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Depositional Sequence Modeling of Late Quaternary Geologic History, West-Central Maine Coast

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ABSTRACT

High-resolution seismic reflection profiles collected in Sheepscot Bay, Maine from 1983-1986 contain evidence of ten distinct seismic facies and subfacies in two seismic depositional sequences. The data base consists of 166 km of boomer data and 413 km of 3.5 kHz tuned-transducer data. These seismic data were supplemented with side-scan sonar, grab samples, and submersible observations. Late Wisconsin glaciation >20,000 to 14,000 years before present (yr B.P.) and deglaciation from 14,000 to 12,500 yr B.P. were responsible for deposition of the first major sequence (G). Included facies are: till, stratified drift, and glaciomarine mud. Sea-level fluctuations, due to glacio-isostasy and glacio-eustasy, caused a marine transgression of deep water in contact with the ice front, followed by highstand (to +70 m locally) and subsequent regression. Regression continued to a lowstand at 50-65 m depth, at which time littoral processes created shorelines along much of coastal Maine. In the study area, the Kennebec River created a lowstand paleodelta. Subsequent transgression shifted paleodelta and estuarine environments landward, recycling sediments. The second depositional sequence (H) contains sediments deposited in several facies during lowstand and Holocene transgression. These are shelf basin and estuarine mud, sometimes containing natural gas, and paleodelta sand and gravel.

The depositional sequence model clarifies the late Quaternary geologic history of the Sheepscot Bay region and the west-central Maine coast in terms of: (1) changing local relative sea level and shoreline positions, (2) location and nature of sedimentary environments in the transition from deglacial to marine and littoral environments, and (3) location and nature of unconformities. This methodology should be useful in similar studies elsewhere.

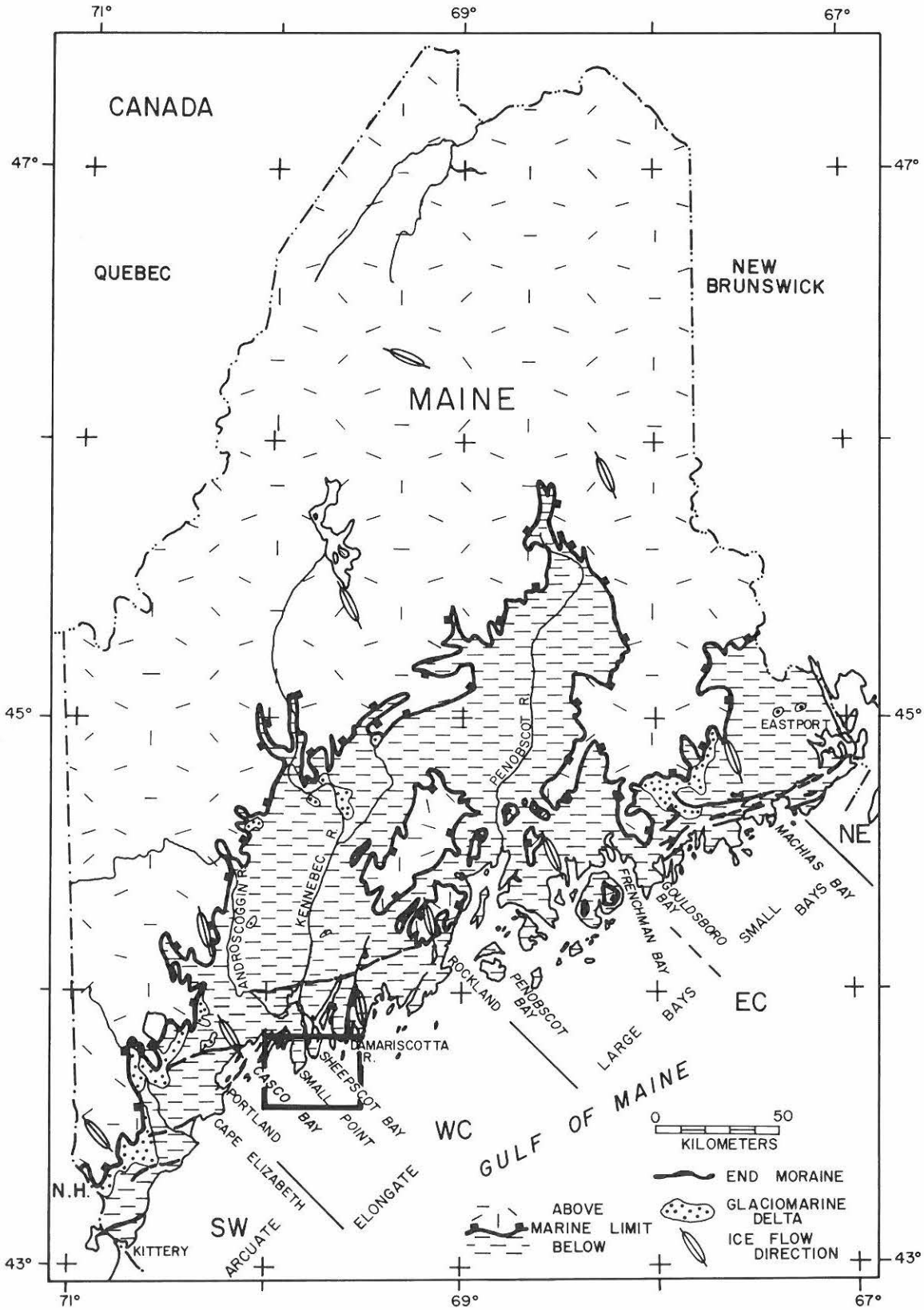

Figure 1. Location map of Maine. The study area is indicated by the bold rectangle. Also shown are the postglacial marine limit, the area of deposition of the Presumpscot Formation (horizontal dashes), major end moraines (bold lines) and glaciomarine deltas (dots) (modified from Thompson and Borns, 1985). The four-fold geomorphic division of the coast is indicated by shore-normal boundary lines (SW, WC, EC, NE).

INTRODUCTION

In this paper we use extensive high-resolution seismic reflection profiling to define late Quaternary stratigraphy in Sheepscot Bay, the inner shelf around Small Point, and eastern Casco Bay, Maine (Fig. 1). These data are coupled with side-scan sonographs, cores, bottom samples, and submersible observations to construct a preliminary evolutionary model of the embayment.

The goals of the present study are to reconstruct deglacial events in west-central coastal Maine in terms of processes and timing, and relationship to ice-margin and sea-level positions. The primary mechanism for this reconstruction is the interpretation of high resolution seismic data through sequence stratigraphy (e.g., Mitchum et al., 1977a,b). Thus, rather than using formation names referred to a type section (lithostratigraphy: i.e., the "Kennebec Formation", "Boothbay Formation", etc.), or attempt a chronostratigraphic designation in poorly-dated but undoubtedly time-transgressive units, we have used the allostratigraphic approach of defining units based on their bounding discontinuities (NACSN, 1983). Therefore, the concentration is on depositional sequences and their boundaries. Finally, the models developed in this study area can be applied elsewhere on the Maine coast and in the Gulf of Maine.

Sheepscot Bay, Small Point, and eastern Casco Bay (informally designated "New Meadows Bay" herein) are located in the west-central (WC) coastal geomorphic compartment of Maine (Fig. 1). It includes the seaward extension of the Sheepscot River and New Meadows River estuaries, strike-parallel embayments within the fjard coastline (Belknap et al., 1986). They are elongated, low relief (<200 m) valleys whose alignment is parallel to the strike of adjacent bedrock peninsulas. Between the two embayments is the paleodelta of the Kennebec River, surrounding Small Point. As with the remainder of the Maine coast, the processes which produced the Quaternary sediments of the study area are the combined actions of glacial erosion and deposition, deglacial and glaciomarine sedimentation, and littoral and marine processes. Isostatic loading and unloading and eustatic sea-level rise (Schnitker, 1974; Stuiver and Borns, 1975; Belknap et al., 1987a) have produced major changes in relative sea level over the past 14,000 years (Fig. 2). These sea-level fluctuations caused major shifts in the shoreline and changes in the location of and processes in sedimentary environments.

During the Late Wisconsin glacial epoch, the Laurentide Ice Sheet flowed in a south-southeasterly direction in this region (Thompson and Borns, 1985), nearly parallel to the strike of the underlying high-grade, metasedimentary, Precambrian to Paleozoic rocks (Osberg et al., 1985). Ice erosion deepened stream valleys in less resistant rocks to produce the present coastal topography. Glacial ice extended into the Gulf of Maine and to Georges Bank at 18,000-20,000 yr B.P. (Pratt and Schlee, 1969; Denton and Hughes, 1981; Figure 2 in Hughes et al., 1985). Based on microfossil and grain-size data in shelf-basin cores (Schnitker 1975, 1986), it is likely that the margin was a floating ice shelf by 16,000 yr B.P. and changed to a calving embayment with a tidewater margin about 14,000 yr B.P. The calving margin reached the present coast between 13,800 and 13,200 yr B.P. (Smith, 1985; Thompson and Borns, 1985; Stuiver and Borns, 1975; Borns, 1973). From then until about 13,000-12,500 yr B.P., ice retreated in contact with deep water, allowing transgression up to the inland marine limit. As isostatic rebound overtook eustatic sea-level rise, the floating ice margin could no longer retreat inland. The grounded ice front produced glaciomarine deltas at the inland marine limit (Thompson, 1982; Thompson and Borns, 1985). The marine limit is found today at elevations from 60 to 132 m above sea level, on a surface subsequently tilted by differential isostatic uplift (Anderson et al., 1984; Belknap et al., 1987a). The marine limit was roughly +70 m in the Sheepscot Bay region. The sea then retreated, resulting in a regressive sequence of shoaling within the glaciomarine mud, capped by

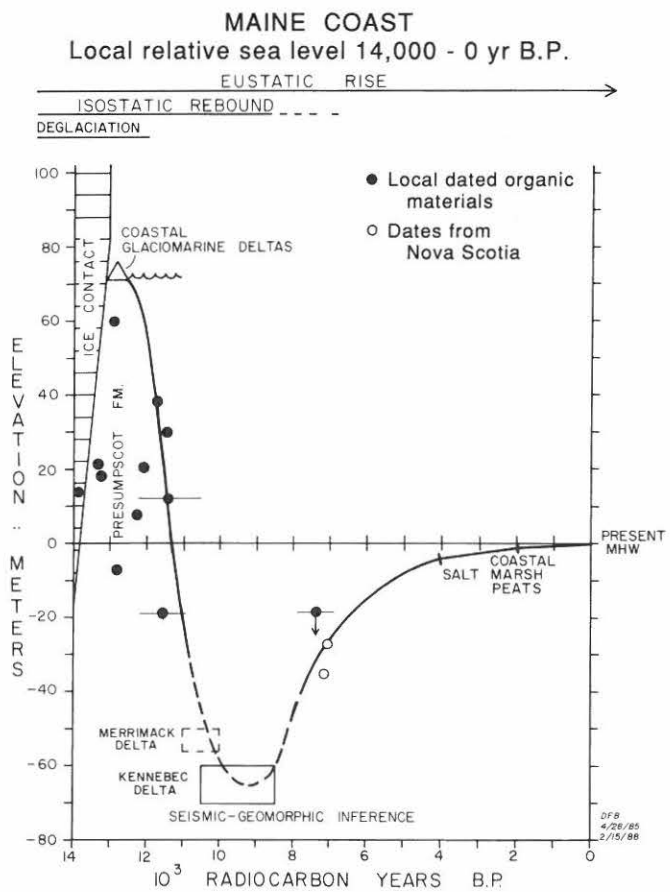

Figure 2. Late Quaternary local relative sea-level curve for the Maine coast, based on ^{14}C-dated shells and salt marsh peats, and seismic interpretation of submerged shorelines and paleodeltas (Belknap et al., 1986; 1987a). Filled circles represent local, dated organic materials, and the open circles represent dates from Nova Scotia. See Belknap et al. (1987a) for details.

littoral sand and gravel. The glaciomarine mud produced within this transgressive to regressive sequence is the Presumpscot Formation, first named by Bloom (1960, 1963). Sea level fell to as low as -65 m (Schnitker, 1974; Belknap et al., 1986, 1987a), producing a series of distinct lowstand shorelines (Shipp et al., this volume). As isostatic rebound decreased (roughly exponentially), eustatic rise became dominant and the ongoing marine transgression began.

Previous marine geologic work in the Sheepscot Bay area has been limited to the study by Schnitker (1974) and the present projects (Belknap et al., 1986, 1987b; Kelley et al., 1987). Schnitker (1974) reported on 3.5 kHz seismic data, three-meter long piston cores, and sediment grab samples. The piston cores in the deeper basins confirm the presence of gas-charged Holocene marine muds. Useful oceanographic data were generated by Graham (1970) using bottom drifters, and by McAlice and Jaeger (1983) using current meter and salinity measurements in the Sheepscot estuary. Fefer and Schettig (1980) summarize marine geologic and ecologic data from the coast of Maine. Seismic reflection profiling studies in other embayments nearby include: Penobscot Bay (Knebel, 1986; Knebel and Scanlon, 1985; Ostericher, 1965); Casco and Saco Bays (Kelley et al., 1986, 1987); Machias Bay and Damariscotta River estuary (Belknap et al., 1987b); Gouldsboro Bay (Shipp et al., 1985, 1987); and the Sheepscot River estuary (Belknap et al., 1986). Oldale et al. (1983) investigated the drowned paleodelta of the Merrimack River offshore of northeastern Massachusetts, and Oldale and Bick (1987) and Oldale and Wommack (1987) have mapped Massachusetts Bay and the northeastern Massachusetts nearshore zone. Birch (1984a,b) conducted a detailed seismic survey of the New Hampshire inner shelf. These later studies have built on preliminary surveys such as Oldale et al. (1973), and Folger et al. (1975).

Quaternary sediments onshore were mapped by Smith (1976). Kellogg (1982) examined coastal sedimentology in regard to archaeological sites in the Boothbay region. Finally, L. K. Fink and students have studied the Popham Beach system for more than a decade (Nelson and Fink, 1978; FitzGerald and Fink, 1987).

METHODS

The primary data base discussed in this report consists of high-resolution seismic reflection profiles. Two systems were used: a Raytheon RTT1000A 3.5 kHz profiler with a 200 kHz fathometer printing simultaneously, and an ORE Geopulse "boomer" system. The Geopulse was run at 105 and 175 joules, filtered between 500-700 and 1500-2000 kHz. The present data set was collected primarily on 10 m research vessels between 1983 and 1986, at less than 10 km/hr, using Loran C as the primary navigational aid. In all, the 1983-1986 data are from 413 km of 3.5 kHz and 166 km of Geopulse lines (Fig. 3).

Bottom sediment samples were collected with a Smith-MacIntyre grab sampler (Kelley et al., 1987). Sediment samples are described in terms of estimated gravel, sand, and mud percentages. Presently, all samples are in the process of detailed analysis. Digitally-rectified side-scan sonar images from an EG&G SMS 960 Sea floor Mapping System were used to site the sediment samples and to correlate with the reflection seismic data (Kelley et al., 1987). In many cases, use of the side-scan sonar and vertical seismic reflection profiles together with sediment grabs resulted in a clearer interpretation of the data than use of a single method would allow. For example, large ripples easily visible on the side-scan data were seen to form an interference pattern in the 3.5 kHz data. Similarly, Kelley et al. (1987, Fig. 18), as well as Birch (1984b) and Knebel (1986) have demonstrated that darker side-scan return patches on smooth bottom correlate with outcrops of the Presumpscot Formation noted in seismic profiles. As another example, grab samples and submersible observations confirm the gravelly sand texture of rippled surfaces (Belknap et al., 1988).

Dives were conducted in 1984 aboard the submersible *Mermaid II*, in conjunction with ROV (Remotely Operated Vehicle) observations; in 1985 aboard the submersible *Johnson Sea Link*; and in 1986 and 1987 aboard the submersible *Delta*. Nineteen kilometers of sea floor were traversed with continuous geologic observation, periodic 35 mm photography, and near-continuous videotape coverage, along lines of existing seismic and side-scan data. These observations confirmed details of remotely-sensed data, such as shape, sediment texture, and dimensions of large oscillation ripples, textural transitions from gravelly sand to mud along the paleodelta slope, and domination of physical versus biogenic reworking (Belknap, 1985; Belknap et al., 1988).

We have used vibracores, bridge borings, and outcrops on nearby land to confirm seismic interpretations (Belknap et al., 1986; Kelley et al., 1986), but as yet no vibracores are available outside the estuaries in the study area. Despite the lack of extensive on-site core data, correlation with cored intervals nearby, bottom observations, and grab samples provide preliminary verification of lithology of seismic stratigraphic units.

RESULTS AND DISCUSSION

Seismic stratigraphic units were identified based on morphology of bounding surfaces, intensity of acoustic contrasts, and intensity, frequency, and geometry of internal reflections. Belknap et al. (1986, Fig. 4; 1987b) discuss this approach in more detail. It is a methodology similar to that described by Mitchum et al. (1977a,b). The units described below are *seismic facies units*, which are contained within one or more *seismic depositional sequences*. These facies and sequences are summarized in Table 1.

Seismic Facies Units

Unit BR is always the lowest unit observed (Figs. 4-8). It has a sharp, high intensity return on a highly irregular surface. No true internal reflections are seen, but hyperbolic surface

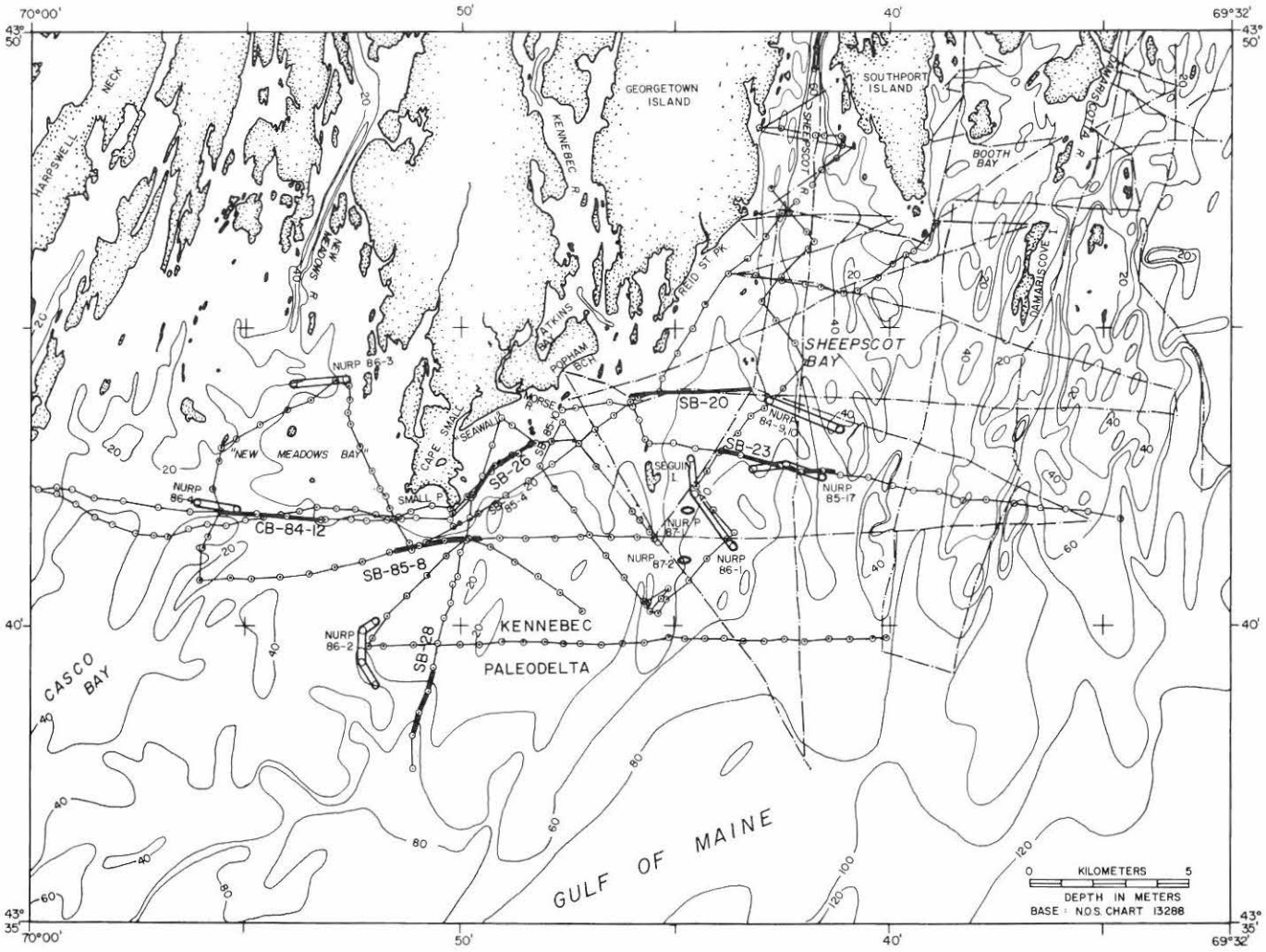

Figure 3. Track lines of seismic reflection profiles and submersible dives for 1983-1986, based on Loran C readings usually at 5 minute intervals. ORE Geopulse lines are circle-dash, Raytheon 3.5 kHz lines are dot-dash, and submersible transects are bold elongated polygons indicated as NURP-85-17, etc.

returns and side echoes are common. Unit BR is interpreted as Paleozoic (or older) bedrock. Numerous outcrops observed on shore, in side-scan data, and on submersible traverses confirm this interpretation. No Mesozoic-Cenozoic rocks (e.g., Tagg and Uchupi, 1966; Uchupi, 1966; Oldale et al., 1973; Birch, 1984a,b; Fader et al., 1977) were noted in the west-central Maine inner shelf.

Overlying bedrock in limited areas is unit T (Figs. 5, 6) which occurs in two subunits - Tm and Tb. It has a sharp upper boundary with a strong reflection, and chaotic internal reflections. Subunit Tm has a mounded geometry (Figure 5) and ranges from two to more than seven meters in thickness. Subunit Tb is a thin (0.5 to >2 m) layer, draping over and ponding in irregularities in bedrock. It is often difficult to distinguish, but occurs as a broadening of the usually sharp bedrock return, with indistinct internal reflections rarely visible within the acoustic pulse. This seismic facies unit is the equivalent of Birch's (1984b) unit 1. Unit T is interpreted as till. Tm is in the form of minor moraines, such as those found on land nearby by Smith (1976), representing de-grounding of marine-based ice. Elsewhere there are larger recessional moraines. For example, in Machias Bay a submerged unit Tm is traceable to the immediately adjacent Pond Ridge moraine onshore (Shipp and Belknap, 1986; Belknap et al., 1987b). Kelley et al. (1986; 1987, Fig. 19) have confirmed the interpretation of till with side-scan images of bouldery surface lags over this unit in Casco Bay. Unit Tb is interpreted as thin drift. Till in nearshore Maine differs from the "lift-off moraines" as seen by King and Fader (1986) in being denser, more reflective, with chaotic internal bedding, and lacking the steep conical form of the latter.

Schnitker (1974) interpreted basal units over much of the study area as till, both dissected by stream erosion and with an undissected morphology. Our present interpretation based on more detailed, higher resolution seismic data is that till is not

TABLE 1. CHARACTERISTICS OF SEISMIC DEPOSITIONAL SEQUENCES AND FACIES, MAINE INNER SHELF.

	Depositional Sequence	Seismic Facies Unit	Characteristics	Interpretation
	7	NG	Strong return, convex upward shape, fuzzy upper surface	Natural gas acoustic wipeout
H	6	SG	Strong, ringing return well stratified, channels and foresets	Holocene sand and gravel
	5	M	Weak to transparent, flat, ponded in basins	Holocene marine basin mud
	4c	GM-P	Strong to moderate surface return, weak to transparent internal reflections, ponded	Glaciomarine mud distal from grounding line, minor effects of ice rafting
	4b	GM-D	Strong, rhythmic bedding, draped	Glaciomarine: interbedded mud, silt and sand
G	4a	GM-M	Weak to transparent, faintly bedded, draped	Glaciomarine sediment or diamicton: grounding line deposit
	3	SD	Strong return, wedge shape, stratified, interfingers with 2 & 4	Stratified drift
	2b	Tm	Strong return, chaotic internal reflections, mound shape	Till, moraine ridge
	2a	Tb	Strong return, indistinct interior reflections	Till, thin drift
Pz	1	BR	Very strong, sharp return, no internal reflections, steep slopes and peaks	Paleozoic bedrock

widely distributed in the area. Schnitker's (1974) "undissected till" is interpreted here as natural gas acoustic wipeout zones (Belknap et al., 1986, 1987b; Schubel, 1974). The "dissected till" is here interpreted as stratified paleodelta deposits (Belknap, 1985). Schnitker (1974) suggested a lowstand at -65 m. While we interpret the seismic records differently, the -65 m depth can be supported by the change from sand and gravel to mud at the seaward edge of the Kennebec paleodelta.

Unit SD is laterally adjacent to and interfingering with unit GM and sometimes till, and overlies bedrock and till. It is stratified and occurs in wedge-shaped bodies that thin seaward. This facies is not prevalent in the study area, but can be found interfingering with GM in Figures 4 and 5 (strike sections), and overlying till in Figure 6. Unit SD is interpreted as ice-contact stratified drift deposited as proximal subaqueous outwash (Rust and Romanelli, 1975) associated with ice grounding lines, analogous to strata found in stratified moraines of Washington County, Maine (Thompson, 1982; Borns, 1986).

Unit GM-M is the basal unit in a thick sequence of highly variable facies overlying bedrock and till (Figs. 4, 8). It is weakly reflective and generally massive, but with occasional stratification which demonstrates its draped nature. It is found only in the thickest sequences of GM and is always overlain by unit GM-D. We interpret this seismic facies as representing massive glaciomarine mud or diamicton, with few dropstones or sorted coarser material. It may represent rapid but uniform initial sub-ice shelf deposition in deep local basins. Knebel (1986) did not differentiate this unit from the rest of the glaciomarine (Qp), but it is clearly evident in his Figure 5. It also has similarities to King and Fader's (1986) "till tongues."

Unit GM-D is highly stratified, with a sharp but smooth upper bounding surface which drapes underlying bedrock or sediment (Figs. 4, 5, 8). It is often rhythmically bedded and concentrically drapes underlying topography. Almost every Geopulse profile in this area demonstrates examples of this seismic facies within the deeper portions of sedimentary basins. It is similar in character to King and Fader's (1986) Emerald Silt Facies A and to parts of Birch's (1984b) unit 2. As shown by Piper et al. (1983), this draping character indicates rapid sedimentation from suspension. The draping nature and rhythmic fluctuations in grain size are typical of deposits from overflow-interflow processes of proglacial environments (e.g., Smith and Ashley, 1985). Unit GM-D is interpreted as draped glaciomarine mud, the equivalent of the lower portions of the Presumpscot Formation found on land. We interpret the environment of deposition of this facies as glaciomarine, at least in part under an ice shelf (Schnitker, 1975, 1986) near the ice grounding line. The rhythmic layering is interpreted as alternating fine sand to silt and silty clay beds, based on correlation with exposures of the Presumpscot Formation at Bunganuc Bluff, in northwest Casco Bay (Kelley and Hay, 1986a, Fig. 44). Variation in water content might also play a part in the variable reflectivity (see Tucholke and Hollister, 1973).

Unit GM-P is an acoustically transparent to faintly bedded seismic facies. It is horizontally bedded in ponded, sometimes discontinuous lenses and, like the draped glaciomarine mud, is very common (Figs. 4, 5, 8). It may correspond to the basal sections of Birch's (1984b) unit 3, and King and Fader's (1986) Emerald Silt Facies B. Its contact with the underlying draped glaciomarine mud is conformable, either parallel or as divergent fill in lows. More than 30 vibracores from Saco Bay, Casco Bay (Kelley et al., 1986), Sheepscot River (Belknap et al., 1986), and Damariscotta River (Belknap and Shipp, 1986) have penetrated into the ponded glaciomarine mud. It is clearly identified as the Presumpscot Formation, on the basis of texture (slightly sandy silty clay to clayey silt with occasional dropstone gravel), stiffness, and blue-gray color. Its environment of deposition is interpreted as distal glaciomarine: a cold, turbid-water calving embayment (Thompson, 1982; Stuiver and Borns, 1975; Bloom, 1960, 1963). The ponded horizontal bedding probably results from density underflows, perhaps low-density turbidity currents, distant from the ice grounding line, produced by slumping and resedimentation of glacial and glaciomarine sediments. Occasional ice-rafted coarse debris produces small hyperbolic returns (a possible example is seen at 45 m depth at time 1235.5

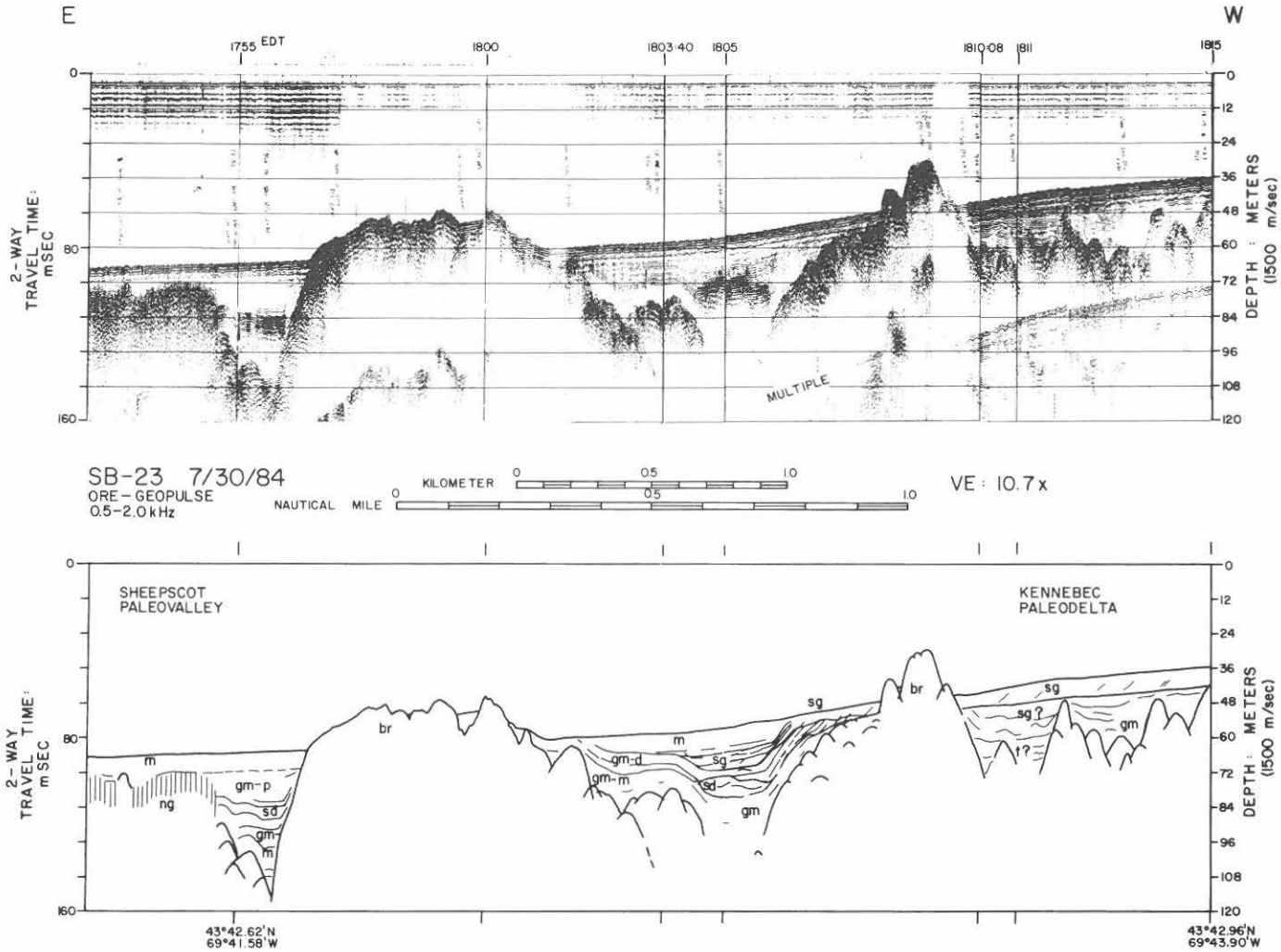

Figure 4. Seismic track line SB-23, geopulse data. Raw data is presented as collected in the upper panel, with navigation time fixes along the horizontal axis, and 2-way travel time along the vertical axis. The lower panel is a line-drawing interpretation. See Table 1 for interpretations of units. The unit gm indicates undifferentiated glaciomarine. The eastern edge of the Kennebec paleodelta occurs at about time 1806 on the record. It is characterized by deltaic foresets and (not obvious) topsets, indicating a sea level at least 50 m below present. To the east (left) lie buried channels of the Sheepscot paleovalley, filled with a basal stratified sand and gravel deposit (tidal channel), and estuarine to marine mud with natural gas. For location see Figure 3.

in Fig. 6). The upper surface of the ponded glaciomarine mud varies in nature with depth below sea level. Above -60 to -70 m, the surface shows a strong reflection, with channels and gullies (Fig. 4). In cores and borings down to at least 30 m depth (Belknap et al., 1986; Kelley et al., 1986) it is oxidized and stiff. This surface reflects subaerial exposure during emergence and creation of a distinct unconformity. Below 70 m depth the upper surface of the glaciomarine mud is less distinct (Fig. 5), becoming paraconformable, and below about 90 m it is conformable with overlying units. This latter relationship is interpreted as continuous marine submergence.

Other deglacial units are known to occur in coastal Maine. Smith (1985, Fig. 2) and Kelley et al. (1987, Fig. 46) portray the generalized stratigraphy of the coast. Ice-contact stratified drift overlies and interfingers with till in a variety of locations. Glaciomarine deltas composed of sand and gravel are found at the inland marine limit, up to 132 m above present sea level (Thompson, 1982; Anderson et al., 1984; Thompson and Borns, 1985). The seaward strata within these landforms interfinger with glaciomarine mud of the Presumpscot Formation and sometimes abut till, as at Pineo Ridge in Washington County, Maine (Miller, 1986). Since these deltas appear to occur uniquely near the marine limit, they do not occur within the seismic data of our study area.

Unit M is acoustically nearly transparent to 3.5 kHz and ORE Geopulse signals (Figs. 4, 5, 7). It caps the basin floors, lapping up onto bedrock outcrop, in an angular or moat-like contact. Submersible observations, grab samples, and piston cores

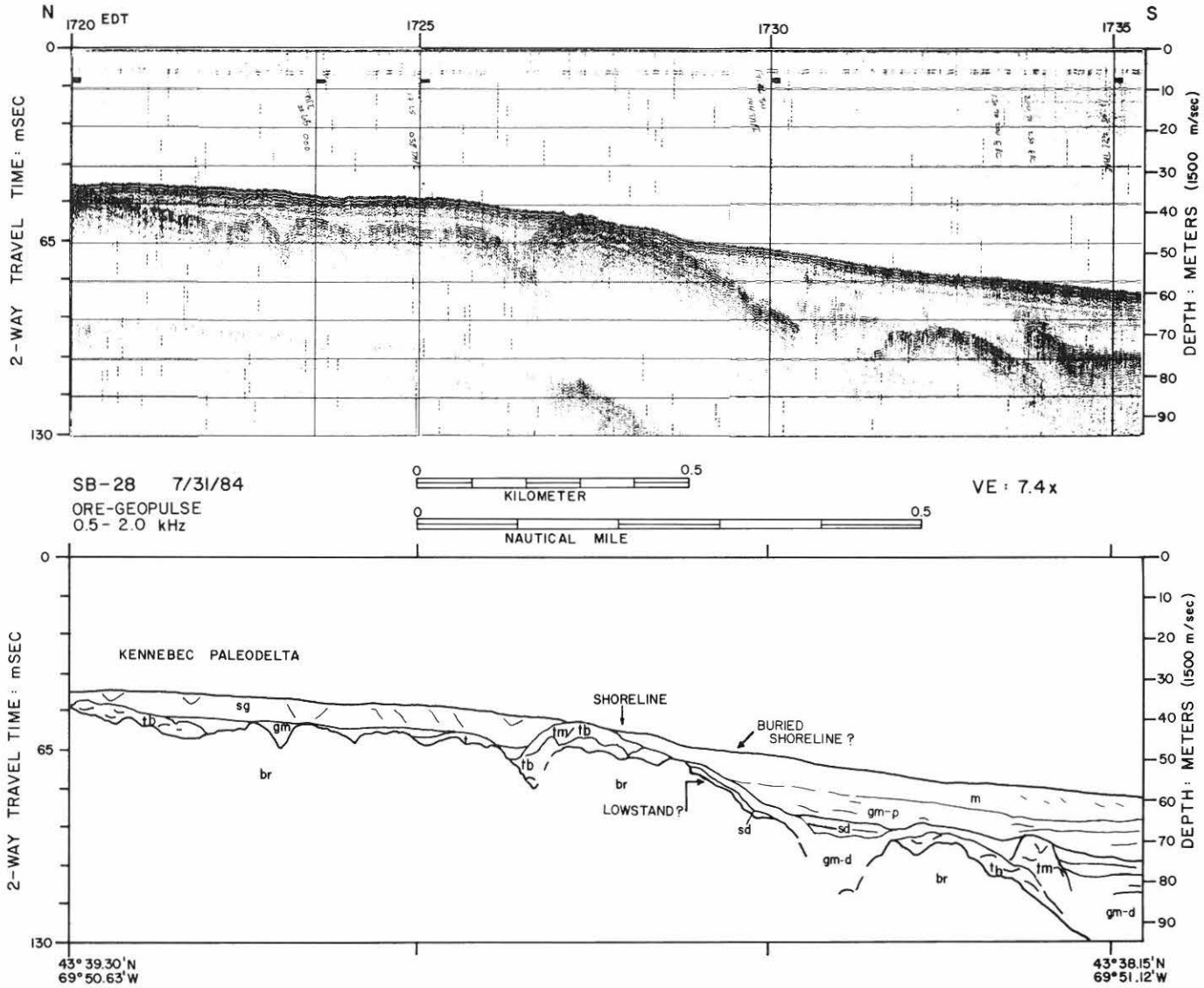

Figure 5. Geopulse profile SB-28, at the southern margin of the Kennebec paleodelta. Till (T) overlies bedrock (BR). Landward of the feature at time 1728 are stratified sand and gravel of the paleodelta, seaward are thick shelf basin muds and glaciomarine units. The feature is interpreted as a lowstand shoreline at a depth of 50 to 65 (?)m, based on geometry, differential preservation of sediments, occurrence at the edge of the delta, and similarity to other features in the depth range elsewhere on the Maine shelf (see Shipp et al., this volume). Location shown in Figure 3.

(Schnitker, 1974) confirm this unit to be Holocene inner shelf basin and estuarine mud. It is not yet possible to distinguish between open marine and estuarine conditions in the mud in the seismic data. Unit M is the equivalent of Knebel's (1986) Qh, the upper part of Birch's (1984b) unit 3, and King and Fader's (1986) LaHave Clay.

Unit SG is a strongly reflecting, well-stratified facies. The 3.5 kHz signal is rapidly attenuated within the upper 5-10 meters of sediment. Commonly, this facies spreads out in convex-upward lobes and fills channels in the underlying glaciomarine sediments (Figs. 4, 5, 7, 8). Bottom grab samples and submersible observations confirm that the surface of this facies is gravelly sand (Kelley et al., 1987). It is often worked into large bedforms such as large oscillation ripples and large sand waves (Belknap et al., 1988). A thin mud drape is an apparently ephemeral surface cover. We interpret steeply dipping clinoforms in unit SG as deltaic foresets (Figs. 4, 7). The expected topset units are occasionally visible, but difficult to distinguish from seismic ringing.

Channels filled with sand and gravel commonly cut 10-20 m into earlier-deposited units (Figs. 6, 8). The sand and gravel supports much steeper slopes than the mud, including distinct

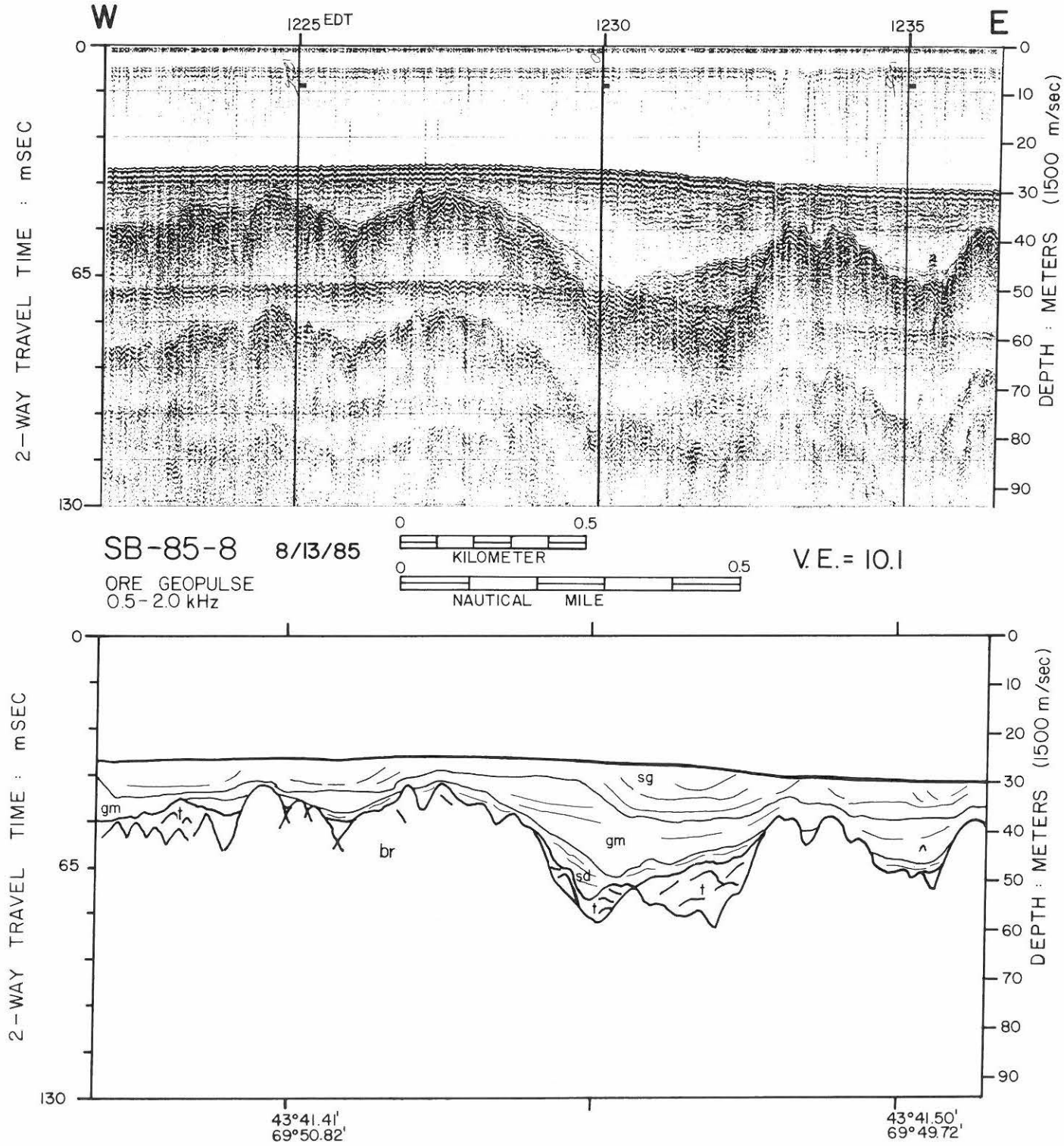

Figure 6. Geopulse profile SB-85-8, in the central paleodelta. Bedrock is covered by till (T) and stratified drift (SD). This is overlain by glaciomarine mud (GM) and paleodelta sands (SG). Location shown in Figure 3.

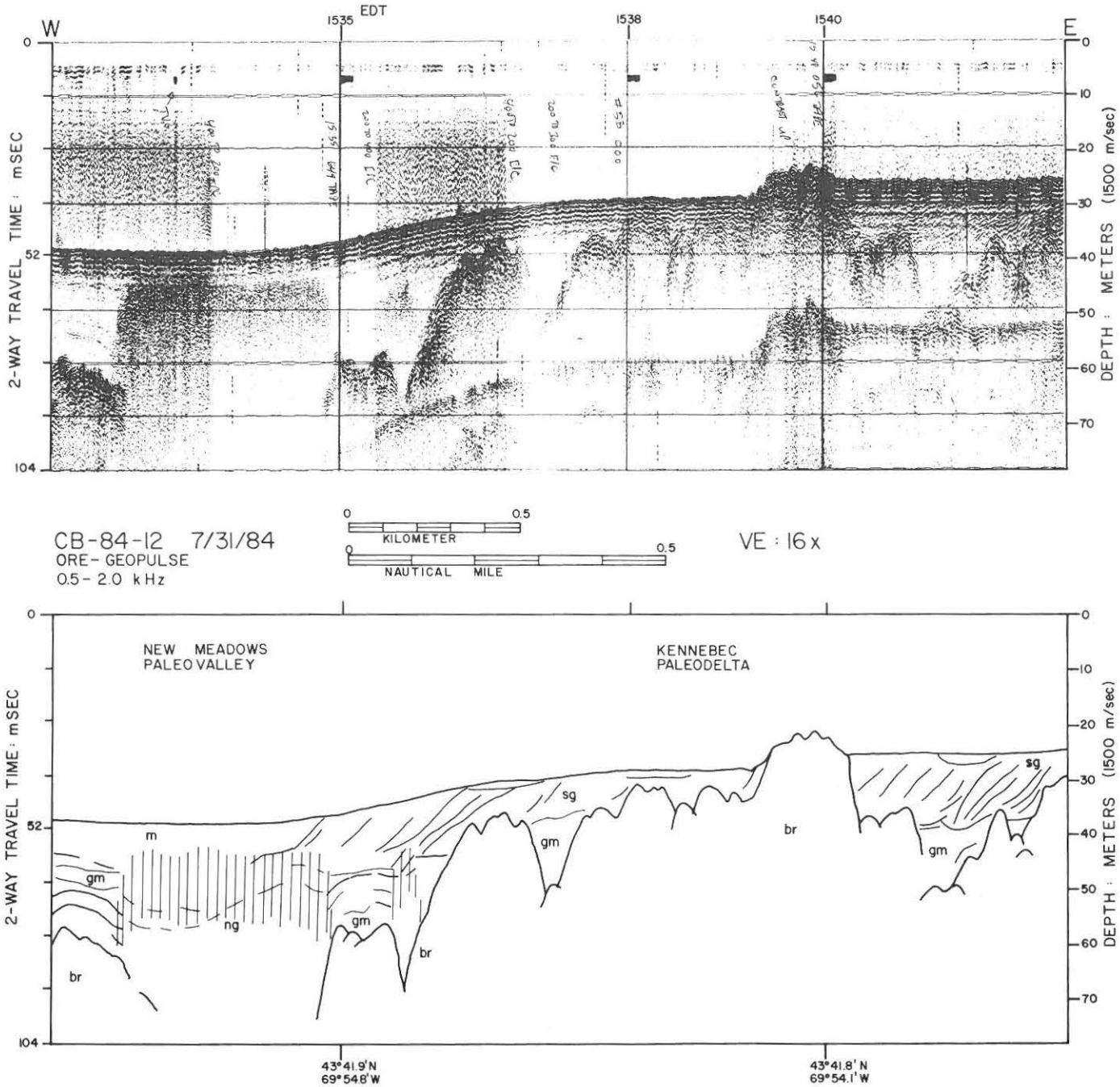

Figure 7. Geopulse profile CB-84-12, at the western margin of the paleodelta. Prominent foresets appear in SG, in later delta lobes produced at sea levels of -35 to -25 m. Location shown in Figure 3.

steps and terraces. Near the present shoreline, sand and gravel shoreface deposits overlie the erosional ravinement surface formed during the Holocene transgression, grading into the littoral units of the present beaches. Sand and gravel may also be present at the toe of foresets, in channels cut into the glaciomarine deposits. We interpret this facies as coarse deposits in tidal paleo-estuarine channels created at the sea-level lowstand. Seismic line SB-23 (Fig. 4) best displays this aspect of SG, but it is likely that it occurs in many of the channel axes, masked by natural gas.

The final seismic signature discerned is unit NG. It is a convex-upward, dark reflection which absorbs the seismic signal almost completely (Fig. 4, 7). It is termed an acoustic wipeout zone. Based on similar signals described elsewhere (e.g., Vilks

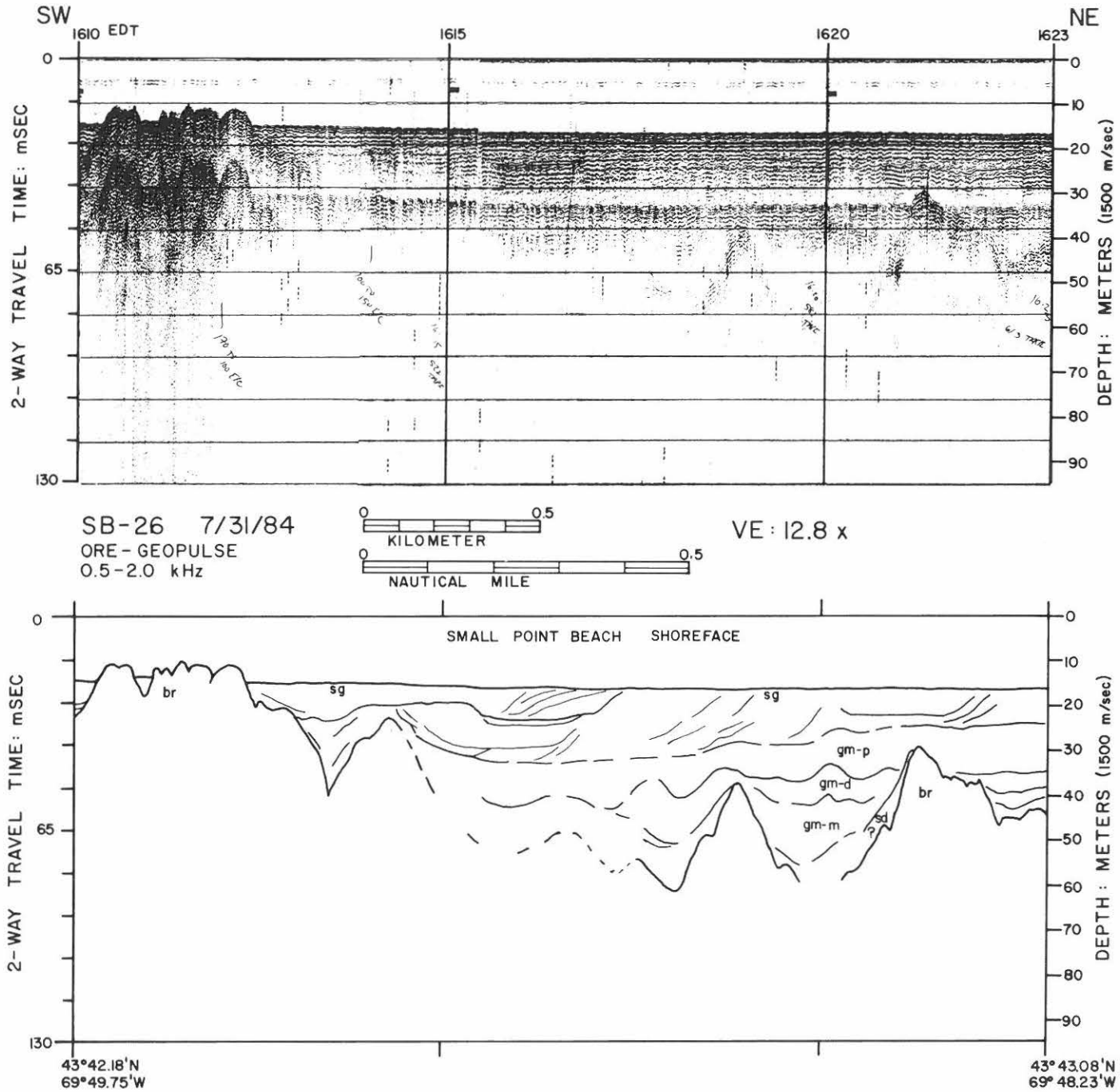

Figure 8. Geopulse profile SB-26, off Small Point (Seawall) Beach. Prominent channels within SG indicate former course of the Kennebec River. Location shown in Figure 3.

et al., 1974; Schubel, 1974), we interpret the cause of wipeout as reflection from natural gas bubbles. Presumably, this gas is largely methane. Cores taken in these zones show gas bubble voids (Belknap et al., 1986) and tend to blow off their caps when stored at sea-level pressure, evolving a flammable gas. Natural gas occurs in shelf basins and over deep valleys cut into glaciomarine sediments when Holocene mud reaches greater than 15 m thickness. In this setting it may evince: (1) buried marsh (Fader et al., 1977), (2) sediment deposited in the organic-rich turbidity maximum zone of paleo-estuarine systems which migrated landward during sea-level rise (Belknap et al., 1986), or (3) an equilibrium of methane production and sediment permeability in a critical thickness of organic-rich mud. Natural gas is also found in lesser concentrations in the glaciomarine sedi-

ments. In any case, the deep natural gas zones are most common offshore of fluvial systems, which suggests terrestrial sources of organic material.

Distribution of Seismic Facies and Sedimentary Environments

Seismic profiles, shown above and in Belknap et al. (1987b), delineate distinct facies changes within the study area. Figure 9 is a simplified map of the seismic facies distribution. Sand and gravel dominates the mouth of the present Kennebec River and surrounding Cape Small. The fan-shaped area, composed of sand and gravel, is bedded in foresets, topsets, and distinct channels. It overlies glaciomarine mud, till, and possibly subaqueous outwash or ice-contact stratified drift. We interpret this association as a lowstand paleodelta of the Kennebec River (Belknap, 1985; Belknap et al., 1986), analogous to the Merrimack Paleodelta offshore of New Hampshire (Oldale et al., 1983). The source of sediment for this lowstand paleodelta is reworked glacial and glaciofluvial sediments within the Kennebec and Androscoggin River drainages. Borns and Hagar (1965) demonstrate that glacial outwash of the Embden Formation, which fills much of the present northern Kennebec River valley, is deeply dissected. The material was reworked locally into the North Anson Formation. It is likely, however, that much of this eroded material was carried down the Kennebec River to the Gulf of Maine. Transport would have been enhanced at sea-level lowstand due to the increased topographic gradient caused by lowering of base level.

Shipp et al. (this volume) show that a prominent series of shorelines occurs along the Maine coast at 50-65 m depth. They were produced at the lowest postglacial sea level, sometime between 10,000 and 8,000 yr B.P. These predominantly erosional shorelines are most common in areas of abundant coarse sediment, such as off Wells and Saco Bay. In most of the

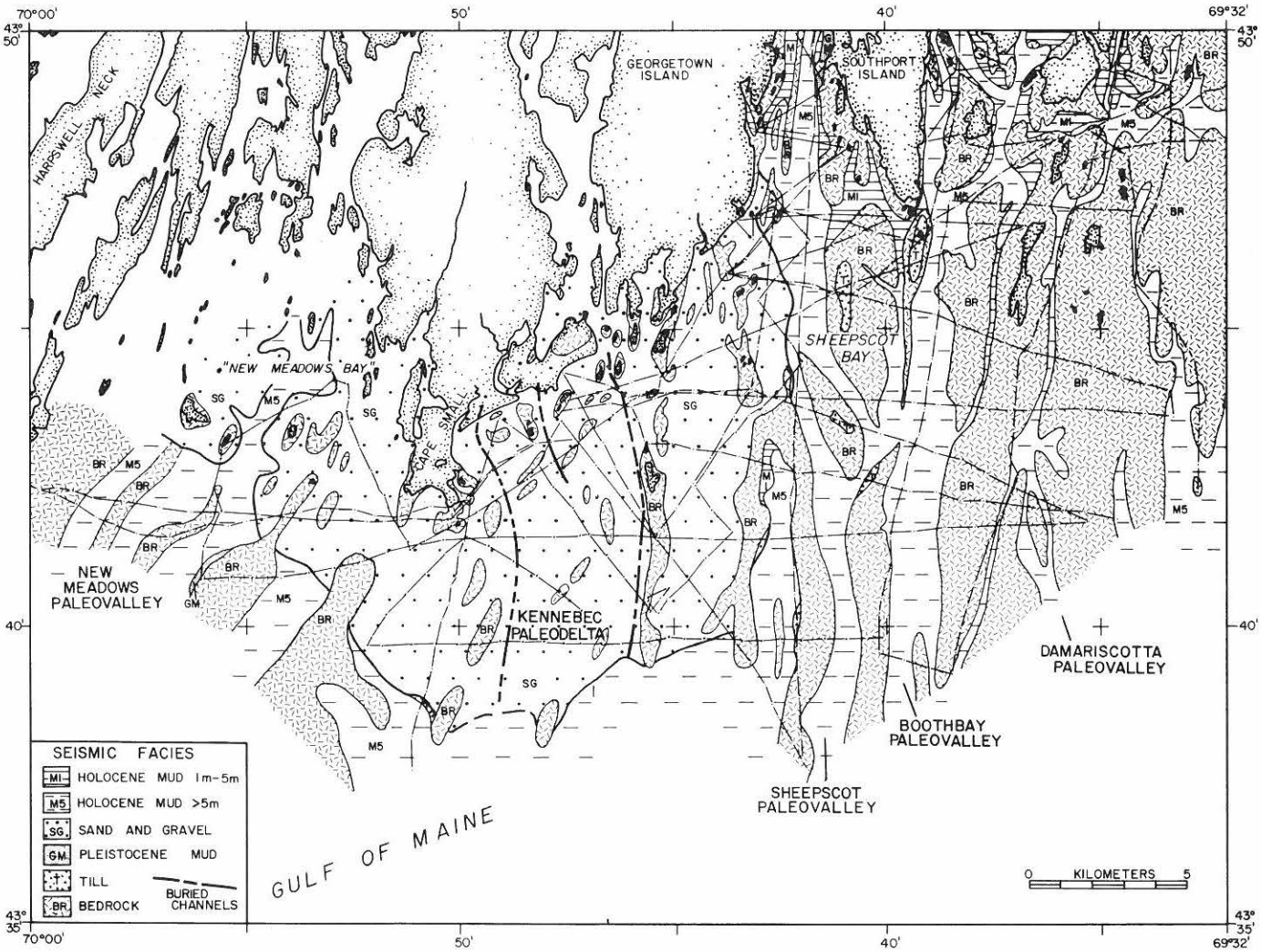

Figure 9. Surficial seismic facies map, compiled from 3.5 kHz and Geopulse data, augmented by side-scan sonar data and sediment samples.

west-central Maine coast such coarse sediment and shorelines are both lacking. At the Kennebec paleodelta, lowstand shorelines are suggested in Figure 5, but along most of the paleodelta margin, prograding lobes at shallower depths (down to -50 m, as indicated by topset-foreset contacts and steep prodelta slopes) predominate (Fig. 7). Evidence for ultimate depth of sea-level fall within the study area is most clearly provided by the depth at which the regressive unconformity becomes a conformity (70-90 m), allowing for a 15-20 m wave base scour zone.

In the central and eastern sections of Sheepscot Bay, sand and gravel are absent; the section is dominated by Holocene mud in basins, separated by bedrock ridges and infrequent till exposures. The basins are underlain by thick sections of glaciomarine mud, both ponded and draped facies. The axes of the basins are dominated by natural gas acoustic wipeouts, obscuring the depth to bedrock and Quaternary sediments. Stratified units in the channel axes suggest early Holocene fluvial or tide-dominated estuarine sediments. Deep mud-filled basins can be traced as coherent paleo-drainage systems into the modern estuaries. Therefore, these are drowned estuaries occupying the shelf valley systems. We identify shelf valleys of the New Meadows and Sheepscot Rivers, and a shelf valley associated with Booth Bay. Belknap et al. (1986, 1987b) discuss the evolution of these estuarine systems with sea-level rise. The sedimentary environments were displaced landward, recycling sediments and shifting depocenters.

The paleo-Kennebec River valley has no offshore extension, rather it terminates in the paleodelta. The Kennebec River may have experienced channel switching during offlap, lowstand, and early onlap phases, leaving several prominent buried channels in the paleodelta (Fig. 9). Upson and Spencer (1964), Tolman et al. (1986), and Kelley and Hay (1986b) have traced lower sea level courses of deranged streams on the Maine coast. The Androscoggin River may have exited through the Brunswick sand plain into Maquoit or Middle Bay in northeastern Casco Bay (Fig. 1) (Kelley and Hay, 1986b). The New Meadows River is another possible channel of the Androscoggin. The Androscoggin abandoned these channels, switching to its present course in Merrymeeting Bay (at the junction of the Androscoggin and Kennebec Rivers, Fig. 1). We speculate that this occurred during Holocene regression and lowstand.

The Kennebec River also switched channels, at least near its present mouth. Profiles SB-26, SB-85-4, and SB-85-10 (Figs. 3, 8, 9), indicate that a major channel exited near the present Morse River Inlet, when sea level was 20-25 m below present (6,000-7,000 yr B.P., Fig. 2). This former channel is filled with laterally prograding sigmoidal clinoforms (Fig. 8). Vibracores within the paleochannels in the Morse River and Atkins Bay penetrate predominantly sand (Belknap et al., 1989).

Also within the paleodelta of the Kennebec River are steps and terraces, interpreted as separate delta lobes produced near lowstand and during rising sea level. Figure 10 shows a prominent step and terrace, with an erosional surface at 18 m below sea level. This erosional zone is within the sand and gravel, well above the level interpreted as glaciomarine (e.g., Fig. 8). The modern beach systems of Popham, Small Point (Seawall), and Reid State Park are composed of sediments reworked from this paleodelta. An erosional unconformity representing the shoreface ravinement surface is clearly visible in Geopulse lines run offshore of these beaches, in a water depth of 10 to 20 m. This eroding shoreface became the predominant subtidal environment after sediment transport to the paleodelta decreased, which was a result of rising sea level, decreasing drainage gradient, and lessened supply of sand upstream. The modern surface of the paleodelta is composed of gravelly, muddy sand, with abundant shell fragments (Belknap et al., 1988). This surface is reworked into rippled scour depressions (Cacchione et al., 1984) and sand waves during winter storms, and receives an organic-rich mud drape during the summer (Belknap et al., 1988).

Sedimentation in the deep basins is by pelagic settling and probably by periodic low-density turbidity currents. Bedrock ridges and the paleodelta are swept relatively clean of fine sediment by tides and waves. In addition, slumping occurs along channel margins, at least within the estuaries (Belknap et al., 1986; Kelley et al., 1986, 1987).

Seismic Depositional Sequences

There are two major depositional sequences as defined by seismic data found within the west-central Maine nearshore, which are probably representative of much of the shallow northern Gulf of Maine. Figure 11 is a schematic representation of these seismic depositional sequences and the seismic facies units contained within them.

The lower boundary of depositional sequence G (glacial-glaciomarine) (Fig. 11) is defined by the basal unconformity on top of Paleozoic bedrock. Its upper surface is an erosional unconformity above 60-70 m below sea level and gradually becomes a paraconformity to conformity below that level. The included seismic facies have been interpreted (above) as till, both in moraines and thin drift, stratified drift, draping proximal glaciomarine mud, and ponded distal glaciomarine mud. The base of the depositional sequence demonstrates onlap (and local concordance) by marine facies, as an ice shelf and later a deep calving marine embayment replaced glacial ice. This onlap terminates at the inland marine limit (Thompson and Borns, 1985). The upper boundary demonstrates erosional truncation, down to at least 65 m below present sea level. This sequence boundary is a result of rapidly falling relative sea level (Fig. 2), which reached a local lowstand as deep as -65 m. Wave base was a further 15-20 m below this level. Below wave base, uninterrupted submergence resulted in more-or-less continuous basin accumulation.

The base of depositional sequence H (for Holocene) is defined by the top of sequence G and its upper surface by modern sea floor. The use of the term Holocene is somewhat arbitrary,

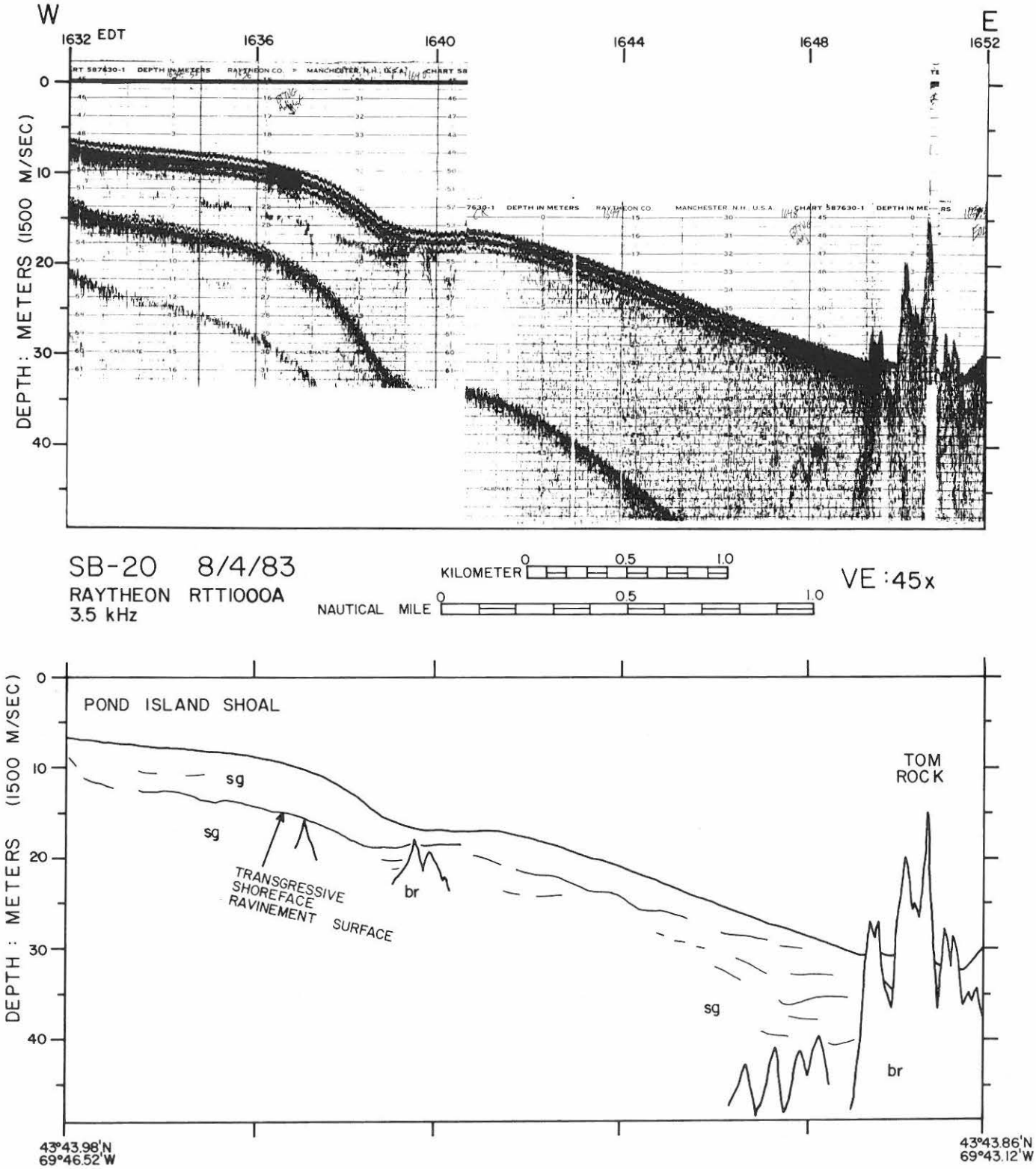

Figure 10. Raytheon 3.5 kHz profile SB-20, off the present mouth of the Kennebec River. A prominent terrace occurs at 18 m, another begins at Tom Rock at 30 m depth. The 18 m terrace is contiguous with the modern shoreface, and is underlain by a strong reflector interpreted as the shoreface ravinement surface.

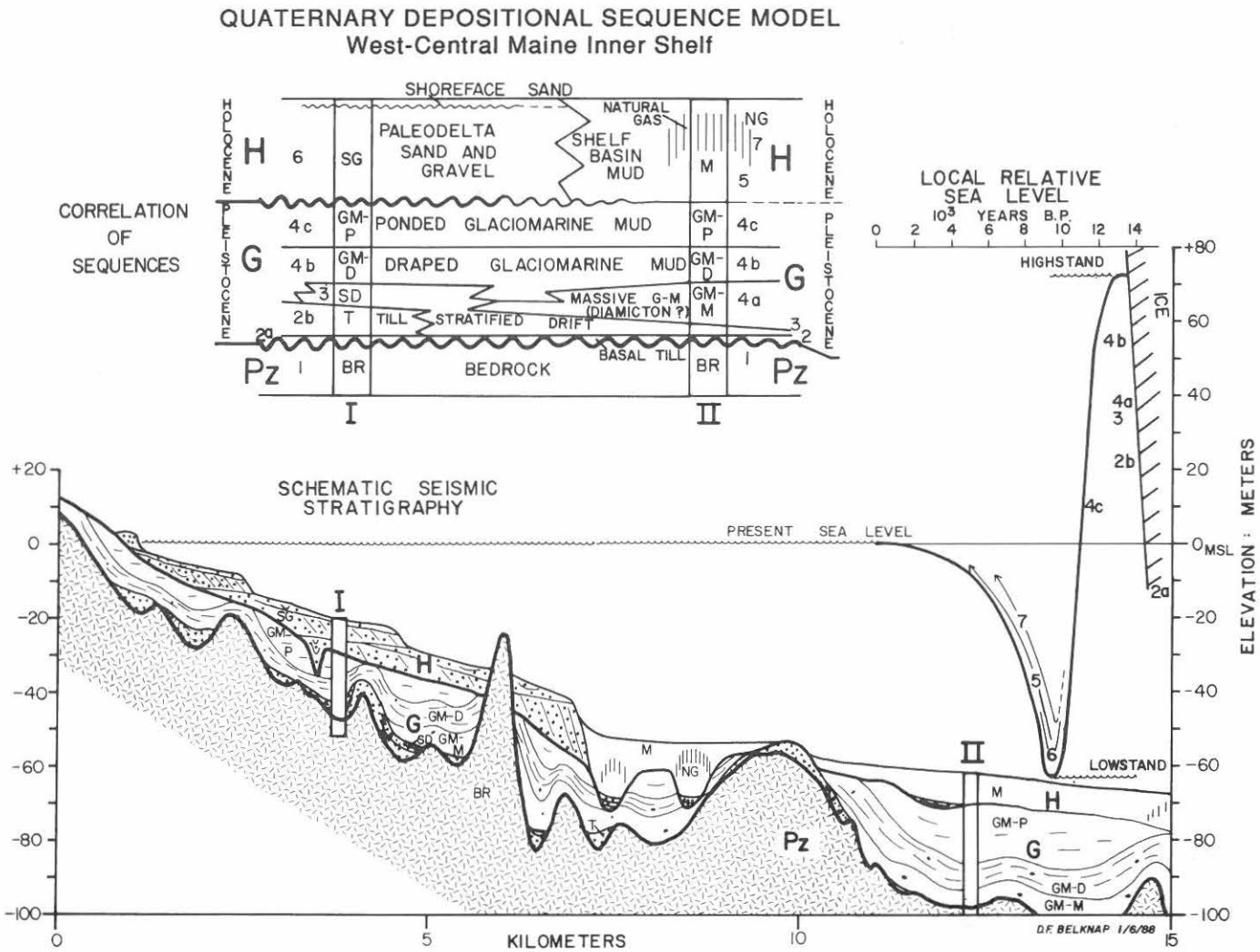

Figure 11. Schematic diagram of seismic depositional sequences, based on the eastern margin of the paleodelta. Sequence G rests on bedrock and its upper surface is an unconformity developed during regression. The unconformity grades to conformity in deep basins. Sequence H starts at the top of G, onlapping during transgression to the present shoreline. Local offlap in H is due to the prograding Kennebec paleodelta.

as the interface of G and H is not well dated, and is time transgressive. The seismic facies included in sequence H are shelf basin mud, paleodelta sand and gravel, and littoral sediments (shoreface, beach, and estuarine). Natural gas deposits occur within the shelf basin or estuarine muds. Baselap (Mitchum et al., 1977a,b) of depositional sequence H is complex. Under the Kennebec paleodelta the interface demonstrates downlap, as the delta prograded. This probably occurred during late stages of falling sea level until the lowstand. The overall deltaic sequence, with a toplap upper surface, demonstrates offlap. In the deeper basins, the relationship is concordant. During sea-level rise the shoreline transgressed the paleodelta, with temporary shoreline stillstands at delta lobes, until littoral erosion became dominant.

CONCLUSIONS

We have defined ten distinct seismic facies units and subunits, based on character and geometry of reflection. They are (Table 1): (1) BR - bedrock, (2) T - till, in two forms: Tb - basal till and Tm - deglacial (recessional) moraines, (3) SD - ice-contact stratified drift, (4a) GM-M - sub-ice-shelf massive glaciomarine mud or diamicton, (4b) GM-D - proximal draped glaciomarine mud, (4c) GM-P - distal ponded glaciomarine mud, (5) M - Holocene mud, (6) SG - sand and gravel, and (7) NG - natural gas. These facies occur within two major depositional sequences (Fig. 11). Sequence G includes the glacial and deglacial Pleistocene units, representing the marine onlap and offlap phases. Sequence H approximates the Holocene transgression.

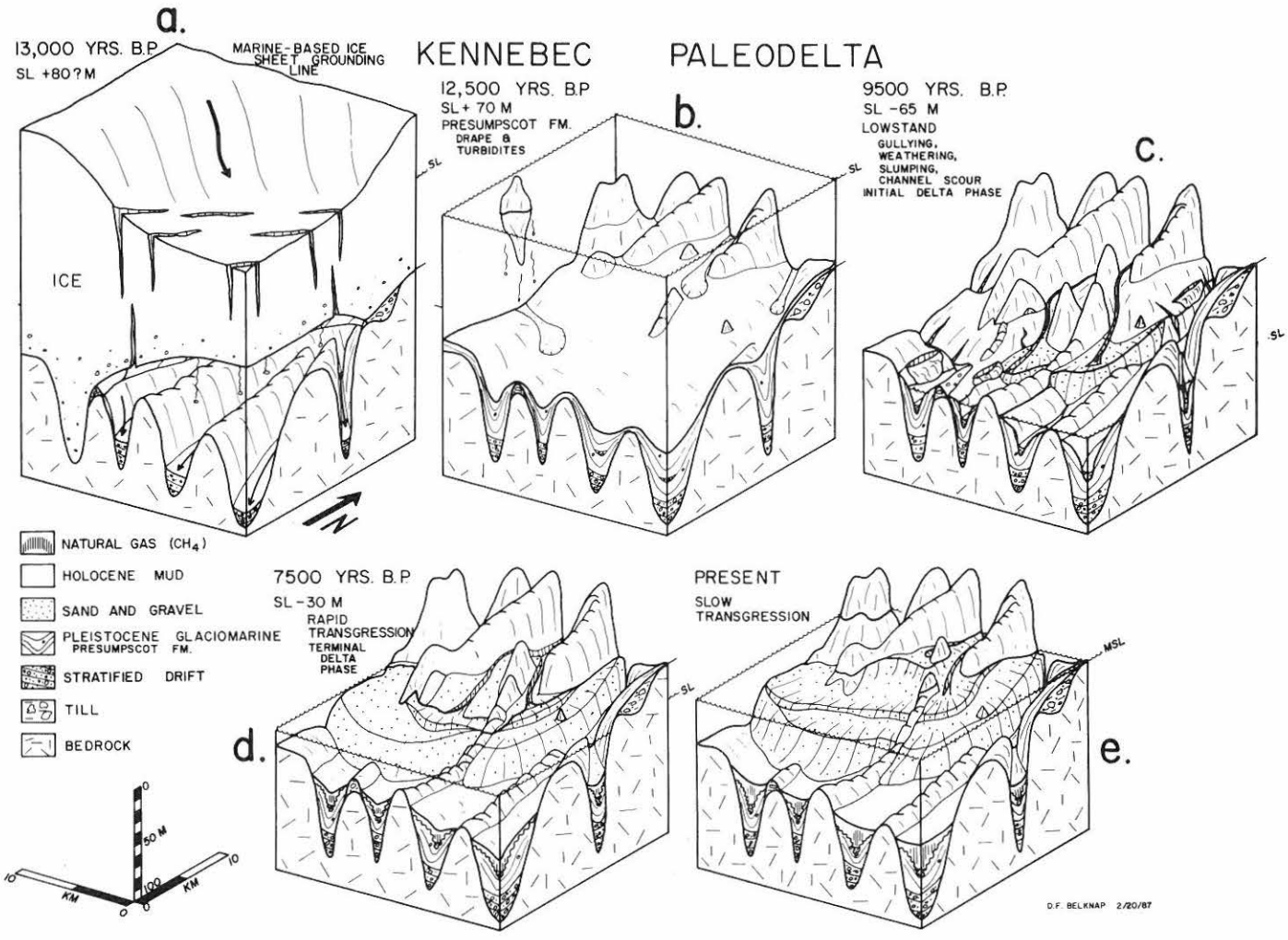

Figure 12. Three-dimensional block diagrams of evolution of the west-central Maine coast during (a) deglaciation, (b) highstand, (c) early Holocene lowstand, (d) early-mid Holocene transgression, and (e) present slow transgression.

We use the term "approximates" because the interface between these two depositional sequences is a time-transgressive surface which varies in nature from an erosional unconformity to a paraconformity and finally a conformity in deeper basins.

The west-central Maine inner shelf has evolved through the combined action of glacial, glaciomarine, littoral, and marine sedimentary processes. Bedrock structure controls the location of major sedimentary basins in the study area. Shelf valleys partially filled with Quaternary sediments can be traced from the 80 m depth contour on the present shelf into the present New Meadows and Sheepscot Rivers, and Booth Bay (Fig. 3). These represent paleo-estuaries which retreated landward during the latest marine transgression, with a concomitant shift of sedimentary environments (Belknap et al., 1986).

Evolution of the Sheepscot embayment - Kennebec paleodelta system is summarized in Figure 12. Local relative sea level fluctuated because of glacio-isostatic and eustatic effects (Fig. 2). This fluctuation caused a transgression of deep shelf environments in contact with retreating glacial ice from 14,000 to 12,500 yr B.P. Figure 12a shows the calving ice margin grounded on bedrock pinning points, depositing till and stratified drift and proximal glaciomarine sediments. Figure 12b is the setting at maximum highstand. A regression occurred from 12,500 to ca. 9,500 yr B.P., resulting in the lowstand setting portrayed in Figure 12c. The sea paused at the lowstand position long enough to cut distinct shorelines at 50-65 m depth (Shipp et al., this volume). A strong unconformity was created at the top of the Pleistocene section during the regression and lowstand, while in the shelf basins sedimentation was more-or-less continuous. Massive influx of sand and gravel from the Kennebec River built a delta during both regression and transgression, prograding over Pleistocene units. As sea level rose and sediment input waned, progressively shallower delta lobes were built on the upper surface of the paleodelta closer to the present Kennebec River mouth (Figure 12d). In the mid-Holocene, shoreface erosion became the dominant process controlling the shoreline.

Slowing of the rate of sea-level rise allowed a strong ravinement surface to form from -20 m to the present shoreface (ca. 6,000 yr B.P. to present; Fig. 2). Figure 12e portrays the present continuing slow transgression.

Establishing the seismic stratigraphic setting has been a first step to understanding the complex geologic evolution of the west-central Maine inner shelf. We are presently undertaking a program of offshore coring in conjunction with continued seismic and submersible studies, in order to better constrain the assumptions on the remote data. In addition, we hope to tie this record with that of the deeper Gulf of Maine and the onshore record to more completely understand the full deglacial history of the region.

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