Retreat of a tidewater margin of the Laurentide ice sheet in eastern coastal Maine between ca. 14 000 and 13 000 ^{14}C yr B.P.

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ABSTRACT

Ice-margin retreat, glacial processes, and associated marine environmental conditions were investigated along a tidewater margin of the Laurentide ice sheet in eastern coastal Maine. The ubiquitous presence of recessional moraines that typically show crosscutting relationships indicates that numerous stillstands and readvances interrupted net ice-margin retreat. Commonly, moraines exhibit changes in orientation associated with elevation changes, suggesting topographic control on retreat. Till, glaciofluvial facies, and glaciomarine sediment are the dominant components of the moraines. These sediments show that the ice terminated in the sea and that glacial meltwater played a key role in their deposition. The most continuous landforms in the study area are the Pond Ridge moraine and Pineo Ridge moraine system. Both landforms crosscut older moraines and contain ice-shove features, and crosscutting of moraines exists within the Pineo Ridge moraine system, indicating that the ice sheet readvanced during their formation. The Pond Ridge moraine and Pineo Ridge moraine system are straighter across elevation changes, compared to other moraines in the area, implying that during their formation, topography had less of a destabilizing influence on ice-margin dynamics. Foraminifera, Ostracoda, and mollusk species in ice-proximal and ice-distal glaciomarine sediments indicate that arctic to subarctic climatic conditions existed during deglaciation. Despite the persistence of a cold environment, 20 km of net ice-margin retreat occurred between ca. 14 000 and 13 000 ^{14}C yr B.P., mostly in contact with the sea.

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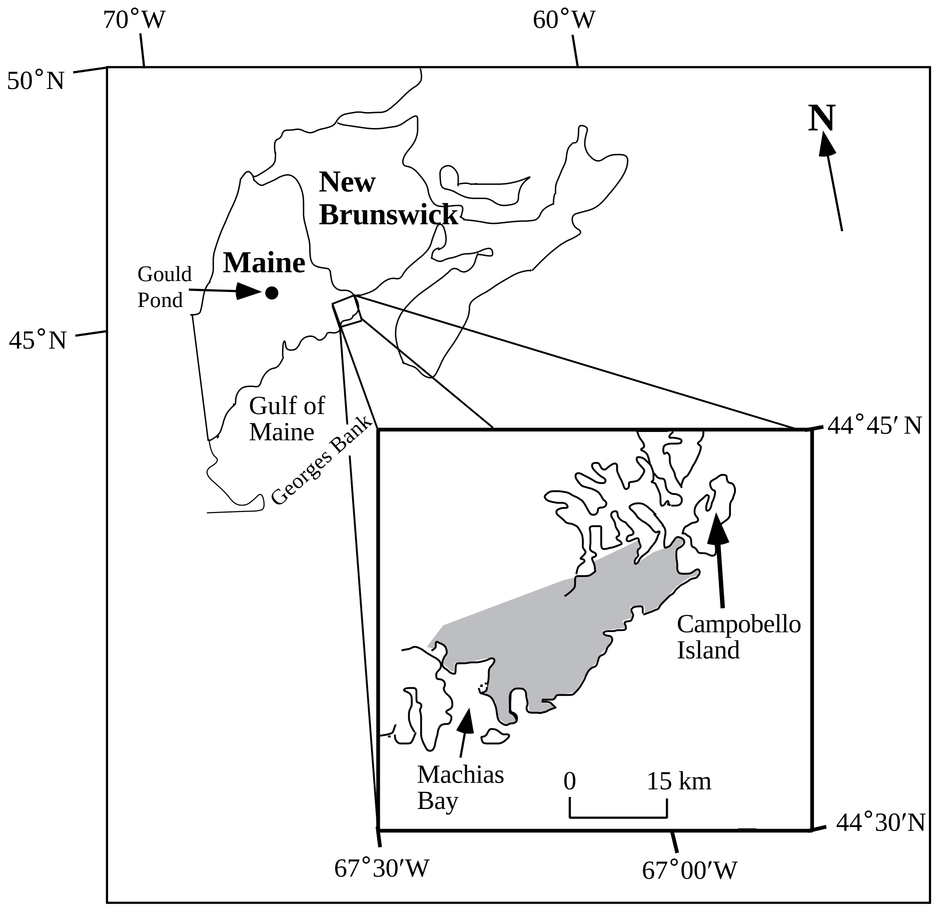


Figure 1. Location of the study area (shaded) in eastern coastal Maine.

INTRODUCTION

Although the basic paleogeography and chronology of deglaciation are reasonably well known for eastern Maine, questions still remain concerning the nature of ice-margin retreat, associated marine paleoclimatic conditions, and factors that may have affected grounding-line dynamics. The high volume of glaciofluvial sediment within the depositional landforms of eastern coastal Maine has been interpreted by many as forming in association with retreat of a temperate, tidewater ice margin (e.g., Ashley et al., 1991). However, there has been a lack of microfaunal or other biotic data from which to infer paleoenvironmental conditions during deglaciation of the coastal region. The conclusions for the present study were based on detailed mapping of glacial retreat along a 40 km section of the Laurentide ice sheet in southeastern Washington County, eastern coastal Maine (Fig. 1). The mapping included analyses of microfauna from ice-proximal and ice-distal deposits in order to reconstruct the associated marine environmental conditions that may have influenced grounding-line dynamics. Previous mapping efforts in the study area were done only in reconnaissance studies (e.g., Leavitt and Perkins, 1935; Borns, 1980; H. W. Borns, Jr., 1994, personal commun.) or focused on specific features (e.g., Pond Ridge moraine, LePage, 1982; Dorion, 1993).

Research in the past two decades has shed light on the dynamics of modern tidewater glaciers (e.g., Powell, 1991; Meier and Post, 1987; Warren, 1992; Powell and Domack, 1995). Although an understanding of these glaciers and their role as a modern analogue for the Northern Hemisphere ice sheets is far from complete, a critical conclusion reached so far is that the dynamics of tidewater glaciers (and therefore marine-based ice sheets?) are a function of several factors, including climate. For example, tidewater glaciers can exhibit unstable behavior that is probably related to basal conditions and fjord geometry (Meier and Post, 1987; Hughes, 1992). Many parts of the eastern Laurentide ice sheet were marine based (i.e., the base of these parts was below relative sea level) and terminated in the North Atlantic Ocean during the last deglaciation. This raises the question, What were the factors that affected retreat of the tidewater margins of the Laurentide ice sheet? The answer will provide insight into the dynamics of large ice sheets, both past and present, and their role in the climate system.

Eastern coastal Maine provides a useful context for the study of interactions between ice sheets and oceans because (1) the study area was covered by a marine transgression during retreat, providing emerged marine deposits and mollusk shells for establishing an AMS (accelerator mass spectrometry) ^{14}C chronology; (2) glacial erosional and depositional landforms are abundant, including linear moraines; (3) moraines provide favorable conditions for blueberry barrens, and thus the locations and morphology of the moraines are commonly easy to delineate; (4) the internal structure of many landforms can be observed within borrow pits; and (5) fresh exposures of the internal structure of surficial units including moraines, glaciomarine clay, and bedrock with glacial erosional features are found along the coastline.

BACKGROUND

Location and Bedrock Geology

The western boundary of the study area is approximately the western side of Machias Bay; the northern boundary is about lat 44°48′N; the eastern edge is the western side of Campobello Island; and the southern side is the coastline (Fig. 1). The bedrock in this area consists of crystalline (e.g., basaltic and rhyolitic flows and tuffs, diabase and gabbro dikes, sills, and irregular small plutons), and sedimentary rocks such as siliceous argillite, shale, and siltstone (Gates, 1975). The bedrock and vertical shear foliation (fracture cleavage) strike between north-northeast and east-northeast. Faults within the area trend either northwest or northeast.

Quaternary Geology

The margin of the Laurentide ice sheet started to retreat from Georges Bank soon after the Last Glacial Maximum (marine isotope stage 2) (Pratt and Schlee, 1969; Borns, 1973; Bacchus, 1993). The margin reached the present coastline of the study area at ca. 14 000 yr B.P. (Schnitker, 1975; Stuiver and Borns, 1975; Thompson and Borns, 1985; Smith, 1985; Smith and Hunter, 1989; this study). The glacial history of the study area is attributed to the Laurentide ice sheet, prior to the development of an independent Appalachian ice cap to the north (Lowell, 1985), on the basis of the deglacial chronology presented in this paper and that in other studies (e.g., Thompson and Borns, 1985; Smith and Hunter, 1989).

Deglaciation of coastal Maine was associated with a marine transgression because the region was isostatically depressed (Bloom, 1960; Belknap et al., 1987). Elevations of topset-foreset contacts in ice-proximal glaciomarine deltas and wave-washed and/or eroded features located throughout coastal Maine indicate that sea level was ~50–60 m asl (above present sea level) during deglaciation of the study area (Thompson et al., 1989). North of the study area, the marine limit—i.e., the maximum extent of the transgression—was ~75 m asl (Thompson and Borns, 1985). Relative sea level along the retreating ice-sheet margin was the result of the interplay between local isostatic rebound and global eustatic sea-level rise. Isostatic rebound appears to have been dominant along the retreating margin of the eastern Laurentide ice sheet, causing relative sea level to fall (Belknap et al., 1987; Thompson et al., 1989; Kelley et al., 1992; Barnhardt et al., 1995; Gray 1996). Despite a falling relative sea level, water depths along the glacial margin would have varied depending on the rate of isostatic rebound, eustasy, and topography. Most present-day elevations in the study area are less than 50–60 m asl and thus would have been below sea level during deglaciation. A few of the highest elevations of the study area are ~75 m asl (one bedrock hill reaches 100 m), and these sites probably would have been slightly above sea level at the time. The average difference between present and inferred former sea level during deglaciation indicates that water depths along the margin would have been less than 50–60 m. Locally, water depths probably were greater in low-lying areas such as Machias Bay. After deglaciation, continued isostatic uplift resulted in a subsequent regression with a lowstand occurring offshore at ca. 11–10.5 ka (–55 m; Barnhardt et al., 1995). Currently, the rate of eustatic sea-level rise dominates over the rate of isostatic uplift along all of coastal Maine (Belknap et al., 1987; Gehrels and Belknap, 1993; Barnhardt et al., 1995).

METHODS

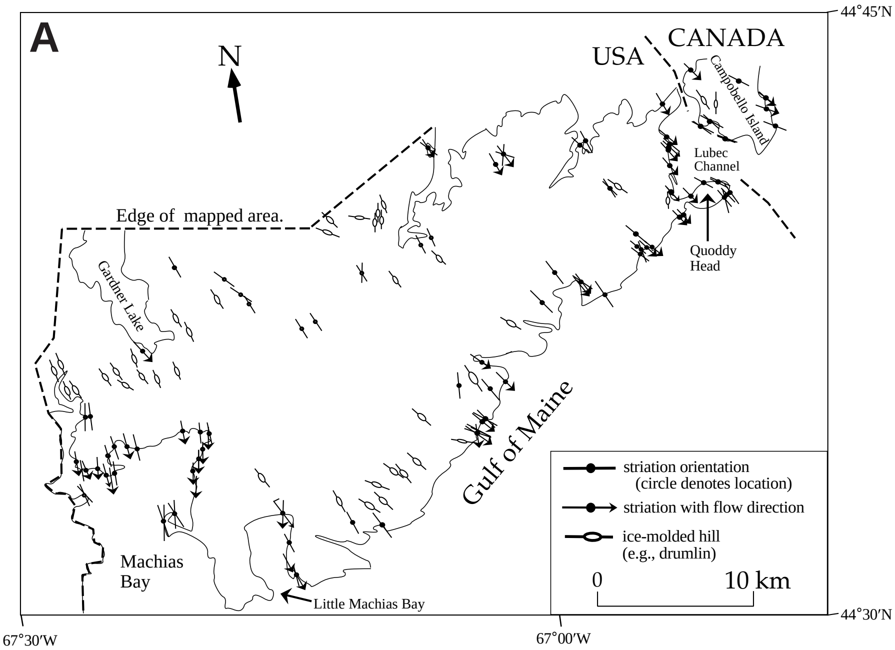
Methodology included mapping the surficial geology, obtaining AMS ^{14}C dates on marine shells from relevant stratigraphic sections, and analyzing marine faunal assemblages from ice-proximal and ice-distal sediments (Kaplan, 1994). Field mapping in 1993 was coupled with detailed aerial-photograph interpretation to compile a surficial geologic map. Mapping was done by using 1:10 000–1:14 000 stereo black and white aerial photographs and 1:24 000 U.S. Geological Survey topographic quadrangles. Most recorded striae were measured on the center of the stoss side of stoss-and-lee features. With rare exception, the crest along the stoss side is oriented in the same direction as the superimposed striae (within ±5°). Conditions under which moraines formed in the study area were inferred from the examination of several exposures. Most of the dated marine fossils were found in eroding coastline exposures. In addition, a core from Lily Lake was obtained by using a 1 m Wright square-rod piston corer (Wright et al., 1984). AMS dates were obtained on individual shells (the only exception is the date from the Lily Lake core, which was on several mollusk shells). Most dates presented in this paper are from samples found in situ, with both valves connected and periostracum attached. Finally, all ^{14}C dates reported here are uncorrected for the marine-reservoir effect because it is unknown for this region during deglaciation. Errors on reported dates may be at least 400–500 yr, the typical value used for samples from the North Atlantic Ocean (cf. Bard et al., 1994).

RESULTS

Ice-Flow Indicators

The erosive imprint of glaciation on the landscape was found useful for inference of processes occurring at the former grounding lines. Ice-flow indicators in the study area are concentrated in local areas such as along Machias Bay and the

Lubec Channel (Fig. 2A). The dominant striae orientation for the entire study area indicates movement toward the southeast (Fig. 2B); almost all streamlined hills also indicate southeastward ice flow (Fig. 2B). Most outcrops exhibit only one direction of ice flow, but locally as many as two different directions are indicated. For example, three sites of superimposed striae in the southwestern part of the study area clearly indicate that southward flow succeeded more southeastward flow (Fig. 2A; also see Fig. 7). Other striae that do not trend southeast are found along the eastern coastline of Machias Bay, where they indicate southward ice flow (Fig. 2A; also see Fig. 7). Striation directions differ around Quoddy Head, a relatively high area (Fig. 2A); the azimuth changes from more eastward (e.g., S76°E, S60°E) on the northeastern side to more southward (e.g., S42°E, S25°E) on the southeastern side.



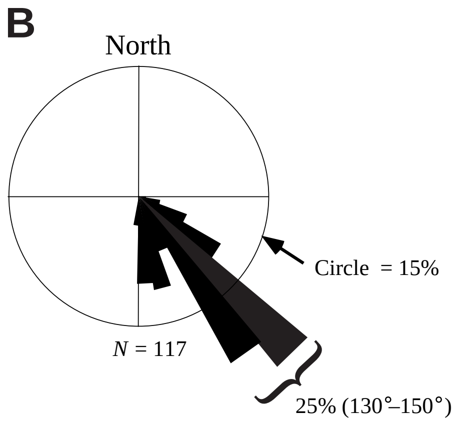


Figure 2. (A) Locations of mapped striae. (B) Rose diagram showing all striae directions measured in the study area. Because there is no evidence that indicates ice flow from the Gulf of Maine toward the northwest onto land, this diagram assumes flow toward the southeast.

Moraines

The most obvious positive glacial landforms in the study area are distinctly linear moraines that formed transverse to ice flow (Figs. 2 and 3). Mapping presented here (Fig. 3) shows a far greater abundance of moraines than that in prior reconnaissance work (e.g., Borns, 1980). The new mapping clarifies the spatial extent and nature of specific features previously studied (e.g., Pond Ridge moraine, LePage, 1982). The geographic pattern of ice-margin retreat was delineated by mapping and correlating moraines (Fig 3). Relative proximity to one another and alignment orientations were the basis for correlation of moraines longer than 1 km in length. Some breaks between moraines have been caused by postglacial stream erosion. Smaller moraines (e.g., <1 km in length) do not exhibit spatial continuity, but their orientations are generally similar to the larger moraines. The Pond Ridge moraine (Leavitt and Perkins, 1935; Borns, 1980; LePage, 1982) is mapped across the southern part of the study area. The trend of Pond Ridge moraine appears to crosscut the southwest trend of the moraines found to the south. The Pond Ridge moraine was previously mapped as curving to the north-northeast near the village of Cutler (Leavitt and Perkins, 1935; Borns, 1980; LePage, 1982), but present mapping shows that it continues essentially eastward toward the coastline and offshore (Fig. 3). Shipp (1989) traced the submarine equivalent of Pond Ridge moraine westward beneath Machias Bay (Fig. 3). It then probably continues on the other side of the bay as a distinct ridge (H. W. Borns, Jr., 1994, personal commun.). Compared to more northern moraines, the Pond Ridge moraine shows less change in trend when crossing the valley partly occupied by Little Machias Bay (Figs. 3 and 4). North of the Pond Ridge moraine, especially northwest and northeast of Little Machias Bay, moraines exhibit changes in orientation that appear to be associated with changes in elevation (Fig. 4). For example, 5–10 km northeast of Little Machias Bay between 0 and 30 m asl, the moraines trend approximately southeast, whereas above 30 m asl, they trend southwest especially as they approach elevations greater than 60 m asl (Fig. 4). Another example of changes in moraine orientation that appear to be associated with changes in elevation is found just northwest of Little Machias Bay. Moraines on the topographic high trend southwest, whereas just to the north of the high area the trend changes to westward on entry to the low-lying area that is Machias Bay.

Across the northern sector of the study area, moraines that are subparallel, quasi-continuous, easily mapped, and closely spaced are mapped as an individual unit termed the Pineo Ridge moraine system (Figs. 3 and 4). Borns (1980) and Thompson and Borns (1985) traced the Pineo Ridge moraine system westward, beyond the study area, and correlated it to the Pineo Ridge moraine-delta complex (Borns, 1980; Miller, 1986; Ashley et al., 1991). Overall, the Pond Ridge and Pineo Ridge moraine system moraines are the most continuous in the study area (Borns, 1980; Kaplan, 1994; Fig. 3, this paper). In addition, the Pineo Ridge moraine system moraines exhibit numerous crosscutting relationships (Fig. 3). Compared to the moraine patterns just to the south, moraines of the Pineo Ridge moraine system usually continue uninterrupted and exhibit less change in their trend where they cross topographic highs and lows, such as Machias Bay (Figs. 3 and 4). The only significant change in the trend of the Pineo Ridge moraine system is in the eastern part of the study area. Northeast of Whiting, the moraines of the Pineo Ridge moraine system and those to the north trend approximately

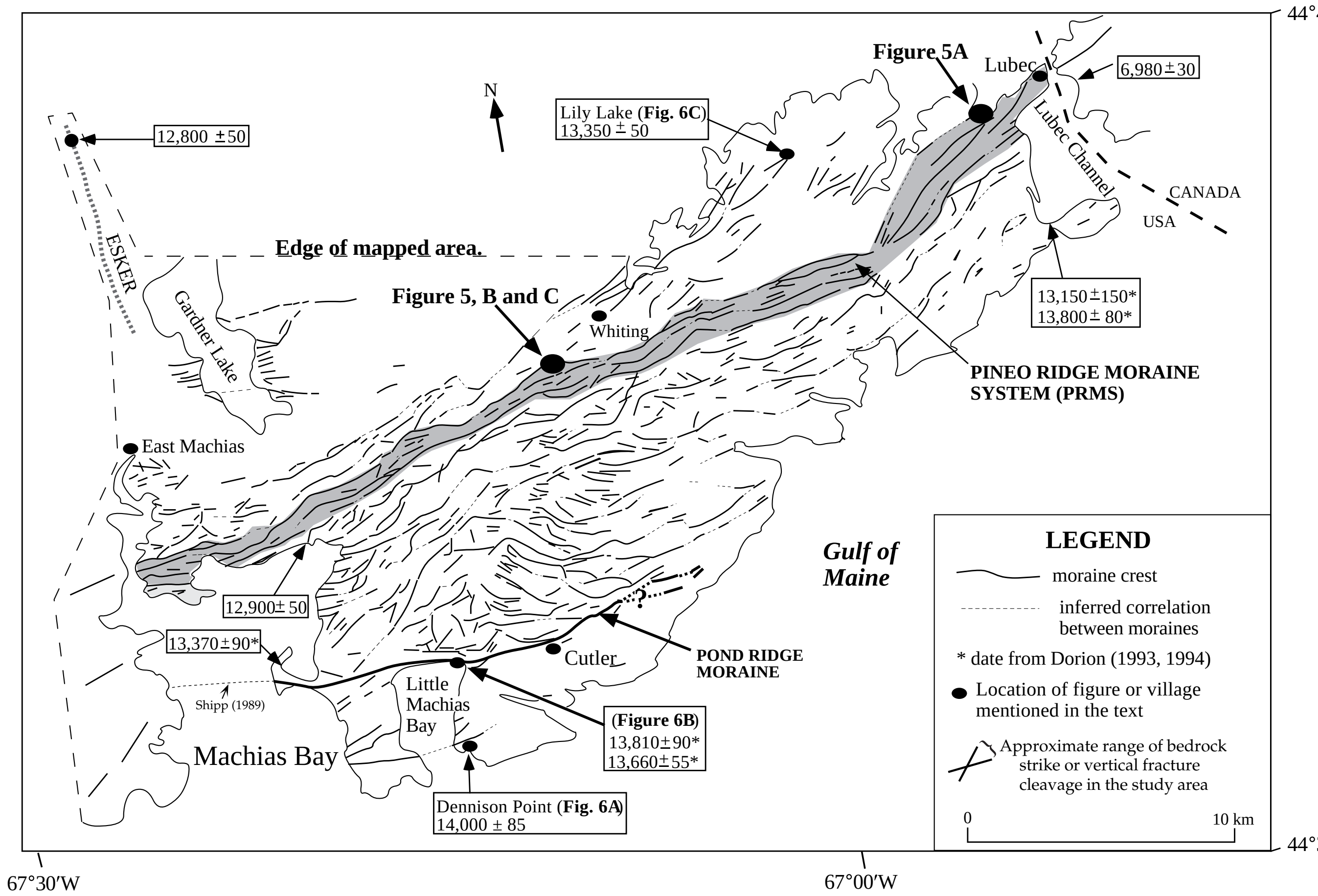


Figure 3. Mapped moraines and correlations, the location of ^{14}C dates (see Table 1), and stratigraphic sections shown in Figures 5 and 6. The Pond Ridge moraine is represented with a thicker line only for illustrative purposes and does not represent a difference in morphology compared to the other moraines.

S45°W, compared to S70°W to the west of Whiting (Fig. 3).

The transition from abundant linear moraines in the southeastern part of the study area to few or none in the northwestern part is distinct throughout the region. In the northwestern part of the study area, locally, no moraines occur between areas of well-defined moraines. This transition apparently represents the proximal edge of a moraine belt that covers all of coastal Maine (Thompson, 1982; Smith, 1982, 1985; Thompson and Borns, 1985).

Internal Stratigraphy and Structure of the Moraines. The environment of moraine deposition and processes at the margin can also be interpreted from the internal stratigraphy and sedimentary structures (see Retelle and Bither, 1989; Ashley et al., 1991; Hunter et al., 1996). The general internal characteristics of study-area moraines were ascertained from detailed examination of two borrow pits within the Pineo Ridge moraine system (Figs. 3 and 5). Unfortunately, no single pit within the study area provided exposures of both the ice-proximal and ice-distal side of a single moraine. Similar stratigraphic information was obtained from other poorly exposed pits in the study area and from sites studied by LePage (1982) within the Pond Ridge moraine. The ice-proximal or up-flow side of the moraine crest contains stratified gravel, sand, and clay beds dipping to the north and northwest ("up ice-flow") and massive diamicton with a fine matrix (Fig. 5A). Gravel and sand beds are both massive and bedded. The axis of the fold shown in Figure 5A (see Fig. 3 for location) strikes southeast, roughly parallel to nearby indicators of southeastward ice flow. At the moraine crest (Fig. 5B), clast-supported diamicton and stratified gravel are the dominant constituents of complexly interbedded deposits. The stratigraphy on the down-flow or ice-distal side of the moraine (Fig. 5C) consists dominantly of gravel, sand, silt, and clay. Diamicton is noticeably lacking, or rarer, in the distal side of the moraine. To summarize, in going from the up-flow or ice-proximal side to the down-flow side of the moraine crest, there is a dominance and then disappearance of diamicton (compare Fig. 5, A–C), followed by a change to completely stratified bedding, characterized by gravel, sand, and clay beds. In addition, a gravel lag commonly mantles the moraines in the study area (Fig. 5A), representing erosion from the regressive offlap (Smith, 1985; Dorion, 1993, 1994).

Chronology

Several AMS ^{14}C ages define the timing of ice-margin retreat across the study area. Five dates were obtained for this study (Fig. 3; Table 1). Five previously published dates (Dorion, 1993, 1994) are presented to help provide a more precise time frame for deglaciation (Fig. 3). Most ^{14}C ages presented in Table 1 and Figure 3 are from samples found in situ, with both valves connected and periostracum attached. The two exceptions, where nothing else was available, include the Hadley Lake sample and the basal date from the Lily Lake core (see Table 2). The Lily Lake sample included several (~20) small, whole

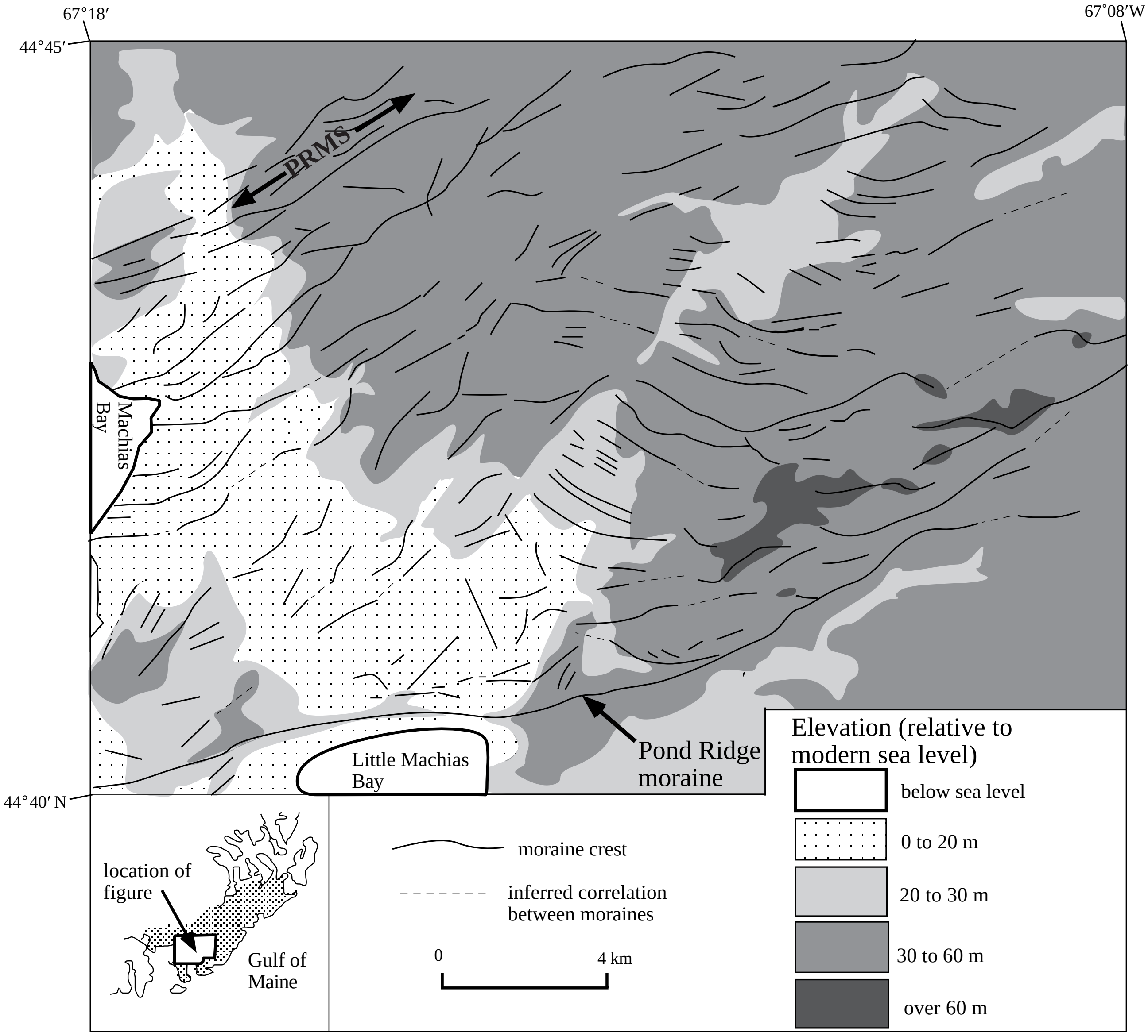


Figure 4. Changes in moraine orientation appear to be associated with elevation changes. Note the Pineo Ridge moraine system (PRMS) in the upper left corner of the map.

shells (one can barely discern the species with the naked eye). Although these shells were not found in situ, the general consistency of their ages within the context of the others (Fig. 3) suggests that they are useful for estimating ages of deglaciation.

All AMS ages from the study area are between 14 000 and 12 800 ^{14}C yr B.P., constraining the general time frame for deglaciation. The one exception, 6980 ^{14}C yr B.P., from Campobello Island is obviously erroneous as the marine transgression was complete by ca. 11 500 ^{14}C yr B.P. (Barnhardt et al., 1995). The oldest shell was collected from clay interbedded with diamicton, in the southernmost part of the study area (Dennison Point, Fig. 6A). The age of 14 000 ± 85 ^{14}C yr B.P. for this shell provides the earliest age for deglaciation of eastern coastal Maine. Two ages for mollusk shells from within the Pond Ridge moraine (Dorion, 1993, 1994) fall between ca. 14 000 and 13 500 ^{14}C yr B.P. Ages that constrain the deglaciation of the northwestern sector of the study area include those from Lily Lake (13 350 ± 50 ^{14}C yr B.P., Fig. 6C) and from Hadley Lake Quadrangle (12,800 ± 50 ^{14}C yr B.P.). Together, the ages indicate that ice retreat from Dennison Point in the southern part of the study area to the northwestern part took ~1000 yr or less. This duration represents an average retreat rate of ~20 m/yr for the average distance of about 20 km across the study area.

Microfaunal Analyses

Microfaunal analyses of fossil assemblages were carried out on samples of glaciomarine sediment, chronologically constrained, from three sites (Table 2, Fig. 6) to better understand the environmental conditions associated with glacial retreat. Samples from two sites, Dennison Point and Pond Ridge moraine, were from glaciomarine clay interbedded with diamicton (Fig. 6, A and B). At Dennison Point, the top diamicton contains striated clasts that are commonly bullet shaped, indicating that it is an ice-proximal deposit (Fig. 6A). The third microfaunal sample is from the basal section of the Lily Lake core (Fig. 6C) where glaciomarine clay is interbedded with sand beds and sand lenses. Core refusal at 1629 cm below lake level prevented collection of information below that depth. At all three sites, the foraminiferal assemblages are almost monospecific; the dominant foraminifera *Elphidium excavatum* forma *clavata* composes between 85% and 93% of the assemblage (Table 2). Ostracoda were found at the Dennison Point site (Table 2, no percentages calculated); the assemblage is more diverse than the foraminiferal assemblage and is dominated by *Cytheromorpha macchesneyi* (T. M. Cronin, 1994, personal commun.).

DISCUSSION

Moraine Formation

The conditions under which the moraines formed are complex, but several general conclusions can be reached on the basis of their stratigraphy and structure (Fig. 5). First, the moraines formed primarily at the grounding line in a marine environment. Supporting evidence includes a change in sediment facies between the ice-proximal and ice-distal side of the moraine crest, stratified beds (although these also occur within subglacially formed eskers), and the presence of marine clay. Maximum elevations for almost all of the moraines in the study area are less than 50–60 m asl, which is the inferred sea level during deglaciation (Thompson et al., 1989); thus almost all moraines originated entirely in a marine environment. Some parts of the highest moraines reach 60–70 m asl and thus may have originated at or slightly above paleo–sea level. It should be noted that the subsequent marine regression after deglaciation may have removed an unknown amount of material from the tops of the moraines (Dorion, 1993, 1994). Second, the presence of stratified beds indicates that meltwater outflow at the margin played a key role in their formation. Third, folds and faults observed in exposures on the ice-proximal side of many of these landforms indicate that the ice was internally active (in contrast to stagnant) when they formed (also see LePage, 1982; Ashley et al., 1991). Additional evidence for glacial activity includes crosscutting relationships between moraines (see discussion section). These general conclusions are in agreement with LePage (1982) for the Pond Ridge moraine and with results of other studies in coastal Maine (e.g., Retelle and Bither, 1989; Ashley et al., 1991; Hunter et al., 1996).

Several processes most likely contributed to the complexity of moraine morphology and stratigraphy. These may have included changes in sediment supply, glacial-hydrologic flow regime, position and angle of an effluent jet relative to the moraine, fluctuations in the position of the grounding line, and remobilization of material after the grounding line had retreated from the site (Retelle and Bither, 1989; Ashley et al., 1991; Hunter et al., 1996). Where diamicton is interbedded with the northward-dipping stratified beds (Fig. 5A), the stratigraphic relationships may indicate that the dipping beds were formed from effluent meltwater streams that were able to flow upslope because of a pressure gradient caused by overlying ice and hydrostatic pressure (Paterson, 1994).

Ice-Margin Retreat

Orientations of moraines and striae were also used to infer the nature of retreat and processes along the grounding line, providing insight into interactions between the ice sheet and the ocean during deglaciation of coastal Maine. An important assumption underlying the reconstruction of ice margins and retreat processes is that mapped moraines formed at the grounding line and not behind it (Fig. 3). This assumption is supported by structural and stratigraphic relationships in the moraines as summarized in the preceding section. The moraines show that deglaciation started in the southern Little Machias Bay and Machias Bay area (Fig. 3). The orientation of the Pond Ridge moraine appears to crosscut the southwest-trending moraine pattern to the south. This crosscutting relationship supports LePage's (1982) conclusion that the Pond Ridge moraine represents a readvance position. North and northeast of the Pond Ridge moraine, the changes in moraine orientations at elevations lower than 30 m asl are inferred to represent the presence of calving embayments or marine reentrants in the ice margin. One marine reentrant existed northeast of the area occupied at present by Little Machias Bay and another around eastern Machias Bay (Figs. 3 and 4). At the higher elevations (e.g., >30 m asl), the changes in moraine orientation are inferred to represent the influence of pinning points, locations where the grounding line was relatively more stable on topographic highs (e.g., >60 m asl, Fig. 4). Northeast of Whiting, the changes in orientation of the Pineo Ridge moraine system and other moraines (i.e., the somewhat more south trend of S45°W compared with S70°W to the west of Whiting) may have been due to a marine reentrant to the east or northeast of the study area (Fig. 1) and/or the location of pinning points, or both. Additional evidence for the existence of a marine reentrant east or northeast of the study area include striae trending more to the east (S60°E–S50°E) along the coastline of the Lubec channel and on Campobello Island (Fig. 2A). Pinning points northeast of Whiting could be inferred because there is higher relief (Kaplan, 1994). Evidently, the grounding line was destabilized in certain areas such as marine reentrants, while at the same time it was stabilized in other areas by the pinning action of higher elevations.

Locally, striae and moraine relationships and morainal patterns indicate that topography also affected ice-flow direction behind the ice margin (Fig. 7). In some areas there is evidence that the near-margin flow direction changed concomitantly with changes in grounding-line orientation (Fig. 7). For example, orientations of striae and small-scale stoss-and-lee bedrock forms (e.g., bedrock outcrop) along the coastlines of Machias Bay and Little Machias Bay differ from the general direction throughout the study area (Fig. 2B), remaining roughly perpendicular to nearby moraine crests even when the latter change orientation. Specifically, the striae azimuths along the eastern coastline of Machias Bay (e.g., S0°E), where moraines trend approximately due east, differ significantly from striae azimuths along the northwestern coastline (e.g., S15°E; Student's t test, 0.001 significance level), where the moraines trend west-southwest (Fig. 7). In the vicinity of Pond Ridge moraine, which trends approximately due east, the youngest striae indicate southward ice flow (Fig. 7). In contrast, the older striae at these localities indicate earlier more southeastward ice flow. LePage (1982) also suggested that a southeastward flow preceded a more southward flow in this area, on the basis of the difference in clast fabric between tills underlying (i.e., predating) and interbedded with the sediments of the Pond Ridge moraine. Elsewhere, changes in striae azimuths around West Quoddy Head peninsula indicate that during ice recession, when the ice thickness had decreased, some divergence of flow occurred around this higher land area (see results section, Fig. 2A).

The effects of topography on glacial processes are most pronounced in the low-lying areas where inferred marine reentrants existed (Fig. 4). Changes in both local ice-flow direction and ice-margin orientation may represent topographic diversion into these lower areas (Fig. 7). Topographic lows probably facilitated ice flow that converged toward a steeper ice cliff within the marine reentrants, which can further increase ice speed and further steepen the ice margin, thus creating a positive feedback. Stromberg (1981) and Hoppe (1973) described what may be analogous glacial processes in southern Sweden. They identified areas of probable calving bays on the basis of moraine and esker orientations, changing striae directions, and clay varve data. They concluded that tidewater glacier margins are frequently unstable and produce rapid flow normal to the ice margin. The rapid flowage, in time, contributes to

A

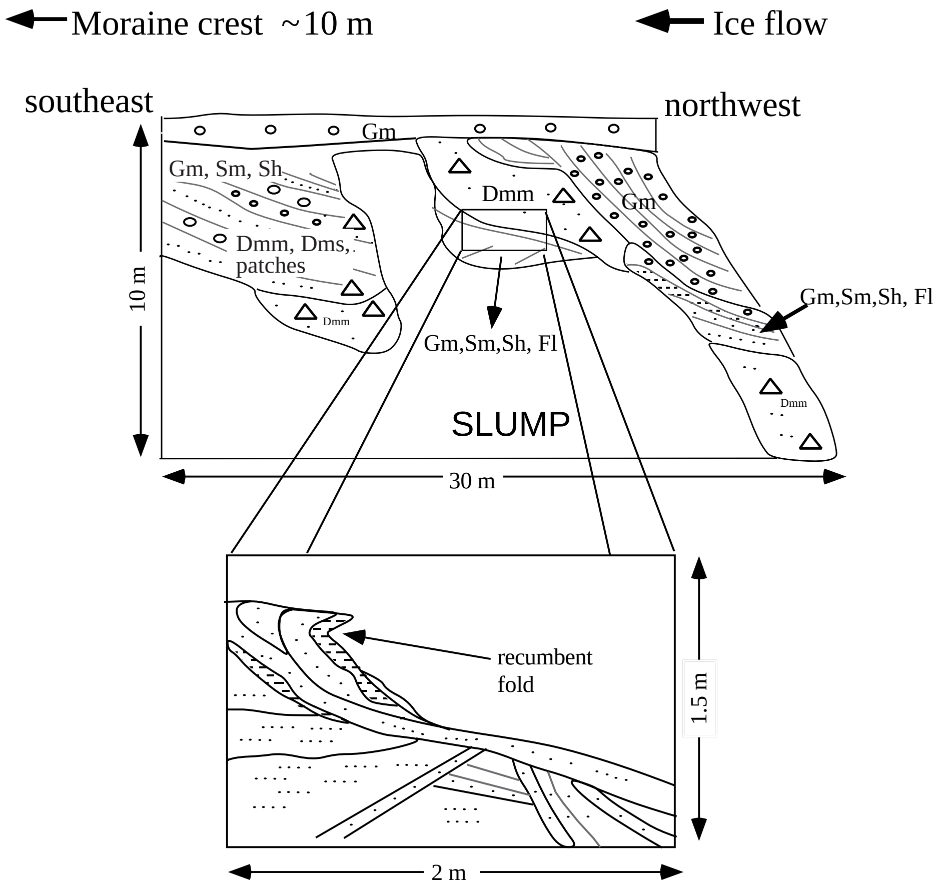
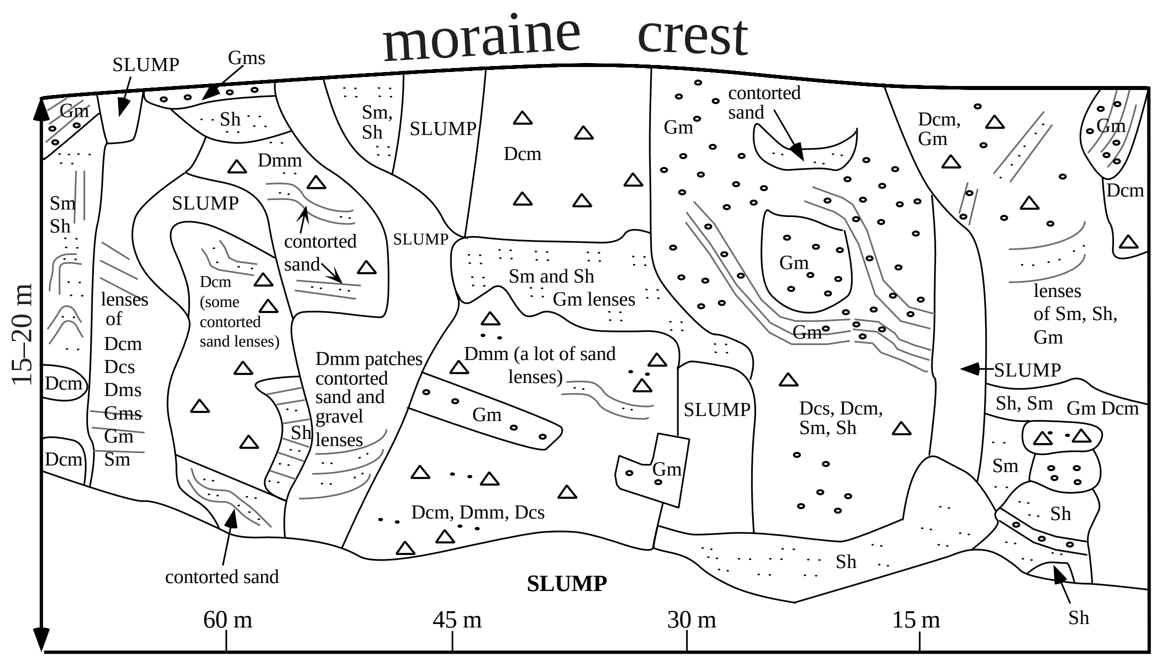


Figure 5. Internal stratigraphy and composition of a typical moraine in the study area (see Fig. 3 for locations). The two illustrated moraines (see text) are mapped as part of the Pineo Ridge moraine system. (A) Cross section of the ice-proximal side (i.e., up-glacier direction) of the moraine, perpendicular to the moraine crest. Note that the fold is oriented approximately transverse to nearby striae (Fig. 2A). Dmm and Gm (see lithofacies code, Fig. 5C) dominate the ice-proximal side of the moraine. (B) Cross section along the moraine crest, looking northwest. The internal stratigraphy is complex with diamicton and gravels being the primary constituents.

B



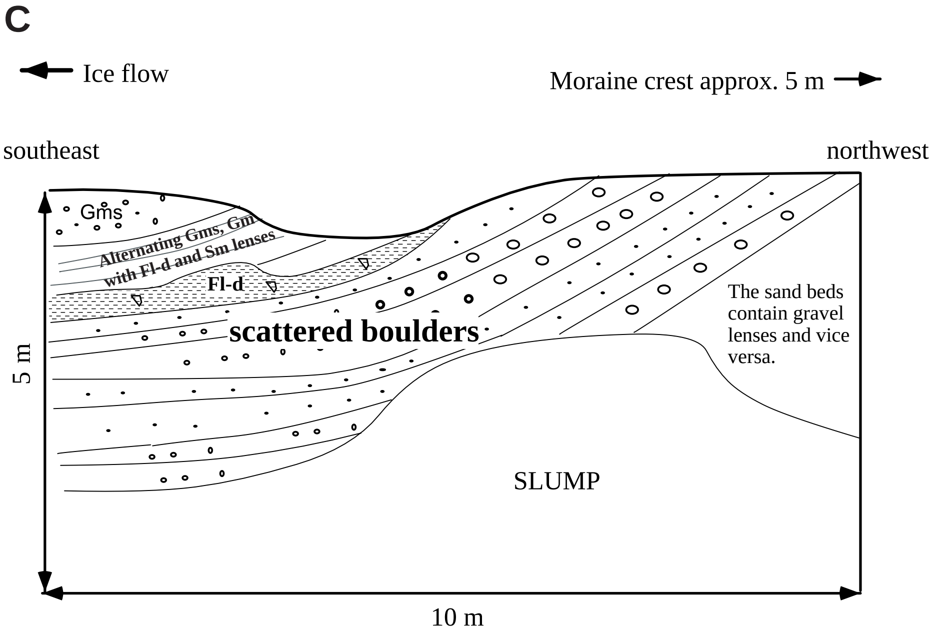


Figure 5. (*Continued*). (C) A simplified cross section of the ice-distal side (i.e., down-glacier or flow direction) of the moraine, perpendicular to the moraine crest. Diamicton is not present, the beds are dipping away from the moraine crest toward the southeast, and dropstone-rich silt and clay are interfingered. Note the decreasing size of gravel clasts, which qualitatively represents decreasing clast size in these dipping beds. Lithofacies code: Dcm—clast-supported diamicton (i.e., lack of fine matrix); Dcs— clast-supported, stratified diamicton; Dmm—matrix-supported (usually sand) massive diamicton; Dms—matrix-supported stratified diamicton; Gm—massive or crudely bedded gravel; Gms—massive, matrix-supported gravel; Sh—sand with horizontal lamination; Sl—sand with low-angle cross-beds; Sm—massive sand; Fl and Fl-d—laminated silt and mud and laminated silt and mud with dropstones, respectively. Triangles—diamicton; ellipsoids—gravel; dots—sand; dashes—silt and mud. Finer boundaries represent orientation of crude stratification or gravel bedding.

high erosion rates and common glacially eroded landforms. There is evidence in the study area that intense glacial erosion also occurred behind the general glacial margin in areas that contained marine reentrants. Striations are more commonly found along the coastline in low-lying areas, such as Machias Bay, where marine reentrants are inferred to have existed (Figs. 2A, 7).

Topography may have affected ice-margin geometry and/or grounding-line dynamics by allowing higher calving rates in deeper water (at lower elevations), changing the bed slope and providing higher-elevation pinning points. Brown et al. (1982) quantified calving rates for the tidewater Columbia Glacier in Alaska and concluded that water depth was the key parameter. Although relative sea level was falling during deglaciation of coastal Maine (rebounding landscape), the more important factor may have been water depth, which would have varied along the ice margin as it retreated across a changing topography. Different sections along the same grounding line would have been retreating in different water depths (for example, northeast of Little Machias Bay [Fig. 4]). The slope of the bed can also have a significant effect on the stability of the grounding line. Retreat upslope (i.e., locally, elevations increasing in retreat direction) is a relatively stable condition for the grounding line, whereas retreat downslope (locally, elevations decreasing in retreat direction) is unstable (Hughes, 1987). The grounding line retreated upslope northeast of Little Machias Bay after it had retreated from Pond Ridge moraine (Fig. 4). In contrast, the grounding

TABLE 1. NEW DATES ON MOLLUSK SHELLS IN THE STUDY AREA

Location	Location: Lat (N)	Location: Long (W)	Age (accession number)*	Species dated	$\delta^{13}C$ (‰)	Setting and signifance
Lily Lake core (see Fig. 6C) (1582–1602 cm)	44°49′45″	67°6′15″	13 350 ± 50 (OS-2151)	*Hiatella arctica*	0.41	Basal lake date (sand interbedded with clay) on shells between 1582 and 1602 cm below lake level. Minimum date on deglaciation of the site.
Dennison Point (see Fig. 6A)	44°38′30″	67°14′30″	14 000 ± 85 (OS-2154)	*Macoma calcarea*	–2.03	Glaciomarine clay interbedded with diamicton. Specimen lived at or near the grounding line.
Campobello Island	44°51′15″	66°58′8″	6980 ± 30 (OS-2153)	*Portlandia arctica*	–8.21	Interbedded sand, glaciomarine clay, and silt; erroneous result.
Hadley Lake Quadrangle (esker pit)	44°45′	67°25′13″	12 800 ± 50 (OS-2155)	*Hiatella arctica*	0.56	Stratigraphic position is unclear. Specimen found in a pile at center of pit. The date is a minimum age for deglaciation of the study area.
Machias Bay Quadrangle (Look's Seafood Cannery)	44°43′13″	67°18′36″	12 900 ± 50 (OS-2152)	*Nucula expansa*	–7.66	Exposure of glaciomarine clay. Minimum date on deglaciation of the site.

Note: All samples, except Hadley Lake Quadrangle locality, were collected in situ. All ^{14}C dates reported here are uncorrected for the marine-reservoir effect.

*All samples were run at National Ocean Sciences AMS Facility, Woods Hole Oceanographic Institution (WHOI), and are normalized to del ^{13}C = –25‰.

line retreated downslope directly north of Little Machias Bay after leaving the Pond Ridge moraine. A modern analogue may be Columbia Glacier; sections of its grounding line are probably unstable partly because of changes in bed slope (Meier and Post, 1987, Hughes, 1992). Hughes (1992) suggested that one section of the Columbia Glacier margin can undergo catastrophic collapse while adjacent sections remain relatively stable.

Topography may have had less of a destabilizing effect on grounding-line dynamics during formation of the Pond Ridge moraine and Pineo Ridge moraine system. There is less apparent change in the trend of the Pond Ridge moraine and Pineo Ridge moraine system, compared to other moraines in the study area, where these features cross the low areas occupied by Little Machias Bay and Machias Bay, respectively (Figs. 3 and 4). There is some evidence that bedrock pinning points may have been important grounding-line stabilizers when the Pineo Ridge moraine system formed. For example, the only major change in Pineo Ridge moraine system orientation, in the eastern part of the study area (Fig. 3), is associated with a bedrock hill and the topographically high area to the northeast that underlies the town of Lubec (Kaplan, 1994). Also, the topographic high of Pond Ridge moraine and Pineo Ridge moraine system may have stabilized the grounding line. A reorganization of the margin prior to and/or during formation of the Pond Ridge moraine and Pineo Ridge moraine system can be invoked to explain the less apparent topographic influence. The margin may have become thicker and steeper, thereby mitigating the destabilizing effect of deeper water depths. The fact that the Pond Ridge moraine and Pineo Ridge moraine system moraines are the only glacial landforms that can be continuously traced across the entire study area (Figs. 3), and westward beyond the study area (Thompson and Borns, 1985; H. W. Borns, Jr., 1994, personal commun.), provides additional evidence for a more linear, less lobate ice margin during their formation.

TABLE 2. MICROFAUNAL ANALYSES (GLACIOMARINE CLAY) AT THREE SITES

Location	Associated AMS ^{14}C ages (yr B.P.)	Microfauna
Dennison Point*	14 000 ± 85 *(Macoma calcarea)*	Foraminifera (*n* = 122) 85.2% *Elphidium excavatum* (Terquem) forma *clavata* Cushman 7.4% *Elphidium excavatum* (Terquem) forma *excavatum* (Terquem) 3.3% *Cassulidina reinforme* 4.1% *Elphidium excavatum* (Terquem) forma *lidoensis* Cushman and *selysenses* (Heron-Allen and Earland) Ostracodes (identified by T. M. Cronin) *Cytheromorpha macchesneyi* Brady and Crosskey *Cytheropteron suzdalskyi* Lev *Cytheropteron elaeni* Cronin *Semicytherura complanata* Bradym, Crosskey, and Robertson *Jonesia simplex* (Norman) *Acanthoccytheris dunekmensis* (Norman)
Pond Ridge Moraine†	13 810 ± 90 *(Nucula expansa)* 13 660 ± 55 *(Macoma calcarea)*	Foraminifera (*n* = 140) 89% *Elphidium excavatum* (Terquem) forma *clavata* Cushman 4.2% *Elphidium excavatum* (Terquem) forma *excavatum* (Terquem) 6.4% unknown
Lily Lake§	13 350 ± 50 *(Macoma calcarea)* Date between 15.92 and 15.82 m	Foraminifera between 15.82 and 15.72 m (*n* = 184) 82% *Elphidium excavatum* (Terquem) forma *clavata* Cushman 12% *Elphidium excavatum* (Terquem) forma *excavatum* (Terquem) 6% Other: *Elphidium excavatum* (Terquem) forma *lidoensis* Cushman and *selysenses* (Heron-Allen and Earland) *Triloculina truncunata* *Cyclogyra* sp. or *Cornuspira* sp. Between 15.92 and 15.82 m (*n* = 45) 93% *Elphidium excavatum* (Terquem) forma *clavata* Cushman 7% *Elphidium excavatum* (Terquem) forma *excavatum* (Terquem) No microfauna found below 15.92 m

*See Table 1 and Figure 6A. The analyses were conducted at the same stratigraphic level as the dated mollusk shell (within 25 cm radius).

†Dates from Dorion (1993, 1994). The glaciomarine clay that was analyzed was not collected from the same exact stratigraphic location as the dated shells. However, both dated shells and the sample used for analyses were from clay that is interbedded with diamicton (Fig. 6B). In addition, the dated shells and glaciomarine clay were collected within 50 m of one another along the same stretch of coastline.

§See Figure 6C for location of samples.

Inferred Readvances in the Study Area

Crosscutting relationships between moraines and moraine patterns as well as ice-push structures indicate that several minor ice readvances occurred during the overall deglacial episode. Borns (1973) and Smith (1982, 1985) suggested that most of the moraines in the coastal moraine belt represent only local fluctuations of grounding lines during net retreat. Smaller moraines in the study area that do not exhibit crosscutting relationships probably represent relatively minor stillstands of the grounding line during retreat, whereas larger crosscutting moraines such as Pond Ridge moraine and Pineo Ridge moraine system are apparently formed from readvances.

Elsewhere, the formation of the Pineo Ridge moraine system has been explained as a composite delta sequence formed in a calving embayment during a stillstand (Miller, 1986; Smith and Hunter, 1989). The significance of the Pineo Ridge moraine system–delta complex is controversial (Borns and Hughes, 1977; Miller, 1986; Smith and Hunter, 1989). Perhaps, the ~100 km ice margin that formed the Pineo Ridge moraine system–delta complex varied in behavior. Some sections of the margin underwent stillstands (e.g., Miller, 1986) while others readvanced. Borns and Hughes (1977) inferred that the Pineo Ridge moraine system–delta complex represented a readvance that correlated with the Port Huron readvance in the Great Lakes region. Although the chronology presented in this study may support that conclusion, additional dates are necessary before such correlations can be made with confidence.

Glacial Retreat Environment

The microfaunal assemblages that occur at Dennison Point, in the Pond Ridge moraine, and the basal section of the Lily Lake core (Fig. 6) indicate that ice-margin recession in the study area occurred in a marine environment characterized by arctic to subarctic climatic conditions. The

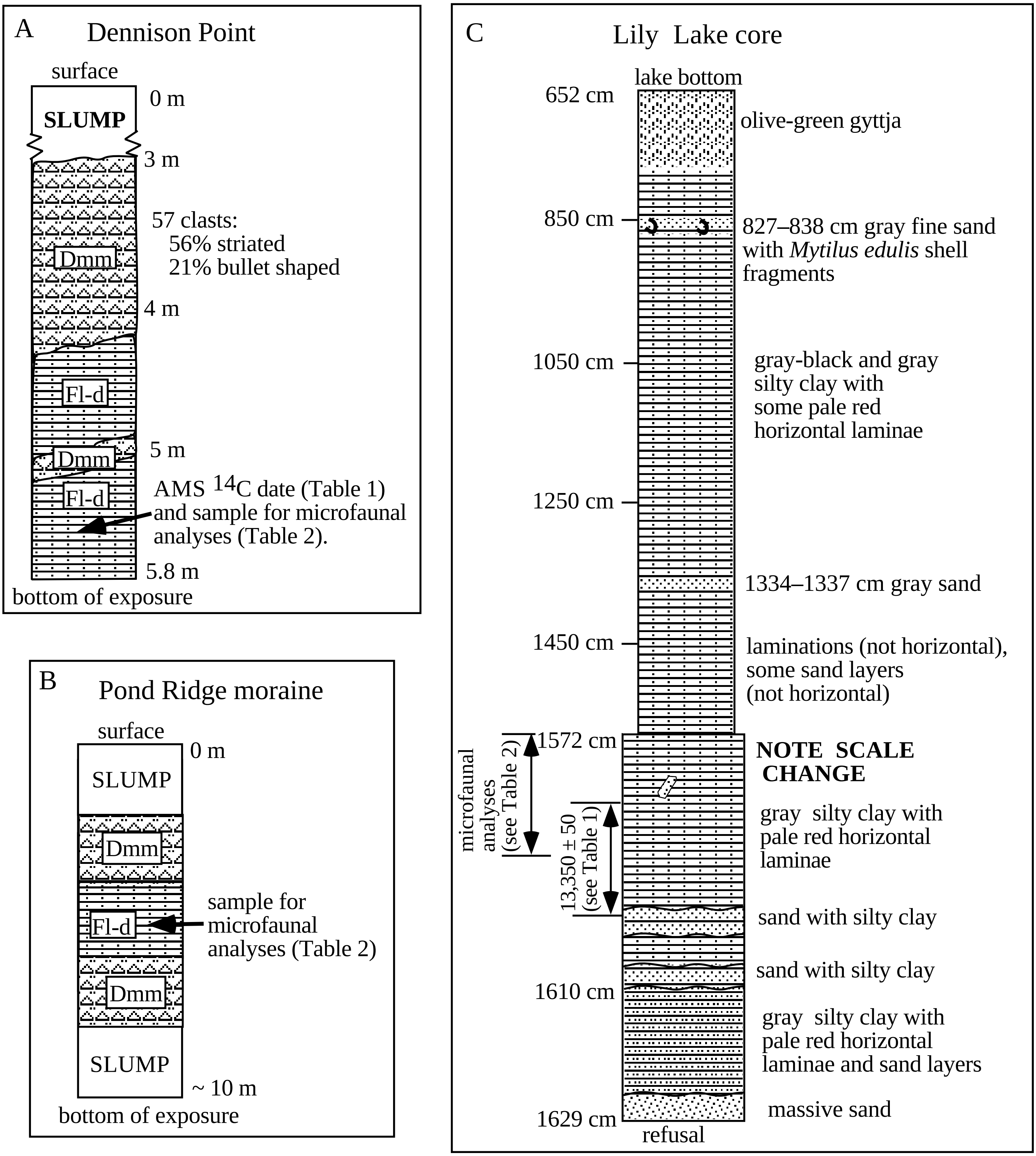


Figure 6. Representative stratigraphic sections at three sites within the study area where samples were collected for microfaunal analyses (see Fig. 3 for locations and Fig. 5C for lithofacies codes). (A) Exposed stratigraphy at the Dennison Point site. Striated and bullet-shaped clasts (>5 cm long axis) are present in the upper diamicton. The lower diamicton had much smaller clasts (<5 cm long axis); thus similar observations were not obtained. (B) Simplified stratigraphy observed within the Pond Ridge moraine (the thickness of the units are schematically drawn as equal because the distance of boundaries from the top were not measured). (C) Stratigraphy observed in the Lily Lake core. The sand lens between 827 and 838 m with *M. edulis* fragments represents the postglacial marine regression over this site. Note the scale change for the lower 50 cm where the samples for the AMS ^{14}C date and microfaunal analyses were collected.

glaciomarine sediments from Dennison Point and Pond Ridge moraine are interpreted as ice-proximal facies (cf. ice-distal or postglacial facies). It cannot be resolved from available data whether the fauna in the basal section of the Lily Lake core lived close to the grounding line (i.e., ice-proximal) or in distal waters that existed after glacial retreat from the site. Hald et al. (1994) described modern marine environments where *Elphidium excavatum* forma *clavata* dominates the benthic foraminiferal assemblage. These environments are characterized by sea-surface temperatures from –1.8 to 1 °C, the persistence of sea ice for a significant part of the year, and salinities almost always several per mil below that of average (35‰) marine water. In Svalbard fjords, this subspecies prefers moderately turbid waters (i.e., moderate sedimentation rates) near the termini of calving glaciers. Currently, *Cytheromorpha macchesneyi* is dominant only in Arctic marine water masses such as the Beaufort Sea and Novaya Zemlya, Russia, where water depth, salinity, and temperatures are usually 10–30 m, 28‰–32‰, and –1.5 to +1.0 °C, respectively (T. M. Cronin, 1994, personal commun.; also see Cronin, 1988). T. M. Cronin (1994, personal commun.) stated that the fossil assemblage at Dennison Point indicates that sea ice probably existed most of the year and that there probably was an ice-free period when warmer, seasonal meltwater flowed from the ice sheet. The dated mollusk species *H. arctica*, *N. expansa*, *P. arctica*, and *M. calcarea* (Table 1) also indicate that relative to the present Gulf of Maine, a much colder marine environment existed during deglaciation (Dyke et al., 1996). These mollusks are found at present in arctic to subarctic water masses, such as in the northwestern North Atlantic Ocean (Dyke et al., 1996).

Other paleoenvironmental studies based on microfaunal, pollen assemblages, and plant macrofossil data support the conclusion that glacial retreat of eastern coastal Maine occurred in an environment much colder than that of the present. Studies of diatom assemblages (Schnitker and Jorgensen, 1990, Popek, 1993) and foraminiferal assemblages (Lusardi, 1992) in cores of glaciomarine sediment from the present Gulf of Maine also suggest that between ca. 14 000 and 13 000 ^{14}C yr B.P., water temperatures were much colder than modern ones. Anderson et al. (1992) used fossil pollen to indicate that tundra conditions existed around Gould Pond, central Maine (Fig. 1), until ca. 9000 yr B.P. Peteet et al. (1994) reached similar conclusions for southern New England, on the basis of pollen and plant macrofossil investigations. They documented tundralike conditions until 12 500 ^{14}C yr B.P. and estimated that mean July temperatures were ~10 °C colder than modern ones.

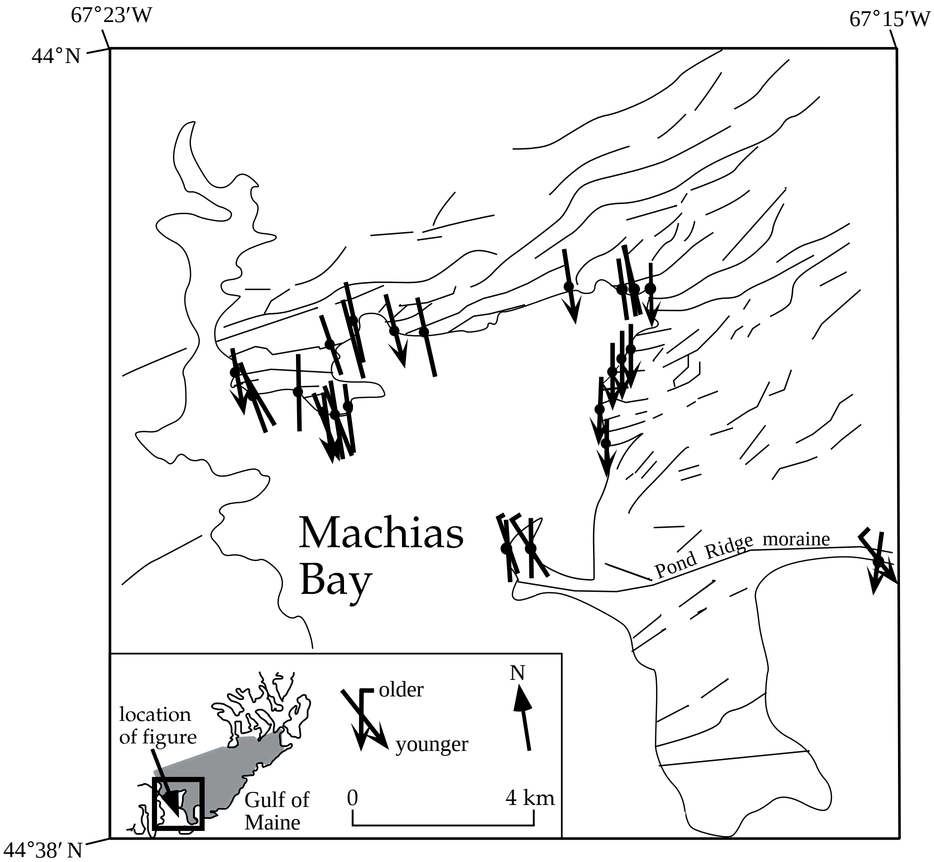


Figure 7. Although the overall ice-flow direction was southeastward and the net ice-margin retreat was northwestward (see Figs. 2 and 3), moraine and striae orientation locally change concomitantly with topographic changes. Along the coastline of Machias Bay, the youngest striae at each site remain perpendicular to nearby moraine crests to the south, even when the latter change in orientation. For example, along eastern Machias Bay, moraines trend approximately west, and the youngest striae indicate that ice flow was due south (S0°E). In contrast, on the northwest side of the bay, moraines trend southwest, and striae show slightly southeastward ice flow (e.g., S15°E). The measurements from the northwestern side of the bay (6 observations) differ significantly at the 0.001 significance level (Student's t test) from the observations on the east side of the bay (11 observations). These observations are used to infer that topography affected both near-margin flow direction and grounding-line orientation (see text for discussion).

The presence of cold-water fauna in glaciomarine sediments in eastern Maine calls into question the previous interpretation of a temperate-tidewater margin and indicates that large sediment accumulations occurred along former ice-sheet margins in arctic to subarctic environmental conditions. The authors of several studies, using present glacial settings in southeast Alaska as an analogy, have argued that the geomorphology and volume of glaciofluvial debris in eastern coastal Maine indicate retreat of a temperate, tidewater margin (e.g., Ashley et al., 1991; Hunter et al., 1996). Their arguments were based on the high volume of glaciofluvial sediment and thus meltwater necessary to have produced the landforms and sediments in eastern coastal Maine. Stewart (1991) and Powell and Domack (1995) noted that present sediment fluxes in cold glaciomarine environments (i.e., arctic and subarctic) are orders of magnitude less than that in temperate settings. Alternatively, periglacial conditions may have existed only at the ice front. However, the studies cited above (Schnitker and Jorgensen, 1990; Anderson; 1992; Lusardi, 1992; Popek, 1993; Peteet et al., 1994) all concluded that distal to the ice margin, the paleoenvironment was colder than that at present. In addition, Kreutz (1994) observed relatively heavy oxygen isotope ratios (e.g., 2‰ to 4‰) in mollusk shells from ice-proximal and ice-distal glaciomarine sediments in the study area. The observed heavy isotope values, despite the abundant evidence for meltwater, imply that cold seawater temperatures prevailed in the Gulf of Maine during deglaciation.

Glaciers in present subarctic settings that produce abundant meltwater (Pfirman and Solheim, 1989) may be an appropriate analogy for deglaciation of coastal Maine. Glaciers in present subarctic settings range from a relatively cold high Canadian Arctic end member, which produces relatively lower meltwater and sediment fluxes, to the relatively warm Svalbard end member (Stewart, 1991). Glaciers in all subarctic environments can have a polythermal or complex thermal regime (Boulton, 1972; Paterson, 1994). Polythermal ice regimes can cause a complex arrangement of frozen and thawed patches at the ice-rock contact, with a well-developed glacial hydrologic system at the bed (Hughes, 1987; Stewart, 1991). This situation is ideal for the entrainment of debris from the substratum at freezing areas and the creation of regelation layers as well as the transport of debris to the ice margin along thawed areas (Boulton, 1972; Hughes, 1987; Iverson, 1991; Stewart, 1991). This scenario, along with the existence of long flow lines that transported debris from source areas, could explain the abundance of glacial landforms and sediment in the coastal region. In summary, the findings presented here do not support previous interpretations of temperate conditions, similar to those of present-day southern Alaska, during deglaciation. However, conditions were probably not as cold and dry as those of the present-day high Arctic. Rather, deglaciation occurred in a cold but meltwater-rich environment similar to that of present-day subarctic Svalbard.

SUMMARY AND CONCLUSIONS

Numerous stillstands and readvances interrupted net retreat of a grounded tidewater margin across southeastern Washington County, Maine. The study area was deglaciated from ca. 14 000 to 13 000 ^{14}C yr B.P., and the overall rate of retreat across the study area was about 20 m/yr. The ubiquitous presence of moraines that frequently exhibit crosscutting relationships allows detailed reconstruction of the paleogeography of tidewater-margin retreat. The moraines, associated sediments, and glacial erosional features in turn offer a basis for inferences regarding the nature of the ice margin and associated glacial processes. Glaciomarine sediment interfingered with ice-proximal and ice-distal facies contains microfauna that can be used as proxies of marine environmental conditions. Moraines contain both diamicton and glaciofluvial sediments, indicating that meltwater outflow from the grounding line played a key role in their formation. The two most continuos glacial landforms are the Pond Ridge moraine and the Pineo Ridge moraine system, which are inferred to represent readvances of the ice margin during general deglaciation.

Several factors are inferred to have influenced the nature of ice-margin retreat and glacial activity along the grounding line. Although relative sea level was falling across a rebounding landscape (Belknap et al., 1987), the present results indicate that the more important factor influencing grounding-line dynamics in eastern Maine was water depth, which varied locally as the grounding line retreated across a rolling topography. The findings from this study suggest that high calving rates in deep water of topographically low areas, in concert with changes in bed slope, caused grounding-line instability. In contrast, pinning points or topographic highs provided locations where the grounding line was relatively more stable. Marine reentrants in topographic lows apparently facilitated ice calving along the ice margin, as they were areas where ice flow converged.

Microfaunal assemblages obtained from ice-proximal and ice-distal sediments are dominated by the foraminifera *Elphidium excavatum* forma *clavata* and ostracode *Cytheromorpha macchesneyi.* These microfauna and the observed mollusk species imply that arctic to subarctic climatic conditions existed during deglaciation. The findings presented here suggest that deglaciation occurred in environmental conditions similar to those of present subarctic Svalbard, i.e., cold but with abundant meltwater.

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