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Late Wisconsinan Deglaciation of Coastal Maine

*Geoffrey W. Smith and Lewis E. Hunter**
Department of Geological Sciences
Ohio University
Athens, Ohio 45701

**Present address:*
Department of Geology
Northern Illinois University
DeKalb, Illinois 60115

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

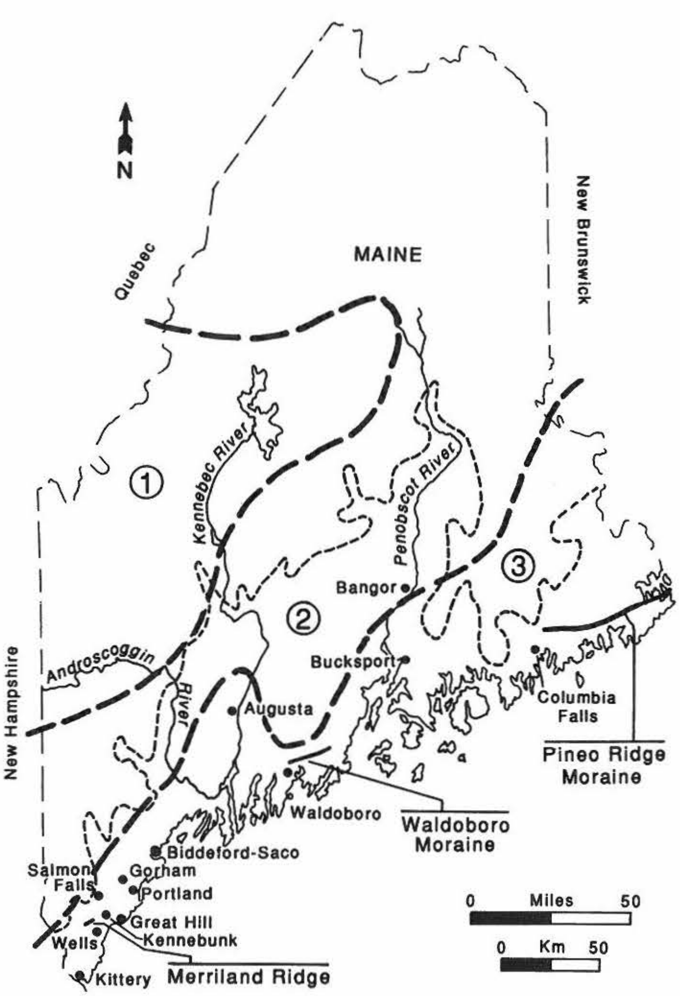

Figure 1. Location map of the Maine coastal zone. Major physiographic sections (dashed lines) include: (1) White Mountains, (2) New England Upland, (3) Seaboard Lowland. Short dashed line is inland limit of late-glacial marine submergence.

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

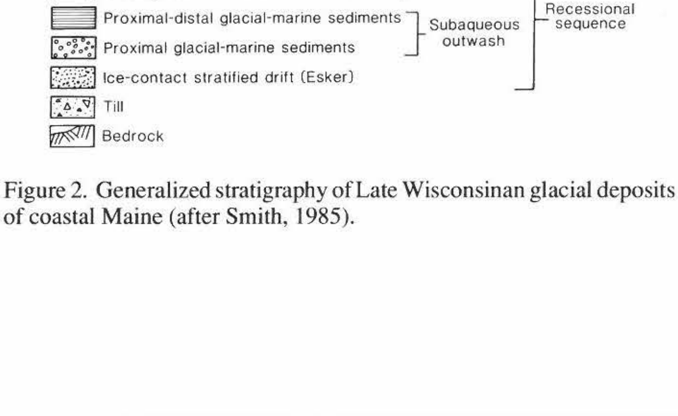

Figure 2. Generalized stratigraphy of Late Wisconsinan glacial deposits of coastal Maine (after Smith, 1985).

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,

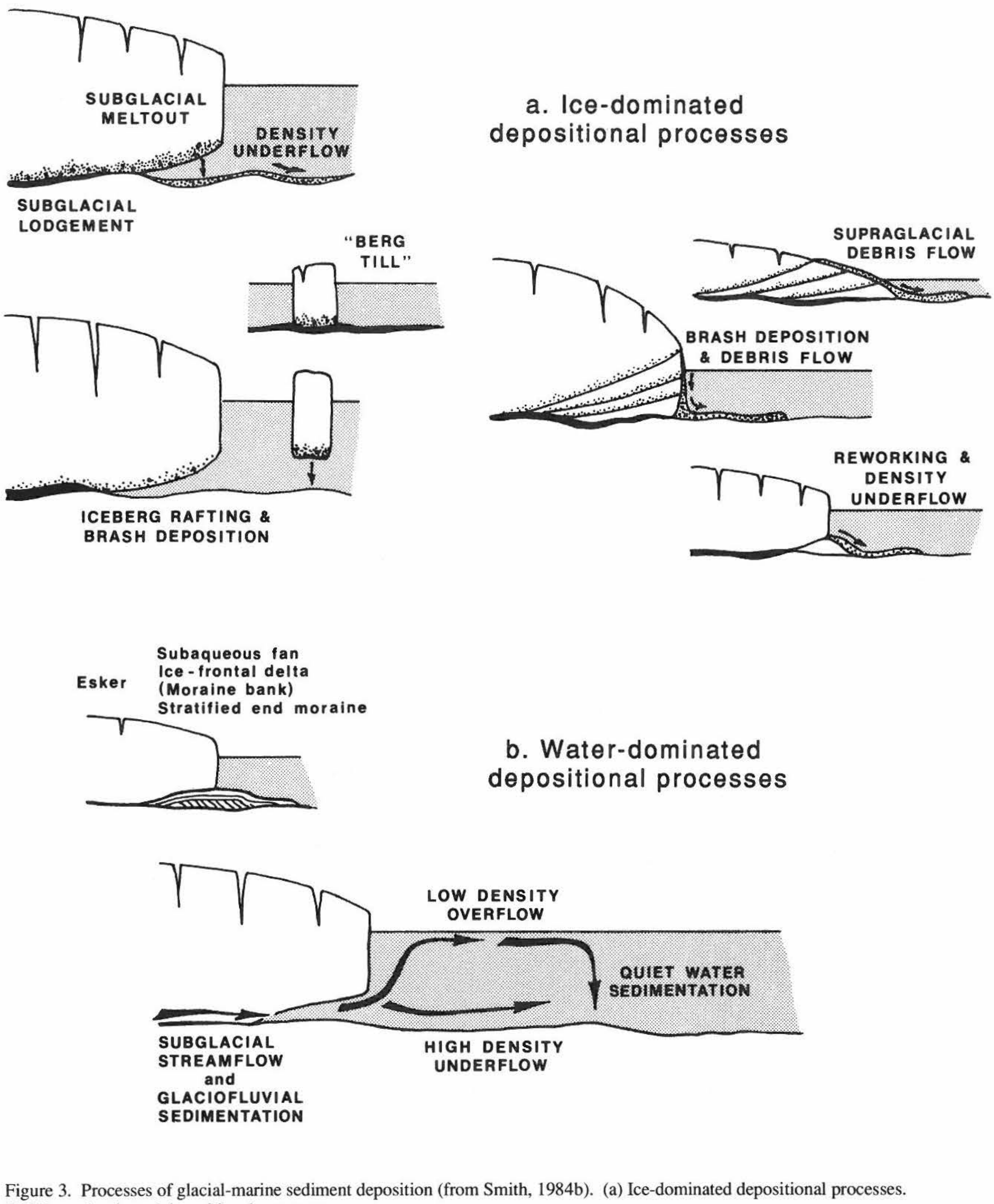

Figure 3. Processes of glacial-marine sediment deposition (from Smith, 1984b). (a) Ice-dominated depositional processes. (b) Water-dominated depositional processes.

(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.

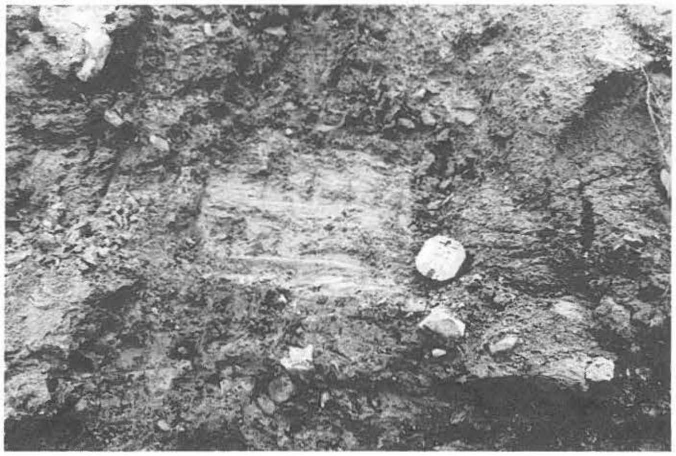

Figure 5. Dms (stratified matrix-supported diamict) lithofacies, Great Hill, York County. Compass for scale.

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

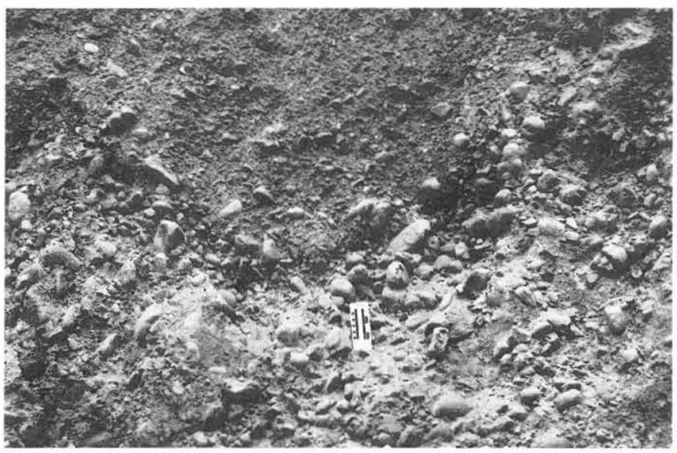

Figure 6. Gmm (massive, matrix-supported gravel) lithofacies, vicinity of Salmon Falls, York County. Upper pebble and cobble gravel in erosional contact with lower cobble gravel. Both gravel units are representative of the proximal subaqueous fan association.

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-

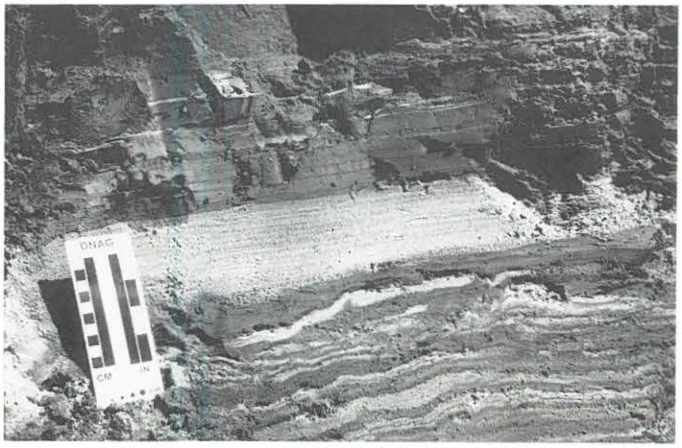

Figure 11. Interbedded fine sand (Ss) and silt (Fl) of the distal subaqueous fan/marine (glacial-marine) mud associations, vicinity of Gorham, Cumberland County. Sand and silt lithofacies overlain by laminated silt (Presumpscot Formation) that grades upward to massive silt (Fm) lithofacies.

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.

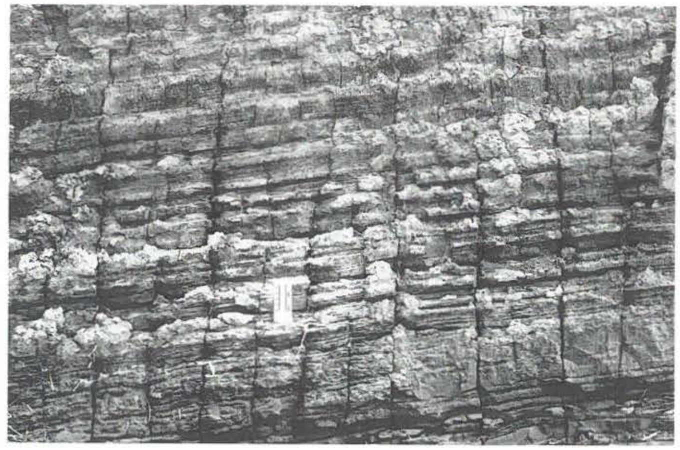

Figure 12. Laminated silt (Fl) of the marine (glacial-marine) mud lithofacies association, vicinity of Portland, Cumberland County.

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,

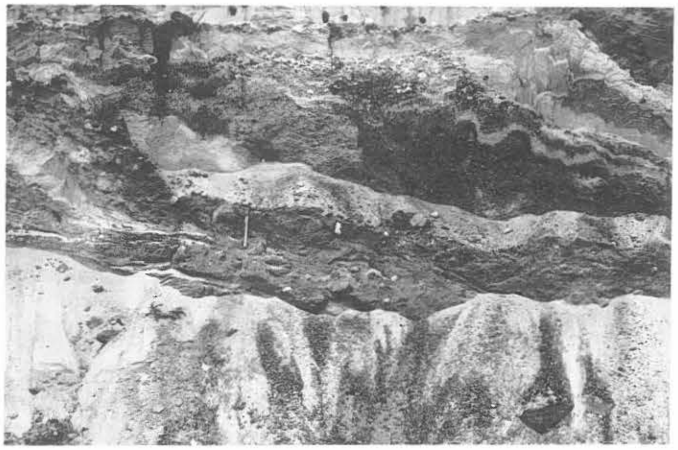

Figure 13. Diamict (Dmm) of the diamict/proximal fan association interbedded with gravel (Gcm), and sand (Ss, Sm) of the proximal subaqueous fan association, vicinity of Salmon Falls, York County. Shovel for scale.

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 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.

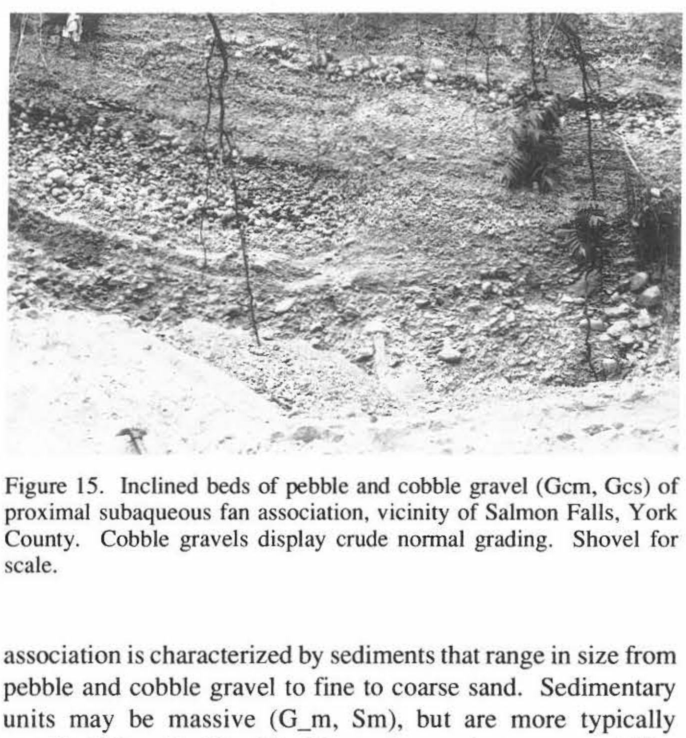

Figure 15. Inclined beds of pebble and cobble gravel (Gcm, Gcs) of proximal subaqueous fan association, vicinity of Salmon Falls, York County. Cobble gravels display crude normal grading. Shovel for scale.

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.

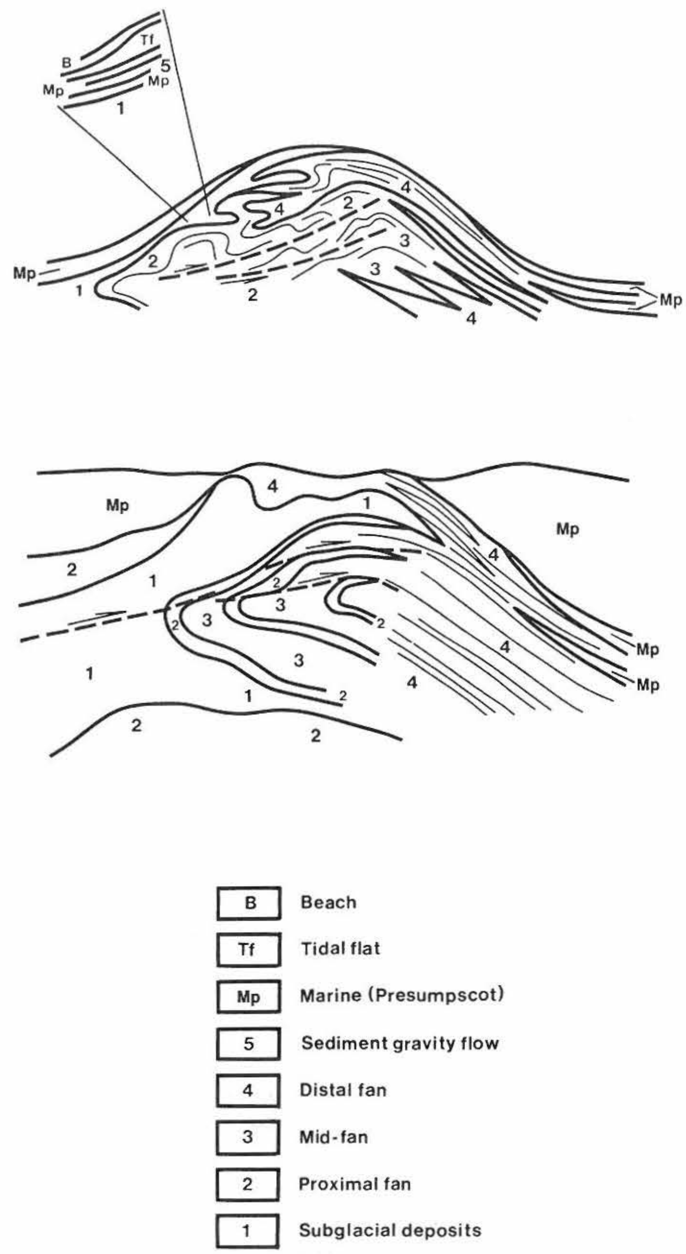

Figure 18. Schematic cross-sections of two coastal end moraines illustrating general stratigaphic relationships and nature of glacial-tectonic deformation. Dashed lines represent major shear planes. Proximal slopes of both moraines are to the left.

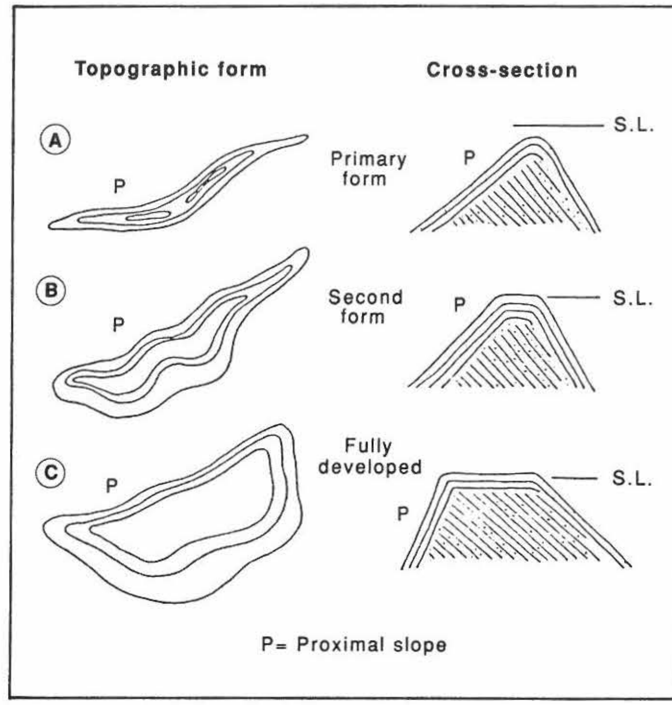

Figure 19. Varieties of ice-frontal delta. (a) Primary form: sharp-crested ridge, not constructed to sea level. (b) Second form: narrow flat-topped ridge, constructed to sea level. (c) Fully-developed delta: broad, flat-topped surface, constructed to sea level (from Gluckert, 1975).

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.

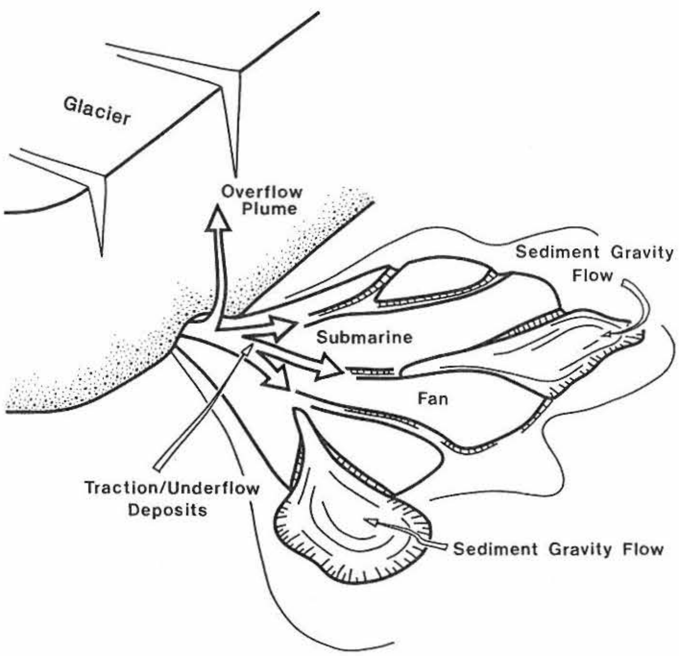

Figure 20. General model for subaqueous fan deposition. Sediment discharge at tunnel mouth consists of traction/underflow deposits and overflow (and interflow) sediments. The former are deposited as proximal to distal subaqueous fan lithofacies. The latter are deposited primarily as distal fan and submarine 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 sheetwash 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;

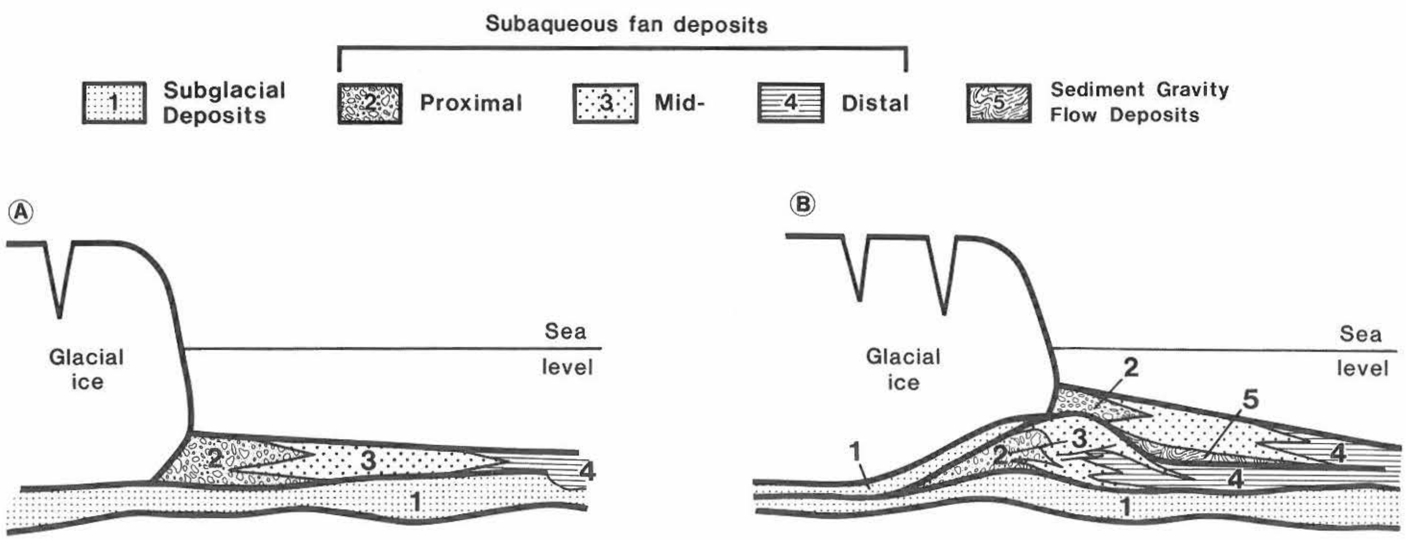

Figure 21. Stratigraphic relationships of subaqueous fan association (a) and subaqueous end moraine association (b). The simple stratigraphic situation of the subaqueous fan association is complicated during minor glacial readvance by glacial-tectonic deformation and thickening of the sedimentary package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

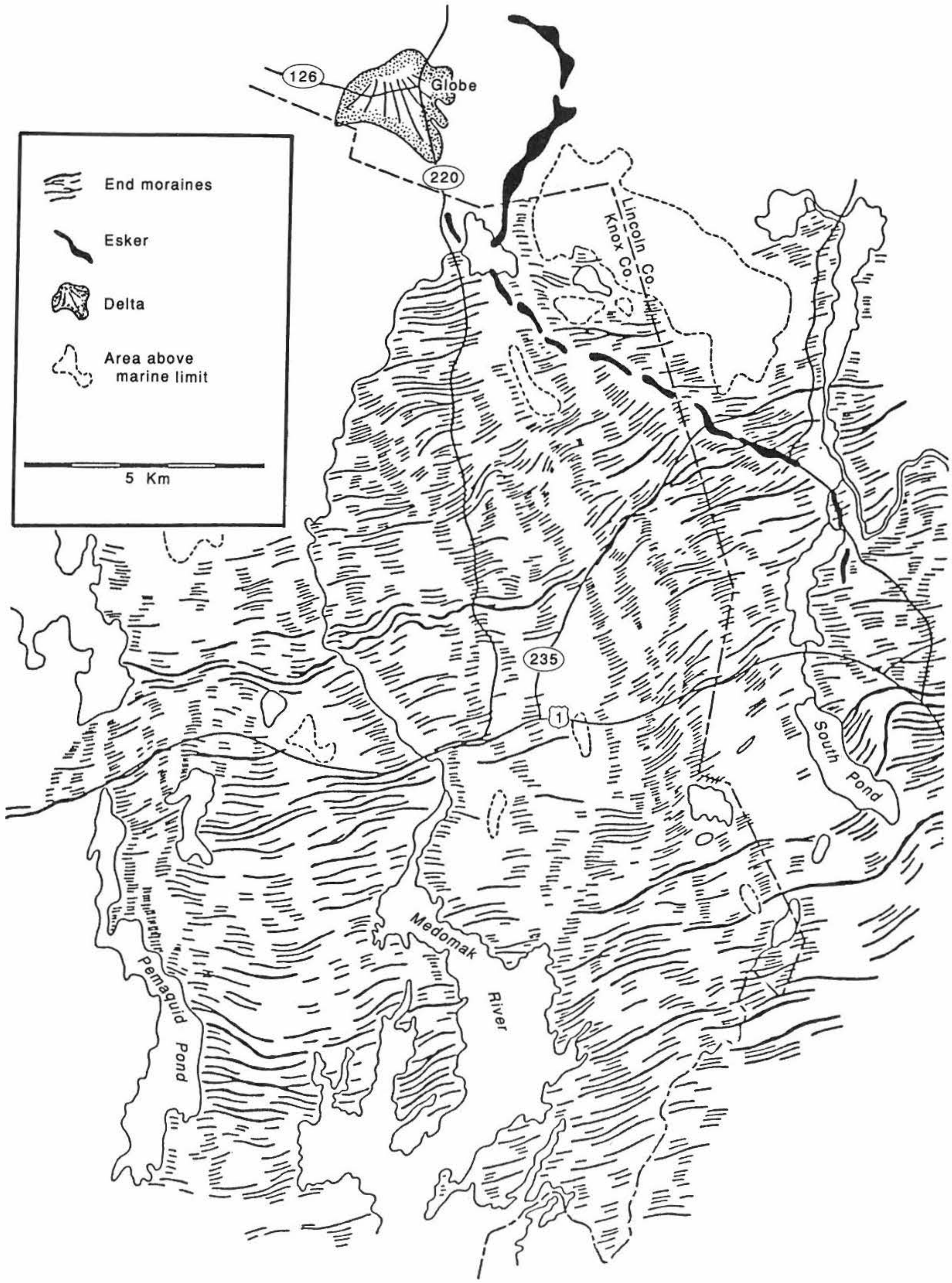

Figure 22. End moraines and glaciofluvial deposits of portions of Lincoln and Knox Counties, Maine.

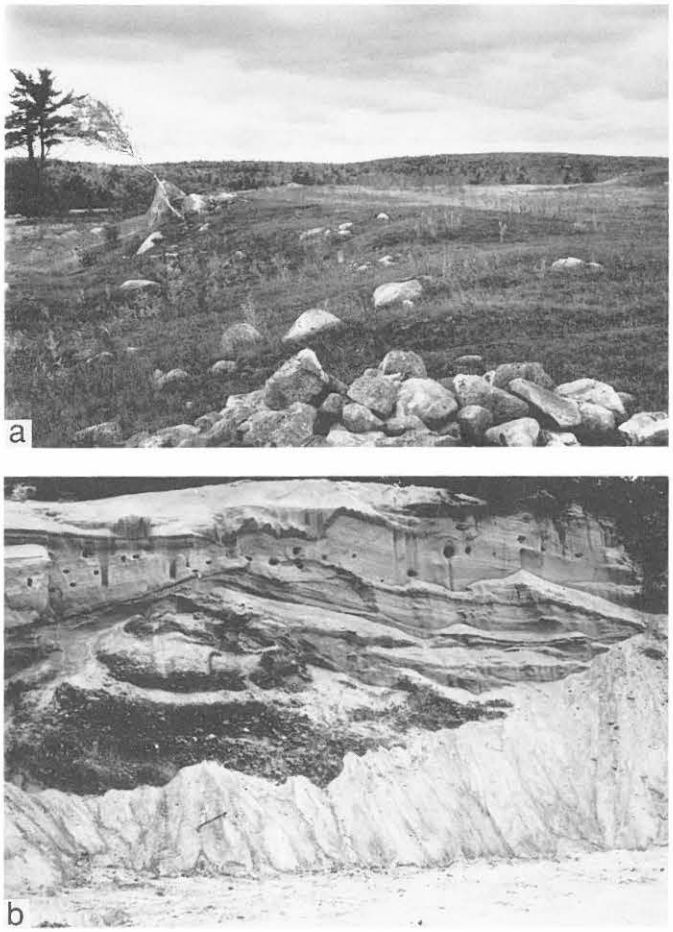

Figure 23. DeGeer moraines. (a) Morphology of DeGeer moraine in the vicinity of Bucksport, Hancock County. Proximal side of moraine is to the right. (b) Gravel core of DeGeer moraine, vicinity of Kennebunk, York County. Moraine is overlain by sand of the subaqueous fan association. Shovel for scale.

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

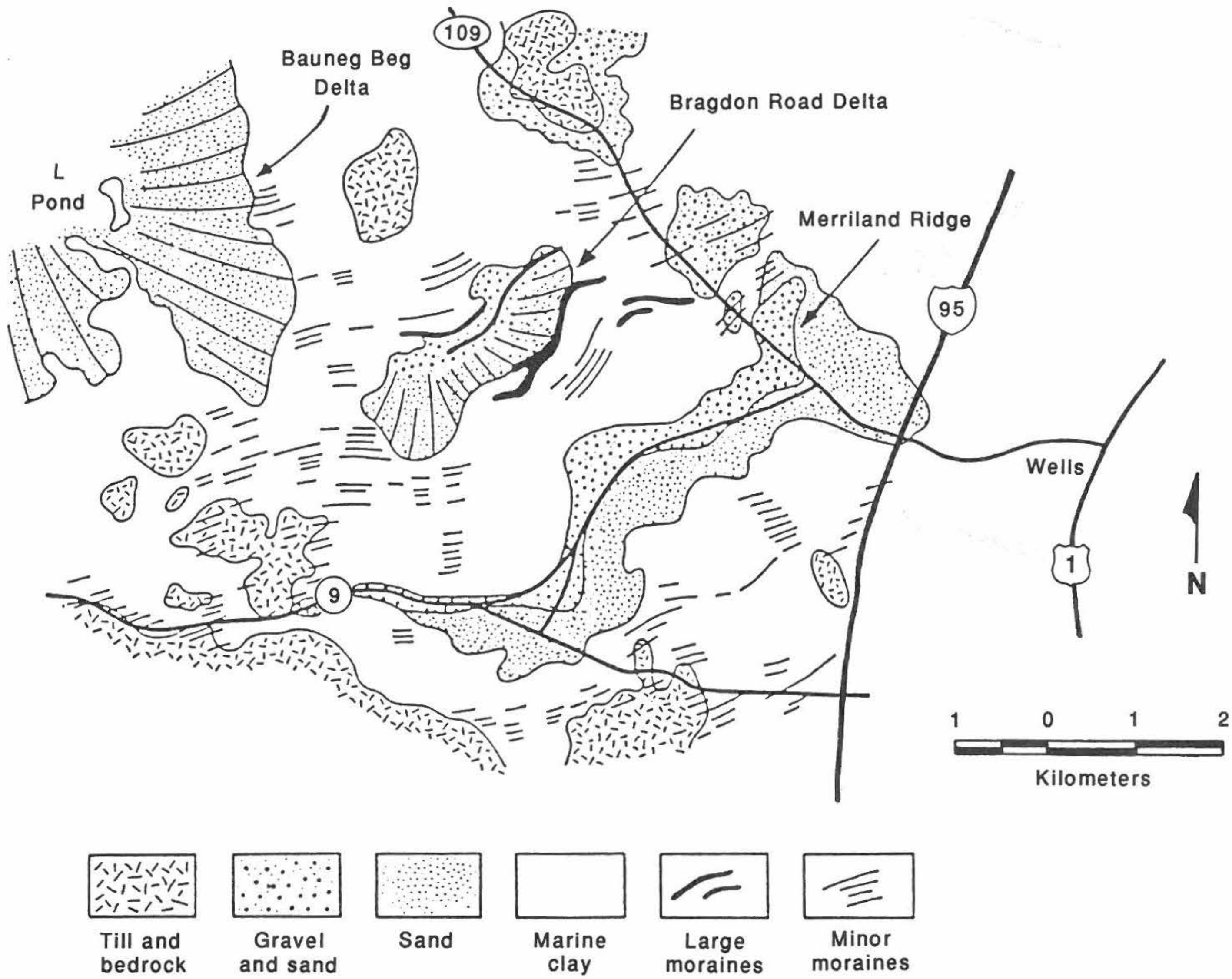

Figure 24. End moraines and glaciofluvial deposits in the vicinity of Wells, Maine (from Smith, 1982).

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.,

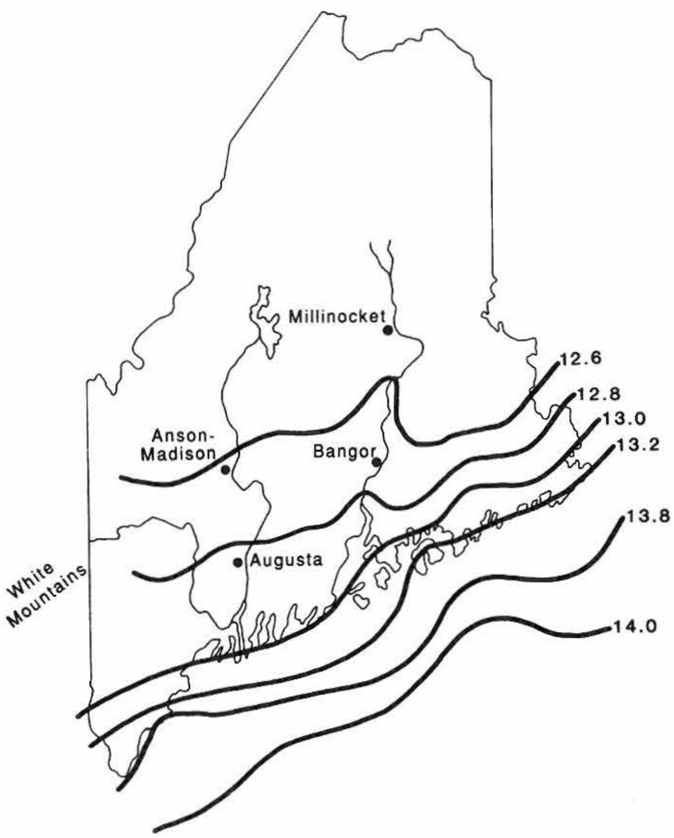

Figure 25. Generalized model of ice retreat from coastal Maine (from Smith, 1985). Location and form of ice-frontal positions based on radiocarbon-dated glacial-marine deposits and distribution of coastal end moraines. Ice-frontal position at 14,000 yr B.P. after Fastook and Hughes (1982). Ages in thousands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

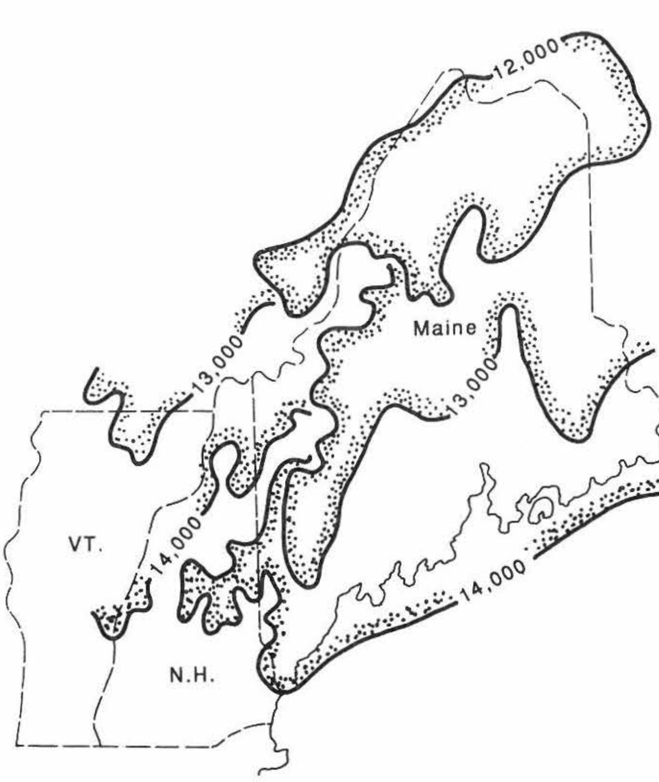

Figure 26. Generalized model of deglaciation of coastal Maine, according to Davis and Jacobson (1985). Ice-frontal positions compiled from Figures 9, 10, 11 of Davis and Jacobson (1985).

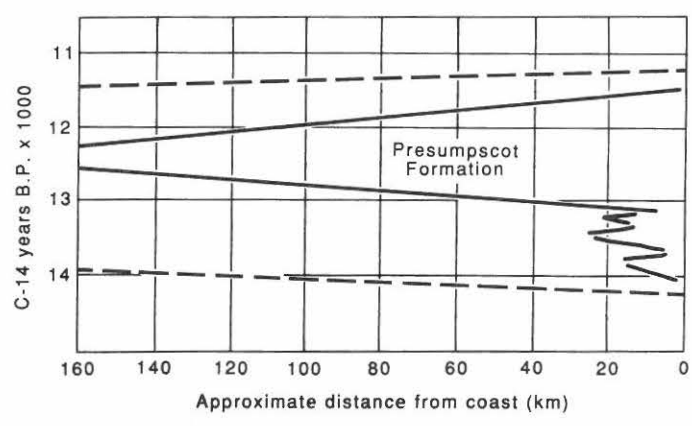

Figure 27. Time-distance curves depicting ice-retreat from coastal Maine. Solid line depicts: ice retreat and coastal submergence (lower line), and isostatic recovery and coastal emergence (upper line) according to Smith (1985). Dashed line depicts: ice retreat and coastal submergence (lower line), and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

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 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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