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Glaciomarine Deltas of Maine and Their Relation to Late Pleistocene-Holocene Crustal Movements

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ABSTRACT

The deglaciation of southern Maine in late Wisconsinan time was accompanied by marine submergence of the isostatically-depressed coastal lowland. Glacial meltwater streams deposited over 100 Gilbert-type deltas into the sea during the transgressive phase of submergence, between about 14,000 and 13,000 yr B.P. These deltas were deposited either in contact with the ice margin or very close to it. They are classified into four categories based on depositional environment: ice-contact deltas (39%), esker-fed ice-contact deltas (30%), leeside deltas (26%), and distal outwash deltas (5%). Most of the deltas for which subsurface data are available accumulated in water less than 80 m deep, and the ice-contact deltas are believed to have formed along a grounded tidewater-glacier margin. Bedrock strike ridges and other hills slowed the retreat of the ice margin, causing brief stillstands during which deltas were deposited adjacent to these topographic highs. The sequence of glaciomarine deltas in Maine probably formed when the transgressing sea had reached its maximum depth and relative sea level was beginning to fall in response to isostatic crustal uplift.

The elevations of the contacts between the topset and foreset beds of 65 deltas in Maine and New Brunswick were measured in order to locate the positions of sea level to which the deltas were graded. These elevations were plotted and contoured to determine whether the late-glacial crustal uplift pattern has been modified appreciably by Holocene crustal movements. The elevation data for southwestern Maine indicate a minimum postglacial tilt of 2.82 ft/mi (0.53 m/km) in the central Kennebec Valley region, with higher elevations to the northwest. Deltas in eastern coastal Maine have anomalously low elevations relative to those in adjacent parts of Maine and New Brunswick. The delta elevations in the eastern region may have resulted from variations in the glacio-isostatic tilt direction across Maine and/or lowering of relative sea level by crustal uplift as the deltas were deposited. Alternatively, these

elevations may define a zone of Holocene crustal subsidence with maximum relative downwarp (at least 40 ft) in the Machias-Eastport area. Contouring of delta elevations also revealed variation in the spacing of the contours between central Maine and the southern extremity of the state. This variation probably resulted from regional differences in the history of late-glacial sea-level change and/or crustal uplift in response to deglaciation.

INTRODUCTION

A variety of recent studies suggest that downwarping of the earth's crust has occurred in parts of southern Maine during Holocene time (Anderson and others, 1984). The effects of the downwarp are clearest along the coast, where the sea has transgressed over man-made structures (Smith and others, this volume). Leveling data indicate that crustal subsidence during the twentieth century has been most rapid -- perhaps as much as 0.9 m/century -- in eastern Washington County (Tyler, this volume). However, until the present investigation, no data were available regarding the possible amount of relative downwarping throughout Holocene time.

Much of southern Maine experienced marine submergence during the retreat of the last (late Wisconsinan) continental glacier that covered New England. This submergence occurred because the earth's crust was still depressed by the weight of the ice sheet as deglaciation began, enabling the sea to flood Maine's coastal lowland in spite of the lower eustatic sea level of late-glacial time. However, isostatic crustal uplift occurred rapidly as the state was deglaciated (Belknap and others, 1987), with the result that glaciomarine deltas formed during the submergence now stand at elevations of up to 422 ft* (128.6 m) above present sea level. We examined these deltas in order to determine whether their present elevations are solely the consequence of uplift resulting from deglaciation, or whether their positions have been modified by Holocene crustal movements. The hypothesis to be tested was that prolonged vertical crustal movement at the rates that appear to be happening today in Washington County should have noticeably changed the elevations of the glaciomarine deltas. A map on which these elevations are contoured should reveal the effects of postglacial tilt, and might also show displacement of the isopleths (equal-elevation contours) resulting from neotectonic crustal disturbance. The preparation of such a map was one of the chief objectives of this study. We also collected information on the origin and stratigraphy of Maine's glaciomarine deltas, and those findings are presented here.

The first part of the investigation (during 1980-1982) was carried out on deltas located between the Maine-New Hampshire border and Penobscot Bay (Figure 1). The stratigraphy of many of the deltas in this region can be examined in gravel pits. Moreover, the central part of the coastal zone appears to have been an area of relative crustal stability during the period covered by Tyler and Ladd's leveling data and perhaps also during most of Holocene time. Thus, this area is a potential source of base-line glacio-isostatic uplift data against which the Washington County region can be compared.

In 1982 the study was extended to the remainder of southern Maine, including Washington and eastern Hancock Counties, as well as adjacent New Brunswick and New Hampshire. A preliminary report by Thompson and others (1983) included a map showing the locations of 69 deltas judged to be suitable for determining the configuration of the uplifted plane of maximum marine submergence. Their map showed the elevation of the contact between the topset and foreset beds in each delta, which indicates the position of relative sea level when the delta was deposited. A number of deltas in remote areas of Washington County did not show exposures of the topset/foreset contact, but in some of these cases it was possible to measure the elevations of the distal ends of meltwater channels on the delta tops. A few prominent wave-cut terraces were also incorporated in the early phase of the study.

The elevation data from which the marine-limit surface was contoured have been refined since the preparation of our 1983 report. We have reevaluated some problematic localities, located and surveyed fresh pit exposures in certain deltas, and utilized new 1:24,000-scale topographic maps. Figure 2 shows the revised contour map of delta elevations, which chiefly comprises deltas whose elevations have been precisely determined by leveling. This map is the focus for much of the discussion that follows. Data from coastal New Hampshire are omitted from Figure 2 because new and more accurate data on the elevations of glaciomarine deltas in this state are presently being obtained by the U.S. Geological Survey (C. Koteff, pers. comm., 1988). The elevations of emerged shorelines that were shown on our earlier map have also been omitted here. Many shorelines fit the elevation pattern defined by the deltas, and thus probably formed at the upper marine limit, but it is difficult to decide exactly which points on shoreline features (wave-cut cliffs and terraces) should be surveyed to define the same water level as the deltas.

PREVIOUS WORK ON GLACIAL ISOSTASY AND GLACIOMARINE DELTAS IN MAINE

The presence of marine clay deposits in southern Maine has been known since the earliest geological survey of the state by C. T. Jackson (1837). Jackson concluded that these deposits reached their present position through uplift of the land, but he

* Elevations and related measurements are expressed in feet to facilitate comparison with elevations on topographic maps.

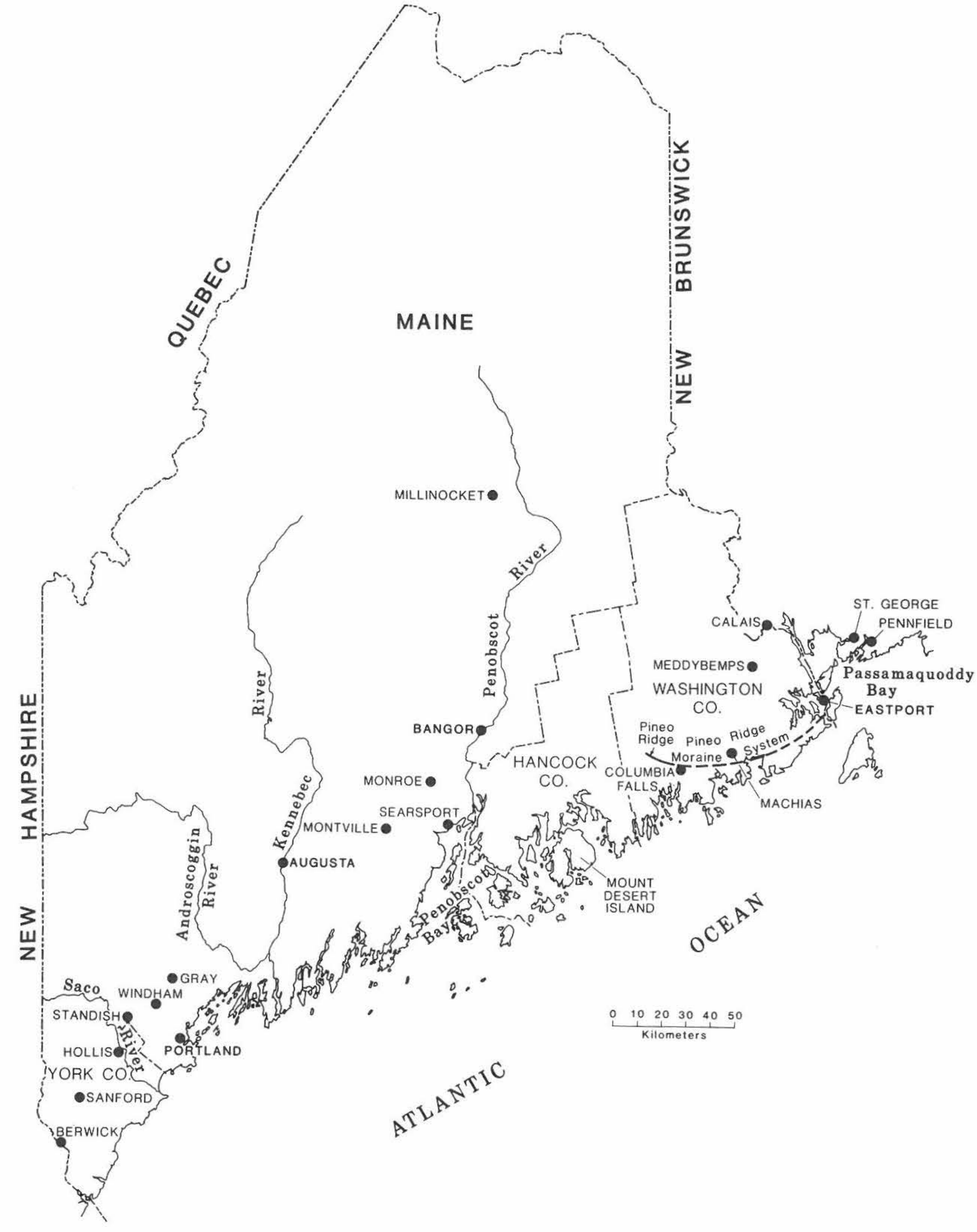

Figure 1. Map showing locations of geographic features mentioned in text.

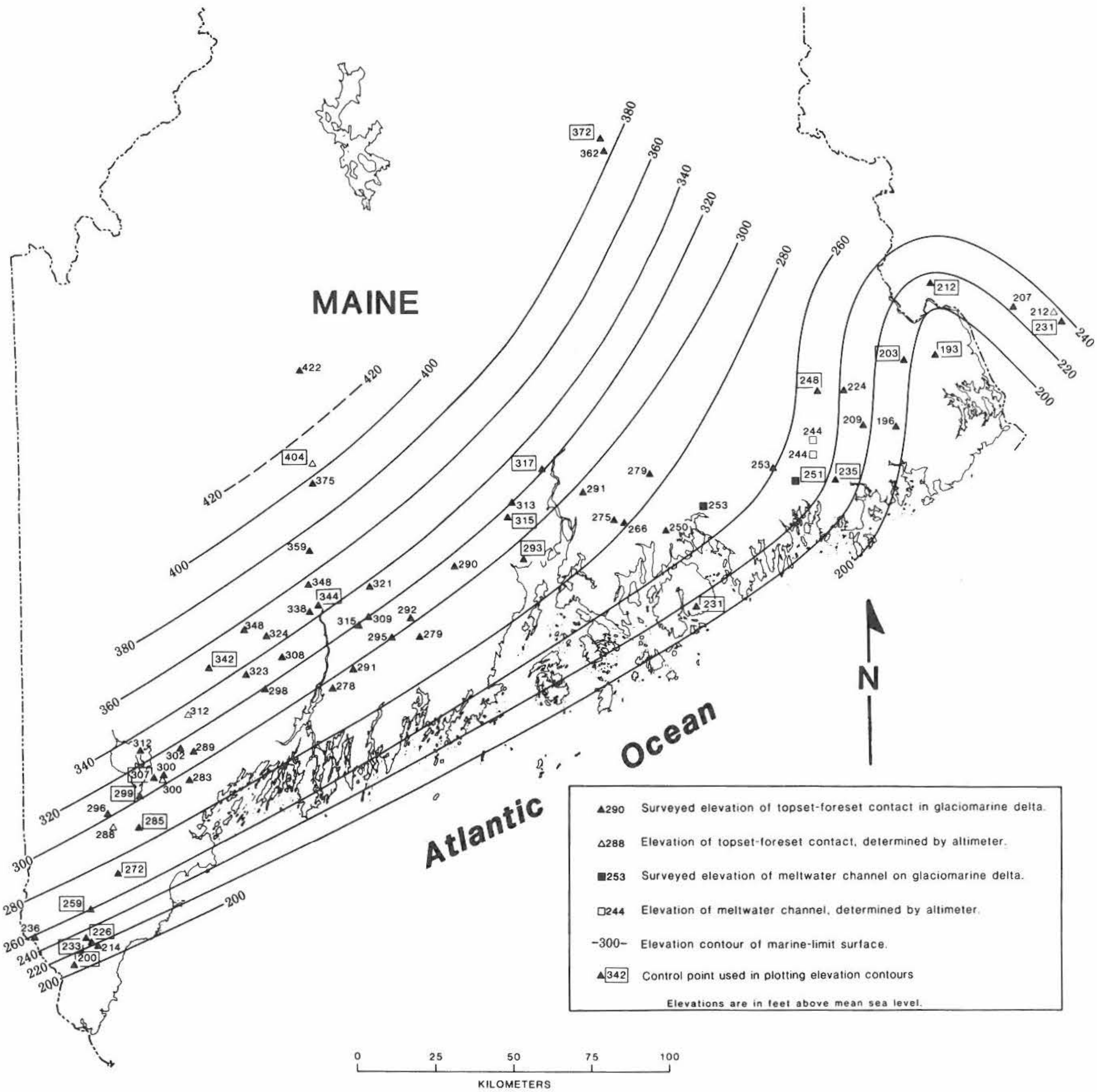

Figure 2. Map showing elevations of surveyed glaciomarine deltas in Maine and adjacent New Brunswick.

was unaware of their glacial origin. The former occurrence of continental glaciation in New England was recognized by the mid 1800's, and shortly thereafter geologists began to study the complex interactions of glacial ice with the changing levels of the land and sea.

As noted by Meyer (1986), the concept of lowering of eustatic sea level by incorporation of sea water into glaciers was not widely accepted until the early 1900's. However, the idea that vertical movements of the continents occur in response to the growth and decay of ice sheets appears to have been favorably received in the United States by the late 1800's (Flint, 1965). The earlier acceptance of the latter theory resulted in part from Gilbert's work on the tilted shorelines of pluvial Lake Bonneville (Gilbert, 1890). Prior to Gilbert's report, Shaler (1874) proposed that the geologically recent depression of the land in Maine, together with the resulting marine submergence, was due to loading by glacial ice. Shaler also estimated the altitude of the marine limit at several places in coastal Maine, and he pointed out that crustal depression had been greater in Maine than in the Boston area. DeGeer (1893) measured the altitudes of raised marine shorelines on Mount Desert Island, and coined the term "isobase" to refer to contours of equal postglacial uplift on his map showing the "last changes of level in eastern North America". This map shows northeast-trending isobases in Maine, with greater uplift to the northwest.

Stone (1890, 1899) provided an extensive description of the glaciomarine deposits of Maine, including the deltas. He observed that several deltas may occur along a single esker system, each delta marking a retreatal position of the ice front. Stone (1899) found the altitude of the marine limit to be about 230 ft (70.1 m) along the present coastline -- a figure that is in close agreement with modern estimates for much of the outer Maine coast (Figure 2). His determination of the location of the inland marine limit (Stone, 1899, p. 484) is also very similar to the findings of later workers. However, although Stone knew that the inland deltas are higher than those along the coast, he did not perceive that the tops of the deltas were graded to the ocean surface. Instead he claimed that they formed "at a considerable depth beneath sea level" (Stone, 1899, p. 482). This opinion was based on the presence of nearby clay and sand deposits, thought to be marine, at altitudes higher than the deltas.

Antevs (1928) conducted the first detailed study of marine-limit elevations in Maine, using topographic maps and barometric measurements of raised beaches and deltas. Over central to southwestern Maine, the pattern of the 200-, 300-, and 400-foot isobases shown on his map is broadly similar to the contour pattern of Figure 2. Few data points were presented for the eastern part of the state, where Antevs extrapolated the 300-foot isobase northeast to the New Brunswick border. Leavitt and Perkins (1935) published a second isobase map, which remained the most recent compilation of marine-limit elevations in Maine until the present study. Their map appears to be derived largely from Antevs' work, but includes some additional elevations in eastern Maine. The isobase pattern shown by Leavitt and Perkins does not incorporate the low (220 ft) delta that appears in the eastern portion of their map. Other deltas and shorelines have subsequently been found at even lower elevations in eastern Maine. Leavitt and Perkins had little data on this remote area, but they assumed that it had experienced less postglacial uplift. The present study considers the alternative possibility that the eastern part of the coastal lowland has subsided following postglacial isostatic uplift.

DESCRIPTION OF GLACIOMARINE DELTAS

Geologic Setting

Radiocarbon dates on shells and seaweed from emerged glaciomarine deposits indicate that southern Maine was rapidly deglaciated between about 14,000 and 13,000 yr B.P. (Thompson and Borns, 1985a, 1985b). During this period the earth's crust experienced residual downwarp from the weight of the late Wisconsinan ice sheet. The crustal depression caused marine submergence that extended far inland from the present Maine coast (Figure 3). Large volumes of sediment from the melting ice were transported into the sea. The silt and clay-size fraction dispersed across the ocean floor, forming a widespread deposit known as the Presumpscot Formation, whereas sand and gravel accumulated along the ice margin as deltas, submarine fans, and stratified end moraines (Thompson, 1982). The distribution of the ice-marginal deposits, from the present coastline to the inland marine limit, and their intertonguing with the Presumpscot Formation, show that the marine transgression was contemporaneous with the generally northward retreat of the ice margin. The submergence was short-lived, since isostatic uplift caused the sea to recede to the present position of the Maine coast by 11,000 yr B.P. (Thompson and Borns, 1985b). Smith (1982) and Thompson (1982) have summarized the stratigraphy and origin of the glaciomarine deposits.

Distribution

This study has identified 100 glaciomarine deltas in Maine (Figure 3), but the actual number may be considerably greater -- perhaps more than 150. There are numerous poorly exposed sand and gravel deposits of uncertain origin, especially near the inland marine limit, that may be deltas.

Many of the largest glaciomarine deltas are clustered near the inland limit of marine submergence in eastern and southwestern Maine (Figure 3). The larger number and size of the deltas in these areas may be partly the result of slower glacial retreat and stillstands as the ice margin became grounded in shallow water near the marine limit. Miller (1986) proposed this explanation for the deposition of the large Pineo Ridge delta complex in the eastern part of the state. In addition, local bedrock lithology probably influenced the amount of rock debris incorporated into the ice sheet, and in turn the volume of deltaic sediments generated by melting of the ice. Most of the large deltas are within or southeast of granitic plutons, which were readily quarried by glacial ice and yielded vast amounts of sand and gravel-size sediment.

The scarcity of deltas along the present Maine coastline probably is the consequence of the greater water depths in that area during late-glacial time, which promoted rapid deglaciation by calving of icebergs into the sea. Stratified end moraines and submarine fans were deposited along the grounding line of the ice sheet, but stillstands of the ice margin were generally too short-lived to permit meltwater deposits to build up to the ocean surface and form deltas graded to sea level.

Although the late-glacial marine transgression extended far up the Kennebec and Penobscot Valleys, few deltas have been found in the formerly submerged area of central Maine. The reason for their scarcity is unclear. Sizable submarine fans do exist along esker systems in this part of the state, and there are a few deltas at the marine limit (Figure 3). Perhaps ice retreat occurred too rapidly for deltas to build up to sea level.

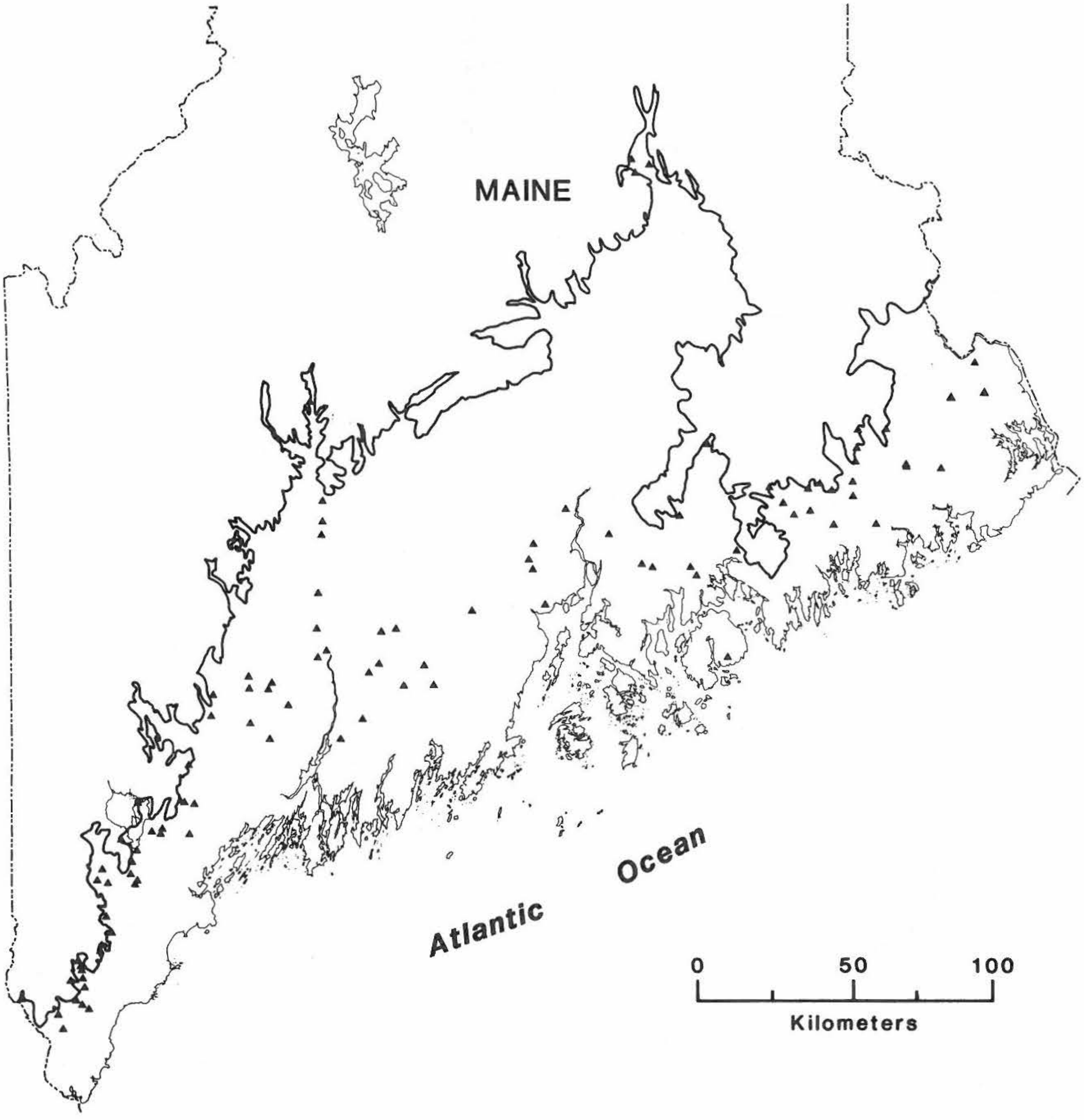

Figure 3. Map showing locations of glaciomarine deltas in relation to the inland limit of late-glacial marine submergence (marine limit from Thompson and Borns, 1985a).

Morphology

In plan view the outlines of the deltas range from the classic fan shape to highly irregular forms. The latter resulted from deposition in contact with decaying ice masses and/or the presence of hilly terrain (including islands) in the area of delta progradation. The Montegail Pond and Meddybemps deltas (Appendix A, Nos. 81 and 94) show good examples of symmetric, arcuate delta fronts (Figure 4). The overall shapes of the deltas are the result of original deposition by glacial meltwater streams. Modifications by marine erosion and redeposition are usually slight and limited to shoreline features on the frontal slopes (discussed below). The proximal sides of the deltas may be steep and very irregular where they were in contact with the ice margin and have collapsed.

The area covered by individual deltas typically is between 0.5 and 15 km^2. Meltwater distributary channels are present on many of the delta plains (delta tops); they are especially

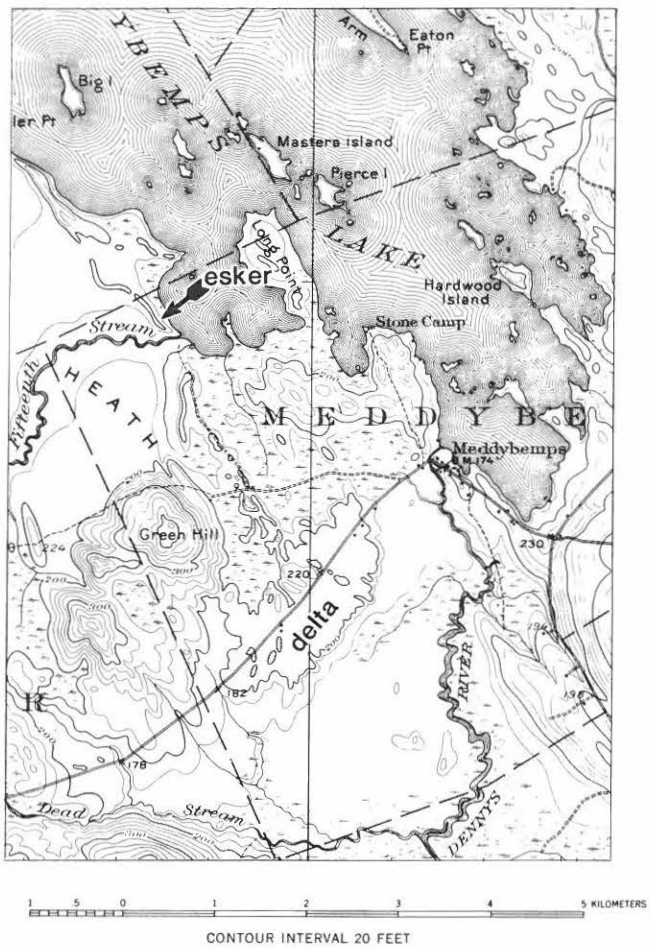

Figure 4. Map showing topography of the Meddybemps esker-fed ice-contact delta.

prominent on some of the deltas in eastern Maine that have been cleared for blueberry cultivation. The channels head at the proximal margins and radiate across the deltas in a seaward direction. Some of them terminate before reaching the front edge of the delta plain, probably because their distal ends have been truncated by marine erosion in the littoral zone. The relationships of channel heads to ice-contact slopes, eskers, and gaps in uplands on the proximal sides of the deltas indicate that the channels were formed by glacial meltwater streams. The channels typically are between 10 and 100 m wide and 1-3 m deep.

Topographic depressions (kettles) occur on most deltas, but vary greatly in size, shape, and abundance. They tend to be larger and more numerous in the central and proximal parts, where sediment was deposited over blocks of stagnant ice that subsequently melted (Figure 5). A few kettles may have originated from melting of icebergs that lodged against delta fronts and were partly or totally engulfed as the deltas prograded around them. The deeper kettles commonly extend down to the water table and contain small ponds, lakes, or bogs.

Figure 5. Northeastward view across the Franklin delta, showing kettle in foreground and uncollapsed delta plain in distance.

Shoreline features, formed during the late-glacial marine submergence, are present on the flanks of some of the deltas. The highest of these features are storm berms that have been found on the seaward edge of several delta plains. A prominent example occurs locally on the south rim of the large delta at Pennfield, New Brunswick (Appendix A, No. 99). The berms are up to about 3 m high.

Wave-cut cliffs and associated terraces have been carved by marine erosion on the sides of some deltas, and may occur on both landward and seaward slopes. Particularly well-developed examples can be seen on the fronts of the Pineo Ridge East, Columbia Falls, and Pennfield deltas (Appendix A, Nos. 80, 88, 99) (Figure 6). In all three of these examples, the base of the wave-cut cliff (which lies along the landward edge of a broad wave-cut terrace) is approximately 5-6 m below the top of the delta. These and other wave-cut cliffs are believed to have formed during the regression of the sea from its highest stand

Figure 6. Wave-cut cliff and terrace on front of the Pennfield delta.

(Thompson and Borns, 1985b). However, the significance of the shorelines remains unclear. They may represent one or more stillstands of sea level, or merely brief erosional episodes resulting from storms. It is also possible that erosion down to wave base carved the terraces before relative sea level fell below the altitude of the delta plains. Because of these uncertainties and the scattered distribution of the shorelines, we have not attempted to correlate them and use them as indicators of crustal movement.

In some instances, deltas have been modified by nonglacial streams during or following the offlap of the sea. These deltas are situated in valleys where downcutting by meteoric streams occurred as relative sea level dropped. Delta surfaces are terraced as a result of this process. The higher terraces may be only slightly lower than the original delta plains, and thus may be mistaken for them.

Stratigraphy

Virtually all of the glaciomarine deltas examined during this study are "Gilbert type" deltas (Gilbert, 1885, 1890). The Gilbert delta forms where streams discharge coarse sediment (sand and gravel) into lake or ocean waters, and the sediment supply is sufficiently large in proportion to water depth that the deposit builds up to the surface of the water body. Some of the coarsest sediment delivered to the delta (commonly gravel) is deposited in distributary streams that cross the delta plain, forming a seaward-thinning sediment wedge consisting of fluvial topset beds.

Sand and gravel that is carried past the distributary mouths and accumulates on the relatively steep delta front forms a succession of foreset beds. As the delta progrades, the topset deposits extend seaward over the older foresets. Therefore, a cross section of a glaciomarine delta in southern Maine (as seen in many gravel pits) typically shows the distinctive association of coarser, sub-horizontal topset beds overlying finer-grained, more steeply dipping foreset beds. The latter in turn have built out over the bottomset beds, consisting of sand and silt deposited on the sea floor (Figure 7). The finest sediment (silt and clay) was dispersed into the ocean and settled to form a muddy deposit known as the Presumpscot Formation (Bloom, 1960). The three units comprising the deltas (topset, foreset, and bottomset beds) are described below.

Topset Beds. The thickness of the topset unit is usually between 0.3 and 3.0 m. The topsets consist chiefly of pebble to boulder gravel in various proportions, and lesser amounts of sand. In many cases the gravel fraction is composed predominantly of cobbles and small boulders that must have been transported in high-energy streams. (Rarely, boulders as large as 1 m in diameter occur in the topset unit.) These gravels are poorly sorted and massive or weakly stratified. Where stratification does exist, it consists of sub-horizontal parallel bedding and small to large-scale cross-bedding. Bedding is most clearly defined where the topset unit contains sand layers. Discontinuous beds of silt to fine sand that occur in the topsets of some deltas may be tidal deposits. However, the tidal range of the late-glacial sea and the degree of tidal influence on the deposition of the glaciomarine deltas have not been evaluated for southern Maine.

Foreset Beds Foreset beds originated as part of the sediment that was carried past the seaward edge of the delta plain accumulated on the delta front. Sediment transport down the face of the delta probably occurred by avalanching, grainflows, and turbidity flows resulting from the rapid influx of sediment at the mouths of distributary streams (J. Boothroyd, pers. comm., 1986). Vertical sections of foreset beds up to 23 m thick have been observed in borrow pits (East Gray delta, Appendix A, No. 32). Individual sets of foreset beds occasionally can be seen to have a lobate form (Figure 8a), and most deltas are thought to consist of a series of overlapping foreset lobes.

The textures of the foreset beds range from very fine sand to boulder gravel. The degree of sorting is variable, but many foresets are composed of poorly sorted pebbly sand to sandy gravel. Stratification is typically well developed except in the proximal parts of ice-contact deltas, where coarse gravel was deposited by high-discharge meltwater streams at the glacier margin. The foresets exhibit parallel bedding within each delta lobe, with individual beds being massive or graded and sharply bounded (Figure 8b). Observed dip angles of foreset beds commonly are in the range of 10-30^{o}.

Locally the foreset beds have been deformed by penecontemporaneous deformation structures of the types described by Reineck and Singh (1980). These include a variety of slump structures, ranging from slight bending and overturning of the foresets to complex folding and faulting where beds have slid down the delta front. Rapid sedimentation and overloading or oversteepening of the foreset slope was responsible for this deformation. "Dish structures" have been observed in the foreset beds of a few deltas. These are ruptures of the bedding, produc-

Figure 7. Cross section through the Globe delta. Pit face exposes topset, foreset, and bottomset beds.

Figure 8. (a) Pit exposure showing lobate foreset beds in the Erskine Academy delta. (b) Pit face showing vertical section through foreset beds in the Erskine Academy delta. Shovel is about 0.5 m long.

ing concave-upward structures, that form by escape of water "during the consolidation and dewatering of quickly deposited sediments which undergo liquefaction and fluidization" (Reineck and Singh, 1980, p. 89). Large dish structures (at least 1.4 m high) have been observed in sandy foreset beds of the Searsport delta (Figure 9; Appendix A, No. 62).

At a few localities delta foresets are interbedded with silty glaciomarine sediment belonging to the Presumpscot Formation. More commonly the latter formation overlies the foreset unit. The contact between these units may be either sharp or gradational. Fossils have not been found in the marine silts that intertongue with the foresets, presumably because the influx of sediment-laden fresh water created a hostile environment for marine organisms while the deltas were being deposited.

Bottomset Beds The bottomset beds are fine-grained proximal sea-floor deposits that accumulated in front of the deltas. Many borrow pits are not deep enough or properly situated to expose bottomset beds; but where these beds can be seen, they are generally sandy, sub-horizontal, and underlie foresets that buried them as the delta prograded. The bottomsets presumably extend seaward from the delta fronts, but here they are concealed by the overlying Presumpscot Formation.

It is difficult to determine the thickness of the bottomset unit from test-hole logs because of the uncertainty in locating boundaries (which may be gradational) between bottomsets and texturally similar foresets. Test borings in southern Maine suggest that the bottomset units may attain a thickness of 10 m, and in places are likely to be considerably thicker (Tolman and others, 1982). Subsurface data also reveal that some deltas have prograded over as much as 18 m of glaciomarine silt-clay.

Depositional Environments

The glaciomarine deltas in Maine can be classified into the following four categories, which are based on depositional environment: (1) ice-contact deltas, (2) esker-fed ice-contact deltas, (3) leeside deltas, and (4) distal outwash deltas. Each of these categories is described below.

Ice-Contact Deltas. Of the 101 deltas that were located during this study, 40 are basic ice-contact deltas. The sediments comprising this class were deposited in the sea along the ice margin, and there is little or no evidence (such as eskers) that the meltwater streams that fed these deltas were confined in englacial conduits. A good example of this category is the Pineo Ridge East delta in Columbia, Maine (Appendix A, No. 80), where distributary channels on the delta plain originate along the proximal margin. This side of the delta has a steep ice-contact slope along part of its length, and elsewhere heads against the Pineo Ridge moraine (Miller, 1986). The latter features mark the position of the ice margin when the delta was deposited; to

Figure 9. Dish structures (center) in fine-grained foreset beds of the Searsport delta. Foresets are truncated by a gravel unit (top) which is probably a nearshore deposit formed by erosion of the delta during regression of the sea.

the north there is only the broad basin occupied by the Great Heath. Thus it is inferred that the Pineo Ridge East delta formed as meltwater streams deposited sediment at the glacier terminus during the stillstand in which the accompanying moraine was formed.

As noted by Crossen (1984), several other characteristics indicating an ice-marginal environment may be seen in the proximal parts of ice-contact deltas. These are: coarse, bouldery gravel, locally containing striated clasts; folds and faults that resulted from collapse adjacent to melting ice; and diamicton lenses (flowtills). Crossen also pointed out the close association between many of the ice-contact deltas in southwestern Maine and strike-controlled bedrock ridges. These deltas were deposited while the ice margin was temporarily pinned against the bedrock highs. Flowtill deposits on the proximal sides of the ice-contact deltas are not as common as might be expected, given the close spatial and temporal association of these deltas with clusters of end moraines formed along the active ice margin. The rarity of flowtill exposures may be due in part to the locations of borrow pits, many of which are not in the ice-contact portions of the deltas, and it may be partly the result of much deltaic sediment having been delivered by englacial streams (discussed below).

Although the simple ice-contact deltas do not connect with esker systems, it has been suggested that meltwater emerging from englacial and subglacial tunnels deposited these deltas (J. Boothroyd, pers. comm., 1986). Gustavson and Boothroyd (1987) demonstrated the importance of tunnel flow in the Malaspina Glacier in coastal Alaska. Most of the stratified sediments at the terminus of this glacier were deposited from fountains and tunnel mouths along the ice margin. The presence of one or more short "tails" of stratified drift extending from the proximal margins of several ice-contact deltas in Maine suggests that tunnel-mouth deposition, rather than subaerial outwash of debris along a broader zone of the ice margin, was at least partly responsible for constructing this type of glaciomarine delta. This model applies to the Windham Hill delta in Windham (Appendix A, No. 26), which is one of the best examples of the ice-contact category. The proximal side of Windham Hill consists mainly of an ice-contact slope, but a short esker ridge extends northwestward from the central part of the delta (Figure 10).

Esker-Fed Ice-Contact Deltas. Many of the other investigated deltas (30 percent) likewise were deposited at the ice margin, and in most respects resemble the ice-contact deltas described above. However, this second category is distinguished by the presence of eskers that connect with the proximal margins of the deltas. The eskers mark the paths of feeder channels in which sediment was transported to the deltas. Many of them are DeGeer-type eskers that were deposited in successive segments. These segments commonly terminate at deltas or submarine fans that formed along short-lived ice-margin positions.

An excellent example of a DeGeer esker and associated series of deltas extends from Augusta north to Smithfield (Caldwell and others, 1985). Here there are six ice-contact deltas along a 40-km interval of the esker system. These and other esker-fed deltas show the same relationship to bedrock topography as the non-esker-fed category. Several of the deltas north of Augusta are located in gaps that penetrate northeast-trending strike-controlled bedrock ridges, suggesting that the deltas were deposited when the ice margin was temporarily pinned against these ridges. The Meddybemps delta (Appendix A, No. 94) is another outstanding example of the esker-fed category (Figure 4).

Leeside Deltas. The third category of glaciomarine delta that occurs in Maine is a distinctive type that comprises 26 percent of the deltas that were examined. They are called "leeside deltas" because they are situated on the lee sides (generally to the southeast) of ridges or ranges of hills that extended above the ocean surface during the marine submergence. Typically these topographic highs have a northeast trend that follows the regional strike of bedrock formations and structures. Debris-laden meltwater streams passed through gaps in the ridges and built deltas on the southeast sides. Eskers locally

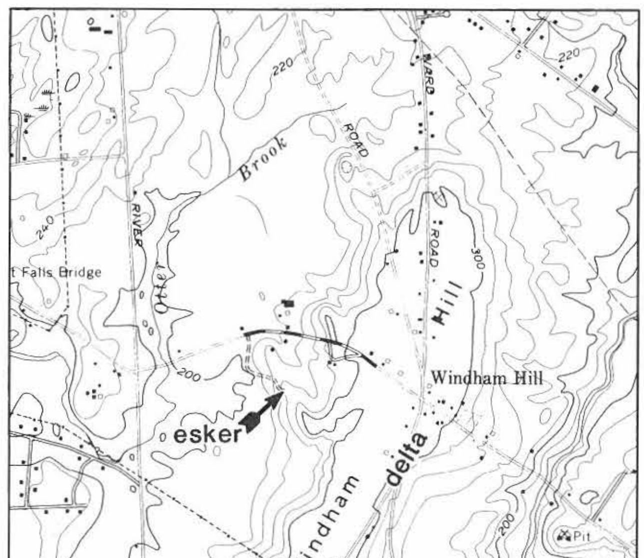

Figure 10. Map showing topography of the Windham Hill ice-contact delta.

occur on the proximal sides of these ridges, and lead to the gaps through which the glacial streams flowed. Some eskers terminate at the gaps, while a few continue through them to the head of the delta on the seaward side.

It is apparent that leeside deltas were localized by terrain factors in much the same way as ice-contact deltas. The regional grain of the topography over much of southern Maine caused the ice margin to become pinned against a succession of ridges. Where sand and gravel was deposited directly in contact with the ice margin, submarine fans -- rather than complete deltas -- often were constructed where the stillstands were too brief for sediment accumulation to reach the ocean surface. However, the leeside deposits are fully developed deltas because most of them were constructed by subaerial streams that reached the sea after crossing hills that stood above the contemporary sea level. This circumstance did not permit submarine fan deposition at the mouths of ice tunnels.

The Irish Hill delta in Monroe (Appendix A, No. 60) is a fine example of the leeside category (Figure 11). The feeder stream for this delta flowed through the narrow col between Irish and Clement Hills and deposited the delta immediately to the east. Another example is the East Gray delta (Appendix A, No. 32). The paths of feeder streams for the latter delta are marked by meltwater channels that cross the ridge extending from Colley Hill southwest to Gray village. A very large kettle -- about 30 m deep -- occurs on the East Gray delta, and kettles are common in the central and proximal parts of other leeside deltas as well. This characteristic prompted Crossen (1984) to call them "ridge and kettle deltas". She concluded that the kettles resulted from collapse of deltaic sediments deposited over ice blocks that had been stranded in the lee of the ridges.

Distal Outwash Deltas. Only five of the deltas located during this study were classified as the distal outwash type, having formed where meltwater streams flowed down a valley and entered the sea at some distance from the ice margin. Three of the distal outwash deltas (Appendix A, Nos. 14, 15, 17) are located in the Hollis-Standish area, at or near the inland marine limit in the Saco River basin. Of these, the North Hollis and South Hollis deltas head in zones of complex ice-contact deposits and may be at least partly esker-fed. The kettled Saco delta likewise may be ice-contact to some degree.

The rarity of valley-train outwash deltas, free of ice-contact influences, is at first surprising in view of the numerous valleys that could have carried outwash to the sea from ice-margin positions above the marine limit. However, this scarcity probably can be explained by the timing and geographic circumstances of ice retreat and isostatic crustal uplift. The inland limit reached by the late-glacial sea was restricted not only by the distribution of lowland areas, but also by the crustal uplift that is thought to have been in progress during deglaciation. Several valleys in the vicinity of the marine limit, such as the upper Kennebec Valley, contain marine clay (Presumpscot Formation) overlain by outwash sand and gravel. The latter deposits were graded to a falling sea level as uplift arrested the marine transgression and

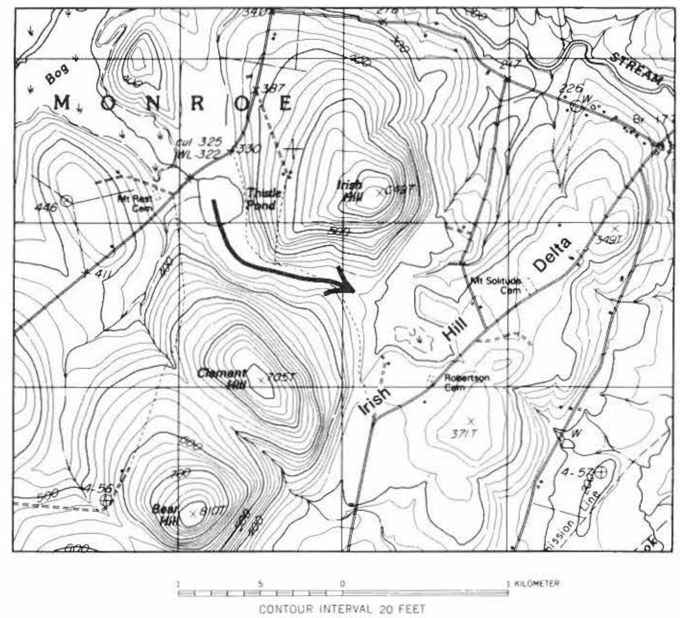

Figure 11. Map showing topography of the Irish Hill leeside delta. Arrow indicates meltwater channel that carried sediment from the glacier margin northwest of the hills. Delta plain is at elevation of about 320 ft; higher areas protruding above the delta surface are till and bedrock.

caused the shoreline to start receding. Thus, fluvial sedimentation and local terracing of earlier glaciomarine deposits replaced the deeper-water fan/delta environment as the receding ice margin separated from the sea along the inland marine limit. If the maximum submergence had persisted longer in central Maine, more outwash deltas could have formed in valleys that drained interior regions.

Water Depth During Deposition of Deltas

In trying to reconstruct the environment in which the glaciomarine deltas were deposited, it is important to know whether the ice margin was grounded or floating. It is improbable that deltas could have been graded to sea level if the glacier margin was a floating ice shelf. In the latter case, the only way to concentrate large volumes of deltaic sediments would have been by deposition at tunnel mouths along the glacial grounding line, where meltwater currents emerged under hydrostatic pressure. This mechanism could have deposited much of the sand and gravel in submarine fans and stratified end moraines (Thompson, 1982); but complete deltas, with topset beds graded to sea level, could not have formed if an ice shelf extended seaward from the grounding line.

Moore (1982) proposed that the glaciomarine deltas in coastal New Hampshire were in fact deposited as "grounding line deltas" beneath a confining ice shelf. However, his model does not explain certain characteristics of Maine deltas that indicate subaerial deposition controlled by sea level. Many of the deltas have broad, flat tops; and meltwater distributary channels are generally graded to a common level along the distal margin of each delta. Moreover, the non-ice-contact leeside and distal outwash deltas are located in settings where Moore's model does not apply.

Crossen (1984) reviewed previous workers' observations on water depths adjacent to grounded versus floating glacier margins in modern environments. The literature suggests that, depending on ice thickness and other local circumstances, some ice margins remain grounded in water as deep as 450 m, while floating ice tongues have been observed in water as shallow as 155 m.

In order to determine the maximum possible water depths in which the Maine deltas were deposited, we examined the results of test borings and seismic-refraction surveys that were carried out on many deltas during gravel aquifer investigations (Tolman and others, 1982; Tepper and others, 1985; Williams and others, 1987). Figure 12 shows the maximum depth to bedrock (or, in a few cases, depth to till) recorded in 53 deltas. The actual water depths in which these deltas formed would have varied at each locality, depending on the sub-delta topography and possible presence of underlying sediments that were too thin to be detected in seismic surveys. Nevertheless, the depth-to-bedrock data approximately indicate the range of water depths. It is apparent from Figure 12 that nearly all of the deltas for which data are available were deposited in maximum water depths of 10-80 m. Depths to bedrock of 20-40 m are most common, while the greatest recorded depth is 104 m (Appendix A, No. 31).

The above data indicate that the glaciomarine deltas were deposited in shallow water, and lead us to conclude that they formed adjacent to the grounded margin of a tidewater glacier. As the ice margin receded inland, shallow conditions at first were encountered locally over topographic highs, and then became more widespread in the vicinity of the inland marine limit. The concentration of large deltas that accumulated as ice retreat slowed in this area is readily apparent on the Surficial Geologic Map of Maine (Thompson and Borns, 1985a).

Relationship of Deltas to Changing Relative Sea Level

It has been demonstrated that most of the glaciomarine deltas in Maine were deposited immediately adjacent or very close to the margin of the late Wisconsinan ice sheet. They formed in a time-transgressive manner as the glacier margin retreated from the coastal lowland between 14,000 and 13,000 yr B.P. They are largely clustered in a zone that extends only about 70 km inland from the outer coast (Figure 3); and the limited available data on the timing of deglaciation (Thompson and Borns, 1985a) suggest that deltas could have been deposited near the marine limit in central Maine at about the same time (close to 13,000 yr B.P.) that deltas were forming much closer to the coast in eastern and southwestern parts of the state. However, it is unlikely that sea level remained static during this period of delta construction, considering the rise in eustatic sea level as global ice sheets melted and the isostatic crustal uplift that was occurring when the ice had retreated from the continental shelf and thinned to a large extent.

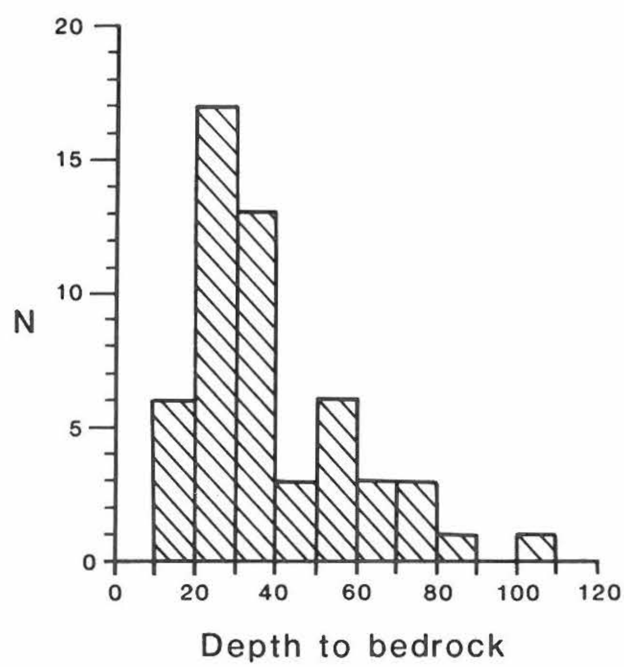

Figure 12. Histogram of maximum recorded depths to bedrock or till beneath glaciomarine deltas.

Several lines of evidence show that nearly all of the Maine deltas were deposited either during the highest stand of late-glacial relative sea level or during the ensuing regression caused by crustal rebound. First, silty ocean-floor sediments (Presumpscot Formation) commonly mantle the delta fronts, but do not cover the tops of any deltas. This would not likely be the case if the deltas had been drowned by a rising sea level soon after their formation. Second, well-preserved kettles and meltwater channels occur on the tops of many deltas; these features would have been destroyed or greatly modified by a postdepositional sea-level rise. Moreover, there is at least one locality (Pineo Ridge East delta) where meltwater channels crossing a delta have been terraced because sea level fell while the channels were still carrying water from the adjacent ice margin. Finally, comparison of the elevations of delta topset/foreset contacts suggests that -- over most of Maine -- they were graded to a regressing sea. These elevation data are discussed below.

The above generalizations hold for the majority of the deltas in Maine, but possible exceptions have been discovered in the vicinity of the Maine-New Hampshire border. At the Berwick delta in southern York County (Appendix A, No. 1) a pit exposure seen in 1986 showed 1.2 m of sand and gravel overlying the topset beds. This sand and gravel unit exhibits sub-horizontal bedding, in contrast to the fluvial cross-bedding of the topsets, and is interpreted as a marine shoreline or nearshore deposit formed by reworking of the topsets. The Berwick delta may have been overwashed during storms, or it may have been drowned if eustatic sea-level rise exceeded the rate of isostatic crustal uplift during the early part of Maine's deglaciation. In either case, the veneer of reworked sand and gravel capping the delta was derived from partial erosion of the topset beds. C. Koteff (pers. comm., 1986) has found similar marine deposits extending across the tops of glaciomarine deltas in coastal New Hampshire. The high shoreline or nearshore deposits, together with the scarcity of meltwater channels on the deltas of this area, suggest that eustatic sea-level rise kept pace with -- or even exceeded -- crustal rebound as the ice margin receded from the southern tip of Maine and adjacent New Hampshire. This situation would have been short-lived, since rebound became dominant and caused relative sea level to fall as the remainder of southern and eastern Maine was deglaciated.

DELTAS AS SEA-LEVEL INDICATORS

Figure 2 shows the locations of deltas that were surveyed to determine the elevation of the upper limit of marine submergence at their respective locations. These elevations are listed in Appendix A. Most of them were precisely determined by leveling from the nearest usable bench mark or surveyed road intersection. The resulting measurements were rounded to the nearest foot. A few elevations recorded by surveying altimeter are also included in Figure 2, but they may be in error by as much as several feet.

The contact between the topset and foreset beds in each delta approximates the water level to which the delta was graded. Depending on local circumstances, this topset/foreset boundary may have developed at a shallow water depth (probably not more than 1-3 m below actual water level); but it certainly is not higher than the contemporary sea level. Gustavson and others (1975) made the following observation regarding topset/foreset contacts in glaciolacustrine deltas:

> "The contact between topset and foreset beds in glaciolacustrine deltas is erosional and has at times been taken to correspond to the elevation of the lake into which the delta was built. Pleistocene lake levels have been recorded to within a fraction of an inch based on this contact. The topset/foreset contact actually records the position of the channel bottom of the stream that supplied sediment to the delta and is always some distance below the actual lake level."

Considering the above observation, the authors surveyed the highest exposed point on each contact and plotted only the maximum values on Figure 2. This procedure was intended to achieve a greater consistency in the elevation data, especially since the influence of tidal range on the topset/foreset contact is unknown.

Sand and gravel pits are common in southern Maine, so at least one exposure of the topset/foreset contact could be found in many deltas. Where suitable exposures are lacking, meltwater channels on the delta tops locally provide an approximate measure of sea level. The channels could have extended into shallow marine waters and thus indicate a minimum sea-level elevation (as do the topset/foreset contacts, which represent the lower limit of channel erosion). On the other hand, the proximal portions of the channels were subaerial and lie above the contemporary sea level. The poorly understood influence of tidal fluctuations during deposition of the deltas is yet another factor to be considered. Because of these complexities, the authors measured meltwater channels only where the topset/foreset contacts were not exposed. The scarcity of bench marks and other points of known elevation hindered the use of channels in remote parts of eastern Maine.

Problems that became evident while measuring delta elevations are summarized here in order to provide a better understanding of the data base used in compiling Figure 2. Parts of some deltas have obviously collapsed during melting of glacial ice masses with which they were in contact. These sites were avoided, but a few other topset/foreset contacts that had been slightly collapsed (and appeared undisturbed) were inadvertently included in the earlier results of Thompson and others (1983). The data obtained from these deltas account for some of the anomalously low elevations on our 1983 map. Sites where this problem was recognized have been resurveyed, often with the benefit of better pit exposures.

Difficulty in the identification of topset/foreset contacts may arise where marine or fluvial erosion has modified deltas subsequent to their deposition. Marine erosion during the offlap of the sea formed wave-cut terraces (commonly veneered with gravel) high on the flanks of some deltas. Excavations in these terraces reveal stratified beach sediments overlying eroded foreset beds. A good example of marine terracing occurs on the delta at Pennfield, New Brunswick (Appendix A, No. 99). A large borrow pit in the southwestern extremity of this delta shows a gravel layer that resembles a topset unit and overlies foreset beds (Figure 13). However, the flat ground surface at the top of the pit can be traced eastward as a wave-cut terrace on the front of the Pennfield delta (Figure 6). The false topset/foreset contact has a maximum elevation of 203 ft (61.9 m), in contrast to a definite contact at 231 ft (70.4 m) in a pit located on the north side of Route 1, 0.8 km east of Pennfield Corner.

The shoreline deposits on terraced delta fronts often can be distinguished from topset beds if there is a good exposure. The contact between these sediments and the underlying foreset beds commonly dips seaward at a steeper angle than a sub-horizontal topset/foreset contact. Bedding within the shoreline unit generally has a steeper dip than topset beds, may be convex-upward, and often terminates against the contact with eroded foresets. In places a lag concentrate of cobbles and boulders occurs along this contact. Finally, the multi-directional, small-scale fluvial cross bedding present in many topset units has not been observed in the shoreline deposits.

Figure 13. Marine shoreline gravel (top) on wave-cut terrace truncating foreset beds of the Pennfield delta.

When spurious topset/foreset contacts of the type described above had been eliminated, there was still considerable discrepancy among some groups of closely spaced elevation measurements. In two cases the surveyed elevations from individual deltas varied by 3 m, with the higher elevations recorded in the proximal parts of the deltas. It is unlikely that crustal rebound would have lowered relative sea level as much as 3 m during construction of the small deltas on which these variations were noted; but it is equally implausible that the topset/foreset contact would dip seaward unless base level dropped as the delta prograded. Reexamination of one of these problematic deltas (in Montville, Figure 1; Appendix A, No. 58) revealed a subtle fluvial terrace on the southeastern part of the delta top. It probably was carved by the nearby stream in early postglacial time, as uplift was just starting to elevate the region above contemporary sea level. The thin veneer of terrace gravel deposited on the eroded delta foresets was mistaken for a topset/foreset association.

Other anomalous delta elevations were the result of foreset beds being mistaken for topsets. This error may occur in shallow pits and other places where only the uppermost part of the foresets can be seen. Individual groups (sets) of parallel foreset beds may be only 1-2 m thick in the upper part of a delta, where they were deposited in shallow water. A vertical succession of these thin foreset lobes can form a larger package (coset) that is several meters thick, as in the Meddybemps delta (Appendix A, No. 94). In places where only the uppermost set is visible, the stratification in the foresets may be confused with the fluvial cross-bedding of the topset unit.

Recognition of the above sources of error has explained some of the unusually low delta elevations obtained earlier in the study. Our revised topset/foreset contact elevations are listed in Appendix A. These data were used to contour the uplifted, time-transgressive marine-limit surface in southern Maine, shown in Figure 2.

SYNTHESIS AND INTERPRETATION OF GLACIOMARINE DELTA ELEVATIONS

Methodology for Contouring and Interpreting Elevations

Sixty-nine deltas were surveyed in order to determine the present elevation of the late-glacial sea level to which they were graded (Appendix A). The coordinates of the surveyed sites are listed in Appendix B. Seven of the deltas were measured by altimeter, and the rest were precisely surveyed. Most of the elevations were obtained from topset/foreset contacts, but meltwater channels were used in four cases where exposures of the topset/foreset contact were lacking.

The delta elevation contours in Figure 2 are isopleths rather than true isobases. This is because the present elevations of the deltas are the consequence of several interacting processes, rather than the simple uplift of a sea-level plane that had been static during ice retreat. Moreover, the contours in Figure 2 are diachronous and should not be construed as representing successive ice-margin positions during deglaciation. Only the highest elevations in any particular area were used in drawing the contours, since these data provide the closest approximation to sea level during the maximum marine submergence. The delta elevation contours are further constrained by the fact that no delta can be higher than the contoured marine-limit surface. A contour interval of 20 ft (6 m) was selected, and the original compilation was done on a 1:500,000-scale map of Maine. Figure 2 is a reduced version of this map.

A number of the deltas shown in Figure 2 are lower than the adjacent contoured surface. Most of these remaining discrepancies are probably due to local variations in former meltwater channel depth above the surveyed topset/foreset contacts, though there may be some complications resulting from the sources of error described above. The delta elevation contours in Figure 2 represent a smoothed surface based on data points judged to be closest to former sea level. In several cases where more reliable survey points superseded earlier data, the new measurements were closer to this surface.

If the crust in southern Maine has simply been uplifted and tilted by postglacial isostatic adjustment, then the delta elevation contours in Figure 2 should be parallel and have an overall northeast-southwest orientation (normal to the flowlines of the late Wisconsinan ice sheet). Large-scale vertical crustal movements that are unrelated to glaciation would be expected to change the spacing and orientation of the contours. However, contour spacing might also vary because of the events related to glaciation, such as changes in eustatic sea level, crustal rebound rate, or the rate of ice-margin retreat while the deltas were being deposited. For example, Figure 14 illustrates the influence that varying uplift rates would have had upon the present elevation gradients of the deltas. The relative importance of these variables is difficult to evaluate. The lack of organic material in the deltas hinders the determination of their chronologic sequence, and the magnitude and rate of eustatic sea-level rise during the

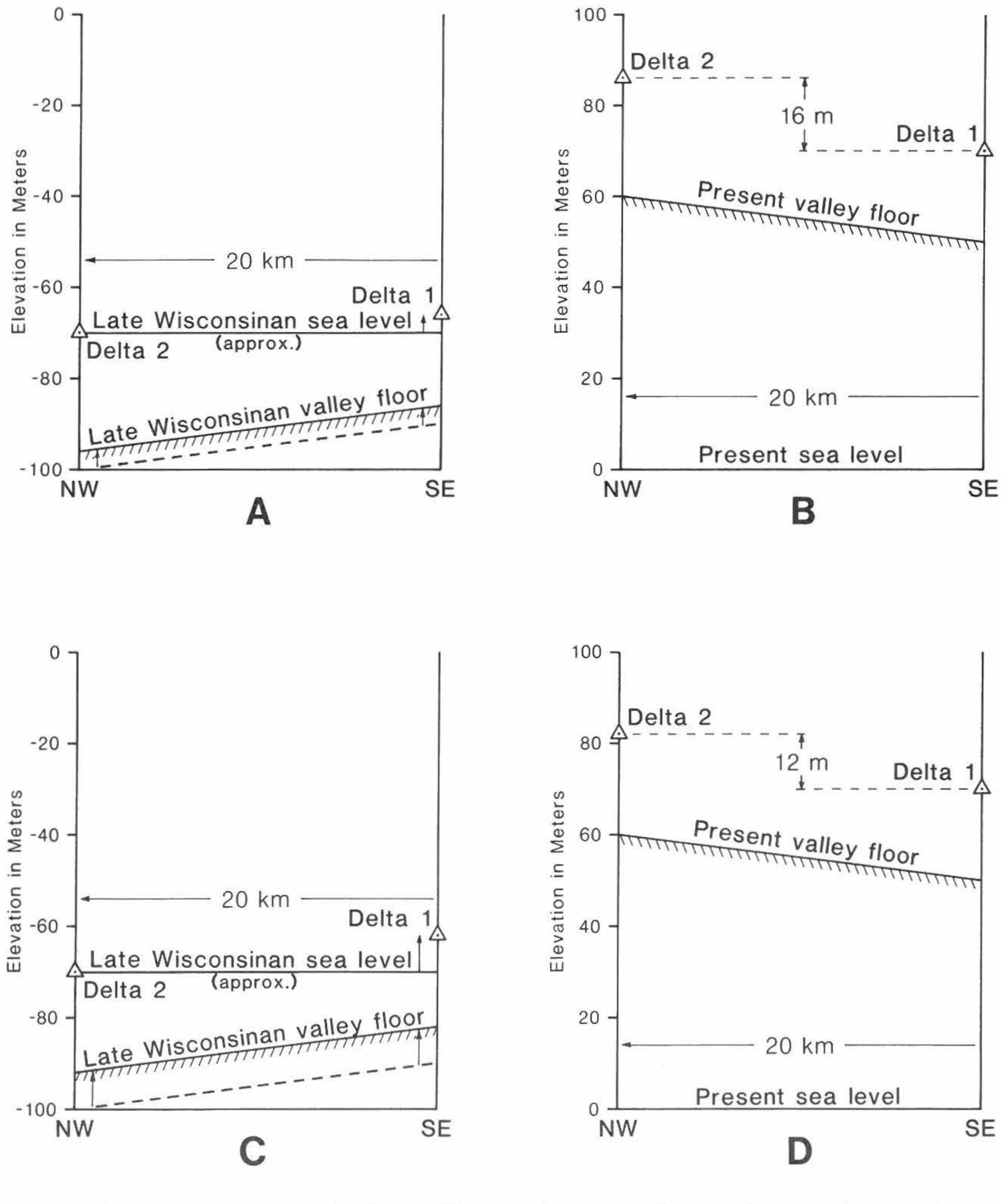

Figure 14. Schematic diagrams showing the effect of different isostatic crustal uplift rates on the present elevation gradient of glaciomarine deltas. (A) Partial rebound occurs between deposition of ice-contact delta 1 at time t_1, and delta 2 at t_2, accompanied by a 20-km retreat of the ice margin during interval t_1-t_2. (B) Rebound is complete. Final elevation gradient between deltas is 0.8 m/km, although differential uplift between points 1 and 2 was 1.0 m/km. (C) Greater rebound occurs during interval t_1-t_2. Other variables are assumed to be the same as in A. (D) Rebound complete. Differential uplift is unchanged from B, but elevation gradient between deltas is only 0.6 m/km. (Estimated position of late-glacial eustatic sea level [approximately -70 m] is from Belknap and others, 1987.)

period of delta formation are unknown. In view of these uncertainties, the isopleth patterns obtained by contouring delta elevations (Figure 2) must be interpreted with caution. Elements of this pattern that may appear to have a neotectonic origin should also be considered as possibly having resulted from the glacial processes outlined above.

In order to test the validity of the contour pattern shown in Figure 2, the delta elevations were subjected to a trend-surface analysis. This analysis included all of the elevation data points except the Norridgewock delta (Appendix A, No. 101), which was discovered near the end of our study and whose elevation has not been precisely determined. First-degree through sixth-degree trend surfaces were fitted to the data. The correlation coefficients for these surfaces range from 0.92 (first degree) to 0.98 (third degree), indicating a close fit to the delta elevations. Figure 15 shows the third-degree trend surface, and Figure 16 is a contour map of the residuals from this surface (deviations of individual data points above or below the trend surface). The results of the trend-surface analysis are compared with Figure 2 in the discussions that follow.

Regional Patterns of Delta Elevations

The delta elevation contours in Figure 2 trend generally northeast-southwest. They are approximately parallel to the coastline over much of southwestern Maine, and normal to the southeastward direction of late Wisconsinan ice flow. As one proceeds into the Penobscot Valley region in central Maine the contours trend more northerly. This shift in direction is assumed to be gradual, but there is little control on the contours north of Bangor, where there are few glaciomarine deltas. The northward curvature across central Maine is based on deltas south and east of Bangor (including some that could not be surveyed) and the isolated pair of surveyed deltas near Millinocket. The northerly trend of delta elevation contours becomes more pronounced in eastern Maine (Figure 2). Here the pattern is complicated by an indentation of the contours. Delta elevations are lower as one proceeds eastward across Washington County, reaching a minimum of 193 ft (58.8 m) near the Canadian border. Then they rise to 231 ft (70.4 m) a short distance farther east in New Brunswick. Possible interpretations of this anomaly are discussed below.

The marine limit surface contoured in Figure 2 has a dip direction of 155° in extreme southwestern Maine, 145° at Augusta, and 120° midway between Bangor and Millinocket. These azimuths are similar to dip directions obtained from the trend surface (Figure 15) for the same areas of the state: 150°, 142°, and 125°, respectively. If these directions indicate the actual postglacial crustal tilt, then the isostatic uplift pattern in southwestern to central Maine suggests that the late Wisconsinan ice sheet radiated across the state from a spreading center located in the direction of southern Quebec. Supporting evidence for this flow pattern is provided by bedrock striation data that show widespread evidence of eastward glacial flow over northern Maine, in contrast to more southerly flow in the southern part of the state (Thompson and Borns, 1985a; Lowell and Kite, 1986).

In southeastern Quebec, elevation contours of the Fort Ann level of glacial Lake Vermont and the ensuing Champlain Sea submergence show patterns similar to Figure 2. The isopleths indicate southeastward crustal tilt (toward western Maine) in the Asbestos area, but curve as they are followed southwestward, showing a more southerly tilt in the Lake Champlain region (Parent and Occhietti, 1988). Work by Shilts (1981) in the Lac Megantic area of southeastern Quebec (near the Maine border) demonstrated that the principal late Wisconsinan glacial flow direction was toward the east-southeast. Again, this is consistent with the southeast tilt in central Maine suggested by Figure 2. A late-glacial residual ice cap persisted in northern Maine (Lowell, 1985), but the degree to which this lingering ice mass affected crustal rebound and sea-level history of areas nearer the coast is not known.

The delta elevation contours in Figure 2 enable a minimum estimate of the amount of crustal tilt resulting from glacio-isostatic uplift. The gradient of the marine-limit surface was measured in the central Kennebec Valley region, where the contours have a relatively uniform orientation. It averages 2.82 ft/mi (0.53 m/km) over a distance of 35.5 mi, measured along a normal passing through Augusta and connecting the 300- and 400-ft contours. A nearly identical gradient of 2.75 ft/mi (0.52 m/km) was obtained from the trend surface in Figure 15. On the latter map we measured the gradient along a normal through the 340-ft contour at Augusta, connecting the 230 and 360-ft contours. The higher trend-surface contours were not included because the curvature near the southwest ends of these computer-generated lines appears to be a boundary effect. Moreover, inclusion of the recently discovered Norridgewock delta (Appendix A, No. 101) probably would cause at least the northwestern part of the trend surface to be slightly steeper than shown in Figure 15.

The marine-limit gradient of 2.82 ft/mi is slightly more than half the uplift gradient of 4.74 ft/mi (0.90 m/km) recorded by surveying glacial Lake Hitchcock deltas in the Connecticut Valley to the west (Koteff and others, 1988). The ice-contact deltas of Lake Hitchcock were graded to an essentially static water level. They provide a good measure of postglacial tilt over a sizable part of western New England, especially if isostatic uplift did not occur until after they had been deposited, as proposed by Koteff and others (1988). These authors found the tilt direction in the Connecticut Valley region to be 159-160°, which is more southerly than the direction that we obtained for the southwestern extremity of Maine (150-155°).

The lesser elevation gradient of the glaciomarine deltas in Maine (compared to the glacial Lake Hitchcock deltas) supports the concept that crustal uplift was in progress and causing relative sea level to fall as southern Maine was deglaciated. Successive deltas were graded to this lowering sea level and presently have a gentler elevation profile than if relative sea level had been static or rising.

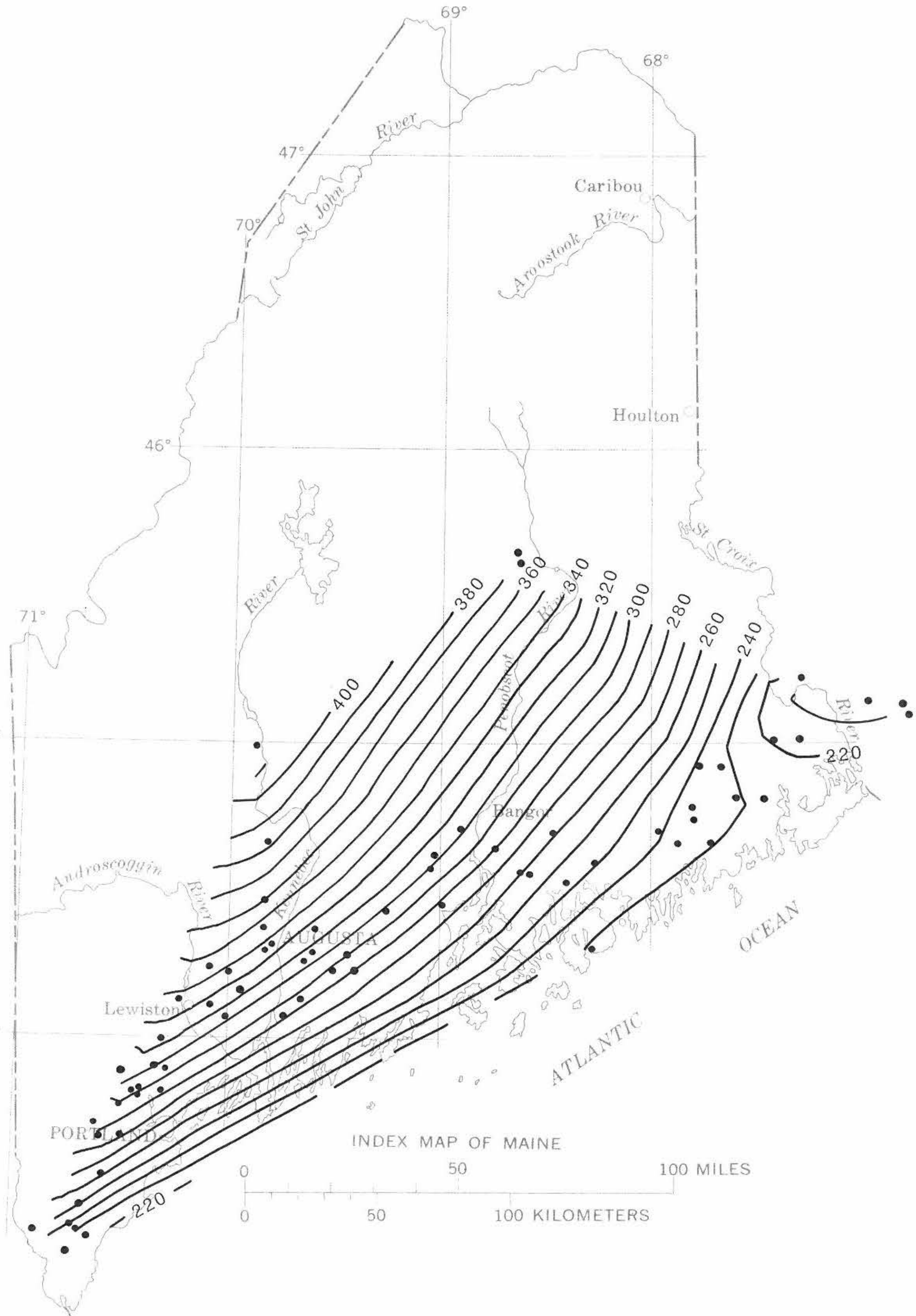

Figure 15. Third-degree trend surface fitted to delta elevations. Dots indicate locations of surveyed deltas (from Appendix B). Contour interval is 10 ft.

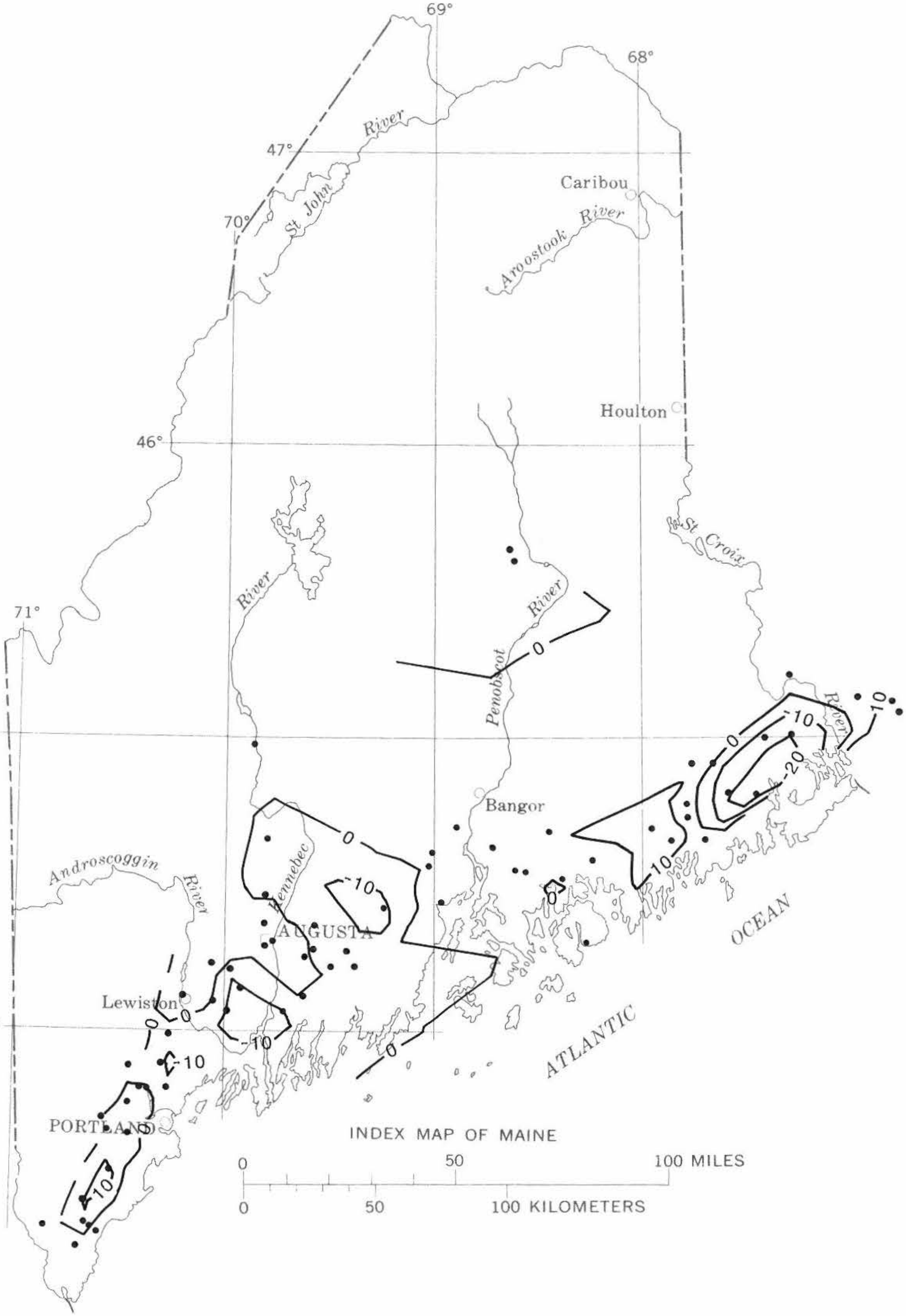

Figure 16. Contoured residuals from third-degree trend surface. Dots indicate locations of surveyed deltas. Contour interval is 10 ft.

Anomalous Delta Elevations in Eastern Maine and Adjacent New Brunswick

The configuration of the delta elevation contours in Figure 2 shows an anomalous embayment in the vicinity of the Maine-New Brunswick border. Although only a few deltas could be located and surveyed on the New Brunswick side of the border, the contour pattern shown in Figure 2 is believed to be generally valid. For example, the 220-ft contour is constrained by the deltas at Columbia Falls (235 ft), Meddybemps (203 ft), St. Stephen, New Brunswick (212 ft), and Pennfield, New Brunswick (231 ft). A similar anomaly is present in the elevation contours of the trend surface (Figure 15). The trend-surface contours in the border region are not as low as in Figure 2, but the residuals from this surface (Figure 16) show a negative deviation of at least 20 ft in eastern Maine.

The marine-limit contours in the Maine-New Brunswick border region are transverse to the end moraines shown by Thompson and Borns (1985a) and Rampton and others (1984). Ice-marginal deposits in this area include the late-glacial Pineo Ridge moraine system (Figure 1), which is a series of deltas, submarine fans, and closely spaced moraine ridges that can be traced eastward across Washington County, Campobello Island, and possibly to the Pennfield delta (Thompson and Borns, 1985b). We propose two explanations to account for the elevation pattern of glaciomarine deltas near the New Brunswick border. The embayment of the contours may have resulted from processes related to deglaciation, including the crustal tilt direction and chronology of ice-margin retreat. Alternatively, the contour pattern may define a zone of postglacial crustal subsidence, the axis of which trends northward through the Eastport-Calais area in Washington County.

Postglacial tilt toward the east-southeast is compatible with Figures 2 and 15, as well as the striation pattern in northern Maine (discussed above). This tilt direction might explain the eastward decrease in marine-limit elevations along the Pineo Ridge moraine system. The Pineo Ridge East and Columbia Falls deltas (Appendix A, Nos. 80 and 88) appear to have formed simultaneously as part of this system, and thus were probably graded to the same sea-level position. However, the upper marine limit determined from the Pineo Ridge East delta is 251 ft (76.5 m), whereas the elevation indicated by the Columbia Falls delta -- 12 km to the east -- is 235 ft (71.6 m). A further decrease in the marine limit as one follows the Pineo Ridge system eastward is suggested by raised shoreline deposits at about 200 ft (61 m) in East Machias. Perhaps these sea-level indicators originally were at the same elevation, and then were differentially uplifted as crustal tilt occurred. The east-southeastward tilt would have been a consequence of the earlier mass distribution of the late Wisconsinan ice sheet in Maine, rather than the late-glacial ice configuration when the Pineo Ridge moraine system was deposited.

The relationship of glaciomarine delta elevations in eastern Maine to those in adjacent New Brunswick is not well known, though the glacial histories of these two areas are very similar. Ice-flow indicators in southern New Brunswick trend south to southeast, and the late Wisconsinan ice margin retreated to the northwest (Rampton and others, 1984). Moraine patterns imply that the Pennfield delta and Pennfield moraine in coastal New Brunswick are the same age as, or slightly younger than, the Pineo Ridge moraine system (Rampton and others, 1984; Thompson and Borns, 1985a). With either a constant or falling relative sea level during formation of these deposits, a dominantly eastward postglacial tilt across the region would have caused the Pennfield delta to be considerably lower than the Columbia Falls delta in the Pineo Ridge system. However, the Pennfield delta is only 4 ft (1.2 m) lower than the one at Columbia Falls. This similarity suggests that the delta elevation gradient in the vicinity of the Maine-New Brunswick border is not simply the result of eastward crustal tilt.

Rampton and others (1984) claimed that relative sea level first rose and then fell as the ice margin receded from the St. George, New Brunswick, area (Figure 1). Their highest delta is the Pennfield delta, with a surface elevation of 73.5 m, followed chronologically by the Utopia (66 m) and McDougall Lake (62 m) deltas as the ice margin retreated northwestward. Gadd (1973) proposed an identical sequence for these deposits. He also described an ice-margin position of the "Bethel-Pocologan Phase", which is intermediate in age between the Utopia and McDougall Lake deltas. During the present study we obtained elevations that agree well with those of Rampton and others (1984), keeping in mind that we surveyed topset/foreset contacts, which are slightly lower than the delta tops. The sea-level positions indicated by the Pennfield, Utopia, and Bethel deltas are at elevations of 231 ft (70.4 m), 212 ft (64.6 m), and 207 ft (63.1 m) respectively (Figure 2). Thus, our results support the concept of lowering relative sea level resulting from crustal uplift as the ice margin withdrew from the Pennfield delta. Perhaps this drop in sea level also was responsible for the low elevations of the deltas north of the Pineo Ridge moraine system in Maine.

The delta elevations shown in Figure 2 are largely the product of glacio-isostasy and related sea-level change, but it is possible that they have been modified by Holocene crustal downwarping in the vicinity of the Maine-New Brunswick border. The contour pattern in Figure 2 is similar to Tyler's pattern of crustal subsidence velocity contours (this volume). The contours on his map trend north-northeast across Washington County and show rates of modern crustal subsidence that increase eastward to a maximum in the Eastport area. If the glaciomarine delta elevations in this part of Maine can be taken as a measure of net Holocene downwarping, Figure 2 indicates that Eastport has subsided 40 ft (12.2 m) or more relative to neighboring parts of the Maine-New Brunswick coast. This estimate is based on comparison of the inferred marine limit at Eastport (190 ft or less) with the 231-235 ft delta elevations at Mount Desert Island, Columbia Falls, and Pennfield, New Brunswick. However, the actual amount of any downwarping might be less, depending on the direction and gradient of glacio-isostatic tilt.

Delta Elevations in Southernmost Maine

The delta elevation contours in Figure 2 are somewhat evenly spaced over most of Maine, but are closer together in the Sanford-Berwick area in the southern extremity of the state. A similar, though less pronounced, pattern is evident from the trend-surface analysis (Figure 15). Figure 2 indicates an elevation gradient of approximately 5.7 ft/mi (1.08 m/km) in this part of Maine, as measured between the 220 and 260-ft contours. The gradient for the corresponding area in Figure 15 is only about 3.1 ft/mi (0.59 m/km). The contour spacing in Figure 2 is thought to be reliable in this part of Maine, since it is based on survey data from a series of deltas with clearly defined topset/foreset contacts.

The close spacing of the elevation contours in the Sanford-Berwick area possibly could indicate a flexure of the marine-limit surface resulting from Holocene crustal subsidence along the coast. Analysis of geodetic leveling data by Tyler (this volume) indicates that modern subsidence may be occurring in this part of Maine at a rate of up to 3 mm/yr. Alternatively, the lesser elevation gradient of the deltas north of Sanford may be due to faster isostatic uplift during deglaciation of the latter area, which would have caused more rapid lowering of relative sea level (Figure 14). Slower ice recession near the inland marine limit, accompanied by falling relative sea level, probably resulted in a further decrease in the delta elevation gradient. Considering the sparse evidence for neotectonic crustal subsidence in southwestern Maine, processes related to glaciation seem to provide the best explanation for variations in delta elevation gradients in this part of the state.

Conclusion

The anomalously low elevations of glaciomarine deltas in eastern Maine may be at least partly the result of Holocene crustal downwarping, though it is also possible that the elevation pattern is the consequence of deglaciation and crustal rebound circumstances. The delta elevation contours (Figure 2) define an apparent trough of subsidence, the axis of which trends northward along the Maine-New Brunswick border. The history of vertical crustal movement through postglacial time is unknown, but the delta elevations show that the marine-limit surface in the Eastport area may have downwarped at least 40 ft (12.2 m) relative to nearby parts of coastal Maine and New Brunswick. The possible subsidence pattern agrees with modern crustal warping trends determined from tide-gauge data and resurveys of bench marks (Anderson and others, 1984; Tyler, this volume). Other types of information, such as modern and historical seismic data (Lepage and Johnston, 1986; Smith and others, this volume) likewise point to eastern Maine as an area of recent crustal unrest. Variations in the direction and spacing of delta elevation contours occur elsewhere in southern Maine, but are likely to have resulted from processes related to ice retreat, isostatic crustal uplift, and changing sea level during late-glacial time.

SUMMARY

The authors have located and examined 101 glaciomarine deltas in the coastal lowland of Maine and adjacent New Brunswick. These deltas formed during the period of marine submergence that accompanied deglaciation of southern Maine. Field evidence indicates that the sea was in contact with the retreating margin of the late Wisconsinan ice sheet throughout the transgressive phase of submergence. Deglaciation of the study area was complete by about 13,000 yr B.P., and isostatic uplift brought the region above sea level between 12,000 and 11,000 yr B.P.

The glaciomarine deltas formed by discharge of sand and gravel into the ocean at the mouths of glacial meltwater streams. They have flat upper surfaces -- locally channeled and kettled -- which were graded to the contemporary sea level, and some of them have been modified by marine erosion and postglacial stream terracing. Virtually all of the deltas are the "Gilbert type", in which sub-horizontal gravelly topset beds overlie seaward-dipping sandy to gravelly foreset beds. The contact between the topset and foreset units in each delta provides a minimum measure of the position of sea level to which the delta was graded. Meltwater distributary channels on the delta plains also enable an approximate determination of former sea level.

The unmodified character of most delta plains and the absence of overlying marine sediments demonstrate that the deltas formed either during the maximum marine transgression or the early part of the ensuing offlap. Exceptions possibly occur in the southern extremity of Maine, where a rising eustatic sea level may have barely overtopped the deltas before being overtaken by crustal rebound. Relative sea level could not have dropped very much as the succession of deltas was deposited, since many of the deltas accumulated in water that was only 20-40 m deep. From these shallow depths it is inferred that the ice-contact deltas formed adjacent to the margin of a grounded tidewater glacier.

The glaciomarine deltas of Maine have been classified into four categories based on their depositional environment: (1) ice-contact deltas; (2) esker-fed ice-contact deltas; (3) leeside deltas; and (4) distal outwash deltas. Ninety-five percent of the investigated examples belong to one of the first three categories, having formed either in contact with the ice margin or separated from it by a narrow ridge or chain of hills. The strike-controlled bedrock topography that typifies much of southern Maine often localized deposition of the deltas. Many of them formed where the ice margin was temporarily pinned against northeast-trending strike ridges.

The authors contoured the elevations of delta topset/foreset contacts and meltwater channels in order to define the present configuration of the uplifted and tilted marine-limit surface, and to determine whether this surface has been warped by Holocene vertical crustal movements. Most elevations were precisely surveyed by leveling, and a few were measured by altimeter where nearby elevation control was lacking. In southwestern Maine the delta elevations increase from 200 ft (61.0 m) near the present coastline to 422 ft (128.6 m) at the inland marine limit (Figure 2). This northwestward increase in elevation is the result of differential glacio-isostatic uplift, which was greater in the direction of increasing former ice thickness.

The delta elevation contours over much of southwestern Maine are somewhat uniformly spaced and have a northeast trend. In the central Kennebec Valley region they indicate a minimum postglacial tilt of 2.82 ft/mi (0.53 m/km) in a direction of 145°. A third-degree trend surface was fitted to the elevation data, and the contour pattern and gradient of this surface generally correspond to those of Figure 2. Comparison with the steeper tilt gradient recorded by glacial Lake Hitchcock deltas in western New England suggests that isostatic uplift was causing relative sea level to fall as the Maine deltas were deposited. This conclusion is supported by stratigraphic evidence of falling sea level during deglaciation.

The marine-limit elevation contours curve to the north in central Maine (Figure 2), perhaps because of a different postglacial tilt direction in this part of the state. However, a sharp indentation of the contour pattern in the vicinity of the Maine-New Brunswick border cannot be readily explained by our present knowledge of the glacial history of the area. Glaciomarine deltas in eastern Washington County are at elevations of 193-224 ft (58.8-68.3 m), in contrast to the higher elevations of deltas to the southwest and northeast. Local variations in the rate of ice-margin recession, accompanied by a changing relative sea level, may have been largely responsible for the observed delta elevations. An alternative explanation is that Holocene crustal subsidence has occurred in eastern Maine. The latter interpretation is not proven by our study, but is compatible with other evidence for downwarping presented elsewhere in this volume. The delta elevations indicate that the net downwarp of the Eastport area may have been 40 ft (12.2 m) or more relative to nearby parts of the Maine-New Brunswick coast.

A minor perturbation of the marine-limit surface occurs in the Sanford-Berwick area of southern York County, where delta elevation contours are more closely spaced than elsewhere in southern Maine (Figure 1). The local steepening of the elevation gradient possibly is the result of downwarping of the coastal zone south of Sanford. Tyler's data (this volume) indicate modern crustal subsidence of up to 3 mm/yr in the same area. However, it is more likely that delta elevations in this part of Maine were controlled solely by the interaction of crustal isostatic adjustment and eustatic sea-level change during deglaciation.

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APPENDIX A. INVESTIGATED GLACIOMARINE DELTAS AND CORRESPONDING SEA-LEVEL ELEVATIONS

Type: Four types of glaciomarine deltas are recognized here (see text for description):
- (1) ice-contact
- (2) esker-fed ice-contact
- (3) leeside
- (4) distal outwash

Elevation: All measurements are surveyed elevations of delta topset/foreset contacts except as noted.
- * = measured by altimeter, with temperature/pressure corrections
- (C) = measurement of distal portion of meltwater channel on delta plain

Delta No.	Name	Location (Town and Quadrangle)	Type	Elevation ft	Elevation m
1	Berwick	Berwick; Somersworth 7.5'	1	200	61.0
2	Five Corners	N. Berwick; Somersworth 7.5'	3		
3	South Lebanon	Lebanon; Rochester 7.5'	4	236	71.9
4	Sand Pond	Sanford; Sanford-Alfred 7.5'	1		
5	Perkins Town	Wells; N. Berwick 7.5'	1	226	68.9
6	L Pond	Sanford-Wells; N. Berwick 7.5'	1	233	71.0
7	Bragdon Road	Wells; N. Berwick 7.5'	1	214	65.2
8	Branch Brook	Sanford; Alfred 7.5'	1		
9	Sanford Airport	Sanford; Alfred 7.5'	2		
10	New Dam Road	Sanford; Alfred 7.5'	1	259	78.9
11	Bernier Road	Sanford; Alfred 7.5'	1		
12	Round Pond	Alfred; Alfred 7.5'	2		
13	Dayton	Dayton-Lyman; Bar Mills 7.5'	2	272	82.9
14	South Hollis	Hollis-Dayton-Lyman; Waterboro-Bar Mills 7.5'	4(?)		
15	North Hollis	Hollis-Limington; Limington 7.5'	4		
16	Plains Road	Hollis; Standish 7.5'	1	288*	87.8
17	Saco	Standish-Limington; Limington 7.5'	4	296	90.2
18	Dingley Spring Road	Gorham; Standish 7.5'	3		
19	Chicopee	Buxton-Gorham; Standish 7.5'	1		
20	Finn Parker Road	Buxton-Gorham; Standish 7.5'	1		
21	Groveville	Buxton; Standish 7.5'	1	285	86.9
22	Sebago Lake	Standish-Gorham; Sebago Lake-N. Windham 7.5'	1	299	91.1
23	Richville	Standish; Sebago Lake 7.5'	1		
24	Raymond Neck	Raymond; Naples 7.5'	1	312	95.1
25	Canal Road	Standish-Gorham; N. Windham 7.5'	1	307	93.6
26	Windham Hill	Windham; N. Windham 7.5'	1	300*	91.4
27	North Windham	Windham; N. Windham 7.5'	2	300	91.4
28	Forest Lake	Cumberland-Windham-Falmouth; Cumberland Center 7.5'	3	283	86.3
29	Libby Hill	Gray; Gray 7.5'	3(?)	302	92.0
30	Gray Village	Gray; Gray 7.5'	1		
31	Crystal Lake-Sabbathday Pond	New Gloucester-Gray; Gray 7.5'	2,3	312*	95.1
32	East Gray	Gray; Gray 7.5'	3	289	88.1
33	Kettlebottom Road	Bowdoin-Sabattus; Lisbon Falls North 7.5'	3	298	90.8
34	Pleasant Hill	Sabattus; Lisbon Falls North 7.5'	2	323	98.5
35	Sawyer Hill	Monmouth; Monmouth 7.5'	2	324	98.8
36	Cemetery Road	Monmouth; Monmouth 7.5'	1		
37	Island Pond	Leeds; Monmouth 7.5'	2	348	106.1
38	Leeds Junction	Greene; Monmouth 7.5'	2		
39	Lake Auburn	Auburn; Lake Auburn East-Lewiston 7.5'	1	342	104.2
40	Twitchell Airport	Turner-Auburn; Lake Auburn East	2		
41	Springer Hill	Litchfield; Purgatory 7.5'	3	308	93.9
42	Call Hill Road	Dresden; Wiscasset 7.5'	1	278	84.7
43	Palmer Hill	Whitefield-Alna; E. Pittston-Wiscasset 7.5'	2	291	88.7
44	Globe	Washington; Jefferson 7.5'	3	279	85.0
45	Mountain Road	Jefferson; Jefferson 7.5'	3	295	89.9
46	Muddy Pond	Washington; Razorville 7.5'	3	292	89.0
47	South Windsor	Windsor; Weeks Mills 7.5'	3(?)	309	94.2
48	Hunts Meadow Road	Whitefield; Togus Pond 7.5'	3(?)	315	96.0
49	Erskine Academy	China; Weeks Mills-China Lake 7.5'	2	321	97.8
50	Meadow Brook	China; China Lake 7.5'	3		
51	Granite Hill	Hallowell; Augusta 7.5'	3	338	103.0

APPENDIX A. CONTINUED.

Delta No.	Name	Location (Town and Quadrangle)	Type	Elevation ft	Elevation m
52	Augusta Airport	Augusta; Augusta 7.5'	2	344	104.9
53	Summerhaven	Manchester-Augusta-Belgrade; Belgrade-Augusta 7.5'	2	348	106.1
54	Belgrade	Belgrade; Belgrade 7.5'	2	359	109.4
55	Smithfield	Smithfield; Norridgewock 7.5'	2		
56	Sand Hill Road	Smithfield; Norridgewock 7.5'	2	375	114.3
57	Concord	Concord; Anson 15'	2	422	128.6
58	Montville	Montville-Searsmont; Morrill 7.5'	3	290	88.4
59	Swan Lake	Swanville; Brooks East 7.5'	1		
60	Irish Hill	Monroe; Brooks East 7.5'	3	315	96.0
61	Twombly Mountain	Monroe; Brooks East-E. Dixmont 7.5'	3	313	95.4
62	Searsport	Belfast-Searsport; Searsport 7.5'	2	293	89.3
63	Hampden	Hampden; Snow Mtn. 7.5'	2	317	96.6
64	Dolby Dam	TA R7 WELS; E. Millinocket 7.5'	1	362	110.3
65	Dolby Pond	TA R7 WELS-Grindstone; Millinocket 7.5'	1(?)	372	113.4
66	East Millinocket	E. Millinocket-Medway; E. Millinocket 7.5'	2		
67	Orcutt Mountain	Bucksport; 'Brewer Lake 7.5'	3	291	88.7
68	Upper Patten Pond	Ellsworth; Branch Lake 7.5'	3	275	83.8
69	West Ellsworth	Ellsworth; Branch Lake 7.5'	3	266	81.1
70	Beech Hill Pond	Otis-Mariaville; Beech Hill Pond 7.5'	1	279	85.0
71	Simmons Pond	Ellsworth-Hancock; Ellsworth 7.5'	2		
72	McFarland Hill	Hancock; Hancock-Ellsworth 7.5'	2	250	76.2
73	Silsby Plain	Aurora; Great Pond 7.5'	3		
74	Jordan Pond	Mount Desert; Seal Harbor 7.5'	1	231	70.4
75	Franklin	Franklin; Sullivan 7.5'	1	253(C)	77.1
76	Denbo Heath	Deblois; Tunk Mtn. 7.5'	2		
77	Pork Brook	T22 MD; Lead Mtn. 7.5'	3		
78	Poplar Hill	Deblois; Schoodic Lake 7.5'	1		
79	Pineo Ridge West	Deblois-T18 MD; Schoodic Lake 7.5'	2	253	77.1
80	Pineo Ridge East	Columbia; Epping 7.5'	1	251(C)	76.5
81	Montegail Pond	T19 MD; Epping 7.5', Tug Mtn.-Wesley 15'	1,2	244*(C)	74.4
82	Crebo Flat West	Deblois-T18 MD-T24 MD; Tug Mtn. 15'	3,2(?)		
83	Crebo Flat East	T18 MD; Tug Mtn. 15'	3		
84	Ben Tucker Mountain	T18 MD; Tug Mtn. 15'	2		
85	Black Brook Ponds	T19 MD; Tug Mtn.-Wesley 15'	2	244*(C)	74.4
86	Rocky Lake	T25 MD; Tug Mtn.-Wesley 15'	1		
87	Sam Hill Barrens	T25 MD; Wesley-Tug Mtn. 15'	3		
88	Columbia Falls	Columbia Falls-Centerville; Columbia Falls 7.5'	1	235	71.6
89	Andy Mountain	Northfield; Wesley 15'	1	209	63.7
90	Bog Lake	Northfield; Wesley 15'	1		
91	Little Seavey Lake	Wesley; Wesley 15'	1(?)	224	68.3
92	Air Line Road	T31 MD; Wesley 15'	1	248	75.6
93	Gardner Lake	East Machias; Hadley Lake 7.5'	2	196	59.7
94	Meddybemps	Meddybemps; Meddybemps Lake East-West 7.5'	2	203	61.9
95	Round Lake	Charlotte; Meddybemps Lake East 7.5'	1	193	58.8
96	Maxwell Crossing	St. Stephen, N.B.; St. Stephen 1:50,000 (21 G/3)	4(?)	212	64.6
97	Bethel	St. Patrick, N.B.; St. George 1:50,000 (21 G/2)	3	207	63.1
98	Utopia	Pennfield, N.B.; St. George 1:50,000 (21 G/2)	1	212*	64.6
99	Pennfield	Pennfield, N.B.; St. George 1:50,000 (21 G/2)	1	231	70.4
100	Baring	Baring-Calais; Meddybemps Lake East 7.5'	1		
101	Norridgewock	Norridgewock; Norridgewock 7.5	2	404*	123.1

APPENDIX B. COORDINATES OF SITES WHERE DELTA ELEVATIONS WERE SURVEYED

Delta			Longitude			Latitude		
No.	UTM EW	UTM NS	Deg	Min	Sec	Deg	Min	Sec
1	356595.2	4791821.0	70.	45.	59.77	43.	16	2.95
3	343608.7	4800616.0	70.	55.	44.07	43.	20.	38.54
5	361704.7	4799667.0	70.	42.	20.48	43.	20.	20.58
6	360183.9	4800674.0	70.	43.	28.87	43.	20.	52.22
7	364266.5	4797868.0	70.	40.	25.23	43.	19.	24.00
10	361604.7	4809665.0	70.	42.	33.99	43.	25.	44.51
13	371089.4	4821169.0	70.	35.	42.33	43.	32.	3.38
16	369856.2	4835983.0	70.	36.	50.06	43.	40.	2.60
17	367801.3	4840461.0	70.	38.	25.65	43.	42.	26.36
21	377944.0	4835479.0	70.	30.	48.83	43.	39.	51.23
22	377731.6	4846118.0	70.	31.	7.02	43.	45.	35.82
24	378650.1	4860539.0	70.	30.	37.74	43.	53.	23.67
25	382955.3	4851467.0	70.	27.	17.79	43.	48.	32.20
26	385418.3	4851022.0	70.	25.	27.26	43.	48.	19.18
27	385460.4	4852638.0	70.	25.	26.66	43.	49.	11.53
28	394230.6	4851623.0	70.	18.	53.57	43.	48.	43.35
29	391740.9	4861268.0	70.	20.	52.00	43.	53.	54.59
31	394646.4	4872124.0	70.	18.	49.56	43.	59.	47.92
32	395645.1	4860413.0	70.	17.	56.50	43.	53.	28.94
33	418981.4	4880096.0	70.	0.	41.86	44.	4.	17.41
34	413018.7	4884614.0	70.	5.	12.51	44.	6.	41.33
35	419856.1	4896983.0	70.	0.	11.81	44.	13.	24.97
37	412878.7	4899260.0	70.	5.	27.56	44.	14.	35.89
39	400867.9	4886827.0	70.	14.	20.34	44.	7.	47.47
41	424614.6	4889957.0	69.	56.	33.79	44.	9.	39.14
42	441352.5	4880300.0	69.	43.	56.55	44.	4.	31.69
43	448211.9	4886330.0	69.	38.	50.36	44.	7.	48.95
44	469192.8	4897026.0	69.	23.	8.56	44.	13.	39.91
45	460460.2	4897012.0	69.	29.	42.15	44.	13.	37.93
46	466401.8	4902967.0	69.	25.	15.73	44.	16.	52.01
47	452781.5	4903440.0	69.	35.	30.30	44.	17.	4.64
48	449860.2	4900708.0	69.	37.	41.10	44.	15.	35.38
49	453724.3	4912960.0	69.	34.	50.81	44.	22.	13.36
51	434140.1	4905087.0	69.	49.	31.91	44.	17.	52.76
52	436778.6	4907013.0	69.	47.	33.70	44.	18.	56.04
53	433652.7	4913629.0	69.	49.	57.84	44.	22.	29.45
54	433976.3	4924176.0	69.	49.	48.01	44.	28.	11.35
56	435218.4	4946267.0	69.	49.	1.81	44.	40.	7.65
57	430750.4	4982449.0	69.	52.	42.41	44.	59.	38.58
58	481050.3	4919666.0	69.	14.	17.10	44.	25.	55.17
60	498358.9	4935491.0	69.	1.	14.38	44.	34.	28.95
61	499676.9	4940365.0	69.	0.	14.61	44.	37.	6.87
62	503203.4	4921811.0	68.	57.	35.04	44.	27.	5.54
63	509404.2	4950594.0	68.	52.	52.58	44.	42.	38.17
64	530094.7	5052949.0	68.	36.	50.06	45.	37.	53.02
65	528788.2	5056842.0	68.	37.	49.56	45.	39.	59.36
67	523290.4	4943346.0	68.	42.	22.68	44.	38.	42.13
68	532756.1	4934305.0	68.	35.	15.09	44.	33.	47.80
69	536247.8	4933471.0	68.	32.	37.05	44.	33.	20.17
70	545090.2	4949172.0	68.	25.	51.32	44.	41.	47.19
72	550098.2	4930980.0	68.	22.	10.17	44.	31.	56.46
74	559955.2	4906652.0	68.	14.	53.79	44.	18.	45.33
75	562071.7	4938514.0	68.	13.	4.53	44.	35.	57.28
79	584770.1	4950950.0	67.	55.	47.86	44.	42.	31.90
80	592432.8	4946676.0	67.	50.	2.62	44.	40.	10.02
81	598252.7	4955520.0	67.	45.	32.30	44.	44.	53.74
85	597712.9	4960193.0	67.	45.	53.61	44.	47.	25.46
88	605336.1	4947159.0	67.	40.	16.61	44.	40.	19.24
89	614599.0	4964701.0	67.	33.	2.04	44.	49.	42.46
91	608074.1	4975962.0	67.	37.	50.32	44.	55.	50.94
92	599410.7	4975985.0	67.	44.	25.28	44.	55.	56.27
93	625107.1	4964411.0	67.	25.	4.19	44.	49.	26.69
94	627612.9	4986310.0	67.	22.	50.32	45.	1.	14.49
95	637358.2	4987626.0	67.	15.	24.22	45.	1.	50.58
96	636112.6	5011030.0	67.	15.	58.17	45.	14.	29.51
97	662968.6	5003039.0	66.	55.	37.54	45.	9.	50.19
98	676321.1	5001880.0	66.	45.	28.45	45.	9.	1.08
99	678437.5	4997972.0	66.	43.	56.77	45.	6.	52.59
101	434876.6	4952381.0	69.	49.	20.16	44.	43.	25.68