Marine sediment record from the East Antarctic margin reveals dynamics of ice sheet recession

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ABSTRACT

The Antarctic shelf is traversed by large-scale troughs developed by glacial erosion. Swath bathymetric, lithologic, and chronologic data from jumbo piston cores from four sites along the East Antarctic margin (Iceberg Alley, the Nielsen Basin, the Svenner Channel, and the Mertz-Ninnis Trough) are used to demonstrate that these cross-shelf features controlled development of calving bay reentrants in the Antarctic ice sheet during deglaciation. At all sites except the Mertz-Ninnis Trough, the transition between the Last Glacial Maximum and the Holocene is characterized by varved couplets deposited during a short interval of extremely high primary productivity in a fjord-like setting. Nearly monospecific layers of the diatom *Chaetoceros* alternate with slightly more terrigenous layers containing a mixed diatom assemblage. We propose that springtime diatom blooms dominated by *Chaetoceros* were generated within well-stratified and restricted surface waters of calving bays that were influenced by the input of iron-rich meltwater. Intervening post-bloom summer-fall laminae were formed through the downward flux of terrigenous material sourced from melting glacial ice combined with mixed diatom assemblages. Radiocarbon-based chronologies that constrain the timing of deposition of the varved sediments within calving bay reentrants along the East Antarctic margin place deglaciation between ca. 10,500–11,500 cal yr B.P., post-dating Meltwater Pulse 1A (14,200 cal yr B.P.) and indicating that retreat of ice from the East Antarctic margin was not the major contributor to this pulse of meltwater.

INTRODUCTION

Cruise NBP0101 of the research vessel/icebreaker (RV/IB) *Nathaniel B. Palmer* conducted a marine geologic and geophysical investigation along the continental shelf of the East Antarctic margin from 58°E to 147°E (Fig. 1). Jumbo piston cores (JPCs) provide high-resolution sediment records of the transition from the last glacial to the Holocene, permitting assessment of the timing and nature of deglaciation. This paper presents swath bathymetric maps (Fig. 1) and radiocarbon-dated lithologic information (Fig. 2) that detail the chronology and processes involved in ice sheet retreat from four regions of the East Antarctic margin: Iceberg Alley (Mac.Robertson Shelf), the Nielsen Basin (Mac.Robertson Shelf), the Svenner Channel (eastern Prydz Bay), and the Mertz-Ninnis Trough (Wilkes Land margin). Each study target is a trough-shaped geomorphic feature that extends across the width of the East Antarctic shelf. As a consequence of their depth (400–1100 m), these troughs preserve a more complete section of sedimentation during deglaciation, in contrast to adjacent shallow parts of the shelf that were still affected by grounded ice and hence sites of nondeposition.

These East Antarctic margin sedimentary records, which are remarkably similar to one another and to a record from the Palmer Deep, Antarctic Peninsula (Domack et al., 2006), support the cohesive response of these regions to climate forcings at the end of the last glacial. Collectively, these data also provide important chronologic constraints for tracing northern versus southern hemisphere origins of deglacial meltwater pulses (Fairbanks, 1989; Bard et al., 1990), a controversial issue (i.e., Clark et al., 2002; Peltier, 2005).

PHYSIOGRAPHY OF ICEBERG ALLEY, NIELSEN BASIN, SVENNER CHANNEL, AND MERTZ-NINNIS TROUGH

The morphology of transverse troughs separated by shallow banks is characteristic of the Antarctic continental margin (Vanney and Johnson, 1979; Johnson et al., 1982; Anderson, 1999). Physiographic maps of troughs in the study areas were made using SEABEAM® multibeam data collected continuously during the cruise and edited onboard by cruise participants. Sub-bottom profiles were collected with a Bathy2000 3.5 kHz, hull-mounted chirp system with signal penetration to depths of ~50–100 m.

Mac.Robertson Shelf–Iceberg Alley and Nielsen Basin

The rugged Mac.Robertson Shelf lies directly west of Prydz Bay, extending 400 km between ~60°E and 70°E (Fig. 1). The shelf is relatively narrow (~90 km wide) and composed of shallow banks (<200 m) cut by three deep troughs that are U-shaped in cross section, with steep sides and relatively flat floors. They were formed by glacial erosion during the Quaternary (O'Brien et al., 1994; ten Brink and Schneider, 1995; Harris and O'Brien, 1996). The intervening banks show evidence of both modern and ancient iceberg turbation (Harris and O'Brien, 1998). Iceberg Alley is linear, ~85 km long and 10–20 km wide. It reaches a depth of 850 m, but most of the

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floor is ~475–575 m deep. The shallow banks on either side are lined with grounded icebergs; hence, its name. Nielsen Basin, bounded by western and eastern Storegg Bank, is deeper and more sinuous than Iceberg Alley, but its inner portion is similarly characterized by a "high relief ridge and valley topography" (Harris and O'Brien, 1998) formed by glacial incision. The deepest region (~1300 m) is located at the innermost part of the shelf, and depths shallow oceanward.

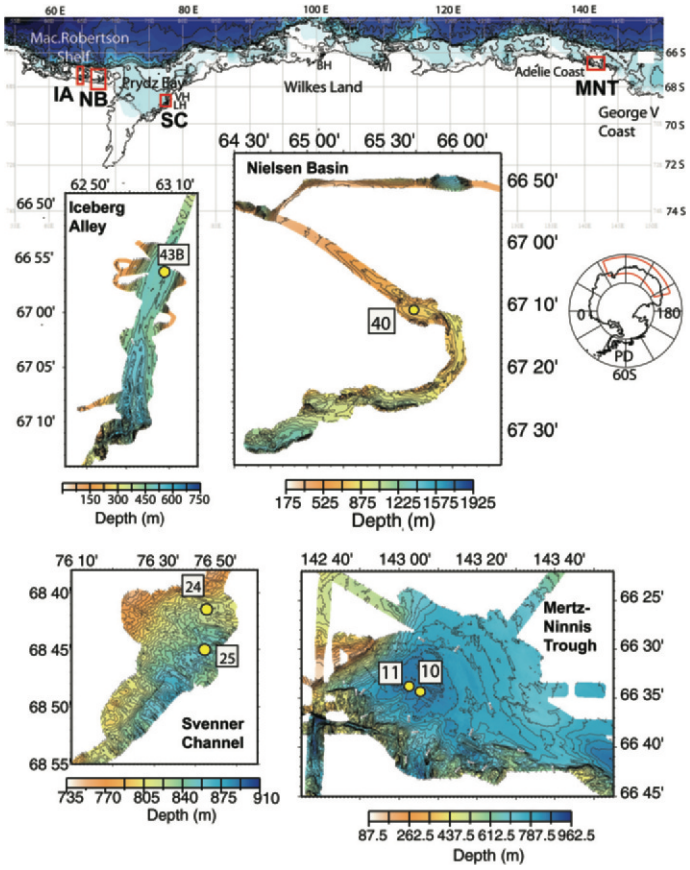

Figure 1. Location map and multibeam swath maps for Iceberg Alley (IA), the Nielsen Basin (NB), Svenner Channel (SC), and the Mertz-Ninnis Trough (MNT). Jumbo piston core sites are noted. General bathymetric chart of the oceans (GEBCO) bathymetric contours are in 500 m intervals; note the transition from light blue to darker blues at the 1000 m contour. LH—Larsemann Hills; VH—Vestfold Hills; BH—Bunger Hills; WI—Windmill Islands; PD—Palmer Deep.

Svenner Channel

Svenner Channel is an elongate (~180 km), glacially eroded trough in eastern Prydz Bay (Fig. 1) (Stagg, 1985). In the outer part of the channel, depths reach ~1000 m, while the deepest parts of the inner channel average ~800–850 m. Formation by glacial erosion is supported by observation of sediment ridges at the seafloor; the trend of these features, which are similar to drumlins or megaflutes, most likely parallels former ice flow direction (O'Brien and Harris, 1996). Directly to the northwest is the Amery Depression, a broad basin averaging 600–800 m water depth. O'Brien and Harris (1996) suggest very rapid grounding zone retreat of the Lambert Glacier across the shelf due to the relatively deep water depths of the Amery Depression and its landward dip. Rapid ice retreat in the Amery Depression would suggest that similarly rapid ice retreat probably took place in Svenner Channel.

Mertz-Ninnis Trough

The Mertz-Ninnis Trough, located along the Wilkes Land Margin between ~142°E and 146°E (Fig. 1), is a deep (>1300 m), linear feature bordered by the Adélie Bank to the west and the Mertz Bank to the northeast; both banks are relatively shallow (200–400 m). Surfaces of the banks show evidence of iceberg turbate structures (Barnes, 1987; Beaman and Harris, 2003) and are the sites of grounded icebergs today. During the Last Glacial Maximum, the Mertz Glacier expanded along the axis of the trough, as indicated by overcompacted sediments at the shelf break (Domack, 1982) and moraines on Mertz Bank

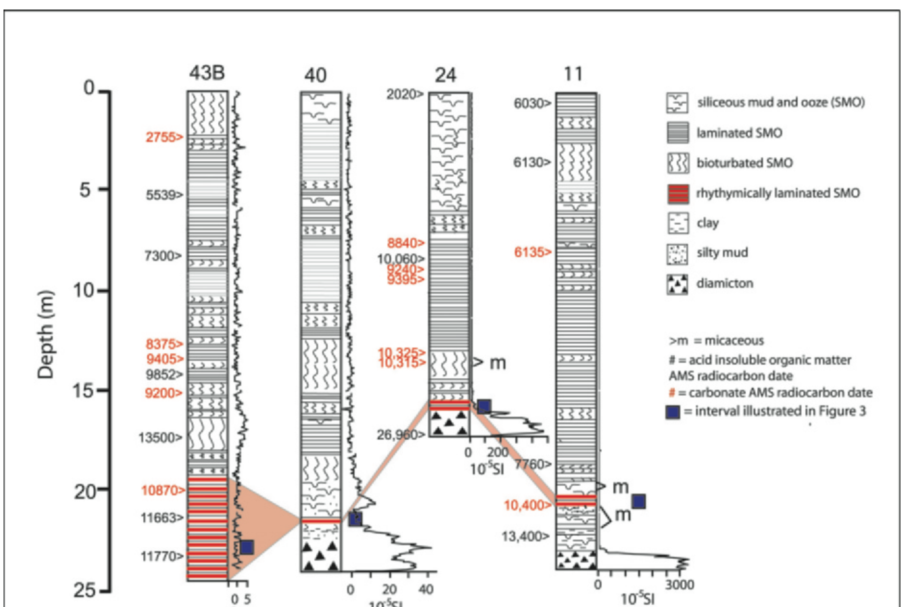

Figure 2. General lithologic columns for cores from JPC11 (Mertz-Ninnis Trough), JPC24 (Svenner Channel), JPC40 (Nielsen Basin), and JPC43B (Iceberg Alley). Uncorrected radiocarbon dates are noted for each core. See Table DR2 (see text footnote 1) for ranges on radiocarbon data. Shading indicates deglacial section described in paper. Curves to right of each lithologic log represent magnetic susceptibility for each core. Magnetic susceptibility was measured onboard ship using a Bartington magnetic susceptibility meter. Note increased magnetic susceptibility in the diamicton.

(Barnes, 1987; Domack et al., 1989; Beaman and Harris, 2003; and McMullen et al., 2006). Beaman and Harris (2003) also note seabed features characterized as "smooth diamicton" and glacially carved megaflutes that document more extensive glacial ice.

CORE DESCRIPTIONS

JPCs were recovered using a 4.5" diameter jumbo piston coring system. The cores were stored under refrigeration and shipped to the Antarctic Research Facility at Florida State University, where they were opened, photographed, described, and sampled (GSA Data Repository, Table DR1[1]).

Iceberg Alley

JPC43B is a 23.96 m core; the upper 19.13 m is alternately laminated and bioturbated biosiliceous ooze and mud (SMO). The lower 4.83 m is distinguished by its varves, rhythmically paired laminations that represent a single year of deposition (Stickley et al., 2005). This section could be as much as 5 m thicker, as ~30 m of post-glacial sediment was observed above a glacially scoured surface. Within each couplet, the lower lamination is diatom ooze, orange to orange-brown, comprised primarily of *Chaetoceros* spp. resting spores deposited during the annual spring bloom. The upper lamination of each couplet is comprised of a mixed diatom assemblage with a higher concentration of terrigenous material, including angular quartz sand, silt, and clay deposited during the summer and fall. Despite the presence of sublaminae, which makes it difficult to determine the exact number of years represented by the deglacial section, a total of 210 couplets (Stickley et al., 2006) were observed in this 4.83 m section, indicating that high productivity and sediment flux lasted for ~200 yr.

Nielsen Basin

JPC40 penetrated to the underlying glacial section (21.55–23.92 m) (Figs. 2 and 3), which is comprised of dark greenish gray silty clay and granule-rich sand. Dropstones are present. The transition to the deglacial section is distinct, with graded couplets, similar to those in Iceberg Alley, from 21.08 to 21.55 m. Thirty-three couplets occur, indicating that this style of deposition lasted for <50 yr, a significantly shorter time frame than in Iceberg Alley. As in JPC43B, alternately laminated and bioturbated SMO overlies the varved section.

Svenner Channel

JPC24 penetrated several meters of underlying glacial sediments (Figs. 2 and 3), greenish gray sandy and silty clay, with sand- to gravel-sized ice-rafted grains and granule-rich clay. The boundary between glacial diamict and overlying biosiliceous Holocene sediments is very sharp; note the color transition from greenish gray sediments to the overlying reddish sediment illustrated in Figure 3. Again, the transition is marked by paired laminations, but only 10 couplets occur between 15.58 and 15.80 m (Fig. 3). The ephemeral nature of the conditions responsible for the formation of these couplets, which in JPC24 likely span no more than a decade, is underscored by the record from nearby JPC25, where the transition, though quite sharp at 8.88 m, is not marked by any varves. And again, the overlying sediments are comprised of alternately laminated and bioturbated SMO.

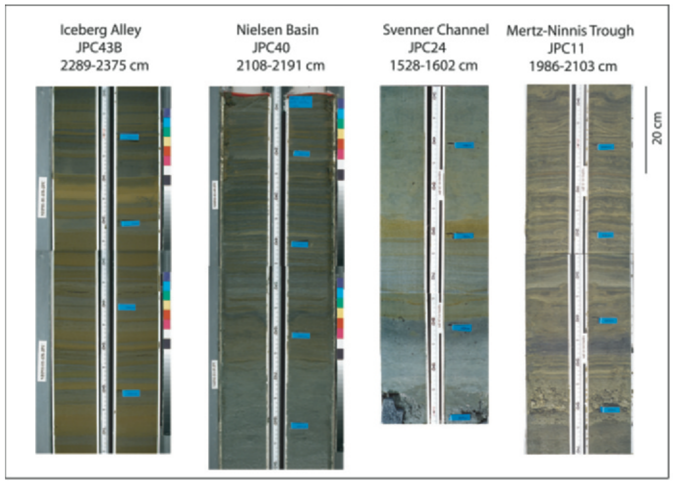

Figure 3. Photographs of laminated sediments deposited at the time of deglaciation, from four cores recovered from the East Antarctic margin. Note the mm- to cm-scale laminae and their regular alternation between an orange to orange-brown lamination and a more gray lamination in JPC43B, JPC40, and JPC24. These rhythmic laminations occur over a 483 cm interval in JPC43B, over a 47 cm interval in JPC40, and over a 22 cm interval in JPC24. In JPC11, paired laminations occur, but the pattern is not as regular.

Mertz-Ninnis Trough

JPC11, a 23.98-m-long core, penetrated glacial sediments, including well-consolidated, mud-supported gravels with a high concentration of ice-rafted material, at 22.92 m (Figs. 2 and 3). The transition to the Holocene is more complex as compared to the other East Antarctic margin cores. A moderately diatomaceous gray, muddy clay with sparsely scattered ice-rafted debris from 22.53 to 22.92 m records this transition and is interpreted as having a meltwater origin. The diatomaceous post-glacial clay is overlain by a diatom-rich, micaceous mud with scattered reddish layers comprised primarily of *Chaetoceros*.

Between 19.90 and 21.58 m, the sediment lithology alternates between bioturbated and thinly laminated SMO. Laminations generally alternate between a *Chaetoceros*-dominated layer and a layer with a more diverse diatom assemblage and a greater concentration of terrigenous material. The species succession, however, is more complicated, and though a seasonal pattern is recognized, it is not repeated as regularly as observed in the other cores (Maddison et al., 2006). Consequently, these sediments are not interpreted as varves. Such data suggest a less isolated setting and greater communication with open waters of the shelf during deglaciation. Also, given the bioturbated intervals that interrupt the finely laminated sequences, it is hard to develop as precise a chronology for deglaciation, particularly in contrast to the uninterrupted and simpler varved sections observed in the other East Antarctic margin cores.

[1]GSA Data Repository item 2006235, Tables DR1 and DR2, core location information and radiocarbon data of core samples, is available on the Web at www.geosociety.org/pubs/ft2006.htm. You can also obtain a copy of this item by writing to editing @geosociety.org.

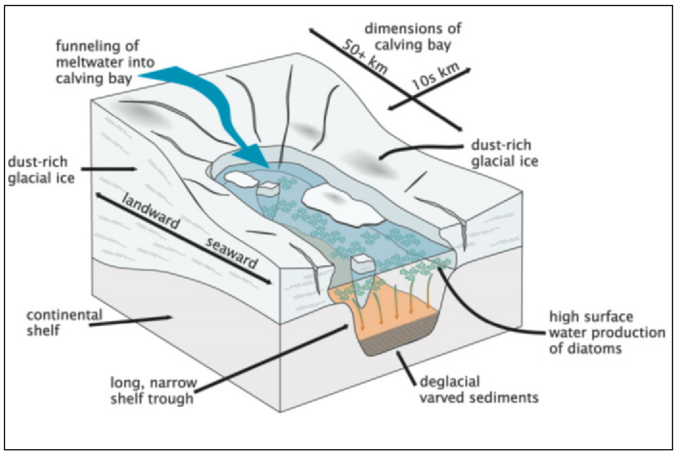

Figure 4. Generalized calving bay reentrant model. Glacial meltwater is funneled into the restricted bay, resulting in high springtime primary productivity. Summer and fall input of more terrigenous material results in a seasonal alternation of downward flux that is preserved at the seafloor as varves.

CALVING BAY REENTRANT MODEL

Model Basics

Hughes (2002) reviewed the history of the term "calving bay" and noted that calving bays—embayments characterized by rapid calving of glacial ice—could speed up ice sheet retreat as the bays migrated landward. More recently, Domack et al. (2006) presented a calving bay reentrant model for the Palmer Deep (western Antarctic Peninsula; Fig. 1) based on regional swath bathymetry that highlighted the importance of the physical configuration of the Palmer Deep as a control on the pattern of ice retreat, with retreat occurring much more rapidly over the deep trough while ice remained grounded on the shallower banks to either side. This style of bathymetry favors the formation of calving bay reentrants during deglaciation as well as the development of sedimentary regimes in which varved sediments were deposited (Fig. 4). Initial deglaciation was focused on these troughs, while ice remained grounded on the sides in a manner similar to that described for the Palmer Deep (Domack et al., 2006).

Varves are described from post-glacial sediments of the Palmer Deep (Domack, 2002; Leventer et al., 2002; Nederbragt and Thurow, 2001; Pike et al., 2001; Domack et al., 2006; Maddison et al., 2005) and, like the Iceberg Alley couplets, are composed of an alternation of primarily *Chaetoceros* resting spore–dominated diatom ooze and siliciclastic-rich diatom mud. A total of 176 couplets were observed in the ~3.5 m deglacial section from the Palmer Deep (Domack, 2002), similar to the number of couplets in the Iceberg Alley section, indicating that both Palmer Deep and Iceberg Alley persisted as calving bay reentrants for about two centuries. The Svenner Channel has the least extensive suite of varves, indicative of a calving bay reentrant that lasted on the order of only a decade. In between these two extremes is the Nielsen Basin, with a calving bay reentrant lasting several decades.

In the Mertz-Ninnis Trough, alternating laminated (but not varved) and bioturbated sediments, which occur at the glacial/interglacial boundary, indicate a different history of deglaciation. The lack of varves in the Mertz-Ninnis Trough is probably related to its geometry. The large-scale and open bathymetry of this trough most likely prevented the formation of both a calving bay reentrant and a restricted oceanographic setting, which resulted in the varved sequences observed at the other sites.

Elevated Primary Productivity, Preservation of Signal

The formation of varves requires the preservation of couplets that record the repetition of an annual cycle of deposition. The calving bay reentrants provided conditions that satisfied both the development of an annual signal and its preservation at the seafloor. Leventer et al. (2002) present a model for the Palmer Deep, applicable to the East Antarctic margin sites, that describes the annual succession of a springtime *Chaetoceros* bloom followed by lower levels of summer production and the added influx of terrigenous material. Here, we further develop that model with a focus on the lower, biogenic half of each couplet, which is dominated by diatoms of the genus *Chaetoceros*, a group commonly interpreted as indicative of high productivity. Convergence of surface slopes in the calving bay reentrants would have focused surface melt, concentrating large volumes of low salinity, low density glacial meltwater via preferential surface flow. Strong stabilization of the upper water column and reduced mixing of phytoplankton out of the photic zone would have enhanced primary production (Leventer et al., 1996). In contrast, along most parts of the coastline, spring and summer meltwater would not have been focused, and may have been diluted relatively quickly by wind-induced mixing, with consequently smaller phytoplankton blooms.

A second consideration is the potential for increased nutrient concentration within the upper water column. Stickley et al. (2005) review potential nutrient sources within the Iceberg Alley setting, including nutrients released from melting sea and glacial ice, and from incursions of Upper Circumpolar Deep Water. We suggest that the dominance of *Chaetoceros* in springtime blooms during the glacial-Holocene transition might have been influenced by the micronutrient iron. Since the early work of Martin et al. (1990), studies document the ability of introduced iron to induce phytoplankton blooms in the Southern Ocean (Coale et al., 2004; Boyd et al., 2000; Boyd and Law, 2001).

We suggest glacial meltwater as the source of iron fertilization within the calving bay reentrant. Melting of the ice margin occurs by surface melt percolating down through firn and refreezing until superimposed ice is formed, or if the melting is extensive enough, an ablation zone develops at the end of the melt season. The ice in the ablation zone melts at the surface, leaving behind and thus concentrating debris, including the dust contained within the ice. As meltwater drains out into the calving bay reentrant, it carries a suspended load with concentrated debris content. This debris includes iron-rich dust, delivered to the ice sheet via aeolian transport (Rea, 1994; Wolff et al., 2006). Under these conditions, each spring, iron-rich meltwater introduced into the surface waters could have fertilized extremely large algal blooms.

The dominance of *Chaetoceros* in the spring bloom lamination is relevant in light of the work of Tsuda et al. (2003), who performed an iron enrichment experiment in the subarctic Pacific that led to a *Chaetoceros*-dominated bloom. They speculate that the high doubling rate of this small centric diatom resulted in the shift from an assemblage previously dominated by a pennate diatom. The situation described by Tsuda et al. (2003) might have parallels to the observations of the massive annual spring sedimentation of *Chaetoceros* resting spores in deglacial sediments from both the East Antarctic margin and the Palmer Deep.

The role of decreased ventilation of the deeper waters in preservation of the seasonally produced signal requires further investigation. In combination with the restricted geographic setting, low-salinity glacial meltwater at the sea surface may have decreased the density of brines resulting from winter sea ice formation, thus preventing them from displacing basin waters. Consequently, basin waters may have had limited ventilation, resulting in anoxia (Willmott et al., 2006), an environment that favors the preservation of laminated facies.

Chronology of Deglaciation

Chronologies for these cores are based on accelerator mass spectrometry (AMS) radiocarbon-dated carbonate material and decalcified organic matter (GSA Data Repository, Table DR2 [see footnote 1]). All samples were analyzed at the Lawrence Livermore National Laboratory Center for Accelerator Mass Spectrometry. Radiocarbon ages were calibrated using CALIB 4.42 (©1986–2004, M. Stuiver and P.J. Reimer).

A chronology for JPC43B, in Iceberg Alley, presented by Stickley et al. (2005), is briefly described below. A total of 11 samples were AMS radiocarbon dated (Table DR1 [see footnote 1]). Radiocarbon ages were calibrated assuming a local reservoir age of 1700 yr ± 200 yr (P. Sedwick, 2004, personal commun.); the reservoir age is based on radiocarbon analyses of kasten cores from Iceberg Alley. The carbonate ages are slightly younger than the decalcified total organic carbon (TOC) dates. One anomalously old date from 17.25 to 17.30 m was disregarded and a second order polynomial was fit to the curve to develop the chronology. The varved section of the core, from 19.13 to 23.96 m, thus dates to ca. 11,200–11,400 cal yr B.P. High sediment accumulation rates inferred from the dated samples at the base of the core support the interpretation of the laminated sediments as varves.

The chronology for JPC24, Svenner Channel, was developed using seven AMS radiocarbon dates (Table DR1 [see footnote 1]). Radiocarbon ages were calibrated assuming a local reservoir age of 1280 ± 200 yr, based on the age of a kasten core top at the same location. Again, the carbonate ages are slightly younger than the decalcified TOC dates. A best-fit line between the shell dates constrains the chronology of the lower part of the core; the varved section of JPC24 (15.58–15.80 m) dates to ca. 10,800 cal yr B.P. In JPC25, the glacial-interglacial transition occurs at 8.88 m. Three dates derived using carbonate shell material place deglaciation at ca. 11,100 cal yr B.P.

Previous work developing radiocarbon-based chronologies for sediment cores from the Mertz-Ninnis Trough has faced difficulties based on the presence of reworked organic material (Domack et al., 1989; Harris and Beaman, 2003); consequently, radiocarbon dates for the Mertz-Ninnis Trough have not been calibrated. Finally, at this time, no radiocarbon data are available for the Nielsen Basin core.

Timing of Deglaciation and Its Significance

Data constraining the timing of deglaciation along the East Antarctic margin are important in terms of their implication for the role of Antarctica in contributing to sea level rise at the end of the last glacial. The Barbados record of sea level during the late Quaternary (Fairbanks, 1989; Bard et al., 1990) documents two events of rapid sea-level rise at the end of the last glacial. The first, termed meltwater pulse 1A (mwp-1A), has been dated to ca. 14,200 cal yr B.P., and the second, meltwater pulse 1B (mwp-1B), began ca. 11,000 cal yr B.P. (Fairbanks, 1989). The source of the water for sea level rise during mwp-1A has been debated, with some researchers suggesting that the meltwater contributing to sea level rise came from the receding Laurentide Ice Sheet (Kennett and Shackleton, 1975; Leventer et al., 1982; Keigwin et al., 1991; Fairbanks et al., 1992; Peltier, 1994) and others suggesting an Antarctic source (Clark et al., 1996, 2002; Weaver et al., 2003; Bassett et al., 2005). Speculation concerning an Antarctic source was initiated by Clark et al. (1996), who reviewed the data indicating a Laurentide meltwater source; their reevaluation indicated that the Laurentide ice sheet could not be the sole source of mwp-1A.

Peltier (2005) recently reviewed this problem and its significance. Pinpointing the source of mwp-1A is crucial for several reasons. First, meltwater discharge, through its impact on oceanic circulation, can influence global climate. For example, Weaver et al. (2003) suggest that an Antarctic source for mwp-1A could have resulted in increased North Atlantic Deep Water production and the initiation of the Bølling-Allerød warm interval. Second, suggestions of an Antarctic source for mwp-1A are based on geophysical models of Earth's response to spatial differences in deglaciation, which can be evaluated through analysis of sea level records (Clark et al., 2002; Peltier, 2005). However, published geologic data from Antarctica do not support an Antarctic source for mwp-1A (Ackert et al., 1999; Conway et al., 1999; Domack et al., 1999; Baroni and Hall, 2004), resulting in a need to reconcile geophysical models and geologic data.

Our East Antarctic margin data constrain the timing of the calving bay reentrant phase of deglaciation and the deposition of the deglacial sediment facies. The data indicate that deglaciation on these portions of the margin began ca.11,500 cal yr B.P., too late to contribute to the sea level rise indicated by mwp-1A, but potentially early enough to contribute to mwp-1B. Other East Antarctic margin marine records provide a similar time frame for deglaciation. Domack et al. (1991) note the initiation of open-marine conditions at Ocean Drilling Project (ODP) Leg 119 Site 740 in the Svenner Channel at 10,700 yr B.P. (dates corrected for Antarctic marine reservoir effect [AMRE] but not calibrated). Work by Harris and O'Brien (1998) and Sedwick et al. (1998, 2001) on the Mac.Robertson shelf places deglaciation at ca. 11,000 yr B.P. (dates corrected for AMRE but not calibrated), based on the initiation of deposition of siliceous muds and oozes. Of particular interest is the Sedwick et al. (2001, p. 223) observation of a bloom of *Chaetoceros* resting spores "associated with the retreat of permanent ice cover" in outer Iceberg Alley at 10,800 yr B.P. (dates corrected for AMRE but not calibrated).

CONCLUSIONS

Multibeam swath bathymetric data and lithologic and chronologic data from JPCs from the East Antarctic margin demonstrate the applicability of a calving bay reentrant model of deglaciation. These cross-shelf troughs were sites of early ice retreat and extremely high productivity, while glacial ice remained grounded on shallower portions of the shelf. High biogenic and terrigenous flux for periods ranging from only a decade to almost two centuries are recorded by varved sediments, in which nearly monospecific *Chaetoceros* layers alternate with slightly more terrigenous layers with mixed diatom assemblages. We propose that springtime blooms of *Chaetoceros* were generated within a stabilized water column characterized by the introduction of iron-rich meltwater into surface waters. Advection within the physically isolated conditions of the calving bay reentrant helped concentrate biogenic flux to the seafloor. The intervening post-bloom summer-fall laminae were formed through downward flux of terrigenous material sourced from the melting glacial ice combined with mixed diatom assemblages. The thickness of the deglacial varved unit—which does not exist at the Mertz-Ninnis Trough site but ranges from tens of centimeters in the Svenner Channel and Nielsen Basin to meters in Iceberg Alley—is driven by the specific configuration of each trough, which controlled the temporal persistence of the calving bay reentrant, ranging from decades to centuries. These records suggest that the Palmer Deep, West Antarctica, and these sites in East Antarctica responded in a similar way to sea level forcing at the end of the last glacial. Radiocarbon data from these cores place deglaciation along the East Antarctic margin between ca. 10,500–11,500 cal yr B.P. These data demonstrate that deglaciation at the study sites followed mwp-1A, thus indicating that East retreat of ice from the Antarctic margin was not the major contributor to this pulse of meltwater.

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