Mantle-plume activity recorded by low-relief erosion surfaces in West Antarctica and New Zealand

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ABSTRACT

Mantle-plume activity has been proposed to explain Neogene and mid-Cretaceous magmatic events, as well as associated tectonism, in West Antarctica; but the arrival time and dimensions of plume influence have been hard to define and are still a subject of debate. Two low-relief erosion surfaces, one in West Antarctica and the other in New Zealand (herein named the Waipounamu erosion surface), provide a way of assessing plume activity by measuring vertical displacements associated with these events. Both surfaces bevel mid-Cretaceous rocks, and both represent prolonged intervals of erosional leveling in a stable tectonic environment. Overlying strata in New Zealand indicate that leveling was near completion in coastal regions by ca. 75 Ma and therefore must have begun around 85 Ma, when New Zealand was beginning to break away from West Antarctica. Fluvial erosion followed by subsidence and marine planation are clearly recorded by these strata, and a similar history seems likely for West Antarctica, accounting for isostatically corrected ice-free bedrock elevations that are well below sea level over much of the region. The absence of uplift at the time of breakup seems incompatible with a plume mechanism for continental breakup. By contrast, the present elevation of the West Antarctic erosion surface records an estimated maximum of ≈3 km of tectonic uplift, associated with alkalic volcanism, beginning at ca. 28–30 Ma. We suggest that this event marks the inception of plume activity in West Antarctica. The resulting structure, the Marie Byrd Land dome, defines an area of plume influence that is smaller than the area defined by geochemistry, but is similar in scale to the Yellowstone plume.

INTRODUCTION

The Marie Byrd Land volcanic province includes 18 large, dominantly trachytic, shield volcanoes of mid-Miocene and younger ages, underlain by an alkalic basalt foundation that is largely ice covered, but apparently up to 5000 m thick where geophysical data are available. Basalt has been erupted throughout the ≈30 m.y. history of the province and has been estimated to make up 65% to 90% of its total volume (LeMasurier, 1990). The province covers a 1000 × 550 km coastal area within the West Antarctic rift system (Figs. 1 and 2), centered over a large structural dome that rises to 2700 m above sea level and 3200 m above isostatically corrected bedrock-surface elevations in adjacent areas. A close connection between dome growth and volcanic evolution is indicated by two long-term patterns of volcanic activity, described below.

The fundamental cause of volcanism has not been obvious. The province lies on an extension of the circum-Pacific "rim of fire," but there has been no subduction here for ≈100 m.y. The Antarctic plate has hardly moved since 85 Ma (Lawver et al., 1991); therefore, mechanisms involving plate dynamics are either ruled out or exceedingly subtle. Although the province lies within a large rift system, comparable in scale to the Basin and Range province, plate reconstructions will admit only minimal amounts of Cenozoic extension (Lawver and Gahagan, 1994). It does not seem likely that passive rifting would produce a large volcanic province in such a region. Nevertheless, the possibility of early Tertiary motion between East and West Antarctica is still being considered as a way of resolving discrepancies in the Cenozoic plate-motion history of the South Pacific (Cande et al., 1995). Until this possibility is eliminated, we cannot dismiss plate motion as a factor in the Neogene volcanism of West Antarctica.

A mantle-plume source has been proposed for this province because Marie Byrd Land basalts are indistinguishable from those on oceanic islands that are widely accepted as products of plume activity (LeMasurier and Rex, 1989, 1991; Hole and LeMasurier, 1994) and because plume activity is consistent with synvolcanic doming. Marie Byrd Land dome dimensions are similar to those of the Yellowstone plume (Parsons et al., 1994), but the familiar linear hotspot tracks are missing because of the long-term absence of plate motion.

The timing and extent of plume influence have become a subject of debate. Based on isotopic and large-ion lithophile element (LILE) characteristics, Hole and LeMasurier (1994) proposed that the effects of plume activity extended far beyond the Marie Byrd Land dome to include possibly the entire West Antarctic rift system. However, it has recently been found that these distinctive geochemical traits extend beyond the rift system, possibly to the New Zealand–Campbell Plateau and Australian continental blocks, and cannot be definitive of a single plume (Blusztajn and Hart, 1995). Buoyant uplift, related to the higher temperature of a plume compared to its surroundings in the mantle (White and McKenzie, 1989), may be the only definitive signature of plume activity in this region.

The Marie Byrd Land dome is defined by the elevations of block-faulted remnants of the West Antarctic erosion surface: a regionally extensive, near-planar surface that truncates mid-Cretaceous and older basement rocks. Dome uplift appears to have begun at the same time as basalt volcanism, at ca. 28–30 Ma; therefore, it has been proposed that this time interval marks the beginning of plume activity (Hole and LeMasurier, 1994). However, Weaver et al. (1994) suggested that plume activity began with mid-Cretaceous magmatism and may have controlled the locus of breakup of New Zealand from West Antarctica. Although nobody has disputed the idea of plume activity in this region, it is clear that further constraints on timing and areal dimensions are needed. Because uplift is usually the first indication of plume activity (Kent et al., 1992), and uplift of the dome is

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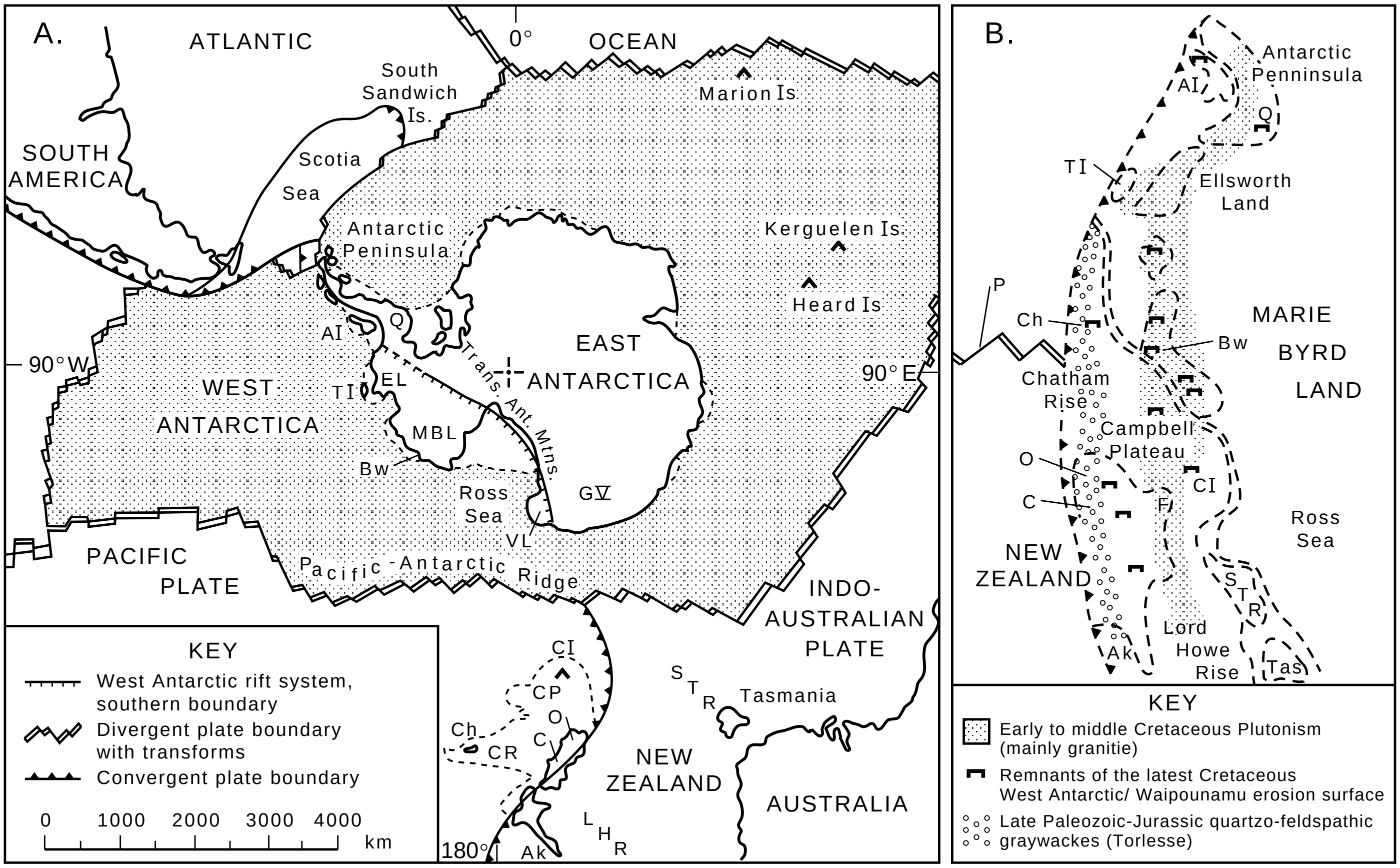


Figure 1. Index maps of Antarctica and the New Zealand–Campbell Plateau block. A: Present-day relationship of the Antarctic plate (dot pattern) to the New Zealand–Campbell Plateau block. B: A plate-tectonic and geologic reconstruction showing spatial and geologic relationships between West Antarctica and the New Zealand–Campbell Plateau block in pre–85 Ma time. Modified from Cooper et al. (1982). Positions of the convergent plate margin and Pacific-Phoenix Ridge at ca. 105 Ma are adapted to this reconstruction from Lawver and Gahagan (1994). The continuity of the Cretaceous plutonic belt and the Torlesse belt shows the overall geologic integrity of the reconstruction. Major localities of erosion-surface remnants are shown to illustrate the basis for considering a prebreakup origin for the West Antarctic erosion surface. Abbreviations for parts A and B are as follows: AI—Alexander Island, Ak—Auckland, Bw—Bowyer Butte, CI—Campbell Island, CP—Campbell Plateau, C—Canterbury, Ch—Chatham Islands, CR—Chatham Rise, EL—Ellsworth Land, F—Fiordland, GV—George V Land, LHR—Lord Howe Rise, MBL—Marie Byrd Land, O—Otago, P—Pacific-Phoenix Ridge, Q—Quilty Nunatak, STR—South Tasman Rise, Tas—Tasmania, TI—Thurston Island, VL—Victoria Land.

defined by the West Antarctic erosion surface, we have studied the surface in detail. Our approach has involved a comparative study of the "Otago peneplain," a Late Cretaceous surface that occurs throughout much of the New Zealand–Campbell Plateau continental block. Plate reconstructions (Fig. 1) suggest either that the two were once a single surface prior to the 85 Ma separation of New Zealand from Marie Byrd Land (Lawver et al., 1991) or that they formed on opposite flanks of a "neo-ocean" soon after breakup. In either case, data from both regions are needed to determine whether these surfaces formed near sea level and whether plume activity was associated with breakup.

THE WEST ANTARCTIC EROSION SURFACE

Morphology and Extent

A low-relief erosion surface, referred to here as the West Antarctic erosion surface, is preserved along a 2500 km coastal belt in West Antarctica (Fig. 2). Remnants of the surface consist of flat mountain summits that bevel granites, volcanic rocks, and metasedimentary rocks of Cretaceous and older ages (Fig. 3) and low-relief unconformities that separate these rocks from overlying late Cenozoic volcanic rocks. Erosional relief on the surface appears to be well under 100 m and is probably less than 50 m in most Marie Byrd Land localities (LeMasurier and Rex, 1983). It is represented by an unconformity in the Jones Mountains (Fig. 2), where it has been surveyed along its entire >30 km of nearly continuous exposure. Maximum measured relief here is 50 m, but relief is generally less than 15 m (Craddock et al., 1964).

No evidence has been found for multiple erosion surfaces. On well-exposed nunataks with 500–700 m of relief there is one summit surface, or near-summit unconformity, and no benches on mountain flanks. However, lateral discontinuity of the surface and abrupt elevation changes from one nunatak to another have been produced by block faulting (Fig. 3).

Erosional dissection of the surface is not extensive. The best example we

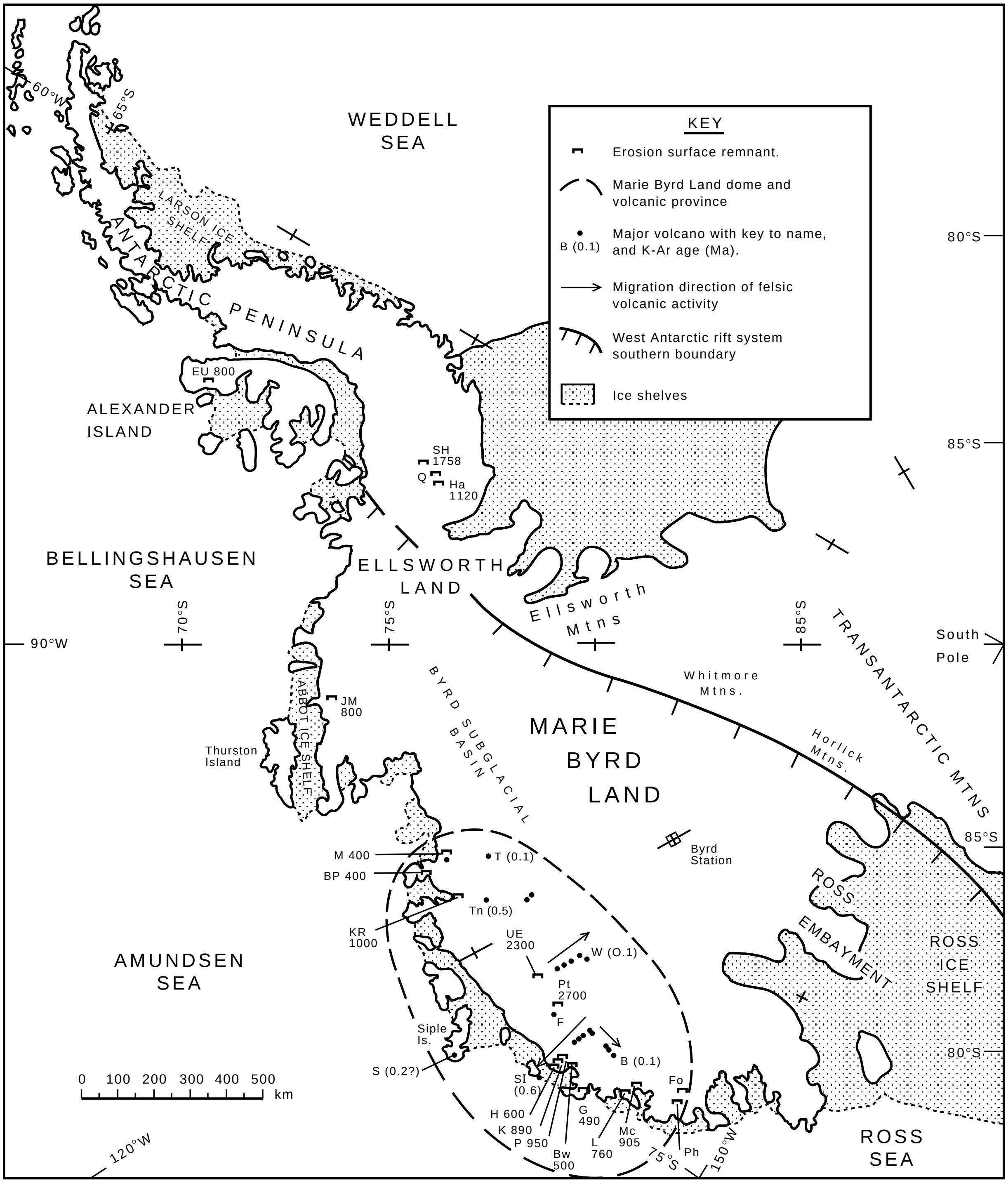


Figure 2. Location map of West Antarctica showing distribution and elevations (unbracketed numbers) of erosion-surface remnants and major volcanoes of the Marie Byrd Land volcanic province. The Marie Byrd Land dome is centered over Mount Petras (Pt). Felsic volcanic activity has migrated away from the center of dome uplift during the past 18 m.y., along rectilinear paths, as shown by arrows. Abbreviations are as follows: B—Mount Berlin, BP—Bear Peninsula, Bw—Bowyer Butte, EU—Elgar Uplands, F—Mount Flint, Fo—Fosdick Mountains, G—Mount Gray, H—Holmes Bluff, Ha—Mount Hassage, JM—Jones Mountains, K—Kouperov Peak, KR—Kohler Range, L—Mount Langway, M—Mount Murphy, Mc—Mount McCoy, P—Patton Bluff, Ph—Phillips Mountains, Pt—Mount Petras, Q—Quilty Nunatak, S—Mount Siple, SH—Sky-Hi Nunataks, SI—Shepard Island, T—Mount Takahe, Tn—Toney Mountain, UE—USAS Escarpment, W—Mount Waesche. References: Craddock et al. (1964)—JM; Laudon (1972)—Ha, SH, Q; Luyendyk et al. (1991)—Fo, Ph; P. A. R. Nell (1990, personal commun.)—EU; Spörli and Craddock (1980)—G, L, LB.

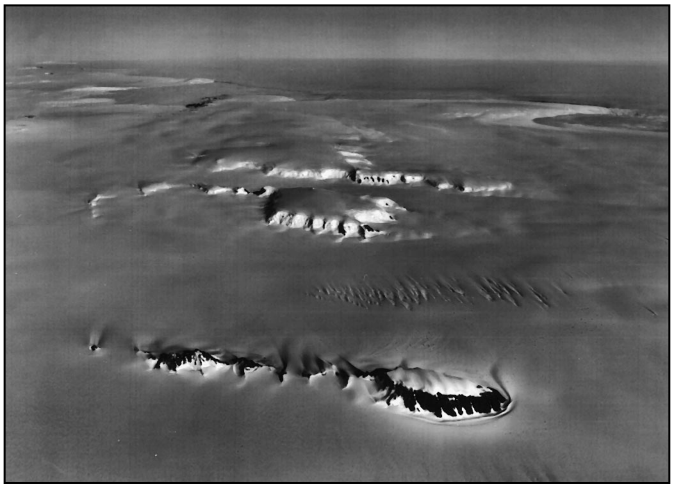


Figure 3. Oblique aerial photograph of the Hobbs Coast of Marie Byrd Land, showing the large areal extent and undissected nature of erosion-surface remnants in this region. View is to the west. The Mount Prince remnant (6 km^2) is in the foreground; Bowyer Butte remnant (≈6 km width; 30 km^2 area) is in the middle ground, 15 km west of Mount Prince. Horst and graben structure is evident from the rectilinear outlines of nunataks, from exposure of the erosion surface at different elevations in adjacent nunataks (see also Figs. 2 and 13), and also from the scale, depth, and rectilinear orientations of subglacial basins (LeMasurier and Rex, 1982; LeMasurier, 1990). Lack of tilting of the surface suggests either that minimal displacement occurred on the faults or that the faults do not flatten with depth.

have found of a paleovalley is at the north end of Holmes Bluff (Fig. 2), where a northeast-trending valley ≈100 m deep cuts through 8.17 Ma basalt flows that rest on the summit surface; this valley is partly filled by a 6.27 Ma basalt flow (LeMasurier and Rex, 1983). This relationship thus records late Miocene valley cutting. In sub-ice topography it is impossible to distinguish clearly between tectonic and glacial erosional effects, but the distinction is important in separating isostatic rebound from tectonic uplift. The lack of sensible glacial-flow patterns in sub-ice topography, the broad areas of undissected erosion-surface exposure, the scarcity of paleovalleys, and lack of dissection of most Miocene and Pliocene shield volcanoes all suggest minimal dissection.

The full extent of the original erosion surface is poorly known. It is well represented by multiple exposures from western Marie Byrd Land to eastern Ellsworth Land. In northern Alexander Island it may be represented by a Late Cretaceous–Paleogene unconformity with 100–200 m relief (P. A. R. Nell, 1990, personal commun.). A striking summit erosion surface occurs along the northern Antarctic Peninsula, but age control is not adequate to support speculation about the age and nature of the surface (D. H. Elliot, 1994, oral commun.)—hence its omission from Figure 2. The query in Table 1 reflects our uncertainty about the relationship, if any, between these Antarctic Peninsula localities (including Alexander Island) and the erosion surface in Marie Byrd Land and Ellsworth Land. In the Transantarctic Mountains, the only low-relief erosion surfaces are the Devonian(?) Kukri peneplain (Grindley and Warren, 1964) and a "strongly dissected peneplain" of unspecified age in northern Victoria Land (Van der Wateren and Verbers, 1992).

Age and Stratigraphic Relationships

The rocks cut by the West Antarctic erosion surface are a complex of Paleozoic and Mesozoic granitic, volcanic, and metaclastic rocks that is dominated by granitoids. Rb-Sr and K-Ar ages of the granitoids range from 90 to 371 Ma, but 90–110 Ma ages are the most common (Adams, 1987; Pankhurst and Rowley, 1991; Weaver et al., 1991, 1994; Mukasa, 1995; and references therein). These are also the youngest rocks beveled by the erosion surface.

The oldest rocks resting on the surface in Marie Byrd Land are hyaloclastite breccias from Mount Petras with K-Ar ages of 22.2 ± 1.6 to 25.3 ± 1.0 Ma (LeMasurier and Rex, 1982, 1983). At Alexander Island (Fig. 2), the overlying volcanic rocks include fresh rhyolite with a Rb-Sr whole-rock isochron age of 62 ± 1 Ma (R. J. Pankhurst, 1990, oral commun.). These data imply a mainly Late Cretaceous to Paleogene age of formation for the unconformity. No sedimentary rocks have been found anywhere on the erosion surface.

THE WAIPOUNAMU EROSION SURFACE OF NEW ZEALAND

Geographic Extent

In New Zealand, a low-relief erosion surface is well preserved over an area in excess of 100 000 km^2. It is conspicuous in upland areas of southern South Island (Figs. 4 and 5), where it occurs as a flat summit surface that truncates upper Paleozoic–Mesozoic schists and graywackes. It is veneered in many areas by fluvial gravel and in others by thin sequences of shallow-marine deposits (Table 1, Fig. 6). In addition, the surface also bevels basement rock at scattered localities on the northern part of the South Island (Grindley, 1980), on the North Island (Cotton, 1938; Korsch and Wellman, 1988), on the Chatham Islands (Campbell et al., 1993), and on Campbell Island (Beggs, 1978). Offshore seismic profiles and scattered drill holes indicate the presence of a low-relief, Late Cretaceous erosion surface (a planar unconformity) beneath the Chatham Rise (Wood et al., 1989), Challenger Plateau (Nathan et al., 1986), Campbell Plateau (Houtz et al., 1967), Otago continental shelf (Carter, 1988), and Canterbury Basin (Field and Browne, 1989) (Fig. 4). These data indicate that the original erosion surface, including currently offshore localities, may have extended over an area as great as 10^6 km^2.

In New Zealand, the surface has been known as the "Cretaceous peneplain" (Benson, 1935) and as the "Otago peneplain" (Cotton, 1949); in each case, an origin by subaerial fluvial erosion has been inferred. In what follows we describe the morphology and stratigraphic relationships of this surface and its sedimentary veneer. Our data show that the surface is a composite of features produced by fluvial erosion, followed by marine transgression and wave planation that took place after the 85 Ma breakup. The term "peneplain" is therefore inappropriate. We propose the name "Waipounamu erosion surface," using the Maori name for the South Island, for this composite erosion surface.

Morphology

The Waipounamu erosion surface is nearly planar to gently rolling and locally incised by paleovalleys. The morphology and stratigraphic relationships of the surface are exceptionally well exposed in the Silver Peaks area, near the east Otago coast (Figs. 5, 6, and 7), which we will use as an exemplary locality. Here, a remarkably flat surface can be traced along ridge crests for 25 km. Parts of this surface show less than 5 m/km of relief, but paleochannels cut into the surface are as much as 40 m deep.

TABLE 1. TIME-SPACE DIAGRAM SHOWING ROCKS, AND INFERRED EVENTS, RELEVANT TO INTERPRETING THE "PENEPLAINS" OF NEW ZEALAND AND WEST ANTARCTICA

Age (Ma)	Alexandra–Central Otago	Naseby-Kyeburn	Aviemore	Horse Range–Oamaru	Dunedin–Silver Peaks	Chatham Islands	Campbell Island	Great South Basin	Ross Sea	Coastal Marie Byrd Land	Antarctic Peninsula	Age (Ma)
10	← • • • • • • • • • • • • folding, reverse faulting, erosional stripping of "peneplain" surface (oblique convergence along plate boundary) • • • • • • • • • • • • →					2–6 Ma basalts	6.6–11.0 Ma basalts			0–30 Ma alkalic basalt and trachyte lava flows, pillow lavas, and hyaloclastite, interbedded with tillite		10
15	ca. 15–18 Ma lacustrine, fluvial, and deltaic clastic sediments and coal (Manuherikia Group)			ca. 10–18 Ma basalt flows rest with angular discordance on 18–20 Ma marine ss. ← • • 13–18 Ma uplift and erosion • • →								15
20	~~~~~ 18 Ma doming and erosion ~~~~~		← • • • • • • • • • • marine regression • • • • • • • • • • →									20
	← • • • ? • MAXIMUM MARINE TRANSGRESSION (23–24 Ma) • ? • • • →											
25		← • • • • • • • 20–25 Ma shallow-marine limestone and greensand • • • • • • • →										25
•				30–40 Ma basalts		ca. 35 Ma basalts						•
•			← • offshore marine sedimentation on broad continental shelf • →							• • • ? • • ? • • • →		•
45			45–50 Ma shallow-marine clastic rocks									45
50											?	50
55				55–60 Ma basaltic tuffs offshore				Late Cretaceous to Oligocene strata		100–300-km-wide continental shelf, ^a1000 m depth at present shelf edge		55
60		Waipounamu erosion surface				ca. 60 Ma basalts					^a1.6 km sediment thickness offshore	60
65				65–75 Ma shallow marine clastic rocks overlie quartzose fluvial conglomerates that occur in paleovalleys. In interstream areas, shallow-marine clastic rocks overlie wave-planed schist basement and graben deposits.						U6 composite unconformity / West Antarctic erosion surface		65
70						70–80 Ma basalts (nonmarine to estuarine)						70
75												75
80						~~~~~~~~~~~~~~~~~~~ unconformity ~~~~~~~~~~~~~~~~~~~						80
85	~~~~~~~~~~~~~~~~ BREAK-UP OF THE MARIE BYRD LAND–NEW ZEALAND–CAMPBELL PLATEAU BLOCK ~~~~~~~~~~~~~~~~							85–95 Ma; 2 km thickness of nonmarine sediment in normal-fault–bounded basins	8 km thickness of strata of inferred Cretaceous age, in fault-bounded basins			85
90				ca. 85–100 Ma nonmarine coarse clastic rocks in normal-fault–bounded basins; possible Marie Byrd Land source for conglomerate clasts in Chatham Islands[†]						ca. 90–110 Ma granitic plutons (correlative with granitic plutonism in western New Zealand)	ca. 100 Ma turbidites	90
•		ca. 100 Ma nonmarine coarse clastic rocks in normal-fault–bounded basins, interbedded with silicic volcanic rocks*										•
•												•
•												•
100				silicic volcanic rocks (ca. 100 Ma)		ca. 100 Ma basalts (including marine pyroclastic rocks)						100
•										Devonian and older metaclastic basement rocks		•
•												•
•				← • • • metamorphism of Haast Schist basement • • • • →								•
280							early–mid Paleozoic metaclastic rocks					280

Note: Arranged (from left to right) as follows: inland New Zealand, coastal New Zealand, continental shelf, and West Antarctica.
*Bishop (1974); Adams and Raine (1988).
[†]Grindley et al. (1977); Nathan et al. (1986); Mitchell (1991).

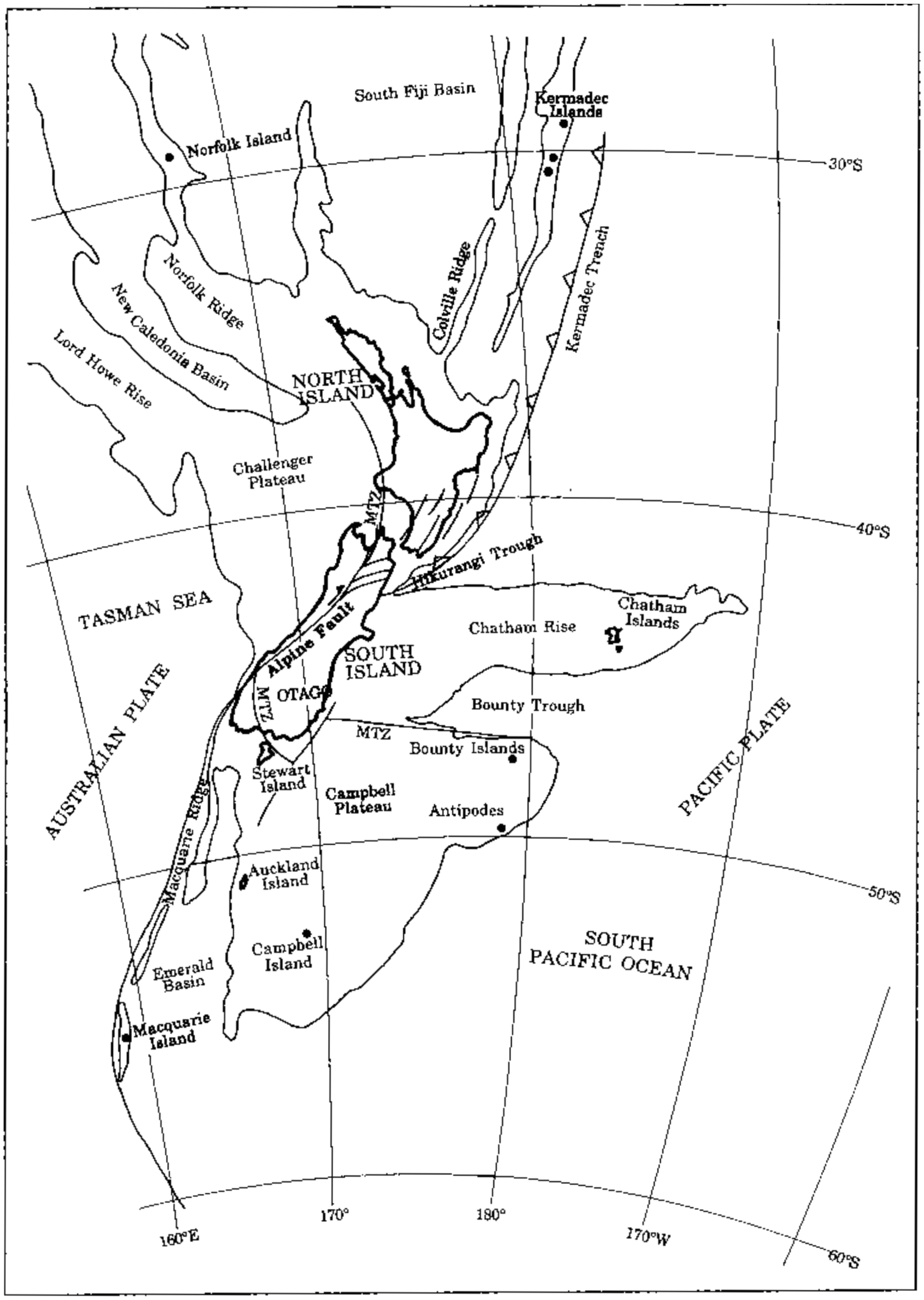


Figure 4. Location map for the New Zealand–Campbell Plateau continental crustal block, outlined by the –2000 m isobath. MTZ is the median tectonic zone, a major terrane boundary.

Farther inland in Central Otago (Fig. 7), the surface is conspicuous in upland areas (>1500 m) where it has been exhumed from beneath sedimentary cover. Miocene paleovalleys are present locally, and relief in these may be as great as 500 m (Douglas, 1986; Stirling, 1990). Remnants of the Waipounamu surface can be recognized as far inland as the Pisa Range (Fig. 7), 90 km southeast of the active plate boundary (Alpine fault). Farther inland yet, uplift and erosion rates are too great to permit preservation of ancient upland surfaces (Adams, 1980).

Although relief on the Waipounamu surface is clearly low (Park, 1906; Cotton, 1938), Pleistocene erosion and Neogene deformation have made its original morphology difficult to quantify (Beanland and Berryman, 1989; Stirling, 1990). Broad open folds deform the surface over much of Otago (Table 1 and Fig. 8), and these must be distinguished from original erosional relief on the surface. These folds are growing structures, many with wavelengths of 20–30 km (Norris et al., 1990). Amplitudes range from 500 m in folds near the coast to 1500 m in the schist-cored anticlines of the Central Otago ranges. They appear to have grown in amplitude as they approach the plate boundary (Fig. 8, inset), eventually to be destroyed by erosion in the zone of rapid uplift within 50 km of the Alpine fault. Quaternary convergence rates have been in excess of 10 mm/yr, and at least 70 km of the Pacific plate has been lost to rapid uplift and erosion along the plate boundary, including a substantial tract of the original Waipounamu

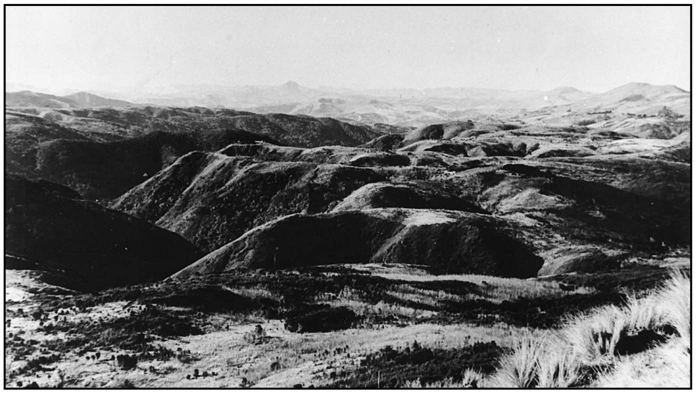


Figure 5. Photos showing the uplifted and exhumed Waipounamu erosion surface, Silver Peaks area, Otago. Stratigraphic columns in Figure 6 were recorded from this region. (A) View looking west across Blueskin Bay estuary, 10 km northeast of Dunedin, showing the exhumed erosion surface along crests of conspicuous flat-topped ridges in the middle distance. The crest of the skyline ridge also approximates the Waipounamu, but lies on the uplifted and dissected western side of the Neogene Waikouaiti fault. Photograph by D. Weston. B: View looking northwest across the Mountain Road–Walker Road area (Fig. 6). Flat-topped ridges are underlain by mid-Mesozoic schist and capped by a discontinuous veneer of shallow-marine sedimentary rocks. Photograph by D. S. Coombs.

erosion surface. The Neogene tectonic history recorded by the Waipounamu surface is the subject of another paper (Landis and LeMasurier, unpub. data).

Age and Stratigraphic Relationships

Pre-Waipounamu Rocks. In most of Otago and South Canterbury, the Waipounamu surface is developed on upper Paleozoic–Mesozoic schists and graywackes. Elsewhere in the South Island, it bevels rocks ranging from Cambrian sedimentary and volcanic rocks to Cretaceous granitoids and Cretaceous graben-fill conglomerates (Table 1). The fault-bounded nature of the graben-fill deposits, their coarse grain size and commonly angular clasts, the presence of large-scale cross-bedding, and absence of extensive upward-fining sequences or slack-water deposits all suggest deposition on alluvial fans and in high-gradient braided-river systems (Bishop and Laird, 1976; Mitchell, 1991). Structural relief on the base of these pre-Waipounamu sequences is at least 1 km, and there is no evidence to indicate that a regionally extensive low-relief erosion surface existed prior to their deposition. We envisage a rugged landscape with at least 1500 m of tectonic relief along some fault scarps in mid-Cretaceous (prebreakup) time.

The Chatham Islands (Table 1, Fig. 4) contain exposures of the pre-Waipounamu rocks closest to West Antarctica, and they record a geologic history similar to that on the South Island. The youngest are mid-Cretaceous graben-fill clastic rocks (Waihere Bay Group; Grindley et al., 1977) that rest unconformably on a Jurassic schist basement (Adams and Robinson, 1977; Campbell et al., 1993). This basin fill contains igneous and metamorphic pebbles of exotic provenance that may have been derived from Marie Byrd Land prior to breakup (Grindley et al., 1977; Dean, 1993). Offshore seismic exploration on Chatham Rise and in the Great South Basin shows a Late Cretaceous regional unconformity beveling both Jurassic metamorphic basement and Cretaceous graben fills, similar to the relationships on land (Wood et al., 1989; Cook and Beggs, 1990).

These rocks clearly predate breakup; the oldest oceanic crust adjoining the Campbell Plateau (Fig. 4) is older than anomaly 34 (82 Ma) (Lawver et al., 1991). Prebreakup rifting can be confidently projected back to ca. 100 Ma (Bishop and Laird, 1976; Carter, 1988). During this early-breakup phase, we envisage high-gradient antecedent river systems spanning the eventual site of separation.

Post-Waipounamu Rocks. In the Dunedin–Silver Peaks area (Fig. 6, Table 1), planar parts of the Waipounamu surface are overlain by glauconitic and fossiliferous shallow-marine sedimentary rocks (e.g., Mountain Road), but adjoining paleovalleys are filled with fluvial deposits overlain by the shallow-marine strata (e.g., Jones Road). This relationship indicates that the paleovalleys were part of the Late Cretaceous landscape and suggests aggradation of river valleys followed by marine transgression. At Brighton (Fig. 7), belemnite- and bivalve-rich sandstone and conglomerate rest on the Waipounamu, whereas only 500 m to the east, the surface is overlain by nonmarine coal measures (McKellar, 1990). Apparently, swampy conditions existed in aggrading river valleys prior to marine transgression. Similar relationships have been described in the eastern Horse Range (Fig. 7, Table 1; Mitchell, 1991; Paterson, 1941).

On the Chatham Islands, a low-relief erosion surface developed on schist basement is overlain by Paleocene shallow-marine rocks (Adams and Robinson, 1977; Campbell et al., 1993), and a similar erosion surface on Campbell Island is overlain by Upper Cretaceous fluvial deposits, overlain in turn by a marine sequence (Beggs, 1978). In these widely separated offshore islands, the Waipounamu surface is planar, as it is in the Silver Peaks–Brighton region.

Diachronous Nature of the Surface. On the South Island, basal deposits on the Waipounamu are Late Cretaceous near the coast (McKellar, 1990) and become progressively younger inland (Bishop, 1974). Paleocene, Eocene, and Oligocene sedimentary rocks (Table 1) rest on the surface between Central Otago and the coast (Suggate et al., 1978). These strata indicate that the Waipounamu is time transgressive, as shown in Table 1. In Central Otago, erosional truncation of lower Tertiary rocks followed by Miocene fluvial and lacustrine deposition record the end of a 50 m.y. interval of marine transgression and wave planation.

DISCUSSION

Reconstruction of the Original Surface

We interpret the beveled nunatak summits and low-relief unconformities of Late Cretaceous–early Tertiary age that occur throughout coastal West

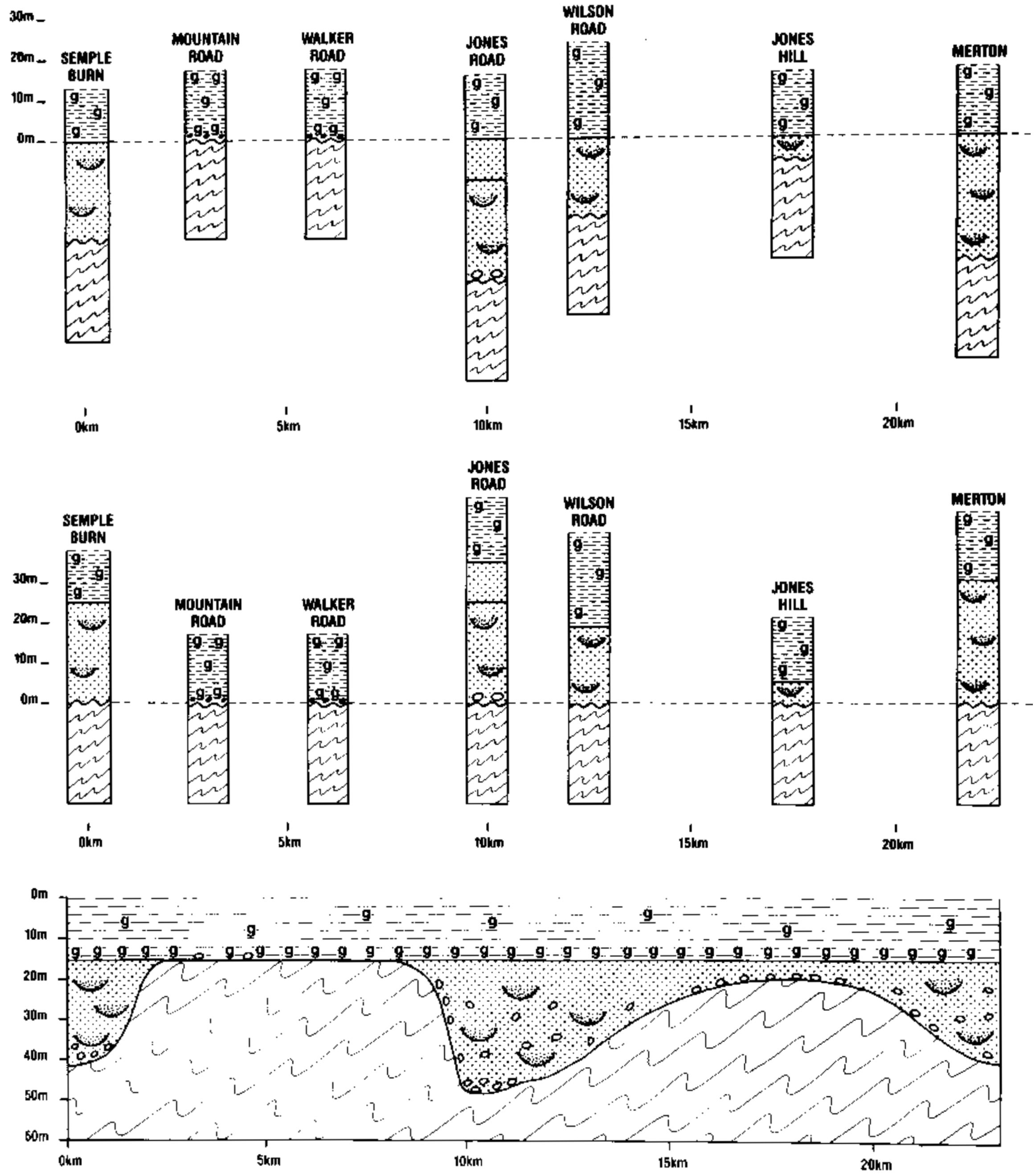


Figure 6. Stratigraphic columns in the Silver Peaks area of east Otago. The Waipounamu surface is developed here on mid-Mesozoic schist and unconformably overlain by thin remnants of a once widespread Cretaceous to Miocene, transgressive sedimentary prism (Table 1). Top row uses the base of the marine section as a datum and is the basis for the schematic cross section. Middle row uses the eroded basement surface as a datum. All three rows show discontinuous nonmarine quartzose sandstones and conglomerates overlain by glauconitic (g) sandstones and siltstones. The Waipounamu erosion surface is the planar unconformity in the bottom panel.

Antarctica to be remnants of a single, originally continuous, low-relief erosion surface. This interpretation is based on the low relief and consistent age range of each individual remnant and on the lack of evidence for more than one surface. However, the data do not support the idea that the West Antarctic and Waipounamu surfaces were a continuous low-relief landscape in prebreakup time. The youngest rocks cut by these surfaces are 85–100 Ma clastic rocks in New Zealand and 90–110 Ma granitoids in Marie Byrd Land (Table 1). The oldest rocks resting on either surface are shallow-marine and fluvial deposits in New Zealand and on the adjoining sea floor. The oldest reasonable paleontologic age for these rocks is late Campanian–early Maestrichtian (ca. 75 Ma), indicating that planation was completed about that time in near-coastal localities (Figs. 7 and 9). We have already noted evidence from New Zealand for high relief and active tectonism in mid-Cretaceous (prebreakup) time, and Weaver et al. (1994)

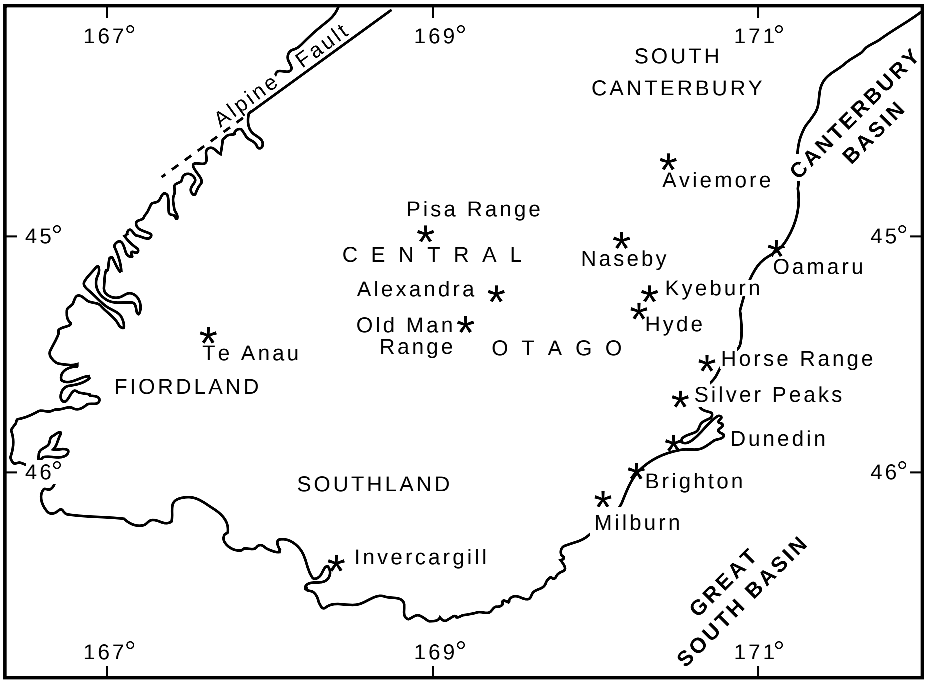


Figure 7. Index map showing localities in Otago and South Canterbury referred to in the text and in Table 1.

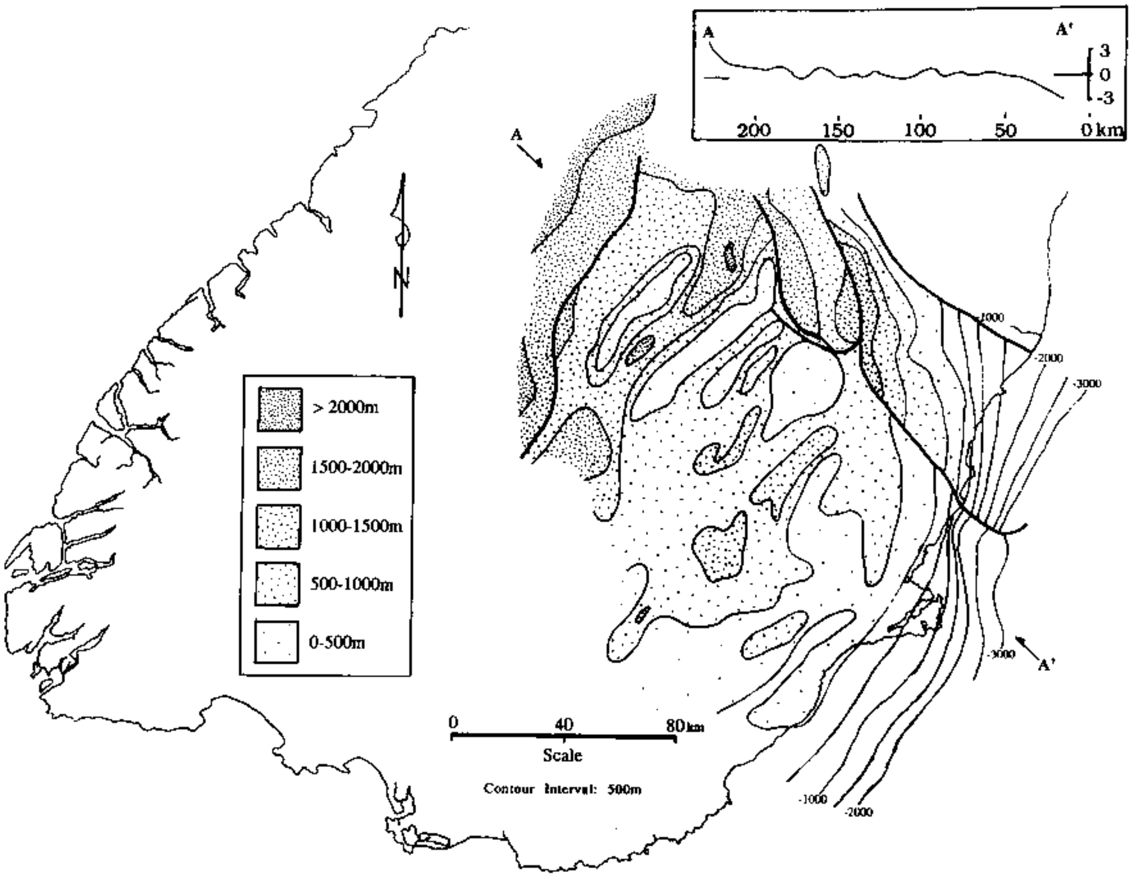


Figure 8. Topographic contours on the Waipounamu erosion surface in southern South Island, and cross section, modified after Benson (1935). The originally flat surface now shows broad open folding related to late Cenozoic convergence along the Alpine fault plate boundary. Many mapped folds have faulted limbs; only major faults are shown here.

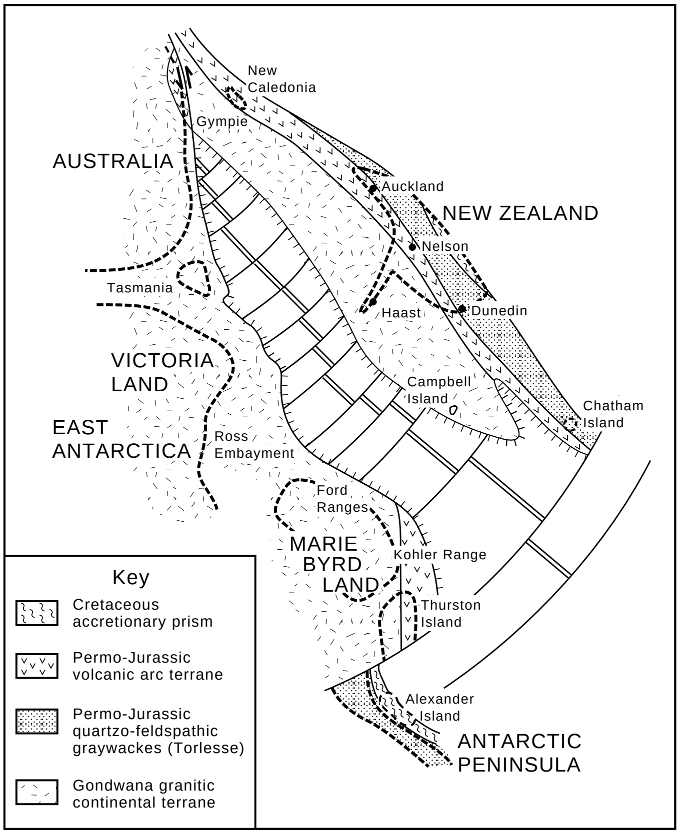


Figure 9. Sea-floor spreading reconstruction of West Antarctica and the New Zealand–Campbell Plateau block around anomaly 32 time (72 Ma), showing proposed relationship of the two blocks when leveling of the Waipounamu and West Antarctic erosion surfaces was near completion on the margins of the two blocks shown. Modified from Grindley and Davey (1982).

described evidence from Marie Byrd Land for regional uplift at ca. 100 Ma. The stable interval during which leveling took place must postdate these events.

Judson and Ritter (1964) estimated that it would take 11–12 m.y. to erode the United States to sea level, at present rates of erosion, with no isostatic rebound. Schumm (1963) included decelerating erosion rates and isostatic response in his estimate of 15–110 m.y. to level the United States. The size of these estimates suggests that the stable environment required to produce the Waipounamu and West Antarctic erosion surfaces must have begun at breakup time, or before, when Marie Byrd Land and New Zealand were still joined. The comparison is reasonable when one considers that the present distribution and scale of these two surfaces is comparable to finding them on the summit of the Appalachians, from Newfoundland to Georgia, and along the summits of the Scandinavian and British Caledonides.

Environment of Planation

The origin of low-relief erosion surfaces is a controversial topic. It has evolved from the peneplain concept (Davis, 1889), involving fluvial erosion graded to sea level, to pediplanation, which involved fluvial erosion graded to interior basins, to marine planation, and back to Neo-Davisian geomorphology. Reviews of the controversies and their current status can be found in Bloom (1991) and Summerfield (1991). For the purposes of our study, it is important to know (1) whether the West Antarctic and the Waipounamu surfaces formed near sea level, (2) when, if ever, they subsided below sea level, and (3) when they were uplifted to their present elevations. Answers to the first two questions are well preserved in New Zealand.

New Zealand. After breakup, tectonic activity was insignificant in most of New Zealand apart from thermal subsidence, minor warping, and basin-margin faulting (Beggs, 1993; Carter, 1988). Available evidence indicates that the climate was warm and moist, weathering was intense (Sherwood et al., 1992), and fluvial erosion reduced mid-Cretaceous mountains to low relief. Late Cretaceous (65–75 Ma) subsidence of the New Zealand landmass, accompanied by marine transgression and aggradation of coastal rivers, is clearly recorded in the sedimentary record (e.g., Fig. 6). These deposits suggest that final leveling of the Waipounamu surface took place during time-transgressive wave planation over large areas of the South Island. There is ample room for debate about how much leveling was fluvial and how much marine; but it is clear that the Waipounamu surface formed at, or very near, sea level. Subsidence continued until late Paleogene time (ca. 25 Ma) and, in spite of fluctuating sea levels, extended the wave-planed surface across the South Island, burying basement rock and fluvial valley fill with a veneer of littoral sand and gravel.

West Antarctica. The low relief and regional extent of the West Antarctic erosion surface are compatible with prolonged fluvial erosion, marine planation, or a combination of the two. However, the possibility that leveling took place by fluvial erosion graded to upland interior basins cut off from the sea does not seem tenable. If leveling began in early-breakup time and continued after separation, the early histories of the West Antarctic and Waipounamu surfaces are the same, or very closely related, and the Waipounamu surface clearly formed near sea level.

On the Antarctic side, there are no sedimentary units to indicate whether subsidence accompanied final leveling of the erosion surface, but its likelihood is evident from comparing postbreakup history in the Ross Sea and its surroundings with that in the Great South Basin of New Zealand. Detailed characteristics of both basins conform closely to a model developed by Dewey (1982) for basin evolution on rifted continental margins. The model begins with rapid subsidence along normal faults in relatively narrow basins; the rapid subsidence is followed by a period of exponential decrease in the rate of subsidence during which sedimentary load is dispersed over an increasingly wide area.

In the Late Cretaceous Great South Basin (Fig. 7), initial sedimentary units are of Cenomanian–Campanian age (85–95 Ma) and comprise thick (2 km) sequences of nonmarine graben fill (Beggs, 1993). Similar, but smaller, rift basins occur widely in the South Island–Chatham Rise region (Wood et al., 1989). With cessation of rifting and onset of flexural warping, coastal subsidence and fluvial aggradation occurred along basin margins. Marine transgression and associated wave planation beveled residual basement highs, reworked the tops of fluvial channel deposits, and in some areas removed the underlying fluvial deposits. The unconformity at the base of this transgressive sequence is the Waipounamu erosion surface.

The Ross Sea (Fig. 1) is underlain by a block-faulted basement of thinned Precambrian–Mesozoic continental crust, containing initial graben fill of inferred Cretaceous age up to 8 km thick (Cooper et al., 1991). These sedimentary rocks are overlain by prograding wedges of Upper Cretaceous–Oligocene strata, which truncate both early rift fill and adjoining basement rock before tapering out onto basement highs (the postbreakup unconformity). Thus, the Ross Sea basins and the Great South Basin both show typical "steer's head" cross sections (Dewey, 1982).

These rift-related sequences and emergent basement highs are mantled unconformably (U6 unconformity of Hinz and Block, 1983) by regionally extensive Oligocene to Recent glacial-marine strata 5–6 km thick (Cooper et al., 1991). The postbreakup and U6 unconformities merge to bevel basement highs as they approach the Marie Byrd Land coast of the Ross Sea, from which they are separated by post-Eocene block faulting. We provisionally interpret the West Antarctic erosion surface to be the uplifted equivalent of this Ross Sea area composite unconformity.

The postbreakup subsidence recorded in the Ross Sea may have been characteristic of the entire rift system. Bedrock throughout most of the rift lies 500–1000 m below sea level, after isostatic adjustment to preglacial levels (Drewry, 1983). Figure 10 shows the Marie Byrd Land dome rising up to >3200 m above this inland regional bedrock surface. Seismic basement has been found at –3000 m (below sea level) under Byrd Station and –4500 m beneath the Byrd subglacial basin, overlain respectively by about 1500 and 2500 m of sedimentary/volcanic rock (Bentley and Clough, 1972), suggesting a history of extension and subsidence like that in the Ross Sea, but data are sparse.

Outboard of the Marie Byrd Land dome, water depths along the coast between the Ross Sea and Antarctic Peninsula are between 500 and 1000 m, similar to the inland elevations. Kimura (1982) showed ≈1.6 km of shelf deposits on the west side of the Antarctic Peninsula; there are no data along the coast of Marie Byrd Land and Ellsworth Land. Combining 1 km of water depth with an estimate of 1–2 km of postbreakup strata yields the 2.3 km of subsidence predicted for nonvolcanic margins at the time of breakup (White and McKenzie, 1989). Much greater subsidence took place within individual basins, and indeed the extraordinary depths of some suggest that the ice itself is basin fill (LeMasurier and Rex, 1983). Most of the extension is believed to have taken place during breakup, at ca. 85 Ma (Lawver et al., 1994). Thus, the Marie Byrd Land dome is surrounded by a region of thin crust (Behrendt et al., 1991) that subsided 500 m or more below sea level, probably while the West Antarctic erosion surface was being formed. It therefore seems likely that the present dome area was also involved in this postbreakup subsidence and was raised from about the –500 m level during Neogene doming.

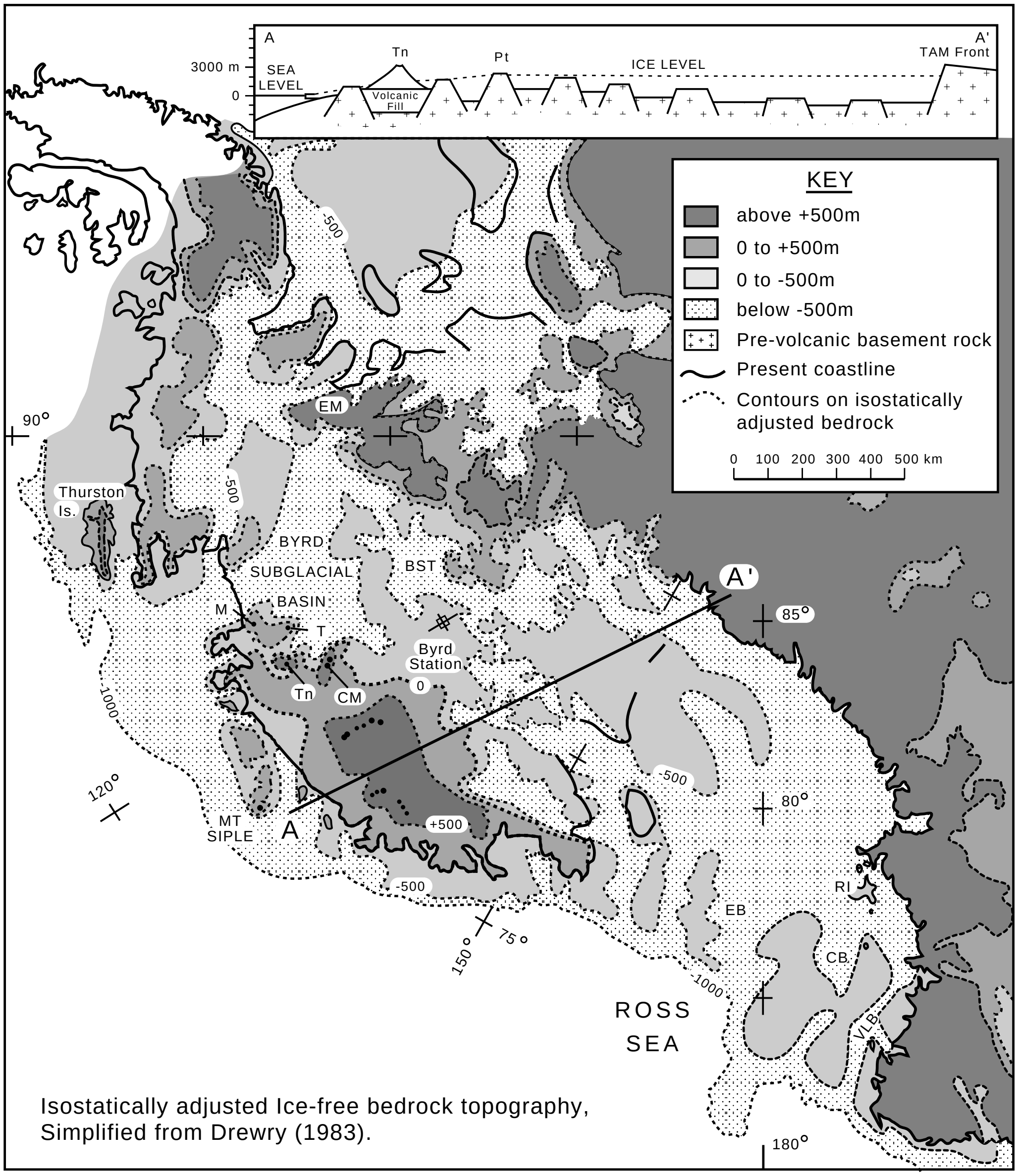


Figure 10. Map of West Antarctica showing isostatically adjusted bedrock surface, after removal of the ice sheet, simplified from Drewry (1983). The Marie Byrd Land dome is illustrated by the 0 m and +500 m contour lines in the center of the map. Our interpretation of dome structure is shown by schematic cross section (inset). Note that (1) except for the dome, virtually the entire rift system is 500 m or more beneath sea level and (2) the high elevations on the dome flanks (Mount Siple, Mount Murphy, Mount Takahe, Toney Mountain, Crary Mountains) are a product of volcanism rather than uplift. Abbreviations are as follows: BST—Bentley subglacial trench (below –2000 m), CB—Central basin, CM—Crary Mountains, EB—Eastern basin, EM—Ellsworth Mountains, M—Mount Murphy, Pt—Mount Petras, RI—Ross Island, T—Mount Takahe, TAM—Transantarctic Mountains, Tn—Toney Mountain, VLB—Victoria Land basin.

Tectonic Significance of Postbreakup Subsidence

The White and McKenzie (1989) model predicts rapid initial subsidence after breakup for nonvolcanic margins, but no subsidence, or even uplift (e.g., Cox, 1989), for the breakup of volcanic margins associated with elevated mantle temperatures that accompany plume activity. Many studies have confirmed that uplift and crustal extension are the first signal of plume arrival at the base of the lithosphere, often preceding volcanism by 10–140 m.y. (Kent et al., 1992). Thus, the subsidence recorded by the Waipounamu and its overlying sedimentary rocks, and strongly suggested by the West Antarctic erosion surface, provides evidence that breakup was unrelated to plume activity.

However, some rifted continental margins do not conform to a simple volcanic vs. nonvolcanic classification (Holbrook and Kelemen, 1993; Mutter, 1993) and the margin of West Antarctica may be one of those. Weaver et al. (1994) described a suite of 95–102 Ma alkalic (A-type) granitoids that occurs within the magmatic complex of Marie Byrd Land. These rocks postdate the subduction-related I-type granitoids within the complex and are coeval with the uplift and early rift activity already described. Weaver et al. (1994) suggested that these granitoids may represent a mantle plume that triggered melting in the lithosphere and later helped control the position of the 85 Ma breakup. Behrendt et al. (1994) described aeromagnetic anomalies that suggest the presence of 10^6 km^3 of volcanic rock beneath the ice, inland of the Marie Byrd Land volcanoes. This rock is thought to be mainly late Cenozoic, with contributions from Jurassic or Upper Cretaceous sources. These examples illustrate that there was significant, and possibly voluminous, igneous activity in West Antarctica that might support a plume-related breakup model, but would create the paradox of having this breakup coincide with subsidence. However, by ca. 105 Ma, a leading segment of the Pacific-Phoenix Ridge (Fig. 1B) had reached the subduction zone offshore of New Zealand, and thereafter, subduction shut off progressively eastward (Bradshaw, 1989; Lawver and Gahagan, 1994; Luyendyk, 1995). Subduction-related magmatism also terminated eastward in Marie Byrd Land (Mukasa, 1995) from 110 Ma at Mount Prince (near Bw in Fig. 2) to 96 Ma in the Kohler Range (KR in Fig. 2). This complex transitional event may have created either (1) a slab-window environment, like that responsible for alkalic volcanism on the Antarctic Peninsula during the past 50 m.y. (Hole et al., 1995), if the Pacific-Phoenix Ridge was actually subducted (Bradshaw, 1989; Lawver and Gahagan, 1994), or (2) a passive rift environment of magma generation if it was not (Luyendyk, 1995). We believe that some sort of magma-generation mechanism related to plate dynamics is a preferable alternative to a mantle plume for the 95–102 Ma interval and should be carefully evaluated in the future.

Postplanation Deformation in West Antarctica

Because of its low relief, known age, and origin near sea level, the West Antarctic erosion surface can be used to measure tectonically induced uplift associated with Neogene volcanism. This provides a method of evaluating plume activity that is independent of geochemistry. Raw elevation data illustrate the magnitude of uplift and areal dimensions of the Marie Byrd Land dome. Patterns of volcanism associated with uplift provide a chronology of dome growth, and the preservation of the surface allows us to estimate the isostatic response to erosional denudation (England and Molnar, 1990). Our biggest problem has been lack of data on basement elevations in grabens; all elevation data are from horst summits and therefore represent maximum uplift at each locality. There is also uncertainty about postbreakup subsidence, and we have therefore used both present sea level and –500 m in our calculations as likely preuplift levels of the erosion surface (cf. Fig. 10).

The Marie Byrd Land dome (Fig. 10) is a composite feature produced by tectonic uplift and constructional volcanism. The structural dome, defined by elevations of the West Antarctic erosion surface, rises to a 2700 m crest at Mount Petras (Pt in Fig. 2). Other erosion-surface exposures are crudely stepped downward to the coast; those 600 m and lower are exposed only within about 30 km of the coast, because farther inland they are buried beneath the ice sheet (Fig. 10, inset). Nevertheless, it seems clear that structural uplift is a major component of dome topography. The highest peaks in the region are late Cenozoic volcanoes (Fig. 10), and these have been ignored in our assessment of uplift.

Two spatial patterns of volcanism are closely related to the doming. (1) The most long-lived pattern involves basalts that rest on the erosion surface. Their ages increase with increasing elevation of the surface (Fig. 11), suggesting that doming and volcanism are coeval and genetically related. (2) Felsic volcanism began at ca. 18 Ma at Mount Flint, on the crest of the dome, and younger centers of felsic activity have been displaced systematically toward the perimeter of the dome along north-south–east-west paths (Fig. 2). This pattern has been attributed to doming of the lithosphere over a hot spot, accompanied by release of differentiated magma from shallow reservoirs along radially propagating relict fractures (LeMasurier and Rex, 1989). These relationships between doming and volcanism, over ≈25 m.y., are the basis for inferring that uplift began with the inception of volcanism at 28–30 Ma.

We have attempted to correct the raw uplift data for isostatic rebound and irregularities of the deformed surface and to compare the results with the White and McKenzie (1989) uplift curves. These curves are based on the calculated uplift or subsidence associated with varying mantle temperatures beneath rifted continental margins. They predict up to 500 m of uplift for cases where mantle temperatures are only hot enough to produce

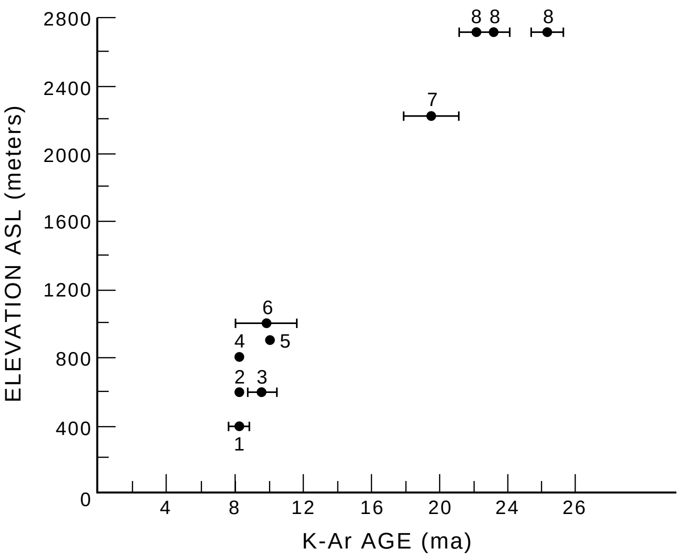


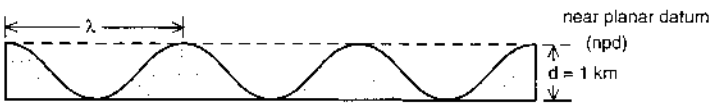
Figure 11. Elevations above sea level (ASL) of erosion-surface remnants vs. age of the oldest volcanic rock resting on the surface in Marie Byrd Land. Locations of numbered points are as follows (letter identifications are in Fig. 2): 1—Mount Murphy (M), 2—Holmes Bluff (H), 3—Bowyer Butte (Bw), 4—Kouperov Peak (K), 5—Patton Bluff (P), 6—Kohler Range (KR), 7—USAS Escarpment (UE), 8—Mount Petras (Pt). Mount Petras and the USAS Escarpment are at the crest of the Marie Byrd Land dome; the remaining points lie on the dome flank. The crudely linear pattern of the plot suggests a 105–122 m/m.y. rate of uplift (LeMasurier and Rex, 1983).

Table 2. METHOD OF ESTIMATING ISOSTATIC RESPONSE TO EROSIONAL UNLOADING, BASED ON DISSECTION OF A NEAR-PLANAR DATUM SURFACE.

Procedure:

1. A single, near-planar datum surface that pre-dates valley cutting is assumed.
2. Model landscape cross-sections are constructed, at right angles to valleys, with an arbitrary depth (d).
3. Cross sections are modelled as sine curves
   (a) Widths of valleys at the divides, λ, are related to depths, d, as: $\lambda = \pi d$
   (b) Widths of divides are modelled as multiples of valley widths.
   (c) Area of each valley, $A = d\lambda/2$*
4. Calculate area of rock removed by erosion (A_r) as the sum of the area of valleys; recalculate A_r as a layer of uniform thickness (t) removed from the model area.
5. Estimate t for Marie Byrd Land dome using local examples of divide width and valley depth.
6. Isostatic response assumes:
   (a) Regional granitic crust (density = 2.7) is buoyed up by mantle (density = 3.3), i.e. Archimedes principle applies.
   (b) Thickness of crustal rock removed by erosion is replaced (isostatically compensated) by an equivalent mass of mantle rock (t_m).

Case 1. Accordant summits.



A (3 valleys) = A_r(rock removed) = $3\left(\frac{d\lambda}{2}\right) = 3\left(\frac{3.1416}{2}\right) = 4.7\ km^2$

Total area of the cross section (A_t) = 3λ (d) = 9.4 km^2; $A_r/A_t = 0.5$
Uniform thickness of rock removed (t_1) can be expressed as 1/2 d = 0.5 km.

Case 2. λ (divides) = λ (valleys)

A (2 valleys) = A_r(rock removed) = $2\left(\frac{3.1416}{2}\right) = 3.1\ km^2$
$A_t = 9.4\ km^2$; $A_r/A_t = 0.33$
Uniform thickness of rock removed (t_2) can be expressed as 1/3 d = 0.33 km

Case 3. λ(divides) = 2λ (valleys)

A (1 valley) = A (rock removed) = $1\left(\frac{3.1416}{2}\right) = 1.6$
$A_t = 9.4\ km^2$; $A_r/A_t = 0.17$
Uniform thickness of rock removed (t_3) can be expressed as 0.17 km.

Case 4. West Antarctic erosion surface, Marie Byrd Land.
d ≃ 300 m, from Holmes Bluff; λvalleys = 0.94 km (from $\lambda = \pi d$)

A: λdivide = 3λvalley ≃ 2.8 km, from Mt. McCoy (Fig. 2)
$A_r = d\lambda/2 = 0.3(.94)/2 = 0.14\ km^2$
$A_t = 4\lambda(d) = 4(.94)(0.3) = 1.13\ km^2$; $A_r/A_t = 0.12$
$t_{4A} = 120$ m
B. λdivide = 6λvalley ≃ 5.6 km, from Bowyer Butte (Figs. 2 & 3)
$A_r = 0.14\ km^2$
$A_t = 7\lambda(d) = 7(.94)(0.3) = 2.0$; $A_r/A_t = 0.07$
$t_{4B} = 70$ m
Uniform thickness of rock removed ≃ <u>95 m</u> (average of 4A and 4B)
Isostatic rebound: $2.7(95) = 3.3(t_m)$; t_m = <u>78m</u> uplift

$$*A = \int_0^{\lambda} \left[\frac{d}{2}\sin\left(\frac{2\pi}{\lambda}x - \frac{\pi}{2}\right) + \frac{d}{2}\right] dx = \frac{d\lambda}{2}$$

alkali basalt, as in Marie Byrd Land, although White and McKenzie (1989) noted that uplift may be greater over the central plume of the hot spot. The maximum uncorrected uplift at Mount Petras is 2700 and 3200 m above the inferred preuplift levels of 0 and –500 m, respectively.

Our method of correcting for isostatic rebound is based on comparing depth and width of valleys with the width of erosion-surface remnants (see Table 2 for procedure). The calculations show that (1) if an erosion surface is preserved at all, the amount of erosion since the surface formed is not likely to be great and (2) large changes in divide width do not produce large variations in the estimate. Our estimate for the Marie Byrd Land dome is based on localities near the coast, where exposures help constrain divide width and valley depth. Following the procedure given in Table 2, we calculate the regional isostatic portion of uplift to be on the order of 78 m.

Recent evidence suggests that the Antarctic ice sheet formed at ca. 35 Ma (Barrett, 1989); therefore, the present erosion-surface elevations should be corrected to an ice-free basis. However, on the map of ice-free Antarctica (Drewry, 1983), the resolution is not good enough to pick elevations of Mount Petras and other erosion-surface localities. Regional ice thicknesses range from >3000 m in subglacial grabens to <1000 m near the coast, and the reliability of data in this region is too low to justify detailed averaging (Drewry, 1983). We have calculated that the removal of 1000 m of ice (density = 0.9) would yield ≈270 m of rebound. After subtracting the

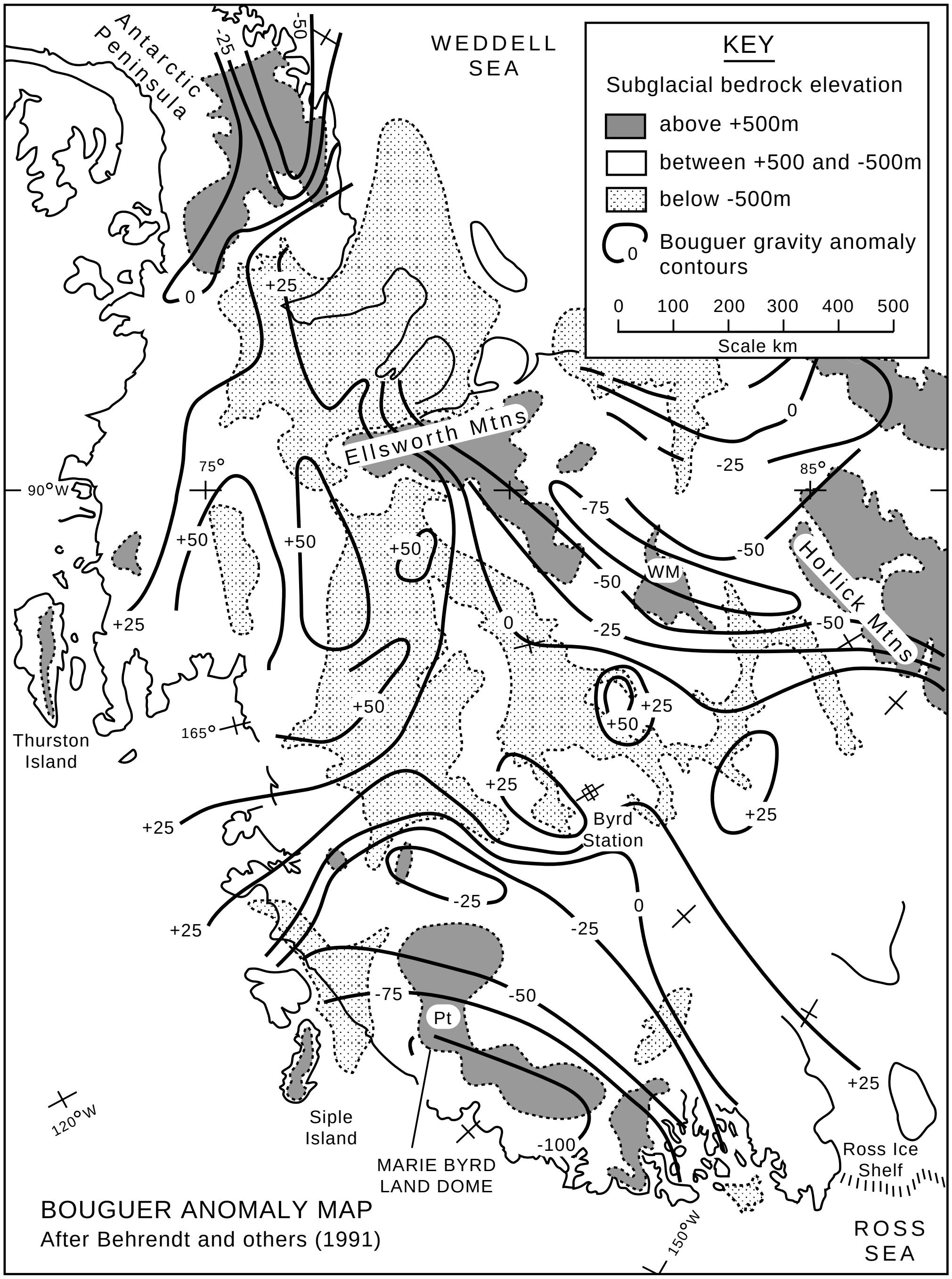


Figure 12. Bouguer anomaly map, after Behrendt et al. (1991), superimposed on present-day bedrock topography (Drewry, 1983). Pt is Mount Petras, at the crest of the Marie Byrd Land dome. Contour interval is 25 mgal.

isostatic portion of uplift and adding the response to deglaciation, the figures for maximum tectonic uplift would be ≈2900 and ≈3400 m above inferred preuplift levels.

The average erosion-surface elevation, between horsts and grabens, would have to be ≈300 m for these corrections to yield the 500 m uplift (i.e., ≈300 – 78 + 270 = ≈500) predicted by White and McKenzie (1989). This consideration implies that the erosion surface would lie at 2100 m below sea level in grabens near Mount Petras, thus requiring ≈4800 m offset on intervening faults. If these are listric faults, such large displacements should be accompanied by backward rotation of the erosion surface, which we do not see (Fig. 3). Furthermore, plate reconstructions (Lawver and Gahagan, 1994) and regional deformation patterns (Wilson, 1995) preclude large amounts (>50 km) of Cenozoic extension in the Ross Sea and West Antarctica. Although there are no data available to calculate average uplift, our results suggest that either (1) uplift greatly exceeds that predicted by the White and McKenzie curves, which could imply dynamic support over a plume head, or (2) if average uplift is only ≈500 m, the faults are planar and steep, permitting large offset with minimal extension. In either case, we are left with two questions: What caused the uplift, and is the dome supported by thick crust or low-density mantle?

Figures 12 and 13 show the gravity data available for this region superimposed on present-day subglacial topography. As expected, the large regional negative Bouguer anomalies are correlated with high bedrock topography and may be evidence of regional isostatic compensation by ei-

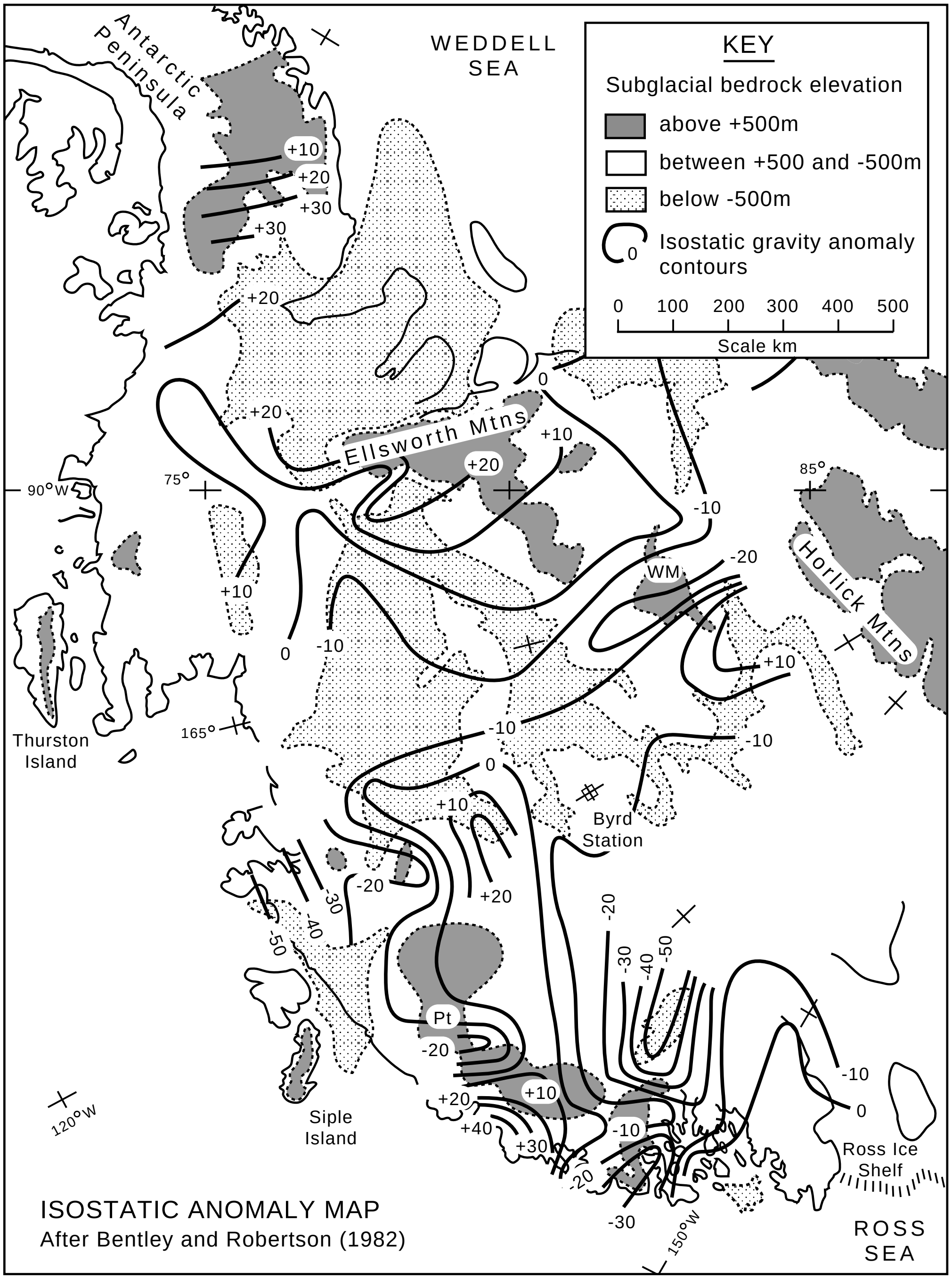


Figure 13. Isostatic anomaly map, after Bentley and Robertson (1982), superimposed on present-day bedrock topography (Drewry, 1983). Pt is Mount Petras. Contour interval is 10 mgal.

ther thick crust or low-density mantle, as in the southern Sierra Nevada (Wernicke et al., 1996). The relevant isostatic anomalies are the positive anomaly on the south side of the Marie Byrd Land dome and, for comparison, the positive anomalies at the base of the Antarctic Peninsula, over the Ellsworth Mountains, and north of the Horlick Mountains. The Marie Byrd Land dome anomaly is the only one of these associated with a large volume of young volcanic rock (LeMasurier and Thomson, 1990). These anomalies were analyzed before the existence of the West Antarctic rift system had been recognized and were tentatively interpreted to be the result of either uncompensated crustal thinning or the intrusion of lower-crustal mafic rocks into the upper crust (Bentley and Robertson, 1982). Neither explanation is compatible with thicker crust beneath the Marie Byrd Land dome, and either would tend to support Pratt-type compensation in the mantle, consistent with plume-related uplift. However, the proportion of Pratt- vs. Airy-type compensation, and whether the dome is totally compensated isostatically, cannot be constrained until seismic determinations of crustal thickness and upper-mantle velocity have been made. From a geologic perspective, we are impressed that uplift was recent and accompanied by extensive volcanism. The buoyant effect of underplating is included in the White and McKenzie (1989) curves, and no other mechanism for recent crustal thickening is apparent; thus buoyant uplift over low-density mantle seems reasonable.

SUMMARY AND CONCLUSIONS

(1) The West Antarctic and Waipounamu erosion surfaces represent a prolonged period of tectonic stability and erosional leveling that must have begun when West Antarctica and New Zealand were still joined and continued long after separation. Deposits resting on the Waipounamu surface indicate that leveling was near completion by ca. 75 Ma in coastal southern New Zealand and the Chatham Islands. Published estimates suggest that 10–15 m.y. is a minimal amount of time to produce a regional erosion surface of very low relief. Although there is clear evidence from both New Zealand and West Antarctica for rifting and high relief at ca. 100 Ma, it seems equally clear that this tectonism must have declined greatly before the 85 Ma breakup in order to provide a reasonable amount of time to produce these surfaces.

(2) Following breakup, final leveling of the Waipounamu surface was accompanied by gradual subsidence and marine transgression. A similar history is suggested for the West Antarctic rift system by the history of the Ross Sea and the isostatically corrected sub–sea-level elevations of large tracts of the rift. Breakup was evidently not accompanied by elevated mantle temperatures commonly associated with plume activity.

(3) The timing and magnitude of dome uplift in Marie Byrd Land is clearly recorded by the West Antarctic erosion surface, beginning at ca. 28–30 Ma and accompanied by block faulting and basalt volcanism. After correcting for post–35 Ma glacial loading and for rebound related to Cenozoic erosional unloading, we calculate that there has been a maximum of ≈3 km of Neogene tectonic uplift. Average uplift cannot be calculated because there are no data on sub-ice elevations of the erosion surface. Comparison of some average-uplift possibilities, and their associated fault offsets, with the White and McKenzie (1989) uplift curves suggests that (1) average uplift is much greater than that predicted for alkalic basalt provinces, implying some dynamic support, or (2) if the average uplift does match the curves (≈500 m), then the faults are planar, with displacements of >4500 m, and have accommodated >2 km of net subsidence of grabens below the preuplift level of the erosion surface. We believe this uplift marks the inception of plume activity in Marie Byrd Land.

(4) We have tried to find constraints on whether uplift of the Marie Byrd Land dome is caused by thicker crust or lower-density mantle compared to the crust and mantle of its surroundings. Reconnaissance geophysical data do not conflict with a plume model, which remains the most geologically reasonable uplift mechanism. However, the extent and nature of isostatic compensation in this region cannot be demonstrated until seismic studies of crustal thickness and upper-mantle velocity have been made.

(5) Neither erosion surface is a true peneplain, in that neither is an end product of fluvial erosion graded to sea level. Thus the terms "Otago peneplain" and "Cretaceous peneplain" are misnomers for what we consider to be a single, time-transgressive surface of fluvial and marine erosion in New Zealand. We propose the new name Waipounamu erosion surface for this feature.

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