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Glacial History of the Antarctic Peninsula from Pacific Margin Sediments[1]

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ABSTRACT

The Pacific margin of the Antarctic Peninsula was drilled during Ocean Drilling Program (ODP) Leg 178 to understand the past 10 m.y. of its glacial history and to test a drilling strategy that might be applied in other regions of the Antarctic margin. This paper offers a mature view of the achievements of drilling, which succeeded in both aims, and focuses on the lessons for subsequent margin drilling and the results of postcruise studies on samples and data from the leg. One of two complementary depositional environments drilled was the topset component of the glacial prograded wedge of the outer continental shelf. Despite poor recovery, a Pliocene–Pleistocene age was found for the major glacial sedimentary sequence groups S1 and S2 previously defined using seismic reflection survey, and a glacial nature and late Miocene age was established for the sampled part of the underlying sequence group S3. Such broad, low-resolution conclusions are possible from shelf drilling using presently available techniques and are useful in terms of Antarctic glacial history, but may represent the limit of what can be achieved if core recovery is poor. The second depositional environment was the glacially derived fine-grained sediment drifts on the upper continental rise, which gave continuous recovery and a high-resolution record of the past 10 m.y. Biogenic opal concentrations showed that the southeast Pacific Ocean (through variations in temperature and sea-ice cover) behaved compatibly with global climate, with a cool late Miocene and warm early Pliocene, cooling to a cold Pleistocene. Nevertheless, ice-rafted detritus confirmed that the Antarctic Peninsula ice sheet was present throughout this period, and clay miner-

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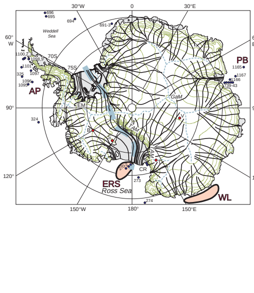
alogy showed that it remained sufficiently large for regular grounding-line migration to the shelf edge. Spectral analysis of sedimentary parameters at drift Sites 1095 and 1096 did not show the dominance of frequencies usually associated with orbital insolation variation. It is therefore uncertain if the glacial–interglacial cyclicity evident in sampled sediments deposited off the Antarctic Peninsula before the late Pliocene–Pleistocene growth of large Northern Hemisphere ice sheets was caused by an orbitally driven or an essentially autocyclic variation in Antarctic ice sheet volume. The matter needs further investigation. An Antarctic Peninsula ice sheet existed for the entire period sampled by drilling, and glacial onset was earlier. The onshore and offshore Antarctic Peninsula records suggest very different times of onset. They, and the oceanic benthic oxygen isotopic record, can be reconciled by positing a moderately glacial Oligocene (extending briefly into the earliest Miocene), a warmer (perhaps nonglacial) early Miocene (to 15–17 Ma), followed by a moderately glacial late Miocene and a more deeply glacial Pliocene–Pleistocene. In addition, in the sense that much of the work reported in this volume is at an interim stage, more insights may yet emerge.

INTRODUCTION

The reasons why the Pacific margin of the Antarctic Peninsula was drilled (during Ocean Drilling Program [ODP] Leg 178) have been described in detail by Barker and Camerlenghi (1999). Briefly, the drilling proposal on which the leg was based was one of a linked series (constructed by ANTOSTRAT Regional Working Groups and refined by an ODP Detailed Planning Group; see also Barker et al., 1998, 1999) intended to elucidate Antarctic glacial history, which otherwise is known only from sparse onshore outcrop and existing offshore drilling (for a review, see Barrett, 1996) and from ambiguous and conflicting low-latitude proxy measurements. ANTOSTRAT proposed drilling at several locations around the Antarctic margin, combining the results by means of a numerical model of glaciation (Huybrechts, 1992, 1993). The Antarctic Peninsula's glacial history was thought to have been shorter than that of East Antarctica, and its ice volume would never have been large. Nevertheless, it was the first region of the Antarctic margin to be drilled of those proposed by ANTOSTRAT. It was selected because of the simplicity of its geology, because its narrow ice catchment and anomalously high precipitation would have given rise to a high-resolution glacial record (covering perhaps the Pliocene–Pleistocene and late Miocene), and because (partly on account of the wealth of available marine seismic reflection data) the main features of the region were known and understood. This last was important because of the need to test the strategy of extracting Antarctic glacial history (and therefore the histories of glacioeustatic sea level change and of ice-volume change of oceanic isotopic composition) by direct drilling of glacially transported sediments before drilling in regions with a longer or more complicated glacial history.

The focus of this paper is on developments since publication of the Leg 178 *Initial Reports* volume (Barker, Camerlenghi, Acton, et al., 1999). A second leg of those proposed by ANTOSTRAT to address the question of Antarctic glacial history has been drilled meanwhile (O'Brien, Cooper, Richter, et al., 2001), in Prydz Bay, East Antarctica (Fig. F1), and related proposals for drilling off Wilkes Land and in the

F1. Antarctic ice sheet ice flow lines, p. 32.



eastern Ross Sea remain under consideration by the advisory structure for ocean drilling. Glacially transported marine sediments have been drilled inshore in the western Ross Sea by the Cape Roberts Project (Cape Roberts Science Team, 1998, 1999, 2000).

Publication in the ODP *Scientific Results* volume is made available to the shipboard scientific party and others working on samples and data from the drilling leg, but the volume is intended to be flexible so as to respond to whatever role it is given. Many Leg 178 workers have used it to publish interim results, intending to undertake further work or to combine data sets for subsequent external publication. Indeed, many have submitted data reports, within which interpretation and speculation are considered inappropriate. Also, of course, workers have not had access to essential data at a sufficiently early stage (for example, the stratigraphic synthesis and the revised calculation of mean composite depths for two of the continental rise sites could not be made available to those working on rise samples). This paper attempts to synthesize our understanding of Antarctic Peninsula glacial history as it now stands as a result of the papers and data reports published here and taking into account the external literature. Also, however, recognizing the interim nature of some of the work, we include comments on the potential value of studies on Leg 178 samples and data not yet accomplished or completed, which might be out of place in a final synthesis but are justified here.

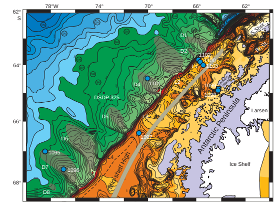
Antarctic Peninsula Topography, Geology, and Climate

The Antarctic Peninsula forms a long, narrow dissected plateau extending southwest from ~63°S and merging into West Antarctica at ~74°S. It gives the appearance of an elevated peneplain at ~900 m at the northern end, rising to 1750 m at ~65°S and remaining at or above that level farther south (Elliot, 1997). Its crest is wider in the south but is dissected everywhere by steep-sided fjords that reach well below sea level, and many offshore islands are found on its western (Pacific) continental shelf. That shelf is typical of most glaciated shelves, in being deeper than low-latitude continental shelves and having a reverse (inward) slope. Within the inshore fjords, water depths commonly reach 1000 m and can exceed 1400 m, whereas depths are typically 300–500 m at the continental shelf edge. The continental slope on the western side is abnormally steep (up to 17°).

The elevated Antarctic Peninsula acts as a major barrier to tropospheric circulation and currently receives a relatively high snowfall, almost four times the continental average (Reynolds, 1981; Drewry and Morris, 1992). Over most of the Antarctic Peninsula the climate is fully polar; modern meltwater-influenced sedimentation is known only from the offshore South Shetland Islands in the far north (e.g., Yang and Harwood, 1997). Yet the permanent ice cover on the peninsula spine is thin, and ice sheet drainage occurs mainly through steep ice falls at the heads of fjords, with grounding lines very far inshore and only rare fringing ice shelves. There is considerable evidence, both direct and indirect, that a grounded ice sheet extended to the continental shelf edge at glacial maxima (e.g., Larter and Barker, 1989; Pudsey et al., 1994; Bart and Anderson, 1995; Camerlenghi et al., 1997b; Barker, Camerlenghi, Acton, et al., 1999). In short, the Antarctic Peninsula seems to have had a small-catchment, small-reservoir, high-throughput glacial regime that should provide a high-resolution climate record.

The Antarctic Peninsula has been the site of subduction of Pacific oceanic lithosphere for at least 150 m.y. and probably since long before the breakup of Gondwana. Its onshore geology, comprising magmatic arc and related volcanic products and deformed and metamorphosed sediments of a series of accretionary prisms (e.g., Dalziel, 1984; Barker et al., 1991; Leat et al., 1995) attests to this history. Outcrop along the spine and on islands along the dissected and overdeepened inner shelf is mostly of plutonic rocks. Part of the middle shelf of the Pacific margin is occupied by a sedimentary basin, thought to have been the upper-slope basin of a subducting margin. A mid-shelf high (MSH) at the outer edge of this basin (Fig. F2) is considered to have been a midslope high, as seen on many subducting margins, and possibly to represent the edge of the rigid overriding plate. The present outer shelf is most probably underlain at depth by accretionary prism material.

F2. Leg 178 sites, p. 33.



Subduction stopped during the Cenozoic with the arrival of a spreading ridge crest at the trench, progressively later northeastward along the margin (Herron and Tucholke, 1976; Barker, 1982; Larter and Barker, 1991a). For example, ridge-crest collision occurred at ~30 Ma south of 67°S, at 3–6 Ma at ~62.5°S, and at a series of intermediate ages in between (see fig. F7 of Barker and Camerlenghi, 1999). North of 62.5°S, where lies the 4700-m-deep South Shetland Trench and an active backarc extensional Bransfield Strait (Br. St. in Fig. F2) (see also Lawver et al., 1995; Barker and Austin, 1998), spreading stopped before the ridge crest reached the margin but subduction probably continues at the trench (Barker and Dalziel, 1983; Maldonado et al., 1994; Kim et al., 1995).

The ridge-crest collision had several significant effects. Subduction-related magmatism ceased some time before collision (Barker, 1982), but a recent phase of alkalic volcanism has been related by some to creation of a "slab window" following collision (e.g., Hole and Larter, 1993), though the age and location of some occurrences do not fit a simple story (Barber et al., 1991). Oceanic magnetic lineations young toward the margin, indicating collision, but a magnetic quiet zone close to the margin probably signifies an abundance of terrigenous sediment prior to collision, sufficient to bury the approaching ridge crest. Deep Sea Drilling Project (DSDP) Site 325 (Hollister, Craddock, et al., 1976), in 3748 m water depth and ~180 km from the continental shelf edge (Fig. F2), shows a loss of terrigenous sediment from 1 to 7 m.y. *after* collision (Hollister, Craddock, et al., 1976; D. Lazarus, M. Iwai, L. Osterman, and D. Winter, pers. comm., 2000), which has been interpreted as sediment deflection caused by uplift of the MSH, thermal (Larter and Barker, 1991a) and possibly also hydrothermal (Larter et al., 1997) in origin, followed by passive-margin-like subsidence. The occurrence of the same process (collision followed by temporary MSH uplift and consequent deflection of terrigenous sediment) along the margin has been used to constrain ages of shelf and rise sediments.

ODP Leg 178 drill sites on the continental shelf and rise were chosen to avoid interference between tectonics (collision and uplift) and climate change—the strategy was to sample younger units in the northeast and older units in the southwest. Site 1103 was an exception to this, but its occupation was forced upon the shipboard party as a result of poor penetration and recovery at other shelf transect sites. Drilling was planned also to avoid the young extensional zone in Bransfield Strait. It was thought at one time that the South Shetland margin (including Bransfield Strait) provided a useful example of the margin before collision, but such a comparison neglects changes in climate with time and the backarc extension in Bransfield Strait, which is not mirrored farther south.

The other significant feature of the Antarctic Peninsula is the elevation of its spine. If that spine is a peneplain, then uplift from close to sea level is implied. Such uplift has not been dated, and several hypotheses have been erected to explain it. It is not simply associated with ridge-crest subduction. A relation is inferred by some between uplift and climate change via the development of a topographic barrier to atmospheric circulation (see Elliot, 1997), so that the time of uplift is important to climate history.

GLACIAL DEPOSITIONAL ENVIRONMENTS AND LEG 178 DRILLING

Four sites (1097, 1100, 1102, and 1103) were drilled on the outer continental shelf during ODP Leg 178, and three sites (1095, 1096, and 1101) were drilled on sediment drifts on the upper continental rise (located in Fig. F2 and described in Barker, Camerlenghi, Acton, et al., 1999). The continental shelf and rise are the two main depositional environments identified by ANTOSTRAT as containing an accessible record of long-term glacial history. In addition, two sites (1098 and 1099) were drilled within Palmer Deep, an inner-shelf basin south of Anvers Island, where an ultra high resolution postglacial (Holocene) sedimentary record is preserved. This paper does not address the results of Palmer Deep drilling, which is the subject of a separate synthesis (Domack, Chap. 34, this volume).

The main aims of Leg 178 drilling, sampling shelf and rise environments for information on glacial history, confirming models of margin evolution, validating the drilling strategy, and identifying some of the technical problems involved, were successfully achieved. It was shown that a glacial environment existed along the Antarctic Peninsula throughout the last 9–10 m.y., with glacial deposition on both shelf and rise but a change at ~4.5 Ma to progradation of the outer continental shelf (Barker, Camerlenghi, Acton, et al., 1999). Predictably, the continuously deposited and more readily recovered sediments of the continental rise have attracted much more attention postcruise than those of the continental shelf.

Continental Shelf and Slope

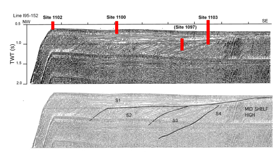
The depositional parts of the high-latitude (glacial) outer continental shelf and slope are underlain by a prograded wedge, most probably largely composed of diamict. This was mostly transported as a low-viscosity, sheared, overpressured basal till beneath low-profile ice streams (the major means of ice sheet drainage) that were grounded to the shelf edge during glacial maxima and was deposited in "trough-mouth fans" at the outlets of major ice drainage areas as topsets on the outer part of the (generally overdeepened and inward sloping) shelf and as foresets on the uppermost slope (Alley et al., 1989; Larter and Barker, 1989, 1991b; Vorren et al., 1989; Boulton, 1990; Bartek et al., 1991; Cooper et al., 1991; Pope and Anderson, 1992; Pudsey et al., 1994; Vanneste and Larter, 1995). During interglacials, the ice sheet grounding line is well inshore and deposition on shelf and slope is slow, comprising mainly fine-grained hemipelagic and diatomaceous muds and oozes. Subsequent ice streams may erode previously deposited shelf sediment, so the shelf record is typically incomplete. On the Antarctic Peninsula Pacific margin between ~63° and 68°S, Pliocene–Pleistocene deposits (the progradational Unit S2 and overlying, usually less progradational Unit S1 of Larter and Barker, 1989) are focused into four lobes (Larter and Cunningham, 1993), within which both topsets and prograding foresets are thickest; Unit (more precisely, sequence group, since S1, S2, and S3 each equate to deposition during many glacial–interglacial cycles) S1 may not be continuous between all lobes. The Antarctic Peninsula continental slope is unusually steep in both the lobe and the interlobe areas (Fig. F2). While most workers assume that deposits within the lobes were transported there by ice streams, which may diverge or migrate in order to allow deposition to take place over a wider area (the conventional view of glacial deposition on shelf and slope), Rebesco et al. (1998) now consider, on the basis of shelf bathymetry, that ice streams flowed across the interlobe areas, transporting the bulk of glacial sediment across the shelf with relatively little deposition there or on the slope, and that the lobes were formed by sediment transported on and deposited from slower-flowing ice-stream flanks. This divergence of view persists but is not addressed further here since it has little or no effect upon the value of shelf, slope, and rise sediments as recorders of glacial history.

Three Leg 178 sites (1100, 1102, and 1103) lie on a transect along the axis of Lobe 1 (Fig. F2), and one (Site 1097) was located between Lobes 3 and 4, where access to deeper layers was better (Shipboard Scientific Party, 1999a, 1999c). All sites were drilled using the rotary core barrel (RCB). Penetration was low at Sites 1100 and 1102. At the other sites, recovery was very low within the upper layers, where the finer-grained matrix of the diamicts was insufficiently consolidated to support the clasts during drilling. Even at greater depths, where the diamicts were more consolidated, recovery improved only to a maximum of 34% in the lower 115 m of Site 1103.

The geometry of the glacial shelf deposits and the achievements of shelf drilling may be summarized with respect to Figure F3, a seismic reflection profile along the axis of progradational Lobe 1, the line of the Shelf Transect (Shipboard Scientific Party, 1999a), with the effective penetration of Site 1097 (from an interlobe area) projected onto it. Briefly, penetration at Site 1100 was only 110 meters below seafloor (mbsf), entirely within sequence group S1 topsets, and recovery was very low. At Site 1102 on the continental shelf edge, penetration was even less (only 15 mbsf), but a camera survey (see the video recording in Barker, Camerlenghi, Acton, et al., 1999), undertaken during a swell-induced pause in drilling, revealed a boulder carapace on the seabed that we interpreted as the result of bottom-current winnowing of fine-grained sediment following disturbance of diamicts by iceberg keels. This concentration of coarse material, we thought, while inhibiting penetration and recovery, might explain the strong seismic reflectors seen at many paleoshelf breaks.

Figure F3 shows that drilling at Site 1103 penetrated seismic sequence group S3 (to 363 mbsf) below an unconformity at the base of sequence group S1; sequence group S2 is absent. In contrast, sequence group S3 was sampled at Site 1097 (to 437 mbsf) below an apparently conformable boundary with sequence group S2 and thin sequence groups S1 and S2. Before drilling, sequence group S3 was considered "preglacial" by some (e.g., Larter et al., 1997), but drilling showed its glacial nature at both sites. Its base was not reached at either site.

F3. Seismic reflection profile I95-152, p. 34.

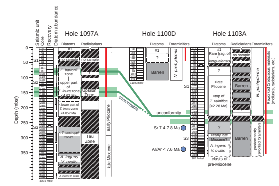


Sequence group S3 has the appearance that it shows in Figure **F3** along the entire margin. It shows parallel to gently divergent bedding on seismic reflection profiles, and preliminary indications are that, unlike the overlying sequence groups S2 and S1, it is not focused into lobes. Dip of sequence group S3 is horizontal in the vicinity of Site 1097 and gently seaward around Site 1103 (Fig. **F3**). If the sequence group S3 bedding is considered as shelf topsets (and the bedding dip in Fig. **F3** results from postdepositional tilt), then the sequence group lacks foresets and is aggradational. In this it contrasts markedly with the overlying, clearly progradational sequence group S2. On many profiles along the margin it is possible to identify a transition to foreset development, with topset truncation in places, formally placed at the base of sequence group S2, as in Figure **F3.** On board ship (Shipboard Scientific Party, 1999a, 1999c), the state of preservation of benthic foraminifers and other shell material was used to distinguish depositional environments for sequence group S3, ranging from subglacial through proximal proglacial to glacial marine at Site 1097, taken as indicating a continental shelf environment. At Site 1103, the same methods showed no evidence of subglacial deposition (a glacial shelf indicator) and the sequence group S3 sediments showed evidence of sorting and downslope movement, leading to a division of opinion on board ship between a shelf and an upper-slope paleoenvironment.

Throughout the continental shelf holes, even within sequence groups S2 and S1, a sparse admixture of biogenic material, mainly diatoms, allowed some age attribution despite poor recovery and some reworking. It seemed likely that, on the Antarctic Peninsula shelf at least, the ice-base processes responsible for terrigenous sediment transport during glacials had usually involved also a degree of erosion and admixture of the biogenic and fine-grained hemipelagic sediment deposited during preceding interglacials. If the microfossils were relatively fresh and whole, making identification easier, then both the time interval between original deposition, erosion, and redeposition and the transport path seemed likely to have been short. In the context of the relatively high-energy environment of the glacial continental shelf, such dating can be considered precise. Thus, the sequence group S3–S2 transition at Site 1097 (considered conformable on the evidence of seismic reflection profiles), was dated as lying within the *Thalassiosira inura* diatom zone, at ~4.5 Ma (although if the sequence group S3/S2 boundary is located beneath the deepest evident foreset, this transition could be slightly older). The lower drilled part of sequence group S3 at Site 1097 and all of sequence group S3 sampled below the S3/S1 unconformity at Site 1103 was considered to lie within the *Actinocyclus ingens* v. *ovalis* diatom zone (6.3–8.6 Ma). It seems reasonable to conclude that S3 is a synchronous depositional sequence group, of which the uppermost part was sampled at Site 1097 but not at Site 1103, as Figure **F3** suggests. The uniform, parallel-bedded character of sequence group S3 on seismic reflection profiles argues against it representing two contrasting depositional environments. Eyles et al. (2001) propose that the sorting detected in some sediments at Site 1103 reflects a continental slope environment, but many others have recorded gravity sliding (resulting in some sorting) on the glacial shelf, down the nose of a "till tongue" ("till delta" or "subglacial delta"), for example (Alley et al., 1989; Vorren et al., 1989; Bart and Anderson, 1995; Vanneste and Larter, 1995). These considerations and the partly subglacial environment of deposition of sequence group S3 sediments at Site 1097 suggest to us that sequence group S3 was everywhere a shelf deposit. It was concluded also that sequence group S3 reflected in some way a less glacial climate than that under which the succeeding sequence groups S2 and S1 were deposited, but it is difficult to be confident about this, given the extremely poor recovery at shallow depths; certainly the recovered sequence group S3 sediments showed a wider range of environments, some of them in a sense less glacial than those of recovered sequence group S2 and S1 sediments, but this is capable of more than one explanation.

The very poor recovery also ensured that the sequence group S2/S1 boundary was not precisely dated at Site 1097. The available diatom ages from sequence group S1 were Quaternary and possibly late Pliocene and from sequence group S2 were Pliocene (Barker, Camerlenghi, Acton, et al., 1999) (Fig. **F4**).

F4. Stratigraphic control on shelf Sites 1097, 1100, and 1103, p. 35.



Postcruise work on data and samples from the drilling leg has been devoted to improving the age constraints on sequence group S3 sediments and the precision with which data from the drill sites could be used to constrain the ages and natures of seismic sequence groups identified in the very large seismic reflection data set from the shelf. In addition, Camerlenghi et al. (in press) undertook a decompaction and backstripping experiment on the sediments along the line of the shelf transect (Fig. **F3**), the axis of deposition of progradational Lobe 1. This experiment, considered preliminary by the authors, is a useful attempt to assess quantitatively the flexural response to sediment loading, thermal subsidence from ridge-crest subduction, and sediment compaction under non-hydrostatic load. The authors conclude that the present overdeepened and inward-sloping shelf, considered to indicate unequivocally a fully glacial regime, developed only within the time of deposition of sequence group S2 and suggests therefore that the environment of deposition of the older sequence group S3 was indeed less glacial. They find these conclusions robust to a range of flexural, thermal, and compaction parameters. The analysis has other implications: neither the parallel bedding of sequence group S3 nor the topsets of the presumed topset/foreset couples of the lower part of sequence group S2 (see Fig. **F3**) are raised to the horizontal within the model; little of their present-day dip is eliminated. This would argue against interpretation of the depositional environment of sequence group S3 as continental shelf. Moreover, also it rules out the topset/foreset interpretation of the basal sequence group S2 features, which are common along the margin and in all other respects resemble features that are characteristic of the sequence group S2/S1 boundary and are common within sequence group S1. One possible explanation lies in the model's assumption of uniform, noncompactible material beneath sequence group S3 all the way along the model profile. To the southeast (inshore) this material is the high-velocity mid-shelf high, where the assumption almost certainly holds, but to the northwest (offshore) it becomes the precollision sequence S4 sediments and their more distal sedimentary equivalents of the precollision accretionary prism. Many accretionary prisms have low velocity and low density toward the trench (e.g., Cochrane et al., 1996) and would compact significantly under load. We should perhaps await more detailed modeling before drawing firm conclusions about the depositional environment of sequence groups S2 and S3 and the implications for paleoclimate.

The rather sparse shipboard constraints on the age of sequence group S3 have been augmented postcruise by two different investigations and by additional work on microfossils (M. Iwai and L. Osterman, pers. comm., 2001) (Fig. **F4**). The assignment of the depth range 320–355 mbsf at Site 1103 to the diatom *A. ingens* v. *ovalis* Zone (6.3–8.6 Ma) is supported in general terms by Sr isotopic ages of 7.4 and 7.8 Ma on barnacle fragments from 262 to 263 mbsf (Lavelle et al., **Chap. 27,** this volume) and by the youngest (7.6 Ma) of a range of $^{40}Ar/^{39}Ar$ ages for volcanic glass fragments from 337 mbsf (Di Vincenzo et al., **Chap. 22,** this volume). Both ages are subject to uncertainties; the barnacles are well preserved and their ages cluster well, but preservation of the fragments is slightly poorer than that of a single fragment from one of the samples that gave an older age. The argon age is merely the youngest of a wide range of ages, each from a group of several small grains of volcanic glass present in low concentrations within the sediments and showing signs of mechanical reworking, and is unsupported by studies on potassium and argon mobility. Nevertheless, the close coincidence of all these ages supports the view that much of the sampled part of sequence group S3 is of late Miocene age. A third study of pollen and spores and nannofossils from the shelf sites (Iwai et al., **Chap. 28,** this volume), has yielded mainly rare forms, unfortunately probably recycled and of little stratigraphic or paleoclimatic value.

Two studies were aimed at making more precise use of data from the drill sites for calibration of the large seismic reflection data set from the continental shelf. At Site 1103, the seismic velocity information provided by downhole logging was of doubtful quality initially, but Moerz et al. (**Chap. 19,** this volume) have undertaken a painstaking reanalysis of the data in a successful attempt to eliminate instrumental uncertainties. Tinivella et al. (**Chap. 16,** this volume) describe tomographic studies of seismic velocity at five points along seismic reflection profile I95-152, which coincides with the shelf transect (Fig. **F3**). The tomographic analysis shows a large velocity increase with depth within topsets and high velocities in underlying foresets. In general, the results suggest that velocity is more strongly related to burial depth than to age or depositional environment, but there are exceptions; in one place (not drilled but equivalent to the section drilled at Site 1097) they show a low velocity within the upper part of sequence group S3, and, at the very shelf edge, foresets have much higher velocities than topsets, even at shallow depth. Where the study can be compared with the downhole log analysis from Site 1103, it shows a strong velocity gradient within the topsets that is not seen in the log data. The optimal correlation of drilling results with seismic reflection profiles is probably made by combining the Site 1103 downhole log velocity data down to ~233 mbsf (below which the only velocity measurements are on sparse shipboard samples) with the tomographic velocities beneath. A study of the large seismic reflection data set from the continental shelf, using the results of drilling, is under way but could not be completed for this volume.

Sediment Drifts on the Continental Rise

A feature of high-latitude glacial sedimentation is that the fine-grained part of the unstable component of sediments deposited from grounded ice on the uppermost continental slope can accumulate within sediment drifts on the upper continental rise (Kuvaas and Leitchenkov, 1992; Tomlinson et al., 1992; McGinnis and Hayes, 1995; Rebesco et al., 1996, 1997; Clausen, 1998), following slumping and turbidity current flow down the slope and suspension as a nepheloid layer within bottom currents. Provided that bottom currents are slow and that the slope residence time of this unstable component of uppermost slope deposits is short compared with a glacial cycle (that is, the instability involves only small-scale mass wasting), the drift sediments can provide an indirect but continuous high-resolution and recoverable record of glacial history. A lesser interglacial sea-ice cover permits primary biogenic production, hence pelagic deposition and dating. Off the Pacific margin of the Antarctic Peninsula these conditions appear to hold; current meter measurements (Camerlenghi et al., 1997a) have revealed relatively slow bottom currents (contour-parallel but generally southwest flowing with mean speed of 6.2 cm/s and not exceeding 20 cm/s over a 10-month period in 1995), there is indirect evidence (e.g., Pudsey, 1992) that glacial-age currents were no faster, and the character of the upper continental slope (slump headwalls and gullying seen on deep-tow boomer profiles—Vanneste and Larter, 1995) indicates that recent instability has been small scale. Both piston cores (Camerlenghi et al., 1997b; Pudsey and Camerlenghi, 1998; Pudsey, 2000; Lucchi et al., in press) and the preliminary results of Leg 178 drilling have verified the preserved and dateable record in the drifts of late Pleistocene glacial–interglacial variability.

Most of the drifts (including those drilled) are separated from the continental slope and from each other (Fig. F2) by a dendritic pattern of channels originating at the base of slope (e.g., Tomlinson et al., 1992), with axes up to 1 km deeper than the drift crests. The channels are maintained by turbidity currents, which carry the coarse silt- and sand-sized components of the unstable upper-slope sediments to the abyssal plain. These drifts are considered an end-member of the drift category, since their topographic distinction from the channels is maintained (it is thought) not by bottom currents (which here are weak), but by erosion by the turbidity currents themselves. Thus, it is possible for interglacial, mainly pelagic sediments to be eroded and resuspended by the turbidity currents on the rise itself (within the channels) and incorporated into the drifts. Rebesco et al. (in press) recently summarized all of the available marine geophysical and marine geological data from the Antarctic Peninsula sediment drifts. Of particular interest is the swath bathymetry in the northeastern area (Drifts 1–3 and part of Drift 4) mentioned originally by Canals et al. (1998), which supersedes the GLORIA side-scan data reported by Tomlinson et al. (1992) and clearly shows both the essentially asymmetric drifts and intervening channels.

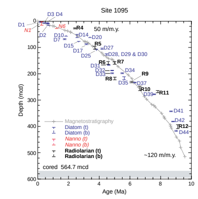
Three sites on the continental rise were drilled during Leg 178 (Fig. F2) (Barker, Camerlenghi, Acton, et al., 1999). Two sites together examined a single drift (Drift 7 of Rebesco et al., 1996). Site 1096 was the closer to the continental shelf, in 3152 m water depth, and sampled an expanded section extending back to 4.7 Ma by combined use of the advanced hydraulic piston corer (APC) and extended core barrel (XCB) to 608 mbsf, with repeated APC coring at shallow depth in order to obtain a more complete section. Site 1095, in 3842 m water depth at a more distal location, where the lower section was more accessible because the upper section was thinner, sampled to ~10 Ma at a maximum depth of 570 mbsf, again with repeated APC coring to improve recovery in the shallow section. At both sites the deep hole was logged almost to maximum depth. This dual-site strategy avoided drilling through a diffuse bottom-simulating reflector seen on seismic reflection profiles at ~700 ms depth and interpreted as caused by silica diagenesis (Barker and Camerlenghi, 1999), thus perhaps obliterating siliceous biostratigraphy. The third site (Site 1101), in 3280 m water depth, sampled a different drift (Drift 4 of Rebesco et al., 1996) back to 3.1 Ma in a single APC/XCB hole to 218 mbsf.

Recovery at Sites 1095 and 1096 was high, after taking into account the multiple coring of the shallow section: Site 1096 recovery = 88%, Site 1095 recovery = 94% down to ~484 m (where recovery dropped sharply), or 82% overall. Recovery at Site 1101 was even higher, 99%. It was possible to obtain a magnetostratigraphic record on board ship from cores and logs and to examine the preserved microfossils. Postcruise magnetostratigraphic studies have been undertaken in more detail, both on U-channel samples of the core (Acton et al., **Chap. 37,** this volume) and by means of a rigorous reexamination of the Geological High-Resolution Magnetic Tool (GHMT) log data (Williams et al., **Chap. 31,** this volume). In addition, Brachfeld et al. (**Chap. 14,** this volume) have examined the magnetic mineral assemblage as a contribution toward high-resolution magnetostratigraphic studies such as the detailed U-channel study of part of the Matuyama Chron at Site 1101 (Guyodo et al., 2001).

The biostratigraphic record was more difficult to interpret because of the likelihood that microfossils had been reworked from the shelf and rise before deposition within the drifts, as a result of the dominance of processes of reworking on the continental shelf during glacial periods and gravity sliding with resuspension on the continental slope. Shipboard efforts have been supplemented by additional studies of the main microfossil groups (pollen and spores and nannofossils by Iwai et al., **Chap. 28;** radiolarians by Lazarus, **Chap. 13;** dinoflagellates by Pudsey and Harland, **Chap. 2;** diatoms by Winter and Iwai, **Chap. 29** [supplemented by a report of identification and occurrence at several sites—Iwai and Winter, **Chap. 35**]; and nannofossils by Winter and Wise, **Chap. 26,** all this volume). The shipboard work, many of these studies, and other work have been brought together into a magnetobiostratigraphic synthesis that focuses mainly on Site 1095, entirely on the rise drifts (Iwai et al., **Chap. 36,** this volume). Not all inconsistencies have been fully resolved in this work, but it is likely to stand as a stratigraphic consensus for some time, affecting all future and unfinished work on samples and data from the drift sites. The revised correlation between multiply cored intervals and the calculation of meters composite depth (mcd) at Sites 1095 and 1096 (Barker, **Chap. 6,** this volume) is similar in retaining uncertainties but is likely to come into common use.

In the magnetobiostratigraphic synthesis (Iwai et al., **Chap. 36,** this volume), the magnetostratigraphy of the continental rise sites (1095, 1096, and 1101) has been revised as a result of U-channel measurements (described in detail by Acton et al., **Chap. 37,** this volume) and (for Sites 1095 and 1096) the spliced mcd scale at shallow depths, and merged with the combined results of the wide range of shipboard and postcruise biostratigraphic studies. Several key discrepancies reported within the Leg 178 *Initial Reports* volume have been reexamined, and either a resolution or a most likely path to resolution has been agreed upon. Two developments are particularly significant. First, drilling at Site 1095 was considered to have encountered a hiatus in sedimentation at ~60 mbsf (Shipboard Scientific Party, 1999b) on the basis of the seismic reflection profile through the site and missing magnetic reversals. However, the revised magnetic reversal stratigraphy is much more complete, and the relatively low resolution of the seismic reflection profile is able to accommodate either continuous deposition or a brief hiatus (of 100–200 k.y., for example, as now proposed). Second, there was within the shipboard stratigraphy described in the Leg 178 *Initial Reports* volume an unresolved discrepancy between diatom (also, subsequently, radiolarian) stratigraphy and the magnetic reversal record at around the Miocene/Pliocene boundary at Site 1095. Following reexamination of all data sets, the magnetic reversal stratigraphy is accepted and the discrepancy is now considered to require most probably a reexamination of published Southern Ocean biosiliceous zonation with particular focus on the occurrence of depositional hiatuses at many sites. The revised age-depth curve for Site 1095 is shown as Figure F5.

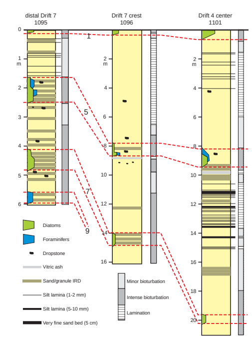
F5. Revised age-depth curve for Site 1095, p. 36.



In addition to the stratigraphy, Iwai et al. (Chap. 36, this volume) also confirm the presence of a diverse assemblage of neritic and benthic diatoms, with particular abundance of the neritic diatom *Paralia sulcata* in upper Miocene (7.5 to 6.7 Ma and slightly younger) sediments at Site 1095, being taken to indicate the existence at that time of a wide and shallow (<100 m) continental shelf. This is within the time of deposition of sequence group S3 sediments on the shelf. *P. sulcata* is also abundant in recovered sediments deposited at Site 1097 on the continental shelf over the same period (M. Iwai, pers. comm., 2001). These occurrences support the conclusion of the decompaction model described above (Camerlenghi et al., in press), that initial glacial overdeepening took place only after the start of deposition of sequence group S2.

Shipboard sedimentologic studies identified a cyclicity in deposition on the drifts that for the late Quaternary could be associated with glacial cycles (Fig. F6); during glacials, with the grounded ice sheet at the continental shelf edge and (presumably) persistent sea-ice cover, drift sedimentation was relatively rapid and comprised gray, terrigenous, finely laminated barren silty clays. Interglacial sediments were thin, brown, diatomaceous, and bioturbated silty clays. Ice-rafted detritus (IRD) was more evident in the bioturbated clays, but the differences in sedimentation rate through the glacial cycle hindered an assessment of absolute rates of ice rafting. At the shallower sites, nannofossils and foraminifers were preserved, at least in the Stage 5 interglacial. This character accords with that seen in piston cores from the drifts (e.g., Pudsey and Camerlenghi, 1998; Pudsey, 2000). Below ~30–50 mbsf at the sites (0.8–2.0 Ma), the alternations of barren, laminated with massive bioturbated and microfossil-bearing beds persist but are less regular. At the distal site (1095) in particular, there are abundant thin silt and mud turbidites within the glacial sequences, and at the two shallower sites (1096 and 1101), nannofossils and foraminifers are preserved in sediments aged between 2.1 and 0.8 Ma, mainly within interglacials.

F6. Uppermost sections of cores at sediment drift Sites 1095, 1096 and 1101, p. 37.



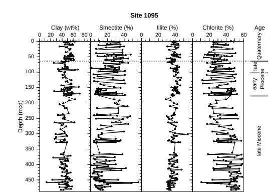
The strong likelihood that the rise sediments had recorded a history of periodic ice sheet advance to the continental shelf edge during glacial cycles, throughout the period examined, has prompted a wide range of detailed postcruise measurements and analyses of sediment features, including IRD (Cowan, Chap. 10, and Hassler and Cowan, Chap. 11, both this volume), grain size (Pudsey, Chap. 12, and Moerz and Wolf-Welling, Chap. 24, both this volume), spectral reflectance (Wolf-Welling et al., Chap. 21, this volume), and a range of other sedimentological, geochemical, and mineralogical parameters (Hillenbrand and Ehrmann, Chap. 8; Hillenbrand and Fütterer, Chap. 23; Kyte and Vakulenko, Chap. 4; Pudsey, Chap. 25; Wolf-Welling et al., Chap. 15, all this volume). Some of these contributions include interpretations of aspects of regional climatic or sedimentologic evolution, but by no means all; their data are in part a substantial resource on which further interpretive work can be based. Hardly any have had access to the consensus stratigraphy or the revised mcd scales noted above.

Two studies have undertaken spectral analyses of downhole log or core parameters. Lauer-Leredde et al. (Chap. 32, this volume) have analyzed several parameters from downhole logs (thorium/potassium ratio and natural gamma at Site 1095 and uranium, neutron-derived porosity, and natural gamma at 1096) and cores (magnetic susceptibility and bioturbated intervals as described on board ship at Site 1095 and chromaticity parameter L* at Site 1096) over much or all of the depths of the deepest hole at each site, and Pudsey (**Chap. 25,** this volume) has included analysis of core magnetic susceptibility and chromaticity parameter a* over three short intervals (0–0.4, 2–3, and 3.7–4.3 Ma) at Site 1095. The results are ambiguous. A wide range of spectral peaks of similar amplitude was found, some of which were interpreted as caused by orbitally induced ("Milankovich") insolation changes, others not. The authors have ignored likely diagenetic effects and have recognized that imperfections in the shipboard stratigraphy could affect the analysis. However, for whatever reason, the orbital cyclicity did not stand out; the questions of when and how the Antarctic Peninsula (or Antarctic) ice sheet and adjacent Southern Ocean responded to orbital forcing are extremely important and remain open (and are discussed further below).

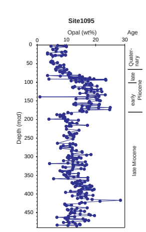
Hillenbrand and Ehrmann (**Chap. 8,** this volume) have used clay mineralogy at Sites 1095 and 1096 to examine the long-term variation in ice sheet behavior. Using a model of the modern glacial–interglacial variation in sediment provenance initiated on board ship (Barker, Camerlenghi, Acton, et al., 1999) and supported by piston core studies by Pudsey (2000) and Hillenbrand (2000) in which mainly downslope, direct transport of sediment high in chlorite during glacials alternates with mainly along-slope (southwestward), indirect transport of sediment high in smectite during interglacials (e.g., Fig. F7), they conclude that periodic migration of the Antarctic Peninsula ice sheet to the continental shelf edge has persisted for the past 9 m.y. with little change. Pudsey (**Chap. 25,** this volume) reports the presence of poorly sorted sand (most probably IRD), mainly within the bioturbated facies, throughout the section at Site 1095, in agreement with these results. Cowan (**Chap. 10,** this volume) describes time variations in IRD (sand fraction) abundance over the past 3.1 m.y. at Site 1101, which are difficult to interpret because non-IRD accumulation rates may have varied. Cowan sees cyclic IRD abundance, with peaks in the later glacial and succeeding interglacial parts of cycles (based on the shipboard lithologic definition of glacial cyclicity), and particularly prominent peaks at 0.88, 1.9, and 2.8 Ma. She identifies those cycles with orbital cyclicity from ~1.9 Ma on. Hassler and Cowan (**Chap. 11,** this volume) investigate IRD (pebble) shape and provenance over the same period, using samples from all drift sites in order to increase total sample number. They suggest a change from mainly englacial and supraglacial IRD transport to ice-base transport at ~0.76 Ma. So late a change in implied glacial state is at odds with most other data and interpretations from the leg.

Hillenbrand and Fütterer (**Chap. 23,** this volume) have examined biogenic opal (mainly from diatoms) in samples from all of the rise drift sites as a measure of paleoproductivity and, by implication, of ocean temperature within the photic zone. They infer a moderately warm late Miocene, a warm early Pliocene, and a cooling between ~3.1 and 1.8 Ma (late Pliocene) to present-day conditions (Fig. F8) and see a sedimentation rate influence on biogenic silica preservation. They assume that sea-ice cover has been a controlling factor throughout, in contrast to Pudsey (**Chap. 25,** this volume), who concludes (from a less precise but similar diatom count at Site 1095) that sea ice was unimportant before the Pleistocene because diatoms occurred in both glacial and interglacial facies at the site. Essentially, both studies show the same

F7. Smectite, illite, and chlorite in Site 1095 samples, p. 38.



F8. Variation of biogenic opal at Site 1095, p. 39.



evolution of shallow ocean temperature at this margin, which is largely compatible also with studies of Southern Ocean climate change (e.g., Abelmann et al., 1990; Kennett and Barker, 1990; Hodell and Venz, 1992; Bohaty and Harwood, 1998) and generally corresponds to global climate change. However, the question of sea-ice cover is one aspect of a concern crucial to Antarctic paleoclimate; in examining features of drift deposition, it is necessary to distinguish between an oceanic and an ice sheet response to orbital forcing. This is discussed further below.

The carbonate preserved at Sites 1096 and 1101 between 2.1 and 0.8 Ma was not interpreted on board ship as an indication that the Polar Front had migrated south of the sites, or disappeared, but its modern prominence in sediments north of this feature was noted. Winter and Wise (**Chap. 26,** this volume) caution against inferences of Polar Front migration or of specific paleotemperatures from nannofossil preservation. Oxygen and carbon isotopic ratios have been measured on *Neogloboquadrina pachyderma* (sinistral), by far the most common foraminifer (Barker et al., **Chap. 20,** this volume), demonstrating the potential of the method in this environment (and, incidentally, showing little or no evidence of elevated shallow-water temperatures). Although most abundant during interglacials, fresh foraminifers (almost always monospecific) are present in sufficient numbers throughout several glacial cycles to allow a single cycle to be examined in detail. For the period between 0.7 and 2.1 Ma, Guyodo et al. (2001) show a convincing correlation of magnetic susceptibility at Site 1101 and the oxygen isotopic record of orbitally induced climate change from ODP Site 677, supporting the view that the interhemispheric responses to orbital insolation change were at that time approximately in phase (a strong correlation between magnetic susceptibility, chromaticity parameter b*, and carbonate content over that period at Site 1101 had been identified on board ship; see fig. F8 of Shipboard Scientific Party, 1999d).

The unusual porosity-depth characteristics at the rise drift sites had been noted on board ship (Barker, Camerlenghi, Acton, et al., 1999), although physical properties data from different sources did not always agree. Volpi et al. (**Chap. 17,** this volume) have reanalyzed the available porosity, bulk density, and seismic velocity information. They identify a combination of measurements on individual samples, vertical seismic profiles, and acoustic tomography data as the optimal data set for use in calibrating the seismic reflection data set from the continental rise and confirm in passing the anomalous (underconsolidated) nature of the drift sediments.

The possibility that the Eltanin asteroid impact (Kyte et al., 1981, 1988), which occurred at ~2.15 Ma (Pliocene) some 1300 km away in the southeast Pacific, affected sedimentation on the rise drifts (Barker and Camerlenghi, 1999) has been pursued by a search for high Iridium concentrations in selected samples from Holes 1096B and 1096C (Kyte, **Chap. 9,** this volume). No strong evidence has been found, but the magnetobiostratigraphic synthesis (Iwai et al., **Chap. 36,** this volume) verifies that the sampled intervals at Site 1096 were correctly identified.

DISCUSSION

The suite of ANTOSTRAT proposals for drilling the Antarctic margin adopted the dual strategy of sampling both the ice-transported diamicts of the continental shelf and slope for a direct but low-resolution record of ice sheet evolution and the derived fine-grained sediments of the continental rise drifts for an indirect but much more detailed record, combining the results from several areas by means of glaciological models of ice sheet development. The major contributions of Leg 178 were to test that strategy and to examine Antarctic Peninsula glacial history. Thus, a discussion of Leg 178 results must consider several topics: lessons for margin drilling, sedimentary processes, and Antarctic Peninsula glacial history. These topics are intertwined, in the sense that the experience of drilling at Leg 178 sites exposed some of the practical difficulties of drawing simple conclusions about paleoclimate, which differ depending upon the depositional environment, the information sought, and the focus of postcruise activity. Discussion is best organized first in terms of depositional environment, then as a consideration of glacial history.

The Shelf—The Problem of Core Recovery

The experience of Leg 178 on the continental shelf was daunting: recovery was very poor at shallow depth and in young sediments and improved only marginally with depth. Downhole logging was only partial, and data interpretation was impeded by poor hole conditions. A magnetostratigraphy was not possible because of the coarse-grained, multidomain nature of the main remanence carriers. Against these difficulties was the recognition that diatoms were incorporated into the tills to an extent sufficient to permit a crude stratigraphic determination, even with poor recovery. The stratigraphy was improved by additional postcruise studies. Most of these difficulties were anticipated and it was known that, even had recovery been perfect, the record of topset sediment accumulation on the outer continental shelf was most probably partial, involving only short periods of rapid deposition and being vulnerable to reerosion. It was clear that only low-resolution questions could be answered by drilling on the shelf.

The low recovery at shallow depth could be explained in terms of the inability of the soft diamict matrix to hold clasts in place during drilling. As the matrix became more consolidated recovery improved, but not to the extent we had hoped. The most likely explanation for low recovery of consolidated topsets on the shelf was a persistent ocean swell, which at times became sufficient to prevent drilling altogether. It was possible to reoccupy a hole, having pulled out for vessel heave or ice, but time was lost and repeated passage of the bottom-hole assembly (BHA) degraded the hole, contributing to premature abandonment and to logging difficulties. Recovery was not improved by varying rotation rate, water pressure, or length cored. The basic RCB drilling tool was used at all shelf sites, for fear of damage to the more delicate APC/XCB tools in these lithologies. The technical limitations, combined with the effects of ocean swell, made fine control of weight-on-bit impossible, which undoubtedly affected recovery. A constant and controllable weight-on-bit was probably a key factor in the high recovery in similar lithologies achieved by the Cape Roberts Project, which drilled in an area remote from ocean swell on fast sea ice using a converted land drill rig with riser (Cape Roberts Science Team, 1998, 1999, 2000). Recovery remote from land might be improved by selecting a time of year when less ocean swell is generated, by changes in drill bit technology, and possibly by active heave compensation. The quality and completeness of log data might be improved by logging while drilling (as undertaken at Site 1167 in Prydz Bay; O'Brien, Cooper, Richter, et al., 2001). It is certainly the case that only low-resolution questions should be asked of shelf drilling. However, answers to such questions could be valuable contributions towards an understanding of Antarctic glacial history.

One additional effect should be noted. ODP Leg 119 in Prydz Bay (Barron, Larsen, et al., 1989), and perhaps common sense, seemed to suggest that till topsets should be more consolidated and indurated than foresets at equivalent burial depth because the topsets would have experienced ice load (even if only an intermittent, tidally driven load by an almost-floating low-profile ice stream). In fact, the tomographic velocity measurements reported here (Tinivella et al., Chap. 16, this volume) suggest the opposite, in that foresets have higher velocities, even at shallow depth. This is potential reassurance for plans to drill foresets where the equivalent topsets are missing, as at the Wilkes Land margin (Fig. F1).

The Rise—Orbital Insolation, Ice Sheet Evolution, and Sedimentary Processes

Postcruise work on the rise drift cores has tended to focus on the glacial cycle—not entirely, but the longer-term studies benefit from a better understanding of the shorter-term processes and variations, which we consider here. Before discussing the problems, it should be mentioned that at the rise sites recovery was very good, logging was not perfect but was extensive and complementary to drilling, and the magnetostratigraphy benefitted from the terrigenous, fine grain–sized remanence carriers, continuous deposition, and high paleolatitude (we had also anticipated higher microfossil abundance and better preservation, but that was perhaps unrealistic). The rarity of calcareous microfossils ruled out the determination of a detailed oxygen isotopic stratigraphy, but that was to be expected so far south of the Polar Front. Indeed, the preservation of (albeit limited) calcareous assemblages at the shallower sites was a surprise. In addition, it is clear that many studies of the properties of cored sediments from the rise drifts are at a preliminary stage, so additional insights may emerge.

The failure of spectral analysis of downhole log and core properties at the continental rise drift sites to show the clear dominance of orbital frequencies (Lauer-Leredde et al., Chap. 32, and Pudsey, Chap. 25, both this volume) has at present a wide range of possible explanations. One class of possibility is that the very uneven rates of sedimentation through the glacial cycle (glacial rates appear much higher than interglacial rates at the more proximal sites on the continental rise) significantly degraded or entirely destroyed evidence of the ice sheet's true orbital sensitivity. Another is that the shipboard stratigraphy, used in both studies published here, was sufficiently in error to mask an orbital sensitivity. A third is that for long periods of ice sheet history, ice sheet volume changes were autocyclic rather than orbitally driven. Enmeshed with these are other considerations: that the sedimentary processes or the ice sheet response may have changed with time, that the behavior of the Antarctic Peninsula ice sheet may have been different from that of the Antarctic ice sheet in general, or that some primary parameters may have been modified by diagenesis.

In considering these possibilities, it is important to distinguish between those properties of drift sedimentation that record ocean circulation and those that reflect the behavior of the ice sheet. It seems inherently likely that changes in ocean circulation, which extends simply and directly as far north as the Polar Front and through mixing processes very much farther, should be driven by orbital insolation changes, even if the ice sheet is not. Ocean temperature and (though perhaps not entirely) sea-ice cover (and thus biogenic production and bioturbation) should reflect this oceanic influence. A detailed oxygen isotopic study of the (sparsely) calcareous Pliocene–Pleistocene rise drift sediments (by using the ice volume effect) could examine in detail the time relation between Northern Hemisphere and Antarctic Peninsula cyclicities.

Autocyclic ice sheet behavior is not unreasonable. At present, the large variation in sea level caused by orbitally induced changes in Northern Hemisphere ice sheet volume through the glacial cycle is a major (perhaps the only significant) cause of Antarctic (and Antarctic Peninsula) ice volume change, involving ice sheet grounding line migration to the continental shelf edge during glacial maxima. If, as Huybrechts' (1992, 1993) glaciological model suggests, the volume of the present Antarctic ice sheet is not sensitive to temperature, and was so also for a few million years before the late Pliocene, when the Northern Hemisphere ice sheets were very small or nonexistent during glacial maxima, it is difficult to see how a strong orbital sensitivity could then have existed (this case is argued in more detail by Barker et al., 1999). In such a circumstance, the small Antarctic Peninsula ice sheet would be sensitive to sea level change generated by changes in the volume of the main Antarctic ice sheet, but such changes would not necessarily be orbitally driven. However, the Antarctic Peninsula ice sheet might also be sensitive to local changes in ocean and atmospheric temperature that *could* be orbitally driven. A possible outcome of this situation would be the presence of orbital frequencies (via the oceanic effect) without their dominance.

If the only cause (of the absence of dominant orbital frequencies in a spectral analysis) was autocyclic ice sheet behavior, then the analysis of late Pliocene and Pleistocene sedimentation (when Northern Hemisphere ice sheets *were* present as drivers of sea level change) should reveal an orbital sensitivity. Neither spectral study (Lauer-Leredde et al., **Chap. 32,** and Pudsey, **Chap. 25,** both this volume) showed this, despite clear nearby evidence to the contrary. Such evidence includes identification of the youngest glacial cycles in the sedimentary record (for example, Fig. **F6**), Cowan's recognition of orbital cyclicity after 1.9 Ma at Site 1101, and Guyodo et al.'s (2001) correlation between magnetic susceptibility and orbital cyclicity at the same site (which was not examined within either spectral study). The absence of clear evidence of dominant orbital cyclicity in late Pliocene and Pleistocene sedimentation on Drift 7 in the spectral analyses suggests that either the sedimentation rates used in these studies are incorrect or that diagenetic or sedimentary processes are distorting the record. Among such sedimentary processes might be (as generally suspected) a gross disparity in glacial and interglacial sedimentation rates (close to the continental slope) together with varying fractions of each cycle for which the grounding line was at the continental shelf edge, changes with ice temperature of the rate of terrigenous sediment supply to the outer slope, or processes such as channel switching, which are common in low-latitude terrigenous sediment transport systems. The existence of such sedimentary processes does not rule out autocyclic ice sheet behavior, but does make it more difficult to detect.

The simple, ideal circumstance is that both the Antarctic and the Antarctic Peninsula ice sheets have remained sensitive to orbital insolation. Such sensitivity could then be demonstrated by a repeated analysis, with a sufficiently improved stratigraphy or by seeking significant ratios of "orbital" frequencies (e.g., Fischer and Roberts, 1991), that would be independent of sedimentation rate. The same study would also test the third hypothesis, that of essentially autocyclic ice sheet behavior.

Antarctic Peninsula Glacial History

The direct evidence of Leg 178 drilling bears on the last 10 m.y. of Antarctic Peninsula glacial history. On the continental shelf, seismic sequence group S3 was penetrated at Sites 1097 and 1103 to 7–8 Ma. On the rise, Drift 7 was penetrated at Site 1095 to ~10 Ma. Both environments show glacial influence to the base of each hole; the onset of glaciation lies deeper in the record. Here, therefore, to provide an up-to-date context for Leg 178 results, we also examine evidence bearing on earlier Antarctic Peninsula glacial history and discuss how it relates to the glacial history of Antarctica as a whole.

It has long been assumed (e.g., Kennett and Barker, 1990; Wise et al., 1992; Barker and Camerlenghi, 1999) that changes in the glacial states of different parts of Antarctica, driven by the same global climate changes, have been essentially in phase but have started from different initial states. For example, Antarctic Peninsula glaciation would always have been less severe than that of the East Antarctic interior, and may have begun later, because of its more northerly location. It may be time for this simplistic view to be reassessed.

Modern Antarctic Peninsula and general Antarctic glacial history are coupled by two powerful factors: the dominantly external drive of sea level change by orbitally induced changes in Northern Hemisphere grounded ice volume and the general circumpolar equalization of ocean temperature and sea-ice cover by the Antarctic Circumpolar Current (ACC). There is some discussion about the time of onset of the ACC (see, for example, Lawver and Gahagan, 1998; Barker, 2001). All estimates make it considerably older than the time extent of recovered Leg 178 sediments (the past 9–10 m.y.), but it may have been younger than the onset of substantial Antarctic glaciation, now generally dated at ~34 Ma (e.g., Wise et al., 1992; Barrett, 1996; Barker et al., 1999; O'Brien, Cooper, Richter, et al., 2001; Zachos et al., 2001). Maximum Northern Hemisphere ice volumes during glacials appear to have grown since ~3 Ma and to have been very large since ~0.8 Ma; Northern Hemisphere ice volume changes seem to have been driven by orbitally induced (Milankovich) insolation changes (e.g., Hays et al., 1976; Ruddiman et al., 1989; Tiedemann et al., 1994). As already discussed, without such external sea level drive, the volume of the earliest Pliocene and late Miocene Antarctic ice sheet may not have been sensitive to orbital cyclicity. Without the thermal equalization effect of the ACC, the climates of different parts of the Antarctic margin may have been more different from each other than they are today.

The suite of ANTOSTRAT drilling proposals used the model of ice sheet evolution developed by Huybrechts (1992, 1993), which examined the effect on ice sheet volume of uniformly increasing the present-day mean annual temperature at sea level around the continent (see also Barker et al., 1999). This model predicts a late-stage development of an Antarctic Peninsula ice sheet, appropriate to its more northerly situation. However, at present the Antarctic Peninsula experiences much greater than the average Antarctic precipitation, perhaps because its 2000-m elevation disrupts tropospheric circulation (Drewry and Morris, 1992). Also, East Antarctic cooling is conveyed to it within the clockwise Weddell Gyre (Reynolds, 1981). It may therefore, while these conditions held, have been more easily glaciated than its latitude would suggest.

There are several suggestions in the onshore regional geology that this is so. More recent studies (Dingle et al., 1997; Dingle and Lavelle, 1998; Troedson and Smellie, 2002; Troedson and Riding, 2002) have greatly refined the interpretation of early glaciations recorded onshore in the South Shetland Islands, off the northwestern Antarctic Peninsula (e.g., Birkenmajer, 1991), but evidence of middle to late Oligocene (26–30 Ma) and earliest Miocene (22–23 Ma) glaciations is confirmed on King George Island. In each case a single glacial episode is proposed, involving subglacial to distal glacial marine deposition, and an intervening late Oligocene (26–24 Ma) interglacial is inferred from associated marine sequences. The presence in these sediments of clasts of rocks outcropping mainly in the Transantarctic and Ellsworth Mountains, beyond the southernmost Weddell Sea (Fig. **F1**), is interpreted in terms of rafting on or within icebergs originating in a glaciated East Antarctica and circulating within an ancestral Weddell Gyre (see also Larter et al., 1997). The South Shetland Islands onshore sediments represent a shallow continental shelf environment and were subaerially exposed recently by rift-shoulder uplift associated with Bransfield Strait opening. Although it is difficult to compare onshore and offshore records, these sediments reveal a level of local glaciation similar in many ways to that of sequence group S3, in which glacial ice is grounded below sea level. Whether there really were very long glacials and interglacials at that time, or they were much shorter and more numerous but with sediments mostly not preserved, is uncertain. Dingle and Lavelle (1998) suggest an age of 30 Ma for glacial onset in the South Shetland Islands region, based on Sr and Ar isotopic ages for the oldest known glacial deposits exposed onshore. Troedson and Riding (2002) speculatively correlate the second glaciation with the isolated high oxygen isotopic event Mi-1 of the earliest Miocene (Miller et al., 1991; Paul et al., 2000). A third glaciation evident in the onshore South Shetland Islands record, the Legru glaciation (Birkenmajer, 1991), is of uncertain age but is considered younger than the other two (J. Smellie, pers. comm., 2001); otherwise, onshore evidence of paleoclimate is missing until the late Miocene. The only other information relevant to the onset of Antarctic Peninsula glaciation comes from onshore exposures of nonglacial sediment of late Eocene age on Seymour Island on the Weddell Sea margin of the northern Antarctic Peninsula (e.g., Francis, 1991; Dingle and Lavelle, 1998) and a temperate setting for shallow marine sediment of Eocene or early Oligocene age sampled during ODP Leg 113 at Site 696 on the South Orkney microcontinent farther east (e.g., Mohr, 1990). Thus, it is possible that East Antarctica became glaciated earlier than the Antarctic Peninsula, but not certain.

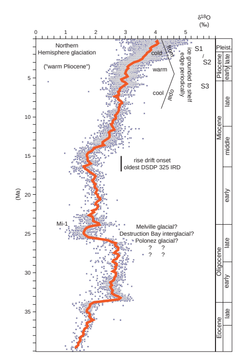
The offshore evidence of glacial onset comes mainly from the distribution of IRD at sites drilled during DSDP Leg 35 (Hollister, Craddock, et al., 1976). Tucholke et al. (1976) noted the earliest evidence of ice rafting in the late early to middle Miocene (15–17 Ma) at DSDP Site 325. The clast petrology accords with a subduction-related origin, thus at a Pacific margin, probably the Antarctic Peninsula. Leaving aside the possibility of downhole contamination by dropstones, this is the earliest offshore evidence of glaciation of any kind for the Antarctic Peninsula. Less directly, working on the seismic reflection character of the continental rise sediment drifts, Rebesco et al. (1997) concluded that the base of their seismic Unit M4, the onset of drift formation that they associated with the onset of glaciation, had an age of ~15 Ma. This and other interpreted ages were based on stratigraphy at DSDP Site 325 and on magnetic anomaly ages along the margin. Results of subsequent Leg 178 drilling on the drifts, while not reaching the base of Unit M4, coincided closely with the estimated ages of overlying units; the base of their Unit M2, at ~650 m close to Site 1096, was provisionally dated at 5 Ma, whereas Hole 1096C reached ~4.7 Ma at 608 mbsf. Similarly, Hole 1095B, which reached ~10 Ma at 570 mbsf, was considered to have just reached the Unit M4/M3 boundary, dated by Rebesco et al. (1997) at 8 Ma. Given these coincidences, a time of glacial onset of ~15 Ma does not seem unreasonable.

There is thus a large discrepancy between the time of onset of Antarctic Peninsula glaciation (defined, arbitrarily, as the onset of sequence group S3-type conditions in which ice is grounded below sea level) as postulated onshore and offshore. One possible explanation for this is that there was an earlier onset throughout the Antarctic Peninsula, at some time around 30 Ma perhaps, and then a return to warmer (nonglacial) conditions immediately after event Mi-1 (in the early Miocene, perhaps at 22 Ma). We need to bear in mind the widespread history of subduction and ridge-crest collision along the Antarctic Peninsula margin between 30 and 10 Ma. Glacial onset might be particularly difficult to identify at an actively subducting margin; oceanic sediments are subducted or deformed within the accretionary prism, and the fate of continental shelf sediments is highly uncertain. The offshore evidence for mid-Miocene glacial onset and glacial conditions (DSDP Site 325, most of the sediment drifts, and the outer continental shelf drilled during ODP Leg 178) comes from parts of the margin that have undergone subduction, culminating in ridge-crest collision, since 30 Ma. This notion of a separate, earlier glacial onset and then a later one is compatible with a simple correspondence to the modern benthic oxygen isotopic record (e.g., Zachos et al., 2001) (Fig. **F9**), which suggests a warmer intervening early Miocene and/or a smaller ice sheet. Offshore evidence for the earlier onset of Antarctic Peninsula glaciation may possibly be seen in sedimentary geometries on seismic reflection profiles crossing the continental margin farther south (Nitsche, 1997, 2000) where ridge crest subduction was earlier, although in those papers the authors identify only sequence groups that correspond closely to those seen (e.g., Fig. **F3**) farther north.

In passing, we should note that the onset of glaciation at a particular margin is one of the targets of the ANTOSTRAT proposal suite, in that its unifying glaciological model (Huybrechts, 1992, 1993; Barker et al., 1999) predicted how large an ice sheet would have to be in order to reach a particular part of the Antarctic margin. Leg 178 did not sample glacial onset, but the unusually high precipitation and high elevation of the Antarctic Peninsula make such information less generally applicable and the small size of the Antarctic Peninsula ice sheet, compared with that of East Antarctica, makes it unimportant in global terms. This target *was* reached by drilling during ODP Leg 188 at Prydz Bay, for example (O'Brien, Cooper, Richter, et al., 2001), where it is much more important.

We come now to the younger part of Antarctic ice sheet evolution, where the evidence from Leg 178 drilling is direct. The record of biogenic opal on the continental rise (Hillenbrand and Fütterer, **Chap. 23**, this volume) (Fig. **F8**) shows a moderately cool late Miocene, a warm early Pliocene, and cooling through the late Pliocene toward Pleistocene conditions similar to today's. Throughout this period, IRD

F9. Benthic oxygen isotopic data, p. 40.



(Cowan, **Chap. 10;** Hassler and Cowan, **Chap. 11;** Pudsey, **Chap. 25,** all this volume), clay mineralogy (Hillenbrand and Ehrmann, **Chap. 8,** this volume) (Fig. **F7**), and similar studies on the continental rise show the persistence of an Antarctic Peninsula ice sheet over the past 10 m.y., even through the warm early Pliocene, sufficiently large for the grounding line to reach regularly to the shelf edge. Neither the drilling results nor the disposition of glacial sediments on the shelf and slope revealed by seismic reflection profiles shows any sign of prolonged deglaciation during this period. Both on the shelf (where the progradational sequence group S2 was deposited) and on the continental rise drifts, terrigenous sedimentation was greater or more rapid (see the drift site sedimentation rate diagrams of Iwai et al., **Chap. 36,** this volume; Fig. **F5**) during the early Pliocene than during the colder late Pliocene and Pleistocene. This supports the observations and interpretation of Barker (1995) that terrigenous deposition is faster because the ice sheet budget is greater; warmer air at the ice margin, holding more water, leads to greater precipitation, which is balanced by the faster flow of warmer, weaker ice. Additional contributary factors might be a reduced area of basal melting (of colder ice) in colder times, and a greater sea-ice extent, making evaporation from the open sea more distant and reducing precipitation on the ice sheet. Such a temperature dependence of terrigenous deposition suggests further that if the Antarctic continent becomes even colder in the future, the margin would become more sediment starved (Barker et al., in press). It supports the suggestion (Barker et al., 1999) that during the late Miocene and Pliocene–Pleistocene, the volume of the Antarctic ice sheet remained essentially insensitive to temperature change.

Onshore evidence of late Miocene to latest Pleistocene Antarctic Peninsula paleoclimate is sparse and widely scattered in occurrence and discontinuous in section, comprising estimates of ice thickness above volcanic effusions and glacial sediments most probably preserved by overlying volcanics (Pirrie et al., 1997; Smellie, 1999) and is essentially compatible with the evidence from drilling.

The decoupling between climate and ice sheet behavior demonstrated by results from the continental rise casts doubt on the shipboard interpretation of the upper Miocene sequence group S3 sampled on the shelf as indicating a "less glacial" environment than the overlying, probably lower Pliocene progradational sequence group S2. The late Miocene interglacials, from the opal evidence (Hillenbrand and Fütterer, **Chap. 23,** this volume), were cooler than those of the succeeding early Pliocene. Certainly, "less-glacial" depositional environments were among those recorded by sediments recovered from sequence group S3 compared with sequence group S2, but recovery of both sequence groups was extremely low and the wider range of sequence group S3 sediments recovered does not extend outside the range of sediments produced within a modern glacial cycle. A change in the range of recorded environments and the striking difference in the geometries of sequence groups S3 and S2 shown in seismic reflection profiles could also have been affected by nonclimatic factors, such as changes in subglacial topography and accommodation space or sediment facies (source lithology and grain-size distribution), as glacial erosion of the Antarctic Peninsula proceeded, or by a change in sediment consolidation resulting solely from nonhydrostatic load. The longer-term decoupling of climate and ice sheet volume that we observe also supports the possibility of decoupling at the higher frequencies of the orbitally induced (Milankovich) insolation variation or equivalent autocyclic variation, as discussed above, which may have affected what was preserved by changing the relative lengths of periods of glacial and interglacial deposition.

The Antarctic Peninsula climate history summarized and hypothesized above is compared in Figure F9 with the oxygen isotopic data collated and displayed by Zachos et al. (2001). The initial correspondence is good; the early (Oligocene and earliest Miocene) glacial episodes recorded onshore are reflected in general terms in the isotopic variation, whether this variation is a result of changes in temperature or ice sheet volume (for a possible resolution of this ambiguity, see also Lear et al., 2000). This suggests a strong coupling of Antarctic Peninsula glaciation to global climate and probably therefore (e.g., Barker et al., 1999) a sensitivity to orbital insolation changes also through this period. Subsequently, renewed middle Miocene cooling, as shown by IRD onset and drift formation, matches the increase in isotopic ratio evident in Figure F9. However, it is difficult to detect in Figure F9 any isotopic evidence for a "warm" early Pliocene, and the volume of the Antarctic Peninsula ice sheet seems to have remained essentially independent of global climate change over the past 10 m.y. This comparison suggests that the behavior of the entire Antarctic ice sheet may at times have departed from a straightforward volume dependence on global climate change and shows the value and significance of information about the intervening early Miocene Antarctic Peninsula paleoclimate, which may have been warmer, possibly nonglacial. The inner shelf basins of the Pacific margin are a possible source of middle and lower Miocene sediments.

CONCLUSIONS

ODP Leg 178 had two principal objectives. The first was to test the viability of drilling the glacial prograded wedge on the continental shelf and the derived sediment drifts on the upper continental rise to obtain a record of Antarctic glaciation. The second was to examine Antarctic Peninsula glacial history over the past 8–10 m.y., a period that included both the "warm Pliocene," proposed by some as having seen substantial deglaciation, and the growth and eventual dominance of the Northern Hemisphere ice sheets. These objectives were its particular contributions to an ANTOSTRAT plan to determine a complete Antarctic glacial history by examining similar sediments in several sectors of the Antarctic margin and combining the results by means of numerical models of glaciation.

The drilling succeeded on both counts, demonstrating the validity of the strategy (while determining its limitations) and obtaining a high-resolution record of Antarctic Peninsula glacial history over the past 10 m.y. In more detail, and mainly as a result of drilling and postcruise work on samples and data, we can conclude the following:

1. Recovery in unconsolidated glacial sediments on the continental shelf was very poor, probably because indurated clasts could not be supported by an unconsolidated fine-grained matrix during rotary drilling. Recovery improved at greater depth but was still poor, probably because ocean swell (perhaps incompletely compensated at times) and the necessary use of crude but robust drilling tools prevented fine control of weight-on-bit. Improved recovery might result from drilling tool developments, better

heave compensation, and drilling during periods of reduced ocean swell, with compensatory use of logging while drilling. Despite the limitations, shelf sediments can provide a useful record of glaciation; a key lesson of Leg 178 is to ensure that only low-resolution questions are asked of ODP-style drilling on the continental shelf.

2. Leg 178 shelf drilling recovered Pliocene–Pleistocene diamicts, reflecting a subglacial environment (compatible with regular grounding line advance to the continental shelf edge) and corresponding to topset beds of seismic reflection sequence groups S1 and S2.
3. Shelf seismic sequence group S3 is also glacial, but a wider range of environments (subglacial, proglacial, and glaciomarine) is represented in recovered sediments. Recovered sequence group S3 (its lower part was not sampled) is of earliest Pliocene and late Miocene age (back to ~7.5 Ma) and is considered to represent paleoshelf (topset) deposition. A decompaction model, supported by the abundances of benthic diatoms on both shelf and rise, suggests that the sequence group S3–age continental shelf was not overdeepened as it is today.
4. Drilling on the continental rise sediment drifts demonstrated clearly that they hold a high-resolution, recoverable record of continental glaciation. During Leg 178 only their upper part (back to ~10 Ma) was sampled, but with excellent recovery. Sediments are mainly fine-grained alternations of interglacial bioturbated biosiliceous silts and clays and glacial laminated, largely barren terrigenous silts and clays. IRD is present throughout.
5. IRD at the rise sites shows that the ice sheet was present throughout the past 10 m.y., and variations in clay mineralogy indicate that the ice sheet grounding line migrated regularly to the continental shelf edge through that period, including the "warm" early Pliocene and the time of deposition of the sampled part of sequence group S3. Biogenic opal variation suggests a cool late Miocene, a warm early Pliocene, and late Pliocene cooling to a cold (as at the present day) Pleistocene, changes that coincide with global climate changes. Thus, the Antarctic Peninsula ice sheet persisted through the "warm" Pliocene, its volume decoupled from longer-term global climate change and sufficiently great for regular grounding line migration to the shelf edge. The possibility of decoupling from global climate of volume changes of the entire Antarctic ice sheet through this period must also be considered.
6. The opal evidence of correspondence between global climate and southwest Pacific oceanic (Antarctic Peninsula offshore) climate casts doubt on the shipboard interpretation of the upper Miocene shelf sequence group S3 as reflecting "less glacial" conditions than the overlying Pliocene sequence group S2. The change in range of recovered facies might have resulted from improved recovery of more consolidated sediments or from changes in conditions of deposition and preservation less simply related to climate.
7. Regular pre-late Pliocene grounding line advance to the Antarctic Peninsula continental shelf edge shows a sensitivity to sea level change that most probably was caused by volume changes of the main Antarctic ice sheet, since there were then no large

Northern Hemisphere ice sheets. Fundamentally, the Antarctic Peninsula ice sheet appears to have acted as a recorder rather than a driver of change, both after and before the development of Northern Hemisphere ice sheets.

8. Spectral analysis of several properties of rise drift sediments (at Sites 1095 and 1096) failed to show the dominance of periodicities corresponding to those of orbital insolation change. The cause of this lack of correspondence is unknown; it could be nonlinear deposition, the inadequacy of the available stratigraphic control, truly decoupled (autocyclic) ice sheet behavior, or a combination of these. The only evidence of sensitivity to orbital insolation change came from IRD, carbonate, and magnetic susceptibility variation at Site 1101 over the past 1.9 m.y. and from correlation between drift sites over the last 200–300 k.y., periods when Northern Hemisphere ice sheet variation drove sea level change. Further investigation of orbital sensitivity is essential, particularly for older periods, and should attempt to distinguish between "oceanic" and "ice sheet" influences, which may have behaved differently.
9. Sedimentation rates on the rise were greater during the warm early Pliocene than during the subsequent, colder late Pliocene and Pleistocene, supporting views that a warmer glacial regime has a greater ice and sediment throughput and suggesting the possibility of sediment starvation around Antarctica should cooling persist.
10. Glacial onset (assumed to be defined as glacial ice grounded significantly below sea level) was earlier than the drilled section. Based on a combination of onshore and offshore data, we suggest an Antarctic Peninsula climate history that included Oligocene (and brief earliest Miocene) glaciation, then was warmer (perhaps nonglacial) until the middle Miocene, then involved renewed glaciation, which has persisted and strengthened to the present day. Such a history would be broadly compatible with global oxygen isotopic ratios measured on benthic foraminifers. An earlier onset than predicted by the glaciological model may have resulted from higher-than-average regional precipitation and from ocean circulation within the Weddell Sea.

In addition, it seems clear that the potential of the rise drift drilling is not yet exhausted. In particular, a wealth of sediment properties measurements is now available to assist further analysis. Some are likely to reflect oceanic influence, others those of the ice sheet. Magnetobiostratigraphic constraints on drift sediment ages have improved and, with possible use of techniques of spectral analysis independent of sedimentation rate, may allow an understanding of the detailed behavior of the Antarctic Peninsula and Antarctic ice sheets over the past 10 m.y. or so.

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REFERENCES

Abelmann, A., Gersonde, R., and Spiess, V., 1990. Pliocene–Pleistocene paleoceanography in the Weddell Sea: siliceous microfossil evidence. *In* Bleil, U., and Thiede, J. (Eds.), *Geologic History of the Polar Oceans: Arctic versus Antarctic:* NATO ASI Ser., Ser. C, 308:729–759.

Alley, R.B., Blankenship, D.D., Rooney, S.T., and Bentley, C.R., 1989. Sedimentation beneath ice shelves: the view from Ice Stream B. *Mar. Geol.,* 85:101–120.

Barber, P.L., Barker, P.F., and Pankhurst, R.J., 1991. Dredged rocks from Powell Basin and the South Orkney microcontinent. *In* Thomson, M.R.A., Crame, J.A., and Thomson, J.W. (Eds.), *Geological Evolution of Antarctica*. Cambridge (Cambridge University Press), 361–367.

Barker, D.H.N., and Austin, J.A., Jr., 1998. Rift propagation, detachment faulting, and associated magmatism in Bransfield Strait, Antarctic Peninsula. *J. Geophys. Res.,* 103:24017–20243.

Barker, P.F., 1982. The Cenozoic subduction history of the Pacific margin of the Antarctic Peninsula: ridge crest-trench interactions. *J. Geol. Soc. London,* 139:787–801.

————, 1995. The proximal marine sediment record of Antarctic climate since the late Miocene. *In* Cooper, A.K., Barker, P.F., and Brancolini, G. (Eds.), *Geology and Seismic Stratigraphy of the Antarctic Margin*. Am. Geophys. Union, Antarct. Res. Ser., 68:25–57.

————, 2001. Scotia Sea regional tectonic evolution: implications for mantle flow and palaeocirculation. *Earth Sci. Rev.*, 55: 1–39.

Barker, P.F., Barrett, P.J., Camerlenghi, A., Cooper, A.K., Davey, F.J., Domack, E.W., Escutia, C., Kristoffersen, Y., and O'Brien, P.E., 1998. Ice sheet history from Antarctic continental margin sediments: the ANTOSTRAT approach. *Terra Antart.,* 5:737–760.

Barker, P.F., Barrett, P.J., Cooper, A.K., and Huybrechts, P., 1999. Antarctic glacial history from numerical models and continental margin sediments. *Palaeogeogr., Palaeoclimatol., Palaeoecol.,* 150:247–267.

Barker, P.F., and Camerlenghi, A., 1999. An approach to Antarctic glacial history: the aims of Leg 178. *In* Barker, P.F., Camerlenghi, A., Acton, G.D., et al., 1999. *Proc. ODP, Init. Repts.,* 178, 1–44 [CD-ROM]. Available from: Ocean Drilling Program, Texas A&M University, College Station, TX 77845-9547, U.S.A.

Barker, P.F., Camerlenghi, A., Acton, G.D., et al., 1999. *Proc. ODP, Init. Repts.,* 178 [CD-ROM]. Available from: Ocean Drilling Program, Texas A&M University, College Station, TX 77845-9547, U.S.A.

Barker, P.F., Camerlenghi, A., and Shipboard Scientific Party, in press. Antarctic glacial history, step 1: the continental margin drilled by ODP Leg 178. *Proc. 8th ISAES:* Wellington, New Zealand.

Barker, P.F., and Dalziel, I.W.D., 1983. Progress in geodynamics in the Scotia Arc region. *In* Cabre, R. (Ed.), *Geodynamics of the Eastern Pacific Region, Caribbean and Scotia Arcs*. Am. Geophys. Union Geodyn. Ser., 9:137–170.

Barker, P.F., Dalziel, I.W.D., and Storey, B.C., 1991. Tectonic development of the Scotia Arc region. *In* Tingey, R.J. (Ed.), *Geology of Antarctica:* Oxford (Oxford Univ. Press), 215–248.

Barrett, P.J., 1996. Antarctic paleoenvironment through Cenozoic times: a review. *Terra Antart.,* 3:103-119.

Barron, J., Larsen, B., et al., 1989. *Proc. ODP, Init. Repts.,* 119: College Station, TX (Ocean Drilling Program).

Bart, P.J., and Anderson, J.B., 1995. Seismic record of glacial events affecting the Pacific margin of the northwestern Antarctic Peninsula. *In* Cooper, A.K., Barker, P.F., and Brancolini, G. (Eds.), *Geology and Seismic Stratigraphy of the Antarctic Margin*. Am. Geophys. Union, Antarct. Res. Ser., 68:75–96.

Bartek, L.R., Vail, P.R., Anderson, J.B., Emmet, P.A., and Wu, S., 1991. Effect of Cenozoic ice sheet fluctuations in Antarctica on the stratigraphic signature of the Neogene. *J. Geophys. Res.,* 96:6753–6778.

Berggren, W.A., Kent, D.V., Swisher, C.C., III, and Aubry, M.-P., 1995. A revised Cenozoic geochronology and chronostratigraphy. *In* Berggren, W.A., Kent, D.V., Aubry, M.-P., and Hardenbol, J. (Eds.), *Geochronology, Time Scales and Global Stratigraphic Correlation*. Spec. Publ.—SEPM (Soc. Sediment. Geol.), 54:129–212.

Birkenmajer, K., 1991. Tertiary glaciation in the South Shetland Islands, West Antarctica: evaluation of data. *In* Thomson, M.R.A., Crame, A.J., and Thomson, J.W. (Eds.), *Geological Evolution of Antarctica*. Cambridge (Cambridge University Press), 629–632.

Bohaty, S.M., and Harwood, D.M., 1998. Southern Ocean Pliocene paleotemperature variation from high-resolution silicoflagellate biostratigraphy. *Mar. Micropal.,* 33:241–272.

Boulton, G.S., 1990. Sedimentary and sea level changes during glacial cycles and their control on glacimarine facies architecture. *In* Dowdeswell, J.A., and Scourse, J.D. (Eds.), *Glacimarine Environments: Processes and Sediments*. Spec. Publ.—Geol. Soc. London, 53:15–52.

Camerlenghi, A., Crise, A., Pudsey, C.J., Accerboni, E., Laterza, R., and Rebesco, M., 1997a. Ten-month observation of the bottom current regime across a sediment drift of the Pacific margin of the Antarctic Peninsula. *Antarct. Sci.,* 9:426–433.

Camerlenghi, A., Rebesco, M., De Santis, L., and Volpi, V., in press. The Antarctic Peninsula Pacific margin: modelling flexure and decompaction with constraints from ODP Leg 178 initial results. *In* Gamble, J., Skinner, D., and Henrys, S. (Eds.), *Proceedings, 8th ISAES*, Royal Society of New Zealand, Wellington.

Camerlenghi, A., Rebesco, M., and Pudsey, C.J., 1997b. High resolution terrigenous sedimentary record of a sediment drift on the Antarctic Peninsula Pacific margin. *In* Ricci, C.A. (Ed.), *The Antarctic Region: Geological Evolution and Processes*. Init. Results, SEDANO Progr., 705–710.

Canals, M., Urgeles, R., Estrada, F., and GEBRAP Team, 1998. Internal structure and seismic facies of the deep-water sediment drifts off northern Graham Land, Antarctic Peninsula: results from a very high-resolution survey. *Ann. Glaciol.,* 27:265–267.

Cape Roberts Science Team (CRST), 1998. Initial report on CRP-1. *Terra Antart.,* 5.

————, 1999. Studies from the Cape Roberts Project, Ross Sea, Antarctica—initial report on CRP-2/2A. *Terra Antart.,* 6(½).

————, 2000. Studies from the Cape Roberts Project, Ross Sea, Antarctica: initial report on CRP-3. *Terra Antart.,* 7.

Clausen, L., 1998. Late Neogene and Quaternary sedimentation on the continental slope and upper rise offshore southeast Greenland: interplay of contour and turbidity processes. *In* Saunders, A.D., Larsen, H.C., and Wise, S.W., Jr. (Eds.), *Proc. ODP, Sci. Results,* 152: College Station, TX (Ocean Drilling Program), 3–18.

Cochrane, G.R., Moore, J.C., and Lee, H.J., 1996. Sediment pore-fluid overpressuring and its effect on deformation at the toe of the Cascadia accretionary prism from seismic velocities. *In* Bebout, G.E., Scholl, D.W., Kirby, S.H., and Platt, J.P. (Eds.), *Subduction, Top to Bottom*, AGU Geophysical Monograph, 96:57–64.

Cooper, A.K., Barrett, P.J., Hinz, K., Traube, V., Leitchenkov, G., and Stagg, H.M.J., 1991. Cenozoic prograding sequences of the Antarctic continental margin: a record of glacio-eustatic and tectonic events. *Mar. Geol.,* 102:175–213.

Dalziel, I.W.D., 1984. Tectonic evolution of a fore-arc terrane, southern Scotia Ridge, Antarctica. *Sp. Pap.—Geol. Soc. Amer.*, 200:1–32.

Dingle, R.V., and Lavelle, M., 1998. Antarctic peninsular cryosphere: early Oligocene (c. 30 Ma) initiation and revised glacial chronology. *J. Geol. Soc.*, 155:433–7.

Dingle, R.V., McArthur, J.M., and Vroon, P., 1997. Oligocene and Pliocene interglacial events in the Antarctic Peninsula dated using strontium isotope stratigraphy. *J. Geol. Soc.,* 154:257–264.

Drewry, D.J., and Morris, E.M., 1992. The response of large ice sheets to climate change. *Philos. Trans. R. Soc. London B.,* 338:235–242.

Elliot, D.H., 1997. The planar crest of Graham Land, northern Antarctic Peninsula: possible origins and timing of uplift. *In* Barker, P.F., and Cooper, A.K. (Eds.), *Geology and Seismic Stratigraphy of the Antarctic Margin* (Pt 2). Am. Geophys. Union, Antarct. Res. Ser., 71:51–73.

Eyles, N., Daniels, J., Osterman, L.E., and Januszczak, N., 2001. Ocean Drilling Program Leg 178 (Antarctic Peninsula): sedimentology of glacially influenced continental margin topsets and foresets. *Mar. Geol.,* 178:135–156.

Fischer, A.G., and Roberts, L.T., 1991. Cyclicity in the Green River Formation (lacustrine Eocene) of Wyoming. *J. Sed. Petrol.*, 61:1146–1154.

Francis, J.E., 1991. Palaeoclimatic significance of Cretaceous–early Tertiary fossil forests of the Antarctic Peninsula. *In* Thomson, M.R.A., Crame, A.J., and Thomson, J.W. (Eds.), *Geological Evolution of Antarctica*. Cambridge (Cambridge University Press), 623–627.

Guyodo, Y., Acton, G.D., Brachfeld, S., and Channell, J.T., 2001. A sedimentary paleomagnetic record of the Matuyama chron from the Western Antarctic margin (ODP Site 1101). *Earth Planet. Sci. Lett.*, 191:61–74.

Hays, J.D., Imbrie, J., and Shackleton, N.J., 1976. Variations in the Earth's orbit: pacemaker of the ice ages. *Science,* 194:1121–1132.

Herron, E.M., and Tucholke, B.E., 1976. Sea-floor magnetic patterns and basement structure in the southeastern Pacific. *In* Hollister, C.D., Craddock, C., et al., *Init. Repts., DSDP,* 35: Washington (U.S. Govt. Printing Office), 263–278.

Hillenbrand, C.-D., 2000. Glaciomarine sedimentation on the continental margins of the Amundsen and Belllingshausen Seas, West Antarctica—indications for paleoenvironmental changes during the Quaternary climatic cycles. *Berichte zur Polarforschung,* 346:1–182.

Hodell, D.A., and Venz, K., 1992. Toward a high-resolution stable isotopic record of the Southern Ocean during the Pliocene-Pleistocene (4.8 to 0.8 Ma). *In* Kennett, J.P., Warnke, D.A. (Eds.), *The Antarctic Paleoenvironment: A Perspective on Global Change* (Pt. 1). Am. Geophys. Union, Antarct. Res. Ser., 56:265–310.

Hollister, C.D., Craddock, C., et al., 1976. *Init. Repts. DSDP,* 35: Washington (U.S. Govt. Printing Office).

Hole, M.J., and Larter, R.D., 1993. Trench-proximal volcanism following ridge crest-trench collision along the Antarctic Peninsula, *Tectonics*, 12:897–910.

Huybrechts, P., 1992. The Antarctic ice sheet and environmental change: a three-dimensional modelling study [Ph. D. thesis, Vrije Universiteit Brussel]. *Berichte zur Polarforschung*, 99:1–241.

Huybrechts, P., 1993. Glaciological modelling of the late Cenozoic East Antarctic Ice Sheet: stability or dynamism. *Geograf. Ann.,* 75A:221–238.

Kennett, J.P., and Barker, P.F., 1990. Latest Cretaceous to Cenozoic climate and oceanographic developments in the Weddell Sea, Antarctica: an ocean-drilling perspective. *In* Barker, P.F., Kennett, J.P., et al., *Proc. ODP, Sci. Results,* 113: College Station, TX (Ocean Drilling Program), 937–960.

Kim, Y., Kim, H.-S., Larter, R.D., Camerlenghi, A., Gamboa, L.A.P., and Rudowski, S., 1995. Tectonic deformation in the upper crust and sediments at the South Shetland Trench. *In* Cooper, A.K., Barker, P.F., and Brancolini, G. (Eds.), *Geology and Seismic Stratigraphy of the Antarctic Margin*, Am. Geophys. Union, Antarct. Res. Ser., 68:157–166.

Kuvaas, B., and Leitchenkov, G., 1992. Glaciomarine turbidite and current-controlled deposits in Prydz Bay, Antarctica. *Mar. Geol.,* 108:365–381.

Kyte, F.T., Zhou, Z., and Wasson, J.T., 1981. High noble metal concentrations in a late Pliocene sediment. *Nature,* 292:417–420.

————, 1988. New evidence on the size and possible effects of a late Pliocene oceanic asteroid impact. *Science,* 24:63–65.

Larter, R.D., and Barker, P.F., 1989. Seismic stratigraphy of the Antarctic Peninsula Pacific margin: a record of Pliocene–Pleistocene ice volume and paleoclimate. *Geology,* 17:731–734.

————, 1991a. Effects of ridge-crest trench interaction on Antarctic-Phoenix spreading: forces on a young subducting plate. *J. Geophys. Res.,* 96:19583–19607.

————, 1991b. Neogene interaction of tectonic and glacial processes at the Pacific margin of the Antarctic Peninsula. *In* Macdonald, D.I.M. (Ed.), *Sedimentation, Tectonics and Eustasy: Sea-level Changes at Active Margins.* Spec. Publ.—Int. Assoc. Sedimentol., 12:165–186.

Larter, R.D., and Cunningham, A.P., 1993. The depositional pattern and distribution of glacial–interglacial sequences on the Antarctic Peninsula Pacific Margin. *Mar. Geol.,* 109:203–219.

Larter, R.D., Rebesco, M., Vanneste, L.E., Gamboa, L.A.P., and Barker, P., 1997. Cenozoic tectonic, sedimentary and glacial history of the continental shelf west of Graham Land, Antarctic Peninsula. *In* Cooper, A.K., Barker, P.F., and Brancolini, G. (Eds.), *Geology and Seismic Stratigraphy of the Antarctic Margin* (Pt. 2). Am. Geophys. Union, Antarct. Res. Ser., 71:1–27.

Lawver, L.A., and Gahagan, L.M., 1998. Opening of Drake Passage and its impact on Cenozoic ocean circulation. *In* Crowley, T.J., and Burke, K.C. (Eds.), *Tectonic Boundary Conditions for Climate Reconstructions.* Oxford (Oxford Univ. Press), 212–223.

Lawver, L.A., Keller, R.A., Fisk, M.R., and Strelin, J.A., 1995. Bransfield Strait, Antarctic Peninsula: active extension behind a dead arc. *In* Taylor, B. (Ed.), *Backarc Basins, Tectonics and Magmatism.* New York (Plenum Press), 315–342.

Lear, C.H., Elderfield, H., and Wilson, P.A., 2000. Cenozoic deep-sea temperatures and global ice volumes from Mg/Ca in benthic foraminiferal calcite. *Science,* 287:269–272.

Leat, P.T., Scarrow, J.H., and Millar, I.L., 1995. On the Antarctic Peninsula batholith. *Geol. Mag.,* 13:399–412.

Lucchi, R.G., Rebesco, M., Busetti, M., Caburlotto, A., Colizza, E., and Fontolan, G., in press. Sedimentary processes and glacial cycles on the sediment drifts of the Antarctic Peninsula Pacific margin: preliminary results of SEDANO-II project. *In* Gamble, J., Skinner, D., and Henrys, S. (Eds.), *Proceedings, 8th ISAES,* Royal Society of New Zealand, Wellington.

Maldonado, A., Larter, R.D., and Aldaya, F., 1994. Forearc tectonic evolution of the South Shetland margin, Antarctic Peninsula, *Tectonics,* 13:1345–1370.

McGinnis, J.P., and Hayes, D.E., 1995. The roles of downslope and along-slope depositional processes: southern Antarctic Peninsula continental rise. *In* Cooper, A.K., Barker, P.F., and Brancolini, G. (Eds.), *Geology and Seismic Stratigraphy of the Antarctic Margin.* Am. Geophys. Union, Antarct. Res. Ser., 68:141–156.

Miller, K.G., Wright, J.D., and Fairbanks, R.G., 1991. Unlocking the Ice House: Oligocene–Miocene oxygen isotopes, eustasy, and margin erosion. *J. Geophys. Res.,* 96:6829–6848.

Mohr, B.A.R., 1990. Eocene and Oligocene sporomorphs and dinoflagellate cysts from Leg 113 drill sites, Weddell Sea, Antarctica. *In* Barker, P.F., Kennett, J.P., et al., *Proc. ODP, Sci. Results,* 113: College Station, TX (Ocean Drilling Program), 595–612.

Nitsche, F.O., Cunningham, A.P., Larter, R.D., and Gohl, K., 2000. Geometry and development of glacial continental margin depositional systems in the Bellingshausen Sea. *Mar. Geol.*, 162:277–302.

Nitsche, F.O., Gohl, K., Vanneste, K., and Miller, H., 1997. Seismic expression of glacially deposited sequences in the Bellingshausen and Amundsen seas, West Antarctica. *In* Barker, P.F., and Cooper, A.K. (Eds.), *Geology and Seismic Stratigraphy of the Antarctic Margin* (Pt 2). Am. Geophys. Union, Antarct. Res. Ser., 71:95–108.

O'Brien, P.E., Cooper, A.K., Richter, C., et al., 2001. *Proc. ODP, Init. Repts.*, 188 [CD-ROM]. Available from: Ocean Drilling Program, Texas A&M University, College Station, TX 77845-9547, USA.

Paul, H.A., Zachos, J.C., Flower, B.P., and Tripati, A., 2000. Orbitally-induced climate and geochemical variability across the Oligocene/Miocene boundary. *Paleoceanography*, 15:471–485.

Pirrie, D., Crame, J.A., Riding, J.B., Butcher, A.R., and Taylor, P.D., 1997. Miocene glaciomarine sedimentation in the northern Antarctic Peninsula region: the stratigraphy and sedimentology of the Hobbs Glacier Formation, James Ross Island. *Geol. Mag.*, 134:745–762.

Pope, P.G., and Anderson, J.B., 1992. Late Quaternary glacial history of the northern Antarctic Peninsula's western continental shelf: evidence from the marine record. *In* Elliot, D.H. (Ed.), *Contributions to Antarctic Research III.* Am. Geophys. Union, Antarct. Res. Ser., 57:63–91.

Pudsey, C.J., 1992. Late Quaternary changes in bottom water velocity inferred from grain size in the Northern Weddell Sea. *Mar. Geol.,* 107:9–33.

————, 2000. Sedimentation on the continental rise west of the Antarctic Peninsula over the last three glacial cycles. *Mar. Geol.,* 167:313–338.

Pudsey, C.J., Barker, P.F., and Larter, R.D., 1994. Ice sheet retreat from the Antarctic Peninsula shelf. *Cont. Shelf Res.,* 14:1647–1675.

Pudsey, C.J., and Camerlenghi, A., 1998. Glacial–interglacial deposition on a sediment drift on the Pacific margin of the Antarctic Peninsula. *Antarct. Sci.,* 10:286–308.

Rebesco, M., Camerlenghi, A., and Zanolla, C., 1998. Bathymetry and morphogenesis of the continental margin west of the Antarctic Peninsula. *Terra Antart.,* 5:715–728.

Rebesco, M., Larter, R.D., Barker, P.F., Camerlenghi, A., and Vanneste, L.E., 1997. The history of sedimentation on the continental rise west of the Antarctic Peninsula. *In* Barker, P.F., and Cooper, A.K. (Eds.), *Geology and Seismic Stratigraphy of the Antarctic Margin* (Pt. 2). Am. Geophys. Union, Antarct. Res. Ser., 71:29–50.

Rebesco, M., Larter, R.D., Camerlenghi, A., and Barker, P.F., 1996. Giant sediment drifts on the continental rise of the Antarctic Peninsula. *Geo-Mar. Lett.,* 16:65–75.

Rebesco, M., Pudsey, C.J., Canals, M., Camerlenghi, A., Barker, P.F., Estrada, F., and Giorgetti, A., in press. Case study 27: sediment drifts and deep-sea channel systems, Antarctic Peninsula Pacific margin, mid-Miocene to present. *In* Stow, D.A.V., Pudsey, C.J., Howe, J., Faugeres, J.-C., and Viana, A. (Eds.), *Deep-Water Contourite Systems: Modern Drifts and Ancient Series, Seismic and Sedimentary Characteristics*, Spec. Publ.—Geol. Soc. London.

Reynolds, J.M., 1981. The distribution of mean annual temperatures in the Antarctic Peninsula. *Br. Antarct. Surv. Bull.,* 54:123–133.

Ruddiman, W.F., Raymo, M.E., Martinson, D.G., Clement, B.M., and Backman, J., 1989. Pleistocene evolution: Northern Hemisphere ice sheets and North Atlantic Ocean. *Paleoceanography,* 4:353–412.

Shipboard Scientific Party, 1999a. Shelf Transect (Sites 1100, 1102, and 1103). *In* Barker, P.F., Camerlenghi, A., Acton, G.D., et al., *Proc. ODP, Init. Repts.,* 178: College Station, TX (Ocean Drilling Program), 1–58.

————, 1999b. Site 1095. *In* Barker, P.F., Camerlenghi, A., Acton, G.D., et al., *Proc. ODP, Init. Repts.,* 178, 1–173 [CD-ROM]. Available from: Ocean Drilling Program, Texas A&M University, College Station, TX 77845-9547, U.S.A.

————, 1999c. Site 1097. *In* Barker, P.F., Camerlenghi, A., Acton, G.D. et al., *Proc. ODP, Init. Repts.,* 178, 1–66 [CD-ROM]. Available from: Ocean Drilling Program, Texas A&M University, College Station, TX 77845-9547, U.S.A.

————, 1999d. Site 1101. *In* Barker, P.F., Camerlenghi, A., Acton, G.D., et al., *Proc. ODP, Init. Repts.,* 178, 1–83 [CD-ROM]. Available from: Ocean Drilling Program, Texas A&M University, College Station, TX 77845-9547, U.S.A.

Smellie, J.L., 1999. Lithostratigaphy of Miocene-Recent alkaline volcanic fields in the Antarctic Peninsula and eastern Ellsworth Land. *Antarct. Sci.,* 11:362–378.

Tiedemann, R., Sarnthein, M., and Shackleton, N.J., 1994. Astronomic timescale for the Pliocene Atlantic $\delta^{18}O$ and dust flux records of Ocean Drilling Program Site 659. *Paleoceanography,* 9:619–638.

Tomlinson, J.S., Pudsey, C.J., Livermore, R.A., Larter, R.D., and Barker, P.F., 1992. GLORIA survey of the Pacific margin of the Antarctic Peninsula: tectonic fabric and sedimentary processes. *In* Yoshida, Y., Kaminuma, K., and Shiraishi, K. (Eds.), *Recent Progress in Antarctic Earth Science:* Tokyo (Terra Sci. Publ.), 423–429.

Troedson, A.L., and Riding, J.B., 2002. Latest Oligocene to earliest Miocene strata of King George Island, South Shetland Islands, Antarctica: stratigraphy, facies analysis and implications for Antarctic Peninsula glacial history. *J. Sedim. Res.*, 72:510–523.

Troedson, A.L., and Smellie, J.L., 2002. The Polonez Cove Formation of King George Island, West Antarctica; stratigraphy, facies and palaeoenvironmental implications. *Sedimentology,* 49:277–301.

Tucholke, B.E., Hollister, C.D., Weaver, F.M., and Vennum, W.R., 1976. Continental rise and abyssal plain sedimentation in the southeast Pacific basin—Leg 35 Deep Sea Drilling Project. *In* Hollister, C.D., Craddock, C., et al., *Init. Repts. DSDP,* 35: Washington (U.S. Govt. Printing Office), 359–400.

Vanneste, L.E., and Larter, R.D., 1995. Deep-tow boomer survey on the Antarctic Peninsula Pacific margin: an investigation of the morphology and acoustic characteristics of late Quaternary sedimentary deposits on the outer continental shelf and upper slope. *In* Cooper, A.K., Barker, P.F., and Brancolini, G. (Eds.), *Geology and Seismic Stratigraphy of the Antarctic Margin.* Am. Geophys. Union, Antarct. Res. Ser., 68:97–121.

Vorren, T.O., Lebesbye, E., Andreassen, K., and Larsen, K.B., 1989. Glacigenic sediments on a passive continental margin as exemplified by the Barents Sea. *Mar. Geol.,* 85:251–272.

Wise, S.W., Jr., Breza, J.R., Harwood, D.M., Wei, W., and Zachos, J.C., 1992. Paleogene glacial history of Antarctica in light of Leg 120 drilling results. *In* Wise, S.W., Jr., Schlich, R., et al., *Proc. ODP, Sci. Results,* 120: College Station, TX (Ocean Drilling Program), 1001–1028.

Yang, S., and Harwood, D.M., 1997. Late Quaternary environmental fluctuations based on diatoms from Yanou Lake, King George Island, Fildes Peninsula, Antarctica. *In* Ricci, C.A. (Ed.), *The Antarctic Region: Geological Evolution and Processes.* Siena, 853–859.

Zachos, J.C., Pagani, M., Sloan, L., Thomas, E., and Billups, K., 2001. Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science,* 292:686–693.

Figure F1. The present-day Antarctic ice sheet with ice flow lines, from Barker et al. (1998), modified to show recent ODP drill sites off the Antarctic Peninsula (AP) and Prydz Bay (PB) and drill sites off Cape Roberts (CR) in the Ross Sea. Also marked are the Transantarctic Mountains (TAM) along the western edge of the Ross Sea (ERS), the Ellsworth Mountains (EM) south of the Weddell Sea, the interior Gamburtsev Mountains (GaM), and the continental margin off Wilkes Land (WL).

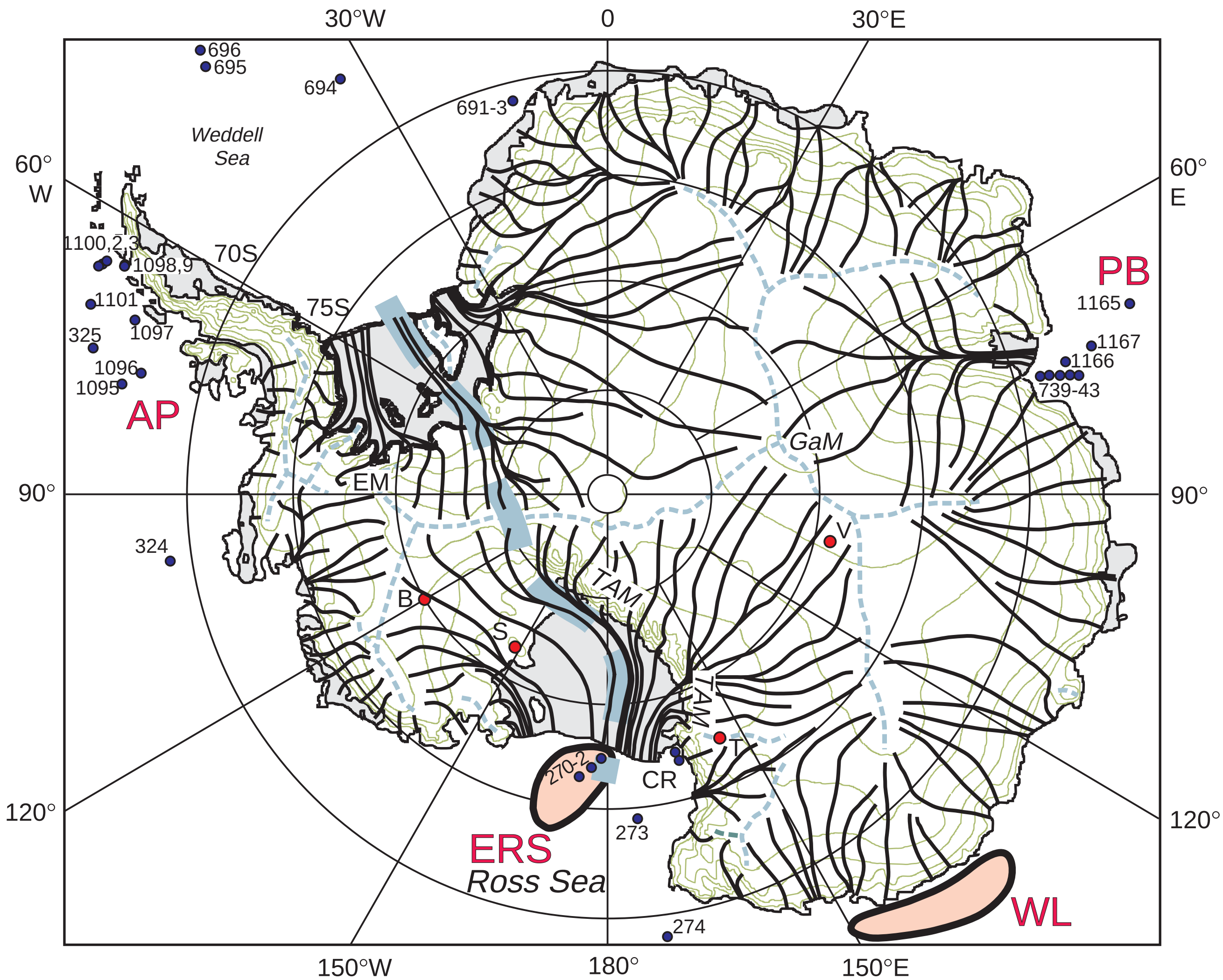

Figure F2. Sites drilled during ODP Leg 178 on the Antarctic Peninsula Pacific margin, with bathymetry from Rebesco et al. (1998) and showing sediment Lobes L1–L4 on the outer continental shelf, Drifts D1–D8 on the upper continental rise, the mid-shelf high (MSH), and DSDP Site 325. Br. St. = Bransfield Strait, S. Shet = South Shetland Islands. Sites 1095 and 1096 lie on Drift D7 and Site 1101 on Drift D4.

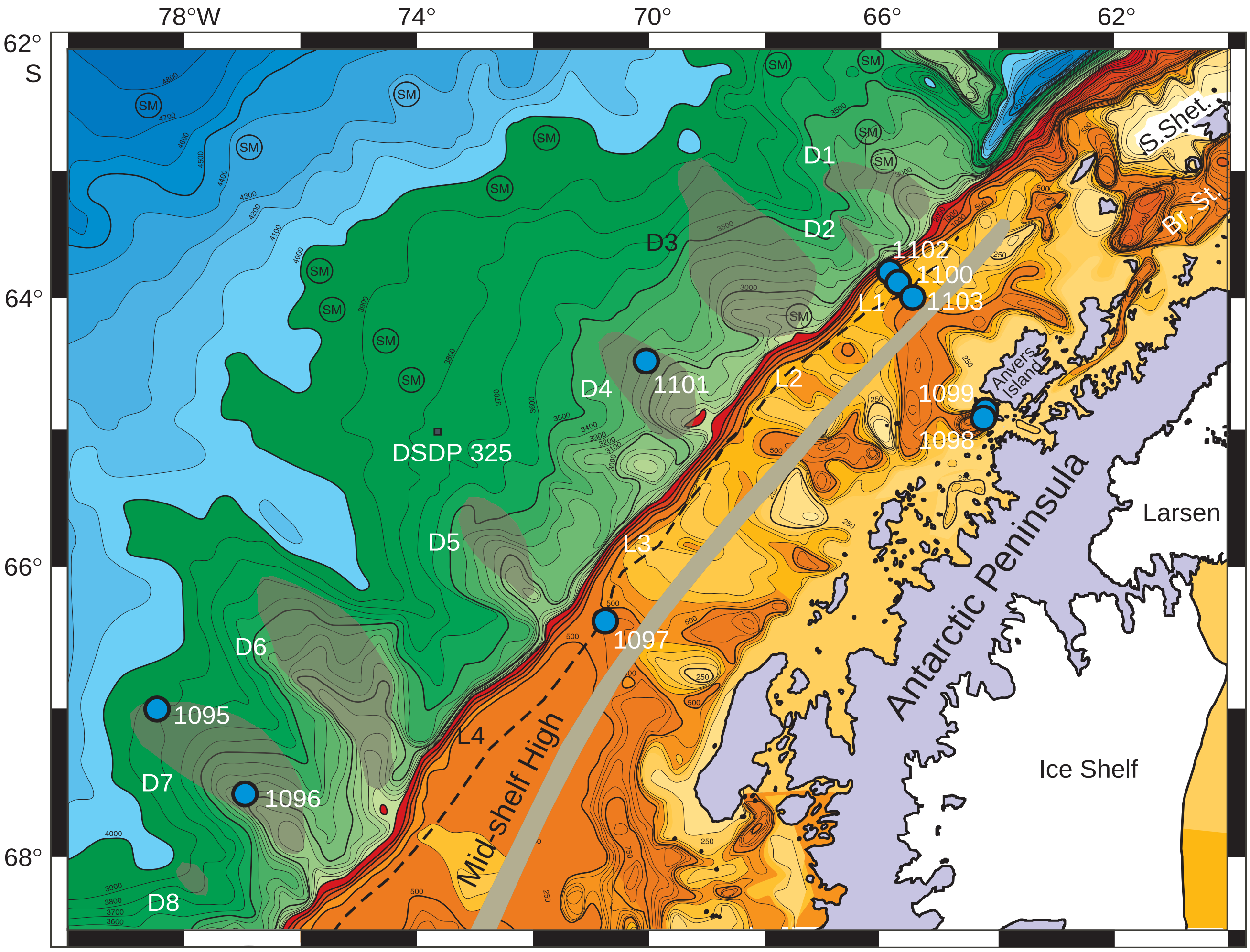

Figure F3. Seismic reflection profile I95-152 along a transect of the outer continental shelf through Lobe L1, with positions and penetration of drill sites marked and schematic projected position and penetration of Site 1097 (from between Lobe L3 and Lobe L4, where sequence group S1 topsets are much thinner). Shaded copy below shows seismic sequence groups S1, S2, S3, and S4. Modified from Barker, Camerlenghi, Acton, et al. (1999). TWT = two-way traveltime.

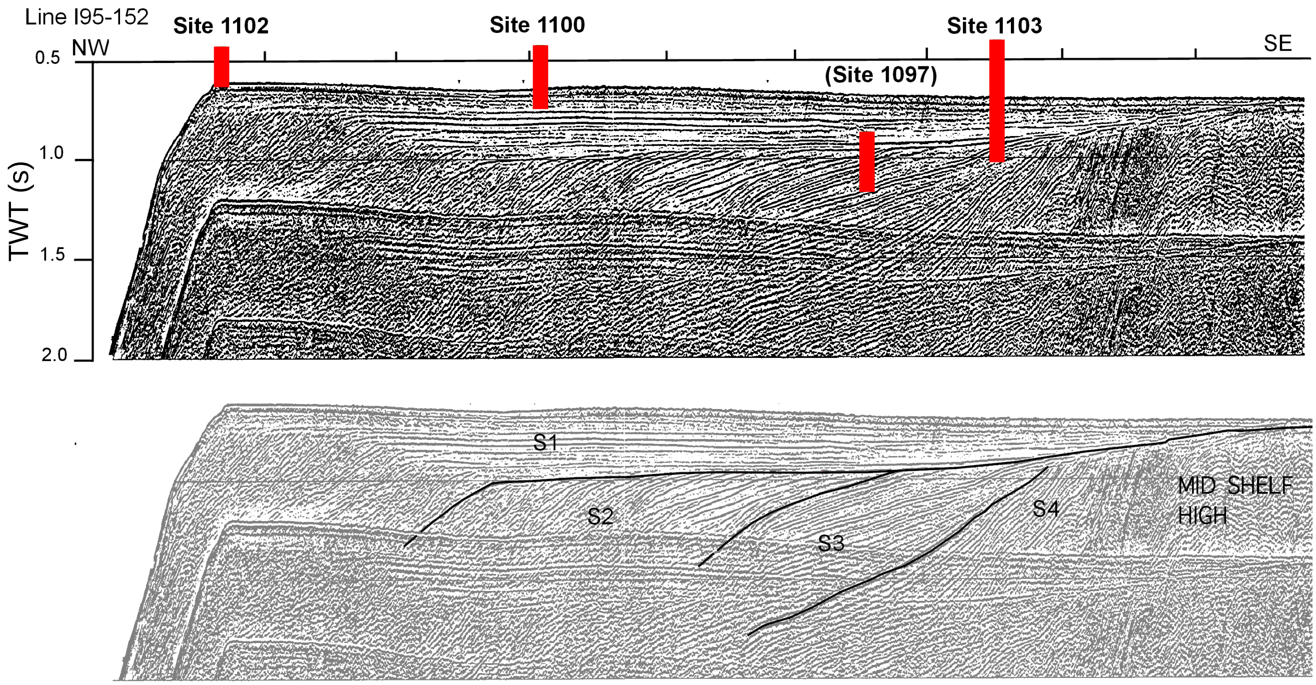

Figure F4. Stratigraphic control on drilled sequences at continental shelf Sites 1097, 1100, and 1103, revised from Barker, Camerlenghi, Acton, et al. (1999) to include additional postcruise constraints on S3 age (Di Vincenzo et al., Chap. 22, and Lavelle et al., Chap. 27, both this volume; L. Osterman and M. Iwai, pers. comm., 2001).

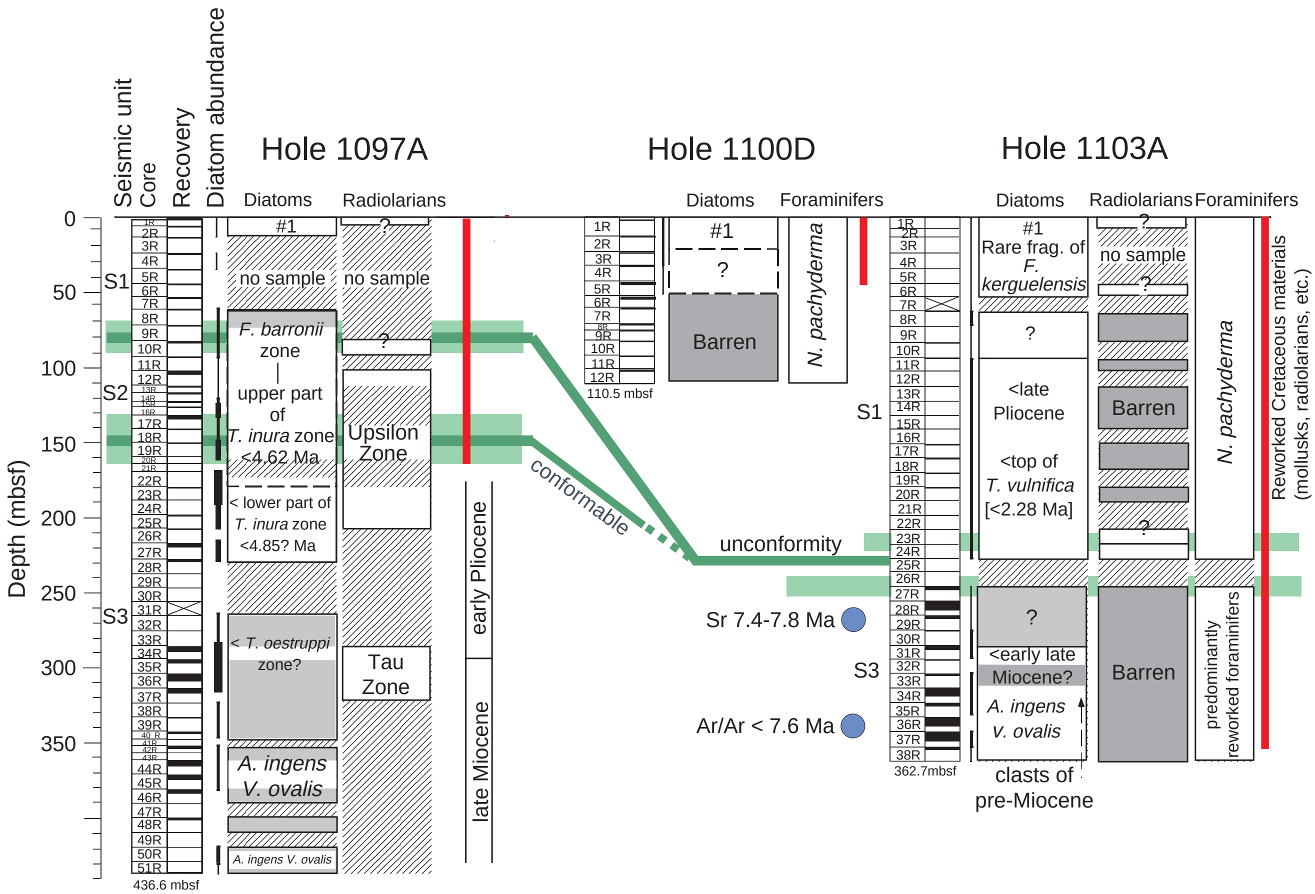

Figure F5. Revised age-depth curve for Site 1095 (from Iwai et al., Chap. 36, this volume), showing the extent of remaining areas of uncertainty and the magnetic reversal stratigraphy now adopted. (t) = top, (b) = bottom.

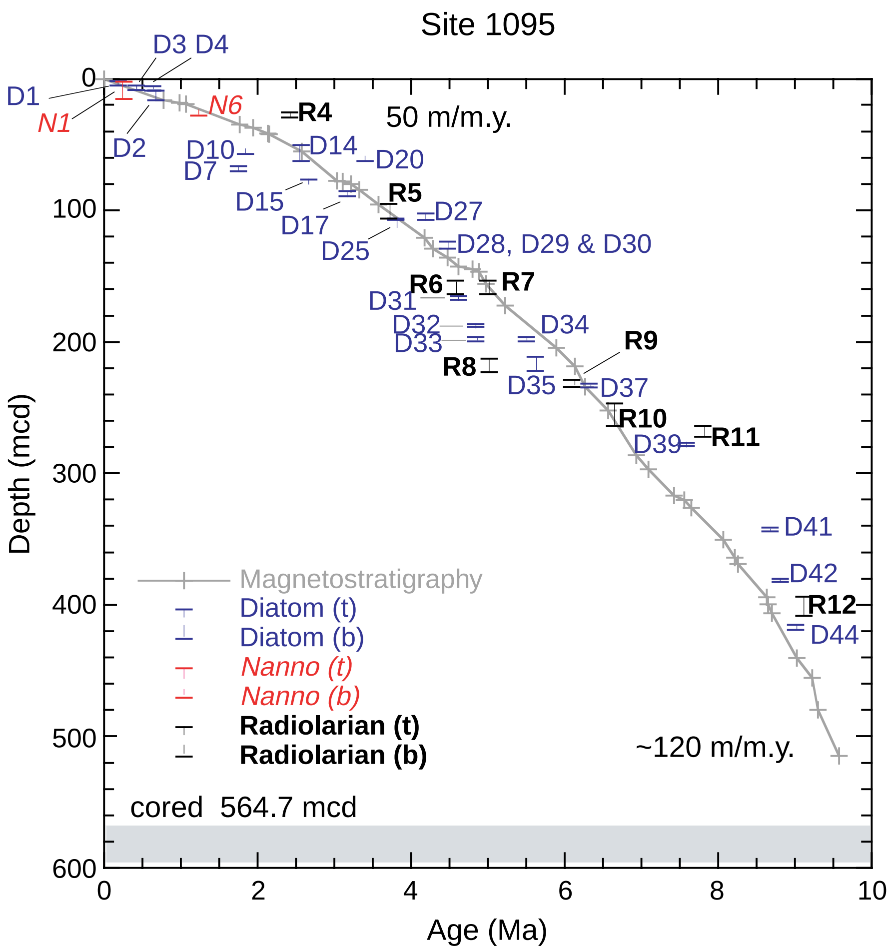

Figure F6. Uppermost sections of cores at continental rise sediment drift Sites 1095, 1096 and 1101, showing lithologic variation over the past 200–300 k.y. and correlations of (interglacial) oxygen isotopic stages 1, 5, 7, and 9, from Barker et al. (1998). IRD = ice-rafted detritus.

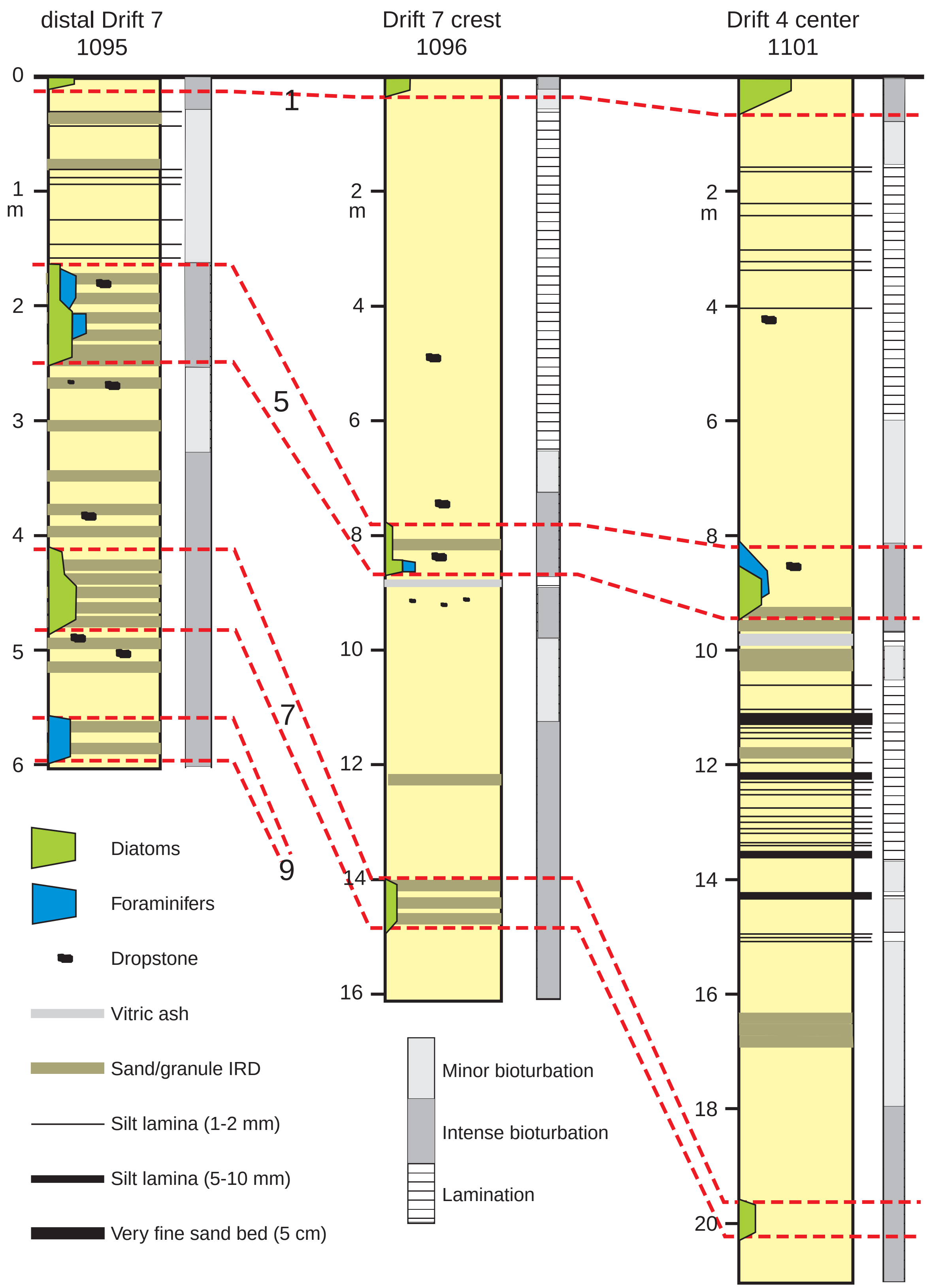

Figure F7. From Figure F2, p. 23, of Hillenbrand and Ehrmann (Chap. 8, this volume) showing the percentages of smectite, illite, and chlorite in samples from Site 1095, plotted against depth. Typical modern glacial and interglacial compositions persist throughout the 9-m.y. section (see text). Compare this evidence of ice sheet volume with the climatic evidence of Figure F8, p. 39.

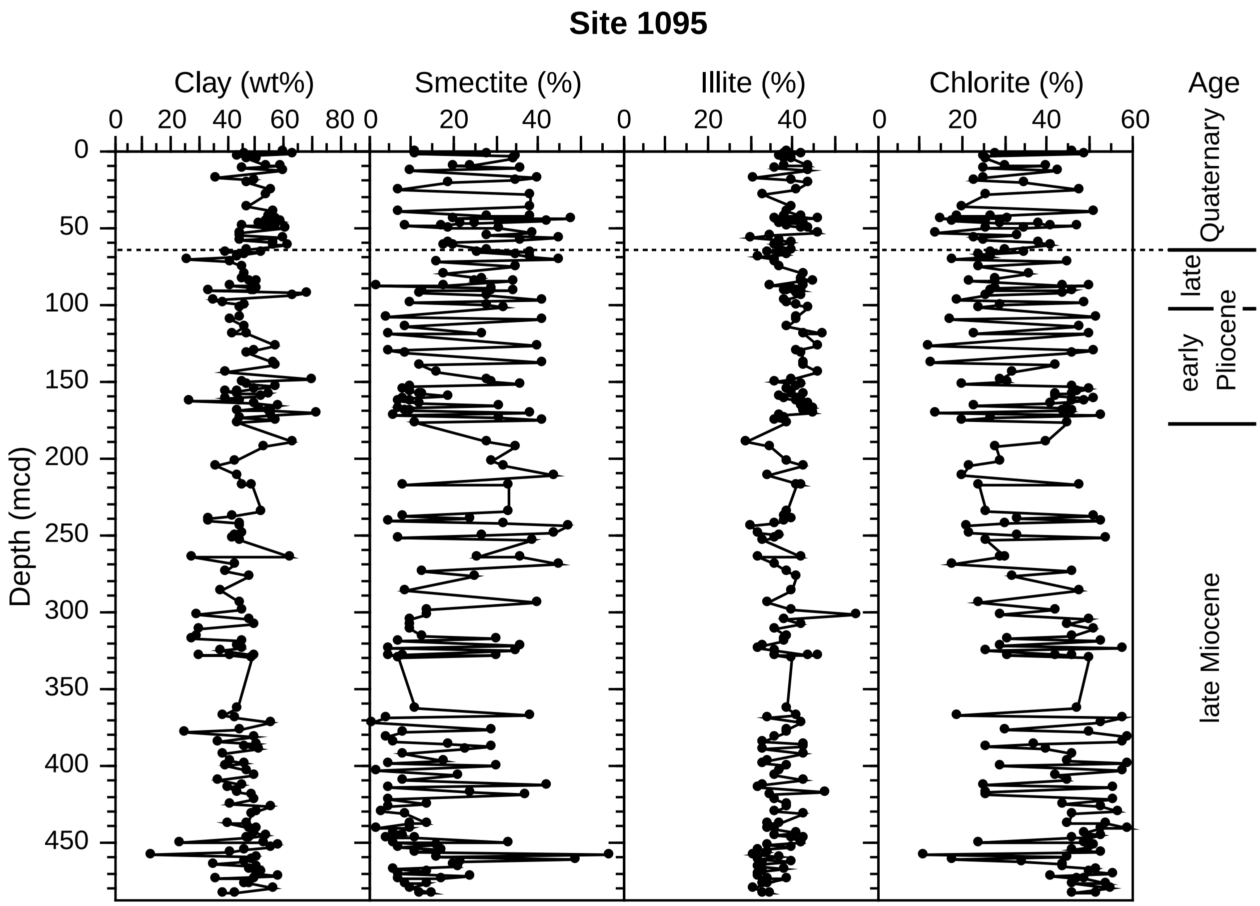

Figure F8. From the data of Hillenbrand and Fütterer (Chap. 23, this volume), showing the variation of biogenic opal at Site 1095, plotted against depth. Most low values are considered to have been from samples of distal turbidites.

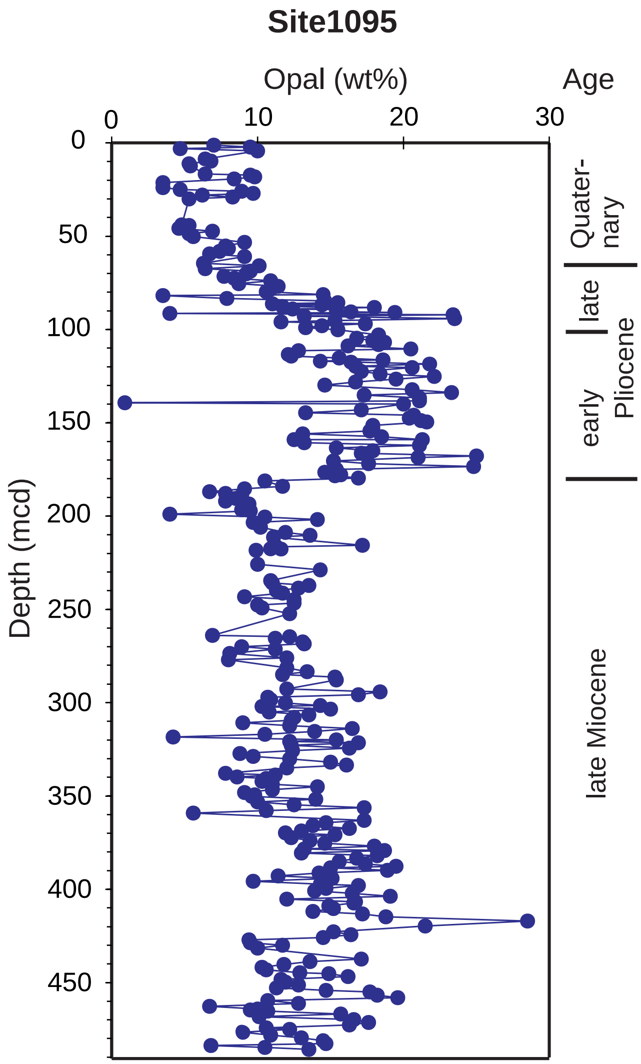

Figure F9. Raw benthic oxygen isotopic data (Zachos et al., 2001) for the last 40 m.y. (and a smoothed version of the same data—a line through points at 200-k.y. intervals, least-squares fit to a five-point running mean) compared with observations of the glacial state of the Antarctic Peninsula, from onshore evidence and from ODP Leg 178. Timescale from Berggren et al. (1995). IRD = ice-rafted detritus.