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# New grounds for reassessing palaeoclimate of the Sirius Group, Antarctica

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Fossil *Nothofagus* leaves from the Meyer Desert Formation of the Sirius Group in the central Transantarctic Mountains have been found with recycled marine diatoms indicating that they are no older than 4.1–3.1 Ma; they have been interpreted as evidence for Pliocene warming and diminution of the East Antarctic Ice Cap. In contrast, Plio-Pleistocene ice-cap stability and sustained polar palaeoclimate has been argued from landforms in Victoria Land and isotopic data from the Southern Ocean. New evidence on this question comes from fossil soils in the Meyer Desert Formation. One of these, the Siesta palaeosol, has a gypsic horizon at a depth of 85 cm as in soils of polar deserts with a mean annual precipitation (MAP) of 120–220 mm. Clastic dykes from a 60–80 cm thick permafrost active layer in the palaeosols are evidence of a mean annual temperature (MAT) of –3 to –11°C. This is warmer and wetter than the Dominion Range today (MAP 36 mm, MAT –39°C) but still a periglacial climate, cooler and drier than southern Chilean moorland and krummholz (MAP 600–7500 mm, MAT 5–8°C), or subantarctic tundra (MAP 400–2500 mm, MAT –5 to +5°C). A Pliocene age for the *Nothofagus* fossils and Siesta palaeosol is compatible with soil development on surfaces that bracket their enclosing deposits. The Dominion Range was significantly warmer and wetter than it is today during Pliocene (*c.* 3.5 Ma) growth of *Nothofagus* and formation of the Siesta palaeosol, but the Beardmore Glacier was still large at that time. Surface soils and palaeosols of the Meyer Desert Formation are evidence of another warmer and wetter episode earlier during the Pliocene (*c.* 5 Ma). Both warm intervals were separated by intervening times of glacial expansion and frigid palaeoclimate.

**Keywords:** Pliocene, Antarctica, Sirius Group, palaeosols, palaeoclimates.

Fossil leaves, insects, seeds and recycled marine diatoms from the Meyer Desert Formation of the Sirius Group in the Dominion Range, central Transantarctic Mountains, have been interpreted as evidence for a warm palaeoclimate (5–10°C mean annual temperature = MAT), and substantial deglaciation during the Pliocene (4.1–3.1 Ma; Wilson 1995; Hill *et al.* 1996; Ashworth *et al.* 1997; Harwood & Webb 1998). However, opponents of this interpretation cite evidence from marine micropalaeontology (Warnke *et al.* 1996) and from Antarctic geomorphology (Marchant *et al.* 1994; Schäfer *et al.* 1999; Denton 1999) for near-modern (± 3°C) frigid palaeoclimates of Antarctica for at least the last 8 Ma, and suggest that the fossil leaves may be middle Miocene or older (Burckle & Pokras 1991; Burckle & Potter 1996). Here we apply to this controversy evidence from soils on, and palaeosols in, the fossil-plant-bearing Meyer Desert Formation of the Dominion Range (Fig. 1).

Soils and palaeosols of high latitudes can yield information about such palaeoclimatic variables as precipitation, from the degree of weathering and leaching of soluble salts (Campbell & Claridge 1987), and palaeotemperature, from the degree and kind of periglacial deformation (Karte 1983). Climatic and ecological zonation of Arctic soils has been well known since the Russian origins of soil science in the late nineteenth century, and twentieth century research established comparable eco-climatic zonation of soils around Antarctica (Bockheim & Ugolini 1990; Bockheim 1995, 1997). Palaeosols also reveal ecological conditions from such features as root traces and the fossils they contain (Retallack 1997). Finally, the ages of geomorphic surfaces, and the sedimentary units that they cap, can be constrained by studies of soil variables such as depth of iron-staining, compared with well-dated chronosequences of ultraxerous and frigid soils extending back millions of years elsewhere in Antarctica (Campbell & Claridge 1987; Bockheim *et al.* 1986; Marchant *et al.* 1994). From this background in Antarctic soil science, the time required for formation of Dominion Range soils and palaeosols, and palaeoclimate under which these soils and palaeosols formed, can now be interpreted, independently of palaeontological and other evidence.

## Materials and methods

Meyer Desert is on the NE flank of the Dominion Range, a large nunatak at the junction of the Mill and Beardmore glaciers, 500 km from the South Pole (Fig. 1, Table 1). We sampled five palaeosols from 8–18 m below the surface soil in a steep fault scarp at two separate localities, representing two sediment-mantled glacial terraces, here called the upper and lower Oliver Platforms. The palaeosols were within glacial sediments mapped as 'Sirius and Dominion drifts' by Denton *et al.* (1989), but both glacial-terrace-capping sedimentary sequences are assigned to the Meyer Desert Formation, Sirius Group, by McKelvey *et al.* (1991). The Sirius Group is a complex sedimentary deposit of different geological age and palaeoenvironments in different parts of Antarctica (Webb *et al.* 1996, 1998). Our observations apply only to the Meyer Desert Formation in the Dominion Range, central Transantarctic Mountains (85°S 166°E).

Three distinct kinds of palaeosols were recognized, and have previously been named the Viento, Peligro and Siesta pedotypes (Retallack & Krull 1998). Viento and Peligro palaeosols (Figs 2 & 3; Table 2) were named for location of their type profiles high on

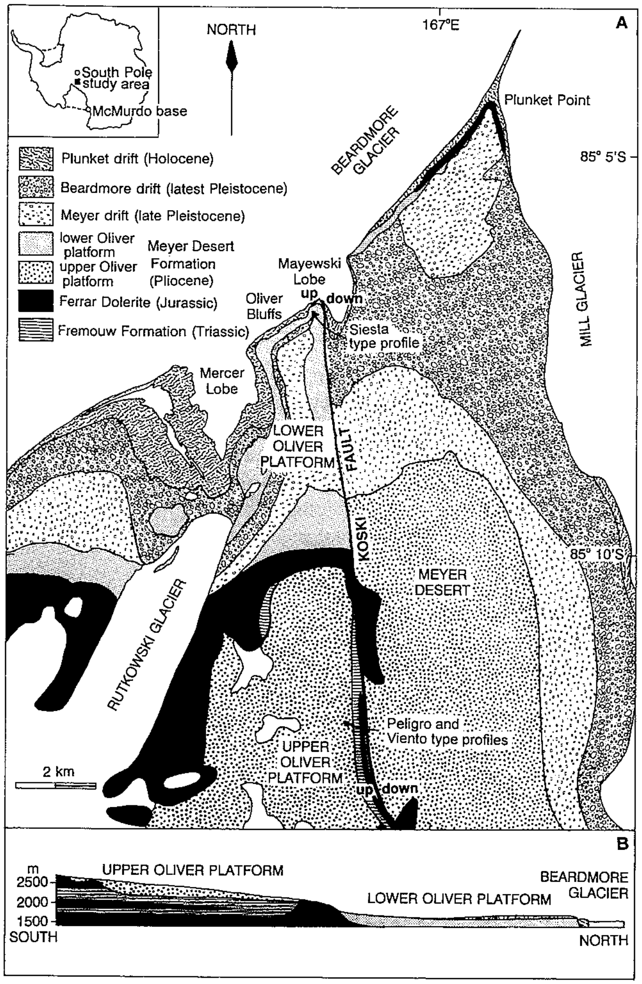


**Fig. 1.** Location of palaeosol type profiles within a map of glacial drifts, sediments, and bedrock in the Dominion Range, Antarctica (after Denton *et al.* 1989; McKelvey *et al.* 1991).

a dangerous and windy scarp of the Koski Fault that exposes sedimentary cover of the upper Oliver Platform, 8 km south of the Beardmore Glacier (85°11.4'S 166°47.1'E). Viento palaeosols also are common in sedimentary cover of the lower Oliver Platform in Oliver Bluffs to the north, capping member 1 and member 2, and 4 m above the base of member 4 of the Meyer Desert Formation of McKelvey *et al.* (1991). The principal locality for *Nothofagus* fossil leaves (Hill *et al.* 1996) is a Viento palaeosol. We have not yet found a Peligro palaeosol in the lower Oliver Platform at Oliver Bluffs, but suspect they were present as a source for reddish fluvial sandstones (member 2 of McKelvey *et al.* 1991) and as a stage in the development of strongly developed palaeosols such as the Siesta palaeosol. The Siesta pedotype (Figs 4–6) was named for location of its type profile out of the wind and warmed by the midnight sun, on a spur of lower Oliver Platform east of Oliver Bluff overlooking the Beardmore Glacier (85°6.8'S 166°43.2'E). Only one Siesta palaeosol was found, at 97 m in member 4, within the same 105 m thick sedimentary sequence of the lower Oliver Platform and Oliver Bluffs containing *Nothofagus* leaves and other fossils at 13–20 m (member 2 of McKelvey *et al.* 1991; Webb & Harwood 1993; Hill *et al.* 1996; Ashworth *et al.* 1997). Recycled marine diatoms associated with the *Nothofagus* fossils have been used to date that part of the succession as Pliocene (3.1–4.1 Ma: Hill *et al.* 1996; Bohaty & Harwood 1998). The Siesta and Viento palaeosols are within the Meyer Desert Formation of the lower Oliver Platform, which McKelvey *et al.* (1991) regard as incised into, and so geologically younger than, Peligro and Viento palaeosols of the Meyer Desert Formation of the upper Oliver Platform (Fig. 1, Table 1).

All the palaeosols studied are buried by several metres of gray till and sand that separate them from the potentially confounding effects of modern soil formation. Such shallow burial in a frigid climatic region has left them friable, unlithified, and little compacted. They are thus exceptionally well preserved and little affected by post-burial alterations that must commonly be considered in the palaeoenvironmental interpretation of palaeosols (Retallack 1997).

Both grain size and mineral composition were determined by point counting (500 points) petrographic thin sections. Thickness of weathering rinds of iron oxide and clay (diffusion sesquans in the terminology of Brewer 1976) were measured using vernier calipers on freshly broken clasts of dolerite only, as there were not sufficiently abundant clasts of sandstone or other lithologies. Molecular weathering ratios (also called molar ratios) were calculated from weight percent oxides of chemical analyses (determined by atomic absorption;

**Table 1.** *Sequence of geological events in upper Beardmore Glacier and Dominion Range*

| Events | Geological age (Ma) |
|---|---|
| **Formation of soil on Plunket drift**<br>Deposition of Plunket drift<br>*Glacial scour deeper into bedrock in modern courses* | <br>0.006–0.007 (Denton *et al.* 1989) |
| **Formation of soil on Beardmore drift**<br>Deposition of Beardmore drift<br>*Glacial scour deeper into bedrock east of Oliver Bluffs* | <br>0.013–0.024 (Denton *et al.* 1989) |
| **Formation of soil on Meyer drift**<br>Deposition of Meyer drift<br>*Glacial scour deeper into bedrock near current line of Oliver Bluffs* | **0.126 ± 0.024 (staining depth age)**<br>0.129–0.182 (Denton *et al.* 1989) |
| **Formation of soil on Meyer Desert Formation of lower Oliver Platform**<br>Deposition of Meyer Desert Formation of lower Oliver Platform (including *Nothofagus* growth in Viento palaeosol and formation of Siesta palaeosol)<br>*Glacial scour of bedrock bench under lower Oliver Platform* | **1.3 ± 0.2 (staining depth age)**<br>4.1–3.1 (Webb *et al.* 1996). including *c.* 1 Ma duration of palaeosol formation |
| **Formation of soil on Meyer Desert Formation of upper Oliver Platform**<br>Deposition of Meyer Desert Formation of upper Oliver platform (including formation of Viento and Peligro palaeosols)<br>*Glacial scour of bedrock bench under upper Oliver Platform* | **4.1 ± 0.8 (staining depth age)**<br>including *c.* 1 Ma duration of palaeosol formation |
| **Formation of soil on high level Sirius Group and bedrock** | **7.7 (3.9–10.3: Kurz & Ackert, 1997)** |

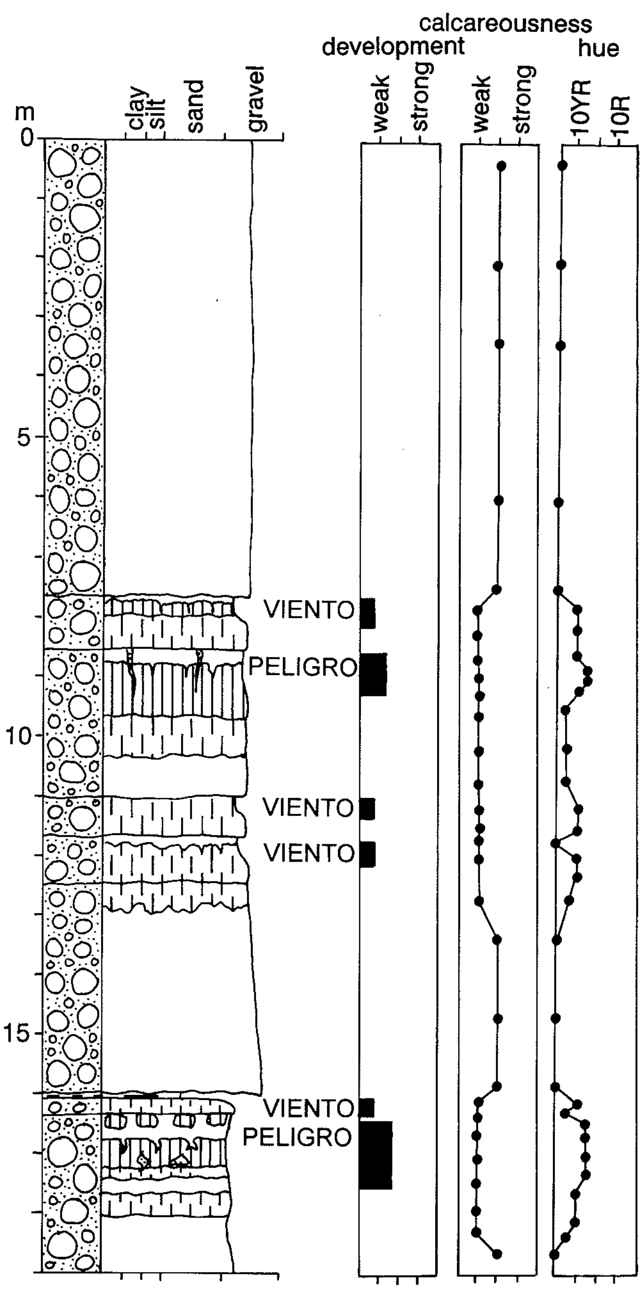


**Fig. 2.** Measured section of palaeosols in the Meyer Desert Formation of upper Oliver Platform, below the crest of the Koski Fault scarp, central Meyer Desert, Dominion Range (from Viento and Peligro type profile locality of Fig. 1). Black boxes represent the position of palaeosols, their width corresponding to degree of development. Scale of calcareousness is from reaction with dilute HCl (Retallack 1997) and hue is from Munsell colour charts. Lithological symbols are as for Figure 3.

Retallack & Krull 1998, table 1) divided by molecular weights of the respective oxides (Retallack 1997). Soil dilation was calculated from chemical analyses of trace elements and bulk densities were determined by the clod method, so that mass transfer and strain could be estimated (method of Brimhall *et al.* 1991). Clay minerals and gypsum were identified by XRD.

The palaeosols have been identified within several different soil classifications (Table 2), including the most recent version of Soil Taxonomy (Soil Survey Staff 1999), which now recognizes frigid soils as Gelisols. The most useful classifications for interpreting the palaeosols remain those used in Antarctic soil mapping (Campbell & Claridge 1987; Bockheim & Ugolini 1990; Bockheim 1997).

## Palaeoprecipitation implications of the palaeosols

Low palaeoprecipitation is compatible with several indications of modest weathering of the Siesta palaeosol. This palaeosol is evidence of palaeoclimate averaged over about a million years, which is the time for its formation inferred from the thickness of its gypsic horizon, depth of staining and other features that are comparable with well dated Antarctic soils (Stage 4 of Campbell & Claridge 1987). The most clayey and ferruginized horizon (20–84 cm) of the Siesta palaeosol is very modestly weathered compared with parent material. It has conservative ratios of $Al_2O_3/SiO_2$, $Al_2O_3/(CaO + MgO + K_2O + Na_2O)$ and Ba/Sr (Fig. 4). Little-weathered grains of hypersthene and rare clasts of limestone persist throughout the profile (Fig. 6b, c), as evidence for limited chemical weathering. The Siesta palaeosol shows little reaction with dilute (10%) HCl compared with its parent material, so has been leached of most carbonate. Clay minerals are mainly illite and chlorite, with only traces of smectite, vermiculite and kaolinite. Weathering rinds (sesquans) on dolerite pebbles within the profile are thin to absent (Figs 4, 6b). Resistate elements (Zr and Ti) decline in abundance upwards in the profile, and are evidence of dilation during soil formation (Fig. 4). Soil dilation and geochemical accumulation of most elements (Fig. 7) are probably due to inputs of fine-grained wind-blown dust in a dry and cold environment.

The Siesta palaeosol has an horizon 12 cm thick with small (1–2 mm) nodules and veins of fine-grained gypsum at a depth of 85 cm (Figs 4,8). This gypsic (By) horizon can be used to make an estimate of palaeoprecipitation. Gypsic horizons are deep within soil in areas with moderate mean annual precipitation (MAP) and shallow in the extremely dry climates of Israel, Morocco and Antarctica (Dan & Yaalon 1982; Campbell & Claridge 1987; Denton *et al.* 1989). Using the relationship between gypsic depth and precipitation established by Dan & Yaalon (1982), the Siesta palaeosol would have received some 120–220 mm MAP. This would be very arid in temperate or tropical latitudes, but is not too dry for seasonal growth of deciduous woody angiosperms at high latitudes where evapotranspiration is low (Walker & Peters 1977). This estimate for palaeoprecipitation is well above the current 30–40 mm MAP in the Meyer Desert of the central Transantarctic Mountains and the Dry Valleys of Victoria Land, where gypsic horizons comparable in thickness to that of the Siesta palaeosol are 28–42 cm from the surface (Bockheim *et al.* 1986; Campbell & Claridge 1987). Inferred palaeoprecipitation for the Siesta palaeosol is less than for Magellanic moors and alpine woodlands of southernmost Chile (MAP 600–7500 mm), which have been proposed as modern analogues for the palaeoenvironment of the Sirius Group (Mercer 1986; Webb & Harwood 1993). It is also less than for tundra of maritime subantarctic regions (MAP 400–2500 mm; Bockheim & Ugolini 1990) and for living *Nothofagus pumilio* and *N. antarctica* of Tierra del Fuego (at least MAP 400 mm; Moore 1983). Although the Siesta palaeosol contains stout root traces of likely woody vegetation, it may have been too dry for growth of *Nothofagus beardmorensis*, which has been found some 77 m lower within the same stratigraphic sequence (Francis & Hill 1996; Hill *et al.* 1996). Perhaps the Siesta palaeosol supported woody

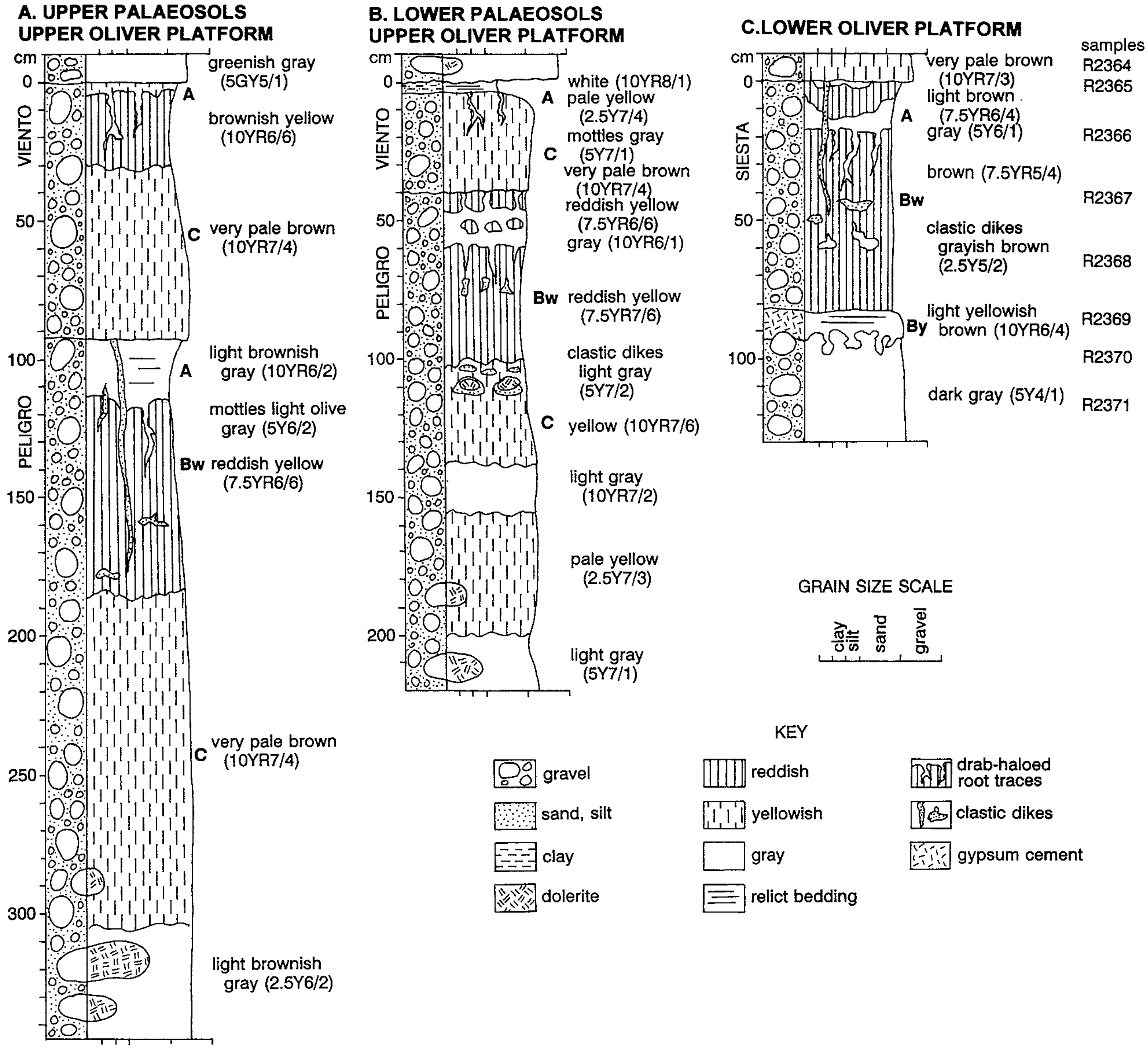


**Fig. 3.** Detailed sections of palaeosols in the Meyer Desert Formation of the Dominion Range, Antarctica: (**a**) upper Viento and Peligro palaeosols of the upper Oliver Platform (Fig. 2); (**b**) lower Viento and Peligro palaeosols (these are the designated type profiles) of the upper Oliver Platform (Fig. 2); (**c**) type Siesta palaeosol in lower Oliver Platform.

**Table 2.** *Pedotypes in the Meyer Desert Formation of the Dominion Range, Antarctica*

| Pedotype | Diagnosis | Campbell & Claridge (1987) stage | Bockheim & Ugolini (1990) classification | FAO 1974 classification | US taxonomy (Soil Survey Staff 1999; Bockheim 1997) |
|---|---|---|---|---|---|
| Viento | Thin (10–30 cm) orange and silty with root traces | 2 (20–100 ka) | Red Ahumisol | Gelic Regosol | Cryorthent |
| Peligro | Thick (60–90 cm) brown and silty with frost cracking | 3 (340–390 ka) | Subantarctic Brown | Cambic Arenosol | Typic Haploturbel |
| Siesta | Thick (80 cm), and brown, with subsurface gypsum-rich layer (By) | 4 (800–920 ka) | Subantarctic Brown | Cambic Arenosol | Gypsic Anhyturbel |

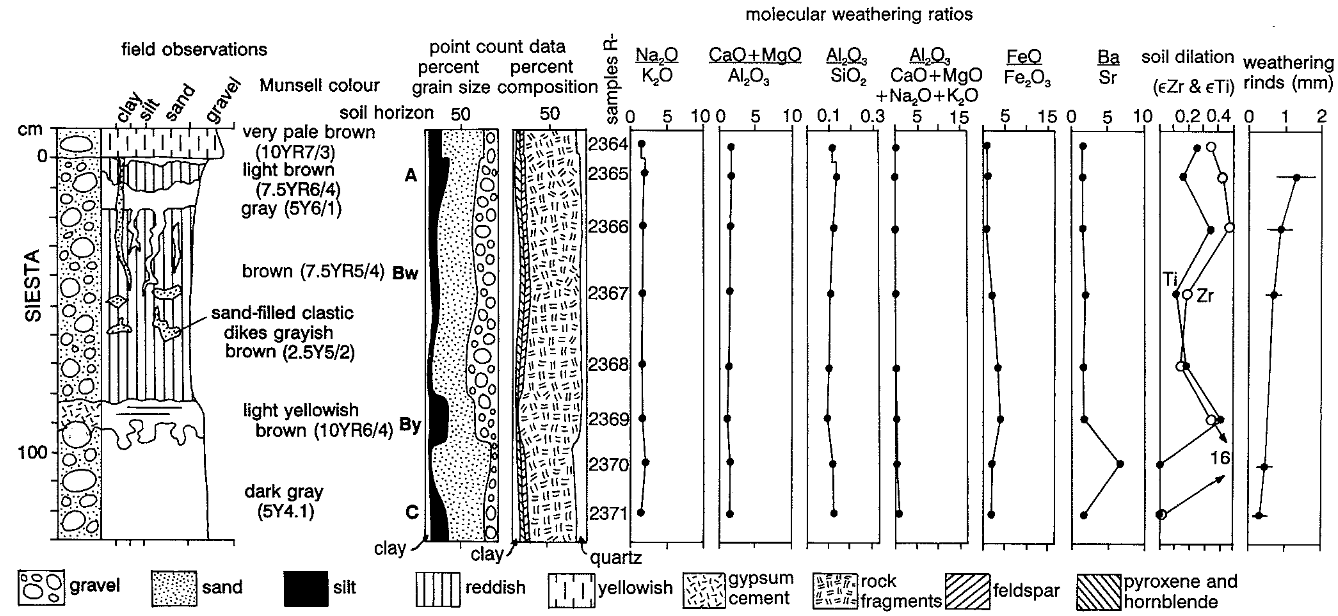


**Fig. 4.** Siesta palaeosol in the Meyer Desert Formation near the northern end of Oliver Bluffs and lower Oliver Platform: measured section, Munsell colours, grain size, mineral composition, molecular weathering ratios, dilation during soil formation estimated from distribution of Zr and Ti, and thickness of weathering rinds on pebbles. Molecular (or molar) weathering ratios are calculated from oxide percent by normalizing to molecular weight (Retallack 1997) and soil dilation is calculated by normalizing to Zr and Ti (as indicated) which are assumed to remain unweathered (Brimhall *et al.* 1991).

plants other than *Nothofagus*, or perhaps *N. beardmorensis* had slightly greater tolerance for aridity than living species of *Nothofagus*?

## Palaeotemperature implications of the palaeosols

Clues to palaeotemperatures during formation of the palaeosols come from sparse drab-haloed root-traces, irregular veins of gray sand, and common frost-cleaved to disintegrated clasts of their glacial till parent materials (McKelvey *et al.* 1991; Retallack & Krull 1998). These are not features of modern soil formation because these palaeosols were on steep slopes and excavated to improve exposure. Sand-filled pockets, dykes and veins that rim or underplate pebbles in both Peligro and Siesta palaeosols of the Meyer Desert Formation are comparable with ice veins and raised stones of soils within an active permafrost layer some 60–80 cm deep. Such deep permafrost active layers are found in regions of Antarctica where mean annual temperature (MAT) is –11 to –3°C and MAP 300–1100 mm (Bockheim 1995). The form of cryoturbation found in palaeosols of the Meyer Desert Formation is most like seasonal frost crack polygons found in regions of MAT <0 to –4°C and coldest month temperature of <– 8°C (Karte 1983).

These palaeotemperature estimates compare favourably with the colder end of estimates from fossils in the Meyer Desert Formation. *Nothofagus* fossils from Oliver Bluffs have asymmetric branches and numerous (about 60), exceedingly fine growth-rings (Francis & Hill 1996). They were probably deciduous prostrate shrubs living close to their absolute temperature limits: i.e. MAT –12 to –15°C, minimum winter temperature (MWT) –22°C, and summer growing season temperature (MST) of 5°C (Hill *et al.* 1996; Francis & Hill 1996). A fossil listroderine tibia from Oliver Bluffs is similar to those of weevils living at MAT –5 to –9°C (Ashworth *et al.* 1997). Although both the weevil and plant fossils are from lower within the same stratigraphic sequence as the Siesta palaeosol, and so represent an earlier geological time, the implications of these fossils for palaeotemperature are not significantly different from the indications of the palaeosol.

Palaeotemperatures estimated here should not be assumed for coastal regions of Antarctica or the surrounding Southern Ocean during the Pliocene because of the high elevation and inland location of the Dominion Range (Bockheim *et al.* 1986). Using a lapse rate of –0.65°C 100 $m^{-1}$ of altitude gives temperatures at present sea level 11°C warmer than in the Dominion Range at its present altitude, 8°C warmer for an altitude in the Pliocene 350 m lower than now (postulated by Gleadow & Fitzgerald 1987) and 3°C warmer if there was a fiord 100 m deep at The Cloudmaker in the lower Beardmore Glacier during the Pliocene (as proposed by Webb *et al.* 1996). Palaeotemperatures estimated from palaeosols and *Nothofagus* fossils combined with this range of lapse rate corrections yield Pliocene MAT –13 to +7°C of air at sea level, warmer than at present for McMurdo Station (MAT –20°C) or nearby Scott Base (MAT –18°C: Campbell & Claridge 1987). Nevertheless the climatic regime of the palaeosols was fundamentally frigid and these palaeosols formed on glacial terraces beside large permanent glaciers (McKelvey *et al.* 1991; Retallack & Krull 1998).

These warmer than current palaeotemperatures based on fossil plants and soils should not be assumed for times other than formation of Siesta and Peligro palaeosols, which are thicker and better developed than surface soils capping the sedimentary sequences containing them (Table 1). Surface soils on both the upper and lower Oliver Platforms are comparable to ultraxerous soils of the Arena Valley in Victoria Land (Fig. 9), but the palaeosols indicate a warmer and wetter palaeoclimate like that of Enderby Land (Bockheim & Ugolini 1990).

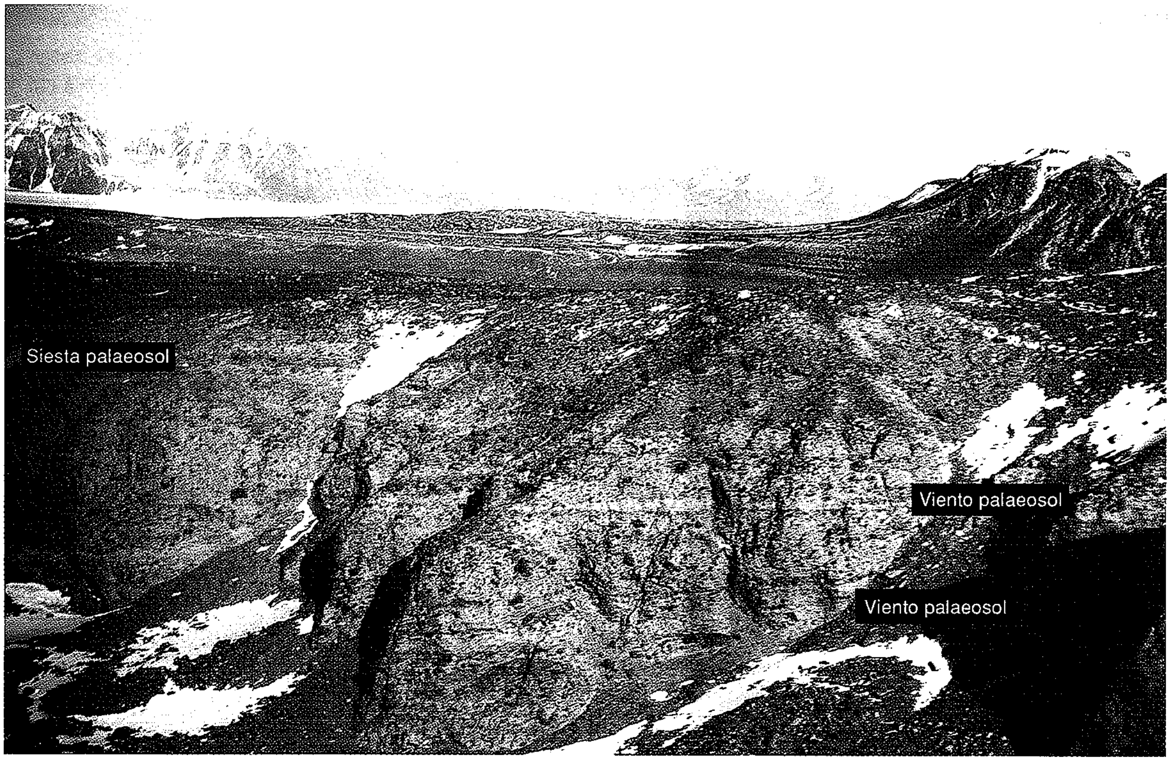
Siesta palaeosol
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Fig. 5.

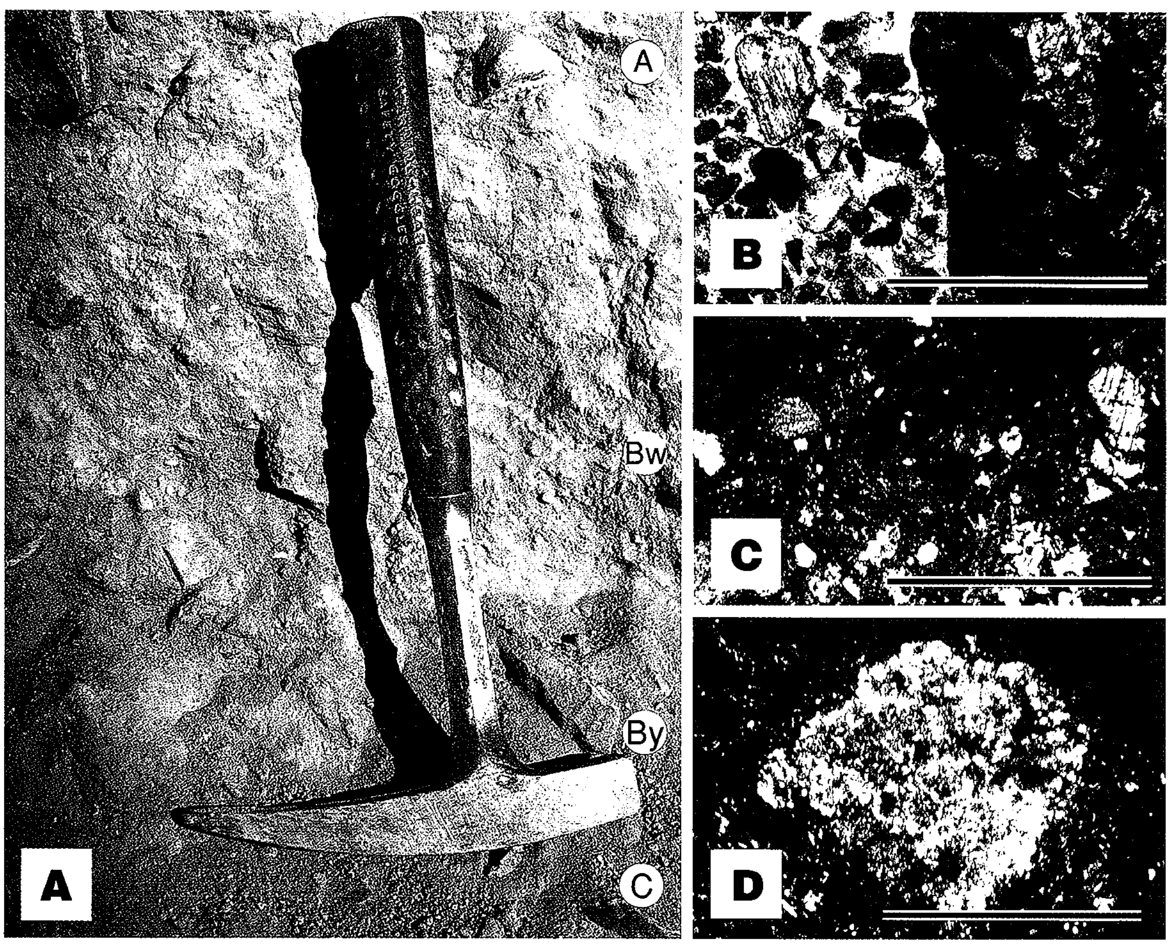
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Fig. 6.

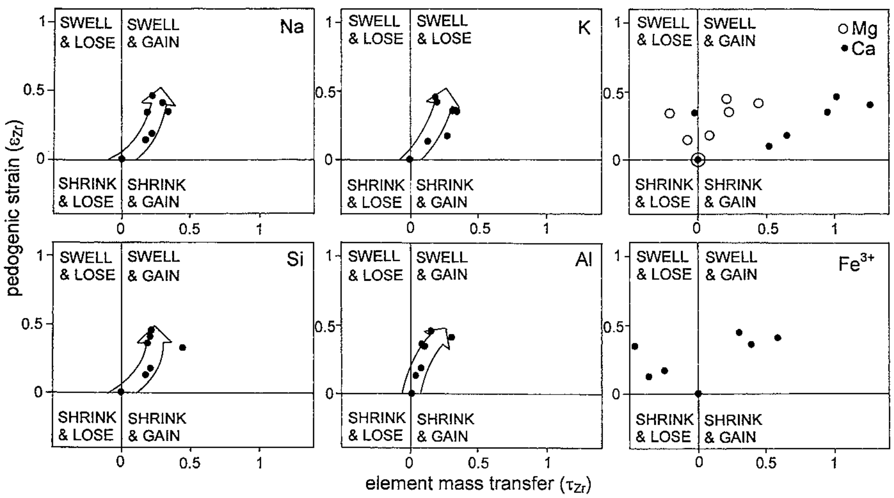


**Fig. 7.** Diagrams illustrating chemical mass transfer and strain, both normalized to Zr as an assumed stable constituent, of the Siesta palaeosol from the Meyer Desert Formation at the northern end of Oliver Bluffs, and lower Oliver Platform, Dominion Range, Antarctica. The diagrams follow equations and conventions of Brimhall *et al.* (1991). The lack of clear weathering trends is evidence of limited weathering and eolian influx in a dry and frigid weathering regime.

This is remarkable because a surface soil on the upper Oliver Platform began to form before glacial incision of lower Oliver Platform and its sedimentary sequence containing *Nothofagus* and the Siesta palaeosol (McKelvey *et al.* 1991). Wind deflation, evident from elaborate, large ventifacts, probably destroyed the original soil that formed on the upper Oliver Platform when *Nothofagus* grew on the aggrading lower Oliver Platform.

The Meyer Desert Formation is thus a record of at least two palaeoclimatic oscillations (Table 1) reflected in alternation of Subantarctic Brown palaeosols with Brown Ahumisol soils (Bockheim & Ugolini 1990). The ultraxerous frigid surface soils were initiated at times of near-modern palaeoclimatic conditions and nearby glacial downcutting, followed by glacial retreat and deposition of lodgement tillites, then formation of palaeosols under conditions more like those of Enderby Land than the central Transantarctic Mountains, and finally a return to near-modern soils and palaeoclimate before renewed glacial downcutting. Palaeoclimatic fluctuations during the Pliocene have long been recognized in northwestern Europe where there were warm palaeoclimatic episodes during the late Zanclean (*c.* 3.5 Ma) and late Messinian (*c.* 5 Ma: Kürschner *et al.* 1996; Utescher *et al.* 2000). Comparable Pliocene palaeoclimatic fluctuations can also be seen in Southern Ocean silicoplankton records (Bohaty & Harwood 1998) and marine isotopic records (Zachos *et al.* 2001), and in seismic reconstructions of Antarctic glacial stratigraphy (Bart 2001).

## Palaeoecological implications of the palaeosols

Roots, prostrate stems and leaves of *Nothofagus beardmorensis* are found within Viento palaeosols, thus providing confirmation that these plants actually grew there (Webb & Harwood 1993; Hill *et al.* 1996; Harwood & Webb 1998). Viento palaeosols are too weakly developed to be informative about palaeoclimate. They represent communities early in plant succession, which can take many centuries in frigid ecosystems (Walker & Peters 1977). Other fossil seed, leaf, beetle, ostracod, snail and fish remains come from a carbonaceous siltstone that represents a local waterlogged habitat (Ashworth *et al.* 1997). All these fossils are stratigraphically below the Siesta palaeosol in a glacial sequence deposited on the lower Oliver Platform.

In contrast, Peligro and Siesta palaeosols contain no fossils, but can be considered trace fossils of mature ecosystems on well drained soils. Both contain drab-haloed root traces that are as thick and deeply penetrating as those of woody plants, probably cushion plants or prostrate shrubs as envisaged by Francis & Hill (1996) for *Nothofagus beardmorensis*. Soils like the Siesta and Peligro palaeosols support cushion plant-lichen communities, for example, the woody plant *Salix arctica* on beaches of Truelove Lowland, Devon Island, Canadian Arctic (MAP 130 mm, MAT −16°C, MST 1.3–13.6°C: Walker & Peters 1977). Low evapotranspiration at low temperature and a short growing season allow growth of mesic, woody angiosperms at such low precipitation. Also generally similar are the soils of Enderby Land, Antarctica (MAP 600 mm, MAT −11°C; Bockheim & Ugolini 1990), where there are no living vascular plants. Both estimates are much less than MAT 5–8°C and summer temperatures of 15°C inferred by previous comparisons with vegetation in southernmost Chile (Mercer 1986). They are also less than MAT −11 to −3°C and MST −2 to +1°C of air over tundra in maritime subantarctic regions (Bockheim & Ugolini 1990). Our concept of well-drained shrub-tundra in a dry frigid climate is supported by a general comparison of palaeosols in the Meyer Desert Formation with Red Ahumisol and Subantarctic Brown Soils (Retallack & Krull 1998). Siesta and Peligro palaeosols lack large scale polygonal cracking and are not much more weathered than

**Fig. 5.** Aerial view of northern Oliver Bluffs and Meyer Desert with Mill Glacier in background, showing orange band of Siesta palaeosol to left and tan band of a Viento palaeosol in main bluff of Meyer Desert Formation, forming lower Oliver Platform.

**Fig. 6.** Annotated profile (a) and thin sections viewed under crossed polars (b–d) of the Siesta palaeosol in the Meyer Desert Formation at the northern end of Oliver Bluffs and lower Oliver Platform, Dominion Range, Antarctica: (**a**) view of excavated profile, with annotations A, Bw, By and C for interpreted soil horizons (see Fig. 3c); (**b**) ferruginized weathering rind (diffusion sesquan in soil terminology) on a pebble of Ferrar Dolerite (to right) in Bw horizon; (**c**) fresh but frost-cracked pyroxene (upper right) and extensively cracked feldspar (right centre) in A horizon; (**d**) fine-grained gypsum forming a nodule in By horizon. Scales indicated for (a) by hammer (handle 25 cm), and for (b–d) by scale bars (all 1 mm).

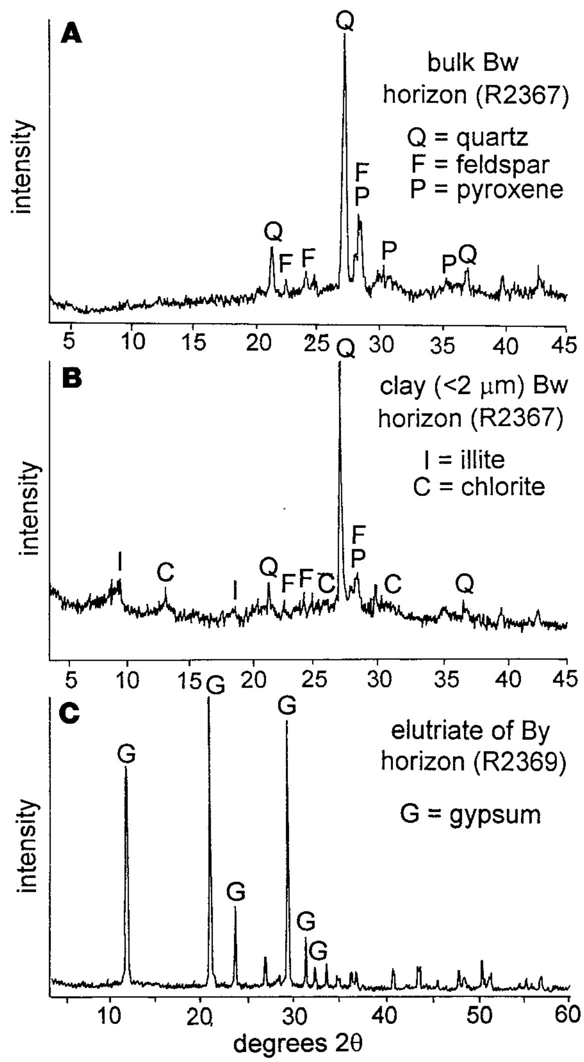


**Fig. 8.** X-ray diffractograms of bulk palaeosols, clay minerals and salt elutriate from the Siesta palaeosol of the Meyer Desert Formation in the northern end of Oliver Bluffs and lower Oliver Platform, Dominion Range, Antarctica.

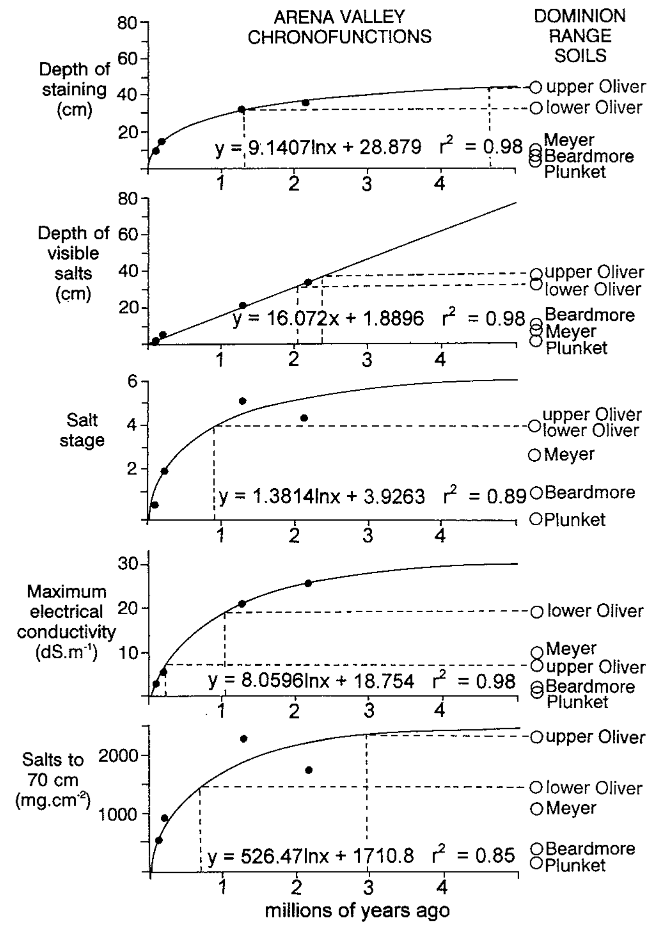


**Fig. 9.** Chronofunctions for soils in the Arena Valley, Victoria Land (filled circles from Marchant *et al.* 1994) used to estimate geological age of soils in the comparably frigid, ultraxerous Dominion Range, Transantarctic Mountains (open circles from Bockheim *et al.* 1986).

soils in the Meyer Desert today (Bockheim *et al.* 1986). The palaeosols also lack ferruginization and chemical weathering found in soils of the Antarctic Peninsula and surrounding islands (Bockheim & Ugolini 1990), or southern Chile (Frederiksen 1988), which have been previously suggested as analogous modern environments (Mercer 1986).

## Duration of palaeosols and sediment accumulation

The degree of development of the various palaeosols of the Meyer Desert Formation reflects their duration of exposure, and has implications for the style and pace of accumulation of these coarse-grained glacial sediments. Both sequences of the Meyer Desert Formation on the upper and lower Oliver Platforms have thick lodgement tillites overlying glacially-striated bedrock (McKelvey *et al.* 1991). Both sequences also have most palaeosols near the top. In the lower Oliver Platform and Oliver Bluffs, fossil *Nothofagus*, beetles, ostracodes and snails are above the basal tillites, and overlain by conglomerates and sandstones with several Viento palaeosols and the Siesta palaeosol near the top of the sequence. In the upper Oliver Platform massive lower tillites are overlain by sandy glacial deposits and then alternations of Peligro and Viento palaeosols. Thus deposition was initially rapid and voluminous following retreat of the glacier that scoured out the bedrock surface. Later, sedimentation slowed to allow growth and burial of plant fossils, and then there were long periods of non-deposition and soil formation. Comparable cycles of glacial scouring, rapid deposition and then soil formation can also be seen in the Sirius Group elsewhere in Antarctica (Webb *et al.* 1996, 1998).

The duration of gaps in sedimentation represented by palaeosols can be estimated by comparison with weathering stages for Antarctic soils devised by Campbell & Claridge (1987). The weathering index is based on depth of staining, reddening, and salt accumulation, which increase through time. By this scale the Siesta palaeosol represents about a million years of soil formation, and the two Peligro palaeosols of the upper Oliver Platform represent a similar length of time (Table 2). More precise estimates could come from chronofunctions for soils of Enderby Land or other areas of Antarctica envisaged to form under comparable conditions to the palaeosols. Such chronofunctions are not currently available.

## Ages of Dominion Range soils and sediments

Soils of Meyer Desert (Fig. 1) show different degrees of soil development related to their geological age, and this also constrains the geological age of the fossil plants and soils. The upper and lower Oliver Platform are bedrock terraces cut by glaciers, and overlain by glacial deposits and palaeosols of the Meyer Desert Formation (Bockheim *et al.* 1986; McKelvey *et al.* 1991; Denton *et al.* 1989). Glacial terrace formation in a region of uplift means that the surface soil of the lower Oliver Platform is geologically younger than the surface soil of the upper Oliver Platform. This is so because the glacier narrowed and deepened to cut the lower terrace as the mountains were uplifted. Furthermore, the sedimentary sequences between the glacial pavement and surface soils are constrained by these soil ages. For example, the Meyer Desert Formation of the lower Oliver Platform with the Siesta palaeosol and *Nothofagus* fossils is younger than the surface soil of the upper Oliver Platform and older than the surface soil of the lower Oliver Platform (Fig. 1, Table 1).

Chronofunctions for depth of staining, depth to soluble salts, electrical conductivity and salt content for well-dated, ultraxerous soils of the Arena Valley, Victoria Land (Marchant *et al.* 1994) can be used to estimate the age of comparable ultraxerous soils on the upper and lower Oliver Platforms (Bockheim *et al.* 1986). Antarctic soils are surprisingly ancient, and may have geologically significant ages of millions of years (Campbell & Claridge 1987). Estimates of the age of Dominion Range soils using these chronofunctions vary depending on the feature used (Fig. 9). Estimates based on salts are probably not reliable because they are related to oceanic proximity, which may have varied considerably over the Neogene (Webb *et al.* 1996). As predicted by overall marine regression, the salt chronofunctions underestimate age. The best age estimates come from depth of staining (Fig. 9). These ages and standard deviations are incorporated in our local chronology (Table 1). Our estimates are within an order of magnitude of estimates from diatom biostratigraphy (Webb *et al.* 1996; Bohaty & Harwood, 1998) and cosmogenic isotopes (Kurz & Ackert, 1997; Schäfer *et al.* 1999). Even maximal errors do not allow ages of more than 6 Ma for the upper Oliver Platform. It may be argued that this approach is compromised by climatic differences noted here between palaeosols and surface soils. However, the controversy is not about colder palaeoclimate, but rather about warmer episodes within the last 8 Ma (Burckle & Pokras 1991; Wilson 1995), and using warmer climate chronofunctions would give markedly younger age estimates for Dominion Range soils. Both soils on, and palaeosols in, the Meyer Desert Formation would be significantly more clayey and chemically more weathered than observed if they were either geologically older than Pliocene, or formed in warmer climates of the mid–early Miocene, Oligocene or Eocene, as has been claimed (Burckle & Pokras 1991; Burckle & Potter 1996; Warnke *et al.* 1996). Tierra del Fuego and the northern Antarctic Penninsula have a frigid landscape with glaciers down to sea level (Frederiksen 1988; Bockheim & Ugolini 1990) as envisaged for Antarctica during Oligocene and Miocene time (Zachos *et al.* 2001), but palaeosols of the warmest and wettest phases of deposition of the Meyer Desert Formation are much less weathered than the soils of Tierra del Fuego and the Antarctic Peninsula.

Our Plio-Pleistocene estimate for soils on the Meyer Desert Formation of the upper and lower Oliver Platforms does not negate the idea that some montane and dry valley landscapes of Antarctica are Miocene or older, as indicated by estimates from soil development (Campbell & Claridge 1987), dating by cosmogenic nuclide accumulation (Ivy-Ochs *et al.* 1995; Brook *et al.* 1995; Kurz & Ackert 1997; Schäfer *et al.* 1999), and radiometric dating of volcanic rocks (Barrett *et al.* 1992; Sugden *et al.* 1995; Armienti & Baroni 1999). A Pliocene (3–4 Ma) age for the Siesta palaeosol and *Nothofagus* fossils has been inferred from diatom and palynological biostratigraphy (Barrett *et al.* 1992; Webb *et al.* 1996; Fleming & Barron 1996; Bohaty & Harwood 1998;Jonkers & Kelley 1998). Such an age estimate is not necessarily undermined by the possibility of eolian deposition of diatoms (Burckle & Potter 1996) or by delivery of diatoms in clasts of sediment ejected from an asteroid impact in the Southern Ocean (Gersonde *et al.* 1997). Eolian deposition of the diatoms is unlikely because the diatoms have been found within clasts too large to be transported by wind (Harwood & Webb 1998). Ballistic delivery after impact of diatomaceous clasts obviates the need for glacial transport from hypothesized Pliocene diatomaceous marine rocks in the Wilkes and Pensacola subglacial basins (Webb & Harwood 1993; Wilson 1995). Marine transgression into East Antarctica behind the Transantarctic Mountains (proposed by Webb & Harwood 1993) would have been a much greater degree of deglaciation than indicated by evidence for frigid palaeoclimate from palaeosols presented here.

## Conclusion

Palaeosols associated with fossil leaves of *Nothofagus* in the Meyer Desert Formation of the Sirius Group in the Dominion Range contain fossil root traces indicating that they supported woody vegetation and so are in place of growth. The palaeosols are not like soils found there today (Bockheim *et al.* 1986), nor like soils associated with *Nothofagus* in Tierra del Fuego (Frederiksen 1988). Instead the palaeosols are more like those of modern Enderby Land, Antarctica (Bockheim & Ugolini 1990) and Devon Island in the Canadian Arctic (Walker & Peters 1977), thus confirming reconstruction of *Nothofagus beardmorensis* as prostrate woody shrubs (Francis & Hill 1996). The age of geologically ancient surface soils of the Dominion Range (Bockheim *et al.* 1986) can be assessed by comparison with well dated soils in the comparably ultraxerous Arena Valley of Victoria Land (Marchant *et al.* 1994). Dominion Range soils formed before and after deposition on the lower Oliver Platform confirm a Pliocene (3.1–4.1 Ma) age of the *Nothofagus* leaves and associated palaeosols previously suggested on the basis of recycled marine diatoms (Webb *et al.* 1996). Furthermore, our soil dating approach suggests two episodes of climatic warming to conditions like those of Enderby Land at about 3.5 and 5 Ma, with intervening and subsequent palaeoclimate comparable to modern conditions. Throughout these palaeoclimatic fluctuations the Dominion Range remained flanked by large glaciers (Fig. 10).

The duration and magnitude of palaeoclimatic warming revealed by palaeosols in the Meyer Desert Formation is thus less than originally inferred from the presence of *Nothofagus* and comparisons with Magellanic moorlands (Mercer 1986). Our interpretation of significant but subdued palaeoclimatic fluctuation is not at variance with indications of relative stability of Southern Ocean palaeotemperatures (Burckle & Potter 1996; Warnke *et al.* 1996) and of polar dry climate and extensive glaciation in other parts of Antarctica for at least the last 8 Ma (Wilson 1995; Sugden *et al.* 1995; Bart 2001).

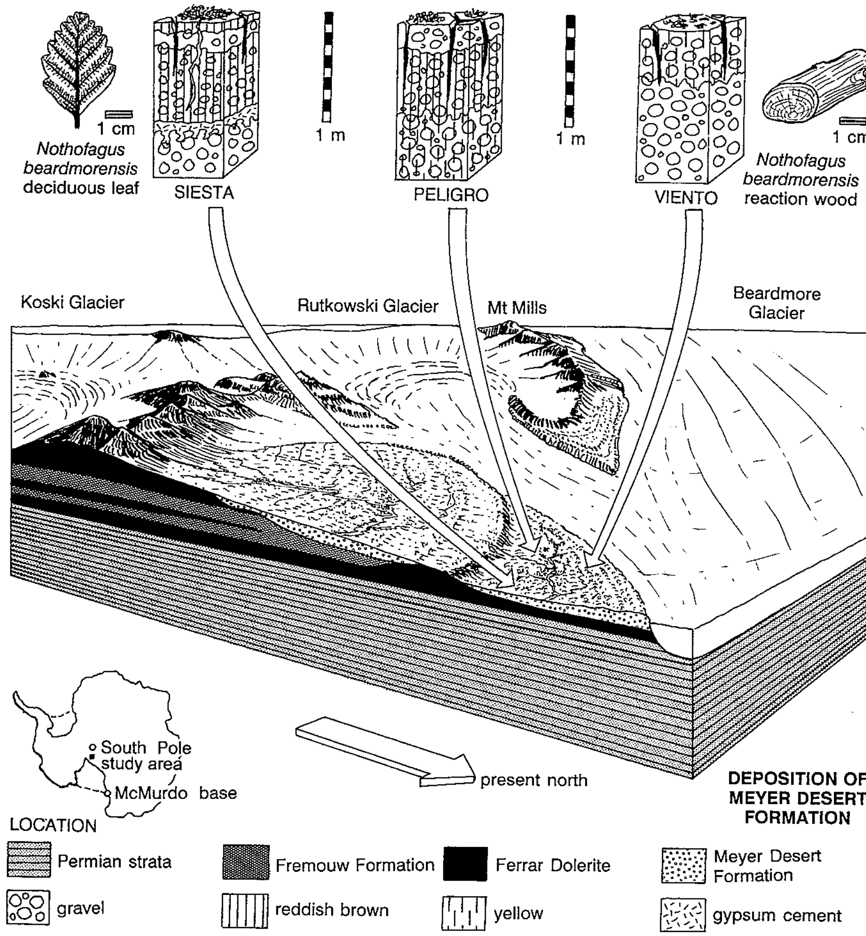


**Fig. 10.** A reconstruction of soils and vegetation of the Dominion Range during deposition of the Meyer Desert Formation on the lower Oliver Platform during the Pliocene (3–4 Ma). Peligro palaeosols are presumed sources of reddish sands and precursors for development toward the Siesta pedotype, but intact profiles have only been found in older stratigraphic levels. The South Pole is 500 km south toward the head of the Beardmore Glacier.

Perceived palaeoclimatic anomalies have led to suggestions of a middle Miocene or older age for the Meyer Desert Formation (Burckle & Pokras 1991;Kurz & Ackert, 1997), but the palaeoclimate of Antarctica was so warm during the middle Miocene and earlier (Warnke *et al.* 1996) that soils very different from the observed palaeosols would be expected (Bockheim & Ugolini 1990). While our study of palaeosols of the Meyer Desert Formation has resulted in a compromise of this palaeoclimatic debate, it also opens up the possibility of obtaining detailed records of Neogene palaeoclimatic fluctuations from comprehensive studies of Sirius Group palaeosols.

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