RICE UNIVERSITY

Evolution of Glacially Derived Freshwater and Overpressure in the Massachusetts Continental Shelf: An Integration of Geophysical and Numerical Methods

by

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A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE

Doctor of Philosophy

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HOUSTON, TEXAS
August 2013
ABSTRACT

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The continental shelf offshore Massachusetts, USA experienced repeated, late Pleistocene glaciations that drove freshwater into shelf sediments and generated overpressure in shelf sediments. To show this, I processed and interpreted high-resolution, multi-channel seismic data that were collected offshore Massachusetts. My interpretations define the extent of a late Pleistocene glaciation and the lithology of the shelf sediments. These interpretations are incorporated into hydrogeologic numerical models of fluid flow and solute (salt) transport.

My interpretations of the seismic data constrain the shelf stratigraphy and the location of a late Pleistocene (Marine Oxygen Isotope Stage 12) ice sheet. The ice sheet extended 100 km farther onto the shelf than the Laurentide ice sheet during the Last Glacial Maximum (LGM). The ice sheet also contained an ice stream that was likely sourced from the Gulf of Maine.

I show that this large, late Pleistocene ice sheet influenced the shelf hydrogeology by generating overpressure and emplacing freshwater into the shelf.
I first interpreted overpressure from high-resolution, full-waveform inversion P-wave velocities obtained from the seismic data. Then I modeled the development of the overpressure with a one-dimensional, finite-difference, fluid flow model that accounts for sedimentation and ice sheet loading. The results of the fluid flow model help to demonstrate that loading from the late Pleistocene ice sheet caused focused fluid flow that created localized zones of overpressure in offshore sediments. The localized overpressure is on the order of 1 MPa.

I simulate the emplacement of freshwater into the shelf sediments that resulted from this large, late Pleistocene ice sheet, as well as several subsequent late Pleistocene ice sheets. I do this with a three-dimensional, finite-element, variable-density model of fluid flow and heat and solute transport that accounts for ice-sheet loading and sea-level change. The model helps explain how the late Pleistocene ice sheets emplaced nearly 100 km$^3$ of freshwater into the sediments. Much of this freshwater is contained within 30 km of the coastline. My results thus integrate seismic interpretations of ice sheet history and shelf stratigraphy into numerical models of fluid flow to show how the glacial history influenced the present freshwater distribution. This freshwater is a potential, non-renewable resource for Massachusetts’s coastal communities.
I have many people to thank along the path of my academic journey. First and foremost, I would like to thank my parents, David and Debbie Siegel. They constantly encouraged me to pursue my studies and continually assured me that I would succeed with anything I put my mind to. Like most parents, they believed in me before I did.

I would like to thank my partner, fiancé, and friend Robert Sepeda. Robert’s support was instrumental in helping me by listening to me, motivating me, loving me, and showing me the values of hard work and organization. Thanks to him, I always look forward to coming home.

I am grateful to the rest of my family who have helped me in countless ways. My brother, Louis Siegel, helped any way he could, even while we were so far apart. I thank my cousins Meg Parascandola and Niccole Berthiaume for always providing warm family welcomes during our visits to New York. I am incredibly appreciative of my grandparents, Sidney and Lucille Siegel and Ed and Anne Snyder, none of whom ever had the educational opportunities I had, but from the very beginning of my education did everything they could to support me.

I thank my advisor Brandon Dugan, who spent many hours teaching me the meaning of concise and reliable scientific research. Brandon provided me with the tools I needed to accomplish this project and move the science forward. I thank Daniel Lizarralde who continually helped me, not just by imparting some of his
incredible knowledge of geophysics, but also for his constant enthusiasm for our work, which greatly motivated me. I thank Mark Person, who provided me with the program that aided in the completion of a significant portion of this thesis and for invaluable thoughts and insights. John Anderson provided me with helpful comments and wonderful conversations. I thank Helge Gonnermann, Dale Sawyer, and Philip Bedient who were kind enough to officially be involved with my thesis process. I am thankful for many meaningful conversations with other professors and staff at Rice, including Steve Danbom, Cin-Ty Lee, Fenglin Niu, Alan Levander and Mary Ann Lebar. I am also thankful to my other colleagues who helped in the collection of the data and with the interpretations, including Whitney DeFoor, Clint Miller and Helen Feng.

And finally, I am very grateful to all my friends throughout my life. They always encouraged me to finish my education and provided me with countless hours of rejuvenation through talk, laughter, travel and food. In semi-chronological order: Sara Andre, Tom Giordano, Craig Wertenberg, Emily Gruen, Jed Byers, Daniel Honegger, Erin Van Campen, Saipriya Choudhuri, Will Babbitt, Jeremy Schwartzbord, Thomas Chow, Melanie Dyjak, Chris Shirley, Chris Walczak, Justin Stigall, Hugh Daigle, Jeni Masi, Alex Witus, Lin Zheng, Lacey Pyle, Howard Pollack, Edgar Alanis, and Clint Miller. I am honored and blessed to have so many wonderful people in my life. To all of you, thank you.
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Freshwater in sedimentary basins throughout Europe, northern North America, and the Northern US Atlantic continental shelf have been shown to be the direct result of past climate cycles of glaciations and sea-level change (Person et al., 2012; Neuzil, 2012; McIntosh et al., 2012). During times of glaciation, increased hydraulic head beneath an ice sheet can force glacial-meltwater deep into basin sediments and beyond the edge of the ice sheet (Bense and Person; 2008; Lemieux et al., 2008c). During times of sea-level lowstands, the exposed continental shelf experiences more meteoric recharge and an enhanced gradient for freshwater flow from onshore (Meisler et al., 1984). Numerical modeling efforts that incorporate these past climate cycles as boundary conditions have been successful at matching the observed salinity patterns in many of these sedimentary basins and demonstrating that much of the freshwater is not in equilibrium with current climate conditions (van Weert et al., 1997; Person et al., 2003; Marksamer et al.,)
2007; Cohen et al., 2010; McIntosh et al., 2011). These types of models are an important step in understanding the emplacement and regional distribution of freshwater, and to assess safe aquifer yields (Kooi and Groen, 2003; Person et al., 2007a). In many offshore environments, however, well data and geophysical data are sparse, and little is known about the distribution of freshwater and its connection to past climate history.

In this dissertation, we focus offshore Massachusetts, USA, where freshwater was discovered on Nantucket Island up to 500 m below sea level (Folger et al., 1978). The observed freshwater is deeper than that predicted by the Ghyben-Herzberg principle (Marksamer et al., 2007). The region was glaciated throughout the Late Pleistocene (Oldale and O’hara, 1984; Uchupi et al., 2001), and the freshwater was likely emplaced by Pleistocene climate cycles. The extent of the freshwater beyond Nantucket Island, however, is unknown as there are no well data further offshore. The emplacement of freshwater offshore Massachusetts by Pleistocene climate cycles represents a potential resource to Massachusetts’s coastal communities, and thus we seek to characterize the extent and volume of freshwater offshore.

We collected a grid of high-resolution, multi-channel seismic data offshore Massachusetts in August 2009 to explore the coupled history of stratigraphy, glaciology, and hydrogeology of the shelf. This allows us to gain a better mechanistic understanding of the freshwater resources by integrating geophysical data and numerical modeling. This dissertation examines the data with an integration of
several numerical techniques: chapter 2 focuses on processing and interpretation of reflection seismic data to show the extent of a late Pleistocene ice sheet on to the shelf; chapter 3 focuses on full-waveform inversion of CDP gathers and a 1D fluid flow model of overpressure; and chapter 4 focuses on predicting salinity with a 3D, finite-element, variable-density, numerical model of fluid flow and heat and solute transport on the shelf.

In chapter 2, we interpret the seismic data and demonstrate the extent of a Marine Oxygen Isotope Stage (MIS) 12 glaciation that extended at least 125 km offshore Massachusetts. We describe several geomorphic features to constrain the presence and possible extent of the glaciation, including: (1) a 50 km wide trough with steeply eroded edges that we interpret as a paleo-ice stream trough; (2) a network of channels that we interpret as sub-glacial meltwater channels; and (3) a back-stepping sediment layer that we interpret as glacigenic sediment deposition during ice sheet retreat. From these analyses, we interpret that the MIS 12 ice sheet extended farther offshore Massachusetts than the Laurentide ice sheet did during the Last Glacial Maximum.

In chapter 3, we model glacially-derived overpressure that may be related to the MIS 12 ice sheet. Our interpretations are based on several geophysical observations including: (1) distinct regions of anomalously high-amplitude seismic reflections (bright spots) with a semi-disturbed seismic facies; and (2) the contrast in velocity profiles determined between the bright spots and adjacent regions with the use of a high-resolution, full-waveform inversion. The bright spots have lower
compressional velocity (190 m/s) when compared to undisturbed regions of the seismic section at the equivalent depth (250 mbsl). We use the contrasting velocities as input to a power-law model to show that the bright spots may have overpressures of 1 - 2 MPa greater than the adjacent, undisturbed sediments.

To determine mechanisms that could produce localized overpressure, we use a 1D fluid flow model that accounts for sedimentation, erosion, and ice-sheet loading. We incorporate the MIS 12 ice sheet and show that it may be a source of localized overpressure via focused fluid flow through permeable clinoform layers. These findings provide the first geophysical evidence of glacial overpressure on the continental shelf offshore Massachusetts and support the conclusions of previous numerical modeling studies that predicted overpressure offshore due to Pleistocene glaciations.

In chapter 4, we present our estimates for the distribution of glacially emplaced freshwater offshore Massachusetts. We show that the MIS 12 ice sheet emplaced as much as 100 km$^3$ of freshwater up to 30 km beyond Nantucket Island, MA. Our estimates are based on a 3D, finite-element, variable-density model of fluid flow and heat and solute transport over the last 3 Ma. Our model incorporates lithologic interpretations from the seismic survey and the MIS 12 ice sheet and sea-level change as climate boundary conditions. Our results indicate that the MIS 12 ice sheet emplaced up to three times as much freshwater offshore than previously estimated that extends 30 km beyond Nantucket Island. We run several sensitivity studies to address the uncertainty of inferred hydraulic properties on predicted
freshwater volume. All of the sensitivity studies show that the late Pleistocene ice sheet results in more freshwater offshore than would be estimated based on the LGM ice sheet extent.
Chapter 2

Geophysical Evidence of a Late Pleistocene Glaciation and Paleo-Ice Stream on the Atlantic Continental Shelf Offshore Massachusetts, USA

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

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1 This chapter is a reformatted version of the article “Geophysical evidence of a late Pleistocene glaciation and paleo-ice stream on the Atlantic Continental Shelf offshore Massachusetts, USA,” Marine Geology, 303-306, doi:10.1016/j.margeo.2012.01.007, copyright (2012) Elsevier, Reproduced with permission of Elsevier Journal.
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 meters 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.

2.1. Introduction

Geophysical evidence of Pleistocene ice sheets and ice streams are critical to reconstruct the timing and extent of Pleistocene glaciations (Anderson et al., 2002; Dyke and Prest, 1987; Dyke et al., 2002), to constrain models of Pleistocene ice sheet
formation (Boulton and Hagdorn, 2006; Denton and Hughes, 2002; Marshall et al., 2002), and to evaluate the influences of Pleistocene ice sheets on subsurface freshwater distribution (Person et al., 2007a; 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; Shaw et al., 2006; Tucholke and Hollister, 1973; Uchupi et al., 2001). 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; Amos and Miller, 1990; Piper, 1988; 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 (Figure 2-1).
including: (1) major erosion observed in near-surface seismic data (Oldale et al., 1974; Uchupi, 1966); (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 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.

Figure 2-1 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 (Knott and Hoskins, 1968; Uchupi, 1966), 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). Red 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). Blue 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.
2.2. Geologic Setting

2.2.1. 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 (Poag, 1978; Schlee et al., 1976).

2.2.2. 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 (Figure 2-1). 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 (Figure 2-1).
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 (Berry and Piper, 1993; Piper and Normark, 1989). To the north, sediments offshore Newfoundland show that the mid-Illinoian ice sheet (marine oxygen isotope stages 8 & 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 (Oldale and O'Hara, 1984; Pratt and Schlee, 1969; Schlee and Pratt, 1970; Uchupi, 1970). 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; McMaster and Ashraf 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 cyclisity changed to an approximately 100 ka cycle with a corresponding larger accumulation of ice volume during glacial cycles (Balco and Rovey, 2010; Williams et al., 1988). 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.

2.3. 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 45in³/105in³, 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 (Figure 2-2). 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 500m. 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 (see Appendix A for more details on processing).
2.4. 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 (Figure 2-3).
Figure 2-3 Seismic line 1 (A-A') uninterpreted (A) and (B) interpreted line drawing of major depositional units. Seismic line 1 (A-A') is located in Figure 2-2. Numbers at the top of the profile indicate the location of crossing seismic lines (Figure 2-2).

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 (Duncan et al., 2000; Goff et al., 2005; Metzger et al., 2000; Monteverde et al., 2008; Nordfjord et al., 2009; Poulsen et al., 1998; Steckler et al., 1999) 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 Figure 2-3) 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 (Monteverde et al., 2008; Poulsen et al., 1998; Steckler et al., 1999). 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 (Figures 2-1, 2-5) (Folger et al., 1978; Hall et al., 1980).

2.4.1. Pre-Pleistocene, Non-Glacial Units

2.4.1.1. 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 meters below sea-level (mbsl) on Nantucket Island in USGS 6001 well data (Figure 2-1). We conclude unit F is Jurassic basement.

2.4.1.2. 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 (Figure 2-3). 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.

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 (Figure 2-3). 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 (Figure 2-4). 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.
Figure 2-4  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.
2.4.1.3. 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 (Figure 2-3). Unit C does not exist in the north, but forms a southward thickening wedge that reaches 400 m in the south (Figure 2-4). We infer unit C was originally thicker in the south, and extended further to the north, but erosion removed much of the sediment (Figure 2-6).

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 (Monteverde et al., 2008; Poulsen et al., 1998; Steckler et al., 1999). 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.
2.4.2. Pleistocene Non-Glacial Units

2.4.2.1. 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 (Figure 2-3). In the north, reflections are sub-horizontal with shallow channels up to 15 meters 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 (Figure 2-4). 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 (Figure 2-5). AMCOR well 6012 is 80 km west of our seismic data. AMCOR well 6012 penetrated Pleistocene sediments from 0-290 meters 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 (Figure 2-5). 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 (Figure 2-1). The projected base of unit B 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 (Figure 2-5) (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.

![Figure 2-5 USGS seismic line 12 data (A) uninterpreted and (B) interpreted with 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 (Figure 2-2).](image)

2.4.2.2. 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.

2.4.2.3. 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 (Figure 2-3). 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 40m. In the south, the sequence thickens to 60 m as U2 deepens (Figure 2-4).

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 six meters 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.

2.4.3. Pleistocene Glacial Units

2.4.3.1. 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 (Figure 2-3). In the south, the unconformity has many incision features that are up to several kilometers wide and tens of meters deep (Figure 2-6).
Figure 2-6 Seismic data showing the U1 unconformity (red 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.
U1 has an overall dip direction to the southeast. However, in its most southeastern extent, the unconformity has a dip direction to the north (Figures 2-3, 2-6). 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) (Figures 2-6 A-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 (Figure 2-6 D). 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.

2.4.3.2. 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 (Figure 2-7). The subunit is the only seismic package that thickens to the south-southwest (Figure 2-4). 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 (Figure 2-3). Horizon 1, a prominent reflection within subunit B2, is correlated through multiple seismic lines (Figure 2-7). 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.

Figure 2-7 Seismic lines 15 and 5 (located in Figure 2-2) showing the deeper portions of glacial unconformity U1 (A-B). 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 (Figure 2-2). (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.
2.5. 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.

2.5.1. 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 (Bart and Anderson, 1996; Vorren et al., 1989), 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). 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 & 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 – 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 (Figure 2-5). The ice stream eroded nearly 100 meters 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 to 15 mm/yr depending on tectonic regime (Iverson and Person, 2012). This implies a minimum of 7,000 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.

2.5.2. 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.

2.5.3. Sub-Glacial Channels

A series of small, channel-like incisions occupy the deepest portions of the U1 unconformity (Figure 2-6). The channels are 5 to 15 meters 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.

2.5.4. 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 widespread 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.
2.6. Discussion

2.6.1. 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 (Figure 2-8) (Knott and Hoskins, 1968; Oldale et al., 1974; Uchupi, 1966; Uchupi, 1970). 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 subsurface depth, we assume the same ice sheet created the features on Georges Bank and offshore Massachusetts.

![Figure 2-8 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 (Figure 2-6).](image)
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 (Figure 2-9) (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 (Figure 2-1) (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 (Figure 2-9). The ice advance, and subsequent erosion, created the trough observed in the offshore Massachusetts seismic data (Figures 2-6, 2-9).

From the distribution of glacial geomorphic features, including the steeply eroded edges and the orientation of the glacial trough (Figure 2-6), a regional picture of the late Pleistocene glaciation in our study region emerges (Figure 2-9). 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 (Figure 2-1) (Dyke et al., 2002; Schlee and Pratt, 1970; Stokes and Clark, 2001). 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 (Figure 2-1) (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. 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.
Figure 2-9 Reconstruction of the late Pleistocene glaciation from our seismic interpretations and from previous observations on northern Georges Bank (Figure 2-8). The morphology of the U1 unconformity, same as (Figure 2-6 D), is shown. Gray circles (labeled A-E) mark the location of steeply dipping erosional surfaces observed in seismic data (Figures 2-6, 2-8). The blue lines with arrows show the inferred direction of ice movement, and the red 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.
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 (Figure 2-9). 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 (Figure 2-9). 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 (Figure 2-6 C) and to the west (Figure 2-6 A) 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 (Balco and Rovey, 2010; Bierman et al., 1999; Ridge, 2004; Stanford, 1993). 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.

2.6.2. Implications for Shelf Pore Water Salinity

Pleistocene ice sheets have a strong control on the emplacement of subsurface freshwater onshore and offshore (Bense and Person, 2008; Cohen et al., 2010; Person et al., 2007a). 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 (Marksamer et al., 2007; Person et al., 2003). 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. Sub-surface freshwater emplacement from ice sheets into permeable units within sedimentary basins and continental shelves can extend 50-100 km beyond the ice sheet margin (Bense and Person, 2008; Marksamer et al., 2007; McIntosh et al., 2011) and to depths of over 1km (Lemieux et al., 2008c).
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).

2.7. 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.
Chapter 3

Glacially Generated Overpressure On the New England Continental Shelf: Integration of Full-Waveform Inversion and Overpressure Modeling

Localized zones of high-amplitude, discontinuous seismic reflections 100 km off the coast of Massachusetts, USA, have P-wave velocities up to 190 m/s lower than those of adjacent sediments of equal depth (250 m below the sea floor). To investigate the origin of these low velocity zones, we compare the detailed velocity structure across high-amplitude regions to adjacent, undisturbed regions through full-waveform inversion. We relate the full-waveform inversion velocities to effective stress and overpressure with a power-law model. This model predicts localized overpressures up to 2.2 MPa associated with the high-amplitude reflections. To help understand the overpressure source, we model overpressure
due to erosion, glacial loading, and sedimentation in one dimension. The modeling results show that ice loading from a late Pleistocene glaciation, ice loading from the Last Glacial Maximum, and rapid sedimentation contributed to the overpressure. Localized overpressure, however, is likely the result of focused fluid flow though a high-permeability layer below the region characterized by the high-amplitude reflections. These high overpressures may have also caused localized sediment deformation. Our forward models predict maximum overpressure during the Last Glacial Maximum due to loading by glaciers and rapid sedimentation, but these overpressures are dissipating in the modern, low-sedimentation-rate environment. This has important implications for our understanding of the continental shelf morphology, fluid flow, and submarine groundwater discharge off Massachusetts, as we show a mechanism related to Pleistocene ice sheets that may be creating regions of anomalously high overpressure.

3.1. Introduction

Continental shelf sediments often contain large regions with pore pressure above hydrostatic pressure (overpressure). These overpressures are often formed by a combination of rapid sedimentation, fluid expansion and migration, and/or phase changes (Bredehoeft and Hanshaw, 1968; Bethke and Corbet, 1998; Hart et al., 1995; Gordon and Flemings, 1998; Swarbrick and Osborne, 1998; Dugan et al., 2003). Understanding and locating overpressure is important as it can affect
subsurface fluid flow, submarine groundwater discharge (SGD), compaction and deformation of sediments, slope stability, offshore drilling strategies, fractures formation, and fluid migration pathways (e.g., Roberts and Nunn, 1995; Gordon and Flemings, 1998; Dugan and Flemings, 2002; Dugan and Sheahan, 2012). Thus, an accurate understanding of overpressure is central to our understanding of offshore fluid-flow regimes and their influence on some continental-shelf processes.

Wells drilled on the New England continental shelf, beneath Nantucket Island, documented unexpected overpressure (7 to 8 m above average sea level) within aquifers that are several hundred meters below sea level (Figure 3-1). This overpressure cannot be produced by the low, local fluvial sedimentation rates and sediment properties in the region (Marksamer et al., 2007). The overpressure may be a relic of Pleistocene sea-level rise and fall and ice-sheet loading (Person et al., 2007a). The continental margin off Massachusetts, USA, was repeatedly glaciated throughout the late Pleistocene (Oldale and O’Hare, 1984; Uchupi et al., 2001; Siegel et al., 2012). Two- and three-dimensional paleohydrologic models simulate how these glacial cycles, in combination with sea-level fluctuations, had a strong influence on the offshore hydrogeologic system (Person et al., 2003; Marksamer et al., 2007; Cohen et al., 2010). In particular, the models show that these glaciation cycles may have emplaced large volumes of subsurface freshwater tens of kilometers beyond the edge of the ice sheets (Person et al., 2003; Marksamer et al., 2007; Cohen et al., 2010). A glacial influence on this hydrogeologic system may also explain low-salinity groundwater observed on Nantucket Island (Folger et al., 1978).
in addition to the observed overpressure. However, there are currently no direct observations that verify the existence or constrain the extent of the overpressure due to glaciation on the continental shelf predicted by numerical modeling.

A high-resolution, multi-channel seismic dataset collected on the continental shelf off Massachusetts, USA shows several regions of localized, relatively high-amplitude, disturbed sediments (bright spots) directly above a marine oxygen isotope stage (MIS) 12 glacial-erosion surface (Siegel et al., 2012) (Figures 3-1 and 3-2). Siegel et al. (2012) show that the continental shelf off Massachusetts was glaciated in the late Pleistocene. Marksamer et al. (2007) show that the Last Glacial Maximum (LGM) may have increased overpressure tens of kilometers beyond the maximum extent of the Laurentide ice sheet. We suggest that the bright spots observed in the seismic data could be regions of localized overpressure generated during late Pleistocene glaciations. To evaluate this possibility, we determine the detailed (2-4 m resolution) vertical velocity structure through the seismic bright spots using a full-waveform inversion approach. We then determine overpressure based on the seismic velocities using an empirical relationship between effective stress and P-wave velocity (Bowers, 1995). We compare this velocity-predicted overpressure to overpressure predicted from one-dimensional (1D) forward models that include ice-sheet loading, erosion, and sedimentation. Our comparison shows that the observed changes in velocities in the bright spots can be explained by localized overpressures that developed from focused fluid flow via permeable pathways due to ice sheet loading from late Pleistocene ice sheets or rapid
sedimentation during LGM ice sheet retreat. Our results thus provide observations to support the predictions of modeling studies that suggest a strong influence of glaciation on modern offshore overpressure (Person et al., 2003; Marksamer et al., 2007; Cohen et al., 2010), while also suggesting that stratigraphy and fluid flow play a role.

3.2. Geologic Setting

The continental shelf off Massachusetts is part of a passive margin that formed from rifting during the breakup of Pangea and the opening of the Atlantic during the late Triassic to early Jurassic. After breakup, the margin continued to subside and fill with sediments from the Cretaceous to the present, forming a thick sedimentary wedge overlying Jurassic basement (Hutchinson et al., 1986). The base of the sedimentary wedge consists of a thick package of Cretaceous to Eocene carbonate mud (Steckler et al., 1999). During the Oligocene and Miocene, siliciclastic sedimentation increased nearly 20 fold, and there was a pronounced shift from primarily aggrading sediments to prograding siliciclastic clinoforms (Unit C in Figure 3-2) (Steckler et al., 1999). The Miocene clinoforms likely graded from mud-prone at the base to silt-prone at the top (Siegel et al., 2012), consistent with lithologic trends observed in Oligocene and Miocene clinoforms off New Jersey (Greenlee et al., 1992).
Figure 3-1 A) Regional base map showing location of seismic lines used in this study along with the location of the maximum extent of the last glacial maximum (line labeled LGM) (Uchupi et al., 2001), and the inferred maximum extent of a MIS 12 ice sheet (black line labeled MIS 12) (Siegel et al., 2012). The box shows the location of figure (B). B) Detailed map showing the location of the seismic lines used in this study along with the location of bright spots observed in the lines (marked with dots). Gray lines show the direction of ice flow for the MIS 12 ice stream. All of the bright spots are approximately located within a trough formed by the paleo-ice stream (Siegel et al., 2012). C) Measured head data comparing shallow water table head in Nantucket well 228 (circles) and head within deeper Cretaceous aquifers in USGS well 6001 (squares). The comparison of the data shows that the aquifers are not related, and that the deeper Cretaceous aquifer has anomalously higher pressure, which is believed to be glacial in origin (Marksamer et al., 2007).
The MIS 12 glacial unconformity (unconformity U1 in Figure 3-2) marks the onset of Pleistocene shelf-crossing glaciations off Massachusetts and the location of an ice stream that was sourced from the Gulf of Maine (Siegel et al., 2012). The ice stream eroded a trough, 100 m deep and 50 km wide, into the underlying Oligocene/Miocene sediments. During ice-stream retreat, a 50- to 100-m-thick deposit of glacigenic sediments, which likely consisted of poorly sorted silts and clays, was rapidly deposited pro-glacially (Unit B2 in Figure 3-2). Several subsequent glacial cycles terminated on the shelf throughout the remainder of the late Pleistocene, including the LGM, which terminated near Martha's Vineyard and Nantucket islands (Figure 3-1) (Oldale and O'Hare, 1984; Uchupi et al., 2001). High sediment input during the Pleistocene, combined with sea-level changes, produced a series of prograding clinoforms (Unit B1 in Figure 3-2). During the last sea-level fall (40 – 30 ka), a shallow sequence boundary was formed (unconformity U2 in Figure 3-2) that is overlain by a combination of late Pleistocene and Holocene siliciclastic sediments (Siegel et al., 2012).
Figure 3-2 Top) Seismic sections that contain bright spots and bright dipping reflections (Locations are indicated in Figure 3-1B). Bottom) Same seismic sections plotted in grayscale with major reflections and interpreted sediment history marked. U1 is an MIS 12 glacial unconformity formed by an ice stream. The bright spots overlie the U1 unconformity and are contained within a 50-75 m thick glacigenic sediment unit that was deposited during ice stream retreat (Siegel et al., 2012). Scale bars on the right show relative amplitude for the seismic data.
3.3. Data Description and Observations

Our geophysical dataset consists of a series of high-resolution, multi-channel seismic (MCS) lines collected off Massachusetts (Figure 3-1A and 3-1B) with the Scripps Institution of Oceanography’s portable seismic system. The source was a 45 in$^3$/105 in$^3$ (737.5 cm$^3$/1720.6 cm$^3$) generator-injector (GI) air gun that produces frequencies up to 200 Hz. The streamer consisted of 48 channels with a group spacing of 12.5 m, a near offset of 50 m, and a far offset of 650 m. The data were sorted into common-depth point (CDP) bins spaced every 6.25 m, yielding an average CDP fold of 24 channels.

We observe bright spots and bright dipping reflections in several seismic lines within the glacigenic sediments of Unit B2 directly above the glacial unconformity (U1) at 300 - 350 mbsl (Figures 3-1 and 3-2). Unconformity U1 is the remnant of an ice stream trough (Figure 3-1B). The bright spots are approximately perpendicular to the flow direction of the ice stream, and trend along the strike of clinoforms in the underlying Miocene stratigraphic unit (Unit C in Figure 3-2). The bright spots have a semi-disturbed seismic character, and relatively higher amplitude compared to adjacent sediments. Discernable reflections within the bright spots, however, correlate with adjacent reflections in the glacigenic sediments just outside the bright spots. The dipping reflections within the underlying Miocene strata also have relatively higher amplitudes that are localized to several tens of meters below the U1 glacial unconformity just below the bright spots (bright dipping reflection in Figure 3-2).
We do not believe that the enhanced reflectivity of the bright spots and underlying dipping reflections is due to the presence of either free gas or gas hydrate. An association of the bright spots with gas hydrate can be ruled out, as methane gas hydrate is not stable at these depths (e.g. Xu and Ruppel, 1999; Phrampus and Hornbach et al., 2012). The presence of free gas, perhaps trapped beneath the unconformity, also seems unlikely based on the lack of frequency-dependent amplitude loss below the bright spots, as would be expected if free gas were present (White, 1975; Taylor et al., 2000; Morgan et al., 2012). This is demonstrated in Figure 3-3, where spectral lines and notches are shifted for data beneath the bright spots relative to the rest of the section, but spectral slopes are unchanged. Similarly, interval velocities in the region of the bright spots are lower than the rest of the section (Figure 3-4), which could indicate the presence of free gas, but the magnitude of the velocity anomaly (100 m/s) is less than what would be expected if free gas were present (Domenico, 1976; Ecker et al., 2000). Thus, while we cannot completely rule out the possibility that the observed bright spots are related to a free gas phase, we believe that this is not likely. For the remainder of this paper we explore an alternative explanation for the origin of the bright spots that is consistent with their distribution and the history of shelf glaciation.
Figure 3-3 A) Seismic section with blue boxes indicating cdp/time ranges over which spectra in B, C, and D are calculated. B) Colored curves: average spectra each for 200 cdps centered on the colored diamond locations in (A) for the time range indicated by the lower blue box in (A); for example, position 1000 (plotted with a thicker curve) is the average of spectra for cdps 900-1100 for the time interval 0.425-0.750s; black curves are average spectra for 200 cdps centered on the diamond locations within the upper box in (A), 0-0.325s. C) Spectra for every cdp calculated for the time range 0-0.325s (the upper box in A). D) Spectra for every cdp calculated for the time range 0.425-0.750s (the lower box in A). All spectra are plotted as decibells down from the maximum power in (C). The effect of a lateral change in anelastic attenuation would appear as a change in spectral slope with frequency, with a steeper slope for higher attenuation regions. This effect is not observed.
3.4. Full-Waveform Inversion

Full-waveform inversion is a modeling tool that estimates detailed velocity structure by predicting the full pre-stack seismic wavefield (Minshull et al., 1994; Sain et al., 2000; Virieux and Operto, 2009). We analyze our dataset using the linearized waveform inversion approach of Kormendi and Dietrich (1991). This type of analysis has been extensively described in the literature, and our particular approach closely follows that of Korenaga et al. (1997), and so we have placed details of our analysis in Appendix B.

![Figure 3-4 A) Smooth interval velocities for all CDPs interpreted across the part of seismic line 1 containing bright spots and the average interval velocity. B) Color map plot of the interval velocities averaged every 50th CDP. This is the same section of seismic line 1 shown in Figure 3-2 (locations is indicated in Figure 3-1B). Most interval velocities are similar; however, there is a noticeable reduction in the velocities between CDPs 800-1100, particularly CDP 1000. The low velocity region is approximately co-located with the location of the bright spots. The interval velocities are the starting model for the full-waveform inversion.](image)
3.4.1. Starting Model

The full-waveform inversion algorithm solves for best-fit models \((m)\) that are linearly close to the starting model \((m_0)\), i.e. solving for short-wavelength velocity perturbations. An accurate inversion result thus requires a starting model that captures the long- to intermediate-wavelength subsurface velocity structure such that reflection travel times are accurately predicted. We use a weighted, damped, least-squares inversion to solve for a smooth starting velocity model based on picked normal moveout (NMO) staking velocities in the seismic data (Lizarralde and Swift, 1999). A smooth inversion seeks to minimize the change in velocity with depth and it creates a model that varies slowly and penalizes rapid change in velocity. NMO velocities were picked every 125 m on prominent horizons that are correlated through the seismic section and are spaced an average 100 ms is time. Three to five adjacent, smoothed interval velocity profiles were averaged and used as the starting model for in the inversion process (Figure 3-4).

3.4.2. Sensitivity Tests

We want to use differences between adjacent velocity profiles, determined from full-waveform inversion, to make inferences on overpressure. It is thus important to have estimates of the uncertainties in these modeled velocity functions. We estimate model uncertainties by testing sensitivity to two main sources of error in the inversion: failure of the 1D assumption (accuracy) and
robustness (precision), which is a function of the inversion's sensitivity to small deviations in the data, potentially due to noise.

We assess the effect of dip on accuracy by comparing full-waveform inversion results from co-located CDPs at the intersection of the perpendicular Lines 1 and 10 (Figure 3-1). Reflection dips range between 0° and 1°, toward the south-southeast, in the depth interval of interest (0-500 mbsl), with a marked increase in dip below the glacial unconformity (U1). The velocities obtained from the full-waveform inversions display a similar pattern, particularly above the U1 unconformity (Figure 3-5A), where both the amplitude and the phase of the velocity/depth functions agree well. Dip introduces both a velocity error and a depth error into the velocity estimate of a particular stratal unit. A simple difference of the two velocity functions thus overestimates the velocity difference of individual units by including both the velocity and depth error due to dip via the misalignment of the peaks and troughs in the velocity profiles. To separate these two effects of dip, we cross correlate the two velocity/depth profiles, applying an 80-m-long, tapered window centered at each depth point, and align each point of the two velocity/depth profiles based on the maximum cross-correlation values so that velocity differences between the same stratal units can be compared. The cross-correlation lags, or depth shift required to maximize cross-correlation at each depth, are thus a measure depth uncertainty as a function of depth (Figure 3-5B), and the difference in inverted velocity between the aligned profiles (Figure 3-5C) is a measure of velocity uncertainty as a function of depth. Above the unconformity, where
reflection dips are low, depth differences between the two profiles are 1 m or less and velocity differences average 15.6 m/s with a maximum difference of 49.0 m/s (2.5%). Below the unconformity, where reflection dip is highest (0.5°-1.0°), the average velocity difference increases to 26.3 m/s, with a maximum difference of 74.5 m/s (3.7%), and depth differences increase to 4-6 m. Despite this deviation, the apparent pattern of velocity variation is similar on both lines. In addition, the bright spots are located above the glacial unconformity were dip is lower, and subsequently less error.

Figure 3-5 A) Full-waveform inversion velocity results for co-located CDPs on crossing lines to test the effects of dip with our assumption of 1D velocity inversion. Line 1 is shown in gray, and line 10 is shown in black. B) Cross-correlation lag with depth based on an 80-m-window. C) The difference in velocity between the two full-waveform inversion results when the velocity profiles are shifted by their cross-correlation lag. The difference is largest just bellow the unconformity (U1) where there is the steepest dip in reflections (Figure 3-2).
We concentrate on the result of the full-waveform inversion through two contrasting regions of the seismic section. CDP 990 traverses the center of the region containing the bright spots, whereas CDP 1400 traverses a section 2.5 km to the north-northeast that does not contain bright spots yet contains similar stratigraphy (Figure 3-2 and 3-6). For both CDP locations, the full-waveform inversion was run on three adjacent CDPs to help quantify the precision of the final velocity model (Figure 3-6). The predicted velocities of adjacent CDPs are similar, with a standard error (standard deviation of the mean) between 0.0 and 32.5 m/s for CDP 990 (average 5.2 m/s) and 0.0 and 19.1 m/s for CDP 1400 (average 6.7 m/s). The error tends to increase below the bright reflection of the glacial unconformity (U1). This is due to higher dips.

The synthetic waveform traces generated with the final 1D stratified Earth model closely match the observed seismic data for the range in P-values used in the inversion (Figure 3-7). The waveform fit is less good at large p-values for the U1 reflection for CDP 990. This may be due to the effect of dip, too coarse of a grid spacing, or an inappropriate parameterization of Vs and density, since the dependence of impedance on changes in Vp, Vs and density becomes more nonlinear with increasing p. However, the fit at all p’s is good over the time interval corresponding to the bright spots, which display a clear increase in amplitude with increasing p for CDP 900 that is well fit by the model (Figure 3-7).
Figure 3-6 A) Full-waveform inversion results for two CDP groups. CDP 1400 does not contain any bright spots, whereas CDP 990 goes through the middle of the bright spot region. Adjacent CDPs within the group all show similar velocity results. The average of the CDPs is shown in bold with the standard error represented by the thin black lines. B-C) Standard error (standard deviation of the mean) of the two CDP groups.
Figure 3-7 Results of full-waveform inversion for seismic line 1 CDPs 990 and 1400 in the $\tau$-p domain showing the recorded seismic data (solid black line) and synthetic seismic data based on the final 1D stratified earth model (dashed red line). For lower p-values, there is an excellent fit between the synthetic and the observed waveforms. The waveforms shown correspond to the full range of time and p-values used for the inversion.
3.4.3. Results

The final velocity model for the CDP group is defined as the average of the three adjacent CDPs (Figure 3-6). The final velocity models show similar patterns that correlate with the observed stratigraphy (Figure 3-8). Sediments just below the seafloor and the O1 reflection have lower velocities, whereas the U1 reflection corresponds to a large velocity increase in both profiles. This indicates that sediment properties are consistent across the seismic section and produce correspondingly similar velocity patterns. The velocity of the bright spots, however, are 190 ± 80 m/s lower than adjacent sediments at the same depth without bright spots (Figures 3-6 and 3-8). This observed velocity difference is larger than the velocity differences of the starting models, the estimated uncertainties due to dip and the inherent precision of the approach. We note that the velocity reduction of 190 m/s (~10%) in the region of the bright spots is small relative to velocity reductions predicted due to even a fraction of a percent of free gas in sediments at these depths (Domenico, 1976). We thus explore overpressure as a mechanism for velocity reduction in the vicinity of the bright spots.
3.5. Velocity-Predicted Overpressure

The velocity of a sedimentary rock in the subsurface is a function of many factors, including its depositional and burial history, lithology, texture, porosity, consolidation state, density, temperature, pore fluid type, and overpressure (Bowers, 1995; Dutta, 2002; Dvorkin and Nur, 2002; Huffman, 2002). In considering the cause of the abrupt lateral change in seismic velocity exhibited by the bright spots, abrupt changes in most of these factors can be discounted. The bright spots occur within a coherent unit characterized by uniformly low impedance everywhere.
except for the isolated, localized bright spots. This indicates lithology is consistent across the unit. Moreover, the coherent unit is inferred to be backfilling, glacifluvial sediments (Siegel et al., 2012), which were shown to have consistent, similar lithology with well data (Kristensen et al., 2008). The velocity could indicate porosity change. Empirical relationships that relate velocity to porosity (e.g. Marion et al., 1992; Erickson and Jarrard, 1998) predict a change in porosity of 0.05 (20% change in porosity), assuming an average porosity of 0.25 in the unit. However, this difference in porosity would not be expected given the sediment type and depositional history is consistent across the seismic section (Hart et al., 1995).

The bright spots are situated above an unconformity that truncates dipping reflectors whose amplitudes are themselves anomalously bright in regions beneath the bright spots. This relationship strongly suggests the influence of overpressure on effective stress. We therefore assume that at a constant depth, observed velocity variations are most likely due to changes in effective stress, which are related to changes in overpressure and the bulk moduli of the materials (Carcione et al., 2003). Dutta (2002) and Bowers (2002) showed that, at a given depth, deviations in observed seismic velocity from that predicted from a compaction trend could be used as an indicator of overpressure.

There are several empirically derived methods that can be used to estimate overpressure from seismic velocity by quantifying the deviation of the observed seismic velocities from a normal velocity trend (Bowers, 1995; Sayers et al., 2002;
Mavko et al., 2009). We use the method of Bowers (1995), which relates effective vertical stress and velocity as,

\[ \sigma' = \left( \frac{V - V_0}{A} \right)^{\frac{1}{C}} \]  

(Equation 3-1)

where \( \sigma' \) is effective vertical stress \((S_v - P_{pore})\), \( V \) is the velocity at depth, and \( V_0 \) is the velocity of unconsolidated, fluid-saturated sediment (1500 m/s). \( S_v \) is the lithostatic stress or total vertical stress, \( P_{pore} \) is the pore pressure, and \( A \) and \( C \) are empirically derived constants that are unique to individual regions. We determine appropriate values for \( A \) and \( C \) via a grid search that seeks values which minimize the difference between the velocity trend predicted by equation (3-1) and the observed trend of average interval velocity (Figure 3-4). We use a range of possible effective stress trends appropriate for this location ranging from the minimum effective stress (pore pressure near-lithostatic) to maximum effective stress (pore pressure near-hydrostatic). For near-lithostatic pore pressure, \( A = 7.55 \) and \( C = 0.29 \); for near-hydrostatic pore pressure, \( A = 0.7 \) and \( C = 0.44 \).

We use equation (3-1) to estimate differences in the effective stress across the seismic section given an observed difference in velocity within the bright spots compared to other regions that are assumed to follow an interpreted normal velocity trend. From the interpreted change in effective stress we predict the corresponding change in pore pressure, which is equivalent in magnitude to the change in overpressure. To express the change in pore pressure \((dp)\) given an
observed change in velocity \((dv)\), we differentiate equation (3-1). The derivative of effective stress with respect to velocity is,

\[
\frac{d\sigma}{dv} = \frac{d}{dv}(S_v - P_{pore}).
\]  
(Equation 3-2)

Lithostatic stress at any given depth is constant and its derivative with respect to velocity is 0. Therefore at a constant depth the change in effective stress with velocity relates to the change in pore pressure,

\[
\frac{d\sigma}{dv} = -\frac{dP_{pore}}{dv}.
\]  
(Equation 3-3)

The right hand side of equation (3-1) can be differentiated with respect to velocity and expressed as:

\[
d\left(\frac{v - v_0}{A}\right)^\alpha = \left(\frac{v - v_0}{A}\right)^\alpha \frac{(v - v_0)^{\alpha-1}}{AC}.
\]  
(Equation 3-4)

The empirical estimate of change in pore pressure with respect to velocity is then:

\[
dP_{pore} = -dv \frac{(v - v_0)^{\alpha-1}}{AC}.
\]  
(Equation 3-5)

where \(dv\) is the change in velocity observed between CDPs at the location of the bright spots.
3.5.1. Results

We infer that the decrease in velocity of 190 ± 80 m/s in the bright spots is due primarily to higher overpressure. Using equation (3-5), the change in velocity can be equated to a difference in overpressure of 0.9 – 2.2 MPa, given the estimated uncertainty in velocity at this depth and the range of scalar parameters A and C. Although there is a considerable range in the estimated overpressure differences, due primarily to uncertainties in A and C, even the minimum estimate represents a significant difference in highly localized overpressure.

Localized zones of relatively high overpressure could develop from a number of mechanisms, including (1) in situ fluid expansion; (2) rapid sedimentation; or (3) lateral pore pressure transfer (Swarbrick and Osborne, 1998; Bowers, 2002; Dugan and Sheahan, 2012). Our study region has been repeatedly glaciated throughout the late Pleistocene, which can lead to sustained overpressures long after glacial retreat (Bense and Person, 2008). Given the glacial history in the region, the location of the bright spots just above the MIS 12 glacial unconformity, and their trend along strike of the underlying sedimentary unit, we suggest that the late Pleistocene ice sheets where the likely source of overpressure and explore this possibility with a simple 1D fluid flow model.
3.6. 1D Fluid Flow Model

To investigate physical mechanisms that could explain the localized overpressure within the bright spots, we employ a 1D, finite-difference, fluid flow model that solves the groundwater flow equation accounting for glacial loading, erosion, and sedimentation. We use a layered, 1D model to represent an idealized transect through the seismic section (Figure 3-2) that contains corresponding lithologic units and erosion surfaces to 1000 mbsl. The model space is discretized with an evolving grid that contains 1 m nodes.

3.6.1. Fluid Flow

We solve a 1D equation for groundwater flow similar to ones solved by Person et al. (2007) and Marksamer et al. (2007) that incorporates ice-sheet loading, erosion, and sedimentation,

\[
K_v \frac{\partial^2 h}{\partial z^2} = S_s \left[ \frac{\partial h}{\partial t} - \frac{\partial}{\partial t} \left( \frac{\rho_i}{\rho_f} \frac{\partial \eta}{\partial t} \right) - \frac{\rho_s - \rho_w}{\rho_f} \frac{\partial L}{\partial t} + P_{ext} \right],
\]

(Equation 3-6)

where \( K_v \) is the hydraulic conductivity in the vertical direction, \( S_s \) is the specific storage, \( \rho_i, \rho_w, \) and \( \rho_b \) are the densities of ice, water, and fluid-saturated sediment, \( \eta \) and \( L \) are the elevations of the ice and sediment relative to sea-level, respectively, and \( P_{ext} \) is an external source term that can be assigned to a particular layer to represent external fluid sources outside of the model domain (Dugan and Germaine, 2008). The coefficient \( \zeta \) in equation 3-6 is the 1D loading efficiency, which defines
the proportion of surface loading that is transferred to subsurface overpressure (Ingebritsen et al., 2007; Lemieux et al., 2008a). This is similar to the three-dimensional pore pressure buildup coefficient (B) defined by Green and Wang (1986) as the change in pore pressure per unit change in applied stress. The exact value of $\zeta$ depends on material properties such as bulk sediment and fluid compressibilities, which we approximate to estimate the general trend of $\zeta$ during loading and unloading. During sediment and ice loading, when the bulk compressibility is orders of magnitude larger than fluid compressibility, $\zeta$ is generally close to a value of 1; thus, during loading in our model, $\zeta$ is set to 0.95. This means much of the increase in overburden leads to a proportional increase in pore pressure. During sediment and ice-sheet unloading, we assume compressibility is 10% of the initial compressibility (Corbet and Bethke, 1992; Stigall and Dugan, 2010), which results in a decrease in $\zeta$; thus, during unloading in our model, $\zeta$ is set to 0.50. This allows glacial overpressure to persist after the ice sheet has retreated.

We solve equation (3-6) using a fully implicit finite-difference approach (Fletcher, 1997) (see Appendix C for model code). The base of the model is a no-flow boundary, and the top boundary is set to be equivalent to either sea-level or the head at the base of the ice sheet when the ice sheet is present; this is consistent with basin-scale models that account for ice-sheet loading (Person et al., 2007a). Many studies have modeled additional effects of ice sheets on fluid flow such as flexure of the crust and the formation of permafrost (Person et al., 2007a; Bense and Person, 2008; Lemieux et al., 2008b); because we are only concerned with a 1D model, and
our study region is near the edge of the ice sheet, we do not account for flexure. Permafrost can reduce the permeability beneath and beyond the extent of the ice sheet. However, because our study region is in a paleo-ice stream trough that exhibited signs of basal melting via sub-glacial drainage channels (Siegel et al., 2012), permafrost would not likely develop.

Hydrostatic pressure ($P_{\text{hydro}}$) is always assumed relative to sea level (which is constant in the model) and defined as:

$$P_{\text{hydro}} = \rho_w g z,$$

(Equation 3-7)

where $g$ is the acceleration due to gravity, and $z$ is the depth below sea level.

Lithostatic stress ($S_v$) is the stress from the overlying water, saturated sediment matrix, and ice sheet when present; lithostatic stress is defined as:

$$S_v = \int_0^z \rho_s g dz,$$

(Equation 3-8)

3.6.2. Model Domain

We simulate four distinct geologic stages consistent with the sedimentation and glacial history determined by Siegel et al. (2012) (Figure 3-9): (1) a slow sedimentation rate during the Oligocene and Miocene; (2) glacial erosion and loading by the MIS 12 ice stream that increases in height to 500 m over a 20 kyr time period; (3) deglaciation and rapid sedimentation rate at the end of MIS 12 during a 10 kyr time period; and (4) moderate sedimentation rate during the late
Pleistocene. The 1D model contains four homogeneous layers to represent this geologic history (Figure 3-9). Layer 1 is an Oligocene/Miocene layer. Layer 2 is a 10-m-thick, high-permeability layer below the U1 unconformity that represents a dipping clinoform that was eroded by the ice stream and continues down dip outside of the model space. Layer 3 is a glacigenic sediment layer. Layer 4 is a Pleistocene/Holocene layer.

![Figure 3-9](image)

**Figure 3-9** Conceptual drawing showing the stratigraphic evolution of the seismic sections that is used for the 1D numerical model (A), (B) 1D model space that is based on observed stratigraphy. Layer 1 in model extends to a depth of 1000 mbsl.

The petrophysical properties we assign to each model layer are based on a combination of available well data and the properties of similar hydrogeologic units used in published hydrogeologic-modeling studies of this area. On the New England continental shelf, fine-grained sediments range in permeability from $10^{-10}$ to $10^{-18}$ m$^2$ (Cohen et al., 2010) whereas clayey silts have an average permeability of $10^{-16.5}$ m$^2$ (Person et al., 2003). Glacigenic sediments such as till can have permeability of
10^{-16} \text{ m}^2 \text{ or lower} \ (\text{Keller et al., 1989}). \ Glacio-lacustrine \ deposits \ on \ the \ island \ of \ Nantucket \ were \ found \ to \ have \ a \ permeability \ of \ 10^{-15} \text{ m}^2 \ (\text{Person et al., 2012}). \ Within \ the \ range \ of \ these \ possible \ permeability \ values, \ we \ assign \ permeability \ to \ each \ model \ layer \ that \ best \ produces \ the \ general \ trend \ of \ overpressure \ we \ estimate \ from \ the \ velocity \ model. \ The \ permeability \ values \ we \ use \ for \ the \ Miocene \ layer, \ the \ clinoform \ layer, \ the \ glacigenic \ layer, \ and \ the \ Pleistocene \ layer \ are \ 10^{-18}, \ 10^{-17}, \ 10^{-18}, \ and \ 10^{-16} \text{ m}^2, \ respectively. \ A \ specific \ storage \ of \ 5.0 \times 10^{-3} \text{ m}^{-1} \ was \ assigned \ to \ all \ layers.

3.6.3. Results

Based \ on \ the \ seismic \ velocities, \ we \ infer \ localized, \ higher \ overpressure \ of \ up \ to \ 2.2 \text{ MPa} \ in \ the \ region \ of \ the \ bright \ spots \ compared \ to \ adjacent \ regions \ of \ the \ seismic \ section \ at \ the \ same \ depth \ (250 \text{ mbsf}), \ and \ to \ underlying \ and \ overlying \ sediment \ layers. \ A \ series \ of \ numerical \ models \ were \ run \ to \ evaluate \ mechanisms \ that \ could \ explain \ the \ localized \ overpressure \ inferred \ within \ the \ bright \ spots, \ which \ are \ located \ within \ the \ glacigenic \ sediment \ layer. \ Glacial \ loading \ by \ an \ ice \ sheet \ during \ MIS \ 12, \ rapid \ sedimentation \ during \ glacial \ retreat, \ and \ moderate \ sedimentation \ during \ the \ Pleistocene \ all \ contribute \ to \ overpressure \ throughout \ the \ 1D \ model \ space \ (Figure 3-10). \ The \ trend \ and \ magnitude \ of \ the \ overpressure \ is \ affected \ by \ the \ permeability \ architecture, \ the \ assumed \ thickness \ and \ duration \ of \ the \ MIS \ 12 \ ice \ sheet, \ and \ the \ time \ scale \ of \ glacigenic \ sedimentation \ during \ ice \ sheet \ retreat, \ but, \ in \ a \ 1D \ model \ space, \ these \ variables \ cannot \ adequately \ create \ a \ localized \ zone \ of \ relatively \ higher \ overpressure \ within \ the \ glacigenic \ sediments.
compared to underlying or overlying units as predicted by the seismic velocities. Thus, an additional mechanism beyond 1D ice sheet and sediment loading is needed to generate localized, higher overpressure within the glacigenic sediment layer.

![Figure 3-10](image)

**Figure 3-10** Results of the 1D numerical modeling shows the general trend in overpressure (solid black line) that is produced from ice sheet loading and sedimentation. The overpressure that is produced by lateral pore pressure transfer (solid gray line) shows the increase in overpressure above and below the glacial unconformity that would result.

### 3.7. Discussion

#### 3.7.1. Sources of Localized Overpressure

Additional sources of overpressure from outside the 1D model domain may lead to localized regions of overpressure by the lateral migration of fluid through
permeable pathways. This is commonly referred to as flow focusing (Yardley and Swarbrick, 2000; Flemings et al., 2002; Dugan and Flemings, 2002; Dugan and Sheahan, 2012). Based on the orientation of the bright spots along strike of the dipping Miocene layers, we infer they are all related to the same clinoform, which could have high permeability (light gray clinoform within the Oligocene/Miocene unit, Figure 3-11). That particular high-permeability clinoform could facilitate fluid migration from down dip where overpressure is higher. We consider two mechanisms that may be sources of overpressure outside our 1D model domain: (1) overpressure generated on the continental shelf from LGM ice-sheet loading and rapid sedimentation near the continental slope (Figure 3-11A) (Marksamer et al., 2007); and (2) overpressure generated shoreward, where ice overburden and subsequently overpressure buildup were greater, which could migrate and increase overpressure near the ice margin where there was less ice overburden (Figure 3-11B) (Boulton et al., 1993; Boulton and Caban, 1995; Bense and Person, 2008; Iverson and Person, 2012).
To evaluate the role of an external overpressure source linked with a high-permeability clinoform (Figure 3-11), we include a pressure source term ($P_{\text{ext}}$ in Equation 3-6) in the high-permeability unit just below the unconformity (Layer 2 in Figure 3-9B). The external pressure source is specified to be 1.5 MPa for a period of 10 kyr, a similar magnitude of overpressure to that predicted near the shelf edge by Marksamer et al. (2007) due to higher sedimentation rates near the shelf edge during LGM glacial retreat. When the additional source term is assigned to the clinoform layer during the LGM, localized (higher than the overlying or underlying layer) overpressure develops in the glacigenic sediments and the clinoform layer, centered on the U1 unconformity. The model-predicted overpressure is approximately where the full-waveform inversion velocities predict the largest region of localized overpressure (i.e., in the bright spots, which are located within the glacigenic sediments). Thus, the assumption of an additional overpressure
source within the clinoform layer in the 1D fluid flow model is the best match to the velocity-predicted overpressure. High overpressure in the underlying clinoform could also be creating the large impedance contrasts in the bright dipping layers (Figure 3-2).

Alternatively, the bright spots may be the result of fluid expulsion from underlying overpressured layers generated by the MIS 12 ice sheet as it retreated and glacigenic sediments were rapidly deposited (Figure 3-11B). Deformation is more likely near the ice sheet margin, where there is typically the highest gradient in subsurface overpressure from ice-sheet loading due to the abrupt changes in ice sheet thickness (Iverson and Person, 2012). Boulton and Caban (1995) show that overpressure generated beyond the edge of an ice sheet can lead to flow focusing in regions where there is a break in a confining unit or permafrost; they noted that these geologic conditions can create fluid expulsion features, referred to as extrusion moraines, and can lead to a release of glacially generated overpressure into the overlying strata. Roberts and Nunn (1995) also show that overpressure can build to a point where fractures occur and fluid is rapidly expelled into overlying sediments. The bright spots may be the remnants of an expulsion feature created by this overpressure where flow was facilitated through a permeable clinoform (Figure 3-11B). This mechanism could also explain the semi-disturbed seismic character of the bright spots.

These results demonstrate that flow focusing from late-Pleistocene, glacially generated overpressure is a possible mechanism for the localized overpressure and
possible sediment deformation features observed in our seismic data. Overpressures encountered in deep aquifers (500 mbsl) on Nantucket are also interpreted to be related to overpressures generated during the LGM via lateral pressure transfer (Marksamer et al., 2007). Together these results highlight that regional sedimentation rate, stratigraphy, and glacial history are all important factors for understanding the occurrence and distribution of overpressure along this previously glaciated margin.

3.7.2. Implications for Submarine Groundwater Discharge

Large submarine groundwater discharge (SGD) fluxes have been inferred off New England from the $^{228}$Ra inventory in the upper ocean (Moore et al., 2008; Moore, 2010). The overpressures predicted by our full-waveform inversion and numerical modeling could be the mechanism driving fluid flow linked to this SGD. Overpressure is a primary driving force for offshore groundwater flow (Dugan and Sheahan, 2012) and thus has a strong control on SGD. Understanding and characterizing SGD is important as it is an integral part of the total nutrient flux to the oceans, which effects marine biological production (Johannes, 1980; Church, 1996; Slomp and Cappellen, 2004). Biological production, in turn, is a significant carbon sink in the global exogenic carbon cycle, which is generally limited by the nutrient supply (Falkowski et al., 2000).

Although we have demonstrated a potential overpressure source that could drive SGD, the magnitude of overpressure is transient due to the cycling of late
Pleistocene ice sheets. Therefore, rates of SGD that may be linked to glacial overpressure likely differ between glacial and interglacial periods.

3.8. Conclusions

A low velocity zone in smoothed interval velocities, further resolved by a full-waveform inversion, shows a distinct, localized region of low velocity that corresponds to relatively high-amplitude, disturbed reflections (bright spots) in a seismic section off Massachusetts, USA. The bright spots occur in glacigenic sediments overlying a MIS 12 glacial erosion surface (U1). By inferring effective stress from seismic velocity anomalies, we interpret a reduced velocity within the bright spots to indicate regions of localized overpressure. The results of our 1D numerical modeling show that glacial loading and sedimentation produce overpressure, but these mechanisms alone will not produce localized overpressure around the U1 unconformity surface. By including additional pressures sources in our 1D model, we showed that a plausible source for localized overpressure may be the migration of overpressures through a high-permeability clinoform directly below the bright spots. Overpressure in an underlying layer may have also caused fluid expulsion and sediment deformation in the region of the bright spots. The overpressure we observed is related to a series of late Pleistocene glaciations that generated overpressure. This overpressure may contribute to SGD in the region. The connection between SGD and glaciations suggests a periodicity to SGD and the corresponding SGD-derived nutrient supply to the oceans.
Influence of Late Pleistocene Glaciations on the Hydrogeology of the Continental Shelf Offshore Massachusetts, USA

Multiple late Pleistocene glaciations that extended onto the continental shelf offshore Massachusetts, USA may have emplaced as much as 100 km$^3$ of freshwater (salinity less than 5 ppt) in continental shelf sediments. To estimate the volume and extent of offshore freshwater, we developed a three-dimensional, variable-density model that couples fluid flow and heat and solute transport for the continental shelf offshore Massachusetts. The stratigraphy for our model is based on high-resolution, multi-channel seismic data. The model incorporates the last 3 Ma of climate history by prescribing boundary conditions of sea-level change and ice sheet extent and thickness. We incorporate new estimates of the maximum extent of a late
Pleistocene ice sheet to near the shelf-slope break. Model results indicate that this late Pleistocene ice sheet was responsible for much of the emplaced freshwater. We predict that the current freshwater distribution reaches depths of up to 500 m below sea level and up to 30 km beyond Martha’s Vineyard. The freshwater distribution is strongly dependent on the three-dimensional stratigraphy and ice-sheet history. Our predictions provide a robust understanding of the distribution of offshore freshwater, which is a potential non-renewable resource for coastal communities along recently glaciated margins.

4.1. Introduction

Freshwater resources are important for agriculture, industry, and personal use; however freshwater volumes, particularly groundwater, are declining (Barlow, 2003). In coastal regions, freshwater resources are particularly vulnerable to declining when climate change issues are considered (Ferguson and Gleeson, 2012). Investigations of freshwater in deep sedimentary basins throughout northern North America show that much of the subsurface freshwater was emplaced during Pleistocene glaciations and is not in equilibrium with current meteoric recharge (Person et al., 2007a; Lemieux et al., 2008a; McIntosh et al., 2012; Neuzil, 2012; Person et al., 2012). These basins were all beneath or near the edge of the Laurentide ice sheet, and enhanced freshwater emplacement associated with ice sheets may explain the origin of the deep freshwater (McIntosh and Walter, 2005; Bense and Person, 2008; McIntosh et al., 2011).
Evidence of glacially emplaced freshwater into the basins comes from many sources. Age data from carbon-14 and noble gases reveal a Pleistocene age for much of the freshwater (Morrisey et al., 2010; Schlegel et al., 2011), and oxygen isotope data reveal isotopically light freshwater which is interpreted as fluid of glacial origin (Vaikmae et al., 2001; McIntosh et al., 2012). In addition, numerical models predict large volumes of sub-glacial meltwater were driven into basin sediments during Pleistocene glaciations (Person et al., 2007a; Person et al., 2012).

Sedimentary basins on the northern US Atlantic continental shelf experienced glaciations in the late Pleistocene in combination with sea-level change throughout the entire Pleistocene (Oldale and O’Hara, 1984; Uchupi et al., 2001; Siegel et al., 2012). Sea-level change is another mechanism that can drive freshwater emplacement (Meisler et al., 1984; Kooi et al., 2000). A combination of climate-driven sea-level change and glaciations has been implicated as the source of freshwater as far as 100 km offshore New Jersey that reaches depths of several hundred meters below the sea floor (mbsf) (Hathaway et al., 1979; Kohout et al., 1988; Cohen et al., 2010). On Nantucket Island, offshore Massachusetts, freshwater extends to depths greater than 500 meters below sea level (mbsl) (Figure 4-1) (Kohout et al., 1977; Folger et al., 1978). The extent of this freshwater offshore Massachusetts is unknown.

Several two-dimensional and three-dimensional numerical modeling studies have predicted freshwater distributions for the continental shelf offshore New England. These models incorporated sea-level change and ice sheets as boundary
conditions in an effort to explain salinity patterns observed below Nantucket Island, and to predict the volume and distribution of offshore freshwater (Figure 4-1C) (Person et al., 2003; Marksamer et al., 2007; Cohen et al., 2010). These studies were based on limited well data and low-resolution seismic reflection data. Recent high-resolution, multi-channel seismic reflection data offshore Massachusetts provide new insights into the continental shelf stratigraphy as well as the seaward extent of glacial deposits. We use these new stratigraphic and glacial interpretations as inputs to a three-dimensional, finite-element, variable-density numerical model that couples fluid flow and heat and solute transport to better estimate freshwater volume beneath the Massachusetts continental shelf. This is the first time it is possible to represent the true three-dimensional stratigraphy offshore Massachusetts based on direct seismic observations. Therefore, we have enhanced capability to understand the impacts of three-dimensional stratigraphy, sea-level cycles, and transient ice-sheet loading on the distribution and the volumes of offshore freshwater. We show that the previously unrecognized extent of a late Pleistocene ice sheet advance resulted in as much as 100 km$^3$ of emplaced freshwater as far as 30 km offshore. The distribution of the freshwater, however, is strongly dependent on the three-dimensional stratigraphy in the continental shelf sediments.
4.2. Geologic Background

4.2.1. Pleistocene Climate

North American glaciations began at the end of the Pliocene, about 2.7 Ma B.P. (Bintanja and van de Wal, 2008; Balco and Rovey, 2010). In the late Pliocene and early Pleistocene, 2.7 – 1.25 Ma B.P., sea-level change had an average amplitude
less than 80 m and a period of 41 ka (Ruddiman et al., 1986; Williams et al., 1988; Huybers, 2007). In the middle Pleistocene, 1.25 – 0.7 Ma B.P., sea-level change gradually increased in amplitude to 120 m and had dominant periods of 41 ka and 100 ka (Huybers, 2007). In the late Pleistocene, 0.7 - 0.012 Ma B.P., sea-level change had an average amplitude of 120 m and a dominant period of 100 ka. In the late Pleistocene, North American ice sheets had their greatest thickness and largest extent (Piper et al., 1994; Bintanja and van de Wal, 2008).

4.2.2. Stratigraphy, Glacial History, and Porewater Geochemistry

Siegel et al. (2012) identified seven stratigraphic units offshore Massachusetts that span the Cretaceous to the Holocene (Figure 4-2). The units show significant variability in stratigraphic thickness down dip and along strike. Unit 1 formed from slow, pelagic sedimentation during the Cretaceous and likely consists of carbonate sands. Unit 2 formed from slow, pelagic sedimentation during the Paleocene and Eocene and likely consists of carbonate mud and sand. Unit 3 formed from increased siliciclastic input during the Oligocene and Miocene (Steckler et al., 1999) and likely consists of silt and clay clinoforms. Unit 4 formed from rapidly deposited glacigenic sediments during the late Pleistocene and likely consists of poorly sorted silts and clays. Units 5 and 6 formed from high siliciclastic input with deposition responding to high-amplitude sea-level change during the late Pleistocene (Metzger et al., 2000) and likely consist of sand, silt, and clay clinoforms. Unit 7 formed from siliciclastic sedimentation and glacial outwash during the late Pleistocene and Holocene and likely consists of sand and silt.
Figure 4-2 A) Uninterpreted seismic line A - A', located in Figure 4-1. B) Interpreted seismic line A - A' showing interpreted age and lithologic units. C) Numerical grid used in three-dimensional modeling; colors correspond to lithologic units in Figure 4-2B. The depth of the grid extends to 3500 m in the south.
Two regional unconformities (U1 and U2) are also present in the study region. U1 (Figure 4-2B) is interpreted as a marine oxygen isotope stage (MIS) 12 glacial unconformity and marks the first Pleistocene shelf-crossing glaciation offshore Massachusetts (Siegel et al., 2012). Several additional glacial advances occurred throughout the late Pleistocene, although their extent onto the shelf was less, probably terminating near the maximum extent of the Laurentide ice sheet during the last glacial maximum (LGM) (Figure 4-1) (Oldale and O’Hara, 1984; Uchupi et al., 2001). U2 is a shallow sequence boundary that formed during the last sea-level fall (40 ka – 30 ka B.P.).

Porewater geochemical analysis from Atlantic Margin Coring Project (AMCOR) and U.S. Geological Survey (USGS) wells offshore New England and New Jersey and on Martha’s Vineyard and Nantucket Island (Figure 4-1) document freshwater that extends nearly 100 km offshore and reaches depths up to 500 mbsl (Kohout et al., 1977; Folger et al., 1978; Hathaway et al., 1979; Kohout et al., 1988). Well USGS 6001 on Nantucket Island shows several stratigraphic zones with varying salinity: from 7.3 – 150 mbsl, salinity is generally less than 1 part per thousand (ppt) in a predominantly sand and gravel layer; from 150 – 360 mbsl, there is an irregular pattern of salinity ranging from 1 – 29 ppt (seawater is 35 ppt), where freshwater is predominantly in sands and brine is in clays; below 360 mbsl there is a zone of freshwater (salinity of 2-3 ppt) in a large sand body. The freshwater in well USGS 6001 is much deeper than that predicted by the Ghyben-Herzberg principle (Kohout et al., 1977; Marksamer et al., 2007). Person et al. (2003) hypothesize that
glaciations on the Massachusetts shelf flushed salt water from permeable units near shore, resulting in the salinity pattern observed in the USGS 6001 well and in the ENW-50 well on Martha's Vineyard (Figure 4-1).

4.3. Previous Continental Shelf Hydrogeologic Modeling Studies

Several modeling studies of the continental shelf offshore Massachusetts simulated fluid flow and solute transport with Pleistocene boundary conditions (e.g., sea-level change, ice sheets, permafrost, pro-glacial lakes) to estimate the offshore distribution of freshwater and to explain the salinity observed in well USGS 6001 (Person et al., 2003; Marksamer et al., 2007; Cohen et al., 2010; DeFoor, 2011). These models used different assumptions of ice sheet extent and timing, sea-level change, and stratigraphy, which resulted in different salinity patterns (Figure 4-1D). One consistency, however, was that all of the studies showed large volumes of freshwater were emplaced offshore due to the combined effects of Pleistocene sea-level change and glaciations.

Person et al. (2003) modeled a two-dimensional transect from onshore Massachusetts through Nantucket Island to the shelf-slope break. They employed a simplified lithology with alternating layers of sand and silt/clay, and they assumed permafrost extended 50 km beyond the ice sheet edge. Sea level was prescribed as a sinusoid with a 100 ka period and a 120 m amplitude. Their models predicted that freshwater recharge during sea-level lowstands was not enough to emplace the deep freshwater observed at well USGS 6001. When they imposed a 1 ka glacial
period during the LGM, freshwater flushed aquifers and their models matched the salinity pattern at well USGS 6001 at 375 mbsl; however, due to the simplified stratigraphy, the model did not match the entire salinity profile at well USGS 6001 (Figure 4-1D).

Marksamer et al. (2007) modeled a similar two-dimensional transect to that of Person et al. (2003), but their stratigraphy included more detail, containing five stratigraphic units based on the lithology at well USGS 6001. They incorporated sediment loading, permafrost, and pro-glacial lakes, and modeled the Laurentide ice sheet during the LGM. Sea level was prescribed as a sinusoid with a 100 ka period and a 120 m amplitude. Their results also confirm that ice sheet loading was necessary to drive freshwater offshore. The refined stratigraphy improved the match between observed and modeled salinity; however, modeled salinity increased with depth, which is not observed in well USGS 6001 (Figure 4-1D).

Cohen et al. (2010) constructed a three-dimensional model to represent the continental shelf from offshore New Jersey to offshore Maine to assess the regional distribution of freshwater. The large model extent prevented the use of detailed stratigraphy, so they only included three lithologic units (medium-coarse sands, fine sands, and silt/clays). They, however, included three Pleistocene glacial cycles (Wisconsin, Nebraskan, and Illinoian), and assumed each had a regional glacial extent identical to the LGM. Sea level was based on an isotopically derived sea-level curve (Imbrie et al., 1984) for the past 200 ka and repeated for the entire
Pleistocene with a 120 m amplitude. Thus, they incorporated short-period (several thousand year) sea-level changes as observed in the isotopic record, however they did not capture the general trend of Pleistocene sea-level change period transitioning from 41 ka to 100 ka. Their results showed that in glaciated regions (i.e., offshore Massachusetts and Maine), freshwater emplacement was primarily in sand, however, there were no localized zones of brine predicted at well USGS 6001 (Figure 4-1D).

4.4. Numerical Modeling

We use a three-dimensional, finite-element, numerical model to simulate variable-density groundwater flow and heat and solute transport for the Massachusetts continental shelf from the end of the Pliocene through the Holocene (3 Ma). Our model includes five late Pleistocene glaciations that extended onto the shelf. This model allows us to explore the complex influences of climate cycles and three-dimensional stratigraphy on offshore freshwater emplacement and distribution. We use PGEOFE, a parallel version of GEOFE, which is a serial, finite-element-based paleohydrogeologic model (see Appendix D) (Person et al., 2007b; Cohen et al., 2010).

4.4.1. Model Domain

Our numerical grid is based on the seismic and lithologic interpretations of Siegel et al. (2012) (Figure 4-2). The base of the model is defined by the depth to the
top of Jurassic basement (Figure 4-2). Seven lithologic units overlie the Jurassic basement (Figure 4-2). The physical properties of these units are based on interpreted sediment type (Table 4-1), and will be referred to as the base-case (BC) model. Units 1 and 2 represent carbonate sands and carbonate muds that thicken to the south. Unit 3 represents Oligocene and Miocene clinoforms and contains three clinoforms that thicken to the south and to the west. Unit 4 represents a glacigenic sediment layer that is thickest in the south. Units 5 and 6 represent late Pleistocene siliciclastic clinoforms. Unit 7 is a layer of uniform thickness that represents Holocene glacial outwash.

We used a grid that contains 32,021 nodes and 172,167 tetrahedral elements that were generated by LaGrit (Gable et al., 1996). The horizontal node spacing averages 2 km while the vertical node spacing ranges from 10 m in the near surface to 400 m in the deeper subsurface. We use the more refined vertical discretization in the near surface to insure there is enough grid resolution in the thin upper model units to avoid numerical dispersion.

<table>
<thead>
<tr>
<th>Unit</th>
<th>Age and Lithology</th>
<th>$k_x (m^2)$</th>
<th>$\phi$</th>
</tr>
</thead>
<tbody>
<tr>
<td>7</td>
<td>Late Pleistocene/Holocene: glacial outwash</td>
<td>$10^{-13}$</td>
<td>0.3</td>
</tr>
<tr>
<td>6</td>
<td>Late Pleistocene: sand, silt, clay clinoforms</td>
<td>$10^{-15}$</td>
<td>0.2</td>
</tr>
<tr>
<td>5</td>
<td>Late Pleistocene: thin clay layer</td>
<td>$10^{-17}$</td>
<td>0.2</td>
</tr>
<tr>
<td>4</td>
<td>Late Pleistocene: glacigenic sediments</td>
<td>$10^{-15}$</td>
<td>0.3</td>
</tr>
<tr>
<td>3</td>
<td>Oligocene/Miocene: Silt/Clay clinoforms</td>
<td>$10^{-17}$</td>
<td>0.2</td>
</tr>
<tr>
<td>2</td>
<td>Paleocene Eocene: Carbonate mud</td>
<td>$10^{-15}$</td>
<td>0.2</td>
</tr>
<tr>
<td>1</td>
<td>Cretaceous: Carbonate sandstone</td>
<td>$10^{-14}$</td>
<td>0.3</td>
</tr>
</tbody>
</table>

Anisotropy: vertical/horizontal permeability ($k_x / k_z$) = 0.1

* Unit 3 contains three clinoforms

Table 4-1 Lithologic Units, Interpreted Age and Lithology, Permeability ($k$), and Porosity ($\phi$) for the Base-Case (BC) models
We solve the following groundwater flow equation, which describes variable-density fluid flow and incorporates the effects of ice sheet loading (Cohen et al., 2010; Person et al., 2012):

\[
\nabla \cdot \left[ k \frac{\rho_o g}{\mu_f} \mu \nabla (h + \rho_r z) \right] = S_s \left[ \frac{\partial h}{\partial t} - \frac{\rho_i}{\rho_o} \frac{\partial \eta}{\partial t} \right],
\]

(Equation 4-1)

where \( \nabla \) is the gradient operator, \( k \) is intrinsic permeability, \( \rho_o \) is density of water at the standard state (10°C, salinity of 0 ppt, and atmospheric pressure), \( g \) is acceleration due to gravity, \( \mu_f \) is viscosity of water at elevated temperature, pressure, and salinity conditions, \( \mu_r \) is relative viscosity \( (\mu_r = \mu_o/\mu_f) \), \( \mu_o \) is viscosity of water at the standard state, \( h \) is hydraulic head relative to a reference datum of modern sea-level (0.0 m), \( \rho_r \) is relative density \( (\rho_r=(\rho_f-\rho_o)/\rho_o) \), \( \rho_f \) is density of water at elevated temperature, pressure, and salinity conditions, \( z \) is elevation relative to a reference datum of modern sea-level (0.0 m), \( S_s \) is specific storage, \( t \) is time, \( \rho_i \) is ice density, and \( \eta \) is ice sheet thickness (see Tables 4-1 and 4-2 for model parameters). Equation 4-1 is similar to a standard equation for variable-density groundwater flow (e.g., Ingebritsen et al., 2007), modified to account for the effects of ice sheet loading \( (\partial \eta/\partial t) \). When the ice sheet is present we assume a loading efficiency of 1, which means the stress from the ice sheet results in an instantaneous increase in subsurface pore pressure below the ice sheet (Lemieux et al., 2008b). We do not account for the effects of sedimentation, which was shown to have little effect on
salinity distribution (Defoor, 2011). The sides and base of the model are no flow boundaries. The initial condition is hydrostatic fluid pressure for the entire model domain.

Fluid density ($\rho_f$) and viscosity ($\mu_f$) are functions of pressure, solute concentration, and temperature. This relationship is defined by thermodynamic equations of state determined by Kestin et al. (1981).

4.4.3. Solute Transport

Solute transport is determined from the advection-diffusion equation (e.g., Ingebritsen et al., 2007; Cohen et al., 2010):

$$\phi \frac{\partial C}{\partial t} = \nabla (\phi D \nabla C) - \vec{q} \nabla C,$$

(Equation 4-2)

where $\phi$ is porosity, $C$ is solute concentration, $D$ is a three-dimensional hydrodynamic dispersion-diffusion tensor for porous medium which is a function of longitudinal and transverse dispersivity and solute diffusivity (see Table 4-2 for model parameters) (Konikow and Grove, 1977), and $\vec{q}$ is the specific discharge vector.

$$\vec{q} = -\frac{k \rho_o g}{\mu_f} \mu_r \nabla (h + \rho_r z).$$

(Equations 4-3)
We model the total dissolved solid concentration and report it as mass fraction (parts per thousand).

The initial porewater salinity is 35 ppt throughout the model. During the model simulation, subaerial surface nodes with downward fluid flow are assigned freshwater salinity (0 ppt), and subaqueous surface nodes with downward fluid flow are assigned seawater salinity (35 ppt). The sides and base of the model have no salinity exchange.

4.4.4. Heat Transport

Heat transport is determined from the conductive and convective-dispersive heat transfer equation (Cohen et al., 2010):

\[
\left[ c_s \rho_s \phi + c_f \rho_f (1 - \phi) \right] \frac{\partial T}{\partial t} = \nabla \left[ \lambda \nabla T \right] - \bar{q} \rho_f c_f \nabla T, \quad \text{(Equation 4-4)}
\]

where \( c_s \) and \( c_f \) are the specific heat capacities of the solid and liquid phases, \( \rho_s \) is the density of the solid phase, \( T \) is temperature, and \( \lambda \) is the thermal dispersion-conduction tensor, which is a function of solid and fluid thermal conductivity, porosity, longitudinal and transverse dispersivities, fluid density, fluid heat capacity, and the specific discharge vector (see Table 4-2 for model parameters) (de Marsily, 1986).

The initial temperature is 4°C at the surface with a gradient of 30°C km\(^{-1}\). During the simulation, subaerial surface nodes have a constant temperature of 5°C,
and subaqueous surface nodes have a constant temperature of 4°C. The base of the model has a constant heat flux of 0.06 W m⁻². The sides of the model are thermally insulated boundary conditions.

<table>
<thead>
<tr>
<th>Model Parameters</th>
<th>Symbol</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Heat capacity of fluid</td>
<td>c_f</td>
<td>4128 J kg⁻¹ K⁻¹</td>
</tr>
<tr>
<td>Heat capacity of solid</td>
<td>c_s</td>
<td>870 J kg⁻¹ K⁻¹</td>
</tr>
<tr>
<td>Longitudinal dispersivity</td>
<td>α_L</td>
<td>100 m</td>
</tr>
<tr>
<td>Transversal dispersivity</td>
<td>α_T</td>
<td>10 m</td>
</tr>
<tr>
<td>Density of solid grain</td>
<td>ρ_s</td>
<td>2500 kg m⁻³</td>
</tr>
<tr>
<td>Density of water at standard state</td>
<td>ρ_o</td>
<td>1000 kg m⁻³</td>
</tr>
<tr>
<td>Density of ice</td>
<td>ρ_i</td>
<td>916 kg m⁻³</td>
</tr>
<tr>
<td>Solute diffusivity</td>
<td>D_m</td>
<td>1.0 x 10⁻⁹ m² s⁻¹</td>
</tr>
<tr>
<td>Solid thermal conductivity</td>
<td>κ_s</td>
<td>2.5 W m⁻¹ K⁻¹</td>
</tr>
<tr>
<td>Fluid thermal conductivity</td>
<td>κ_f</td>
<td>0.58 W m⁻¹ K⁻¹</td>
</tr>
<tr>
<td>Specific storage</td>
<td>S_s</td>
<td>10⁻³ m⁻¹</td>
</tr>
</tbody>
</table>

Table 4-2 Parameters Used in All Simulations

### 4.4.5. Sea-Level Boundary Condition

The sea-level boundary condition is applied by assigning surface nodes above sea level a hydraulic head equal to the surface elevation (Figure 4-3) while surface nodes below sea level are assigned a hydraulic head equal to sea level. We initialize the model with a sinusoidal sea-level curve for 1 Ma (Figure 4-4A). Then we prescribe an idealized, 2 Ma Pleistocene sea-level curve (Figure 4-4B). For the Pleistocene sea-level curve we assume sea-level fall is 80% of a cycle in the early and middle Pleistocene and 85% of a cycle in the late Pleistocene (Lisiecki and Raymo, 2007). Periodicity is 41 ka in the early Pleistocene, 100 ka in the late Pleistocene, and a combination of the two in the middle Pleistocene. Maximum sea-level fall (relative to modern) is 85 m in the early Pleistocene, 120 m in the late
Pleistocene, and varying with a range between 85 m and 120 m the in the middle Pleistocene.

Figure 4-3 Surface elevation of three-dimensional grid and the location of the cross sections shown in Figures 4-6, 4-7, 4-10, 4-11, and 4-12. The upper right hand corner (north-east corner) is 1-3 meters above sea level. The dashed lines are elevation contours in 10 m intervals. Location of the grid is shown in Figure 4-1.
Figure 4-4 A) Sinusoidal sea-level curved used to initialize models. B) Idealized 2 Ma Pleistocene sea-level curved used in all models. C) Stacked marine oxygen isotope data (Lisiecki and Raymo, 2005), which is a proxy for the general trends in sea-level period and amplitude in the Pleistocene.

4.4.6. Ice Sheet Boundary Condition

We prescribe the ice sheet boundary as a time series of ice sheet thickness and extent that represents the late Pleistocene glacial history on the shelf based on geophysical interpretations of an MIS 12 ice sheet that reached the shelf edge (Siegel et al., 2012) and the LGM ice sheet (Uchupi et al., 2001). We assume that the MIS 12 glaciation was followed by four glacial advances identical in extent to the LGM
(Figure 4-5A). Hydraulic head at the base of the ice sheet is 90% that of the ice sheet height (Boulton et al., 1995). The height profile of the ice sheet is a polynomial expression (van der Veen, 1998):

\[ \eta = H \sqrt{1 - \left(\frac{\xi}{L}\right)^2}, \]  

(Equation 4-5)

where \( \eta \) is the height of the ice sheet at a particular location (node) which is a distance \( \xi \) from the center of the ice sheet. \( H \) and \( L \) are the maximum ice sheet thickness and length at a particular time step and reflect the change in size of the ice sheet over a glacial cycle. The maximum thickness prescribed for the MIS 12 ice sheet is 1200 m (Figure 4-5B) based on the minimum ice sheet thickness (275 m) interpreted at the shelf edge (Siegel et al., 2012). The maximum thickness prescribed for the remaining ice sheets is 500 m (Figure 4-5B) based on numerical models of the LGM (Marshall et al., 2002). The ice sheets build from the northern edge of the model and advances to the south. The late Pleistocene ice sheets were only on the shelf for several 10s of thousands of years, so a minimum amount of crustal depression from the stress of the ice sheet is expected. The model nodes are depressed 20% of the ice sheet height (Cohen et al., 2010).
Figure 4-5  A) Ice sheet time series showing the assumed extent of the late Pleistocene ice sheets onto the continental shelf past the northern edge of the model domain.  B) Cross-sectional profile of maximum ice sheet height and extent onto the shelf for the Last Glacial Maximum (LGM) and Marine Oxygen Isotope Stage (MIS) 12 ice sheets.

4.5. Results

4.5.1. Sea-Level Change

Our initial simulation with the base-case model parameters (Table 4-1) considers the impacts of Pleistocene sea-level change (Figure 4-4) as a freshwater-forcing mechanism on the shelf in the absence of glaciations (Figure 4-6A). Sea-level change results in 18 km³ of predicted, present-day freshwater (salinity less than 5
ppt) in the shelf system (1.1% of total fluid volume). This estimate assumes freshwater-saturated sediments have a porosity of 0.2. The northern edge of the model domain has salinity less than 5 ppt (Figure 4-6A), and the average salinity of the model domain is 24.2 ppt (Table 4-3). Salinity also correlates with the surface elevation, with the eastern side of the model (20 m higher elevation) being fresher than the western side (Figure 4-7A).

The average salinity of the system decreased with time after each sea-level cycle (Figure 4-8). The time of highest average salinity corresponds to the time of highest sea-level (dashed lines in Figure 4-8), however, the time of highest average salinity lags behind the time of highest sea-level by an average of 15 ka in the late Pleistocene. This shows that the shelf hydrologic system takes time to respond to sea-level forcing.

The magnitude of response to sea-level forcing varies in the proximal and distal portions of the model. To illustrate this, we focus on two points through time within Unit 2 (Figure 4-9), proximal point P1 and distal point P2 (locations shown in Figure 4-6). Salinity with time at P1 has a large magnitude of change with each sea-level cycle, ranging from 30 ppt during a sea-level highstand to 10 ppt during a sea-level lowstand (Figure 4-9G). Salinity with time at P2 becomes fresher with each sea-level cycle; however, the magnitude of change with each sea-level cycle is less than 2 ppt from a sea-level highstand to a sea-level lowstand (Figure 4-9G). The magnitude of change in hydraulic head relative to change in sea level is similar in
the proximal and distal points. The average hydraulic head of -20 m to -30 m reflects the long-term trend of late Pleistocene sea level ranging from -120 m to 0 m. The average shore-normal flow rates across distal and proximal near surface sediments (Units 6 and 7) are similar. They are highest during sea-level lowstands and reach up to 0.2 m/a.
Figure 4-6 Predicted, present-day salinity for base-case permeability (Table 4-1) with: A) no ice sheet; B) single ice sheet advance during the Last Glacial Maximum (LGM); and C) ice advance during Marine Oxygen Isotope Stage (MIS) 12 followed by four additional ice advances to the margin of the LGM (all glacial cycles shown in Figure 4-5). Dashed black lines are 5 ppt salinity contour. Arrows indicate the maximum extents of the MIS 12 and LGM ice sheets. The black dots (P1 and P2) mark the locations where salinity and hydraulic head are displayed through time (see Figure 4-9). Location of cross sections is shown in Figure 4-3.
Figure 4-7 Predicted, present-day salinity for base-case permeability with: A) no ice sheet; B) single ice sheet advance during the Last Glacial Maximum (LGM); and C) ice advance during Marine Oxygen Isotope Stage (MIS) 12 followed by four additional ice advances to the margin of the LGM (all glacial cycles shown in Figure 4-5). Dashed black lines are 5 ppt salinity contour. Location of cross sections is shown in Figure 4-3.
Figure 4-8 Average salinity for entire model domain from 1.25 Ma to present for base-case permeability (Table 4-1) with no ice sheets. Dashed lines mark sea-level highstands. Plot shows lag in response of sub-surface hydrology to sea-level inundation.
Figure 4-9 A-C) Late Pleistocene sea-level curve used in all models. Ice sheet extent onto shelf used for boundary conditions of: D) no ice; E) LGM ice sheet; and F) all glacial cycles. G-I) Salinity; and J-L) hydraulic head through time at a specific proximal point (P1) and distal point (P2) within Unit 2 (carbonate mud) for the specific boundary conditions of sea level and ice sheets shown above. Time series for the proximal point (P1) is shown in black and the time series for the distal point (P2) is shown in gray. Figure 4-6 shows the location of the two monitoring points.
4.5.2. Late Pleistocene Ice Sheets

To explore the effects of Pleistocene ice sheets, several models were run with sea-level cycles (Section 5.1) but also including an ice sheet boundary condition (Figure 4-5). The LGM ice sheet (last ice sheet cycle shown in Figure 4-5A) results in 35 km$^3$ of predicted, present-day freshwater offshore (2.2% of total fluid volume) (Table 4-3). This is nearly double the amount of freshwater predicted in the simulation that only considered sea-level change. The head induced by the LGM ice sheet drives freshwater 5 km farther offshore in the deeper sediments as compared to the sea-level only simulation (Figure 4-6B). The salinity pattern beyond the maximum extent of the LGM, however, shows little difference from the sea-level only simulation (Figures 4-6A and B; 4-7A and B; 4-9G and H). The average salinity of the system is 23.5 ppt, less than 1 ppt lower then the sea-level simulation (Table 4-3). These results are consistent with Cohen et al. (2010), who showed that glacially emplaced freshwater does not extend far beyond the edge of the LGM ice sheet.

The MIS 12 ice sheet, followed by four additional late Pleistocene glacial cycles (all glacial cycles shown in Figure 4-5A) results in 100 km$^3$ of predicted, present-day freshwater offshore (6.3% of total fluid volume) (Table 4-3). The freshwater extends up to 30 km beyond Martha's Vineyard (Figure 4-6C). This is more than five times as much freshwater than the sea level only simulation and almost three times as much freshwater than the LGM ice sheet simulation. The
average salinity of the system is 19.3 ppt, and the shelf has a broad region of brackish water (5-30 ppt) (Figure 4-6C).

<table>
<thead>
<tr>
<th>Ice Boundary</th>
<th>Freshwater Volume (km$^3$)</th>
<th>Average Salinity (ppt)</th>
</tr>
</thead>
<tbody>
<tr>
<td>No Ice</td>
<td>18</td>
<td>24.2</td>
</tr>
<tr>
<td>LGM Ice Sheet</td>
<td>35</td>
<td>23.5</td>
</tr>
<tr>
<td>All Glacial Cycles</td>
<td>100</td>
<td>19.3</td>
</tr>
</tbody>
</table>

Table 4-3 Results of Base Case Model

4.5.3. Sensitivity Study

We conduct a sensitivity study to address the effects of permeability on freshwater distribution. Permeability has a strong control on the emplacement and preservation of Pleistocene meltwater in glaciated sedimentary basins as low permeability confining units prevent flushing of underlying aquifers after ice sheet retreat (McIntosh et al., 2012; Person et al., 2012). Permeability was assigned to the base-case model based on lithology interpreted from seismic data. We, however, lack direct well control; therefore there is uncertainty in our assigned lithology and permeability.

To evaluate the impact of permeability architecture on freshwater distribution, we run models with different permeability (Table 4-4) (Figure 4-10). Sensitivity study 1 (SS1) tests the effects of a lower-permeability confining layer at the surface (Unit 7). Sensitivity study 2 (SS2) tests the effects of a higher-permeability carbonate aquifer (Unit 1). Sensitivity study 3 (SS3) tests a simplified lithology (alternating layers of high and low permeability), which resembles
previous models (Person et al., 2003, Cohen et al., 2010). We did not vary specific storage or porosity in our sensitivity studies. Specific storage, unlike permeability, does not vary by several orders of magnitude within continental shelf sediments and porosity, unlike permeability, does not have a first order control on fluid flow.

### Table 4-4 Horizontal Permeability (m$^2$) Used For Sensitivity Studies

<table>
<thead>
<tr>
<th>Unit</th>
<th>Lithology</th>
<th>BC</th>
<th>SS1</th>
<th>SS2</th>
<th>SS3</th>
</tr>
</thead>
<tbody>
<tr>
<td>7</td>
<td>glacial outwash</td>
<td>$10^{-13}$</td>
<td>$10^{-16}$</td>
<td>$10^{-13}$</td>
<td>$10^{-16}$</td>
</tr>
<tr>
<td>6</td>
<td>sand, silt, clay clinoforms</td>
<td>$10^{-15}$</td>
<td>$10^{-15}$</td>
<td>$10^{-15}$</td>
<td>$10^{-14}$</td>
</tr>
<tr>
<td>5</td>
<td>thin clay layer</td>
<td>$10^{-17}$</td>
<td>$10^{-17}$</td>
<td>$10^{-17}$</td>
<td>$10^{-14}$</td>
</tr>
<tr>
<td>4</td>
<td>glaciogenic sediments</td>
<td>$10^{-15}$</td>
<td>$10^{-15}$</td>
<td>$10^{-15}$</td>
<td>$10^{-16}$</td>
</tr>
<tr>
<td>3</td>
<td>Silt/Clay clinoforms</td>
<td>$10^{-17}$/10$^{-16}$/$10^{-17}$</td>
<td>$10^{-17}$/10$^{-14}$/10$^{-17}$</td>
<td>$10^{-17}$/10$^{-14}$/10$^{-17}$</td>
<td>$10^{-14}$</td>
</tr>
<tr>
<td>2</td>
<td>Carbonate mud</td>
<td>$10^{-15}$</td>
<td>$10^{-15}$</td>
<td>$10^{-15}$</td>
<td>$10^{-14}$</td>
</tr>
<tr>
<td>1</td>
<td>Carbonate sandstone</td>
<td>$10^{-14}$</td>
<td>$10^{-14}$</td>
<td>$10^{-13}$</td>
<td>$10^{-16}$</td>
</tr>
</tbody>
</table>

Anisotropy: Vertical/Horizontal Permeability ($k_z / k_x$) = 0.1

* no anisotropy

### 4.5.3.1. Effects of Confining Unit Permeability at the Surface (SS1)

The lower-permeability confining unit at the surface (SS1) results in no predicted, present-day freshwater due to sea-level change only, and reduces the amount of freshwater due to sea-level change and the LGM ice sheet with 10 km$^3$ of predicted, present-day freshwater offshore. Less freshwater infiltration due to sea-level change is expected, as the lower-permeability layer reduces vertical infiltration in the near-surface layer, and prevents freshwater from entering the system during sea-level lowstands. Cohen et al. (2010) observed a similar pattern in their three-dimensional sensitivity study where including low-permeability confining layers resulted in the least volume of freshwater infiltration. Conversely, there is 109 km$^3$
of predicted, present-day freshwater from the MIS 12 and late Pleistocene glacial cycles (Table 4-5), with freshwater extending nearly as far as the base-case model (Figure 4-10B). In this case, in the lower-permeability confining layer, the MIS 12 ice sheet produced a larger gradient in hydraulic head in the vertical direction than the base case model. Consequently, even though the permeability is less, vertical fluid velocity and freshwater infiltration was sustained throughout the glacial cycle. In addition the lower-permeability layer helps preserve freshwater that was emplaced by the MIS 12 ice sheet by reducing the exfiltration rate of freshwater from all underlying units after ice sheet retreat (Figure 4-10B).

4.5.3.2. Effects of Carbonate Permeability (SS2)

The inclusion of high-permeability carbonate sands without anisotropy (SS2) results in the emplacement of more freshwater under all boundary conditions compared to the base-case models. The predicted, present-day freshwater volume is 26, 58, and 104 km$^3$ from sea-level change, the LGM, and the MIS 12 ice sheet boundary conditions (Table 4-5). There is only a small change, however, in the amount of freshwater for the late Pleistocene glaciation boundary condition as compared to the base-case model (Figure 4-10C). The increase in stored freshwater is due to the high-permeability carbonate sands (Unit 1) which have an order of magnitude higher fluid velocity and thus more freshwater is advected offshore during sea-level lowstands.
4.5.3.3. Effect of Simplified Lithology (SS3)

The simplified lithology (SS3) of alternating layers of high permeability ($10^{-14}$ m$^3$) and low permeability ($10^{-16}$ m$^3$) results in the smallest volume of freshwater in the system after the MIS 12 and late Pleistocene glaciations. The predicted, present-day freshwater volume with the MIS 12 ice sheet boundary conditions is 70 km$^3$ (Table 4-5), with the maximum extent being 25 km offshore (Figure 4-10D). This difference is primarily due to the lack of a large carbonate aquifer (Unit 1) that the previous models had and which preserves glacially-derived freshwater.

<table>
<thead>
<tr>
<th>Ice Boundary</th>
<th>SS1</th>
<th>SS2</th>
<th>SS3</th>
</tr>
</thead>
<tbody>
<tr>
<td>No Ice</td>
<td>0.0</td>
<td>26</td>
<td>18</td>
</tr>
<tr>
<td>LGM Ice Sheet</td>
<td>10</td>
<td>58</td>
<td>55</td>
</tr>
<tr>
<td>All Glacial Cycles</td>
<td>109</td>
<td>104</td>
<td>70</td>
</tr>
</tbody>
</table>

*Table 4-5 Freshwater Volume in Sensitivity Studies (km$^3$)*
Figure 4-10 Results of base-case model and sensitivity tests (Tables 4-1 and 4-4). A), C), E), and G) show the permeability assigned to each model unit. B), D), F), and H) show the corresponding, present-day salinity after sea-level change and all five glacial cycles. The dashed black lines show location of 5 ppt salinity contour. Thin black lines mark model unit boundaries. Location of cross sections is shown in Figure 4-3. A and B) Base-case model results. C and D) Effect of confining unit permeability at the surface (SS1). E and F) Effects of carbonate permeability (SS2). G and H) Effect of simplified lithology (SS3).
4.6. Discussion

4.6.1. Extended Ice Sheet Scenario

The difference in predicted freshwater between the base-case model with all the glacial cycles, including the MIS 12 ice sheet, and the base-case model with the LGM ice extent is likely due to: (1) enhanced freshwater infiltration, where compared to the LGM ice sheet, the MIS 12 ice sheet was thicker (i.e., a larger driving force) and had a more expansive areal extent (i.e., a larger region effected by glacial recharge); and (2) remnant overpressure (hydraulic head in excess of hydrostatic) after the MIS 12 ice sheet retreat. There is an order of magnitude increase in freshwater infiltration during the MIS 12 glacial advance to near the shelf-slope break. Freshwater was emplaced throughout the shelf with 376 km$^3$ of emplaced freshwater by the time the ice sheet reaches its maximum extent (425 ka B.P.). This is consistent with continental-scale models of the Laurentide ice sheet, where ice-sheet loading produces large volumes of freshwater that infiltrate and charge deep basins (Lemieux et al., 2008a). The four late Pleistocene glacial cycles following the MIS 12 glaciation each recharge several 10s km$^3$ of freshwater into the system, however, the majority of the freshwater offshore at present was emplaced by the MIS12 ice sheet. Lemieux et al. (2008a) noted that permafrost could influence freshwater infiltration rate as it creates an impermeable layer at the surface. We did not include the formation of permafrost because of the proximity to the ocean,
which would moderate the temperature, keeping it above freezing and thus preventing the formation of sustained permafrost regions.

MIS 12 ice-sheet loading causes an increase in overpressure in the permeable sediments above Unit 5. The overpressure decreases to the south as ice thickness decreases, and consequently, creates a seaward overpressure gradient that yields increased seaward-directed fluid flow in shelf sediments. This drives freshwater offshore. McIntosh et al. (2011) predicted a similar pattern in the near-surface sediments of the Michigan basin during glaciations, which contributed to freshwater flushing of sedimentary units below and beyond the extent of the Laurentide ice sheet.

Overpressures that develop due to loading from the MIS 12 ice sheet are preserved below low permeability confining layers (Units 3 and 5) long after ice sheet retreat (400 Ka B.P.) (Figures 4-9L; 4-11A and B). When the MIS 12 ice sheet was at its maximum (425 ka B.P.), nearly 1000 m of excess hydraulic head developed in the sediment where ice overburden was greatest (see Figure 4-5B for MIS 12 ice sheet profile). By the time the ice sheet has fully retreated and sea level is at a highstand (400 ka B.P.), overpressure in sediments above the confining layers has nearly dissipated and hydraulic head is equivalent to sea level (Figure 4-11A and B). Sediments below the confining layers, however, maintain elevated overpressure after the ice sheet has fully retreated (Figure 4-11A and B). This occurs from the migration of overpressure into the deeper sediments during MIS 12 ice sheet retreat. McIntosh et al. (2011) observed a similar phenomenon of glacially
derived overpressure persisting in deeper confining units in their model after ice sheet retreat.

Figure 4-11 A and B) Overpressure after MIS 12 ice sheet retreat (400 ka B.P.). Most of the overpressure is below Unit 3, a low-permeability clinoform layer. C and D) Salinity after MIS 12 ice sheet retreat (400 ka B.P.). Black lines mark model stratigraphy, units 3 and 5 have a low permeability clinoform. Location of cross section is shown in Figure 4-3.

Remnant, glacially derived overpressure helps preserve emplaced freshwater because the overpressure drives fluid flow towards the surface, and thus prevents downward intrusion of seawater during highstands. Remnant overpressure is higher in the northwest region of the model because the northwest region contains a thicker section of a low permeability clinoform in Unit 3 as
compared to the northeastern region (Figure 4-11A and B). The difference in remnant overpressure results in a substantially different salinity pattern where the northwestern region has more freshwater and a greater offshore extent of freshwater compared to the northeastern region (Figure 4-11C and D).

The results of the sensitivity studies further highlight the important effect of an extensive MIS 12 ice sheet for the emplacement of a large volume of freshwater offshore Massachusetts. In all three sensitivity studies, the addition of the MIS 12 ice sheet increased the amount of offshore freshwater (Table 4-5). For SS1 and SS2, the predicted extent of freshwater from the MIS 12 ice sheet at present is primarily contained within 30 km of Martha’s Vineyard. This is similar to the base-case model (Figures 4-10A-C). This is twice as much freshwater as predicted based on a simplified lithology (SS3) and as predicted based on the LGM ice sheet extent only (Table 4-5). Thus, our knowledge of an extensive MIS 12 ice sheet and our improved three-dimensional model geometry allow us to better constrain probability for a relatively large volume of non-renewable freshwater to exist offshore Massachusetts.

### 4.6.2. Seafloor Topography and Sea-Level Change

Seafloor topography (elevation and relief) has a direct control on the infiltration of freshwater and the distribution of the mixing zone between freshwater and saltwater because: (1) seafloor elevation determines exposure during a lowstand, where a lowstand exposes the shelf to increased meteoric
recharge (Meisler et al., 1984); and (2) seafloor relief controls the rate of shoreline transgression during sea-level rise, where rapid transgression leads to density instabilities (high density saltwater overlying low density freshwater) that lead to advection and a broad zone of brackish water (Kooi et al., 2000). In simple, cross-sectional numerical models, these processes proved effective at emplacing freshwater and creating broad zones of brackish water offshore. In our three-dimensional model, however, we observe these processes occurring in the northsouth (down-dip/shore-normal) and east-west (along-strike/shore-parallel) directions because of variations in the three-dimensional seafloor elevation of 20 m across the northern section of the model (Figure 4-3 and 4-7). For the base-case model with sea-level change only, freshening of the sediments occurs for a longer period of time on the northeastern side of the model during each sea-level lowstand, and consequently the northeastern side of the model contains fresher water (Figure 4-7A).

The east-west difference in elevation across the model domain also results in different average salinity across the region that experienced ice sheet loading (Figure 4-7C). The eastern half of the model has an average salinity of 10.1 ppt and the western half has an average salinity of 18.6 ppt, even though the cross section experienced the same magnitude of ice sheet loading. This further shows that topography has a strong control on the freshwater distribution. Previous, two-dimensional, cross-section models were not able to explore the subtle effects of sea-floor elevation on freshwater distribution after glaciations.
Figure 4-12 Salinity profiles of near surface sediments during sea-level rise from 15 – 0 ka B.P. Arrows mark the direction of vertical fluid flow. The red lines mark sea level, and the red triangles mark the shoreline location. Location of cross sections is shown in Figure 4-3.
Sea-floor topography also affects the distribution of freshwater in near surface sediments (Units 6 and 7) as it engenders convection cells up to 10s of kilometers in length and up to 100s of meters in depth that redistribute fresh and saline water (Figure 4-12). The convention cells develop from density differences that are a direct result of the sea-floor topography. Seafloor relief causes isotherms in the near surface sediments to dip (i.e., they parallel the sea floor), which can induce thermally derived density differences and convection in high permeability (>10^{-15} m^2) sediments (Wilson and Ruppel, 2007). For that reason, convection cells do not develop in the sensitivity test (SS1), which contains a low-permeability layer at the surface (Figure 4-10B). Sea-floor elevation controls shelf exposure time, where sea-level lowstands cause dense saline water to move downward, which can induce salinity derived density instabilities and convection (Kooi et al., 2000). For that reason, we observe convection cells on the exposed portion of the shelf (Figure 4-12A and B).

The convection cells in the base-case model develop periodically across the model space both above and below sea level, however, they are more numerous when the shelf is inundated. The periodic development of the convection cells in the upper sediment layers effects the freshwater distribution in two ways: (1) during a sea-level lowstand, there are regions of the model with subaerial surface nodes that are not freshening as fluid is being discharged in these areas and therefore freshwater cannot infiltrate; and (2) during a sea-level highstand, there are previously freshened regions of the model with subaqueous surface nodes that are
not becoming more saline as fluid is being discharged in these areas and seawater cannot infiltrate. This suggests that areas of high permeability offshore may have freshwater that was emplaced during a sea-level lowstand and preserved, as convection cells that developed subsequently during a sea-level highstand prevented the infiltration of seawater into the sediments.

**4.6.3. Current Freshwater Distribution**

The discovery of an MIS 12 glaciation that extended to near the shelf-slop break suggests that there is an increased likelihood of freshwater far offshore Massachusetts; however, three-dimensional variation in seafloor elevation and the three-dimensional distribution of stratigraphy (clinoform confining units and carbonate sands) strongly influenced the emplacement and distribution of freshwater through time. Most of the present-day freshwater is in the northern section of the model domain within 30 km of Martha’s Vineyard (Figure 4-13). However, subtle differences in sea-floor elevation resulted in more freshwater on the eastern half of the model than the western half (Figure 4-13), even though there was more freshwater in the western half of the model after the MIS 12 ice sheet retreat (Figure 4-11). The three-dimensional ice sheet thickness was not explored. Due to the effects of ice sheet gradient on sub-surface fluid flow, we infer that the three-dimensional ice sheet thickness would also have an impact on freshwater emplacement and distribution. Future studies, including offshore drilling, should focus on understanding the complex relationship between three-dimensional shelf
stratigraphy, three-dimensional sea-floor topography, and three-dimensional ice sheet growth to better predict freshwater distribution. A better understanding of the freshwater distribution and dynamics may help constrain: (1) the generation of microbial methane (Martini et al., 1998; Schlegel et al., 2011); (2) the availability of freshwater resources (Kooi and Groen, 2003; Person et al., 2007a); (3) submarine groundwater discharge and the nutrient supply to the ocean over time (Slomp and Cappellen, 2004); and (4) the safety and long-term storage of nuclear waste (Iverson and Person, 2012; Vidstrand et al., 2013).
Figure 4-13 Plan view of predicted, present-day freshwater distribution for the base-case model after sea-level change and all five glacial cycles. Contours are thickness in meters of the continental shelf sediments that are filled with freshwater. The model domain is marked with the thick black lines and shaded light grey.

4.7. Conclusions

Three-dimensional, variable-density, numerical modeling that couples fluid flow and heat and solute transport on the continental shelf offshore Massachusetts provides evidence that an MIS 12 ice sheet followed by four late Pleistocene glacial
cycles resulted in as much as 100 km$^3$ of present-day freshwater offshore. Most of the freshwater is within 30 km of Martha’s Vineyard and shallower than 500 mbsl. Sensitivity studies that address permeability also show that the MIS 12 ice sheet resulted in as much as three times the volume of freshwater as compared to models that did not include an MIS 12 ice sheet. We demonstrate that more freshwater exists offshore than previously estimated; this is due to better constraints of the glacial history and inclusion of detailed three-dimensional topography and lithology. The freshwater that exists offshore is an important, non-renewable resource for Massachusetts’s coastal communities, however, geochemical data are needed to further determine and map the distribution of the freshwater and to fully address its utility as a resource.
Conclusions

Interpretations from geophysical data that determined the extent of an MIS 12 ice sheet and the detailed stratigraphy and lithology offshore Massachusetts were integrated with numerical modeling that couples climate history to predict the distribution of glacially-emplaced freshwater and the generation of glacially-derived overpressure. This work was an integrated approach to demonstrate the complex relationship between late Pleistocene glaciations, stratigraphy, and hydrogeology on the continental shelf offshore Massachusetts.

Our results show that an MIS 12 ice sheet reached the shelf edge, extending over 100 km further offshore Massachusetts than the Laurentide ice sheet during the Last Glacial Maximum. The hydrodynamic effects of the MIS 12 ice sheet on the shelf were responsible for generating overpressure and emplacing freshwater, however, the occurrence of clinoforms within Oligocene/Miocene sediments had a
strong control on the distribution of the overpressure and freshwater. The Clinoforms that were overrun by the MIS 12 ice sheet may have focused fluid flow that generated localized overpressure up to 1-2 MPa and caused sediment deformation features, which are observed in the seismic data. The clinoforms also contributed to the preservation of 100 km$^3$ of freshwater that was emplaced by the MIS 12 ice sheet by confining the freshwater in underlying carbonate aquifers. The extent of the clinoforms offshore is not uniform, and thus the extent of freshwater is not uniform.

This coupled behavior is important and will help contribute to future studies of glaciology, hydrogeology, and climate studies as: (1) we provide geophysical evidence for the marine extent of a pre-LGM ice sheet, which may help calibrate models of Pleistocene ice sheet cycles and determine the direction of paleo-ice stream flow; (2) we provide geophysical evidence of localized overpressure, which may be contributing to submarine groundwater discharge which effects the nutrient supply to the oceans; and (3) we provide numerical predictions of freshwater volume, which may help predict potential, non-renewable resources for Massachusetts coastal communities. Future studies, however, should be mindful of the connection we demonstrated between climate history, stratigraphy, and fluid flow, as often all of these parameters are not constrained.


Balco, G., Rovey, C. W., II., 2010, Absolute chronology for major Pleistocene advances of the Laurentide Ice Sheet, Geology, 38(9), 795-798.


Boulton G.S., Dongelmans, P., Punkari, M., Broadgate, M., 2001, Palaeoglaciology of an ice sheet through a glacial cycle: the European ice sheet through the Weichselian, Quaternary Science Reviews, 20, 591-625.

Boulton, G., Hagdorn, M., 2006, Glaciology of the British Isles Ice Sheet during the last glacial cycle: form, flow, streams and lobes, Quaternary Science Reviews, 25, 3359-3390.


Huuse, M., Lykke-Andersen, H., 2000, Overdeepened Quaternary valleys in the eastern Danish North Sea: morphology and origin, Quaternary Science Reviews, 19, 1233-1253.

Huybers, P., 2007, Glacial variability over the last two million years: an extended depth-derived age model, continuous obliquity pacing, and the Pleistocene progression, Quaternary Science Reviews, 26, 37-55.


Uchupi, E., Mulligan, A. E., 2006, Late Pleistocene stratigraphy of Upper Cape Cod and Nantucket Sounds, Massachusetts, Marine Geology, 227, 93-118.


Vidstrand, P. Follin, S., Selroos, J., Näslund, J, Rhén, I., 2013, Modeling of groundwater flow at depth in crystalline rock beneath a moving ice-sheet margin,
exemplified by the Fennoscandian Shield, Sweden, Hydrogeology Journal, 21, 239-255.


Appendix A: Seismic Processing

All data were collected aboard the R/V Endeavor in August of 2009.
(http://terra.rice.edu/department/research/dugan/EN465_Cruise_Report.pdf)

This appendix provides detail on the processing used to produce the stacked sections for interpretation. All data are available at the University of Texas Institute for Geophysics. (http://www.ig.utexas.edu/sdc/cruise.php?cruiseIn=en465)

A.1 Acquisition and Processing Geometry

The seismic source was Scripps’ generator-injector (GI) air gun (45/105 in³) firing with 6 s intervals. Data were recorded for 4 s after each shot. The streamer length was 600 m, with 48 channels spaced at 12.5 m and a near offset of 55.25 m. The streamer was towed at a depth of 3 m and the GI gun at a depth of 2 m with a nominal ship speed of 4 kts. Shot locations were obtained from a GPS instrument aboard the R/V Endeavor; nominal shot spacing was 10 to 12 m. Common depth point (CDP) bins were assigned every 6.25 m, yielding CDPs with a 22-28 trace fold. Data processing was performed on CDP gathers.

A.2 Filtering and Amplitude Control

We used a bandpass filter with poles of 3-6-120-240 Hz. A trace mute was applied to eliminate noise in the upper 10s of ms of the far offset traces. True amplitude recovery was used with a 5 dB/s gain. This increased the amplitude in the
signal from deeper reflections, and reduced the amplitude contrast between the sea-
floor reflection and deeper reflections.

**A.3 Multiple Reduction**

We collected data on water depths ranging from 25 m to 125 m, resulting in a large amount of energy from water-bottom multiples interfering with true reflections. I used two methods to eliminate some of the energy generated by water-bottom multiples: predictive deconvolution in the slope-intercept (tau-p) domain and energy rejection in the frequency-wave number (f-k) domain.

Data are transferred from the time domain to the tau-p domain because coherent events that interfere with each other in the time domain (water-bottom multiples and primary reflections) are separated and isolated in the tau-p domain. In addition, multiples in the tau-p domain are periodic, whereas multiples in the time domain are only periodic at zero offset; thus, filters such as predictive deconvolution that operate on the periodicity and predictability of multiples are more effective at removing multiples in the tau-p domain (Stoffa, 1989). After predictive deconvolution is applied to the transformed data, an inverse tau-p transform is used to return the data to the time domain.

Frequency-wave number (f-k) filtering can be considered a form of velocity filtering. Data are transformed via a Fourier transform to components of frequency (f) and wave number (k). The two parameters are related by the apparent velocity
of the signal (Kearey et al., 2002). In the f-k domain, signals of a certain apparent velocity will be separated and isolated from signals of a different apparent velocity.

I apply a constant normal moveout (NMO) velocity similar to water (1,480 m/s) to CDP gathers. This makes water-bottom multiples a flat reflection horizon in the time domain, which is a nearly infinite apparent velocity. I then do an f-k transform the NMO corrected gathers. In f-k space, nearly infinite apparent velocity corresponds to a wavenumber of zero. I remove energy centered near wavenumber of zero (water-bottom multiples). After energy rejection, data are transformed back to the time domain followed by an inverse NMO correction; leaving a CDP gather with energy subtracted that corresponded to the water-bottom multiples.

**A.4 Stacking and Post-Stack Processing**

The last phase of processing is to produce a stacked section. This involves an interactive velocity analysis on CDP gathers to determine stacking (RMS) velocities for each seismic line. Picking ranged between every 50 to every 200 CDPs. By applying an NMO correction prior to stacking, true reflections sum together, further increasing the signal to noise ratio.

I used a post-stack predictive deconvolution filter on each stacked section to further reduce the water-bottom multiple. The filtered, stacked section is then depth migrated with a Kirchhoff depth migration using smooth interval velocities obtained from my stacking velocity picks. The final processing steps for each stacked section
were an automatic gain control with an operator length of 800 ms and Wiener Levinson frequency-distance (f-x) deconvolution, which helps clean the signal horizontally.
Appendix B: Full Waveform Inversion

We follow the methods of Kormendi and Dietrich (1991), which apply the Fréchet derivatives of Dietrich and Kormendi (1990) and use a conjugate gradient algorithm to iteratively solve the non-linear inverse problem. The inverse problem seeks a 1D stratified earth model \( m \) that minimizes the cost function \( S(m) \);

\[
S(m) = \frac{1}{2} \left( \| d_{\text{syn}} - d_{\text{obs}} \|_D^2 + \| m - m_0 \|_M^2 \right),
\]

(Equation B-1)

where \( d_{\text{obs}} \) is the observed CDP gather, \( d_{\text{syn}} \) is the synthetic wavefield, \( m_0 \) is the starting model, and \( \| \| \) are weighted L² norms. A fully elastic synthetic wavefield is calculated from \( m \) using the reflectivity algorithm of Kennet and Kerry (1979) with layer parameters of P- and S-wave velocities, density, and attenuation.

The inversion scheme iteratively solves for P-wave velocity updates, and S-wave velocity and density are updated at each iteration based on empirical relationships (Hamilton, 1978; Castagna et al., 1985; Korenaga et al., 1997).

The inversion is performed in the frequency/ray-parameter (w-p) domain. CDP gathers were first transformed to the intercept-time/ray-parameter (τ-p) domain using the approach outlined in Korenaga et al. (1997) and then transformed to the w-p domain using a Fourier transform. The finite aperture of the MCS streamer limits the range of ray parameters in the data, and the range limit varies as a function of τ. We calculated the expected range of observed p as a function of τ using a smoothed interval velocity determined from semblance analysis (Figure 3-
4). We limited the p-range of the data for three discrete $\tau$ windows such that the maximum $p$ included in the inversion is slightly less than the maximum expected $p$. For example, for the upper 0.25 s of data, where the greatest $p$-values are present, we use a maximum $p$-value of 0.35. This method achieves a resolution in the velocity field on the order of one fourth of the wavelength, yielding a resolution of 2 - 4 m given the frequency range of the seismic source used for this survey.

**B1. Source Wavelet**

An accurate estimate of the source wavelet is critical for the inversion, since the cost function is based on a sample-for-sample difference of the observed and synthetic data, and the synthetic data are determined as the theoretical response of the 1D stratified Earth model convolved with the estimated source. We follow the methods of Korenaga et al. (1997) and Minshull et al. (1994) to estimate the source wavelet using the recording of the first arrival (P) and first multiple (M) in the near-offset trace. The first arrival is the convolution of the source and the seafloor response function, and the recorded multiple is the negative convolution of the source with the seafloor response function twice,

\[
P(t) = s(t) * r(t) \quad \text{(Equation B-2)}
\]

\[
M(t) = -s(t) * r(t) * r(t) \quad \text{(Equation B-3)}
\]

where $P(t)$ and $M(t)$ are the primary and multiple time series, $s(t)$ is the source signal, $r(t)$ is the seafloor response function, and $*$ denotes the convolution operator. Dividing equation B.3 by equation A.2 yields
\[ r(t) = - \frac{M(t)}{P(t)}. \]  \hspace{1cm} \text{(Equation B-4)}

The source signal can be obtained from equation (B.2) and (B.4) as:

\[ s(t) = \frac{P(t)}{r(t)}. \]  \hspace{1cm} \text{(Equation B-5)}

Equations (B.4) and (B.5) are solved in the frequency domain. To enhance the signal of the primary and the multiple, we stack the near-offset traces from 20 adjacent shots. The exact primary and multiple travel times are calculated for the given water depth and source-receiver offsets and are used to window the primary and the multiple arrival times prior to deconvolution. The result of the deconvolution is then used as the source for the full-waveform inversion.
Appendix C: MATLAB Code Used for 1D Overpressure Modeling

This appendix provides detail on the Matlab code used to solve 1D overpressure. The code solves an equation that accounts for sedimentation and ice sheet loading:

\[
K_z \frac{\partial^2 h}{\partial z^2} = S_s \left[ \frac{\partial h}{\partial t} - \zeta \frac{\rho_i}{\rho_f} \frac{\partial \eta}{\partial t} - \zeta \frac{\rho_b - \rho_w}{\rho_f} \frac{\partial L}{\partial t} - P_{ext} \right],
\]

(Equation C-1)

where \( K_z \) is the hydraulic conductivity in the vertical direction, \( S_s \) is the specific storage, \( \rho_i \), \( \rho_w \), and \( \rho_b \) are the densities of ice, water, and fluid-saturated sediment, \( \eta \) and \( L \) are the elevations of the ice and sediment relative to sea level, respectively, and \( P_{ext} \) is an external source term which can be assigned to a particular layer to represent external fluid sources outside of the model domain (Dugan and Germaine, 2008). This groundwater flow equation (C-1) is solved with a fully implicit scheme which provides a solution that is unconditionally stable (Fletcher, 1997). The finite difference solution was compared to an analytical solution that describes the formation of a thick sedimentary package with a constant sedimentation rate overlying an impermeable base (Bredehoeft and Hanshaw, 1968; Gibson, 1958).

FD_MOD2.m

%----------------------------------------------------------------------------------------%  
% 1D pore pressure model for:  
% a) sediment loading and unloading  
% b) ice sheet loading and unloading  
% c) variable permeability  

Subroutines and Functions used in program:

- (Fi_Iterate) is used to calculate pore pressure with each iteration
- (k_effective) calculates the effective permeability for the model domain
- (lith_D) determines lithologic stress
- (ES_Matrix) calculates the effective stress

par_file  % par_file contains all input parameters

%%% Create Constant Vectors %%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%

depth=1:1:td;  % depth increment of model (1 meter)
hydro=depth*dw*9.8;  % hydrostatic pressure relative to top of model
                    % (current sealevel)

PoreTime=(zeros(1,d1-1),(d1:1:td))*(dw*9.8);
LithTime=lith_D(d1,td,dw,ds);

%%% Pore Pressure For Time Steps %%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%

MEST1=ES_Matrix(PoreTime, LithTime);  % maximum effective stress for t1

%%% Initial Pore Pressure at time 1: %%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%

poreT1=hydro(:,d1:td);  % pore pressure at T1, same as hydro, no loading

%%% Pore pressure for time t2: %%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%

delt=(t1*365*24*60*60)/(d2-d1);  % size of time step per grid =
                                % total time / change in elevation

del_sed=-1;  % sediment erosion per time step
del_ice=(hi-d1*(dw/di))/(d2-d1);  % ice growth per time step above min-
                                   % flotation thickness

%%% ice must be more than flotation thickness %%%%%%%%%%%%%%%%%%%%%%%%%%

if hi < d1*(dw/di)
    del_ice=0;
    disp('Ice Less Than Floatation Thickness')
end
Bi=BL;  % B for ice loading  
Bs=BL;  % B for sediment loading  

x=poreT1/(dw*9.8)-depthT1;  % initial pore pressure  
x=(x,x(length(x)));  % base node for no flow boundary  

ki=ones(1,length(x))*k1;  % vector of permiabilities  

% Create confining layer:  
if clt == 0  
    disp('No Confining Layer')  
else  
    for i = (d2-d1)+dcl:1:(d2-d1)+clt+dcl  
        ki(i)=clp;  % node has permiability set by "clp"  
    end  
end  

% loop for iterations: decrease grid size from erosion & load ice sheet  
for n = 1:1:(d2-d1)  
    HI=x(:,2:length(x));  % remove grid cell due to erosion  
    ki(ki(:,2:length(ki)));  % remove grid cell due to erosion  
    ke=keffective(ki, 1);  % calculate new k effective  
    % adjust boundary conditions  
    HI(1)=(d1*(dw/di)+n*del_ice)*(di/dw)-(d1+n);  
    hi=hi<d1*(dw/di)  
    HI(1)=0;  % Head of seafloor if no ice is present:  
end  

FI_Iterate  

N=(d2-d1);  
% fill pore pressure matrix:  
file1=((zeros(1,d1-1+n),x(1,1:length(x)-1)+depth(d1+n:td)))*(dw*9.8);  
PoreTime=(PoreTime; file1);  

% fill lithology matrix:  
file1=lith_D(d1+n, td, dw, ds);  
file1=file1+del_ice*n*9.8*di;  
LithTime=(LithTime; file1);  

end  

x=x(1,1:length(x)-1);  % remove base boundary  

ki=ki(ki,1);  

depthT2=depth(d2:td);  

x=x+depthT2;  % add hydrostatic head  

poreT2=x*(dw*9.8);  % pore pressure in (Pa)  

% set eroded seds pore pressure to 0  
for l = 1:1:length(PoreTime(:,1))  
    for i = d1:1:d2-1  
        PoreTime(l,i)=0;  
    end  
end
MEST2=ES_Matrix(PoreTime, LithTime);  % maximum effective stress for t2

%--- Pore presure at time t3:
%-----------------------------------------------

% *this is done in two stages:
% A) unoad ice sheet for time (t2)
% B) load glacigenic sediments for time (t2)
%

display ('ice sheet unloading')

Bi=BU;  % reduced B for deglaciation
Bs=BL;

%--- (A) ---------------------
delt=(t2*365*24*60*60);  % size of time step per grid = total time

del_sed=0;
del_ice=(hi-d2*(dw/di));  % decline in ice to floatation thickness

HI=poreT2/(dw*9.8)-depthT2;
HI(1)=0;  % top node has head of seafloor
HI=(HI,HI(length(HI))));  % base node for no flow boundary

ki=(ki,ki(length(ki))));  % initial k from previous time step
ke=keffective(ki, 1);

% fill pore pressure matrix:
file1=zeros(1,d2-1),x)*(dw*9.8);
PoreTime=(PoreTime; file1);

% fill lithology matrix:
file1=lith_D(d2, td, dw, ds);
LithTime=(LithTime; file1);

%--- (B) ---------------------

Bs=BL;
delt=(t2*365*24*60*60)/(d2-d3);  % size of time step per grid =  
% total time / change in elevation 

del_sed=1;  

% initial pore pressure 

x=(poreT3A/(dw*9.8)-depthT2);  

% base node for no flow boundary 

ki=ki(1:length(ki));  % initial k from previous time step 

for n = 1:1:(d2-d3)  

Hl=(0,x);  

% adjust boundary conditions: 

Hl(1)=0;  % new node has head of sea floor 

ki=(k2, ki);  % new node has perm of k2 

ke=keffective(ki, 1);  

end 

depthT3=depth(:,d3:td);  

% fill pore pressure matrix: 

file1=((zeros(1,d2-1-n),x(1,1:length(x)-1)+depth(d2-n:td)))*(dw*9.8);  

PoreTime=(PoreTime; file1);  

% fill lithology matrix: 

file1=lith_D(d2-n, td, dw, ds);  

LithTime=(LithTime; file1);  

end 

depthT3=depth(:,d3:td);  

x=x(1,1:length(x)-1);  % remove extra base node 

ki=ki(1,1:length(ki)-1);  

x=x+depthT3; 

x=x*(dw*9.8);  

poreT3=x;  

MEST3=ES_Matrix(PoreTime, LithTime);  % maximum effective stress for t3 

%--- Pore pressure at time t4:  

%------------------------------------------------------------- 

display ('Pleistocene sedimentation') 

Bs=BL;  

delt=(t3*365*24*60*60)/(d3-d4);  

del_sed=1;  

del_ice=0;  

x=(poreT3/(dw*9.8)-depthT3);  % initial pore pressure
x=(x,x(length(x)));     % base node for no flow boundary
ki=(ki,ki(length(ki)));  % initial k from previous time step

for n = 1:1:(d3-d4)

    HI=(0,x);

    % adjust boundary conditions:
    HI(1)=0;  % new node has head of sea floor

    % determine new effective stress:
    ki=(k3,ki);  % new node has perm of k3
    ke=keffective(ki, 1);

    % artificial layer loading
    if n >= ((d3-d4)-3)
        for i=(d2-d3)+1+n+do:1:(d2-d3)+1+n+do+lo
            HI(i)=HI(i)+mo;
        end
    end

    FL_Iterate

    % fill pore pressure matrix:
    file1=((zeros(1,d3-1-n),x(1,1:length(x)-1)+depth(d3-n:td)))*(dw*9.8);
    PoreTime=(PoreTime; file1);

    % fill lithology matrix:
    file1=lith_D(d3-n, td, dw, ds);
    LithTime=(LithTime; file1);

    end

    x=x(1,1:length(x)-1);     % remove extra base node
    ki=ki(1,1:length(ki)-1);
    depthT4=depth(:,d4:td);
    x=x+depthT4;
    x=x*(dw*9.8);
    poreT4=x;

    MEST4=ES_Matrix(PoreTime, LithTime);  % maximum effective stress for t4

    %--- Lithologic Pressure For Time Steps -------------------------------%
    %---------------------------------------------------------------%

    lithT1=(depthT1-d1)*ds*9.8+d1*dw*9.8;
    lithT2=(depthT2-d2)*ds*9.8+hi*di*9.8;  % note, density of ice is used

    if hi < d1*(dw/di)
        lithT2=(depthT2-d2)*ds*9.8+d2*dw*9.8;
    end
lithT3=(depthT3-d3)*ds*9.8+d3*dw*9.8;
lithT4=(depthT4-d4)*ds*9.8+d4*dw*9.8;
%
% Effective Stress For Time Steps
lithT1=(depthT1-d1)*ds*9.8+d1*dw*9.8;
lithT2=(depthT2-d2)*ds*9.8+d2*dw*9.8;
lithT3=(depthT3-d3)*ds*9.8+d3*dw*9.8;
lithT4=(depthT4-d4)*ds*9.8+d4*dw*9.8;
%
% Plot Results
esT1=lithT1-poreT1;
esT2=lithT2-poreT2;
esT3=lithT3-poreT3;
esT4=lithT4-poreT4;
figure(1)  % pore pressure predicted from model
subplot(2,4,1)
pore=poreT1;
depthP=depthT1;
lith=lithT1;
es=esT1;
max_ES=MEST1;
sf=d1;
S_plot1
title('initial condition')
subplot(2,4,2)
sf=d2;
S_plot2

subplot(2,4,3)
pore=poreT3;
depthP=depthT3;
lith=lithT3;
es=esT3;
max_ES=MEST3;
sf=d3;

subplot(2,4,5)
plot([-10000000,12000000],(d2,d2),'r--')
subplot(2,4,6)
plot([-10000000,12000000],(d2,d2),'r--')
S_plot1
title('ice retreat & sedimentation')
plot([-1000000,12000000],[d2, d2],'r')
S_plot2
plot([-1000000,12000000],[d2, d2],'r')

subplot(2,4,7)

S_plot2
plot([-1000000,12000000],[d2, d2],'r')

subplot(2,4,4)
pore=poreT4;
depthP=depthT4;
lith=lithT4;
es=esT4;
max_ES=MEST4;
sf=d4;
S_plot1
title('pressent')
plot([-1000000,12000000],[d3, d3],'k')
plot([-1000000,12000000],[d2, d2],'r')

subplot(2,4,8)
S_plot2
plot([-1000000,12000000],[d3, d3],'k')
plot([-1000000,12000000],[d2, d2],'r')

figure(3)
%--- Deviatoric Stress

figure(2) % changes in effective stress predicted from model

%--- Change in Effective Stress from T3 to T4
% subplot(1,3,1)
% del_es=esT4(:,(d3-d4)+1:length(esT4))-esT3; % change in effective stress
% plot(del_es,depthT3,'k','linewidth',2)
% set(gca,'YDir','reverse')
% axis([-10000,10000,0,600])
% hold on
% plot((0,12000000),(d3, d3),'k')
% plot((0,12000000),(d2, d2),'r')

subplot(1,3,1)
del_st=MEST4(:,d4:td)-esT4; % difference in max effective stress and present
plot(del_st,depthT4,'k','linewidth',2)
set(gca,'YDir','reverse')
axis([0,10000000,0,600])
title('Max ES - Present ES')
hold on
plot((0,120000000),(d3, d3),'k')
plot((0,120000000),(d2, d2),'r')

subplot (1,3,2) % Predicted Porosity
Ph0=0.4; % initial porosity
BC=5E-8; % Bulk Compressibility
mes=MEST4(1,d4:td); % Maximum Effective Stress
Ph=Ph0*exp(-BC*mes);

plot(Ph,depthT4,'k','linewidth',2)
set(gca,'YDir','reverse')
axis((0.25,.5,0,600))
title('Predicted Porosity')
xlabel('porosity')

% Han's Empirical relationship
C=.4;
%Pe=(lithT4-poreT4)*.01;

%Tosaya's empirical relationship for shaley sandstones
Vp=5.8-8.6*Ph-2.4*C

subplot (1,3,3)
plot(Vp,depthT4,'k','linewidth',2)
hold on
set(gca,'YDir','reverse')
axis((1.0,2.5,0,600))
title('Predicted Velocity')
xlabel('velocity (km/s)')

FI_Iterate.m

% FI_iterate
%
% model input = HI (head initial (m) )
% model output = x (head out (m) )

%delx=1;

%--- Calculate K Effective (ke) -------------------------------
%------------------------------------------------------------------------------------------
\[ Tx = \left( \frac{ke \times dw \times 9.8}{vis} \right) \]
\[ S = S_s \]

\[ \% Tx = Tx \times 0 + K; \% Note, special case just for benchmark \]

\[ \% constant used in numerical approximation: \]
\[ \alpha = \left( \frac{delt}{(delx^2)} \right) \times (Tx/S); \% Fluid transport term \]
\[ \beta = Bi \times (di/dw) \times \text{del_ice}; \% Ice loading term \]
\[ \gamma = Bs \times ((ds-dw)/dw) \times \text{del_sed}; \% Sed loading term \]

\[ \% make d matrix \]
\[ d = (HI(2) + \alpha(1) \times HI(1) + \beta + \gamma); \]
\[ \text{for } i = 3: \text{length(HI)} - 2 \]
\[ d(i-1) = HI(i) + \beta + \gamma; \]
\[ \text{end} \]
\[ d(\text{length}(HI) - 2) = HI(\text{length}(HI) - 1) + \beta + \gamma; \]

\[ \% make values for compressed tridiagonal matrix \]
\[ a = -\alpha'; \]
\[ b = (1 + 2 \times \alpha)'; \]
\[ b(\text{length}(b)) = (1 + \alpha(\text{length}(\alpha))); \]
\[ c = -\alpha'; \]
\[ x = \text{TDMAsolver}(a, b, c, d); \]

\[ \% fill in ends of X which are fix as the boundary condition \]
\[ \text{file1} = (); \]
\[ \text{for } i = 1: \text{length}(x) \]
\[ \text{file1}(i+1) = x(i); \]
\[ \text{end} \]
\[ \text{file1}(1) = HI(1); \]
\[ \text{file1}(\text{length}(HI)) = x(\text{length}(x)); \]

**keffective.m**

function (ke) = keffective (ki, delx)

\[ \% Output (ke) is the effective permeability based on the array with \]
\[ \% intrinsic permeability at each node (ki) \]
\[ \% (ke) will have two less nodes than (ki) \]

\[ ke = (); \]
\[ \text{for } i = 2:1: \text{length}(ki) - 1 \]
\[ \% ke(i) = \left( \frac{3 \times \text{delx}}{((\text{delx}/(ki(i-1))) + (\text{delx}/ki(i)) + (\text{delx}/(ki(i+1))))} \right) \]
\[ ke(i) = \left( \frac{2 \times \text{delx}}{((\text{delx}/(2 \times ki(i-1))) + (\text{delx}/ki(i)) + (\text{delx}/(2 \times ki(i+1))))} \right); \]
\[ \text{end} \]
\[ ke = ke(:,2: \text{length}(ke)); \]

end
Appendix D: PGEOFE Parameter Files

This section outlines the parameter files used in PGEOFE. These files are used to define the boundary conditions of: sea-level change, climate temperature, and ice sheet extent. They are also used to define material properties and initial conditions. There are a total of 15 parameter files, all of which are listed below.

This program is currently set up to run on the Rice University Department of Earth Science Cluster called “Sapphire”. PGEOFE is currently compiled and located at:

/home/js22/pgeofe/pgeofe-2.5.p4-bigred/pgeofe

MATLAB visualization scripts, and data output files are located on the Linux machine called “asha”:

/home/js22/pgeofe/pgeofe-2.5.p4-bigred/pgeofe

1) mv3d_v1.control: Simulation control file
-Contains the model run time, time step size, model unit properties such as permeability, porosity, and specific storage.

2) mv3d_v1.coord: Node coordinates
-Node number, X, Y, and Z coordinates

3) mv3d_v1_clin.elements: Elements connectivity
-Defines model tetrahedral elements

4) mv3d_v1.salt.init: Initial condition
-Contains the initial salinity, head, and temperature

5) mv3d_v1.boundaries: Boundary conditions
- Contains heat flux at base, surface temperature and side boundary conditions

6) mv3d_v1.column: Column data

7) mv3d_v1.area: Area of top nodes
   - Node number, X, Y, and Z coordinates off all surface nodes

8) mv3d_v1.surfmesh: Surface mesh data

9) mv3d_v1.surfhydro: Surface hydrology initial conditions

10) mv3d_v1.lake: Lake connection and data
    - Not used as we do not account for pro-glacial lakes

11) climate_warm.dat: Climate time series
    - Contains the temperature time series

12) mv3d_v1.LP5_sea_level: Sea level time series
    - Sea level at every time step

13) icesheet.dat: Ice sheet time series
    - Ice sheet height (H) and length (L) at every time step

14) margin_LGM2.dat: Ice sheet margin data
    - X and Y coordinates for ice sheet margin

15) mv3d_v1_LP1.ice: Ice sheet data
    - Maximum ice sheet height, length, and flow direction