EAST ANTARCTIC ICE SHEET SENSITIVITY TO PLIOCENE CLIMATIC CHANGE FROM A DRY VALLEYS PERSPECTIVE

BY

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ABSTRACT. A case is made for the stability of the East Antarctic Ice Sheet during Pliocene time from landscape development and surficial sediments in the Dry Valleys sector of the Transantarctic Mountains. The alternate hypothesis of Pliocene meltdown requires atmospheric temperatures 20°C above present values, late Pliocene ice-sheet overriding of the Transantarctic Mountains, and possible rapid late Pliocene mountain uplift of 1000-3000 m. The geomorphological results suggest that these conditions were not met in the Dry Valleys region. Rather, Pliocene mean annual atmospheric temperatures were at most only 3° to 8°C above present values; ice-sheet overriding occurred in Miocene time (>13.6 Ma); Pliocene glacier expansion was limited; and Pliocene surface uplift was only about 250 to 300 m. These conclusions are based on field studies in Taylor and Wright Valleys, in the western Asgard Range, and in the Quartermain Mountains. The chronology comes from numerous 40Ar/39Ar dates on in-situ volcanic ashes that occur in stratigraphic association with unconsolidated diamictons in the western Dry Valleys, basaltic lava flows interbedded with widespread tills in Taylor Valley, and re-worked basaltic clasts in alpine moraines in east-central Wright Valley. The combined evidence from the Dry Valleys region indicates that slope evolution was severely restricted throughout Pliocene time, and has been so since at least the middle Miocene. The implication is that most of the Dry Valleys landscape is relict and that it reflects ancient erosion, possibly under semi-arid climate conditions, prior to middle-Miocene time.

The Antarctic Ice Sheet and Pliocene Global Change

The modern Antarctic Ice Sheet consists of about 30 x 10^6 km^3 of ice spread over an area of 13.6 x 10^6 km^2 (Drewry et al. 1982; Drewry, 1983; Oerlemans and van der Veen, 1984). It is divided by the Transantarctic Mountains into separate components in West and East Antarctica (Fig. 1).

The marine West Antarctic Ice Sheet (3.3 x 10^6 km^3; 6 m sea-level equivalent) rests on a rugged bedrock floor (Fig. 2), much of it well below sea level and likely to remain submerged if deglaciation and isostatic compensation were completed (Drewry, 1983). Ice thickness is greatest in central West Antarctica, where subglacial troughs exceed -2000 m. In the interior, there is a complex of ice-surface divides, domes, and saddles of generally low elevation. Mountains that fringe central West Antarctica protect the grounded marine ice. Most surface topographic features dovetail into ice streams, which drain 90 percent of the interior and are the most dynamic components of the ice sheet (Hughes, 1977; Bindshadler, 1991).

The terrestrial East Antarctic Ice Sheet (26 x 10^6 km^3; 60 m sea-level equivalent) rests on a base that is predominately above sea level (Fig. 2). Some subglacial basins extend below sea level, although many would rise above sea level if the ice sheet were removed and isostatic compensation completed (Fig. 3a) (Drewry, 1983). Of particular importance to our discussion are the Wilkes and Pensacola subglacial basins, both situated inland of the Transantarctic Mountains (Fig. 3b). On one flank the East Antarctic Ice Sheet is dammed behind the Transantarctic Mountains, but elsewhere it terminates either in the ocean or in floating ice shelves. It is higher and has a smoother surface topography than the West Antarctic Ice Sheet; central domes reach 3200-4000 m elevation (Fig. 1) (Drewry, 1983). Inland flow discharges into the Filchner-Ronne Ice Shelf or into major outlet
glaciers that pass through the Transantarctic Mountains, whereas coastal flow discharges locally by ice streams into fringing ice shelves or into the Southern Ocean. Between some of the major outlet glaciers, local domes and ice ridges feed glaciers that pass through the mountains or terminate in the western Dry Valleys. One dome that is important to our discussion is the local Taylor Dome, situated just inland of the Dry Valleys sector of the Transantarctic Mountains (Fig. 1). The Taylor Dome (formerly the McMurdo Dome) reaches 2450 m elevation on the periphery of the East Antarctic Ice Sheet. It feeds Taylor and Wright outlet glaciers, which terminate in the Dry Valleys (Drewry, 1982).

Ice shelves fringe much of coastal Antarctica. Most occur in protected embayments or are anchored at pinning points. The two largest, the Ross (536 x 10^3 km^2) and Filchner—Ronne (532 x 10^3 km^2) Ice Shelves, occur in deep embayments and are fed by ice flow from both the West and East Antarctic Ice Sheets.

Today the Antarctic Ice Sheet gains mass by accumulation of about 2.2–2.7 x 10^{15} kg a^{-1} (Giovinetto and Bentley, 1985; Doake, 1985). Unlike the Greenland Ice Sheet (and most former Northern Hemisphere ice sheets), the Antarctic Ice Sheet does not now have surface melting ablation zones. As a result, it currently loses mass almost entirely by basal melting beneath ice shelves and by calving of icebergs. In the absence of summer-temperature control of surface melting margins, the primary factors controlling the area and volume of the ice sheet are thought to be eustatic sea-level oscillations (which cause grounding-line advance and retreat; Denton and Hughes, 1981;
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Fig. 2. Subglacial topography beneath present-day Antarctic Ice Sheet, without isostatic adjustments for the removal of the ice sheet. Adapted from Denton et al. (1991) and based on Drewry (1983).

Stuiver et al. 1981), changes in interior accumulation (which today apparently varies with mean atmospheric temperature above the ground inversion; Robin, 1977), changes in ice shelves (and hence in possible backstress on grounded ice), and changes in ice dispersal into the surrounding ocean from ice streams (Bindshadler, 1991).

A basic problem of global change involves the response of the Antarctic Ice Sheet to warmer-than-present polar climates, including that of a greenhouse world. This problem can be attacked by developing numerical models based on the dynamics of the present ice sheet (Huybrechts, 1993), and then subjecting these models to the forcing expected in a greenhouse world. An understanding of ice-sheet behavior can also come from examining the response of the Antarctic Ice Sheet during past intervals of warmer-than-present climate. In this regard, it has long been assumed that the East Antarctic Ice Sheet has been a stable feature on the surface of the Earth since middle-Miocene time (Shackleton and Kennett, 1975; Savin et al. 1975; Kennett, 1982; Miller et al. 1987). But this paradigm has been severely challenged by the discovery of reworked marine diatoms in outcrops of the Sirius Group along much of the Transantarctic Mountains (Fig. 4). These deposits are taken to indicate that the East Antarctic Ice Sheet underwent dramatic deglaciation during Pliocene warm intervals (Fig. 3b). The problem is important because it involves nothing less than the evolution of polar ice sheets and the global climate system.

Our purpose here is to examine the proposition that the East Antarctic Ice Sheet indeed collapsed during intervals of excess Pliocene warmth. We start by reviewing the hypothesis of Pliocene de-
Pliocene Deglaciation Hypothesis

A major shift in the climate system of the Earth occurred in the Pliocene Epoch between 2.9 and 2.2 Ma ago (Shackleton, 1993). Late Cenozoic ice ages began on a global scale during this decline, first with a 41,000-yr and then a 100,000-yr pulsebeat. Prior to 2.9 Ma ago, repeated intervals of warmer-than-present climate marked Pliocene time, and large ice sheets were limited to Antarctica. The most recent of these warm intervals occurred about 3.0 Ma ago, when high-latitude air and sea-surface temperatures in the Northern Hemisphere were elevated due to increased meridional oceanic heat transfer (Dowsett et al. 1992). Other evidence for Pliocene warmth during this and earlier intervals comes from such diverse paleoclimatic indicators as European flora (van der Hammen et al. 1971; Grube et al. 1986); the lack of significant ice-rafted detritus in North Atlantic Ocean cores (Ruddiman and Raymo, 1988);
terrestrial plant fossils in the Arctic (Funder et al., 1985; Matthews, 1989); African pollen assemblages (Bonnefille, 1976, 1983), micromammals (Wesselman, 1985), and Bovids (Vrba, 1988a, 1988b); molluscs along the western margin of the North Atlantic (Stanley, 1985); diatoms, radiolaria, and silicoflagellates from the Pacific and Southern Oceans (Hays and Opdyke, 1967; Ciesielski and Weaver, 1974); Andean flora (Hooghiemstra, 1993); and some interpretations of the marine oxygen-isotope record (Prentice and Denton, 1988; Shackleton, 1993). Warmer-than-present Antarctic conditions are inferred from early-to-mid Pliocene benthic foraminifers in cores from DVDP sites 10 and 11 in eastern Taylor Valley (Ishman and Rieck, 1992); from early Pliocene diatoms and silicoflagellates in the subantarctic Southern Ocean (Ciesielski and Weaver, 1974; Abelman et al., 1990); and from microfossils in raised marine deposits in the Westfold Hills (Pickard et al., 1988).

The deglaciation hypothesis calls for Pliocene meltdown of the East Antarctic Ice Sheet at 3.0 to 2.5 Ma ago (Fig. 3b) (Harwood, 1983, 1986; Webb and Harwood, 1987; Barrett et al., 1992) and also during earlier Pliocene warm intervals (Webb et al., 1984; Webb and Harwood, 1991). This hypothesis carries with it the implication that the East Antarc-
Fig. 4. Location of Sirius Group outcrops and erratics in the Transantarctic Mountains. References are as follows: (1) Mercer (1968); (2) Doumani and Minshew (1965); (3) Mayewski (1975); (4) LaPrade (1984); (5) McGregor (1965); (6) Claridge and Campbell (1968); (7) Elliot et al. (1974); (8) Mercer (1972); (9) Barrett and Elliot (1973); (10) Prentice et al. (1986); (11) Denton et al. 1991; (12) J. Anderson, unpublished; (13) Faure and Taylor (1981); (14) Barrett and Powell (1982); (15) Brady and McKelvey (1979); (16) H.W. Borns Jr., unpublished; (17) Brady and McKelvey (1983). Fossil Nothofagus wood occurs in the northern Dominion Range deposit. Adapted from Denton et al. (1991).
Fig. 5. The Dry Valleys region is bounded by the Mackay Glacier to the north and Ferrar Glacier to the south. Map is from U.S. Geological Survey, McMurdo Sound 1:1,000,000 sheet, 1974.
Fig. 6. (a) Sketch map showing location of drill cores, place names, and stratigraphic sections used in text. Place names are numbered as follows: 1, Conrow Glacier; 2, Bartley Glacier; 3, Meserve Glacier; 4, Hart Glacier; 5, Goodspeed Glacier; 6, Rhone Glacier; 7, Matterhorn Glacier; 8, Lacroix Glacier; 9, Seuss Glacier; 10, Canada Glacier; 11, Commonwealth Glacier; 12 Rhone Platform; 13, Thomson moraine; 14, Lake Vanda; 15, Lake Brownworth; 16, Lake Bonney; 17, Hart Ash; 18, Prospect Mesa; 19, Wright Upper Glacier; 20, Wright Lower Glacier; 21, Arena Valley Ash.
(b) Sketch map showing locations of figures and plates used in this paper and other papers in this volume. Arrows indicate direction of view for each photograph. Numbers refer to figures as listed below. 1. is Fig. 9 (top) of Denton et al. (1993); 2, is Fig. 10 (top) of Denton et al. (1993); 3, is Fig. 9 (bottom) of Denton et al. (1993); 4, is Fig. 23 (top) of Denton et al. (1993); 5, is Fig. 11 (top) of Denton et al. (1993); 6, is Fig. 6 of Denton et al. (1993); 7, is Fig. 16 of Denton et al. (1993); 8, is Fig. 17 of Denton et al. (1993); 9, is Fig. 13 (top) of Denton et al. (1993); 10, is Fig. 13 of Denton et al. (1993); 11, is Fig. 2 of Wilch et al. (1993b) and Fig. 20 of Denton et al. (1993); 12, is Fig. 25 of Denton et al. (1993); 13, is Fig. 10 (bottom) of Denton et al. (1993); 14, is Fig. 11 (bottom) of Denton et al. (1993); 15, is Figs. 6 and 7 of Wilch et al. (1993b) and Fig. 26 of Denton et al. (1993); 16, is Fig. 7 of Marchant et al. (1993b) and Fig. 27 of Denton et al. (1993); 17, is Fig. 5 of Hall et al. (1993) and Fig. 28 of Denton et al. (1993); 18, is Fig. 29 of Denton et al. (1993); 19, is Fig. 30 of Denton et al. (1993); 20, is Fig. 24 (top) of Denton et al. (1993); 21, is Fig. 23 (bottom) of Denton et al. (1993); 22, is Fig. 11 of Hall et al. (1993); 23, is Fig. 1 of Hall et al. (1993); 24, is Fig. 4 of Hall et al. (1993); 25, is Fig. 10 of Hall et al. (1993); 26, is Fig. 4 of Marchant et al. (1993b); 27, is Fig. 11 of Marchant et al. (1993b); 28, is Fig. 7 of Marchant et al. (1993a); 29, is Fig. 2 of Marchant et al. (1993a); 30, is Fig. 3 of Marchant et al. (1993a); 31, is Fig. 6 of Marchant et al. (1993a); 32, is Fig. 4 of Marchant et al. (1993a); 33, is Fig. 5 of Marchant et al. (1993a); 34, is Fig. 8 of Marchant et al. (1993a).
ctic Ice Sheet is susceptible to climatic warming of the magnitude last reached in the Pliocene. By analogy, Barrett et al. (1992) implied that the ice sheet could again nearly disappear in a greenhouse world a few degrees warmer than present.

The deglaciation hypothesis is critically dependent on the inferred age and origin of reworked marine diatoms in Sirius Group outcrops along the Ross Sea sector of the Transantarctic Mountains (Fig. 4) (Webb et al. 1984; Harwood, 1986). The implication is that the diatoms originated in marine basins, and that subsequently the East Antarctic Ice Sheet expanded to cover the basins, override the Transantarctic Mountains, and re-work marine microfossils into the Sirius Group. The only realistic location for the basins is in the interior of East Antarctica—hence the argument that the ice sheet must have been sufficiently small to expose the Wilkes and Pensacola Subglacial Basins to the sea. Diatom flora that include elements indicative of warm (2°–5°C) interior East Antarctic seas are assigned ages varying from early to late Pliocene on the basis of stratigraphic ranges in subantarctic deep-sea cores (Harwood, 1986). Isotopic dating of a volcanic ash layer in the CIROS-2 core in the Ferrar Glacier trough (Figs. 5 and 6) confirms the biostratigraphic age control for certain critical diatoms in the age range of 2.5 to 3.0 Ma ago (Barrett et al. 1992).

The deglaciation hypothesis would be strengthened on two counts if the Pliocene paleotopography was greatly different from the present. The first is that the present-day Wilkes and Pensacola Subglacial Basins must have been much deeper than they are today to allow marine flooding (and hence marine diatom growth) consequent on Antarctic deglaciation. Huybrechts (1993) calculates that deglaciation must go nearly to completion with far more recession than shown in Figure 3b before the Pensacola basin near the South Pole becomes ice free; under these conditions isostatic compensation would have lifted the Pensacola and most of the Wilkes Subglacial Basins above sea level.

The deglaciation hypothesis also would be strengthened if the Pliocene-age Transantarctic Mountains were considerably lower than now. This would make it much easier to explain overriding by an East Antarctic Ice Sheet that emplaced marine microfossils into the Sirius Group at or after 2.5 Ma ago (Elliot et al. 1991). Overriding of lower mountains by a relatively thin temperate ice sheet would not require polar ice shelves in the Ross Sea, nor the existence of a West Antarctic Ice Sheet (Huybrechts, 1993). Because *Nothofagus* fossil wood occurs along with marine diatoms in Sirius Group deposits now at 1800 m elevation near Beardmore Glacier, the overriding ice sheet is inferred to have advanced under temperate conditions into scrub vegetation with *Nothofagus* (Webb et al. 1987; Webb and Harwood, 1987; McKelvey et al. 1987, 1991; Carlquist, 1987; Webb and Harwood, 1991; Hill et al. 1991). If it is assumed that Pliocene mountain elevations were indeed lower so that the *Nothofagus* scrub grew near sea level, then the postulated late-Pliocene/early Pleistocene *Nothofagus* growth would imply atmospheric warming of 20°–25°C above present temperatures (Barrett, 1991, p. 46), a value also suggested for early Pliocene time (Webb and Harwood, 1991, p. 215). Without uplift, the Pliocene temperature in Antarctica necessary to allow *Nothofagus* growth at the current high elevation of the fossil wood-bearing beds would have been at least 30°–35°C warmer than at present. In this regard, Pliocene-Pleistocene mountain uplift of 1000–3000 m has been postulated from faulting of Sirius Group deposits near Beardmore Glacier and from the current high elevation of the *Nothofagus* wood beds (Webb and Andreasen, 1986; Webb and Harwood, 1987), as well as from the "youthful-appearing" rift shoulder scarp along the mountain front (Behrendt and Cooper, 1991).

Pliocene uplift, but of much less magnitude, also is inferred from benthic foraminifers in DVDP cores 10 and 11 in eastern Taylor Valley (Ishman and Rieck, 1992). A mountain barrier created by tectonic uplift is credited with forcing late Pliocene climate change and transforming the character of the Antarctic Ice Sheet from temperate to polar (Behrendt and Cooper, 1991).

The possible Pliocene collapse of the East Antarctic Ice Sheet has been taken up by other branches of earth science. High sea levels (35 ± 18 m) of mid-Pliocene age (~ 3.0–3.5 Ma ago) on the Atlantic coastal plain (Orangeburg, Chippenham, and Thornburg Scars) have been related to melting of the East Antarctic Ice Sheet (Dowsett and Cronin, 1990; Krantz, 1991). Wide oscillations of a dynamic East Antarctic Ice Sheet are used to explain fluctuations in the marine-oxygen isotope record and in biogenic productivity noted in some deep-sea cores (Abelman et al. 1990; Krantz, 1991; Ishman and Rieck, 1992). A multi-disciplinary study (PRISM) designed to assess the Pliocene warm peak at 3.0 Ma ago (Cronin and Dowsett,
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1991) has incorporated East Antarctic ice collapse into an emerging global reconstruction (Dowsett et al. 1992).

The Pliocene deglaciation hypothesis has testable implications. First, the East Antarctic Ice Sheet must have overtopped (or nearly overtopped) the Transantarctic Mountains to emplace Sirius Group outcrops on mountain surfaces now at high elevations. The amount of thickening required depends critically on the surface elevation of the mountains at the time of overriding. For reasons already described, the deglaciation hypothesis is more tenable if the Transantarctic Mountains underwent extensive Pliocene-Pleistocene surface uplift. Second, the deglaciation hypothesis requires that mountain overriding occurred after the oldest part of the stratigraphic range of the youngest diatom included in Sirius deposits. This means that overriding was younger than 3.0 Ma ago (Barrett et al. 1992) and perhaps coincided with the Pliocene climatic deterioration and consequent growth of Northern Hemisphere ice sheets centered at 2.5 Ma ago (Elliot et al. 1991).

The third implication is of climatic warming so substantial that Pliocene environments as far south as 86°S latitude were at least subantarctic and perhaps even southern Patagonian in character (Mercer, 1986). Subantarctic diatoms in Sirius rocks at Reedy Glacier imply sea-surface temperatures of 2–6°C in the Pensacola Subglacial Basin near the South Pole, and Nothofagus fossil wood in Sirius deposits at Beardmore Glacier implies air temperatures at least 20°C above present-day values if the mountains were much lower than now. Further, the only plausible mechanism to cause deglaciation is to melt the ice sheet down from the surface; this requires extensive surface melting ablation zones that in turn demand a climatic warming of 17°–20°C (Huybrechts, 1993). It has been suggested that surging collapse of the East Antarctic Ice Sheet may have caused deglaciation of the inland basins, leading to warming (Harwood, 1986). However, Huybrechts (1993) concludes that alternate instability mechanisms (marine ice-sheet instability, large-scale surging, creep instability) are unlikely to have caused East Antarctic deglaciation.

The Dry Valleys

The Dry Valleys sector of the Transantarctic Mountains (Figs. 5 and 6) is particularly suitable for testing predictions of the deglaciation hypothesis for several reasons. First, it contains high-elevation Sirius outcrops with enclosed marine Pliocene diatoms, including a deposit on Mt. Feather, one of the original sites used to postulate the deglaciation hypothesis (Webb et al. 1984). Second, it includes the CIROS-2 core, where the presence of critical diatoms that also occur in Sirius deposits is tied to an 40Ar/39Ar date of a volcanic ash layer (Barrett et al. 1992). Third, the Dry Valleys feature other cores (DVDP 10 and 11 in Taylor Valley, CIROS-1 at the mouth of Ferrar Valley) and sections (Prospect Mesa in central Wright Valley) that contain microfossils that have been used to infer warmer-than-present Pliocene marine paleoenvironments (Webb, 1972, 1974; Barrett, 1992; Ishman and Rieck, 1992; Prentice et al. 1993). Fourth, widespread surficial Pliocene and Miocene drifts and volcanic ashes are still well exposed in the Dry Valleys because Quaternary glacier expansions were so small. These factors make it possible to address issues of Pliocene paleoclimate and ice-sheet overriding of the mountains. A cautionary note is necessary, however, because the Transantarctic Mountains potentially have a complex history of denudation and surface uplift. Therefore, results from individual tectonic blocks, including the Dry Valleys, do not necessarily apply to the mountains as a whole.

The Dry Valleys feature the largest tract of ice-free terrain in the Transantarctic Mountains. Long trunk valleys now largely free of ice (Taylor, Wright, and the Victoria system) cross the mountains from the inland ice sheet to the Ross Sea. These valleys are flanked by the Quartermain Mountains and by the Asgard and Olympus Ranges. The Wilson Piedmont Glacier blocks the mouths of Wright and Victoria Valleys. The Taylor Dome, a peripheral dome of the East Antarctic Ice Sheet, feeds local East Antarctic ice into Taylor and Wright outlet glaciers, which flow into the western valleys. Inland ice flow is drained north around the Dry Valleys by Mackay Glacier and south by Mulock Glacier.

Today the Dry Valleys are cold, windy, and hyper-arid. An excess of sublimation over precipitation at all elevations yields a distinct precipitation deficit (Chinn, 1980). For example, beside Lake Vanda at 123 m elevation on the floor of Wright Valley, mean annual temperature is -19.8°C (Schwerdtfeger, 1984), and mean annual precipitation is 10 mm water equivalent (Keys, 1980). Snowfalls are light; they can occur at any time of year.
but are concentrated in the summer. Alpine glaciers occur where wind-blown snow is concentrated into pre-existing theater-shaped embayments. Alpine glaciers are largest near the coast where snowfall (and therefore wind-concentrated accumulation) is greatest. Local glaciers are small, or even nonexistent, in alcoves close to the polar plateau. Because they are small and occur in a very cold environment, these alpine glaciers are frozen to underlying beds and are nearly free of debris (Meserve Glacier in Wright Valley has a basal temperature of -18°C; Bull and Carnein, 1968). In contrast, the larger outlet glaciers can attain basal melting conditions; for example, the deepest ice of Taylor Glacier is wet based (Robinson, 1984). Today, 90% of glacier ablation in the Dry Valleys is by sublimation and only 10% by melting (Chinn, 1980). Small meltwater streams, all at relatively low elevations, feed enclosed lakes in Taylor and Victoria Valleys. In Wright Valley, the Onyx River flows westward along the valley floor from Lake Brownworth (dammed and fed by Lower Wright Glacier) to Lake Vanda (enclosed in the lowest part of the valley). Meltwater from alpine glaciers on the south valley wall also feeds the Onyx River.

In the remainder of this paper we discuss those aspects of the Dry Valleys glacial and tectonic record that reflect on the problem of Pliocene ice-sheet stability. We start with landscape analysis of the Dry Valleys morphology, then move on to a consideration of late Tertiary landscape submergence and uplift, and conclude with a discussion of the surficial volcanic and glacial deposits that rest on the landscape.

**Landscape Analysis**

**General Statement**

Landscape analysis of the Transantarctic Mountains is the foundation for deciphering the late Tertiary behavior of the East Antarctic Ice Sheet dammed along the inland mountain flank. This is because late Tertiary glacial deposits resting on the landscape surface can not be interpreted in terms of ice-sheet history without a knowledge of the topographic evolution of the mountains. Evaluating the evidence for ice-sheet overriding of the Dry Valleys sector of the Transantarctic Mountains is particularly important, because the deglaciation hypothesis requires such overriding after 3.0 Ma (Barrett et al. 1992) to emplace high-elevation Sirius deposits. Further, extensive Pliocene/Pleistocene uplift invoked to explain both the high present-day elevation of many Sirius outcrops, as well as the high shoulder scarp of the Transantarctic Mountains, should be evident in the sequence of landscape development.

The Transantarctic Mountains comprise gently tilted blocks of sedimentary rocks (Devonian-Triassic Beacon Supergroup) which overlie Precambrian-Devonian basement and are intruded and capped by Jurassic dolerites and basalts (Ferrar Supergroup). They form the faulted, uplifted, and tilted shoulder of the highly asymmetric West Antarctic intracontinental rift system (Fig. 7). In southern Victoria Land they rise steeply from the coast to elevations of 2000 m (Dry Valleys block) to 4000 m (Royal Society block). The offshore Victoria Land basin is 10–14 km deep (Cooper and Davey, 1987a, 1987b), giving a vertical displacement of 15–20 km.

The timing of tectonic events is not clear, because the mountains themselves contain only sparse post-Jurassic deposits and because sedimentary units in the offshore Victoria Land basin are poorly dated. It is thought that there were at least two phases of rifting activity in the Ross Sea (Tessensohn and Wöhrer, 1991; Cooper et al. 1991; LeMasurier and Rex, 1991). The first occurred in the Mesozoic (possibly coincident with the separation of Antarctica and Australia about 90 Ma ago) and was marked by diffuse crustal thinning. The second began in the Eocene (Fitzgerald et al. 1986) and continued to the present, although the details are poorly known. The second phase involved uplift and tilting of the Transantarctic Mountains, subsidence of the rift system in the Ross Embayment, and alkali magmatism beginning about 25 Ma ago (Kyle, 1990).

The spatial-temporal patterns of denudation (rock uplift relative to the landscape surface) and the resulting flexural isostatic response (along with the interactions of morphotectonics, denudation, and ice-sheet behavior) are potentially very important in the evolution of the Transantarctic Mountains. New results from fission-track thermochronology show that denudation pulses along the 3000-km length of the Transantarctic Mountains were complex (Fitzgerald, 1992). The implication is that individual blocks within the Transantarctic Mountains may have had differing denudational histories. Here we consider the denudational and surface uplift history of the Dry Valleys block, one which is bounded by transform faults beneath Mackay Glacier on the north and Ferrar Glacier on the south (Fig 5). Identical fission-track den-
udation profiles for the north walls of Wright and Ferrar Valleys (Fitzgerald et al. 1986) suggest that the Dry Valleys sector of the Transantarctic Mountains has behaved as a single tectonic block since the early Tertiary.

The Dry Valleys Landscape

Landscape Antiquity. Traditionally, the Dry Valleys landscape has been attributed to glacier erosion combined with salt weathering and wind deflation. In this scenario, the major east-west trending valleys were cut by outlet glaciers (Selby, 1971, 1990; Denton et al. 1984). Slow backweathering of the initial steep glacial walls by salt weathering and wind deflation in a frigid, arid environment produced rectilinear slopes below free faces, tent-like ridges bounded by rectilinear slopes, or rectilinear slopes beneath plateaus (Selby, 1971, 1974). Theatern-shaped basins in intervening mountain ranges between east-west trending valleys are attributed to ancient cirque glaciation (Selby, 1971, 1974; Wilson, 1973; Denton et al. 1984) followed by prolonged backwearing of rectilinear slopes and free faces, again by salt weathering and wind deflation in a cold and rainless desert (Selby and Wilson, 1971; Wilson, 1973). In the high sandstone-and-dolerite mountain ranges of the western Dry Valleys, this backwearing left residual buttes and mesas.

As an alternate hypothesis, we suggest here that the Dry Valleys landscape is largely inherited from a climatic regime that existed prior to the imposition of the present cold, hyper-arid desert environment. This new hypothesis is based on the enormous antiquity of Dry Valleys landscapes and the
realization that significant slope evolution has not occurred since near the end of the middle Miocene.

For some time it has been evident that the east-west trending trunk valleys are ancient. Microfossils at the base of Dry Valleys Drilling Project (DVDP) core 11 in eastern Taylor Valley are late Miocene in age (Ishman and Rieck, 1992) (Figs. 5 and 6). In central Wright Valley the diatoms in DVDP core 4A in Lake Vanda are middle-to-late Miocene in age (Brady, 1979, 1982; Prentice et al. 1993). From 87Sr/86Sr ratios, reworked shells in Jason glaciomarine diamicton beside Lake Vanda are estimated to be 9±1.5 Ma old, and in-situ shells (Chlamys tusensis) in the Prospect Mesa are estimated to be as much as 5.5±0.4 Ma old (Prentice et al. 1993). The lowermost sediments in the CIROS-1 core taken from a deltaic environment off-shore of Ferrar Valley are early Oligocene in age (Barrett, 1989); the implication is that a proto-Ferrar Valley existed by that time.

But the key breakthrough for our reevaluation was the realization of the enormous antiquity of some Dry Valleys slopes and high-elevation planation surfaces. The benches of middle Taylor Valley have remained essentially unmodified since at least early-to-mid Miocene time, as demonstrated by numerous 40Ar/39Ar whole-rock ages of surface basanite volcanic deposits (Wilch et al. 1993). In the western Dry Valleys, the evidence comes from the undisturbed nature of old till sheets, moraines, and colluvial deposits on rectilinear valley-wall slopes that rise above planation surfaces or flattish valley floors. New laser-fusion 40Ar/39Ar dates of single feldspar crystals from volcanic ashes in the Asgard and Quartermain Mountains point to remarkable slope stability since middle-Miocene time. Details are given elsewhere in this volume (Marchant et al. 1993 a, b). Little-to-no slope evolution has occurred in the high western Asgard Range during at least the last 13.6 Ma. In Arena Valley in the Quartermain Mountains, the preservation of Miocene- and Pliocene-age ashes on steep valley slopes shows that the major bedrock landforms are relict and that little slope evolution/colluviation has occurred during at least the last 11.3 Ma. Likewise, the morphology of alpine-glacier lateral moraines on the south wall of east-central Wright Valley is preserved nearly intact, despite the fact that some of these moraines date to >3.7 Ma in age (Hall et al. 1993).

The implication of these data is that the rectilinear near slopes and much of the overlying colluvium in the Dry Valleys region are relict. Overall, we infer that most Dry Valleys slopes are close to moribund under the present hyper-arid cold-desert environment and that valley floors have undergone only minor erosional modification since middle-to-late Miocene time.

The current processes that shape the Dry Valleys landscape, particularly at elevations above 1500 m, feature salt weathering and wind deflation (Selby, 1971, 1974). These processes have modified pre-existing bedrock forms only modestly during at least the last 11.3–13.6 Ma (Marchant et al. 1993 a, b). Salt weathering and wind deflation are particularly evident on exposed sandstone bedrock; comparison with adjacent colluvium-mantled slopes suggests that 0.5–3 m of bare sandstone bedrock commonly has been removed by these processes since late Miocene time (Marchant et al. 1993 a, b). Glacial erosion is now minimal. Outlet glaciers may scour their beds where they are wet based (Robinson, 1984), but cold-based alpine glaciers are protective rather than erosional (Holdsworth, 1970). Small meltwater streams originate beside some glaciers below 1500 m elevation at the height of summer in the central Dry Valleys region and feed enclosed lakes on valley floors (Chinn, 1980). The largest stream is the Onyx, which flows inland seasonally along the floor of Wright Valley from Lake Brownworth near the Wilson Piedmont Glacier to Lake Vanda. Some meltwater streams cut ravines and form small and thin alluvial fans, but the discharges are rarely high enough to move more than sand-sized particles. In general, only wind deflation, salt weathering, rockfall, minor colluviation, and minor erosion by small streams are now actively altering the landscape.

We thus postulate that the Dry Valleys landscape is largely inherited from a previous climatic regime. But what was that regime? To answer this question, we compare the relict Dry Valleys landscape with possible counterparts elsewhere on the planet. The semi-arid platform deserts of the Colorado Plateau (Oberlander, 1989) and southwestern Jordan adjacent to the Dead Sea Rift (Osborn, 1985) exhibit landscapes nearly identical with the upland surfaces of the Dry Valleys region. Denudation of the near-horizontal strata of platform deserts yields a tabular landscape with rock plains that head in escarpments and support isolated mesas and buttes. The cuesta is the dominant component of platform deserts, which are therefore
Fig. 8. Detached buttes rising above a flat-lying pediment in Monument Valley on the Colorado Plateau. We postulate that such landscapes as these on the Colorado Plateau are very similar, and perhaps formed by the same mechanisms, as the Dry Valleys upland landscapes shown in Figures 9 and 10.

termed arid or semi-arid cuestaform landscapes (Oberlander, 1989). We now describe cuestaform landscapes as a basis for comparison with Dry Valleys morphology.

Platform Desert Cuestaform Landscapes. A semi-arid cuestaform landscape develops by backwearing and parallel recession of the cuesta scarp, accompanied by expansion of canyon systems and headward growth of pediment rock plains (Oberlander, 1977; Doehring, 1977; Cooke et al. 1993). The active element of the scarp is the caprock substrata, which erodes back to cause successive collapses of unsupported caprock. The substrata are generally (but not always) less resistant to weathering than the caprock. They form a rampart leading from the base of the cap rock (commonly a rectilinear slope) to the rock plain. This rampart can be rilled or smooth, talus-covered or bare. The substrata rampart can begin at the base of the caprock face or, more commonly, the break in slope at the top of the rampart can occur in the substrata below the caprock face. Denudation agents on the rampart include slope wash, creep, spring sapping from groundwater, salt weathering, and granulation. Flash floods transport the material eroded from the scarps along canyon floors and across pediments. On the Colorado Plateau, Cenozoic scarp retreat ranges from 0.5 to 7 km/Myr⁻¹.
Fig. 9. **Top:** oblique aerial view looking WNW of the upper planation surface with the inselberg of Shapeless Mountain (2739 m) in the background. The main escarpment bounding the surface is 600 to 900 m high. The scalloped front of this scarp is the result of backwearing by box canyons in the Olympus Range, each bounded by a rectilinear slope topped by a free face. The irregularities on the upper planation surface are likely to be the result of areal scouring beneath overriding ice.

**Bottom:** an oblique aerial view looking WNW across the western Asgard Range toward Shapeless Mountain. In the background, detached buttes and mesas, remnants of the upper planation surface, occur at the western edge of main escarpment. In the foreground are similar buttes, together with cols (wind gaps) that occur along the mountain divide in the western Asgard Range.

(Schmidt, 1989). Embayments and headlands develop because of varying resistance of the cuesta scarp to backwearing (for example, due to lithologic or joint-density). Retreating headlands can leave detached mesas and buttes, or inselbergs, that rise above an expanding pediment rock plain (Frontispiece, Fig. 8). Rapidly retreating embayments may develop headward to form canyons with stubby tributaries. The ramparts of cuesta scarps can change form during recession as new erodible substrates are exposed at different levels (or old erodible substrates terminate) during general scarp recession. Moreover, climate changes can cause alternating talus development and stripping on ramparts (Gerson, 1982). This can produce talus flatiron relics with intertalus gaps along the front of retreating ramparts. Talus formation or stripping has differing climatic thresholds, depending on rock types and many other factors. Therefore, under some climatic conditions ramparts can be free of talus, whereas under other climatic conditions the ramparts can be covered with talus. Finally, climatic variations can alter the shape of the cap rocks from domed to cliffed, if there are variations in the importance of slopewash erosion.

The role of groundwater sapping of the substrata may be of primary importance in scarp re-
cession and canyon development (Ahnert, 1960; Laity and Malin, 1985; Howard and Kochel, 1988; Laity, 1988). It has been suggested that groundwater flowing through the cap rock can converge to cause increased substrata sapping, thus enhancing headward migration. In fact, Laity and Malin (1985) proposed that the erosional process of groundwater sapping (and hence undercutting of cap rock) is instrumental in producing networks of theater-headed valleys on the Colorado Plateau. Where sapping by laterally flowing groundwater is dominant, the canyons have flat floors, hanging valleys, steep walls, theater-shaped heads, and constant width; drainage extension by groundwater sapping exploits joint patterns and is particularly important on the dipslope side of a cuesta where groundwater is channelled preferentially. In contrast, where overland flow predominates, the valley heads are tapered, and the networks are more arborescent than those cut by the groundwater sapping. There can be markedly asymmetric erosional patterns on opposing sides of a divide. The scarp along the backslope can be deeply indented by prominent canyons that extend headward because of forward spring sapping. The scarp along the updip slope can be either relatively straight or moderately indented. When headward-cutting valleys breach the divide, the cap rock can...
be removed to form low cols (wind gaps) through the divide.

Dry Valleys Escarpment Landscapes. The detailed morphologic forms of the relict upland surfaces of the Dry Valleys bear a remarkable resemblance to morphologic forms in semi-arid cuestaform landscapes of platform deserts (Frontispiece; Figs. 8, 9, and 10). The similarities start with the fact that the high-elevation, western Dry Valleys bedrock is composed of near-horizontal strata (Beacon sedimentary formations, Ferrar dolerite sills, and Kirkpatrick extrusions) resting on a crystalline basement complex. The denuded landscape cut into this bedrock is tabular. It shows several extensive rock plains separated by a prominent scarp, with isolated mesas and buttes. The escarpments, which are the marker features of the landscape, show resistant cap rocks over ramparts (rectilinear slopes). Flat-floored valleys with stubby tributaries have cut toward drainage divides. The resulting indentations are asymmetric, both in the western Asgard and Olympus Ranges and in the Quartermain Mountains. North-facing valley networks are longer, with more frequent stubby tributaries, than south-facing networks. In the Asgard Range, headward cutting has in places removed cap rock, leaving cols (wind gaps). In places in the Olympus...
Range, headward cutting has removed cap rock all along the divide, and side-wall expansion has removed intervalley ridges to leave residual buttes and mesas. These features closely resemble the cuestaform landscapes of semi-arid platform deserts, except that the Dry Valleys uplands exhibit cuesta scarps more prominently than the gentle dip slopes, which are largely buried by inland ice. Therefore, it is preferable to refer to these uplands as escarpment landscapes.

Because of the remarkable similarities in the detailed morphology of the Dry Valleys uplands with the platform terrain of the Colorado Plateau (Oberlander, 1989) and of southwestern Jordan near the Dead Sea Rift (Osborn, 1985), it is tempting to suggest that the upper-level platform surfaces of the Dry Valleys also reflect denudation under semi-arid conditions. This would be consistent with the suggestion of Selby (1974) that the straight slope segments in the Dry Valleys represent a Richter-slope development which is most likely in an arid or semi-arid environment with little vegetation.

However, an arid or semi-arid climate is not prerequisite for the evolution of platform structure, because it is possible for escarpments to develop in other climatic environments as long as accumulating debris is removed efficiently. For example, the
Bottom: meltwater channel and pothole system cut along the western flank of a sandstone butte in the western Asgard Range, viewed southward. Cliffs alongside tributary channels in foreground are about 30 to 40 m high. We postulated that these channels were cut by subglacial meltwater streams beneath northward-flowing ice that overrode the western Asgard Range.

Valley Systems. Deep transverse valleys graded to near sea level are cut into the upper-level landform assemblage of the Dry Valleys region. The overall valley systems, when reconstructed beneath adjacent glaciers, are broadly arborescent (Fig. 14). They feature long trunk valleys that extend inland to near the Transantarctic Mountains crest, where the inland ice sheet now submerges the western mountains. Some valleys, such as the middle part of Wright Valley (Figs. 15 and 16), have flattish floors and are nearly straight. Others, such as central Taylor Valley (Figs. 11, 17) and east-central Wright Valley (Fig. 16), are sinuous and narrow, with truncated or interlocking spurs. The mouths of Ferrar, Taylor, Debenham, and Mackay Valleys are below sea level, while the central portions of Wright and Taylor Valleys are close to sea level. Several of the deeper valleys show elevated valley escarpment terrains cut in horizontally bedded sandstone in both southeastern and northwestern Australia show canyons, scarps, and theater-shaped embayments; yet they probably developed in a humid climate (Young, 1983, 1987). Also, Moon and Selby (1983) argued that the form of the great Namib escarpments and the western edges of the southern Africa plateau, both of which probably developed largely in semi-arid conditions, are controlled by rock mass strength. This implies a mode of origin which is not dependent on climate. In this regard, Augustinus and Selby (1990) argued that characteristic Dry Valleys slopes are composed of cliffs in strength equilibrium above Richter denudation slopes.

Fig. 13. Top: the Labyrinth in upper Wright Valley, viewed eastward down Wright Valley. We postulate that the Labyrinth was cut by subglacial meltwater streams.
EAST ANTARCTIC ICE SHEET SENSITIVITY

(Note that for technical reasons Figures 23 and 24 are out of sequence).

Fig. 23. Top: oblique aerial view looking southward up the Arena Valley, Quartermain Mountains. Moraine loops in foreground represent incursions of southward-flowing Plio-Pleistocene lobes of Taylor Glacier into lower Arena Valley. Well-developed rectilinear slopes occur on the valley walls. Far in the background is Mt. Feather, which at 2985 m elevation is the highest peak in the Quartermain Mountains. The Sirius Group occurs at 2650 m elevation on the northeast flank of this mountain.

Bottom: oblique aerial view of part of the Rhone platform on the north wall of central Taylor Valley, looking northeast. In the center of the photograph, the Thomson moraine overlies a volcanic complex showing two eruptive events, one isotopically dated at about 3.5 Ma and the other at about 3.0 Ma (Wilch et al. 1993b).

benches. The rock floors of central and eastern portions of Wright and Victoria Valleys today have a reverse inland slope.

The western valley floors rise inland in theater-shaped steps over resistant strata. Some steps have steep risers; some have degraded risers. The heads of all the major valleys are steep walled, semi-circular indentations cut into high planation surfaces. For example, the Taylor-Ferrar valley system has a major step at Cavendish Rocks; other steps occur beneath upper Taylor Glacier. The deep portion of Wright Valley extends far inland to
Fig. 24. **Top:** oblique view of the Arena Valley Ash deposit. The *in-situ* ash overlies a buried ventifact desert pavement at 1410 m elevation in central Arena Valley. Laser fusion $^{40}$Ar/$^{39}$Ar analyses of 18 individual crystals yield an age of 4.343±0.108 for the Arena Valley Ash (Marchant et al. 1993b, c, d).

**Bottom:** *in-situ* volcanic lapilli on top of a well-preserved desert pavement on the Rhone platform near the Thomson moraine in Fig. 23. $^{40}$Ar/$^{39}$Ar analyses indicate that this lapilli (and hence the underlying desert pavement) is 2.97±0.14 Ma (Wilch et al. 1993b).

semi-circular steps that rise abruptly up to the level of the Labyrinth at the head of North and South Forks; another steep semi-circular step at Airdevron six Icefalls separates the Labyrinth surface from an upper planation surface. Victoria Valley has a steep, theater-shaped head now covered with Victoria Upper Glacier.

All the deep transverse trunk valleys have small, high-level tributaries that are cut into an upper-level planation surface. These tributaries grade to an intermediate planation surface or to valley benches and are now left hanging above the major trunk valleys. These hanging tributaries are stubby and broad; some have steep, theater-shaped heads. Only a few major tributaries feed directly into the deep transverse valleys. Taylor Valley has two major tributaries (Beacon and Arena); both enter upper Taylor Valley from the south. Wright Valley has only one major tributary (Bull Pass), which hangs above the central north wall. Victoria Valley has several major tributaries. One is Barwick Valley, which has a steep semi-circular head cut into an upper planation surface. Others are Balham and McKelvey Valleys, which have tapered heads and fluvial notches at the inland escarpment.

Under what climatic regime were the valley systems dissected into the upper platform surfaces? It is obvious from the interlocked and truncated spurs, as well as from the winding valley segments, that fluvial erosion was important in eroding the middle and eastern reaches of Wright, Taylor, and
Ferrar Valleys (Victoria Valley does not preserve spurs). These valley segments are cut in basement rocks and lie well below the high planation surfaces. The valley walls here are generally long, rectilinear, and free of extensive colluvium. This is consistent with slope formation in a weathering-limited environment. The inland reaches of the valleys approach the high-elevation planation surfaces and are cut into slightly dipping Beacon sandstone and Ferrar Dolerite. Here the valley slopes and patterns closely resemble those on the upper planation surfaces that we suggest were cut in a semi-arid environment. It is thus possible that the valley systems were cut through the Transantarctic Mountains under a semi-arid climatic regime, at least in part. Both groundwater and surface flow may have been important in the backwasting necessary to cut the tributary box canyons and the abrupt steps of the western valley floors. But the drainage basins (both surface and groundwater) of these valley systems were sufficiently extensive to support overland stream flow on the basement rocks in the deep lower valleys. The formation of weathering-limited rectilinear slopes in the crystaline and metamorphic rocks of the lower valleys is also consistent with a semi-arid environment. However, it is also possible that the valleys were cut under a more humid climate that allowed development of weathering-limited slopes. This would be consistent with the fossil impression of a Nothofagus leaf in the CIROS-1 drill core that penetrated Oligocene deltaic and glacial sediments at the mouth of Ferrar Glacier trough (Barrett, 1989; Mildenhall, 1989).

Glacial Erosion. Although we suggest that escarpment landscapes are widespread in the Dry Valleys, there also is evidence that glaciers were important in landscape evolution. Middle-Miocene tills rest on the floors of theater-headed valleys in the western Asgard Range and in the Quaternary Mountains (Marchant et al. 1993 a, b), and perhaps on the floors and walls of Wright Valley (Hall et al. 1993). Spurs in Taylor and Wright Valley are
truncated. Victoria and central Wright Valleys are overdeepened and straightened. In Taylor Valley, it is possible that the relict valley represented by sidewall benches was straightened by axial glaciers before fluvial erosion of the inner winding valley (itself with truncated spurs). Cliffed troughs occupy some of the main valleys, such as those of the Mackay and Miller Glaciers (Fig. 12). Other cliffed troughs truncate valley benches in Wright, Victoria, and Taylor Valleys. All are marked by straightened valley sections and the sharp truncation of the gentle upper slopes. Areal scouring is another glacial feature. Typically, landscapes of areal scouring have a relief of tens of meters and bear evidence of roches moutonnées or streamlined forms. Areal scouring occurs on the dolerite of the upper planation surface, the valley benches, and the rolling hills and lowlands of the coast. In several places areal scouring is associated with major subglacial meltwater channel systems, for
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Fig. 16. Oblique aerial view of central and lower Wright Valley, showing straightened valley section in foreground and truncated spurs near the valley mouth. A prominent valley bench occurs in the lower left corner; compare with similar benches in Fig. 17. US Navy, VX8-6 photograph, TMA 1564, f-31, 162.

The CIROS-1 core taken from the seafloor offshore of Ferrar Glacier trough shows intermittent glaciation as early as early Oligocene time (Barrett, 1989). We thus infer that glaciers filled the developing Dry Valleys drainage systems intermittently since early Oligocene time.

Transantarctic Mountains Evolution

The hypothesis that landscape evolution of the Dry Valleys block extends far back into the Tertiary under warmer-than-present climate that allowed fluvial planation and downcutting carries...
with it the implication that the pattern of denuda-
tion and its effects can be tied into the tectonic
evolution of the Transantarctic Mountains. This is
because denudation pulses of backwearing and
fluvial erosion would reflect changing base levels
at the mountain front. It should be pointed out
that such linkage would not occur under the tradi-
tional hypothesis of landscape development of the
Dry Valleys because salt weathering, wind defla-
tion, and glacial erosion in a rainless, frigid desert
have no obvious linkage to base-level changes in-
duced by tectonism. In fact, in this environment
uplift could actually slow erosion by elevating the
landscape surface into a colder regime. It also fol-
lows that under our interpretation of paleocli-
mate, the linkage between erosion and tectonism
would cease with the imposition of a hyper-arid,
cold-desert climate by the end of middle-Miocene
time. Overall, a knowledge of elevation and shape
changes of the land surface is of essential impor-
tance in interpreting late Cenozoic ice-sheet be-
havior (particularly ice-sheet overriding) from gla-
cial drifts that rest on the Dry Valleys landscape.

Fig. 17. Oblique aerial view of middle and lower Taylor Valley, looking to the east. Mount Erebus and McMurdo Sound are
in the background. Prominent benches and truncated spurs occur in the center of the photograph. Remnant fluvial spurs sepa-
rate two lobes of Lake Bonney, which fill the old fluvial valley. US Navy, VXE-6 photograph; TMA 540, f-31, 006.
Geomorphological Map. To exploit the postulated linkage between denudation and tectonic evolution, we constructed a new geomorphological map at a scale of 1:250,000 (Fig. 18) (see Map 1, fold out on back cover). This map, which covers some 7000 km² of the Dry Valleys region, ties together the relict erosional features into overall patterns that show the sequencing of landscape development. The Dry Valleys landscape was classified initially on the basis of form, with further subdivision on the basis of inferred process. Thus the primary classification on the morphogenetic map is into categories such as planation surfaces, cliffs, and rectilinear slopes. The secondary classification distinguishes the types of cliffs, for example, those caused by glacial activity and those formed by subaerial processes. Inevitably, there are many difficulties involved in such an approach, the most important of which is that similar forms may result from different processes. Nevertheless, the results are internally consistent and, even with the inherent limitations, produce new insights. The mapped landforms and their relationship to the tectonic history of the Dry Valleys block of the Transantarctic Mountains are discussed in detail elsewhere (Sugden et al. 1994). Here we present only the main findings.

Three main landform associations are highlighted on the geomorphological map. The first comprises the planation surfaces and erosional remnants which form the highest tablelands, summits, and coastal lowlands. The surfaces rise in steps from the coast; three levels are separated by clear escarpments (Fig. 19). The highest surface, commonly underlain by Ferrar dolerite, occurs at elevations of 2000–2400 m on the inland flank of the Dry Valleys and is bounded on its coastal side by a scarp 500–1200 m high. An intermediate surface occurs at an elevation of 1700–1850 m on interfluves between the Dry Valleys and is increasingly dissected towards the sea until it is terminated by the abrupt 1600-m downfaulted mountain front above the coastal lowland. This lowland, which is generally below 200 m elevation and could be a downfaulted part of the intermediate surface, slopes gently seaward.

Taken together, the pattern of the surfaces, escarpments, and associated landforms demonstrates that the landscape originated by scarp retreat from the coast. Scarp retreat is well illustrated by the way the buttes and inselbergs in the Asgard and Olympus Mountains are progressively smaller and more restricted with increasing distance from the main scarp front. Such backwearing with a series of erosional scarps parallel to the coast is duplicated in southern Africa and in Australia, where it reflects pulses of erosion penetrating inland from the coast due to base-level changes. Given suitable cap rocks, more than one pediplain surface can result from an initial base-level change.

The second landform association consists of the valley systems. The major trunk valleys extend through the faulted mountain front. The sides of these valleys are largely rectilinear in form, although usually they have a concave slope at the foot rather than an abrupt break of slope. These rectilinear valley sides pass uniformly across the mountain front, suggesting that faulting occurred prior to the onset of the present cold-desert climate that terminated all but minor landscape modification. The frontal fault zone truncates an inner bench of Taylor Valley, implying that some faulting postdated initial valley cutting. A final point is that the bedrock floors of Victoria and Wright Valleys now have a reverse, inland slope. One explanation is that tectonic backtilting occurred after the rivers ceased to flow so that they could not excavate new channels to the changing base level.

The third landform assemblage is glacial. The significance of the glacial landforms in the context of the geomorphological map is that glaciers have

See MAP 1, fold-out on back cover

Fig. 18. Geomorphological map showing the distribution of the landscape types and the relationship to reconstructed river valleys.
not contributed the main features of the landscape in the Dry Valleys region, except for a few troughs. However, glacial erosion, although relatively minor, has affected all surfaces from the highest plateaus to the valley bottoms and coastal lowlands. This implies that glacial modification either accompanied or followed the main stages of landscape dissection.

Landscape Subsidence and Uplift. A major implication of this landscape analysis and the limiting ages for the old surfaces is that the Dry Valleys morphology was essentially formed prior to the end of middle-Miocene time and that many of the erosion surfaces are probably much older (Marchant et al. 1993a, b). We postulate that the imposition of a cold-desert environment halted major landscape development in the Dry Valleys by the end of middle-Miocene time. Thus any subsequent elevation changes must be inferred from sediments superimposed on the preexisting landscape surface.

We recognize substantial subsidence (>400 m) of the ancient landscape surface after denudation had effectively ceased, followed by surface uplift of about 300 m since 3.5 Ma ago. The evidence for landscape subsidence comes from fjord sediments and from the present elevation of fluvial valley segments. The inner, lower fluvial segment of Taylor Valley is now close to (or even partly below) sea level. The floor of the Lake Bonney basins, which are fluvial features of this inner valley (Fig. 17), is less than 30 m above sea level (C.H. Hendy, written communication, 1993). Between Canada and Suess Glaciers, the floor of this inner valley is 90 m below sea level (Mudrey and McGinnis, 1975). The preservation of spurs and sinuosity suggests that this inner fluvial valley has not been modified significantly by glacial erosion. Hence the implication is of subsidence of Taylor Valley since fluvial erosion of the narrow, inner valley. Likewise, Ferrar Valley shows truncated and smoothed spurs that extend to near sea level. The bedrock floor of lower Ferrar Valley is now 165 m below sea level. Certainly Ferrar Valley has experienced much more glacier erosion and straightening than has Taylor Valley. Nevertheless, fluvial features are preserved, and the situation is compatible with submergence.

Perhaps the best evidence for landscape subsidence comes from Wright Valley. Fjord sediments in central Wright Valley date to 9±1.5 Ma (Jason glaciomarine diamicton now at 3–250 m elevation) and mid-to-early Pliocene (Prospect Mesa gravels now at 165 m elevation) (Webb, 1972, 1974; Prentice et al. 1993). The fjord(s) were about 100 m deep, and hence reached to about the present-day 300-m contour (Prentice et al. 1993). Any fjord in central Wright Valley must have flooded east-central Wright Valley, simply because it affords the only access for seawater. The floor of east-central Wright Valley, now at 190–270 m elevation, shows evidence of fluvial erosion in the form of interlocking valley spurs. Hence the fjord(s) flooded a fluvial segment of Wright Valley to a present-day elevation of about 300 m. The suggestion is of valley submergence of as much as 300 m, with an unknown but comparatively small adjustment for sea-level change.

If it is assumed that the Dry Valleys formed a coherent tectonic block from north to south, then the combined data from Wright and Taylor Valleys suggest a total subsidence in excess of 400 m. The evidence from Wright Valley fjord sediments indicates that this subsidence had occurred prior to 9±1.5 Ma ago, and that it persisted until sometime after 3.5 Ma ago (Hall et al. 1993). This is consistent with interpretations of benthic foraminifers in eastern Taylor Valley that a deep but shallowing fjord existed between at least ~6.0 Ma and 3.4 Ma ago (Ishman and Rieck, 1992).

Following substantial subsidence, the Dry Valleys landscape surface was uplifted about 300 m in the last 3.5 Ma without reactivating the frontal faults now exposed above sea level. The evidence comes from surface deposits and sediment cores. In Wright Valley, the Pliocene fjord surface, represented in part by the Prospect Mesa Gravels, reached to the present-day 300-m contour, marked by the truncated ends of lateral moraines of small alpine glaciers that flowed down the south valley wall and calved into the fjord (Hall et al. 1993). The youngest of these moraines dates to < 3.5 Ma ago (Hall et al. 1993); this limiting fjord age is consistent with the date based on foraminifers from the C. tufisensis shell beds at Prospect Mesa in central Wright Valley (Webb, 1972, 1974). Thus, we conclude that subsequent to 3.5 Ma ago, the fjord drained and the valley-floor threshold was lifted to its current elevation of 270 m (Calkin, 1974). The magnitude of the uplift is such that it cannot be explained by Pliocene sea level changes (Haq et al. 1988; Hall et al. 1993).

In Taylor Valley, 40Ar/39Ar analyses of numerous in-situ subaerial cinder-cone deposits of Strombolian-style eruptions provide limits on sur-
Fig. 20. McMurdo Volcanic Group outcrops in Taylor Valley. Volcanic outcrops (shaded) include both in-situ and remobilized volcanic rocks. Age estimates of volcanic events are listed at outcrop localities along with lowest in-situ outcrop elevation. See Table 1 for description of age and elevation data (adapted from Wilch et al. 1993a).

Tectonic and Structural Relations. Two lines of evidence suggest that denudation of the Dry Valleys region was associated with rock uplift that was greatest near the coast, declining inland. One is the tilt of the basement surface (Kukri peneplain), which is assumed to have been approximately horizontal at the time of its formation before being covered by a uniform layer of Beacon sediments (Gunn and Warren, 1962). In the inland parts of Taylor and Wright Valleys, this basement surface is exposed at an elevation of 700 and 900 m, respectively. It rises toward the coast at an angle of 2°–3°, reaching elevations of about 1800 m near the mountain front. The basement is downfaulted at the mountain front, as shown by its relationship with particular dolerite sills in a series of fault blocks (Fitzgerald, 1992). This pattern of maximum uplift near the faulted mountain front, declining inland, is mirrored by results of apatite fission-track analysis. Gleadow and Fitzgerald (1987) plotted the dislocation of a pre-uplift, approximately horizontal isochron, which reveals the same faulting with a downthrow of at least 1650 m. As mentioned above, the fact that rectilinear valley sides cut across the faults suggests that much of the faulting was completed before the valleys were vacated by rivers and before the imposition of a hyper-arid, cold-desert environment.

These structural relationships give some insight into the depth of denudation. Making certain assumptions, which are discussed elsewhere (Sugden et al. 1994), we suggest that a wedge of rock some 4–5 km thick has been eroded from the mountain front, declining to about 1 km at the western rim of the Dry Valleys, 75 km inland. Apa-
Table 1. Summary of $^{40}$Ar/$^{39}$Ar Isotopic Age, Paleomagnetic, and Elevation Data from McMurdo Volcanic Group rocks in Taylor Valley

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<th>Sample</th>
<th>Total Gas Age (Ma)</th>
<th>Plateau Age (Ma)</th>
<th>%39 Isotope Correlation Age (Ma)</th>
<th>$^{40}$Ar/$^{39}$Ar Age (Ma)</th>
<th>%39 Pol. Elev. (m)</th>
<th>Elev. (m)</th>
<th>K/Ar Age (polarity)</th>
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<td>TWV87032</td>
<td>2.65 ± 0.19</td>
<td>2.63 ± 0.06</td>
<td>64.5 ± 0.21</td>
<td>296.2 ± 12.0</td>
<td>100.0</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>TWV87037</td>
<td>3.03 ± 0.45</td>
<td>3.08 ± 0.28</td>
<td>96.4 ± 1.16</td>
<td>290.0 ± 13.8</td>
<td>100.0</td>
<td>602</td>
<td>1250</td>
<td>Eroded pyroclastic and lava flow deposits.</td>
</tr>
<tr>
<td>TWV87042</td>
<td>3.38 ± 0.45</td>
<td>3.52 ± 0.20</td>
<td>70.5 ± 0.55</td>
<td>293.9 ± 15.1</td>
<td>100.0</td>
<td>86.3</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TWV87047</td>
<td>3.23 ± 0.43</td>
<td>3.18 ± 0.33</td>
<td>88.8 ± 3.05</td>
<td>297.5 ± 3.5</td>
<td>100.0</td>
<td>94.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TWV87071</td>
<td>1.61 ± 0.14</td>
<td>1.65 ± 0.10</td>
<td>87.1 ± 1.75</td>
<td>297.6 ± 15.5</td>
<td>100.0</td>
<td>382</td>
<td>1677</td>
<td>Multiple vent sites including partially</td>
</tr>
<tr>
<td>TWV87081</td>
<td>2.58 ± 0.25</td>
<td>2.20 ± 0.18</td>
<td>53.2 ± 1.45</td>
<td>295.0 ± 14.5</td>
<td>87.9</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>TWV87082</td>
<td>3.09 ± 0.13</td>
<td>3.00 ± 0.08</td>
<td>196.5 ± 1.9</td>
<td>295.6 ± 19.5</td>
<td>100.0</td>
<td>324</td>
<td>754</td>
<td>Eroded pyroclastic and lava flow deposits.</td>
</tr>
<tr>
<td>TWV87074</td>
<td>2.87 ± 0.61</td>
<td>3.03 ± 0.21</td>
<td>69.2 ± 3.15</td>
<td>290.2 ± 3.7</td>
<td>100.0</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>TWV87073</td>
<td>3.07 ± 0.45</td>
<td>3.08 ± 0.28</td>
<td>96.4 ± 1.16</td>
<td>290.0 ± 13.8</td>
<td>100.0</td>
<td>62.5</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Note: Locations are described relative to nearby alpine glaciers (cross reference with Fig. 2). Listed total gas, plateau, and isotope correlation ages with associated analytical uncertainties (2 standard deviation units). Ages calculated using decay constants recommended by Steiger and Jaeger (1978).

The fission-track analysis also suggests that an eroded wedge of comparable size has been removed in the last 55–60 Ma (Fitzgerald, 1992).

By tectonic relationships and dating, we suggest a hypothesis of landscape evolution tied to the tectonic history of the Dry Valleys block of the Transantarctic Mountains (Fig. 21). The initial stage is rifting and the creation of a new base level along the line of the rift. At this time, the flat, relatively undisturbed surface created by Kirkpatrick Basalt, an equivalent of which today comprises inselbergs rising above the upper planation surface at Allan Nunatak. The second stage is planation that extends inland from the new coastline and subsequently from the axis.
EAST ANTARCTIC ICE SHEET SENSITIVITY

Fig. 21. Model of landscape evolution in the Dry Valleys region showing the situation (a) at continental break-up, (b) during planation under semi-arid conditions, (c) after valley cutting by rivers, and (d) following further minor uplift at the mountain front. Apatite fission-track analyses suggest surface uplift of about 5 km since mountain formation (Gleadow and Fitzgerald, 1987). Our analyses (based on differential uplift of the Kukri Peneplain) suggest rock uplift of 4 km near the mountain front and 2 km inland. Downcutting and backwearing by rivers and glaciers kept pace with tectonic uplift, and valleys were graded to sea level. Denudation since rifting has been 4 to 4.5 km at the coast and 1 km 75 km inland.

of major river valleys. Scarp retreat may have been stimulated by successive changes in base level to form the upper and then the intermediate surface. Alternatively, these surfaces may have resulted from the single base-level change, with different escarpments reflecting lithological con-

traits. The third stage is the dissection of the planation surfaces by the downcutting of valleys to near sea level. Presumably much of this downcutting accompanied faulting and rock uplift of the mountain front. The fourth stage is submergence of the lower reaches of the main valleys by the sea, followed by the subsequent uplift of the landscape surface to present-day elevations.

Our reconstruction supports the view that the Transantarctic Mountains are associated with asymmetric rifting and resulted from the separation of the last fragments of Gondwana about 50–55 Ma ago (Fitzgerald, 1992). The Ross Sea basin was thinned by extension and formed the lower plate, while the Transantarctic Mountains formed the upper plate.

The model of landscape evolution is at an early stage and is discussed elsewhere (Sugden et al. 1994). Nevertheless, the initial conclusions are important. Denudation removed a wedge of rock from the coastal zone and the process was accompanied by rock uplift, mainly along the mountain front. Almost all the denudation was completed by the middle Miocene. This reconstruction agrees with structural and tectonic evidence, as well as with the wider plate tectonic processes that would be expected following the final breakup of Gondwana.

Miocene Ice-Sheet Overriding of the Dry Valleys Region

Denton et al. (1984) argued that a thick ice sheet overrode the Dry Valleys sector of the Transantarctic Mountains, flowing northeastward across major pre-existing valleys. Two overriding episodes were postulated. Denton et al. (1984) further suggested that the Peleus and Asgard tills in Wright Valley and the western Asgard Range, respectively, represent early phases of East Antarctic ice-sheet expansion that led to overriding. Based on the assumption that the Peleus till postdated Prospect Mesa gravels in central Wright Valley, Denton et al. (1984) placed the younger overriding episode in the Pliocene.

Here we again argue for ice-sheet overriding from a suite of glacial erosional features that are superimposed on the Dry Valleys landscape. Most of these features are of relatively small scale, and therefore the resulting imprint on the landscape is minor. The features include zones of areal scouring that are particularly evident on dolerite of the upper plateau surface, on rolling topography near
the coast, and on valley benches. They also include stoss-and-lee features (with lee-side potholes) on sandstone cols in the western Asgard Range and Quartermain Mountains. They include channel systems on areally scoured terrain and across mountain divides (Sugden et al. 1991). Finally, they encompass widespread stripping of surficial sediments from the landscape that left large areas of bare bedrock and remnant patches of drift and colluvium. From our studies in the western Asgard Range and in the Quartermain Mountains we now recognize only one episode of massive overriding (Marchant et al. 1993 a). We concur that Asgard and Peleus tills represent an outlet glacier that filled Wright Valley during the initial phase of overriding. Ice passing across the western Asgard Range from the Taylor Valley drainage system merged with Asgard/Peleus ice. During maximum overriding, coalesced ice masses submerged the Dry Valleys. The major adjustment that we have made in the earlier concept involves the chronology of overriding. On the basis of our new volcanic ash chronology in the western Asgard Range and in the Quartermain Mountains, we now believe that overriding occurred in the middle Miocene, sometime between 14.8/15.2 Ma and 13.6 Ma (Marchant et al. 1993 a). This age is based on 40Ar/39Ar analyses of individual volcanic crystals removed from in-situ ash deposits that occur in stratigraphic association with Asgard till. It is in accord with our conclusion that Peleus till in east-central Wright Valley is >3.8 Ma (Hall et al. 1993).

Because the relative timing of glacial events and landscape subsidence is unknown, there remain several possible explanations for ice-sheet overriding of the Dry Valleys region. One extreme is that a large and thick ice sheet, similar to that reconstructed by Denton et al. (1984), overrode the Transantarctic Mountains with surface elevations similar to present values. This reconstruction came from the fact that the younger of two postulated overriding episodes was assumed to be Pliocene in age by Denton et al. (1984) from the inferred stratigraphic position of Peleus till at Prospect Mesa. An argument was then made that Pliocene landscape elevations were similar to present values, and the reconstruction was made on this basis.

The concept of massive Pliocene overriding of the Dry Valleys region needs revision on several counts. The first is that our new data show that the overriding was, in fact, middle Miocene in age (Marchant et al. 1993 a, b); this renders irrelevant any use of the Pliocene landscape surface as the basis for reconstructing an overriding ice sheet. The second is that we now recognize an episode of considerable landscape subsidence in the evolution of the Transantarctic Mountains, which occurred at least as early as late Miocene time and probably considerably earlier. This opens the possibility that ice-sheet overriding may be due, at least in part, to lowering of the western thresholds, which would promote greatly increased ice flow into the Dry Valleys. There is as yet no geologic evidence that a subsequent rise of these western thresholds cut off overriding ice. Rather, overriding ice had dissipated from the Dry Valleys prior to 13.6 Ma ago (Marchant et al. 1993 a), whereas the documented surface uplift that drained the Wright fjord and shallowed the Taylor fjord was Pliocene/Pleistocene in age. This surface uplift began after 3.5 Ma ago in Wright Valley (Hall et al. 1993) and about 3.4 Ma ago in eastern Taylor Valley (Ishman and Rieck, 1992). However, the geologic record is fragmentary, and there may well have been an early phase of surface uplift prior to 3.5 Ma that has not yet been recognized.

A final point is that increased accumulation, although it may have played a minor part in initiating overriding, probably was not a dominant factor for two reasons. One is that the environment during expansion was a cold desert as shown by the paleoclimatic data from the western Asgard Range and the Quartermain Mountains (Marchant et al. 1993 a, b). Another is that, although the larger alpine glaciers in eastern Wright Valley expanded to flow into the Asgard trunk glacier during the initial phase of overriding, the small snowbanks and glacierets in the western Asgard Range did not. Rather, Asgard ice lobes from the trunk glacier in Wright Valley expanded southward into these valleys (just as today, lobes from Taylor Glacier project southward into Beacon and Arena Valleys in the Quartermain Mountains). The implication is that accumulation during the early phase of overriding was not sufficiently heavy to cause alpine glaciers to expand out of these basins and to merge with the trunk glacier. This puts severe limits on the degree to which increased accumulation on the nearby ice sheet caused overriding.

Huybrechts (1993) points out the difficulties from a glaciological viewpoint for a massive ice-sheet overriding of the present Transantarctic Mountains. Such a reconstruction requires mutually contradictory paleo-environmental condi-
EAST ANTARCTIC ICE SHEET SENSITIVITY

tions. Namely, polar conditions are required for expansion of grounded ice across the Ross Sea floor, yet warmer-than-present conditions are required for the greatly increased accumulation needed for interior thickening. Moreover, increased accumulation rates would not cause interior ice-surface elevations to rise significantly, because of changed ice temperatures and reduced areal extent. The implication of these modelling experiments is that overriding is difficult to achieve with the present topography, unless the climate system operated in a completely different way (which our paleoclimatic data suggest was not the case). It follows that any evidence for mountain overriding may, in fact, reflect landscape subsidence and be in itself a clue to the evolution of the Transantarctic Mountains.

Within these constraints, there are at least two possible explanations for overriding. The first is simply that landscape subsidence alone was sufficient to submerge the Dry Valleys region by inland ice. Another explanation for overriding is that a polar climate allowed ice shelves and then a grounded ice sheet to form in a shallower-than-present Ross Sea. Low portions of the Transantarctic Mountains (such as the subsided Dry Valleys sector) were submerged by an expanded Antarctic Ice Sheet that resulted from such grounding. This ice sheet need not have been nearly as high as that reconstructed by Denton et al. (1984) in order to override a Dry Valleys land surface that was considerably lower than now, and thus would not have required greatly increased interior accumulation.

The dissipation of overriding ice from the Dry Valleys that led to the present ice-free conditions occurred in cold-desert environmental conditions and was accomplished largely by sublimation prior to 13.6 Ma ago (Marchant et al. 1993 a, b). One possible reason for such recession is that the overriding ice sheet deepened its bed by stripping sediments in the Ross Sea and by cutting major troughs through the Transantarctic Mountains. Once the ice sheet was then perturbed, it retreated by grounding-line recession from this deeper bed back to its present configuration. Another possible reason is that there may have been an early phase of surface uplift prior to 3.5 Ma that has not yet been recognized in the geologic record. Any such uplift of the western rim of the Dry Valleys could have greatly reduced the inflow of East Antarctic ice and initiated the Taylor Dome on the inland flank of the Dry Valleys.

We are not able to resolve these issues with the data at hand from the Dry Valleys. Answers may come from investigations in the Royal Society Range to the south (where areally scoured bedrock from overriding ice extends over a considerable elevation range) and in the Convoy Range to the north (which is marked by extensive areal scouring and channel systems from overriding ice). Meanwhile, we can draw one conclusion that is central to the present discussion. Namely, ice-sheet overriding of the Dry Valleys region, no matter what its cause, antedated 13.6 Ma ago and hence occurred in the Miocene, not in the Pliocene as required by the deglaciation hypothesis.

Limited Pliocene Glacier Expansion

The extent of Pliocene glacier expansion in the Dry Valleys is a key to evaluating the deglaciation hypothesis. An important requirement of this hypothesis is that ice-sheet expansion across the Transantarctic Mountains occurred after 3.0 Ma ago (Barrett et al. 1992), in order to emplace late Pliocene marine diatoms stripped from interior marine basins into Sirius Group outcrops, some of which occur on the high inland erosion surfaces of the Dry Valleys. We have direct glacial geologic control, coupled with numerical dates, that places limits on the Pliocene expansion of Taylor Glacier (Wilch et al. 1993; Marchant et al. 1993 b; Marchant et al. 1994) and of alpine glaciers on the south wall of central Wright Valley (Hall et al. 1993). In addition, we have complementary evidence that large tracts of the Dry Valleys remained free of glacier ice throughout the Pliocene Epoch (Marchant et al. 1993 a, b, c, d).

Taylor Glacier

Taylor Glacier originates at Taylor Dome and extends for nearly 50 km into Taylor Valley, where it terminates in Lake Bonney at about 50–60 m elevation. A Pliocene longitudinal profile of Taylor Glacier (Fig. 22) comes from two key moraine sequences (Figs 23, page 175, and 25). The first is on the north valley bench in Taylor Valley below Mt. J.J. Thomson (Fig. 26) (Wilch et al. 1993). The second is in Arena Valley in the Quartermain Mountains on the south side of upper Taylor Glacier (Fig. 23) (Marchant et al. 1993 b).

Figure 26 shows the distribution of glacial drift sheets and McMurdo Volcanic basalts on the Rhone platform. The key outcrop that limits Plio-
Reconstructions that assume plastic flow closely approximate the present ice-surface profile of Taylor Glacier and show that a 325 m thickening at Arena Valley (Taylor IVa, Quaternary maximum) yields a corresponding ice-surface rise at Taylor Dome of about 160 m (Marchant et al. 1994). Similarly, the maximum Pliocene-age drift, represented by Taylor IVb in Arena Valley and the Thomson moraine in Taylor Valley, yields a corresponding ice-surface rise at Taylor Dome of about 250 m. This implies limited Pliocene East Antarctic Ice Sheet expansion in this sector of Antarctica. Location of cross sections shown in map insert.

cene ice extent occurs on this bench near the Matterhorn Glacier. Here a complex of pyroclastic and lava-flow deposits of the McMurdo Volcanic Group extends from 1250 m to 602 m elevation (Wilch et al. 1993). Glacial drift(s) on the valley bench can be traced westward along the north valley wall. The configuration of this drift(s) shows that it was deposited by an expanded Taylor Glacier. The Thomson moraine, which marks the upper margin of this drift sheet, passes across the surface of the volcanic complex at 1082 m elevation. Based on the whole-rock $^{40}$Ar/$^{39}$Ar ages of underlying volcanics, the Thomson moraine has a maximum age of 2.97 ± 0.14 Ma. It represents the greatest expansion of Taylor Glacier in the last 3.47 Ma, because numerous $^{40}$Ar/$^{39}$Ar ages show that the volcanic complex on which the moraine rests extends at least this far back into Pliocene time. A minimum age for the Thomson moraine is afforded by a $^{40}$Ar/$^{39}$Ar whole-rock date of 2.71 ± 0.14 Ma ago for the portion of Rhone volcanic cone on the valley bench near Rhone Glacier above 900 m that remains pristine and has not been overridden by glacier ice. This pristine portion of the Rhone cone is lower than, and up glacier from, the upper limit of the Thomson

Figs 23 and 24, see pp. 175–176.
moraine. Hence, deposition of the Thomson moraine must have antedated the cone, which began to erupt at about 2.71 Ma. Therefore, we conclude that the Thomson moraine and associated drift is between 2.7 and 2.97 Ma in age. We note, also, that the maximum Quaternary expansion of Taylor Glacier is marked by a line of erratic boulders that crosses the base of the Rhone cone at 900 m elevation. This represents the greatest ice extent since the eruption of the Rhone cone ceased about 1.50±0.05 Ma ago (Wilch et al. 1993).

Figure 27 (see Map 2, fold out on back cover) shows a series of drift sheets, most with boulder-belt moraines, in lower Arena Valley in the Quartermain Mountains of the western Dry Valleys re-
region. These drift sheets were deposited by a southward-flowing marginal lobe of Taylor Glacier (Marchant et al. 1994). A drift chronology comes from surface-exposure ages of boulders from moraine crests, using in-situ cosmogenic $^3$He and $^{10}$Be. The results show that the drifts range from $>$4.4 Ma old (Quartermain I) to about 100,000 yr old (Taylor II). Taylor IVb drift has a minimum age of about 2.1 Ma. Taylor IVa drift predates 1.3 Ma.

Table 2 shows the tentative correlation among drifts deposited by Taylor Glacier in upper and lower Taylor Valley. Taylor IVb and Thomson drifts are correlated on the basis of limiting numerical ages and of relative position within drift sequences. Figure 22 shows the longitudinal profile of Taylor Glacier when it stood at the Taylor IVb and Thomson ice limits. This profile represents the maximum expansion of Taylor Glacier in at least the last 3.47 Ma (Rhone platform) or 4.4 Ma (lower Arena Valley). The results show that the rise of the ice surface during the Taylor IVb/Thomson expansion was $<$250 m at the Taylor Dome and about 475 m near Arena Valley; the ice surface reached to 1082 m elevation at the Thomson moraine. To place this Pliocene expansion in perspective, we compare it in Figure 22 with the documented Quaternary advances of Taylor Glacier. The earliest and greatest such advance (Taylor IVa in Arena Valley and the Rhone cone erratic line near Mt. J. J. Thomson; Marchant et al. 1993 b; Wilch et al. 1993b) occurred in early Quaternary time, after 1.5 Ma ago. It is evident from Figure 22 that the maximum Pliocene expansion of Taylor Glacier was only marginally greater than the largest Quaternary advance. It is not clear whether this minor difference represents a real but small paleoclimatic difference or whether it is somehow related to the Pliocene/Pleistocene surface uplift of about 300 m documented for the Dry Valleys sector of the Transantarctic Mountains.

The Pliocene reconstruction of Taylor Glacier has a bearing on the interpretation of the Sirius Group. Sirius Group sediments crop out in the Dry Valleys region at Table Mountain (Barrett and Powell, 1982), Mt. Feather (Brady and McKelvey, 1979, 1983), just south of Mt. Fleming, and just north of Shapeless Mtn. (McKelvey, 1991) (Fig. 18). The deposit at 2650 m elevation on the north flank of Mt. Feather was one of the original Sirius outcrops used to establish the deglaciation hypothesis (Webb et al. 1984; Harwood, 1986), because it contains a diverse diatom flora that includes mid-to-late Pliocene elements. This deposit

See MAP 2, fold-out on back cover.

Fig. 27. Glacial geologic and surficial geomorphic map of Arena Valley. From Marchant et al. (1993b).

Fig. 26. Map of surficial deposits on Rhone platform from Wilch et al. (1993b). Light gray areas delineate Thomson drift.
Table 2. Suggested Correlations of Pliocene-Pleistocene and Older Drifts in Lower Arena, Central Taylor, and Eastern Wright Valleys and in the Western Asgard Range

<table>
<thead>
<tr>
<th>Lower Arena Valley</th>
<th>Central Taylor Valley</th>
<th>Eastern Wright Valley</th>
<th>Western Asgard Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>Taylor II drift</td>
<td>Bonney Drift</td>
<td>Alpine II</td>
<td></td>
</tr>
<tr>
<td>Taylor III drift</td>
<td>Taylor III drift</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Taylor IVa drift</td>
<td>Taylor IVa drift</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Taylor IVb drift</td>
<td>Thomson moraine</td>
<td>Alpine III</td>
<td></td>
</tr>
<tr>
<td>Quartermain I till</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Quartermain II till</td>
<td></td>
<td>Pelcus till</td>
<td>Asgard till</td>
</tr>
</tbody>
</table>


is perched on an old geomorphologic surface at the top edge of the steep Beacon Valley headwall (Fig. 18) (Brady and McKelvey, 1979, 1983). It consists of local rocks and rests on striated sandstone bedrock. The implication drawn by Webb et al. (1984) from the enclosed marine diatoms is that the Wilkes Subglacial Basin inland of the Dry Valleys was flooded with marine waters in mid-to-late Pliocene time. Subsequent overriding of the Transantarctic Mountains by a wet-based ice sheet would be required to emplace the Sirius outcrop on Mt. Feather. Barrett et al. (1992) implied that overriding occurred after 2.8 ± 0.3 Ma ago.

The Pliocene glacial-geologic reconstruction of Taylor Glacier affords an independent test of whether mid-to-late Pliocene marine diatoms could actually have been emplaced in the Mt. Feather Sirius outcrop at 2650 m elevation by the East Antarctic Ice Sheet (and hence whether Wilkes Basin was flooded in mid-to-late Pliocene time). Mt. Feather is the highest peak of the Quartermain Mountains, which lie between the upper reaches of Taylor and Ferrar Glaciers. Taylor Glacier originates at Taylor Dome, which rises 150 m above the East Antarctic ice-sheet surface and merges with a broad ice divide that extends far inland to Dome Circé (Drewry, 1982). Such an ice configuration suggests that the interconnected Taylor and Ferrar Glaciers monitor local and regional East Antarctic ice-sheet fluctuations.

The independent test involves a comparison between the maximum Pliocene-Pleistocene ice-surface elevation of the Taylor and Ferrar Glaciers and the elevation of the Sirius outcrop on Mt. Feather. If the maximum Pliocene-Pleistocene ice-surface elevation lies below 2650 m elevation, then the Sirius Group outcrop could not have been deposited during Pliocene-Pleistocene time by East Antarctic ice. If, on the other hand, upper Taylor and Ferrar Glaciers thickened to at least 2650 m, then this would allow emplacement of the Sirius Group outcrop on Mt. Feather by East Antarctic ice that submerged the Quartermain Mountains during Pliocene/Pleistocene time.

Figures 23 and 25 show the location of the Arena Valley moraines relative to the Mt. Feather Sirius outcrop and to the Thomson moraine in central Taylor Valley. Figure 22 shows the reconstructed Pliocene longitudinal profiles for Taylor Glacier based largely on the Thomson moraine and Taylor IVb drift, from near Mt. J. J. Thomson and in Arena Valley, respectively. The Thomson/Taylor IVb longitudinal profile represents the surface of Taylor Glacier sometime between 2.71 and 2.97 Ma ago, a position not exceeded in at least the last 3.47–4.4 Ma. Because it is physically connected with Taylor Glacier, Ferrar Glacier almost surely showed the same behavior.

The key point of Figures 22, 23, and 25 is that the upper reaches of Taylor Glacier did not thicken nearly enough in the last 3.47–4.4 Ma to engulf Mt. Feather. Rather, ice lobes from Taylor Glacier pushed only into the mouths of Arena (and Beacon) Valley. This conclusion is consistent with the implication of surface volcanic ash deposits that most of Arena and Beacon Valleys remained free of ice in Pliocene time (Marchant et al. 1993a,b,c,d). For example, in Arena Valley, an in-
situ ashfall layer (Arena Valley Ash) dated to 4.34 Ma ago rests on a desert pavement (Marchant et al. 1993c); in Beacon Valley, a volcanic ashfall deposit dated to 10.4 Ma ago fills a relict thermal contraction crack that marks the same surface as today (Marchant et al. 1993d,e).

Thus we conclude that the mid-to-late Pliocene diatom flora in the Mt. Feather Sirius outcrop could not have been emplaced by an expansion of the East Antarctic Ice Sheet that engulfed the Quartermain Mountains.

Wright Valley Alpine Glaciers

The Goodspeed, Hart, Meserve, Bartley, and Conrow Glaciers head in short theater-shaped valleys in the eastern Asgard Range and flow down the south wall of east-central Wright Valley. The areal distribution, stratigraphy, and limiting 40Ar/39Ar dates of lateral drift sheets delineate the maximum mid-to-late Pliocene areal extent of these glaciers.

Figure 28 (see Map 3, fold-out on back cover) shows the distribution of surficial deposits in east-central Wright Valley. Peleus till, at >3.8 Ma old, is stratigraphically lowest and represents the last expansion of East Antarctic ice through Wright Valley. Drift units (Alpine I, Alpine II, Alpine III, and Alpine IV) flanking the alpine glaciers on the south valley wall are younger than Peleus till and record subsequent alpine-glacier fluctuations. On the basis of 40Ar/39Ar dates of reworked basalt clasts enclosed in these drifts (Hall et al. 1993), Alpine IV drift (the oldest alpine drift) is >3.7 Ma old, and Alpine III drift is <3.5 Ma old. Advanced soil development (weathering Stage 5 of Campbell and Claridge, 1987) of Alpine III drift suggests a probable Pliocene or early Quaternary age (Hall et al. 1993). Finally, Wright, Onyx, and Loop drifts, as well as Alpine I and II drifts, are Quaternary in age (Hall et al. 1993).

These data point to the striking conclusion that Wright Valley alpine glacier tongues have expanded by only several hundred meters, to the Alpine III drift limits, in the past 3.5 Ma. For glacier size, this result is compatible with that for Taylor Glacier. Taylor Glacier (100 km in length) advanced only through lower Taylor Valley and into lower Arena Valley within the last 3.47 Ma years; the maximum Pliocene extent barely exceeded the maximum Quaternary advance. Wright Valley alpine glaciers (4–6 km in length) advanced only several hundred meters in the past 3.5 Ma years; again the maximum position achieved in the past 3.5 Ma (Alpine III drift) barely exceeded the Quaternary maximum extent (Alpine II drift).

Ice-free Terrain From Surficial Ash Deposits

40Ar/39Ar dated, in-situ volcanic ashfall deposits in contraction cracks, avalanche deposits, or on well-preserved desert pavements show that large tracts of the Dry Valleys remained free of ice during the Pliocene Epoch (Figs 29 and 30). The Arena Valley Ash, which rests on a desert pavement at 4140 m elevation in central Arena Valley, indicates that ice-free conditions existed at this site 4.34 ± 0.108 Ma ago (Fig. 24) (Marchant et al. 1993c). Similarly, the Hart Ash, which overlies colluvium on the south wall of lower Wright Valley between the Hart and Goodspeed Glaciers (Hall et al. 1993), indicates that ice-free conditions existed at this site at 3.9 ± 0.3 Ma ago. The areal distribution of these ashfall deposits shows that alpine glaciers in lower Wright and Arena Valleys could not have been much more extensive at times of Pliocene ashfall than they are now. In addition, the preservation of in-situ ashfall deposits on valley floors and walls indicates that erosive wet-based glaciers could not have extended across ashfall sites since at least mid-Pliocene time, otherwise ash deposits would show evidence of significant reworking or else lie buried beneath Pliocene-age till, which is not the case (Marchant et al. 1993c). In particular, the Arena Valley Ash limits Pliocene expansion of upper Ferrar Glacier during the last 4.34 Ma. The Arena Valley Ash is located about 4 km north of Ferrar Glacier. A low bedrock threshold (1600 m elevation) at the head of Arena Valley now prevents Ferrar Glacier from spilling into upper Arena Valley and covering the Arena Valley Ash. In places, this bedrock threshold lies only 30 m above the present ice surface of Ferrar Glacier (Marchant et al. 1993 b). Ancient tills and glacial erosional features that occur along this threshold indicate that wet-based ice from Ferrar Glacier advanced northward into Arena Valley at
times prior to 7.4 Ma ago (Marchant et al. 1993 b). However, the preservation of the in-situ Arena Valley Ash suggests that no wet-based ice advanced across this threshold and into central Arena Valley during at least the last 4.34 Ma ago. These data are consistent with limited glacier expansion during Pliocene time.

Summary
Our results from Taylor and Wright Valleys are consistent in showing relatively restricted Pliocene glacier expansion, relative to current glacier size. The large Taylor Glacier expanded into lower Taylor and Arena Valleys whereas the much smaller Wright Valley alpine glaciers simply advanced down the valley wall. In both cases these advances were only slightly more extensive than the maximum Pleistocene advance. Thus a noticeable re-
result is that there was little discrepancy between Pliocene and Pleistocene glacier behavior. The implication is that any Pliocene thickening of sectors of the East Antarctic Ice Sheet adjacent to the Dry Valleys was minimal. This is consistent with the volcanic ash data, which shows that wide tracts of the western Dry Valleys adjacent to the inland ice sheet remained free of glacier ice throughout the Pliocene Epoch (Marchant et al. 1993 a, b).

**Cold-desert Pliocene Paleoclimate**

**General Statement**

One important requirement of the deglaciation hypothesis is that Pliocene air temperatures were sufficiently warmer than at present (20°–25°C above present-day values if the Beardmore Sirius deposits were near sea level) to allow growth of *Nothofagus* trees or shrubs at 85°S latitude (Barrett, 1991; Webb and Harwood, 1991), and also to allow surface melting zones to be superimposed on the East Antarctic Ice Sheet (Huybrechts, 1993). In Antarctica, an atmospheric warming of 20°C would place the 0°C isotherm high on the Dry Valleys landscape. Since the 0°C isotherm occurs at approximately the position of the snowline on mountain glaciers north of the Antarctic Convergence, such warming would introduce melting ablation zones on Dry Valleys glaciers, which therefore would have left a temperate glacier landform assemblage, including extensive outwash plains (with coarse gravel), kame terraces, and massive moraines (with enclosed lodgement and meltout till). In addition, one would expect evidence of water activity on steep hillslopes, including debris flows and stream gullies. With this background, we now examine the landform assemblage and landscape surfaces of the Dry Valleys region for evidence of Pliocene paleoclimate.

**Dry Valleys Hillslopes**

The rectilinear slopes in the Dry Valleys are largely relict and had basically achieved their current form by the end of Miocene time. There is little evidence of Pliocene development of these recti-
linear slopes beyond the minor weathering that is occurring today. Thus, in our view, the evolution Dry Valleys hillslopes shows no evidence of enhanced Pliocene warmth accompanied by increased slope development. Our evidence, which comes from the high western Asgard Range and Quatermain Mountains and from eastern Wright Valley, is reviewed by Marchant et al. (1993 a, b) and Hall et al. (1993). In summary, Peleus till, which is >3.8 Ma old, rests on the south rectilinear wall of central Wright Valley to 1150 m elevation with little disturbance. Pliocene alpine moraines also rest on this rectilinear slope with little modification. In the western Asgard Range, in-situ Miocene and Pliocene glacial sediments and colluvium cover portions of rectilinear slopes. In Arena Valley in the Quatermain Mountains, Pliocene and Miocene drifts and ash-avalanche deposits rest on essentially unmodified rectilinear slopes. A comparison of adjacent exposed and colluvium-covered segments of rectilinear slopes suggests that current weathering processes have generally removed no more than 0.5–3.0 m of exposed bedrock since late Miocene time.

Terrestrial-Marine Comparisons

Background. Our purpose here is to compare inferred paleo-marine conditions in the Dry Valleys with our paleoclimatic interpretations derived from the terrestrial stratigraphic record. We wish to determine if there was substantial atmospheric warming compatible with growth of Nothofagus 600 km to the south at Beardmore Glacier. Or did cold-desert environments persist, indicating only modest warming at best? The necessary terrestrial-marine comparisons to address these issues can be made for crucial Pliocene intervals in Wright Valley and also in Ferrar and Taylor Valleys.

Pliocene marine water bodies occupied eastern Ferrar Glacier trough, eastern Taylor Valley, and Wright Valley. Foraminiferal and diatom biostratigraphy has been carried out for the following marine cores: CIROS-1 and CIROS-2 (Ferrar Glacier trough; Harwood, 1989; Webb, 1989), DVDP-10 and DVDP-11 (eastern Taylor Valley; Ishman and Rieck, 1992), and DVDP-4A (Wright Valley; Brady, 1982) (see Fig. 6). In addition, there is biostratigraphic control for the Prospect Mesa section that exposes Pliocene marine sediments (Prospect Mesa gravels) in central Wright Valley (Webb, 1972, 1974; Prentice et al. 1993). The biostratigraphic chronologies permit valley-to-valley correlation of marine sediments, and equally important, paleo-water temperatures can be inferred from the microfossil assemblages in the cores and the Prospect Mesa gravels.

One example comes from central Wright Valley. Webb (1972, 1974) concluded from the benthic foraminiferal assemblage in Prospect Mesa gravels (Pecten gravel) that bottom-water temperatures in a shallow Wright Valley fjord (~100 m paleo-water depth) were within the range of -2°C to +5°C, and perhaps as much as +10°C. The associated diatom assemblage, along with the absence of coccolithophores, is taken to indicate water temperature of 0°C to <3°C (Burckle and Pokras, 1991; Prentice et al. 1993). Prospect Mesa gravels contain the mid-Pliocene foraminifer marker species Ammoelpidiella antarctica (Trochoelpidiella onyx of Webb, 1972, 1974), which in Taylor Valley core DVDP-10 is taken to occur at 3.4–3.8 Ma ago (Ishman and Rieck, 1992). Burckle et al. (1986) found late Pliocene diatoms in Prospect Mesa gravels now taken to suggest an age of 2.5 to 3.0 Ma from diatom biostratigraphy (Prentice et al. 1993). However, the 87Sr/86Sr ratios of two shells (C. tuftsensis) assumed to be unaltered from Prospect Mesa gravels suggest a date of 5.5±0.4 Ma (Prentice et al. 1993), which is considerably older than the age estimates from biostratigraphic ranges of the enclosed microfossils.

Another example comes from the CIROS-2 marine core from Ferrar Glacier trough (Barrett et al. 1992). This core reveals the Thalassiostra insigna/T. vulnifica and T. vulnifica diatom zones. In Southern Ocean subantarctic cores, T. insigna and T. vulnifica occur together at 2.5–3.1 Ma; T. vulnifica spans the range 2.2–3.1 Ma. A volcanic ash in the CIROS-2 core interval containing T. vulnifica yielded single crystal and glass laser-fusion 40Ar/39Ar dates of about 2.8±0.3 Ma in age. This is taken as the age of the T. vulnifica zone in Ferrar Glacier trough. Most important, however, the age is also taken to date marine flooding of interior East Antarctic marine basins, because Tinsigna and T. vulnifica occur together in Sirius Group outcrops in the Transantarctic Mountains. From other diatoms also enclosed in Sirius outcrops, these interior marine water bodies are inferred to have had surface temperatures of 2°C–5°C (Harwood, 1986). The implication of such relatively warm interior water bodies is that similarly warm water must have occurred in McMurdo Sound and the Dry Valleys fjords. By this scenario, the Dry Val-
ley regions would be nearly surrounded by relatively warm marine water. Not only would these water bodies include McMurdo Sound and its extension into Ferrar, Taylor and Wright Valleys, but also a flooded interior Wilkes Basin on the western mountain flank, replacing part of the present-day East Antarctic Ice Sheet. Such a paleogeographic reconstruction is consistent with the growth of Nothenofagus in the Transantarctic Mountains as far south as 85°S latitude alongside Beardmore Glacier.

Wright Valley. Our terrestrial-marine comparison in Wright Valley depends on the relationship of the Alpine III and IV moraines, discussed in detail by Hall et al. (1993), to the Prospect Mesa gravels of Pliocene age. The stratigraphic position of Peleus till (which is widespread on the floor of central and east-central Wright Valley) relative to Prospect Mesa gravels is crucial to the interpretation of Wright Valley glacial and marine history. At Prospect Mesa in central Wright Valley where a piece of Peleus till rests on Prospect Mesa Gravel, Denton et al. (1984, 1991) assumed that this overlying Peleus till was in-situ and therefore postdated Prospect Mesa gravels. By this interpretation, the marine flooding of Wright Valley (represented by Prospect Mesa gravels) antedated through-valley Peleus glaciation. But it is now thought that the piece of Peleus till that rests on the Prospect Mesa gravels at Prospect Mesa almost surely slumped down the north valley wall (Prentice et al. 1993). The implication is that the widespread Peleus till on the valley floor is, in fact, older than the subsequent fan deposits below Bull Pass (including the Prospect Mesa gravels). This reinterpretation is based on irregular fabrics and platy structure in the retransported Peleus till, taken together with the discovery of two new fossil pecten localities on portions of adjacent fan surface not covered with Peleus till.

If correct, this new interpretation of the stratigraphy at Prospect Mesa means that marine flooding of Wright Valley represented by Prospect Mesa gravels occurred after Peleus glaciation, not before. It also means that the Prospect Mesa gravels and the Alpine IV/III drifts occupy the same relative stratigraphic position in Wright Valley; both are younger than Peleus till and both antedate extensive soil development (weathering stage 6 of Campbell and Claridge, 1987). The Prospect Mesa gravels are mid-to-late Pliocene in age from benthic foraminifers and from diatoms, and 5.4 Ma in age from $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of two C. tuftensis valves. From $^{40}\text{Ar}/^{39}\text{Ar}$ dates of reworked basalt clasts, Alpine IV drift is >3.7 Ma old, and Alpine III drift <3.5 Ma old (Hall et al. 1993).

As evident in Figure 28, there is a striking difference in the areal pattern of Alpine III and IV drifts as compared to Alpine II drift alongside Conrow, Bartley, Meserve, Hart, and Goodspeed Glaciers on the south wall of Wright Valley. In all cases, Alpine II drift forms complete terminal loops as well as lateral moraines. In no case except for Goodspeed Glacier does either Alpine III or Alpine IV drift form terminal loops. In fact, Alpine III and IV drifts are strikingly absent on the valley floor and comprise prominent lateral moraine complexes with lower ends that terminate at 250–300 m elevation on the south valley wall. In two cases (Conrow and Bartley left lateral), flat and irregular tongues of colluviated drift extend downslope from the lateral moraines to near the valley floor.

Given the available chronologic and stratigraphic data, our favored explanation for the lack of terminal loops is that the Alpine III and IV glaciers calved into a fjord that flooded Wright Valley to an elevation of about 300 m (Hall et al. 1993), and was drained by the time Alpine II drifts were deposited. The downslope termination of Alpine III and IV lateral moraines approximately mark the former shoreline on the south valley wall, where the alpine glaciers had calving termini. The flat and irregular drift tongues below these lateral moraines represent downslope slumps of drift. The chronologic implication of this reconstruction is that a marine water body existed in Wright Valley both >3.7 Ma (Alpine IV drift) and <3.5 Ma (Alpine III drift). This explanation is consistent with the presence of an early-to-mid-Pliocene fjord in Wright Valley inferred from the Prospect Mesa gravels.

If our explanation of the lack of terminal Alpine III/IV loops is correct, then we can reconstruct the approximate extent and depths of the associated marine water body in Wright Valley. If we assume little tilting of Wright Valley from Pliocene/Pleistocene surface uplift, then marine waters extended westward from McMurdo Sound into North and South Forks. Near Prospect Mesa the marine water depth was about 135 m deep, a value consistent with the 100 m paleo-depth estimated from foraminiferal fauna (Webb, 1972, 1974). The marine water body was narrow and shallow in eastern Wright Valley where truncated fluvial spurs are preserved on both valley walls. At the valley...
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threshold near present-day Lake Brownworth, marine water depth was only 20–30 m.

The evidence that comes from the physical character of the alpine drifts suggests that the Alpine III/IV glaciers were cold-based and existed in a cold-desert environment. There are no temperate glacier features, such as basal till, striated bedrock (even where the bedrock has a protective sediment cover), kame terraces, bulky moraines (with enclosed till layers), waterlain till at former grounding lines, or coarse gravel outwash above the former marine shoreline. Moreover, there is remarkably little glaciomarine sediment on the valley floor. Rather, the complex of alpine glacial deposits is typical of the present cold-desert regime. The drift sheets are composed of coarse gravel with numerous ventifacts, almost no striated stones, and no enclosed lodgement or meltout till layers. Alluvial fans deposited by the meltwater streams are similar to modern fans. The existence of cold-based ice is consistent with the fact that the alpine glacier tongues have not carved troughs on the rectilinear valley wall. This overall argument for cold-desert conditions from former alpine glaciers also is in accord with little-to-no slope development in central Wright Valley since deposition of Peleus till and Hart Ash (Hall et al. 1993). We argue, then, that temperate ice tongues with surface melting zones did not exist even at sea level, during deposition of Alpine III and IV drifts when a fjord occupied Wright Valley.

Ferrar and Taylor Valleys. From ⁴⁰Ar/³⁹Ar chronologies, we can link the key T. vulnifica zone in the CIROS-2 core in Ferrar fjord (Barrett et al. 1992) with our terrestrial sequence (Wilch et al. 1993). The ash layer in the T. vulnifica zone is about 3.0 Ma old (2.8 ± 0.3 Ma) (Barrett et al. 1992). From the ⁴⁰Ar/³⁹Ar chronology of the Thomson moraine at 1082 m elevation near Mt. J. J. Thomson in lower Taylor Valley (Wilch et al. 1993), along with the surface-exposure chronology of a drift sequence in lower Arena Valley (Brook et al. 1993), we have reconstructed the longitudinal profile of Taylor Glacier at 2.71–2.97 Ma ago. The results show that Taylor Glacier, which drains the peripheral Taylor Dome of the East Antarctic Ice Sheet, was then at its maximum extent in the past 3.47 Ma, rather than in a collapsed state (Marchant et al. 1993b; Wilch et al. 1993b).

What was the paleoenvironment in Taylor Valley at ~3.0 Ma ago? We can answer this question again with data from near Mt. J. J. Thomson and in Arena Valley. The upper moraine limit of Thomson drift at 1082 m elevation rests on a 1–1.5 m-thick deposit of lapilli with included volcanic bombs (Fig. 24, p. 176) (Wilch et al. 1993b). The preservation of delicate spines on all lapilli pieces shows that the deposit is in-situ. This lapilli deposit makes up part of a large volcanic complex of pyroclastic and lava-flow deposits. This lapilli rests on a well-developed, in-situ desert pavement formed of an interlocking mosaic of gravel-sized ventifacts of dolerite and granite. Volcanic bombs encased within the lapilli deposit yielded an ⁴¹Ar/³⁹Ar date 2.97±0.14 Ma. We take this to be the age of burial of the desert pavement, because in-situ lapilli rests on the pavement surface and infills cavities between ventifacts (Wilch et al. 1993b).

The existence of this desert pavement, similar in all respects to modern desert pavements, indicates that the Dry Valleys were arid at ~3.0 Ma ago. Our analysis of the landforms in Arena Valley alongside upper Taylor Glacier indicates that the environment was also polar as well as arid (Marchant et al. 1993b). The evidence comes from the physical characteristics of Taylor IVb drift (the basis of the longitudinal profile of Taylor Glacier at ~3.0 Ma). This drift is composed predominately of boulder moraines with a matrix of loose gravel with numerous ventifacts and no striated clasts. In all respects except surface weathering, these Taylor IVb moraines are identical to moraines of Taylor II and III drifts deposited in Arena Valley during late Quaternary time (Marchant et al. 1994; 1993b). We thus infer that Taylor IVb moraines were deposited by a cold-based lobe of the Pliocene Taylor Glacier (Marchant et al. 1993b). This inference is supported by the lack of meltwater channels, fans, or lacustrine sediments associated with Taylor IVb drift. It is also in accord with the conclusion of Marchant et al. (1993c) that polar desert conditions have persisted in Arena Valley for at least the past 4.34 Ma from an in-situ ashfall layer resting on an desert pavement in Arena Valley. Taken together, these data indicate that the average temperatures during the Pliocene in Arena Valley were no more than 3°C above present values (Marchant et al. 1993b,c,d).

Surficial Ash Deposits

Under the present hyper-arid cold-desert climate, the surface of unconsolidated deposits in the Dry Valleys region shows well-developed ventifact
pavements, interlocking contraction cracks (sand wedges of Marchant et al. 1993d), and coarsegrained gravel lags. Sand wedges form only in cold continental climates, where mean annual temperatures are below 0°C (Pewé, 1959, 1966; Berg and Black, 1966; Black, 1976) and ventifact pavements with associated gravel lags generally form only in arid climates. The discovery of Miocene- and Pliocene-age volcanic ash in stratigraphic association with these morphologic forms allows construction of detailed paleoclimate records. We have identified and dated over 50 different surficial ash deposits in the Dry Valleys region (Marchant et al. 1993a,b,c,d). Most of these ash deposits occur within relict sand wedges and are of Miocene and Pliocene age. \(^{40}\)Ar/\(^{39}\)Ar analyses of single volcanic crystals and glass shards removed from ashfall deposits within relict sand wedges in the western Asgard Range indicate cold climate conditions at 15.0 Ma (DMS-91-21), 13.6 Ma (DME-91-41), 13.5 Ma (DMS-89-143), 12.0 Ma (DMS-90-38 B), 10.5 Ma (DMS-89-132 B), and 10.0 (DMS-90-36 B). Numbers in parenthesis refer to ash sites discussed in Marchant et al. (1993 a, b). Likewise, dated ashfall deposits that rest on desert pavements indicate arid conditions in the western Asgard Range at 14.8/15.2 Ma (DMS-90-124 B) and in Arena Valley, Quartermain Mountains, at 4.34 Ma (DMS-86-86 B) (Marchant et al. 1993a, b).

The lack of numerous clay-sized grains within Dry Valleys ash deposits is consistent with persistent cold-desert conditions throughout Pliocene time. In-situ volcanic ash deposits in the Dry Valleys region contain less than 10% clay-sized grains, and volcanic crystals lack evidence of chemical weathering (Marchant et al. 1993a,b,c,d). This is because where suitable climate conditions cause glass to be unstable at the ground surface it quickly alters to clay. The rate at which surficial ash deposits weather to clay minerals depends on atmospheric temperature and the abundance of pore water (rates are increased at high atmospheric temperatures and high pore-water pressures; Lowe and Nelson, 1983; Lowe, 1986). For example, under humid temperate conditions in New Zealand, which are compatible with growth of *Nothofagus*, volcanic ash deposits older than about 50,000 years are >60% clay (Birrell and Pullar, 1973; Lowe and Nelson, 1983; Lowe, 1986). The absence of clay-sized grains in Miocene and Pliocene surficial ash deposits suggests that the warm climate conditions required for *Nothofagus* growth in the Transantarctic Mountains could not have existed in the western Dry Valleys region during Pliocene time. In fact, the lack of significant weathering within over 50 Miocene- and Pliocene-age ashes in the Dry Valleys region suggests that cold-desert conditions have persisted during the last 15.0 Ma (Marchant et al. 1993a).

**Quaternary Analogue**

A striking result of our reconstructions is the remarkable resemblance between Pliocene and Quaternary glacier behavior in the Dry Valleys. This is illustrated by the drift chronologies of Taylor Glacier and Wright Valley alpine glaciers. In all cases, Pliocene drifts occur just outside (or just above) the outermost Quaternary moraine. These results, then, do not show a striking dissimilarity that might be expected if the Pliocene climatic regime were markedly different from the Quaternary regime. Not only were the advances of similar magnitude, but the drift characteristics were similar.

The youngest Quaternary drifts (Taylor II, Taylor III, Alpine II) all fall within, or close to, the time span covered by the Vostok ice core in central East Antarctica, which suggests that the mean annual temperature was only about 2°C warmer in the penultimate interglaciation than it was during the Holocene (Lorius et al. 1985). This is consistent with the notion that the Dry Valleys remained a cold desert in late Quaternary time. The similarities in areal distribution and physical characteristics of Pliocene and late Quaternary drifts and moraines imply similar cold-desert paleoenvironments in the Pliocene. This is consistent with other terrestrial indicators (slope stability, ash-covered desert pavements, ash-filled contraction cracks, and lack of temperate ice landforms and sedimentary units). Given this situation, the Quaternary conditions may afford insights into Pliocene paleoclimate and glacier behavior. One recurring suggestion is that under the late Quaternary hyper-arid cold-desert conditions, the fluctuations of Dry Valleys polar glaciers were controlled primarily by changing accumulation rates (Denton et al. 1989). The same fundamental control may have applied to the Pliocene glacier system in the Dry Valleys if it also existed in a hyper-arid cold-desert environment.

**Overview**

From terrestrial-marine correlations, we conclude
that the geomorphic evidence argues strongly for cold and dry climatic conditions in the Dry Valleys adjacent to the marine water bodies with microfossils indicative of warmer-than-present temperatures. The persistence of cold-desert conditions suggests that paleo-water temperatures of the Pliocene Wright Valley fjord were at the low extreme of the ranges estimated by Webb (1972, 1974) and Prentice et al. (1993) from the Prospect Mesa gravels. There is no evidence from the terrestrial record at ~3.0 Ma for paleoclimate compatible with an adjacent inland marine water body with surface temperatures of 2°–5°C. Pliocene expansion of the Dry Valleys glacier system was of limited extent. Reconstructed ice-surface profiles show that the maximum Pliocene thickening of Taylor Glacier was only slightly larger than the modest Quaternary thickening. Finally, the terrestrial record inferred from volcanic ashfall deposits and the physical characteristics of drift sheets gives no hint of temperatures elevated enough to produce the melting ablation surfaces necessary for ice-sheet collapse (Huybrechts, 1993). On the contrary, Taylor Glacier, which drains Taylor Dome on the periphery of the East Antarctic Ice Sheet, reached its maximum Pliocene extent in the last 3.47 Ma at close to 3.0 Ma ago.

Conclusions

We applied a geomorphologic test in the Dry Valleys of whether the East Antarctic Ice Sheet remained robust or melted down during intervals of excess Pliocene warmth. The geomorphologic results do not meet the predictions of the meltdown hypothesis. Ice-sheet overriding of the Dry Valleys region occurred in the middle Miocene, not in the late Pliocene/early Pleistocene. Pliocene expansion of the Dry Valleys glacier system was modest and large tracts of the western Dry Valleys remained free of ice; such limited Pliocene expansion means that the East Antarctic ice could not have emplaced the mid-to-late Pliocene marine diatoms in the high-elevation Mt. Feather Sirius outcrop (one of the original sites used to propose meltdown; Webb et al. 1984). Pliocene surface uplift of the Dry Valleys block of the Transantarctic Mountains was minimal (about 300 m) rather than extensive (1000–3000 m). Hyper-arid cold-desert environmental conditions have persisted since at least the end of middle Miocene time. Any Pliocene warming was modest and did not approach the values deemed necessary for meltdown (Huybrechts, 1993). The implication of these geomorphologic data is that the East Antarctic Ice Sheet remained robust throughout Pliocene-Pleistocene time. But a distinct weakness of the geomorphologic approach is that a well documented alternative mechanism of emplacing marine Pliocene diatoms in high-elevation Sirius Group outcrops has yet to emerge.

The geomorphologic history of the Dry Valleys is interlocked with the tectonic evolution of the Transantarctic Mountains. Denudation and backwearing under a semi-arid or highly seasonal humid environment accompanied renewed rifting and rock uplift that began about 55 Ma ago. The resulting Transantarctic topography features planation surfaces with erosional remnants that together form an escarpment landscape. These surfaces rise in steps from the coast, with three levels separated by scarps. The intermediate surface is terminated by the downfaulted mountain front. The high-level planation surfaces are dissected by valley systems graded to near sea level. Much of this downcutting, which was accomplished by fluvial and glacial erosion, was accompanied by faulting and rock uplift of the mountain front. Following imposition of a hyper-arid, cold-desert climate by the end of middle-Miocene time, which effectively halted denudation, there occurred subsidence of the landscape, followed by about 300 m of Pliocene-Pleistocene surface uplift. One noteworthy conclusion is that all Sirius Group outcrops in the Dry Valleys now occur as remnants on the oldest elements of the landscape. The geomorphologic implication is that they represent early phases of glacial activity in the denudation of the Transantarctic Mountains. Another conclusion is that ice-sheet overriding of the Dry Valleys region might have been a consequence of mountain subsidence.

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MAP 1.

Fig. 18. Denton et al. p. 181.

Geomorphological map showing the distribution of the landscape types and the relationship to reconstructed river valleys. The main landscape elements start with a high original surface that supports large, isolated mountain remnants. Intermediate surfaces are cut into the high rock platform and separated from it by a prominent cuesta, which is the marker feature of the landscape. The isolated buttes, needles, and mesas of the intermediate surfaces form a spectacular inselberg landscape. The present valley system is incised in this landscape and forms a dendritic network with a sinuous plan profile. All major valleys have high-level tributaries, some graded to the intermediate surface and others to valley benches. The landscape is minor. Small cirque basins occur incised and, in places, valleys have been widened and straitly between episodes of stream incision and v occurs in central Taylor Valley, where truncated spout on the valley walls and sinuous stream channels occ
nches. The imprint of glacial erosion on this semi-arid occur incised sporadically along some rectilinear slopes d and straightened. Glacial modification occurred inter-

vission and valley downcutting. The best evidence for this incised spurs and straightened valley sections occur high tunnels occur at the valley bottom (see Figs 11 and 17).
MAP 2.
Fig. 27. Denton et al. p. 190.
Glacial geologic and surficial geomorphic map of Arena Valley. From Marchant et al. (1993b).

Fig. 7. Marchant et al. p. 276.
Glacial geologic and geomorphic map of Arena Valley, Quartermain Mountains.
Glacial geologic and surficial geomorphic map of east-central Wright Valley. From Hall et al. (1993).

Surficial geologic map of east-central Wright Valley showing the alpine glacier drifts, and Ross Sea drifts.