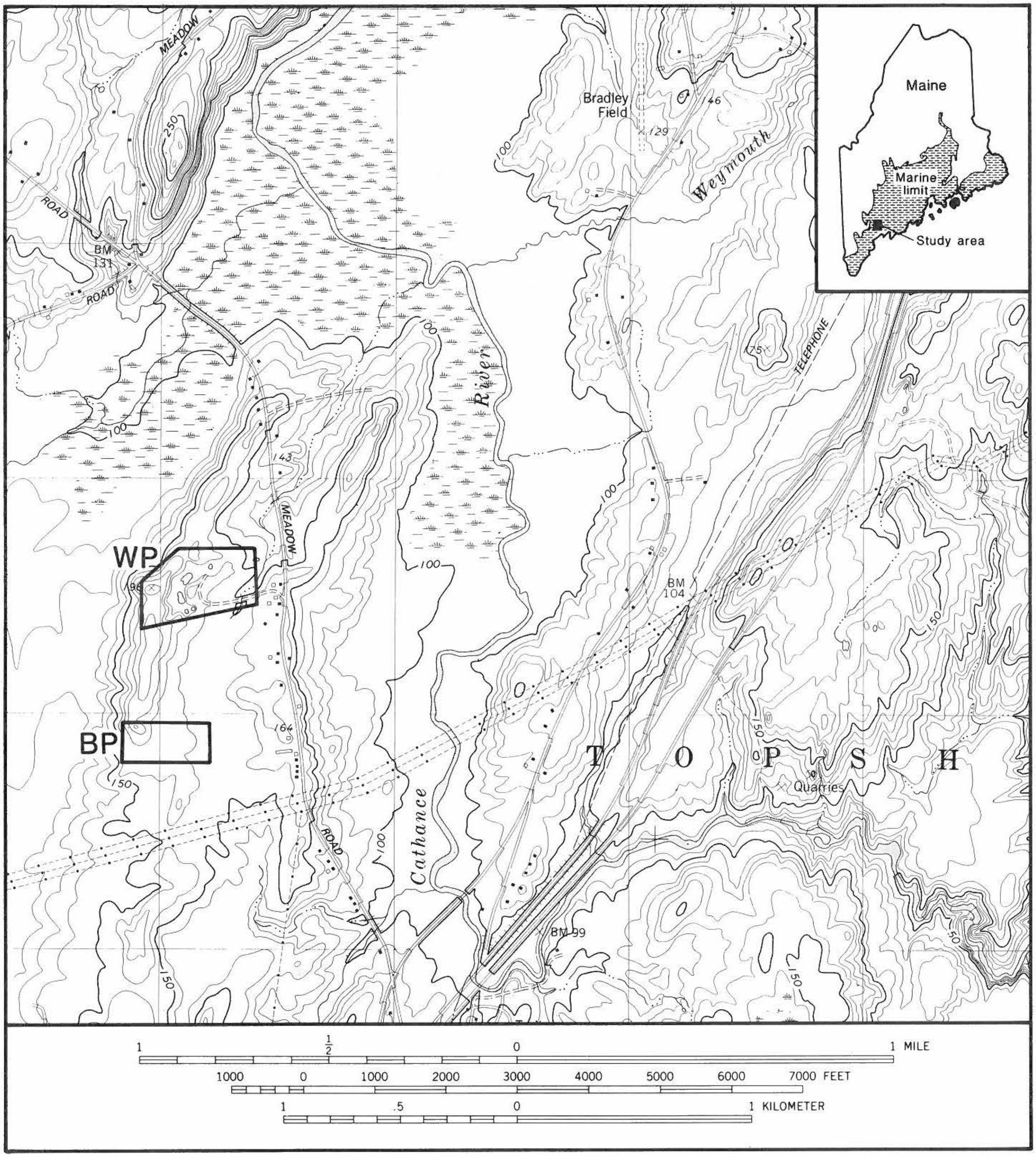


Figure 1. Location map of the study area, Topsham, Maine. Letters refer to Webber pit (WP) and Bisson pit (BP). Shading on inset map refers to area of late Wisconsinan marine submergence.

north behind the moraine complex. Additional exposures are cut into a flat to rolling plain that descends to the Cathance River lowland to the east.

The Bisson pit is located approximately 500 m due south of the Webber pit along the axis of the same bedrock ridge (Fig. 1). Excavation along the northern margin of the pit has exposed the bedrock ridge while southern excavations are approximately 20 feet lower along the plain contiguous with the lower exposures in the Webber pit.

Bedrock Geology

The bedrock within the area is the Richmond Corner Member of the Cushing Formation (Hussey, 1981), consisting of quartz-plagioclase gneiss with rusty weathering schistose compositional layering. The Cushing Formation, part of the Cambro-Ordovician Casco Bay Group, lies in fault contact with the Silurian Vassalboro Formation (Hussey, 1981). The trace of the fault is located along the northeast-trending lowland to the west of the study area (Fig. 1).

Surficial Geology

At its maximum, the late Wisconsinan Laurentide ice sheet extended beyond the present coastline of Maine onto the continental shelf (Schnitker, 1974; Fastook and Hughes, 1982; Belknap, 1987; Belknap et al., 1986, 1989; Oldale, 1989). It is generally believed that recession from that maximum position began around 17,000 to 15,000 yr B.P. (Tucholke and Hollister, 1973). Stuiver and Borns (1975) estimate that the ice margin reached the coast around 13,500 yr B.P. Smith (1985), however, places the retreating margin at the coastline as early as 13,800 yr B.P. in the southwestern corner of the state, and inland of the marine limit by 12,800 to 12,600 yr B.P.

The retreat of the marine-based ice sheet was interrupted by numerous, perhaps annual (cf. Meier, 1984) forward oscillations recorded by many end moraines, both small and large, and submarine outwash fans (Thompson, 1979, 1982; Smith et al., 1982; Smith and Hunter, this volume). The ice-marginal deposits are distally and laterally intercalated with fine-grained, locally fossiliferous glaciomarine sediments. The marine silt and clay, referred to as the Presumpscot Formation (Bloom, 1960, 1963), exhibits a distinctive smooth to gently rolling topography as a veneer on the end moraine and outwash fan deposits. Marine submergence of the coastal zone occurred from the time of deglaciation (ca. 13,000 to 12,600 yr B.P.) until isostatic emergence of the coastal zone was complete by ca. 11,500 yr B.P. (Smith, 1985). Based on data from a marine limit elevations map (Thompson et al., 1983), it is estimated that the water depth over the Topsham area at the time of deglaciation was approximately 30 to 45 meters (90 to 140 feet). Water depths over the site were probably greatest immediately following deglaciation due to isostatic depression. Accelerated uplift after ice retreat produced a shoaling of the coastal zone, reflected in the stratigraphy by intertidal and beach sediment overlying the fine-grained silt and clay.

In this study, we propose a preliminary model for glacial and glaciomarine sedimentation in the lower Androscoggin Valley. The model uses vertical and lateral lithofacies changes and facies associations as primary data. Interpretation relies heavily upon studies conducted on distribution of sediments and related processes in modern glaciomarine environments (e.g. Gilbert, 1982; Powell, 1981, 1983, 1984; Mackiewicz et al., 1984; Dowdeswell, 1987). A model for the deposits associated with deglaciation of coastal Maine is also presented by Smith and Hunter elsewhere in this volume.

Methods

Stratigraphic logging of deposits was accomplished in the field by methods as described by Eyles et al. (1983). Individual lithofacies are identified in the field by texture, sedimentary structures, and stratigraphic relationships (Table 1). Lithofacies logs were constructed to show vertical changes in the sedimentary succession; several cross-sections of pit faces were mapped to show lateral and vertical lithofacies changes. Facies associations (cf. McCabe et al., 1986) used in this report reflect the assemblage of lithofacies within morphostratigraphic units such as moraines or submarine fans.

Maps of the gravel pits were produced by both plane table and alidade, and pace and compass methods. Till fabrics were measured according to methods proposed in Andrews and Smithson (1966). Particle size was measured by wet and dry sieving for the coarser than 4.5 phi fraction. The finer than 4.5 phi fraction was analyzed using a Micromeritics Sedigraph 5000 D housed at the University of Massachusetts.

RESULTS

Webber Pit - Description of Sections

Section A. Section A is situated on a transverse cut approximately 50 m long and 7 m high in the northwest corner of the Webber pit (Fig. 2). The stratigraphy consists mainly of alternating bodies of diamicton and gravelly sand and sand layers. At least 3 layers of diamicton are present in the section (Fig. 3a). Diamicton bodies in this section and in the pit in general, range from tabular to wedge shaped. Upper and lower bounding surfaces, when viewed parallel to the moraine axis, are commonly sharply defined and horizontal; however, some lower bounding surfaces of some of the bodies are slightly arcuate to concave upwards. In longitudinal section, some diamicton bodies thicken in wedge shape towards the ice proximal (north) face. Some units display an upturning or flexure towards the south or free face of the moraine ridge.

The uppermost diamicton unit (Dmm1, Fig. 3a) is a tabular block up to 1.6 m thick with a sharp basal contact. The matrix is olive (5Y 4/3, damp) and compact with a well-developed

TABLE 1. A LITHOFACIES SCHEME FOR THE LOWER ANDROSCOGGIN RIVER VALLEY, MAINE (AFTER EYLES ET AL., 1983; MIALL, 1978)

Facies Code	Lithofacies	Sedimentary Structures	Interpretation	Facies Assemblage
Dmm	massive, matrix-supported diamicton	none	lodgement till	End moraine, submarine fan
(s)			sheared	End moraine
(c)			current reworked	End moraine
(r)			resedimented	End moraine
Dms	matrix-sup. diamicton	stratified	debris flow deposits	End moraine, submarine fan
(r)			resedimented	End moraine, submarine fan
(s)			sheared	End moraine
Gm	massive or crudely bedded gravel	horizontal bedding, pebble imbrication	turbidite, lag deposits, tractive flow	Submarine fan, shallow marine, end moraine
Sm	massive sand	none	turbidite, tractive flow	Submarine fan, shallow marine, submarine plain, end moraine
(s)		sheared		End moraine, submarine fan
Sp	planar crossbedded sand, (med. to v. coarse, may be pebbly)	solitary (alpha) or grouped (omikron) planar crossbeds	turbidite, tractive flow	Submarine fan, end moraine
Sr	rippled sand (v. fine to coarse)	ripple marks of all types	ripples (lower flow regime)	Submarine fan, end moraine
Sh	horiz. lam. sand, (v. fine to v. coarse, may be pebbly)	horizontal laminations	planar bed flow (l. and u. flow regime)	Submarine fan, end moraine, shallow marine
Sg	graded sand, (v. fine to v. coarse, may be pebbly)	graded	turbidite	Submarine fan, submarine plain, end moraine
(s)		sheared		Submarine fan, end moraine
(d)	with dropstones			Submarine fan
Fl	clay, silt, sand	laminated	turbidite, overflow plume	Submarine fan, shallow marine, end moraine
(d)	with dropstones			Submarine fan, submarine plain
Fm	massive clay and silt	none	turbidite, overflow	Submarine plain, shallow marine, end moraine
(d)	with dropstones			Submarine plain

horizontal to sub-horizontal fissility. Mean grain size of eight samples of matrix sediment in this block showed a range from 3.2 to 4.2 phi units (very fine sand to coarse silt). Sorting values range from 2.4 to 3.2 (poorly to very poorly sorted). Sand contents in the block range from 25 to 55% (cf. Fig. 3d). In general, no distinct trends in grain size distribution are discernible within the unit.

Clasts in this unit are widely dispersed through the matrix and range from pebbles to boulders 70 cm long. Clast lithologies are dominated by local metapelitic and metavolcanic rocks. Several black gabbro boulders occur in the diamicton. These presumably were transported by ice from the Androscoggin Lake complex 40 km to the north-northwest. Clast fabric (n = 25 stones) is distinctly bimodal with one strong north to north-northwest grouping (340° to 360°) and a strong 280° to 300° trend (Fig. 3b). No clay skins or silt caps were observed around the clasts.

Lower in the section (at approximately 2 m from the base) a one meter-thick slab of diamicton (Dmm2, Fig. 3a and Fig. 4a) has essentially the same color and grain size distribution as the uppermost diamicton, Dmm1. However, the lower unit is more complex structurally. The upper surface of the body lies in sharp contact with the overlying sand unit. Along this surface a cluster of 5 striated "bullet-shaped" boulders lies with long axes trending from N3°E to N16°W. In addition to the thin fissility similar to that of the upper block, the lower diamicton contains thin subhorizontal to curvilinear sandy partings and lenses (up to 10 cm long) and several arcuate concave upward fractures (up to 1.5 m long) that is in places highlighted by a thin bed of medium to coarse sand. The sand lenses are texturally similar to those sand beds overlying and underlying the diamicton.

Approximately 4 m from the base of section A (Fig. 3a), a second type of diamicton Dmm(r) is exposed. This unit is variable in texture and structure. The unit ranges from a com-

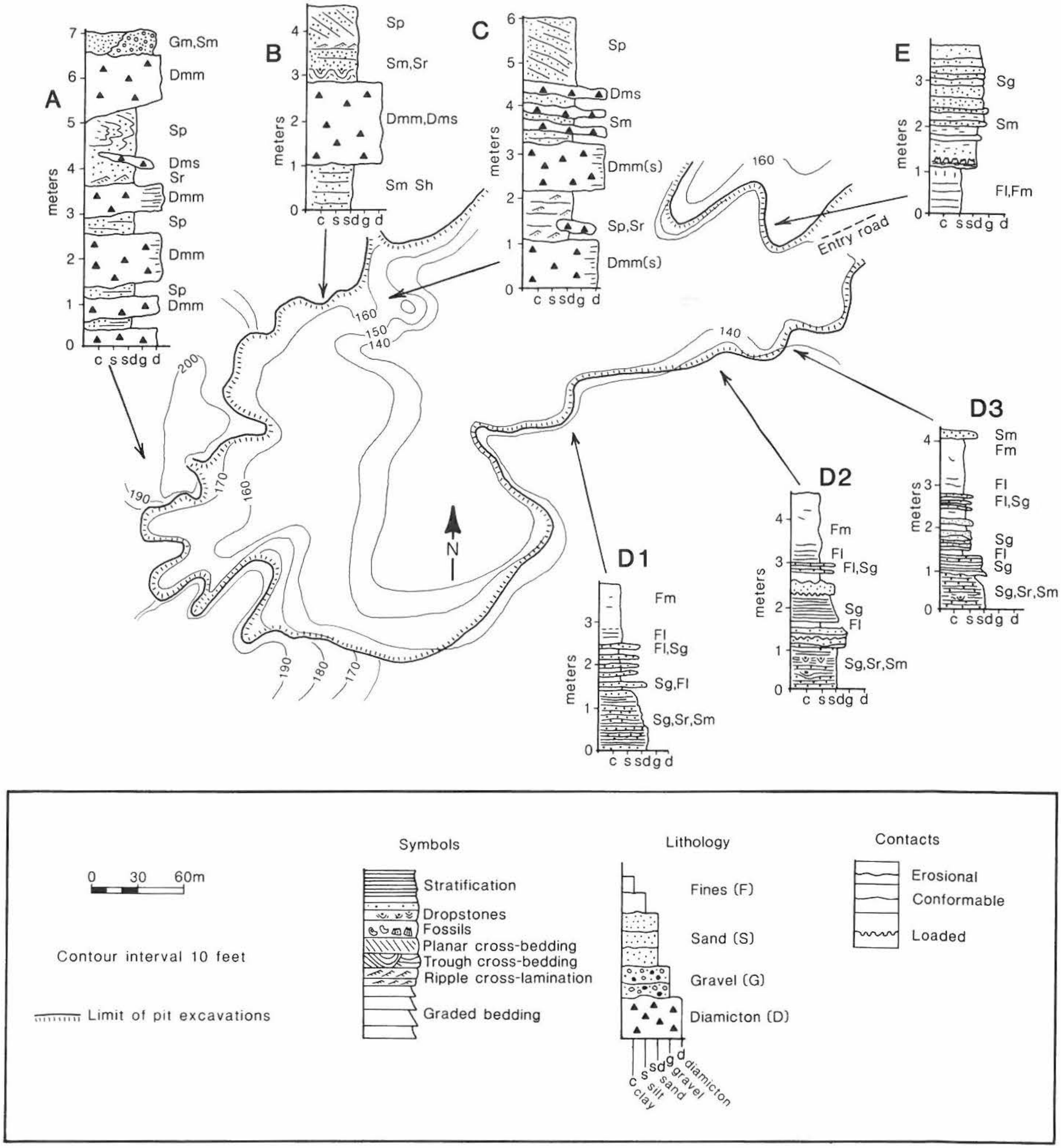


Figure 2. Map of the Webber pit with lithofacies logs and locations of sections described in text. Stratigraphic symbols after Eyles et al. (1983) and Miall (1978) (see Table 1).

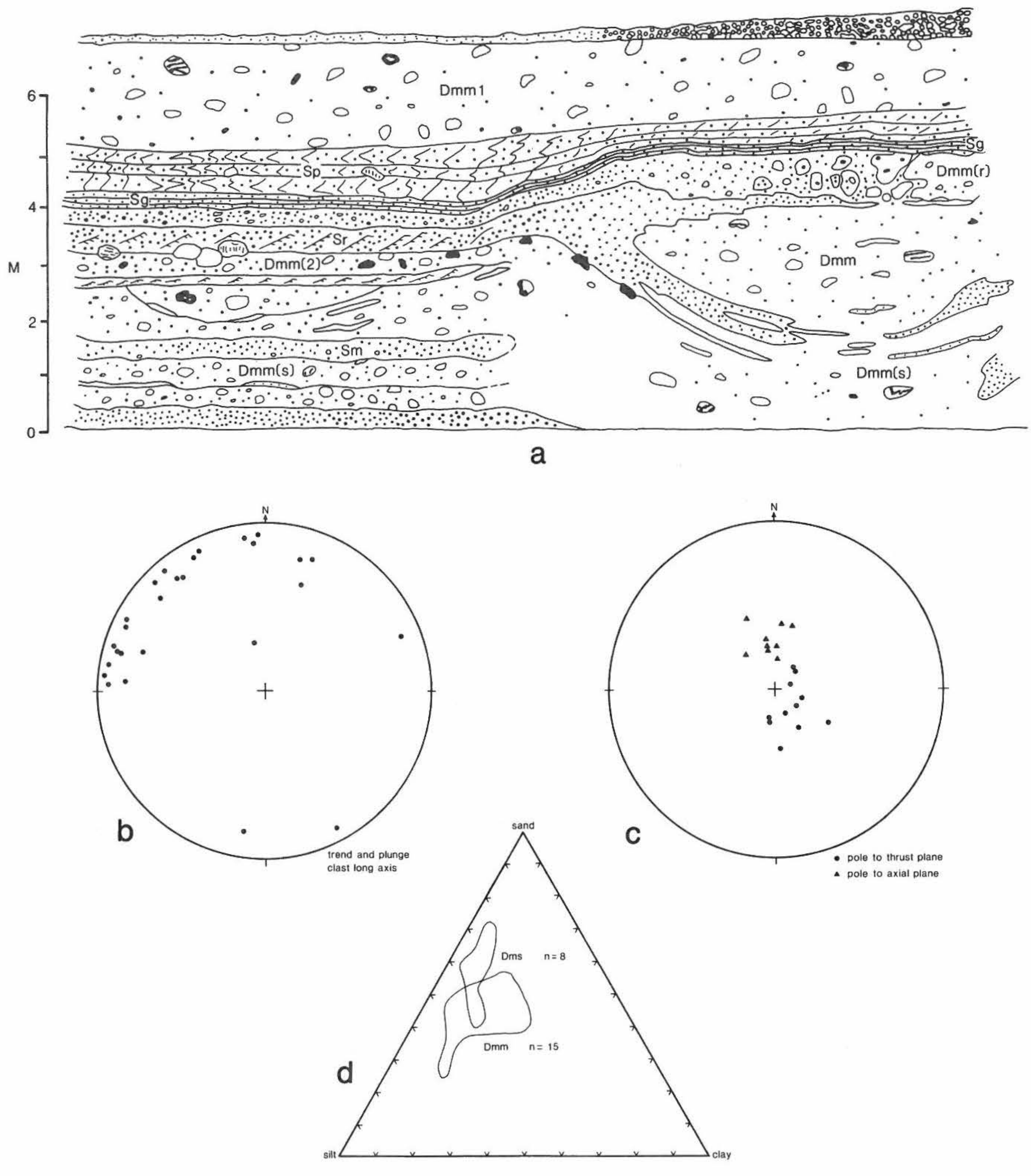
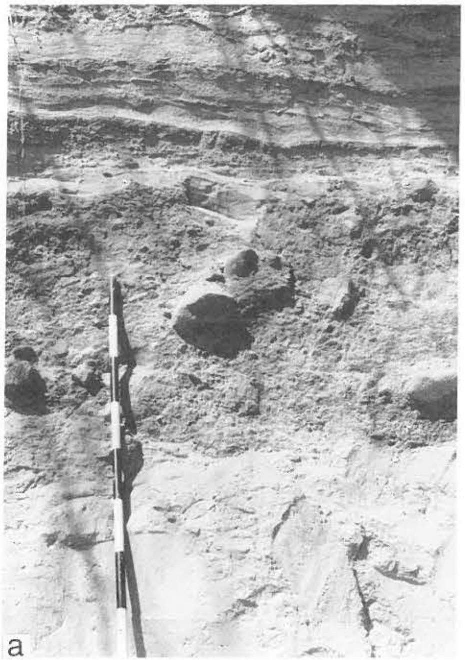
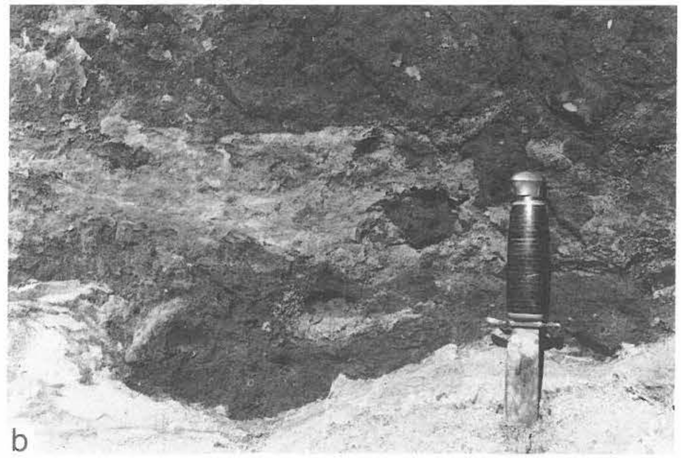


Figure 3. (a) Lithofacies diagram of section A in Webber pit. Stratigraphic symbols after Eyles et al. (1983) and Miall (1978; see Table 1). (b) Till fabric of unit Dmm1 plotted on 3-D stereoprojection. (c) Stereonet projection of structural features in deformed sand unit (Sr, Sp, Sg) beneath Dmm1. (d) Sand-silt-clay ternary plot of diamictons in Webber Pit.





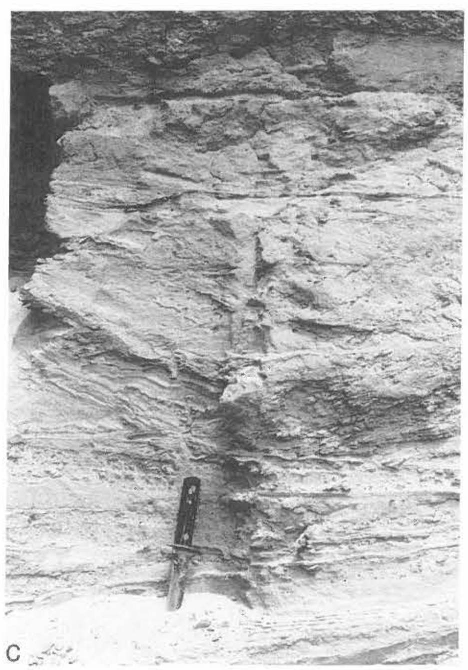


Figure 4. Photographs of section A, Webber pit. (a) Facies Dmm beneath a bedded sand unit. (b) Facies Dmm(r). Clast of diamicton is surrounded by poorly sorted coarse sand which is in turn incorporated in 0.5 m thick layer of diamicton. (c) Facies Sp, Sh deformed by folds and thrust faults beneath facies Dmm1.

pact, olive, matrix-supported diamicton (similar to Dmm1) to poorly sorted, bedded sand containing rounded clasts of diamicton (Fig. 4b). Approximately 20 m west of the massive portion of the diamicton body, the bedded sand contains suspended intraclasts of laminated mud and sand. Sand content was measured as high as 70% by weight in the diamicton matrix in this unit and gravel content was also higher than facies Dmm.

Interstratified sand beds in the exposure make up several facies (Sm, Sh, Sp, and Sr, Table 1). Mean grain size is approximately 2.5 phi (fine to medium sand); sorting coefficients average 0.85 (moderately sorted). Below the upper diamicton unit (Dmm1) in section A lies 1.0 to 1.6 m of planar cross-bedded, plane-bedded, and graded sand. Several of the sand beds contain Type A and Type B ripple-drift cross-lamination (Jopling and Walker, 1968) draped by fine silt laminae. The sand is folded and thrust-faulted in the upper meter (Fig. 4c) and undeformed at the base. Fold hinges are subparallel to the outcrop face (east-west) and axial planes of the folds dip to the south. Thrust fault planes dip gently to the north-northwest (Fig. 3c). Dropstones are found at several locations in the sands.

At approximately 3 m from the base of the section, rippled fine to medium sand overlies facies Dmm2 (Fig. 3a). Careful excavation of the contact between the diamicton unit and the rippled sand reveals that the upper surface of the till was fluvially eroded, presumably by the currents that deposited the bedded sand. The till surface displayed molding and rounding. Rounded clasts of till were found in scour pits on the undulating till surface. Laminations in the overlying sand conformed to the topography on the till surface, producing low angle cross-bedding from the sand infilling.

Section B. A series of cuts in the central portion of the ridge (Fig. 2) exposes various longitudinal and transverse views through deformed diamicton facies (Dmm(s) and Dms(s)), and massive sand (Sm(s)), which are overlain by structureless mud and thin horizontal-bedded sand and silt (Fig. 5).

Several diamicton layers are exposed in this section. The thickest layer is a massive unit that forms a wedge that thickens

from the crest of the moraine towards the proximal face (Fig. 5a). A prominent north-dipping foliation is outlined by sand stringers up to 5 cm thick. Striated boulders up to 50 cm long are oriented with long axes parallel to the sandy foliation. Grain size analyses of the matrix (Adams, 1984) show ranges similar to values from facies Dmm exposed in section A (45 to 60% sand, 20 to 40% silt, 15 to 20% clay). The massive diamicton overlies a deposit approximately 1.5 m thick made up of thin diamicton layers and sand beds that are recumbently folded. The limbs of the fold (approximately 10 m long) dip northward, parallel to the foliation and boulder fabric in the overlying unit. The fold nose closes tightly to the south. Several meters of medium to coarse, massive and plane-bedded sand (Sm, Sp, Fig. 5b) and graded to massive pebbly gravels underlie the diamicton layers and sand beds. The upper contact of the sand and gravel dips steeply northward (approximately 30^{0}) under the diamicton while the lower contact is sharp and horizontal. The section is capped by blocky structureless silt and poorly sorted silty sand similar to the uppermost sand beds in section A.

Section C. Section C is located in the northwestern corner of the Webber Pit (Fig. 2). The section, approximately 7 m high, exposes several layers of diamicton (Dmm(s), Dms; Fig. 2) separated by beds of medium to coarse sand and granular sand (Sp, Sr). Two layers of diamicton at the base of the section are olive-gray and similar in structure and composition to the resedimented diamicton (Dmm(r)) in section A. The diamicton units in C are massive in the center of the layer and contain sandy lenses and partings along both the upper and lower margins of the deposit. Grain size distribution of one sample from the center of the deposit (58% sand, 34% silt, 8% clay) is coarser grained than most of the facies Dmm, yet not as coarse as lithofacies Dmm(r) in section A (Fig. 3a). At the top of the lower diamicton unit, a large angular clast of the material is apparently detached from the diamicton and is contained within the overlying sand. Sandy lenses at the top and base of the main unit are deformed into sinuous wrinkles. Sand in these lenses is the same color and texture as the sand in the underlying unit and was probably incorporated during emplacement of the diamicton.

From approximately 3.50 to 4.50 m in this section, another diamicton lithofacies (Dms) includes thin continuous beds of sandy diamicton 5 to 10 cm thick separated by poorly sorted silty sand beds. The sandy beds contain intraclasts of the diamicton. Both the diamicton layers and the sand layers dip to the south.

Beds within the sand are massive, graded, plane-bedded, and ripple-drift cross-laminated (Type A), medium to coarse sand. The graded beds range in thickness from 1 to 50 cm. Grain size analysis from the top and bottom of a graded bed shows a

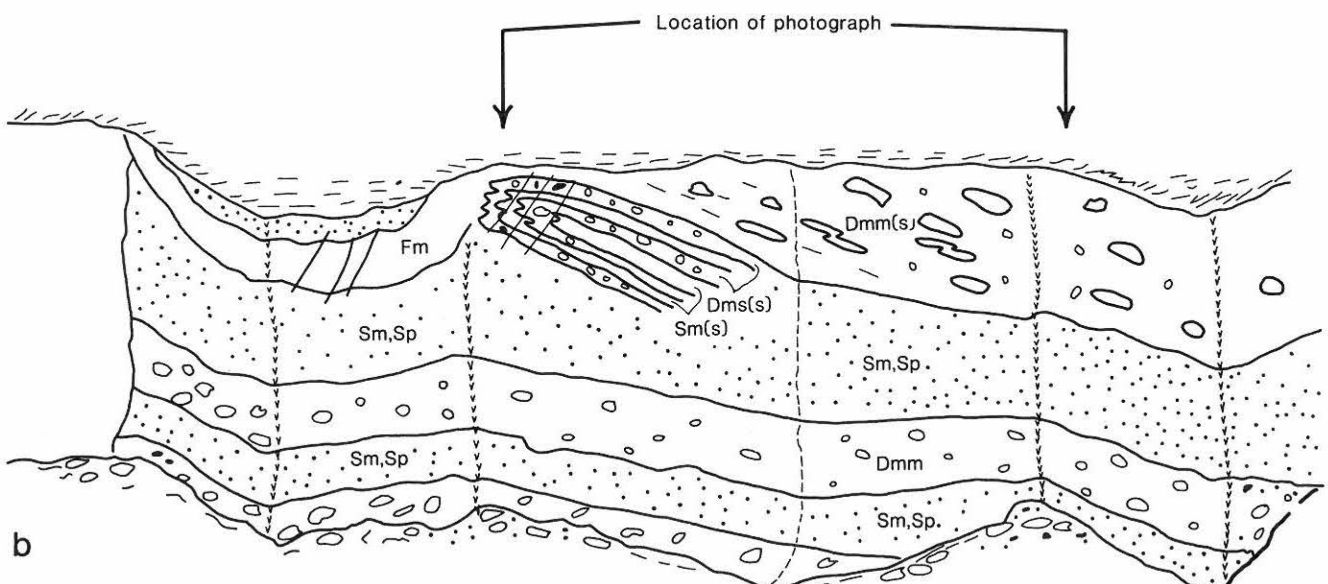


Figure 5. Section B, Webber pit. (a) Photograph of middle of section (approximately 5 m in height). (b) Outcrop sketch of section B with lithofacies codes for major units. Location of photograph 5a is shown by arrows.

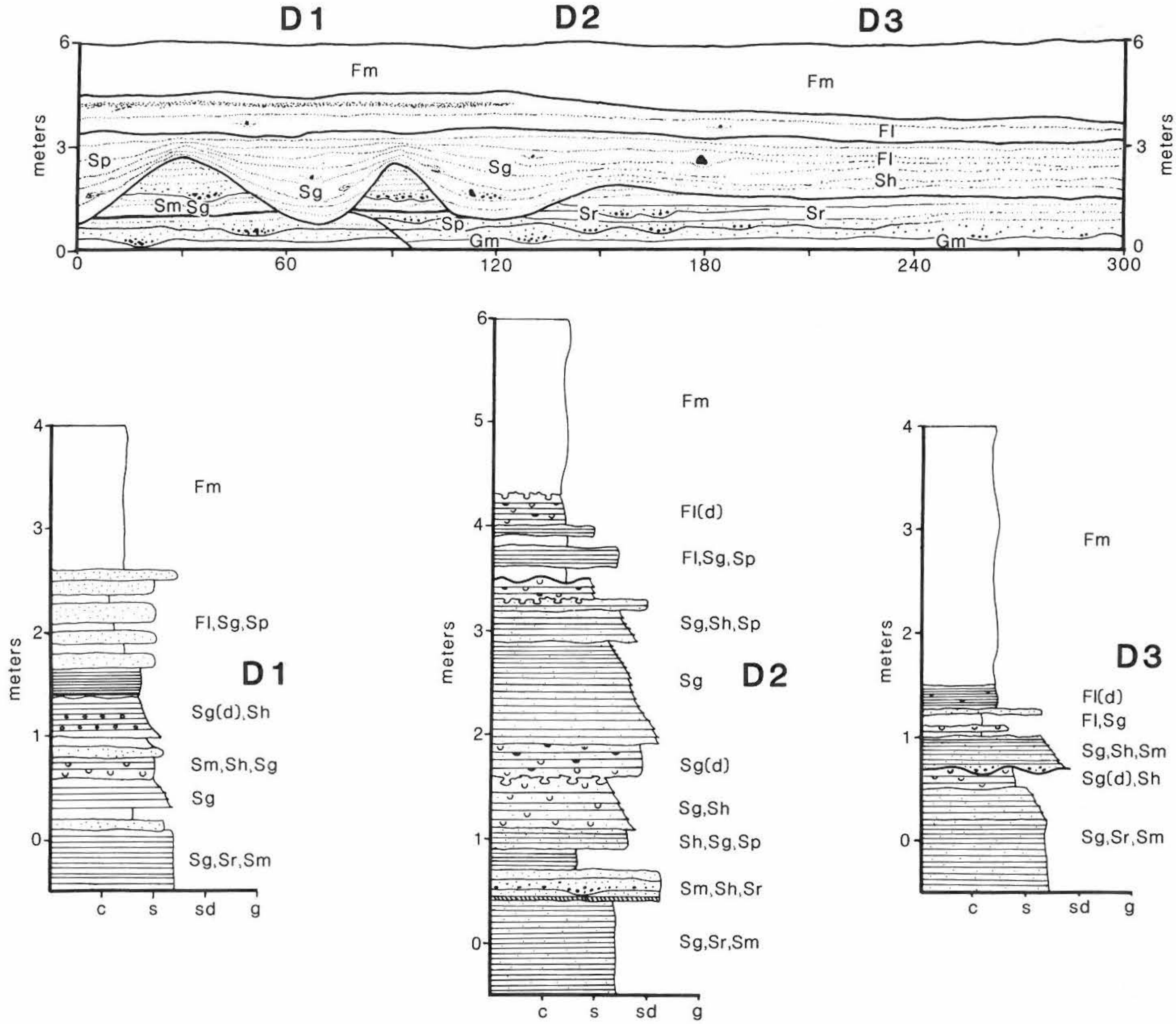


Figure 6. Lithofacies diagram and lithofacies logs within section D in Webber pit. Location of logs indicated above the diagram. Stratigraphic symbols after Eyles et al. (1983) and Miall (1978) (see Table 1).

mean grain size difference of 0.5 phi-units. The sand beds are sheared and cut by thrust planes that strike approximately N80°W and dip to the northeast. Overturned and thrusted silty drapes on top of ripple-drift cross-lamination display a shear rotation that parallels the orientation of the thrust planes, i.e., shear direction towards the south.

The top of the section is composed of south-dipping cross-bedded gravelly sand (Sp) that grades back to the moraine crest. The sand unit fines upward from gravelly sand at the base to coarse granular sand at the top of the section.

Section D. This section is exposed along a wall, from 4 to 7 m high, that extends along the southern margin of the Webber Pit (Fig. 2). In general, the section consists of a fining-upward sequence ranging from gravel facies to interlaminated mud and fine sand (Fig. 6).

The base of section D is dominated by coarse sand and gravel exhibiting channel scour and fill (Gm, Sm, Sp, Sg, Sr; Fig. 6). The channels are up to 30 m wide and 2 m deep. Generally, the channel fill deposits are structureless sand or gravel (Fig. 7a). However, larger channels are filled with plane-bedded and

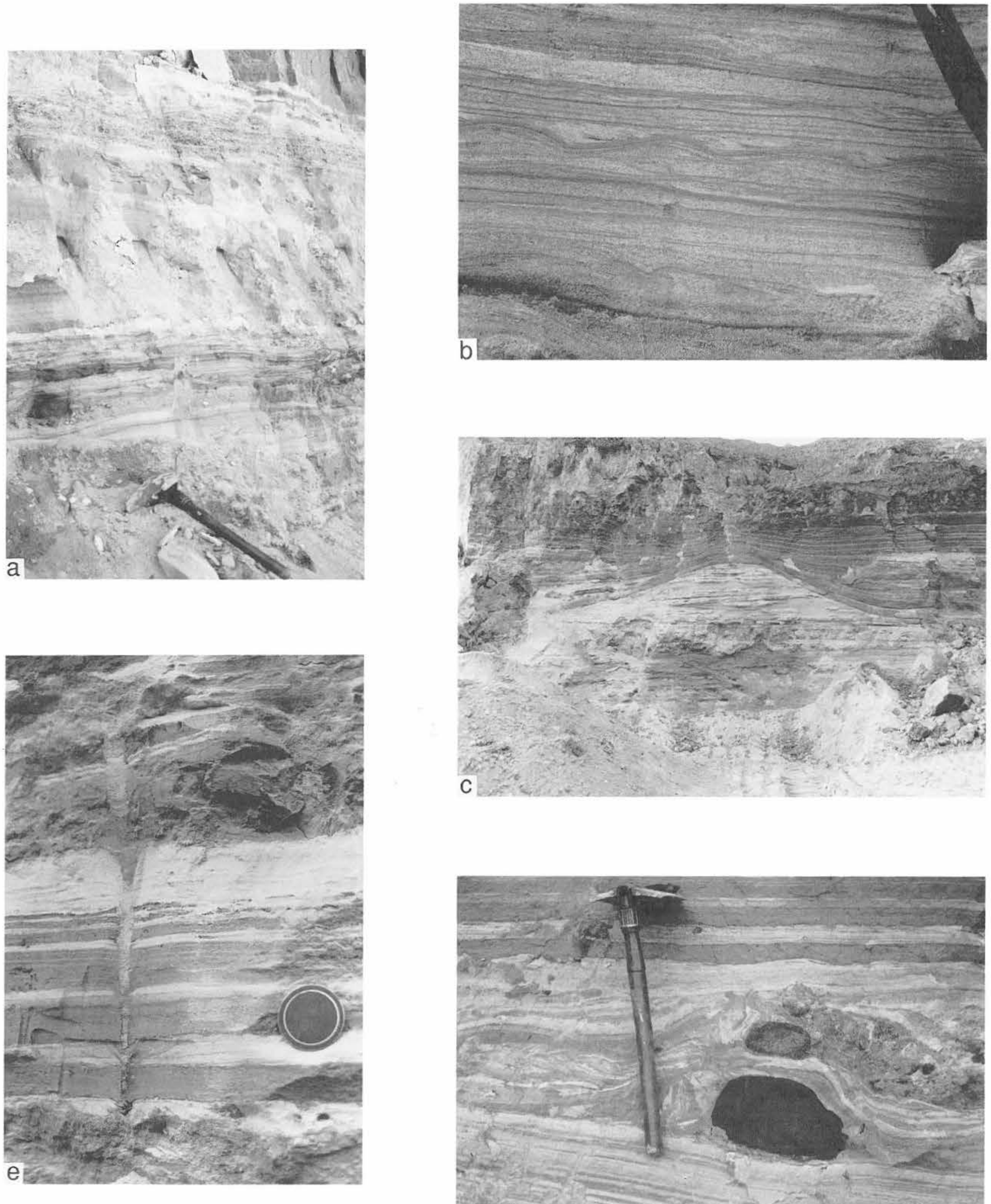


Figure 7. Photographs of section D within the Webber pit. (a) Gravel channel deposits overlain by massive and horizontal bedded sands at the base of the section. (b) Rippled and ripple-drift cross-laminated sand. (c) Rhythmically bedded sand and mud lithofacies draping a truncated fan lobe. (d) Laminated sand and mud containing a lens of ice rafted diamicton with accompanying soft sediment deformation. (e) Vertical burrow (?mollusc) cutting through laminated sand and mud.

planar cross-bedded sand. Above the channeled sand and gravel, the section contains interbedded graded sand and gravel beds with erosional scour contacts and gravel lag deposits. Sets of graded and planar-bedded sand range in thickness from 10 cm to 1 m. The interbedded sand and gravel is, in turn, overlain by rippled and ripple-drift cross-laminated sand (Fig. 7b) with sandy silt interbeds. Sand and silt beds range in thickness from 5 to 60 cm.

A distinct unconformity separates sand and gravel-dominated lithofacies near the base of the section from the overlying rhythmically bedded sand and mud lithofacies (Figs. 6, 7c). At approximately the 2 m level of the section (Fig. 6) the sand layer thickness is approximately equal to silt layer thickness. The sand is moderately- to moderately-well sorted while the silt laminae are very poorly to extremely poorly sorted. Sand and silt layers are laterally continuous for tens of meters. The proportion of mud increases upward within individual sand beds and, in general, up-section. Individual laminae thicken as they fill channels incised through underlying coarser facies (Fig. 7c) and thin as they drape over topographic highs. Soft sediment deformation is common especially where the laminated sand and silt drape over steep sloping channel sides. Isolated pebbles and diamicton lenses are common (Fig. 7d) in both sand and silt layers. Vertical burrows (Fig. 7e), probably formed by bivalves, are found in several horizons. The laminated sand and mud grades transitionally into structureless blocky marine mud (Fm) above the 3 m level (Fig. 6).

Section E. Approximately 2.7 m of sandy and muddy sediment overlies massive mud in section E in the northeastern area of the Webber pit (Fig. 2). The underlying mud (Fm) is massive blue gray plastic silty clay with horizons stained with black sulfide. The silty sand beds are intercalated with the clay near the top of the massive mud unit. Within the interbedded silt and clay, a rich *in situ* assemblage of invertebrate fossils is found. Paired valves of *Mytilus edulis* with *Balanus* attached on the upper valve lie on a pavement of articulated and broken shells including *Macoma calcarea* and *Hiatella arctica*. A sample of *Mytilus* and *Balanus* from this horizon at 47 m (152 ft) asl yielded a date of 12,820±120 yr B.P. (SI-7017; Retelle and Konecki, 1986a). The fossiliferous horizon is overlain by a 3 cm thick silty medium to coarse sand followed by 70 cm of massive to laminated brown sandy silt with lenses and laminae of medium sand. The upper 1.35 m of the section consists of beds of massive, moderately sorted sand with 0.5 to 1.0 cm laminae of silt. The deposit thins against the slope of the topography to the north.

Bisson Pit - Description of Sections

Section X. This exposure is located in the northwestern corner of the Bisson pit (Fig. 8) along a south facing hillslope. Bedrock crops out at the northern margin of the pit. Large boulders are common in the drift and on the surface in this area. At least 1 m of cross-bedded to planar bedded sand (Sp, Sh) occurs at the base of this section. This unit is overlain by approximately 1.5 m of faintly fissile, compact, olive (5Y 5/3) diamicton (Dms, Fig. 8). This diamicton matrix contains from 41 to 59% sand, 36 to 42% silt, and 5 to 17% clay. The range of mean grain size (3.9 to 5.5 phi) and sorting (2.1 to 3.7 phi) values are slightly finer-grained than other diamictons analyzed from the Webber pit, although there are abundant lenticular inclusions of well-sorted medium to coarse sand throughout the unit. A clast fabric on 25 stones shows a strong east-west orientation, perpendicular to regional ice flow direction. Clasts in the deposit, in general, lack silt caps and clay skins. The section is capped by approximately 2.5 m of well-sorted trough cross-bedded and planar-bedded sand. The dip of the bedding in the pit fans out from southwest to southeast, generally parallel with the topographic expression of the landform.

Sections Y and Z. These sections are lower in elevation and approximately 50 and 100 m to the southeast of section X, respectively, on relatively flat topography south of the bedrock ridge (Fig. 8). Section Y consists of approximately 4.5 m of vertical exposure in bedded and structureless sand and mud. Section Z, approximately 50 m to the east, exposes approximately the same stratigraphy as the upper 3 m of Y (Fig. 9). The bottom unit of the exposed stratigraphy is bedded medium to coarse sand with thin interlaminated silt. Sand layers are up to 20 cm thick in this part of the section and consist of massive, graded, and multiple graded beds (Fig. 10a). Most beds extend laterally for several tens of meters, however some beds pinch out or thin against bedrock or drift highs. Single sandy and muddy beds and sets of beds and laminae can be correlated between sections Y and Z. Dropstones are present, but not in great abundance. No *in situ* invertebrate fossils were found, however vertical burrows are present in some sand beds. Silt layers separating the sand beds are from 1 to 3 cm thick and are predominantly light olive-gray (5Y 6/2, damp) and structureless, although some contain 1-2 mm thick sand interlaminae. Other fine silt layers which occur in the section are dark olive-gray (5Y 4/2) and finer grained.

In general, sand thickness decreases up-section while mud thickness increases. Sand laminae up to 0.5 cm thick, separated by fine-grained silty clay layers 1 to 2 cm thick, are found up to approximately 3 m from the base of section Y (Fig. 9). Additionally, several continuous sand layers, each several grains thick, occur in the middle of the silty laminations (Fig. 9b). Some of the silty layers demonstrate a fining-upward trend (90-92 cm, Fig. 9b); others do not. This facies is transitional to the mud-dominated facies above and is designated Sm/Fm on the lithofacies log.

Above the 3 m level in section Y (and the 1.5 m level in section Z), the stratigraphy is dominated by fine grained laminated and structureless mud (Fig. 10b). Sand laminae thin up-section. Several 0.3 to 0.5 cm continuous sand laminae are found in the laminated mud facies (Fl) before disappearing completely (Fig. 10c). The mud laminae consist of alternating light olive-gray (5Y 6/2, damp) and olive-gray (5Y 4/2) couplets.

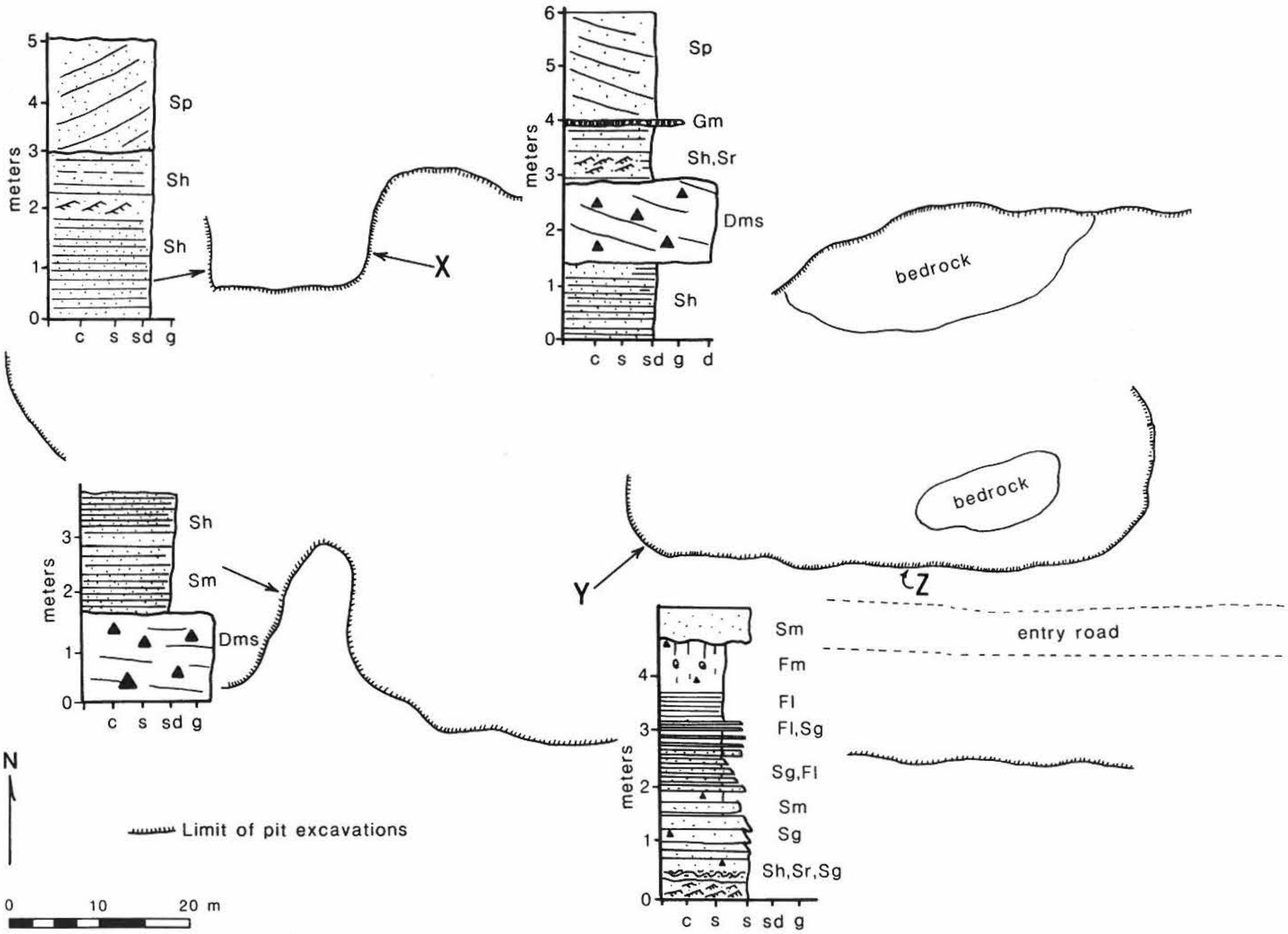


Figure 8. Map of the Bisson pit with lithofacies logs and locations of sections described in text.

Approximately 25 gray-olive couplets are represented in the sections. The lighter gray portion of the couplet in general thins up-section from 4 cm at the base of facies Fl, to several mm thick at the top. The light portion of the couplet commonly grades up to the dark portion of the couplet or is faintly draped by the darker. The contact between the dark olive-gray portion of the couplet and the overlying light layer is generally sharp. Some minor changes in grain size were measured through the light and dark olive-gray couplets (Fig. 9b). Sand content is slightly higher in the light layer; clay content is up to 10% higher in the dark layers. The light layers in the 175 to 190 cm level of the section (Fig. 9b) show fining-upward trends, while particle size within the dark layer remains relatively constant and slightly finer grained.

The laminated mud is overlain by a massive, generally structureless fine-grained mud (Fm). In general, the contact between lithofacies Fl and Fm is gradational. Texturally, lithofacies Fm is similar to the dark olive member of the laminated mud couplets, with low sand contents (1 to 5% sand,

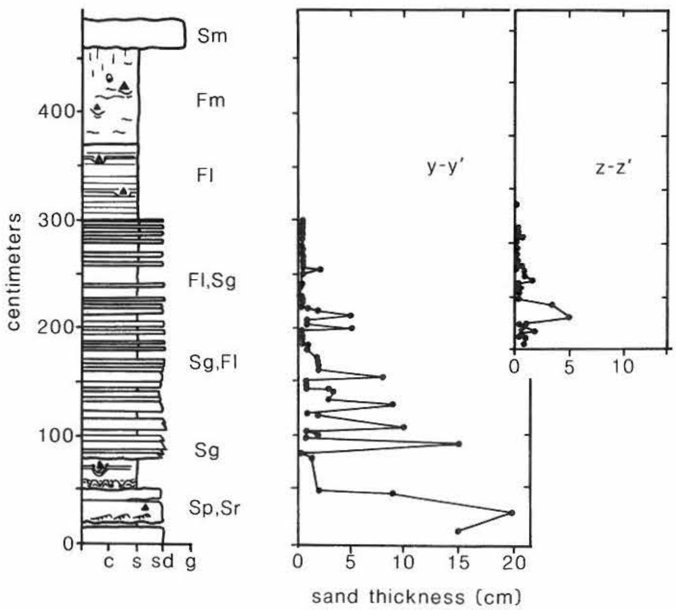


Figure 9. (a) Lithofacies log of section Y with measurements of sand bed thickness for sections Y and Z within the Bisson pit.

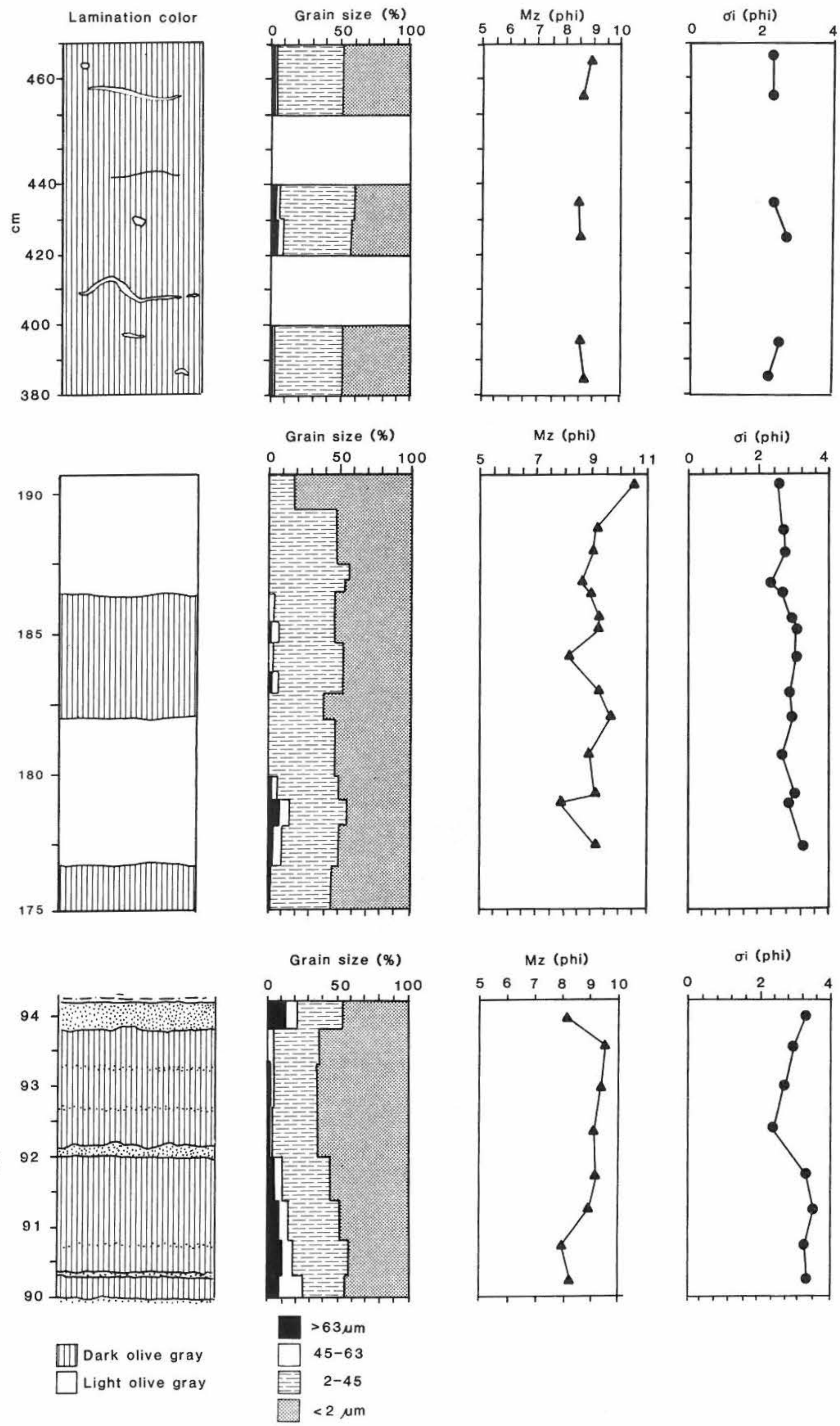


Figure 9. (b) Grain size analyses for selected zones of section Y.

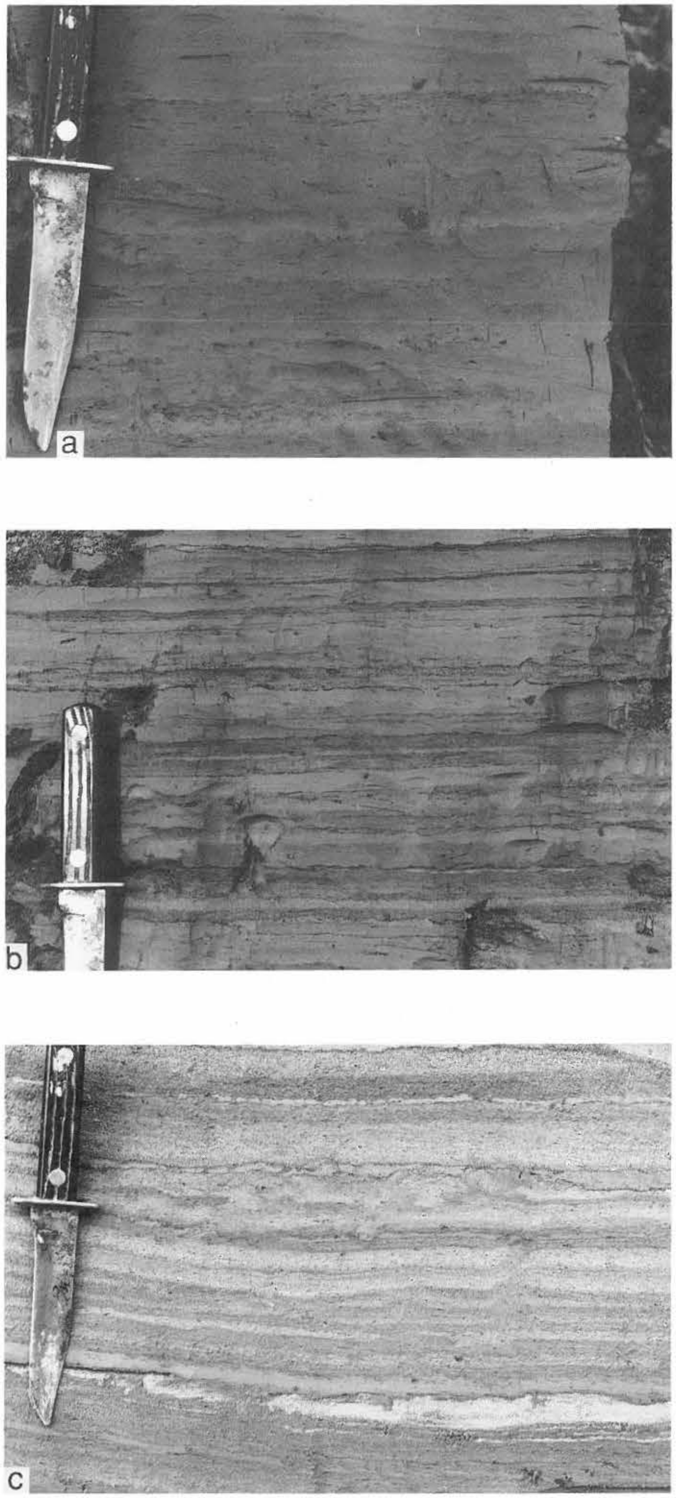


Figure 10. Photographs of sections Y and Z within the Bisson pit. (a) Faintly laminated silt unit with 0.5 cm sand bed, 175 cm level in section Z. (b) Laminated gray and dark olive-gray silts (lithofacies Fl) separated by 0.5 to 1.0 cm sand laminae, 90 cm level in section Z. (c) Multiple graded sand beds separated by 1 cm silt laminae, 80 cm from the base of section Y.

45 to 50% silt, and 40 to 50% clay; Fig. 9b). Thin, faint, light gray laminae (ca. 1 mm) also occur in the mud, but in minor proportion. Dropstones are found in this unit, but not in abundance. Sand grains are dispersed throughout the sediment, and do not occur in discrete laminae. Invertebrate fossil remains and related trace fossils are also found in this unit, however shells are commonly disintegrated or leached away leaving molds. Exposures of this unit commonly display blocky jointing in the upper meter. Manganese/iron oxide staining that probably results from ground water flow along the vertical joint planes gives a steel-gray to black coloration to jointed surfaces of this sediment.

DISCUSSION

Facies Assemblages

The deposits in the Webber and Bisson pits can be organized into four facies assemblages that, in this paper, are named for the specific morphostratigraphic unit in which they occur. These include the end moraine, submarine fan, submarine plain, and shallow marine facies assemblages. The facies assemblages contain lithofacies, such as Sp or Sm, common in both the end moraine and submarine fan assemblages. This is attributable to similar processes that were operating across facies assemblage boundaries in the glaciomarine environment. A model of the relationship between lithofacies and facies assemblages in this study area is shown in Figure 11.

End Moraine Facies Assemblage

The end moraine facies assemblage consists primarily of diamicton (lithofacies Dmm, Dmm(r), and Dms) interbedded with gravel and sand units, and is a composite feature constructed by glacial and fluvial processes. Structures within the diamicton support deposition and deformation by active ice and include oriented boulders, recumbent folds, and subhorizontal and arcuate foliation which may possibly be thrust faults. In addition, folds and thrust faults found in the interbedded sand probably result from ice-thrust deformation. The diamicton layers in the end moraine deposits are not continuous layers, but are tabular and wedge-shaped, suggesting disturbance or transport by ice shove. Diamicton bodies abruptly upwarped downglacier suggest deformation by loading. Andrews and Smithson (1966) and Andrews (1975) suggest that cross-valley moraines on Baffin Island are formed by the squeezing of water-soaked till from beneath the ice margin pinned on the proximal side of the moraine. Similarly, the drift ridge in front of the rapidly advancing Hubbard Glacier in southeastern Alaska consists of fine proglacial sediment incorporated and later squeezed out in front of the glacier in its advance across the fiord (L. Mayo, public commun., GSA, 1986).

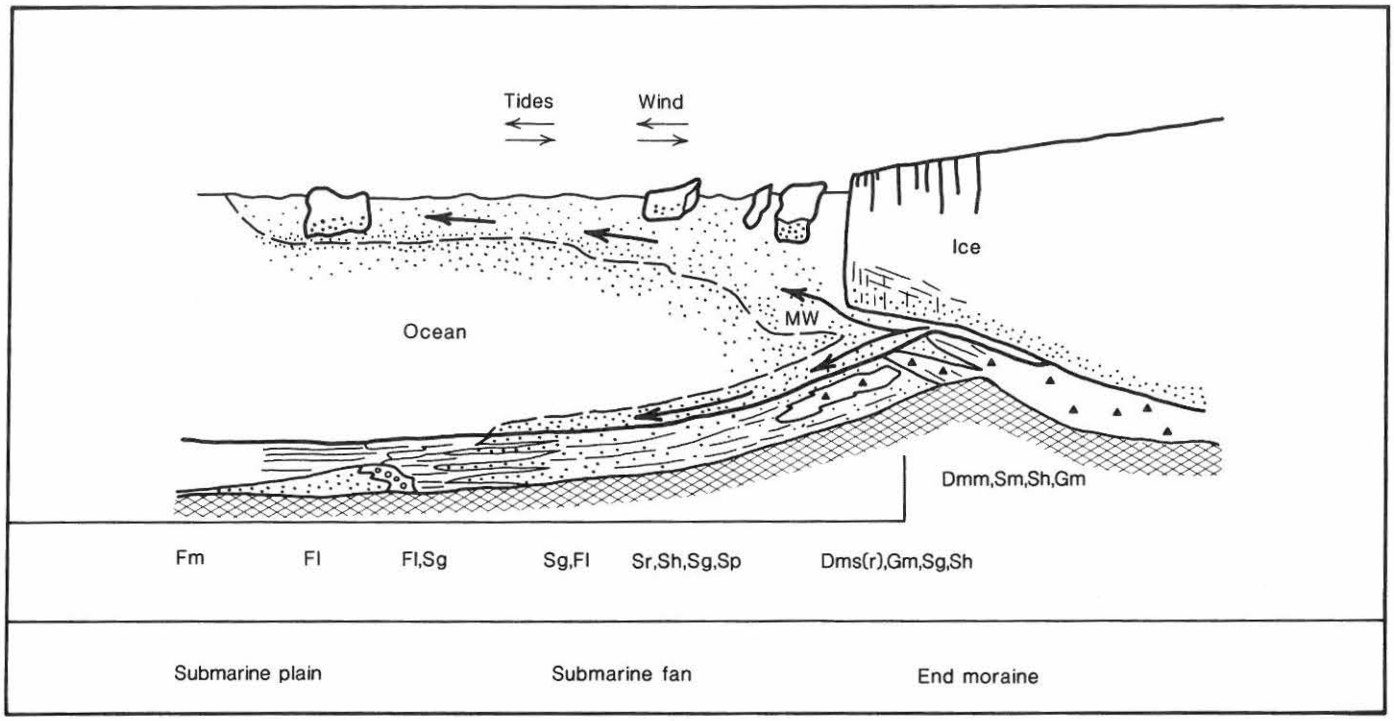


Figure 11. Proposed model of the relationship between lithofacies and facies associations for the lower Androscoggin River valley, Maine.

There are probably several origins for the diamicton in the end moraine complex. Some of the lithofacies Dmm is probably undisturbed lodgement till. In some locations, the diamicton referred to as Dmm(r) is nearly the same texturally, however fissility is usually lacking and thin sand partings and lenses are present. It is possible that this facies represents lodgement till which has been remobilized by ice-shove or slumping and has incorporated sand during transport. Diamicton lithofacies Dms is interstratified with sand and appears to be reworked along the margins of the deposit. These diamictons are interpreted to be either flow tills or debris flow deposits that originated either on a debris-covered ice foot or along the crest of an end moraine.

The occurrence of the sand and gravel facies within the moraine is attributed to meltwater deposition from a submarine outwash fan adjacent to, or draping over the moraine crest. The glaciofluvial sediment may also have been deposited in a subglacial regime in an esker tunnel or conduit (Sharpe, 1987) that fed the submarine fan. Multiple intercalations of the fluvial sediments with massive diamicton in the ridge exposures suggest that the ice margin was fluctuating near the grounding line. Studies by Stemen (1979), Jong (1980), Lepage (1979, 1982), and Smith (1982) demonstrate that in general, the coastal moraines in Maine are composed of any available materials at the ice front, and most commonly stratified drift. Thus, relative size of the moraines and complexity of the stratigraphy is probably a function of the available sediments and the amount of oscillation of the ice front. Multiple oscillations of the ice margin, and not rapid retreat, are necessary to produce complex stratigraphy and larger moraines. The occurrence of current-scoured pits and molding on till surfaces underlying coarse sandy beds suggests that the ice margin withdraws or the glacier bed decouples to allow sandy beds to be deposited. The accumulation of stratified diamicton in debris flows also requires temporary removal of the grounded ice margin from the moraine ridge or slumping of the moraine front before reincorporation into the moraine by the ice front. Such marginal fluctuations are common for tidewater glaciers and have been documented on a yearly cycle for the Columbia Glacier in southeastern Alaska (Meier, 1984).

The end moraine facies described in this study may be compared to the morainal bank lithofacies described by Powell (1981) from Glacier Bay, Alaska. In Glacier Bay, small moraines (push moraines) may be formed by advances of the ice margin in winter over previously deposited sediments on the fiord floor. The sediments can be a mixture of fiord-floor mud, gravel, and various diamicton facies. Larger morainal banks are generally constructed when ice retreat is halted at bedrock pinning points on the fiord floor or in a constriction or bend in the valley (Powell, 1981, 1983, 1984). The construction of the moraine in the Webber Pit was likely enhanced by slowed retreat as the tidewater glacier margin was pinned on the bedrock ridge (Retelle and Konecki, 1986b).

Submarine Outwash Fan and Submarine Plain Facies Assemblages

As is common with most of the facies assemblages described for this area, the submarine fan, submarine plain, and end moraine share common elements (Fig. 11 and Table 1). Where meltwater streams flowed from subglacial tunnels across end moraines at the sea floor, submarine outwash fans were constructed (cf. Rust and Romanelli, 1975; Rust, 1987; Sharpe, 1987). Along moraine crests, submarine outwash sediments may simply be draped over the top of the moraine or grade to a tunnel source along the moraine crest. Additionally, if the glacier margin readvances over the fan, diamicton (lodgement till or flow till) or other sediments may be emplaced over the top or shoved into a moraine ridge. The ice proximal zone of the fan complex can have a wide variety of lithofacies due to fluvial, glacial, marine, and gravity processes acting in the area (Powell, 1984). Within the submarine fans exposed in both the Webber and Bisson pits, bodies of diamicton, probably redeposited lodgement till, lie conformably within fluvial beds and probably represent slumping from the end moraine. Slumps can be caused in this environment by glacier push, berg calving, earthquakes, or current-induced slumping of water saturated sediment (Powell, 1983; Schwab and Lee, 1983).

The dominant sediment types of the submarine fan assemblage are the sand and fine sediment (mud) lithofacies. Theoretically, strong subglacial meltwater traction currents with high suspended sediment concentration (Powell, 1984; Mackiewicz, et al., 1984) disperse coarse gravel to sand from tunnel mouths across fan delta surfaces and along the sea floor until the inertia of the current can no longer be sustained within the marine waters of the fiord or sea bottom. Differences in density between the saline marine waters and the sediment-charged meltwaters would likely be reduced during the melt season because of the increase of freshwater near the ice front during the melt season (Gilbert, 1982; Stewart, 1982; Domack, 1984). At the same time, finer, and less dense sediments would separate from the tractive underflow and disperse at levels in the water column that are of equal density as either interflows, or overflows. Additionally, underflows may be generated from slumping of sands that pile up at the fan head or along the moraine. All three phases of flow have been observed in the modern glaciolacustrine environment (cf. Gustavson, 1975; Smith, 1978). In the marine setting, overflow plumes issuing from subglacial sources have been observed in many fiords and inlets (Sharma and Burrell, 1970; Hoskin and Burrell, 1972; Gilbert, 1982; Powell, 1983; Elverhoi, 1984; Mackiewicz et al., 1984). Powell (1984) and Mackiewicz et al. (1984) point out that interflows and underflows are not well understood in the modern environment and are modeled from observations in Pleistocene sediment assemblages (cf. Rust and Romanelli, 1975).

In the Webber and Bisson pits, ice frontal position, and hence distance from the ice margin to site of deposition, can be constrained for coarse fan sediments. Coarse-grained gravel and current-bedded sand facies (Sr, Sp) were presumably deposited by traction currents issuing from meltwater flow at a tunnel mouth, or from turbidity currents flowing down the moraine bank or submarine fan. The meltwater source probably originated within 200 to 300 m of the exposure. Northward-sloping topography behind the northern margin of the Webber pit would provide an upsloping ramp, precluding underflow to the southern margin of the pit once the grounded ice margin had retreated a short distance north of the moraine-fan complex. Some sands, particularly those dispersed within the laminated silt layers and one- to two-grain-thick layers in the silts may be transported distally by overflow currents (cf. Gilbert, 1982). However, repetitive sequences of laterally extensive massive to graded sandy beds (Figs. 7b,c) require a topographic gradient back to a source at a fan head. Elverhoi (1984) describes a similar topographic control on laminated glaciomarine sediment deposition in front of the Kongsvegen glacier in Spitsbergen and interprets the interlaminated muds as deposited from an overflow plume.

Laminated fine grained sediments (lithofacies Fl) represent the transition from coarse laminated sand deposits on the fan to more massive clayey silt above (lithofacies Fm). Decreased occurrence of sand laminae up-section indicates the removal of proximal meltwater source, although lateral to the central axis of the fan, the deposition of fine, laminated sediments was probably coincident with coarse-grained deposition. The graded nature of the light gray member of the laminated silt couplet may reflect either distal turbidite deposition or sedimentation from an overflow plume with some cyclical sorting of the fine sediment. Similar laminated muds in the glaciomarine environment have been interpreted as varves (Stevens, 1985, 1986), rhythmites (Domack, 1984), and cyclopels (Mackiewicz et al., 1984).

Stevens (1985) interprets the occurrence of laminated glaciomarine muds in Sweden to represent annual ice-proximal deposition of a fine-grained sediment couplet from an overflow plume. Size separation, which defines the varves, results from the stratification of the marine water body during the meltwater runoff season. The coarser-grained sediments of the couplet would reflect increased meltwater discharge and better developed stratification in spring and early summer, while the darker, finer laminae would represent decreased runoff and sedimentation of floccules when the pycnocline was higher in the water column.

Domack (1984) also suggests that rhythmically laminated sediments in the Puget Sound Lowlands may be the result of seasonal deposition in proximal locations from underflow and overflow. Thickness and grain size of beds and laminations and variations within them may reflect seasonal and diurnal changes in meltwater flux and other cyclical processes such as tidal currents (Domack, 1984).

Laminated fine-grained sediment couplets in Muir Inlet, Alaska (Mackiewicz et al., 1984, Cowan et al., 1988) termed cyclopels, are typically 0.5 to 1.0 cm thick and are interpreted to be deposited from suspension loads of interflow and overflow

plumes at a rate of 2 to 3 cycles per day, accounting for up to 9 meters of cyclopel mud per year. Variabilities of cyclopel thickness at any location in the fiord and between locations were attributed to plume width and extent from the ice, which is governed by a combination of strength and timing of both meltwater flux and tidal currents (Mackiewicz et al., 1984).

Laminated fine sediments in the Webber and Bisson pits display general thinning- and fining-upward trends that have been documented in other areas discussed above. Similarly the gray, or lighter, component of the couplets is generally the coarser of the two laminae and displays a fining-upward trend within the lamination and up-section. Very thin (mm), light laminae commonly occur in the upper, apparently structureless, unit assigned to the submarine plain lithofacies. This suggests that possibly the light layer may be the result of size sorting in the overflow plume as suggested by Stevens (1985). Light-colored laminae appear to be more graded, though still poorly sorted, indicating a crude grain size organization in the water column. Decreasing influence of the plume is shown by the thinning of the light laminae except during conditions which allow the plume to extend farther from the ice front. This could occur during periods of increased melt and/or low or receding tide (Mackiewicz et al., 1984). "Normal" distal sedimentation comprises the deposition of finer mud, which is the dark laminae of the couplet. This may be accomplished by flocculation of fine material when the freshwater plume is less extensive and fine clay-size particles can interact with saline waters.

Shallow Marine Facies Assemblage

Distribution of lithofacies in this assemblage primarily reflects the underlying parent material, topography of the landform and surrounding landforms, and processes of erosion and deposition which are likely a function of water depth. Subtidal, lagoonal, and beach environments are included in this facies assemblage. Water depth or relative sea level in the region is related to isostatic rebound and eustatic sea level changes. At this stage of retreat of the ice sheet, isostatic effects dominate, and emergence of the formerly submerged marine terrain occurs (Stuiver and Borns, 1975; Belknap et al., 1986; Belknap, 1987). Sediments range in size from coarse gravels to fine massive muds. Along the moraine crest in the Webber pit (sections A and B) a veneer of poorly sorted, sandy, openwork gravel overlies deposits of diamicton. Broken shell fragments, most commonly those of *Mytilus edulis*, are commonly found in the gravel. Poorly sorted, massive, silty sand drapes the slopes of the moraine and interfingers with massive silt along the moraine. Sand of a similar texture overlies massive marine mud along the southern margin of the Webber pit and also in most of the exposures in the Bisson pit.

The most complex exposure of shallow marine sediments is composed of massive and structureless silty clay grading upward to massive sand beds separated by thin silt laminae. The rich in situ fauna at the base of the sands is distinctly a shallow water, or intertidal, fauna consisting of mussels with attached barnacles. Black sulfidic staining in the fine sediments in this zone is common in tidal marsh or lagoonal sediments and represents a micro-reducing environment associated with the decomposition of organic matter associated with the organisms in the fossil assemblage. Similar assemblages are documented from local deposits in the Pejepscot area of Topsham (Attig, 1975) and in many localities in the state (Stuiver and Borns, 1975; Smith, 1985), and represent important relative sea level data if the assemblage can be associated with a shoreline deposit.

CONCLUSIONS

The glacial and glaciomarine stratigraphy exposed in the Webber and Bisson pits is organized into four facies assemblages: end moraine, submarine fan, submarine plain, and shallow marine. Individual lithofacies within each facies assemblage are often common to other assemblages, demonstrating the transition in sedimentary processes between the ice margin and the adjacent sea floor.

The end moraine assemblage is a composite of interbedded subglacial, submarine outwash, and mass flow deposits which contain evidence of deformation by an active, fluctuating, grounded ice margin.

The submarine fan sediments originate at fan heads along the former ice margin and grade distally to the sea floor. Along the ice margin the fan contains submarine outwash gravel and sand and diamicton deposited by mass flow. Distal and lateral to the ice margin, fan sediments consist of sand deposited by turbidity current, and possibly underflow and rhythmically bedded silt deposited from overflow plume suspension. Structureless fine-grained sediments comprise the marine plain assemblage and represent sedimentation from suspension distal and lateral to the effects of underflow currents and overflow plumes.

The shallow marine assemblage is a texturally variable but ubiquitous deposit developed during regression of the inland sea. The deposits occur either in transition with the underlying marine units or lie unconformably over an erosional contact.

ACKNOWLEDGMENTS

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Stratified Waterlain Glacigenic Sediments and the "New Sharon Soil," New Sharon, Maine

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ABSTRACT

Quaternary deposits exposed in the Sandy River valley at New Sharon, Maine, include stratified waterlain glacigenic sediments which are comprised of interlayered fine-grained sand, silt, clay, and diamicton. At another location, 30 m downstream from this section, is the site of an organic-bearing unit, previously described as a two-till locality.

The glacigenic sediments are interpreted primarily as debris-flow deposits with interbedded basal till, which accumulated in a proglacial lake and were subsequently overidden by advancing Late Wisconsinan ice.

The origin of the organic-bearing unit is unclear. However, while excavation of the unit did not reveal a two-till stratigraphy, it did expose glacigenic sediments similar to the section upstream, overlying organic-bearing silt and peaty, sandy gravel. Previous interpretations of the New Sharon stratigraphy and their regional significance based on the "New Sharon soil" warrant re-evaluation.

INTRODUCTION

The Sandy River valley at New Sharon, west-central Maine, contains some of the best exposures of Quaternary deposits in New England. It is also the site of the only surface locality on the New England mainland where organic material is reported to occur between older- and younger-aged tills (Caldwell, 1959, 1960; Schafer and Hartshorn, 1965; Mickelson et al., 1983; Stone and Borns, 1986; Borns et al., 1987). The stratigraphy is considered by many workers to be critical to understanding the glacial history of New England, in particular the number of regional ice advances which have occurred in the area.

There are several exposures of glacigenic sediments along the Sandy River from New Sharon downstream to Mercer. At one of these exposures (Site A, Fig. 1), there is a section of stratified waterlain glacigenic sediments comprised of thinly-bedded silt, fine-grained sand, clay, and diamicton, as well as thickly-bedded diamicton. Approximately 30 m downstream from Site A is another section which contains an organic-bearing unit, but is presently covered by stream alluvium (Site B, Fig. 1, this paper; locality C of Caldwell, 1959).

The organic-bearing unit at New Sharon has been described by others as a brown silt and a buried soil (Caldwell, 1959, 1960), buried non-glacial sediments (Borns and Calkin, 1977), a sub-aerial weathering profile containing organic debris (Mickelson et al., 1983), a nonglacial soil (Oldale and Eskenasy, 1983), the New Sharon soil (Hanson, 1984), an organic silt layer (Koteff and Pessl, 1985), and a paleosol (Stone and Borns, 1986). Some of these workers, as well as others, have suggested regional correlation of till stratigraphy from southern New England to southeastern Quebec, and have based part of their correlation on the New Sharon stratigraphy (Borns and Calkin, 1977; Oldale and Eskenasy, 1983; Hanson, 1984; Koteff and Pessl, 1985; Thompson and Borns, 1985a; Stone and Borns, 1986; Dredge and Thorleifson, 1987; Vincent and Prest, 1987).

Although regional correlations are not presented in this paper, the earlier interpretations of the New Sharon stratigraphy are being re-evaluated (Weddle, in prep.). The purpose of this paper is to discuss the origin of the glacigenic sediments at Site A and their relation to the organic-bearing unit at Site B (Fig. 1).

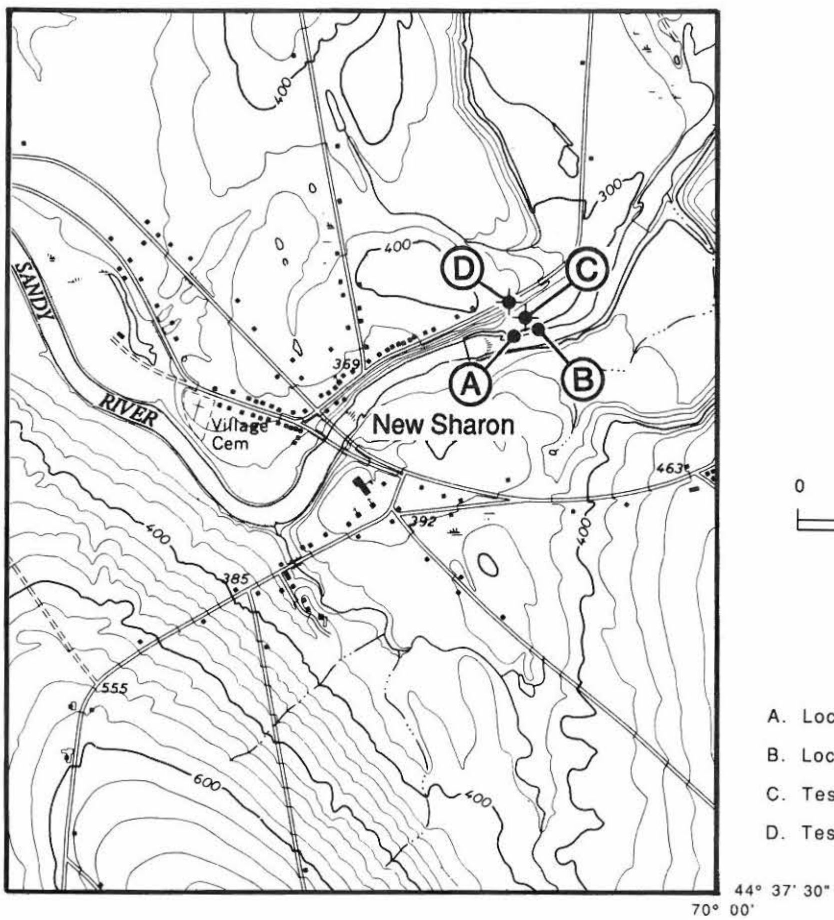


Figure 1. Study site. (A) indicates location of stratified waterlain glacigenic sediments. (B) indicates location of organic-bearing unit (from New Sharon 7.5 minute USGS topographic map).

PREVIOUS WORK

Caldwell (1959) was the first to refer specifically to the sections along the Sandy River at New Sharon. However, Clapp (1906, 1908) referred to an old till exposed at Norridgewock, a town approximately 8 km northeast of New Sharon. Unfortunately, the location was not accurately described, and could not be found by the author. The Norridgewock site was mentioned by Leavitt and Perkins (1935), but was not exposed at the time of their study. However, the section at New Sharon was reportedly known by Dr. E. H. Perkins to contain organic macrofossils (H. W. Borns, Jr., pers. commun., 1986), though it is not mentioned in Leavitt and Perkins (1935). Dr. Joseph M. Trefethen, Maine State Geologist Emeritus, was a field assistant to Perkins on the survey for the Leavitt and Perkins (1935) volume, and recalls Perkins' mention of New Sharon as an organic-bearing site (J. M. Trefethen, pers. commun., 1987).

Caldwell (1959, 1960) presented the earliest comprehensive work on the sections at New Sharon. Locality C of Caldwell (1959), from which wood was collected and radiocarbon dated at >38,000 yr B.P. (Y-689, Caldwell, 1960; W-910, Schafer and Hartshorn, 1965), is described from top to bottom as follows: 0.75 m of blue-gray till, over 0.1 m of gray varved clay, over 0.15 - 1.0 m of brown, sandy organic-bearing silt, over 0.25 - 0.8 m of green-gray silt with its upper surface oxidized 0.15 - 0.3 m deep, over 0.3 m of green-gray till. Caldwell (1959) suggested that the varved clay was deposited in a proglacial lake, which existed at the margin of the Late Wisconsinan ice sheet during its advancing phase, and was subsequently overidden by that ice. Caldwell (1960) tentatively assigned a Late Wisconsinan age to the upper till unit, and an Early(?) Wisconsinan age to the lower till unit. Later, Caldwell and Pratt (1983), and Caldwell and Weddle (1983) suggested that Middle Wisconsinan aged till may also be present at New Sharon.

Borns and Calkin (1977) drilled a continuous core (Fig. 1) through glacigenic sediment at New Sharon, recording multiple-till layers; however, the boring did not penetrate the organic-bearing sediment. They also submitted wood for radiocarbon dating that yielded an age >52,000 yr B.P. (Y-2683), and suggested two alternative explanations for the multiple-till stratigraphy. The units could represent local, ice-marginal fluctuations, or they could be of regional stratigraphic importance, representing successive, major ice advances across Maine.

The latter interpretation is favored by Borns and Calkin (1977), and they tentatively assigned Early, Middle, and Late Wisconsinan ages to units at New Sharon, correlative with the Quaternary stratigraphy of southeastern Quebec described by MacDonald and Shilts (1971). This stratigraphy and correlation is also proposed by Stone and Borns (1986) to account for the New Sharon section in the regional Pleistocene stratigraphy of New England.

However, Thompson and Borns (1985a) suggest that much of the section at New Sharon could be pre-Late Wisconsinan in age, and Borns et al. (1987) describe the section at New Sharon as consisting of a Late Wisconsinan till overlying an organic silt, in turn underlain by two till units. This latter stratigraphic description is not consistent with the earlier assignments by Stone and Borns (1986) and by Borns and Calkin (1977) of two regionally significant tills, described from the test boring, which presumably overlie the organic-bearing unit and the lower till described by Caldwell (1959).

Figure 2. Stratified waterlain glacigenic sediments along the Sandy River at New Sharon (Site A on Fig. 1).

SITE A

Facies Description

The section exposing the glacigenic sediments (Site A on Fig. 1) occurs along the north bank of the Sandy River (Fig. 2) and is approximately 4 m thick. A schematic stratigraphic section of the exposure showing facies is represented in Figure 3. The sequence is characterized by two general facies recognized at the site; diamictons and fine-grained units (Fig. 3, Table 1).

Diamictons

The diamictons (Figs. 3, 4, Table 1) vary in grain size from clay-sized particles to clasts mostly 1 - 3 cm in length, with some boulders greater than 0.5 m in length. Most clasts are subrounded to subangular, and many are striated. The color of the diamictons is uniformly olive-gray (Munsell color 5Y 5/2). Metasiltstone and metasandstone are the dominant clast lithologies, with less common granite, granite-gneiss, gabbro, metavolcanic, quartz, and metaconglomerate clasts also present.

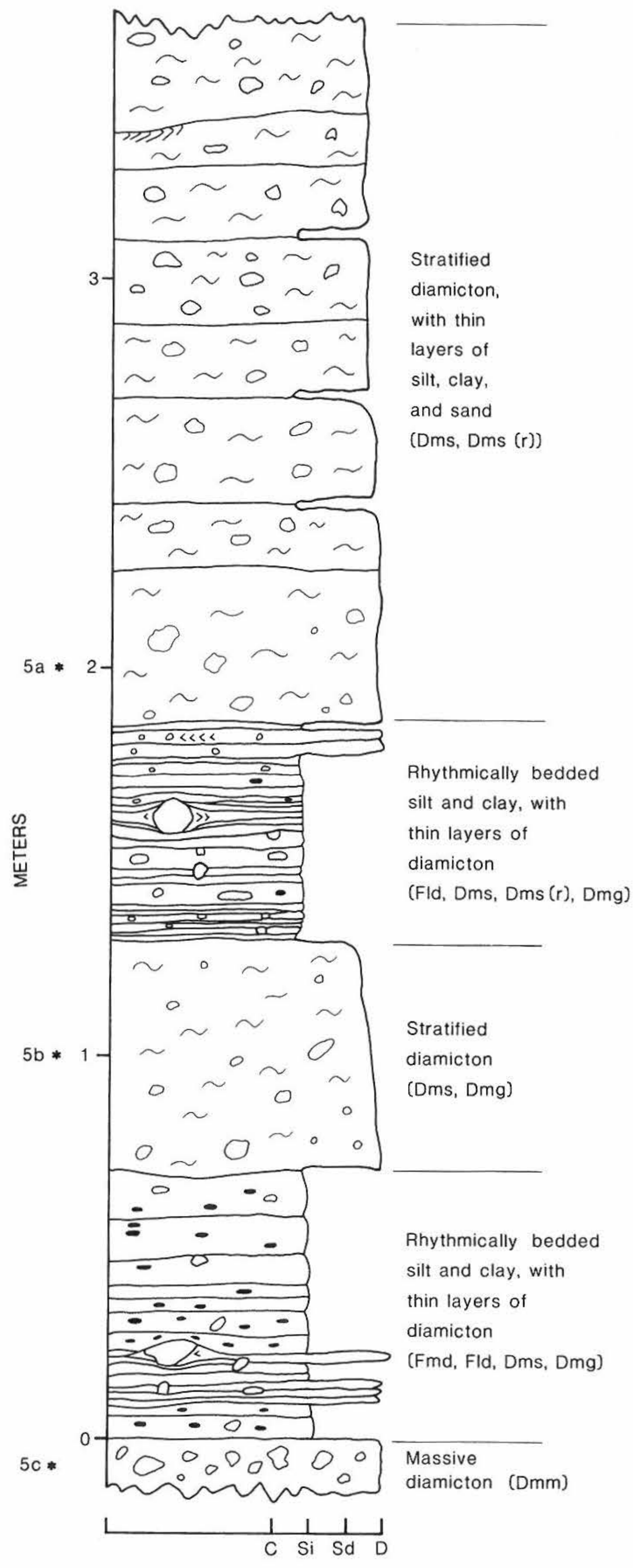
The diamictons generally are matrix supported, with crude stratification in most units. Normal grading is also present, and stringers of fine-grained sand are found in places. Thickness of the diamicton layers varies from a few cm to about 1 m. The contacts between beds are generally sharp and conformable, although rare erosional contacts can be found.

Pebble orientations were measured from discrete diamicton layers, and their fabrics were plotted (Fig. 5). Clasts with long to intermediate axial ratios of at least 3:2 were measured, and the trend and plunge of the long axis was recorded, plotted on an equal-area net (lower hemisphere projection), and contoured using a computer program written by C. E. Corbato (Ohio State University) as modified by D. E. Lawson (Cold Regions Research and Engineering Laboratory). Petrofabric orientations were evaluated statistically using the eigenvalue method of Mark (1973, 1974) as modified by D. E. Lawson. In this method, the significance value (S_1) of the fabric represents the strength of clustering about a mean axis, where a value of 1.00 is uniform and 0.00 is random.

Figures 5a and 5b (Site A on Fig. 1; Fig. 3) have preferred orientations along a northwest-southeast trend, but 5a is polymodally distributed, with weaker orientations along east-northeast – west-southwest trends. The S_1 values for 5a and 5b are 0.60 and 0.70, respectively. Fabric 5c, from the lowest unit exposed at Site A, has a strong preferred NW-SE trend. Clast

TABLE 1. FACIES DESCRIPTION
(after Eyles et al., 1983).

CODE	FACIES	DESCRIPTION
Dmm	Matrix-supported, massive diamicton.	Structureless mud, sand, pebble admixture.
Dms	Matrix-supported, stratified diamicton.	Obvious textural segregation or sorting within unit. Stratification more than 10% of unit thickness.
Dms(r)	Dms with evidence of resedimentation.	Flow noses present; diamicton contains deformed silt/clay laminae and stringers, and rip-up clasts. Grading present; clast fabric random or parallel to bedding. Erosion and incorporation of underlying units.
Dmg	Matrix-supported, graded diamicton.	Unit exhibits vertical grading in matrix or clast content.
Fld	Fine-grained, laminated unit.	Laminated silt and clay with dropstones.
Fmd	Fine-grained, massiv unit.	Massive silt and clay with dropstones.



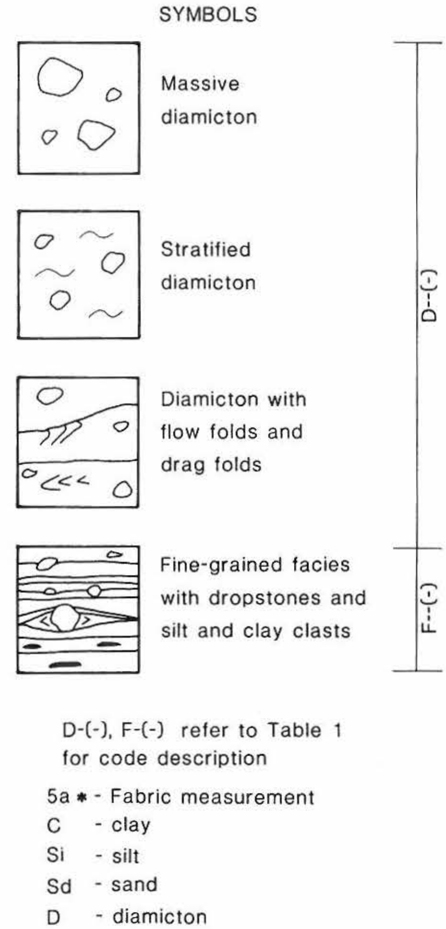
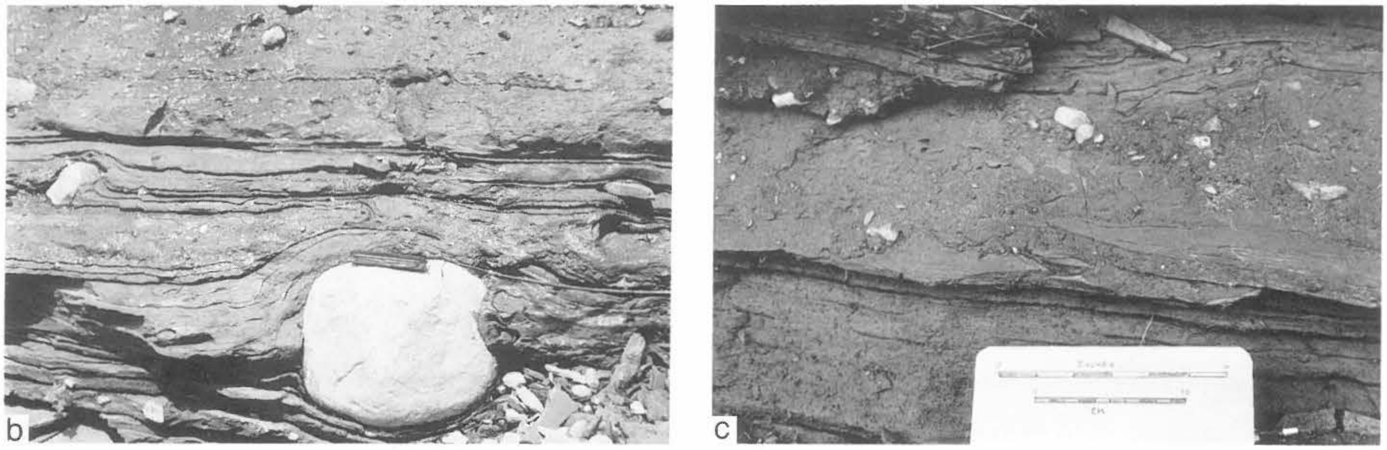
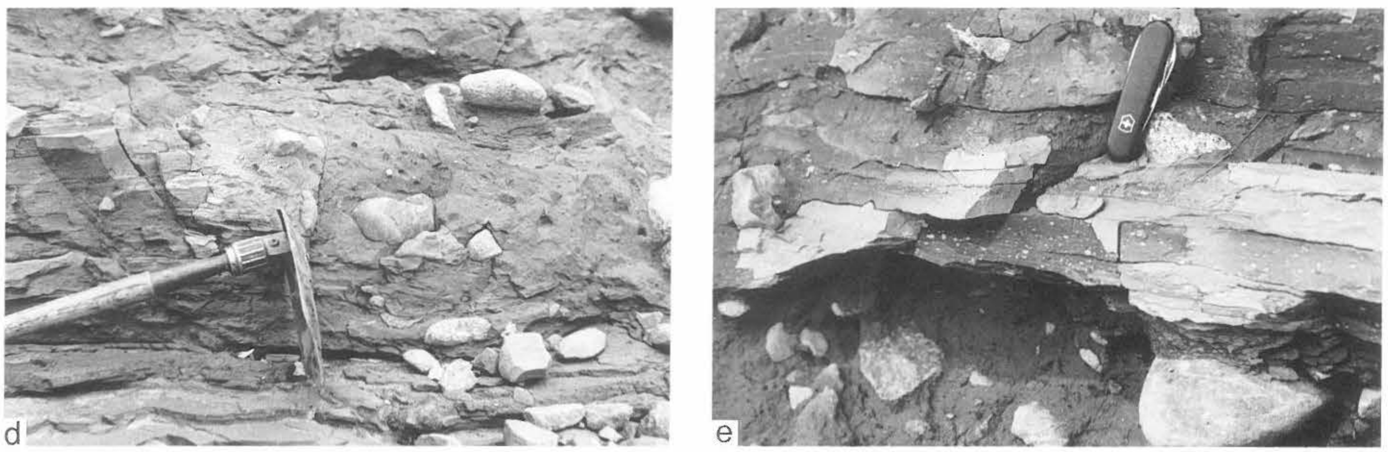


Figure 3. Schematic stratigraphic section of sediments exposed at New Sharon (Site A).

Figure 4. Features related to sediment flow and subaqueous deposition at New Sharon. (a) Flow nose in fine-grained deposits.





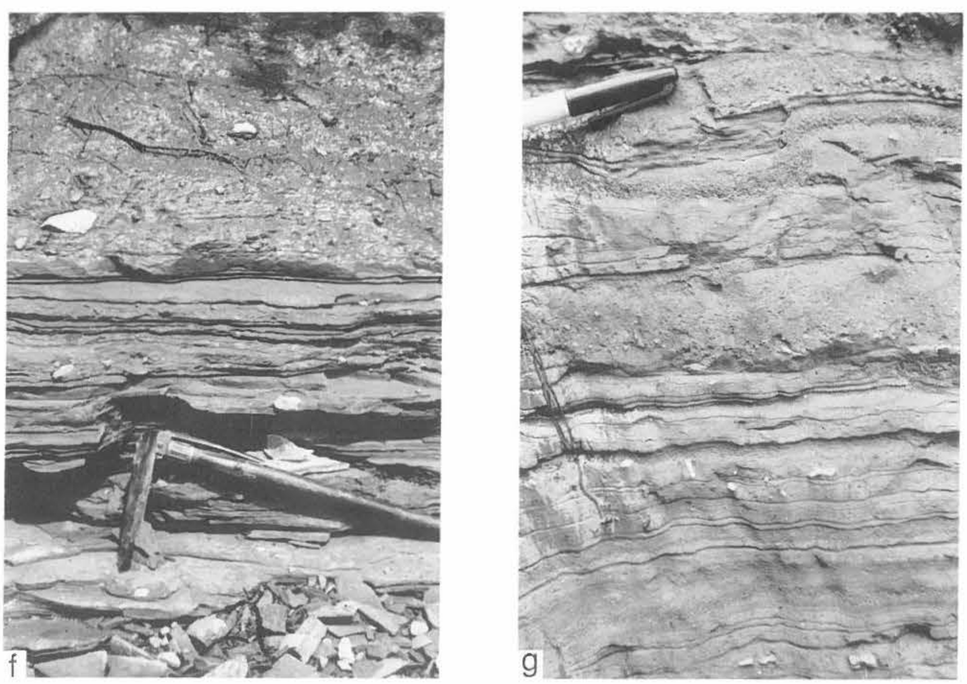


Figure 4 (continued). Features related to sediment flow and subaqueous deposition at New Sharon. (b) Non-striated outsized clast (dropstone?) in fine-grained deposits. (c) Monolithologic, subangular rip-up clasts of fine-grained unit in diamicton. (d) Erosional contact between fine-grained unit (below) and diamicton unit (above). (e) Dropped silt clasts in fine-grained units. (f) Conformable contact between fine-grained unit (below) and diamicton unit (above). (g) Graded beds in fine-grained unit.

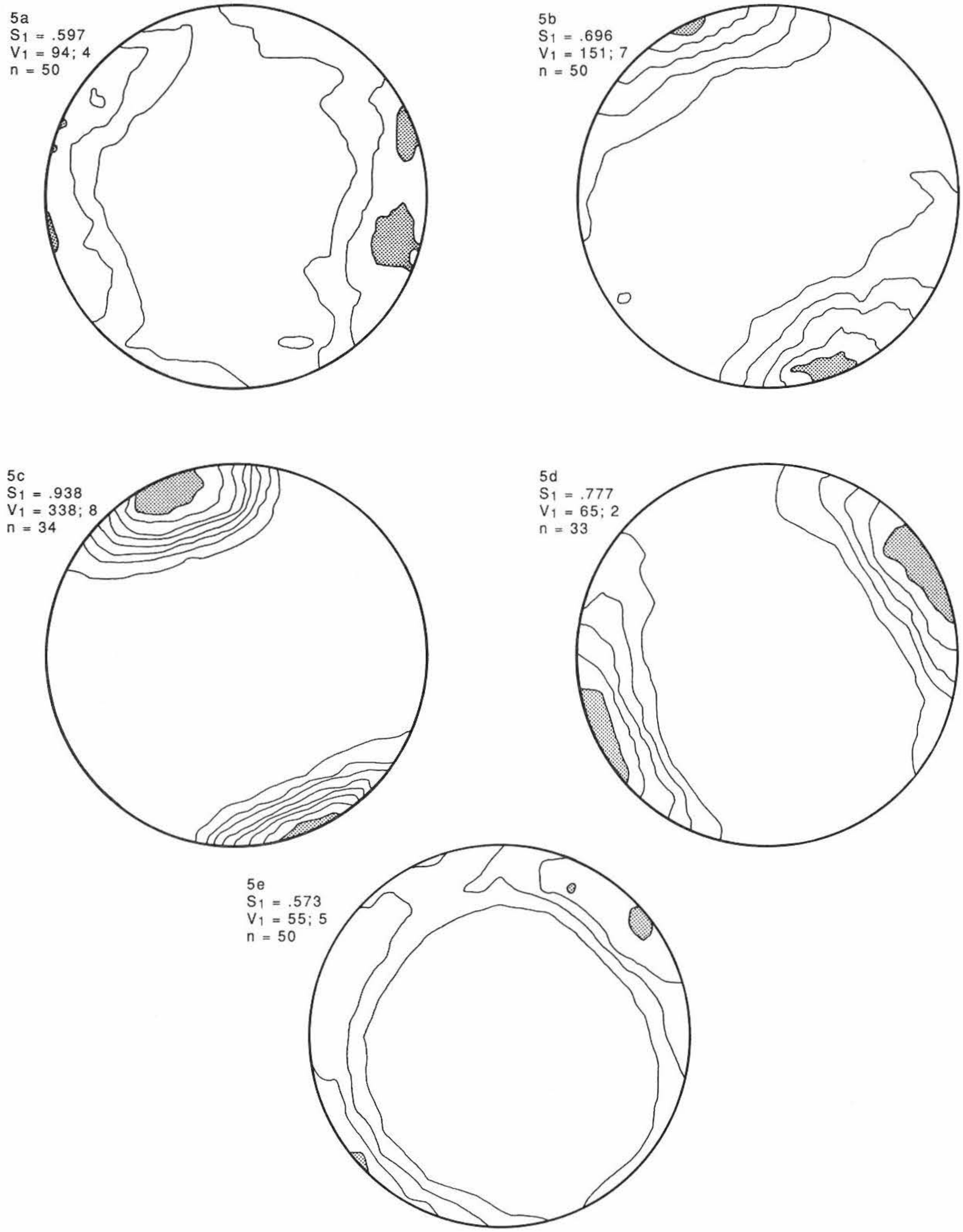


Figure 5. Petrofabrics of diamictons from Site A at New Sharon (5a-c); fabrics 5d and 5e are from similar units on opposite bank of river from Site A. S_1 = significance of fabric strength; V_1 = trend and plunge of azimuthal vector; n = number of measurements. Contour interval = 2 sigma.

axial trends cluster in a tight grouping with a preferential NW plunge, and the fabric has an S_1 value of 0.94.

Fabrics 5d and 5e are from stratified diamictons which occur on the opposite bank from site A. These fabrics are also polymodally distributed, with NE-SW trends, and the fabrics of clasts in 5d and 5e have a preferential NE plunge. The S_1 values for 5d and 5e are 0.78 and 0.57, respectively.

Fine-Grained Facies

The fine-grained facies (Figs. 3, 4, Table 1) consist of laminated clay, silt, and fine-grained sand, as well as more massive silt and clay layers, up to 3 cm thick. The colors of the units in this facies vary from gray to dark gray (5Y 5/1 and 5Y 4/1). There is no noticeable gradational color change between or within the fine-grained units or the stratified diamicton units. Diamicton layers less than 1 cm thick are commonly interbedded with the fine-grained units. Thin beds and laminae of diamicton often occur as the basal portion of thinly-bedded, fining-upward rhythmic sequences of diamicton, fine-grained sand, silt, and clay. Outsized clasts, and non-lithified clasts of silt and fine-grained sand (0.1 - 0.5 cm) also occur in this facies.

The rhythmic nature of the sediments at site A has been noted by other workers (Caldwell, 1959), and the sediments have been described as varves. The repetition of diamicton, silt, and clay in these sediments suggests some cyclic deposition of these units. An embedded Markov chain analysis was performed, testing for randomness in the sequence of sediments, utilizing the quasi-independence model of Powers and Easterling (1982). A chi-square (X^2) test of significance of the quasi-independence transitional frequency of facies results in X^2 of 18.09. With 19 degrees of freedom, the probability of any significant periodicity in the sequence is less than 50% (Table 2).

Harms et al. (1982) believe that analyzing for Markovian processes works best in systems which have several facies and many repeated sequences. Due to the small exposure of this section and lack of many repetitive sequences, the use of this test may not be applicable to the units described here.

Interpretation

The sediments exposed at this section at New Sharon are interpreted primarily as the products of sediment-flow deposition in a subaqueous environment, and to a lesser degree by basal processes. The fine-grained units are interpreted as the result of sediment flows (debris flows, turbidity flows), and rain-out following a sediment flow. Ashley et al. (1985) have distinguished between slump-generated surge rhythmites and annual rhythmites. Following their criteria to distinguish between annual- and surge-rhythmites, the fine-grained sediments at New Sharon are graded, they do not occur as discrete clay and silt beds forming couplets (unlike varves), and the clay beds in the rhythmites are not of consistent thickness. These qualities in context with the Markov chain analysis suggest the rhythmic nature of the sediments results from randomly occurring events, and that the sediments are not annual deposits.

Horizontal layering of several meters in lateral extent, outsized clasts within the fine-grained units, draped beds over the upper surface of some interbedded thick diamicton units, normally-graded thin beds of medium- to fine-grained sand, and very rare isolated cross-bedding in fine-grained sand layers within the unit suggest a subaqueous depositional environment for the fine-grained facies.

Stratigraphic sequences found along Lake Erie, Ontario, similar in description to the New Sharon deposits, have been interpreted as the result of basal melt-out beneath the floating terminus of the ice sheet (Driemanis, 1979, 1982; Gibbard, 1980), and by other complex processes (Driemanis et al., 1987). Ashley et al. (1985) also have described how stratified sediments can form subglacially by basal melt-out processes. Several interpretations have been proposed for interbedded diamictons and rhythmic sediments near Toronto, Ontario (Eyles and Eyles, 1983; Sharpe and Barnett, 1985; Sharpe, 1987, 1988a,b). Other workers have presented studies on how different glacial deposits can be characterized according to properties developed during their deposition, enabling the worker to distinguish between different modes of deposition associated with glaciation (for example, Shaw, 1977, 1982, 1983; Boulton, 1979; Lawson, 1979a,b, 1981a,b, 1982; Boulton and Deynoux, 1981; Haldorsen and Shaw, 1982; Eyles et al., 1983; Eyles and Miall, 1984; Gravenor et al., 1984; Ashley et al., 1985; Goldthwait and Matsch, 1988). A common theme in the papers cited focuses on the complex nature of the glacial environment, and that specific distinguishing criteria must be evaluated with the general characteristics of the sediments before interpretations of origins of glacial deposits can be made. Studies by Boulton (1971), Mills (1977; 1984), Lawson (1981a,b), Haldorsen and Shaw (1982), Shaw (1982), and Gravenor et al. (1984) are relevant to the glacigenic sediments at New Sharon. These authors point out that pebble fabrics in lodgement or melt-out tills may have a preferential alignment and may show an imbrication that dips up-glacier. Melt-out till may show transverse clast orientation and the up-glacier dip of clasts may not be as steep as in lodgement till. Clast fabrics in sediment-flow deposits generally

TABLE 2. EMBEDDED MARKOV CHAIN ANALYSIS.

	Quasi-Independence Transition Frequency						
Facies	1	2	3	4	5	6	Σ
1) Dmm	---	.45	.13	.70	.28	.45	2.01
2) Dms	.45	---	.21	1.14	.45	.73	2.98
3) Dms(r)	.12	.21	---	.32	.13	.21	0.99
4) Dmg	.69	1.13	.31	---	.69	1.13	3.95
5) Fmd	.28	.45	.13	.70	---	.45	1.59
6) Fld	.45	.73	.21	1.13	.45	---	2.97
Σ	1.99	2.97	0.99	3.99	1.58	2.97	

degrees of freedom = 20
$X^2 = 18.09$
$P > 0.5$

show a poorly-defined preferred orientation, or none at all. Pebbles in flows may show a polymodal distribution, and principal-axis orientations could lie either parallel or transverse to the direction of sediment flows.

Furthermore, Lawson (1979b) has shown from studies at the Matanuska Glacier in Alaska that the strength of clustering about the mean axis of pebble orientations varies considerably between melt-out till and sediment flows. The S_1 values from clasts in basal zone ice range from 0.77 to 0.99, with a mean of 0.86. For melt-out till, S_1 values range from 0.75 to 0.89, with a mean of 0.82, and for sediment-flow and ice-slope colluvium deposits, values range from 0.49 to 0.70, with a mean of 0.57. The close values for the strength of basal zone ice and melt-out till fabrics suggests that melt-out till fabric is generally preserved during the melt-out process.

For Pleistocene till in Edmonton, Canada, Shaw (1982) shows that the primary clast fabric for relatively undisturbed melt-out till is largely preserved. While the grand distribution of the samples has a fabric strength of 0.81, the mean is 0.61 and individual distributions show preferred orientations at variance from ice-flow direction. Samples from within the main body of the unit which retained preferred orientations parallel to ice-flow direction show strength values ranging from 0.54 to 0.84, with a mean of 0.69. Samples from locations adjacent to or in disturbed or reworked zones show orientations inconsistent with ice-flow direction, with strength values ranging from 0.46 to 0.63, and a mean of 0.55. Also, Haldorsen and Shaw (1982) show that for Pleistocene till in Astadalen, Norway, melt-out or flow till S_1 values range from 0.53 to 0.66, with a mean of 0.58. For the Norwegian tills, the orientation of the fabrics from melt-out till shows preferred alignment parallel to last ice movement, whereas deposits of equivocal melt-out or flow till origin show preferred orientation oblique to regional ice movement.

Eyles et al. (1988) have discussed Late Pleistocene subaerial debris-flow deposits near Banff, Canada, which show shallow imbrication either up- or down-flow direction, and fabrics both transverse and parallel to flow direction. The strength of these fabrics range from 0.47 to 0.91, with a mean of 0.64. For a glacigenic diamicton sequence in British Columbia, Parker and Hicock (1988) have presented a range of fabric strengths from lodgement till (0.47 - 0.66, mean 0.57) and from glacigenic and subaquatic flow (0.43 - 0.73, mean 0.52). Montane glacial diamictons from Alberta have been studied by Levson and Rutter (1988) and have been subdivided into several categories. The strength of the fabrics of some of these units has been determined; for basal till (lodgement and melt-out) the range is 0.64 - 0.78, mean 0.69, for englacial melt-out till the range is 0.50 - 0.57, mean 0.53, and for subaerial flow tills the range is 0.47 - 0.50, mean 0.48.

Studies of glacial deposits in northeastern Illinois by Johnson et al. (1985), Hansel et al. (1987), and Hansel and Johnson (1987) provide information on fabric strength of late Wisconsinan till in that region. Fabrics from diamictons described as subglacial till range from 0.55 to 0.95, with a mean of 0.77. Of interest, however, is the slight difference in fabric strength of units interpreted as lodgement till compared to basal melt-out till. In these units, fabric strength of lodgement till varies from 0.65 to 0.94 (mean 0.80), whereas basal melt-out till varies from 0.55 to 0.95 (mean 0.81). Johnson et al. (1985) caution that a single criterion cannot be used to differentiate diamicton resulting from lodgement, melt-out, or sediment flow.

Dowdeswell et al. (1985) in comparing clast fabrics of ancient and modern glacigenic deposits suggest that lodgement till fabric is less strong than basal melt-out till fabric. These data from lodgement till at the front of Icelandic glaciers have a mean fabric strength of 0.69. However, Boulton and Deynoux (1981) and Mills (1984) suggest that scatter in basal tills is a sign of post-depositional deformation. Dowdeswell and Sharp (1986) indicate that fabric data from modern glacial sedimentary environments can be useful in interpreting the depositional environment of Quaternary glacial deposits. They note that there is a general reduction in fabric strength and increase in clast dip associated with the transition from melt-out tills, through undeformed and deformed lodgement tills, to sediment flows. Rappol (1985), however, suggests that clast fabrics from Pleistocene sediments in Allgau, Germany, do not clearly distinguish between subglacial till and debris flow deposits. Finally, May et al. (1980) have reviewed the need for caution in the interpretation of till fabrics, and according to Dowdeswell et al., (1985), if field related criteria are not considered, fabric analyses have limitations due to ambiguity of interpretation.

The diamicton petrofabrics from Site A, in general, are interpreted as orientations unrelated to basal deposition. Three fabrics from the glacigenic sediments (Figs. 5a, 5d, and 5e) have orientations transverse to the last regional ice-flow direction, which is indicated by NW-SE trending striations and streamlined forms in the study area (Caldwell, 1959, 1986; Thompson and Borns, 1985b; Weddle, 1987). One fabric (Fig. 5b) crudely trends along the regional ice-flow direction, but the contour pattern lacks a strong single maximum and clast plunge is preferentially down-ice, contrary to what might be expected for a lodgement or basal melt-out till. Fabric 5d has a moderately strong fabric with moderate clast scatter, but is oriented transverse to ice-flow direction. This orientation could be a result of compressive flow. Alternatively, the transverse fabrics could represent debris flow material deposited into the lake by fans along the valley sides. The S_1 values of these fabrics (5a,b,d,e) are less than uniformity and vary from 0.57 to 0.78, with a mean value of 0.66, somewhat lower than the means of basal tills presented earlier. While these values are within the lower range of basal till, in particular fabrics 5b and 5d, and may represent post-depositional deformation of basal till, the diamictons must be evaluated within the sedimentary and stratigraphic context in which they occur, and these considerations suggest that fabrics 5a and 5e are not reflective of lodgement or basal melt-out processes. While fabrics 5b and 5d are more representative of subglacial processes, their stratigraphic positions suggest alternative modes of origin to basal deposition.

However, the diamicton at the base of the section at Site A (Fig. 5c) has a fabric that reflects basal processes. The lower contact of this unit is not exposed and those contact relations are not known, but the unit contains striated, bullet-shaped clasts, the axes of which have NW-dipping imbrication, and plot as a single maximum with a measure of fabric strength at 0.94.

The general dominance of polymodal orientation and moderate to large scatter rather than a single maximum of the long axes of clasts in diamictons, occurrence of outsized clasts and draping of layers over the clasts, generally sharp and conformable contacts between units, flow folds and drag folds, rip-up clasts, graded beds, and silt and sand clots in laminated beds suggest the sediments at Site A are primarily the result of subaqueous deposition into a proglacial lake, interrupted occassionally by ice marginal fluctuations when basal till was deposited. An environment similar to the model proposed by Evenson et al. (1977) for the lower Catfish Creek till at Plum Point, Ontario, is envisioned as the environment of deposition of the glacigenic sediments at Site A at New Sharon.

SITE B

Organic-Bearing Unit (New Sharon Beds)

The organic-bearing unit is located at site B, approximately 30 m downstream from site A (Fig. 1). This unit is now covered by stream alluvium, but was re-excavated in 1985 to examine the relation between it and the glacigenic sediments it was reported to be associated with (Caldwell, 1959; Weddle, 1986). This excavation provided almost 3 m of vertical exposure and the stratigraphy varied from that originally described by Caldwell (1959).

The trench exposed 0.5 m of gray, layered sediment, similar to the sediments at site A, sharply overlying 1 m of intensely deformed organic-bearing silt (containing wood fragments) and interbedded silty pebbly sand, which abruptly overlies 1.5 m of massive sandy gravel (Fig. 6). The gravel, which contains fragments of peat, is cut by a 3-cm wide, subvertical, gray (2.5Y 5/0) west-dipping diamicton stringer. It has been proposed that the organic-bearing silt and the gravel be referred to informally as the New Sharon Beds (Weddle, 1988).

The sandy gravel immediately below the silt is unoxidized, but approximately 1 m below this contact the color of the gravel changes abruptly but gradationally downward from gray (2.5YR 6/0) to reddish yellow (5YR 6/8 - 7.5YR 6/8). The organic-bearing silt does not display any gradational color change from its surface downward, nor is there any obvious weathering or soil profile developed on its surface. The silt is comprised of thinly laminated, alternating layers of very dark to dark grayish brown (2.5Y 3/2 - 4/2), and dark gray to olive gray (5Y 4/1 - 4/2) fine sandy silt and silty pebbly sand. The silt layers are complexly deformed (Fig. 7); in places they are cut by east-over-west thrusts. However, the deformation in the New Sharon Beds does not extend upward into the overlying gray layered sediments.

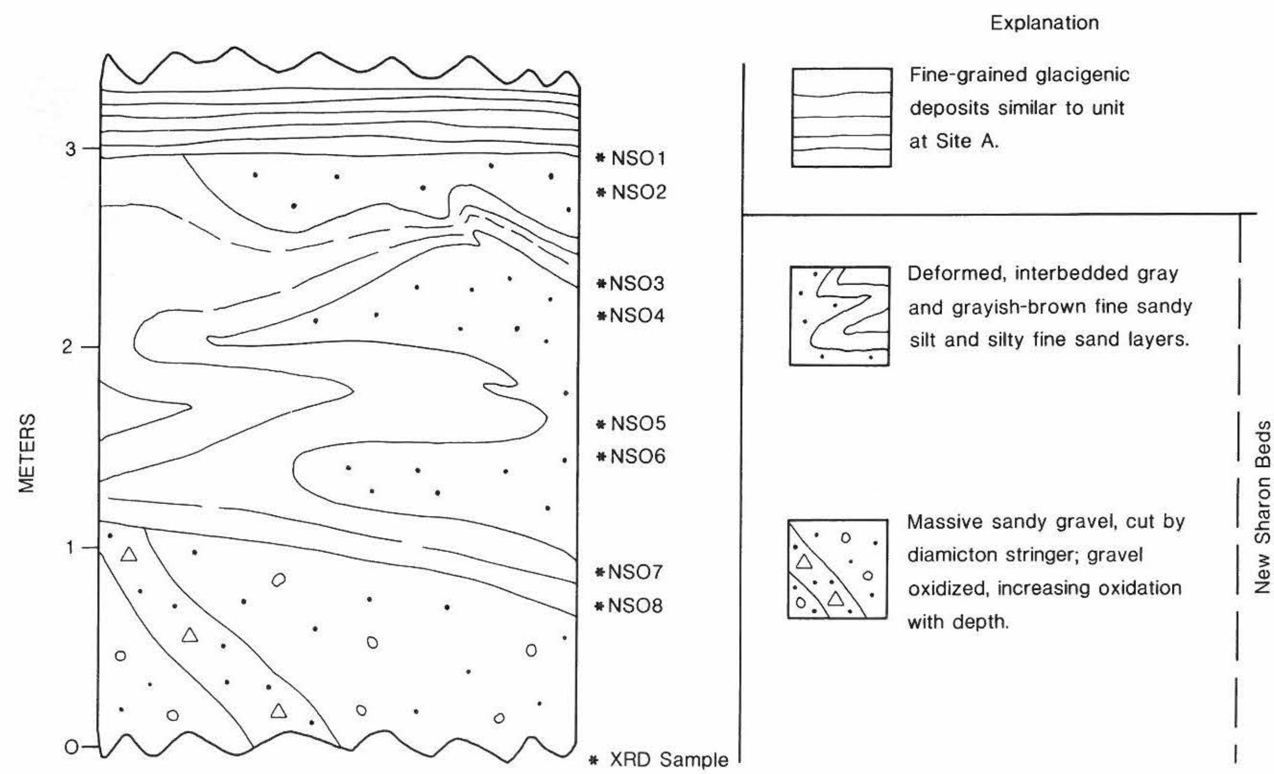


Figure 6. (a) Schematic diagram of the New Sharon Beds, excavated 24 July 1985, at New Sharon, Maine (Site B on Fig. 1).

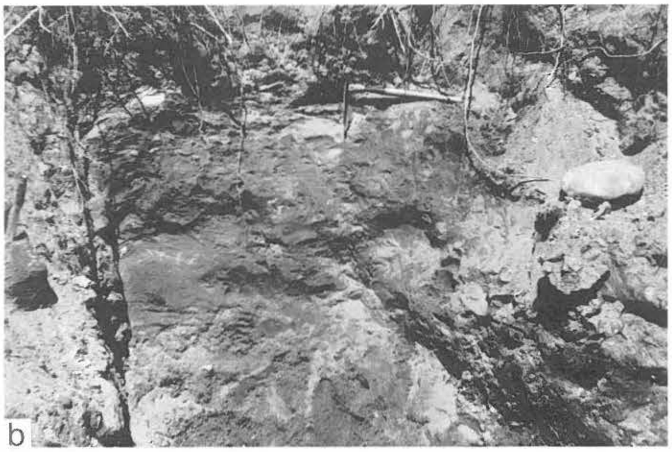


Figure 6 (continued). (b) Photograph of excavation (north face), entrenching tool approximately 0.5 m in length.

Figure 7. Close-up of deformed sand layers in the New Sharon Beds at New Sharon, Maine (north face).

Other than as the thin stringer which cuts the gravel of the New Sharon Beds, diamicton was not encountered in the excavation. However, trenching had to be ceased at 3 m depth due to ground water seepage into the pit. Single-channel seismic refraction lines run parallel to the trench indicate velocities of 1800-2250 m/s at depths greater than 3 m, which suggests that a very compact non-lithified unit underlies the gravel. Bedrock velocities (greater than 4500 m/s) occur at a depth of approximately 24 m (Weddle and Caldwell, 1986; Weddle, 1986).

Eight samples were collected from the silty unit, four from gray, and four from grayish-brown layers. These were analyzed by X-ray diffraction (XRD) for clay-mineral identification and evidence of alteration following the methods described by Jackson (1956) and Carroll (1970). Samples were treated with hydrogen peroxide to eliminate organic material. Oriented mounts of the less than 1 micron fraction of the clay percentage of the samples were prepared by centrifuging the clay fraction, mounting the material on a glass slide by vacuum suction, and allowing the slide to air dry. Slides were scanned on a Phillips x-ray diffractometer, from 32 to 2 degrees 2 theta at 2°/min, at optimum MA, KV, and range. Samples were run air-dried, treated with ethylene glycol, K-saturated, and heated to test for mineral alteration.

Two sample analyses are shown in Figure 8, one from each color variation of the silt. Fe-chlorite reflections are represented by the 14, 7, 4.7, and 3.5 angstrom peaks; illite/mica reflections are represented by the 10, 5, and 3.3 angstrom peaks. The failure of any noticeable expansion of the 14 angstrom peak after ethylene glycolation, or any significant collapse of peaks after K-saturation or low heat treatment in any of the eight samples indicates that altered chlorite and illite are not present in the samples.

Identification of pollen grains and macrofossil fragments from the New Sharon Beds was done by Dr. Robert Nelson, Colby College. The silt contained wood fragments and abundant alder, pine, and spruce pollen, and very little hardwood pollen (R. Nelson, pers. commun., 1985). The analyses are consistent with earlier analyses indicating a climate cooler than present, more representative of an interstadial deposit than an interglacial deposit, although probably harsher conditions than the closed spruce-pine forest indicated by the earlier pollen analyses (Caldwell, 1960; Caldwell and Weddle, 1983; Appendix A, this report).

During the fall of 1987, the Water Resources Division of the U.S. Geological Survey office in Augusta drilled a test boring at New Sharon as part of its statewide ground water monitoring program. The boring location is shown on Figure 1. Surface elevation at the test boring site is approximately 94 m above sea level, and the surface elevation of the New Sharon Beds (Site B) is approximately 89 m above sea level. The boring penetrated almost 12 m of glacigenic sediments similar in appearance to the deposits at Site A. The drilling was conducted using a hollow-stem auger, and samples were collected by split-spoon sampling; the complete log of the boring is given in Appendix B. The boring did not penetrate the New Sharon Beds, but at the depth equivalent to their surface elevation, glacigenic material similar to that at Site A was recovered. The boring was terminated when a boulder was encountered which could not be penetrated. The seismic data previously mentioned confirms that boring refusal was not at bedrock, but rather at a boulder.

DISCUSSION

The numerous references by others (Caldwell, 1959, 1960; Borns and Calkin, 1977; Mickelson et al., 1983; Oldale and Eskenasy, 1983; Hanson, 1984; Koteff and Pessl, 1985; Stone and Borns, 1986) to the weathered appearance of the New Sharon Beds is not substantiated by the 1985 excavation and subsequent XRD analyses. This is significant because some of these workers proposed regional correlations based in part upon the reported presence of a buried soil at New Sharon. No visible

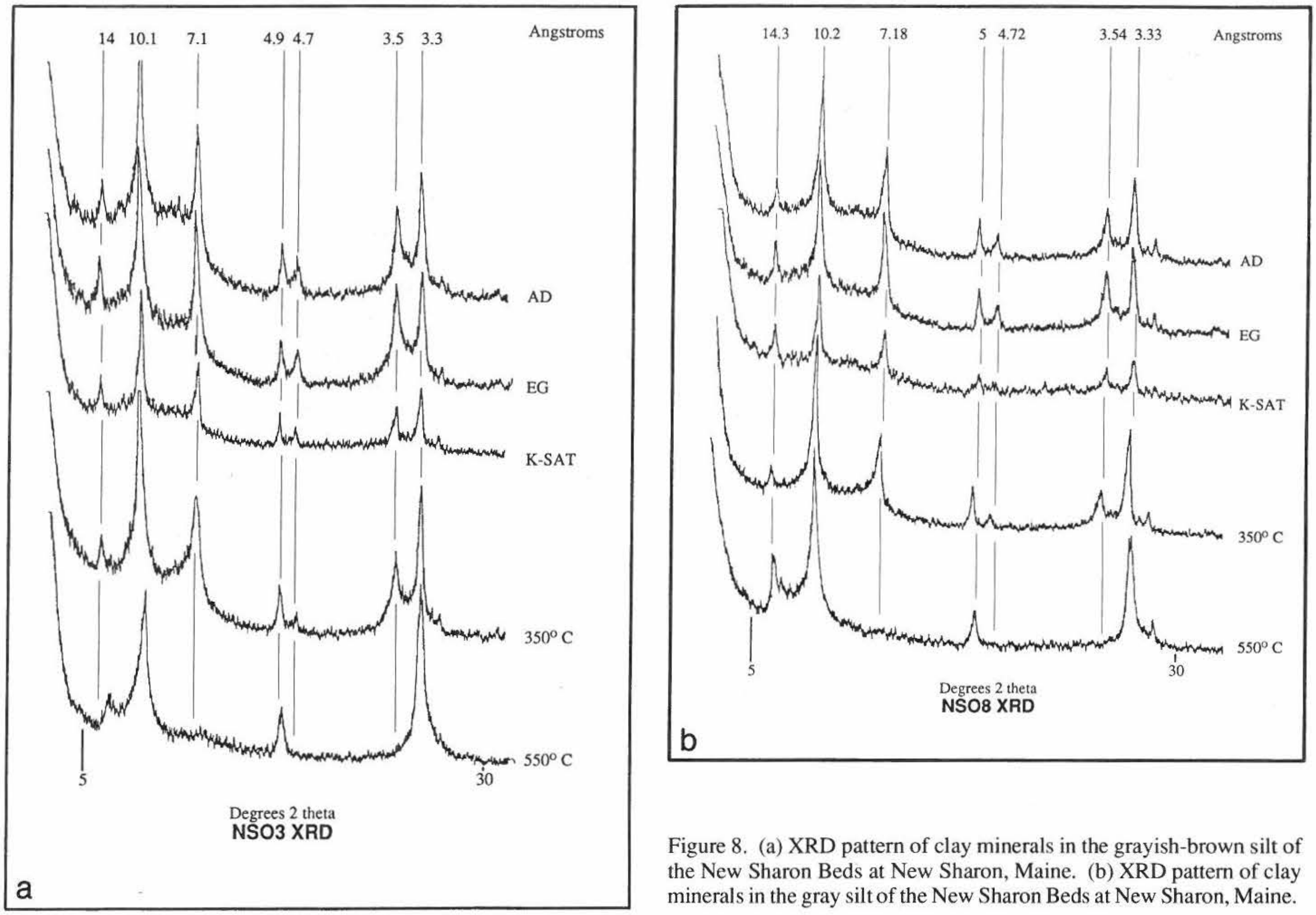


Figure 8. (a) XRD pattern of clay minerals in the grayish-brown silt of the New Sharon Beds at New Sharon, Maine. (b) XRD pattern of clay minerals in the gray silt of the New Sharon Beds at New Sharon, Maine.

soil profile or surface weathering horizon is present in the silt or the gravel, although the gravel has an oxidized appearance with depth, intensifying downward.

Koteff and Pessl (1985) describe a two-till section at Nash Stream, New Hampshire, comprised of upper and lower till units, separated by stratified drift. In places, the lower till at Nash Stream is overlain by lacustrine sediments, and the lacustrine sediments vary in color from pale yellow to very pale brown, with a pinkish cast. This suggests that some degree of oxidation of the lacustrine sediments may have occurred. Where the lower till is exposed at the surface, it is deeply oxidized. This oxidation of the lower till and lacustrine sediments is interpreted to be a function of a lengthy interglacial or interstadial weathering interval (Koteff and Pessl, 1985). The Nash Stream section is regarded by Koteff and Pessl (1985) as an appropriate reference section for the two-till stratigraphy of southern New England.

In southern New England, two-tills, an upper and lower till, have long been differentiated by several characteristics (Schafer and Hartshorn, 1965; Pessl and Schafer, 1968; Koteff and Pessl, 1985). One of these characteristics is the presence of an oxidized zone in the lower till which decreases in intensity with depth from the surface. Newton (1978, 1979) described the mineralogical characteristics of the oxidized zone of the lower till, and interpreted it as a paleosol, based primarily on differences in weathering of clay minerals from the upper till and the lower till. He also found that the clay minerals in the weathering profile are characterized by the alteration of chlorite and illite to vermiculite and mixed-layer illite-vermiculite.

Newman et al., (1987) conducted similar clay-mineral studies on tills from drumlins in Boston Harbor, Boston, Massachusetts. They described two distinct tills, upper and lower units, distinguished along with other criteria by a weathering horizon in the lower till, and degree of alteration of chlorite to vermiculite in the horizon. They interpreted the alteration to be due to soil development, and the more intensely weathered lower till to be of greater age than the less intensely weathered upper till.

The gravel unit from the New Sharon Beds may be outwash associated with an earlier glaciation, or an interstadial stream gravel. However, the oxidation associated with the gravel at New Sharon is unlike the oxidation of the lacustrine deposits or lower till at Nash Stream, or the oxidized zone on the lower till

of southern New England. Whereas the oxidation intensity decreases with depth in the lower till of southern New England, the oxidation intensity increases with depth in the gravel of the New Sharon Beds.

Assigning an age to the oxidation of the gravel, or suggesting an interglacial or interstadial weathering interval to account for the oxidation is unjustified because no studies have been done to evaluate the oxidation of the gravel from New Sharon. Furthermore, the XRD analyses of the silts from the New Sharon Beds indicate that there has been no soil development associated with that unit. In particular, chlorite and illite alteration to vermiculite and mixed-layer illite/vermiculite as noted by Newton (1978, 1979) in the southern New England lower till is lacking in the New Sharon Beds. The New Sharon Beds contain no record of ever being subjected to weathering comparable to that proposed for the lower till of southern New England.

CONCLUSION

The stratified waterlain glacigenic sediments at New Sharon are interpreted as the product of sedimentation into a proglacial lake. From New Sharon to its junction with the Kennebec River, the Sandy River is a northeastward-flowing tributary. If preglacial drainage in the Sandy River valley was similar to present-day drainage, southward advancing ice in the Kennebec River valley could have blocked drainage of the northeastward-flowing tributary, creating an ice-dammed proglacial lake in the Sandy River valley.

The glacigenic sediments at Sites A and B represent part of a sequence of proglacial materials of lacustrine and subglacial origin, deposited as debris flows and turbidity flows, and to a lesser extent by basal processes. The lack of significant weathering horizons or significant breaks in deposition between any of the units suggests the deposits are the product of a single glacial event, most likely of early Late Wisconsinan age, and constitute part of the regional Late Wisconsinan surface drift of the area. Initial observations by Caldwell (1959), and more recent studies (Caldwell and Weddle, 1983; Weddle, 1985, 1986; Weddle and Caldwell, 1984, 1986; Weddle and Retelle, 1988), indicate these glacigenic sediments were subsequently overridden by the main phase of Late Wisconsinan ice.

The relation between the New Sharon Beds and the glacigenic sediments is not completely clear. Glacigenic sediments occur stratigraphically above the New Sharon Beds. Pollen from the New Sharon Beds contains specimens from aquatic plants (Caldwell and Weddle, 1983; R. Nelson, pers. commun., 1985), and the thinly bedded deformed silt layers of that unit may be lacustrine. Alternatively, although a thorough facies analysis was not conducted on the New Sharon Beds, the stratigraphy exposed in the trench may represent glacial outwash or stream alluvium with overlying flood-plain deposits.

A further complication is that the New Sharon Beds appear to be more complexly deformed than the overlying glacigenic sediments. This deformation could be accounted for by emplacing the New Sharon Beds in the stratigraphy by ice shove (glacial rafting), possibly in a manner similar to that described by Ruszczynska-Szenajch (1987), who has subdivided glacial rafts into three genetic groups. These include (1) glaciotectonic rafts that have been detached and transported by mechanical action of the glacier, without being incorporated within the ice body; these may be subdivided into (a) squeezed-out or pressed-out rafts, (b) dragged rafts, and (c) pushed rafts; (2) glacio-erosional, or glacio-dynamic rafts, that have have been detached by freezing onto a glacier and then transported within the ice body; and (3) rafts of composite origin formed by these processes. Without detailed study of the New Sharon Beds, glaciotectonic origin cannot be determined with precision, but must be considered.

The east-over-west thrusts in the New Sharon Beds are not in accord with regional Late Wisconsinan ice-flow indicators shown by Caldwell (1959, 1986), Thompson and Borns (1985b), and Weddle (1987). However, they may be attributed to early Late Wisconsinan ice initially controlled by topography during ice advance, and deforming proglacial deposits as it advanced. Deformation of proglacial deposits has been described by Kalin (1971), Gripp (1979), Rabassa et al. (1979), Humlum (1985), Kruger (1985), Van Der Meer et al. (1985), Boulton (1986), Drozdowski (1987), Eybergen (1987), Van Der Wateren (1987), Van Gijssel (1987), and Croot (1988) in discussing the formation of ice-push moraines in modern and ancient deposits, and by Smith (1988) on the effects of a rapidly advancing glacier on sedimentation in a proglacial lake. Sharp (1985) mentions glacitectonically deformed sediments in Iceland in which a thrust found in the core of a push-moraine originates at the contact between cohesive silt and peat and underlying sand and gravel. A more detailed discussion of glacitectonic features and structural geology of the New Sharon sections can be found elsewhere (Weddle, in prep.).

The lack of similar organic-bearing material at any of the other sections along the Sandy River and the deformation associated with the New Sharon Beds suggests that their presence at Site B may be a local phenomenon and that they are allochthonous. It may be that the lowest diamicton at Site A is correlative with the the diamicton under the New Sharon Beds reported by Caldwell (1959). However, he described a weathered appearance to the surface of that unit, and there is no indication of weathering on the surface of the lowest diamicton at Site A. Furthermore, the petrofabric of the lowest unit at Site A corresponds to the regional trend of the Late Wisconsinan ice. Caldwell (1959), however, did note the presence of lineations on the bedding surface of fine-grained deposits above the New Sharon Beds but at the base of the overlying till. He attributed the lineations to overriding ice because they are approximately parallel with regional striations. The New Sharon Beds could then represent a transported block of older material within a sequence of stacked Late Wisconsinan deposits.

This Late Wisconsinan age assignment to the sequence of exposed deposits at New Sharon is contrary to the Quaternary

stratigraphy for Maine as suggested by Borns and Calkin (1977), and Stone and Borns (1986). In particular, there is no direct evidence for Early or Middle Wisconsinan drift deposits at New Sharon as described by Borns and Calkin (1977), and by Stone and Borns (1986), and as alluded to by Caldwell and Pratt (1983), Caldwell and Weddle (1983), Weddle (1985), and Thompson and Borns (1985a).

Thompson and Borns (1985a) suggest that the wood found at New Sharon was killed by ice, and that this constrains the age of the overlying sediments to the age of the wood. The author has not found wood in growth position, nor was the wood found by Caldwell (1959), or Borns and Calkin (1977), reported to be in growth position. However, only material found in growth position should be considered to date the enclosing sediment (Vogel, 1980). If it is not collected in growth position, its age should never be equated with the enclosing sediment (LaSalle and David, 1988). The wood in the New Sharon Beds is not in growth position, and it is likely that the New Sharon Beds are allochthonous.

Finally, it appears that there is a subsurface stratigraphy at New Sharon of at least 24 m thickness. Knowing the stratigraphy beneath the New Sharon Beds only by its seismic velocity is not particularly useful. However, the U.S. Geological Survey test boring showed that the stratigraphy at Site A extends to a depth of at least 6 m below the surface exposure of the New Sharon Beds. These units are clearly derived from a single glacial event, one in which the New Sharon Beds may have been incorporated by glacial rafting during the advancing phase of the Late Wisconsinan ice.

Determining the subsurface stratigraphy through more detailed test borings might resolve whether the New Sharon Beds are truly an allochthonous or autochthonous deposit. The occurrence and lithology of any older, underlying units could also be determined. Re-evaluation of the earlier interpretations of the stratigraphy is warranted. No older, underlying till is currently exposed at New Sharon, although it may occur in the subsurface. The eroded bluffs of glaciolacustrine sediments and diamicton expose deposits of Late Wisconsinan age.

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APPENDIX A. RESULTS OF POLLEN ANALYSES FROM THE NEW SHARON BEDS BY PERCENTAGE.

Sample	NS-4*	NS-12*	PL 60-17**	New Sharon 1[+]	New Sharon 2[+]
TREES AND SHRUBS					
Abies	5.7	3.3	3.5	-	trace
Pinus	27.4	33.1	37.9	16.8	35.5
Picea	61.8	58.0	44.4	4.4	17.5
Larix	-	trace	-	-	-
Tsuga	-	trace	-	-	-
Betula	trace	trace	trace	1.7	4.1
Alnus	3.8	5.5	13.2	50.8	29.8
Salix	-	-	-	1.0	trace
Acer negundo	-	-	-	trace	trace
Corylus	-	-	-	2.7	trace
Ostrya	-	-	-	-	trace
Tilia	-	-	-	-	trace
Ulmus	-	-	-	-	trace
Quercus	trace	-	trace	-	-
Carya	trace	trace	-	-	-
Myrica	-	-	-	1.0	-
Cornus	-	-	-	1.4	-
Prunus	-	-	-	trace	-
Ericaceae	trace	-	trace	1.0	trace
NONARBOREAL POLLEN					
Poaceae	trace	1.6	1.0	7.7	3.4
Cyperceae	3.0	1.6	1.0	5.4	3.0
Potamogeton	-	-	-	trace	-
Cruciferae	-	-	-	trace	-
Thalictrum	trace	trace	trace	trace	trace
Umbelliferae	-	-	-	-	trace
Urtica	-	-	-	trace	-
Chenopodiinea	-	-	-	trace	trace
Monotropa	-	-	-	-	trace
Rosaceae	-	-	-	-	trace
Caryophyllaceae	-	-	-	trace	-
Artemisia	-	-	trace	trace	trace
Nemopanthus	-	-	trace	-	-
Nuphar	trace	-	-	-	-
Unknown	-	trace	-	4.8	1.5
SPORES					
Ferns	1.2	trace	11.5	8.7	4.1
Equisetum	trace	-	1.3	-	-
Lycopodium	1.3	1.0	-	trace	3.9
Sphagnum	1.3	trace	3.1	11.5	trace

* - D. W. Caldwell, pers. commun., 1985
** - Jan Terasmae, pers. commun. to D. W. Caldwell, 1960.
[+] - R. E. Nelson, pers. commun., 1985.
Trace is less than 1.0%; (-) indicates not observed.

These data are compiled from personal communications to the author or made available to the author with the permission of the individuals involved. In the communications, tentative interpretations have been presented and are included here in paraphrased format, or as partially quoted sections from the letters.

The communication from Terasmae to Caldwell (1960) has been quoted by others (Caldwell, 1960, Borns and Calkin, 1977). The interpretive section reads as follows: "..., hemlock is absent or rare. In that regard these assemblages are similar to those found in the Scarborough Beds at Toronto, and the St. Pierre Beds at St. Pierre, Quebec. All assemblages indicate cold climate for the area concerned and hence, should be termed interstadial unless assemblages indicating much warmer climate are also found in the same sediment sequence. It is interesting to postulate a correlation between the New Sharon interval (New Sharon Beds of this paper) and the St. Pierre Beds, now dated at about 65,000 years B.P."

The communication from Nelson to Weddle (1985) is paraphrased as follows: "...both samples are dominated by alder, pine, and spruce, although the overwhelming dominance of spruce reported by Caldwell and Weddle (1983) is not present. Pine and alder account for about 75% of the total pollen in each sample, with spruce being nowhere near as major a component as it was in the earlier counts. Both samples are indicative of at least partly open country, possibly parkland but more likely open forest. The eastern alders are all shrubs and prefer moist to wet substrates. Neither they nor the pines are shade tolerant. The fact that the alders are so important in both counts indicates that at least locally there were nonforested conditions, though possibly just a local boggy area on a small stream floodplain. The lack of major amounts of grass and sedge pollen is important, since they would normally dominate floodplain environments; in this case, the alders were apparently dense enough to suppress most herbaceous undergrowth. The composites, though minor elements in the pollen flora, are small pollen producers but are also most commonly found growing on sunny and well-drained sites. The dogwood pollen could well be derived from *Cornus canadensis*, a boreal shrublet found far north of here today. The few scattered grains of *Corylus* and *Acer negundo* may merely be indicative of the presence of rare in-

dividuals at the limits of their distribution. The individual grains of *Tilia* and *Ulmus* could be due to either long-distance transport or possibly reworking; either way, single grains are not considered significant.

In sum, the assemblage would probably be of an environment perhaps comparable to uplands of modern northern Maine or adjacent Canada, beyond the elevational or latitudinal limits of most hardwoods such as maple and oak. Labrador or Newfoundland might be reasonable approximations. The conditions indicated are definitely interstadial rather than interglacial, although harsher (probably colder) than the closed spruce-pine forest indicated by the counts of Terasmae and Caldwell that were reported in Caldwell and Weddle (1983). The nonforested areas seem to have been dominated by alder thickets, otherwise there would be more grass and sedge pollen as well as other herbaceous types. It is not clear why pine is dominant over spruce in these counts; possibly it may represent an environment drier than present. Eastern white pine tends to favor well-drained substrates, such as outwash terraces or delta surfaces.

The plant macrofossil record is represented by a single seed of pondweed (Potamogeton), a single carbonized base of a spruce needle (either *Picea mariana* or *P. glauca*), a single sedge seed (*Carex*), and a very abraded small spruce cone scale (either *Picea mariana* or *P. glauca*). These are the species that form treeline in the north today and currently reach their southern limits in southern New England."

Dr. Richard Jagels (Forest Biology, University of Maine) examined wood fragments extracted from the New Sharon Beds in 1985 and found that they were too dessicated to be identifiable.

APPENDIX B. FIELD BORING LOG FOR 1987 USGS TEST BORING AT NEW SHARON, MAINE*.

Depth (m)	Sample**	Blows/15 cm	Description and comments
0 - 3.9 m			Cuttings; brown to tan, silty, rounded medium- to coarse-grained sand; drilling rate change at 2.1 meters (m); stream alluvium.
			Driller notes boulders; cuttings, tan to light gray, silty, rounded fine- to medium- grained sand, drilling rate change at 3.9 m; stream alluvium.
3.9 - 11.4 m	SS1	4,6,6,23	Sample interval 4.5 - 5.1 m; compact, gray, fine, sandy, slightly clayey silt, few subangular pebbles; diamicton.
	SS2	4,4,4,6	Sample interval 5.4 - 6.0 m; same as SS1, gray diamicton. (5.0 m depth equivalent to surface elevation of New Sharon Beds at Site B; no change in drilling rate between SS1 and SS3).
	SS3	4,4,5,8	Sample interval 6.9 - 7.5 m; same as SS2, dark gray diamicton.
	SS4	5,5,5,6	Sample interval 8.5 - 9.1 m; Interbedded gray, fine-grained sand, silt, and clay (0 - 0.3 m); compact gray diamicton, same as SS3 (0.3 - 0.6 m).
	SS5	5,6,14,35/4	Sample interval 10.0 - 10.6 m; same as lower part of SS4, compact gray diamicton.
	SS6	4,7,100/2	Sample interval 11.2 - 11.4 m; same as SS5, compact gray diamicton.
11.4 m			Boring terminated; driller notes refusal on boulder.

* - Surface elevation approximately 94 meters above sea level.
** - Samples collected by split-spoon sampler (SS#) advanced 60 cm by the weight of a 63.5 kg hammer dropped 76 cm; number of blows recorded every 15 cm interval.

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Deglaciation of the Upper Androscoggin River Valley and Northeastern White Mountains, Maine and New Hampshire

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ABSTRACT

The mode of deglaciation of the White Mountains of northern New Hampshire and adjacent Maine has been a controversial topic since the late 1800's. Recent workers have generally favored regional stagnation and downwastage as the principal means by which the late Wisconsinan ice sheet disappeared from this area. However, the results of the present investigation show that active ice persisted in the upper Androscoggin River valley during late-glacial time. An ice stream flowed eastward along the narrow part of the Androscoggin Valley between the Carter and Mahoosuc Ranges, and deposited a cluster of end-moraine ridges in the vicinity of the Maine-New Hampshire border. We have named these deposits the "Androscoggin Moraine." This moraine system includes several ridges originally described by G. H. Stone in 1880, as well as other moraine segments discovered during our field work. The ridges are bouldery, sharp-crested, and up to 30 m high. They are composed of glacial diamictons, including flowtills, with interbedded silt, sand, and gravel. Stone counts show that most of the rock debris comprising the Androscoggin Moraine was derived locally, although differences in provenance may exist between moraine segments on opposite sides of the valley.

Meltwater channels and deposits of ice-contact stratified drift indicate that the margin of the last ice sheet receded northwestward. Early penetration of mountain passes by esker-forming glacial streams was accompanied and followed by cutting of meltwater channels and deposition of ice-contact fluvial and lacustrine sediments as the uplands were uncovered. Lakes were ponded between the ice margin and north- to west-facing slopes. During the late stage of deglaciation, larger volumes of meltwater deposits accumulated on valley floors.

Most of the features formed by glacial meltwater in the study area could have developed in association with stagnant ice. However, the Androscoggin Moraine, together with moraines described by R. F. Gerath in the Berlin-Gorham area, supports the continued presence of active ice in the upper Androscoggin River basin when the Presidential, Carter, and Mahoosuc mountain ranges had emerged from the late Wisconsinan ice sheet. Glacial striation trends and previous studies in the White Mountains suggest that an ice lobe spilled into the Androscoggin River valley from the Connecticut River basin to the north and west. On the basis of the limited available radiocarbon data, the Androscoggin Moraine is believed to have been deposited by the Laurentide ice sheet at about 14,000 yr B.P.

INTRODUCTION

The manner in which the White Mountains were deglaciated has long been one of the most debated issues in the study of New England glacial geology. Much of the research in this region, especially in New Hampshire, has been directed toward the following topics: the configuration of the receding late Wisconsinan ice margin, the scale and relative importance of stagnation versus active ice flow in the marginal zone of the waning ice sheet, and whether an ice cap or alpine glaciers persisted in the mountains subsequent to their emergence from the continental ice. The theories concerning these problems have been summarized by Goldthwait (1916), Goldthwait and Mickelson (1982), Gerath et al. (1985), and Waitt and Davis (1988). Gerath et al. (1985) pointed out that much detailed geologic mapping remains to be done to improve our understanding of White Mountain glacial history. In the northern part of the mountains, the only intensive mapping studies are those by Goldthwait (1940, 1970) in the Presidential Range and Gerath (1978) in the Berlin-Gorham area.

This report discusses the style of deglaciation in the northeastern White Mountains and adjacent western Maine. A review and synthesis of previous work are followed by the results of our investigations in a section of the Androscoggin River valley that straddles the Maine-New Hampshire border. The principal study area extends from Bethel, Maine, west and north to the origin of the Androscoggin River at the outlet of Umbagog Lake (Fig. 1).

Recent surficial geologic mapping by W. B. Thompson in the Maine portion of the Androscoggin River valley led to the discovery of a large cluster of glacial end moraines in the vicinity of the state line. Part of this moraine system had been briefly described by Stone (1880, 1899) during his work for the U.S. Geological Survey. Thompson (1983, 1986) concurred with Stone's identification of the ridges as moraines and assigned the name "Androscoggin Moraine" to this group of deposits.

The Androscoggin Moraine is significant for two reasons. First, it provides information about the style of deglaciation in the mountains along the Maine-New Hampshire border. The moraine may indicate persistence of active ice in the upper Androscoggin and Connecticut River basins when flow had ceased elsewhere in the mountains. Second, the Androscoggin Moraine appears to be the most clearly defined of the few moraines that have been described in the White Mountains and is the only known example of a cross-valley moraine. Problems discussed here include the age of the moraine system and whether it was deposited by the Laurentide ice sheet or by a northern Appalachian ice mass subsequent to marine incursion in the St. Lawrence Lowland. We also consider ice-flow directions indicated by glacial striations and streamlined hills, end moraines reported by previous workers, the sequence of meltwater deposits, and the relationships of these features to the regional history of deglaciation in the mountains of northern New Hampshire and western Maine.

PREVIOUS STUDIES IN THE UPPER ANDROSCOGGIN RIVER BASIN AND ADJACENT AREAS

Until recent years, little detailed work was done on the glacial geology of the upper Androscoggin River basin. Hitchcock (1878b) mapped glacial striations and "modified drift" (water-laid deposits) in New Hampshire, producing a series of rudimentary surficial geologic maps encompassing the whole state. During the late 1800's and early 1900's, most research centered on the possible existence of an ice cap or system of alpine valley glaciers in the higher parts of the White Mountains.

Packard (1867) envisioned an ice cap that "discharged local glaciers into the principal valleys radiating from the central peaks." He thought that one of these glaciers flowed northward down the Peabody River valley (Fig. 1) and joined the "Androscoggin glacier" at Gorham. From observations on striation trends, Vose (1868) expressed the same opinion concerning the ice flow direction in the Peabody Valley. Hitchcock (1878a) refuted this theory on the basis of stoss-and-lee erosional forms on bedrock, which indicated ice flowing southward up the Peabody Valley. Vose also tested the possibility of local glaciation in the Androscoggin River valley by measuring striation trends in the east-west section of the valley between Bethel and Gorham (Fig. 1). He found striations indicating eastward flow and concluded that a valley glacier had extended at least to Bethel. Vose distinguished the proposed valley glaciation from the "general operation" (presumably of continental ice) upon all of New England, but acknowledged that the relationship between these events was unknown.

Advocates of local glaciation in the White Mountains subsequent to emergence from the most recent continental ice sheet included Agassiz (1870a,b), Hitchcock (1878a), and Upham (1904). These geologists described the Bethlehem Moraine, and inferred that it was deposited by ice flowing from the Franconia Range northward into the Ammonoosuc River valley. This alleged cluster of moraine segments is located east of Littleton, New Hampshire, about 20-30 km west of the present study area. Upham (1904) claimed that the Bethlehem Moraine was formed adjacent to an ice cap, rather than the valley glacier suggested by Agassiz and Hitchcock. Goldthwait (1916) pointed out numerous flaws in the local-ice theories, and the vague, inconclusive nature of the evidence that had been advanced in support of this model. However, the possibility of a late-glacial residual ice cap on the higher peaks of the White Mountains was entertained as recently as the 1950's (Flint, 1951).

The Bethlehem Moraine was also at the center of early debate about whether deglaciation occurred by regional stagnation or the northward retreat of an active ice margin. Goldthwait (1938) concluded that widespread stagnation and downwastage was the chief mode of glacial retreat from the White Mountains. He relegated the Bethlehem Moraine to being just "a zone of massive kettled outwash." On the other hand, Lougee (1940) emphatically upheld the Bethlehem deposits as moraine ridges,

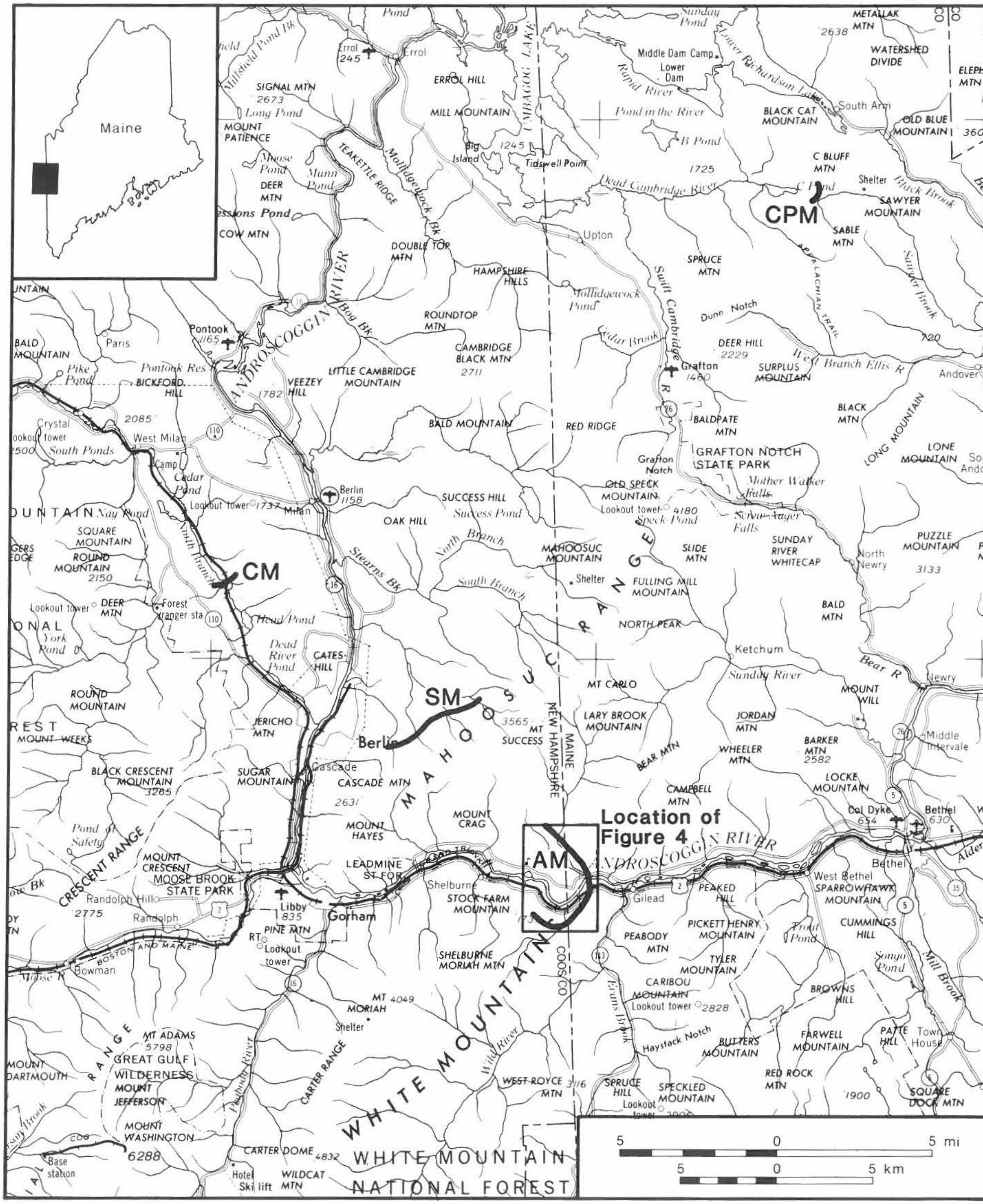

Figure 1. Map showing locations of principal geographic features and end moraines mentioned in text. AM: Androscoggin Moraine system (outer boundary). CM: Copperville Moraine. CPM: C Pond Moraine. SM: Success Moraine.

his chief evidence apparently being the morphology and bouldery surfaces of these deposits. Lougee agreed with Crosby (1934) that the moraine system was produced by a glacial readvance. The readvance theory was supported by exposures in the Connecticut River valley west of Littleton showing two tills separated by thrust-faulted sand and gravel.

The concept that widespread ice stagnation and downwastage began early in the deglaciation of the White Mountains and was the chief mode of ice retreat has generally prevailed in recent years. Goldthwait and Mickelson (1982) advocated this model on the strength of similarities between glacial features in the White Mountains and those of the Glacier Bay area in southeast Alaska. Much of the recent research in the White Mountains has centered on determining the time of development of Presidential Range cirque glaciers relative to overriding of the mountains by the late Wisconsinan continental ice sheet (Goldthwait, 1940, 1970; Bradley, 1981, 1982; Gerath and Fowler, 1982; Fowler, 1984; Waitt and Davis, 1988). The cirque controversy is not central to our discussion of the upper Androscoggin River valley region and thus will not be considered further.

On the Maine side of the border, north and northeast of the Androscoggin Moraine, Leavitt and Perkins (1935) briefly described glacial features in the Mahoosuc Range. They proposed that a high-level ice-contact lake drained eastward through Mahoosuc Notch (south of Mahoosuc Mountain, Fig. 1) as the glacier margin began to withdraw from these mountains. Leavitt and Perkins also named glacial Lake Cambridge, which they described as first draining south through the Grafton Notch spillway. This lake had several lower outlets as the ice-margin receded and finally emptied westward into the upper end of the Androscoggin River valley.

Caldwell (1974) reported on the results of reconnaissance surficial mapping in western Maine. He noted the presence of an end moraine (which he named the "C Pond Moraine") in a narrow pass through the northeastern Mahoosuc Range (Fig. 1). Caldwell described this moraine as being hummocky, strewn with large boulders, and incised by a 6 m-deep meltwater channel.

ICE FLOW DIRECTIONS IN THE UPPER ANDROSCOGGIN BASIN

Two types of ice flow indicators have been recorded in the study area: glacially streamlined hills, and striations and grooves on bedrock outcrops. Figure 2 shows the ice flow directions measured by earlier workers and the present authors. The striations and grooves usually indicate ice flow trends, rather than absolute directions. However, roche moutonnée and crag-and-tail forms demonstrate that the regional flow of glacial ice was toward the southeast or east. In most places this information enabled a reasonably certain flow direction to be assigned to the striations.

The streamlined hills are composed of till and bedrock in varying proportions. They show a wide range of sizes and include southeast-trending drumlinoid hills, crag-and-tails, and fluted till ridges. Gerath (1978) mapped large numbers of these features on the northwest side of the Mahoosuc Range in New Hampshire (Fig. 2). They also occur in adjacent western Maine, where drumlin axes trend approximately 130° (Caldwell, 1974). This belt of streamlined hills is a distinctive terrain and can be seen in high-altitude imagery (Fig. 3).

Goldthwait (1940) found that striations and grooves in the Presidential Range (southwest corner of Fig. 2) indicate an average ice flow direction of 140°. This is in accord with the general southeast trend of striations and streamlined hills throughout the present study area. However, there are significant exceptions to the regional trend, as described below.

The Androscoggin River valley abruptly narrows and curves to the east at Gorham, where it cuts across the northeast-trending mountain chain including the White Mountains and Mahoosuc Range (Fig. 1). It is evident from Figure 2 that glacial ice flow was locally channeled in conformance to this section of the valley. Numerous striations trending east to east-southeast have been recorded between Gorham and Bethel, commencing with the observations of Vose (1868) and Hitchcock (1878a). A prominent roche moutonnée indicating eastward ice flow can be seen adjacent to the Portland-Montreal pipeline, on the south flank of Hark Hill in Shelburne (Fig. 4). This eastward flow was topographically controlled and probably occurred as the late Wisconsinan ice sheet thinned over the mountains.

Farther up the Androscoggin River valley, between Gorham and Errol, glacial striations measured by the authors generally trend 135-155°. Thus they conform to the regional ice-flow pattern documented by Gerath (1978) and earlier workers. We had conjectured that striations in the upper Androscoggin basin might indicate a local south to southwestward flow of late-glacial ice stranded in the lowland between the mountains of northernmost New Hampshire and the Mahoosuc Range-Blue Mountains to the southeast. However, the only anomalous striations found during our reconnaissance are located on outcrops where Route 26 skirts the base of Errol Hill and Mill Mountain (Fig. 2). Here there are two outcrops that show east-trending striations in addition to the usual southeast-trending set. One site has a clearly older 103° set on the sheltered lee side of the outcrop, and a younger 135° set on the stoss surface. A similar relationship exists at the other site, where there are four striation sets that seem to indicate a progressive shift in ice-flow direction from 73° to 135°.

The eastward flow in the Errol area is thought to be unrelated to local topography, since it would have been easier for glacial ice to flow southeast along the Mollidgewock Brook valley than to cross Errol Hill and Mill Mountain. Perhaps this event was contemporaneous with the early and widespread eastward ice flow across northern Maine that was described by Lowell and Kite (1988).

Figure 2. Map showing glacial striation localities (with known or inferred ice-flow directions) and representative trends of glacially streamlined hills. Data on New Hampshire hills are from Gerath (1978); those on Maine hills are from Caldwell (1975b,c). Dots on arrows indicate sites where striations were measured; flagged arrows represent older flow directions. Sources of data: (1) Vose, 1868; (2) Hitchcock, 1878a,b; (3) Goldthwait, 1940; (4) Goldthwait et al., 1951; (5) Caldwell, 1975a,b,c; (6) Gerath, 1978; (7) measurements by the authors.

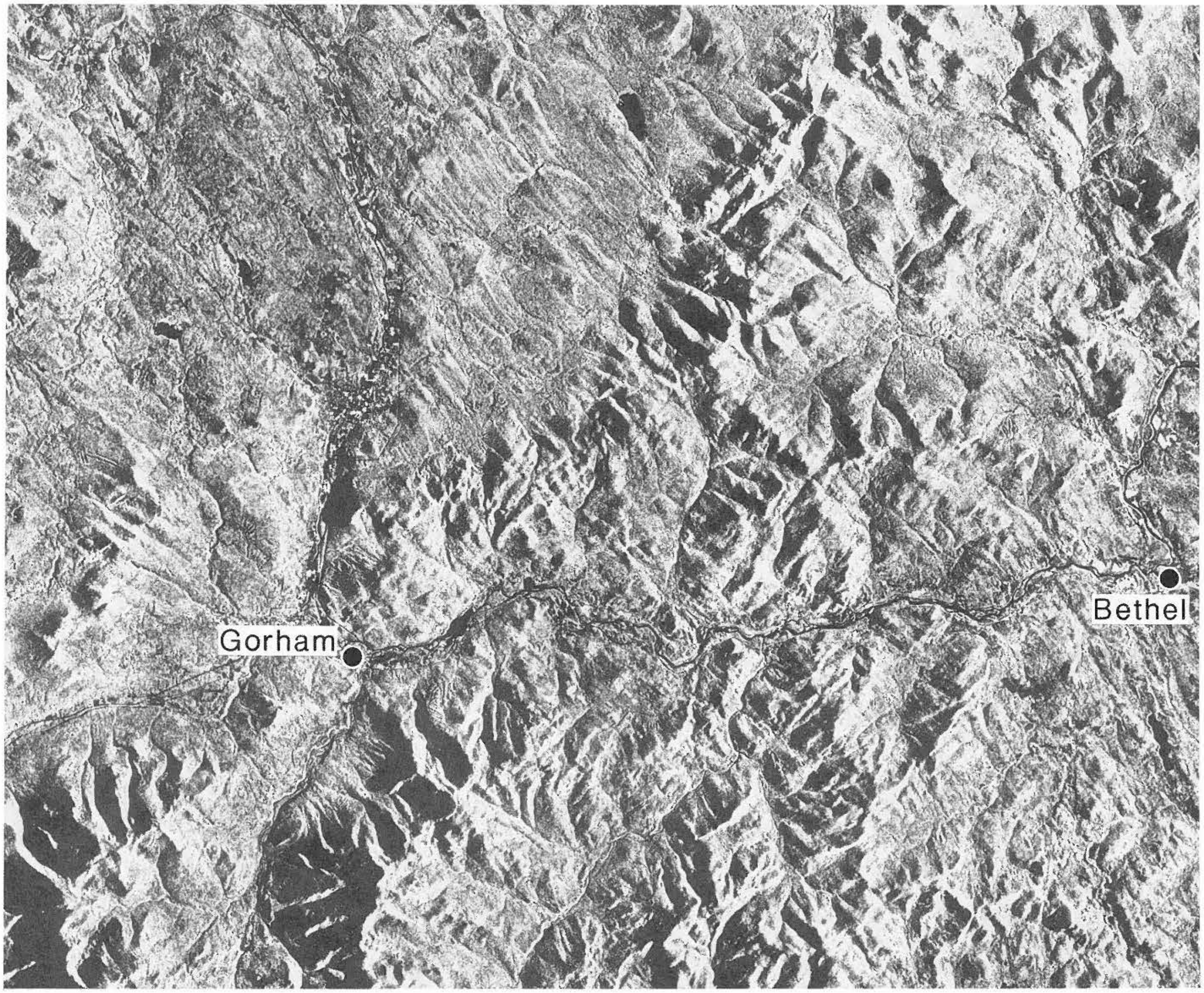

Figure 3. Side-looking airborne radar image of the study area. Note glacially streamlined hills on the proximal side of the Mahoosuc Range (northeast of Gorham).

The southward deflection of ice flowing up the Peabody River valley (south of Gorham) has already been mentioned. Other topographically controlled deflections from the regional trend of glacial flow have been recorded on the north side of the Presidential Range by Goldthwait (1940). His map shows northeast-trending striations on the lower northwest slope of Mt. Adams (Fig. 2). Goldthwait also mentioned east-trending grooves on Kelton Crag, located on a northern spur of Mt. Madison. Both of these localities are on the side of Randolph Valley, which is a deep east-west trough that cuts across the divide between the Connecticut and Androscoggin River basins.

THE ANDROSCOGGIN MORAINE

The Androscoggin Moraine overlaps the border between Gilead, Maine, and Shelburne, New Hampshire, in the Androscoggin River valley. Altogether it consists of at least 21 moraine segments, many of which are grouped in clusters of en-echelon ridges (Fig. 4). The arcuate patterns formed by these ridges, as well as the curvature of several individual ridges, indicate that the ice margin along which they were deposited was convex down-valley. Each of the moraine segments shown in Figure 4

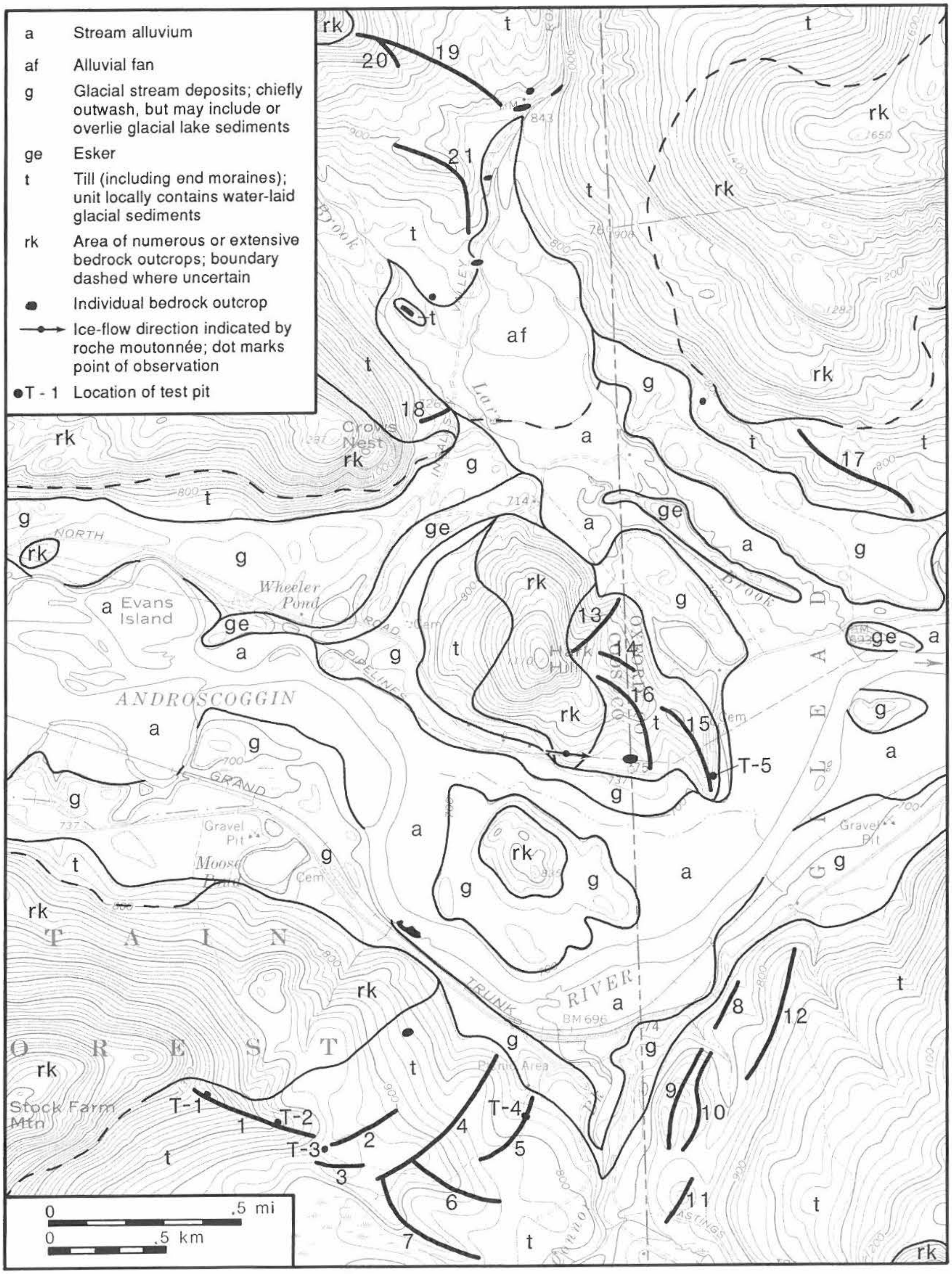

Figure 4. Surficial geologic map of the southeastern part of the Shelburne 1:24,000 quadrangle. Numbered lines indicate crests of moraine ridges (Androscoggin Moraine system).

is numbered for reference purposes in the discussion that follows.

Previous Research

G. H. Stone was probably the first to describe part of the Androscoggin Moraine. In an article concerning the "Androscoggin glacier" he gave a detailed morphologic description of prominent moraine ridges (9, 10, 11 on Fig. 4) on the south side of the valley (Stone, 1880). He noted that some of the morainal sediments "show signs of water-wash, with a loose structure as of gravelly upper till" (p. 300). Stone also mentioned the composite ridge (moraines 14-16) that projects southeast from Hark Hill. He illustrated this ridge in a later article, describing it as the "terminal moraine of the local Androscoggin glacier" (Stone, 1899, p. 274).

Upham (1904) briefly commented on the Androscoggin Moraine. He considered it to be possibly the most clearly defined moraine in the White Mountains, in contrast to the "promiscuous morainic belt" at Bethlehem. On the other hand, Leavitt and Perkins (1935) disputed Stone's (1899) identification of the Hark Hill ridge as an end moraine. They first described it as "a mass of bouldery till ... deposited in the angle between the Androscoggin ice tongue and one coming down Ingalls River valley" (p. 41). Later in the same volume, Leavitt and Perkins called the deposit a kame terrace (p. 118). They claimed that no corresponding ridge exists on the south side of the Androscoggin River valley, so evidently they were unaware of Stone's 1880 article.

Description of the Moraine System

Morphology. A study of the Androscoggin Moraine was begun in 1982. The authors carried out field work in the early spring and late fall, when the absence of leaf cover increased visibility in the densely wooded terrain. The first objective was to determine whether the ridges described by Stone are in fact moraines. Some of them are oriented parallel to regional or local ice-flow directions, which suggested that they might be drift tails. Stone recognized this possibility when he investigated the ridge extending southeast from Hark Hill: "It bears N20°W, which was so near the direction of the flow of the continental glacier that I carefully examined the northern end of the deposit to see if it was a 'tail' to a spur of the hill" (Stone, 1880, p. 301).

Our field work confirmed that both the deposits noted by Stone, and other nearby ridges, are moraines. They differ morphologically from drift tails in one or more of the following respects: (1) arcuate shape; (2) orientation transverse or oblique to ice flow; (3) presence of composite, multi-crested ridges; and (4) abrupt truncation against higher till or bedrock hills. The latter relationship is particularly noticeable on moraines 1 and 16 (Fig. 4). These high, sharp-crested moraines are very narrow in contrast to Stock Farm Mtn. and Hark Hill, and there are abrupt breaks in slope where they terminate against the sides of these hills.

The moraine segments described by Stone (1880) (14-16 and 9-11) are oriented transverse to the Androscoggin River valley. Their crests rise in elevation from as little as 720 ft* (219 m) where the moraine system is breached by the river to a maximum of 880-900 ft (268-274 m) on either side of the valley. These moraines are steep-sided, generally sharp-crested, and rise abruptly 30 m or more in places. The proximal sides of moraines 9 and 16 are especially high and steep. Ridges 9 and 10 comprise at least two closely-spaced moraines with intervening kettles and short connecting ridges. Many large boulders (>1 m) are strewn along the ridges; one boulder on the proximal slope of moraine 16 is 8 m in diameter. The lithologies of these boulders are identical to rock types (discussed below) that outcrop immediately upvalley from the moraine system. Extensive search did not reveal bedrock outcrops along any of the moraine ridges shown in Figure 4.

The ridge (1) that protrudes eastward from Stock Farm Mtn. is one of the most striking segments of the Androscoggin Moraine (Fig. 5). It reaches a maximum elevation of about 1260 ft (384 m) and has a very sharp crest that stands at least 30 m above the adjacent terrain. This moraine is accessible via a logging trail that leaves an old section of U.S. Route 2, about 0.35 km west of the state line (Fig. 4). The trail follows the crest of a lower moraine (5) and then passes several other morainal deposits (2-7) before ascending the high ridge of moraine 1. The lower moraines form a cluster of nested and locally overlapping(?) ridges. Some of them are more clearly defined than they appear on the topographic map, but they are not as prominent as moraine 1.

Figure 5. View southwest across the Androscoggin River valley. Arrows indicate moraine ridge (no. 1 on Fig. 4) projecting eastward from Stock Farm Mtn. (middle distance) and pine-covered moraines 14-16 extending from Hark Hill (foreground).

*Elevations are expressed in feet for ease of comparison with USGS topographic maps.

Other bouldery moraines with varying degrees of topographic expression have been discovered in addition to the ones previously reported by Thompson (1983, 1986). The northernmost are the group east of Mt. Cabot (Fig. 4). The highest member of this group (moraine 19) consists of several till ridges that rise as a series of en-echelon steps to a maximum elevation of 1240 ft (378 m), which is essentially the same elevation as the upper end of moraine 1 on Stock Farm Mtn. Moraine 21 comprises two closely spaced ridges that are not distinguished by the topographic contours.

Composition. There are few exposures of the glacial sediments in the Androscoggin Moraine system. Shallow cuts along logging trails and observations in shovel holes revealed loose, sandy, stony, light olive-gray to olive-colored diamicton. This material appears identical to ablation and resedimented facies of the typical late Wisconsinan surface till seen in crystalline-rock terrain over much of New England (Koteff and Pessl, 1985). A borrow pit exposure in moraine 18 (next to Ingalls Valley Road, Fig. 4) is badly slumped, but includes sand and gravel overlying and interbedded with bouldery till.

Five test pits were excavated by backhoe to depths of 1.8-2.4 m in order to examine the near-surface stratigraphy of the Androscoggin Moraine and collect samples for granulometric analysis and provenance studies. Four pits were located along the logging trail on the east side of Stock Farm Mtn., and one where North Road crosses the south end of moraine 15 adjacent to Hark Hill (T1-T5 in Fig. 4).

Test pit 1 showed 2.1 m of sandy, loose, bouldery diamicton. The exposed material has been thoroughly oxidized to a yellowish-brown color (10YR-5/8). It contains scattered lenses of thinly laminated, olive-gray silt and sand to 5 cm in thickness, which exhibit minor deformation. The diamicton is interpreted as till (possibly flowtill) from which most of the fines have been winnowed by meltwater.

Test pit 2 exposed 2.4 m of mostly massive, stony, sandy-silty diamicton. Pervasive oxidation is limited to the zone of modern soil development in the uppermost 1 m. Below this zone the diamicton is olive-colored (5Y-5/3). A small, deformed pod of stratified sand was seen near the bottom of the pit. The material in this exposure is chiefly till, though it may have been resedimented (as flowtill) from the adjacent ice margin.

Test pit 3 revealed a varied section in which 1.0 m of massive, pebbly to bouldery, sandy diamicton overlies 0.8 m of massive to stratified, sandy-silty diamicton interbedded with lenses of laminated silt and sand. The silt lenses are variably slumped and distorted, and contain scattered stones. The color of the nonoxidized diamicton below the modern soil zone is 5Y-5/3 (olive). We interpret the gravelly, washed upper diamicton unit to have formed as a debris flow off the ice margin in the area where moraine 2 abuts the lower end of moraine 1. The underlying stratified diamicton unit likewise was deposited in an unstable ice-marginal environment in which resedimentation occurred.

Test pit 4 exposed 1.8 m of massive, sandy-silty, olive-gray (5Y-5/2) diamicton. As in the previously described test pits, stones up to boulder size are common. The diamicton is equated with the typical surface till of the study area. It is rather compact, and water-laid sediments and collapse structures such as those seen in test pits 1-3 are lacking. The till in test pit 4 may be a different facies -- perhaps a lodgement facies -- or just a single, relatively homogeneous debris flow emplaced off the ice margin.

Test pit 5 was located in the floor of a small overgrown borrow pit in moraine 15. It showed 2.1 m of section, the upper 0.6 m of which is loose, oxidized sandy gravel. This gravel is separated by a sharp contact from the underlying nonoxidized unit, which is a well-stratified sequence consisting of three sub-units. From top to bottom, these are: (1) 10 cm of thinly laminated silt containing a few pebbles and cobbles; (2) 41 cm of moderately fissile, massive, silty-sandy, olive-gray (5Y-5/2) diamicton with scattered pebbles and cobbles; and (3) thinly laminated silt-sand with local deformation (adjacent to dropstones) and lenses of pebbly diamicton. Contacts between these sub-units are indistinct and defined mainly by the presence or absence of lamination. The gravel unit in test pit 5 is outwash, while the lower unit probably is a ponded deposit that received subaqueous debris flows (flowtills) from the adjacent glacier margin. Many large boulders occur on the moraine surface, but none were seen in the test pit.

Table 1 shows the combined percentages of silt and clay in bulk sediment samples collected from the unweathered zones in the test pits. The increase in fines from pit 1 to pit 5 is attributed to two factors. First, meltwater removed nearly all of the silt and clay from the gravelly till in moraine 1, high on the side of Stock Farm Mtn. Some of the tills in pits 2-4 retain their fines, and they locally contain interbeds of silty water-laid sediments. Second, silty beds are especially abundant in pit 5, where the morainal sediments were deposited in a ponded environment in the valley bottom.

The morphologic and stratigraphic evidence presented above demonstrate that the ridges of glacial sediment shown in Figure 4 are end moraines. The interbedded flowtills and water-laid deposits, locally contorted bedding, and angular stones scattered through the stratified units are evidence of the dynamic depositional environments and abundant meltwater that occur at the margins of temperate glaciers.

TABLE 1. WEIGHT PERCENTAGE OF SILT AND CLAY FRACTION (PERCENTAGE PASSING #200 SIEVE) IN BULK SAMPLES FROM TEST PITS 1-5

T-1	2.2 %
T-2	13.1
T-3	34.5
T-4	38.0
T-5	46.0

Provenance. Large areas of bedrock are exposed upvalley from the Androscoggin Moraine (Fig. 4). Billings and Fowler-Billings' (1975) map of the Gorham 15-minute quadrangle shows three principal rock types in the 5-km section of the Androscoggin River valley just west of the moraine system. In decreasing order of abundance, these are: Littleton Formation (Dl), consisting mostly of high-grade paragneiss, schist, and quartzite; medium-grained, gray biotite quartz diorite (qd); and Concord quartz monzonite (co), a medium-grained gray rock containing both muscovite and biotite. These rock types can be seen in road cuts along U.S. Route 2 in Shelburne. Much of what Billings and Fowler-Billings (1975) considered to be Littleton Formation in the Shelburne area was later reassigned to the Rangeley, Smalls Falls, and Madrid Formations by Hatch and Moench (1984) and Moench (1984).

Most of the boulders in the Androscoggin Moraine are the same rock types that outcrop immediately upvalley. Coarse, variably rusty, two-mica gneisses and schists of the Rangeley Formation (formerly Billings and Fowler-Billings' Dlg member of the Littleton Formation) and biotite quartz diorite are particularly common. For example, moraine 4 contains many large boulders (1-3 m), with a great concentration on the proximal side. Most of these boulders are the quartz diorite that outcrops as near as 0.25 km to the northwest. The area of outcrop of this intrusion extends across the northeast slope of Stock Farm Mtn. and is somewhat larger than shown on Billings and Fowler-Billings' map.

Table 2 presents the results of stone counts on 100-stone samples from test pits 1, 2, 4, and 5. Clasts of the Rangeley, Smalls Falls, Madrid, and Littleton Formations were not differentiated as such, because it was not always possible to distinguish these formations in hand specimen. We have retained Billings and Fowler-Billings' (1975) names and symbols for lithologies listed in Table 2 so that comparisons can be made with these authors' map of the Gorham quadrangle.

Some major differences among the stone counts are apparent in Table 2. Most of the variation occurred between test pits 1 and 2 (both located in moraine 1) and pits 4 and 5, which were located nearer the center of the valley in moraines 5 and 15 (Fig. 4). It is likely that moraine 1, on Stock Farm Mtn., would have received debris from the southern margin of the Androscoggin ice tongue. This theory is supported by the relative amounts of granite pegmatite and quartz diorite in the test pits. The somewhat greater percentage of pegmatite in test pits 1 and 2 may reflect the 0.3-km wide pegmatite outcrop on the south side of the valley, 2 km west-northwest of these pits (Billings and Fowler-Billings, 1975). The abundant quartz diorite in test pits 4 and 5 probably came from the outcrop area of this distinctive rock unit northeast of Stock Farm Mtn. The sample from test pit 4, containing 59 percent quartz diorite, was taken immediately downvalley from the diorite outcrops.

Differences in percentages of other rock types among the test pits are not easily explained. The Concord quartz monzonite, for example, outcrops widely to the west and northwest of the Androscoggin Moraine, and yet is considerably more abundant in moraine 1 than in moraines 5 and 15. A noteworthy variation also exists in the combined percentages of basaltic and rhyolitic dike rocks, which total 16 percent in test pit 5, but only 0-2 percent in the other pits. The dike lithologies in test pit 5 are heterogeneous, suggesting that these rocks were not simply derived from a single nearby outcrop. Billings and Fowler-Billings (1975) compiled a map of dikes in the Gorham quadrangle. This map shows clusters of dikes on the north side of the Androscoggin River valley at Gorham and in the Berlin area. Perhaps the moraines projecting from Hark Hill (Fig. 5) are enriched in dike rocks because ice on the north side of the valley could have incorporated a greater percentage of erratics from the Gorham-Berlin dike swarms. Alternatively, the local abundances of dikes on Billings and Fowler-Billings' map may be an artifact of the extensive outcrop in those areas (particularly at Berlin).

Correlations. Correlation of the Androscoggin Moraine segments on opposite sides of the valley is tenuous because of their range of elevations and the gaps caused by glacial and Holocene stream erosion. The earliest ice-margin positions are marked by moraines 8-12 and 17, which were built to maximum elevations of 850-890 ft (259-271 m) (Fig. 4). A single tongue of eastward-flowing ice extended across the Androscoggin River valley when these moraines were deposited. Moraines 6 and 7 may correlate with members of the 8-12 group, but meltwater flowing down the Connor Brook valley eroded the lower ends

TABLE 2. STONE COUNTS FROM TEST PITS

Lithologic names followed by symbols are those of Billings and Fowler-Billings (1975). One hundred stones were counted from each pit.

Lithology	No. of Stones Pit 1	Pit 2	Pit 4	Pit 5
Diorite (dw)	4	6	0	0
Granite pegmatite (p)	6	4	0	2
Concord quartz monzonite (co)	20	18	2	4
Quartz diorite (qd)	26	25	59	46
Diorite (dn)	2	0	0	0
Biotite quartz monzonite (bqm)	1	4	2	5
Basalt (dike rocks)	0	2	2	13
Rhyolite (dike rocks)	0	1	0	3
Undifferentiated igneous rocks (including White Mountain Plutonic/Volcanic Suite)	2	1	2	2
Mica schist and micaceous quartzite (Dlsg)	27	25	7	2
Calc-silicate rocks (Dlb, Dll)	0	1	0	0
Biotite-muscovite gneiss (Dlg)	10	8	16	19
Quartz conglomerate (?) (Dlc)	0	1	0	0
Undifferentiated metasedimentary rocks (mostly granofels)	0	0	8	2
Amphibolite (Ammonoosuc Volcanics - Oam)	0	0	2	2
Biotite gneiss (Ammonoosuc Volcanics - Oam)	2	4	0	0

of these ridges, destroying any former connections with the latter moraines.

The other moraines east of Stock Farm Mtn. (1-5) are younger than 6-12, and probably are contemporaneous with the moraines projecting southeast from Hark Hill (14-16). The steeply sloping crest of moraine 1 is believed to closely reflect the slope of the youngest ice margin recorded by the moraine system. This steep gradient (144 m/km), together with the apparent truncation of moraines 6-7 by 4, suggest that the ice margin thickened and readvanced slightly when the younger moraines were constructed.

The Androscoggin River valley ice lobe probably divided into three smaller sublobes as the ice receded westward past Hark Hill and Crows Nest (Fig. 4). While one sublobe deposited the moraines between Stock Farm Mtn. and Hark Hill, another extended through the gap between Hark Hill and Crows Nest, forming moraine 13 and possibly 18. The third sublobe spilled across the saddle between Crows Nest and Mt. Cabot, and deposited the 19-20-21 group. The steep profile (up to 213 m/km), bouldery crest, and upper elevation limit of moraine 19 are similar to those of moraine 1. These two moraines could have formed at the same time, when Crows Nest was beginning to emerge as a nunatak.

RELATIONSHIP OF MELTWATER CHANNELS AND DEPOSITS TO REGIONAL PATTERN OF DEGLACIATION

When glacial meltwater deposits of the northern White Mountains are considered together with those of the Mahoosuc Range-Blue Mountains in western Maine, a southeast-to-northwest pattern of deglaciation is evident. The retreat of the late Wisconsinan glacier from the study area was marked by the following events: (1) deposition of eskers and subglacial cutting of meltwater channels when glacial streams first penetrated mountain passes opening to the southeast; (2) subaerial channel cutting and minor glaciofluvial sedimentation as the uplands were deglaciated; (3) formation of ice-contact lacustrine deposits where meltwater was ponded between the glacier margin and north- to west-facing slopes; and (4) deposition of outwash and/or glaciolacustrine sediments on valley floors when local deglaciation had reached an advanced stage. Two or more of the above events occurred in the indicated sequence in some valleys, but these events overlapped in time across the area as a whole. Gerath (1978) described similar meltwater environments in the Berlin-Gorham region.

Eskers

Englacial meltwater drainage systems developed as the last ice sheet thinned over the mountains of northern New Hampshire and western Maine. Inspection of the state surficial maps (Goldthwait et al., 1951; Thompson and Borns, 1985a) shows that several long esker systems followed hydraulic gradients leading to notches through the mountains. Examples of notches that were conduits for esker-forming streams include Crawford Notch at the south end of the Presidential Range (Goldthwait, 1940), Grafton Notch and Sawyer Notch in the northeast part of the Mahoosuc Range (Caldwell, 1975c), and other mountain passes northeast of the study area (Thompson and Borns, 1985a). Judging from the deep, narrow bedrock gorges found where some esker segments terminate at notches in western Maine, intense subglacial erosion occurred as the streams passed through these gaps. Wherever evidence of meltwater flow direction has been found in the eskers, it reveals that the streams depositing them flowed in directions ranging between south and east.

High-Elevation Meltwater Channels, Glacial Lakes, and Related Deposits

Streams issuing from the ice margin probably continued to carve channels on the floors of mountain passes as the latter sites were deglaciated. However, the relative importance of proglacial versus subglacial channel cutting is uncertain. Eskers are very small or absent at the heads of some passes such as Evans Notch (discussed below), so the meltwater channels in these places may have been cut mainly by proglacial streams. Elsewhere, hillside channels developed along the irregular margin of the thinning ice sheet. Examples were described by Gerath (1978) and Goldthwait and Mickelson (1982).

Ephemeral ice-contact lakes were ponded in topographic reentrants between the ice margin and emerging mountain slopes during deglaciation. Deposits associated with the lakes are partly deltaic, but may also include lake-bottom and glaciofluvial sediments. Although they are mostly small and not abundant, these lake deposits indicate a generally northwestward recession of the ice margin. None of the ponded deposits known in the study area are located southeast of their corresponding spillways.

Several meltwater drainage systems are described below to illustrate the pattern of ice retreat from the uplands.

Evans Notch-Wild River Area. Shortly before deposition of the Androscoggin Moraine system, thinning and recession of glacial ice in the high mountains just south of the Androscoggin River caused meltwater to pour southward into the Saco River basin. One of the principal drainage routes was the deep bedrock gorge at Evans Notch, just east of the Maine-New Hampshire border (Fig. 6, E-1). Ponding may have occurred behind the 1400-ft (427 m) divide in this notch, but only a few small fluvial or deltaic deposits of sand and gravel were graded to the floor of the meltwater channel that crosses the divide.

Withdrawal of the ragged ice margin from Evans Notch permitted meltwater to drain northeast along the Wild River into the Androscoggin River valley. During part of this phase, the glacier margin lay against the high ridge between the Andros-

Figure 6. Map showing meltwater drainage paths (indicated by arrows) in the uplands, including possible spillways for ice-contact glacial lakes, and outlets for glacial Lake Bethel: B - Bowman divide; C - Copperville area; E - Evans Notch-Wild River area; H - Hunters Pass; LB - glacial Lake Bethel; LC - glacial Lake Cambridge; M - Mahoosuc Range-Success Moraine area.

coggin Moraine and the Wild River. Two meltwater channels cut across this ridge just east of the state line (Fig. 6, E-2). A small ice-contact delta(?) is graded to the north end of one channel, whose elevation is about 1110 ft (338 m). The other channel is at approximately 1090 ft (332 m). The authors noted a bedrock outcrop and lag concentrate of boulders on the floor of the latter channel, which is crossed by Hastings Trail. Drainage from both outlets plunged down the steep slopes to the southeast and into the Wild River.

Glacial Lake Cambridge. Near the northern limit of the study area, Grafton Notch was an early outlet, at about 1550 ft (472 m), for meltwater draining south through the Mahoosuc Range (Fig. 6, LC-1). The prominent, flat-floored channel passing through the notch is located west of Route 26. A segmented esker that terminates just north of the notch was deposited when ice still filled the lowland northwest of the Mahoosucs. Some of the small, poorly exposed kames flanking the esker are high enough to be deltas graded to the Grafton Notch channel, but their identity is uncertain. Ponded meltwater draining through the notch would have constitued the earliest and highest level of glacial Lake Cambridge, which was named by Leavitt and Perkins (1935).

Recession of the ice margin caused Lake Cambridge to drain eastward through Dunn Notch (Fig. 6, LC-2), at an elevation of about 1450 ft (442 m). Deltaic topset and foreset beds were seen in a gravel pit located south of Cedar Brook on the west side of Route 26 (Fig. 1). The level of the topset/foreset contact in this delta corresponds to the Dunn Notch spillway.

Further deglaciation may have resulted in eastward drainage of Lake Cambridge across the C Pond Moraine (Fig. 1, CPM) and down through Sawyer Notch. The elevation of the lowest possible spillway across the moraine is likewise approximately 1450 ft (442 m). This spillway could not have drained the lake for very long; slight additional recession of the ice from the Upton area opened lower outlets at 1270-1330 ft (387-405 m) into the Androscoggin River valley northwest of Hampshire Hills (Fig. 1). The demise of Lake Cambridge presumably occurred when it emptied directly into the head of the Androscoggin River near Errol. Additional field work is needed to verify the lake spillways and to determine the regional significance of the C Pond Moraine (originally reported by Caldwell, 1974).

Success and Copperville Moraines. Progressive northwestward recession of the ice margin is also recorded by meltwater channels and deposits in the part of the Mahoosuc Range near Berlin. Initial drainage occurred at high levels soon after the peaks emerged from the ice. Leavitt and Perkins (1935) and Gerath (1978) described Mahoosuc Notch (Fig. 6, M-1) as one of the earliest drainage channels. Its elevation is about 2470 ft (753 m).

Gerath (1978) reconstructed the deglaciation of the New Hampshire portion of the Mahoosuc Range. He found that retreat of the ice margin down the proximal side of the range allowed meltwater to drain southwestward into the Androscoggin River valley, carving a series of channels that are lower to the northwest (Fig. 6, M-2,3,4). Gerath discovered thick ice-contact deposits in the township of Success and called them the "Success Moraine" (Fig. 1, SM). He described this moraine as "a complex mass of outwash and glaciolacustrine sediment which formed in a depositional environment confined between a broad, retreating Laurentide ice front and the north slopes of the Mahoosuc Range" (Gerath et al., 1985, p. 23).

The Success Moraine reaches elevations of slightly over 1600 ft (488 m). On topographic maps it does not show obvious ice-marginal ridges, and sedimentation in ponded water played a major role in its formation. The moraine is a lacustrine ice-contact morphosequence in the terminology of Koteff and Pessl (1981). Gerath (1978) described an exposure that indicates that the ice margin advanced over the proximal side of the Success Moraine. It shows 6 m of till overlying 12 m of deformed glaciolacustrine sediments. Gerath concluded that the moraine was deposited at the terminus of an active ice tongue that persisted in the Androscoggin River valley, but that the lower meltwater channels a short distance to the northwest formed as the ice tongue stagnated. Gerath noted similarities between these channels and modern analogs adjacent to stagnating ice in Alaska.

The Copperville Moraine (Fig. 1, CM), also named and first described by Gerath (1978), is located northwest of Berlin in what he called the "Copperville through-valley." This valley crosses the divide at about 1070 ft (326 m) between the Connecticut and Androscoggin River basins. The Copperville Moraine is a small transverse ridge of stratified sand and gravel, situated approximately 3.5 km northwest of the divide. Gerath (1978) interpreted it as being the collapsed head of a fluvial ice-contact morphosequence that was built across the divide and into the Dead River valley to the southeast. The moraine was thought to have been deposited adjacent to active ice flowing from the Connecticut River basin and is somewhat lower and younger than the Success Moraine. Although it was identified as a fluvial deposit, some meltwater ponding probably occurred on the proximal side of the Connecticut-Androscoggin divide (Fig. 6, C).

Glacial Lakes South and West of Gorham. Field evidence suggests that meltwater was ponded in the lower part of the Peabody River valley (Fig. 1). About 2.5 km south of Gorham village, next to where Route 16 crosses the river, a cut bank shows a diamicton unit (probably till) overlying rhythmically bedded lacustrine sediments. This exposure was briefly mentioned by Crosby (1934) and subsequently described by Gosselin (1971), Goldthwait (1972), and Davis et al. (1988). It is believed to represent overriding of glacial-lake beds by an ice tongue advancing southward from the Androscoggin River valley. Until more work is done on this section, several interpretations must be considered. The deposits seen here may have formed during either the advance or recession of the late Wisconsinan ice sheet; and it is also possible that the lacustrine sediments predate the last glaciation.

Other features besides the Copperville Moraine demonstrate that meltwater from ice in the Connecticut River basin spilled eastward into the Androscoggin River valley. One of the higher meltwater channels is Ice Gulch (also known as Icy Gorge), a 40-m deep gorge illustrated by Goldthwait and Mickelson (1982). This channel heads at the 2490-ft (759 m) gap called Hunters Pass (Fig. 6, H). It drained meltwater from ice northwest of the Crescent Range.

West of Gorham village, there is a deep east-west glacial trough between the Presidential Range to the south and the Crescent Range to the north. Known as "Randolph Valley" (Goldthwait, 1940), it straddles the drainage divide at Bowman, whose elevation is about 1500 ft (457 m) (Fig. 6, B). This divide separates the Moose River, which joins the Androscoggin at Gorham, from Israel River, which flows westward to the Connecticut River. Randolph Valley channeled meltwater into the Androscoggin River valley from a broad ice lobe in the upper Connecticut River basin.

Gerath (1978) found that meltwater draining through Randolph Valley deposited the Moose Brook delta in a small glacial lake at Gorham. Gerath also mapped a younger outwash fan from the same source in Moose Valley. These deposits indicate that glacial streams continued to transport sediment eastward through Randolph Valley until the Gorham area was essentially ice-free. Gerath (1978) concluded that the upper Connecticut River basin probably held the latest active ice in the region.

Goldthwait et al. (1951) and Goldthwait and Mickelson (1982) mapped a series of "ice-contact deposits" west of the Bowman divide. Goldthwait (1940) described these deposits as gravel hummocks that are lower in elevation to the east. Descending sets of meltwater channels occur on both sides of Randolph Valley at the proximal end of this series of kames (Goldthwait and Mickelson, 1982). Lougee (1939) published a photograph of a pothole in the bedrock floor of one channel. Perhaps the kames are chiefly fluvial deposits, in which case they may be contemporaneous with the outwash carried down the Moose River valley to Gorham. However, the present authors saw delta foreset beds in some of the deposits west of Bowman, and Lougee (1939) considered this to be an area of deltaic sedimentation. Both fluvial and lacustrine environments may have existed in turn as the Israel River valley was deglaciated.

Other Ice-Marginal Lakes. In addition to the examples discussed above, there probably are other places where meltwater was ponded between mountain slopes and the ice margin. Lougee (1939) compiled a map showing an extensive network of lakes that he thought existed along the glacier margin as it receded northwestward from the upper Connecticut River basin in New Hampshire. Evidence for some of these water bodies, such as glacial Lake Ammonoosuc, has been discussed in the literature (Goldthwait, 1916; Lougee, 1940). Certain other lakes proposed by Lougee (1939) are not supported by field evidence (R. P. Goldthwait, pers. comm., 1987). Nevertheless, the distribution of ice-contact deposits and associated meltwater channels in the areas described above indicate a consistent pattern of northwestward thinning and retreat of an irregular ice margin from the mountains of northern New Hampshire and adjacent Maine.

Meltwater Deposits along the Androscoggin River

The upper Androscoggin River valley contains both glaciofluvial and glaciolacustrine deposits. The fluvial deposits include scattered esker segments, but mostly comprise dissected outwash which is locally kettled and collapsed from deposition over dead ice. Gerath (1978) has described the meltwater deposits from the state line upstream to Milan, New Hampshire, so this area will not be discussed further.

Geologic mapping in the Androscoggin River valley of western Maine has not progressed to the point where we can reconstruct the sequence of meltwater deposits in detail. However, gravel-aquifer studies by Williams et al. (1987) suggest the presence of a single long outwash sequence extending from Gorham east to Bethel. The elevations of the dissected outwash remnants become gradually lower downvalley, from about 840 ft (256 m) at Gorham to 700 ft (213 m) at Bethel. There is an overall decrease in particle size and degree of ice-contact collapse along this section of the valley. Bouldery, poorly-sorted gravel in pits at Gorham contrasts with the well-stratified sand and pebble gravel of the Bethel area. In places, however, a greater abundance of kettles, together with kames or ice-channel fillings at higher elevations, may indicate the presence of older sequence heads that were modified by the later drainage from New Hampshire.

The meltwater from the upper Androscoggin River valley emptied into a glacial lake at Bethel (Fig. 1). Evidence for this lake includes a delta at West Bethel. A topset/foreset contact in the delta was prominently exposed along the south bank of the Androscoggin River following the spring flood of 1987. The elevation of the contact (which indicates the former lake level) is approximately 690 ft (210 m). Exposures of delta topset and foreset beds were also seen in borrow-pit exposures, and lake-bottom sediments were penetrated by an observation well (well 14-2 in Williams et al., 1987). Moreover, several test holes at Bethel have encountered glaciolacustrine sand, silt, and clay (Prescott, 1980). We propose to call the lake which existed in this area "glacial Lake Bethel." Lake Bethel extended into the low, swampy valleys south of Bethel village. It initially drained south through outlets into the Crooked River basin (Fig. 6, LB-1,2,3). One of the best-defined of the probable spillways (LB-2) is a channel at about 690 ft (210 m), which carried water into the northeast corner of Songo Pond.

Deglaciation of the Bethel lowland eventually allowed Lake Bethel to drain northeast along the Androscoggin River valley. The last impoundment of the lake may have occurred behind a till barrier that probably blocked a narrow segment of the valley in Newry, and which has since been incised by the modern river (Fig. 6, LB-4).

DISCUSSION

There has been much controversy as to whether the White Mountains were deglaciated mainly by regional stagnation and downwastage of the ice, or by progressive retreat of an active ice margin. Johnson (1941) attempted to reduce the confusion by pointing out that "normal ice retreat" and "down-wasting" are not mutually exclusive processes. Thinning by ablation of the terminal region of a terrestrial ice sheet can result in recession of the glacier margin regardless of whether the ice is stagnant or actively flowing. Moreover, the configuration of the margin would be irregular in a high-relief terrain such as the White Mountains, even if deglaciation occurred by recession of active ice.

Goldthwait and Mickelson (1982) advocated the concept of early and pervasive glacial stagnation as the last ice sheet thinned over the White Mountains. They also noted that there was a northwestward progression in the uncovering of valleys. Gerath et al. (1985) discussed the importance of topography in controlling the dynamics of the thinning ice sheet and concluded that the Copperville and Success Moraines formed at the edge of ice streams that persisted in valleys open to the northwest.

Meltwater channels and ice-contact deposits like those described in the previous section of this report commonly form in association with stagnant ice (Goldthwait and Mickelson, 1982), and stagnation presumably did occur in valleys that were cut off from ice flow. However, our work on the Androscoggin Moraine supports the conclusion that active ice was present in the northeastern White Mountains until an advanced stage of local deglaciation, when the Presidential and Mahoosuc Ranges had largely emerged from the late Wisconsinan ice sheet. The position of the Androscoggin Moraine suggests that it is slightly older than the Success Moraine, but certainly younger than the meltwater drainage through higher channels to the south. Thus we infer that an irregular but still-active ice margin retreated across the Maine mountains south of the Androscoggin River just prior to deposition of the moraine. Local stagnation eventually occurred as ice masses separated over the mountains, especially in the lee of the Mahoosuc Range, but ice flow continued in the upper Androscoggin River basin.

The limited striation evidence indicates that the ice which deposited the Androscoggin Moraine and stratified moraines of the Berlin-Gorham area did not originate in the lowland at the head of the Androscoggin River. Instead, it is believed to have flowed from the west and northwest as part of an ice lobe in the upper Connecticut River basin. This lobe extended through the Copperville and Randolph Valleys and deposited the moraines described above. Gerath (1978) found evidence that ice persisted in the Connecticut River basin, shedding debris and meltwater into the Androscoggin River valley until the latter area was almost totally ice-free.

No field evidence has been discovered from which to reconstruct the ice-surface profile across the mountains of northernmost New Hampshire when the Androscoggin Moraine was deposited. However, J. L. Fastook (Quaternary Studies Institute, Univ. Maine) developed a preliminary computer model of the ice profile extending from the moraine north-northwestward across the mountains for a distance of 85 km. Several combinations of ice-floor elevations and glaciological conditions were tested using Fastook's model, and each of the reconstructed ice profiles was high enough to cover all (or nearly all) of the mountainous terrain upglacier from the Androscoggin Moraine. Therefore, it seems likely that the moraine was deposited by ice flowing out of the upper Connecticut Valley region, though how far to the north this flow originated is uncertain.

The rarity of moraines in the White Mountains and western Maine leads us to question why the Androscoggin Moraine system formed in the first place. It could be the result of a local glacial readvance, but there is no independent evidence to support this theory. Nor is there any reason why a readvance capable of depositing such a large moraine cluster would not be recorded elsewhere in the study area.

Examination of the regional topography shown on the Lewiston 1:250,000 quadrangle reveals another possible explanation for the Androscoggin Moraine. The highest part of the White Mountains extends from the Franconia Range northeast to the Presidentials and thence to the Mahoosuc Range and Blue Mountains in Maine. Together, these ranges form a continuous northeast-trending mountain chain that obstructed the southeast flow of the late Wisconsinan ice sheet. The ice was funneled into narrow valleys that penetrate the mountains and eroded deep troughs such as Franconia Notch and Crawford Notch. Along a broad section of the mountain front including the Presidentials and Mahoosucs, ice flow was directed into the constricted segment of the Androscoggin River valley east of Gorham. The conditions were similar to drainage of modern ice caps by outlet glaciers (Embleton and King, 1975). Accelerated flow occurred in the narrow ice stream extending down the Androscoggin River valley. A balanced condition was attained for an uncertain length of time, during which the ice margin stood at the Maine-New Hampshire border and deposited the Androscoggin Moraine. The large boulders and considerable volume of sediment in the moraine ridges attest to the erosion caused by this ice stream.

It remains to determine the age of the ice-marginal deposits discussed here and the relationship of ice retreat in the study area to the deglaciation history of northern New England and adjacent Quebec. Davis and Jacobson's (1985) synthesis of radiocarbon dates shows that the receding late Wisconsinan ice margin reached the inland marine limit in southwestern Maine by about 13,000 yr B.P. A reconnaissance of ice-contact deposits and meltwater channels indicates that the glacier margin then receded generally northwest from the marine limit to the White Mountains (W. B. Thompson, unpublished data). On the basis of these inferences, the upper Androscoggin Valley region may have been deglaciated sometime after 13,000 yr B.P. This chronology is compatible with Gerath's (1978) estimate that the Gorham area became ice-free between 12,600 and 12,100 yr B.P., but probably is too young for reasons discussed below.

If the study area was deglaciated after 13,000 yr B.P., there is a question as to whether the recessional deposits were formed by the main Laurentide ice sheet or a residual Appalachian ice mass that existed along the Maine-Quebec border in late-glacial time (Lowell and Kite, 1988). This ice mass began to detach from the Laurentide ice sheet prior to 13,000 yr B.P., as a marine calving bay encroached upon the ice stream flowing down the lower St. Lawrence Valley (Dyke and Prest, 1987). The Champlain Sea expanded into the St. Lawrence Lowland southwest of Quebec City no later than about 12,000 yr B.P., by which time the separation of the Appalachian ice mass was complete (Dyke and Prest, 1987; Parent and Occhietti, 1988). However, it is doubtful that this remnant ice extended far enough to the southwest to affect northern New Hampshire or adjacent western Maine. Numerous end moraines and other ice-contact deposits in the Sherbrooke-Asbestos area of Quebec indicate a progressive northwestward recession of the Laurentide ice margin from the northern tip of New Hampshire (Parent and Occhietti, 1988).

A difficulty with proposing ice retreat from the study area after 13,000 yr B.P. would be the lack of agreement with the deglaciation model of Davis and Jacobson (1985). These authors depicted the White Mountains and Mahoosuc Range as protruding from the thinning ice sheet by 14,000 yr B.P., and deglaciation of northernmost New Hampshire and the Boundary Mountains (on the Maine-Quebec border) by 13,000 yr B.P. Dyke and Prest (1987) likewise inferred that the ice margin lay a short distance north of the Quebec-New Hampshire border at 13,000 yr B.P. This sequence favors construction of the ice-marginal deposits in the northern White Mountains by the Laurentide ice sheet at about 14,000 yr B.P. Davis and Jacobson's model also shows a broad area of remnant ice over southwestern Maine that was cut off from the ice sheet northwest of the emerging mountains. No unequivocal field evidence has been found to support the existence of such a large remnant ice mass in this part of Maine, although striations at several localities south of the Androscoggin River indicate a late shift in glacial flow to the south and even to the south-southwest (W. B. Thompson, unpublished data).

We conclude that the upper Androscoggin River valley and adjacent mountains were deglaciated by the Laurentide ice sheet before 13,000 yr B.P., and perhaps as early as 14,000 yr B.P. This is based on Davis and Jacobson's (1985) analysis of New England radiocarbon dates, the ice-margin chronology proposed by Dyke and Prest (1987), and the synthesis by Parent and Occhietti (1988) of studies in southeastern Quebec. A possible discrepancy remains between deglaciation of the study area by 14,000 yr B.P. and a radiocarbon date indicating the presence of an active ice margin at Kennebunk in southwestern coastal Maine as recently as 13,200 yr B.P. (Smith, 1985). There are very few dates on actual ice-margin positions in northern New England. A more consistent pattern of ice retreat should become apparent as additional dates are obtained and combined with the results of intensive field studies.

SUMMARY

Early investigators of the deglacial history of the White Mountains often described small parts of this region, such as the Bethlehem Moraine or the deposits of glacial Lake Ammonoosuc. Some of the ensuing controversies over the style of deglaciation narrowly focused on these deposits, with the result that the overall history of retreat of the late Wisconsinan ice sheet from the mountains remained poorly documented. Arguments about the presence or absence of local glaciers were followed by equally intense debate concerning the mode of ice recession.

Recent workers such as Goldthwait and Mickelson (1982) and Gerath et al. (1985) have taken a broad perspective on the deglacial history of the White Mountains and have generally favored regional stagnation and downwastage as the principal means by which the ice disappeared. However, a study of the Berlin-Gorham region by Gerath (1978) and the present authors' work on the Androscoggin Moraine suggest that active ice flow persisted in the upper Androscoggin and Connecticut River basins until the Presidential, Carter, and Mahoosuc Ranges were largely deglaciated. The Androscoggin Moraine resulted from topographic funneling of ice flow and development of a rapidly flowing ice stream in the constricted east-west segment of the Androscoggin River valley. Gerath (1978) and Gerath et al. (1985) concluded that moraines in the Berlin-Gorham area likewise were deposited by ice streams. However, the latter moraines are morphosequences of stratified drift, in contrast to the sharp-crested ridges of bouldery till with interbedded meltwater deposits that comprise the Androscoggin Moraine.

Meltwater channels and deposits of ice-contact stratified drift reveal that systematic northwestward recession of the continental ice margin occurred throughout the northern White Mountains and adjacent western Maine. Retreat of an active ice margin was the likely mode of deglaciation in this region, especially on the northwest side of the White Mountains and Mahoosuc Range, though there was widespread stagnation of detached ice masses in the lee of the mountains.

Deglaciation of the study area probably occurred by retreat of the Laurentide ice sheet as early as 14,000 yr B.P. Borns (1985) similarly concluded that the nearby Frontier Moraine on the Maine-Quebec border was deposited along the Laurentide ice margin prior to 13,000 yr B.P. A contrasting model would deglaciate the upper Androscoggin River valley by progressive northwestward retreat of an ice margin that is thought to have lain near the Maine coast as recently as 13,200 yr B.P. This interpretation suggests that the Androscoggin Moraine and other ice-marginal deposits in the study area were deposited after 13,000 yr B.P. and may have been formed by a residual Appalachian ice mass. The latter model is problematic because field studies in southeastern Quebec have shown that remnant Appalachian ice on the Maine-Quebec border did not extend far enough to the southwest to affect the White Mountains, and that the Laurentide ice margin retreated progressively northwest

from the northern tip of New Hampshire (Parent and Occhietti, 1988). Moreover, Davis and Jacobson's (1985) interpretation of New England radiocarbon dates and the paleogeographic reconstructions by Dyke and Prest (1987) indicate separation of the late Wisconsinan ice sheet over the White Mountains and Mahoosuc Range by 14,000 yr B.P., and total deglaciation of our study area by 13,000 yr B.P.

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Late-Glacial Dunes, Ventifacts, and Wind Direction in West-Central Maine

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ABSTRACT

A late-glacial complex of eolian deposits, along with wind-modified till clasts and bedrock surfaces, was studied in a 100 km^2 belt along the Kennebec River near Madison, Maine. Of the 64 dunes mapped, most are longitudinal and the remainder are wind-shadow dunes. The eolian sand was derived from a sandy outwash deposit occupying the Kennebec River valley. This report discusses the areal extent, volume, and provenance of the sand, the uniformity of dune type, the orientation of longitudinal dune axes, and azimuths of flutes and cusps on wind-modified bedrock and fixed till-clast surfaces. These data indicate that the complex was formed primarily by west-northwest winds approximately 12,900 to 12,600 yr B.P. while the outwash was actively accumulating and before vegetation became well established in this region.

INTRODUCTION

The purpose of this study is to determine the nature and significance of the eolian features of the Anson quadrangle in west-central Maine: (1) to describe the distribution, morphology, and internal structure of the deposits, (2) to determine their provenance, (3) to describe the type and distribution of wind-modified rock surfaces, (4) to determine the time of formation of the sand deposits and the rock-surface modifications, and (5) to determine the direction of transport of the deposits.

The area of study was chosen because of the relatively wide distribution and good exposure of its eolian deposits, its accessibility, and its recorded glacial stratigraphy. The study area was limited to that part of the upper Kennebec River valley contained within the southern three-quarters of the Anson 15-minute quadrangle (Figs. 1 and 2). Hills in the northwest portion of the quadrangle reach 440 m above sea level and to the south the Kennebec River is approximately 49 m above sea level.

Many geologists, including True and Hitchcock (in Hitchcock, 1861), Stone (1886, 1899), Leavitt and Perkins (1935), Trefethen (1949), Allen (1955), Caldwell (1959, 1965), Hanley (1959), Bloom (1960), and Prescott (1966, 1969), have reported the presence of postglacial eolian deposits and/or ventifacts in localities throughout Maine. Others, including Perkins and Smith (1925), Leavitt and Perkins (1935), Smith (1964), and Borns and Hagar (1965a, 1965b), have reported them in the upper Kennebec River valley (Fig. 1). Some have suggested that the effective wind direction was either west, northwest, or north, but none has placed the period of formation of the eolian deposits and ventifacts into a precise glacial chronology. McKeon (1972) provides a more detailed review of these earlier studies.

LATE-GLACIAL STRATIGRAPHY

The provenance of the eolian deposits and the chronology of the eolian activity in the study area is closely related to the late-glacial stratigraphy of the region. The stratigraphy of the northern part of the Kennebec River valley, as defined by Borns and Hagar (1965a, 1965b), indicates the following sequence of events. As the margin of the last Wisconsin ice sheet retreated northwestward in Maine, the sea transgressed the deglaciated and isostatically depressed region and deposited the clays and

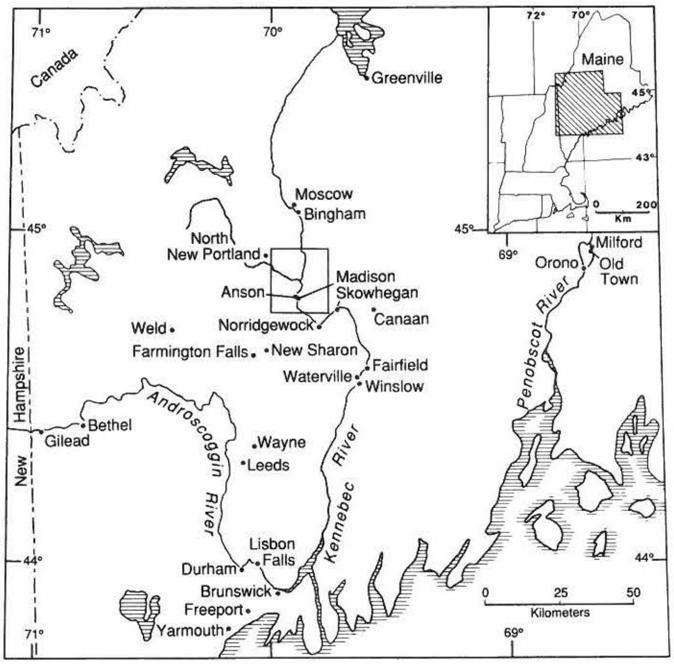

Figure 1. Map showing the location of the study area. Eolian features have been reported in or near all towns shown. For references, see text.

silts of the Presumpscot Formation. Locally, marine fossils found at the base of the Presumpscot Formation near its maximum altitude of 135 m in the Kennebec River valley at Embden Pond (Fig. 2, location 1), yielded a date of 12,930±200 yr B.P. (Y-1477, Borns and Hagar, 1965b). This date is thought to closely approximate the time of maximum marine transgression of the Kennebec River valley. There is some debate about this date since it is reported as 13,020±240 yr B.P. in Stuiver and Borns (1975). [Smith (1985) has noted that this date complicates the ice retreat pattern based on other regional dates and suggests that it may be too old.] As the rate of regional isostatic uplift exceeded the rate of eustatic rise of sea level, sandy outwash of the Embden Formation (Borns and Hagar, 1965b) was transported to the receding shoreline and was deposited gradationally on the Presumpscot Formation. The prograding Embden Formation formed a continuous plain within the Kennebec River valley south approximately to Waterville (Fig. 1). The late glacial sea had regressed at least to the present coast by about 12,300 yr B.P. (Stuiver and Borns, 1968). Since Waterville is halfway between the study area and the present coast, it is estimated that the southern limit of the Embden Formation stood in the sea at Waterville approximately 12,600 yr B.P. By this time the outwash ceased to accumulate, probably because the ice margin had withdrawn to a position north of Greenville where its meltwater streams could no longer discharge sediment into the Kennebec drainage. As uplift continued, both the Embden and Presumpscot Formations were deeply incised by the early Kennebec River. Within this incision, coarse gravels (the "North Anson Formation" of Borns and Hagar, 1965b) were deposited unconformably on the Presumpscot Formation. Presently, the river is cutting these postglacial stream terrace gravels (Fig. 2).

Note that throughout this article "Embden Formation" is used to refer to the coarser upper section of what is presently referred to as the Presumpscot Formation. While the Presumpscot Formation is largely glaciomarine, its upper sediments grade from estuarine to subaerial.

EOLIAN DEPOSITS

Much of the conventional terminology used to describe eolian deposits has been developed from work in modern, warm climate deserts. Although this terminology is used herein, it is recognized that the dune area of this study is small in scale and more rugged in topography when compared to those of the world's modern deserts. Another important distinction is that the eolian deposits of the Anson area were produced in a periglacial environment. The term dune is used for these deposits. Note that Koster (1982), in his report on the eolian deposits of the Netherlands, uses the term dune only if there is a height difference of a least five meters; any eolian ridge less than five meters in height he calls "undulations".

Eolian deposits, first defined on low altitude panchromatic aerial photographs and soil maps, were examined and sampled during the summer of 1971.

Distribution

The eclian deposits are not distributed randomly throughout the study area. All are found east of the western limit of the Embden Formation as mapped by Borns and Hagar (1965b) within the Kennebec River valley, with the exception of two localities southwest of the confluence of the Carrabassett and the Kennebec Rivers (Fig. 2). Eolian deposits in the Carrabassett River valley are on, or less than 0.5 km south of, the Embden Formation. Of the 64 dunes identified in the study area, 50 are east of the present Kennebec River, and all but three of the remainder are in the Carrabassett River valley. In general, more dunes are to the east than to the west of a chain of hills 100 m in relief that trends north-south between the eastern limit of the Embden Formation and Wesserunsett Lake (Fig. 2). The range in altitude of all dunes is 85-218 m above sea level, and the mean is 138 m above sea level. With respect to the glacial stratigraphy, one dune rests upon the erosional surface of the Presumpscot Formation, nine on the Embden Formation, and one on the postglacial stream terrace gravels. The remaining 53 dunes rest upon till and/or bedrock.

Morphology

There are three basic types of eolian deposits present in the study area: longitudinal dunes, wind-shadow dunes, and blanket

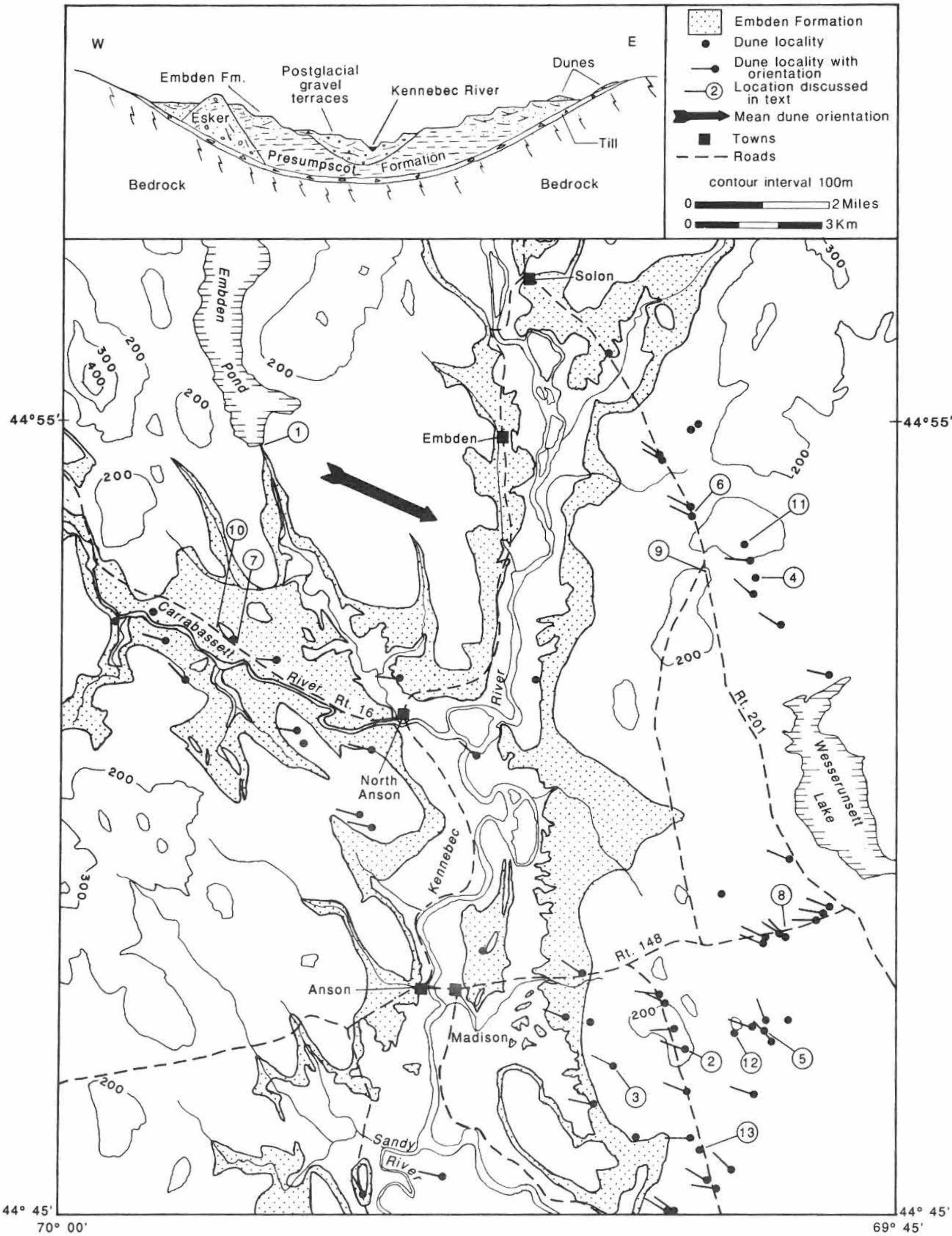

Figure 2. Dune locations and orientations, general topography, location of late-glacial outwash, and generalized cross section of the Kennebec River valley within the study area. (Base map adapted from the 1:62,500 scale Anson quadrangle of the U.S.G.S., and generalized cross section modified from Borns and Hagar, 1965a. The 100 m elevation contour has been omitted for clarity.)

Figure 3. A well formed, stable, and vegetated longitudinal dune trending west-northwest to east-southeast, center to lower left, rests upon the Embden Formation outwash in the Carrabassett River valley 2 km west of North New Portland near Route 145 (Fig. 1). This dune is about 400 m in length and 4 m in width.

Figure 4. An unstable and partially vegetated wind-shadow dune on the southeast flank of Ward Hill (Fig. 2, location 2).

TABLE 1. APPROXIMATE DIMENSIONS IN METERS OF 10 LONGITUDINAL DUNES IN THE ANSON AREA. LISTED IN ORDER OF DECREASING AREA. (-) INDICATES NOT MEASURED.

	Height	Length	Width	Length/Width Ratio
	4	380	35	10.9
	3	165	33	5.0
	7	80	44	1.8
	1	100	16	6.3
	-	100	13	7.7
	2	33	25	1.3
	3	31	16	1.9
	2	33	13	2.5
	2	50	7	7.1
	1	33	6	5.5
Mean	2.8	101	21	5.0

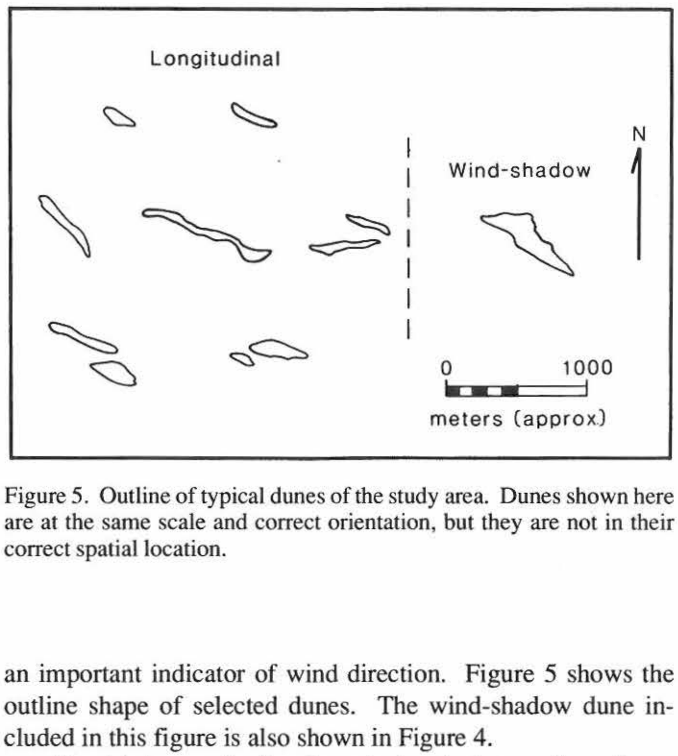

Figure 5. Outline of typical dunes of the study area. Dunes shown here are at the same scale and correct orientation, but they are not in their correct spatial location.

sand. The most useful eolian deposits for interpreting wind direction are longitudinal dunes (Fig. 3). In plan view these dunes vary from linear to sinuous forms with length/width ratios from 11 to 1.3 (Table 1). These sand ridges are generally symmetrical in cross section and have flank slopes dipping 5 to 20^{o}.

Approximately one quarter of all dunes identified are of the wind-shadow type (Fig. 4). These dunes, unlike the relatively uniform longitudinal dunes, are irregular in plan, often with length/width ratios close to 1.0, and have maximum diameters up to 800 m. The volume of the average wind-shadow dune is five to ten times that of the average longitudinal dune. Very few of the wind-shadow dunes have a shape that is an indication of wind direction; however, their location relative to an obstacle is an important indicator of wind direction. Figure 5 shows the outline shape of selected dunes. The wind-shadow dune included in this figure is also shown in Figure 4.

The third type of eolian deposit, the blanket sand, conforms to the underlying topography as a more or less continuous sheet of sand less than one meter thick. Because of its planar form, the blanket sand is not an adequate indicator of wind direction. Furthermore, because they are more susceptible to destruction by erosion and by agricultural tilling, blanket sands have not been preserved as well as the longitudinal and wind-shadow dunes.

There is a fair correlation between type of eolian deposit and local relief. Longitudinal dunes generally are found on the remnants of the Carrabassett and Kennebec River valley out-

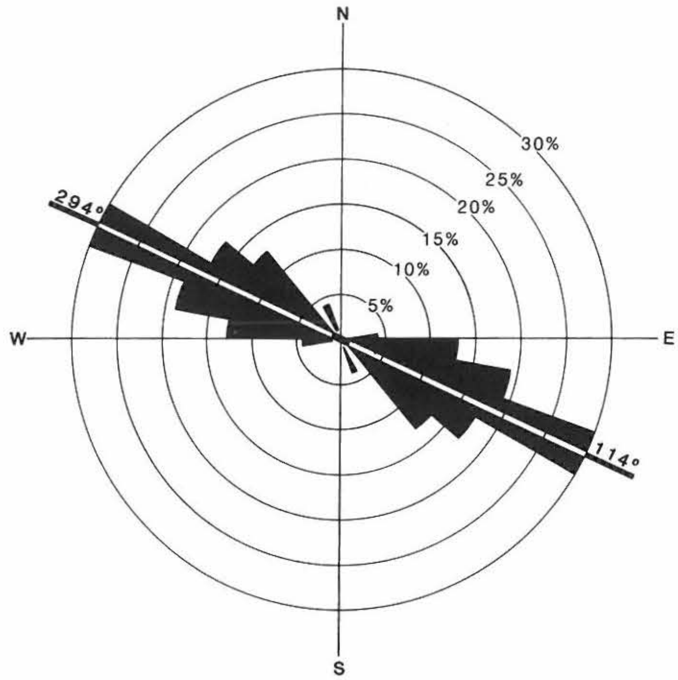

Figure 6. Rose diagram showing the long axis orientation of 47 dunes within the Anson quadrangle. Diagram is based on the summation of 10° class intervals.

wash plains and on the lower portions of the west flanks of the north-south trending chain of hills to the east of the Kennebec River (Fig. 2). While some of the wind-shadow dunes are located on the west slopes, most are on the east slopes of these same hills. Although recognized only in limited exposures, a thin blanket sand appears to have mantled most of the study area to the east of the Embden Formation and hence does not show a topographic preference.

The orientation of the long axis of most longitudinal and a few wind-shadow dunes was recorded to $\pm 2^{\circ}$; the mean orientation of these 47 dune azimuths is 294-114° with a standard deviation of 17°. The small variance of the distribution, as best shown in a rose diagram (Fig. 6), implies that the dunes were formed by a uni-directional wind rather than bi- or multi-directional winds.

Internal Characteristics

The color, general mineralogy, roundness, and average textural parameters of the dune sands are as follows. Unweathered, dry, and organic-free samples of dune sand range in color from light brownish gray (10 YR 6/2) to light yellowish brown (10 YR 6/4). In general, the sand is composed of a high proportion of quartz with lesser amounts of feldspar and other minerals, including approximately five percent heavy minerals ($\rho > 2.95$). Selected quartz grains in the 1.0 to 2.0 ϕ (0.5 - 0.25 mm) size range were found to be either subrounded or rounded, according to the Russel-Taylor-Pettijohn descriptive classification (Mueller, 1967). The roundness increases with increasing distance east from the Embden Formation. Sand samples weighing 100-150 gm and collected with a 2.5 cm diameter soil auger from the upper meter of 52 of the 64 eolian deposits were sieved mechanically using standard wire mesh sieves, 0.5 - 4.0 ϕ (0.71 mm - 0.0625 mm), at 0.25 ϕ intervals. Cumulative grain size curves were plotted for each sample. Figure 7 shows curves for three representative dunes. The mean textural parameters and descriptive equivalents (Mueller, 1967; Folk, 1968) are shown in Table 2.

The internal sedimentary structure of eolian deposits usually provides a strong indication of the effective wind direction. Exposures suitable for the study of bedding characteristics are few in the study area. However, the deposits can be described as either tabular planar bedded, trough bedded, or structureless. The most prominent bedding type, tabular planar, was observed in exposures of three dunes (Fig. 2, locations 2, 3, and 4). The beds range from 0.5 cm to 2 cm in thickness, are subparallel, laterally persistent for three or four meters, and distinguished on the basis of slight variation in grain size. Unfortunately, bedding of this type is not helpful in determining transport direction. The second type, trough bedding, is a good indicator of transport direction, but it was observed only in the upper one meter of one dune (Fig. 2, location 5), where the long axes of the troughs trend northwest-southeast. The structureless blanket sand, which has an average thickness of 0.5 m, shows ungraded, massive bedding which presents no indication of transport direction. In summary, the internal structures of the eolian deposits in the study area are not reliable indicators of wind direction at the time of deposition.

WIND-MODIFIED ROCK SURFACES

Ventifacts

Ventifacts found in the area vary in size and lithology, but their distinct textural features resemble those of ventifacts from diverse environments as described by others. A few pebbles, smaller than 4 cm in their longest direction, showed extensive modification on all sides, forming triquetrous or "brazil nut" ventifacts, a type common to the loose stones of desert pavements. Sharp (1949) maintained that "many generalizations concerning ventifacts have been made with respect to smaller stones," and that the features of larger ventifacts may differ greatly from those found on smaller ventifacts. In this study only the larger (>8 cm) ventifacts were selected for the purpose of recording the various features resulting from wind abrasion. Most ventifacts were less than 30 cm in length but the largest ventifact found was approximately 1 meter in diameter (Fig. 8d). Most of these ventifacts showed one or more of the following features.

Polished Surfaces. A few quartz clasts are highly polished to a waxy luster (Figure 8a). The quartz grains of some coarse-

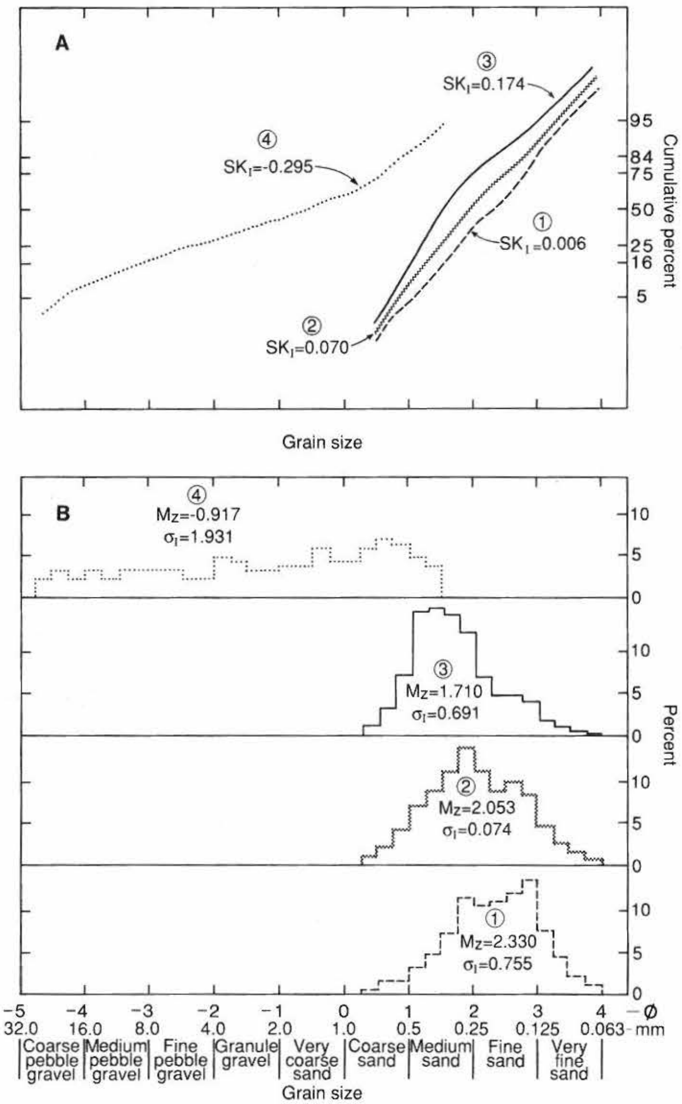

Figure 7. Grain size distribution curves and textural parameters for three representative dune sand samples from the Anson quadrangle. Sample 1 is approximately one standard deviation finer in mean size than the mean of all dunes. Sample 2 is closest in mean size to the mean of all dunes. Sample 3 is approximately one standard deviation coarser in mean size than the mean of all dunes. M_Z is mean grain size. σ_I is inclusive graphic standard deviation, SK_I is inclusive graphic skewness. Sample 4 is representative of the outwash of the Embden Formation. (a) Log probability scale, (b) Histograms in percent in each 0.25 ø class interval. Both samples 1 and 2 are described as moderately sorted, fine-grained sands that have symmetrically skewed platykurtic distributions. Sample 3 is described as a moderately well-sorted, medium-grained sand that is fine skewed and mesokurtic.

grained (>3 mm) granites and quartz veins in other clasts are similarly polished.

Smoothed Surfaces. Most fine-grained (<3 mm) ventifacts are smoothed, not polished, to a dull finish on some cusp, flute, and pit-hollow surfaces.

TABLE 2. MEAN TEXTURAL PARAMETERS AND DESCRIPTIVE EQUIVALENTS FOR ALL 52 DUNE SAND SAMPLES ANALYZED.

	Mean grain size ø	Sorting σ_I	Skewness Sk_I	Kurtosis K_G
Mean	2.048	0.757	0.105	1.024
Description	fine grained	moderately sorted	fine skewed	platykurtic
Std. Dev.	0.310	0.107	0.118	0.137
Median	1.928	0.813	0.082	0.886
Minimum	1.233	0.544	-0.212	0.474
Maximum	2.623	1.081	0.375	1.298

Facets. Less than one out of ten wind-modified clasts have facets. Of these, most clasts have only one or two facets, which are usually flat or slightly concave and only rarely display sharp interfacial edges.

Differentially Abraded Surfaces. Veins, phenocrysts, and laminae commonly project above the general rock surface (Fig. 8b and 8c). Usually these features are found best developed on one side of the clast, which was generally inclined from the local horizontal surface by 30 to 60°.

Fluted Surfaces. Many of the clasts display flutes, or subparallel, discontinuous, shallow grooves on surfaces inclined less than 30° with the present horizontal.

Pitted Surfaces. These surfaces of alternating smoothed hollows and more resistant raised areas are common among coarse-grained (>3 mm) granites. The pitted surfaces are developed through accelerated abrasion of feldspar grains (Fig. 8d).

Ventifacts with these features were found at many localities within the Carrabassett River valley and in the hills to the east of the Embden Formation in the Kennebec River valley. All ventifacts found in this study area are modified till clasts located at the contact of the Late Wisconsin till and the overlying eolian sand. Close inspection of the sand/till interface at one section (Fig. 2, location 3) shows that approximately 50 percent of the till clasts, most of which are fixed rigidly in the upper till surface, display one or more of the ventifact features as described above. There is no indication of migration by frost action of the till clasts into the sand. Examination of a small pit, dug in the till to a depth of approximately 0.6 m, suggests that the distribution of clasts is homogeneous near the surface. Thus, whereas many clasts at the surface showed signs of modification by wind-driven sand, no distinct surface concentration of clasts similar to a deflation surface was observed.

In desert environments the features of wind abrasion listed above typically may be found on all sides of non-fixed ventifacts, but in the Anson area they are found only on the north-to-west face and/or on the top of the fixed ventifacts. Some of these features, especially the fluted surfaces, are indicative of effective

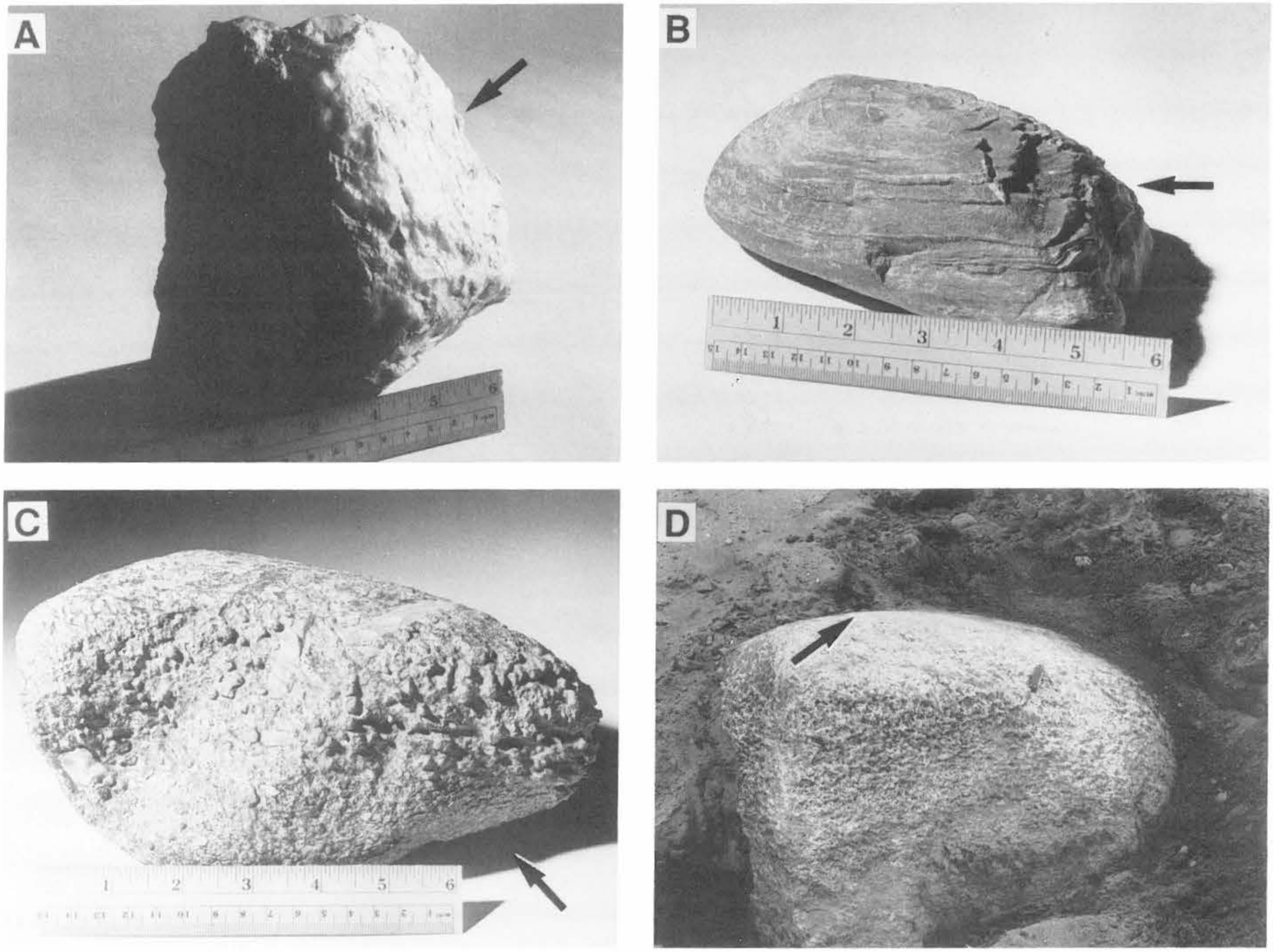

Figure 8. Ventifacts from the study area. Arrows indicate effective wind direction. (a) Highly polished quartz clast from till surface near location 6 (Fig. 2). (b-d) Differentially abraded till-clast ventifacts. (b) Quartz veins and laminae stand out on the right side, all other sides are not wind worn. (c) Resistant garnets protrude on front faces. Top surface is slightly fluted, all other sides are not wind worn. (d) Pitted granite boulder excavated from dune sand at location 6 (Fig. 2). Pocketknife provides scale.

wind direction. Measurements were made of directional features of fixed, till-clast ventifacts at three sites (Table 3).

Bedrock Surfaces

At three other sites, bedrock outcrop surfaces display wind abrasion features similar to those observed on some ventifacts. Directional trends of these features were also measured at each site (Table 3). The most modified outcrop is fine-grained, glacially smoothed, graywacke metasandstone of the Silurian Sangerville Formation (Pankiwskyj, 1979), located on the gentle northwest-facing slope of a broad hill mantled by till and eolian sand (Fig. 2, location 8). Its surface displays both numerous open-ended, one to two centimeter long flutes superposed on glacial striations and shallow grooves (Fig. 9a) and fan-like, two to three centimeter long cusps (Fig. 9b). Both flutes and cusps were found on the top and northwest side of the low outcrop, but not on the southeast side. The alignment of the trough-like flutes provides the trend of formation while the laterally asymmetrical pointed cusps, and the selective distribution on the outcrop surface of both flutes and cusps, provides the direction of formation (Table 3).

DISCUSSION

Much of the traditional literature on the geomorphology and sedimentology of eolian deposits deals with the sands and related features of modern deserts. Thanks to researchers like Niessen and Koster (1984), who assembled an annotated bibliography of periglacial sand dunes and eolian sand sheets, deposits such as those found in the Anson area now can be interpreted with more appropriate analogues.

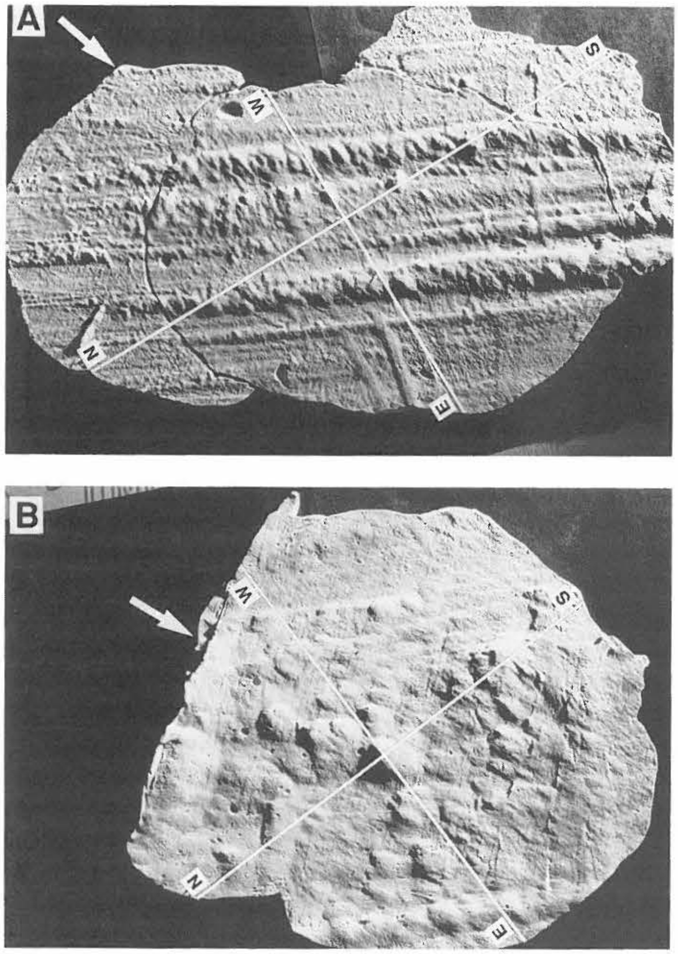

Figure 9. Plaster impressions of wind-modified bedrock surfaces at location 8 (Fig. 2), photographed with low angle illumination. At the outcrop these features are difficult to photograph. Arrow indicates effective wind direction. (a) Note flutes (WNW-ESE) are developed within glacial striations (NNW-SSE). Cast is 0.4 m in length. (b) Cusps, with vertices pointing upwind. Cast is 0.25 m in length.

Dune Orientation and Wind Direction

Dunes in the study area are classified as either longitudinal or wind-shadow according to their morphology and distribution. Effective winds are the sand-moving winds; the term effective wind direction is used here to refer to these winds. The writer contends that the effective wind direction parallels the mean orientation of the long axes of the dunes (Fig. 6). Since the majority of long-axis orientations measured were from longitudinal dunes, the assumed direct relation between longitudinal dune-axis orientation and wind direction must be verified. Verstappen (1968) reviewed several theories on the genesis of longitudinal dunes and found that "all the theories have in common the belief that a marked dominance of one wind direction is essential to their formation." Smith (1949) also noted that the elongation of longitudinal dunes represents the alignment of former winds, but warned that there is not universal agreement in this theory. For example, Bagnold (1942) and McKee (1966, 1979) believe that longitudinal dunes are oriented along the vector of two slightly divergent wind directions. Nevertheless, there is general agreement that the orientation of longitudinal dunes approximately parallels effective wind direction.

TABLE 3. AZIMUTHS OF DIRECTIONAL FEATURES ON WIND-MODIFIED ROCK SURFACES.
Features measured include cusps, flutes, interfacial ridges, and normals to facets.

Location (Figure 2)	Type of Surface	Number Measured	Mean Azimuth, by location ($\pm 2^o$)
3	Till Clasts	6	286^o
6	Till Clast	1	304^o
7	Till Clast	1	282^o
8	Bedrock	130	294^o
9	Bedrock	5	305^o
10	Bedrock	40	310^o
Mean of 6 locations			297^o

It is emphasized that most of the Anson area dunes are similar in shape and grain size, and dissimilar in size and extent, to classical longitudinal dunes. Folk (1971a), in reference to the longitudinal dunes of "vast inland deserts," listed their consistent parallelism over large areas, uniform lateral spacing, and occasionally bifurcating form, as features of this dune type. Cooper (1958) reviewed the literature on inland longitudinal dunes and described some dunes that are as much as 100 m high and continuous for 100 km. In summary, almost all research on inland longitudinal dunes has been carried out in deserts where the individual dunes are typically 10 to 50 times larger than those of the Anson area. Nevertheless, the subparallel orientation, lateral symmetry, and linearity of the dunes of the study area warrant a longitudinal classification; hence, the mean orientation of these dunes defines the trend of the effective wind.

Wind-Modified Rock Surfaces and Wind Direction

The morphology and distribution of facets, pits and cusps on ventifacts has long been recognized as being directly associated with wind direction (e.g. Sharp, 1964). Observations of modern ventifacts enabled Thiesmeyer and Digman (1942) to assert that "the longitudinal dimension of these (ventifact) furrows is generally parallel to wind direction." Within the Anson quadrangle, this relationship is also apparent since the average trend of directional features on the ventifacts (Table 3) agrees with the mean dune orientation to within 3^o (Fig. 6).

Careful measurement of the azimuth of the surface features of a till-clast ventifact should provide the effective wind direction, but it must also be established that the till-clast has not been displaced by frost action since eolian modification. It was

observed that only a few small (<8 cm) ventifacts were not fixed in the till surface at most localities. Thus, it is assumed that the majority are *in situ*. Even if some of the till-clasts had been moved post-ventifaction by frost action, the measurements may still be significant, for Harris (1969) found that while the plunge of a till clast may be affected by freeze-thaw, its trend is not.

The flutes and cusps found on bedrock surfaces in the study area are similar but less well developed than those reported by Denny (1941), Segerstrom (1962), Minard (1966), and Thornbury (1969). Table 3 shows that the orientation of wind-modified rock surface features on bedrock closely approximates that of similar features on till-clasts. Clearly, wind-modified bedrock surfaces rather than clast surfaces are preferred as fixed indicators of wind direction, yet accounts in the literature of the use of wind-modified bedrock surfaces as indicators of wind direction are comparatively few. In New England there are two reports on the use of bedrock outcrops in Massachusetts to determine late-glacial wind directions (Hartshorn, 1961; Koteff, 1964).

Provenance of Eolian Deposits

There are only three possible sources of the eolian deposits of the study area: the Presumpscot Formation (a marine silt and clay), the Late Wisconsin till, and the Embden Formation (a sandy outwash). The Presumpscot Formation clearly was not the primary source of the eolian sands for the following reasons: (1) its grain-size distribution in weight percent for sand/silt/clay is approximately 7/39/54 (Caldwell, 1959), which shows first, that too few sand size particles are present to yield such sizable eolian deposits and second, that because of the high clay content it is highly resistant to eolian erosion; (2) only one of the 64 mapped dunes rests upon its erosional surface; and (3) its surface exposure often is very limited in areas west-northwest of the dunes, especially in the Carrabassett River valley. In some cases till has been reported as a source for eolian sands (Flint, 1971; Embleton and King, 1968). If till was the primary source of the eolian deposits in the study area, then the deposits should be found on both sides of the Kennebec River valley, rather than just on the east side, and signs of deflation such as a till-clast pavement should exist at the upper till surface. Mechanical grain-size analysis of one till sample from this area showed that it is 42 percent sand (0 - 4 ϕ or 1.0 - 0.625 mm). This grain-size range includes approximately 93 percent of all eolian sand particles analyzed. Eolian deposits, including both dunes and blanket sands, cover an area east of the Kennebec River valley, within the Anson quadrangle, of approximately 100 km^2 and average at least one half meter in thickness. Thus, the minimum total volume of all eolian deposits is about 5×10^7 m^3. If the till was the primary source of these sands, then the till surface over the same area would have had to deflate, on the average, by 1.2 m to produce this volume. Not only is such extensive deflation highly unlikely, but no evidence indicating any deflation was found.

Of the three possible sources, only the Embden Formation could provide a sufficient amount of sediment of the necessary grain-size distribution to form the eolian deposits of the study area. Furthermore, other workers have reported on the derivation of similar glacio-eolian deposits from adjacent valley outwash plains both as inferred for Pleistocene age deposits (e.g. Fernald, 1960) and as observed for Holocene deposits (e.g. Black, 1951 and Péwé, 1975).

Interpretations Based on Textural Parameters

Comparison of textural parameters, computed from the cumulative grain-size curves, indicates not only the similar nature of the many sand samples from the study area, but also supports the basic assumption that the sands are of eolian origin.

The mean size of the eolian deposits of the study area (2.048ϕ or 0.242 mm) indicates clearly that these deposits cannot be interpreted as loess deposits, which are primarily silt- and clay-size deposits (mean size 0.06 to 0.02 mm) and which are transported in suspension over long distances (Reineck and Singh, 1980). The term cover sand may be appropriate for the Anson area eolian sands, since they do drape an irregular topography as in some parts of the Netherlands (Ruegg, 1983) and their suggested niveo-eolian origin (De Jong, 1967) fits the assumed periglacial depositional environment. However, the Anson sands are slightly coarser in mean size than the cover sands described either in Denmark (Kolstrup and Jorgensen, 1982) or in the Netherlands (Ruegg, 1983).

Ahlbrandt (1979) published plots of textural parameters for 506 eolian sand samples from five continents. A comparison of the mean values obtained for the Anson area (Table 2) with Ahlbrandt's plots shows that while the Anson eolian sands are coarser and more poorly sorted than the 291 coastal dune sand samples, they fall very close to the mean of the 175 inland dune sand samples in all four textural parameters (grain size, sorting, skewness, and kurtosis).

Friedman (1961) and Moiola and Weiser (1968) submit that a beach sand can be distinguished from a dune sand by plotting mean grain size versus skewness. A plot of this type suggests that the deposits sampled in the Anson quadrangle are indeed dune sands and not beach or water-worked sands (Fig. 10a). Friedman (1961) and Moiola and Weiser (1968) agree that dune sands may not be as readily distinguished from river sands as from beach sands, but they found that a combination plot of mean size and standard deviation best separates the two environments (Figure 10b).

More recent work has shown that such graphical comparison of textural parameters does not always adequately permit determination of a given sand sample's depositional environment whether it be alluvial, littoral, or eolian. Ahlbrandt and Fryberger (1981) suggest that textural data are useful primarily as complementary evidence for an eolian interpretation. The assumed eolian origin for the sands of the Anson area is not

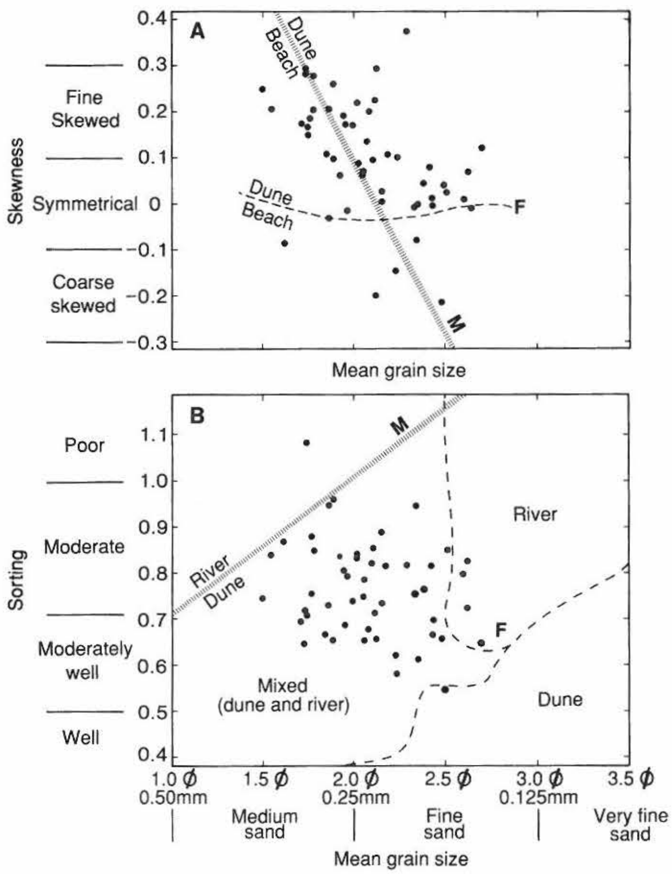

Figure 10. Combination plots of textural parameters of dune sand samples with implied environmental boundaries by others (F=Friedman, 1961, M=Moiola and Weiser, 1968). (a) Inclusive graphic skewness versus mean grain size. (b) Inclusive graphic standard deviation (sorting) versus mean grain size.

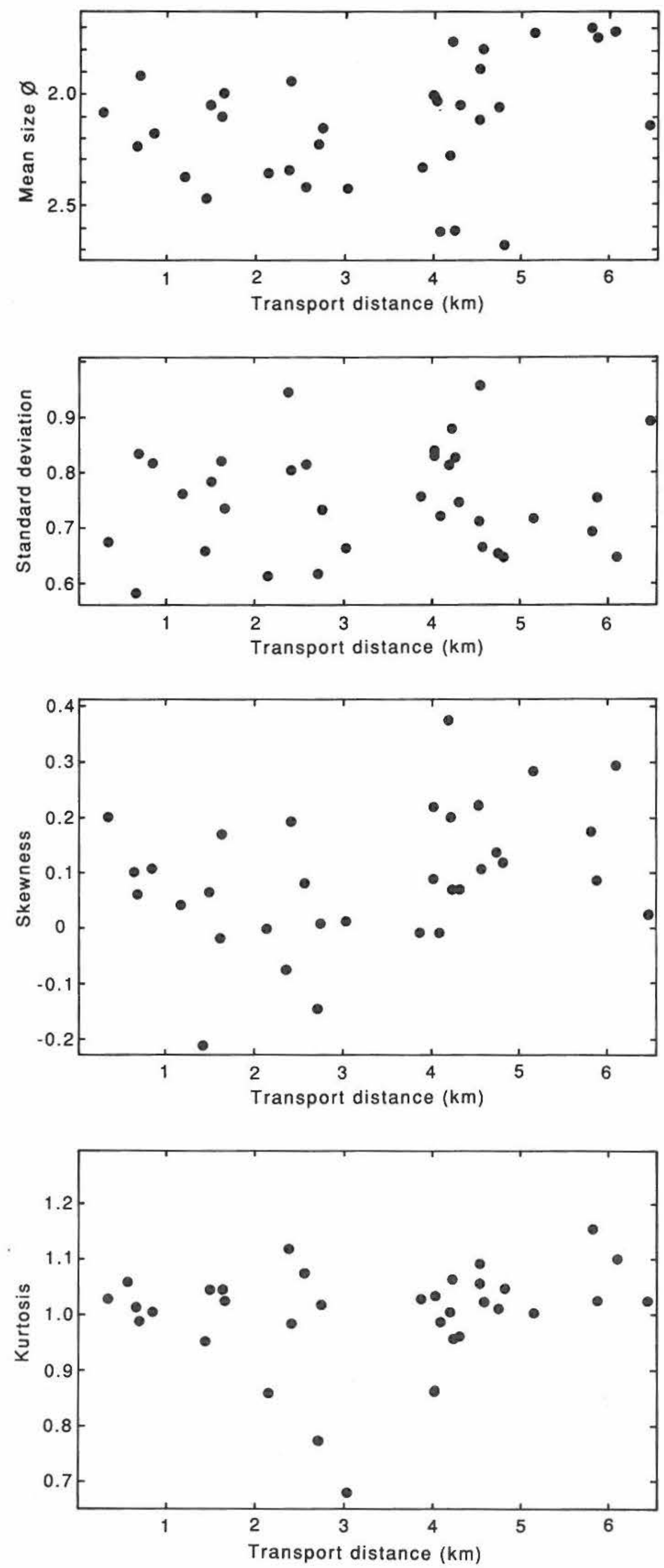

Figure 11. Plots of the four textural parameters of dune sand samples versus transport distance from the eastern limit of the outwash source.

discounted when their textural parameters are compared with published data.

The eolian sand's textural parameters could be expected to change with increasing transport in the direction defined by the mean orientation of the longitudinal dune axes and consistent positioning of the wind-modified rock surfaces. Figure 2 shows that sands of dune localities to the southeast of the Carrabassett River valley apparently have been transported up to 3 km farther from the eastern limit of the Embden Formation than the sands of dune localities to the north and south.

Plots of textural parameters versus transport distance, measured east-southeast of the Embden Formation (Fig. 11), show only poor correlations between: (1) decreasing mean size and increasing transport distance, (2) decreasing standard deviation, or improved sorting, and increasing transport distance, and (3) increasing kurtosis and increasing transport distance. Only skewness, which increases slightly with increasing transport distance, shows a relatively significant change as the result of eolian transport of up to 6 km.

Two factors may explain the apparent insensitivity of the textural parameters of the Anson sands to change with increasing transport -- too short a transport and/or imprecise sampling. Folk (1971b) has shown that there are consistent and significant differences in the textural parameters of sands of one dune sub-environment and those of the next (i.e. dune crests versus dune flanks) for the longitudinal dunes of the Simpson Desert, Australia. Generally, longitudinal dunes were sampled in this study at mid-length of their long axes near the dune crest, and wind-shadow dunes were sampled near their geometric center. The auger sampling technique used in this study samples several beds and thus smooths inter-bed textural variations and also may sample more than one dune sub-environment (Ahlbrandt and Fryberger, 1981).

Chronology of Eolian Activity

No datable material was found at the till surface below the eolian sands during the field study, so an absolute chronology of eolian activity within the study area cannot be established. However, the dates of eolian activity can be based on indirect evidence. The Embden Formation, the outwash which was the primary source of the eolian deposits, prograded into the regressing late-glacial sea in the Kennebec River valley near North Anson no earlier than approximately 12,900 yr B.P. (Borns and Hagar, 1965b). By about 12,600 yr B.P. the outwash had stopped accumulating and the early Kennebec River had begun to incise the outwash plain. It was during this interval that conditions were probably most favorable for eolian activity because of the availability of sediment.

Flint (1971) briefly summed up the periglacial eolian history typical of many areas:

> Most of the ancient dunes recorded in the literature of the Pleistocene were built in the lee of outwash bodies.... Some accumulated during outwash aggradation, probably during deglaciation; others did not form until postglacial dissection of the outwash had begun. Later both source and dunes became fully clothed with vegetation and dune building came to an end.

If dunes of the study area were built after the dissection of the Embden Formation, then the outwash plain surface might show evidence of deflation. Since no such pavement was found, it is suggested that the eolian sediments were derived from the outwash primarily while it was accumulating or approximately 12,900-12,600 B.P.

Modern and Historical Eolian Activity. An obvious indicator of modern eolian activity in the Anson area is the active, partially vegetation-free surface of some dunes. The stable dunes usually are covered by a thick humus layer and occasional pine trees. Other indicators of modern activity include the burial of wire fences and the migration of dune sand into wooded areas and across road segments (Fig. 2, locations 11, 12, and 13 respectively). Evidence of limited historical eolian activity is

Figure 12. Buried soil horizon exposed in a 6 m high section of a wind-shadow dune (Fig. 2, location 5). A radiocarbon date of charred wood fragments (arrow) immediately above the soil surface implies that this soil surface was buried by dune activity about 1700 AD.

provided by a radiocarbon date, 245 ± 60 yr B.P., or A.D. 1705, (SI-970), of a 2 cm-thick bed of charred wood chips in the uppermost layer of a soil profile developed on and buried by the sand of a single wind-shadow dune (Fig. 12).

Modern Wind Directions. The only official weather bureau station within the study area, at Madison (Fig. 2), provided a 25 year record for prevailing wind direction, by compass octants, showing the average annual wind direction as northwest (Martin, 1930). A summary of the May-to-October daily wind records of four stations of the Maine State Forestry Department, located at Canaan, Moscow, New Sharon, and Weld (all within 35 km of the study area), shows that most winds of 6 m/sec and greater velocity are from the northwest. Thus the modern mean annual direction, to the nearest compass octant, of winds capable of presently transporting the sand in the study area is also northwest. The modern and historical eolian activity is nevertheless considered to be a limited eolian activity rather than a complete erosion and redeposition of the late-glacial deposits.

It is most likely that the late-glacial influx of vegetation severely limited eolian activity. The general vegetation cover of the immediate region of the study area was generally treeless tundra from the recession of the sea (about 13,000 yr B.P.). By about 11,000 yr B.P., trees were sparsely scattered about to create a mixed woodland landscape (Bonnichsen et al., 1985). The oldest bog-bottom date in the region, from below the vegetation zone in a core from Muddy Pond near Waterville, is 12,350 ± 300 yr B.P. (Davis, 1969). There is no proof that there was a period of significant eolian activity at any other time during either the late-glacial or Holocene period in this area. If there was significant eolian activity during the Holocene, then the distinct longitudinal dunes would have to have been built in conflict with the thick forest cover that, as shown by pollen diagrams, was present throughout the Holocene. Alternatively, a regional fire could have removed the forest cover and allowed

the deposits to reactivate. However, no extensive evidence of either buried trees or organic matter, such as a soil developed on till, was found beneath the eolian sand in dune sections.

An additional piece of evidence, the absence of weathering, shows that the period of eolian activity is basically late-glacial. No exposed weathering rind was found where flutes and cusps cut bedrock striations, indicating that the bedrock was not subjected to a significant amount of weathering between retreat of the overlying ice sheet and the deposition of the overlying eolian sand.

It is instructive to compare the proposed length of this period of primary eolian activity of approximately 300 years with the length of time required to form ventifacts. Hickox (1959) found that it was possible to form small ventifacts in 10 years; however, Sharp (1964) found that medium-coarse grained granitic to dioritic clasts showed no wear in 10 years by a sand with a mean size between 2 and 1 ϕ (0.25 and 0.4 mm) in a region where the wind velocity was greater than 6 m/sec for 17 percent of the time. Kuenen (1960) has shown in laboratory experiments that with a constant wind force of 11 m/sec and sand of 2 to 1.75 ϕ (0.25 to 0.30 mm) grain size, it takes 24,000 hours to effect a six percent weight loss on a quartzite ventifact. At present the Anson area has winds of this velocity less than one percent of the time from May to October. At this rate it should take about 550 years to form such a ventifact under present wind conditions in the Anson area. However, the ventifacts of the study area were not modified to such a well developed state as those of Kuenen. A qualitative appraisal of this evidence, then, indicates that it is possible that ventifacts found in the Anson area formed in the proposed 300 year period.

The maximum intensity of eolian activity during this interval was probably in the late summer and early autumn months when the ground surface was unfrozen, relatively dry, and free of snow (Embleton and King, 1968). It was not necessary that the climate be strictly arid, but only that the upper surface of the outwash plain be relatively dry. However, the eolian activity may not have been restricted to just the summer and autumn months. The blanket sand present may be analogous to the "cover sands" described in the low countries of northwestern Europe, where they are interpreted as niveo-eolian in origin (West, 1968; Flint, 1971). If similar conditions existed in the study area, then substantial amounts of sand may have been transported in the winter months, too, and hence the west-northwest wind direction reported herein may be more than a seasonal one.

CONCLUSIONS

The data and observations as reviewed here lead to the following conclusions for the Anson quadrangle of the upper Kennebec River valley in west-central Maine: (1) the late-glacial outwash, the Embden Formation, was the primary source of the eolian sands; (2) orientation of dune axes, azimuths of directional features on wind-modified till-clast and bedrock surfaces, and geographic distribution of eolian deposits are indicative of a west-northwest effective wind direction; (3) a period of major eolian activity occurred about 12,900 to 12,600 yr B.P.; and (4) the effective late-glacial winds during this period were from approximately the same direction as at present, that is, west-northwest.

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Late Wisconsin Glacial Geology of the Eastern Portion of Mount Desert Island

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ABSTRACT

The areal extent and nature of glacial ice in the Gulf of Maine at the late Wisconsin maximum are the subject of ongoing debate. A contribution can be made to this issue by detailed mapping of glacial erosional and depositional features on the highest point on the Maine coast, Mount Desert Island. Such mapping showed (1) basal till on valley floors, (2) plucked bedrock and erosional cirques on the bedrock ridges, (3) subglacial meltwater channels and P-forms on ridges, (4) striated and polished bedrock surfaces on the highest mountain tops, (5) regional full-glacial south-southeastward ice flow yielding to late-glacial ice flow to due south, and (6) interbedded glacial and marine deposits. Evidence that the marine waters remained below 127 m elevation is found in cores of postglacial lacustrine sediments.

From this evidence the following glacial conditions are inferred: (1) basal ice melting and deposition below 90 m, (2) basal ice melting, freezing, and erosion above 90 m, and (3) meltwater saturation of the ice-bed interface. It is likely that the conditions changed over time, but that most of the erosion occurred as the ice streamed into the Gulf of Maine during the late phases of glaciation.

INTRODUCTION

With regard to glacial history, the Gulf of Maine (Fig. 1) lies in a critical position between southern New England and the Maritime Provinces of Canada. The controversy centers on the maximum extent of glaciation in the Gulf. Glacial geologists in southern New England have suggested that the last glacial ice extended well offshore into the Gulf. This is in sharp contrast to the suggestions of some glacial geologists in Canada who envision a late Wisconsin ice margin just off the present-day coast (Grant, 1977). Recent work in Nova Scotia (Stea and Finck, 1984; Wightman, 1980) shows that ice covered that area and perhaps a compromise between the minimum and maximum models is a more accurate estimate of the ice position in the Gulf of Maine. Although a direct test of this awaits detailed mapping and dating of deposits currently underwater, investigations of the highest point on the present coastline can provide insight. If this area was subjected to intense glacial erosion and was overrun by a thick ice mass, then a late Wisconsin terminal position well offshore would be favored. If, on the other hand, the highest areas were not covered, then a terminal position near the present shoreline would seem more likely. Glacial erosional features, glacial deposits, and bedrock weathering studies provide evidence that active glacial ice covered the highest point adjacent to the Gulf of Maine. The limits of this ice cover, thus, must have extended beyond the present coastline.

Shaler (1889) conducted the first systematic geologic study of Mount Desert Island. He noted that glacial erosion features indicate that the last glacial ice covered the hills, and concluded that wave erosional features extended to the summits of the island.

Later, Raisz (1929) adopted fluvial explanations for the topography of Mount Desert Island. He described its mountains as a dissected monadnock surrounded by a peneplain; the advancing glacial mass followed the existing drainage system, enlarging the valleys only. He also suggested that the asym-

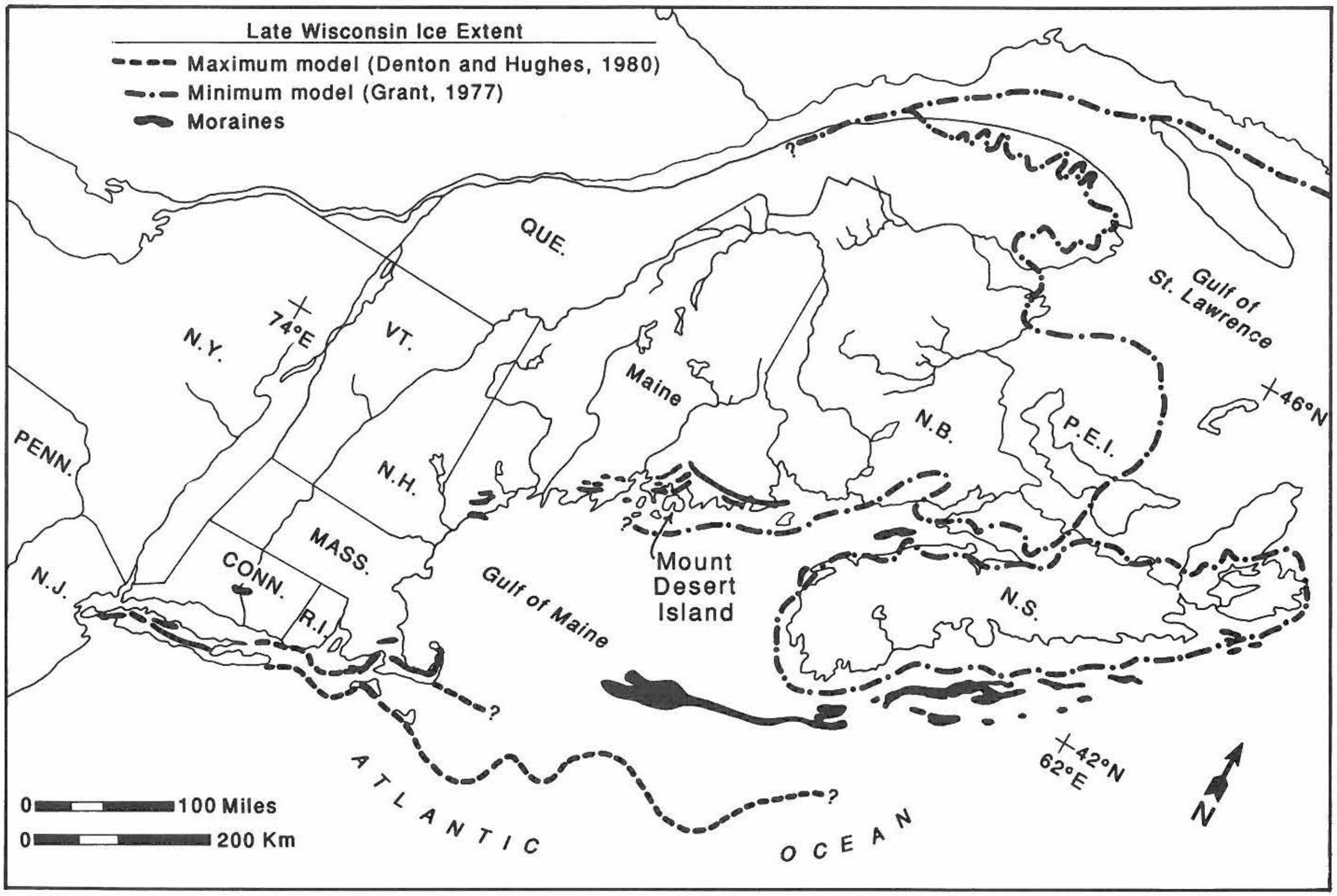

Figure 1. Mount Desert Island and its regional setting.

metrical profile of ridges on Mount Desert Island resulted from thick glaciers crossing the island to the southeast.

More recently, Chapman and Rioux (1958) studied jointing and sheeting of granite on the island. They suggested that the structure of the island developed in a predominantly periglacial environment and resulted from a mature-stage fluvial drainage system. Subsequent glacial erosion modified the trend of the valleys from N10°E to N15°W.

Field work for the present study was limited to a 25 km^2 portion of the eastern part of Mount Desert Island (Fig. 2). A granite bedrock core provides 460 m of relief through which a series of north-south trending valleys are cut. These valleys extend completely through the granite bedrock into the metamorphic rocks which surround the core (Chapman, 1974). One valley floor to the west of the study area lies 45 m below present-day sea level and forms the only fjord, Somes Sound, on the east coast of the United States. Field work for this project was conducted primarily during the summer of 1978 (Lowell, 1980). An understanding of the glacial erosion features, both large-and small-scale, and glacial deposits permits inference of former glacial conditions and ice limits.

GLACIAL EROSION FEATURES

Features of glacial erosion at several scales provide the most important set of data to reconstruct former glacial activity. Large-scale features studied here include mountain asymmetry, troughs, cirques, and meltwater channels. Small-scale features include striations and friction cracks; stoss-and-lee forms are intermediate in scale.

Large-Scale Ice-Flow Indicators

Asymmetrical mountain profiles are the largest indicators of ice-flow direction; north slopes are gentle and smooth, and south slopes are steep. Southern cliffs are 30-160 m high and consist of bedrock steps at sheeting-plane locations that are 1-5 m apart. West-to-east mountain profiles also show this asymmetry. Gentle and smooth western mountain slopes grade with a continuous, even surface to flat mountain tops, whereas the eastern slopes drop abruptly from the tops. Mountains displaying these directional imprints include Dorr Mountain, Champlain Mountain, Enoch Mountain, Halfway Mountain, The

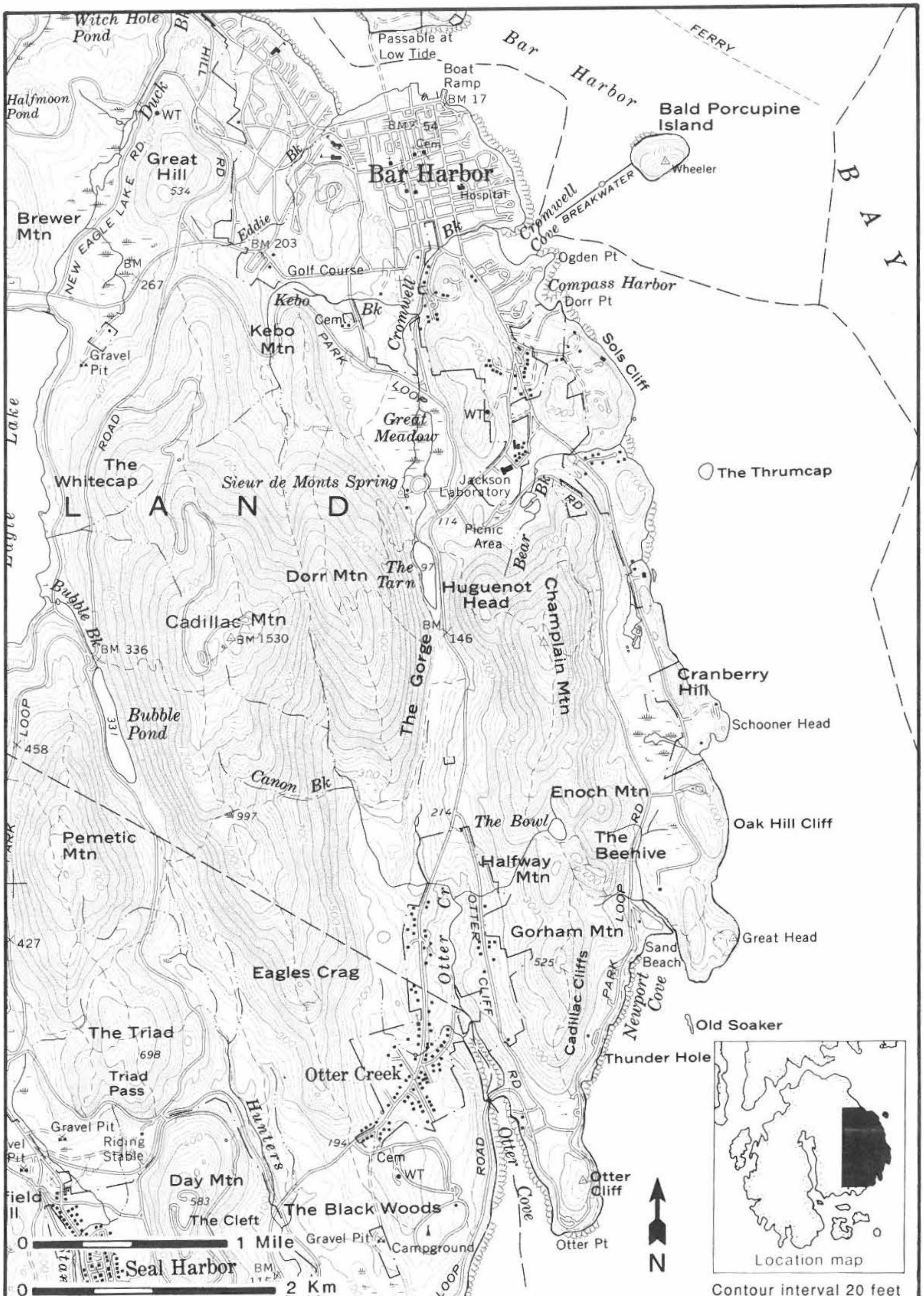

Figure 2. Topography and place names of study area.

Beehive, and Huguenot Head (Fig. 2). The last four are examples of roche moutonnées.

Stoss-and-lee forms, superimposed on the mountain profiles, are smaller but more widespread indicators of ice-flow direction. Typically 1-2 m high, these forms occur throughout the study area where bedrock sheeting and topography conditions are favorable. On the north side of Dorr Mountain (Fig. 2), a faint series of eastward trending stoss-and-lee forms are superimposed on a strong southward trending set. The large cliffs and lee faces are products of glacial plucking, whereas the smoother slopes result primarily from glacial abrasion.

Small-Scale Ice-flow Indicators

Striations provide a means of identifying former ice-flow directions and variations in those directions. In this study, distinction is made between striation direction and striation trend. Striation trend refers to the striation orientation measured with respect to north. This value always falls between 0^{o}, due north, and 180^{o}, due south. Striation direction is assigned only when field evidence permits a unique determination of ice-flow polarity. Stoss-and-lee forms 5-10 cm high are the primary field evidence used to separate the two. Age associations between multiple striations are determined through the use of geometric and crosscutting relations.

Within the study area, striations trend northwest-southeast to north-south (Fig. 3) with minor variations reflecting topographic control. For example, on Dorr Mountain striations partly wrap around the ridge parallel to ridge alignment.

At the few exposed striation locations along Otter Creek (Figs. 2, 3), striations trend parallel to valley trend, but show changing flow directions. Other examples of this phenomenon occur on the north shore of Compass Harbor (Fig. 3). Here, recent removal of unconsolidated sediment from the bedrock surface has revealed four striation sets. The oldest striation direction is 130^{o}, and subsequent directions are 155^{o}, 185^{o}, and 215^{o}. A few meters seaward of this location only the oldest, strongest striation sets of 130^{o} and 155^{o} remain; the faint, younger striations have been removed by wave erosion of the bedrock surface. In the northern part of the study area, the ice flow changed from northwest-southeast to north-south.

In addition to striations on near-horizontal surfaces, striations occur on vertical and near-vertical surfaces throughout the study area. These striations do not always trend horizontally across the vertical rock surface; many striations dip at some angle to the horizontal. For these striations, the trend recorded was the rock surface strike.

Although extensive striation mapping was not conducted on Cadillac Mountain (Fig. 3), spot checks at the summit showed striations trending northwest-southeast. This ubiquitous occurrence of striated surfaces in the study area indicates strong glacial abrasion at all elevations. This pattern requires a complete ice cover that was erosive at its base. If intense erosion can only occur some distance behind the margin of an ice sheet (Sugden and John, 1976), then Mount Desert Island must be at that distance from the former ice margin. The ice-flow patterns followed topography to some extent and became reoriented from a northwest-southeast to a north-south direction during the later stages of flow.

Friction cracks are widespread in the study area. They are best developed on the southern end of ridges aligned parallel to ice flow. For example, The Beehive lies south of Enoch Mountain (Fig. 2) and displays the highest density of friction cracks in the study area.

The friction cracks in the study area are of three different types. Crescentic gouges, comprising most of the friction cracks, range in width from 5 to 150 cm and in depth from 0.5 to 15 cm. A typical gouge measures 25 cm wide and 3 cm deep. Rare lunate fractures are smaller than crescentic gouges; typically they are 10 cm wide and 2 cm deep. Chattermarks also occur in limited numbers and range from 3 to 10 cm in width. Crescentic gouges and lunate fractures furnish information about ice-flow trend. A line bisecting the horns of these features parallels ice-flow direction, and the primary fracture dip determines the flow direction (Harris, 1943). Typical directions of inferred ice-flow range between 130^{o} and 170^{o}, with extreme values from 10^{o} to 245^{o}. At locations where friction cracks and striations occur together, the trends of both typically agree within 10^{o}, with a maximum difference of 50^{o}. Slocum (1978) reported similar observations from a detailed study on the north slope of The Beehive. There he showed that 78 percent of 151 measured friction cracks indicate ice-flow direction between 160^{o} and 190^{o}.

Glacial Troughs

The pronounced topography of Mount Desert Island exists because of a highly resistant granite core. Processes of fluvial erosion have been more effective in removing the surrounding, weaker rocks than in removing the hornblende granite (Chapman and Rioux, 1958). However, fluvial processes alone are not an adequate explanation for the high relief present today. Ten or twelve through-going valleys cut the resistant bedrock almost to, and in one case below, present-day sea level; this represents up to 400 m of relief. One example of such relief, Otter Creek valley, will be described below in detail and the evidence used to support a model of trough valley formation.

Description of Otter Creek Valley. Otter Creek valley, which extends from Otter Cove through The Gorge to the Great Meadow and Bar Harbor (Figs. 2, 4a-d), cuts deeply into the granite bedrock and lies between Huguenot Head and Champlain Mountain on the east, and Dorr and Cadillac Mountains on the west. The 7.5 km north-south profile (Fig. 4e) is low and flat with the watershed divide at an elevation of 40 m located 600 m south of The Tarn. Lowell (1980) presents a more detailed description of the valley profiles.

In an east-west direction, the valley displays different widths at different locations (Fig. 4a,b,c). At Great Meadow the

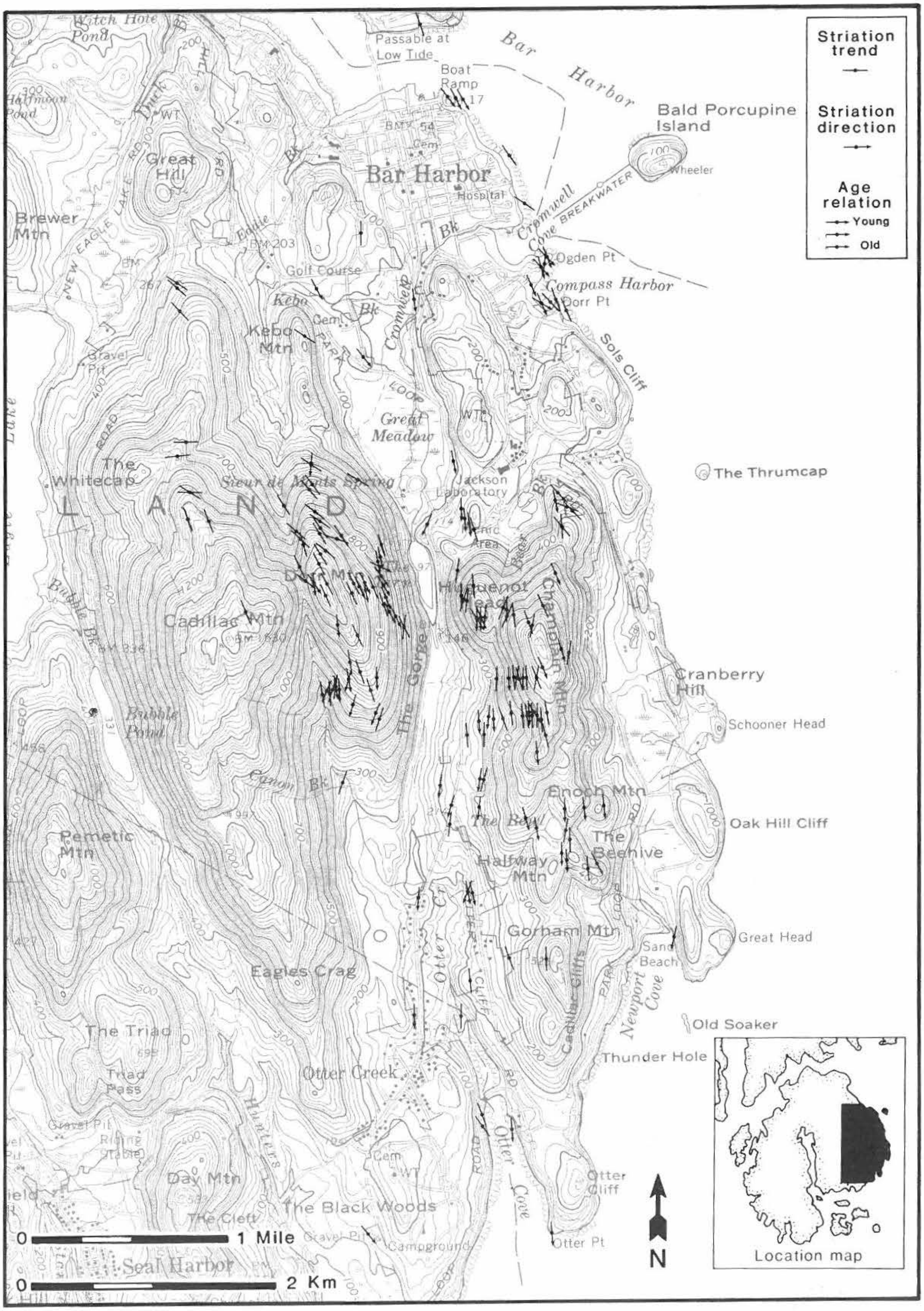

Figure 3. Striation map of study area.

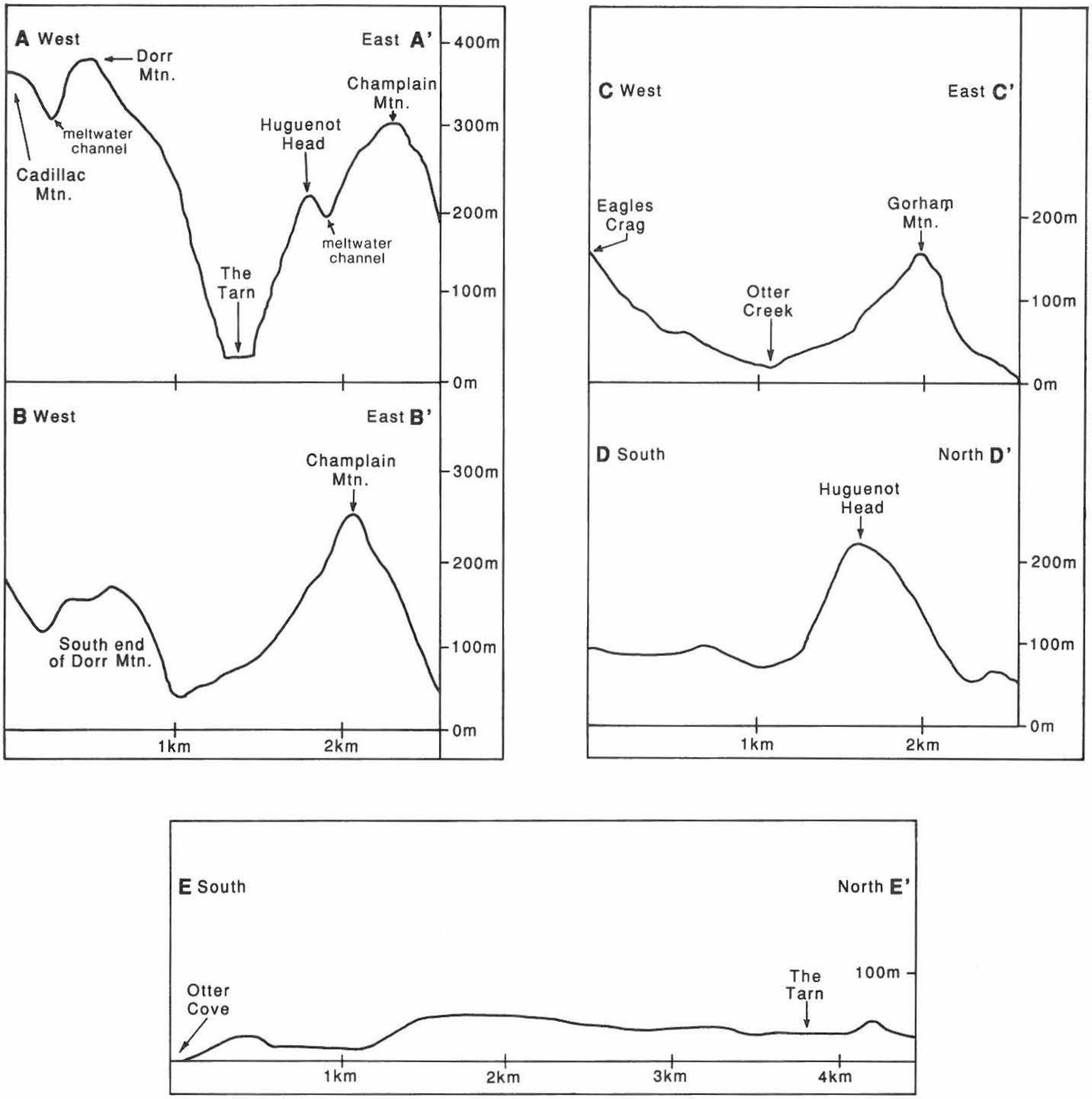

Figure 4. Cross valley (A-A', B-B', C-C') and longitudinal (D-D', E-E') profiles of Otter Creek valley. See Figure 6 for location of profiles.

valley is 900 m wide, whereas to the south at The Tarn the valley attains its narrowest width of 150 m (Fig. 4a). Farther south, the valley widens in a step-like manner, first to 600 m wide, then to 1.3 km (Fig. 4b), and finally to 1.5 km (Fig. 4c).

Otter Creek valley also varies in the shape of its cross-valley profiles. Near The Tarn, the valley cross-section shows a flat floor between steep walls (Fig. 4a). Two meltwater channels (see below) accentuate the relief, but the 2-m deep lake (The Tarn) is not in itself the cause of the flat floor. To the south, an asymmetrical profile is present (Fig. 4b). However, south of Dorr Mountain, valley sides have nearly symmetrical profiles (Fig. 4c). Only a small part of the profile near the bottom of Otter

Creek valley may result from unconsolidated deposits. From north to south in the valley, steep symmetrical walls (Fig. 4a) are replaced with asymmetrical walls (Fig. 4b), which give way to gentle symmetrical walls (Fig. 4c).

Model of Formation. Otter Creek valley, with its through valley form, glacial striations throughout, parabolic cross-section, deeply-incised bedrock, irregular floor, and open ends, matches the definition of an open trough (Sugden and John, 1976, p. 129). Open-ended glacial troughs, particularly those of Mount Desert Island, were assigned the term dorr (Chadwick, unpub. manuscript). The term is reintroduced here to describe Otter Creek valley and other open-ended troughs.

The dorrs on Mount Desert Island have been attributed to glacial modification of existing fluvial valleys. Raisz (1929, p. 140) suggested that submature to mature fluvial topography resulted from uplift and stream rejuvenation. Subsequent glacier ice flowed through fluvial saddles and eroded the valleys to their present depths. Further, Chapman and Rioux (1958) suggested that mature fluvial development produced valleys trending N10^{o}E and that glacial erosion subsequently caused a shift of the valleys to a N15^{o}W trend.

However, these theories which attribute the formation of Otter Creek valley to fluvial erosion are inadequate for several reasons. First, the small drainage area available on Mount Desert Island would not provide the water necessary to incise the valleys to their present depth. Second, fluvial erosion could not produce the flat longitudinal or U-shaped cross-valley profiles present today. Third, it is unlikely that fluvial erosion would produce several parallel, closely spaced troughs that exist across Mount Desert Island. Therefore, an alternate explanation for the overall topography of Otter Creek valley and other dorrs on Mount Desert Island is that they were formed by glacial activity.

Shaler (1889, p. 1005-1009) suggested that glacial exploitation of bedrock weaknesses caused the trough formation. Convergent ice flow in the northern portion of the valley increased glacial erosion, whereas divergent ice flow in the southern portion of the valley lessened glacial erosion (Shaler, 1889). Sugden (1968, 1974) and Sugden and John (1976) refer to this as selective erosion. In addition to ice-flow speed, selective erosion also depends on areal changes in the thermal conditions of glacial ice and bedrock structure. Glacial erosion of bedrock depends largely on quarrying activity, as abrasion is not a major factor in overall surface lowering (Boulton, 1974, p. 63). The depth of quarrying or plucking depends on exploitation of structural weaknesses in bedrock. For granite bedrock, quarrying can be quite important because the major structural weaknesses are widely spaced sheeting planes and joints. Overburden removal by quarrying sets up internal stresses that can cause dilation cracks in the bedrock thus allowing continued quarrying (Lewis, 1954; Harland, 1957; King, 1970). The process allows continued production of bedrock fractures during glaciation. This constitutes the positive feedback mechanism King (1970) envisioned.

In addition to rock structural weaknesses, glacial erosion may involve subglacial water. The effects of basal meltwater on erosion are difficult to assess, but may be considerable (Boulton, 1974, p. 63). Holtedahl (1967) demonstrated fluvial erosion at trough heads. The presence of meltwater at the glacier bed has several other important functions: (1) Confined water exerts hydrostatic pressure that controls the effective pressure of the overlying glacier ice. Effective pressure controls whether glacial erosion or glacial deposition occurs (Boulton, 1972, 1974). (2) Water freezing within bedrock fractures loosens the rock. (3) Water freezing onto basal ice incorporates loose rock into the glacier bed. (4) Water in subglacial cavities controls the stress distribution in bedrock. For subglacial cavities filled with water, overburden stress is transmitted equally to the rock, but in subglacial cavities without water, stresses are distributed unequally to the rock and fracture is likely (Boulton, 1974, p. 63).

Finally, subglacial topography influences glacial erosion. Valley floors, with limited relief, are not effective erosion locations. However, bedrock protrusions or escarpments are easily eroded because two or more bedrock surfaces are already free. Crosby (1928, p. 1170) noted that "...a glacier, if it is to do much erosive work by plucking, must have an escarpment to gnaw at or its (erosion) output will be limited mainly to impalpable rock floor..." Linton (1963) describes this as the ability of a glacier to "bite down," and uses the term to describe glaciated terrain because it is so characteristic of glaciation. One location that glaciers "gnaw at" is the down-ice (lee) side of a resistant mountain range. Erosion, first occurring on the down-ice side of the mountain range, would remove material at the escarpment site. This erosion site would migrate opposite to ice-flow direction and would eventually form a valley or trough, and for a small mountain range, erosion could cut through the entire range to produce a dorr. Formation of such a dorr might require several glaciations, such that each successive cycle of erosion would begin where the previous one left off. I propose that Otter Creek valley (Fig. 3) is a dorr resulting from selective glacial erosion of the Cadillac Mountain granite pluton.

The size of the valley reflects equilibrium conditions between ice discharge and erosion. For a given ice flow, there is a certain equilibrium size; a smaller ice discharge will have minor erosional impact. Haynes (1972) has reported this relationship in the troughs of outlet glaciers. The southern symmetrical portions of Otter Creek valley result from complete rock removal and thus complete adjustment. However, in many sections the valley profile indicates that an equilibrium size has not been reached (Fig. 4). Huguenot Head (Fig. 3) was the site of active erosion during the last glaciation and represents the trough head. The Tarn represents a smaller dorr within Otter Creek valley. Future glacier activity will erode Huguenot Head backward to produce an effective cross-sectional area that can discharge ice through Otter Creek valley. Evidence of erosion by this plucking process indicates that pressure melting conditions existed at least as high as Huguenot Head (222 m).

Some estimation of the conditions under which the intense erosion occurred can be gathered from the work of Rothlisberger and Iken (1981). They indicate that optimum conditions for plucking are sufficiently and suitably jointed bedrock, large amplitude of water-pressure fluctuations, and a high basal shear stress (for cavity formation). The granite bedrock of Mount Desert Island has abundant fractures, and the evidence of glacial plucking is common. Thus, if the mechanism suggested by Rothlisberger and Iken (1981) operated on Mount Desert Island, then we can infer that basal conditions included large amplitude water-pressure fluctuations and a high basal shear stress.

Although erosion seemingly must occur some distance back from the ice margin, the present understanding of the processes of plucking and the exact conditions present on Mount Desert Island do not allow quantification of the distance between the ice margin and the site of the plucking. However, because of the desirability of large water-pressure fluctuations and high basal shear stress, I suggest that the intense plucking described here occurred as the ice underwent accelerated flow during draw-down conditions that have been suggested for the Gulf of Maine region (Denton and Hughes, 1981). The conditions suggested by Rothlisberger and Iken (1981) are consistent with that view.

Glacial Cirques

Description. Several topographic features that exhibit the morphology of basins or theaters occur on Mount Desert Island's mountains. The basins are entirely bedrock expressions and, except for one basin, they have no associated deposits. Steep faces on south-trending ridge lines mark the backwalls of these basins and bedrock arms extend from the faces to form the sides of the basins. Although the basins have similar morphology, overall dimensions vary.

The best-developed basin lies between The Beehive and Enoch Mountain (Figs. 2, 5) and faces east. The west and north walls of the basin rise steeply from the basin floor for 55 m, and the south wall, 50 m high, grades into the gentle north slope of The Beehive. Superimposed on the steep walls are 1 to 5 m high stoss-and-lee forms oriented north-south.

Three basins of different size lie on the south ridge of Dorr Mountain (Fig. 5). The smallest is situated at an elevation of 170 m. The south ridge forms the east and north walls of the basin, and rockslide debris rests against the basin's 20 m high north wall. A second bedrock ridge forms a common boundary for the west wall of this basin and the east wall of a second, larger basin. The north wall of the second basin rises from the basin floor, and within 5 m of the wall top, bedrock steps display striations with a 150° trend as do striations north of and above the basin. A third basin contains the first and second basins; however, definition of the third basin is poor. All of these interrelated basins face south from Dorr Mountain.

Another basin on the north end of Champlain Mountain (Fig. 5) faces southeast. Similar in morphology but somewhat smaller in size than the other basins, this basin floors at 150 m above sea level. A bedrock lip and three walls enclose the 20 m by 20 m partially wet floor. In the center portion of the wall, P-forms (Sudgen and John, 1976) are etched into the wall to a height of 5 m above the floor. Along the top of the south wall, friction cracks trend southeast parallel to the basin trend. All of these basins have similar morphology, orientation and associated features, thus they likely have a similar origin.

Discussion. The significance of the above features requires a critical examination of their origin. First, however, classification of the basin is necessary. Evans and Cox (1974, p. 151) have defined the morphology of a cirque as:

> A hollow open downstream but bounded upstream by the crest of a steep slope (headwall), which is arcuate in plan around a more gently sloping floor. It is "glacial" if the floor has been affected by glacial erosion while part of the headwall has developed subaerially, and a drainage divide was located sufficiently close to the top of the headwall for little or none of the ice that fashioned the cirque to have flowed in from outside.

The term cirque commonly implies a present or former cirque glacier (Haynes, 1968; Flint, 1971; Davies, 1972; Embleton and King, 1975; Sudgen and John, 1976, p. 199). Raisz (1929, p. 159) suggested a multi-stage glacial sequence for formation of the Amphitheater on Penobscot Mountain west of the study area. Although the morphology of the basins in the study fit the definition of a cirque, there is no evidence to indicate subaerial development. Therefore, the basins may be called cirques, but the application of the term glacial does not apply as defined above because of the lack of subaerial headwall development.

Except for the basin near The Beehive (Fig. 5), the small size of the basins in the study area casts some doubt on their having been sites of active cirque glaciers. Moreover, all these basins face south or southeast at relatively low elevations. Thus they are not likely sites to produce glacier ice. The lack of similar features facing northward suggests climate did not control their development. Therefore, an alternative for the cirque glacier hypothesis is needed.

The presence of glacial erosional features such as striations and friction cracks in, near, and above the basins indicate active erosion at these locations during the last glaciation. Furthermore, the presence of P-forms on one basin wall indicates the presence of subglacial meltwater (Dahl, 1965; Holtedahl, 1967) which aids glacial quarrying. The orientation of these basins allowed removal of rock in the down-glacier direction. Ice, flowing at a full-bodied stage, would conform to the topography of the bedrock ridges, enclosing the ridge without a cavity. As the ice moved past the southern end of a ridge, a tensile force would be created at the ice-bedrock interface (T. J. Hughes, pers. commun., 1979). This tensile force could remove loose portions of bedrock from the ridge. As the tensile force is concentrated at that position, bedrock erosion would also be concentrated, causing the formation of an indentation or basin on the southern

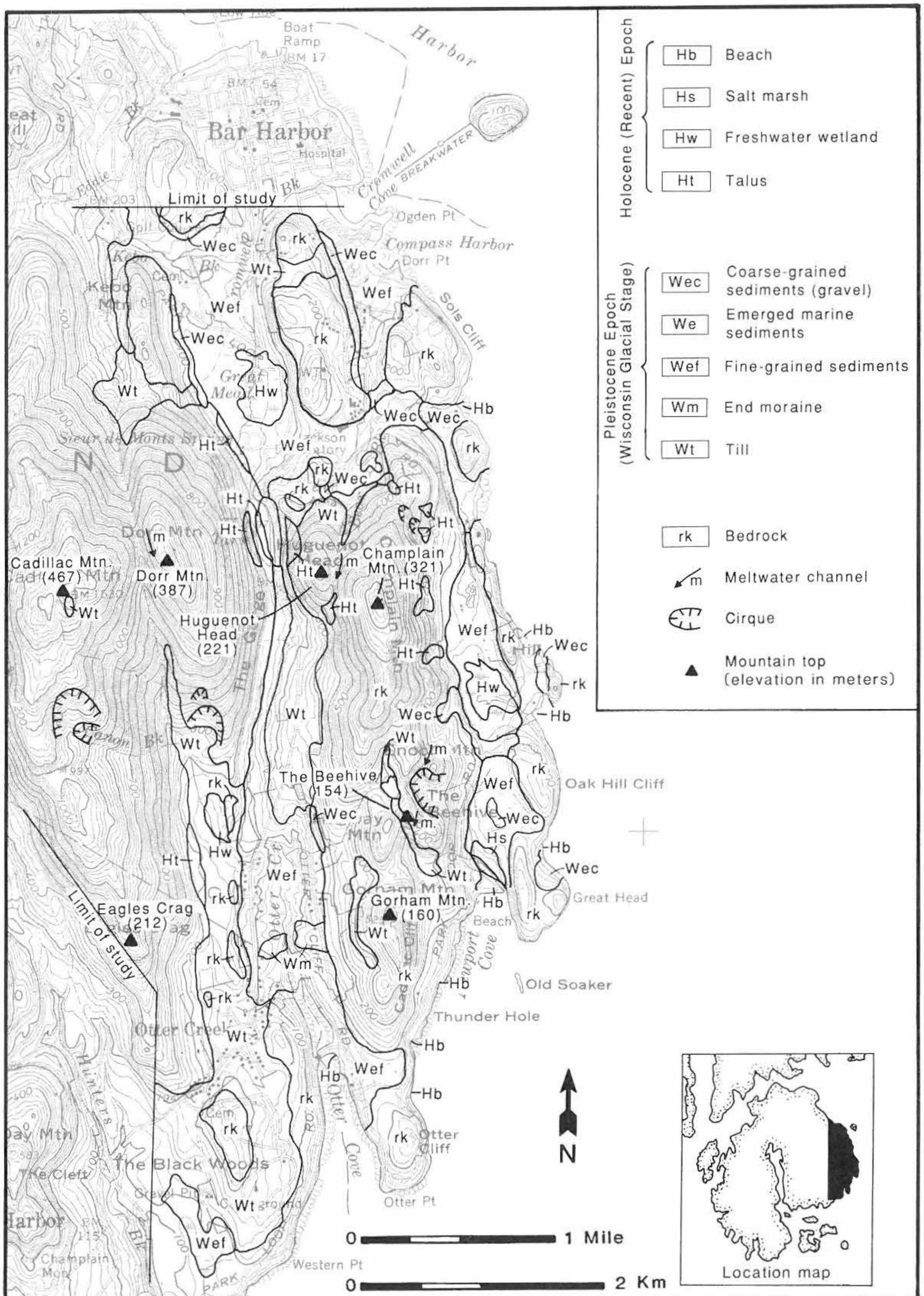

Figure 5. Surficial geologic map of the study area.

or lee ends of the ridges. Therefore, since an alpine glacier origin is unlikely, I conclude that the cirques in the study area are glacially produced beneath an ice sheet where pressure melting and plucking could occur.

Glacial Meltwater Channels

Description. Two major and two minor gorges are cut into bedrock at high elevations between ridges of the study area. The largest gorge, located between Cadillac and Dorr Mountains (Figs. 2, 5), displays a striated and polished east wall above an elevation of 322 m. Below this elevation, bedrock surfaces are either fresh and angular where rockfall and rockslide are active or it is smooth and undulating. The mass wasting activity produces an accumulation that partly covers the gorge floor.

Below the debris is an unknown thickness of diamicton with a flat east-west upper surface. The diamicton floor drops to the north or south from the central portion of the gorge floor. Debris is also accumulating against the west wall 150 m west of the east wall. Fresh, angular surfaces extend to the top of the west wall where the surface is smooth and planar. Just above the top of the wall a 200 m wide platform extends west to the western summit of Cadillac Mountain (Fig. 2).

A similar but smaller gorge lies between Huguenot Head and Champlain Mountain (Fig. 2). Both possess striated and polished surfaces adjoining smooth, undulating water-worn surfaces, active rockslide and rockfall, flat diamicton-covered floors, and platforms situated to the west of the gorges.

Two smaller gorges, similar in form to each other, occur near The Beehive (Fig. 5); the first, located between The Beehive and Halfway Mountain, has a floor that is 100 m wide at 135 m elevation. Striated and polished bedrock walls extend only 3 to 4 m above the floor, marking the vertical extent of the gorge. The diamicton cover along the floor of the gorge extends north into the depression containing the pond called The Bowl (Fig. 2) and extends south into a V-shaped continuation of the gorge. Within the continuation, an underfit stream erodes diamicton. The second gorge lies between the east side of Champlain Mountain and Enoch Mountain (Figs. 2, 5). Vertical gorge walls trend southeast in the northern part of the gorge. However, within 40 m to the south, the walls trend south. The floor of the gorge is at 110 m elevation and is diamicton covered. The southern end of the gorge terminates abruptly at the cirque north of The Beehive. The gorge opens into the top portion of the north wall of the cirque. These two small gorges both possess diamicton-covered floors and bare rock walls similar in morphology to the larger gorges described above.

Discussion. Derbyshire (1962) classified three major types of glacial drainage channels on a genetic basis: marginal, submarginal, and subglacial. Some characteristics of subglacial channels include: flat diamicton-covered floors, steep ice-molded walls, occasional gradient reversals, and irregular longitudinal profiles (Sissons, 1960, 1961; Derbyshire, 1962; Sugden and John, 1976). The gorges on Mount Desert Island exhibit flat diamicton-covered floors, oversteepened walls that allow rockslide activity, striated and polished surfaces, and smooth and undulating surfaces. Therefore, I conclude that these gorges are subglacial meltwater channels.

The transition between striated and polished surfaces, and smooth undulating surfaces produced by fluvial erosion, marks the boundary between active ice and water. In order for the transition to be situated on the channel walls, the ice and water must both be present simultaneously; this implies active ice. Furthermore, in order for diamicton deposits to remain on the channel floors, its deposition must have occurred during late ice dissipation to have avoided erosion by glacial meltwater. Therefore, final ice dissipation probably involved a thin, active ice cover. Since the meltwater must have flowed down to the 322 m elevation, the ice thickness must have been higher than that elevation.

SURFICIAL GLACIAL DEPOSITS

The surficial glacial deposits (Fig. 5) have a depositional pattern reflecting the basal glacial conditions and the subsequent marine invasion and deglaciation. In the sections that follow, the term till is used since the diamictons here are all interpreted to be glacial in origin.

Till

Areal Occurrence. The most extensive surficial glacial deposit in the study area is till, and the distribution of till is largely dependent on elevation and topography. Below elevations of 90 m, a nearly continuous till cover lies on the floor of Otter Creek valley (Fig. 5), whereas above this level till is rare and exists only as small patches in depressions. The higher till patches vary from a few square meters in size to areas as large as the depression containing The Bowl (Fig. 5). Although a continuous till cover does not exist at higher elevations, scattered erratic pebbles, cobbles, and boulders are present.

Till in the study area generally forms a blanket cover with little or no surface morphology. The present till morphology is erosional, not depositional. Intermittent streams, with channels that originate in zones of bedrock weakness, have cut through the till blanket separating it into discrete lobes. Its thickness is typically 3-4 m and reflects the underlying bedrock topography; bedrock knobs ranging from 10 to 100 m long protrude through the thin till cover. The lobes do not extend completely across the valley; a break-in-slope near 50 m elevation marks the lower end of the till lobes. At this location, emerged marine sediments overlie the till deposits.

The only location where the till forms independent topography is near Otter Cove where it forms a moraine (Fig. 5). Fine-grained emerged marine deposits (see below) also overlay this moraine which has stratified lenses intermixed with the till.

Description. To obtain representative samples of the till, five trenches were dug along existing scarps in Otter Creek

valley (Fig. 6); Lowell (1980) provides a complete description of trench location, stratigraphy, and sample information. In summary, these show that a textural change occurs between different locations in the valley. Till exposed in the north or central portion of the valley is compact to very compact, light olive to olive color, and silty to sandy in texture. Average grain-size distribution for this till is 10% gravel, 36% sand, and 54% silt and clay. However, till exposed in the southern portion of the valley is loose, light olive to tan in color, and sandy in texture. Average grain-size distribution for this till is 20% gravel, 49% sand, and 31% silt and clay.

Till fabric (TF) analyses were performed in five of the trenches (Lowell, 1980). Six till fabrics, obtained by measuring 50 stones with a minimum long (a-axis) to intermittent axis ratio of 2:1, are plotted as rose diagrams (Fig. 6). Whereas TF-6 was taken from a horizontal face to verify sampling procedure, all other measurements were taken from vertical faces.

TF-1, from 1.8 m depth, shows a random or weak bimodal fabric, whereas TF-2, from 3.5 m depth, shows a strong preferred east-west orientation. TF-4 and TF-6, from a depth of 1.4 to 1.6 m, both show a northeast-southwest trend; these two fabrics come from 1 m above a bedrock surface that bears striations trending 22^{o}.

Samples of 100 randomly selected cobbles and pebbles from trenches 2-5, and of 25 stones each from other exposures constitute the clast data (Fig. 7). For each stone the following observations were recorded: length of three mutually perpendicular axes (a,b,c), roundness, presence or absence of fracture, presence or absence of striations, striation pattern, and clast lithology. The b:a axis ratio and the c:b axis ratio were calculated; these ratios allowed placement on a Zingg (1935) diagram, which gave a shape classification of each stone. The clast analysis provides insight into the origin of the till.

The total clast sample (650 stones) shows the following trends (Fig. 7). Disk-shaped clasts are slightly more numerous than spherically shaped clasts (38% compared to 28%). Clast roundness shows a normal distribution with a slight preference toward rounded clasts. Fractured clasts, those displaying at least one angular surface, represent 85% of the sample population. Striated clasts, representing 29% of the total sample, are further classified according to striation pattern (random, 39%; sub-random, 30%; sub-parallel, 20%; and parallel, 6%). Classification of the till clasts by lithology is based on mapped bedrock units (Chapman, 1974). Gabbro and diorite rocks comprise 43%; hornblende granite 18%; the metasedimentary Bar Harbor Formation 14%; and the metamorphic Ellsworth Formation 12% of the sample clasts.

Discussion. The nature of the till deposits suggests a subglacial origin. The oriented fabrics suggest a change from compressive conditions in the center of the valley, which produced transverse fabrics, to extensional conditions in the southern portion of the valley, with fabrics parallel to ice flow. This agrees with the stress conditions necessary for effective erosion in the trough. The shape, striated surface, and fractured portions of the clasts also indicate a subglacial origin for the till (Boulton, 1978). The change in texture reflects active entrainment and deposition.

Given this origin, we can infer some subglacial conditions based on the areal distribution of the till. Since the primary sites of deposition are located below 90 m above sea level and sites above that elevation are generally till free, that elevation may mark a transition in basal conditions. At the lower elevations, the thicker ice is more likely to be the site of entrainment and deposition. The higher elevations are erosion and entrainment sites only. The distribution of erosion features supports this concept (Fig. 8).

Emerged Glacial Marine Deposits

Emerged glacial marine deposits are the youngest glacial deposits in the study area, and their distribution is restricted to the floor of Otter Creek valley (Fig. 5). This study has noted the distribution and field description of these deposits, but no attempt was made to investigate, in detail, the late-glacial marine chronology and its extent.

Areal Distribution. Emerged marine deposits are primarily restricted to the lower portions of Otter Creek valley. An eroded bench in a 10 m thick deposit near 85 m elevation on Canon Brook (Fig. 2) may mark the upper limit of the last marine invasion in Otter Creek valley. Along the floor of the valley, emerged marine deposits form a nearly continuous cover from Otter Cove in the south to Compass Harbor in the north (Fig. 5); this gives Otter Creek valley its flat floor. Increased relief in the southern portion of the valley results from active gully erosion of up to 10 m in the emerged marine deposits. In most trenches and test pits, the thickness of the marine deposits is less than 50 cm. The varying thickness of the emerged marine deposits results from, as well as influences, topography.

Physical Description. Emerged marine deposits exhibit a wide range of textures. Commonly, the deposits are light tan to olive in color, silt-sized, compact, and massive. Grain-size analyses of the sediment show that the material is 3 percent sand, 87 percent silt, and 10 percent clay. This silt locally contains dropstones up to 20 cm in diameter. However, other exposures of the emerged marine sediments show textures ranging from fine sands to boulder deposits. For example, trench 5 contains a well-sorted pebble-cobble gravel and well-sorted coarse sand over a stony till unit.

Relation to Other Deposits and Discussion. The stratigraphic relationship between emerged marine deposits and glacial till deposits varies. In many localities the marine deposits lie above the tills. However, in trench 4 in the moraine in Otter Creek valley (Fig. 6), marine silt and clay are interbedded with a loose sandy till. Furthermore, folded clay beds, isolated, tilted clay stringers and clay balls, and abrupt unit contacts indicate glaciotectonic contortion of marine sediments. Near the junction of Route 3 and Otter Cliff Road (Fig. 2) an abandoned borrow pit shows marine silt and clay at the base of the pit under a loose sandy till. This marine unit extends south from the pit to

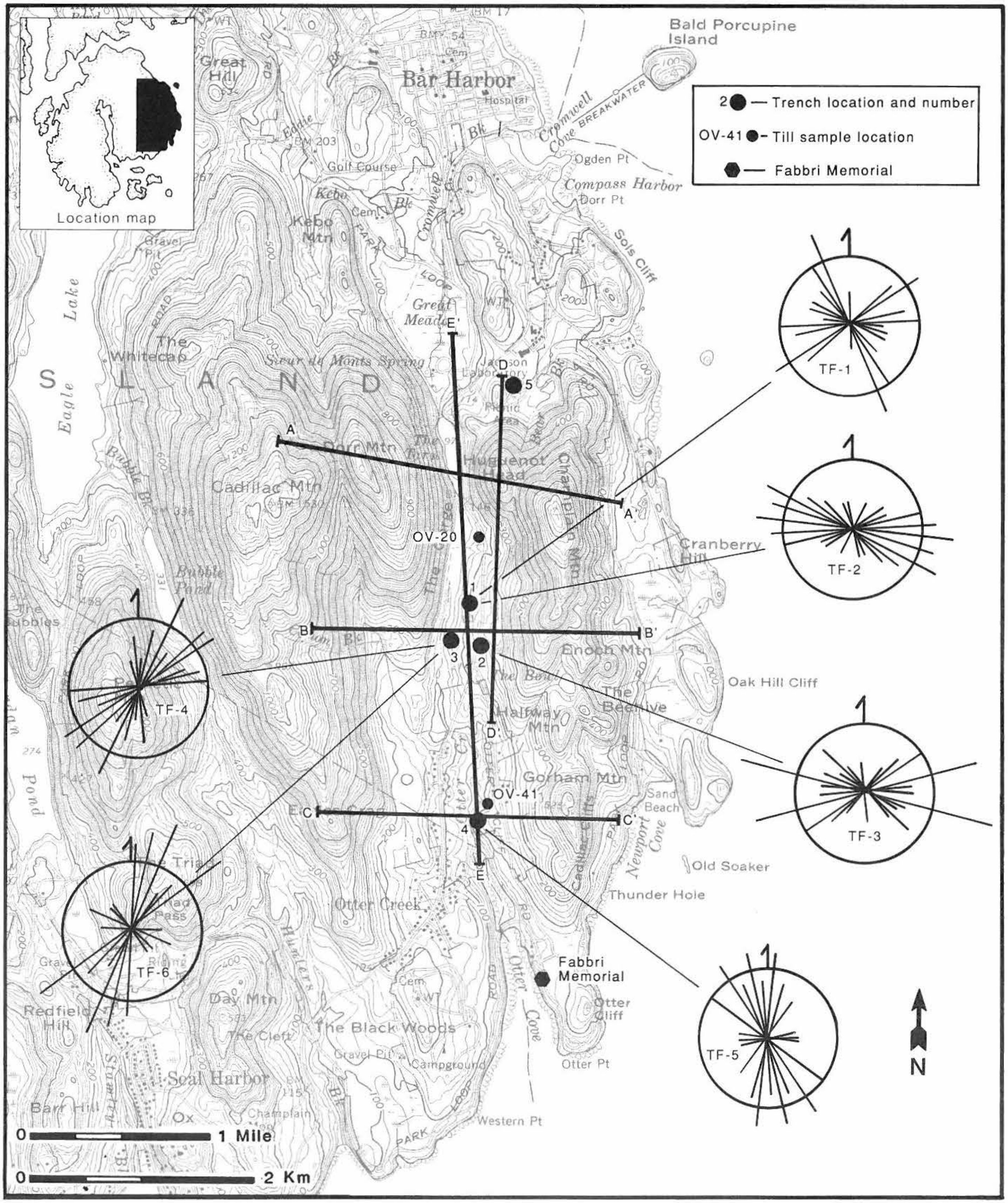

Figure 6. Locations of till fabrics (TF) and trenches. Circle on rose diagrams represents 10% of count (n=50). Heavy lines are locations of profiles shown in Figure 4.

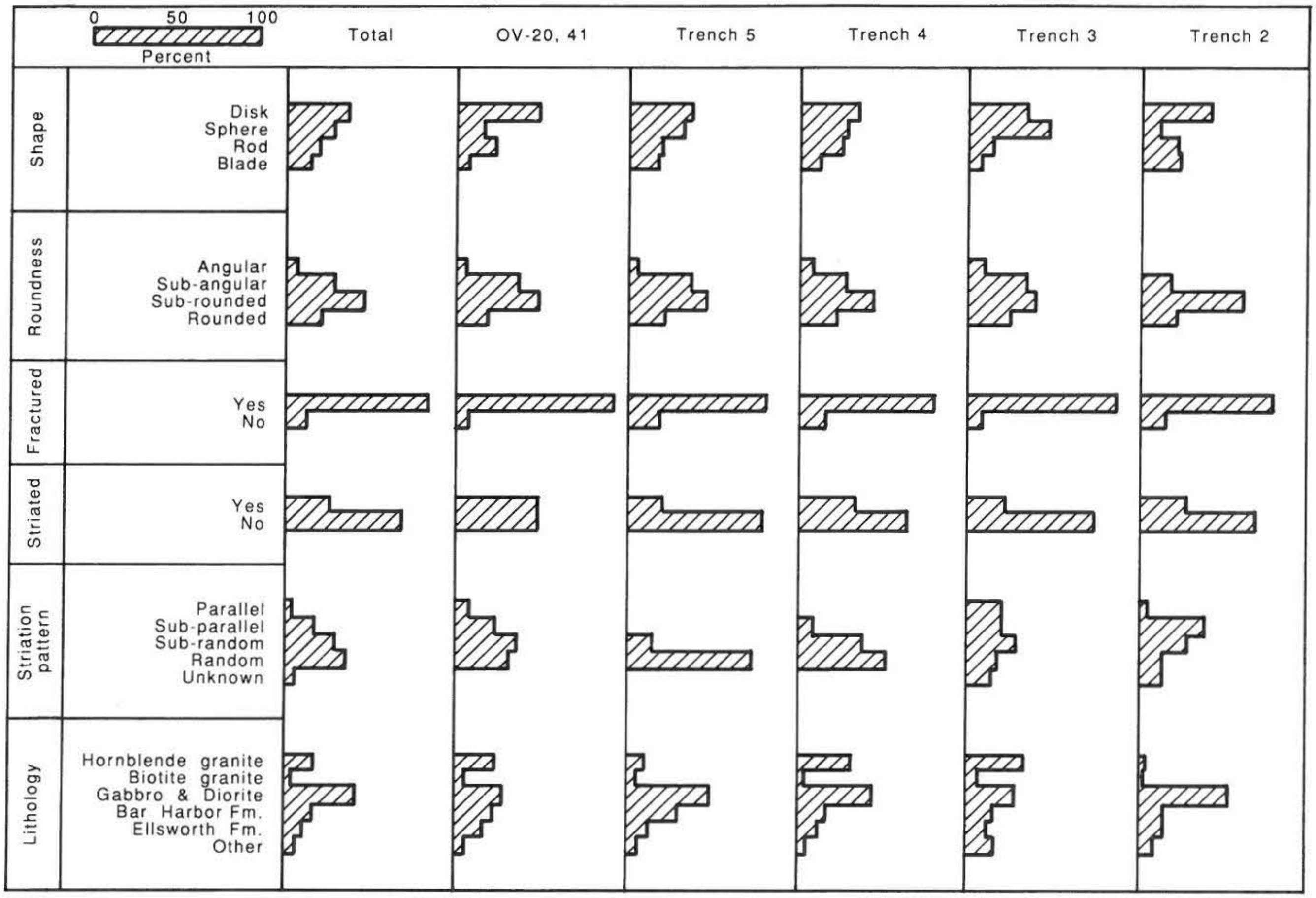

Figure 7. Clast analysis of various till samples from study area.

become the surface unit to Otter Cove, and it is the same unit comprising the uppermost material in trench 4. Marine sediments both over and incorporated into till indicate that till deposition occurred contemporaneously with the marine sediments. The inclusion of clay in till and the areal distribution of the deposits suggest a slight readvance of glacier ice into the marine waters.

An exposure in Otter Cove (Fig. 2), located at Fabbri Memorial (Fig. 6) illustrates the stratigraphic relationship. The uppermost unit of imbricated cobble gravel overlies a compact clayey silt that contains dropstones. The clayey silt grades downward to a silty sand resting conformably on sandy till. Slump deposits extend from the exposed till to striated bedrock 6 m lower. This section displays emerged marine deposits overlying till and striated bedrock and reflects the general sequence of events.

Postglacial Lacustrine Sediments

Description of The Bowl Core. The Bowl (Fig. 2), which has a surface elevation of 127 m, was cored to provide chronological control on deglaciation and to provide a maximum elevation of marine submergence. Drainage into the pond is from the north along a bedrock gully containing till that extends under the pond.

Sounding of water and probing of the sediment showed that bedrock topography and sediment accumulation control the lake-bottom configuration. The northwest portion of The Bowl, although adjacent to steep bedrock slopes, is shallow because of accumulation of sediment derived from the inlet stream. The deepest location sounded (10.7 m of water) is east of the inlet and 55 m from the north shore to the east of the inlet location; this was the coring site.

At the coring site, a three-section 7.5 m core was obtained with a 5 cm modified Davis piston corer through pond ice, and retained in the aluminum core tubes until extruded in the laboratory. The core stratigraphy is described below from bottom to top. Dents in the core tube suggest that refusal occurred on a stone rather than on bedrock; therefore, the total depth of sediments overlying bedrock is unknown, but it is at least 7.5 m. The lowest material, which is interpreted as till, underlies a gray sandy-silty material, above which lies a sand unit that contains rare fresh-water diatoms of the genera *Stauroneis*, *Tabellaria*, and *Frustulia*. The sand unit underlies a mixture of sandy gravel

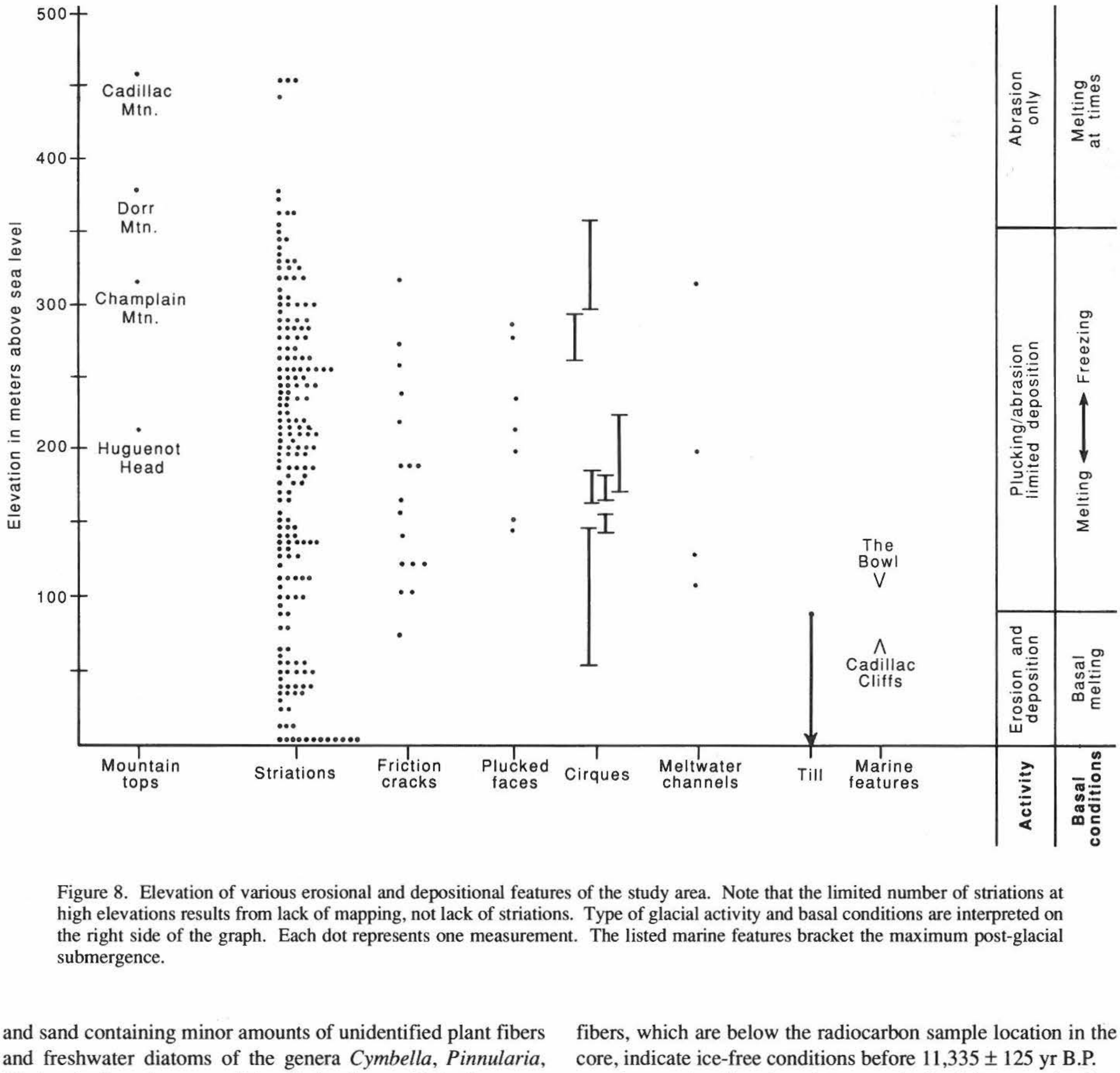

Figure 8. Elevation of various erosional and depositional features of the study area. Note that the limited number of striations at high elevations results from lack of mapping, not lack of striations. Type of glacial activity and basal conditions are interpreted on the right side of the graph. Each dot represents one measurement. The listed marine features bracket the maximum post-glacial submergence.

and sand containing minor amounts of unidentified plant fibers and freshwater diatoms of the genera *Cymbella*, *Pinnularia*, *Diploneis*, *Gyrosignma*, and *Fragilaria*. A gray silt that becomes darker and laminated away from the base overlies the sandy gravel and sand. Freshwater diatoms of the genera *Cymbella*, *Fragilaria*, *Gomphonema*, and *Pinnularia* were identified in a sediment sample taken from below the lowest distinct laminations within the gray silt unit. Within the laminated gray silt unit is a 2 cm thick zone of plant fibers that underlies gray-tan silts that grade upward into organic gyttja.

Radiocarbon Age and Discussion. In order to obtain a radiocarbon date, 10 cm of material was removed from the lowest organic-rich (6%) layer of The Bowl core. The date (11,335 ± 125 yr B.P.; SI-4043) reflects the beginning of significant postglacial organic accumulation. Diatoms and plant fibers, which are below the radiocarbon sample location in the core, indicate ice-free conditions before 11,335 ± 125 yr B.P.

A second radiocarbon date was obtained from nearby Sargent Mountain Pond. Sargent Mountain Pond, at 355 m elevation, lies between Penobscot Mountain and Sargent Mountain, 5 km due west of The Bowl. During coring, it was discovered that the coring apparatus could not cut through a lake-bottom mat of aquatic moss; as pressure was applied, the moss mat pushed through the soft underlying sediments and acted as a strainer which allowed disrupted sediment into the core tube. At a depth where the underlying sediment was firm enough to resist the moss, the core tube cut through the moss and retrieval of undisturbed gravel was possible. The depth of recovered gravel was 9.8 m below the water-sediment interface in 3.5 m of water. The total carbon of the gravel sample was used for radiocarbon dating

and yielded a date of 13,230 ± 360 yr B.P. (SI-4042). The large errors resulted from the low (~1%) carbon content (Robert Stuckenrath, written commun., 1979). This date only constrains the glacial history by showing that ice must have covered this elevation until at least 13.2 ka (i.e. subaerial conditions could not have existed at 21 ka). A third radiocarbon date of 12,250 ± 160 (Y-2241; Stuiver and Borns, 1975) from the west side of Mount Desert Island dates a portion of the marine invasion.

A consideration of these dates suggests the following chronology for Mount Desert Island. Sometime prior to 13.2 ka, the highest portion of the island became exposed. Subsequent to this, around 12.2 ka, marine waters rose against the island and left deposits. From the spatial relationship of these deposits it seems likely that the marine waters removed all the remaining ice from the valley floors. Significant organic accumulation in the lacustrine environment lagged the marine invasion on the order of 1 ka.

The Bowl core has a second important aspect. The exclusive occurrence of freshwater diatoms indicates that the marine invasion was restricted to elevations below 127 m.

SIGNIFICANCE OF THE EVIDENCE FROM MOUNT DESERT ISLAND

From the nature of glacial erosion features and glacial deposits on Mount Desert Island, I conclude that a late Wisconsin ice sheet overran the island. The absence of weathering zones or tors on higher mountains, as well as the presence of freshly striated bedrock and plucked faces near the summit of Cadillac Mountain, indicates that ice covered Cadillac Mountain to sufficient thickness to produce erosional features and striated bedrock surfaces at all altitudes. Bedrock quarrying occurred as meltwater locally froze onto the glacier base. The age of this ice cover is assigned to the late Wisconsin because postglacial lacustrine sediments yield dates of 11,355 ± 125 (SI-4043) and 13,250 ± 360 (SI-4042) yr B.P., and because detailed bedrock surface studies (Lowell, 1980) indicate no differences in length of bedrock weathering since the last ice cover. These dates bracket deglaciation at Mount Desert Island only. The available evidence does not indicate when this glacial cover began.

Specific late Wisconsin ice conditions deduced from the field evidence are: basal melting in the valley bottoms that allowed deposition of basal till, a melted ice base over the ridges that allowed localized freezing and quarrying, and ice covering Cadillac Mountain and filling adjacent bays to a thickness of at least 600 m (Fig. 8).

As eustatic sea level rose, sea water rapidly removed ice from the valleys through tidal flushing within the valley confines. Interbedded marine and glacial sediments indicate minor fluctuations of glacier ice into marine waters during deglaciation. Maximum marine submergence must must have been less than 127 m. Final deglaciation of Mount Desert Island was controlled by a marine transgression.

ACKNOWLEDGMENTS

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Late Quaternary Glacial History of Mt. Katahdin and the Nunatak Hypothesis

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ABSTRACT

Most workers agree that Mt. Katahdin was covered by ice at some time, but no indisputable evidence exists to support the common notion that cirque glaciers postdated continental ice recession from Mt. Katahdin. Erratics of northern provenance near its summit indicate that Mt. Katahdin was overridden by continental ice centered north of the mountain. These unweathered erratics, as well as generally unweathered bedrock and weakly developed soils on the Table Land, suggest that upland areas were covered by continental ice during the late Wisconsinan. Theoretical ice-surface profiles constructed along flow lines that lead inland from a dated ice-front position near the present coastline also suggest that overriding occurred during late Wisconsinan time. Roches moutonnées with chattermarked stoss sides facing down-cirque and ragged sides facing up-cirque, along with till of a northern provenance on the Northwest Basin floor, suggest that erosion and deposition by continental ice were the last late Wisconsinan events that occurred on Mt. Katahdin. The lack of end moraines on all cirque floors suggests that cirque glaciers did not reform following ice sheet recession. Lateral moraines on the east and south flanks of Mt. Katahdin were formed by continental ice during the final stages of deglaciation when Mt. Katahdin became a nunatak. The nunatak hypothesis suggests that some life forms survived late Quaternary glaciations on nunataks, refuges not completely covered by ice. However, for at least part of the late Wisconsinan, Mt. Katahdin was probably completely inundated by an ice sheet, therefore offering little or no support for the nunatak hypothesis.

INTRODUCTION

A long-standing controversy in Quaternary geology concerns the glacial history of mountainous areas in northeastern North America, and whether these areas supported isolated icecaps or cirque glaciers in late Wisconsinan time (Flint, 1951). A more recent controversy concerns the vertical and lateral extent of late Wisconsinan ice sheets in the northern hemisphere (Ives, 1978; Denton and Hughes, 1981; Andrews, 1982), which also has important implications for the "nunatak hypothesis" (Ives, 1974).

The nunatak hypothesis states that certain plant and animal species survived the late Quaternary ice ages on nunataks, or in refuges that were not completely inundated by ice. In the strict sense, "nunatak" is an Innuit word meaning a mountain peak surrounded on all sides by glacial ice (Ives, 1974). Fernald (1925) and E. Dahl (1946, 1955) used the nunatak hypothesis to explain anomalous present-day distributions of certain groups of vascular plants as due to survival in refuges in northeastern North America and the Scandinavian mountains, respectively. More detailed botanical work and investigations of invertebrates and mammals on both sides of the North Atlantic Ocean during the twentieth century further strengthened the nunatak hypothesis. Over the last two decades, biogeographical studies in the High Arctic of North America have extended application of the nunatak hypothesis to the overall history of North Atlantic flora and fauna. However, as the biological community strengthened its support for the nunatak hypothesis, geologists and physical geographers on both sides of the North Atlantic argued that all mountain tops in Scandinavia and in many parts of North

America were completely inundated by continental ice during at least the major glaciations of the late Quaternary (R. Dahl, 1963; Flint, 1943). Thus, many geologists and physical geographers have demonstrated unequivocal evidence for glaciation in areas formerly believed to be refugia, and suggest that biologists must look elsewhere for an alternative explanation for anomalous distributions of plants and animals (cf. Hoppe, 1968). More recently, however, glacial geologists have identified some areas in the Maritime Provinces of Canada that appear to have remained ice-free during the late Wisconsinan (Brookes, 1970, 1977; Grant, 1977a, b; Waitt, 1981).

Ives (1974, p. 607) summarized a variety of reasons for testing of the nunatak hypothesis, which should lead to a better understanding of the following: (1) the evolution of life-forms in the arctic regions, (2) the history of glaciation of high and middle-high latitudes, (3) rates of subspeciation, (4) patterns and rates of plant migration, (5) rates of weathering in cold climates, and (6) effectiveness of glacial erosion. Ives (1974, p. 607) further pointed out that progress in any of the above mentioned fields should have an impact on scientific understanding of these fields as they apply to alpine areas at lower latitudes. Thus, characteristics of boulder weathering and soil development at high altitudes, such as in the Rocky Mountains (Birkeland, 1973), and at high latitudes, such as on Baffin Island (Birkeland, 1978), should have application to studies of similar geomorphic processes on lower mountains in the mid-latitudes, such as on Mt. Katahdin.

The purpose of this paper is to reassess previous work and present new data and interpretations about the glacial history of Mt. Katahdin, a highland area in north-central Maine. Evidence against post-ice sheet mountain glaciation on Mt. Katahdin is only summarized here and is presented in more detail elsewhere (Davis, 1976, in prep.). This paper also focuses on evidence for the probable overriding of Mt. Katahdin by continental ice during late Wisconsinan time, and thus contributes to the controversy surrounding the nunatak hypothesis.

PREVIOUS WORK

At 1605 m (5267 ft), Mt. Katahdin is the highest of a group of monadnocks that form outliers of the Longfellow Mountains (Fig. 1), which constitute an extension of the White Mountains into Maine. With a local relief of about 1450 m, Mt. Katahdin is one of the largest massifs in the Appalachians. The mountain is composed largely of granite and is part of a large Devonian pluton that intruded lower and middle Paleozoic sedimentary and volcanic rocks (Griscom, 1966; Hon, 1980). The upper portion of the Katahdin massif is marked by a broad and rolling upland known as the Table Land, which is surmounted by a number of peaks (Figs. 2, 3). Several deep cirques are carved into the mountain flanks, and shape a rugged and steep topography below the Table Land (Figs. 4, 5). Tree line on Mt. Katahdin currently reaches to about 1100 to 1300 m, and the upper portions of the massif are covered by tundra and block fields.

Quaternary research concerning Mt. Katahdin began when Hitchcock (1861, p. 265) suggested that an ice sheet had never entirely covered the mountain, as indicated by the sharp outline of the Knife Edge arête (Figs. 5, 6). Packard (1867, p. 239) also believed that Mt. Katahdin always had been a nunatak because he failed to find erratic boulders above an altitude of about 1200 m. However, De Laski (1872, p. 29) found large fossiliferous and "chalk flint" erratics ("weighing a pound or more") within about 100 m of Pamola Peak and suggested that Katahdin's main summit was molded by an overriding ice sheet. In a detailed description of the physical geography of Mt. Katahdin, Hamlin (1881, p. 221) recognized that North and South Basins were cirques and that the long ridge east of Basin Ponds was a moraine (Figs. 2, 4). Subsequently, Tarr (1900, p. 447) attributed the steep appearance of South Basin to headwall erosion by a cirque glacier that persisted after the most recent recession of continental ice. Tarr (1900, p. 443) also classified the long ridge east of Basin Ponds as a terminal moraine deposited by local cirque glaciers that he envisioned as flowing eastward off Mt. Katahdin. Finally, Tarr (1900, p. 436) reinforced a suggestion that continental ice covered Mt. Katahdin at some time in the past, because he found erratics scattered about summit areas.

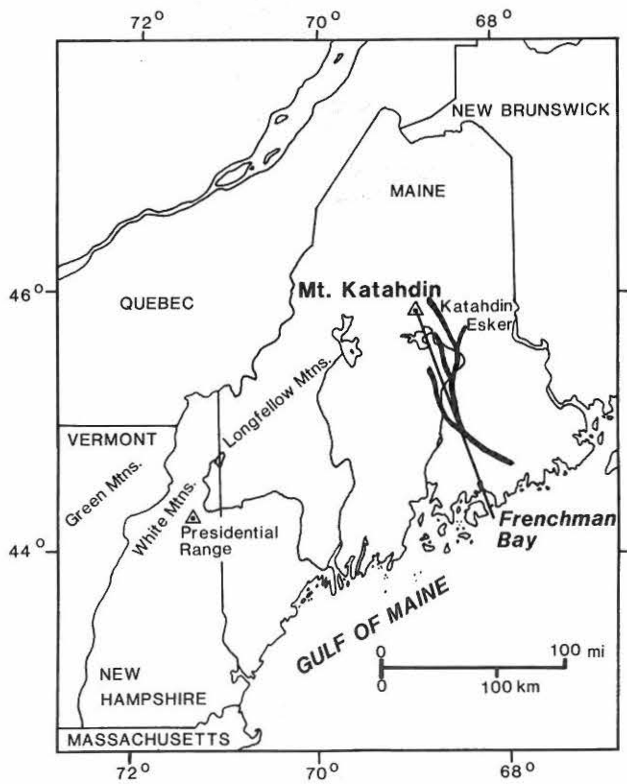

Figure 1. Location map of Maine and surrounding areas. Note location of Katahdin esker system and of transect for theoretical ice-surface profile between Mt. Katahdin and Frenchman Bay shown in Figure 19.

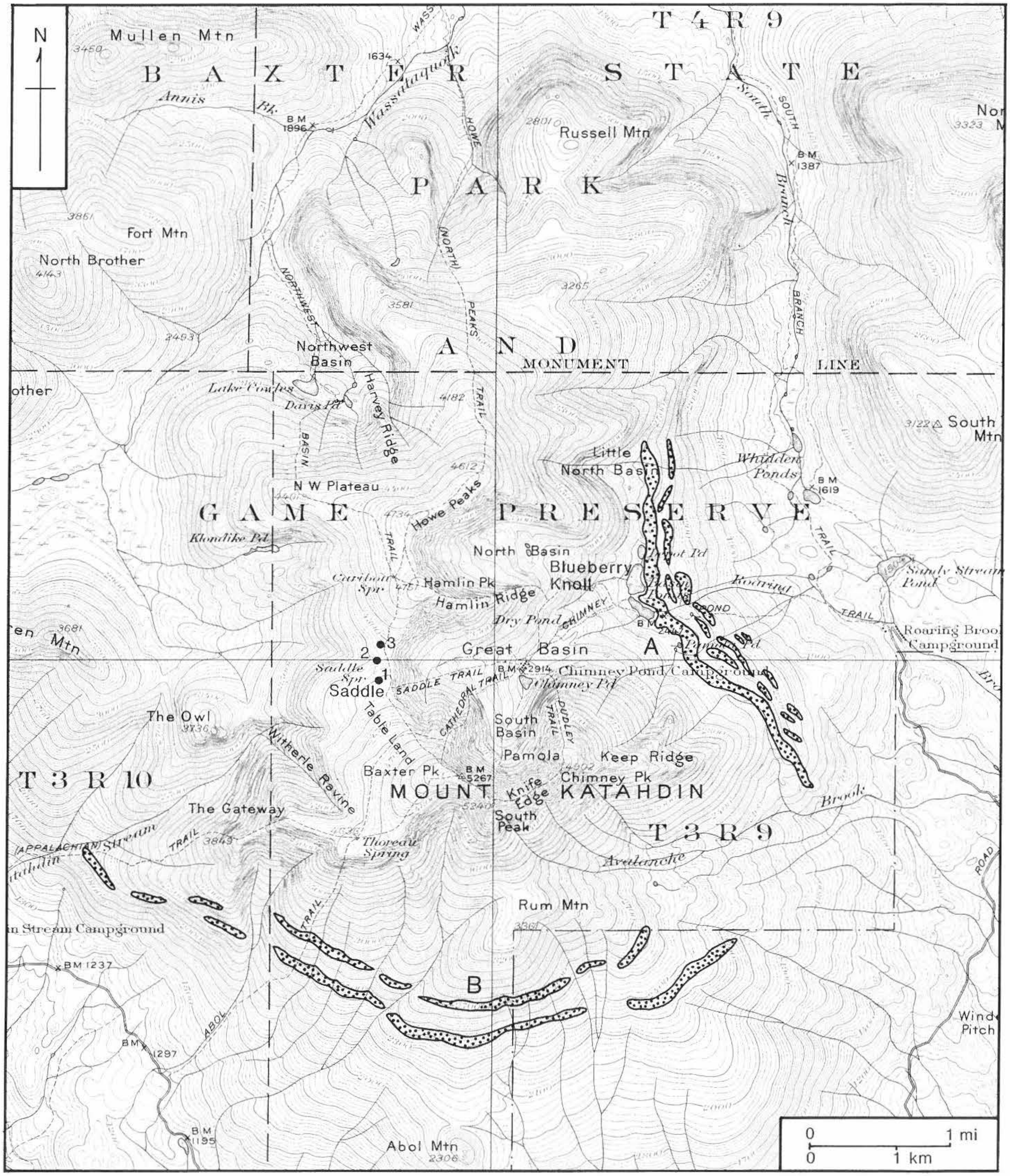

Figure 2. Topographic features of Mt. Katahdin. Major moraine systems are outlined and labeled: A - Basin Ponds moraine system, and B - lateral moraine system on the south slope of the mountain. Soil pit sites on Table Land labeled 1, 2, and 3. From U.S. Geological Survey 15-minute Mt. Katahdin, Maine quadrangle.

Figure 3. View to the northwest of Table Land from north slope of Baxter Peak, with Great Basin at the right.

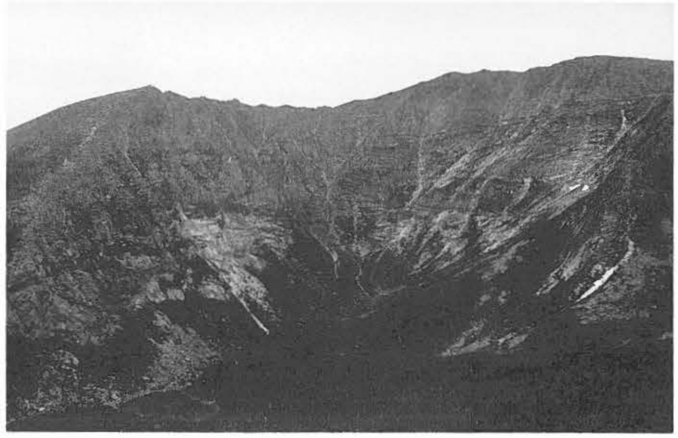

Figure 5. The headwall of South Basin cirque. On the skyline are the Knife Edge at the center, Baxter Peak on the right, and Pamola Peak on the left. Horizontal sheeting noticeable on the lower part of headwall.

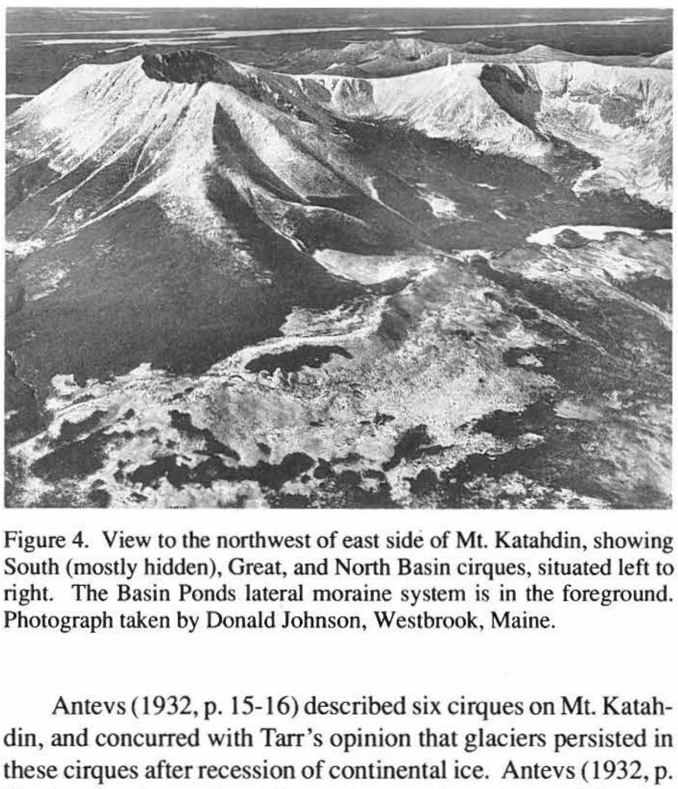

Figure 4. View to the northwest of east side of Mt. Katahdin, showing South (mostly hidden), Great, and North Basin cirques, situated left to right. The Basin Ponds lateral moraine system is in the foreground. Photograph taken by Donald Johnson, Westbrook, Maine.

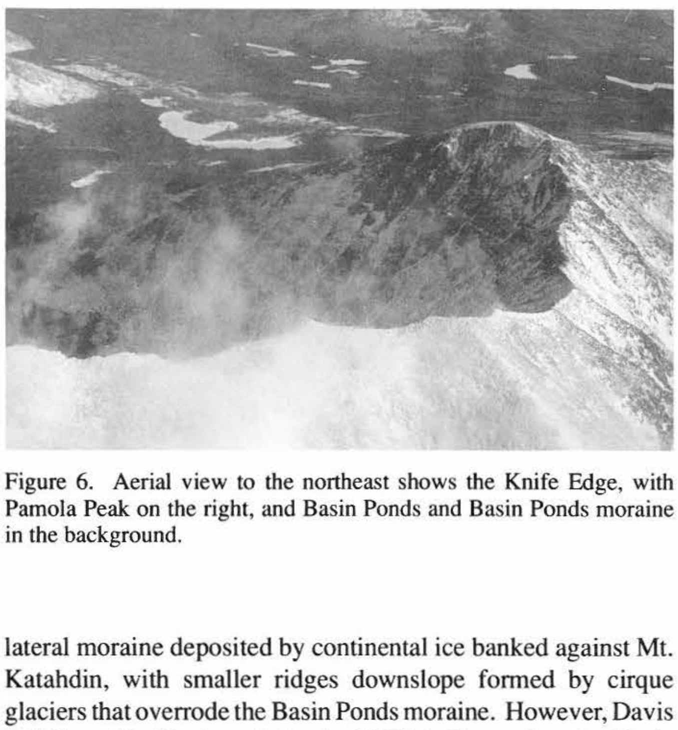

Figure 6. Aerial view to the northeast shows the Knife Edge, with Pamola Peak on the right, and Basin Ponds and Basin Ponds moraine in the background.

Antevs (1932, p. 15-16) described six cirques on Mt. Katahdin, and concurred with Tarr's opinion that glaciers persisted in these cirques after recession of continental ice. Antevs (1932, p. 8) also found erratic pebbles close to the cairn marking the summit of Baxter Peak (Fig. 2). Finally, Antevs interpreted the ridge east of Basin Ponds as a medial moraine deposited between Katahdin cirque glaciers on the west and an active continental ice tongue on the east. More recently, Caldwell (1959, 1966, 1972, 1980; Caldwell and Hanson, 1982), Thompson (1960a,b, 1961), and Davis (1976, 1978, 1983; Davis and Davis, 1980; Davis et al., 1980; Waitt and Davis, 1988) have described glacial features on Mt. Katahdin. Caldwell (1966, p. 52) supported Antevs in his interpretation of the long ridge east of Basin Ponds as a medial moraine. However, because of its morphology, the ridge was considered by Thompson (1961, p. 470-473) to be a lateral moraine deposited by continental ice banked against Mt. Katahdin, with smaller ridges downslope formed by cirque glaciers that overrode the Basin Ponds moraine. However, Davis (1976, p. 81; Davis and Davis, 1980) believes that the Basin Ponds moraine and the ridges downslope were formed entirely by a tongue of active continental ice receding into the trough east of Mt. Katahdin. Davis (1976, p. 69; 1978) also found large, faceted, striated erratics within 50 m of Baxter Peak, indicating that the mountain was overridden by continental ice at some time in the past.

Elsewhere in New England, various conclusions have been reached concerning the Quaternary history of mountain areas. Goldthwait (1970) summarized evidence against post-Wisconsinan reoccupation of cirques subsequent to continental ice in the Presidential Range of the White Mountains in New

Hampshire (Fig. 1). This evidence consisted of: (1) glacial grooves and roches moutonnées, indicating uphill ice movement, preserved on or at the base of cirque headwalls; (2) drift of northern provenance on cirque floors; and (3) the lack of lateral or end moraines in the cirques. Although Bradley (1981) suggested that large boulders of schist were transported out of north-facing valleys by post-ice sheet cirque glaciers in the Presidential Range, Fowler (1984), Gerath et al. (1985), Davis and Waitt (1986), Davis and Thompson (1988), and Waitt and Davis (1988) attribute these boulders to postglacial mass wasting. In the Green Mountains of Vermont (Fig. 1) Wagner (1970, 1971) considered small ridges at valley mouths to be end and lateral moraines, and the products of late-Wisconsinan mountain glaciers. However, Waitt and Davis (1988) note that the valleys above these ridges are not headed by cirques, and suggest that the ridges are merely hummocky ground moraine deposited by the waning ice sheet. Finally, Borns and Calkin (1977) did not find evidence for post-ice sheet cirque glaciers in the Longfellow Mountains in west-central Maine.

EVIDENCE AGAINST POST-ICE SHEET MOUNTAIN GLACIATION

Nearly all previous workers have thought some of the cirques and separating arêtes, especially those on the east side of Mt. Katahdin, are too fresh to have survived overriding by a continental ice sheet (Tarr, 1900; Antevs, 1932; Thompson, 1960a,b, 1961; Caldwell, 1966, 1972, 1980; Caldwell and Hanson, 1982; Caldwell et al., 1985). Yet sharpness of arêtes and steepness of cirque headwalls are not necessarily diagnostic of post-ice sheet cirque glaciation. For example, ragged arêtes and well-formed cirques in the North Cascade Range of Washington were scarcely altered by burial beneath hundreds of meters of the late Wisconsinan Cordilleran ice sheet (Waitt, 1975, 1977). Rather, the east-facing cirques of Mt. Katahdin (Figs. 4, 5, 6) have been freshened by postglacial mass wasting, which is controlled by the strong vertical jointing of the bedrock (Fig. 7) (Davis, 1976). Also, summit areas of the Katahdin massif consist of granophyric rock (Hon, 1980) that provides a resistant cap which protects the less resistant pluton core from erosion (Caldwell and Hanson, 1985). Evidence against post-ice sheet cirque glaciation on Mt. Katahdin is summarized below.

Roches Moutonnées

The floor of Northwest Basin (Fig. 2) contains at least three roches moutonnéees whose grooved, chattermarked stoss ends face downvalley and whose ragged ends face the cirque headwall. The long axes of these forms and glacial grooves etched into them trend south-southeast, slightly oblique to the trend of the cirque. Thus, these features suggest that a cirque glacier did not occupy Northwest Basin after continental ice receded.

Cirque-Floor Drift

Mt. Katahdin is formed of the Katahdin granite, biotite granite to granophyre, which intruded lower Paleozoic sedimentary and volcanic rocks that surround the massif (Osberg et al., 1985). Thus, sedimentary and volcanic stones in drift on Mt. Katahdin are erratic. The composition of drift in Northwest Basin indicates that any post-ice sheet cirque glaciers there, as envisioned by Antevs (1932), would have flowed opposite to the south or southeast flow of continental ice. Gravel to pebble size stones in a thin till sheet on the floor of Northwest Basin are dominantly granite, but 22 to 38 percent of the pebbles are sedimentary and volcanic erratics (Table 1). Many of these erratic stones are derived from the lower Devonian Seboomook Formation and Matagamon Sandstone, two of the bedrock units north of Mt. Katahdin. Many of the granite clasts also may have been derived from the mountain flank northwest of the cirque, as Katahdin granite bedrock extends about 10 km north of Northwest Basin (Osberg et al., 1985). The presence of erratics indicates that the last glacier to occupy Northwest Basin was the ice sheet flowing up the cirque, because any post-ice sheet cirque glacier should have redistributed much of the imported drift back downvalley. Thus, erratic drift corroborates the southeastward ice-flow direction inferred from roches moutonnées in Northwest Basin.

Erratic pebbles in thick drift are not as abundant on the floors of the three east-facing cirques (Fig. 4) as they are in Northwest Basin cirque, yet some counts show as much as 20 percent sedimentary and volcanic clasts (Table 1). Numerous erratics

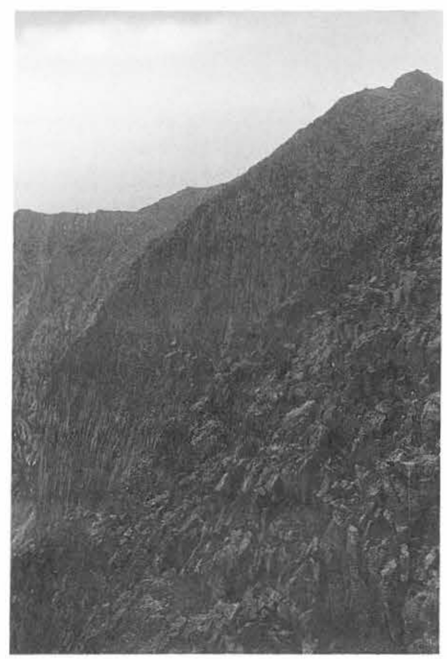

Figure 7. Prominent vertical jointing on upper part of the South Basin headwall. South Peak is on the right and Chimney Peak is on the left.

TABLE 1. ERRATIC STONES COUNTED IN THE PEBBLE FRACTION OF TILL, MT. KATAHDIN.[1]

Location	Sample Number	Number counted	Percent erratic[2]	Location	Sample Number	Number counted	Percent erratic[2]
Northwest Basin	K85	100	38	Basin Ponds moraine, south	K91	50	44
	K86	50	30		K92A	50	34
	K87	50	22		K92B	50	34
	K88	50	28		K93	50	24
					K94	50	26
North Basin	K11	100	15				
	K42	50	20	Moraines east of Basin Ponds	K71	50	42
	K47	100	13	moraine	K72	50	12
	K56	100	9		K81	25	48
(Blueberry Knoll)	K43	50	16		K82	50	64
	K44	50	14		K83	50	68
Great Basin	K10	200	2	Moraines, south slope Katahdin,	K96	50	42
	K32	200	3	east	K97	50	48
	K33	200	4	Moraines, south slope Katahdin,	K98	50	32
	K34	100	3	west	K99	50	24
	K35	200	2				
	K55	200	2	Table Land (near Hamlin Peak)	K6	200	10
					K27A	100	6
South Basin	K31	100	4		K27B	100	13
	K54	200	5		K28	100	5
Basin Ponds moraine, north	K57	50	20	Table Land (Northwest Plateau)	K89	100	14
	K58	50	24				
	K59	50	18	Table Land (near the Saddle)	K7	100	11
	K60	50	14		K8	100	10
	K61	50	10		K9	300	5
	K62	50	32		K19	100	9
	K63	50	14		K22	100	8
	K64	50	26		K23	100	9
	K65	50	32		K24	100	10
	K66	50	16		K25	100	11
	K67	50	22				
	K68	50	24	Table Land (north side	K15	100	8
	K69	50	40	Baxter Peak)	K16	100	6
	K70	50	32		K17	100	7
					K18	100	7

[1]Summarized from detailed counts by lithology in Davis (1976, Appendix B; specific locations on his Fig. 13).
[2]Includes 31 distinguished varieties of sandstone, siltstone, shale, metasandstone, chert, slate, mica schist, chlorite schist, and volcanic rocks.

among the cobble-boulder fraction are also easily distinguished in trail-cuts in these cirques. Thompson (1961, p. 470) and Caldwell (1966, p. 53) interpret the lower abundance of erratics on the floors of the east-facing cirques to mean that cirque glaciers removed material after continental ice disappeared. An alternative explanation is Boulton's (1974, p. 55; 1975, p. 15) basal-freezing and regelation mechanism, which suggests maximum erosion of bedrock on the up-glacier side of obstructions and maximum deposition of till on the down-glacier side. The continental ice flowing over Mt. Katahdin incorporated mainly granite as it flowed up the northwest mountain flank and consequently deposited mainly granite clasts in the east-facing cirques, which were hollows in the glacier bed.

That all the drift on east-facing cirque floors could have been derived from postglacial mass wasting is also unlikely because the cirque walls are too steep to have accumulated much drift. Later mass wasting from the steep rock slopes would inevitably also consist mostly of angular local-rock debris, as mapped mass-wastage deposits near valley sides and cirque walls indeed do (Davis, 1976). Thus, the small abundance of erratics in the east-facing cirques is not necessarily the result of mountain glacier activity after ice sheet recession.

Moraines

The distinct Basin Ponds moraine (Hamlin, 1881; Tarr, 1900; Antevs, 1932; Caldwell, 1972; Davis, 1976; Thompson and Borns, 1985a) extends for about 5 km approximately on contour at 740 m (2400 ft) altitude across the eastern slope of Mt. Katahdin as a continuous hummocky ridge that dams Basin, Depot, and Pamola Ponds (Figs. 2, 4). The moraine consists of mixed-lithology till, but also contains many rounded granitic boulders as large as 6 m in diameter. The Basin Ponds moraine has been interpreted varyingly as a terminal moraine deposited by late Wisconsinan mountain glaciers flowing eastward from cirques floored in granite on Mt. Katahdin (Tarr, 1900), as a medial moraine between such mountain glaciers and continental ice to the east (Antevs, 1932; Caldwell, 1966, 1972), or as a

lateral moraine marking the maximum late Wisconsinan continental ice limit against an ice-free Mt. Katahdin (D. Grant, pers. commun., 1976; Caldwell and Hanson, 1986). However, the morphology of the moraine and the erratic component of its till do not support the first two ideas. The third idea suggests that Mt. Katahdin remained a nunatak throughout the late Wisconsinan and perhaps longer, and will be addressed later in this paper.

The Basin Ponds moraine contains many large granitic boulders; however, the pebble fraction is 10 to 44 percent erratic, suggesting that all of the material could have been transported by an ice sheet flowing around the mountain, rather than by local glaciers from the east-facing cirques. The moraine does not descend in a convex form eastward as expected from a local-glacier moraine; rather it is convex westward (upvalley) and roughly follows a contour along the east slope of Mt. Katahdin (Davis, 1976, p. 55). The moraine extends both north and south beyond the mouths of the three east-facing cirques. There is too little space between the moraine and Keep Ridge for a hypothetical former cirque glacier or even a large drainage channel (Figs. 2, 4). Therefore, the Basin Ponds moraine must have been built by late Wisconsinan continental ice occupying the valley east of Mt. Katahdin, when the mountain was an emergent nunatak; both the lithology and the form of the moraine are wrong for deposition by cirque glaciers flowing eastward from the mountain.

Several smaller moraines lie on the mountain flank just downslope from the Basin Ponds moraine (Figs. 2, 4). Some of these are arcuate, but generally they are parallel to each other and to the Basin Ponds moraine. These smaller ridges contain 42 to 68 percent erratic pebbles (Table 1), even more than the Basin Ponds moraine. Moreover, these moraines lie downslope of the Basin Ponds moraine below Keep Ridge and below Little North Basin. These moraines are significant because they lie south and north of the three large east-facing cirques and could not possibly have been built by cirque glaciers. Thus, for several reasons, these smaller ridges were not formed by cirque glaciers that overrode the Basin Ponds moraine as suggested by Thompson (1961), but rather are recessional moraines built by the west margin of the continental glacier as it receded downslope and eastward from the Basin Ponds moraine.

Several closely spaced ridges that lie along the south slope of Mt. Katahdin (Fig. 2), where cirques are absent, are believed by Davis (1976) and Davis and Davis (1980) to be lateral moraines. These nearly continuous ridges extend for about 8 km and appear to descend in altitude from 730 m (2400 ft) in the east to 570 m (1850 ft) in the west. Caldwell et al. (1985) suggested that these ridges are kame terraces. However, the ridges are composed of till and many erratics, some with striated and faceted faces, recovered from pits dug in the ridges. The high percentages of erratic pebbles (Table 1), including red slate not found higher on the mountain, indicate that the ridges were built by a tongue of continental ice flowing around Mt. Katahdin. Caldwell identified these red erratics as derived from the Capens Formation, a Silurian red and green slate that crops out about 30 km due west and about 80 km northwest of Mt. Katahdin (Caldwell and Davis, 1983, p. 84). These erratics and the absence of cirques on the south side of the mountain make the deposition of these ridges, whether they are lateral moraines or kame terraces, by local glaciers impossible (Davis, 1983).

Near the mouth of Great Basin and South Basin, a hummocky landform named "Bears Den moraine" by Tarr (1900, p. 442) encloses Dry Pond (Fig. 2). Tarr interpreted this feature as a late-Wisconsinan end moraine deposited by a cirque glacier flowing from South Basin. The landform is a depression surrounded by mounds rather than a continuous morainal ridge. The feature appears to have been cut by a postglacial stream, which concentrated large and rounded granitic boulders as a lag deposit by piping out fine particles. Thus, this landform offers no proof of cirque glaciation (Davis and Davis, 1980).

Thompson (1961, p. 471) and Caldwell (1966, p. 55) interpreted a smooth and subdued mound east of Chimney Pond as an end moraine deposited by a cirque glacier in South Basin after disintegration of the continental ice sheet. The mound is littered with many rounded granitic boulders as large as 3.5 m in diameter; only 2 percent of the pebbles are erratic. However, the mound has neither a ridge form like the Basin Ponds moraine nor is it arcuate, and thus offers no proof of cirque glaciation.

Blueberry Knoll at the mouth of North Basin (Fig. 8) has been interpreted as an end moraine of a cirque glacier (Caldwell, 1966, p. 53) and as a rock glacier deposit (Thompson, 1961, p. 472). The knoll is covered by till composed mainly of granitic boulders as large as 3 m in diameter, but the pebble fraction is as much as 16 percent erratic (Table 1). Although the upper part of North Basin displays transverse furrows and ridges, conical pits, and a fan-like lobe diagnostic of rock glaciers described by Wahrhaftig and Cox (1959), Blueberry Knoll does not exhibit these features. Seismic profiles show an average thickness of about a meter of till over bedrock on Blueberry Knoll, but a much

Figure 8. North Basin cirque floor. View to the east from Hamlin Peak. Arrow points to Blueberry Knoll. Water bodies from west to east include an unnamed pond, upper Basin Pond, Depot Pond, Whidden Ponds, Sandy Stream Pond, and Katahdin Lake.

greater thickness of till on the adjacent cirque floor (P. T. Davis, 1976; unpublished data, 1977). Therefore, Blueberry Knoll is not an end moraine, but is most probably a bedrock high capped by ice sheet till.

The only genuine, indisputable moraines lie outside the cirques of Mt. Katahdin. No lateral and looped end moraines lie on cirque floors (Davis, 1976, p. 46); rather the cirques contain only formless till that is overlapped by bouldery talus and protalus ramparts that do not extend far from steep valley walls. A suggestion that ridge-like deposits in cirques have been subdued by erosion (Thompson, 1961, p. 471) is inconsistent with the excellent preservation of the Basin Ponds moraine and other lateral moraines on Mt. Katahdin (Fig. 4). The speculation by Tarr (1900) that moraines lie hidden beneath dense forest does not hold for North Basin, the floor of which is nearly devoid of vegetation. Indeed, Davis (1976) made many traverses across all cirque floors with air photos in hand, but did not find morainal ridges in any cirque basin. Caldwell's (1972, Fig. 1) map exaggerates selected drift ridges on cirque floors by ignoring hundreds of other hummocks within and outside the Mt. Katahdin cirques.

Discussion

Caldwell et al. (1985, p. 55) suggested that Northwest Basin has a different deglacial history than the east-facing cirques on Mt. Katahdin. However, taking into account its relatively small size appropriate to its somewhat lower altitude, Northwest Basin had a large potential accumulation area for reception of "blowover" snow, and its northwest aspect favored a low rate of ablation. Caldwell et al. (1985, p. 55) also suggested that a marine calving bay to the east could have provided precipitation sufficient to reestablish mountain ice at Mt. Katahdin. Although calving bay dynamics could partly explain readvances of lowland continental ice (Thomas, 1977), probably no combination of increased precipitation and wind-drifted snow could have compensated for the dramatic rise in temperature that pushed the snowline hundreds of meters above cirque floors at Mt. Katahdin during ice sheet deglaciation (Davis, 1976; Davis and Davis, 1980). Thus, all erosional and depositional evidence on Mt. Katahdin suggests that post-ice sheet, mountain glaciation did not occur on Mt. Katahdin. All of the erosional features, drift, and moraines are most readily accounted for by the sequence of ice sheet glaciation and deglaciation following the last significant mountain glaciation.

EVIDENCE FOR ICE SHEET OVERRIDING MT. KATAHDIN

Goldthwait (1970) summarized evidence for continental ice overriding the highest summits in the Presidential Range (Fig. 1) of the White Mountains in New Hampshire: (1) stoss and lee forms on bedrock exposures, (2) fresh, polished, and striated bedrock surfaces, (3) fresh erratics in drift, and (4) weakly developed soils. On Mt. Katahdin similar evidence suggests that the highest summits were also inundated by continental ice. Stoss and lee forms and far-traveled, north-derived erratics lying within 45 m of Baxter Peak (Fig. 2) show that Mt. Katahdin was overridden by a continental ice sheet at some time in the past (Tarr, 1900; Antevs, 1932; Davis, 1976, p. 69). Moreover, the freshness of erratics and polished bedrock surfaces on Mt. Katahdin, the thinness and poor development of soils, and theoretical ice profiles extrapolated back from recessional ice positions on the Maine coast indicate that a late Wisconsinan ice sheet buried Mt. Katahdin (Davis, 1976, p. 69-75; Davis, 1978, 1983). This evidence for ice sheet overriding of Mt. Katahdin, probably during the late Wisconsinan, refutes the nunatak hypothesis and is summarized below.

Grooves, Striations, and Stoss - Lee Forms

Hitchcock (1861, p. 387) did not observe striated or polished surfaces on the summit areas of Mt. Katahdin, therefore he concluded that the mountain had never been overridden by a continental ice sheet. Although preservation of striated and polished surfaces might not be expected due to intense physical weathering in an alpine environment, Davis (1976) noted an undulatory, grooved, glacially abraded surface not related to any joint plane in the bedrock on Cathedral Ridge arête at about 1400 m altitude. Although Hitchcock (1861, p. 388) noted that the northwest side of Mt. Katahdin resembles a large stoss slope and the southeast side appears to be a jagged lee slope, he passed off these observations as coincidental. The shapes of these surfaces could be primarily controlled by the underlying resistant granophyric rock capping the Katahdin pluton (cf. Hon, 1980). Alternatively, the steep headwalls of the three east-facing cirques on Mt. Katahdin could owe their origin in part to ice sheet erosion (T. Hughes, pers. commun., 1976). The same glacier basal-freezing and regelation mechanism of Boulton (1974, 1975), used to address erratic distributions on cirque floors earlier in this paper, could explain much of the cirque headwall erosion by subglacial cavitation processes during ice sheet overriding. Likewise, sharp arêtes preserved from pre-late Wisconsinan cirque glaciation could also be freshened by an overriding ice sheet.

Drift and Erratics on Summits and Table Land

De Laski (1872), Hamlin (1881), Tarr (1900), Antevs (1932), and Davis (1976) found unmistakable erratics near the highest summit of Mt. Katahdin, Baxter Peak, and concluded that at some time in the past a continental ice sheet overrode the massif. Caldwell (1959, 1966, 1972) did not find definitive erratics on the summit areas of Mt. Katahdin, but did find non-granitic stones on the summits of some of the lower mountains in the region. Although Caldwell and Hanson (1982) once believed that Mt. Katahdin was overridden by continental ice,

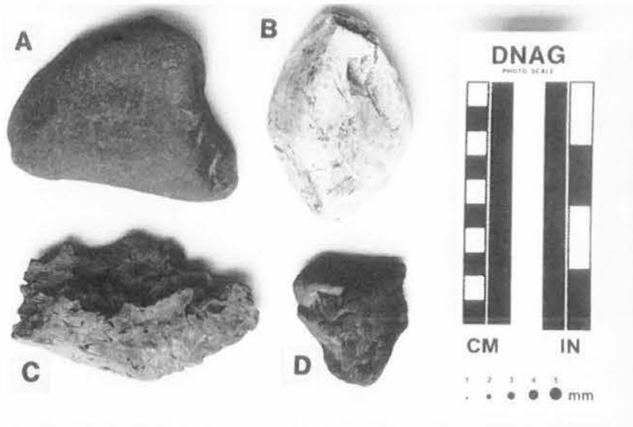

Figure 9. Erratic stones found on the Table Land's north slope of Baxter Peak, Mt. Katahdin. (a) Faceted, striated clast of dark gray, massive shale. (b) Well rounded clast of talc schist, perhaps derived from more than 200 km distance in Quebec. (c) Striated clast of fossiliferous brown siltstone, probably derived from the Seboomook Formation at least 50 km to the northwest. (d) Rounded clast of metalliferous hematite, also possibly derived from as far away as Quebec.

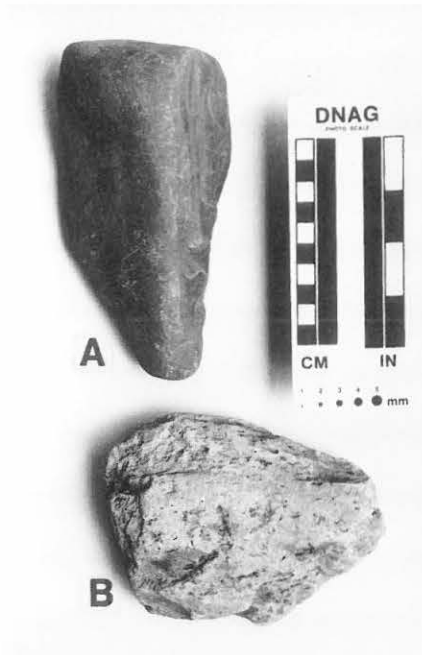

Figure 11. Erratic clasts found in till near the Saddle, Mt. Katahdin. (a) Bullet-shaped, striated, faceted clast (16-cm long axis). (b) Faceted clast (13-cm long axis) of pitted, light gray, fissile shale.

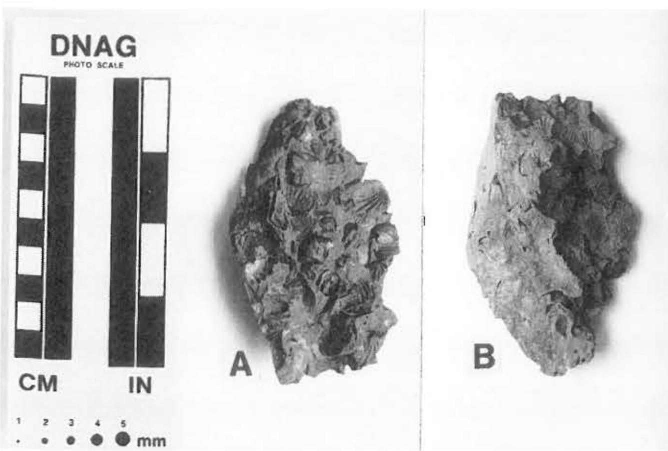

Figure 10. Striated, faceted, erratic stone of fossiliferous brown siltstone shown in Figure 9c. Found near the Saddle, Mt. Katahdin, probably derived from the Seboomook Formation.

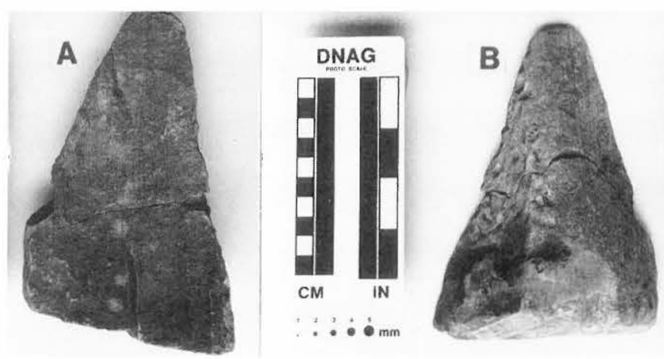

Figure 12. Striated, faceted, erratic clast of layered sedimentary rock found 100 m northwest of Baxter Peak, perhaps derived from the Matagamon Sandstone.

they now feel that all the erratics on the broad upland of Mt. Katahdin are only xenoliths eroded from the summit granophyric facies (Caldwell and Hanson, 1986). They suggested that all of these erratics represent a single rock unit, the Matagamon Sandstone, which they feel was probably intruded by the Devonian Katahdin pluton.

However, many of the erratics found on Mt. Katahdin between the Saddle and Baxter Peak (Fig. 2) represent rock types other than the Matagamon Sandstone, and many show definite signs of glacial abrasion. Examples of four erratic cobbles (6 to 10 cm long axes) are shown in Figure 9. Figure 10 displays both the broken face (a) exposing a rich fauna of brachiopods and the smooth faceted side (b) of the erratic cobble from the Seboomook Formation shown in Figure 9c. In Figure 11a the bullet-shaped, striated, faceted clast (16 cm long axis) appears to be derived from the same rock type as the clast shown in Figure 9a. The faceted clast (13 cm long axis) of pitted, light gray, fissile shale shown in Figure 11b does not exhibit any sign of metamorphism that one might expect from a xenolith weathered from a chilled zone of country rock overlying a granitic pluton. Figure 12 also exhibits a cobble of gray siltstone with a "glacial shape" and sedimentary layering exposed on a broken face (a), and tiny brachiopods displayed on the striated, faceted surface (b). This cobble could have been derived from the Matagamon

Figure 13. Large erratic cobble of lichen-covered, rounded quartzite, found within a block field about 50 m north of Baxter Peak.

Sandstone, and its shape and surface features suggest glacial transport. Although larger erratics were observed by Davis (1976) on the uplands of Mt. Katahdin, the largest retrieved was a 7-kg cobble (18 x 12 x 12 cm) of lichen-covered, rounded quartzite (Fig. 13), found within a block field about 50 m north of Baxter Peak. Although all stones at the surface of block fields on Mt. Katahdin are heavily covered with lichen growth, the erratics are generally more rounded than are the dominant granitic blocks, and therefore easily distinguished.

Erratic stones in the pebble fraction (2 to 5 cm long axis) of till exposed in trail-cuts and in burrow pits on the uplands of Mt. Katahdin only range from 5 to 11 percent (Table 1); however, an overriding ice sheet was probably eroding and entraining primarily granitic material across the Katahdin massif. Also, as pointed out by Caldwell et al. (1985), a large portion of ice-sheet flow should have been around Mt. Katahdin rather than across the top. Indeed, Lowell (1985), Kite et al. (1981), and Hyland (1981, 1986) summarized striation data from lowlands north and west of Mt. Katahdin that support divergent ice-sheet flow around the mountain. However, many of the stones in the pebble counts also reveal striated, faceted surfaces, and thus testify to an ice sheet overriding Mt. Katahdin at some time.

Although glacial erosional features and glacial drift including erratics provide clear evidence for ice-sheet overriding of alpine areas, other geological evidence is less definitive. For example, block fields and related mountain-top detritus (felsenmeer), tors and related weathering forms, soil profiles with extensive clay mineral development, and chemical weathering of bedrock and debris are open to multiple interpretations which follow (E. Dahl, 1955; R. Dahl, 1966a,b, 1967; Ives, 1958a,b, 1966).

Block Fields

E. Dahl (1955) and Ives (1957, 1958a,b, 1966) suggested that block fields require a long time to develop, and therefore are indicative of alpine areas that have remained ice-sheet free throughout the late Quaternary glaciations. However, R. Dahl (1966a, b, 1967) found erratics within block fields in the Narvik Mountains of northern Norway, and suggested that either the block fields have been overridden by continental ice or, more probably, that block fields had developed rapidly during postglacial time.

Block fields are found over much of Mt. Katahdin, from the summits to the floor of North Basin (Fig. 2). The angular character of the blocks on Mt. Katahdin is due in part to the well-jointed granitic bedrock. In addition, many forms of patterned ground are common both on the Table Land and on the floor of North Basin, and include sorted circles, polygons, nets, stripes (block streams), and steps (block terraces), as classified by Washburn (1980).

Although block fields and patterned ground occur on the floor of North Basin at altitudes as low as 950 m (Fig. 8), the best-developed forms appear on the Table Land at altitudes between 1300 and 1500 m (Figs. 14, 15, 16, 17). The character of block fields and patterned ground on the Table Land is a function of altitude, slope angle, and underlying bedrock structure. The effect of slope angle on block fields and patterned ground is well-illustrated on the northern slope of Baxter Peak (Fig. 2). The best developed sorted polygons and most highly concentrated block fields occur on slope angles of less than 5^{o}, with polygon lengths ranging between 4 and 12 m and widths between 1 and 6 m (Fig. 14). Elongated polygons and sorted stripes are characteristic of steeper gradients between 5^{o} and 10^{o}

Figure 14. Sorted polygons at an altitude of 1350 m on the north slope of Baxter Peak, Mt. Katahdin. View is looking east.

Figure 15. Sorted stripes (block streams) at an altitude of 1330 m on the northeast slope of Baxter Peak, Mt. Katahdin. View is looking north across Great Basin.

Figure 17. Block field at an altitude of 1360 m on northeast slope of Baxter Peak above Cathedral Ridge, Mt. Katahdin. Cairn is one meter high.

Figure 16. Block terrace at an altitude of 1480 m on north slope of Baxter Peak, Mt. Katahdin. Note rucksack at center of photograph for scale.

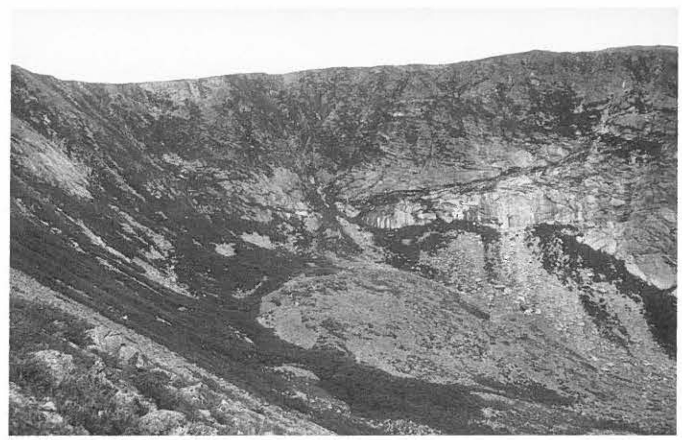

Figure 18. View to northwest of upper part of North Basin, with vertical cliff on right side of headwall and fan-shaped, fossil rock glacier on cirque floor below headwall.

(Fig. 15). In these areas polygons extend to 25 m in length and 10 m in width. Finally, sorted steps (Fig. 16) occur along planes of structural weakness in the bedrock. Slope angles also appear to have some influence on the size of blocks comprising these features. Slope angles less than 10° are generally characterized by angular blocks less than 50 cm long (Fig. 17), whereas gradients exceeding 10° concentrate blocks more than 6 m long (Fig. 16). Large rounded to subrounded erratics occur at all elevations on the Table Land, in both block fields and till.

Dating by relative degree of block field and patterned ground development is useless on Mt. Katahdin. Well-sorted polygons were found on the Table Land and on the rock glacier at the head of North Basin (Fig. 18). However, pits dug on the Table Land did not reveal contacts between till and block fields, although erratics were found mixed within the block fields. The survival of block fields and patterned ground under overriding "cold-base" ice is a possibility (cf. Sugden, 1977, 1978). However, the occurrence of well-formed block fields and patterned ground on a rock glacier (Fig. 18), which must have been active during the early postglacial, implies that the block fields and patterned ground have developed since late Wisconsinan deglaciation.

Rock Weathering and Soil Development

The degree of weathering of granitic boulders has been used to determine glacial chronologies in the Arctic (cf. Birkeland, 1978) and in the alpine environment (cf. Birkeland, 1973). However, on Mt. Katahdin grussified granitic boulders are found everywhere, including on the Table Land, in South Basin, on Blueberry Knoll, and on the Basin Ponds moraine. Perhaps the weathered grus on Mt. Katahdin is common in part because of

the open-framework, coarse-grained, easily weathered texture of the lower Doubletop facies of the Katahdin granite (cf. Hon, 1980). In any case, such a distribution of weathered granitic boulders does not provide even a relative age differentiation for glacial deposits on Mt. Katahdin.

However, limited chemical weathering of exposed bedrock and erratics may provide evidence for late Wisconsinan ice sheet glaciation of the uplands of Mt. Katahdin. Surface pitting and weathering rind development is incipient on the granophyric bedrock and boulders on the Table Land. Moreover, erratic cobbles on the Table Land are fresh in appearance, with nearly nonexistent weathering rinds. For example, the slab cut from the erratic quartzite cobble shown in Figure 13 does not exhibit more than a 1-mm thick weathering rind on its rim. Typical weathering rind thicknesses on granitic rocks in alpine environments of the western United States and in Arctic Canada are on the order of 1 to 2 cm thick for pre-late Wisconsinan age glacial deposits (Birkeland, 1973; Locke, 1985). Also, surface weathering pit depths may commonly reach 10 to 15 cm for these older deposits, whereas on Mt. Katahdin weathering pits are minimal. Thus, rock weathering of granitic rocks on the Table Land supports a late Wisconsinan age for ice sheet overriding.

Preliminary data from three soil profiles exposed in pits dug on the Table Land (Table 2) also suggest that little weathering has occurred on the summit areas of Mt. Katahdin since deglaciation. Testifying to the youthful development of these soils are the following: 1) soil oxidation depths less than 50 cm; 2) thin (less than 12 cm) and poorly developed humic (A) horizons (grayish, yellow-brown, dry Munsell colors); 3) lack of either cambic (color) or argillic (textural) B horizons (dull, yellow-orange, dry Munsell colors), with less than a 2 percent increase in clay relative to the humic horizons or parent material, and weak, single-grained structures. In contrast, typical late Wisconsinan soil profiles in arctic and alpine environments should have much greater oxidation depths, much thicker and better developed, brownish black humic horizons, redder and more

TABLE 2. GENERALIZED SOIL PROFILE DATA FOR THREE SITES ON TABLE LAND, MT. KATAHDIN.

Site[1]	Depth(cm)	Horizon[2]	Dry Munsell Color	Clay %	Structure
1	0-10	A	10 YR 5/2	6.5	single-grained
	10-18	C1ox	10 YR 6/4	10.3	"
	18-50	C2ox	10 YR 6/4	8.5	"
	50-56+	Cn	10 YR 5/4	5.0	"
2	0-8	A	10 YR 5/2	7.7	single-grained
	8-15	C1ox	10 YR 6/4	10.2	"
	15-38	C2ox	10 Yr 6/4	8.9	"
	38-51+	Cn	10 YR 5/4	6.3	"
3	0-12	A	10 YR 5/3	8.8	single-grained
	12-44	Cox	10 YR 6/3	10.6	"
	44-62+	Cn	10 Yr 5/3	7.9	"

[1]See Fig. 2 for locations of soil pit sites.
[2]Follows nomenclature of Birkeland (1984).

clay-rich subhumic horizons, and subangular to angular blocky structures (Birkeland, 1973, 1978). Perhaps more detailed analyses of clay minerals by X-ray diffraction (Birkeland, 1984) or of hornblende etching by microscopy (Locke, 1979) could reveal better developed weathering characteristics; however, these preliminary soil data argue for a late Wisconsinan glaciation of the Table Land.

Theoretical Ice-Surface Profiles

A theoretical ice-surface profile constructed from esker data in Maine by Shreve (1985a,b) suggests that Mt. Katahdin was a nunatak, with the late Wisconsinan ice margin only reaching to about 1100 m. However, there are no geomorphic features on Mt. Katahdin that relate to an ice margin at this altitude. Also, Shreve (1985a, p. 639) assumed that the entire length of the Katahdin esker system formed simultaneously. Such an assumption may be erroneous, as continental ice in Maine likely underwent deglaciation by stagnation-zone retreat (Koteff, 1974). Thus, morphosequences (Koteff and Pessl, 1981) of eskers and other ice-contact deposits were probably laid down sequentially, rather than simultaneously, as the ice margin retreated. A few hundred years may have been required to build the Katahdin esker system by segments, with the entire length never undergoing subglacial flow at any one time. Therefore, Shreve's (1985a,b) ice-surface profile reconstruction may be a minimum reconstruction by several hundred meters where it passes near Mt. Katahdin. Thus, the age of ice-sheet overriding of Mt. Katahdin can not be firmly established from field data.

An alternative theoretical ice-surface profile (Fig. 19) suggests that continental ice covered Mt. Katahdin at least during the late Wisconsinan maximum. Figure 19 shows an ice-surface profile constructed to depict reasonable crustal, topographic, and glacial parameters. The profile is constructed along a flow-line that trended from Mt. Katahdin to Frenchman Bay (Fig. 1), where the late-Wisconsinan ice margin stood about 12,700 yr B.P. (Thompson and Borns, 1985a), based on field data by Borns (1973), Smith (1982, 1985), Thompson (1982), and Thompson and Borns (1985b). The ice-surface profile is conservative in estimating ice thicknesses because a "warm-base" with a 0.7 bar shear stress is assumed. The ice-surface profile is also conservative in ice thickness approximations because the continental ice sheet was far more extensive at the maximum Wisconsinan about 18,000 yr B.P., when the ice margin extended well out onto the continental shelf (Fig. 1). Thus, although the ice-surface profile reconstruction suggests that the ice sheet was about 400 to 500 m thick over Mt. Katahdin 12,700 yr B.P., the ice sheet was probably much thicker during the maximum late Wisconsinan.

More sophisticated ice-surface profiles constructed by Borns and Hughes (1977), Hughes (1981), Fastook and Hughes (1982), Hughes et al. (1985), and Hyland (1986) also suggest that Mt. Katahdin was covered by continental ice during late Wisconsinan time. In contrast, ice-surface profile reconstruc-

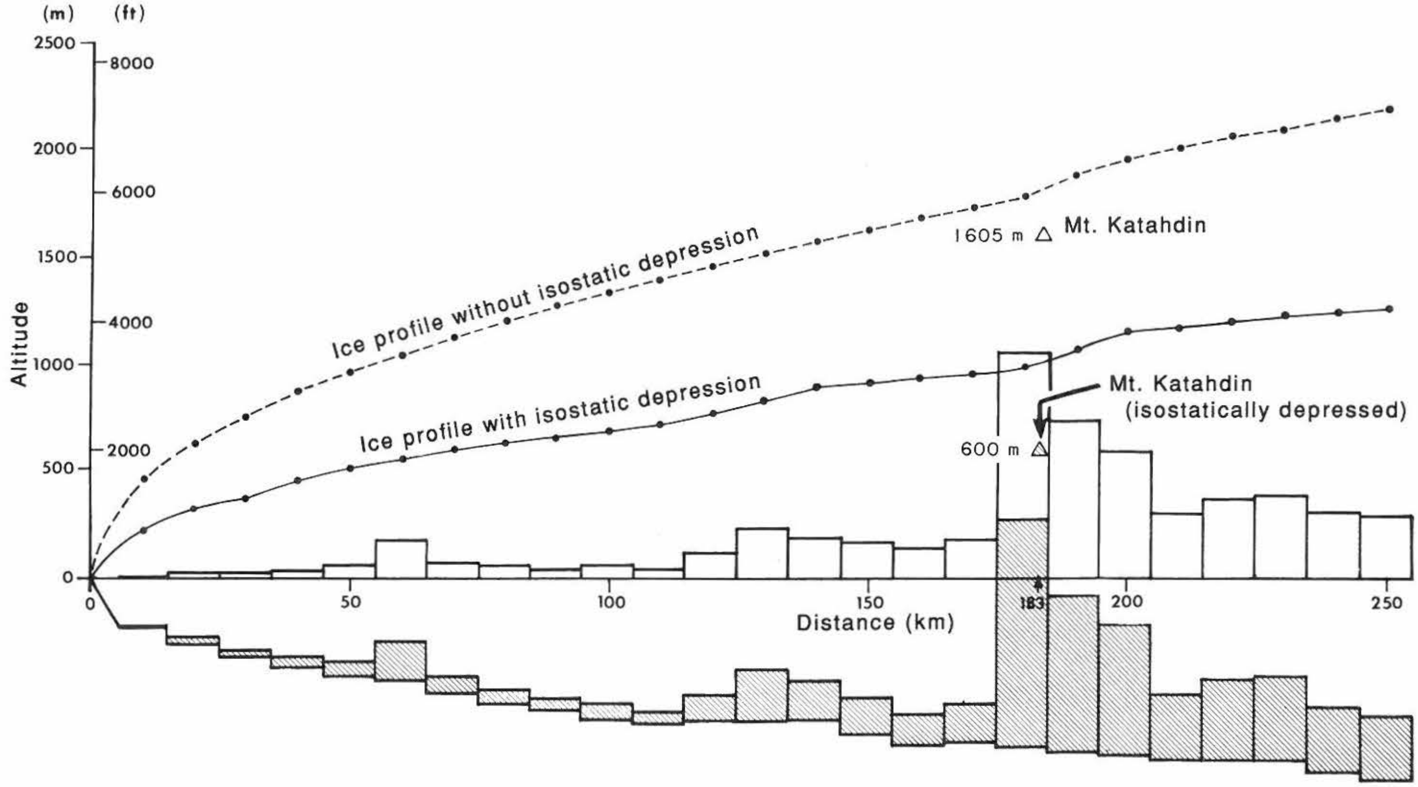

Figure 19. Late Wisconsin ice-surface profile along N20°W transect from Frenchman Bay to north of Mt. Katahdin (see Fig. 1 for location). This parabolic profile is derived from the empirical formula $h_m = 4.0\ x^{1/2}$, where h_m is the thickness of ice in meters (Nye, 1952, p. 529). This formula assumes a glacier basal shear stress of 0.7 bar, characteristic of a "warm-base" glacier. The profile also provides a correction factor for bedrock topography (Nye, 1959, p. 502). The dashed line profile and blank topographic blocks represent an undepressed crust. The solid line profile and shaded topographic blocks simulate an isostatically depressed crust. Elliptical ice-surface profiles derived from empirical formula in Paterson (1972, p. 889; 1981) also show Katahdin covered with ice during the late Wisconsin.

tions for the Laurentide ice sheet in the western Canadian Arctic (Beget, 1987) and in the middle part of the United States (Beget, 1986; Clayton et al., 1985) suggest far lower basal shear stresses than the 0.7 bars applied in Figure 19. However, in these areas, underlying bedrock and debris are more clay-rich, thereby more easily deformed or more prone to increased subglacial pore pressure (Boulton and Jones, 1979) than is terrain underlying a continental ice sheet in Maine.

Discussion

Caldwell and Hanson (1986) suggested a complex explanation for erratic stones on the summit areas of Mt. Katahdin involving weathering of xenoliths from the granophyric caprock. However, striated, faceted erratic clasts up to 30 cm long derived from a variety of sources to the north, some probably as far away as Quebec, have been found within till and block fields on Mt. Katahdin's Table Land. Many of these erratics are fresh and unweathered, suggesting a late Wisconsinan age of ice sheet glaciation. Stoss and lee forms on the uplands of Mt. Katahdin are open to various interpretations; however, granophyric bedrock and boulders are generally fresh without marked surface pitting or thick weathering rinds and, thus, are also suggestive of a late Wisconsinan age. Although block fields are usually not useful for estimating ages for deglaciation of alpine areas, on Mt. Katahdin the presence of well-developed patterned ground on a fossil rock glacier, which was certainly active during the waning phases of deglaciation, suggests that block fields may have developed rapidly during postglacial time. Moreover, soils are far too weakly developed on the Table Land to suggest any age older than the late Wisconsinan. Most theoretical ice-surface profiles constructed for Maine show Mt. Katahdin covered by continental ice from the late Wisconsinan maximum well into the deglacial cycle about 13,000 yrs B.P. Erratics found on the summit areas of Mt. Katahdin by De Laski (1872), Hamlin (1881), Tarr (1900), Antevs (1932), and Davis (1976) require "overtopping by continental ice" (Flint, 1971, p. 596). Relative dating methods which employ rock weathering and soil profile studies suggest that an ice sheet overrode Mt. Katahdin during the late Wisconsinan. Although numerical dating of deglacial events in alpine areas is difficult to interpret (Davis and Davis, 1980), theoretical models may be useful complements to field data.

SUMMARY AND THE NUNATAK HYPOTHESIS

Cirques were carved prior to the last overriding of Mt. Katahdin by a continental ice sheet. Sharpness of arêtes and steepness of cirque headwalls are not necessarily diagnostic of post-ice sheet cirque glaciation. Rather, well-preserved roches moutonnées and drift with striated, faceted erratics on cirque floors suggest that an invading ice sheet was the last ice to occupy the cirques of Mt. Katahdin. All moraines on Mt. Katahdin occur outside the cirques and mark ice sheet margins during late Wisconsinan deglaciation.

Unweathered erratics of northern provenance, generally unweathered bedrock and boulders, and weakly developed soils on the Table Land, along with theoretical ice-surface profiles, suggest that Mt. Katahdin was overridden by an ice sheet during the late Wisconsinan. The occurrence of erratics in block fields and the preservation of patterned ground on cirque floors suggest that these features may have developed during postglacial time. Thus, Mt. Katahdin does not provide any supporting geological data for the nunatak hypothesis.

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Postglacial Evolution of Drainage in the Middle and Upper St. John River Basin, Maine and New Brunswick

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ABSTRACT

Lacustrine and fluvial deposits in the St. John River valley indicate a complex postglacial history, particularly for the Siegas Terrace. Ice-dammed lakes developed during deglaciation, but gave way to drift-dammed lakes in many parts of the drainage basin. The largest of the drift-dammed lakes, Lake Madawaska, persisted until approximately 10,000 yr B.P. Peat formed between 10,000 and 9,000 yr B.P. on portions of the emerged lake bottom before being buried by alluvial deposition. Most dated abandoned-channel deposits formed between 10,100 and 7,700 yr B.P., when the St. John River channel actively migrated across the old Lake Madawaska bottom. A second period of channel instability may have occurred after 3,000 yr B.P. Dates from overbank sand and silt show deposition continued on the Siegas Terrace through 800 yr B.P. No mappable alluvial surface occurs above the Siegas Terrace, hence the highest terrace along this reach of the river is late Holocene in age.

The major controls acting on the postglacial evolution of the drainage basin probably were isostasy, climate, and the nature of cross-valley drift and bedrock obstructions. We speculate that these controls influenced other drainage basins in the region, but to different degrees and with different effects. It is likely that postglacial events varied greatly from basin to basin. While there has been little study of other alluvial deposits in northern New England and adjacent Canada, the complex history of the upper and middle St. John River suggests that systematic, basin-wide studies of alluvium in other river basins should be quite rewarding.

INTRODUCTION

Floodplains and terraces represent a badly neglected part of the late Quaternary history of Maine. For a number of reasons, most surficial geologists working in the state aim their research in other directions, such as glacial geology or sea-level change. The lack of work on postglacial alluvial events should not be attributed to the absence of a significant record, because alluvial deposits are well developed along many reaches of Maine rivers. Most of these deposits contain abundant organic material for radiocarbon dating, but the events they record will be deciphered only through systematic, basin-wide studies of alluvial stratigraphy and geomorphology.

This paper reports on the first study to concentrate on the postglacial alluvial events in northern New England and adjacent Canada. The study was conducted in the St. John River basin, upstream from Grand Falls, New Brunswick (Fig. 1), where a complex postglacial history is recorded by a variety of deposits.

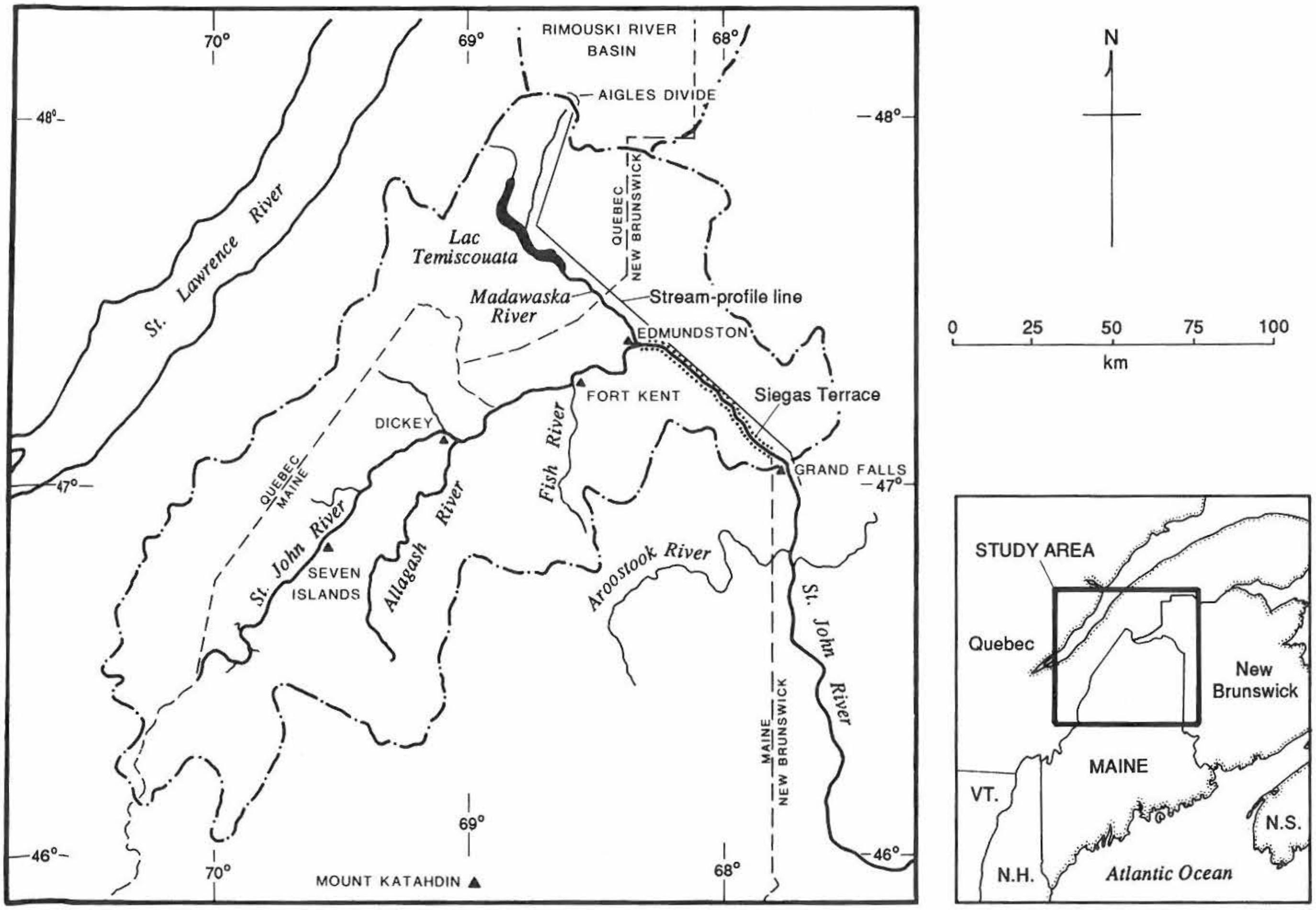

Figure 1. Selected locations in the St. John River basin. Profile line refers to position of transect in Figure 3. Figure 4 shows the location of the basin relative to eastern Canada and northeastern U.S.A.

The purpose of this paper is to arouse interest in studies of alluvial deposits throughout the region by reporting the postglacial history of the St. John River and discussing some of the most important controls influencing the evolution of drainage.

Much of the field data used in this study was obtained during the Maine Geological Survey's surficial geology mapping of various 15 minute quadrangles in northern Maine. Field and laboratory methods are discussed at length by Kite (1979, 1983). In general, field work consisted of automobile, foot, or canoe traverses, and sampling and description of stratigraphic sections exposed by stream erosion or man-made excavations. Interpretation of soil-survey maps (Arno, 1964) and vertical aerial photos aided field mapping. Particle-size analyses were conducted in the Department of Geology and Geophysics, University of Wisconsin at Madison. New radiocarbon dates were obtained from 36 samples from the field area, supplemented by seven previously published dates from the basin. Radiocarbon dating was performed at the Smithsonian Institution Radiation Biology Laboratory in Rockville, Maryland.

STRATIGRAPHY AND GEOMORPHOLOGY

The study area lies within the St. John River Lowlands, a small physiographic unit with low mountains and hills, generally less than 500 m in altitude (Denny, 1982). A varied array of sediments and landforms occur in these lowlands (Fig. 2). Although ice-contact landforms are common in large valleys, the widest reaches of these valleys are dominated by a dated sequence of lacustrine deposits, peat, and alluvium. The best developed postglacial landform is the Siegas Terrace, which occurs between Edmundston and Grand Falls (Fig. 1). The Siegas Terrace is the highest postglacial fluvial surface in this reach, occurring 9-12 m above typical summer river levels.

Many exposures of sediments in the Siegas Terrace and other terraces reveal up to 10 m of gray lacustrine silt that is virtually unfossiliferous and displays rhythmites 0.2 to 3.0 cm thick. Some of these lacustrine rhythmites are overlain by till and predate the last glaciation (Prescott, 1973; Lowell and Kite, 1986a), but a significant thickness of the rhythmites is

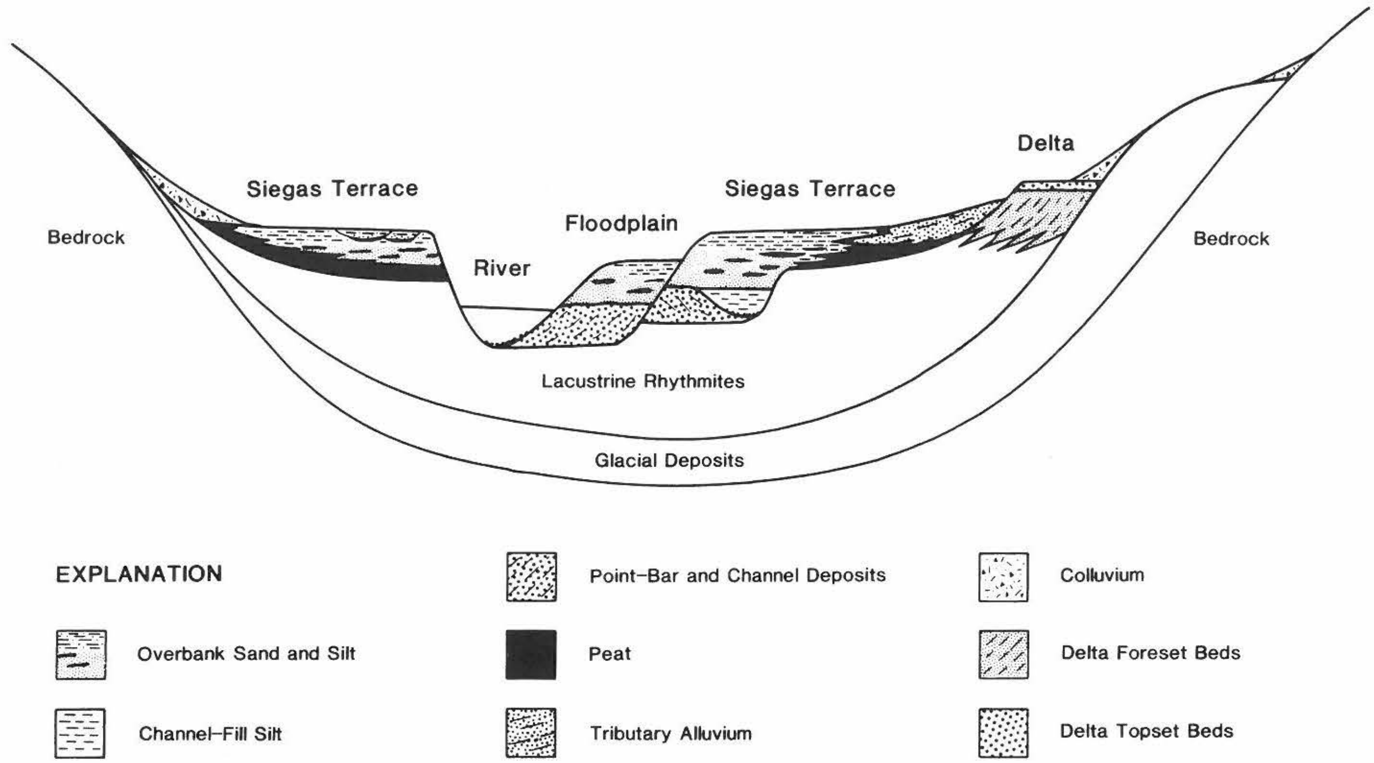

Figure 2. Idealized cross-section through the Siegas Terrace. See Figure 1 for general location of the terrace. Not to scale.

TABLE 1. MAJOR FACIES IN THE SIEGAS TERRACE STRATIGRAPHY.

Facies	Characteristics	Facies Codes (after Miall, 1982)
River-channel and point-bar gravel	Sandy pebble to cobble gravel; commonly cemented by iron and manganese oxides at contact with over-lying channel-fill silt; horizontal bedding and imbrication or planar crossbeds; sand beds dominant in some outcrops. (1.3 to 5 m thick.)	Gm, Gp, Sp
Channel-fill silt	Gray silt to clayey silt; abundant plant fragments and bivalve shells; poorly developed bedding, except at base of one late Holocene deposit; mottles and vivianite replacing wood fragments common. (1.0 to 4.6 m thick.)	Fcf
Overbank silt	Fining upward sequences of sand and silt; abundant wood fragments and thin organic lenses; inverse-graded, minor Sr, Ss, Sp, and normally graded bedding; extensive bioturbation; scour features common near floodplain/terrace surface. (Up to 7.0 m thick.)	Fl, Fm, Sl

demonstrably postglacial. A conformable contact with overlying postglacial peat or alluvium occurs at many exposures near the valley sides, but several exposures near the valley center show an erosional unconformity between the lake silts and overlying alluvium. Nearly all exposures of lacustrine sediments occur upstream from thick cross-valley drift accumulations that probably served as dams for these lakes. Small sand-and-pebble beach deposits have been used to reconstruct lake levels in areas without forest cover (Kiewiet de Jonge, 1951; Kite, 1979), although beaches are not common in the upper St. John River basin.

The sedimentary character of alluvium along the St. John River depends upon the type of material from which it was derived, and upon the alluvial facies in which it was deposited. Three facies dominate the Siegas Terrace alluvial stratigraphy: river-channel and point-bar gravel, channel-fill silt, and overbank silt and sand (Table 1; Fig. 3). Old gravel deposits in the Siegas Terrace are somewhat finer grained than adjacent modern St. John channel alluvium, but tributary channel alluvium appears not to have varied in particle size. The best-exposed alluvial unit is gray channel-fill silt, analogous to modern-day abandoned-channel (oxbow) deposits. Channel-fill silt is very

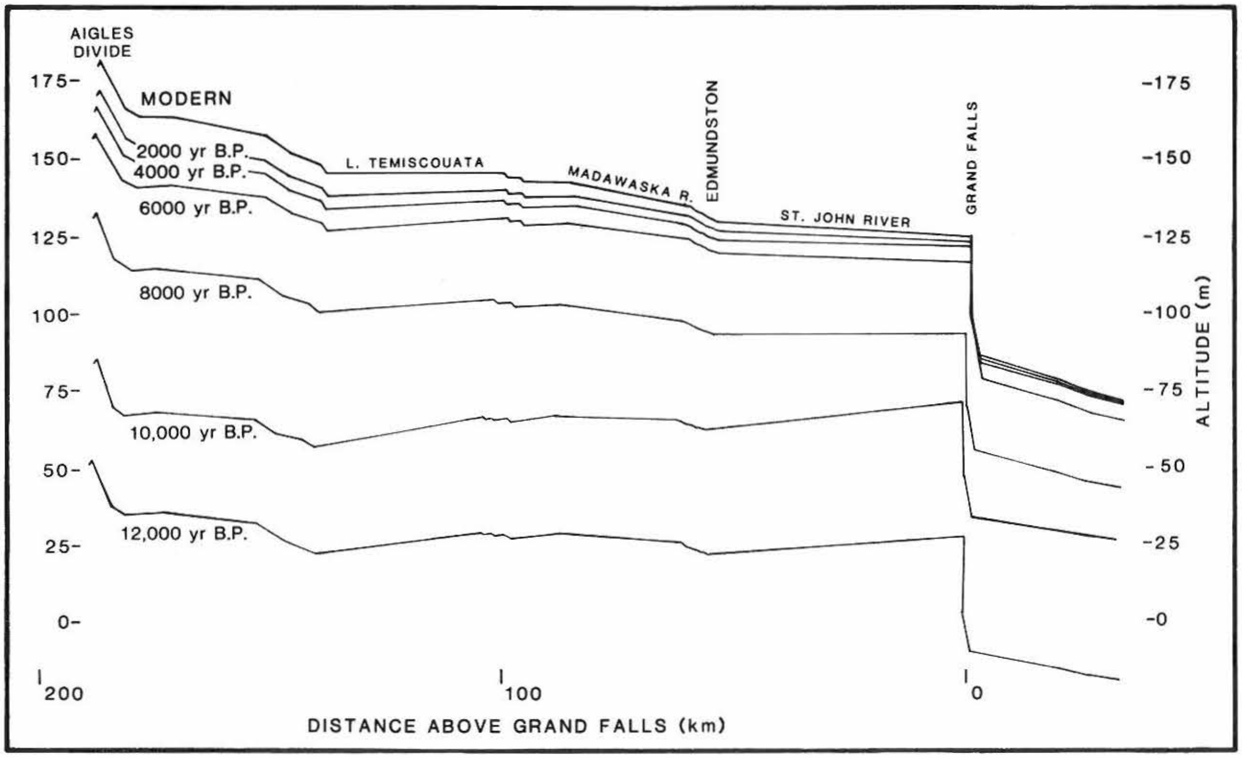

Figure 3. Stream-profile reconstruction: Grand Falls to the Aigles divide, based on regional isobase reconstructions of Andrews and Tyler (1977). Figure 1 shows position of the profile line. Although considerable rebound occurred between 12,000 and 10,000 yr B.P., maximum differential rebound along this transect occurred between 10,000 and 8,000 yr B.P. Isobase reconstructions by Quinlan and Beaumont (1982) yield similar stream-profile reconstructions.

similar to lacustrine silt, except that it has abundant plant fragments and bivalve shells, and has less well-developed bedding. Overbank sand and silt includes abundant wood fragments and thin organic lenses. Over half of the beds in overbank deposits show inverse grading, but soil-profile development and bioturbation has obliterated most bedding in the silty upper 1 m of Siegas Terrace alluvium. The total thickness of postglacial alluvium in the Siegas Terrace ranges from 4 to 10 m.

GEOLOGICAL HISTORY

Forty-three radiocarbon dates allow assignment of absolute ages to postglacial events in the upper St. John River basin. Most dates were obtained from organic sediments within alluvial deposits or peat in the Siegas Terrace. Kite (1983) and Kite and Stuckenrath (1986) give descriptions of the stratigraphic setting and the nature of each sample used for radiocarbon dating.

After deglaciation, ca. 12,000 to 11,000 yr B.P., drift-dammed lakes developed in many valley reaches previously occupied by ice-dammed lakes (Kite et al., 1982; Kite, 1983; Lowell and Kite, 1986a). Radiocarbon dates obtained from lacustrine sediments in the Siegas Terrace indicate that the largest of these lakes, Lake Madawaska, lasted from deglaciation until ca. 10,000 yr B.P. If rhythmic laminae in sediments of other now-extinct lakes are annual in origin, these lakes persisted for lengths of time comparable to Lake Madawaska.

Alluvial deposition did not begin in much of the St. John River valley until after the drift-dammed lakes drained. In at least two localities, peat beds accumulated on the emerged Lake Madawaska bottom for 500 to 900 years before alluvial aggradation began. However, numerous channel-fill silt lenses formed between 10,100 and 7,700 yr B.P., indicating that active channel migration coincided with early Holocene alluviation on the exposed lake bottom.

Middle Holocene alluvium is not as well dated as earlier and later deposits. No channel-fill silt lenses have been found that date between 7,700 and 3,000 yr B.P., suggesting that overbank deposition dominated the middle Holocene. In addition, differences of up to 4,500 years between radiocarbon dates from the same overbank horizon make it difficult to assess the true age of many overbank deposits and indicate that much of the floodplain alluvium has been reworked. If dates from apparently reworked alluvium are ignored, calculated sedimentation rates were nominally lower during the middle Holocene than during the early or late Holocene (Kite, 1983).

Three channel-fill lenses date after 3,000 yr B.P., although late Holocene silt lenses are less common than analogous early Holocene deposits. The dated organic sediments with the

highest elevation show that deposition occurred on the Siegas Terrace after 800 yr B.P. This late Holocene deposition on the highest alluvial terrace suggests an extremely large flood, at least 4 m above the level of the highest historical flood (6,400 m^3/s). Shallow erosional channels on much of the surface of the Siegas Terrace may have formed in such a flood. If the channel and flood-plain configuration of the St. John River has not changed significantly since this flood, discharge may have exceeded 18,000 m^3/s) during this large flood. However, the validity of this discharge estimate is dubious because the event may have been the result of ice-jam flooding, in which high water levels may not correlate directly with high discharge. Moreover, the ages and elevations of dated abandoned-channel lenses suggest that an increase in channel-incision rates occurred during the late Holocene, so flooding of a magnitude similar to modern floods may have deposited the high-level alluvium prior to accelerated channel incision during the last 800 years (Kite, 1983).

CONTROLS OF POSTGLACIAL DRAINAGE EVOLUTION

The major controls acting on the postglacial evolution of the drainage basin probably were extrinsic factors, such as climate and glacially-induced isostasy, and intrinsic factors such as the geometry, structure, and distribution of cross-valley drift or bedrock obstructions. The glacial geology of the study area shows that the upper and middle St. John River basin was last glaciated by a residual ice cap that developed during disintegration of the Laurentide Ice Sheet (Hughes et al., 1985; Thompson and Borns, 1985; Lowell and Kite, 1986b). Most radiocarbon dates obtained from the oldest organic sediment in lake bottoms within or near the St. John River basin are between 10,900 and 11,700 yr B.P. (Davis and Jacobson, 1985), but the time lag between actual deglaciation and the onset of organic accumulation in these lakes may be substantial (Davis and Davis, 1980). These basal lake-sediment dates are consistent with the glacial chronology established in coastal Maine and the St. Lawrence Lowlands, so we assume that the lake-bottom dates are close approximations of the onset of ice-free conditions. It appears that glacial ice persisted until ca. 10,000 yr B.P. at high elevations within 250 km of the St. John River basin (David and Lebuis, 1985), but it is unlikely that the relatively low mountains in the drainage basin were able to support glaciers after disintegration of the residual ice cap.

Assuming deglaciation occurred between 12,000 and 11,000 yr B.P., both isostasy and climate in the St. John River basin would have been strongly influenced by the persistence of the Laurentide Ice Sheet in the St. Lawrence Valley, only 50 km to the northwest. Reconstructions of isobases throughout the region (Andrews and Tyler, 1977; Quinlan and Beaumont, 1982) suggest the weight of the nearby ice sheet delayed some of the rebound in the western headwaters of the drainage basin until after 10,000 yr B.P. Although basin-wide isostatic rebound was greater immediately after deglaciation, differential rebound between the southeastern and northwestern ends of the drainage basin altered stream gradients more between 10,000 and 8,000 yr B.P. (Fig. 3). Streams and valleys trending perpendicular to isobases would experience the greatest changes in gradients because of differential rebound at this time. Accordingly, the competence and capacity for sediment transport in these reaches, which are well above the marine limit, were influenced by isostatic rebound long after deglaciation. Moreover, the delayed differential rebound probably caused westward shifting of some drainage divides (Kite, 1983).

The climate of the St. John River basin at 11,000 yr B.P. was significantly different from that of today. The landscape was dominated by cold tundra (Davis and Jacobson, 1985) associated with periglacial conditions (Roy and Lowell, 1980; Kite, 1983). Clearly, the landscape was cold relative to today; however, because of the lack of direct or indirect field evidence, reconstructing the former moisture regime is difficult. Modern analogy shows that high-discharge floods are rare in tundra environments of eastern North America. If streams in the modern tundra environment are good analogs, then streams in the upper St. John basin probably lacked sufficient energy to erode large drift dams until after 11,000 yr B.P.

A plausible reconstruction of the Polar Front (dominant summer storm track) at 11,000 yr B.P. suggests that moist maritime tropical air masses would seldom extend into the St. John River basin (Fig. 4). The Polar Front was situated farther south, constrained by the ice sheet to the west and the cold meltwater-laden currents of the late Pleistocene Labrador Current to the east. Under this scenario, the early postglacial St. John River basin was removed from areas of heavy winter snowfalls, intense mid-latitude cyclones, or tropical cyclones that have caused all of the major historic floods in the basin. At 11,000 yr B.P., small tributaries would have been sites of colluviation and associated periglacial alluviation, but the absence of large floods left tributary streams unable to transport much of the colluviated sediment into larger streams.

Many of the variations in the pollen record for New England and adjacent Canada since 10,000 yr B.P. can be attributed to diseases and the arrival of immigrating species (Davis, 1976; Davis et al., 1980), but some climatic variations have been determined. Davis et al. (1980) suggest that the climate of the White Mountains in New Hampshire between 9,000 and 5,000 yr B.P. was 2°C warmer (mean annual temperature) and 400 mm drier (mean annual precipitation) than at present. Although the precipitation estimate seems too low for the St. John basin (400 mm represents a greater than 40% decrease in precipitation), it is probable that a warm, dry episode occurred in the study area during the early to middle Holocene. If so, larger streams would have experienced smaller floods because of less winter snowfall and fewer mid-latitude cyclones; small tributaries may have experienced relatively large floods produced by local air-mass thunderstorms.

A final climatic consideration concerns the rate at which ice-cored drift obstructions melted in a tundra environment.

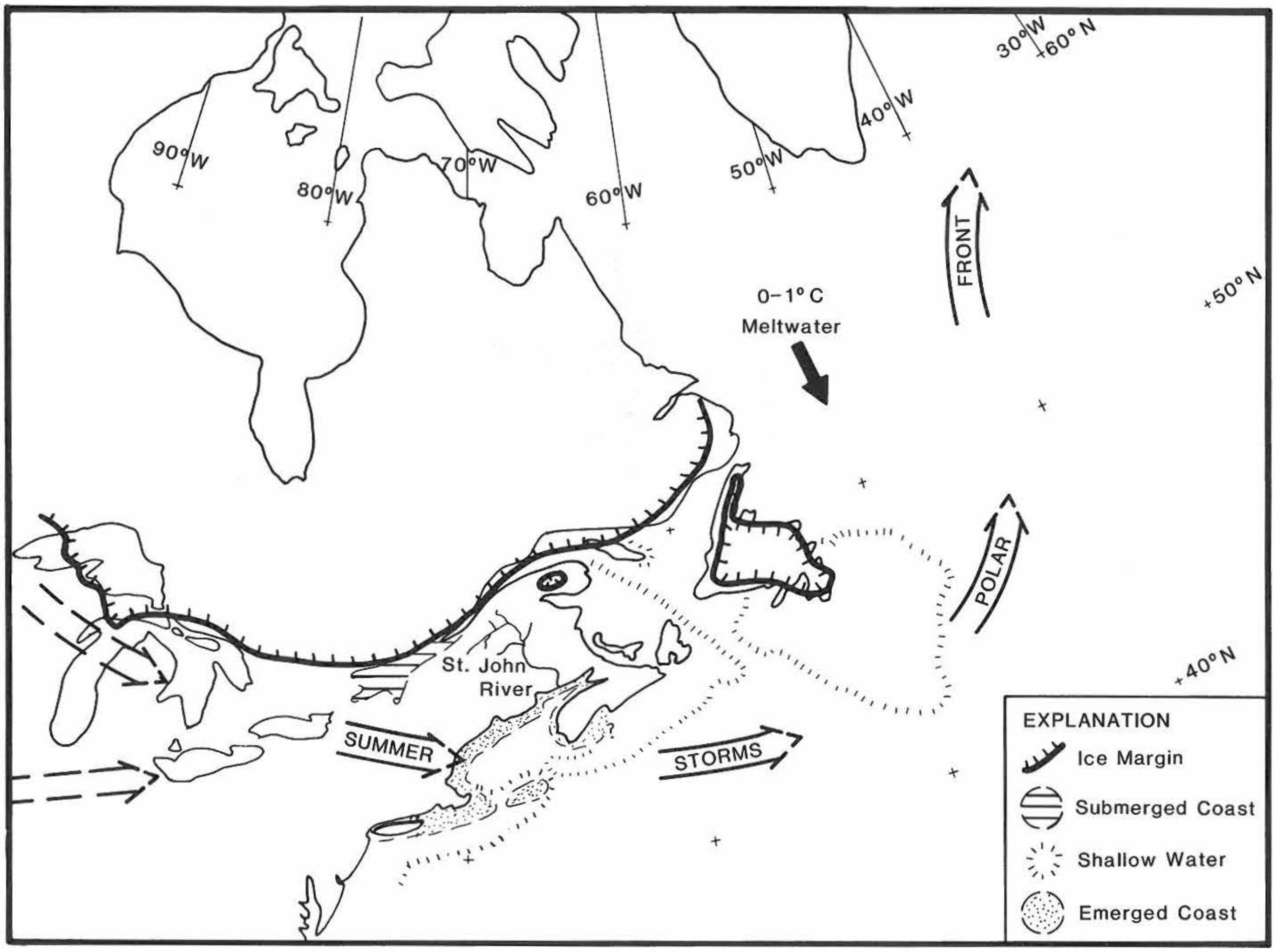

Figure 4. Likely position of the Polar Front (dominant summer storm track) at 11,000 yr B.P. (after Kite, 1983). Position of the Polar Front follows Ruddiman and McIntyre (1973) over the Atlantic Ocean, but follows the southern limit of tundra over New England and adjacent Canada, as shown by Davis and Jacobson (1985). Ice margins follow David and Lebuis (1985) for the Gaspé Peninsula and Dubois and Dionne (1985) for the rest of southern Quebec.

Ice-cored drift may have persisted for centuries (or longer) after the disappearance of glaciers (Ostrem, 1965). This ice-cored drift may have clogged major drainageways and retarded the evolution of the fluvial system.

With or without ice cores, numerous drift accumulations clogged the St. John River and its major tributaries after deglaciation. In general, the history and original character of these drift obstructions is poorly understood. Mapping of eroded remnants suggest that many of these drift accumulations typically had considerable length and included large boulders that could armor outlet channels and protect them from incision. Moreover, although drift accumulations blocked the eventual paths of many rivers in the region, the initial postglacial drainage in some of these rivers was temporarily routed through other channels, commonly floored by bedrock. Accordingly, we hypothesize that many drift dams persisted for hundreds or thousands of years after deglaciation.

The down-cutting St. John River has encountered many bedrock obstructions, creating rapids or waterfalls that act as local base levels. The most important of these knickpoints is the bedrock sill at Grand Falls, where a drift-floored channel was abandoned before 9,800 yr B.P., in favor of a bedrock-floored channel (Kite and Borns, 1980; Kite et al., 1982). Today, the St. John River cascades 23 m at the falls and drops another 15 m in the 1.5 km long gorge downstream. Upstream from the falls, the gradient of the river is less than 0.15 m/km. The configurations of the channel and the bedrock sill are such that if the falls retreat a few hundred meters upstream, then the sill would be completely eroded. In turn, this event would lower local base level by at least 15 to 20 m and trigger severe erosion of the Siegas Terrace as far upstream as the next shallow bedrock surface at Edmundston.

It appears that a similar bedrock sill occurred on the Aroostook River, near its confluence with the St. John (Bailey, 1894).

Thick silt rhythmites, similar to Lake Madawaska sediments, occur in the Presque Isle area (Prescott, 1972), upstream from a deeply incised bedrock gorge on the lower Aroostook River. These silt rhythmites indicate that the now-eroded bedrock obstruction blocked drainage long enough for substantial lake sediments to accumulate.

IMPLICATIONS FOR OTHER BASINS IN THE REGION

The complex history of postglacial drainage evolution in the middle and upper St. John River basin may not be representative of events in other upland basins in New England and eastern Canada, because the relative importance and timing of controlling factors differed. Knowledge of Holocene alluvial events in these regions is extremely limited, and we can only speculate about which controls were most important.

Glacial and bedrock geology are two sources of great variability from one drainage basin to the next. Deglaciated landscapes are not in equilibrium with fluvial processes, so the alluvial histories of many streams in the region have been dominated by adjustments to the fluvial system. One of the most striking aspects of rivers in the region is that very few truly alluvial reaches exist. Most stream reaches have not yet reached grade, nor will they for many future millennia. Many Holocene adjustments are straightforward, such as erosion of bedrock knickpoints and drift obstructions or deposition in glacially scoured lake basins. Other adjustments to the fluvial system are more complex, such as bed-load transport in channels flowing over a landscape with great variability in available clast sizes. Anyone studying Holocene alluvial events in New England and adjacent Canada must be familiar with the glacial history and the distribution of glacial deposits in their study area.

In general, climatic controls would have acted simultaneously on adjacent fluvial systems, but the apparent time-transgressive nature of climatic and floral events (Davis and Jacobson, 1985) suggest that events might be nonsynchronous over large regions. Moreover, widespread climatic events would produce varied and complex responses in the diverse glaciated drainage basins of the region (Patton, 1981). Different types of flood events dominate drainage basins with different size and runoff characteristics, so one should not expect adjacent streams to have identical Holocene histories. For example, the frequency of intense local thunderstorms will be much more important to the evolution of small drainage basins with little runoff storage capacity than to large basins or basins with substantial runoff storage in lakes or wetlands.

Isostatic controls acted simultaneously in adjacent fluvial systems, but certainly they did not always produce identical responses. Streams oriented parallel to northeast-southwest trending isobars experienced little direct effect from differential isostatic rebound. In contrast, gradients on southeast-draining streams have steepened throughout the Holocene, while those on northwest-draining streams have decreased. The maximum changes in gradient were probably no more than 0.4 m/km (Kite, 1983), an insignificant amount on most reaches of small streams, but a substantial change on many reaches of large rivers like the St. John.

Isostatic and eustatic sea-level changes have been major controls of alluvial chronologies in reaches of streams at or below the marine limit. An upstream influence of sea-level controls can be documented at many outwash deposits that were graded to relict deltas or other shoreline features. Caution must be exercised in extending sea-level controls far inland because the effects of sea-level fluctuation commonly disappear upstream from the first major bedrock knickpoint above the marine limit. Postglacial sea-level changes have been too rapid and too short-lived for their direct effects to reach the upper reaches of large basins.

In conclusion, the rich postglacial history of the middle and upper St. John River basin indicates that fluvial systems in the region did not assume a modern character immediately after deglaciation. We do not know whether specific events in the St. John River were typical of other upland basins in the region, but we strongly believe that the under-researched landforms and stratigraphy in other river basins also record complex and significant postglacial histories. Unfortunately, very little of the postglacial evolution of drainage can be sorted out until surficial geologists in the region stop treating alluvium as overburden obscuring the glacial geology. Postglacial alluvium merits much more attention than it has received to date.

ACKNOWLEDGMENTS

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