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1. Leg 188 Synthesis: Transitions in the Glacial History of the Prydz Bay Region, East Antarctica, from ODP Drilling[1]

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ABSTRACT

Drilling during Leg 119 (1988) and Leg 188 (2000; Sites 1165–1167) of the Ocean Drilling Program (ODP) provides direct evidence for long- and short-term changes in Cenozoic paleoenvironments in the Prydz Bay region. Cores from across the continental margin reveal that in preglacial times the present shelf was an alluvial plain system with austral conifer woodland in the Late Cretaceous that changed to cooler Nothofagus rainforest scrub by the middle to late Eocene (Site 1166). Earliest recovered evidence of nearby mountain glaciation is seen in late Eocene–age grain textures in fluvial sands. In the late Eocene to early Oligocene, Prydz Bay permanently shifted from being a fluvio-deltaic complex to an exclusively marine continental shelf environment. This transition is marked by a marine flooding surface later covered by overcompacted glacial sediments that denote the first advance of the ice sheet onto the shelf. Cores do not exist for the early Oligocene to early Miocene, and seismic data are used to infer the transition from a shallow to normal depth prograding continental shelf with submarine canyons on the slope and channel/levees on the rise.

Cores from the continental rise at Site 1165 show long-term (millions of years) early Miocene and younger decreases in sedimentation rates as well as short-term (Milankovitch periods) cyclicity between principally biogenic and terrigenous sediment supply—resulting from the cyclic presence of onshore glaciers and changes in ocean circulation. Middle Miocene transitions include rapid decreases in sedimenta-

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tion rates, increased ice-rafted debris, shifts in clays and other minerals, and regional erosion of the slope and rise. These transitions may reflect enhanced glacial erosion and reduced glacial meltwater from progressively colder ice. At this time, seismic data show that depocenters began to shift from the outer continental rise to the base of the continental slope coincident with the initial stages of the glacial erosion and overdeepening of the continental shelf.

During the late Miocene to early Pliocene there was a transition to greater subglacial activity on the shelf and more pronounced cyclic facies variations on the continental rise. At this time, severe glacial morphologies initiated on the shelf with the erosion of Prydz Channel and other troughs by fast-moving ice and the deposition of overcompacted glacial diamictons by slow-moving ice on adjacent banks. The Prydz trough-mouth fan also began to form with alternating deposition of debris flows (ice at shelf edge) and muddy units (reduced ice) (Site 1167). The fan also records a transition during the late Pleistocene for times younger than 780 k.y. when short-term glacial variations continued but ice reached the shelf edge only a few times.

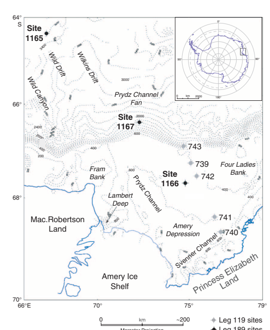
Both short-term and long-term transitions characterize the Cenozoic evolution of the Prydz Bay region from the Cretaceous nonglacial to late Neogene full-glacial paleoenvironments. These transitions are known only from ODP cores, and further insights will require additional drilling.

INTRODUCTION

Three sites (1165, 1166, and 1167) were drilled during Ocean Drilling Program (ODP) Leg 188 in the Prydz Bay region of Antarctica (Fig. F1) to achieve three principal objectives:

1. Date the earliest evidence of glacial activity in the region and decipher the Paleogene environment of Antarctica by drilling on the continental shelf;
2. Link Oligocene and younger events in the East Antarctic Ice Sheet with changes in the Southern Ocean by drilling a transect of holes across the continental shelf, slope, and rise; and
3. Acquire a record of late Miocene and younger ice advances to the shelf edge and interglacial periods by drilling into the trough-mouth fan built by advances of the Lambert Glacier–Amery Ice Shelf.

F1. Prydz Bay region, p. 28.



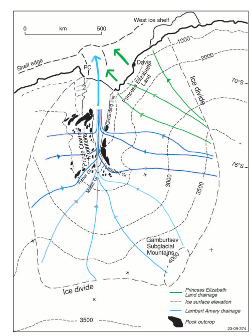
Leg 188 is also notable for the many technical advances made in successfully acquiring the desired cores under unusually harsh operating conditions. This was the first Antarctic drilling leg to work without the aid of a support ship and conduct logging-while-drilling operations in polar-glacial sediments. Site 1165 is the deepest drill hole in Antarctica to date, and Site 1167 is the first drill hole to sample deeply into a polar trough-mouth fan. These accomplishments and others lay the path for future Antarctic one-ship drilling operations to more fully decipher south polar paleoenvironmental history.

The East Antarctic Ice Sheet has a long-lived record of growth and decay since the Paleogene (Barron et al., 1991) and is now the largest ice mass on Earth. The ice sheet has been a significant driving force in global climates and sea level changes and in ocean and atmospheric circulation. The long-term cooling of the world's oceans and climates over the past 50 m.y. is clearly recorded in the ocean $\delta^{18}O$ records from around Antarctica and elsewhere. However, the variable isotopic records from the deep ocean have not yet been fully and unequivocally linked to equivalent proximal (i.e., nearshore) records of glacier volume fluctuations.

Drilling in Prydz Bay was part of a carefully coordinated effort by the Antarctic Offshore Stratigraphy Project (ANTOSTRAT) to drill several widely separated segments of the Antarctic continental margin where large glacier fluctuations are inferred in order to achieve a circum-Antarctic Cenozoic glacial history. The first drilling leg of this effort, ODP Leg 178 (Barker, Camerlenghi, Acton, et al., 1999; Barker, Camerlenghi, Acton, and Ramsay, 2002), was drilled across the Antarctic Peninsula margin. Sedimentary sections were sampled in similar paleodepositional environments to those drilled during Leg 188 in Prydz Bay (O'Brien, Cooper, Richter, et al., 2001). However, the onset of glaciation in the Antarctic Peninsula region is thought to be younger (middle Miocene age onset) and more sensitive to short-term changes (Barker and Camerlenghi, 2002) than the earliest stages of glaciation in the Prydz Bay region (late Eocene) (Shipboard Scientific Party, 2001a).

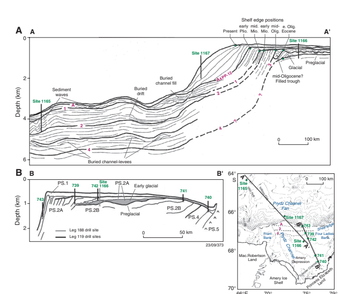
The Lambert Glacier–Amery Ice Shelf drainage system now drains into Prydz Bay from the south, and this system is believed to have the longest-lived record for the Cenozoic Antarctic Ice Sheet (Fig. F2). Leg 188 was drilled in that part of Prydz Bay that would likely sample the first advances of the ice sheet. ODP Leg 119 had previously drilled a transect of holes in Prydz Bay (Barron, Larsen, et al., 1989, 1991), but these holes did not penetrate fully into and through the early glacial sections to reach preglacial Cenozoic rocks, as Leg 188 did at Site 1166.

F2. Lambert Glacier drainage basin, p. 29.



The geologic records obtained from the drilling are predictably more complete and continuous at the continental rise drill site (Site 1165) than on the continental shelf and slope. We recovered a mostly complete section of early Miocene and younger hemipelagic and pelagic rocks at Site 1165. On the shelf, large parts of the section have been eroded, as is evident from seismic reflection data and from the Leg 119 and Leg 188 drill cores, which comprise rocks from subaerial, shallow-marine, and subglacial paleoenvironments (Fig. F3) (Barron et al., 1991; Shipboard Scientific Party, 2001a). The sediments eroded from the continental shelf and onshore regions during the Neogene and Quaternary were transported by grounded glaciers to the continental shelf edge, where they form the debris flow deposits sampled on the continental slope at Site 1167.

F3. Seismic profiles across the continental margin, p. 30.



The resulting morphology of the Prydz Bay continental margin, with highly eroded shelf (Cooper et al., 1991a; O'Brien and Leitchenkov, 1997), trough-mouth fan on the slope (O'Brien et al., 1999), and canyons and large drift deposits on the rise (Kuvaas and Leitchenkov, 1992), reflects the cumulative Paleogene and younger influence of ice and ocean currents on paleodepositional environments. Similar morphologies are observed on other parts of the Antarctic margin (e.g., Cooper et al., 1991b) and were also drilled in part on the Antarctic Peninsula during Leg 178 (Barker, Camerlenghi, Acton, et al., 1999; Barker, Camerlenghi, Acton, and Ramsay, 2002).

In this paper we incorporate highlights from the initial drilling reports (O'Brien, Cooper, Richter, et al., 2001) with postcruise studies to focus on some key thematic issues related to Leg 188 science objectives.

REGIONAL SETTING

Geologic and Glaciologic Features

Prydz Bay lies at the seaward end of the Lambert rift graben, which extends ~500 km south of Prydz Bay. The adjacent rift-flank mountains (e.g., Prince Charles Mountains) (Fig. F2) are mostly ice covered and have a visible relief of up to 3500 m. Most of the basement rocks of the region are high-grade Precambrian metamorphic and intrusive rocks (Tingey, 1991; Mikhalsky et al., 2001) with isolated onshore exposures of Paleozoic granite, mafic dykes, Permian sediments, Paleogene volcanics, and Oligocene and younger glacial sediments (Tingey, 1991; Hambrey and McKelvey, 2000a, 2000b).

The offshore area is underlain by the Prydz Bay sedimentary basin, which contains as much as 12 km of sediment (Cooper et al., 1991a) and is part of the extensive rift system that also includes the Lambert Graben. The broad pattern of ice and sediment movement in the region is controlled by the Lambert Graben. The rifting history extends back to at least Cretaceous (Truswell, 1991) and probably back to Permo-carboniferous (Arne, 1994). Sediments that probably occupy the Lambert Graben crop out in the Amery Oasis and were described by Fielding and Webb (1995) and McLoughlin and Drinnan (1997a, 1997b). During Leg 119 a pre-Cenozoic sedimentary section was sampled in the Prydz Bay Basin and included an undateable floodplain redbed unit similar to Triassic sediments onshore (Site 740) and a middle Aptian alluvial coal-bearing sequence (Site 741) (Turner and Padley, 1991). Core samples from the Mac.Robertson shelf indicate nearby outcrops of Jurassic and Early Cretaceous sedimentary rocks (Truswell et al., 1999). Cenozoic rocks were found at all Leg 119 shelf sites, and like Leg 188 Site 1166, they include late Neogene glacial diamictites with interbedded diatomaceous muddy units that unconformably overlie early Oligocene glaciomarine rock or pre-Cenozoic units.

A diverse suite of sediment was sampled during Leg 188, as detailed in O'Brien, Cooper, Richter, et al. (2001). In general, rocks from the shelf (Site 1166) include upsection nonglacial silts, massive sands, shallow-marine claystone, diamictons, and diatomaceous muddy units. On the slope (Site 1167), debris flow units separated by muddy layers were recovered; on the rise (Site 1165), hemipelagic and pelagic sediments were found. As we discuss below, Leg 188 rocks include components recycled from the onshore and offshore shelf sedimentary sections.

The Lambert Graben is now covered by the Lambert Glacier–Amery Ice Shelf ice drainage system, which drains ~16% of the East Antarctic Ice Sheet or 10% of all Antarctic ice outflow (Allison, 1979; Fricker et al., 2000) (Fig. F2). The drainage area includes the Gamburtsev Subglacial Mountains (Fig. F2), which may be a nucleation point for earliest Antarctic glaciation (Huybrechts, 1993) and the Prince Charles Mountains, which show evidence of glaciation since at least the Oligocene (Hambrey and McKelvey, 2000a, 2000b). Ice flows into Prydz Bay principally from the Lambert–Amery system, but there are small volumes that flow in from glaciers along Princess Elizabeth Land on the southeastern side of Prydz Bay and minor amounts from Mac.Robertson Land north of the Amery Ice Shelf (Fig. F2).

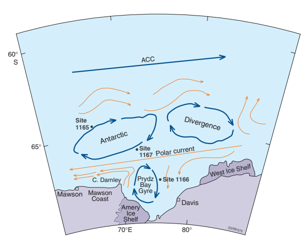
Offshore Morphology and Oceanography

The Prydz Bay continental shelf, like other polar shelves, is deepest (>800 m water depth) adjacent to the coast near the front of the Amery Ice Shelf and in large cross-shelf troughs (e.g., Prydz Channel). The seabed beneath the southern end of the Amery Ice Shelf reaches depths of 2500 m below sea level (Fricker et al., 2000). Water depths are shallowest near the continental shelf edge and on the top of banks adjacent to the large cross-shelf troughs (e.g., Four Ladies Bank, which is as shallow as 200 m). In nearly all areas, the water depths exceed the common 100-m average water depth of nonpolar shelves. Results from Leg 188 drilling help to explain the uncommon morphology of the shelf, as discussed below.

The Prydz Channel Fan lies at the mouth of Prydz Channel and forms a large seaward bulge in bathymetric contours (Figs. F1, F2). The fan extends to water depths of ~2700 m with a slope of ~2° and is as thick as ~1100 m (O'Brien et al., this volume). O'Brien and Harris (1996) infer that the Prydz Channel and Prydz Channel Fan developed initially in the early to middle Pliocene when ice volumes and ice flow directions changed, leading to the Lambert–Amery system forming a fast-flowing ice stream that cut Prydz Channel and deposited debris on the upper slope. The Prydz Bay continental slope is cut by submarine canyons that generally trend north to north-northeast on the east side of the Prydz Channel Fan and north to north-northwest on the west side of the fan. Elongate sediment-drift deposits of several hundred meters relief lie adjacent to canyons on the rise and also trend generally northward (e.g., Wilkins and Wild Drifts) (Fig. F1). Site 1165 was drilled into the flanks of the Wild Drift, adjacent to Wild Canyon, which has its head on the continental slope off Mac.Robertson Land (Kuvaas and Leitchenkov, 1992).

Modern ocean circulation in the Prydz Bay–Cooperation Sea region has many components. Surface circulation in the bay is a cyclonic gyre with cold-water inflow from the east over the general area of Site 1166 and outflow along the west side of Prydz Bay (Fig. F4) (Smith et al., 1984; Wong, 1994). There is relatively little high-saline deep water due to Prydz Bay's geography and bathymetry (Smith et al., 1984). These two factors lead to little downslope bottom water current activity beyond the shelf edge. The deepwater movements on the continental slope and rise (e.g., over Sites 1167 and 1165, respectively) are attributed to the three large-scale ocean systems: the Polar Current, moving west near the shelf edge; the Antarctic Divergence, producing cyclonic gyres over the slope and inner rise; and the Antarctic Circumpolar Current, moving east over the outer rise and beyond. The interaction of the near-seafloor currents (e.g., downslope density currents and along-slope ocean currents) is believed to control slope and rise sediment deposition in the Prydz Bay region (Kuvaas and Leitchenkov, 1992), as elsewhere such as the Antarctic Peninsula region (Rebesco et al., 1997).

F4. Present-day ocean currents in the Prydz Bay region, p. 31.



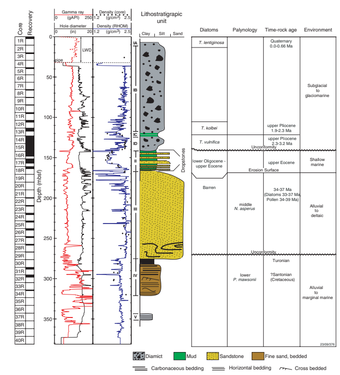
EAST ANTARCTIC GLACIAL TRANSITIONS

Mesozoic to Paleogene: Glacial Initiation

A primary objective of Leg 188 drilling was to decipher the glacial history of the East Antarctic Ice Sheet in the Prydz Bay region, which is considered a drainage area for the earliest Cenozoic ice sheets. Leg 119 drilling in Prydz Bay (Barron, Larsen, et al., 1989, 1991) established that a grounded glacier covered at least part of Prydz Bay (Sites 739 and 742) during early Oligocene. Leg 188 reconfirmed the Leg 119 assertion and acquired the data to show that the transition from nonglacial to glacial paleoenvironments occurred in the late Eocene or earlier. Further, Leg 188 drilling recovered lithologic and biostratigraphic data to help document the morphologic evolution of the Prydz Bay continental margin, which resulted from Cenozoic glaciation.

Until the early Oligocene, the Prydz Bay mid-continental shelf around Sites 1166 and 742 was part of a subaerial alluvial plain that filled the Lambert rift system including the Prydz Bay Basin. Seismic facies and links to core data support this concept (**Erohina et al.,** this volume; **Handwerger et al.,** this volume). Cretaceous sediments were recovered at Sites 741 and 1166 in Prydz Bay (Fig. **F5**). Site 741 yielded alluvial sediments consisting of channel sandstone beds, floodplain siltstone and claystone, and sporadic coal beds (Turner and Padley, 1991). Palynomorphs indicate an Early Cretaceous (?Albian) age for these units (Truswell, 1991). Black to gray carbonaceous claystone and siltstone were recovered from lithostratigraphic Units IV and V at Site 1166, giving a Turonian–?Santonian microflora (*Phyllocladidites mawsonii* Zone) (**Macphail and Truswell,** this volume a) (Fig. **F5**). Geophysical logs suggest some interbedded sandstone units. The combination of lithologies, palynomorphs, and organic matter (**Claypool et al.,** this volume) indicates deposition in an alluvial plain to marginal marine lagoonal setting. Mesozoic sediments intersected at Site 1166 contain Cretaceous (Turonian) palynofloras dominated by gymnosperms with rare dinoflagellate cysts, indicating a coastal plain environment with a conifer woodland and rare freshwater swamps analogous to vegetation near the northern limit of tree growth in the modern Arctic (**Macphail and Truswell,** this volume a). The fossil assemblage implies year-round moderate humidity and temperatures. There is no evidence of Cretaceous glaciation from Leg 188 cores.

F5. Site 1166 summary data, p. 32.



The only other late Mesozoic to early Paleogene rocks recovered offshore in the region are core samples containing recycled palynofloras from the outer continental shelf near Mac.Robertson Land. These samples indicate nonmarine sedimentation on this part of the margin from the Jurassic to the Early Cretaceous (Aptian?) (Truswell et al., 1999), predating Prydz Bay ODP sections.

At Site 1166, an unconformity separates the Cretaceous and middle–late Eocene units (Fig. **F5**). On seismic data, this unconformity shows almost no relief and only occasional truncation (**Erohina et al.,** this volume) but nonetheless represents a break of at least 47 m.y. (top of the *P. mawsonii* Zone to the base of the *Nothofagidites asperus* Zone) (Young and Laurie 1996). Sediments deposited during this hiatus may be preserved elsewhere in Prydz Bay and the Lambert Graben because **Quilty** (this volume) and Quilty et al. (1999) report that fragments of the Cretaceous bivalve *Inoceramus* are reworked into Cenozoic sediments. Stilwell et al. (2002) reported whole Cretaceous bivalves in boulders within Cenozoic glaciomarine sediments near Beaver Lake in the northern Prince Charles Mountains. These fragments indicate marine conditions in the Lambert Graben at some stage during Cretaceous.

In the middle to late Eocene, alluvial or deltaic sediments were again deposited (Fig. **F5**) (Site 1166 Unit III of Shipboard Scientific Party, 2001c; Unit PS2A2 of **Erohina et al.,** this volume). Their texture, downhole logging signature, and distribution suggests the upper part of the unit was deposited mostly in fluvial channels and also in flood or tidal basins or lagoons (Fig. F5). The lowest part of the alluvial-deltaic unit contains contorted fragments of carbonaceous sands that have yielded Cretaceous palynomorphs. These fragments are likely remnants of fluvial bank collapse blocks. The unit contains a middle to late Eocene palynoflora that suggests the Prydz Bay coastal plain was covered by a low-growing scrub of gymnosperms and angiosperms (*Nothofagus*) analogous to stunted rainforest scrub vegetation presently growing near the alpine limits of similar floras in Tasmania and Patagonia (**Macphail and Truswell,** this volume a). The flora implies humid conditions and probably microthermal (cool–cold) temperatures at sea level. More precise temperature estimates are not possible because the major components of this flora can tolerate a wide range of temperatures. With the shift of vegetation to cooler-climate species in middle to late Eocene, the first lithologic evidence for nearby glaciation in the rift-flank mountains to the south is seen in sand grain surface textures in late Eocene massive sand units (Strand et al., 2003).

In the late Eocene to early Oligocene the central part of Prydz Bay (e.g., Sites 739, 742, and 1166) permanently shifted from a fluviodeltaic complex with rainforest scrub to an exclusively marine continental shelf environment. The massive fluvial sands of Unit III at Site 1166 grade upward into alternating sand–clay layers in Unit II (Fig. F5) with increasing marine dinoflagellates (**Macphail and Truswell,** this volume a) and signal a change to tidal influence and a relative subsidence of the area to near sea level. The alluvial or delta plain sediments (Site 1166; Unit III) are cut by an erosion surface with relief of ~50 m in places (**Erohina et al.,** this volume). Glaciomarine mudstone and sandstone containing diatoms, dinoflagellates, and lonestones onlap the erosion surface in the area around Site 1166. The glaciomarine unit (Unit II of Shipboard Scientific Party, 2001c; Unit PS.2A1 of **Erohina et al.,** this volume) is onlapped by glaciomarine diamictites encountered at Site 742 (**Erohina et al.,** this volume; Hambrey et al., 1991).

The succession of Paleogene events in this part of Prydz Bay are as follows:

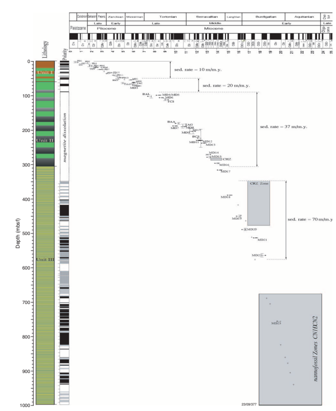
1. Deposition of an alluvial fan or delta with braided streams and some glacially derived detritus from the Lambert Graben hinterland;
2. Relative sea level fall and cutting of an erosion surface;
3. Relative sea level rise, producing an onlapping sequence of marine sediments;
4. Delivery of lonestones into Prydz Bay by floating ice; and
5. Glacier advance into Prydz Bay, depositing till and waterlain till (Hambrey et al., 1991).

The available paleontological evidence points to these events occurring around the Eocene/Oligocene boundary. The palynomorphs within the Paleogene alluvial and glaciomarine units at Site 1166 (Units III and II) (Fig. F5) bracket the ages of 33.9 and 39.1 Ma, based on age estimates of the *N. asperus* Zone in eastern Australia (**Macphail and Truswell,** this volume a; Young and Laurie, 1996). Diatoms yield age estimates of 33–37 Ma (Shipboard Scientific Party, 2001c). Paleontological evidence at Sites 739 and 742 yield similar ages; however, Sr isotope ages on shell fragments from Site 739 yield ages of 22.7–29.3 Ma (Thierstein et al., 1991). Lavelle (2000) reported revised Sr dates for these shells of ~33 Ma. Such an age would resolve the discrepancy; however, his brief report has not been followed by full publication.

Oligocene–Middle Miocene

Rocks from times between the early Oligocene and early Miocene were not recovered during ODP Prydz Bay drilling but were cored during ODP Legs 119 and 120 on the 1000-km distant Kerguelen Plateau. There, principally pelagic marine units were sampled (Barron, Larsen, et al., 1989; Wise, Schlich, et al., 1989). Seismic reflection data (e.g., Fig. F3) and modeling provide the only evidence that the early Oligocene continental shelf continued to prograde and gradually deepen to normal continental shelf water depths as a result of shelf erosion, increasing ice loading onshore, and sediment loading of the continental slope (e.g., ten Brink et al., 1995). During this period, sea level changes may have resulted in periodic subaerial erosion of the shelf with sediments being carried down canyons on the slope into widespread channel-levy systems and drift deposits of the continental rise (e.g., Kuvaas and Leitchenkov, 1992; Cooper et al., 2001).

The complete section of drift deposits to the basal regional reflector P3 (which is the same as reflector P2 of Kuvaas and Leitchenkov, 1992) was drilled at Site 1165 (**Handwerger et al.,** this volume). The Site 1165 cores establish that the early to middle Miocene was a time of abundant terrigenous sedimentation on the continental rise, with sediments rapidly building the Wild Drift on the flank of the long-established Wild Canyon system, which originates from near the western side of Prydz Bay (Figs. **F1, F6**). The relative amounts of downslope and along-slope sediment transport cannot be established unequivocally from drill cores. But regional sediment distribution patterns from seismic reflection data (e.g., Fig. **F3**) show that the continental slope and rise are underlain by thick sediment bodies and large and extensive canyon systems that have been in the same place throughout most of the Cenozoic; these observations indicate active nearby sediment source(s). The Lambert Graben is a large, long-lived Paleozoic and younger rift valley that is now, and likely has been, a conduit for a large drainage basin—further indication that Site 1165 sediments were likely derived principally from onshore Prydz Bay.



F6. Age-depth plot, Site 1165, p. 33.

Studies in the Lambert rift-flank Prince Charles Mountains show that polythermal glaciers existed in at least late Oligocene to early Miocene and likely provided abundant suspended sediments to the shelf areas and beyond (Hambrey and McKelvey, 2000a, 2000b). The biogenic–terrigenous cyclicity in the continental rise sediments at Site 1165 in the early to middle Miocene may indeed record the variable terrigenous sediment supply from onshore glacier fluctuations (Shipboard Scientific Party, 2001a; **Rebesco,** this volume). The palynological record for Site 1165 indicates that southern beech vegetation (*Nothofagus*) may have existed nearby into early Miocene (**Macphail and Truswell,** this volume b), suggesting that climates and glacial conditions were more temperate than today. **Pospichal** (this volume) reports scattered warmer-water nannoplankton and thin chalk horizons through the Miocene at Site 1165, suggesting the intrusion of warmer waters into the area.

The continental shelf, which before the middle Miocene prograded principally by sediment bypass and erosion of the shelf's Paleogene topset strata, began to aggrade by deposition of topset strata that formed glacial banks. The aggrading initiated as the shelf subsided more rapidly than before under the increasing inner-shelf erosion, greater flexural loading by widespread onshore ice build up, and glacially derived slope sediments (e.g., ten Brink et al., 1995) (Fig. **F3**). The bypass sediments are those on the continental rise in lithostratigraphic Unit III

(999–308 meters below sea floor [mbsf]) at Site 1165. This is based on identification of Permian, Jurassic, Cretaceous, and Paleogene recycled pollen and spores from samples below 807 mbsf (**Macphail and Truswell,** this volume b) and on the rare occurrence of recycled phytoliths (291–601 mbsf) that are comparable to those in modern and equivalent older trees/shrubs, grass, and ferns (**Thorn,** this volume).

On the rise the early to middle Miocene sediments of Site 1165 Unit III are predominantly fine-grained mudstones with a high proportion of quartz and detrital feldspar, probably in the silt size fraction (Shipboard Scientific Party, 2001b). The main clay minerals in samples are illite and kaolinite with minor chlorite. Mechanical weathering of igneous and metamorphic rocks typically delivers illite to sedimentary basins (Ehrmann et al., 2003), so is not unexpected in sediments adjacent to a glaciated landmass. Kaolinite, however, is typically a product of chemical weathering. Its presence in continental rise sediments off Prydz Bay that postdate the initial early Oligocene glaciation suggests that either kaolinite was recycled from older sedimentary basins on the shelf or that chemical weathering continued onshore through the early Miocene.

The input from shelf sedimentary basins to the continental rise in response to climatic change, ice buildup, and increased erosion becomes clear in the middle Miocene section at the distinct boundary between lithostratigraphic Units III and II at 307 mbsf at Site 1165 (Fig. **F6**). Total clay minerals increase as detrital plagioclase decreases, diatoms become more common, and the number of lonestones also increases (O'Brien, Cooper, Richter, et al., 2001). Also, glauconite grains, likely from the shelf and reworked pseudomorphs and tests of Paleogene benthic foraminifers, appear in the sediments (**Quilty,** this volume). There is further evidence of climatic/depositional change at this time from the abrupt uphole disappearance of sponge spicules and orosphaerid radiolarians that occur from 290 to 520 mbsf (**Quilty,** this volume). The above transitions in conjunction with multiple strong unconformities in seismic profiles point to erosion of the outer shelf in the Prydz Bay and Mac.Robertson areas and, thus, a major expansion of ice.

Middle Miocene to Pleistocene: Establishment of the Polar Ice Sheet

The depositional systems of the continental margin began to change dramatically at about the middle Miocene (i.e., 14–16 Ma; ~300–400 mbsf), coincident with progressively declining sedimentation rates and other lithologic changes at Site 1165 (Fig. **F6**). The morphologic inferences come principally from regional seismic data (**O'Brien et al.,** this volume) and model studies (e.g., ten Brink et al., 1995) and are supported by the new drilling data, especially those from Site 1165. At about middle Miocene time, depocenters began to shift landward from the continental rise and beyond to the continental slope and base of slope (Fig. **F3**). The shift coincided with increases in clays, diatom content, and ice-rafted debris (IRD) at Site 1165 at ~300 mbsf (Shipboard Scientific Party, 2001b)—a shift that points to coarser sediment components being deposited elsewhere (e.g., landward of the rise) and to more icebergs crossing the site.

In the middle–late Miocene, shelf and slope progradation increased and shifted toward the middle of Prydz Bay rather than the earlier even distribution along the shelf edge (Fig. **F3**) (Cooper et al., 2001). These shifts in the geometry of the margin coincide on the rise at Site 1165 with the first occurrence of ?recycled glauconite at 213 mbsf (**Quilty,** this volume), notable decreases in grain density, and porosity at ~200 mbsf and with additional uphole increases in IRD concentrations above ~150 mbsf (Shipboard Scientific Party, 2001b). Other more subtle changes are noted in the Site 1165 cores over this interval (e.g., 300–150 mbsf) and include increased organic carbon, increased color reflectance, higher silica in pore waters, and greatly diminished magnetite. Collectively, these variations point to changes in erosion processes, sediment-source locations, and lithologic controls on secondary circulation processes and diagenesis.

In this middle–late Miocene transition, the continental shelf was initially overdeepened by glacier erosion (e.g., La Macchia and De Santis, 2000), and many slope canyons were filled and buried, resulting in more distal continental rise areas such as Site 1165 receiving diminishing sediment (Figs. **F3, F6**) (**O'Brien et al.,** this volume). Prior to the change, temperate glaciers eroded onshore and intermittently on the shelf, resulting in widespread erosion of a normal water depth shelf during lowered sea levels.

We envision that after the change, larger and (?)colder glaciers episodically extended far onto the continental shelf. Onshore, the Lambert Glacier system oscillated between warm and cold states. In the warm state, the Lambert Graben was a huge fjord with glaciers flowing from the edges and a mixture of open water, sea ice, and icebergs in the basin proper (Hambrey and McKelvey, 2000a, 2000b; **Whitehead and Bohaty,** this volume). In the cold state, glacial maximum glaciers in the graben merged into a major axial ice stream. The ice stream eroded troughs beneath fast-moving areas and deposited banks beneath slow-moving ice on parts of the shelf. This resulted in localized erosion in deep trough areas of the inner shelf, Mac.Robertson shelf, and Prydz Trough during times of glacier expansions in colder paleoclimates.

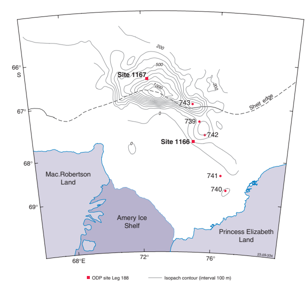
In the late Miocene and early Pliocene, shelf morphology was further modified strongly by focused erosion troughs and deposition banks as a direct result of glacier fluctuations. The widespread shelf erosion is documented by the early Oligocene to late Miocene hiatus at Site 739 and the early Oligocene to Pliocene unconformities documented at Sites 1166 and 742 (Barron, Larsen, et al., 1989; Shipboard Scientific Party, 2001c). Massive and overcompacted glacial diamictons that form Four Ladies Bank lie above these unconformities. The shelf banks have been sampled at Sites 739, 742, and 1166 and their subglacial origin deciphered from subglacial sand grain textures in the diamictons (Strand et al., 2003) and broad bank/trough morphologies on the shelf (O'Brien et al., 1999). Periods of reduced ice cover and open water on the Prydz Bay shelf since the late Miocene are documented at all ODP shelf drill sites by the presence of diatomaceous muddy units (e.g., Site 1166) (**Whitehead and Bohaty,** this volume), but such units are only a small fraction of the late Neogene recovered core.

On the rise the late Miocene and Pliocene are marked by decreasing sedimentation rates and a number of hiatuses in the section at Site 1165 (Fig. **F6**) (**Warnke et al.,** this volume). The early Pliocene section (~34–50 mbsf) has low IRD concentrations and relatively low kaolinite levels (**Warnke et al.,** this volume; **Grützner,** this volume). Upsection increases in kaolinite and IRD at 32 mbsf (~3.5 Ma) (Florindo et al., 2003a) are accompanied by an increase in the grain size of magnetic minerals; these increases point to enhanced detritus eroded from sedimentary basins on the shelf and basement outcrops on the shelf and onshore.

The Neogene record of terrestrial plants in the Prydz Bay region is sparse, with most palynomorphs at Sites 1165 and 1167 likely recycled (**Macphail and Truswell,** this volume b; **Thorn,** this volume). There are a few occurrences of species of the genus *Coptospora* plus *Phyllocladites* and *Nothofagidities* that may not be recycled and may be indicators of Miocene-age tundra-style vegetation similar to that suggested for the Ross Sea region (Raine, 1998). These pollen have no modern counterparts, so climatic interpretation is uncertain.

The last major episode of late Neogene glacial erosion and sedimentation in the Prydz Bay region created the Prydz and Svenner Channels on the shelf and the Prydz Channel Fan on the upper slope. Widespread erosion of the shelf and slope is marked by seismic reflection surface A of Mizukoshi et al. (1986), which is the same as PP-12 of **O'Brien et al.** (this volume). O'Brien et al. (1995) tie surface PP-12 to ODP Site 739 (105.9–130 mbsf) to infer that the surface is of early Pliocene age. Prior to the erosion of PP-12, the shelf prograded and aggraded nearly evenly across Prydz Bay, but after, sedimentation concentrated in the Prydz Channel Fan. On the shelf, post-PP-12 sediment sections are flat-lying and aggrade beneath the Four Ladies Bank, whereas in Prydz Channel coeval sediments are thin and strongly progradational (Fig. **F7**). Changes in the reflection geometries in Prydz Channel point to the development of an ice stream beneath which most debris was transported in a mobile, basal layer to the shelf edge. On the bank, slower-moving ice deposited more basal till.

F7. Sediment above reflection surface PP-12, p. 34.



Pleistocene and Younger: The Young Fluctuating Ice Sheet

Site 1167 is the first deep drill hole in an Antarctic trough-mouth fan, and the cores give an expanded view of glacier fluctuation history for the past 1–2 m.y. The majority of the fan comprises poorly sorted pebbly, clayey sands and diamictons deposited as debris flows; however, the upper 5 m of the fan is principally hemipelagic muds and occasional turbidites (Golding, 2000; Shipboard Scientific Party, 2001d; Passchier et al., 2003). The succession of facies is consistent with predictions based on shallow cores from tough-mouth fans found on other glaciated margins (Vorren and Laberg, 1997). The debris flow intervals are up to tens of meters thick and are separated by thin mud units and rare sand and gravel beds. The debris flows are thought to be derived from slumping of subglacial debris that melted out at the shelf edge when the Lambert Glacier grounded there, and the muds were formed by hemipelagic settling during periods when the ice had retreated from the shelf edge (Shipboard Scientific Party, 2001a).

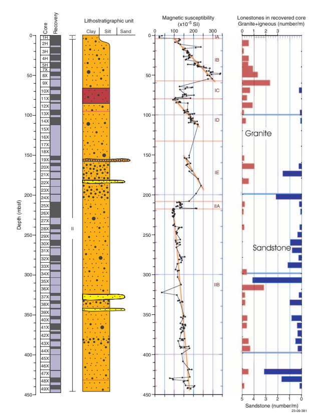
During periods when ice retreated from the shelf edge, the fan surface was also probably eroded and reworked followed by deposition of some mud units. Such erosion can be seen by the development of erosion surfaces within the fan (**O'Brien et al.,** this volume). Some of these surfaces can be mapped throughout the fan in seismic reflection data. Several of these surfaces intersect Site 1167, where they are close to mud intervals or changes in sediment composition, reflected in geophysical logs, mineralogy, and magnetic properties (**O'Brien et al.,** this volume). It is not possible to definitively tie the drill core records from the slope (Site 1167) to the rise (Site 1165) because the sedimentary section thins between the sites and there is a hiatus at Site 1165 that corresponds to the lower half of Site 1167 (**Warnke et al.,** this volume). But, as described below, cyclicities are observed in cores from the slope and rise to further suggest advances and retreats of the ice sheet.

Fan construction occurred principally between the early Pliocene and late Pleistocene, with only 34 m of sediment (<5%) younger than 780 k.y. (i.e., younger than the Brunhes/Matuyama boundary, noted in the cores) (Shipboard Scientific Party, 2001d). The upper meter of the fan has an age of 36.9 ± 3.3 ka at 0.45 mbsf (Theissen et al., 2003). The oldest sediments that could be dated using nannoplankton and strontium isotope dating are from a depth near 217 mbsf and have an age of ~1.1 Ma (M. Lavelle, pers. comm., 2001). Drilling reached 447 mbsf and did not reach the bottom of the fan (i.e., reflector PP-12) (**O'Brien et al.,** this volume), which is thought to be early Pliocene. The fan stratigraphy shows a decreasing number of inferred muddy horizons upsection, indicating a reduction in the number of advances to the shelf edge in the late Pleistocene. In fact, there may have been as few as three advances of the ice sheet to the shelf edge in the past 780 k.y. (**O'Brien et al.,** this volume). On the shelf, the Last Glacial Maximum grounding zone wedges of the Amery Ice Shelf are arranged around the flanks of Prydz Channel no more than 120 km north of the present edge of the Amery Ice Shelf and far from the shelf edge (Domack et al., 1998; O'Brien et al., 1999).

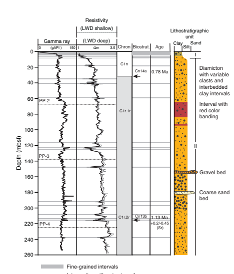
Site 1167 cores exhibit systematic changes in pebble composition, clay mineralogy, magnetic properties, and geophysical log readings that may also be related to ice sheet variations. At 217 mbsf, there is a pronounced uphole change from sandstone- to granite-dominated clasts (Fig. **F8**). This change corresponds to an abrupt decrease in smectite abundance, an increase in magnetic susceptibility and spectrophotometer values, and an increase in gamma log values (indicating more potassium-bearing minerals). These changes suggest a shift from erosion of sedimentary basins on the shelf by sedimentary pebbles, recycled clays, and organic matter, finer-grained magnetites typical of redbeds in Prydz Bay Basin (Domack et al., 1998) to erosion of offshore and onshore basement rocks. The changes occur at ~1.1 Ma at a boundary equivalent with regional seismic unconformity PP-4.

Further evidence of prior ice fluctuations comes from above 217 mbsf at Site 1167. Magnetic susceptibility values form an uncommon "sawtooth" pattern whereby values start high then fall linearly before abruptly jumping to higher values again (Fig. **F9**) (Scientific Party, 2001d). Though susceptibility values do not fall to the "sediment-rich" values seen below 217 mbsf, they still may be explained by progressive increases in sedimentary detritus during cycles of ice volume increase. These systematic cyclic variations, however, do not correspond exactly to depositional cycles indicated by the interbedding of muds and debris flow deposits. Rather, susceptibility cyclic variations typically occur over several phases of debris flow and mud deposition, implying a longer-term fluctuation of ice volume than that of the advance and retreat cycles indicated by the sediment facies. Another explanation is that the sawtooth trends may result from secondary alteration of magnetic minerals below unconformities that represent relatively long breaks in sedimentation. Although the sawtooth patterns cannot now be fully explained, we believe it is linked to changes in ice volume that have acted throughout late Neogene.

F8. Magnetic susceptibility, and lonestones, Site 1167, p. 35.



F9. LWD data, Site 1167, p. 36.



EVIDENCE FOR NEOGENE WARM PERIODS

Evidence for warm periods in Antarctica during the Neogene has produced considerable debate (e.g., Webb et al., 1984; Warnke et al., 1996). The Prydz Bay region and Prince Charles Mountains have outcrops of late Miocene and early Pliocene deposits that probably formed under relatively warm conditions (Whitehead et al., 2001; Whitehead and McKelvey, 2001), and the Leg 188 drill sections also provide fragmentary evidence of warm episodes.

Pospichal (this volume) notes the sporadic presence of warmer-water nannoplankton and thin chalk horizons throughout the Miocene at Site 1165. Whitehead and Bohaty (2003) report Pliocene diatom assemblages at Sites 1165 (16–22 mbsf) and 1166 indicative of sea-ice concentrations less than those of today on the continental rise and on the shelf. Samples from these sites cover the range of ~1.95–3.2 Ma. The youngest evidence of warmer conditions is a few warmer-water foraminifers in the early Pleistocene section of Site 1165 (**Quilty,** this volume) and nannoplankton in mud units at Site 1167 at 37.4 mbsf (~780 ka) and 218 mbsf (~1.13 Ma) (M. Lavelle, pers. comm., 2001). The IRD signal at Site 1165 inferred from the >250-µm size fraction also suggests warmer conditions in the early Pliocene (**Grützner**, this volume; **Warnke et al.,** this volume).

These observations suggest that oceanic temperatures in the Miocene and Pliocene were warmer than at present and that brief intrusions of warm water in the early Pleistocene were accompanied by nannoplankton and foraminifers. Warmer conditions are indicated by open-water facies deposited in the Prince Charles Mountains during the Pliocene–Pleistocene 250 km from the present Amery Ice Shelf edge (~3.1–1.0 Ma) (Whitehead and McKelvey, 2001). Lower Pliocene (4.5–4.1 Ma) diatomaceous deposits in the Vestfold Hills contain diatom assemblages that suggest summer temperatures were 1.6°–3.0°C warmer than today (Quilty, 1993; Whitehead et al., 2001). The low sea-ice concentrations derived from diatom assemblages (J.M. Whitehead and Wotherspoon, pers. comm., 2003) and the warm-water periods inferred from high abundances of silicoflagellates (*Dictyocha*) at Site 1165 (Whitehead and Bohaty, 2003) are consistent with isotopic estimates of warmer conditions than present during the early Pliocene (Hodell and Venz, 1992).

GLACIAL AND BIOGENIC CYCLES ON THE SHELF, SLOPE, AND RISE

The Leg 188 cores from all sites in Prydz Bay show some evidence of cyclic patterns on a wide range of timescales, from millions to tens of thousands of years (Shipboard Scientific Party, 2001a). Our usage of "cyclic patterns" ranges from noting patterns that are regular and systematic (e.g., periodic variations at Milankovitch frequencies) to patterns that have widely variable data parameters and that may or may not have good temporal control (e.g., stacked lithostratigraphic units and unconformities and variable physical properties of sediments—with variable quality age control). Our usage means similarity but not exact repeatability of data parameters in subsequent cycles. Such cyclicity is common in drill core from around Antarctica and elsewhere and, where preserved, provides a useful guide to inferring depositional paleoenvironments and processes that are common to dynamic glacial and oceanographic systems.

On the continental rise, recovery was very good, as were downhole logging records, thereby yielding good records of cyclic patterns. Age dating in this setting was hampered by extensive silica diagenesis, which resulted in loss of silicious microfossils at great depth and dissolution of magnetite in the upper part of the hole. Both losses meant degradation of age control in these parts of the section, which in turn affected the ability to decipher accurate periodicities for cyclic patterns. The rarity of calcareous microfossils on the rise due to large water depths and high polar latitudes prevented the definition of the oxygen isotope stratigraphy, useful for calibrating ages and deciphering paleoceanographic Milankovitch periodicities. On the continental slope, logging-while-drilling data provided surprising and useful information on cyclic patterns, but the absence of diatoms in the fan deposits precluded making reliable estimates of cyclic periodicities. On the continental shelf, the poor core recovery and the large natural variability and incompleteness of the sediment section permitted only sporadic observation of cyclic patterns and no resolution of periodicities.

Site 1165

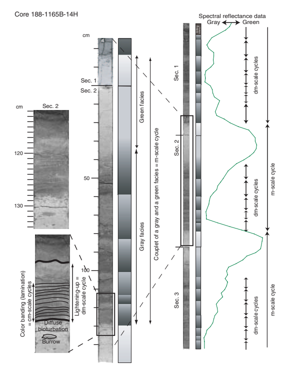
Cyclic patterns are best documented in the continental rise drift deposits from Site 1165. Below ~60 mbsf, the patterns can be visually observed as systematic fluctuations between darker- and lighter-colored sediments that have different physical properties and biogenic contents (Figs. F10, F11). The fluctuations occur throughout the hole, but are masked above 70 mbsf by the uniform brown color of the sediment and masked below ~400 mbsf by low color contrast between dark-colored terrigenous and biogenic units. In these sections of the core, the cyclic patterns are best seen in the laboratory and downhole logging measurements, respectively.

Warnke et al. (this volume) illustrate the cyclic patterns in high-resolution measurements from the upper 50 m of the section, and these are recorded in clay-mineral assemblages, grain-size distribution, spectrophotometer reflectance curves, and $\delta^{18}O$ measurements (in the upper 17 mbsf, where calcareous microfossils are found). The authors do not attempt to establish cyclic periodicities but do relate the fluctuations in parameters to a dynamic behavior of the ice margin in Prydz Bay. A tentative identification of marine isotope stages (MIS) 19 and 71 is proposed, consistent with interglacial–glacial variations.

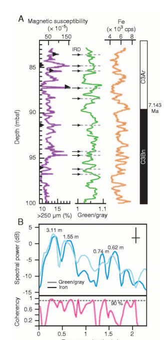
In the upper part of the core in the late Miocene–age section, the cyclic pattern is visually evident and was described initially by the Shipboard Scientific Party (2001b) as alternations between a lighter-color greenish gray facies and a darker-color gray to dark gray facies. Color variations may indicate changes in Fe^{2+} (green) and organic carbon (gray) (Potter et al., 1980). These cyclic patterns are also seen in the lightness factor values from the spectrophotometer (Damuth and Balsam, this volume). The lighter facies have lower density and magnetic susceptibility values and greater biogenic (principally diatoms) content than the darker facies. The patterns in the upper part of the Site 1165 core are more thoroughly described by Rebesco (this volume).

Cyclic patterns in the lower part of the core appear as systematic and coincident spikes in resistivity (positive) and gamma ray (negative) values (Fig. F12). The spikes occur at cemented beds with lower clay content (Facies III-2)—beds that are inferred to have had higher biogenic

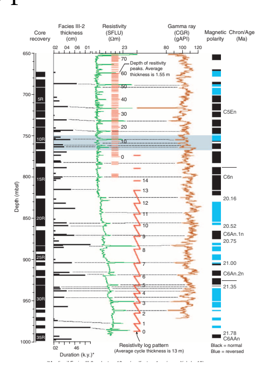
F10. Cyclic sedimentation in the late Miocene, Site 1165, p. 37.



F11. Milankovitch-frequency fluctuations, Site 1165, p. 38.



F12. Cyclicity in logging data, Site 1165, p. 39.



content, now diagenetically altered, than encasing beds (Shipboard Scientific Party, 2001b). Initial onboard studies suggested the cemented beds were calcareous, but postcruise studies (Williams et al., 2002) indicate that they are predominately silicious. The cyclic patterns throughout the hole demonstrate that sedimentation since the early Miocene (bottom of hole) has systematically varied between principally terrigenous and biogenic sources.

The cyclic patterns appear to fall in the range of Milankovitch periodicities, within the limits of errors in the age vs. depth curve for the hole (Fig. **F11**). For the interval 83–100 mbsf (late Miocene) the shipboard-determined periodicities in the lightness and bulk density values are ~94, 41, 31, 21, and 18 k.y. (Shipboard Scientific Party, 2001b). Postcruise studies of the same interval (6.76–7.21 Ma) by Grützner et al. (2003) incorporate additional high-resolution measurements of color spectra, X-ray fluorescence (XRF), Fe, grain size of IRD, and magnetic susceptibility. They use cross-spectral analyses to refine the periodicities to 84.1, 41.9, 20.0, and 16.8 k.y. and note that the correlation of spectral peaks with astronomic cycles of obliquity (41 k.y.) and precession (19–23 k.y.) suggests an orbital origin for the observed variations in color and iron concentration.

For the lower section of the hole (~675–999 mbsf; ~18 to ~22 Ma), the postcruise studies (Williams et al., 2002; T. Williams, pers. comm., 2004) include analysis of the cyclicity in the downhole resistivity and gamma radiation values (Fig. **F12**). They suggest that the regular spacing of Facies III-2 rocks with resistivity peaks in the downhole logs could also be astronomically paced. Between 18 and 20 Ma the average thickness of cycles is 1.55 m, which gives a periodicity of ~15–23 k.y., depending on sedimentation rates. Between 20 and 22 Ma, the average cycle thickness of 13 m gives a periodicity of ~135 k.y., based on the age model in Fig. **F6,** but may be shorter with their alternate age model. Studies of lithologic cyclicity for nearly the same time period in Antarctica are reported for the Cape Roberts area of the Ross Sea adjacent to the Transantarctic Mountains (Naish et al., 2001). Here the site is proximal to the front of an East Antarctic outlet glacier, and they tie the cycles to orbital influences at periods of 40 and 125 k.y.

Cyclic patterns of terrigenous and biogenic units in continental rise drift deposits at Site 1165 adjacent to the East Antarctic Ice Sheet can be compared with similar cyclic sediments in rise drift deposits at ODP Site 1095 adjacent to the Antarctic Peninsula (West Antarctic Ice sheet). At Site 1095 initial shipboard analyses of alternating terrigenous (gray) and biogenic (brownish) units pointed to the variations having Milankovitch periodicities (Shipboard Scientific Party, 1999). However, detailed postcruise studies by Lauer-Leredde et al. (2002) and Pudsey (2002) established that the observed lithologic cyclic patterns are far from regular and included Milankovitch as well as other periodicities ranging from ~20 to 110 k.y. The irregularity is due in large part to the inherent variability of different age models for the drill site (e.g., magnetostratigraphy and diatom/radiolarian biostratigraphy). Pudsey (2002) notes that in the late Miocene section (6.2–6.4 Ma), which is nearly equivalent to the period that Grützner et al. (2003) studied at Site 1165, the irregularity may be related to drainage basin geology and tectonic uplift masking a well-defined orbital cyclicity. The Prydz Bay region is an older passive margin setting (e.g., Cooper et al., 1991a) than the Antarctic Peninsula, and depositional patterns near Site 1165 have not been greatly affected by basin tectonics since the early Miocene; therefore, such processes would not mask a climatic cyclic pattern.

Site 1166

Cyclic patterns in cores from Site 1166 on the continental shelf are poorly defined due to generally low core recovery and the highly varied lithologies encountered (Fig. F5). Lithologic cyclicity is observed in the late Eocene to early Oligocene as alternating nonmarine massive sands and silty sands of logging Unit 4A (163–195 mbsf). Here, the cyclic pattern may result from variations in fluvial/tidal environments resulting from changing sea levels or autocyclic processes in the depositional environment. In the early Pliocene and younger age glacial section (0–135 mbsf), three possible cyclic intervals in grain density are noted and may represent three advances and retreats of the continental ice sheet (Shipboard Scientific Party, 2001c). Isotopic measurements ($\delta^{18}O$) on foraminifers from this section by **Theissen et al.** (this volume) show large variations that are consistent with glacial–interglacial changes in ice volume/temperature, but the values are too sparse to establish a clear cyclic pattern.

Site 1167

Site 1167 on the continental slope is marked by irregular cyclic patterns in lithologies, magnetic susceptibilities, and $\delta^{18}O$ values at widely differing scales. The age control is, however, inadequate to establish periodicities.

Cyclic lithologies at Site 1167 are denoted by thick debris flow units (as thick as ~50 m) alternating with thin mud horizons (as thick as 3 m). The irregular cyclic pattern can be seen in cores but is more evident in the continuous downhole logs that were recorded in the upper 260 m of the section (Fig. F9). In the logs, the thin and poorly recovered mud layers appear as coincident spikes in resistivity and gamma ray values and delineate ~15 debris–mud cycles of variable thickness (Shipboard Scientific Party, 2001d). In this part of the hole, as noted above, a cyclic sawtooth pattern was identified in shipboard magnetic susceptibility values within four broad intervals (50–80 m thick) that are characterized by linear upward-decreasing susceptibility values (Shipboard Scientific Party, 2001d) (Fig. F9). These intervals are bounded by mud layers. Cyclic patterns are also evident in $\delta^{18}O$ measurements on planktonic and benthic foraminifers and are of variable periodicity superimposed on a nearly linear uphole increase in $\delta^{18}O$ values (**Theissen et al.,** this volume). Even where measurements are sparse, below 50 mbsf, cyclic variations are systematic, sometimes large, and sometimes (but not always) bounded by mud horizons. Above 50 mbsf, **Theissen et al.** (this volume) tentatively identify MIS 16–21, in which each stage has a cyclic period of about 4–8 m, and is interpreted as a glacial-interglacial cycle.

The cyclic patterns at Site 1167 can generally be explained by fluctuations in oceanographic and depositional paleoenvironments in response to changes in the Antarctic cryosphere and depositional source areas over the past 1–2 m.y. Periods of increased ice volumes, when glaciers extended to the continental shelf edge, are times of enhanced sedimentation and debris flows, and contrast with periods of diminished ice and low sedimentation and resulting mud layers. Isotopic measurements in slope debris flow units also support this general concept. Yet many detailed questions for Site 1167 remain unanswered, including

1. Why are some cyclic patterns linear sawtooth, such as magnetic susceptibility, yet others more commonly vary smoothly over similar time periods?
2. Why don't cyclic patterns, such as those for susceptibility and $\delta^{18}O$ measurements, consistently match apparent paleoclimate markers (e.g., mud layers as possible markers of interglacials), as commonly postulated?
3. And why are there strong systematic cyclic patterns for a variety of parameters in a potentially chaotic depositional environment (e.g., slope debris flows)?

The drilling data from Site 1167 provide the first detailed deep measurements from an Antarctic trough-mouth fan to help answer these questions. The new data illustrate, however, that prior conceptual models of ice sheet depositional processes may not apply to the world's largest glacier outlet system in Prydz Bay. The cyclic patterns are real, but the regional data are still inadequate to explain them.

Prydz Bay Cyclic Patterns

Leg 188 drilling provides further evidence that paleoenvironments and paleoclimates for the Prydz Bay continental margin have strong cyclic components throughout the Cenozoic since earliest glacial times. There is a long-term cyclicity over millions of years in the climatic and ocean cooling, as seen in the changes in palynology (e.g., Site 1166) (Macphail and Truswell, this volume a), the systematic increases in $\delta^{18}O$ isotope values (Site 1167) (Theissen et al., this volume, 2003), and ice volume and sediment delivery as reflected by systematic increases in IRD and decreases in sedimentation rates since the middle Miocene (Site 1165) (Fig. F6) (Shipboard Scientific Party, 2001a). There is also a shorter-term cyclicity over tens of thousands of years in glacial and interglacial periods reflected in changing lithologic properties across the continental margin. These include the fluctuations in physical properties and diatom contents of stacked strata on the shelf (Site 1166) (Shipboard Scientific Party, 2001c), the alternating debris flows and mud layers on the slope (Site 1167) (Shipboard Scientific Party, 2001d), and systematic fluctuations between terrigenous and biogenic facies on the rise (Site 1165) (Shipboard Scientific Party, 2001b). The glacial and interglacial cyclicity is seen also in variations in $\delta^{18}O$ isotope measurements at all three Leg 188 drill sites (Theissen et al., this volume, 2003; Warnke et al., this volume), but is best documented for Leg 188 at Site 1167.

The explanations for the cyclic patterns are equivocal based solely on Leg 188 data because there are many contributing and interrelated factors that have been widely discussed regarding cyclicity—factors that include changing ocean currents and temperatures, sea levels, ice volumes, terrigenous erosion rates, biogenic productivity, CO_2 levels, and others. The relative importance of these factors changes with location and time. On the continental shelf, the late Eocene cyclic sand–shale sequences point to controls of changing sea levels, but in the early Oligocene and again in the late Neogene the greatest control on cyclic patterns was glacial extent. In the early Miocene, the cyclic patterns on the continental rise may have been largely controlled by distribution of abundant fine-grained terrigenous debris by downslope and bottom-contour ocean currents, whereas in the middle Miocene and younger times there was an increasing relative effect of biogenic productivity and climatic cooling (with resulting cold-ice cover and lower terrigeneous sediment supply) that dominates cyclic patterns. On the flanks of the Kerguelen Plateau (~1000 km northeast of Prydz Bay), cyclic terrigeneous–biogenic lithologic patterns in the late Miocene and younger sections at ODP Sites 745 and 746 result from variable biogenic productivity and terrigenous sediment supply (Ehrmann et al., 1991; Ehrmann and Grobe, 1991). Generally, for the late Neogene, the greatest controls on cyclic patterns throughout the Prydz Bay region are from volume changes of glacier ice affecting erosion and redistribution of sediments and from paleoceanographic changes modulating ocean-biogenic productivity.

DIAGENESIS AND SEDIMENTARY ENVIRONMENTS

Sediments recovered during Leg 188, like those from most drilling legs, have been altered by diagenetic processes resulting from fluid circulation, geothermally induced changes, and other factors, some of which may be related to paleoclimate and paleoceanographic changes (e.g., changes in biogenic productivity, water temperature, and ocean chemistry). Leg 188 cores show commonly observed diagenetic effects (e.g., sediment compaction and microbial generation of methane) as well as diagenetic effects found only in sections with high biogenic productivity (e.g., silica dissolution and diagenesis, dissolution of magnetite, and precipitation of carbonates). Diagenesis in the cores is a secondary effect not unique to the high-latitude environment, but in the absence of a direct record (now altered), the diagenetic record gives further evidence of changes in paleoenvironments (e.g., **Quilty,** this volume) that at times had relatively large silica biogenic productivity common to cold-water settings.

Silica-Related Diagenesis

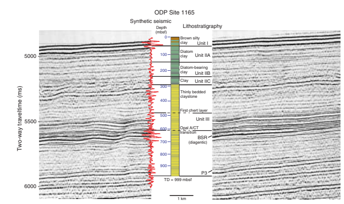
Primary silica is noted at all Leg 188 drill sites in the form of biogenic material (e.g., diatoms, radiolarians, and sponge spicules) and magmatic products (e.g., quartz sand). Here, we focus on the biogenic component because it is more sensitive to dissolution and reprecipitation than the magmatic component.

Continental rise sediments at Site 1165 contain relatively large concentrations of biogenic silica, as much as 30% diatoms in smear slides, whereas sediments from the shelf and slope Sites 1166 and 1167 contain relatively low concentrations or none, reflecting differences in productivity across the continental margin (Shipboard Scientific Party, 2001a). Diagenesis of silica (i.e., dissolution and reprecipitation) is best documented at Site 1165 by the downhole transition from sediments with diatoms (opal-A) above 606 mbsf to sediments with silicified horizons (inferred opal-CT and/or chert) and calcified layers below this depth (Shipboard Scientific Party, 2001a). From studies in other parts of the world where similar downsection silica phase transitions are documented (e.g., Bering Sea, Sea of Japan, and offshore California [USA]) (Hein and Obradovic, 1989), the depth to the transition depends on subsurface temperatures and relative concentrations of biogenic and terrigeneous material, with higher temperatures and larger biogenic silica concentrations facilitating the transition.

At Site 1165, the full silica transition may take place over a 150-m-thick depth zone, which is marked by the first appearance of chert at 492 mbsf, loss of all diatoms below 606 mbsf, abrupt shifts in downhole logging measurements (e.g., velocity, density, resistivity, and porosity) at 610 mbsf, and a zone of low recovery that extended to ~650 mbsf with alternating soft (clay/silt) and hard sediment layers. (Shipboard Scientific Party, 2001b; Williams et al., 2002). The hard layers are likely a combination of chert, based on common chert fragments in sediment residues from below ~598 mbsf (Quilty, this volume), calcified horizons, based on rock samples and downhole logging (Shipboard Scientific Party, 2001b), and opal-CT, inferred from silica transitions in other areas noted above.

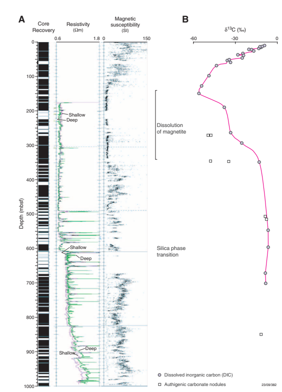
In seismic reflection data across Site 1165, the full silica transition zone is denoted by a distinct 150-ms-thick band of high-amplitude reflections (Fig. F13). The band can be traced across the Prydz Bay continental rise. Within the band is a strong reflection that results from the abrupt change at 606 mbsf and appears to be a strong bottom-simulating reflection (BSR) like diagenetic BSRs (i.e., different from gas hydrate BSRs) seen elsewhere, such as off the Antarctic Peninsula, where the BSR and inferred diagenetic boundary is the décollement surface for slumps and slides (Volpi et al., 2003). The above diagenetic evidence points to silica microfossil concentrations (likely mostly diatoms) being sufficiently large to result in creation of the thin hard silicious layers by dissolution and reprecipitation. Early Miocene depositional environments were likely marked by periods of high biogenic silica production/deposition during cooler periods and/or periods of lower terrigeneous supply. The record of younger paleoenvironments comes from the unaltered cores above the silica transition.

F13. Seismic reflection profile, Site 1165, p. 40.



The diagenetic effects of high silica concentrations in sediments and pore waters at Site 1165 are recorded also by the dissolution of magnetite and reduction of magnetic susceptibility within the upper part of the sedimentary section (94–362 mbsf) (Fig. F14A). This was noted initially by the Shipboard Scientific Party (2001b) and explained by Florindo et al. (2003b). The effect, although not restricted to polar environments, is yet another indicator of the silica-enhanced and "open circulation" sedimentary section with a nearby magnetite-rich sediment source that characterizes the continental rise at this location. The combination indicates adequate nearby sources of both terrigenous and biogenic source materials to further explain the cyclic facies patterns thought to be due to ice sheet fluctuations (Shipboard Scientific Party, 2001a).

F14. Diagenetic effects, Site 1165, p. 41.



Silica diagenesis, although important at Site 1165, is not recognized as a factor at either Site 1166 (shelf) or Site 1167 (slope). Silicious microfossil concentrations are relatively small and localized to fine-grained marine sections at Site 1166, and silicious microfossils are found only in the upper 5 mbsf at Site 1167 (Shipboard Scientific Party, 2001c, 2001d). The low concentrations and small values for silica in pore waters point to silica dissolution or nondeposition on the slope, possibly due to extended sea-ice cover (Shipboard Scientific Party, 2001d).

Carbon-Related Diagenesis

Carbon-related diagenesis in Leg 188 cores is seen in recycled carbon (e.g., coal) and in secondary features resulting from in situ diagenesis (e.g., authigenic carbonate nodules). The diagenesis results from processes of different types characteristic of the varied paleoenvironments of the Prydz Bay margin, ranging from relatively cold deep-ocean to relatively warm lagoonal.

Organic carbon concentrations in Leg 188 rocks vary from <0.1 wt% in hemipelagic deposits on the continental rise to a maximum of 9.2 wt% in lagoonal deposits on the continental shelf; there is evidence of biogenic and recycled terrestrial carbon at all drill sites (Shipboard Scientific Party, 2001b, 2001c, 2001d). The carbon concentrations and subsurface temperatures directly affect the diagenetic processes related to microbial degradation, methanogenesis, and, in places, diagenetic formation of authigenic carbonates (**Claypool et al.,** this volume). The carbon and oxygen isotopic compositions of the authigenic carbonates indicate the subsurface conditions when this carbonate formed. **Claypool et al.** (this volume) report such stable isotopic measurements for authigenic carbonate nodules from Sites 1165 and 1166.

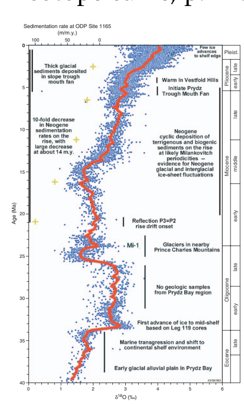
Comparison of the stable isotopic measurements with pore water isotopic values for the cores at these sites gives evidence of prior depositional environments that are quite different. The siderite nodules from lagoonal carbonaceous shales at Site 1166 (shelf) have carbon and oxygen isotopic compositions consistent with nodule growth during early stages of methanogenesis. The authigenic siderite from Site 1165 (rise) apparently grew in or just below the sulfate reduction zone that now extends to 150 mbsf but in the early Miocene, when sedimentation rates were 8–10 times greater, likely extended only to 10–20 mbsf. Similarities in $\delta^{13}C$ depth profiles for dissolved inorganic carbon and authigenic carbonates (Fig. **F14B**) and the volume percent of carbonate in these carbonates are consistent with authigenic carbonate formation within the upper part of a methane-charged sediment drift, from dissolved carbonate derived in part from anaerobic methane oxidation.

Gas hydrates were anticipated at Site 1165 but not found, likely because of currently unsuitable conditions resulting from changes in paleodepositional environments that altered sedimentation rates. **Claypool et al.** (this volume) suggest that as sedimentation rates decreased since the early Miocene the sulfate reduction zone thickened, thereby destabilizing methane hydrate due to lowered concentrations of dissolved methane. They also note that some authigenic carbonate nodules have high $\delta^{18}O$ values that could result from high $\delta^{18}O$ concentrations in pore waters due to decomposition of methane hydrate at the time of nodule formation. Hence, hydrates may at one time have existed at Site 1165, but are now not found here.

CONCLUSIONS

Leg 188 drilling (O'Brien, Cooper, Richter, et al., 2001) has added much proximal evidence to that collected during prior Leg 119 drilling in Prydz Bay (Barron, Larsen, et al., 1989, 1991). These drilling legs have established that changes in ice volume/extent affected sediment erosion and redistribution, which resulted in cyclic patterns observed in the drilled Cenozoic sections. These effects are more prominent since the early Miocene than before. Leg 188 drilling in Prydz Bay has augmented the initial discoveries of Leg 119 to identify and/or better define key paleoenvironmental transitions in the region since the late Mesozoic (Fig. **F15**). The transitions outline a general climate cooling since the late Eocene, increasing ice cover to today's polar setting, a trend that is consistent with that illustrated by global oxygen isotopic curves showing cooling of the world's oceans (e.g., Zachos et al., 2001).

F15. Prydz Bay events vs. global oxygen isotope curve, p. 42.



In detail, the transitions exhibit systematic variability over times of millions to tens of thousands of years, suggesting also that regional- to local-scale processes have been active. Leg 188 drilling provides the first complete transect of drill sites across the Antarctic margin including shelf, slope, and rise sections, to yield new detailed data on these depositional paleoenvironments.

Key transitions in the region deciphered from Leg 188 drilling studies include the following:

1. The Prydz Bay shelf evolved from a largely low-lying pre-Oligocene alluvial to deltaic system to a shallow-marine setting until about the middle Miocene, when shelf progradation to aggradation coeval with erosional shelf overdeepening commenced.
2. Vegetation onshore changed from temperate conifer woodland to cool *Nothofagus* scrub by the late Eocene with *Nothofagus* possibly lasting into the early Miocene before the loss of vegetation in the late Neogene.
3. The first evidence for ice near the shelf is in the late Eocene, with grounded ice there in the early Oligocene, some floating ice in the early Miocene, and ice buildup in the middle Miocene, based on glacial sediments and IRD from the continental shelf and rise. A change from distributed to focused ice flow on the shelf occurred in the late Miocene to early Pliocene with the carving of Prydz Trough and others. No cores exist in Prydz Bay for the early Oligocene to early Miocene to decipher ice volumes for this period.
4. On the continental rise and slope, pre-Miocene canyons and channel/levy systems were enhanced by large sediment drifts initiated in the late Oligocene to early Miocene, and canyons were covered by thick slope-fan deposits starting in the middle to late Miocene. Sedimentation rates on the rise decreased gradually by ~10-fold from the early Miocene to Quaternary, with the most rapid decline since 14–16 Ma (middle Miocene) during a period when IRD increased.
5. On the slope, abrupt shifts in lonestone compositions, clay contents, and other core properties as well as shifts in regional depositional patterns heralded changes in ice/sediment source areas in Pleistocene times.
6. Leg 188 Site 1165 shows evidence of relatively rapid fluctuations between times of principally biogenic and principally terrigenous sedimentation with a glacial component, and in some cores with adequate age control the fluctuations occur at or near Milankovitch periodicities.

The combination of seismic reflection, drilling, and onshore geologic records from Prydz Bay continental margin illustrate that the Cenozoic morphologic and paleoenvironmental histories were influenced by sedimentologic and oceanographic processes such as sea level changes and others common to low-latitude nonpolar margins in preglacial and earliest-glacial times (i.e., pre-early Oligocene) and by those such as subglacial erosion common to high-latitude polar margins in full-glacial periods (i.e., middle to late Miocene and younger).

The intermediate history is mostly unknown for lack of cores, but where known in the early to middle Miocene at Site 1165 and onshore back into Oligocene time in the Prince Charles Mountains, the record is one of transitions with both long-term components (e.g., shift in depocenters, reorganization of ocean currents, and increase in onshore ice) and short-period ones (e.g., ice volume and subglacial water fluctuations and biogenic productivity variations). Perhaps during the early to middle Miocene, the margin of Prydz Bay was similar to that of east Greenland today, as suggested also by Hambrey and McKelvey (2000a, 2000b). Regardless, large volumes of glacially influenced sediment were eroded from Prydz Bay onshore and shelf regions and redistributed down canyon systems into deepwater areas during this period. These post-early Miocene transitions point to increasing ice, decreasing onshore erosion, relative increases in biogenic productivity, and increasing erosion of the shelf, leading to today's polar environment.

A major feature of the Prydz Bay record that emerges from the new drilling and other studies is the effect of the progressive long-term cooling trend on the geological record. The record on the continental shelf suggests the first signs of Cenozoic glaciation were stunted vegetation and glacially abraded sand grains, followed by ice-rafted clasts in the marine realm. Glaciomarine deposition and erosion and subglacial deposition then followed. Early Miocene glaciation produced huge quantities of fine sediments that bypassed the shelf and were deposited in drifts on the rise. We suggest that this was because of abundant meltwater outflow from the early Miocene ice sheets. Progressive cooling during the middle Miocene (14 Ma) (Florindo et al., 2003a) led to expansion of the ice that started eroding the sedimentary basins of the shelf and Lambert Graben. Cooler ice produced less meltwater and less suspended sediment delivery to the rise, but ice at sea level produced more icebergs and, hence, more ice rafting of debris. This trend continued through the late Miocene and Pliocene. Further cooling shifted the locus of maximum snow accumulation from the interior of the continent to the coastal fringe, so, during the early Pliocene, coastally sourced ice deflected the Lambert Glacier westward, forming the Prydz Channel Ice Stream. The combination of low precipitation, cold ice, and deep shelf erosion eventually led to the cessation of advances to the shelf edge by the Lambert–Amery Ice Stream during the middle Pleistocene (**O'Brien et al.,** this volume).

This overall trend to cold, less active ice sheets through Neogene was punctuated by interludes of warm conditions that included extreme advances of the Lambert Glacier–Amery Ice Shelf system during the Pliocene.

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Figure F1. Index map of Prydz Bay region showing bathymetry, names of features, and sites drilled during ODP Legs 188 (this report) and Leg 119 (Barron, Larsen, et al., 1989, 1991). Modified from O'Brien, Cooper, Richter et al., 2001).

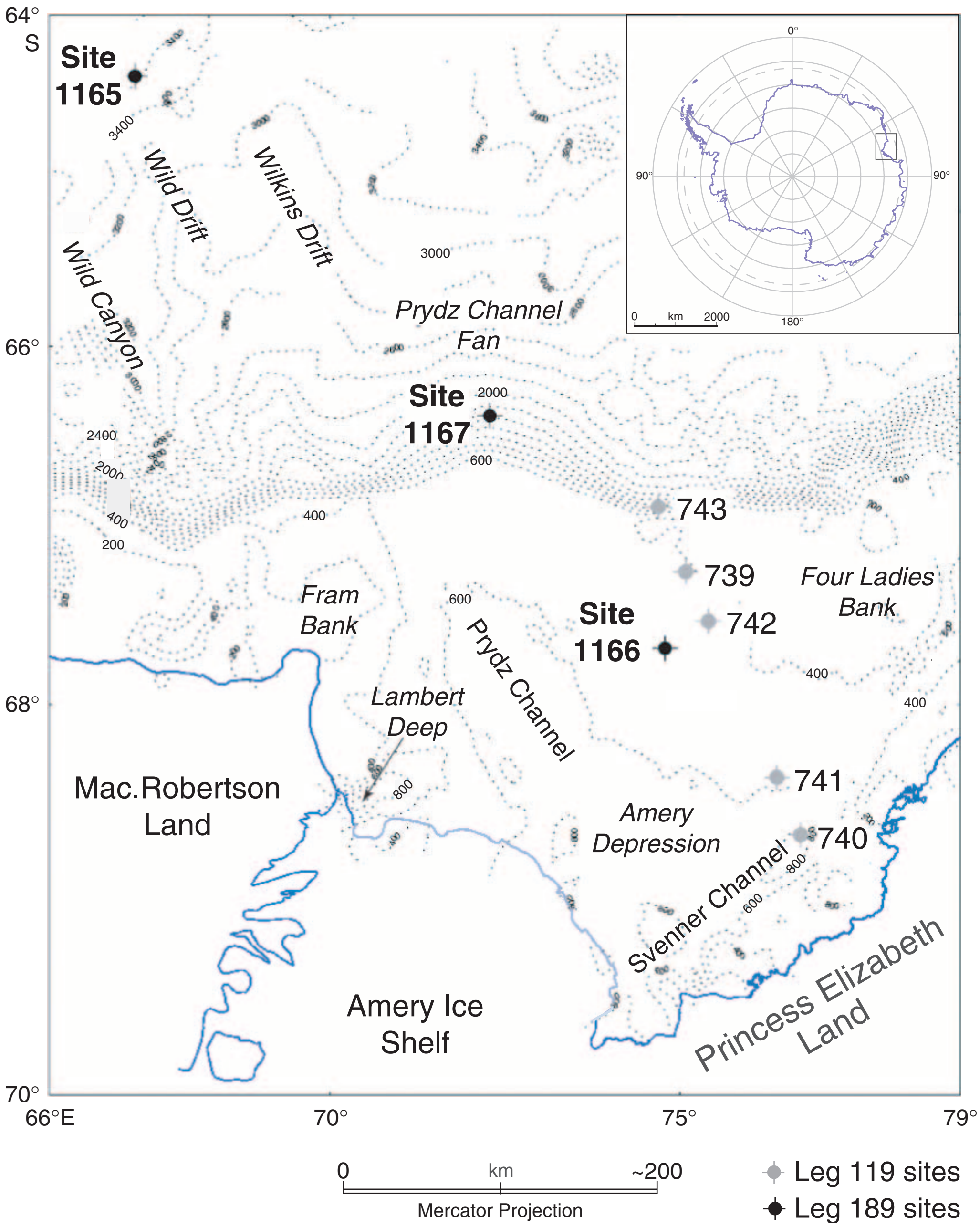

Figure F2. The Lambert Glacier drainage basin, East Antarctica, showing locations of the exposed mountains and the Gamburtsev Subglacial Mountains. Arrows show directions of ice flow today (onshore) and in prior times (offshore) when the ice sheet was greatly expanded. Ice surface elevations are in meters (modified from Hambrey et al., 1991). Gl. = Glacier. Dashed line PC offshore is the edge of the Prydz Channel.

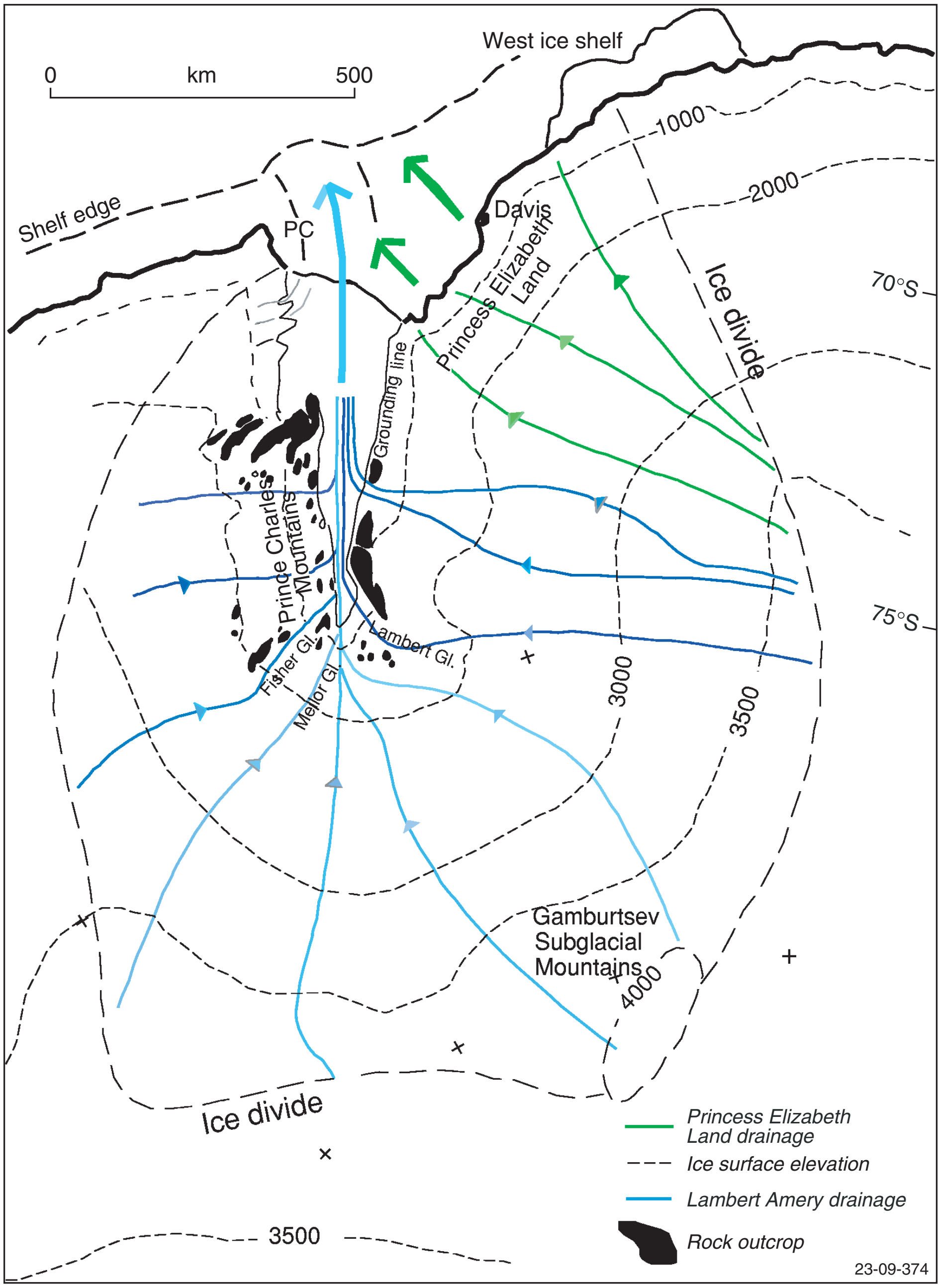

Figure F3. Diagrammatic sections based on seismic reflection profiles across the continental margin and ODP Leg 119 and 188 drill sites. **A.** Transect A–A′ across Leg 188 drill sites from Cooper et al. (2001). Ages are based on correlation with Leg 188 drill sites and ODP Leg 119 Sites 739 and 742 near Site 1166. Approximate locations of paleoshelf edges are marked by dots. **B.** Transect B–B′ across Leg 119 drill sites, modified from Cooper et al. (1991a).

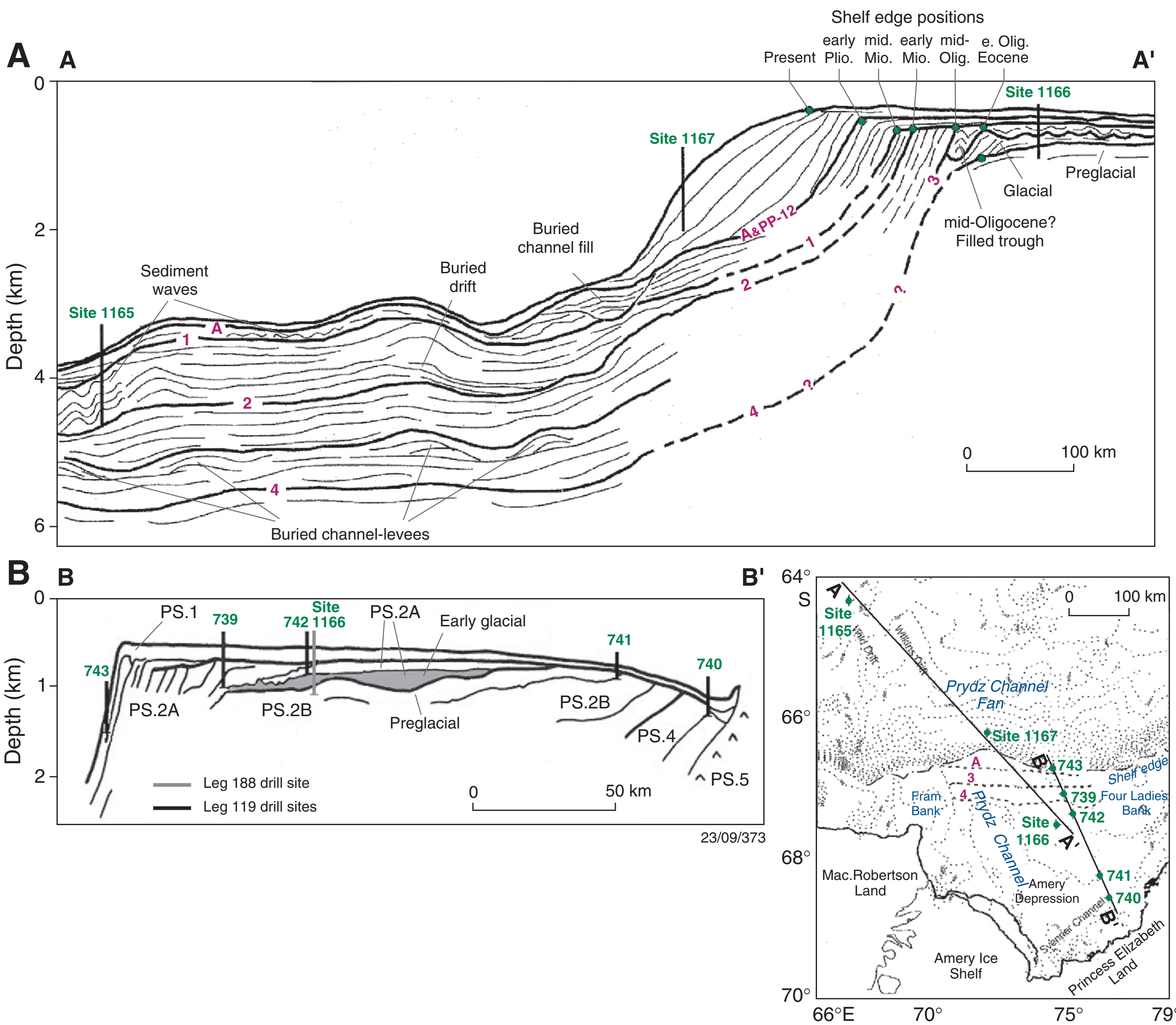

Figure F4. Generalized map of present-day ocean currents in the Prydz Bay region (modified from Smith et al., 1984). ACC = Antarctic Circumpolar Current. The Polar Current and ACC are major surface current systems that extend into Antarctic Deep Water, and their variable interactions partly control sediment distribution across the continental margin.

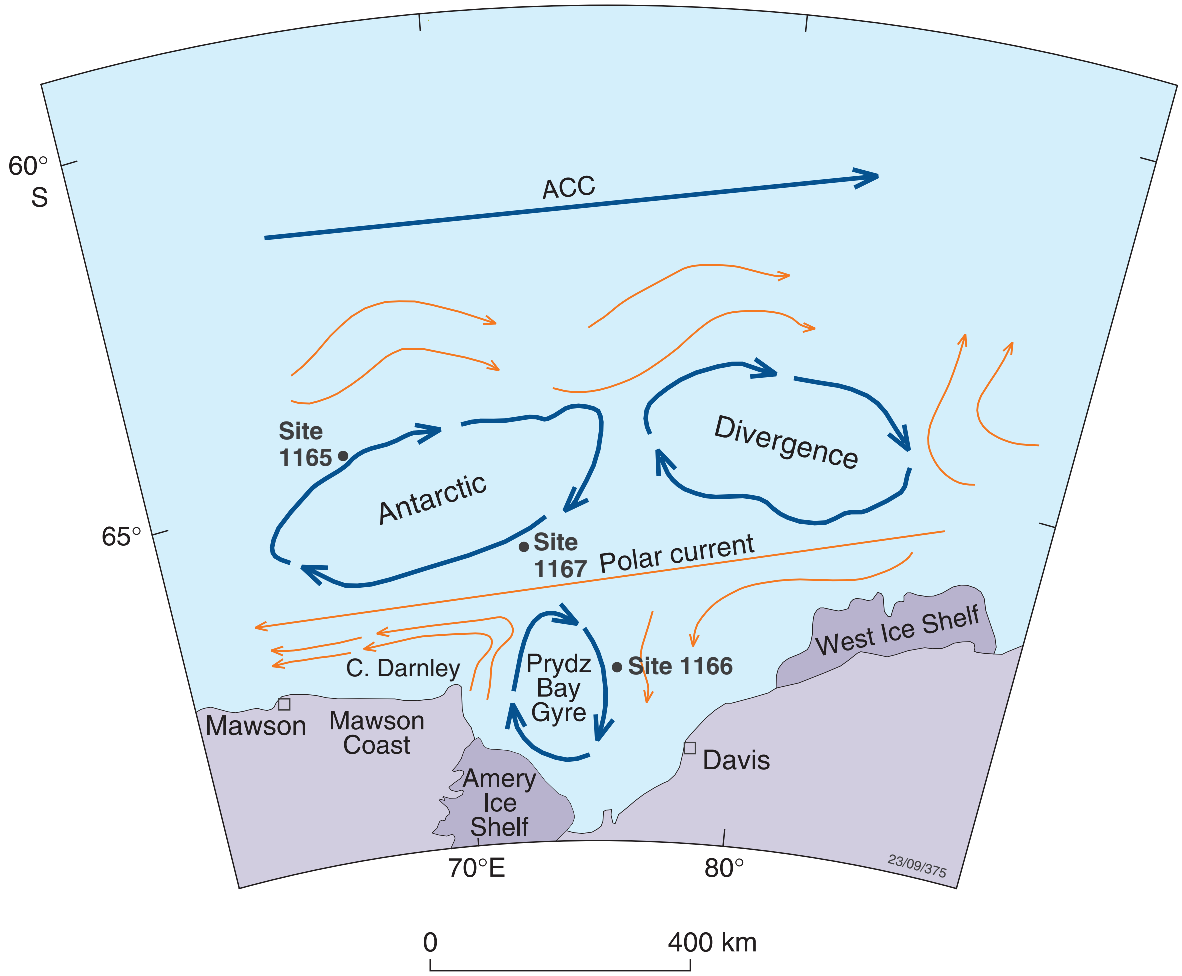

Figure F5. Composite section for Site 1166 showing core recovery, downhole logging data, lithostratigraphic units, diatom distribution, palynology zonation, and diatom ages (modified from Shipboard Scientific Party, 2001c). Palynologic ages are from Macphail and Truswell (this volume a). LWD = logging while drilling.

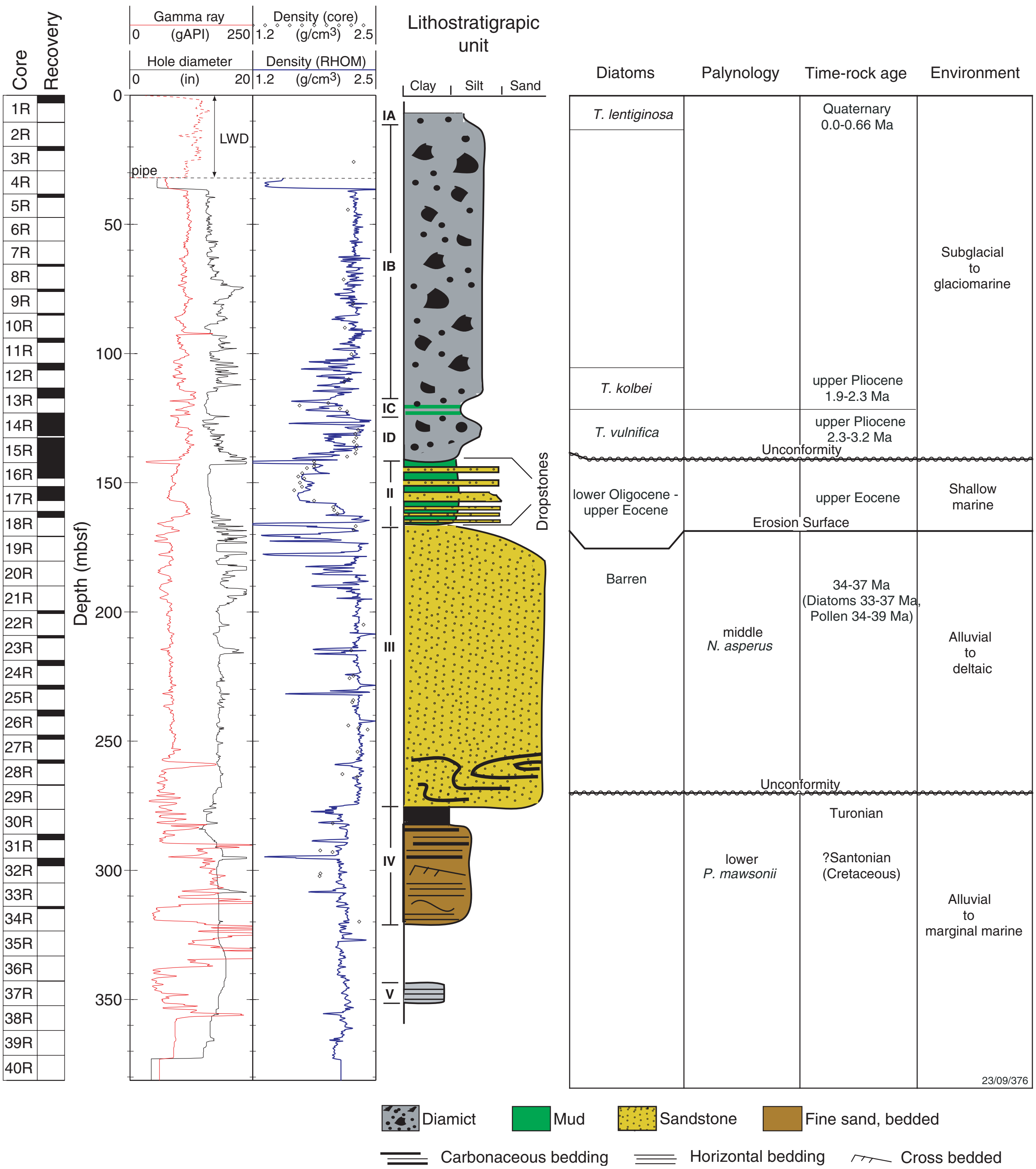

Figure F6. Lithostratigraphic column and age-depth plot for ODP Site 1165 from Florindo et al., 2003a.

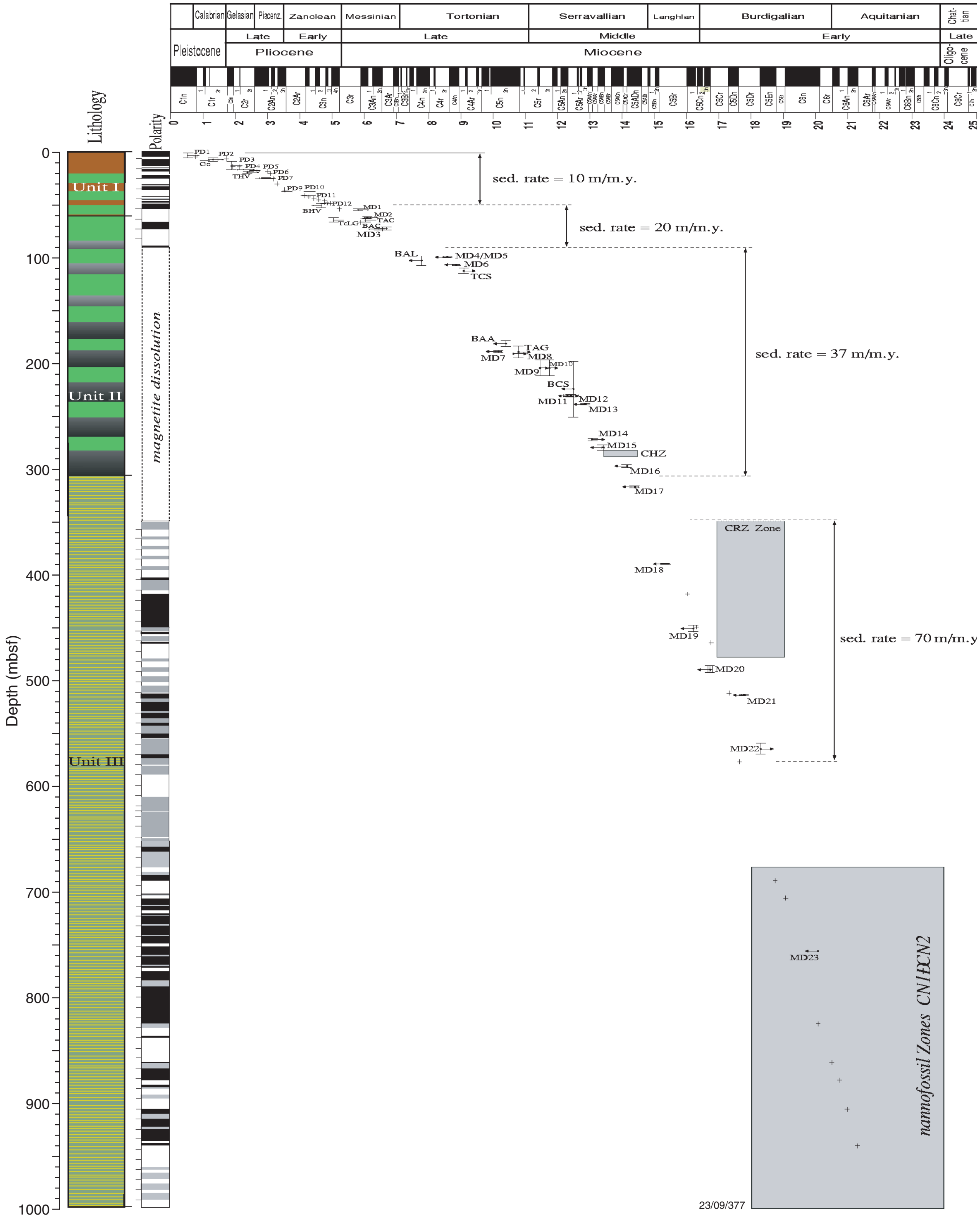

Figure F7. Isopach map of all sediment that lies above reflection surface PP-12 (post-late Miocene), with contours in milliseconds two-way traveltime (from Shipboard Scientific Party, 2001a). Sediment deposition is mostly concentrated in the Prydz Channel Fan. Elsewhere in Prydz Bay, post-PP-12 sediments are thin and have been highly eroded at various times.

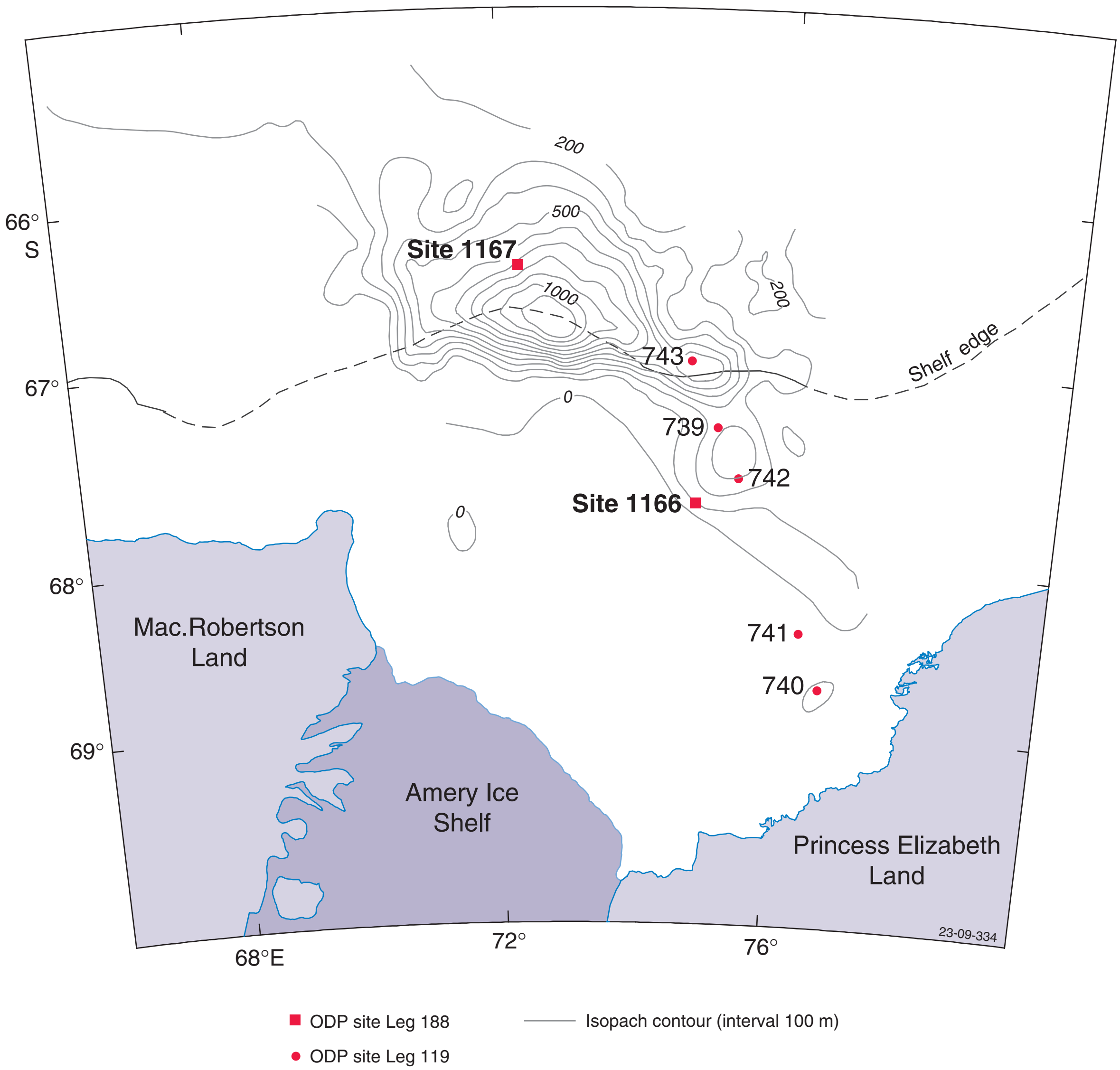

Figure F8. Composite section for Site 1167 showing core recovery, lithostratigraphic units, magnetic susceptibility data, and distribution of granite and sandstone lonestones (modified from Shipboard Scientific Party, 2001d).

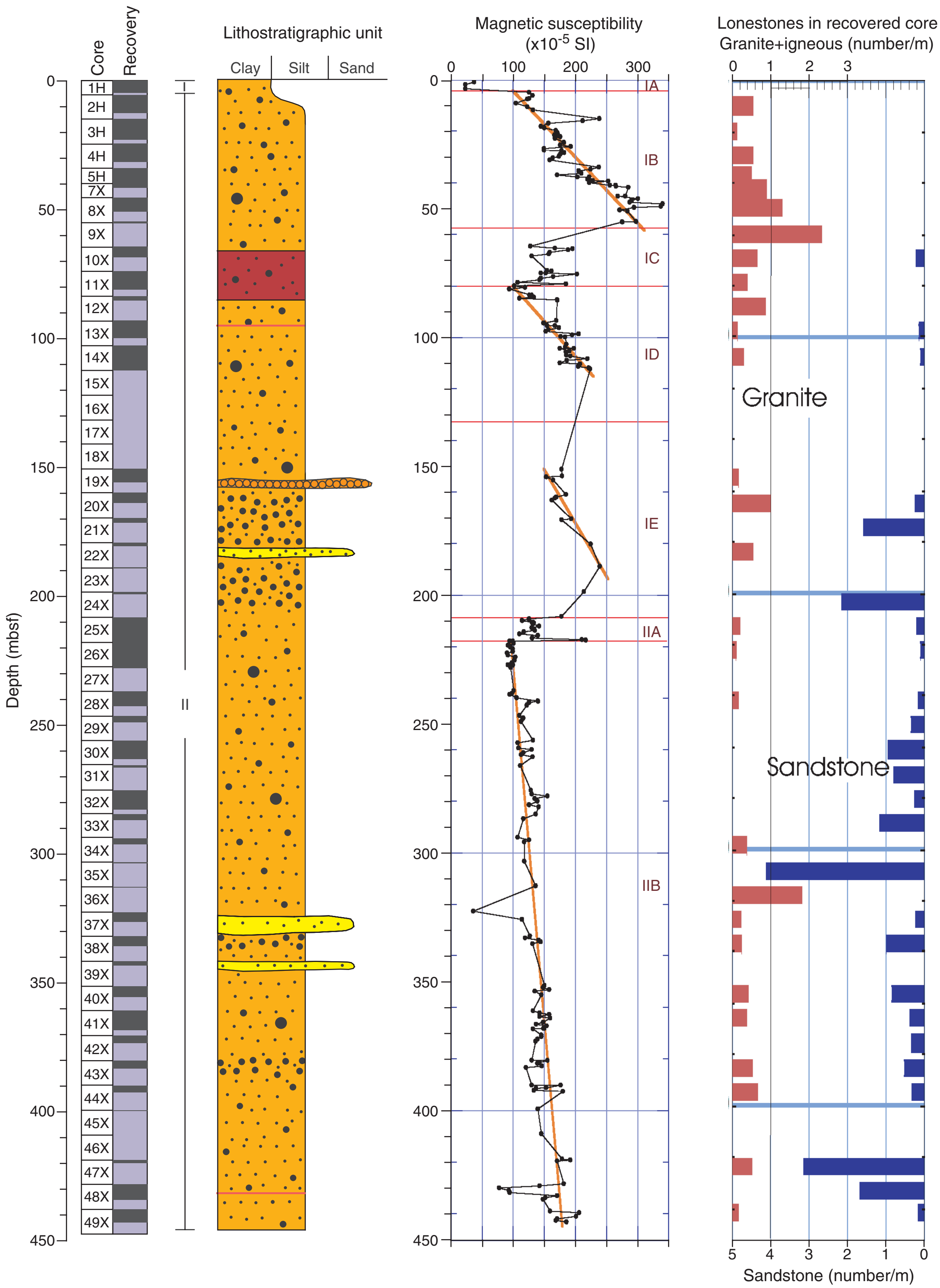

Figure F9. Composite section for the upper part of Site 1167 showing logging-while-drilling data, magnetic chrons, ages, lithostratigraphic units, and inferred fine-grained clay intervals (modified from Shipboard Scientific Party, 2001d). Also shown are correlations of regional reflectors PP-2 to PP-4 with the drill site (from O'Brien et al., this volume). The 1.13-Ma Sr isotope age is from M. Lavelle (pers. comm., 2001).

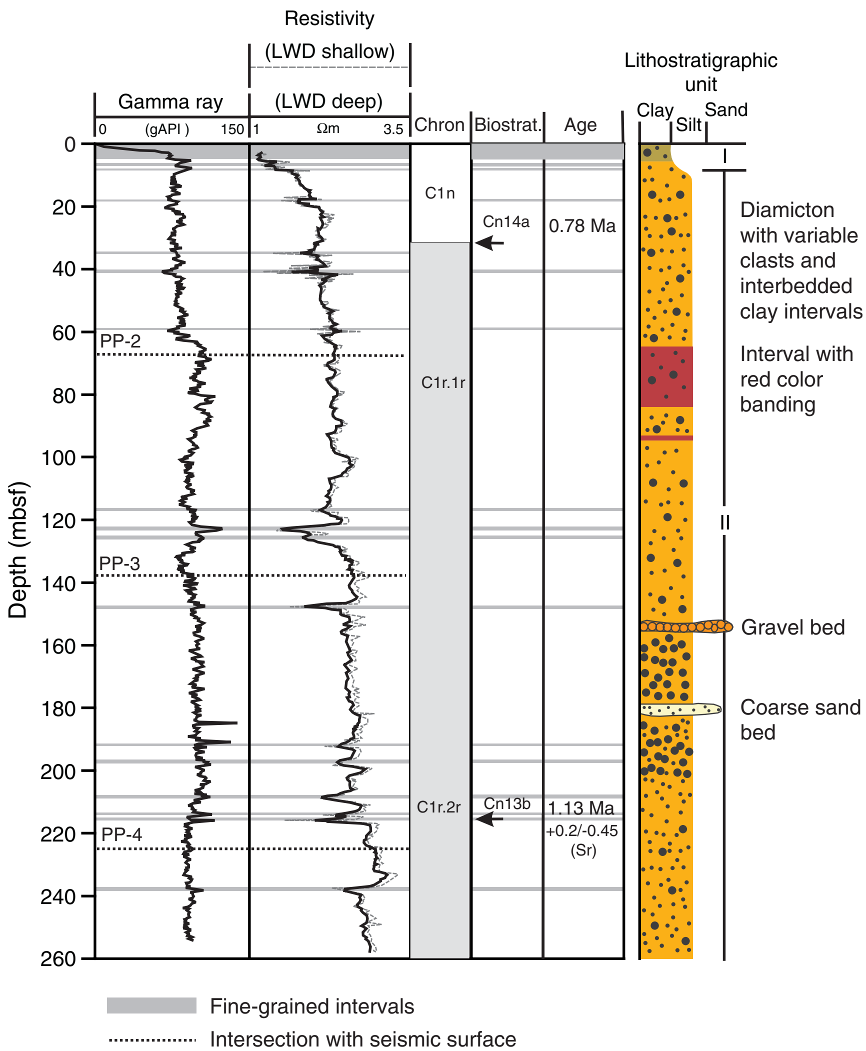

Figure F10. Composite diagram showing a detailed example of cyclic sedimentation in the upper Miocene section of Site 1165 (from Rebesco, this volume). The centimeter-scale cycles are characterized by color banding and lamination as well as by changes in sediment physical properties.

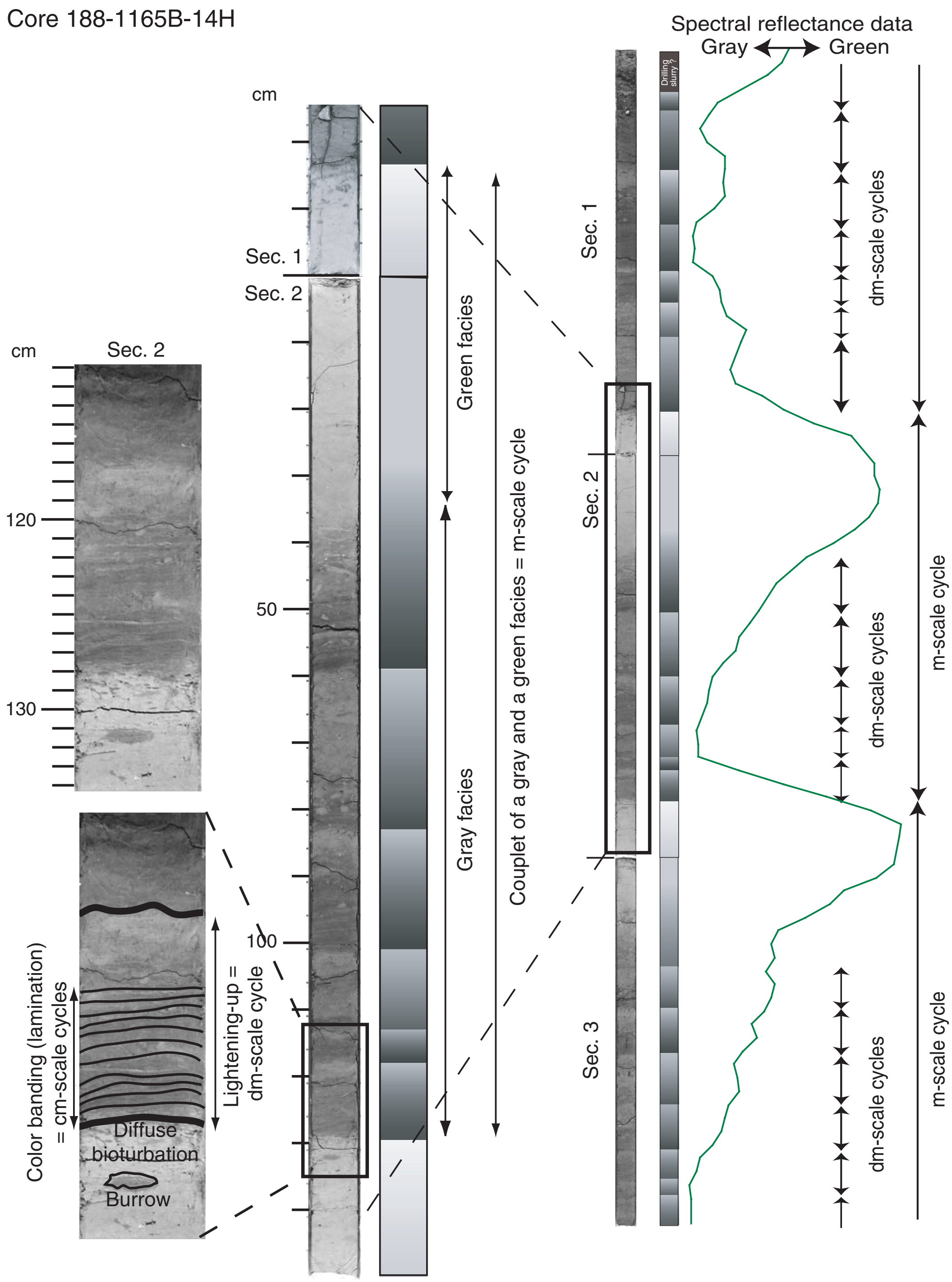

Figure F11. Data curves for the upper part of Site 1165 (80–100 mbsf) illustrating apparent Milankovitch-frequency fluctuations (from Grützner et al., 2003). **A.** data plots of magnetic susceptibility, green/gray color ratio, and Fe-intensity data. Triangles = samples with >10% sand-sized particles (i.e., >250 µm), arrows = ice-rafted debris (IRD) layers. Magnetostratigraphic age control is shown at right. **B.** Cross-spectral analyses of green/gray ratios and Fe records in depth domain. Peaks shown correspond to cycles of 84.1, 41.9, 20.0, and 16.8 k.y. durations.

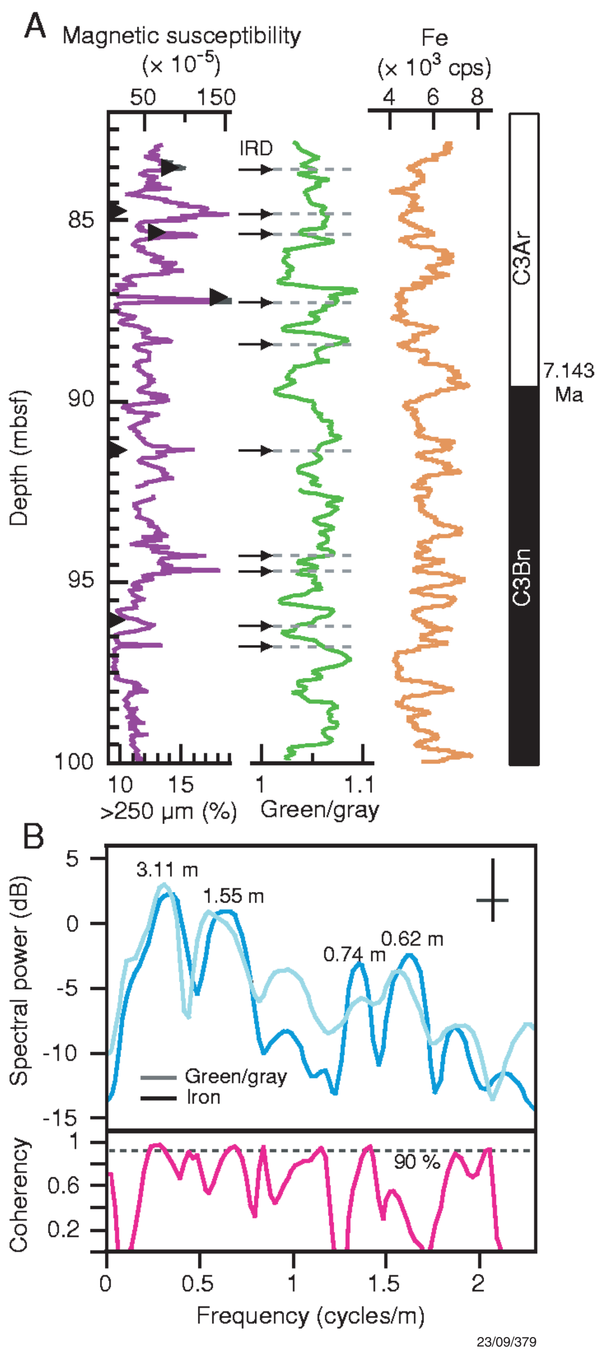

Figure F12. Data profiles for the lower part of Site 1165 illustrating cyclicity in downhole logging data from Williams et al. (2002, pers. comm., 2004). The figure shows core recovery, Facies III-2 (predominately siliceous claystone), resistivity data, gamma radiation data, magnetostratigraphic boundaries, and vertical spacing of resistivity peaks (i.e., cycle thickness). SFLU = spherically focused resistivity, CGR = computed gamma ray.

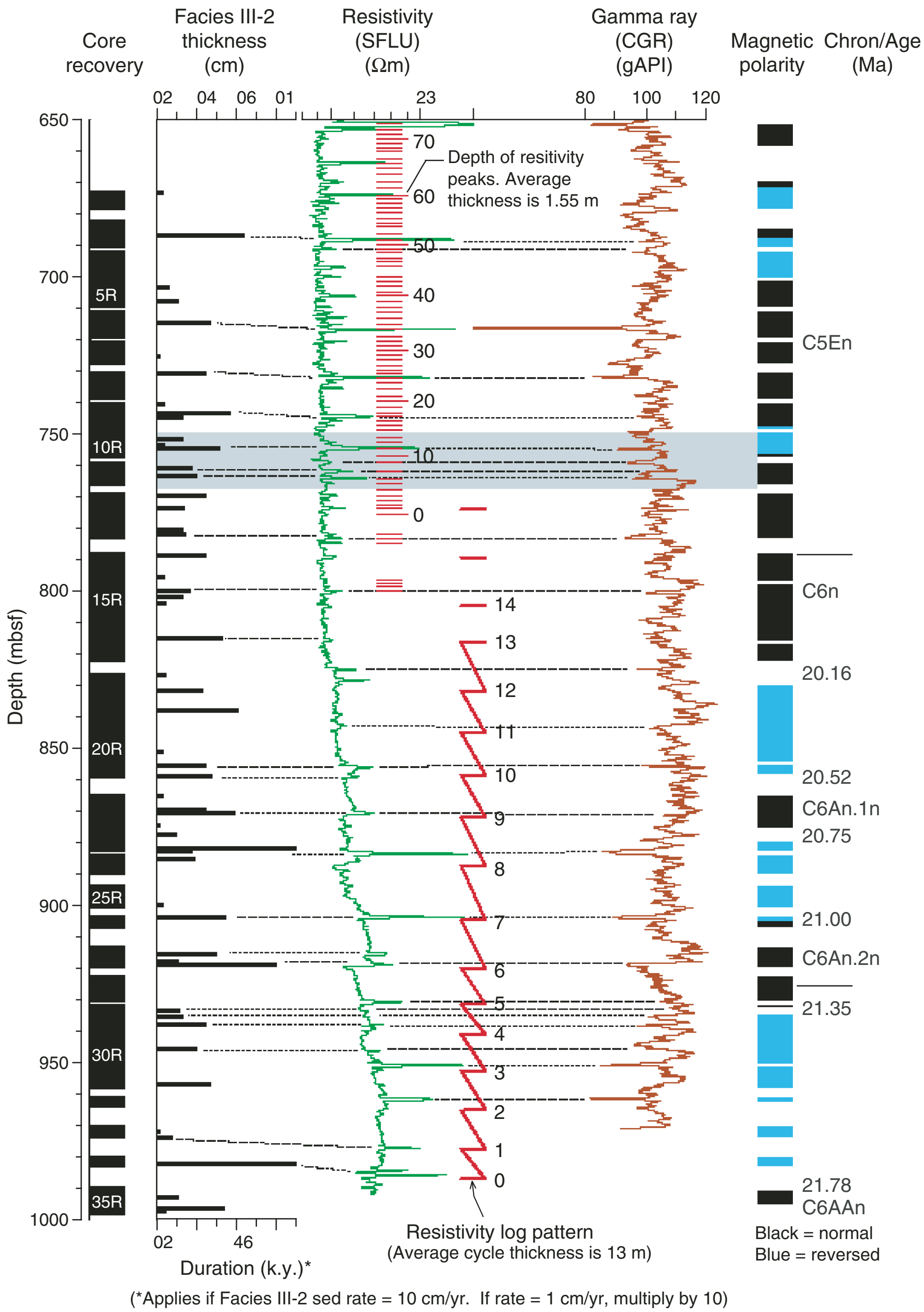

Figure F13. Seismic reflection profile with lithostratigraphy and synthetic seismic trace for Site 1165 (modified from **Handwerger et al.,** this volume).The strong band of reflections at ~5600 ms results from alternating hard and soft layers found at the diagenetic transition from opal-A to inferred opal-CT silica. The largest reflection packet at ~5650 ms is a regional bottom-simulating-reflection (BSR) that results from the abrupt physical property changes at 608 mbsf, a depth below which diatoms are not found. P3 is a regional reflector discussed in the text. TD = total depth.

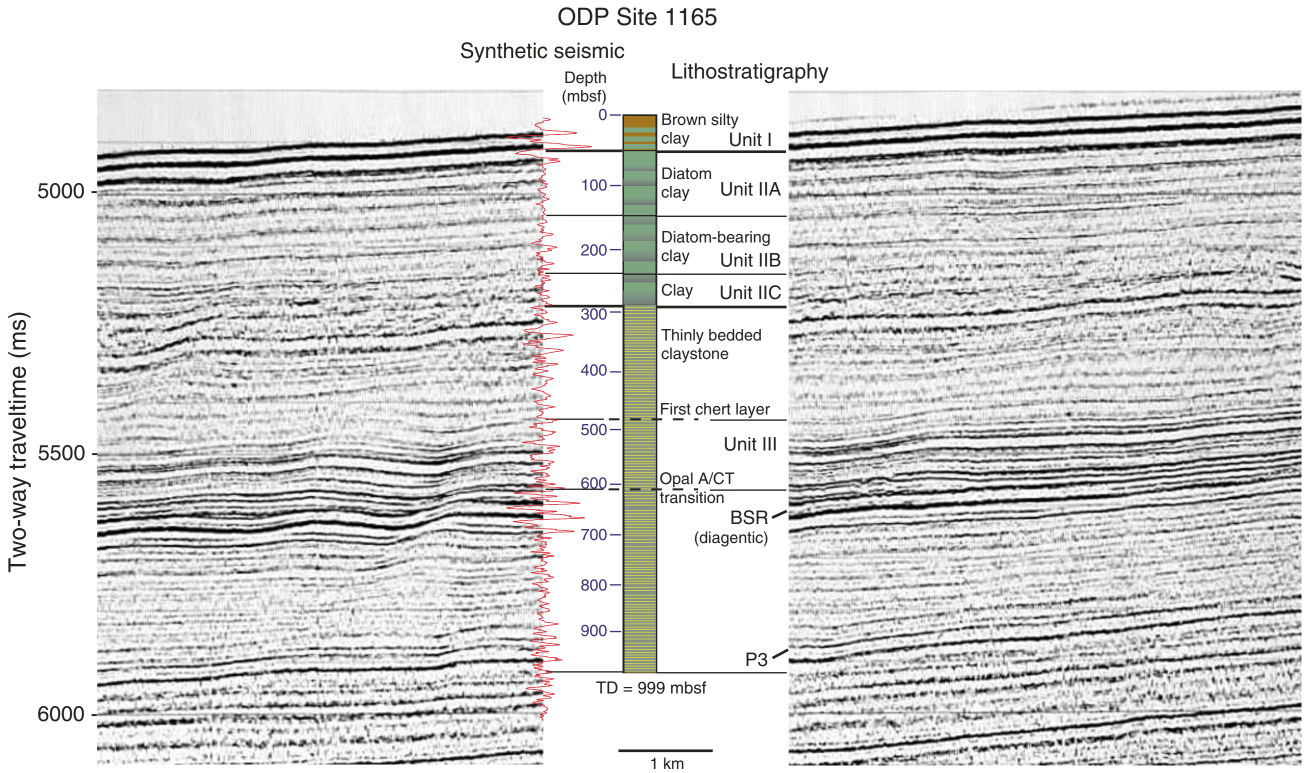

Figure F14. Downhole profiles from Site 1165 illustrating diagenetic effects on silica, magnetite, and carbon in varied paleoenvironments. **A.** Downhole resistivity log and shipboard multisensor track (MST) magnetic susceptibility measurements from Shipboard Scientific Party (2001b) and **B.** Isotopic $\delta^{13}C$ measurements from Claypool et al. (this volume).

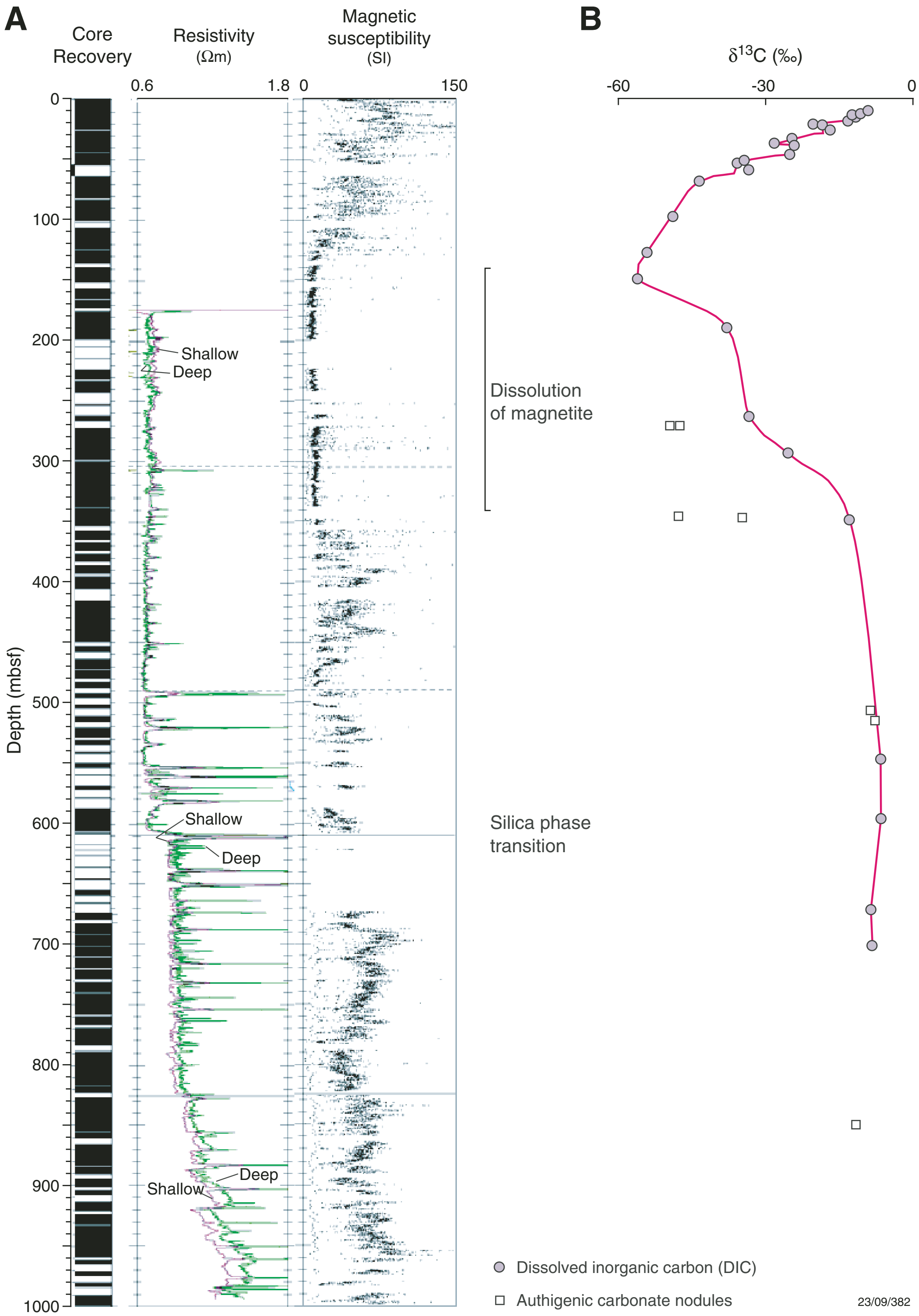

Figure F15. Summary diagram showing geologic and glacial events in the Prydz Bay region, compared with the global oxygen isotope curve of Zachos et al. (2001). Sedimentation rates are from Shipboard Scientific Party (2001b) and Florindo et al. (2003a). Transitions in the Prydz Bay region have both long-term and short-term components like those of the isotopic curve. Long-term decreases in the middle Miocene and younger sedimentation rates on the continental rise mimic the pronounced cooling trend of the global oceans.