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The Pagodroma Group – a Cenozoic record of the East Antarctic ice sheet in the northern Prince Charles Mountains

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Abstract: The northern Prince Charles Mountains overlook the western side of the 700 km long Lambert Glacier–Amery Ice Shelf drainage system. Within these mountains, at Amery Oasis (70°50'S, 68°00'E) and Fisher Massif (71°31'S, 67°40'E), the Cenozoic glaciomarine Pagodroma Group consists of four uplifted Miocene and Pliocene–early Pleistocene formations here named the Mount Johnston, Fisher Bench, Battye Glacier and Bardin Bluffs formations. These are composed of massive and stratified diamicts, boulder gravels and minor laminated sandstones, siltstones and mudstones. Each formation rests on either Precambrian metamorphic rocks, or on Permo-Triassic fluvial strata. The unconformity surfaces are parts of the walls and floors of palaeofjords. The Miocene Fisher Bench Formation exceeds 350 m in thickness at Fisher Massif, where the yet older Miocene (or Oligocene) Mount Johnston Formation overlies basement rocks at up to 1400 m above sea level. Individual formations contain either Miocene diatoms, or else Pliocene–early Pleistocene diatom-foram assemblages. The diamicts are interpreted as fjordal ice-proximal or ice-contact sediments, deposited seawards of tidewater glacier fronts located some 250 to 300 km inland of the present ocean margin. Each formation records an ice recession following a glacial expansion.

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Key words: Antarctic glacial history, Cenozoic stratigraphy, fjordal sedimentation

Introduction

Considerable controversy exists concerning the Cenozoic glacial history of Antarctica. The East Antarctic offshore stratigraphical record, through drilling, is becoming established from

i) DSDP Leg 28 in the Ross Sea (Hayes & Frakes *et al.* 1975),

ii) ODP Legs 119 (Barron & Larsen *et al.* 1989, 1991) and 188 (Shipboard Scientific Party 2000) to Prydz Bay, and

iii) the Cape Roberts and earlier projects in the western Ross Sea (Barrett 1989, Hambrey & Wise 1998, Fielding & Thomson 1999).

However, the onshore record is sparse and incomplete. This paper describes the distribution and stratigraphy of thick Cenozoic glaciomarine sequences of the Pagodroma Group (Hambrey & McKelvey 2000a) exposed within the northern Prince Charles Mountains of the Lambert Glacier–Amery Ice Shelf region of Antarctica, where they has been preserved by uplift.

The strata were laid down by ancestral phases of the Lambert Glacier–Amery Ice Shelf drainage system. This ice-drainage originates deep in the interior of the continent, but unlike many ice drainage basins in the Antarctic, it has a decipherable glacial history preserved in sedimentary successions along its flanks. Furthermore, this system drains approximately 1 million km^2 (Allison *et al.* 1985, McIntyre 1985) or 9–10% of the East Antarctic ice sheet by area, making it arguably the largest ice-drainage system in Antarctica. Resolving the history of this system through investigations of the Pagodroma Group, and the thick prograded sequence offshore in Prydz Bay, will largely reveal the history of the ice sheet as a whole.

The current controversy concerning the Antarctic ice sheet revolves about whether it has had a stable or unstable history (Stoeven 1997, Miller & Mabin 1998). It has been widely assumed that the East Antarctic ice sheet has been a stable feature since about mid-Miocene time (Shackleton & Kennett 1975, Denton *et al.* 1993, Kennett & Hodell 1995). However, reworked marine diatoms recovered from the Sirius Group, a sparse but widespread glacigenic sequence in the Transantarctic Mountains, have been interpreted as indicating major deglaciation of East Antarctica during global Pliocene warming (Webb *et al.* 1984, Harwood & Webb 1998). This warming is supported by the presence of a *Nothofagus* (southern beech) flora in some Sirius Group strata (Hill *et al.* 1996).

The northern Prince Charles Mountains extend between the latitudes of approximately 69°30' to 71°50'S, bordering the western side of the Lambert Glacier–Amery Ice Shelf drainage

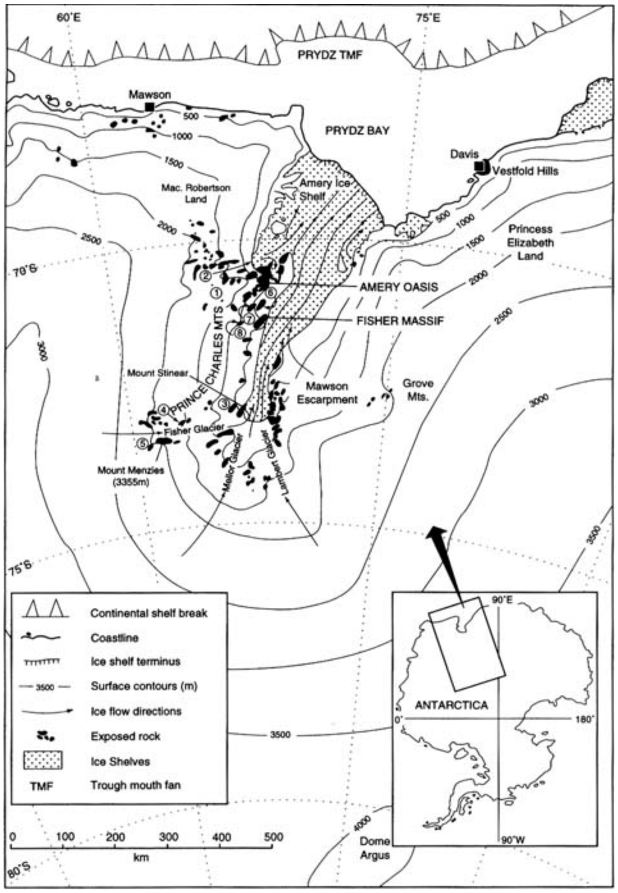

Fig. 1. Geographical setting of the Lambert Glacier–Amery Ice Shelf system and the Prince Charles Mountains within Antarctica. The two main areas of investigation, the Amery Oasis and Fisher Massif, lie on the western flank of the Lambert Glacier–Amery Ice Shelf system.
1. Gorman Hills, 2. Charybdis Glacier, 3. Mount Rymill, 4. Mount Scherger, 5. Mount Mather, 6. Mount Meredith, 7. Mount Lanyon, 8. Mount Collins. Map based upon Hambrey & McKelvey (2000b), and Drewry (1983).

system (Fig. 1). Krebs (1997) has shown that the present grounding line of the system occurs about 170 km south of Amery Oasis, adjacent to Mount Stinear in the southern Prince Charles Mountains. The 700 km long Lambert Glacier–Amery Ice Shelf drainage system occupies the Lambert Graben, a long-lived tectonic feature dating back possibly to Permian time (Federov *et al.* 1982, Stagg. 1985). Hambrey (1991) emphasized that the Lambert Graben has always been one of the most important ice-discharge routes for the East Antarctic ice sheet. However, today only 42% of the ice front currently in Prydz Bay is actually derived from the inner reaches of the Lambert Glacier–Amery Ice Shelf drainage basin. The provenance of much of the ice and the debris it carries, is from tributary ice streams entering along the length of the drainage system.

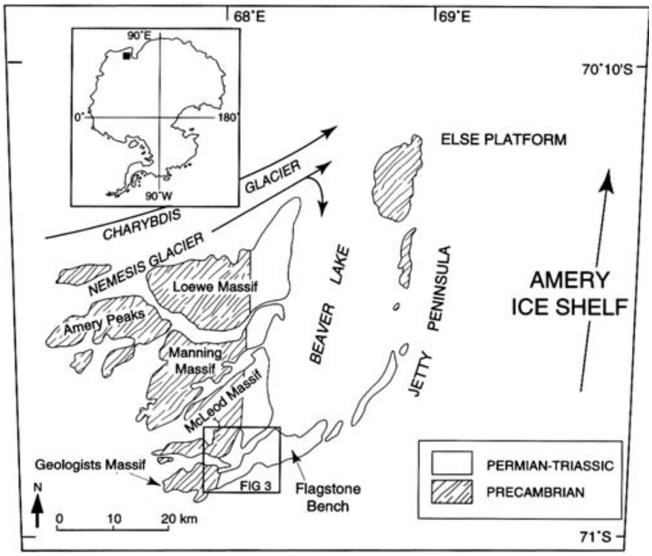

Fig. 2. Sketch map of the Amery Oasis, northern Prince Charles Mountains.

The geological setting of the Pagodroma Group

The crystalline basement of the northern Prince Charles Mountains consists of Proterozoic granulite facies rocks intruded by intermediate and granitic plutons at around 500 Ma (Tingey 1991, Mikhalsky *et al.* 1996, Stephenson & Cook

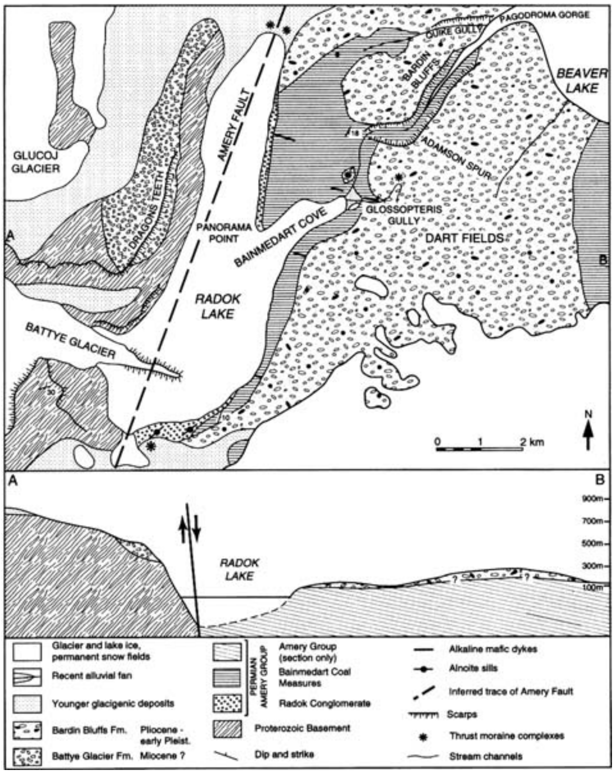

Fig. 3. Salient physiographical and geological features of the Radok Lake–Beaver Lake region of the Amery Oasis.

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1997). The age of the granulite facies metamorphism is *c.* 1000 Ma (Black *et al.* 1987, Henson & Zhou 1995).

In the Amery Oasis late Permian and Triassic Amery Group strata (Fielding & Webb 1996, McLoughlan & Drinnan 1997), of fluvial origin and at least 2500 m thick, abut the Precambrian basement along the Amery Fault at Radok Lake (Figs 2 & 3). The Amery Group dips gently east-south-east at about 10° to 20° and is the only known *in situ* Upper Palaeozoic–Mesozoic sequence in the Prince Charles Mountains, and indeed, in Mac. Robertson Land. The Amery Fault constitutes the western boundary of the Beaver Lake Graben, which is situated high on the western margin of the much larger Lambert Graben. The Amery Fault is the only major fracture whose extent is well documented in the region. However, on stratigraphical evidence, McLoughlan & Drinnan (1997) have demonstrated that major normal faults with throws of the order of 1000 m cut the Amery Group. To date, no firm field evidence of Cenozoic faulting has been presented.

Bardin (1982) described in considerable mineralogical detail a 60 m thick sequence of glacial sediments at Pagodroma Gorge in the Amery Oasis, and drew attention to similar strata further south at Mount Meredith and Mount Collins, and at Fisher Massif. Drewry (1981) applied the name Pagodroma Tillite informally to glacigenic sequences of the Amery Oasis and interpreted the strata Bardin described from Pagodroma Gorge as "two clearly different till units separated by a transitional horizon of fluvioglacial deposits". Bardin & Belevich (1985) gave details of a benthic diatom flora from the Pagodroma Gorge strata, and suggested a post-Early Miocene age. At the Dragons Teeth overlooking Radok Lake in Amery Oasis (Fig. 3), Bardin & Kolosova (1988) described the exposed part of a 120 m sequence of glacigene strata. Bardin & Chepaljiga (1989) identified the marine mollusc *Hiatella arctica* (L.) from dropstone-bearing silty sandstone, which they considered to be of Miocene age, cropping out alongside Beaver Lake near the base of the Pagodroma Gorge sequence, (see Whitehead & McKelvey 2001, figs 2, 3 & 5).

McKelvey & Stephenson (1990) attributed the Pagodroma Tillite to a major late Pliocene or early Pleistocene expansion of the Lambert Glacier–Amery Ice Shelf system. Adamson & Darragh (1991) recorded several mollusc occurrences in Pagodroma Tillite exposed along Pagodroma Gorge. They drew attention to pronounced local relief on the sub-Pagodroma Tillite erosion surface, and in addition showed that the present Pagodroma Gorge continued as an incision across the bottom of Beaver Lake to at least 230 m water depth. McKelvey (1994) summarized a Pagodroma Tillite sequence at least 350 m thick on the flanks of Fisher Massif, and drew attention to a separate sequence estimated to be 70 m thick, occurring some 400 m higher near the southern crest of the massif.

Offshore drilling to within 90 km of the Amery Ice Shelf by Leg 119 of the Ocean Drilling Program (Barron & Larsen *et al.* 1989, 1991), penetrated a discontinuous glaciomarine Eocene/Oligocene to Quaternary sequence approaching 500 m in thickness. The late Miocene to Quaternary intervals recovered from four drilling sites (Hambrey *et al.* 1991) must be at least in part correlative with the Pagodroma Tillite sequences of the northern Prince Charles Mountains.

Hambrey & McKelvey (2000a, 2000b) raised the status of the Pagodroma Tillite (formation) to that of Pagodroma Group, and demonstrated the bulk of the Pagodroma Group to be composed of fjordal ice-proximal diamicts, laid down by the receding tidewater glacier front of ancestral phases of the Lambert Glacier–Amery Ice Shelf system.

Stratigraphy of the Pagodroma Group

Four Pagodroma Group formations are recognized at Amery Oasis and Fisher Massif. These are:

Fisher Massif

(a) Fisher Bench Formation. > 335 m. Miocene.
(b) Mount Johnston Formation. > 215 m. Lower Miocene, or Oligocene?

Amery Oasis

(c) Bardin Bluffs Formation. > 60 m. Pliocene.
(a) Battye Glacier Formation. > 130 m. Miocene?

None of these formations is complete, each being overlain unconformably by either a thin mantle of sandy gravel (interpreted as till from dry-based glaciers) of Pleistocene age, or else younger deflation pavements. The Battye Glacier,

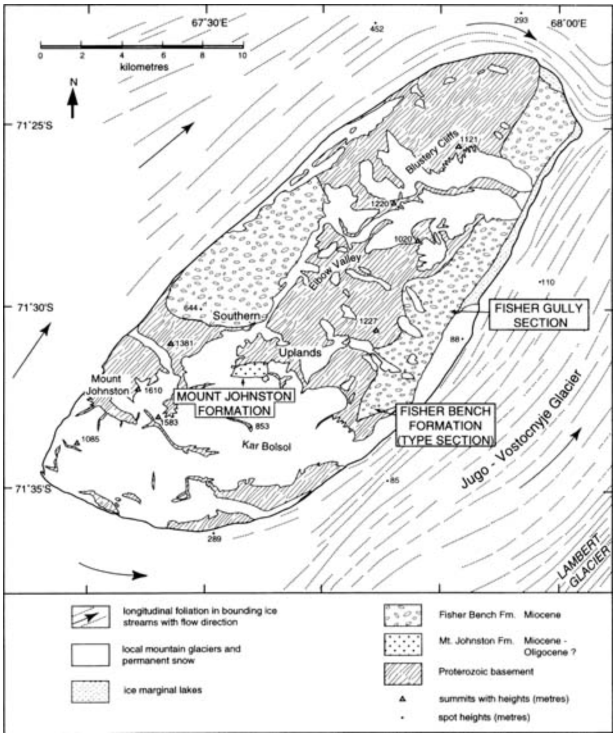

Fig. 4. Salient physiographical and geological features of Fisher Massif.

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Fig. 5. The Southern Uplands, and the south-eastern face of Fisher Massif. View looking westwards. A terrace of Fisher Bench Formation (centre, and right) rests unconformably, at 270 m a.s.l., upon Proterozoic metamorphic basement. The Fisher Bench Formation probably exceeds 335 m in thickness, but here only the lowermost 285 m are exposed. In the foreground the Jugo Vostocnyje Glacier (*c.* 100 m a.s.l.) flows from left to right. Kar Bolsol Glacier cirque is on left. The planar unconformity surface (short arrows), a palaeofjord floor, is essentially horizontal. However, in the most western exposure (left) this floor becomes inclined and rises steeply (long arrow) as part of the palaeofjord wall. Asterisks indicate the top and bottom of the Fisher Bench Formation type section. The Southern Uplands skyline (*c.* 1300–1470 m a.s.l.) contains scattered Mount Johnston Formation sequences, representing perhaps 215 m of stratigraphic thickness.

Fig. 6. Southern Uplands, Fisher Massif. View looking eastwards. Scree-mantled Mount Johnston Formation, *c.* 70 m thick, unconformably overlying an irregular erosion surface (arrowed, and immediate foreground) cut on Proterozoic metamorphic basement. Although poor exposure precludes measurement of this section, the indistinct trace of stratification of the Mount Johnston Formation is discernable through the scree mantle. In right middle distance the unconformity surface has been stripped of Cenozoic strata. Small outliers (< 20 m thick) of Mount Johnston Formation cap at irregular intervals the crest of the snow-flanked ridge (*c.* 1362 m) in centre background. Skyline figure (encircled) indicates scale.

Fisher Bench and Mount Johnston formations rest unconformably upon Proterozoic metamorphic basement. The Bardin Bluffs Formation overlies, with angular unconformity, the Upper Permian Radok Conglomerate and Bainmedart Coal Measures of the Amery Group (Fig. 3). In the following account, the sediment facies are described according to a non-genetic classification developed by Moncrieff (1989), and modified by Hambrey (1994, table 1.1). The depositional environments identified herein are from a detailed study of the Pagodroma Group's sedimentology by Hambrey & McKelvey (2000a). The term diamict is used non-genetically for poorly sorted sediment with a wide range of particle sizes and shapes. It embraces both diamicton (non-lithified) and diamictite (semi-lithified), both of which are present in the Pagodroma Group.

Fisher Massif

Fisher Massif is a south-west–north-east trending range measuring about 30 km by 12 km, and bounded by the Jugo-Vostocnyje Glacier and another unnamed tributary ice stream, both flowing north-eastwards from the interior of Mac. Robertson Land, down to join the Amery Ice Shelf (Fig. 4). The range is largely ice-free except for the southern end where the Kar Bolsol cirque glacier flows eastwards to join the Jugo-Vostocnyje Glacier. The ice surface surrounding Fisher Massif falls from *c.* 780 m on the south-west margin to below 100 m on the north-east flank of the range. Two broad undulating terraces of the Fisher Bench Formation, and ranging between 500 to 800 m in altitude, extend along much of the western and eastern flanks of Fisher Massif (Fig. 4). A small ice-free valley, Elbow Valley separates the Northern Uplands dominated by Blustery Cliffs (*c.* 1100 m), from the Southern Uplands dominated by Mount Johnston (1610 m). The Mount Johnston Formation caps part of the Southern Uplands, overlooking the Kar Bolsol Glacier (Figs 5 & 6).

On Fisher Massif, the Pagodroma Group successions are stratified, albeit sometimes indistinctly, and are clearly flat-lying. In altitude, their bases range from *c.* 100 m to *c.* 1480 m. In general, good exposures of the Pagodroma Group at Fisher Massif are few, as all but the steepest slopes are mantled by loose scree, and in flatter areas the sequence is either hidden beneath a thin blanket of Pleistocene sandy gravel or alternatively, ablated tesselation pavements.

Mount Johnston Formation

Type section (Fig. 7a)
Lat. 71°31'43"S; Long. 67°34'43"E. (top of section.)
Locality: Southern Uplands, Fisher Massif.

Thickness
Maximum thickness of individual measured sections 54 m. Cumulative thickness considerably greater, perhaps more

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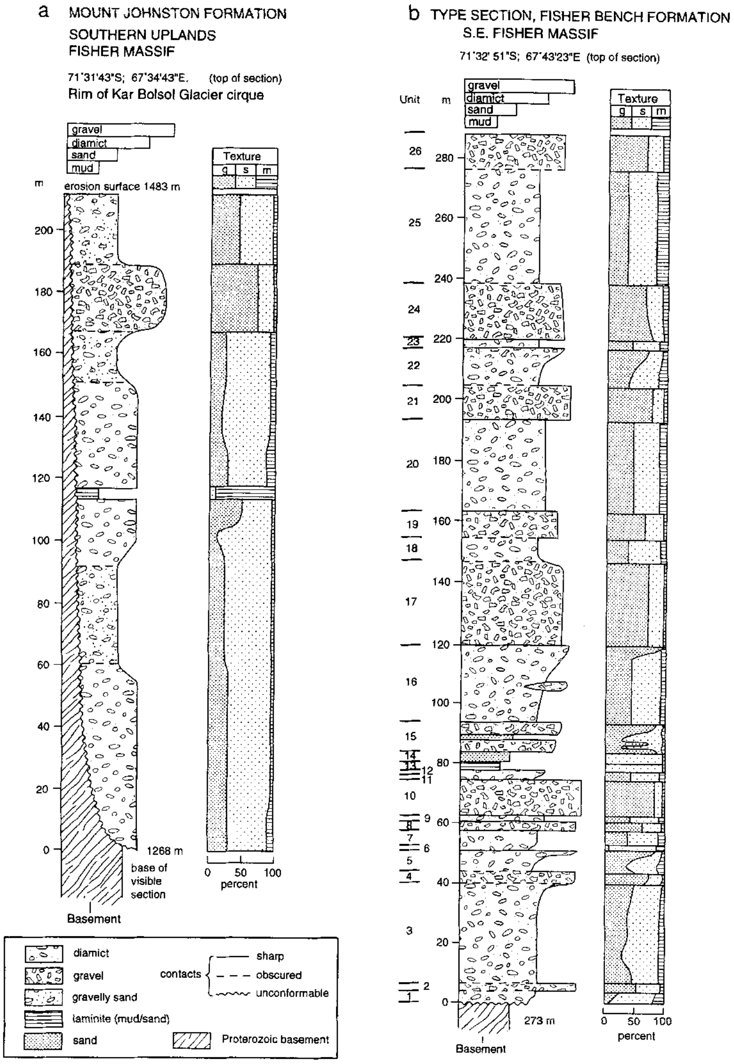

Fig. 7. Type sections of **a.** Mount Johnston Formation, and **b.** Fisher Bench Formation, at Fisher Massif. Texture based on visual estimates in the field, of gravel, sand and mud.

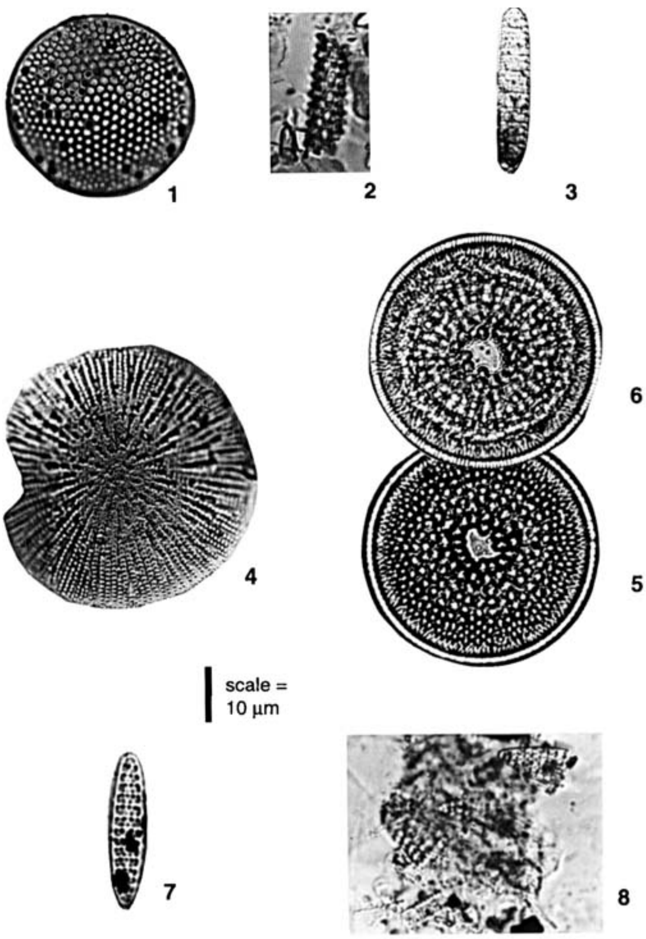

Fig. 8. Diatoms, Pagodroma Group. 1. *Stephanopyxis splendidus*, Mount Johnston Formation. 2. *Pyxilla reticulata* (fragment), Mount Johnston Formation. 3. *Denticulopsis simonsenii*, Fisher Bench Formation. 4. *Actinocyclus actinochilus*, Bardin Bluffs Formation. 5. *Actinocyclus ingens* var. *nodus*, Fisher Bench Formation. 6. *Actinocyclus ingens* var. *nodus*, Fisher Bench Formation. 7. *Fragilariopsis kerguelensis*, Bardin Bluffs Formation. 8. *Fragilariopsis kerguelensis*, (fragment within a diatomaceous aggregate) Bardin Bluffs Formation.

than 215 m. Formation top not preserved.

Facies

Undifferentiated massive clast-rich sandy diamict, boulder gravel, and boulder gravel sand. Stratification is indistinct, on a several metre-thick scale. A 1 m thick laminite interval occurs *c.* 117 m above base, at 71°31'49"S, 67°32'03"E.

Formation age

Early Miocene or older. The presence of the diatoms *Pyxilla reticulata* and *Stephanopyxis splendidus* (Fig. 8), which biostratigraphically overlap in the Oligocene, together with the absence of the younger middle Miocene species found in the Fisher Bench Formation suggests a early Miocene or even Oligocene age.

The flat-lying Mount Johnston Formation overlies an unconformity surface that rises in altitude from 1268 m to 1480 m over a distance of one kilometre. The present top of the Mount Johnston Formation is an erosional, ice-smoothed surface, and no individual measured section through the unit exceeds 54 m.

Anisotropic depositional fabrics (Fig. 9) are rare and indicate palaeo-iceflow directions mainly in the north-west–south-east quadrants. However, the bulk of the Mount Johnston Formation is interpreted as ice-proximal fjordal sediment, with the occasional anisotropic fabrics suggesting ice-grounding events (Hambrey & McKelvey 2000a).

Fisher Bench Formation

Type section (Fig. 7b).

71°32'51"S, 67°43'23" E. (top of section.)

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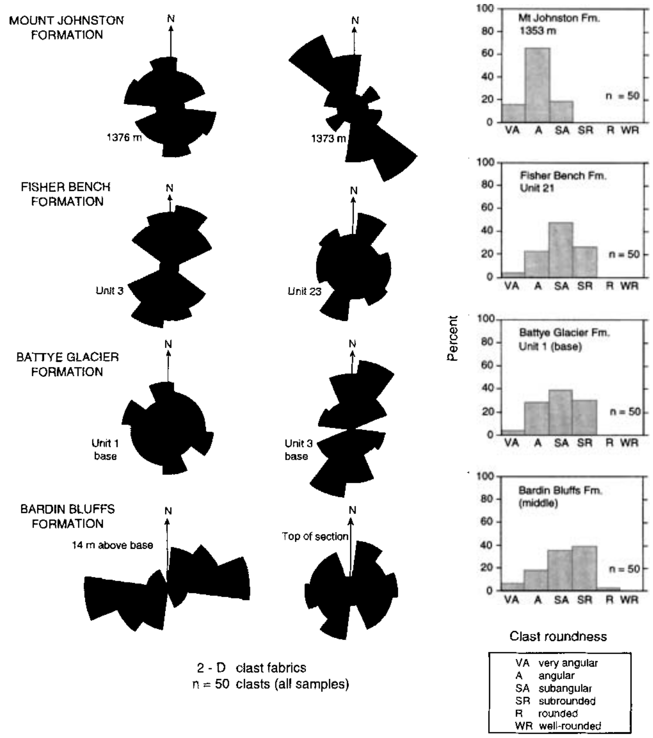

Fig. 9. Two-dimensional clast fabrics, measured in the horizontal plane, for diamict and gravelly sand facies of the Pagodroma Group formations. Stratigraphical positions within the formations are indicated by either altitude above sea level, or else by the unit numbers within the type sections. Similarly, Power's roundness distributions for diamict and gravel are illustrated.

Fig. 10. Fisher Massif, south-eastern flank. View looking south-south-west. Approximately 180 m of scree-mantled Fisher Bench Formation diamict and gravel overlie unconformably, at 270 m a.s.l., the Proterozoic metamorphic basement. The horizontal unconformity surface (arrows), part of a palaeofjord floor, to the right becomes steeply inclined where it rises to become the palaeofjord wall. Recent mudslides (light coloured) streak both the Pagodroma Group and the basement cliffs. The flat-topped Shaw Massif, 1356 m, (centre-right background) is 50 km distant. Mount Willing is on extreme right.

Locality: south-east Fisher Massif (Figs 4 & 5).

Other representative sections.

71°32'30" S, 67°45'16"E. (top of section)
Locality: south-east Fisher Massif.
71°30'08"S, 67°51'20"E. (top of section)
Locality: Fisher Gully, Fisher Massif (Fig. 4).

Thickness

A minimum of 285 m occurs in the type section. At least another 50 m of scree-covered Pagodroma Group overlies the measured part of the section. Formation top not seen. Laiba & Pushina (1997) state the thickness to be 320 m.

Facies

Massive clast-rich sandy diamict, boulder gravel and sandy gravel, in alternating units up to 40 m thick. Minor pebble gravel, sand and laminite.

Formation age

Miocene. Laiba & Pushina (1997) describe a middle–late Miocene benthic diatom flora from a laminite unit *c.* 70 m above the formation base. From the base of the formation (Fig. 7b, Unit 1) this current investigation recovered a middle Miocene (14.2–12.5 Ma) diatom flora (Fig. 8).

The type section is moderately well exposed, rising above 170 m high cliffs of metamorphic basement. The unconformity surface here is planar and horizontal, being a palaeofjord floor remnant. About 1 km to the south-west, this same surface curves and rises steeply towards the west for over 300 m (Figs 5 & 10); the surface here representing part of the palaeofjord wall.

Although alternating units of sandy diamict and sandy boulder gravel predominate in the type section, only 1.5 km to the north, at 71°32'30"E, 67°45'16"S, a 277 m sequence contains only sandy diamict, apart from a 0.5 m sandy laminite at about 98 m above the base of the section.

Approximately 7.5 km north-east along the bench from the type section, in Fisher Gully (Fig. 4), nearly 100 m of clast-rich sandy diamict are overlain by *c.* 180 m of poorly exposed diamict. Here the base of the succession does not crop out, and the relationship with the Fisher Bench Formation type section is unclear.

North-east again from the Fisher Gully section, in a series of 5-m high cliffs (at about 71°29'48"S, 67°51'58"E.) diverse facies of the Fisher Bench Formation are well exposed. Thin sharp-bounded diamicts are interbedded with laminites (Hambrey & McKelvey 2000a, figs 13, 14). The latter contain dropstones at some horizons, and in one instance the laminites have been modified to intraformational pebble breccias. In this area, many of the Pagodroma Group exposures have been deformed and disrupted, as a result of either later movement of an enlarged Jugo-Vostocnje Glacier against the base of the terrace, or else partial gravity collapse of the succession.

The horizontal stratification and only occasional development of anisotropic depositional fabrics (Fig. 9) supports the interpretation that the sandy diamicts are ice-proximal fjordal deposits, deposited close to the (tidewater) ice front of a receding ancestral Lambert Glacier–Amery Ice Shelf drainage system (Hambrey & McKelvey 2000a). The sharply bounded and thinner diamicts, commonly associated with laminites and thin sands (e.g. Fig. 7b, units 4, 12 & 15), together with those to the north of the Fisher Gully section, are interpreted as sub-aquatic gravity flows, reworked from glacially deposited fjordal diamict. The laminites record deposition of suspended sediment either under the influence of fjordal tides, in a similar manner to the cyclopsams and cyclopels of Powell (1984), or as distal turbidites in the fjords. The boulder gravels in the type section, dominated by local basement material, we consider to represent fjord-wall collapse debris mixed with basal glacial debris.

Amery Oasis

Amery Oasis covers an area of *c.* 1800 km^2 (Adamson *et al.* 1997). It extends eastwards from the rugged Geologists, McLeod, Manning and Loewe massifs to Else Platform and Jetty Peninsula, which border the eastern shores of the 50 km long ice-covered and tidal Beaver Lake (Fig. 2). Much of the Beaver Lake basin is back-filled by a southward flowing arm of ice sourced from the Nemesis and Charybdis glaciers (see Adamson *et al.* 1997, fig. 3). Beaver Lake is joined to the smaller freshwater Radok Lake via the 7.5 km long Pagodroma Gorge (Fig. 3). A shallow arm of Radok Lake, Bainmedart Cove, is contiguous with the head of Pagodroma Gorge. The 346 m deep Radok Lake is the deepest lake (excluding sub ice-sheet lakes) in Antarctica (Adamson *et al.* 1997). Battye Glacier occupies a deep trough, separating Geologists Massif from McLeod Massif, and extends across Radok Lake as a floating ice tongue 2 km long.

The Pagodroma Group in the Amery Oasis consists of the Battye Glacier Formation (Figs 3 & 11) and the younger Pliocene–early Pleistocene Bardin Bluffs Formation (Figs 3 & 12). The Battye Glacier Formation extends west from Radok Lake, whereas the Bardin Bluffs Formation is widely distributed to the east of the lake. However, apart from at Bardin Bluffs in Pagodroma Gorge, good exposures of either formation are few.

Battye Glacier Formation

Type section (Figs 11 & 13).
70°51'51"S, 68°57'11"E. (top of section.)
Locality: Dragons Teeth, Amery Oasis.

Thickness
171 m. Formation top not preserved. Pleistocene sandy boulder gravel overlies disconformably.

Fig. 11. Battye Glacier Formation, overlooking western shore of Radok Lake at the Dragons Teeth. View looking south-west. Approximately 160 m of Miocene? glaciogenic diamict and gravel overlie unconformably, at 315 m a.s.l., the Proterozoic metamorphic basement. Asterisks indicate top and bottom of Battye Glacier Formation type section. Note the planar and horizontal unconformity surface (short arrows). The Battye Glacier Formation thins further westwards to eventually abut the rising (palaeofjord wall) basement surface (between long arrows).

Facies
Massive clast-rich sandy diamict, boulder gravel and gravelly sand with indistinct coarse (5 to 20 m) stratification. Minor laminites present.

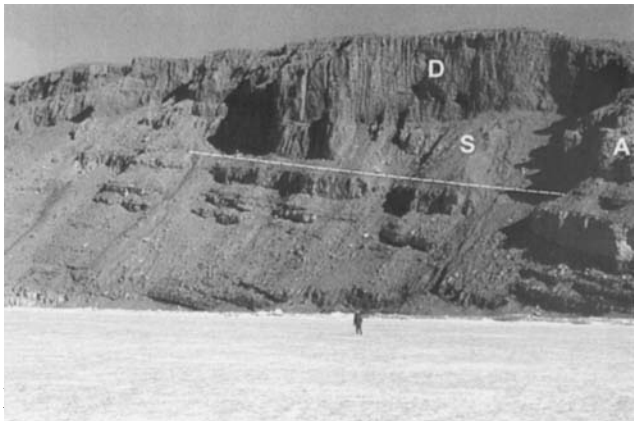

Fig. 12. Bardin Bluffs Formation, Bardin Bluffs, Pagodroma Gorge. View looking north-west. Approximately 30 m of indistinctly stratified thick-bedded (up to 5 m) glaciogene diamicts (D) and basal silty sands and silts (S) rest unconformably on gently dipping Permian Bainmedart Coal Measures of the Amery Group (see also Bardin 1982, fig. 138.2, p. 1071). Dotted line indicates approximate position of the contact. The Cenozoic strata infill a palaeovalley trending sub-parallel to the bluffs. The Amery Group outcrop A, (right), is part of the near wall of the exhumed palaeovalley

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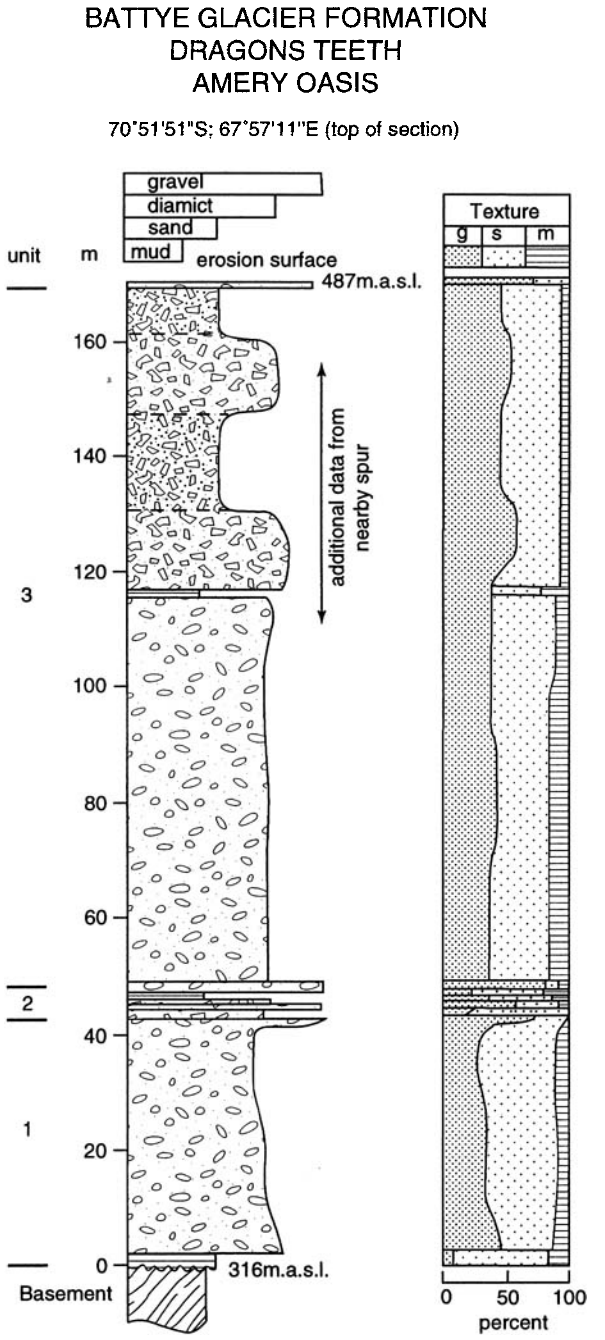

Fig. 13. Type section of the Battye Glacier Formation, Dragons Teeth, Amery Oasis. See Fig. 7 for explanatory key.

Formation age

Not known, probably Miocene. No age data for the Battye Glacier Formation are yet at hand. Sparse non-age diagnostic bivalve detritus has been recovered. We suggest that this formation broadly equates with the Miocene Fisher Bench Formation because of the similar palaeogeomorphic settings of the two units and the comparable altitudes of their bases. Laiba & Pushina (1997) similarly suggest this correlation.

Part of the Battye Glacier Formation, probably from within our designated type section of the unit (Fig. 11), has been described by Bardin & Kolosova (1988). This scree-draped succession at Dragons Teeth extends northwards along strike for at least 7 km, resting upon the smoothed and planar-horizontal unconformity surface cut on the Proterozoic basement. This surface we believe to be part of a palaeofjord floor. Over a distance of 1.5 km westwards, the Battye Glacier Formation thins and then abuts, at *c.* 600 m a.s.l., the rising unconformity surface, which we interpret to be the lower part of the adjacent palaeofjord wall (Fig. 11).

The massive clast-rich sandy diamicts comprising Units 1 and 3 (Fig. 13) we consider to be ice-proximal fjordal deposits (Hambrey & McKelvey 2000a). Anisotropic fabrics recorded at a few horizons (Fig. 9) may suggest some grounding of ice, moving in north-east–south-west and north–south directions.

Apart from these diamicts, which when viewed from a distance display indistinct stratification, the minor facies (i.e. Fig. 13, Unit 2) include a few laminated silty fine sandstones with dropped pebbles and cobbles. These strata are often interbedded with thin (< 1.5 m) sharp-bounded diamicts exhibiting gravel lags on top (Hambrey & McKelvey 2000a figs 12C, 12D, 15). The thin diamicts are of subaquatic gravity-flow origin, and the laminated silty sands we consider to be either cyclopsams of tidal origin or distal turbidites, similar to those in the Fisher Bench Formation. At the base of the type section, and along strike, a well-sorted fine sand (< 75 cm) contains at least two species of broken, but otherwise well-preserved, bivalve fragments up to 12 mm across.

Three km west-south-west of Dragons Teeth, and about 1.5 km south of the Glucoj Glacier (i.e. at 70°51'05"S, 67°53'11"E), both the Pagodroma Group and the widespread mantle of Pleistocene sandy gravel are absent. Well-sorted pebble- and boulder gravels up to 40 cm thick occupy sinuous shallow depressions cut into the metamorphic basement. The gravel clasts range from rounded to well-rounded. Large ripple and mound-like bed-forms are preserved and a fluvial origin is apparent. A weathering profile or skeletal soil at least 0.5 m deep is developed on the basement rocks adjacent to the gravels. Close by stand scattered basement tors. The relationship of these gravels and the nearby weathering features (tors, palaeosol) relative to the Pagodroma Group is not yet known.

Bardin Bluffs Formation

Type section

70°48'38"S, 68°08'44"E. (top of section.)
Locality: Bardin Bluffs, Pagodroma Gorge, Amery Oasis (Figs 3 & 12).

Thickness

Exceeds 65 m. Top not preserved.

Facies

Predominantly massive or weakly stratified diamict and massive gravelly sand. Minor laminite and silty fine sandstone. About 12.5 m of silty sand and silt, containing relatively minor ice-rafted debris, occur at the base of the type section (Bardin & Belevich 1985, Whitehead & McKelvey 2001). This basal facies is known only in the immediate vicinity of Bardin Bluffs, and at Adamson Spur (Fig. 3).

Formation age

Pliocene–early Pleistocene. A foraminiferal and radiolarian fauna from *c.* 11.5 m above the base of the formation in the type section constrains the age to between 5 and 1 Ma (Whitehead & McKelvey 2001). Bardin & Belevich (1985) describe a benthic diatom flora from a slightly older stratum in the same section (Whitehead & McKelvey 2001, fig 4) that includes *Actinocyclus actinochilus*, a mid-Pliocene to Recent species (Laiba & Pushina 1997).

Bardin Bluffs is here named after the late Russian geologist V. I. Bardin, who first drew attention to the Cenozoic sequences in Amery Oasis (e.g. Bardin & Kolosova 1983, 1988). The Bardin Bluffs Formation is widespread in the Amery Oasis, both to the east and south-east of Radok Lake. Within Pagodroma Gorge the best exposures are at Bardin Bluffs and on Adamson Spur, although at the latter locality glaciotectonic soft-sediment deformation, including thrusting and overfolding, prohibits recognition of a reliable stratigraphic section. The Bardin Bluffs Formation overlies with angular unconformity, Permian fluvial strata of the Amery Group. The unconformity surface descends gently from *c.* 180 m overlooking the northern part of Radok Lake, to near sea level on the western shore of Beaver Lake, immediately north of Pagodroma Gorge. Well-developed glacial striations on the unconformity surface occur above the south-east shore of Radok Lake (at 70°52'11"S, 68°01'21"E) and on the south side of Bainmedart Cove at GR 65003533 (Antarctic Division 1990). These trend 010° and 355° respectively.

Massive or weakly stratified clast-rich sandy diamict and massive gravelly sand are by far the most abundant facies. At Bardin Bluffs the former facies constitutes virtually the whole section, except for the basal 12 m of silty fine sands and silts (with scattered clasts), described by Bardin & Belevich (1985, p. 78, fig. 2a) and Whitehead & McKelvey (2001).

About 1 km north-east of Bardin Bluffs, overlooking the mouth of Pagodroma Gorge, two other facies not seen elsewhere in the Pagodroma Group crop out. Overlying and abutting unusually rugged local topography on the unconformity surface, an admixture of Amery Group blocks and boulders, in diamict, produces a spectacular megabreccia facies. About 300 m further to the north-east, at GR 70024045 (Antarctic Division 1990), and stratigraphically only slightly lower (< 10 m), a shelly muddy sand with sparse dropstones crops out within a few metres of the base of the Bardin Bluffs Formation (Whitehead & McKelvey 2001, fig 3). Topographically this exposure is the lowest recorded in the region, being less than 4 m a.s.l.

Fabric data from the sandy diamicts at Bardin Bluffs (Fig. 9) are ambiguous, but on the basis of other sedimentological criteria these strata are clearly ice-proximal fjordal deposits (Hambrey & McKelvey 2000a). Preferred orientations in one case are east–west, whereas the palaeovalley which encloses that part of the basal Bardin Bluffs Formation trends north–south, suggesting the fabric is slope-related. Some ice contact with bedrock is indicated by striations on the sub-Bardin Bluffs Formation surface, e.g. those overlooking the southern shore of Bainmedart Cove. However whilst such glacial abrasion may have been contemporaneous with deposition of the recessional Bardin Bluffs Formation, it is quite possibly older, being an artefact of the cutting of the fjord floor during a glacial expansion (see below).

Micropalaeontology

Palaeontologically, the Pagodroma Group strata contain only sparse and probably reworked microfloras and microfaunas. However, from three localities relatively numerous microfossils have been recovered. These localities are:

a) A silty fine sandstone-siltstone interval within the basal diamict unit of the Fisher Bench type section (i.e. Unit 1, Fig. 7b).

b) The laminite sampled by Laiba & Pushina (1997), located either in the type section of the Fisher Bench Formation or close by, along strike. We assume this laminite to be Unit 13 of the type section (Fig. 7b).

c) A silty fine sandstone approximately 11.5 m above the base of the Bardin Bluffs Formation at Bardin Bluffs (Whitehead & McKelvey 2001, Fig. 4, sample 90-7).

The significant features of the microfossil assemblages recovered from (a) and (c) and the reasons why we consider them to be *in situ* are outlined below.

a) Base of Fisher Bench Formation

The Middle Miocene age of the basal Fisher Bench Formation is indicated by the presence of the diatoms *Actinocyclus ingens* var. *nodus* Baldauf with a last appearance datum (LAD) of 12.5 Ma, and *Denticulopsis simonsenii* Yanagisawa et Akiba (referred to as *Denticulopsis hustedtii* (Simonsen et Kanaya) Simonsen in earlier literature). The first appearance datum (FAD) of the latter is 14.2 Ma. These diatom datums are from the Southern Ocean zonation of Harwood & Maruyama (1992), and have been revised to the Berggren *et al.* (1995) palaeomagnetic timescale. The relatively high abundance of diatoms within a restricted stratigraphical interval of fine-grained sediment, and the absence of biostratigraphical and biogeographical mixing in this assemblage suggests it to be an *in situ* one, which occupied a fjordal marine environment during times of reduced ice cover.

c) Bardin Bluffs Formation

This assemblage (sample 90-7) contains foraminifera (24 genera and 32 species), large diatoms, and one radiolarian species. Also present are ostracods (two species), macrofossil debris, fossil wood fibres and faecal pellets. The foraminifera are abundant, well preserved, and the tests are small (< 125 microns). In terms of test numbers, benthic taxa constitute about 80% of the foraminiferal population. The planktonic element is represented by *Neogloboquadrina pachyderma* (Ehrenberg) and *Globigerina* sp. cf. *G. antarctica* Keany & Kennett. Kennett (1978, 1980) noted that the first appearance of *N. pachyderma* is diachronous, appearing progressively later in the Neogene record from lower to higher latitudes. Kennett & Srinivasan (1983) assigned a maximum age range of late Miocene (N16 ~ 10 Ma) to Recent for *N. pachyerma*. Berggren (1992) observed that at ODP 747 (Kerguelen Plateau, ~55°S) this species first appears in Chron 4 (~ 7 Ma) and at ODP 751 (~59°S), in Chron 4A (~8 Ma). *Neogloboquadrina pachyderma* is present in the early Pliocene (4.8 to 4 Ma) of DVDP 10 and 11 (Taylor palaeofjord in McMurdo Sound, ~ 78°S), but is not present in the underlying deep-water marine early Pliocene and late Miocene sediments of these drill holes (Ishman & Rieck 1992). In view of the latitude of the Bardin Bluffs Formation (70°S), the occurrence of *N. pachyderma* favours an age no older than early Pliocene (< 5.2 Ma). *Globigerina* sp. cf. *G. antarctica* is the dominant of the two planktonic species present. Keany & Kennett (1972) and Kennett (1978, 1980) indicate that this species ranges through the Matuyama Reversed Epoch and has a last occurrence in Jaramillo Event sediments (C1r.1n). This equates with an age span of 2.6 to 1.07–0.99 Ma, or late Pliocene–early Pleistocene (Cande & Kent 1995). The extant radiolarian *Antarctissa denticulata* is also present in sample 90–7 and indicates an age no older than early Pliocene. The presence of these three planktonic taxa indicates that the lower part of the Bardin Bluffs Formation (i.e. Member 1 of Whitehead & McKelvey 2001) was deposited at sometime within an age range of early Pliocene to early Pleistocene. The relatively abundant planktonic foraminiferal assemblage and less common single radiolarian species point to the existence of normal marine surface waters without a permanent ice shelf or semi-permanent sea-ice cover.

Bardin & Belevich (1985, plate 1) illustrated that their predominantly benthic diatom flora from close to the base of the Bardin Bluffs Formation (i.e. beneath sample 90-7) also includes planktonic diatoms. The extant planktonic diatom *Actinocyclus actinochilus* (Ehrenberg) Simonsen is illustrated in photo 4 of their plate 1. This species has a FAD in the late Pliocene at *c.* 3.1 Ma (Bohaty *et al.* 1998). *Actinocyclus actinochilus* fragments also occur in sample 90-7. Thus these data suggest these Bardin Bluffs strata are no older than late Pliocene. Overlying diamicts (i.e. basal Member 2 of Whitehead & McKelvey 2001) contain the diatom *Fragilariopsis kerguelensis* (O'Meara) Hustedt (Fig. 8) which has a FAD of 3.2 Ma, again consistent with the *Actinocyclus actinochilus* age of < 3.1 Ma for the basal Bardin Bluffs Formation.

Discussion

Four features of the Pagodroma Group deserving further comment are (a) the sub-Pagodroma Group surfaces, (b) the tectonic setting of the four formations, (c) the glacial history they each record, and (d) the distribution of equivalent strata elsewhere in East Antarctica.

The sub-Pagodroma Group surfaces

The erosion surfaces preserved beneath the Mount Johnston, Fisher Bench and Battye Glacier formations are remnants of fjordal glacial troughs cut into the Precambrian basement rocks since (a), they are overlain by unequivocal glaciomarine facies (Hambrey & McKelvey 2000a,) and (b), in two instances parts of the fjord walls are preserved.

The youngest erosion surface, underlying the Pliocene–early Pleistocene Bardin Bluffs Formation, and cut into Amery Group strata, is a composite one. A younger smoothed surface gently inclined eastwards towards Beaver Lake is superimposed upon, and has nearly completely removed an older weathered and more rugged surface (Whitehead & McKelvey 2001). Only the lowest parts of the latter remain, in the form of the bottoms of several small valleys, notably at Bardin Bluffs, Adamson Spur, Glossopteris Gully and perhaps Quike Gully. It is possible that the three former were once contiguous, being remnants of the same valley. It is the younger smoothed component of the erosion surface that exhibits the (previously mentioned) north–south trending striations on Amery Group sandstones above Bainmedart Cove, and overlooking the south-eastern shore of Radok Lake. In contrast the older weathered part of the erosion surface, suggested by Adamson *et al.* (1997) to be fluvially cut, lacks striations.

The precise palaeogeographical setting of this composite erosion surface remains unclear. The later smoothed and striated component clearly represents a fjord floor but no associated wall remnants are apparent. It is possible that part of the palaeofjord wall was located in the vicinity of the Dragons Teeth and subsequent Pleistocene glacial excavation of the Radok Lake basin removed the original feature.

Tectonic considerations

The fact that the Pliocene–early Pleistocene Bardin Bluffs Formation occurs at a lower altitude than the Miocene Fisher Bench Formation (and also the probably contemporaneous Battye Glacier Formation), and that in turn, the Fisher Bench Formation similarly occurs at a lower altitude than the older Mount Johnston Formation clearly identifies Neogene tectonism. However, the precise amounts and rates of uplift are not presently known. Clearly the fossiliferous strata at the base of the Fisher Bench Formation (Fig. 7b, Unit 1) can only indicate an overly conservative minimum of 270 m of uplift since 14.2 to 12.5 Ma. The real uplift figure must be considerably greater, for the bathymetry of the basal Fisher Bench strata is not known, and an original bathymetric setting of several hundreds of metres is not improbable, given the depths of modern fjords. Furthermore, since at least 285 m of younger marine strata overlie the fossiliferous strata, then minimum amounts of uplift in the order of perhaps 550 to 600 m are to be contemplated. Laiba & Pushina (1997, p. 982) suggest a bathymetry of *c.* 50 m for their benthic diatom flora, which we estimate they recovered from about 77 m above the base of the section. Based on these figures, a minimum uplift of the order of 400 m is indicated by the Laiba & Pushina flora. Although the lower Miocene or possibly Oligocene age of the Mount Johnston Formation needs refining, this formation clearly predates the mid-Miocene Fisher Bench Formation. Consequently, the fact that the unconformity surface beneath the Mount Johnston Formation is overlain by glaciomarine strata at nearly 1400 m attests to a minimum uplift of that order since before the mid-Miocene (i.e. before *c.* 14.2 to 12.5 Ma).

Glacial history

The cutting by grounded ice of individual palaeofjords, each now overlain by the Mount Johnston, Fisher Bench and Bardin Bluffs formations respectively, we interpret as the record of three separate and major expansions of ancestral phases of the Lambert Glacier–Amery Ice Shelf drainage system (Hambrey & McKelvey 2000b). ODP drilling of the thick (> 500 m) but incomplete early Oligocene to Holocene glaciomarine sequences in Prydz Bay, confirms that the ice fronts of such glacial expansions reached far out across the continental shelf, beyond the mouth of Prydz Bay, and contributed to the development of a major trough-mouth fan (Hambrey *et al.* 1991, O'Brien & Leitchenkov 1997, Shipboard Scientific Party 2000). The subsequent deposition of the Pagodroma Group proximal glaciomarine strata onto the palaeofjord floors records the recession of the floating ice fronts to positions at least 250 km and 300 km south from inner Prydz Bay, in Pliocene and Miocene times respectively (Hambrey & McKelvey 2000b). Given that the present day Lambert Glacier–Amery Ice Shelf system drains *c.* 9% of the area of the Antarctic ice sheet, it follows that ice front advances and recessions, over such distances, strongly suggest considerable volume variations of the Miocene and Pliocene East Antarctic ice sheet. Furthermore, should the little-known Cenozoic strata at Mount Stinear in the southern Prince Charles Mountains (see below) prove to be of similar glaciomarine origin, then recession of a Lambert Glacier–Amery Ice Shelf drainage system a further 450 km southwards into the Lambert Graben must be contemplated. Such an event would have involved greater volume variations of the East Antarctic ice sheet.

Other Cenozoic sequences

Elsewhere in the northern Prince Charles Mountains, beyond the Amery Oasis and Fisher Massif, the distribution of Pagodroma Group strata is poorly known. Cenozoic strata at Mount Meredith (Fig. 1) are thickly blanketed by scree, and no good exposures are known. Similar strata, as yet uninvestigated, occur at glacier level immediately south-east of Mount Lanyon (Fig. 1). A very few erratics of Cenozoic diamict have been found in the vicinity of the Gorman Hills, and on moraines bordering the Charybdis Glacier (Fig. 1).

Approximately 160 km to the south of Fisher Massif, Cenozoic strata in the southern Prince Charles Mountains abut either side of the tabletop crest of Mount Stinear (Fig. 1) (Whitehead *et al.* 2000). The scree-draped sequence on the eastern side is approximately 110 m thick. Here, the geological setting and sediment facies appear to be very similar to those of the Mount Johnston and Fisher Bench formations at Fisher Massif, suggesting at Mount Stinear the presence of an uplifted fjordal sequence.

Nearby, diamicts cap mounts Rymill and Scherger (Fig. 1), and further to the west across the Fisher Glacier at Mount Menzies Cenozoic formations of diamict with minor sandstone and laminite are widespread. Similar diamicts also occur at Mount Mather (Fig. 1). Initial investigations suggest terrestrial glacial deposition (Whitehead *et al.* 2000). However, the temporal relationship between these Cenozoic strata in the southern Prince Charles Mountains and the Pagodroma Group is not yet known.

Elsewhere in East Antarctica, appreciable thicknesses of little known Tertiary glacigenic strata, probably correlative (at least in part) with the Pagodroma Group, occur onshore notably in the Vestfold Hills, in Enderby Land, and in the Sør Rondane Mountains of Dronning Maud Land. In the Vestfold Hills the early Pliocene Sørsdall Formation (Adamson & Pickard 1986, Quilty *et al.* 2000, Whitehead *et al.* 2001), at least 10 m thick, consists of sandy diatomite, diatomaceous sand and minor sandstones and diamicton. Deposition in a shallow marine embayment is envisaged (Whitehead *et al.* 2001). In Enderby Land, the Tula Moraine (Hayashi 1990) is widely distributed in low coastal settings in the Mount Riiser–Larsen region. At least 30 m thick, this diamict is a basal till of Pleistocene or older age. In westernmost Enderby Land on the Soya Coast of Lutzow-Holm Bay, the Osen Glacial Bed of Yoshida (1983) crops out in the Skarvnes area and consists of at least 10 m of lodgement till occupying a shallow glacial trough. Its age is not known.

Within the Sør Rondane Mountains, the setting of the oldest glacigenic sequences together with the exposure ages of the underlying basement rocks (Moriwaki *et al.* 1992), tempts direct comparison with Pliocene and Miocene strata of the Pagodroma Group. However these Sør Rondane strata record terrestrial glacial deposition and so are more comparable palaeogeographically to the terrestrial sequences of the southern

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Prince Charles Mountains (Whitehead *et al.* 2000), rather than to the Pagodroma Group.

Thus throughout East Antarctica it is only the Sirius Group sequences of the Transantarctic Mountains which by way of their settings, thicknesses and presumed antiquity, appear to be comparable to the Pagodroma Group. Stroeven (1997) emphasizes that the youngest Sirius Group sequences, of a presumed Pliocene age indicated by recycled or *in situ* diatom floras (Webb *et al.* 1996, Harwood & Webb 1998), lie within the current ice drainage system. This contrasts with the older Sirius Group sequences, which often occur at higher altitudes and in locations which are not compatible with the present drainage topography. A similar comparison can perhaps be made between the Mount Johnston Formation and the three younger formations of the Pagodroma Group. However in as much as it is wholly marine, and consists of such relatively thick Cenozoic successions, the Pagodroma Group appears at present to be unique in the onshore geology of East Antarctica.

Conclusions

The newly defined Pagodroma Group in the northern Prince Charles Mountains includes the Mount Johnston, Fisher Bench, Battye Glacier and Bardin Bluffs formations. Each is separated spatially, but all are dominated by diamicts of proximal glaciomarine origin. Their ages range from early Miocene or Oligocene, through middle Miocene to Pliocene–early Pleistocene, on the basis of well-preserved marine microfossils (foraminifera, a radiolarian, diatoms). The oldest formation (the Mount Johnston Formation) has been uplifted *c.* 1400 m, whereas the base of the youngest formation (the Bardin Bluffs Formation) still remains close to sea level. Each formation of the Pagodroma Group rests on a glacially eroded surface of considerable relief, which in most localities is recognisable as part of the floor or lower wall of a palaeofjord. The erosion of the palaeofjords is a consequence of glacial expansion. The individual formations, each overlying a palaeofjord floor, record subsequent glacial recession. Thus the Pagodroma Group demonstrates ice–front fluctuations of the Lambert Glacier–Amery Ice Shelf drainage system, on a scale of several hundred kilometres in the Neogene (Hambrey & McKelvey 2000b). This concept is compatible with the model of a dynamic East Antarctic ice sheet in the Pliocene (and in earlier times), proposed by Webb *et al.* (1984) and Harwood & Webb (1998), through studies of the Sirius Group in the Transantarctic Mountains.

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