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Cenozoic glacigene sedimentation and erosion at the Menzies Range, southern Prince Charles Mountains, Antarctica

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ABSTRACT. The Menzies Range in the southern Prince Charles Mountains, Antarctica, records at least four intervals of Cenozoic terrestrial glacigene sedimentation, and two periods of glacial erosion. The oldest Cenozoic strata, here named the Pardoe Formation, are >240 m thick, and consist of variable diamicts with subordinate sandstones and minor laminated lacustrine siltstones. The Pardoe Formation overlies a rugged erosion surface cut into Precambrian basement. Two subsequent Cenozoic sequences are here named informally the Trail diamicts and the younger Amphitheatre diamicts. The latter infilled the lower regions of an extremely rugged erosion surface, many components of which still dominate the present topography. The palaeodrainage of this erosion surface is markedly discordant with that of the older erosion surface underlying the Pardoe Formation. These three depositional events and the two associated erosion surfaces record warmer climates and increased snow accumulation under conditions of temperate wet-based glaciation. During the excavation of the sub-Amphitheatre diamict erosion surface, the East Antarctic ice sheet was either absent, further inland or the height of its surface relative to the Menzies Range was considerably lower than at present. The fourth and youngest depositional episode, recorded by a veneer of boulder gravel distributed along the northern flank of the Menzies Range, is from dry-based glacier ice, and assumed to be <2.6 Myr.

1. INTRODUCTION

During the Australian National Antarctic Research Expedition (ANARE) 1997/98 field season, a reconnaissance survey was carried out in the southern Prince Charles Mountains to locate and study Cenozoic landforms and strata in the Menzies Range. The aim of this investigation was to determine the glacial thermal regimes indicated by the Cenozoic landforms and strata, and then infer from them the attendant Cenozoic palaeoclimate.

It is well established that temperature largely determines the way in which glaciers move, erode and deposit sediment (Drewry, 1986; Hambrey, 1994; McLane, 1995). Glaciers may be identified as either polar or temperate, although a continuum, via a subpolar or polythermal category, exists between these two models (Hooke, 1998). The bases of temperate glaciers are considered to be at the pressure-melting point of ice, and the glaciers are thus wet-based. Such glaciers have a very considerable capacity to erode their substrates and transport large volumes of sediment. In contrast, the colder polar glaciers are dry-based. These achieve little erosion, and transport negligible detritus (Hambrey, 1994; Bennett and Glasser, 1996). The East Antarctic ice sheet is traditionally considered to be polar and dry-based. However, the base of the thicker parts of the East Antarctic ice sheet is now known to be at the pressure-melting point and so essentially is wet-based. Fisher Glacier, at the head of the Lambert Glacier–Amery Ice Shelf drainage system, is one such example (Budd and others, 1970, 1971). But beneath thinner parts of the East Antarctic ice sheet, dry-based conditions persist, and so the ice sheet overall is best considered to be polythermal. Analysis of the various Cenozoic landforms and strata in the Menzies Range will indicate whether polar or temperate glaciers were operative, and so give some indication of the past palaeoclimates.

1.1. Geological setting

The southern Prince Charles Mountains are located up to 750 km inland from the coast of Prydz Bay, on either side of the Lambert Glacier–Amery Ice Shelf drainage system (Fig. 1). They consist of widely scattered nunataks composed largely of Precambrian metasediments and gneisses (Tingey, 1982, 1991) mantled by little-known Cenozoic glacigene sediments (Trail, 1964). Within the southern Prince Charles Mountains, the Menzies Range trends east–west (Fig. 1) and is ~65 km long. The range is bounded on its northern side by the eastwards-flowing Fisher Glacier. Fisher Glacier is ~2100 m thick, and its surface elevation descends along the range from 2100 to 1600 m (Morgan and Budd, 1975). The southern side of the range is bounded by the northeastwards-flowing Geysen Glacier, most of the surface of which is above 2000 m. Both glaciers merge at the eastern end of the range, and flow down into the head of Lambert Glacier, ~110 km further east. Two unnamed glaciers carry ice from Geysen Glacier through the Menzies Range to Fisher Glacier, and so

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separate off Mount Bayliss (~2025 m) and Mount Mather (~2340 m) from the central part of the range.

At the centre of the Menzies Range lie the two highest peaks, Mount Menzies (~3355 m) and Pardoe Peak (~2920 m), which are also the highest peaks in the Prince Charles Mountains. On either side of the central Menzies Range, deep but short cirque valleys open onto Geysen and Fisher Glaciers. Those on the southern side are back-filled with ice from Geysen Glacier. In contrast, those opening onto Fisher Glacier are essentially devoid of ice, except where Fisher Glacier ice flows laterally down into their mouths. The largest two of these cirque valleys join immediately northwest of Mount Menzies, to form the Amphitheatre (informal name). This composite cirque has sheer walls 1000–1500 m high (Fig. 1c).

The latitude and elevation of the Menzies Range results in a frigid climate. Average air temperatures are <–18°C in summer and <–35°C in winter (extrapolated from Allison, 1998). During the 1997/98 field season, liquid water was observed in only two ice-covered shallow ponds. The modern glacial conditions are dry-based, with negligible apparent erosion and deposition.

1.2. Previous work

Few details of the Menzies Range geology are known. The greatest part of the range consists of Archaean metasediments, including metaquartzites and metapelites, intruded by Proterozoic tholeiitic (now lower amphibolite facies) sills and dykes (Tingey, 1982, 1991). Complex deformation includes recumbent folding. Mount Bayliss consists in the main of yet older Archaean augen gneiss, cataclastic granite and amphibolite (Tingey, 1982, 1991).

Cenozoic glacigene strata, the subject of this study, blanket much of the Archaean basement (Trail, 1964). Air-photo interpretation of the central Menzies Range by Derbyshire and Peterson (1978) revealed numerous moraine ridges, benches and trimlines, diverse-patterned ground and ice-cored moraines. These Cenozoic strata at Menzies Range constitute, at least in part, the proximal or basin-edge facies of a Lambert Glacier–Amery Ice Shelf basin sequence. The distal facies of the basin has been recently investigated in Prydz Bay, some 800 km to the north, by Ocean Drilling Program (ODP) 188 (Shipboard Scientific Party, 2001). The Pagodroma Group of the northern Prince Charles Mountains (Hambrey and McKelvey, 2000b) occupies an intermediate position in the basin.

2. STRATIGRAPHY

On the central Menzies Range, three Cenozoic sequences have been recognized and are here defined. These are, in order of decreasing age:

(1) Pardoe Formation, >240 m thick

(2) Trail diamict, >100 m thick

(3) Amphitheatre diamict, >10 m thick.

At present, only the Pardoe Formation is given formal stratigraphic status. It is not yet possible to designate and describe adequate type sections for the two succeeding sequences, so they are referred to informally as the Trail and Amphitheatre diamicts. More fieldwork will almost certainly lead to subdivision of the Trail diamict, with the recognition of at least one more stratigraphic unit (see section 3).

Most of the central Menzies Range Cenozoic strata are non-lithified. For this reason, exposure is generally poor, as the sequences are mantled by loose scree, and horizontal surfaces are modified by patterned ground development. The provenance of the Cenozoic strata appears to be dominated by the local Precambrian basement. However, further knowledge of the basement geology is required to confirm this.

Place-names within the Menzies Range are few. Pardoe Bench, the Amphitheatre and Thursday Cirque are informal names (Fig. 1), only recently submitted to the Australian Antarctic Place Names Committee. For convenience of description and discussion, we have identified numerically 22 locations and indicated their positions in Figure 2.

2.1. The Pardoe Formation

The Pardoe Formation is at least 240 m thick at location 1 (Fig. 3a and b). However, the best exposures so far known are within a 155 m section at location 8 (73°24′ S, 61°54′ E), which is here designated the type section (Fig. 4). The formation contains three contrasting facies. Variable diamicts are the predominant facies, with subordinate amounts of sandstone and laminated siltstone. These facies allow recognition within the formation of four members, all present in the several sections examined at locations 2, 3 and 8 (Fig. 3b).

Member 1, a matrix-supported diamict (~60% clasts), ranges in thickness from 15 m in the type section to 1.5 m at location 2, and only ~0.1 m at location 3. The average clast size is 2–20 cm.

Member 2 shows considerable variation, consisting of silts, fine sands, slightly lithified sandstone, and minor clast-rich muddy diamict (in the sense of Moncrieff, 1989). The unit ranges in thickness from 35 m in the type section (Fig. 4) to ~0.5 m at location 6. The sandstone tends to be massive, whereas the silts and fine sands vary from massive to laminated. The top of Member 2 is a marked disconformity, and at several localities both Members 1 and 2 have been completely removed by this erosion. For instance, at Pardoe Bench (i.e. location 2), the two members overlie the rugged unconformity surface at an altitude of 2100 m (Fig. 3b, cross-section 2). Member 3 overlies. However, in a closely adjacent section the same two members, underlain by basement, again occur at the base of Member 3, but at an altitude of 2300 m. The lack of intervening exposure makes it difficult to definitely establish the stratigraphic relationships between the two sections. We consider the relationships illustrated in Figure 3b (cross-section 2) to be the most likely. Implicit in this interpretation is that Member 1, mantling the rugged unconformity surface, passes both vertically and laterally to Member 2. Figure 3b suggests nearly 200 m of Member 2 strata have been removed. At location 6, Member 2 appears to rest directly upon basement.

Members 3 and 4 are similar indistinctly stratified diamicts. However, Member 3 averages ~50% clasts, whereas Member 4 averages ~70%. In both instances, the clasts are usually of cobble grade, although a few scattered blocks in Member 4 exceed 2 m in diameter. At location 8, Member 3 is ~110 m thick, and an incomplete Member 4 exceeds 45 m. At location 2, Member 3 is 150 m thick, and an incomplete Member 4 exceeds ~70 m. At location 1, only Member 4 is evident, and here it probably exceeds 200 m in thickness. The base of the sequence is not exposed. At locations 2 and 8, the basal part of Member 3 contains abundant soft-sediment clasts of laminated silt and fine sand, almost certainly

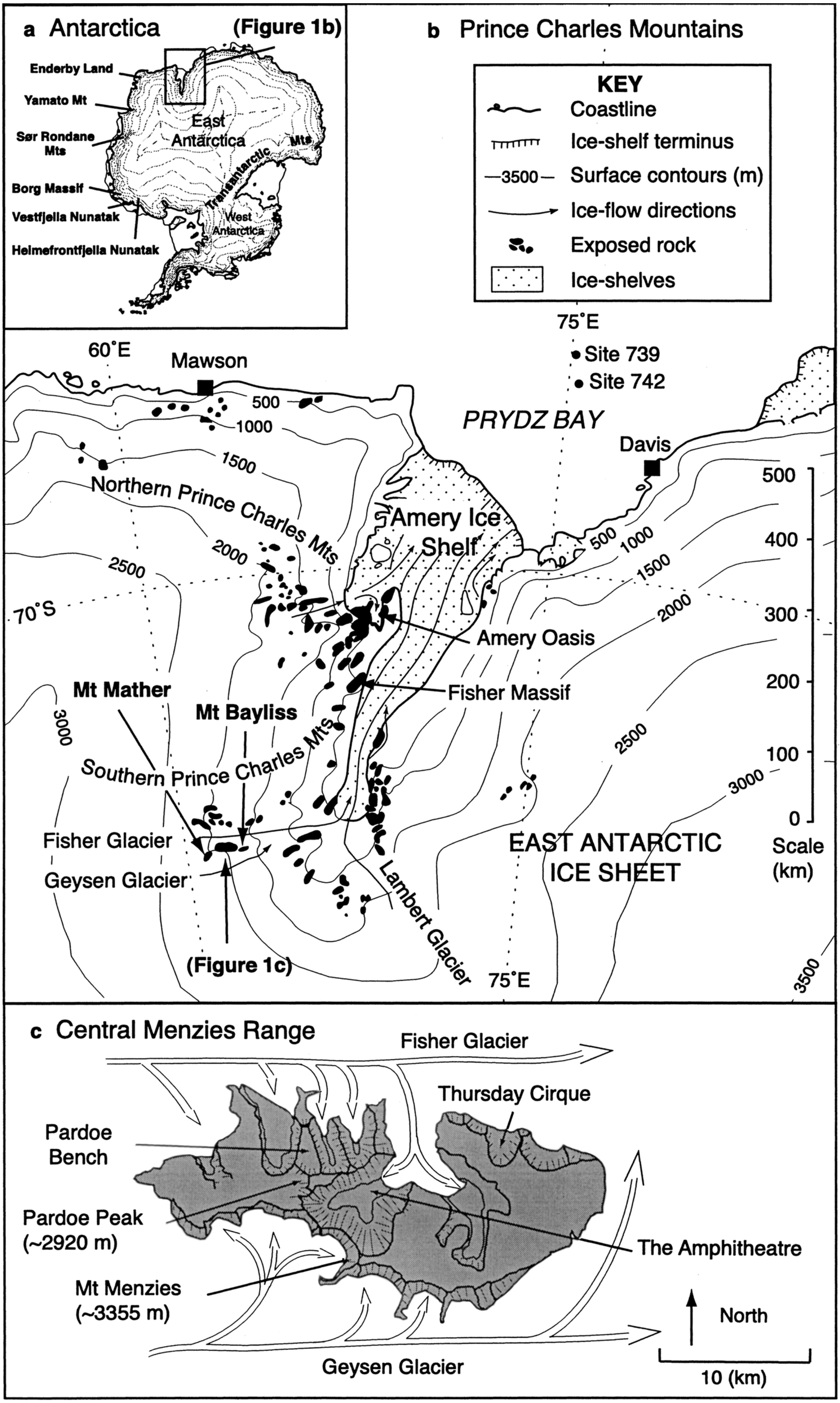

Fig. 1. (a) Location of Lambert Glacier–Amery Ice Shelf drainage system. (b) Prince Charles Mountains and the Lambert Glacier–Amery Ice Shelf drainage system. (c) Central Menzies Range.

derived from the underlying Member 2. At the base of Member 3 in the type section (Fig. 4), a 15 m thick diamict interval contains in places up to 80% of such intraformational clasts.

2.2. The sub-Pardoe Formation erosion surface

The Pardoe Formation overlies a rugged erosion surface cut into the Precambrian basement rocks. On the Menzies Range the altitudes of the exposed parts of this unconformity surface (i.e. locations 1, 2, 3, 5, 6, 7 and 8) range from <1700 to ~2800 m. Overall, these altitudes increase to the south and east, in the direction of the summit of Mount Menzies. In all cases, the unconformity surface is fresh and non-weathered. The precise palaeogeographic relationships between the individual surface remnants at the various localities are not always evident. However, some reconstruction is possible. Locations 1–3 reveal cross-sections of a single east–west-trending palaeovalley (Fig. 3b). This exhumed feature we refer to as the Pardoe palaeovalley. Its axis descends westwards over a distance of 6 km, from a height of 2150 m at location 3 to <1700 m at location 1 where the palaeovalley floor is obscured by Quaternary boulder gravels. At location 2, striations cut on the unconformity surface are compatible with the east–west trend of the palaeovalley axis.

At location 2, where the Pardoe Formation exceeds 200 m, the floor of the palaeovalley in cross-section is notably asymmetrical (Fig. 5). The southern wall is steep, whereas the northern wall is more gently inclined, and benched (Fig. 3b). Close by, to the southwest, the shallow cross-section of another palaeovalley is apparent at locality 6 (Fig. 2). This section is ~0.75 km across and is infilled with ~75 m of the Pardoe Formation. The palaeovalley floor, here at ~2500 m elevation, appears to descend gently northwest, towards the Pardoe palaeovalley. We surmise these two features were originally contiguous.

The palaeogeographic relationships between the unconformity surfaces preserved at locations 4, 5 and 8 are unclear. However, all are parts of palaeovalleys that drained to the north or northwest. The configuration of the basement rocks and the Pardoe Formation at location 8 is that of a buttress unconformity.

2.3. The Trail diamict

The Trail diamict ranges from cobble-grade, poorly sorted conglomerate to clast-rich, sandy diamict. These sediments mantle the eastern and western flanks of the central Menzies Range as an incomplete blanket, at altitudes of 1600–2200 m (Fig. 2c). A few boulders, within the diamict, exceed 3 m in diameter. The Trail diamict rests on Precambrian basement and has nowhere been observed to directly overlie the Pardoe Formation (Fig. 6). However, the abundance of clasts and coarse grade of the Trail diamict are reminiscent of Member 4 of the Pardoe Formation (Fig. 3b). Further work may show both to be stratigraphically the same unit, the Trail diamict constituting the youngest and much more widespread part of the Pardoe Formation

The Trail diamict is probably >100 m thick. It is difficult to make an accurate estimate, as the base of the unit is only exposed at the thinned (by erosion) edges of the blanket, along the northeastern margin of the central part of the range. Exposure is poor, for the diamict surface has been modified greatly by patterned ground development, and all scarps are mantled by scree. The surface morphology, includ-

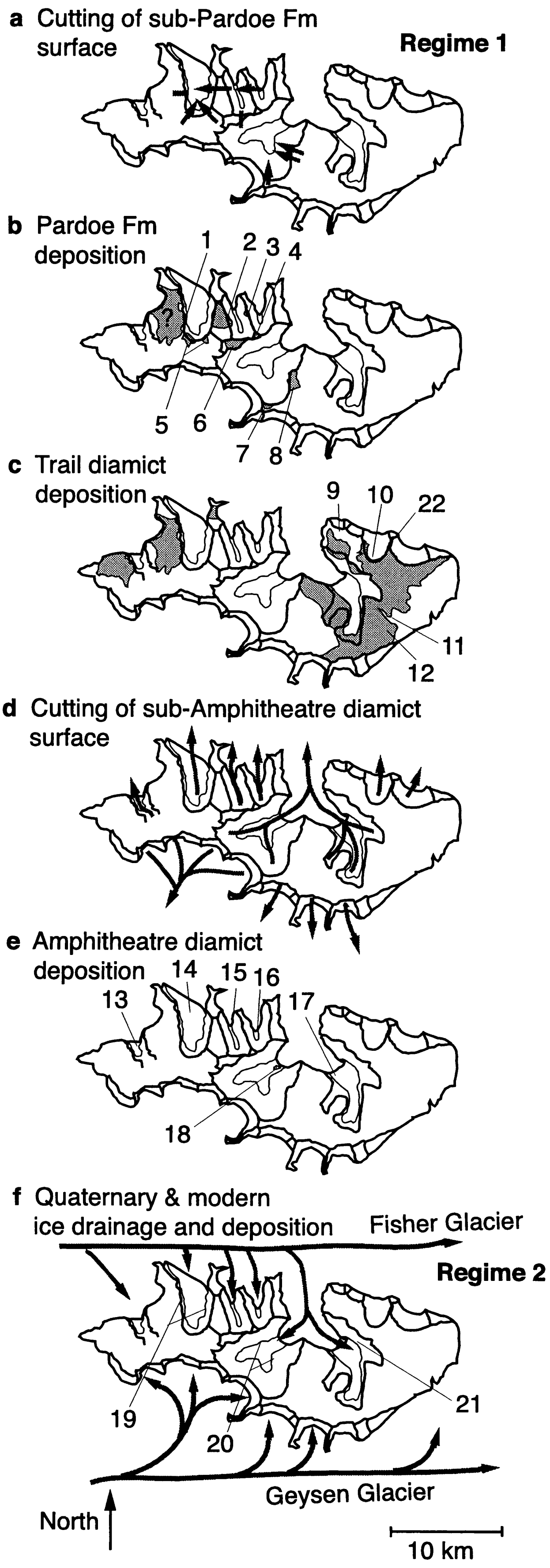

Fig. 2. Cenozoic stratigraphic and ice-flow history of the central Menzies Range. Pre-Quaternary deposits are hachured. Grey arrows summarize Cenozoic ice-flow directions. Black arrows indicate Quaternary–modern flow directions. Locality numbers are indicated. Maps (a–e) summarize Regime 1, that of pre-Quaternary temperate glaciation. (f) Quaternary dry-based polar glaciation and modern ice drainage.

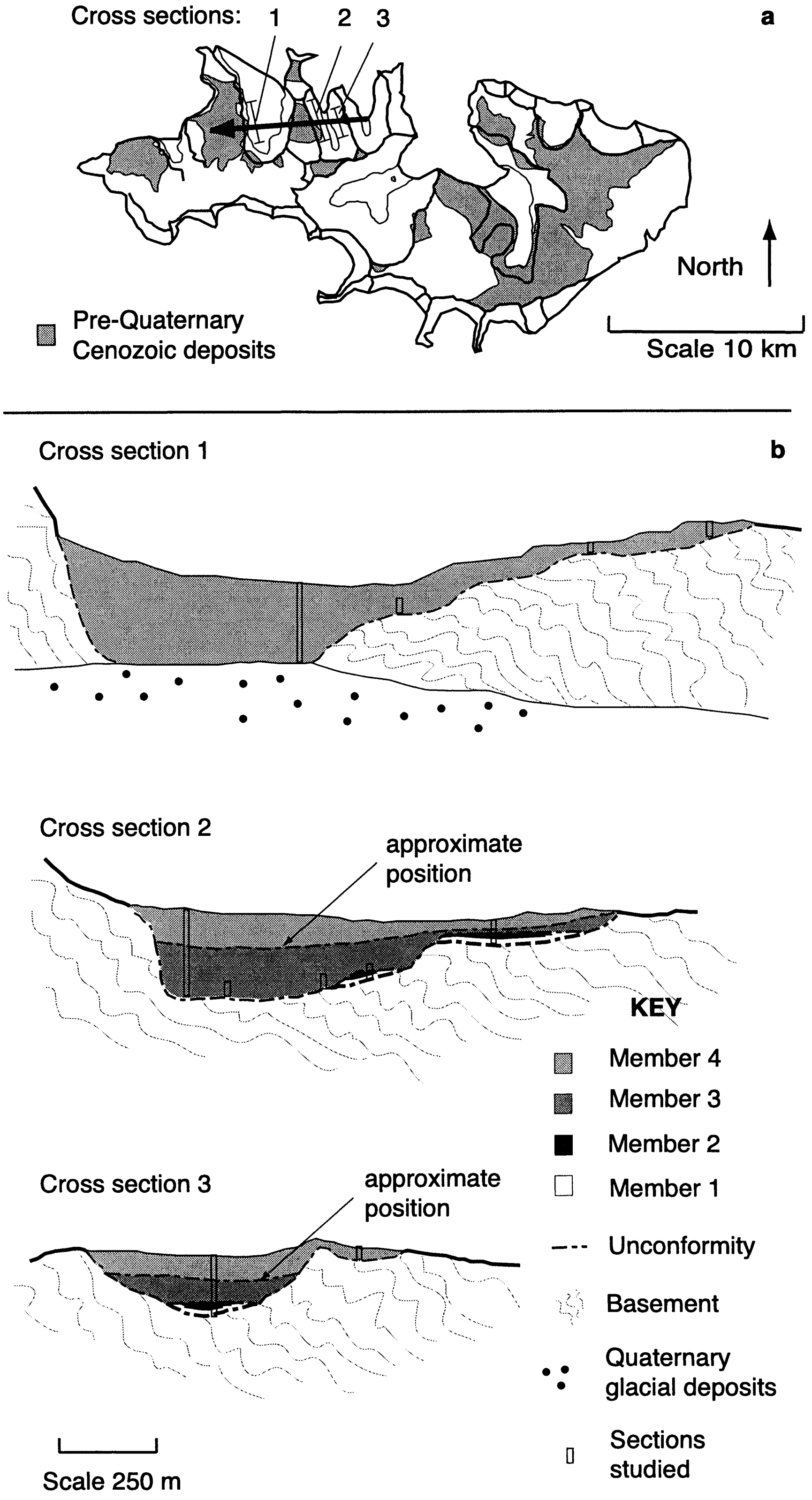

Fig. 3. (a) Distribution of pre-Quaternary deposits on the central Menzies Range. Arrow indicates axis of the Pardoe palaeovalley. Location of cross-sections illustrated in Figure 3b indicated. (b) Pardoe palaeovalley cross-sections, and locations of Pardoe Formation measured sections.

ing moraine ridges, indicates more than one depositional episode. There is further evidence of this at location 11 (Fig. 2c), where in the floor of an erosional depression in the Trail diamict the chemically weathered surface of a diamict unit is exposed, representing an interval of ice retreat and subaerial weathering. Non-weathered Trail diamict overlies. The basement surface underlying the Trail diamict is exposed intermittently along the northeastern margin of the central Menzies Range, at altitudes of 2000–2080 m (Fig. 6).

There are few data available from which to determine the configuration of the sub-Trail diamict erosion surface. From along the northeastern margin (i.e. locations 9, 22 and further eastwards) the surface falls gently southwards to less than ~1680 m, before rising again towards the crest of the Menzies Range. This suggests an elongate shallow depression, oriented northwest–southeast, and descending gently towards Fisher Glacier. However, in the back wall of Thursday Cirque a northeast–southwest-trending palaeovalley within the

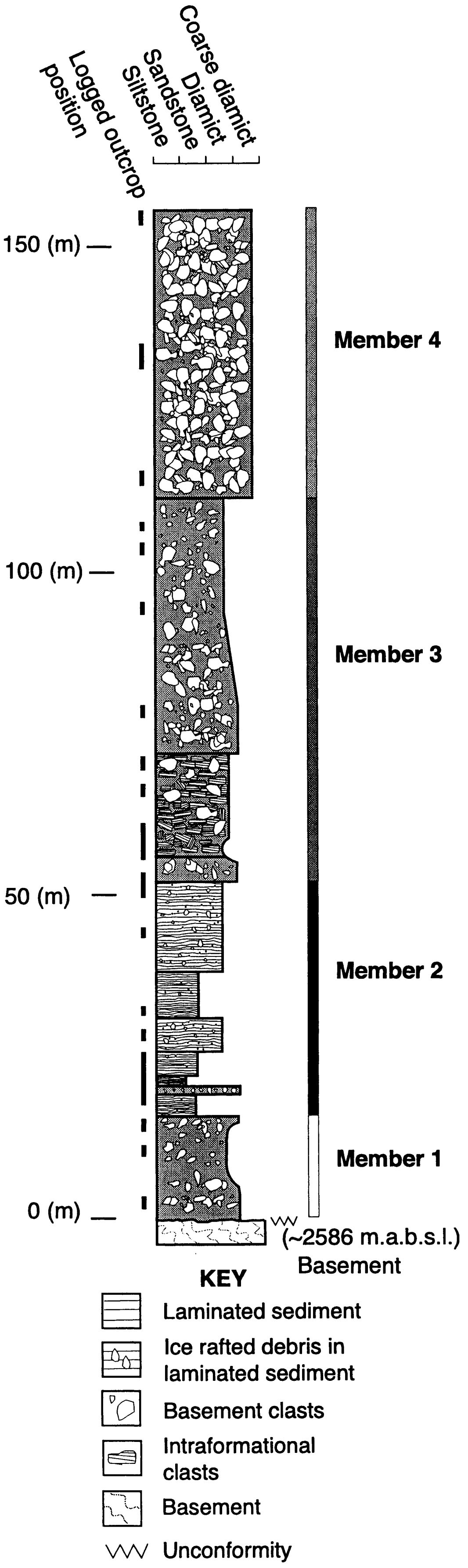

Fig. 4. The Pardoe Formation type section at location 8 (73°27′ S, 61°54′ E).

erosion surface is apparent (Fig. 2c, location 10). This smaller feature, ~80 m deep, appears to trend at a high angle to the northwest-oriented, shallow regional depression defined by the sub-Trail diamict erosion surface.

2.4. Amphitheatre diamict

In the floor of the Amphitheatre are scattered exposures of matrix-rich pebble grade diamicts with sub-rounded clasts (Fig. 2e, location 18). A noticeable feature is the presence of plutonic, as well as metamorphic, clasts. The matrix varies from fine to silty sand, and in places is indistinctly laminated. It has a distinctive greyish-green colour, a feature not seen in other Menzies Range diamicts. The Amphitheatre diamict is markedly more matrix-rich than those of the Pardoe Formation and the Trail diamict. Thin lenses of pebble conglomerate and sand are rare. These outcrops are the only ones known so far of the Amphitheatre diamict. It is not possible to estimate the stratigraphic thickness, but ~10 m is evident in exposures. We assume the Amphitheatre diamict also occupies the floors of the cirque valleys to the west and east of the Amphitheatre (Fig. 2e, locations 13–17), concealed beneath coarse Quaternary boulder gravels.

2.5. The sub-Amphitheatre diamict erosion surface

The erosion surface upon which the Amphitheatre diamict rests is part of a rugged landscape that includes the Amphitheatre and the other (north-facing) ice-free cirque valleys to the west and east (Fig. 2e, locations 13–17). Similar cirques developed on the southern side of the range, are presently infilled with ice from Geysen Glacier and are also part of the same surface. With a relief of up to 1500 m, this landscape is much more rugged than that underlying the Trail diamict. It is deeply incised into both the sub-Trail diamict erosion surface and that beneath the Pardoe Formation. The palaeodrainage directions of the erosion surface beneath the Amphitheatre diamict (Fig. 2d) are markedly discordant with that beneath the Pardoe Formation (Fig. 2a).

2.6. Quaternary cobble–boulder gravels

A veneer (<4 m) of cobble and boulder gravels, in the sense of Moncrieff (1989) and Hambrey (1994) and devoid of matrix, overlies the floors and lower walls of the ice-free cirque valleys, including the Amphitheatre. As such, this veneer defines a spectacular trimline along the north side of the range (see Tingey, 1991, fig. 23) (Fig. 7). Within the Amphitheatre the altitude of the trimline decreases gently southwards, from ~1880 m near the valley mouth to ~1720 m, 3 km up-valley. The gravels also drape the northern slopes of Mount Bayliss and similarly extend along the base of Mount Mather. It is not known whether they occur on the southern side of the Menzies Range.

3. DISCUSSION

Clearly the Pardoe Formation, Trail and Amphitheatre diamicts, and the Quaternary boulder–cobble gravels on the Menzies Range record at least four separate intervals of glacigene sedimentation. Furthermore the deposition of the Pardoe Formation and the Amphitheatre diamict both followed periods of deep glacial erosion. The discordance of the ice palaeodrainages recorded by these two periods of erosion is notable. Both these erosion surfaces, and the Pardoe Formation and the Amphitheatre and Trail diamicts, indicate conditions of wet-based alpine glaciation such as occur today in temperate alpine climates. However, the youngest depositional event, the Quaternary cobble–boulder gravels, was from polar dry-based ice, similar to that of the present

Fig. 5. Pardoe Bench or location 2 (~2100 m), viewed looking westwards from location 3. The cross-valley profile of the Pardoe palaeovalley, infilled with >200 m of Pardoe Formation, is indicated (arrows). The steep southern wall of the palaeovalley contrasts with the gently sloping and benched northern wall. The lobe of ice (bottom right) is lateral ice of Fisher Glacier flowing southwards down into a cirque valley. Height to bench surface ~450 m.

glacial regime. Several features of the Cenozoic geology of the Menzies Range merit further discussion.

3.1. The Pardoe palaeovalley

Since the floor of the Pardoe palaeovalley descends steadily westwards over a distance of 6 km, there can be little doubt that the ice within it similarly flowed westwards. This contrasts with the present eastward flow of nearby Fisher Glacier. The valley-floor remnant perched at location 6 appears to be part of a former tributary that fed down into the westwards-flowing palaeovalley ice.

It is not clear why the cross-section of the Pardoe palaeovalley is asymmetrical, as portrayed by the steep southern wall and contrasting gently inclined and benched northern wall (Fig. 3b, cross-section 2; Fig. 5). It is probable the northern wall records glacier narrowing and increased down-cutting in response to uplift. The contrasting steep southern wall may be a consequence of the proximity of Pardoe Peak, which would have caused increased snow accumulation along the southern side of the palaeovalley, and thus greater glacial erosion in this vicinity. Fault movement coincident with the southern wall is also a possible factor, but there is as yet no firm field evidence.

Fig. 6. Trail diamict overlying Precambrian basement atop the eastern wall of Thursday Cirque at ~2060 m. View looking north-northeast. Encircled figure indicates scale. Fisher Glacier (~1600 m), left background.

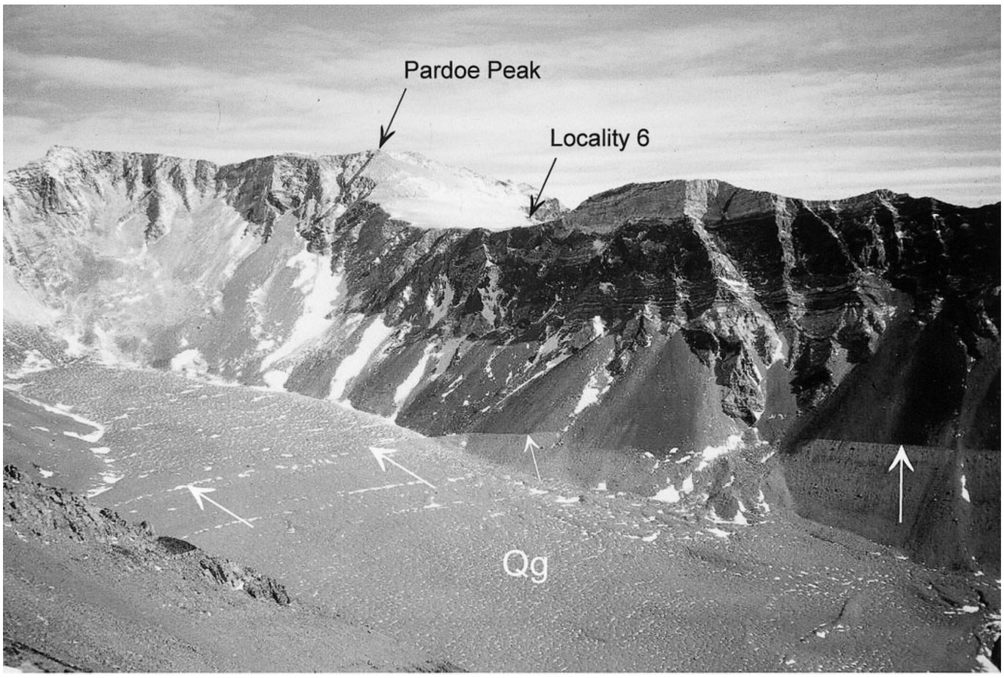

Fig. 7. View looking westwards into the southwestern head of the Amphitheatre. Pardoe Peak (2981 m) overlooks a perched cross-valley profile (location 6) oriented approximately orthogonal to the trend of the Amphitheatre floor, some 600 m below. The floor of the Amphitheatre ascends gently, over 4 km, from 1500 m (bottom right) to 1820 m (left middle-distance). Arrows indicate trimline delineated by Quaternary gravels (Qg) mantling the floor and far wall of the Amphitheatre. Rock outcrops (left foreground) are 700 m above the floor of the Amphitheatre.

3.2. Depositional setting of the Pardoe Formation

The various diamict facies of the Pardoe Formation clearly indicate wet-based (temperate) glacial sedimentation. Within the diamicts, the fine matrix (30–50%) and the sub-rounded or sub-angular clast shapes point to abrasion by basal-sliding ice. The presence of soft-sediment intraformational clasts within the oldest Member 3 diamicts further illustrates basal (i.e. wet-based) glacial erosion. That Member 4 alone infills Pardoe palaeovalley at location 1 suggests erosion of at least Member 3 prior to Member 4 deposition.

Within Member 2 the absence of obvious cross-stratification and only moderate sorting in the sandstone, fine sands and silts suggests a low-energy, aquatic regime. The presence of ice-rafted debris, as in the thin clast-rich muddy diamicts of Member 2 in the type section (Fig. 4), signals an ice-influenced climate. The rugged morphology of the sub-Pardoe Formation erosion surface and the apparent complete absence of marine fossils suggest an initial alpine glaciolacustrine setting.

3.3. Quaternary dry-based cobble–boulder gravels

The veneer of boulder and cobble gravels, mantling the lower north-facing slopes of the Menzies Range, records the change from temperate, wet-based glaciation to cold, polar dry-based glaciation. The spectacular trimlines along the north side of the range (e.g. Fig. 2f, locations 19–21; Fig. 7) record a major episode of south-flowing lateral ice of Fisher Glacier, presumably during the Last Glacial Maximum. However, it is likely that more than one pulse of Quaternary dry-based ice is recorded about these slopes of the Menzies Range. The several discontinuous moraine ridges diversifying the boulder gravel veneer north of the trimline (Derbyshire and Peterson, 1978, fig. 2), suggest subsequent similar events or at least significant still-stands.

3.4. Palaeoglacial considerations

Polythermal conditions occur in Antarctica today where the ice sheet and its tributary glaciers are at the pressure-melting point, where the:

(1) ice is thick,

(2) surface temperature is relatively high,

(3) ice velocities are high, producing frictional heat, and

(4) snow accumulation is light or moderate (Sugden and John, 1984).

Column and flowline numerical models indicate that deeper parts of the Lambert Graben tributary glaciers are at pressure-melting point (Budd and others, 1970). Even during the Last Glacial Maximum there may have been 40–60% basal thaw in this region (Hughes, 1998). However, recent deposits exposed in the Sør Rondane Mountains associated with East Antarctic ice-sheet basal slip are thin (Hasegawa and others, 1992), and low terrigenous sedimentation rates occur in Prydz Bay (Domack and others, 1991), suggesting that modern glacial erosion and deposition is minor. This contrasts with the wet-based landforms and deposits preserved on the Menzies Range and indicates a major transition in glacial conditions associated with decreasing:

(1) surface temperature,

(2) ice accumulation rates,

(3) ice-flow velocity, and

(4) frictional heat at the base of the ice (lower sliding velocity, less basal shear stress and less basal sediment entrainment).

The Cenozoic temperate glacial events recorded on the Menzies Range identify a markedly warmer Cenozoic climate than presently exists. Furthermore the large amount

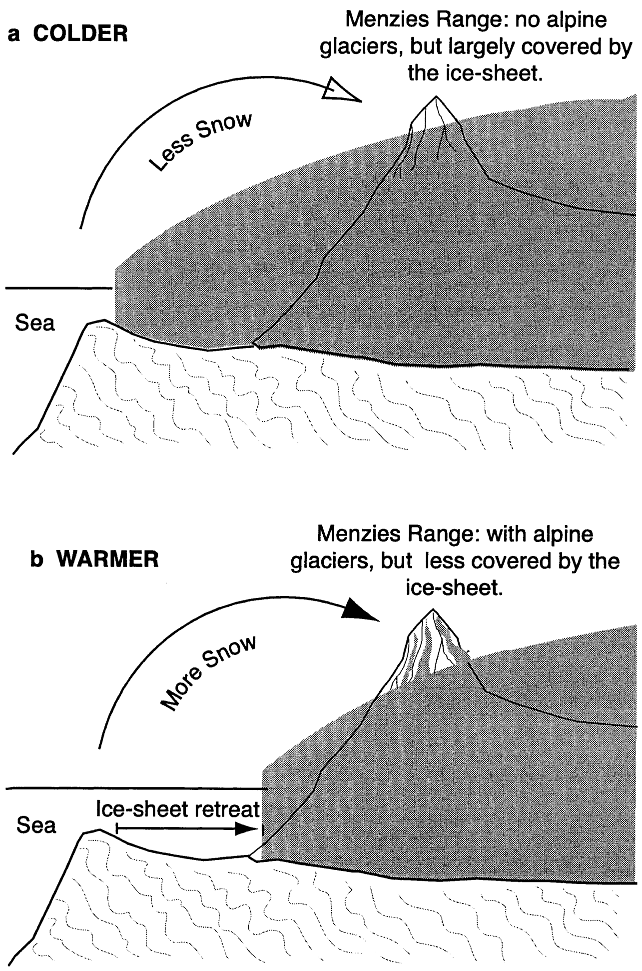

Fig. 8. (a) Elevated surface of the East Antarctic ice sheet encroaches onto the Menzies Range. Conditions of polar (dry-based) glaciation inhibit snow precipitation and thus prevent development of alpine glaciers on the Menzies Range. (b) Warmer conditions diminish the ice sheet and so lower its surface. Closer proximity to open waters promotes snow precipitation and the development of alpine glaciers at the Menzies Range.

of glacial excavation represented by the erosion surfaces beneath the Pardoe Formation and the Amphitheatre diamict indicates higher snow precipitation at that time, for no appreciable (i.e. erosionally competent) alpine glaciers exist today on the Menzies Range.

The elevation of the ice sheet surrounding the Menzies Range prior to the Quaternary is difficult to assess. The current presence of Geysen Glacier ice back-filling the cirque valleys on the southern side of the range strongly suggests a lower ice-sheet level at the time of their cutting. However, should the Trail diamict have been deposited by ice from the Geysen and Fisher glaciers, then a higher regional ice-sheet level at that time is indicated. Alternatively, the present setting of the Trail diamict may reflect subsequent tectonic uplift.

It is now well established that during cold glacial intervals snow precipitation in Antarctica decreases, due to reduced evaporation in the surrounding marine environment. This in turn leads to the retreat of alpine glaciers on Antarctic nunataks and coastal ranges, but there is only a relatively minor decline in the thickness of the interior ice sheets (Jouzel and others, 1989). In contrast, during a period of pronounced warming and sea-level rise the surface of an ice sheet is drawn down because of increased coastal rafting. This means that the ice levels in the Fisher and Geysen Glacier valleys would be lowered, as would also happen throughout the fjords of the Lambert Glacier–Amery Ice Shelf drainage system. Here because of the increased calving, the Amery Ice Shelf's coastal margin would recede southwards. The consequent deeper penetration of ocean waters into the fjord system would result in increased marine evaporation and snow precipitation throughout the northern and southern Prince Charles Mountains, similar to the Antarctic coastal response to climate warming envisaged by Domack and others (1991) and Zwally (1994). Such a scenario accounts for the development of competent wet-based glaciers on a Menzies Range then surrounded by lower ice surfaces than at present (Fig. 8). Alternately, alpine glaciers may have occurred on the Menzies Range prior to the existence of the East Antarctic ice sheet and created the landforms preserved.

The Menzies Range region currently experiences some of the lowest snow accumulation in Antarctica (<20 mm w.e. a^{-1}) (Giovinetto and Bentley, 1985). The snow-accumulation rate would have been much higher during alpine glaciation of the Menzies Range. Alpine glaciers occur on the Antarctic Peninsula where snow accumulation exceeds 400 mm w.e. a^{-1} (Giovinetto and Bentley, 1985). Similar, or even higher, snow-accumulation rates may have occurred during alpine glacial erosion on the Menzies Range.

3.5. The age of the Menzies Range Cenozoic landforms and strata

It is surmised that the earliest Cenozoic ice on Antarctica was in the form of alpine glaciers on the Gamburtsev Mountains, and other mountainous regions in the Antarctic interior such as the Prince Charles Mountains (Drewry, 1975; Hambrey and Barrett, 1993; Huybrechts, 1993). These alpine glaciers eventually coalesced and developed into the ancestral East Antarctic ice sheet (Drewry, 1975; Hambrey and Barrett, 1993), from which much ice flowed down into Prydz Bay via the Lambert fjord system (Barker and others, 1998; Hambrey and McKelvey, 2000a). The oldest evidence for Antarctic glaciation comes from late Eocene to early Oligocene glacial sediments in Prydz Bay, at ODP sites 739 and 742 (Hambrey and others, 1994). Since the Menzies Range lies some 650 km south of the latitude of these two sites, it follows that it too became glaciated at about that time.

Approximately 250 km to the northeast of the Menzies Range, in the northern Prince Charles Mountains, palaeontological evidence indicates that Cenozoic glacial marine strata there range from Lower Miocene (or older) to Plio-Pleistocene (Hambrey and McKelvey, 2000b; McKelvey and others, 2001; Whitehead and McKelvey, 2001). Elsewhere in East Antarctica, Cenozoic temperate glaciated landforms and terrestrial strata, similar to those of the Menzies Range, have been described from the Transantarctic Mountains (Webb and Harwood, 1991; Webb and others, 1996; Stroeven, 1997), western Dronning Maud Land (Holmlund and Näslund, 1994) and further east in the Sør Rondane (Moriwaki and others, 1992b) and Yamato Mountains (Yoshida, 1983).

Extensive terrestrial wet-based glacial deposits, similar to the Trail diamict, were deposited in the Transantarctic Mountains as the Sirius Group. Palaeontological age data from the Sirius Group range from Miocene to Plio-Pleistocene (Webb and others, 1996). However, radiometric age data indicate that some Sirius Group strata are older, though still of Neogene age (Sugden and others, 1995; Marchant and others, 1996). In the Sør Rondane Mountains, the Stage 3 and 4 tills of Moriwaki and others (1992b) also resemble the setting and facies of the Trail diamict and are thought to be Pliocene in age.

Wet-based glacier cirques and valleys, such as the Amphitheatre and other cirque valleys of the Menzies Range, are preserved throughout East Antarctica. However, some of these features have formed at different times. The excavation of Wright Valley in the Dry Valleys region of southern Victoria Land occurred >9 Myr ago (Prentice and others, 1993), whereas alpine glacial valleys in the Sør Rondane Mountains date from the mid- to late Pliocene (Moriwaki and others, 1992b). Similar features also occur ~300 km east of the Sør Rondane Mountains, in the Yamato Mountains (Yoshida, 1983). Other pre-Quaternary alpine glacial valleys in Dronning Maud Land occur at Borg Massif (Brunk, 1989) and beneath the East Antarctic ice sheet, near Vestfjella and Heimefrontjella nunataks (Holmlund and Näslund, 1994). It seems reasonable therefore to attribute the temperate landforms and strata of the Menzies Range to the period of temperate glaciation that produced these other post-40 Myr glacigene features in East Antarctica. A precise age assessment of any of the Menzies Range landscape features and the related stratigraphic sequences is not yet possible.

An upper age limit is suggested for the temperate glacial landforms and strata, by the youngest strata in the Menzies Range which are a veneer of dry-based glacier cobble–boulder gravels. The datum identified globally at 2.6 Myr and considered to herald the onset of bipolar glaciation (Morrison and Kukla, 1998) is also considered to mark the change from wet-based and polythermal glacial erosion and deposition to cold polar dry-based conditions (Moriwaki and others, 1992a; Webb and Harwood, 1993; Wilson, 1995). An alternative mid-Miocene date has also been given for this change in glacial conditions (Sugden and others, 1999). We therefore consider the two periods of Cenozoic erosion on the Menzies Range, and the deposition of the Pardoe Formation and the Amphitheatre and Trail diamicts, at least preceded the 2.6 Myr datum and subsequent deposition of Quaternary boulder–cobble gravels. Finally, given the established pre-Pleistocene antiquity of many of the temperate glacial landscape features and strata in East Antarctica, we think it unlikely the similar features in the Menzies Range represent one or a number of the episodes of Antarctic Pleistocene warming identified by Scherer (1991), Barrett and others (1998), Bohaty and others (1998) and Scherer and others (1998).

4. CONCLUSIONS

The Cenozoic geology of the Menzies Range records two very different climatic regimes. Glacial landforms and deposits produced under conditions of temperate wet-based glaciation on the Menzies Range postdate ~40 Myr. The subsequent veneer of polar dry-based glacier gravels dates from 2.6 Myr or later. At least four intervals of Cenozoic terrestrial glacigene sedimentation and two periods of glacial erosion can be identified. The oldest Cenozoic strata have been named here the Pardoe Formation, which is >240 m thick and consists of diamicts with subordinate sandstones and laminated siltstones. The Pardoe Formation overlies a rugged erosion surface cut into Precambrian basement by alpine glaciation. Two subsequent Cenozoic sequences are here named the Trail diamicts and the younger Amphitheatre diamicts. The latter infill the lower regions of an extremely rugged alpine glacial erosion surface, many components of which still dominate the present topography. These three depositional events and the two associated erosion surfaces record warmer climates and increased snow accumulation under conditions of temperate wet-based glaciation. During alpine glaciation that excavated the sub-Amphitheatre diamict erosion surface, the East Antarctic ice sheet had not yet formed, was further inland or its surface elevation relative to the Menzies Range was considerably lower than at present. The fourth and youngest depositional episode, recorded by a veneer of boulder gravel distributed along the northern flank of the Menzies Range, is from dry-based glacier ice that formed under glacial conditions like today's.

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REFERENCES

Allison, I. 1998. Surface climate of the interior of the Lambert Glacier basin, Antarctica, from automatic weather station data. *Ann. Glaciol.*, **27**, 515–520.

Barker, P. F. *and 8 others*. 1998. Ice sheet history from Antarctic continental margin sediments: the ANTOSTRAT approach. *Terra Antartica*, **5**(4), 737–769.

Barrett, P. J., C. Fielding and S. W. Wise. 1998. Initial report on CRP-1, Cape Roberts Project, Antarctica. *Terra Antartica*, **5**(1), 187.

Bennett, M. R. and N. F. Glasser. 1996. *Glacial geology: ice sheets and landforms.* Chichester, etc., John Wiley & Sons.

Bohaty, S. M., R. P. Scherer and D. M. Harwood. 1998. Quaternary diatom biostratigraphy and palaeoenvironments of the CRP-1 drillcore, Ross Sea, Antarctica. *Terra Antartica*, **5**(3), 431–453.

Brunk, K. 1989. Large-scale geomorphologic–glaciological mapping of the arid high-polar Borgmassivet, New Schwabenland, Antarctica. *Ber. Polarforsch.* 66, 85–102.

Budd, W., D. Jenssen and U. Radok. 1970. The extent of basal melting in Antarctica. *Polarforschung, Band 6*, **39**(1), 1969, 293–306.

Budd, W. F., D. Jenssen and U. Radok. 1971. Derived physical characteristics of the Antarctic ice sheet. *ANARE Interim Rep., Ser. A(IV), Glaciology* 120. (Meteorology Dept. Publ. 18.)

Derbyshire, E. and J. A. Peterson. 1978. Photo-geomorphic map of the Mt. Menzies nunatak, Prince Charles Mountains, Australian Antarctic Territory. *Z. Gletscherkd. Glazialgeol.*, **14**(1), 17–26.

Domack, E. W., A. J. T. Jull and S. Nakao. 1991. Advance of East Antarctic outlet glaciers during the hypsithermal: implications for the volume state of the Antarctic ice sheet under global warming. *Geology*, **19**(11), 1059–1062.

Drewry, D. J. 1975. Initiation and growth of the east Antarctic ice sheet. *J. Geol. Soc. London*, **131**(3), 255–273.

Drewry, D. 1986. *Glacial geologic processes.* London, Edward Arnold.

Giovinetto, M. B. and C. R. Bentley. 1985. Surface balance in ice drainage systems of Antarctica. *Antarct. J. U.S.*, **20**(4), 6–13.

Hambrey, M. J. 1994. *Glacial environments.* London, University College Press.

Hambrey, M. J. and P. J. Barrett. 1993. Cenozoic sedimentary and climatic record, Ross Sea region, Antarctica. *In* Kennett, J. P. and D. A. Warnke, *eds. Antarctic paleoenvironment: a perspective on global change. Part 2.* Washington, DC, American Geophysical Union, 91–124. (Antarctic Research

Series 60.)

Hambrey, M. J. and B. C. McKelvey. 2000a. Major Neogene fluctuations of the East Antarctic ice sheet: stratigraphic evidence from the Lambert Glacier region. *Geology*, **28**(10), 887–890.

Hambrey, M. J. and B. C. McKelvey. 2000b. Neogene fjordal sedimentation on the western margin of the Lambert Graben, East Antarctica. *Sedimentology*, **47**, 557–607.

Hambrey, M. J., W. U. Ehrman and B. Larsen. 1994. The Cenozoic sedimentary record of the Prydz Bay continental shelf, East Antarctica. *Terra Antartica*, **1**(2), 399–402.

Hasegawa, H., S. Iwata and N. Matsuoka. 1992. Observations of clayey till and underlying glacier ice in the central Sør Rondane Mountains, East Antarctica. *In* Yoshida, Y., K. Kaminuma and K. Shiraishi, *eds. Recent progress in Antarctic earth sciences*. Tokyo, Terra Scientific Publishing Co., 679–681.

Holmlund, P. and J.-O. Näslund. 1994. The glacially sculptured landscape in Dronning Maud Land, Antarctica, formed by wet-based mountain glaciation and not by the present ice sheet. *Boreas*, **23**(2), 139–148.

Hooke, R. LeB. 1998. *Principles of glacier mechanics*. Upper Saddle River, NJ, Prentice Hall.

Hughes, T. J. 1998. *Ice sheets*. New York, etc., Oxford University Press.

Huybrechts, P. 1993. Glaciological modelling of the Late Cenozoic East Antarctic ice sheet: stability or dynamism? *Geogr. Ann.*, **75A**(4), 221–238.

Jouzel, J. *and 9 others*. 1989. A comparison of deep Antarctic ice cores and their implications for climate between 65,000 and 15,000 years ago. *Quat. Res.*, **31**(2), 135–150.

Marchant, D. R., G. H. Denton, C. C. Swisher, III and N. Potter, Jr. 1996. Late Cenozoic Antarctic paleoclimate reconstructed from volcanic ashes in the Dry Valleys region of southern Victoria Land. *Geol. Soc. Am. Bull.*, **108**(2), 181–194.

McKelvey, B., M. Hambrey, D. Harwood, M. Mabin, P. Webb and J. Whitehead. 2001. The Pagodroma Group. The Neogene record in the northern Prince Charles Mountains of a dynamic Lambert Glacier and East Antarctic ice sheet. *Antarct. Sci.*, **13**(4), 455–468.

McLane, M. 1995. *Sedimentology*. New York, Oxford University Press.

Moncrieff, A. C. M. 1989. Classification of poorly sorted sedimentary rocks. *Sediment. Geol.*, **65**(1–2), 191–194.

Morgan, V. I. and W. F. Budd. 1975. Radio-echo sounding of the Lambert Glacier basin. *J. Glaciol.*, **15**(73), 103–111.

Moriwaki, K., Y. Yoshida and D. M. Harwood. 1992a. Cenozoic glacial history of Antarctica — a correlative synthesis. *In* Yoshida, Y., K. Kaminuma and K. Shiraishi, *eds. Recent progress in Antarctic earth science*. Tokyo, Terra Scientific Publishing Co., 773–780.

Moriwaki, K., K. Hirakawa, M. Hayashi and S. Iwata. 1992b. Late Cenozoic history in the Sør Rondane Mountains, East Antarctica. *In* Yoshida, Y., K. Kaminuma and K. Shiraishi, *eds. Recent progress in Antarctic earth science*. Tokyo, Terra Scientific Publishing Co., 661–667.

Morrison, R. and G. Kukla. 1998. The Pliocene–Pleistocene (Tertiary–Quaternary) boundary should be placed at about 2.6 Ma, not at 1.8 Ma. *GSA Today*, **1**(8), 9.

Prentice, M. L. *and 6 others*. 1993. Late Neogene Antarctic glacial history: evidence from central Wright Valley. *In* Kennett, J. P. and D. A. Warnke, *eds. Antarctic paleoenvironment: a perspective on global change. Part 2*. Washington, DC, American Geophysical Union, 207–250. (Antarctic Research Series 60.)

Scherer, R. P. 1991. Quaternary and Tertiary microfossils from beneath Ice Stream B: evidence for a dynamic West Antarctica ice sheet history. *Palaeogeogr., Palaeoclimatol., Palaeoecol.*, **90**, 395–412.

Scherer, R. P., A. Aldahan, S. Tulaczyk, G. Possnert, H. Engelhardt and B. Kamb. 1998. Pleistocene collapse of the West Antarctic ice sheet. *Science*, **281**(5373), 82–85.

Shipboard Scientific Party. 2001. Leg 188 summary: Prydz Bay–Cooperation Sea, Antarctica. *In* O'Brien, P. E., A. K. Cooper and C. Richter, *eds. Ocean Drilling Program. Proceedings*. College Station, TX, Ocean Drilling Program, Texas A & M University, CD-ROM. (Init. Repts. 188.)

Stroeven, A.P. 1997. The Sirius Group of Antarctica: age and environments. *In* Ricci, C. A., *ed. The Antarctic region: geological evolution and processes*. Siena, Museo Nazionale dell'Antartide, 747–761. (Terra Antarctica Publication.)

Sugden, D. E. and B. S. John. 1984. *Glaciers and landscape. Second edition*. London, Edward Arnold.

Sugden, D. E., G. H. Denton and D. R. Marchant. 1995. Landscape evolution of the Dry Valleys, Transantarctic Mountains: tectonic implications. *J. Geophys. Res.*, **100**(B7), 9949–9967.

Sugden, D. E. *and 6 others*. 1999. Landscape development in the Royal Society Range, southern Victoria Land, Antarctica: stability since the mid-Miocene. *Geomorphology*, **28**, 181–200.

Tingey, R.J. 1982. Geologic evolution of the Prince Charles Mountains — an Antarctic Archean cratonic block. *In* Craddock, C., *ed. Antarctic geoscience*. Madison, WI, University of Wisconsin Press, 455–464.

Tingey, R. J., *ed.*. 1991. *Geology of Antarctica*. Oxford, Clarendon Press.

Trail, D. S. 1964. The glacial geology of the Prince Charles Mountains. *In* Adie, R. J., *ed. Antarctic geology*. Amsterdam, North Holland Publishing Co. Interscience, 143–151.

Webb, P.-N. and D. M. Harwood. 1991. Late Cenozoic glacial history of the Ross embayment, Antarctica. *Quat. Sci. Rev.*, **10**(2–3), 215–223.

Webb, P.-N. and D. M. Harwood. 1993. Pliocene fossil *Nothofagus* (southern beech) from Antarctica: phytogeography, dispersal strategies and survival in high latitude glacial–deglacial environments. *In* Alden, J., J. L. Mastrantonio and S. Ødum, *eds. Forest development in cold climates*. New York and London, Plenum Press, 135–165. (NATO ASI Series A: Life Sciences 244.)

Webb, P.-N., D. M. Harwood, M. C. G. Mabin and B. C. McKelvey. 1996. A marine and terrestrial Sirius Group succession, middle Beardmore Glacier–Queen Alexandra Range, Transantarctic Mountains, Antarctica. *Mar. Micro-Palaeontol.*, **27**(1–4), 273–297.

Whitehead, J. M. and B. C. McKelvey. 2001. The stratigraphy of the Pliocene–lower Pleistocene Bardin Bluffs Formation, Amery Oasis, northern Prince Charles Mountains, Antarctica. *Antarct. Sci.*, **13**(1), 79–86.

Wilson, G. S. 1995. The Neogene East Antarctic ice sheet: a dynamic or stable feature? *Quat. Sci. Rev.*, **14**(2), 101–123.

Yoshida, Y. 1983. Physiography of the Prince Olav and Prince Harald Coasts, East Antarctica. *Natl. Inst. Polar Res. Mem., Ser. C* 13.

Zwally, H. J. 1994. Detection of change in Antarctica. *In* Hempel, G., *ed. Antarctic science: global concerns*. Berlin, Springer Verlag, 126–143.

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