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# Geophysical evidence of a late Pleistocene glaciation and paleo-ice stream on the Atlantic Continental Shelf offshore Massachusetts, USA

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ABSTRACT

Interpretations of seismic reflection data collected offshore Massachusetts, USA, reveal the first conclusive geophysical evidence of a pre-Wisconsinan glaciation that extended beyond the limits of the Last Glacial Maximum (LGM) in the region. The data image numerous glacial geomorphic features that define the extent of a paleo-ice stream, including: (1) a regionally distributed erosion surface that forms a 50 km wide trough, with steeply eroded sidewalls (4°–18°) and nearly 100 m in relief at the margins; (2) a network of sub-ice sheet meltwater channels; and (3) a transparent, glacigenic seismic unit. The orientation of the paleo-ice stream trough indicates that the ice stream flowed to the south-southwest, toward the shelf break. This suggests that the ice stream formed further to the north, where it appears that Georges Bank (southeast of the Gulf of Maine, USA) redirected ice flow. Limited well data constrain the glacial erosion event (up to 300 m below sea-level) to occur within the Pleistocene. The glacial event represents a time of larger ice volume on the northern Atlantic continental shelf, as compared to the LGM; thus, we suggest that the event corresponds to marine oxygen isotope stage 12 (late Pleistocene) when the first major Pleistocene shelf-crossing glaciation began offshore southeastern Canada. These geophysical constraints on a late Pleistocene glaciation offshore Massachusetts have important implications for: (1) models of the Laurentide Ice Sheet, as the geomorphic evidence of pre-LGM ice streams are difficult to characterize yet account for most of the ice sheet's mass flux; and (2) the pore water salinity pattern offshore New England, as sedimentary basins near an ice sheet margin often contain large volumes of glacially emplaced freshwater.



## 1. Introduction

Geophysical evidence of Pleistocene ice sheets and ice streams are critical to reconstruct the timing and extent of Pleistocene glaciations (Dyke and Prest, 1987; Anderson et al., 2002; Dyke et al., 2002), to constrain models of Pleistocene ice sheet formation (Denton and Hughes, 2002; Marshall et al., 2002; Boulton and Hagdorn, 2006), and to evaluate the influences of Pleistocene ice sheets on subsurface freshwater distribution (Person et al., 2007; Cohen et al., 2010). The Wisconsin ice sheet has been well documented throughout the northern US and Canadian Atlantic shelf with the use of bathymetric data, sediment cores, and near-surface seismic data (Schlee and Pratt, 1970; Tucholke and Hollister, 1973; Uchupi et al., 2001; Shaw et al., 2006). Near-surface seismic data, multibeam data, and sediment cores have helped to characterize the distribution of the Wisconsin ice sheet; however, pre-Wisconsinan ice sheet deposits are too deeply buried to be characterized with these shallow-imaging methods (Piper, 1988). In many areas, the evidence of pre-Wisconsinan glaciation has been removed by erosion during the Wisconsinan glaciation (Giosan et al., 2002). Many studies have dated deposits of glacially eroded sediments on the continental slope offshore southeastern Canada to infer the timing of pre-Wisconsinan ice sheets (Alam and Piper, 1977; Piper, 1988; Amos and Miller, 1990; Piper et al., 1994). Although these methods establish the timing of Pleistocene glacial cycles offshore, they do not provide conclusive evidence for the location and extent of ice sheet erosion on the continental shelf.

Offshore Massachusetts, USA, is a prime region to constrain the offshore extent of Pleistocene glaciations as it marks the transition from the repeatedly glaciated Gulf of Maine to the proglacial continental shelf offshore New Jersey, USA. Many studies of shallow, well-preserved glacial features have been used to infer the maximum extent of the Wisconsin ice sheet offshore Massachusetts (Fig. 1) including: (1) major erosion observed in near-surface seismic data (Uchupi, 1966; Oldale et al., 1974); (2) the distribution of gravel in surface sediment samples (Pratt and Schlee, 1969; Schlee and Pratt, 1970); and (3) moraines and glacio-tectonic structures on Martha's

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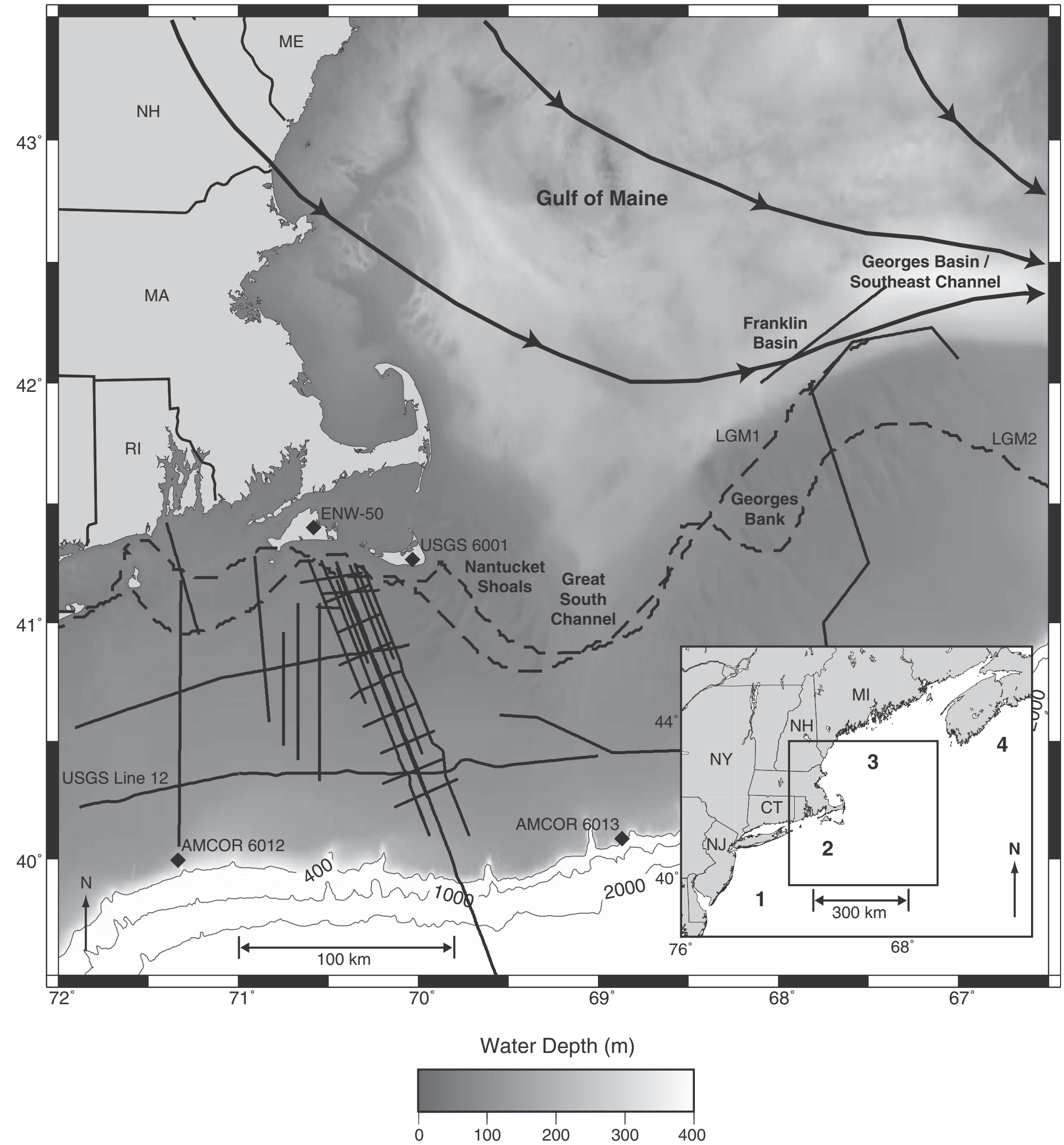


**Fig. 1.** Regional basemap showing seismic lines (solid black lines) and wells (diamonds) used in this study. Black lines mark location of single-channel, shallow seismic profiles (Uchupi, 1966; Knott and Hoskins, 1968), multi channel USGS seismic profiles (Schlee et al., 1976), and high-resolution, multi-channel seismic reflection profiles. Black diamonds mark the locations of AMCOR and UCGS wells (Folger et al., 1978; Hathaway et al., 1979). Black dashed lines mark the inferred boundary of the maximum extent of the Laurentide Ice Sheet from previous studies, LGM1 (Uchupi et al., 2001), LGM2 (Schlee and Pratt, 1970; Tucholke and Hollister, 1973). Black lines with arrows show LGM ice flow direction (Shaw et al., 2006). Inset basemap shows location of study region offshore New England, USA, with the four geographic regions used to define the Pleistocene glacial history: (1) offshore New Jersey; (2) offshore Massachusetts; (3) offshore Maine (Gulf of Maine); and (4) offshore southeastern Canada.

Vineyard, MA, USA and Nantucket Island, MA, USA (Oldale and O'Hara, 1984; Uchupi and Mulligan, 2006). These studies did not interpret pre-Wisconsinan ice sheet margins.

In August, 2009, we collected a high-resolution, multi-channel seismic survey offshore Massachusetts as part of an Integrated Ocean Drilling Program (IODP) site survey. Here, we present evidence for the extent of a late Pleistocene glaciation offshore Massachusetts based on our interpretations of the seismic data. The data image a regionally extensive erosion surface that has glacial–geomorphic features consistent with a paleo-ice stream trough. The direction of ice-stream flow was to the south-southwest and indicates that the ice stream originated to the north, near Georges Bank (southeast of the Gulf of Maine). Georges Bank contains similar glacial geomorphic features that we assume are contemporaneous with the glacial features offshore Massachusetts. From this we develop a regional interpretation for a paleo-ice stream that extended from the Gulf of Maine to offshore Massachusetts, and we infer a potential ice sheet margin. The interpreted glacial extent has a geometry similar to the Wisconsin ice sheet. Our interpretation of the seismic data suggests that the late Pleistocene glaciation extended farther south across the Massachusetts continental shelf than previously thought. This has important implications as it can help constrain hydrogeologic models used to predict the emplacement of sub-surface freshwater, and glacial models used to predict the formation of Pleistocene ice sheets.

## 2. Geologic setting

### 2.1. Continental shelf

Our study region lies on the passive continental margin offshore Massachusetts. The formation of the continental shelf began with rifting of the Atlantic during the late Triassic to early Jurassic. The rifting formed basement rock comprised of a series of normal fault blocks

(Hutchinson et al., 1986). As the basement rock subsided, a thick sedimentary wedge accumulated from the Cretaceous through the present. The result was a passive margin with a sedimentary wedge up to 14 km thick (Schlee et al., 1976; Poag, 1978).

### 2.2. Pleistocene glacial history

We define four geographic regions on the Atlantic continental shelf to address its Pleistocene glacial history: (1) offshore New Jersey; (2) offshore Massachusetts; (3) offshore Maine (Gulf of Maine); and (4) offshore southeastern Canada (Fig. 1). To the south of our primary study region, offshore New Jersey shows little evidence of Pleistocene glaciations (Carey et al., 2005). North of our study region, offshore Maine and offshore southeastern Canada contain abundant evidence of multiple Pleistocene glaciations (Piper et al., 1994). Offshore Massachusetts, there is limited information on Pleistocene glaciations and their extent beyond the Laurentide Ice Sheet during the LGM, which reached the islands of Martha's Vineyard and Nantucket (Fig. 1).

Much of the understanding of Pleistocene ice sheets on the Atlantic continental shelf comes from interpretations of glacially derived sediments on the continental slope offshore southeastern Canadian. Piper et al. (1994) interpreted variations in sediment type and thickness observed in a 20 m piston core sample obtained on the continental rise offshore southeastern Canada as evidence for numerous episodes of glaciation in the late Pleistocene. They noted that isotope stages 14 and 12 were the first widespread Pleistocene glaciations on the continental shelf as indicated by the significant erosion of Cretaceous to Tertiary sediments, and their deposition on the slope. Interpretations of seismic reflection data on the continental slope offshore Nova Scotia also show that isotope stage 12 was the first wide-spread Pleistocene glacial event (Piper and Normark, 1989; Berry and Piper, 1993). To the north, sediments offshore Newfoundland show that the mid-Illinoian ice sheet (marine oxygen isotope stages 8 and 6) covered most of Grand Banks and reached the shelf break, extending farther than the Wisconsin ice sheet (Alam and Piper, 1977; Huppertz and Piper, 2009). These studies suggest that during the early Pleistocene, glaciations did not reach the southeastern Canadian shelf, as there are few glacially derived sediments. Thus, the onset of shelf-crossing glaciations appears to have started in the middle to late Pleistocene.

The Gulf of Maine also contains evidence of repeated Pleistocene glaciations. Core data from the Gulf of Maine show several episodes of glacial erosion, till deposition, and glacial outwash (Pratt and Schlee, 1969; Schlee and Pratt, 1970; Uchupi, 1970; Oldale and O'Hara, 1984). Interpretations from low-resolution seismic data offshore Massachusetts show Pleistocene glacial unconformities and large, buried channels associated with glacial scouring (Knott and Hoskins, 1968; McMaster and Ashraf, 1973a, 1973b). Glacio-tectonic thrust structures identified on sea cliffs on Martha's Vineyard and Nantucket also reflect repeated Pleistocene glaciations (Oldale and O'Hara, 1984). These studies show evidence of multiple Pleistocene ice sheet advances; however, the ability to constrain maximum glacial extent offshore Massachusetts has been limited by the available data.

The marine oxygen isotope record supports the interpretations of geologic data on the Atlantic continental shelf and shows that the Pleistocene contained multiple episodes of widespread ice accumulation (Giosan et al., 2002). In the late Pleistocene, glacial cyclicity changed to an approximately 100 ka cycle with a corresponding larger accumulation of ice volume during glacial cycles (Williams et al., 1988; Balco and Rovey, 2010). The longer period of ice accumulation during the late Pleistocene led to an increased ice extent onto the continental shelves, as indicated in the geologic record (Piper et al., 1994). Thus, sedimentologic and isotopic data indicate that the largest Pleistocene advances of ice sheets onto the Atlantic continental shelf appear to have been more common in the late Pleistocene.

## 3. Data and methods

We collected a grid of high-resolution, multi-channel seismic data offshore Massachusetts using the Scripps Institution of Oceanography's portable seismic system. The system employs a 45 in.$^3$/105 in.$^3$, generator–injector (GI) air gun source that produces frequencies up to 200 Hz, and a digital streamer with 48 hydrophone groups spaced at 12.5 m. We collected seven north–south (dip) lines, and 11 east–west (strike) lines (Fig. 2). Seismic processing included outside trace muting, bandpass filtering, true amplitude recovery, velocity filtering in the f–k and radon transform domains, deconvolution in the tau–p domain, normal moveout (NMO) correction, post-stack deconvolution, Kirchhoff post-stack depth migration, and deconvolution in the f–x domain. In 50–150 m water depth, the water bottom multiple has been sufficiently suppressed by our processing. In 25–50 m water depth, the water bottom multiple interferes with true reflections, and interpretation is more difficult in the shallow section. Stacking velocity analysis was performed every 500 m. Interval velocities, based on the stacking velocities, were used for depth migration. Our data achieve a horizontal and vertical resolution of 4–9 m, assuming peak frequency of 100 Hz and an average compressional wave velocity of 1800 m/s.

Data processing produced clear, interpretable images, enabling us to identify prominent reflections, infer depositional units, and constrain time stratigraphic sequences up to 800 m in depth. We relate sediment deposition styles to glacial and non-glacial processes.

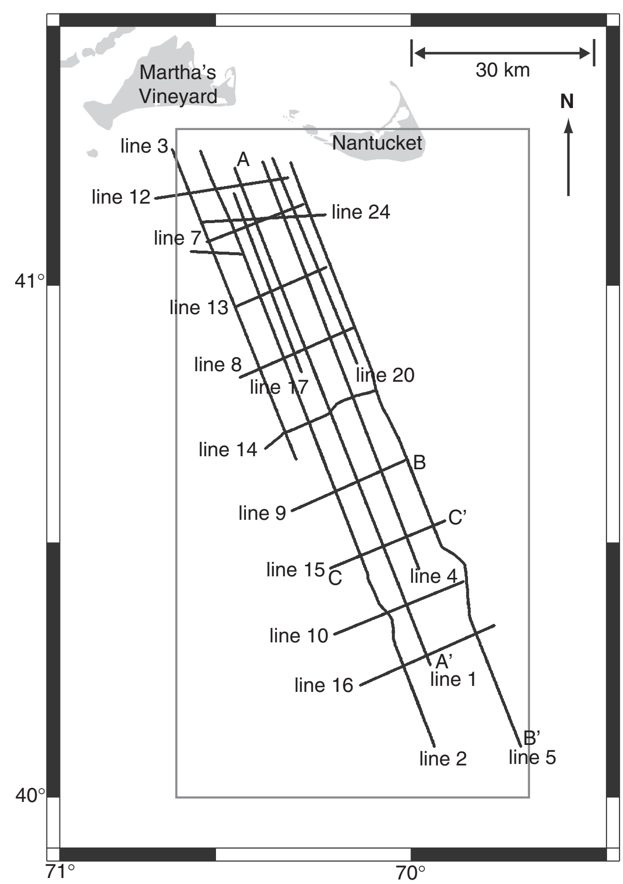


**Fig. 2.** Basemap showing location of high-resolution, multi-channel seismic reflection lines collected for this study offshore Martha's Vineyard and Nantucket Islands, USA. Seismic lines are labeled by their acquisition line number. The gray box defines the region of the isopach maps (Fig. 4).

## 4. Observations, interpretations, and age estimates

Six stratigraphic units (A, B, C, D, E, and F) and two regional unconformities (U1 and U2) are identified based on their seismic character, amplitude, and bounding surfaces (Fig. 3).

There are no direct well ties to the seismic lines we acquired; thus age estimates rely on correlation of observed stratigraphic architecture offshore Massachusetts with dated stratigraphic architecture interpreted on high-resolution seismic reflection profiles collected offshore New Jersey (Poulsen et al., 1998; Steckler et al., 1999; Duncan et al., 2000; Metzger et al., 2000; Goff et al., 2005; Monteverde et al., 2008; Nordfjord et al., 2009) and on shallow, low-resolution seismic profiles collected offshore Martha's Vineyard (Knott and Hoskins, 1968; McMaster and Ashraf, 1973a). The units interpreted from the New Jersey seismic data are tied to Ocean Drilling Program (ODP) sites, which provide good age control for the New Jersey seismic units. Thus, the dated seismic data offshore New Jersey help us estimate the age of seismic units offshore Massachusetts based on observed similarities in their seismic sequence architectures. For example, the deeper, pre-glacial sediments offshore Massachusetts change from a predominantly parallel-aggradational reflection package (unit D in Fig. 3) to a progradational reflection package with greater dip (unit C in Fig. 3). This change in reflection stacking pattern is also observed offshore New Jersey, where it is interpreted to represent a 20 fold greater siliciclastic sediment supply during the Oligocene and Miocene relative to the Cretaceous and Eocene (Poulsen et al., 1998; Steckler et al., 1999; Monteverde et al., 2008). In addition, we estimate age by correlating seismic surfaces observed in USGS seismic data offshore Massachusetts with nearby Atlantic Margin Coring Project (AMCOR) well data (Hathaway et al., 1979) and USGS well data (Figs. 1 and 5) (Folger et al., 1978; Hall et al., 1980).

### 4.1. Pre-Pleistocene, non-glacial units

#### 4.1.1. Unit F — Jurassic basement

Unit F is the deepest unit interpreted, and represents the limit of penetration from the seismic source. The unit contains few internal reflections. A high impedance contrast produces the top, high-amplitude bounding reflection, where unit F has an average P-wave velocity of 5 km/s and the overlying strata have an average P-wave velocity of 2–3 km/s. The unit is interpreted as acoustic basement based on its high velocity. McMaster and Ashraf (1973a) observed acoustic basement of similar seismic character, depth, and dip in seismic profiles southwest of Martha's Vineyard. Folger et al. (1978) describe Jurassic basalt at 460 m below sea-level (mbsl) on Nantucket Island in USGS 6001 well data (Fig. 1). We conclude unit F is Jurassic basement.

#### 4.1.2. Units E and D — Cretaceous to Eocene carbonate mud

Unit E is bounded below by unit F. It is mainly transparent and contains few continuous, sub-horizontal reflections that have a dip direction to the south (Fig. 3). The unit is interpreted as Cretaceous as it overlies the Jurassic basement. Cretaceous strata offshore Martha's Vineyard (McMaster and Ashraf, 1973a) and offshore New Jersey (Steckler et al., 1999) display similar seismic character to unit E. Steckler et al. (1999) identified the Cretaceous–Tertiary boundary reflection ranging in depth from 0.5 km near shore to 1.5 km near the shelf break offshore New Jersey, a similar depth range of unit E.

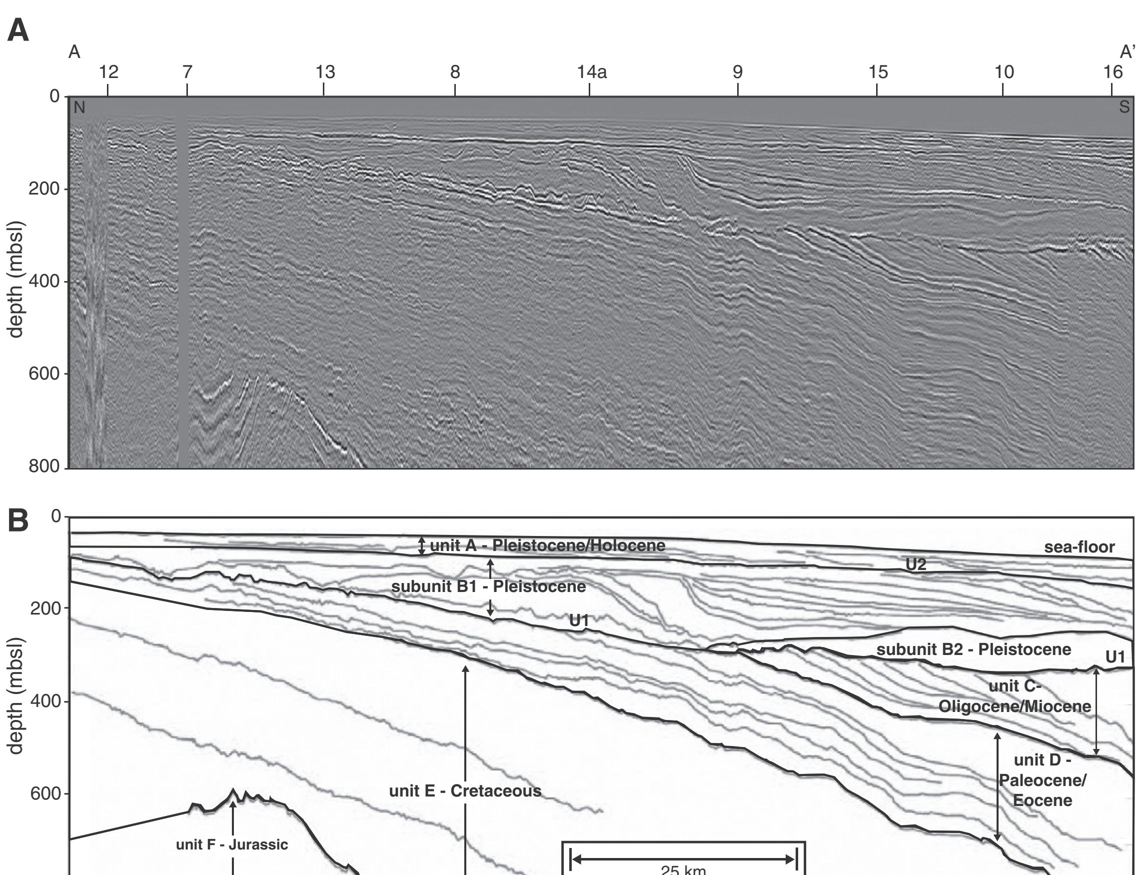


**Fig. 3.** (A) Uninterpreted seismic line 1 (A–A′) and (B) interpreted line drawing of major depositional units. Seismic line 1 (A–A′) is located in Fig. 2. Numbers at the top of the profile indicate the location of crossing seismic lines (Fig. 2).

Unit D overlies unit E, and is bounded above by the U1 unconformity and unit C. Within unit D, seismic reflections are continuous, sub-horizontal, and high-amplitude (Fig. 3). The reflections have an aggradational stacking pattern and a dip direction to the south, with an average dip of 0.35°. Unit D thickens from 50 m in the north to 300 m in the south (Fig. 4). The high-amplitude, continuous, aggrading reflection pattern of unit D is similar to reflection patterns of late Paleocene and early Eocene strata observed in the New Jersey seismic data (Steckler et al., 1999).

Steckler et al. (1999) determined that the late Cretaceous to early Eocene represented a time when sedimentation was low, subsidence was high, and carbonate precipitation dominated sediment input. This produced parallel, sub-horizontal reflections. Given the interpreted age of units E and D, and their correlation with the stratigraphic architecture offshore New Jersey, the units are inferred to be dominated by carbonate mud.

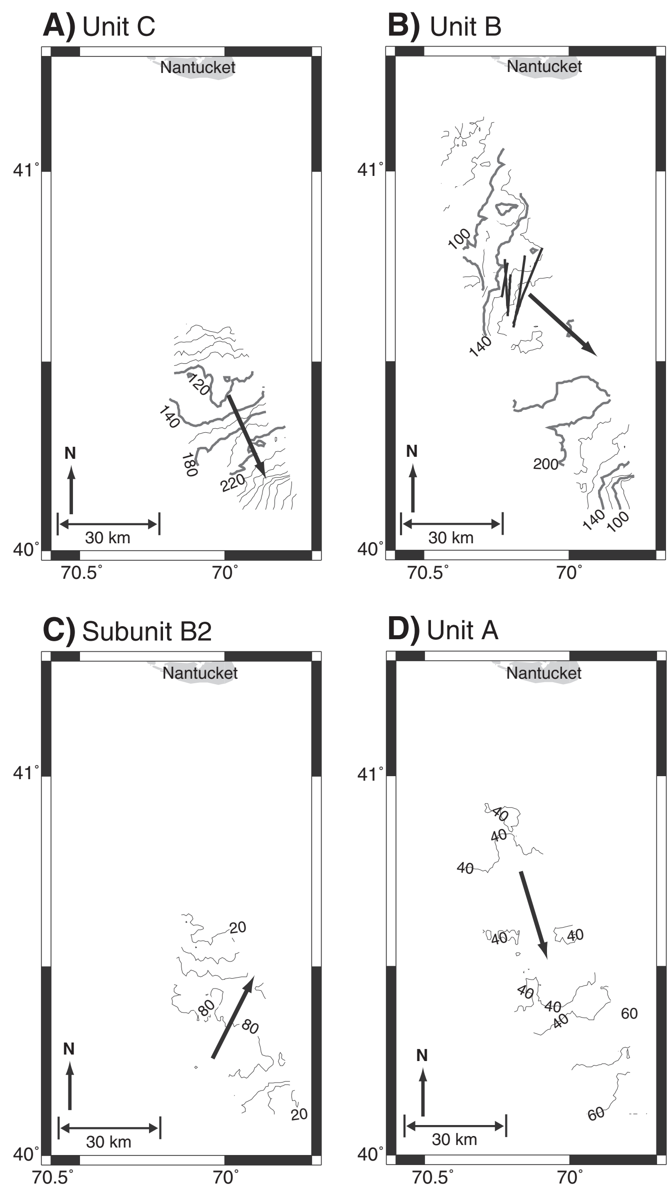


**Fig. 4.** Isopach maps of seismic stratigraphic units (A) unit C, (B) unit B, (C) subunit B2, and (D) unit A. Contours are thickness in intervals of 20 m. Bold contours are labeled for clarity. Arrows show our interpreted direction of sediment outbuilding based on orientation of internal reflection geometry. (C) Black lines mark mapped locations of unit B clinoform rollovers which are also used to estimate direction of sediment outbuilding.

*4.1.3. Unit C — Oligocene and Miocene siliciclastics*

No major correlative surface separates unit D from unit C; however, a change in their seismic stacking patterns exists. Unit C is bounded above by an unconformity (U1). Unit C reflections have higher amplitudes and steeper dip (0.6°) than unit D and form a progradational package (Fig. 3). Unit C does not exist in the north, but forms a southward thickening wedge that reaches 400 m in the south (Fig. 4). We infer unit C was originally thicker in the south, and extended further to the north, but erosion removed much of the sediment (Fig. 6).

The change from aggradation (unit D) to progradation (unit C) likely corresponds to an increase in sediment supply or an increased frequency of eustatic rise and fall. In this case, the observed change in stratigraphic architecture correlates with an increase in sediment supply during the Oligocene and Miocene observed offshore New Jersey, when the siliciclastic sediment input increased 20 fold (Poulsen et al., 1998; Steckler et al., 1999; Monteverde et al., 2008). The observed change from aggradation to progradation also correlates with a global pattern of continental margin stratigraphic architecture during the Oligocene and Miocene (Bartek et al., 1991). Based on unit C's similarity to the stratigraphic architecture of Oligocene and Miocene strata observed offshore New Jersey, we conclude the unit to be of Oligocene and Miocene age.

*4.2. Pleistocene non-glacial units*

*4.2.1. Unit B — Pleistocene siliciclastics*

Unit B is underlain by a regional unconformity (U1) and overlain by a regional unconformity (U2). We divide the unit into two subunits: B1 and B2. B1 comprises the majority of the unit, and contains prominent reflections that are considered to be non-glacial siliciclastic sediments, as they exhibit a progradational clinoform morphology. B2 is a mainly transparent layer that overlies the deeper portions of the U1 unconformity in the south and is considered glacial in origin, as the sediments build-out landward.

Reflections in subunit B1 form high-amplitude clinoforms (Fig. 3). In the north, reflections are sub-horizontal with shallow channels up to 15 m deep and up to 300 m wide. Clinoforms with components of aggradation and progradation overlie and downlap onto the transparent layer in the south. Individual clinoforms within B1 can be correlated through most of the data, which allows us to map clinoform rollovers. Clinoform rollovers have a north-northeast strike, indicating progradation to the east-southeast (Fig. 4). Seaward of the clinoform rollovers, reflections are sub-horizontal. Metzger et al. (2000) describe a similar, non-glacial, Pleistocene depositional character in seismic data offshore New Jersey. They interpreted the seismic architecture to be the result of high sediment input with deposition responding to sea-level change. Based on these similarities, we suggest unit B1 is Pleistocene.

A Pleistocene age is also consistent with age estimates from AMCOR and USGS well data, which are used to infer the depth of the base of Pleistocene across our seismic data. We map several seismic horizons observed in USGS line 12 that correlate with AMCOR wells 6012 and 6013 (Fig. 5). AMCOR well 6012 is 80 km west of our seismic data. AMCOR well 6012 penetrated Pleistocene sediments from 0 to 290 m below the seafloor (mbsf) and Miocene sediments at ~290 mbsf; Pliocene sediments were absent (Hathaway et al., 1979). AMCOR well 6013 is located 60 km east of our seismic lines. It penetrated Pleistocene sediments to 300 mbsf (Hathaway et al., 1979). In USGS line 12, a mapped horizon correlates with the Miocene–Pleistocene boundary in AMCOR well 6012. This horizon is assumed to represent the base of the Pleistocene in USGS line 12. The base of the Pleistocene horizon crosses our high-resolution seismic data at a depth of approximately 450 mbsl in the southern extent of our seismic lines (Fig. 5). This is approximately 50 m below the base of unit B. We also project the mapped base of unit B to well ENW-50 on Martha's Vineyard, approximately 20 km north of line 1 (Fig. 1). The projected base of unit B

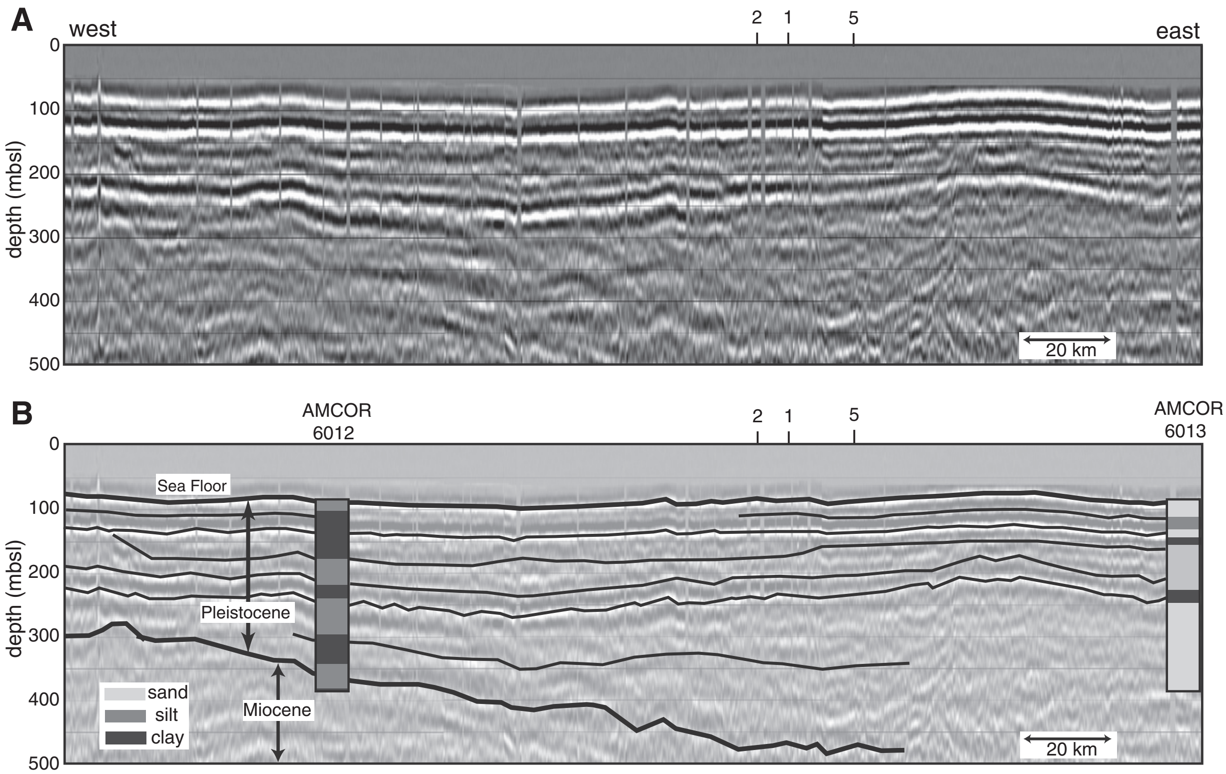


**Fig. 5.** (A) Uninterpreted USGS seismic line 12 data and (B) interpreted USGS seismic line 12 with interpreted lithology from AMCOR well data. The deepest mapped horizon is assumed to correlate with the base of AMCOR 2012 that separates Miocene sediments from Pleistocene sediments. Numbers at the top of the profiles mark the location of crossing seismic lines (Fig. 2).

approximately coincides with the base of Pleistocene sediments on Martha's Vineyard identified by Hall et al. (1980).

Martha's Vineyard and Nantucket Island well data show that Pleistocene sediments are predominantly sand inter-bedded with thin layers of silt (Folger et al., 1978; Hall et al., 1980). Pleistocene sediments in AMCOR well 6013 were sand-dominated, while Pleistocene sediments in AMCOR 6012 were silt- and clay-dominated (Fig. 5) (Hathaway, et al., 1979). Based on these regional well data, we interpret unit B is composed of layers of sand, silt, and clay of various thickness.

*4.2.2. U2 unconformity*

U2 is a shallow, erosional unconformity that parallels the seafloor. The unconformity is a high-amplitude reflection that separates unit B reflections below from sub-horizontal, downlapping unit A reflections above.

The shallow depth of U2 suggests it is a young feature. Studies offshore New Jersey (Duncan et al., 2000; Goff et al., 2005; Nordfjord et al., 2009) identified a shallow sequence boundary that is similar to U2; the sequence boundary formed during the last sea-level fall approximately 40 ka–30 ka. Based on the similarity of U2 with the sequence boundary offshore New Jersey, we conclude U2 is the sequence boundary formed during the last sea-level fall, and it separates Pleistocene deposits below from late Pleistocene to Holocene deposits above.

*4.2.3. Unit A — Late Pleistocene and Holocene siliciclastics*

Unit A is bounded by a large, regional unconformity (U2) below and the seafloor above. The sequence contains low-amplitude, sub-horizontal reflections that prograde to the south (Fig. 3). Reflections in the north have small channel incisions tens of meters wide and up to 10 m deep. In the north, the sequence is thin with a near-constant thickness of approximately 40 m. In the south, the sequence thickens to 60 m as U2 deepens (Fig. 4).

We conclude that unit A represents deposition during the late Pleistocene and Holocene. Studies of shallow sediment structure offshore New Jersey (Duncan et al., 2000; Goff et al., 2005; Nordfjord et al., 2009) identified a similar late Pleistocene to Holocene sequence. Shallow sediment core data offshore Massachusetts show sediment accumulation rates between 0.625 mm/yr–1.25 mm/yr (Bothner et al., 1981). These sedimentation rates place the base of unit A (40 mbsf) between 32 ka and 64 ka, which is consistent with our age estimate of late Pleistocene to Holocene.

Well data show near-surface sediments offshore New Jersey are sandy silt and clayey sand (Metzger et al., 2000). Shallow sediment samples offshore Massachusetts document that sediments in the top 6 m grade from sand in the northeast to 70% silt and clay in the southwest (Bothner et al., 1981). Based on the regional well data, we infer unit A consists of sandy silt and clayey sand.

## 4.3. Pleistocene glacial units

*4.3.1. U1 unconformity*

U1 is a regional unconformity that ranges in depth from 50 m to 350 mbsl. It is a high-amplitude reflection that truncates units D and C (Fig. 3). In the south, the unconformity has many incision features that are up to several kilometers wide and tens of meters deep (Fig. 7).

U1 has an overall dip direction to the southeast. However, in its most southeastern extent, the unconformity has a dip direction to the north (Figs. 3 and 6). U1 contains several steep, erosional edges (4°–18°) with relief up to 100 m in the southeast (line 5) and 80 m in the west (lines 9 and 14) (Fig. 6A–C). The structure map of U1 shows a 50 km-wide trough with steeply eroded edges forming the southeastern and western margins of the trough and a gentler relief forming the northeastern margin (Fig. 6D). The trough represents nearly 100 m of erosion from units C and D (Paleocene through Miocene), and potentially unit B (Pleistocene). We interpret U1 as a Pleistocene erosional unconformity as it truncates unit C of Miocene age, is overlain by subunit B2 of Pleistocene age, and is shallower than the base of Pleistocene identified with AMCOR and USGS well data.

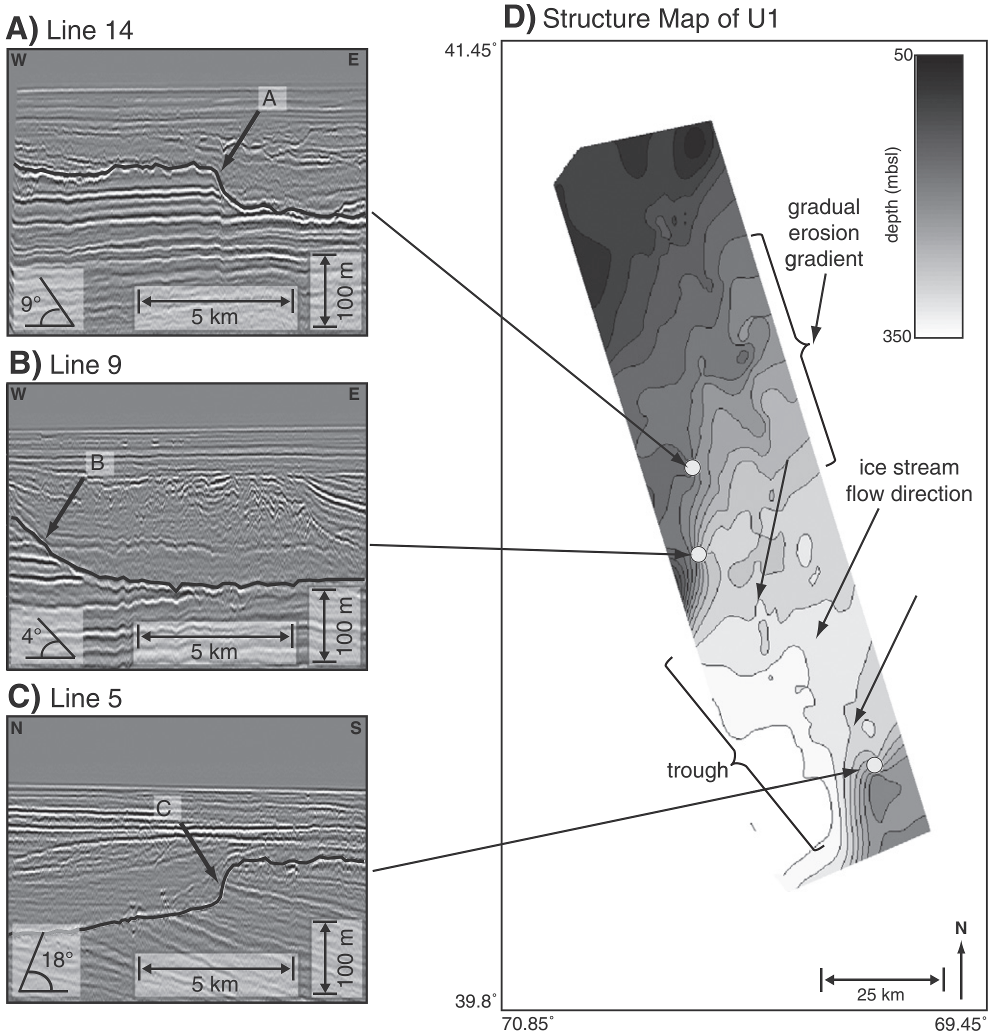


**Fig. 6.** (A–C) Seismic data showing the U1 unconformity (black horizon) and the steeply dipping erosional surfaces. The steeply dipping surfaces mark the margin of a paleo-ice stream. The black arrows show the location of the steeply dipping erosion surfaces plotted on the adjacent structure map. (D) Structure map of the U1 unconformity. Contours are depth (mbsl) in intervals of 20 m. The deep trough is assumed to be the remnant of a paleo-ice stream, with ice flowing in a south-southwest direction.

#### 4.3.2. Subunit B2 — Pleistocene glacigenic sediments

Subunit B2 overlays U1 in the deeper portions of the unconformity. It consists of a 25–75 m thick, mainly transparent, seismic reflection package with few discernible internal reflections (Fig. 7). The subunit is the only seismic package that thickens to the south-southwest (Fig. 4). Reflections on individual seismic lines have an apparent dip direction to the north; overlying and underlying sediments have an apparent dip direction to the south (Fig. 3). Horizon 1, a prominent reflection within subunit B2, is correlated through multiple seismic lines (Fig. 7). This surface has a dip direction to the north-northeast and forms a large, wedge-shaped package that thickens to the south-southwest. Other reflections within subunit B2 are assumed to have a similar dip direction. This geometry is indicative of landward progradation of subunit B2. The north-northeast dip direction of reflections within subunit B2, as well as the landward sediment building, suggests a different mechanism and sediment source for this subunit when compared to overlying and underlying units whose surfaces dip to the south and prograde basinward.

## 5. Interpretation of a late Pleistocene glaciation and ice stream

Based on our observations, we suggest the U1 unconformity records an episode of glacial erosion offshore Massachusetts. U1 has characteristics associated with glacial processes including; (1) it is wide-spread; (2) it contains a broad trough with a width-to-depth ratio similar to glacial troughs; (3) it is bounded above by glacigenic sediments (subunit B2); and (4) it contains a network of channels interpreted as sub-glacial drainage channels.

### 5.1. Glacial erosion and a paleo-ice stream

The geomorphic features of U1 suggest a glacial origin. The unconformity extends across our seismic study region, nearly to the continental shelf edge. Glacial erosion surfaces on continental shelves are often recognized by their basin-scale, regionally-distributed erosion (Vorren et al., 1989; Bart and Anderson, 1996), where erosion from the ice sheet truncates the underlying sedimentary layers of the continental shelf.

The erosion associated with U1 extends to nearly 400 mbsl, far too deep for subaerial erosion during a sea-level lowstand. Even if subsidence were fast (0.05 mm/yr) during the Pleistocene (Carey et al., 2005), allowing 100 m of subsidence, the unconformity would still be at subaqueous depths during a sea-level lowstand. Several edges of the unconformity dip steeply (4°–18° dip) and exhibit nearly 100 m of relief. This is more relief than is typical of many lowstand fluvial valleys on continental shelves, and the width-to-depth ratio (50 km wide–100 m deep) is not consistent with river valleys (100s of m wide–10s of m deep) (Anderson et al., 1996; Goff et al., 2005).

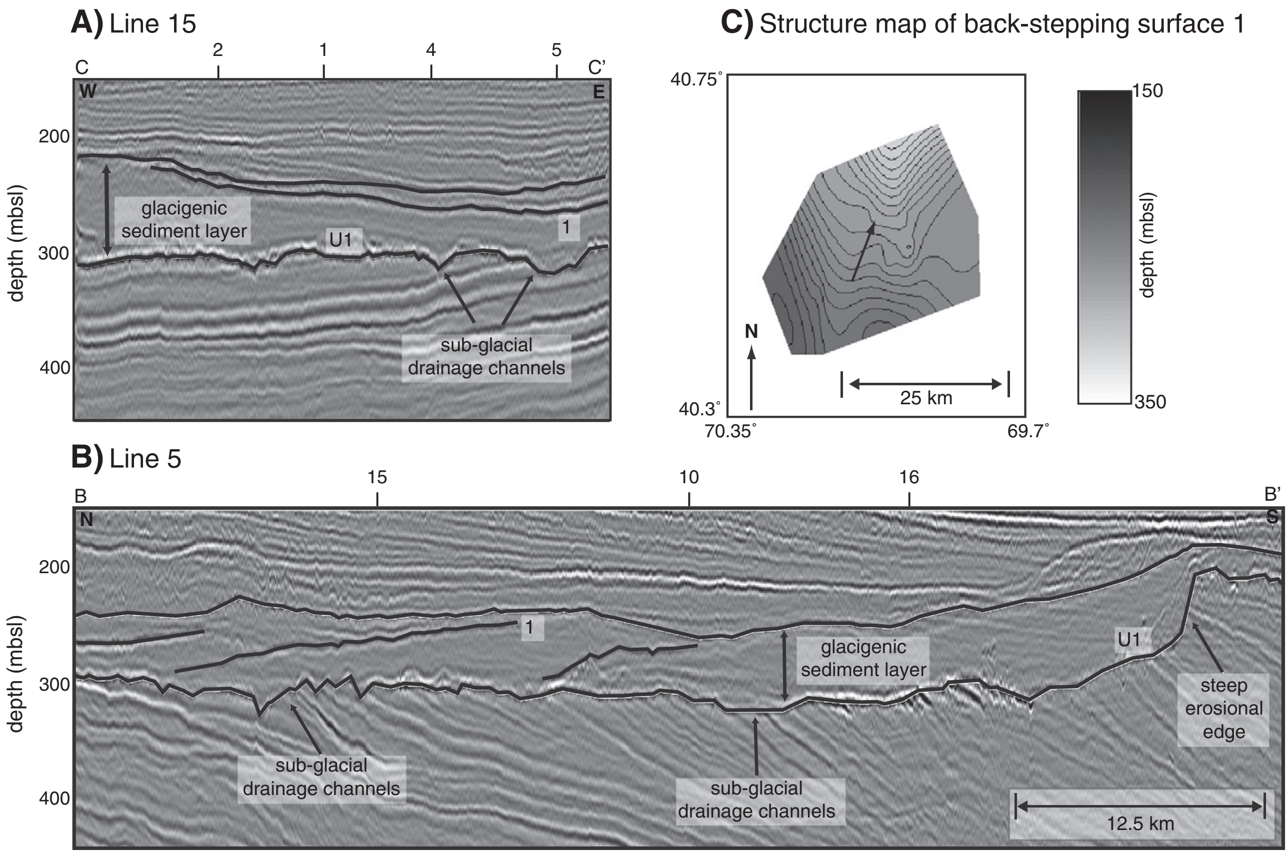


**Fig. 7.** (A–B) Seismic lines 15 and 5 (located in Fig. 2) showing the deeper portions of glacial unconformity U1. Large channels are interpreted to indicate sub-glacial drainage networks. We interpret the transparent layer above the unconformity as back-stepping glacigenic sediments. Note that these surfaces have a dip direction that is opposite to the underlying and overlying sedimentary units. Numbers at the top of the profiles mark the location of crossing seismic lines (Fig. 2). (C) Structure map of horizon 1 within the glacigenic sediment layer, contours are in intervals of 5 m. The surface dips to the north-northeast, as indicated by the black arrow.

Moreover, there is little evidence of an opposing, steeply eroded, high-relief wall; the steeply eroded sediment edges are only seen in the southeast (line 5) and west (lines 9 and 14). This lack of symmetry further indicates these are not lowstand fluvial valleys.

The trough identified in the structure map of U1 contains several geomorphic features commonly associated with ice streams, and therefore, we interpret it as the record of a paleo-ice stream. The width of the trough (approximately 50 km) is consistent with that of an ice stream (Stokes and Clark, 1999). Higher-latitude troughs have been observed with widths ranging from 50 to 150 km (Bart and Anderson, 1996; Wellner et al., 2001). The steep erosional edges are interpreted to represent the edge of the paleo-ice stream and mark the shear zone that forms between the fast flowing ice stream and slow moving ice sheet (Stokes and Clark, 1999). This also explains the asymmetric nature of the trough, as the ice stream may have more strongly eroded the eastern edge of the trough. Ottesen et al. (2005) show that several paleo-ice stream troughs offshore Norway are asymmetrical, where steep ridges are only observed on one side of the trough.

The direction of the paleo-ice stream flow is inferred to be to the south-southwest based on the orientation of the trough and the dip of the trough floor (Fig. 5). The ice stream eroded nearly 100 m of units C and D. This is a significant amount of erosion, but not uncommon for glacial erosion (Hallet et al., 1996; Smith and Anderson, 2010), particularly near the ice-sheet margin where glaciers are wet-based and upflow of groundwater can induce low effective-stress conditions (Boulton et al., 2001). Glacial erosion rates are thought to vary between 0.05 and 15 mm/yr depending on tectonic regime (Iverson and Person, 2011). This implies a minimum of 7000 years to erode 100 m of sediments within the trough, well within a typical Pleistocene glacial period (Williams et al., 1988).

The large amount of sediments eroded by the ice stream, and the interpreted direction of ice stream flow, suggests there should be a large accumulation of sediments (a trough mouth fan) on the continental slope south of the ice stream trough (Stokes and Clark, 1999). Inspection of bathymetry south of the paleo-ice stream trough does not reveal the topographic features of such a deposit. The absence of such a deposit, however, does not preclude the presence of an ice stream (Stokes and Clark, 2001). Given the limitations of our 2D seismic survey and its extent, it is reasonable to suggest an ice stream was responsible for the formation of the trough given the geomorphic features present.

*5.2. Glacigenic sediments*

The seismic character, back-stepping sediment fill pattern, north-northeast dip direction, and location of the transparent unit (subunit B2) all suggest a glacial origin for subunit B2. The lack of prominent, mappable surfaces, and the lack of sediment sampling prevents detailed analysis of the unit. For simplicity, we refer to the unit as glacigenic, as it is inferred to have a glacial origin, but the mechanism of sedimentation is not well constrained.

We suggest glacifluvial backfill as a possible mechanism for subunit B2. This process is often observed in glacial tunnel valleys. As an ice sheet retreats, channelized sub-glacial meltwater delivers eroded sediments to the ice sheet margin. The sediments then build landward as the ice sheet retreats (Kristensen et al., 2008). Glacifluvial backfill can be recognized in seismic data by back-stepping clinoforms and a sometimes transparent or chaotic seismic facies (Huuse and Lykke-Andersen, 2000; Praeg, 2003). The proposed ice stream offshore Massachusetts would have retreated to the north-northeast. Backfilling sediments would thus build to the north-northeast; similar to the back-stepping clinoforms observed in subunit B2. Glacifluvial backfill explains the north-northeast dip direction and landward sediment building observed in subunit B2 as related to the retreating ice stream.

### 5.3. Sub-glacial channels

A series of small, channel-like incisions occupy the deepest portions of the U1 unconformity (Fig. 6). The channels are 5 to 15 m deep and up to several kilometers wide. Some of the channels have a sub-horizontal base and have steep sides; however, we do not know their true 3D structure. The channel size and spacing is considerably smaller than the seismic line spacing, thus, individual channels cannot be correlated throughout the seismic data. Despite the limitations in characterizing the true shape of the channels, they are observed almost entirely in the trough, which suggests they are related to the process of streaming ice that created the trough. Streaming ice, in general, occurs because of: (1) a decrease in friction at the base of an ice stream from basal melting (Boulton et al., 2003), which would produce a network of sub-ice drainage channels; or (2) sliding sustained from soft, deformable sediments at the base of an ice stream (Alley et al., 1986), which would produce elongated incisions such as mega-scale glacial lineations.

We interpret the observed features to be the result of sub-glacial drainage. The large volume of glacigenic sediments overlying the unconformity suggests a wet-based ice sheet with channelized sub-ice sheet meltwater (Kristensen et al., 2008). The potentially flat base suggests constant fluid pressure across the channel as fluid flow was primarily driven by ice overburden pressure (Clark and Walder, 1994). Ice sheets and ice streams, with a low hydraulic gradient, often produce anastomosing networks of flat-based channels (Fountain and Walder, 1998). In addition, sub-ice sheet channels can form with deformation taking place over soft, clayey sediment, allowing the formation of channelized sub-ice sheet meltwater (Clark and Walder, 1994). The unconformity truncates unit C, which is likely a clay-prone substrate comprising the lower, more distal portion of prograding sediments.

Alternatively, changes in the substrate beneath an ice sheet can affect the amount of water available for channelized erosion; high meltwater infiltration rates in softer sediments result in less fluvial erosion (Lowe and Anderson, 2003). We assume, based on the consistency of the seismic reflections and our interpretations of units C and D as continuous siliciclastics and carbonate mud respectively, that the lithology does not change abruptly where channels occur, and thus does not control channel formation.

Another explanation for the observed features may be mega-scale glacial lineations, which occupy the outer portions of glacial troughs (Wellner et al., 2001; Wellner et al., 2006), similar to what we observe offshore Massachusetts. The dimensions of the channels are consistent with the dimensions of mega-scale lineations observed in troughs offshore Canada (Shaw et al., 2006) and offshore Norway (Ottesen et al., 2005), where lineations were observed to vary between 0.1 and 3.0 km in width, and reach depths of 15 m. Mega-scale glacial lineations, however, tend to have regular spacing, which we do not observe in the incisions in U1. In addition, mega-scale glacial lineations are often characterized as incisions in a bed of deforming till overlying the glacial unconformity. In our data, the incisions are part of the unconformity, not in an overlying till layer. A till layer on the order of a few meters thick, however, would not be imaged with the resolution of our seismic data. In addition, the underlying Miocene strata may be deformable enough to allow lineations to occur. Thus, the possibility of glacial lineations cannot be ruled out entirely.

### 5.4. Timing

From our seismic-based interpretations, we can estimate that the ice sheet was situated farther seaward on the Massachusetts shelf than the LGM position. The larger ice extent suggests that this glaciation may coincide with a previously recognized Pleistocene glacial advance beyond the LGM ice sheet limits (Berry and Piper, 1993; Piper et al., 1994). The marine oxygen isotope record shows stage 16 as the largest glacial period in the Pleistocene with the greatest ice accumulation (Williams et al., 1988). Stage 12 also shows more ice accumulation than the LGM. Sediment and seismic data from offshore southeastern Canada show that wide-spread glaciations did not cross the shelf until the beginning of the late Pleistocene, around marine oxygen isotope stage 12 (Berry and Piper, 1993; Piper et al., 1994). The large glaciation we interpret offshore Massachusetts is likely related to the glaciations observed offshore southeastern Canada as it also represents the first major glacial event on the continental shelf in the region. Thus, we suggest that the glaciation is late Pleistocene and coincides with marine oxygen isotope stage 12, when large-scale glaciation in North America began to reach the shelf break.

## 6. Discussion

### 6.1. Regional correlation of glacial features

The high-relief, erosional boundary (U1) offshore Massachusetts contains similar stratigraphic features to those observed on the northern side of Georges Bank (Fig. 8) (Uchupi, 1966; Knott and Hoskins, 1968; Uchupi, 1970; Oldale et al., 1974). Oldale et al. (1974) concluded that the erosional surfaces along the northern edge of Georges Bank were the result of glacial erosion in the Pleistocene as an advancing ice sheet reached the topographic high of Georges Bank. Based on the similarity in steep erosional relief and

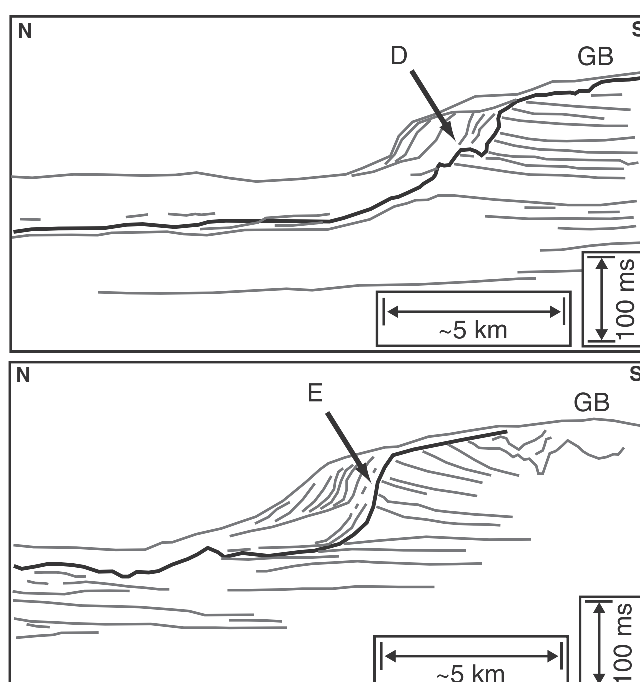


**Fig. 8.** Redrawn interpretations from Knott and Hoskins (1968) of seismic lines that cross the northern edge of Georges Bank (marked GB). These seismic lines display similar, steeply eroded sediments to what we observe offshore Massachusetts (Fig. 6).

subsurface depth, we assume the same ice sheet created the features on Georges Bank and offshore Massachusetts.

The large trough we observe in U1 is similar in size, scale, and orientation to basins that surround Georges Bank. Core data show Pleistocene sediments greater than 80 m thick fill basins on the northern edge of Georges Bank (Fig. 9) (Schlee and Pratt, 1970; Uchupi, 1970). Seismic data image a large unconformity, interpreted as a glacial erosion surface at the base of Pleistocene sediments in the Franklin Basin, just north of Georges Bank (Fig. 1) (Oldale et al., 1974). Uchupi (1970) suggested that Georges Bank acted as a barrier to ice sheet flow, and redirected ice movement to the south. As the ice flowed to the south, it eroded and over-deepened the basins around Georges Bank. Ice streams tend to form in basins because of increased basal melting from the greater ice overburden (Boulton et al., 2003). The deep basins of Georges Bank may have facilitated the development of an ice stream that was consequently diverted to the south-southwest by the topographic high of Georges Bank. We suggest that as the ice stream advanced south-southwest past Georges Bank, it continued to advance to the south-southwest and eroded shelf sediments offshore Massachusetts (Fig. 9). The ice advance, and subsequent erosion, created the trough observed in the offshore Massachusetts seismic data (Figs. 6 and 9).

From the distribution of glacial geomorphic features, including the steeply eroded edges and the orientation of the glacial trough (Fig. 6), a regional picture of the late Pleistocene glaciation in our study region emerges (Fig. 9). We assume a similar initial direction of ice advance for the late Pleistocene glaciation as the Laurentide Ice Sheet during the LGM, which advanced to the south-southeast across the Gulf of Maine (Fig. 1) (Schlee and Pratt, 1970; Stokes and Clark, 2001; Dyke et al., 2002). During the LGM, the topography of the Gulf of Maine channeled ice towards the southeast; however, the ice was redirected to the east, over-deepening the Southeast Channel and forming an ice stream that ran along the northern edge of Georges Bank (Fig. 1) (Shaw et al., 2006). During the late Pleistocene glaciation, we infer that a greater volume of ice flowed through the Gulf of Maine and forced an additional ice stream to develop near the western edge of Georges Bank, which then directed ice movement to the south.

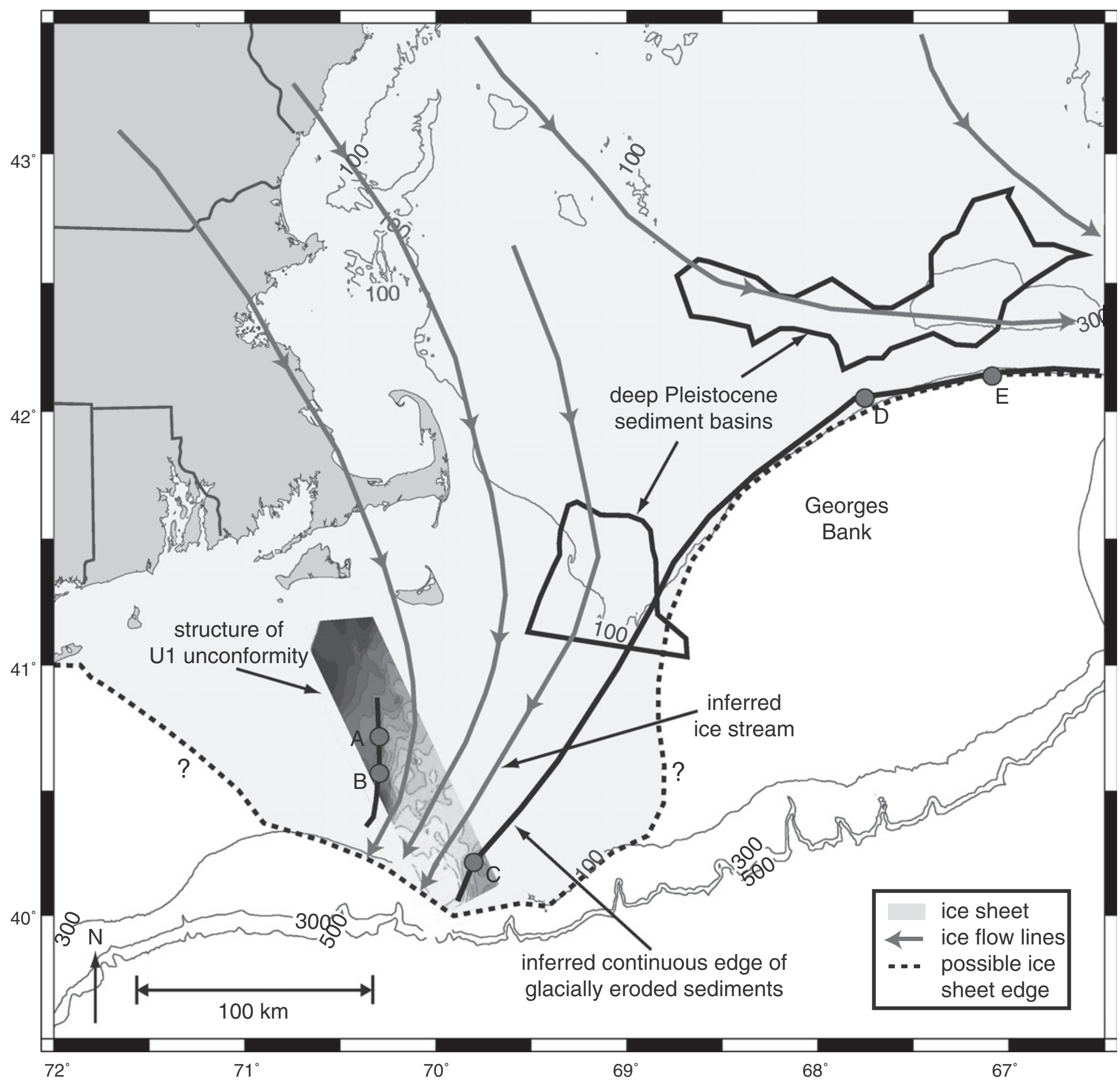


**Fig. 9.** Reconstruction of the late Pleistocene glaciation from our seismic interpretations and from previous observations on northern Georges Bank (Fig. 8). The morphology of the U1 unconformity, same as (Fig. 6D), is shown. Gray circles (labeled A–E) mark the location of steeply dipping erosional surfaces observed in seismic data (Figs. 6 and 8). The gray lines with arrows show the inferred direction of ice movement, and the black lines mark the inferred continuous edge of the ice stream. We assume the ice originally advanced in a similar direction to the Wisconsin ice sheet and was redirected to the south-southwest when it encountered Georges Bank. The redirected ice formed an ice stream that continued to advance to the south-southwest and carved the deep trough and steeply eroded sediments observed offshore Massachusetts. The dashed black line marks the possible ice sheet edge beyond the ice stream margins.

Assuming the ice stream reached its maximum extent during a sea-level lowstand (sea-level 100 m below present), the ice stream was grounded in 250 m of water. This water depth implies a minimum ice thickness of 275 m for the ice to remain grounded, assuming ice density is 90% of sea-water density.

There is no definitive evidence of the maximum extent of the late Pleistocene ice sheet in our study region; however, based on the location of the ice stream, it is reasonable to assume a southern extension of the ice sheet to near the shelf break just beyond the extent of our seismic data, more than 125 km offshore Massachusetts (Fig. 9). Shaw et al. (2006) showed that the Wisconsin ice sheet that streamed around the northern edge of Georges Bank was also the ice margin. It is possible that the ice stream we propose was also the ice sheet margin along the western edge of Georges Bank. Though, as the ice traveled farther south to offshore Massachusetts, the ice sheet may have continued to widen and erode the shelf sediments (Fig. 9). This is indicated by the continuation of the glacial erosion beyond the edge of the ice stream, as the glacial erosion surface (U1) extends further to the south (Fig. 6C) and to the west (Fig. 6A) beyond the steeply eroded edges.

The regional picture of the late Pleistocene ice sheet is consistent with marine-based studies of Pleistocene ice sheets offshore southeastern Canada (Berry and Piper, 1993; Piper et al., 1994). Evidence from land also shows late Pleistocene ice sheets achieving a greater ice extent than the LGM (Stanford, 1993; Bierman et al., 1999; Ridge, 2004; Balco and Rovey, 2010). Pre-Illinoian ice sheets are interpreted to have advanced 40 km farther than the Laurentide Ice Sheet in New Jersey (Stanford, 1993; Ridge, 2004). Balco and Rovey (2010) dated till sequences in Missouri, USA, that recorded several Pleistocene advances of the Laurentide Ice Sheet; they show that major advances of the Laurentide Ice Sheet were more common after the mid-Pleistocene transition to longer periods of ice accumulation. Our regional interpretation of a late Pleistocene ice sheet advance beyond the limits of the LGM is consistent with marine- and shore-based studies of the Laurentide Ice Sheet. Our observations provide further geophysical evidence for the extent of a late Pleistocene ice sheets offshore Massachusetts and the location of a paleo-ice stream.

### *6.2. Implications for shelf pore water salinity*

Pleistocene ice sheets have a strong control on the emplacement of subsurface freshwater onshore and offshore (Person et al., 2007; Bense and Person, 2008; Cohen et al., 2010). Geochemical analysis from USGS and AMCOR wells offshore Massachusetts show freshwater at depths that are greater than expected for equilibrium with present sea-level (Folger et al., 1978; Hathaway et al., 1979). The observed freshwater cannot be explained by modern topography and sea-level conditions, and likely was emplaced by sub-glacial recharge from the Laurentide Ice Sheet (Person et al., 2003; Marksamer et al., 2007). Several numerical modeling studies have attempted to predict the amount of freshwater emplaced offshore based on Pleistocene climate cycles; however, these studies only imposed the LGM for their glacial boundary condition (Person et al., 2003; Marksamer et al., 2007), and under-estimated the extent of at least one late Pleistocene glaciation interpreted from our data. Subsurface freshwater emplacement from ice sheets into permeable units within sedimentary basins and continental shelves can extend 50–100 km beyond the ice sheet margin (Marksamer et al., 2007; Bense and Person, 2008; McIntosh et al., 2011) and to depths of over 1 km (Lemieux et al., 2008). Thus, the regional extent and potential timing of a late Pleistocene ice sheet offshore Massachusetts enhances our ability to predict the emplacement of sub-surface freshwater offshore. Using an increased ice extent for Pleistocene glaciations from this study in a 2D numerical model doubled the amount of freshwater predicted offshore Massachusetts (DeFoor, 2011).

## 7. Conclusions

We used high-resolution, multi-channel seismic data to characterize a late Pleistocene glaciation that extended 125 km offshore Massachusetts. The age was estimated based on correlations with AMCOR and USGS well data offshore Massachusetts and adjacent seismic data offshore New Jersey. We suggest that the event may be related to oxygen isotope stage 12 when the first extensive Pleistocene glaciation was shown to cross the continental shelf offshore southeastern Canada.

The late Pleistocene glaciation is characterized by a regionally distributed erosion surface, and contains a 50 km wide trough with steep erosional edges. The trough is interpreted as the record of a paleo-ice stream. The base of the ice stream trough contains many small, incised features interpreted to be a network of sub-glacial drainage channels. Meltwater delivered sediments to the retreating ice stream margin, which produced glacigenic backfill sediments that overlay the unconformity.

The glacial geomorphic features we observe are similar to features observed on the northern side of Georges Bank. We assume the features on Georges Bank and offshore Massachusetts are contemporaneous and define the regional distribution of a late Pleistocene ice sheet. The basins of Georges Bank facilitated the development of an ice stream with Georges Bank acting as a barrier to ice flow and redirecting ice movement to the south-southwest. The ice stream over-deepened basins around Georges Bank and eroded the glacial trough we observe offshore Massachusetts.

Our observations have important implications for understanding Pleistocene glacial cycles on the Atlantic continental shelf. The late Pleistocene ice stream was previously unknown and can be used to constrain models of Pleistocene ice sheet accumulation and ice sheet mass flux. In addition, the greater extent of the late Pleistocene ice sheet offshore Massachusetts as compared to the LGM, suggest a potentially greater volume of emplaced sub-glacial meltwater in the shallow shelf sediments than previously predicted.

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