Antarctic Science 18 (1), 83–99 (2006) © Antarctic Science Ltd Printed in the UK DOI: 10.1017/S0954102006000083

A review of the Cenozoic stratigraphy and glacial history of the Lambert Graben–Prydz Bay region, East Antarctica

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Abstract: The Cenozoic glacial history of East Antarctica is recorded in part by the stratigraphy of the Prydz Bay–Lambert Graben region. The glacigene strata and associated erosion surfaces record at least 10 intervals of glacial advance (with accompanying erosion and sediment compaction), and more than 17 intervals of glacial retreat (enabling open marine deposition in Prydz Bay and the Lambert Graben). The number of glacial advances and retreats is considerably less than would be expected from Milankovitch frequencies due to the incomplete stratigraphic record. Large advances of the Lambert Glacier caused progradation of the continental shelf edge. At times of extreme glacial retreat, marine conditions reached > 450 km inland from the modern ice shelf edge. This review presents a partial reconstruction of Cenozoic glacial extent within Prydz Bay and the Lambert Graben that can be compared to eustatic sea-level records from the southern Australian continental margin.

Received 1 December 2004, accepted 15 July 2005

Key words: glaciation, ice sheet, Milankovitch, Prince Charles Mountains

Introduction

In recent years there has been considerable investigation of the Cenozoic geology of the Lambert Graben and Prydz Bay region over a broad latitudinal transect ~1000 km long between latitudes 64.5–73.5°S, and extending from the continental rise to ~450 km inland. In consequence, the region has become a major source of information on the Cenozoic evolution and glacial history of Antarctica. Offshore studies include seismic survey and sediment recovery through Australian, Russian, and United States research cruises, and the international Ocean Drilling Program Legs 119 and 188 (Barron *et al.* 1991, Leitchenkov *et al.* 1994, Shipboard Scientific Party 2001, Cooper & O'Brien 2004, Leventer *et al.* 2004). Quaternary geological features and sediments in Prydz Bay have been mapped and sampled (Domack *et al.* 1998), and characteristics of modern sediments have been recorded by Quilty (1986) and Harris *et al*. (1997, 1998). Shore-based Cenozoic stratigraphic research has also been undertaken in ice-free coastal areas and inland deglaciated regions (e.g. Pickard 1986, Quilty *et al.* 2000, McKelvey *et al.* 2001). This paper summarizes the Cenozoic stratigraphy and associated erosion surfaces in the vicinity of Lambert Graben and Prydz Bay. Furthermore the glacial history of the Lambert Graben–Prydz Bay region is compared to the eustatic sea-level record from Australia, and with deep-sea isotope studies, both of which provide a partial proxy for East Antarctic Ice Sheet (EAIS) volume changes.

Study area

This paper reviews the Cenozoic stratigraphy of the Antarctic continent and margin from 60°E to 80°E. The region includes the continental rise, slope and shelf, covering both Prydz Bay and the adjacent Mac. Robertson Shelf, and also the Lambert Graben and Prince Charles Mountains (PCMs) (Fig. 1).

Structural setting

The Lambert Graben and Prydz Bay basin constitute a dominantly north–south structural depression, which arose as an aulacogen within the triple junction formed due to rifting between India and Antarctica in the Late Palaeozoic and Early Mesozoic (Stagg 1985, Cooper *et al.* 1991a). Prydz Bay (Fig. 1) marks the seaward end of the Lambert Graben, a major structural feature, which allows access of marine waters into the East Antarctic interior, particularly during periods of reduced ice-sheet volume. The Lambert Graben–Prydz Bay basin is some 900 km long, and progressively narrows (from 1000 to 50 km) and deepens (from 200 to 1600 m below sea-level) farther inland (Fedorov *et al.* 1982, Kurinin & Aleshkova 1987, Harris *et al.* 1997) (Fig. 2). The region is heavily glaciated, the Lambert Graben being fully covered by the Amery Ice Shelf (Phillips *et al.* 1996) north of a grounding line at *c.* 73°40'S. This ice shelf is fed by the Lambert Glacier and numerous tributary glaciers, which together drain 13% (by surface area) of the EAIS, the largest catchment area in East

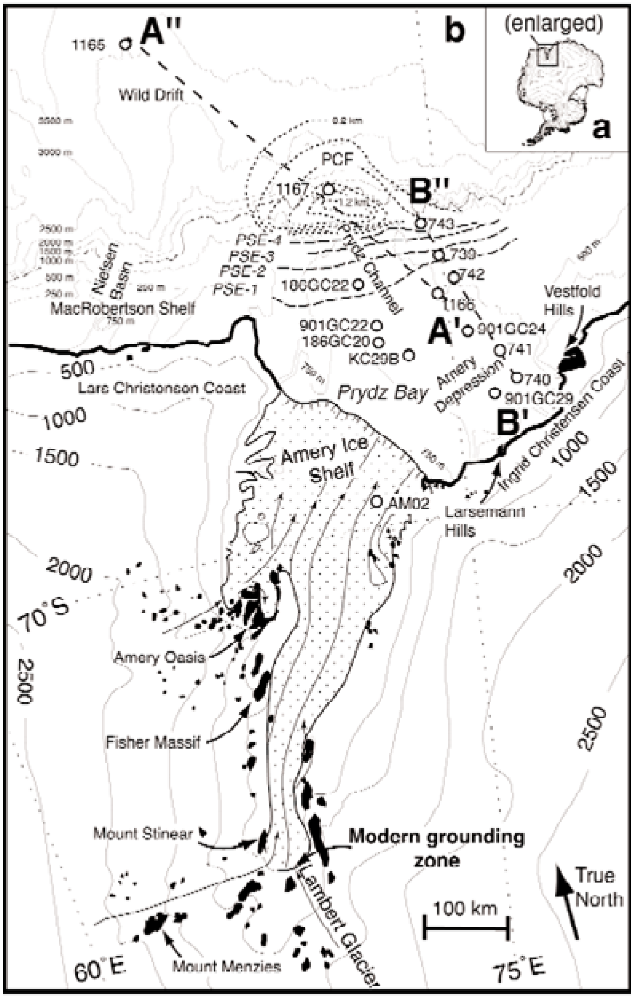

Fig. 1.**a.** Antarctica, with **b.** the Prydz Bay–Lambert Graben region in detail. Ice sheet topography and sea-floor bathymetry are contoured (units in metres). The modern glacial grounding zone occurs ~450 km from the seaward edge of the Amery Ice Shelf. Locations mentioned in the text are named and illustrated: core (circles), coastal and mountain sites (black). The position of two seismic profiles (Fig. 3) are marked by wide dashed lines, between the points A'', A', B'' and B'. The past position of the continental shelf edge moved (from PSE-1 to -4) and is marked by medium-width dashed lines (Cooper *et al.* 1991b, Leitchenkov *et al.* 1994). Progradation of the shelf edge culminated with deposition of the PCF and creation of the modern continental shelf edge (~1000 m bathymetric contour). Depocenter contours of the PCF, from 0.2 to 1.5 km sediment thickness, are marked by narrow-width dashed lines (from Leitchenkov *et al.* 1994).

Antarctica (Hambrey 1991). Uplifted areas on the rift margin are now largely ice-free regions constituting the PCMs, a series of nunataks and ranges of up to 3355 m in altitude (Fig. 1). The major ice-free regions include the Amery Oasis, Fisher Massif, Mount Stinear and Mount Menzies. Ice-free areas also occur along the Prydz Bay coast in the Vestfold and Larsemann Hills.

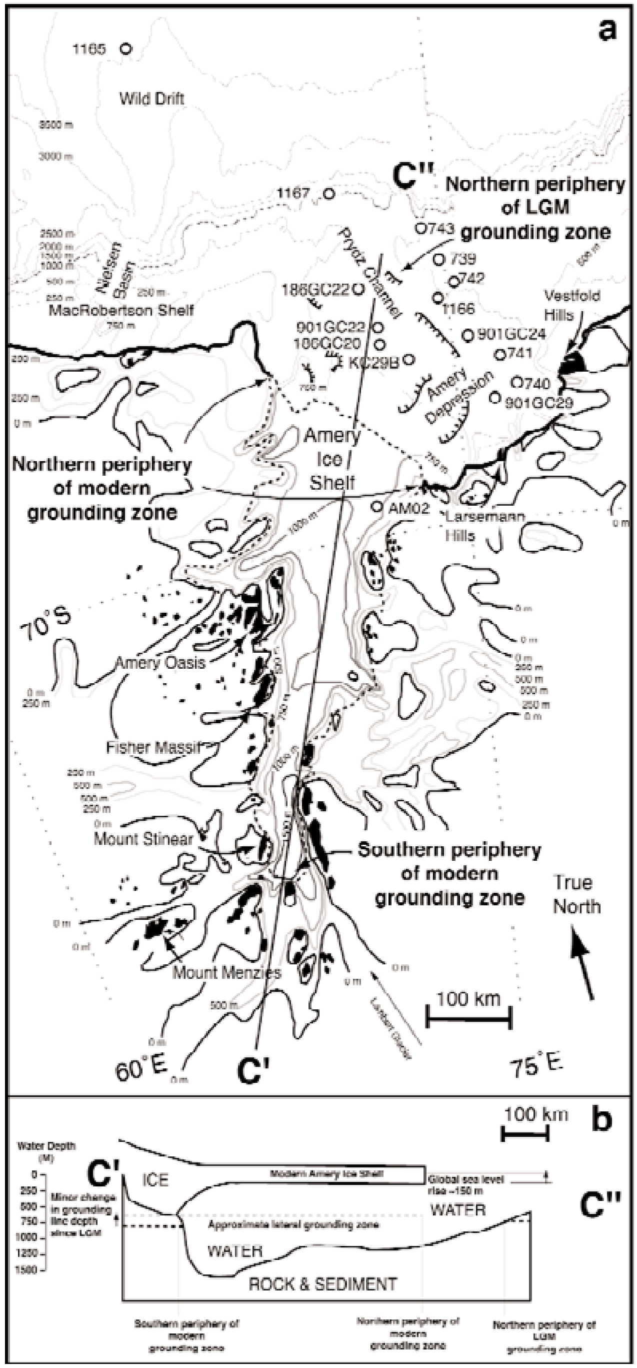

Fig. 2. a. Sea-floor bathymetry and sub-glacial topography of the Prydz Bay–Lambert Graben region (modified from Kurinin & Aleshkova 1987, A. Ruddell, personal communication 2000). The Lambert Glacier, and glaciers from the adjacent Ingrid and Lars Christensen Coasts, converged and grounded around the eastern and western peripheries of Prydz Channel in Prydz Bay during the Last Glacial Maximum (LGM) (Domack *et al.* 1998). The Amery Ice Shelf may have extended into the Lambert Graben during the LGM, but its inland extent at this time is unknown. **b.** The section C'' to C' diagrammatically illustrates the retreat in position of the northern periphery of the grounding zone since the LGM (note: the modern ice shelf is not to scale). The modern Amery Ice Shelf covers much of the Lambert Graben.

Prydz Bay lies seaward of the Amery Ice Shelf and extends to the continental shelf edge. West of Prydz Bay is the Mac. Robertson Shelf, dissected by deep north–south oriented submarine valleys (e.g. Nielsen Basin). The inner part of Prydz Bay is the Amery Depression, which descends to depths of about 800 m but contains deeper closed depressions (to 1100 m) in the south-west (Taylor & Leventer 2003). Extending north-west from the Amery Depression to the shelf edge is a trough 150 km wide and 600 km deep, known as the Prydz Channel (Quilty 1986, O'Brien & Leitchenkov 1997) (Fig. 1). Sea floor contours bulge seaward from the northern end of Prydz Channel indicating the presence of a trough mouth fan, so typical of many glaciated margins, and termed the Prydz Channel Fan (PCF) (Shipboard Scientific Party 2001). Beyond the PCF, the continental slope and rise feature north-west oriented sediment ridges formed by turbidity current redeposition and subsequent contour current transport of Prydz Bay sediment (Kuvaas & Leitchenkov 1992). The more westerly of these sediment ridges is the Wild Drift (Shipboard Scientific Party 2001), which lies some 1000 km north of Mount Stinear and Mount Menzies of the southern PCMs.

Regional geology

Prydz Bay

Depositional units within Prydz Bay have been identified from seismic data (Stagg 1985, Cooper *et al.* 1991a, 2001, Leitchenkov *et al.* 1994, O'Brien *et al.* 1995, O'Brien & Harris 1996, Harris *et al.* 1997, Erohina *et al.* 2004) (Fig. 3). The Lambert Graben–Prydz Bay basin sediment fill rests unconformably on Proterozoic metamorphic basement (seismic acoustic unit PS.5) (Cooper *et al.* 1991a). Permian and Mesozoic continental deposits initially accumulated on this surface within a rift basin. In southern Prydz Bay red sandstone and siltstone of probable Mesozoic age are up to 5 km thick (seismic unit PS.4 of Cooper *et al.* (1991a)). During the later Mesozoic in Prydz Bay, alluvial and restricted lagoon and marine sediments were deposited (unit PS.2B; Cooper *et al.* 1991a, Erohina *et al.* 2004). Within these strata *in situ* Early Cretaceous (Albian) fern pollen have been recovered from ODP Site 741 and similarly a Late Cretaceous (Turonian–?Santonian) palynoflora occurs at ODP Sites 742 and 1166 (Truswell & Macphail 2001).

Seismic profiles also link the Cenozoic stratigraphy across Prydz Bay down to the Wild Drift on the continental rise; however, correlation is not straightforward. The Cenozoic succession spans from the Palaeocene to Recent, but major glacial generated disconformities cause this record to be fragmentary. Furthermore, sequences that appear continuous on low-resolution seismic profiles can, with higher resolution seismic data, be shown to also contain unconformities. On the inner shelf only a thin Cenozoic layer (unit PS.1) unconformably overlies Mesozoic and Palaeozoic strata (Cooper *et al.* 1991a). Towards the outer shelf, the Cenozoic sequence thickens and includes at least three units (units PS.1, PS.2A2, and PS.2A1 of Cooper *et al.* (1991a) and Erohina *et al.* (2004)). Intermittent Cenozoic glacial sedimentation has caused the continental slope to prograde seawards, culminating in deposition of the PCF (Cooper *et al.* 1991a). Northwards of the continental shelf edge in water depths beyond the direct influence of glacial erosion, the Wild Drift contains a relatively complete Early Miocene to Recent stratigraphic record (Florindo *et al.* 2003).

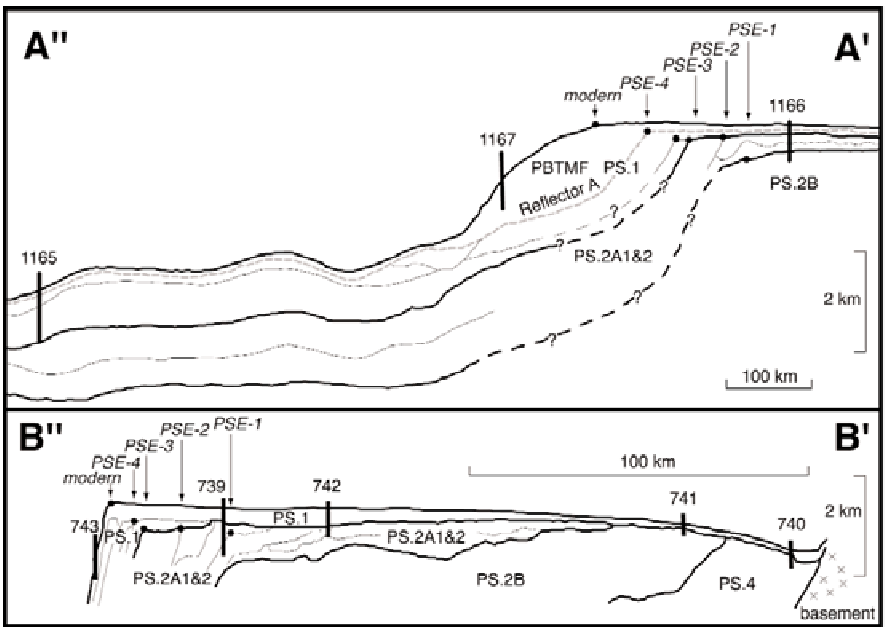

Fig. 3. Seismic profiles from A'' to A' and B'' to B' illustrating seismic units (PS.4 to PS.1), palaeo-shelf edges (PSE-1 to -4), and position and depth of ODP drill sites (modified from Cooper & O'Brien 2004).

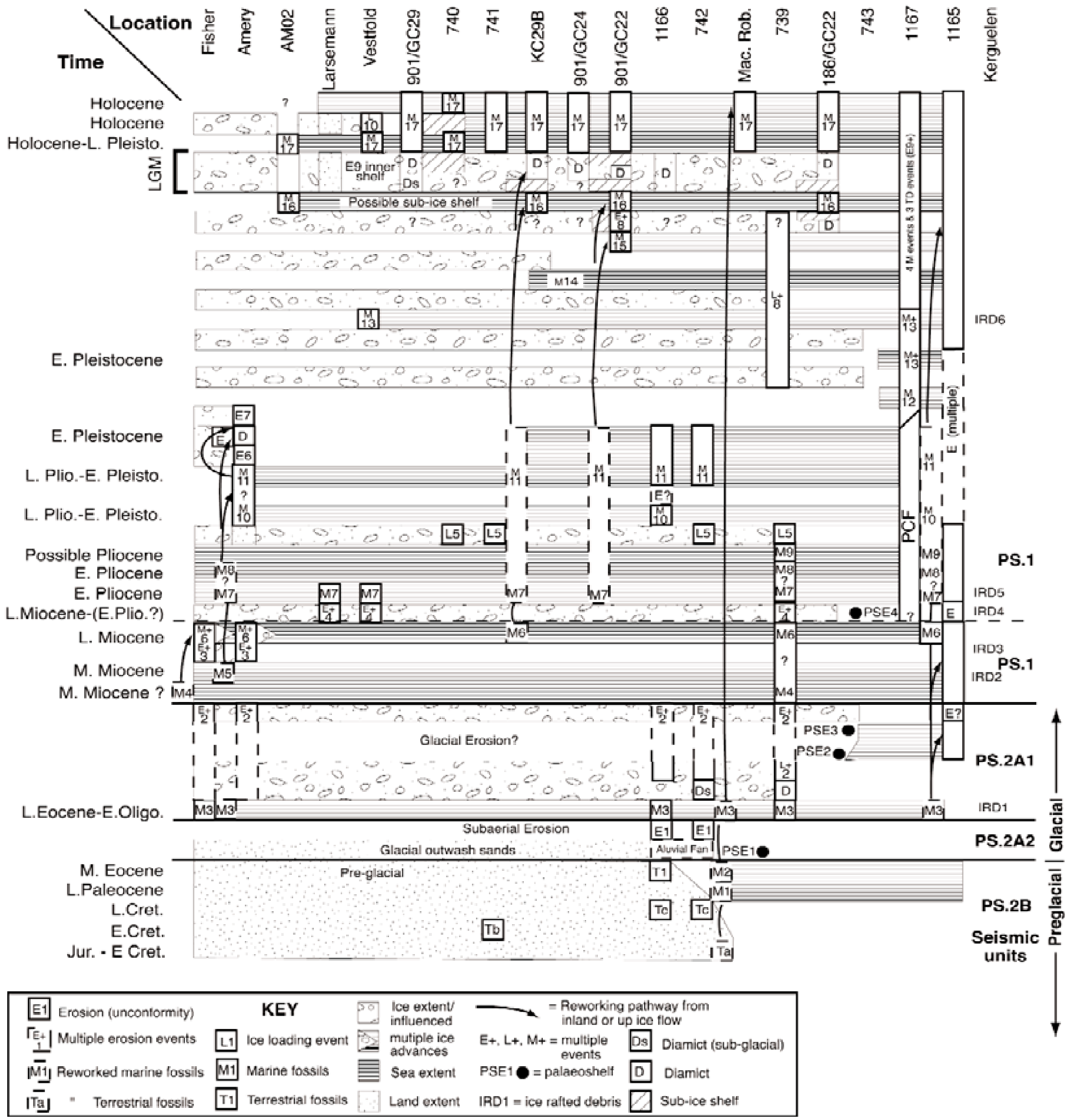

Fig. 4. Illustrative synthesis of the Cenozoic stratigraphy of the Prydz Bay–Lambert Graben region and adjacent Mac. Robertson Shelf, and continental rise (Site 1165). Locations have been arranged from left to right with decreasing latitude, which approximates to increasing distance from the Antarctic Ice Sheet. Geologic age increases down the page. Boxes indicate events as tabulated in Table I (T = terrestrial, M = marine, E = erosion, L = ice loading). These are numbered in chronological order through the Cenozoic (e.g. M1, M2, etc). Three Mesozoic terrestrial events are also illustrated and have been labelled Ta, Tb, and Tc. The solid outlined boxes around the event labels indicate there is *in situ* evidence for these events. Boxes with dashed outlines indicate those events identified from reworked microfossils (arrows indicate the younger strata where the reworked fossils occur). It has been assumed that the reworked fossils were glacially transported from deposits originally farther inland and have been illustrated slightly landward of the location where they were found. The extent of terrestrial environments, grounded ice and marine sediments are indicated by patterned symbols (as in the key). Blank regions are unresolved due to lack of data. Seismic units and the approximate age and position of palaeo-shelf edges are indicated. Iceberg events north beyond Prydz Bay are evident within the Kerguelen ice rafted debris (IRD) record (far right) and may correspond to events in the Prydz Bay–Lambert Graben region.

Lambert Graben / Prince Charles Mountains

The Proterozoic metamorphic basement of the PCMs decreases in grade from granulite facies in the north, to green schist facies in the south (Tingey 1991, Mikhalsky *et al.* 2001). Permo–Triassic Amery Group strata outcrop in the Amery Oasis (Webb & Fielding 1993, McLoughlin & Drinnan 1997) and consist of fluvial deltaic sandstone, siltstone, shale, coal and conglomerate.

The PCMs are partially draped with Oligocene to Plio–Pleistocene marine sediments of the Pagodroma Group (Hambrey & McKelvey 2000a, 2000b). These strata occur up to 450 km inland from the Prydz Bay coast on Mount Stinear (Whitehead *et al.* 2000) and were deposited in a fjordal marine environment within the Lambert Graben during times of glacial recession (Hambrey & McKelvey 2000b). *In situ* and glacially reworked marine fossils enable biostratigraphic dating of the Pagodroma Group's Bardin Bluffs and Battye Glacier formations on the Amery Oasis (250 km inland), and the Fisher Bench and Mount Johnston formations on Fisher Massif (300 km inland) (McKelvey *et al.* 2001, Whitehead *et al.* 2004). During glacial advances Lambert Graben ice eroded the formations and underlying basement and re-deposited the sediment on the continental shelf and slope of Prydz Bay (Barker *et al.* 1998). The four formations preserved in the northern PCM are erosional remnants occupying the floors and abutting the walls of former fjords (McKelvey *et al.* 2001). Farther inland, 600 km from the Prydz Bay coast, on the Mount Menzies Range, relict landforms and mantling Cenozoic strata record several phases of alpine glacial erosion and deposition. However, these events have not yet been dated (Whitehead & McKelvey 2002). Erratics of Cenozoic subglacial diamict also occur in the Grove Mountains, east of the PCMs and Lambert Graben. These may contain palynological evidence for a terrestrial flora but the age is not well constrained (Fang *et al.* 2004). Late Quaternary glacial events in the PCMs have caused little landscape modification and deposited only small moraines preserved above the present regional ice surface (Whitehead & McKelvey 2002). During the Quaternary the Amery Ice Shelf either retreated considerably, or else open-marine sediment was advected southwards beneath the shelf ice, as sediment of this age have been there recovered using short gravity corers (Hemer & Harris 2003).

Mac. Robertson Shelf

The Mac. Robertson Shelf has no adjacent onshore outcrop of Cenozoic strata and has not been drilled, but cores taken during Australian National Antarctic Research Expeditions (ANARE) research cruises in 1993, 1995 and 1997 (O'Brien *et al.* 1995, Harris *et al.* 1997) have yielded evidence of a considerable sedimentary cover (Truswell *et al.* 1999, Quilty *et al.* 2000). In contrast to the north–south orientation of the Lambert Graben–Prydz Bay depression, the Mac. Robertson Shelf is dominated by east–west trending rift basins that formed prior to the separation of Antarctica and India (O'Brien *et al*. 2004). These rift basins hold terrestrial sediments containing palynomorphs of Early and Middle Jurassic, Early Cretaceous and Middle–Late Eocene age (O'Brien *et al*. 1995, Truswell *et al*. 1999, Quilty *et al*. 2000, Truswell & Macphail 2001, Macphail & Truswell 2004). The basins also contain marine strata with fossils of possibly Late Cretaceous, and definite Late Palaeocene and Middle Eocene, and Late Eocene–Oligocene ages (Quilty *et al.* 1999, 2000). Younger submarine valleys, such as the Neilsen Basin dissect the rift basins. The older Mac. Robertson Shelf strata (and their associated fossils) have slumped from the valley sides and so have been reworked into Quaternary deposits (Quilty *et al.* 1999).

Vestfold Hills / Larsemann Hills

The Vestfold Hills and Larsemann Hills consist of Archaean and Proterozoic gneisses, respectively, upon which thin and often fossiliferous Pliocene and Quaternary sediments occupy depressions (Adamson & Pickard 1986a, 1986b, Pickard *et al*. 1988, Hirvas *et al*. 1993, Gore 1997, Quilty *et al.* 2000, Harwood *et al.* 2000, Hodgson *et al.* 2001). The oldest known Cenozoic strata in the Vestfold Hills are the mid-Pliocene Sørsdal Formation, which contains a diverse marine fossil flora and fauna (Pickard 1986, Feldmann & Quilty 1996, Quilty *et al.* 2000, Harwood *et al.* 2000, Whitehead *et al*. 2001, Fordyce *et al.* 2002). Thin patches of similar-aged fossiliferous marine sediments occur in the Larsemann Hills (Quilty *et al.* 1990, McMinn & Harwood 1995). Other younger Cenozoic strata in both areas attest to repeated glaciation, and several marine transgressions and regressions (e.g. Hirvas *et al.* 1993).

Stratigraphic review

This overview of the Cenozoic stratigraphy of the Lambert Graben and Prydz Bay has been synthesized from previous work in the region. Evidence for deposition, erosion and ice-loading comes from seismic data, sediment cores, and outcrop sections. Original deposit, sequence, unit, and boundary names are retained wherever possible. However, to further aid clarification when placing in chronological order diverse events of similar ages, an additional labeling system has been created to emphasize:

marine deposits (M1, M2, etc),

terrestrial deposits (T1, etc),

unconformities (E1, etc), and

ice loading events (L1, etc) (Fig. 4; Table I).

The symbol "+" denotes multiple events that cannot be separated.

Table I. Events (T = terrestrial, M = marine, E = erosion, L = ice loading) are listed in chronological order. The symbol '+' denotes multiple events, which cannot be dated separately due to low dating resolution. The 'event age' is constrained by the oldest and youngest age range evident for that event. Evidence includes location, the age, datum source, other age-related information, and supporting references. The 'event age' for erosion and ice loading surfaces are constrained by the age range of relatively older and younger marine deposits.

Event	Event age	Location	Age source	Other information	Age	Ref.
M17+	< 3000 yr BP	Modern Marine	14C	SMO-1	< 3000 yr BP	31
L10	~3000–1500 yr BP	Site 740	14C		~3000-1500 yr BP	29
"		Vestfold Hills		Chelnok Glaciation (lateral ice)		30
M17+	< 11 930 yr BP	Site 740	14C	SMO-1	~5485 yr BP	29
"		Site 740	14C	SMO-1	~11 140 to 9480 yr BP	29
"		ODP Site 741	14C	SMO-1	10 600 cal. yr BP*	29
"		MacRob. Shelf	14C	SMO-1	> 3100 yr BP*	28
"		AM02	14C		6548 ± 60 yr BP	23
"		MacRob. Shelf	14C	SMO-1	> 8600 yr BP*	28
"		MacRob. Shelf	14C	SMO-1	> 10 700 cal. yr BP*	27
"		core 901/GC29	14C	SMO-1	11 650 cal. yr BP*	26
"		core 901/GC24	14C	SMO-1	11 930 cal. yr BP*	25
T		Larsemann Hills	lichen 14C	Ice free during LGM	LGM	24
E9	22 170–11 930 yr BP	Inner shelf/ periphery Prydz Channel		Erosion/moraine deposition	LGM associated	25,29
M16	22 170–21 680 yr BP	AM02	14C	SMO-2	21 680 ± 160 yr BP	23
"		KC29B	stratigraphy	SMO-2		22
"		186/GC22	14C	SMO-2	~22 170 yr BP	21
E8+	1.13–0.02217 Ma	901/GC22		unconformity		21
M15	> 33 590 yr BP	901/GC22		SMO-3		21
E/L?		Prydz Channel		terrigenous glacial influence		32
M14		Prydz Channel	14C	SMO-4	infinite 14C	32
E8/L8+	<1.13 Ma	Site 739		ice loading (event IV)		10
"		Site 1167		multiple debris flows		13
M13+	1.13–0.78 Ma	Site 1167	nanno., palaeomag.		0.9–0.78 Ma	in 19
"		Site 1167		multiple fine grained deposits	1.13–0.78 Ma	13
M12–13		Vestfold Hills	TL, amino acids	Davis interglacial	> 1or > 0.3 Ma?	20
M12	~1.13 Ma	Site 1167	Sr dating of forams		~1.13	in 19
M11–12	2.0–0.9 Ma	Site 1167	nanno.		2– 0.9 Ma	13
E7	< 2.6 Ma	Amery Oasis	diatom	erosion of Bardin Bluffs Fm	< 2.6 Ma	17
E6	2.6–0.9 Ma	Amery Oasis	rew. diatom	within Bardin Bluffs Fm	2.6–0.9 Ma	4
M11	2.5–1.8 Ma	Site 1166	diatom		2.5/2.1–2.0/1.8 Ma	15
"		Site 742	diatom		2.5–1.8 Ma	18
M10–11	2.6–1.8/0.99 Ma	Amery Oasis	diatom/foram	Bardin Bluffs Fm	2.6–1.8/0.99 Ma	17
M10	3.2–2.5 Ma	Site 751	diatom		~3.2–3.0 Ma	16
"		Site 1166	diatom		3.2/2.7–2.7/2.5	15

* = reservoir corrected age, TL = thermoluminescence

References

1. Truswell & Macphail (2001)
2. Quilty *et al.* (1999)
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6. Barron & Mahood (1993)
7. Whitehead *et al*. (2003)
8. Baldauf & Barron (1991)
9. Cochrane & Cooper (1991)
10. Solheim *et al.* (1991)
11. Harwood *et al*. (2000)
12. Mizukoshi *et al*. (1986)
13. O'Brien *et al.* (2004)
14. McMinn & Harwood (1995)
15. Whitehead & Bohaty (2003)
16. Bohaty & Harwood (1998)
17. Whitehead & McKelvey (2001)
18. Mahood & Barron (1996)
19. Cooper & O'Brien (2004)
20. Hirvas *et al*. (1993)
21. Domack *et al.* (1998)
22. Taylor & Leventer (2003)
23. Hemer & Harris (2003)
24. Burgess *et al.* (1994)
25. O'Brien & Harris (1996)
26. Taylor & McMinn (2002)
27. Taylor & McMinn (2001)
28. Rathburn *et al*. (1997)
29. Domack *et al*. (1991)
30. Adamson & Pickard (1986a)
31. e.g. Taylor *et al.* (1997)
32. Leventer *et al.* (2004)

The ages of the diverse Cenozoic events have been largely constrained through biostratigraphy for Palaeocene to Middle Pleistocene features, and by ^{14}C dating for those of the Late Pleistocene to Holocene. The resolution of these two dating techniques is vastly different, and thus it is possible to identify more individual events from the younger record using ^{14}C dating. Due to the comparatively lower dating resolution obtained from biostratigraphy the events of similar age documented in this review are only broadly correlatable. In reality, depositional and erosional events alternate at a much finer time resolution (e.g. Milankovitch frequencies during glacial and interglacial cycles (Naish *et al.* 2001)) than biostratigraphy can resolve. The Cenozoic events identified here constitute the least number of marine incursions and glacial advances evident from the Lambert Graben–Prydz Bay Cenozoic record. Both the temporal and spatial extent of the stratigraphic record is summarized in Fig. 4, illustrating past variations in

Table I. (continued) Events (T = terrestrial, M = marine, E = erosion, L = ice loading) are listed in chronological order. The symbol '+' denotes multiple events, which cannot be dated separately due to low dating resolution. The 'event age' is constrained by the oldest and youngest age range evident for that event. Evidence includes location, the age, datum source, other age-related information, and supporting references. The 'event age' for erosion and ice loading surfaces are constrained by the age range of relatively older and younger marine deposits.

Event	Event age	Location	Age source	Other information	Age	Ref.
L5	4.1–2.5 Ma	Site 742	diatom	prior M11 deposition	>1.8	10
"		Site 739	diatom	ice loading M8-9 (event III)	<4.1	10
M9?	Pliocene ?	Site 739	diatom			8
M8	4.1–3.6 Ma	Site 739	diatom	Zone NSOD-14-15	3.9–3.6 Ma	8
M7–8	4.9–3.7 Ma	Site 739	diatom	Zone NSOD12	4.9–3.7 Ma	8
"		Amery Oasis	rew. diatom	Bardin Bluffs Fm	5.8–3.6 Ma	4
M7	4.2–4.1 Ma	Vestfold Hills	diatom	Sørsdal Fm	4.2–4.1 Ma	11
"		Larsemann Hills	rew. diatom		4.9–4.1 Ma	14
E4+	6.3–4.1 Ma	Prydz Bay		PP-12 seismic reflector		13
"		Prydz Bay		A seismic reflector		12
"		Site 1165		not ice unconformity	6.4–5.9 Ma	5
"		Fisher Massif	diatom	erosion of Fisher Bench Fm	< 10.7 Ma	from 4
"		Amery Oasis	diatom	B/n Battye Gl. & Bardin Bluffs Fm	10.7–2.6 Ma	from 4
"		Vestfold Hills	diatom	beneath Sørsdal Fm	> 4.1 Ma	11, 4
"		Site 739	diatom	ice loading M4-6 (event II)	< 6.3 Ma	10
M6	10.7–8.5 Ma	Fisher Massif	diatom	Fisher Bench Fm	12.1/10.7–8.5 Ma	4
"		Amery Oasis	diatom	Battye Gl Fm	10.7–9.0 Ma	7
E3+	10.7–9.0 Ma	Amery Oasis	diatom	associated Battye Glacier Fm		7
M5	12.1–11.5 Ma	Amery Oasis	rew. diatom	Bardin Bluffs Fm	12.1–11.5 Ma	4
M4-6	14.2–6.3/6.2 Ma	Site 739	diatom		14.2–6.2/6.3 Ma	8
M4	14.5–12.5/12.1 Ma	Fisher Massif	(rew.?) diatom	Fisher Bench Fm	14.5–12.5/12.1Ma	4
E2+	37–9.0 Ma	Prydz Bay		R1 seismic reflector		9
"		Site 739		b/n M3 & M4-6	35.9–6.2 Ma	8
"		Amery Oasis	diatom	Amery Erosion surface	>9.0 Ma	7
"		Fisher Massif	diatom	beneath Fisher Bench Fm	>8.5 Ma	4
M3	37–30 Ma	Site 739	diatom		35.9–34.8 Ma	6
"		Site 1165	diatom		37–33 Ma	5
"		Amery Oasis	rew. diatom	Bardin Bluffs Fm	>30 Ma	4
"		Fisher Massif	(rew.?) diatom	Mt Johnston Fm	35 – 33 Ma	4
E1	L.Eocene–E.Oligo.					3
T1	M. Eocene	Site 1166	palynology			1
M2	M. Eocene	MacRob. Shelf	rew. foram			2
M1	L. Paleocene	MacRob. Shelf	rew. dinoflagellate	Zone P4		2
Tc	L. Cretaceous	Site 742, 1166	palynology		Turonian-?Santonian	1
Tb	E. Cretaceous	Site 741	palynology		Albian	1
Ta	Jurrasic-E. Cret.	MacRob. Shelf	palynology			1

* = reservoir corrected age, TL = thermoluminescence

References

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7. Whitehead *et al.* (2003)
8. Baldauf & Barron (1991)
9. Cochrane & Cooper (1991)
10. Solheim *et al.* (1991)
11. Harwood *et al.* (2000)
12. Mizukoshi *et al.* (1986)
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14. McMinn & Harwood (1995)
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18. Mahood & Barron (1996)
19. Cooper & O'Brien (2004)
20. Hirvas *et al.* (1993)
21. Domack *et al.* (1998)
22. Taylor & Leventer (2003)
23. Hemer & Harris (2003)
24. Burgess *et al.* (1994)
25. O'Brien & Harris (1996)
26. Taylor & McMinn (2002)
27. Taylor & McMinn (2001)
28. Rathburn *et al.* (1997)
29. Domack *et al.* (1991)
30. Adamson & Pickard (1986a)
31. e.g. Taylor *et al.* (1997)
32. Leventer *et al.* (2004)

glacial conditions within East Antarctica's largest catchment area.

Paleocene marine deposition (M1)

On Mac. Robertson Shelf recycled Late Palaeocene (Zone P4) marine fossils (foraminifera and dinoflagellates) are recorded by Quilty *et al.* (1999).

Middle Eocene terrestrial (T1) and marine deposition (M2)

On the Mac. Robertson Shelf recycled Middle Eocene foraminifera, pollen and spores have been recorded by Quilty *et al.* (1999). Middle Eocene terrestrial pollen and spores and marine dinocysts have also been identified in Prydz Bay, from Site 1166 (Truswell & Macphail 2001, Macphail & Truswell 2004) and further study is currently underway.

Late Eocene–Early Oligocene glacial onset and subaerial exposure (E1)

Glaciation in Prydz Bay probably commenced at ~40 Ma, when Antarctica changed from being essentially ice-free to having a large ice sheet (that may have been comparable to Pleistocene ice volumes). The onset of continent-wide glaciation is evidenced by deep-sea oxygen isotopes and Mg/Ca ratios (Billups & Schrag 2003). A package of sediment identified from Prydz Bay seismic data, PS.2A (Cooper *et al.* 1991b) (Fig. 3), may contain the 'transitional zone' between pre-glacial and glacial conditions in Prydz Bay (O'Brien *et al.* 2001a). The zone, drilled during ODP Leg 188, displays a facies change from glaciofluvial deltaic sands (Strand *et al.* 2003) to glaciomarine sediment (M3), and is interpreted as a Late Eocene–Early Oligocene marine transgression (O'Brien *et al.* 2001a). Facies change within the unit has been subdivided into PS.2A2 and PS.2A1, with an intervening unconformity (E1) (Erohina *et al.* 2004). The E1 erosion surface is not considered glacial, but rather records subaerial exposure during a eustatic sea-level low (Erohina *et al.* 2004).

Late Eocene–Early Oligocene glacial retreat and marine deposition (M3)

A significant Antarctic iceberg calving event occurred at ~36 Ma, depositing ice rafted debris (IRD) on the Kerguelen Plateau, ~1200 km north of Prydz Bay (Breza & Wise 1992). IRD deposition did not occur here again until the Late Miocene (Breza 1992).

It is as yet unclear how the Eocene IRD event is related to marine deposition that occurred at a similar time in Prydz Bay and the Lambert Graben (M3). This marine event is evidenced by *in situ* diatom-bearing marine deposits dated between 35.9–34.8 Ma occurring at Site 739 (Baldauf & Barron 1991, Barron & Mahood 1993), and dating between 37–33 Ma at Site 1166 (Florindo *et al.* 2003). Glacially reworked diatoms > 30 Ma occur in the Bardin Bluffs Formation, and *in situ* or glacially reworked diatoms, 35–33 Ma, occur in the Mount Johnston Formation (Whitehead *et al.* 2004). These indicate that marine conditions occurred at least as far inland as Fisher Massif in the Lambert Graben. This marine incursion broadly correlates with glacial-marine deposition in Prydz Bay (within PS.2A1 of Erohina *et al.* (2004)), and also perhaps with the Mac. Robertson Shelf recycled Late Eocene–Early Oligocene microfossils recorded by Quilty *et al.* (1999).

Post-Late Eocene–Early Oligocene progradation of the continental shelf edge (PSE-1 to 3) and glacial erosion ($E2^+$)

Initial glaciation of the continental shelf compacted underlying M3 marine sediments, but neither reached the palaeo-shelf break, nor caused extensive shelf-edge progradation (Leitchenkov *et al.* 1994, Cooper & O'Brien 2004). Subsequent glacial erosion has apparently removed the Late-Eocene–Early Oligocene to Middle Miocene record from much of the Lambert Graben and inner Prydz Bay. This produced the erosion surface $E2^+$ on the continental shelf (seismic reflector R1 of Cochrane & Cooper (1991)), that is the record of multiple glacial advances (Fig. 3).

After the Late Eocene or Early Oligocene, glacial erosion excavated deep valleys on the continental shelf of Prydz Bay during an undated relative sea level low stand (Cooper *et al.* 2001). Erosion by the Lambert Glacier out to the palaeo-shelf edge caused deposition of sediment beyond the shelf break, and a major phase of northwards continental shelf progradation prior to the Late Miocene (O'Brien *et al.* 2001b). The seismically identified succession of palaeo-shelf edges of Cooper *et al.* (1991b), and Leitchenkov *et al.* (1994), occur between progradational sequences and define intervals of glacial retreat (Fig. 3). In summary the oldest three palaeo-shelf edges are:

PSE-1; of Cooper *et al.* (1991b); OPSE of Leitchenkov *et al.* (1994). Oligocene diatom-bearing sediments at Site 739 have been compacted by ice loading during glacial advance northwards beyond *PSE-1* (Event I of Solheim *et al.* 1991).

PSE-2; of Cooper *et al.* (1991b). Prior to the formation of this shelf break, one and perhaps two identifiable sequences (Fig. 3) were deposited beyond PSE-1 (Leitchenkov *et al.* 1994, Cooper *et al.* 1991b). Deep-sea sediment oxygen isotopes and Mg/Ca ratios indicate that the Antarctic ice volume was small at 26 Ma (Billups & Shrag 2003), and this may signal when PSE-2 or one of the other palaeo-shelf breaks formed.

PSE-3; of Cooper *et al.* (1991b); PSE-1 of Leitchenkov *et al.* (1994). The sequence bounded by PSE-3 and a fourth palaeo-shelf break (PSE-4) reflects glacial variability. Drilled at ODP Site 739, this sequence includes both diamict and other marine sediment broadly dated between the Middle and Late Miocene (Baldauf & Barron 1991). Evidence for similar aged deposits and glacial unconformities occurs throughout the Lambert Graben and Prydz Bay region and is discussed below.

Middle and Late Miocene marine deposition (M4–M6) and glacial erosion (E2+–E5)

There is evidence for at least three separate marine events in the Lambert Graben and Prydz Bay through the Middle and Late Miocene (M4, M5 and M6). The M4 event is only evident as reworked diatom fossils (14.5–12.5/12.1 Ma) within the Fisher Bench Formation. Similarly, the M5 event is evidenced by reworked diatoms (12.1–11.5 Ma) in the Bardin Bluffs Formation (Whitehead *et al.* 2004). The Late Miocene Fisher Bench and Battye Glacier formations (10.7 to 8.5 Ma and 10.7 to 9.0 Ma respectively) of the Pagodroma Group constitute M6. A more precise correlation of these two formations is not yet possible. The M6 marine event may be associated with a warm climatic interval 10–8.5 Ma (centred at ~8.5 Ma) identified from deep-sea oxygen isotope and Mg/Ca ratio data (Billups & Schrag 2002). Beyond Prydz Bay, younger Late Miocene (7.5 to 7.0 Ma) strata have been recovered from the Wild Drift at ODP Site 1165 and exhibit evidence for 41 ka orbitally controlled volume oscillations in the EAIS (Grützner *et al.* 2003). At ODP Site 739, Middle and Late Miocene (14.2–6.2 to 6.3 Ma; Baldauf & Barron 1991) sediments span the interval between PSE-3 and PSE-4. However, this low-resolution age span encompasses those ages suggested for M4, M5 and M6.

Collectively, the Miocene marine deposits help constrain the age and minimum extent of glacial retreat and marine incursion into the Lambert Graben. Associated disconformities are also broadly datable and can be used to gauge the extent of glacial advance. There is evidence for at least three regional Miocene erosion surfaces ($E2^+$, $E3^+$ and $E4^+$).

Erosion surface ($E2^+$)

This widespread Middle Miocene erosion surface, probably a consequence of the contemporaneous increase in Antarctic ice volume (Billups & Schrag 2002) is superimposed upon and has obliterated numerous older surfaces cut during multiple glacial advances since the Late Eocene–Early Oligocene. The surface is manifested at:

i) at Fisher Massif, beneath the Fisher Bench Formation, (> 8.5 Ma, Whitehead *et al.* 2004),

ii) at Amery Oasis, beneath the Battye Glacier Formation (Amery Erosion Surface), (> 9.0 Ma, Whitehead *et al.* 2003),

iii) between Early Oligocene and Middle Miocene sediments at ODP Site 739 (Barron *et al.* 1991),

iv) on top of unit P2.2A, post Late Eocene–Early Oligocene in age (Erohina *et al.* 2004).

Erosion surface ($E3^+$)

Two, perhaps local, erosion surfaces ($E3^+$) are indicated within the Battye Glacier Formation at Beaver Lake from (i) reworked diatomaceous sediments near the base of the formation and (ii) an erosion surface separating two lithological members (Whitehead *et al.* 2003) and may be Late Miocene in age (10.7–9.0 Ma). More regional glacial advances that may be associated with the $E3^+$ surfaces caused the ice compaction of Late Eocene–Early Oligocene sediments at ODP Site 739 (L3) (Solheim *et al.* 1991). Other events that may be contemporaneous with $E3^+$ could have occurred during the Middle Miocene glacial expansion (15–13 Ma), which created a transition in sediment texture and colour on the continental rise Site 1165 (O'Brien *et al.* 2001a). Changing hydrological conditions also caused a brief disconformity here at ~14 Ma (Florindo *et al.* 2003), and may be contemporaneous with glacial advance.

Erosion surface ($E4^+$) and ice loading (L4)

In the Late Miocene the continental shelf of Prydz Bay began to be deepened landward (De Santis *et al.* 2001), due to multiple glacial advances. Evidence for these glacial advances includes erosion surfaces and ice loading events that occurred during the time interval 10.7–4.1 Ma, and are as follows:

i) Ice loading < 6.3 Ma of Late Miocene (6.2/6.3–4.9 Ma, Baldauf & Barron (1991)) sediments at ODP Site 739 (ice load event II of Solheim *et al.* (1991)),

ii) Erosion surfaces beneath Early Pliocene Sørsdal Formation in the Vestfold Hills that formed > 4.1 Ma (Harwood *et al.* 2000),

iii) Glacial erosion of the Fisher Bench Formation < 10.7 Ma at Fisher Massif,

iv) Glacial erosion of the Battye Glacier Formation, and creation of the sub-Bardin Bluffs Formation surface between 10.7 and 1.8/0.99 Ma in the Amery Oasis. The younger age range of this surface may also correlate with later glacial erosion surfaces discussed below.

Beyond the influence of glacial erosion, a change in hydrological conditions created a disconformity on the continental rise at ODP Site 1165 in the Late Miocene (6.4 and 5.9 Ma) (Florindo *et al.* 2003).

Late Miocene or Early Pliocene initiation of the Prydz Channel and Prydz Channel Fan (PCF)

The cutting of the $E4^+$ surface culminated in the channelized erosion of the Prydz Bay floor, creating the Prydz Channel and depositing the PCF beyond the palaeo-continental shelf break (PSE-4) (i.e. PSE-4 of Cooper *et al.* (1991b); PSE-2 of Leitchenkov *et al.* (1994)) (O'Brien *et al.* 2004). The $E4^+$ erosion surface is marked by seismic reflector 'A' of Mizukoshi *et al*. (1986); PP-12 of O'Brien *et al.* (2004)) that ties to an Early Pliocene (~4 Ma) interval at ODP Site 739 (105.9–130 mbsf) (O'Brien *et al.* 1995). Therefore the PCF and PSE-4 post date Early Pliocene (Leitchenkov *et al.* 1994, O'Brien *et al.* 1995), or at least Late Miocene, strata at Site 739 (Cooper *et al.* 2001).

Computer modelling suggests that uplift of the PCMs could have channelized the Lambert Graben ice flow (Taylor *et al.* 2004). Increasing ice flow from the south-east into Prydz Bay may also have contributed to the cutting of the Prydz Channel and caused a westerly deflection of the (then) Lambert Glacier, which influenced the position of the PCF (O'Brien *et al.* 2001a). Deposition of the PCF changed the geometry of the continental shelf from a linear to convex prograding edge (O'Brien *et al.* 2001a) and shifted the depocentre towards the continental shelf break, away from the continental rise (Florindo *et al.* 2003).

Early Pliocene marine deposition (M7–M8)

In situ and glacially reworked fossils identify at least two Early Pliocene marine events (M7 and M8) in the Lambert Graben and Prydz Bay. The broad age control on some sequences however, prevents their assignment specifically to either M7 or M8. Examples include:

i) M7–M8 glacially reworked Pliocene fossils dating between 5.8–3.6 Ma, within the Bardin Bluffs Formation at the Amery Oasis (Whitehead *et al*. 2004),

ii) M7 reworked marine fossils in the Larsemann Hills (Quilty *et al*. 1990, McMinn & Harwood 1995) between 4.9–4.1 Ma in age,

iii) The M7 *in situ* marine Sørsdal Formation of the Vestfold Hills (Harwood *et al.* 2000), with a revised age of between 4.2–4.1 Ma (Whitehead *et al.* 2004),

iv) At ODP Site 739 (Baldauf & Barron 1991) three *in situ* Pliocene horizons are present. The older two beds accumulated within the M7–M8 Pliocene intervals. The younger bed, M9, is of uncertain Pliocene age. The diatom biostratigraphic evidence for M7, M8 and M9 at ODP Site 739 is as follows:

- M7 was deposited between 4.9–3.7 Ma from the revised age of bed NSOD 12 (in Baldauf & Barron (1991)), based on the presence of the diatom *Fragilariopsis praeinterfrigidaria* (FO 4.9 Ma and LO 3.7 Ma).
- M8 was deposited between 3.9–3.6 Ma from the revised age of bed NSOD 14–15 (in Baldauf & Barron (1991)), based on the presence of the diatoms *Thalassiosira lentiginosa* (FO 4.9 Ma) and *Fragilariopsis praecurta* (LO 3.6 Ma).
- M9 is a possible Pliocene bed (the Pliocene diatom *F. praeinterfrigidaria* is present, but this may be reworked and thus the M9 event could be younger than this diatom).

v) A largely complete and *in situ* Early Pliocene marine deposit also occurs on the continental rise at ODP Site 1165 (Florindo *et al.* 2003). Silicoflagellate fossils indicate that three warm water events occurred here at 4.8–4.5 Ma, ~4.3 Ma and ~3.7 Ma (Whitehead & Bohaty 2003) and a diatom proxy indicates there was less sea-ice than today (Whitehead *et al*. 2005).

Pliocene glacial advance (E5) and ice loading (L5)

There is evidence for glacial advance at < 4.1 Ma resulting in the compaction of the Early Pliocene sediments at ODP Site 739 (event III of Solheim *et al.* (1991)), and the cutting of the undulating unconformity surface E5, visible in seismic data from across the shelf (O'Brien & Harris 1996). This ice loading occurred before Late Pliocene–Pleistocene (> 1.8 Ma) marine deposition (Mahood & Barron 1996) at ODP Site 742 (Solheim *et al.* 1991).

Late Pliocene–Early Pleistocene marine deposition (M10–M11)

At least two Late Pliocene marine events (M10 and M11) are recorded in the Lambert Graben and Prydz Bay region. The oldest, M10, occurs at ODP Site 1166 and is between 3.2/2.7–2.7/2.5 Ma in age (Whitehead & Bohaty 2002). A terrigenous bed of glacially derived sediment, and possible older unconformity, separates M10 from M11 at Site 1166 (Florindo *et al.* 2003). The M11 marine event at Site 1166 was deposited 2.5/2.1–2.0/1.8 Ma (*Thalassiosira kolbei* diatom zone of Harwood & Maruyama (1992)) (Whitehead & Bohaty 2002). This may correlate with a similar-aged bed at Site 742 (2.5–1.8 Ma) (Mahood & Barron (1996) in Florindo *et al.* (2003)). Either of the M10 or M11 marine events in Prydz Bay may correlate with the Bardin Bluffs Formation in the Amery Oasis 2.6–1.8/0.99 Ma (Whitehead & McKelvey 2001).

Late Pliocene–Early Pleistocene glacial erosion (E6–E7)

Glacial advance (E6) < 2.6 Ma resulted in the entrainment of older (5.8–3.6 Ma) fossils in the Bardin Bluffs Formation caused soft sediment deformation and created a disconformity between the Upper and Lower members of the formation (Whitehead & McKelvey 2001, Whitehead *et al.* 2004). Subsequent glacial advances (E7) caused erosion of both members of the Bardin Bluffs Formation (Whitehead & McKelvey 2001). However, minimum ^{10}Be rock exposure ages from bedrock in the Amery Oasis mountain peaks (above 1200 m elevation) indicate that they have remained ice-free for much of the last ~2.3 ± 0.3 Ma, and similarly for Fisher Massif (above 1260 m elevation) for the last 1.8 ± 0.2 Ma (Fink *et al.* 2000).

Early–Middle Pleistocene marine deposition (M11–M13$^+$)

At the PCF multiple open marine depositional events occurred, between intervals of glacial advance to the continental shelf edge, during the Early–Middle Pleistocene. During one marine event, locally called the Davis Interglacial (Hirvas *et al.* 1993), the ice sheet retreated to the current Antarctic coastline of the Vestfold Hills, and subsequent to the middle Pleistocene there was a declining frequency of glaciation to the shelf edge. The following evidence describes this history in greater detail:

Prydz Channel Fan

At ODP Site 1167 the upper 447.5 mbsf of the PCF was drilled with 40% recovery (O'Brien *et al*. 2004). The logged upper 260 mbsf revealed large debris flows interbedded with 16 minor fine-grained beds. The debris flows record when the glacial ice front was at the continental shelf break, causing slope instability, whilst the interbeds are open marine sediments. The section between 217–32 mbsf at Site 1167 spans the interval of ca. 1.13 to 0.78 Ma (O'Brien *et al.* 2004) and includes eleven discrete marine events collectively constituting M13^{+}. The following age data (Cooper & O'Brien 2004) are pertinent:

i) Calcareous nannofossils from the interval between 228–218 mbsf were deposited between 2–0.9 Ma (either M11 or M12),

ii) Sr dating of *Neogloboquadrina pachyderma* from ~217 mbsf records an event at ~1.13 Ma (M12),

iii) At 37.4 mbsf calcareous nannofossils date between 0.9–0.2 Ma and are further constrained to between 0.9–0.78 Ma through magnetostratigraphy,

iv) At 32 mbsf a polarity reversal marks the Bruhnes–Matuyama boundary at 0.78 Ma,

v) Above 5.25 mbsf diatoms of the *T. lentiginosa* Zone indicate an age of < 0.66 Ma,

vi) The uppermost 0.45 mbsf is indicated by electron spin resonance to be < 36.9 ± 3.3 ka (Theissen *et al.* 2003).

Vestfold Hills 'Davis Interglacial' marine deposition (M13^{+})

A few deci-centimetres of marine sediment, bounded by diamicts (Gore, personal communication 2005), were deposited in the Vestfold Hills during the relatively warm 'Davis interglacial', currently dated at 1.0–0.3 Ma by thermoluminescence and amino acid techniques (Hirvas *et al.* 1993). This age is currently being re-evaluated (Quilty *et al.* unpublished data).

Pleistocene glacial erosion (E8^{+}) and ice loading (L8^{+})

Glacial advance extended to ODP Site 739 and caused the sediment compaction we here identify as L8^{+} and equate with loading event IV of Solheim *et al.* (1991). This event may also correspond with multiple glacial advances to the continental shelf edge between 1.13–0.78 Ma, which deposited debris flows within the interval 217–32 mbsf of the PCF at Site 1167 (O'Brien *et al.* 2004). Three younger glacial events deposited similar debris in the upper 32 m of the PCF since < 0.78 Ma (O'Brien *et al.* 2004). These advances occurred prior to, and reached beyond the Last Glacial Maximum extent (Domack *et al.* 1998). However, they represent a reduction in glacigenic sedimentation out to the continental shelf break that has been related to a number of factors, such as progressive over-deepening of the inner continental shelf, reduction in ice volume due to climatic cooling and an altered interaction between Milankovich cycles and ice residence time (O'Brien *et al.* 2004). During or subsequent to the E8+/L8+ glacial advances in the Amery Oasis, the elevation of the EAIS has not been sufficient to override the upper surface of the Battye Glacier Formation (~800 m elevation), where minimum ^{10}Be rock exposure ages indicate this area has largely remained ice free since 0.95 ± 0.09 Ma or 0.86 ± 0.1 Ma (Fink *et al.* 2000).

Pleistocene marine deposition (M14–16)

Short cores (< 10 m) of Quaternary diamict (subglacial), silty clay and granulated facies (sub-ice shelf) separated by subordinate diatomaceous muds and oozes (open marine or sub-ice shelf advected marine sediments) have been recovered from Prydz Bay (e.g. SMO-1, -2, -3 and -4 of Domack *et al.* (1998), Taylor & Leventer (2003), and Leventer *et al.* (2004)). SMO-2, -3 and -4 we here identify as depositional events M16, M15 and M14. An infinite ^{14}C dating age was obtained for SMO-4 (M14) in the Prydz Channel (Leventer *et al.* 2004), implying a real age >50 ka (beyond the limit of the dating method (Bowman 1990)) or incorporation of older reworked organic material. Infinite dates were also obtained for SMO-3 and -2 (Leventer *et al.* 2004). Other ^{14}C dating indicates SMO-3 (M15) to predate 33 590 yr BP and SMO-2 (M16) to approximate to 22 170 yr BP (uncorrected for reservoir effects) (Domack *et al.* 1998). Discrepancies in the SMO-2 ages (i.e. 22 170 yr BP or an infinite ^{14}C dating ages) need future clarification, as these dates may represent distinctly different aged units. The stratigraphical relationship between SMO-3 (M15), SMO-4 (M14) and Early Pleistocene marine events (M11–M13^{+}) also need future clarification.

An unconformity (E8) occurs between SMO-3 (M15) and SMO-2 (M16) in Prydz Bay gravity core 901/GC22 (Domack *et al.* 1998). These beds, and overlying terrigenous sediment, contain reworked Pliocene (4.9–1.9 Ma) and Late Miocene (9.0–8.5/6.4 Ma) diatoms (Domack *et al.* 1998, Taylor & Leventer 2003), spanning the age of the M6 to M11 marine events. Neogene sediment eroded from inner Prydz Bay, including deposits similar to the Sørsdal Formation, provided a likely source for these reworked diatoms. It is unlikely that they were eroded from the Pagodroma Group because there is little evidence for Late Quaternary ice sheets over-riding Pliocene and Miocene deposits in the PCMs (Mabin 1991, Adamson *et al.* 1997).

SMO-3 (M15) and SMO-2 (M16) have previously been interpreted as sub-ice shelf deposits in the Prydz Channel (Domack *et al.* 1998), whilst later work indicated that the SMO-2 diatoms constitute an open marine flora (Taylor & Leventer 2003). New geochemical data from SMO-2 may reconfirm the initial sub-ice shelf interpretation (Leventer, personal communication 2005). Core AM02 from beneath the Amery Ice Shelf (~80 km landward of today's ice edge) may equate with SMO-2, with diatom deposition beginning 21 680 ± 160 yr BP and continuing to at least 6548 ± 60 yr BP at the core top (uncorrected ^{14}C dates of Hemer & Harris (2003)). Alternatively, this deposit has been interpreted as post-LGM, and the ~21 680 yr BP date held to be an artefact of bioturbation or recycling of older organic or other carbonaceous matter (Hemer & Harris 2003). However we consider the age authentic and interpret the post-LGM age (6548 ± 60 a BP (uncorrected)) of the core top a consequence of subsequent sediment starvation beneath the Amery Ice Shelf or a loss of surface sediment layers that can occur with the coring method used.

Last Glacial Maximum glacial erosion (E9)

The LGM is recognized globally as the peak of the last climatic cooling and glacial ice advance, which started and ended abruptly 24.0–18.0 ka (North Atlantic dates) (Bard 1999), whilst in Antarctica associated features of younger age represent last 'local glacial maximum'. During the LGM, the Amery Ice Shelf grounded half way across Prydz Bay (Fig. 2) (Harris *et al.* 1997, Domack *et al.* 1998). Elsewhere along the nearby coast the ice sheet advanced little, but may have covered the southern Vestfold Hills (Fabel *et al.* 1997). The Larsemann Hills remained largely ice-free through out the LGM (Burgess *et al.* 1994, Hodgson *et al.* 2001).

LGM ice in Prydz Bay did not solely come from the Lambert Glacier, but included significant inputs from the adjacent Ingrid and Lars Christensen Coasts (O'Brien & Harris 1996). Glacial erosion occurred within the inner shelf of the Amery Depression (O'Brien & Harris 1996), and deposited subglacial sediments along the periphery of Prydz Channel (O'Brien *et al.* 1999), whilst the area directly above the Prydz Channel was covered by an ice-shelf (Domack *et al.* 1998). Sub-ice shelf related intervals have been illustrated in Fig. 4, whilst a more detailed stratigraphical analysis of the various sedimentary facies deposited in Prydz Bay during the Late Pleistocene can be found in Domack *et al.* (1998).

Glacial moraines throughout the PCMs illustrate past thicknesses and profiles of the Cenozoic EAIS, in particular adjacent to the Lambert Glacier and its tributaries (Derbyshire & Peterson 1978, Mabin 1991). Ice volume data and exposure age dating of these moraines will enable past ice occupancy of the Lambert Graben and Prydz Channel to be more accurately determined (White, unpublished data). In the southern PCMs, moraines associated with the LGM on the Menzies Range occur 250 m above modern ice levels (Derbyshire & Peterson 1978, Whitehead & McKelvey 2002). The alpine Battye Glacier within the Amery Oasis had begun retreating from the northern edges of Radok lake subsequent to the LGM, at ~10 000 yr BP (minimum ^{10}Be moraine exposure age) (Fink *et al.* 2000).

Table II. Eustatic sea-level low stands, during Antarctic glacial advances, created sub-aerial erosion surfaces on Australia and most likely similar aged glacial erosion surfaces in Antarctica. Possibly eustatic sea-level low stands are evident from hiatuses (H) within the Great Australian Bight (Li et al. 2004). These can be broadly related to glacial erosion and ice loading events within the Prydz Bay-Lambert Graben region. Brackets denote one or more Australian hiatuses that may relate to one or more Antarctic erosion or ice loading events. Those hiatuses within the age range of Antarctic marine events, may either pre-or post-date the marine event.

Event	Event Age	Event Li *et al*. (2004)	Age
M17+	< 3000 y rBP		
L10	~3000–1500 yr BP		
M17+	< 11930 yr BP		
E9	22 170–11 930 yrs BP		
M16	22 170–21 680 yrs BP		
M15	> 33 590 yrs BP		
M14			
E8/L8+	1.13–0.02217 Ma		
M13+	1.13–0.78 Ma		
M12	~1.13 Ma		
		Hiatus 15	1.5
E7	< 2.6 Ma		
E6	2.6–0.9 Ma		
M11-12	2.0–0.9 Ma		
		Hiatus 14	2.5
M10	3.2–2.5 Ma		
L5	4.1–2.5 Ma		
		Hiatus 13	3.5
M9	possible Pliocene		
M8	4.1–3.6 Ma		
M7	4.2–4.1 Ma		
		Hiatus 12	4.5
		Hiatus 11	6
		Hiatus 10	7
E4+	10.7–4.1 Ma		
M6	10.7–8.5 Ma	Hiatus 9	9.3
E3+	10.7–9.0 Ma		
M5	12.1–11.5 Ma	Hiatus 8	11.5
M4	14.5–12.5/12.1 Ma	Hiatus 7	13.5
		Hiatus 6	14.8
		Hiatus 5	16.4
		Hiatus 4	18.7
		Hiatus 3	20.5
		Hiatus 2	22.3
		Hiatus 1	23.8
E2+	37–12.1 Ma		
M3	37–30 Ma		
E1	L.Eocene–E.Oligo.		
T1	M. Eocene		
M2	M. Eocene		
M1	L. Paleocene		

Late Pleistocene–Holocene marine deposition (M17$^+$) and minor glacial advance (L10)

The onset of open-marine conditions within Prydz Bay and deposition of siliceous mud and ooze of SMO-1 (M17) indicate the retreat of the Lambert Glacier following the LGM. The sudden change from glacigene erosion to marine deposition (SMO-1) in inner Prydz Bay suggests that LGM ice retreat was rapid (O'Brien & Harris 1996). This occurred due to ice decoupling, lifting the grounded ice off the sea-floor (O'Brien & Harris 1996) as global sea-level rose. If ice had occupied inner Prydz Bay (including the Amery Depression) during the LGM, decoupling would have been further accentuated by the inland sloping topography of Prydz Bay and the Lambert Graben, which promoted retreat of the grounding zone far inland (Fig. 2). ^{14}C dating (corrected for reservoir effects) establish the initial age of SMO-1 (M17) deposition across Prydz Bay at:

i) 10 600 calibrated (cal.) yr BP (ODP Site 740) (Domack *et al.* 1991),

ii) > 10 700 cal. yr BP (cores GC1 and GC2) (Taylor & McMinn 2001),

iii) 11 650 cal. yr BP (core GC29) (Taylor & McMinn 2002),

iv) 11 930 cal. yr BP (core 901/GC24) (12 680 ± 110 yr BP (uncorrected)) (O'Brien & Harris 1996).

Marine sediments exposed in the Vestfold Hills 12 000–13 000 yr BP reflect their emergence from beneath the retreating EAIS (Bird *et al.* 1991, Fabel *et al.* 1997). Glacial retreat on the Amery Oasis also enabled the onset of organic lake deposition 12 400 cal. yr BP (Wagner *et al.* 2004). The deposition of SMO-1 is interrupted at ODP Site 740 by an interval of terrigenous deposition thought to represent a brief Amery Ice Shelf advance (Domack *et al.* 1991). This is considered to have been synchronous with a minor advance in the Sørsdal Glacier across the Vestfold Hills at *c.* 3000–1500 yr BP (Domack *et al.* 1991), termed the Chelnok Glaciation (Adamson & Pickard 1986a) (L10). Following this event, open marine deposition resumed at Site 740 and persists to the present day. Five different modern sediment facies are currently being deposited throughout Prydz Bay and have been largely correlated to the geomorphology of the sea-floor (Harris *et al.* 1998). In general biogenic siliceous mud and ooze is concentrated in deep basins, such as the Amery Depression, due to water currents. More terrigenous facies occur at shallower depths where they are often reworked by icebergs. Calcareous gravels occur in the shallowest areas and are strongly influenced by bottom water currents.

Comparison with Australian eustatic sea-level record

A comprehensive record of sedimentation and hiatuses spanning the last 25 m.y. has been documented from the Great Australian Bight, offshore from the southern Australian margin (Li *et al.* 2004). If the hiatuses there are related to low eustatic sea-levels, it is expected that similar aged hiatuses should be observed from near the East Antarctic margin, due to glacial erosion during corresponding ice sheet advances. Both the Lambert Graben–Prydz Bay and southern Australia are considered to have had little tectonic activity through the last 25 m.y. Table II compares Li *et al.* (2004) hiatus history with the record in our study. It is clear that the level of dating resolution in the Prydz Bay–Lambert Graben region is lower than that of southern Australia, however, some similarities have emerged:

1) E2$^+$ may broadly correlate with hiatuses 1 to 6 and maybe 7,

2) M4 pre- or postdates hiatus 7. If hiatus 7 predates M4, it may correlate with E2$^+$,

3) M5 pre- or postdates hiatus 8. If hiatus 8 postdates M5, it may correlate with E3$^+$,

4) M6 pre- or postdates hiatus 9, thus hiatus 9 may correlate with E3$^+$ or E4$^+$, respectively,

5) E4$^+$ may broadly correlate with hiatuses10 to 12 and maybe hiatus 9,

6) L6 may broadly correlate with hiatus 13,

7) A glacial advance may separate M10 and M11, which would correlate with hiatus 14,

8) E6 or E7 may broadly correlate with hiatus 15.

Summary

Sediments and landforms in the Lambert Graben and Prydz Bay region record a significant East Antarctic glacial history. The history of Prydz Bay has been synthesized with the integration of seismic, sedimentologic and chronostratigraphic information following Ocean Drilling Program Leg 119 (Cooper *et al.* 1991a, Hambrey *et al.* 1991, 1994, Ehrmann *et al.* 1992, Leitchenkov *et al.* 1994). New information has also been obtained from Ocean Drilling Program Leg 188 in Prydz Bay. However, Prydz Bay is only the seaward edge of the larger Lambert Graben catchment, such that the Cenozoic strata from Prydz Bay must at least in part correlate with Pagodroma Group outcrops along the Lambert Graben margin (McKelvey *et al.* 2001). The current synthesis integrates these data sets and other Cenozoic information (including Quaternary records) from Mac. Robertson Shelf, the Larsemann Hills and Vestfold Hills.

The Prydz Bay–Lambert Graben region provides, in part, a temporal and spatial history of East Antarctic Ice Sheet advance and retreat. The Cenozoic geological record spans discrete intervals of time from pre-glaciation through to today. There is evidence for at least 17 intervals of marine deposition and 10 intervals of glacial erosion and sediment compaction. The true history is inevitably more complex, but the geological record is incomplete due to glacial erosion. Coarse dating resolution for most deposits has also caused them to be 'lumped' as evidence for similar aged events, when they may actually be different (Table I). An example of this may occur in the Battye Glacier Formation, which contains three distinctive diatom deposits (all dated between 10.7–9.0 Ma and called M6). Thus any similar labelled events may not actually correlate, but share a similar age range.

Significant Cenozoic changes within the Lambert Graben–Prydz Bay region can be summarized as follows. Pre-glacial Prydz Bay was a terrestrial forested environment, with the first evidence for glaciation occurring as glacial outwash sands on the continental shelf of possible Late Eocene age. Transgressive marine conditions inundated Prydz Bay and much of the Lambert Graben in the Late Eocene–Early Oligocene. The remaining Cenozoic record consists of marine deposits interspersed between horizons or erosion surfaces that provide evidence for glacial advance across the continental shelf. Extreme glacial advances caused the progradation of a linear continental shelf edge; however, this changed to a convex edge at an interval between the Late Miocene and Early Pliocene. This change created many modern bathymetric features associated with deepening of the inner continental shelf (Amery Depression and Prydz Channel) and the deposition of the PCF. Prior to the Late Pliocene–Early Pleistocene the Lambert Graben–Prydz Bay had marine glaciers that lacked ice shelves, but had an ice cliff terminus, and released much terrigenous sediment into the marine environment. A transition to the current ice shelf terminus, low terrigenous depositional environment occurred during or after the Late Pliocene. In the Middle Pleistocene there was a decline in frequency of ice advance to the continental shelf edge. During the LGM the Lambert Glacier advanced partway across the continental shelf, but did not reach the shelf edge.

The stratigraphic overview has identified several unresolved questions about the timing and significance of particular geological events, such as:

i) How many continental shelf progradational sequences and palaeo-shelf edges are there, and how old are these? In particular, when did PCF deposition commence?

ii) How does the Middle Miocene transition in sediment texture and colour on the continental rise (ODP Site 1165) relate to changing deposition on the continental shelf?

iii) How does the gradual Neogene decrease in deposition rate at Site 1165 relate to changing glacier-thermal regime?

iv) How do discrete IRD events, detected on the Kerguelen Plateau, relate to EAIS glacial advance and retreat?

Further study of sediments within Prydz Bay and on the PCMs may help address these questions, as will improved dating chronology (e.g. revised biostratigraphies and cosmogenic and amino-acid dating calibration). Future study of Pagodroma Group correlatives in the PCMs may identify further evidence for episodes of glacial retreat from potential fossil bearing deposits on mounts Stinear, Rymill, Lanyon, Meredith and Clemence Massif. The drilling of the progradational sequences at the shelf-break may help reconstruct the Early Oligocene to the Middle Miocene history of this region. Drilling of Mac. Robertson Shelf may recover *in situ* Cenozoic sediments that pre-date Antarctic glaciation.

Acknowledgements

This paper was written through support from ARC Discovery Grant DP0345161. We thank John Gibson, Damian Gore, Amy Leventer and Eugene Domack for reading and critically commenting on earlier versions of this manuscript. P.E. O'Brien publishes with permission of the Chief Executive Officer, Geoscience Australia.

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