Antarctica at the close of a millennium. *Royal Society of New Zealand Bulletin 35*: 2002, 453–469 453
1–877264–06–7

Plenary paper

Tectonics and landscape evolution of the Antarctic plate since the breakup of Gondwana, with an emphasis on the West Antarctic Rift System and the Transantarctic Mountains

PAUL FITZGERALD
Department of Earth Sciences
Syracuse University
Syracuse, NY 13244, USA

Abstract Landscape evolution of the Antarctic continent since the breakup of Gondwana has been controlled in the first order by changes in the location and configuration of plates. Following breakup, Antarctica moved toward its present polar location, and became geographically and then thermally isolated, which led to the development of the present-day hyper-arid, cold polar climate. The thermo-tectonic history of the Transantarctic Mountains (TAM) can be constrained since the Jurassic using apatite fission-track thermochronology to delineate timing and patterns of exhumation events that occurred in the Early Cretaceous, Late Cretaceous, and Cenozoic. These exhumation events are related to regional tectonic events: (1) the initial breakup between Australia and Antarctica in the Early Cretaceous; (2) the main phase of extension between East and West Antarctica in the Late Cretaceous accommodated on low-angle extensional faults; (3) propagation southward of seafloor spreading from the Adare Trough into continental crust underlying the western Ross Sea in the early Cenozoic, which likely acted as the trigger for the flexural uplift of East Antarctic lithosphere to form the TAM.

Landscape evolution of the TAM has been constrained using a combination of geomorphology, thermochronology to constrain the exhumation history, dating of ash deposits and surface exposure dating, as well as offshore drilling. Landscape evolution of the Dry Valleys block in Antarctica is a good example of the interaction of tectonics and climate. Following the onset of uplift of the TAM and change in base-level beginning in the early Cenozoic, escarpment retreat, formation of planation surfaces, and downcutting by fluvial systems with relatively little subsequent glacial modification produced a landscape that has changed very little in the last c. 15 m.y.

Keywords tectonics; landscape evolution; Antarctica; Transantarctic Mountains; West Antarctic Rift System

INTRODUCTION

The vast majority of the Antarctic landscape is presently undergoing extremely slow change and modification as >98% of the Antarctic continent is covered by ice. The development of this landscape since Gondwana breakup was ultimately controlled by the tectonic evolution of the Antarctic plate, beginning with rotation and translation of the West Antarctic microplates and formation of the Weddell Sea. The rifting of continents away from Antarctica, beginning with Africa and proceeding in a clockwise direction around Antarctica, led to the physical isolation of Antarctica. This physical isolation, plus the present polar position of Antarctica, led to development of seaways and the circum-Antarctic current, thermal isolation, climate change, and the present-day cold polar environment (Lawver et al. 1992). While there is little doubt about the nature of the landscape today, controversy surrounds aspects of Neogene landscape evolution, especially the timing of the transition from "warm" dynamic ice sheets to the present cold, hyper-arid environment with a cold, stable ice sheet (e.g., Miller & Mabin 1998).

The objective of this paper is to present an overview into a number of factors controlling the development of the Antarctic landscape and the tectonic evolution of parts of Antarctica. This will include the major stages in the breakup of Gondwana, the tectonic evolution of the West Antarctic Rift System (WARS) and the TAM, and the erosional history of the Dry Valleys block during the Cenozoic. This paper will concentrate on the tectonics and landscape evolution of the Dry Valleys block, as most information from along the TAM is available from this area. As such, one must be cognisant that as the Dry Valleys block is presently free of through-going outlet glaciers, it may not be completely representative of the rest of the TAM.

GONDWANA: CONFIGURATION AND BREAKUP

The fit of the large continents within Gondwana is well-established (e.g., Lawver et al. 1998) as three continents, Australia, India, and Africa, fit well against the present-day rifted margins of Antarctica (Fig. 1A). Africa and South America fit tightly together across a closed Atlantic Ocean. The Gondwana reconstruction is less well constrained along the Transantarctic margin of East Antarctica, where there is considerable uncertainty over the number and position of some of the microplates within the West Antarctic region (e.g., Dalziel & Elliot 1982; Storey 1996; Storey et al. 1998). The Ellsworth-Whitmore Mountains crustal block (EWM), Antarctic Peninsula, Thurston Island, and Marie Byrd Land are generally accepted as the main West Antarctic microplates. The Haag Nunatak, Berkner, and Filchner microplates are also used in reconstructions (e.g., De Wit et al. 1988), and Marie Byrd Land can be thought of as comprising a west and east portion (Divenere et al. 1996). Paleomagnetic data, geometry, and geological constraints are used to constrain the fit of these microplates. As Antarctic Peninsula, Thurston Island, and Marie Byrd Land contain a large proportion of subduction-related rocks, these blocks are retained in approximately their present location along the paleo-Pacific margin of Gondwana in the majority of reconstructions (e.g., Lawver et al. 1992). To obtain their Gondwana pre-breakup positions, the Antarctic Peninsula

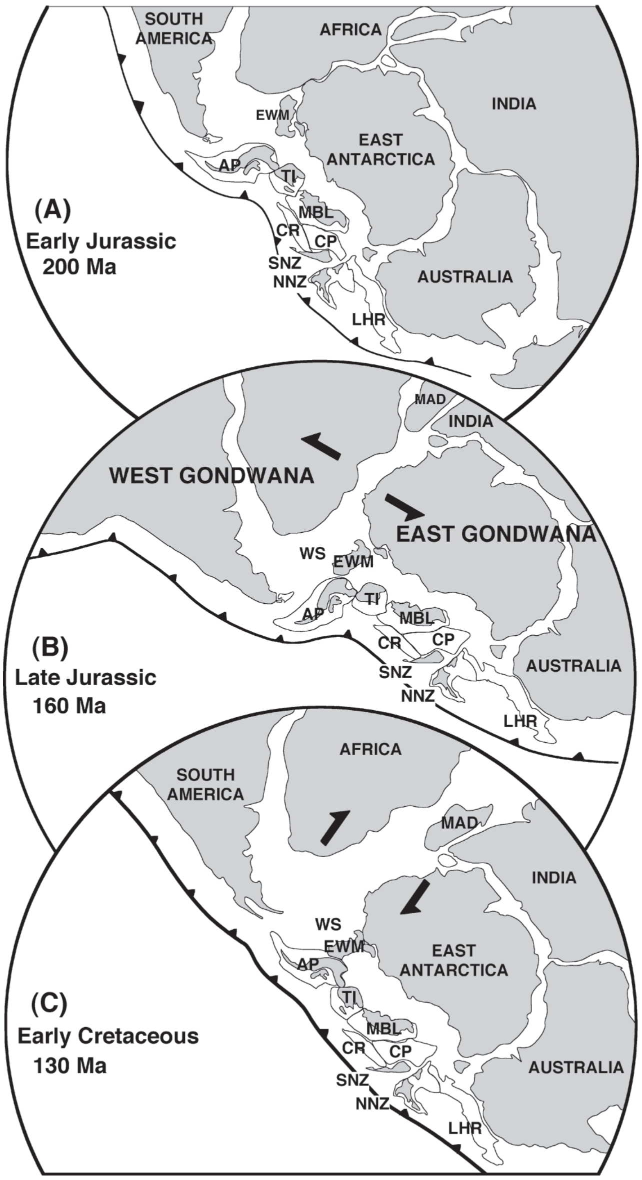


Fig. 1 Gondwana tight fit reconstruction and breakup model. Continent and microplate positions are from Lawver et al. (1992, 1998) with other information from Storey (1996). AP, Antarctic Peninsula; TI, Thurston Island; MBL, Marie Byrd Land; CR, Chatham Rise; CP, Campbell Plateau; SNZ, southern New Zealand; NNZ, northern New Zealand; LHR, Lord Howe Rise; WS, Weddell Sea.

is rotated anticlockwise from its present-day position with respect to East Antarctica and located along the western margin of the southern portion of South America. The Thurston Island block is rotated (Grunow et al. 1991) into its pre-breakup position, and the two microplates of Marie Byrd Land are restored along a strike-slip fault (Divenere et al. 1995). Campbell Plateau, Chatham Rise, and North and South Islands of New Zealand can be reconstructed to Marie Byrd Land.

The initial rifting stage in the breakup of Gondwana (Fig. 1B) involved right-lateral transtension as East Gondwana (Antarctica, Australia, India, New Zealand) and West Gondwana (South America and Africa) moved apart with stretching beginning in the north and propagating southward (Lawver et al. 1992). Initial breakup was also associated with plume-generated magmatism and rotation and translation of microplates. Plume-related magmatism, with a head likely located in southeastern Africa (Cox 1988; White & MacKenzie 1989; Storey 1996), led to the emplacement of huge within-plate mafic and felsic magmatic provinces in many Gondwana continents as well as Antarctica. The mafic part of the magmatic province forms a long linear belt from Australia (Tasman province), through Antarctica (Ferrar and Dronning Maud Land provinces) to South Africa (Karoo province). Ar/Ar dating in these provinces gave dates of 175 ± 18 Ma (Tasman; Hergt et al. 1989), 177 ± 2 Ma (Ferrar; Heimann et al. 1994), and 182 ± 2 Ma (Karoo; Hooper et al. 1993). This short eruption time is similar to other continental flood basalt provinces. Granites contemporaneous with this basaltic volcanism and considered to be rift-related occur in West Antarctica (Storey et al. 1988) and southern South America (Chon-Aike or Tobífera province) (Gust et al. 1985). As the long linear nature of the basaltic rocks exposed in the Ferrar and Tasman provinces, subparallel to the Pacific margin of Gondwana at that time, is not compatible with classic circular plumes, Cox (1988) proposed a hot line rather than a hot spot. Elliot (1992) suggested that Jurassic magmatism occurred along a number of rifts that intersected at the Dufek Massif. Zones of weakened lithosphere were postulated to propagate away from nodes of magmatism (e.g., Dufek Massif, Beardmore Glacier region, southern Victoria Land (SVL)), providing pathways for lateral migration of magmas derived from lithospheric sources. More recently, Storey & Kyle (1999) suggested that the Weddell Sea region was the site of a large Gondwana megaplume causing domal uplift and formation of a triple junction. In this case, production of magma batches, due to different degrees of plume-lithosphere interaction, migrated along zones of crustal weakness to form the various large igneous provinces. It has even been suggested that before breakup, this Gondwana breakup plume may have caused or expedited formation of the early Mesozoic Gondwanide foldbelt, due to the buoyancy of a hot plume acting on the downgoing slab and causing it to flatten (Dalziel et al. 1999).

The Weddell Sea area is the focus for rotation and translation of the microplates such as the EWM, a displaced part of the Gondwanide fold belt. The EWM has a structural trend perpendicular to the TAM and has long been considered to have rotated into its present position (Schopf 1969). Although the original position of the EWM is controversial, it most likely originated from near southeastern Africa (e.g., Curtis & Storey 1996). Rotation of West Antarctic microplates must have been accomplished by c. 165 Ma (the time of opening of the Weddell Sea), and rotation of the EWM was finished by 175 Ma (Grunow et al. 1987), before translation of the EWM into their present position. Rotation of microplates did not apparently involve the production of oceanic crust (Marshall 1994) but may have occurred as block rotation with controlling faults concealed beneath Mesozoic sedimentary basins in the Weddell Sea (Storey 1996). A good example of a sedimentary basin that formed during the initial rifting stage is the Latady Basin on the east side of the Antarctic Peninsula (Rowley et al. 1983). Stretching also occurred in the Ross Embayment (WARS) during initial stages of Gondwana breakup, as shown by Jurassic tholeiitic magmatism (Ferrar Dolerite and

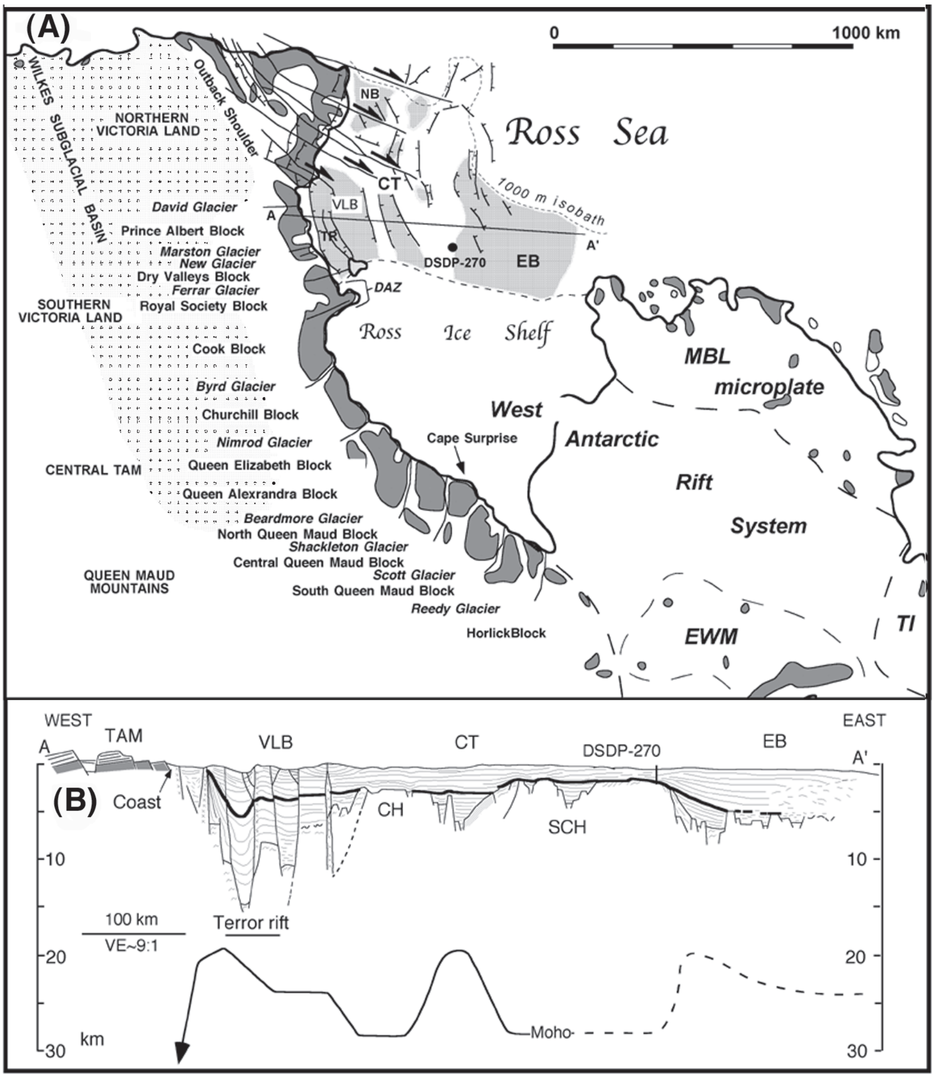


Fig. 2 **A**, Map of the Transantarctic Mountains and West Antarctic Rift System. Structural features along the TAM and in the West Antarctic Rift System are from Tessensohn & Wörner (1991), Fitzgerald (1992), Fitzgerald & Baldwin (1997), and Salvini et al. (1997). DAZ = Discovery accommodation zone from Wilson (1999). Basin positions in the Ross Sea (light grey) are from Davey & Brancolini (1995) and includes North Basin (NB), Eastern Basin (EB), Central Trough (CT), and the Terror Rift (TR) lying within the Victoria Land Basin (VLB). The approximate outline of the Wilkes subglacial basin is shown inboard of the TAM in a dot pattern. West Antarctic microplates are after Dalziel & Elliot (1982); see Fig. 1 for abbreviations. **B**, Generalised crustal profile A–A′ (in A) across the Ross Sea is after Cooper et al. (1991). Early rift grabens lie beneath regional unconformity U6 (heavy line). Late rift-faults and intrusive rocks deform the VLB and some small basement grabens. Note intra-basement reflector (dashed line) on eastern edge of VLB, possibly a detachment fault extending under the entire basin. DSDP site 270 is projected onto this cross-section.

Kirkpatrick Basalt) (Elliot 1992), as well as extension between the Lord Howe Rise and northern New Zealand (Lawver et al. 1992).

The continental margins around Antarctica young in a clockwise direction, starting in the Weddell Sea, which formed initially as a back-arc basin in the Late Jurassic (Fig. 1B) (Lawver et al. 1991). In the Early Cretaceous (Fig. 1C), the Gondwana breakup stress regime changed from dominantly north–south between East and West Gondwana to dominantly east–west with the two-plate system being replaced by a multiple plate system (e.g., Lawver et al. 1992). This change in stress regime dramatically affected the tectonics of the region. For example, there was large-scale ductile deformation concentrated along shear zones in gneissic and magmatic rocks in the Antarctic Peninsula (Storey et al. 1996), and existing sedimentary basins such as the Latady Basin underwent thin-skinned deformation and inversion (Kellogg & Rowley 1989). Separation also began between India and Antarctica in the Early Cretaceous (Lawver et al. 1991). Initial stretching began between Australia and Antarctica in the Early Cretaceous (Stagg & Willcox 1992), but seafloor spreading did not begin until c. 95 Ma (Cande & Mutter 1982; Veevers et al. 1990). By c. 110 Ma, the microplates of West Antarctica had nearly reached their present location with respect to East Antarctica. By the Late Cretaceous, Antarctica had reached its final polar location and configuration, and the final stage of breakup was completed when New Zealand (Campbell Plateau) rifted from Marie Byrd Land at 84 Ma (e.g., Stock & Molnar 1987; Lawver et al. 1991).

TECTONIC EVOLUTION OF THE WEST ANTARCTIC RIFT SYSTEM AND THE TRANSANTARCTIC MOUNTAINS

Architecture and tectonic evolution of the WARS

The WARS encompasses the Ross Sea, the area under the Ross Ice Shelf, and part of the area under the West Antarctic Ice Sheet (Fig. 2). Except for the area under the Ross Sea, the WARS is not that well known, although recent airborne geophysical surveys (e.g., Bell et al. 1998) are beginning to delineate more features of this huge rift system. A series of asymmetric basement grabens (North Basin, Victoria Land

Basin, Central Trough, Eastern Basin) separated from each other by basement highs (Coulman High and Central High) underlie the Ross Sea (e.g., Davey & Brancolini 1995). The sedimentary basins of the Ross Sea have been formed by crustal extension since Gondwana breakup. Basin formation began with development of a rift zone associated with Jurassic tholeiitic magmatism, subsidence, and possible strike-slip motion in the Ross Sea coeval with emplacement of Jurassic magmatic rocks in the TAM (e.g., Schmidt & Rowley 1986; Elliot 1992; Wilson 1993). However, the main phase of extension (and hence crustal thinning) in the WARS between East and West Antarctica was in the Late Cretaceous between c. 105 and 85 Ma (Lawver & Gahagan 1994, 1995). Post-early Oligocene extension has only occurred within a narrow zone within the Victoria Land Basin (e.g., Cooper & Davey 1985; Cooper et al. 1987, 1991).

The total amount of extension between East and West Antarctica within the Ross Embayment is poorly constrained. The magnitude of extension has been estimated by a variety of methods: (1) plate reconstruction (e.g., Stock & Molnar 1987); (2) simple one-layer stretching models assuming an initial pre-extension crustal thickness (c. 35 km) compared to the measured present crustal thickness (17–27 km) (Fitzgerald et al. 1986; Lawver & Scotese 1987; Behrendt & Cooper 1991; Bentley 1991; Cooper et al. 1991; Davey & Brancolini 1995), suggesting c. 400 km of extension; (3) paleomagnetic studies (Divenere et al. 1994), which suggest a substantially greater amount of extension (c. 440–1820 km). However, paleomagnetic constraints are hampered by the lack of correlative mid-Cretaceous poles and by the large uncertainties on rotation amounts between West Antarctic microplates and East Antarctica, reducing the precision when determining the amount of the translation between East and West Antarctica (Luyendyk et al. 1996). Most reconstructions (e.g., Cooper et al. 1991; DiVenere et al. 1994) use estimates of c. 100% when removing the post-mid-Cretaceous extension between East and West Antarctica.

The mechanism of extension in the Ross Embayment during Late Cretaceous extension is not well known, as this area of extended crust is now completely submerged. Regional crustal architecture, plus characteristics of basement and sedimentary breccia at DSDP site 270 (Ford & Barrett 1975), are consistent with rocks having undergone progressive deformation during exhumation along low-angle detachments (Fitzgerald & Baldwin 1997). Fitzgerald & Baldwin (1997) suggested that Cretaceous extension between East and West Antarctica was accommodated predominantly by detachment faulting. Metamorphic basement recovered at DSDP site 270 has been both ductilely and brittlely deformed and sedimentary breccia is locally derived, formed by erosion of previously brecciated ductilely deformed basement (Ford & Barrett 1975). Angular clasts within the syn-tectonic breccia contain ductile fabric that had been subsequently overprinted by brittle deformation, and the breccia itself has also undergone brittle deformation. The basement thus records an apparent temporal progression of ductile to brittle deformation that is also found in footwalls of major detachment faults, formed as the originally deep (5–15 km) lower plate rocks are drawn outward and upward from beneath the brittlely extending upper plate. Apatite fission-track data from calc-silicate gneiss recovered from site 270 has a mean age of 103 ± 11 Ma, suggesting that this rock was exhumed in the mid Cretaceous during regional extension. Recently dredged mylonitic gneisses from west of Cape Colbeck, Edward VII Peninsula, also have fission-track and $^{40}Ar/^{39}Ar$ rapid cooling ages of mid–Late Cretaceous age (Baldwin et al. 2001), suggesting rapid exhumation at that time. However, these dredged samples appear to have a history more closely aligned to western Marie Byrd Land. There, subduction-related tectonics, active for most of the Paleozoic and Mesozoic, switched abruptly to extensional tectonics following subduction of the Pacific-Phoenix spreading centre at c. 105 Ma (e.g., Weaver et al. 1992, 1994). Approximately 20 m.y. of extensional tectonics in New Zealand, Marie Byrd Land, and the Ross Embayment followed (Bradshaw 1989; Lawver & Gahagan 1994), before seafloor spreading began between Marie Byrd Land and Campbell Plateau at c. 84 Ma (e.g., Mayes et al. 1990). Cretaceous extension in New Zealand associated with fragmentation of the Gondwana margin led to development of the Paparoa metamorphic core complex (Tulloch & Kimbrough 1989), exhumed rapidly at c. 100 Ma. At about the same time, metamorphic rocks in the Fosdick Mountains metamorphic complex were also exhumed from mid-crustal depths (Richard et al. 1994). While described by some as a metamorphic core complex (Weaver et al. 1992), the Fosdick Mountains lack the low-grade metamorphic overprint and the structural relationships of ductile to brittle deformation in lower plate and upper plate rocks similar to other core complexes (Luyendyk et al. 1996).

Architecture and tectonic evolution of the TAM

The TAM define the western edge of the Mesozoic–Cenozoic intracontinental WARS and the eastern margin of the East Antarctic Craton. This is a boundary that has experienced repeated tectonism since rifting during the Neoproterozoic, followed by transpression and subduction-related magmatism in the Mid–Late Cambrian. Terrane accretion and further transpression in the middle Paleozoic moved the convergent margin oceanward, before rifting processes began again beginning with Jurassic tholeiitic magmatism accompanying initial Gondwana fragmentation (e.g., Elliot 1975; Dalziel & Elliot 1982; Dalziel 1992). The TAM thus define a fundamental lithospheric boundary that has a profound crustal anisotropy due to these repeated cycles of tectonism. The TAM stretch for c. 3500 km across Antarctica (Fig. 2), are typically 100–200 km wide, and reach elevations locally in excess of 4500 m. Structurally, the TAM can be envisaged as a series of simple asymmetric tilt blocks, typically with a longitudinal dimension of a few hundred kilometres. The blocks are divided by transverse structural features such as transfer faults, grabens, or in a few more complicated areas, by accommodation zones, typically with these transverse structural features being occupied by major outlet glaciers (Gould 1935; Fitzgerald 1992; Wilson 1999). The front of the TAM, the Transantarctic Mountain Front (TMF, Barrett 1979) is a zone of major normal faulting, with the onland portion typically extending c. 20–30 km inland from the coast, with 2–5 km of displacement down to the coast (Fig. 3). The TMF does not coincide with the entire area coastward of the frontal escarpment of the TAM; rather, the TMF seems to extend approximately one-half to three-quarters across the region from the coast toward the frontal escarpment.

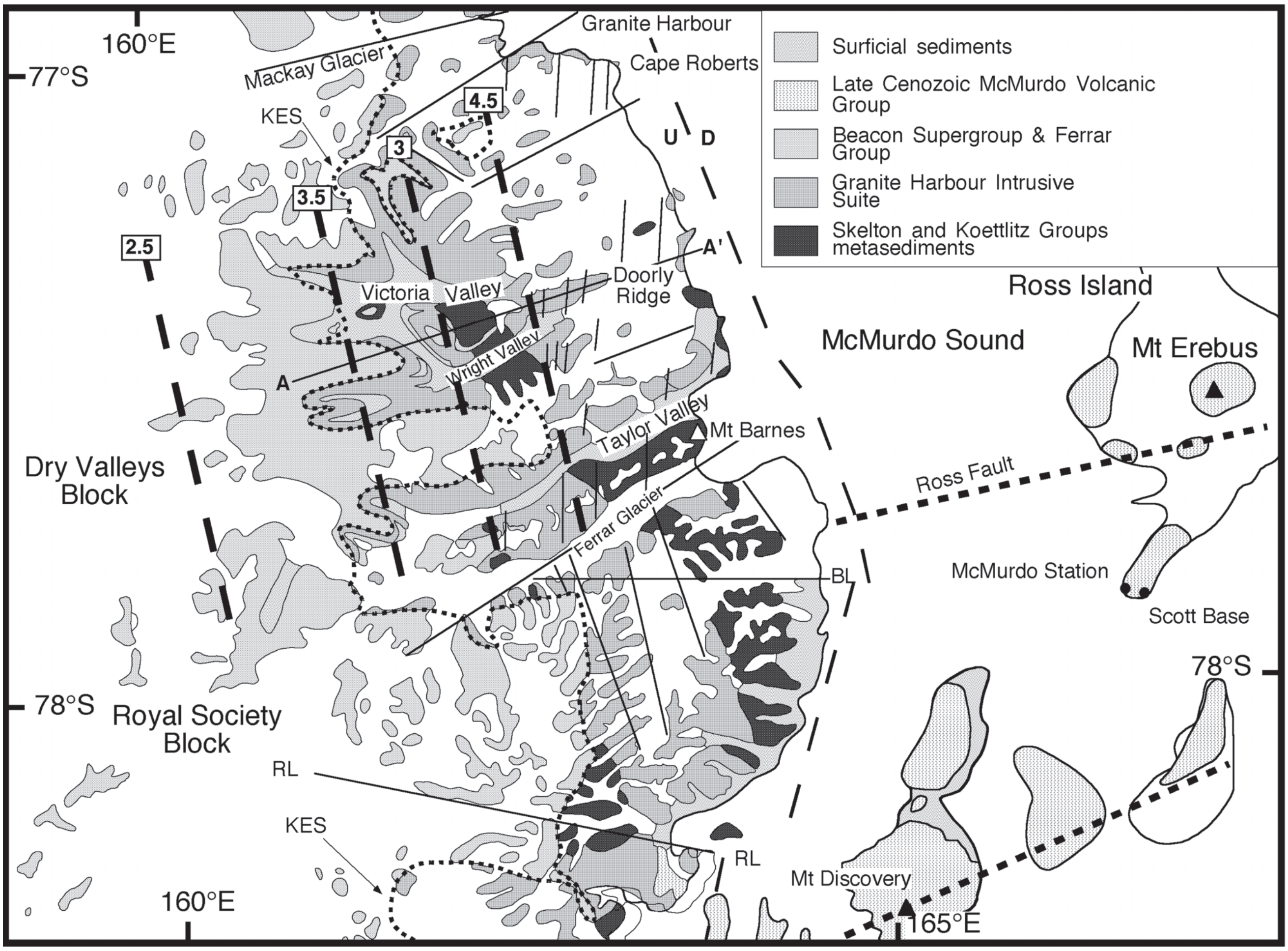


Fig. 3 Generalised geological map of SVL. Geology and location of faults and lineaments are after Gunn & Warren (1962), McKelvey & Webb (1962), Warren (1969), Findlay et al. (1984), Gleadow & Fitzgerald (1987), Fitzgerald (1992), and Wilson (1995, 1999). Offshore faults are from McGinnis et al. (1985), Damaske et al. (1994), and Barrett et al. (1995). Approximate "rock uplift" contours (marked by black dashed lines with rock uplift amount in kilometres from 4.5 to 2.5) for the Dry Valleys block are modified from Fitzgerald (1992); see Fig. 4.

The overall geology of the TAM is relatively simple (e.g., Elliot 1975). Regional basement is composed primarily of late Proterozoic–Cambrian metamorphic rocks and Cambrian–Ordovician granitoids of the Granite Harbour Intrusive Suite. Basement was deformed during the Cambrian–Ordovician Ross Orogeny that preceded and accompanied intrusion of granitoids. Exhumation of the basement following the Ross Orogeny produced the low-relief Kukri Peneplain (Gunn & Warren 1962) or Kukri Erosion Surface (KES), which is unconformably overlain by Devonian–Triassic glacial, alluvial, and shallow marine sediments of the Beacon Supergroup (Barrett 1991). In the Jurassic, sills of the Ferrar Dolerite dated at 177 ± 2 Ma (Heimann et al. 1994) intruded basement and sedimentary cover contemporaneous with their extrusive equivalent, the Kirkpatrick Basalt (e.g., Elliot 1975). There is a c. 160 Ma gap in the onland geologic record between the Jurassic tholeiitic magmatism and the late Cenozoic alkaline volcanism of the McMurdo Volcanic Group (LeMasurier & Thomson 1990). In a few places along the range, for example the Taylor Valley (Fig. 3), Miocene and younger shallow marine sediments are present, indicating that c. 400 m of subsidence occurred before deposition of these marine sediments (Sugden et al. 1995a). The outcrop pattern of the TAM reflects its simple tilt block structure as Kirkpatrick Basalt is limited to the inland flank of the range, whereas basement representing deeper crustal levels is exposed primarily along the coastal sector extending inland, usually only along major outlet glaciers. There are exceptions to this simple outcrop pattern. For example, in the Miller Range near the Nimrod Glacier, basement is present on the inland flank of the range, and in northern Victoria Land (NVL) the geology is dominated by the accretion of terranes since the late Cambrian (e.g., Stump 1995 and references therein).

The faulting across the TMF is poorly documented as faults in general are difficult to find (e.g., Wilson 1995; Turnbull et al. 1999) especially within the granitic basement. Beacon Supergroup rocks have been found downfaulted to the coast in only one place along the TAM, Cape Surprise in the central TAM (Barrett 1965; Miller et al. 1999). Recently, Beacon Supergroup rocks have been drilled offshore SVL at a depth of 825 m below seafloor in the Cape Roberts drillhole #3, indicating some 3 km of offset relative to Beacon Supergroup strata inland (Cape Roberts Science Team 2000). The ridge east of Mt Doorly in SVL (Fig. 4) records step-faulting across the TMF, in general down to

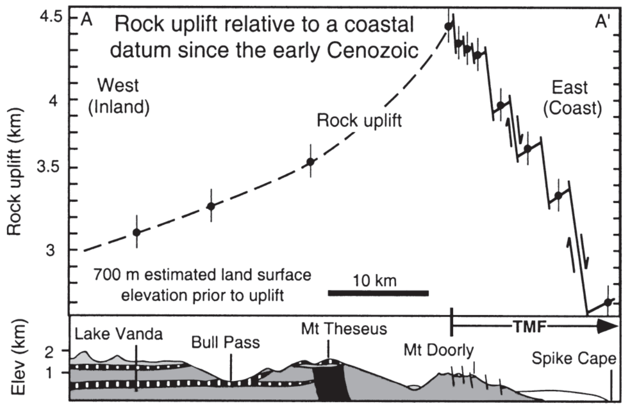


Fig. 4 Overall structure of the TAM in the Dry Valleys area shown using rock uplift plotted across the TAM in comparison to a geological cross-section (A-A′ as marked in Fig. 3). Cross-section is modified from a cross-section originally in McKelvey & Webb (1962). Geological symbols as for Fig. 3, except Beacon Supergroup (light grey) and Ferrar Dolerite sills (white lines on black background) have been separated. Diagram is modified from Gleadow & Fitzgerald (1987), with additional data from Fitzgerald (1992). Position of faults is from field mapping along the ridge east of Mt Doorly or from the offset in AFT ages. As faults offset a partial annealing zone formed before the early Cenozoic onset of "rapid" exhumation, faults must be Cenozoic in age.

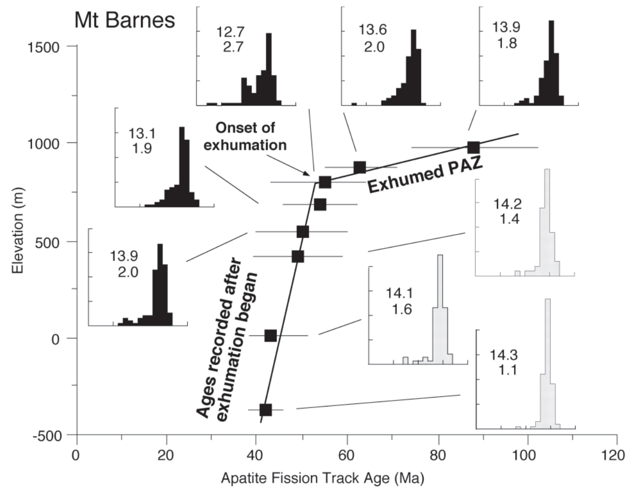


Fig. 5 AFT profile from Mt Barnes (eastern end of Kukri Hills) showing the "classic" shape of an exhumed PAZ (gentle slope with confined track length distributions indicative of significant annealing). The "break in slope" defines the transition (55 Ma) from a time of relative thermal and tectonic stability to a time of "rapid" exhumation (steep slope with confined track length distributions indicative of rapid cooling), which defines an apparent exhumation rate.

the east (coastward) across faults with throws from 10's to 100's of metres. The cumulative offset along the Doorly Spur is 800 m, with c. 2 km of throw across the onland portion of the TMF from the axis of maximum uplift to the coast (Gleadow & Fitzgerald 1987; Fitzgerald 1992). This offset compares to a cumulative offset across the TMF in the central TAM near Cape Surprise of 3–5 km (Barrett 1965; Miller et al. 1999). The strike of faults within the TMF in SVL is NNE through NE (Wilson 1995), oblique to the NNW strike of the TAM there, indicating that extension across the TMF is not orthogonal but includes a component of right-lateral strike-slip movement as well (Fitzgerald 1992; Wilson 1992, 1995). The rift margin, at least in SVL, is therefore an en-echelon array of oblique slip faults with an accompanying set of linking transverse structures. Wilson (1995) determined from fault kinematic solutions two regimes during the Cenozoic—a dominantly normal-sense phase occurring first, followed by a later phase of dominantly dextral shear. The timing of this strike-slip motion is difficult to constrain, especially from onland information, as it is known in SVL only to postdate faulting which has a dominantly normal (i.e., dip-slip) sense of motion (Wilson 1995). Salvini et al. (1997) proposed a tectonic model based on seismic interpretation of structures in the Ross Sea, linking these with onshore northwest–southeast structures crosscutting NVL. In their model, regional northwest–southeast dextral strike-slip faults overprint an older Mesozoic extensional event (i.e., the main phase of extension in the Late Cretaceous) which formed the main north–south-trending basins in the Ross Sea (Davey & Brancolini 1995). In the early Cenozoic, subsidence and magmatism were localised in the western Ross Sea (e.g., Davey & Brancolini 1995 and references therein), accompanying the initial uplift of the TAM (e.g., Gleadow & Fitzgerald 1987). This was followed by reactivation along northwest–southeast faults forming pull-apart basins, positive and negative flower structures, and push-up ridges caused by the northwest–southeast right lateral motion beginning c. 30 Ma and continuing to the present.

Uplift and exhumation history of the TAM

Because of the onland geologic gap in the TAM of c. 160 m.y. since the Jurassic, plus the fact that coring of the adjacent sedimentary basins in the Ross Sea has penetrated sediment only as old as late Eocene (Barrett 1996; Cape Roberts Science Team 2000), the uplift history of the TAM has to a large part depended upon the application of apatite fission-track thermochronology (AFTT). Fundamental to applying AFTT to this problem has been the sampling strategy employed, with the collection of vertical sampling profiles across the TAM, and the identification of an exhumed apatite partial annealing zone (PAZ) within those vertical profiles. The recognition of exhumed PAZs in these vertical profiles (Fig. 5), with the identification of a "break in slope" representing the base of an exhumed apatite PAZ, allows better constraints on the timing, amount, and rate of exhumation (e.g., Gleadow & Fitzgerald 1987; Fitzgerald & Gleadow 1990; Brown 1991). A break in slope marks the transition from a time of relative thermal and tectonic stability to a time of rapid cooling due to exhumation. Periods of exhumation in the TAM are usually separated by definitive periods of relative thermal and tectonic stability.

Before proceeding with discussion of the uplift and exhumation history of the TAM, one should be cautious of using thermochronology to constrain "uplift". Thermochronologic systems such as AFTT record cooling, they do not record exhumation, rock uplift, nor surface uplift (see England & Molnar 1990; Summerfield 1991; Fitzgerald et al. 1995). We assume that the cooling recorded within such systems is a result of exhumation, which may be caused by a change in base level due to rock uplift or basin subsidence. Rock uplift must exceed exhumation in order to generate surface uplift resulting in the formation of mountain ranges. However, cooling information can reasonably be interpreted as providing information about exhumation, and combining thermochronologic with geologic constraints may allow constraints about rock uplift and surface uplift to be derived. Surface uplift is the displacement of the Earth's surface with respect to a fixed reference frame such as mean sea level or the geoid. Rock uplift is the displacement of rock with respect to mean sea level or the geoid. Exhumation is erosion and/or tectonic denudation resulting from displacement of rock with respect to the Earth's surface. Surface uplift is equal to rock uplift minus exhumation (England & Molnar 1990; Summerfield 1991).

Since the Cambrian, the greatest amount of exhumation along the TAM occurred following the Ross Orogeny when metamorphic and granitic basement forming the core of a subduction-related mountain belt was eroded as much as 15–20 km (Capponi et al. 1990) to form the remarkably subdued KES. However, erosion to form the KES had little to do with the formation of the TAM as we see them today. Subsequent to the formation of the KES, the sediments of the Beacon Supergroup were deposited along the axis of the present TAM. The tectonic setting of the Beacon Supergroup has been described as a back-arc terrain (Dalziel & Elliot 1982) and a foreland basin (Collinson 1991), although it is apparent that the tectonic setting varied in space and time (Barrett 1991). The depositional basins of the Beacon were flanked by highstanding topography on either side, implying that since the close of Beacon sedimentation there has been a topographic reversal (Barrett 1981), as the TAM are now highstanding while the regions on either side are basins.

The situation in the TAM is similar in part to the topographic reversal in the southwest United States in the middle Tertiary following the Laramide Orogeny. Paleocurrent directions in early Tertiary alluvial "rim gravels" and finer grained equivalents along the Mogollon Rim of the Colorado Plateau indicate a source to the south (Peirce et al. 1979), which was at that time a highland due to crustal thickening, thrusting, and magmatism accompanying the Laramide Orogeny. Subsequent to the deposition of the "Rim Gravels," a topographic reversal has occurred. The region to the west and south is now the Basin and Range Province, a lowland brought about by extension and crustal thinning in the mid Tertiary, firstly by extension along low-angle detachment faults and then high-angle basin and range faulting (Spencer & Reynolds 1989). In this analogy, the TAM would occupy the location where the "rim gravels" are, and the WARS would represent the Basin and Range Province. One difference is that the Colorado Plateau-Basin and Range Province system has topographic reversal on only one side, whereas in Antarctica, the areas of the Ross Embayment and Wilkes subglacial basin have both apparently undergone topographic reversal.

Vertical changes or uplift along the TAM as a result of Jurassic magmatism is still poorly constrained. There was probably some surface uplift from inflation of the crust due to intrusion of dolerite sills and extrusion of basaltic flows, and there was at least 500 m of relief in the TAM as recorded by basaltic flows in the Allen Hills area (Elliot 1992). However, the uplift to form the present-day TAM has occurred since the Jurassic. AFTT records three main periods of exhumation along the TAM since the Jurassic, with exhumation periods initiated in the Early Cretaceous, Late

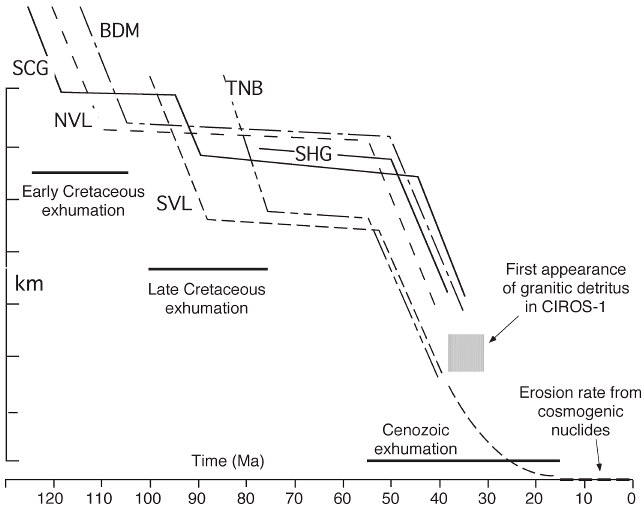


Fig. 6 Schematic diagram showing the variation of exhumation events along the TAM at different localities; see text for details. SCG, Scott Glacier area; BDM, Beardmore Glacier area; SHG, Shackleton Glacier area; SVL, southern Victoria Land; TNB, Terra Nova Bay; NVL, northern Victoria Land. A relative scale only is shown for exhumation as the amount of exhumation at any one locality (e.g., Dry Valleys; see Fig. 4) will vary across the range. Early or Late Cretaceous exhumation are not always present throughout an area (e.g., Scott Glacier region; see Fitzgerald & Stump 1997).

Cretaceous, and early Cenozoic (Fig. 6). In overview, these three episodes of exhumation appear relatively uniform along the range, but in detail, trends and subtleties emerge that may have an important bearing on the tectonic evolution of the range. The major period of exhumation accompanying rock uplift that formed the TAM we see today began in the early Cenozoic (Gleadow & Fitzgerald 1987; Fitzgerald & Gleadow 1988; Fitzgerald 1992). The TAM can be thought of as the erosional remnants of the upturned flexure of strong East Antarctic lithosphere that started to form in the early Cenozoic. However, the edge of the flexure is a fault zone (TMF) rather than a single major lithospheric-penetrative fault, as has been suggested (Stern & ten Brink 1989).

In NVL, the onset of Early Cretaceous exhumation is poorly constrained but began at 125–110 Ma and is of 1–2 km magnitude, as shown by AFT age profiles from the Admiralty Mountains (Redfield 1994). Regional AFT data from NVL indicate that the early Cenozoic exhumation began at c. 55 Ma (Fitzgerald & Gleadow 1988). Along the southeast coast near Mt Murchison and the mouth of the Tucker Glacier, exhumation since the early Cenozoic is c. 9 km, although across most of NVL the amount of exhumation is c. 4 km, decreasing westward. A mid-Cretaceous thermal event at c. 100 Ma, termed the Rennick Thermal Event and associated with formation of the Rennick Graben, has reset Rb-Sr and K-Ar dates in Jurassic magmatic rocks, partially reset AFT ages, and produced anomalous paleomagnetic pole positions in Jurassic rocks (e.g., Kreuzer et al. 1981; Elliot & Foland 1986; Fitzgerald & Gleadow 1988; Fleming et al. 1993; Schäfer & Olesch 1998). Profiles collected from peaks near the Priestley and Campbell Glaciers inland from Terra Nova Bay record Late Cretaceous exhumation (Balestrieri et al. 1994, 1997), indicating that exhumation there began at c. 85 Ma, and was c. 1.1 km in magnitude at Mt Nansen. Nearer the Terra Nova Bay coast, exhumation begun in the early Cenozoic is evident in the AFT data, although at Mt Monteagle AFT ages appear to have been reset by the mid-Cenozoic Meander Instrusives (age 38 Ma) (Balestrieri et al. 1997).

In SVL, Cretaceous exhumation has only been revealed in a few vertical profiles in the Kukri Hills (Fitzgerald 1995). Although the onset of this Cretaceous exhumation is poorly constrained, it began at c. 100 Ma and appears to continue through to the early Cenozoic in these few localities, but at quite a slow rate of c. 63 m/m.y. In contrast to the pattern of Cenozoic exhumation, which is always greater nearer the coast decreasing inland, this Cretaceous exhumation appears to be greater on the western (inland) side of the Kukri Hills, decreasing towards the coast until it is undetectable near the coast. Numerous vertical profiles from near the inland edge of the TMF clearly indicate the onset of early Cenozoic exhumation at c. 55 Ma (Gleadow & Fitzgerald 1987; Fitzgerald 1992), typically in the order of 4.5 km near the inland edge of the TMF.

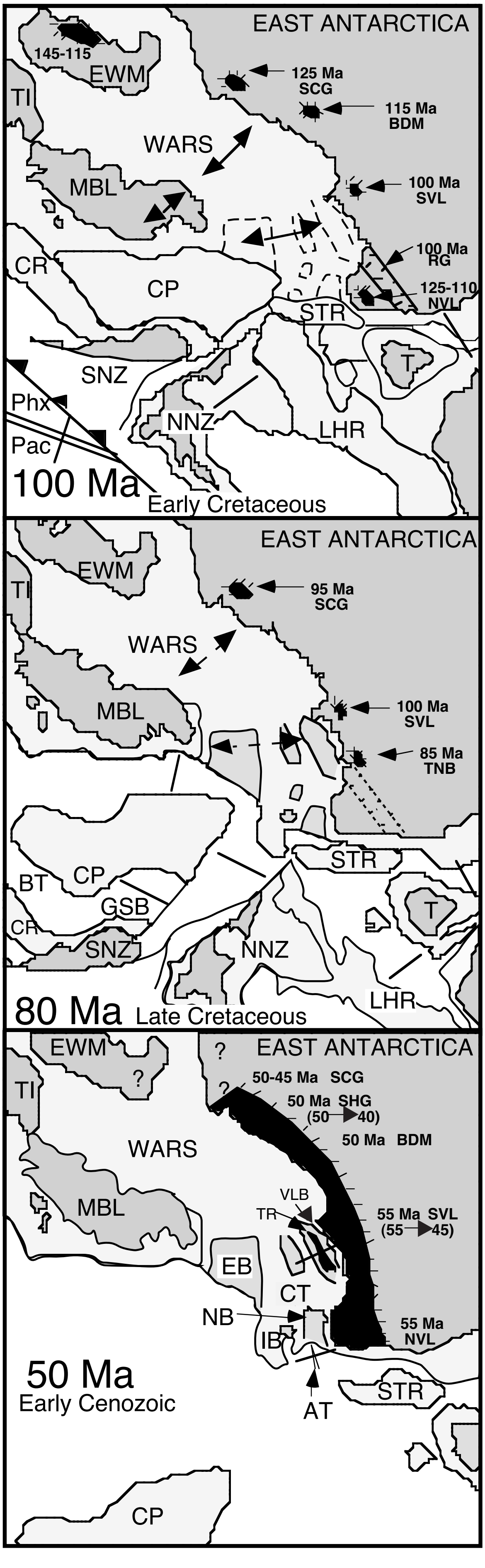
In the central TAM, studies have been undertaken in the Nimrod-Beardmore Glaciers region, near the Shackleton Glacier, and along the Scott Glacier. In the Beardmore region, Early Cretaceous exhumation is seen in a profile from Moody Nunatak indicating the onset of exhumation at c. 115 Ma. Along the Beardmore Glacier, exhumation begun in the early Cenozoic (50 Ma) dominates (Fitzgerald 1994), and is c. 5.5 km in magnitude at the Cloudmaker c. 70 km inland, but greater at the coast. There is no recorded Cretaceous exhumation in the Shackleton Glacier region. However, early Cenozoic exhumation begins at c. 50 Ma in this region close to the coast, but near the inland flank of the TMF it begins at c. 40 Ma (Miller et al. 1999). In the Scott Glacier region, periods of exhumation begin at c. 125, c. 95, and at c. 45 Ma (Fitzgerald & Stump 1997). Early and Late Cretaceous exhumation in the Scott Glacier region is in the order of 1–2 km each, depending on location, whereas exhumation since the early Cenozoic is c. 6.5 km at the coast decreasing to c. 3 km, c. 50 km inland where the frontal escarpment is reached.

Two subtle temporal trends in the TAM Cenozoic exhumation patterns are apparent. Firstly, exhumation appears to have begun earlier in Victoria Land than farther south along the TAM. The onset of early Cenozoic exhumation is c. 55 Ma in NVL and SVL, c. 50 Ma in the Beardmore Glacier area and the Shackleton Glacier, and c. 45 Ma in the Scott Glacier region. This trend is subtle within the precision of the data, and the location of the AFT age profile with respect to the coast may also influence the record of when early Cenozoic exhumation at a particular location is recorded. This is because the onset of exhumation may get younger inland, as well as southward. This inland younging trend may be apparent in the Shackleton Glacier when the break in slope is c. 50 Ma near the coast and c. 40 Ma at Mt Munson near the frontal escarpment, as defined by the northern wall of the Prince Olav Mountains (Miller et al. 1998). This same trend is also seen along the Kukri Hills in SVL (Fitzgerald 1995). This inland younging trend is most likely the result of escarpment retreat at a rate of c. 2 km/m.y., with the retreat rate apparently slowing dramatically c. 10 m.y. after the onset of early Cenozoic exhumation.

Relationship of exhumation in the TAM to regional tectonic events and models for the uplift of the TAM

The trends and patterns of Early Cretaceous, Late Cretaceous, and Cenozoic exhumation recorded along the TAM (Fig. 7) provide constraints and ideas for tectonic models that can then be tested using other techniques. The fact that there are distinct exhumation events through time within the TAM suggests that different mechanisms of uplift are responsible. To date, almost all studies have concerned themselves with mechanisms that attempt to explain or model the major period of exhumation of the TAM, correlative to the onset of uplift of the present-day TAM, begun in the early Cenozoic.

Early Cretaceous exhumation recorded in the Admiralty Mountains in NVL is most likely related to the initial stretching between Antarctica and Australia, which began at c. 125 Ma (Stagg & Willcox 1992). The Rennick thermal event at c. 100 Ma has been associated with the formation of the Rennick Graben as a failed rift (Roland & Tessensohn 1987; Fleming et al. 1993). Formation of the Rennick Graben is related to the initial seafloor spreading between Antarctica and Australia, which began at 95 ± 5 Ma (Veevers et al. 1990) or c. 95 Ma (Cande & Mutter 1982). Initial spreading between Australia and Antarctica was very slow (4.5 mm/yr) until c. 45–49 Ma (Cande & Mutter 1982; Veevers et al. 1990). Other Early Cretaceous exhumation events along the TAM are more problematic in their relationship to regional tectonics given their greater distance from Australia-Antarctic seafloor spreading. In addition, in both SVL (Fitzgerald 1995) and the Nimrod-Beardmore region (Fitzgerald 1994) Early Cretaceous exhumation appears to be greater in magnitude on the inland flank of the TAM. In the Nimrod-Beardmore region, the greater magnitude of exhumation on the inland flank of the TAM may be related to the apparent folding of the KES into a broad southerly plunging syncline. This syncline has a north–south-oriented axial trace through the Bowden Névé, with c. 2 km of relief on the flanks (Barrett et al. 1970; Barrett & Elliot 1973). The age of this syncline is most likely Cretaceous (Fitzgerald 1994). However, rather than being a syncline *sensu stricto*, the synformal shape of the KES (which was originally inferred from extrapolation of Beacon Supergroup stratigraphy) may simply appear to form a syncline due to Cretaceous uplift on its western (inland) flank followed by Cenozoic uplift on its eastern (coastward) flank. If this is the case, then subsequent erosion following block tilting (up to the west) along the inland flank of the TAM in the Early Cretaceous may be what is being recorded by the AFT profiles at Moody Nunatak and the Kukri Hills. What regional tectonic event caused this Cretaceous exhumation is difficult to ascertain. However, it is either related to the initial phase of the main period of extension in the WARS (more likely) or incipient extension to form a proto(?)-Wilkes Land Basin (less likely), possibly in association with Rennick faulting as an inboard right-stepping en-echelon rift system to the Rennick Graben. It must have been an en-echelon feature as the Rennick Graben most likely extended through to the Ross Sea before being cut-off by Cenozoic uplift of the TAM (Cooper & Davey 1985). The postulated Cretaceous proto-Wilkes Basin was then later overprinted by Cenozoic flexure of the eastern edge of the East Antarctic Craton that gives the Wilkes Basin today its characteristic shape (Stern & ten Brink 1989). This tectonic scenario is by



necessity speculative; however, the asymmetry of Early Cretaceous exhumation, while subtle, allows the suggestion, and it is a concept that is testable. Early Cretaceous exhumation at the southern end of the TAM in the Scott Glacier region may be related to rock uplift within the EWM (Fitzgerald & Stump 1997). Rock uplift there may have been caused by early rifting within the WARS between EWM and Marie Byrd Land as the EWM moved into its present position with respect to East Antarctica, which occurred sometime between 175 and 110 Ma (Grunow et al. 1991).

Late Cretaceous exhumation (i.e., exhumation initiated after 97 Ma) along the TAM and revealed in AFT profiles at Terra Nova Bay and the Scott Glacier is most likely related to the main phase of extension between East and West Antarctica. This main phase of extension has only been constrained as between 105 and 85 Ma (Lawver & Gahagan 1994, 1995). Initiation of extension between East and West Antarctica followed on from subduction of the Pacific-Phoenix spreading ridge along the Pacific margin outboard of New Zealand and then Marie Byrd Land. Subduction of the spreading ridge led to extension in New Zealand (Bradshaw 1989) and the switch from subduction-related magmatism in Marie Byrd Land to rift-related anorogenic magmatism at c. 100 Ma (Weaver et al. 1994). As a result of this ridge subduction, continued divergent motion between the Pacific and Antarctic plates resulted in New Zealand being transferred from the Antarctic plate to the Pacific plate as a new spreading ridge, the Antarctic-Pacific ridge, developed between New Zealand and Antarctica (Lawver & Gahagan 1994; Davey & Brancolini 1995). The subduction of the Phoenix-Pacific spreading centre along the Pacific margin of Gondwana leading to Late Cretaceous extension in the WARS is similar to the tectonic situation along the western coast of the United States in the Oligocene when the Pacific-Farallon spreading centre was subducted off the western United States leading to widespread extension in the Basin and Range Province (Atwater 1970; Coney 1987). It is thus possible that Late Cretaceous exhumation in the TAM is a result of upper plate block movement above low-angle detachment faults (Fitzgerald & Baldwin 1997). Fitzgerald et al. (1986) proposed that the mechanism for the main phase of uplift of the TAM beginning in the early Cenozoic was related to differential partition of strain into different lithospheric levels under the Ross Embayment and TAM, with strain partition accommodated on low-angle detachment faults. The Fitzgerald et

Fig. 7 Paleogeographic reconstructions of WARS and TAM region in the Early Cretaceous (100 Ma), Late Cretaceous (80 Ma), and early Cenozoic (50 Ma) (after Lawver et al. 1992), showing the tectonic evolution of the region. "Uplift" events accompanying recorded exhumation along the TAM are shown in a solid pattern with ages shown indicative of initiation of exhumation for that region. Cenozoic exhumation gets younger to the south along the TAM and also from the coast inland. Extension directions (after Davey & Brancolini 1995 and Cande et al. 1998) are shown by double headed arrows, dashed where extension had just ceased. See text for the relationship of exhumation along the TAM to regional tectonic events. Abbreviations as for Fig. 1 and 6, plus Phx, Phoenix plate; Pac, Pacific plate; T, Tasmania, STR, South Tasman Rise; BT, Bounty Trough; GSB, Great South Basin; IB, Inselin Bank; EWM, Ellsworth-Whitmore Mountains crustal block; RG, Rennick Graben; VLB, Victoria Land Graben; TR, Terror Rift; EB, Eastern Basin; NB, North Basin; CT, Central Trough; AT, Adare Trough.

al. (1986) model is invalid as the mechanism for Cenozoic uplift of the TAM because that model was developed when it was assumed that the timing of uplift of the TAM and the timing for the majority of extension in the Ross Embayment was synchronous. However, it is now known that the main period of extension in the Ross Embayment was in the Late Cretaceous rather than the early Cenozoic. In addition, that model also fails as "nett" rock uplift generated by detachment models for passive margin development (Lister et al. 1991) is in fact much less than that recorded in the TAM in the Cenozoic. However, "nett uplift" generated by detachment models is similar to the amount of exhumation recorded in the TAM during the Early or Late Cretaceous. Exhumation in the Late Cretaceous is only seen in two regions along the TAM (Terra Nova Bay and Scott Glacier region), possibly because appropriate crustal levels that contain low-temperature thermochronologic evidence for Cretaceous exhumation has been eroded off the front of the TAM by later Cenozoic exhumation. Cretaceous exhumation along the Kukri Hills in SVL, initiated at c. 100 Ma, may actually be more closely related to the Late Cretaceous extension within the WARS than Early Cretaceous exhumation observed in either NVL, the central TAM, or Scott Glacier. This is because exhumation in one profile in the Kukri Hills continues through the Late Cretaceous to the Early Cenozoic.

Since the major phase of the uplift of the TAM, initiated in the early Cenozoic, was discovered (Fitzgerald et al. 1986; Gleadow & Fitzgerald 1987), relating it to a regional tectonic event has always been difficult and unsatisfactory. The Eocene was a time of major plate reorganisation in the Indian Ocean. Spreading in the Tasman Sea ceased at c. 55 Ma, but rapid spreading between Antarctica and Australia did not begin until c. 45 Ma (Cande & Mutter 1982). Cenozoic extension occurred between East and West Antarctica, but was limited to the Terror Rift within the Victoria Land Basin on the western side of the Ross Embayment (Cooper et al. 1991). However, recent geophysical observations in the Adare Trough (Cande et al. 1998; Cande 1999), a prominent graben and fossil spreading centre c. 100 km northeast of Cape Adare in NVL, offer perhaps the first reasonable explanation for why the TAM were uplifted beginning in the early Cenozoic and why the uplift along the TAM appears to get younger to the south. Magnetic anomalies in the Adare Trough indicate spreading between East and West Antarctica occurred at least from 40 to 28 Ma, as synthetic models matched the observed magnetic anomalies if a spreading rate of 12 mm/yr between chron 18 (40 Ma) and chron 10 (28 Ma) is assumed. Spreading occurred east of the Inselin Bank in the Inselin Trough from 61 to 55 Ma, but jumped west of the Inselin Bank to the Adare Trough at chron 24 (53.5 Ma) (Cande et al. 1998; Cande 1999). Although magnetic anomalies in the Adare Trough have only been found back as far as 43 Ma, detailed ship-track coverage required to resolve the complete history of the Adare Trough, the Balleny Corridor, and related features is as yet far from complete (S. Cande, J. Stock pers comm. 1999). As there appears to be no significant fanning of anomalies in the Adare Trough, it is possible that a similar amount of extension, at least 144 km, also took place in the western Ross Sea.

Recent geophysical models for the uplift of the TAM all describe the uplift of the TAM as a lithospheric flexure of strong East Antarctic lithosphere (Stern & ten Brink 1989; van der Beek et al. 1994; Busetti et al. 1999). The Stern & ten Brink (1989) model for TAM uplift, plus its subsequent modifications (Bott & Stern 1992; ten Brink et al. 1993, 1997), model the TAM as a flexural uplift of strong East Antarctic lithosphere incorporating rift-related heat sources, erosion, and the Veining-Meinesz uplift effect to raise and maintain the TAM. In addition to the uplift driven by mantle density differences under the TAM versus the WARS, Bott & Stern (1992) demonstrated that tensional stresses resulting from low densities could also produce additional flexural displacement across a range-parallel master fault. The other geophysical models explain the uplift of the TAM as due to flexure of the strong East Antarctic lithosphere, but as a result of lithospheric necking rather than faulting (van der Beek et al. 1994; Busetti et al. 1999). The Stern and ten Brink flexural model requires a trigger that causes a lithospheric break along the TAM and hence the initiation of flexure of strong East Antarctic lithosphere against the weaker WARS lithosphere, in effect changing a continuous plate model into a broken plate model. Ten Brink et al. (1997) suggested that this trigger was a result of transtensional motion along the front of the TAM in the early Cenozoic. However, the kinematics of the TMF following initiation of uplift indicate that faulting there was dominantly normal-sense shear followed by dominantly dextral strike-slip oblique shear (Wilson 1995), and strike-slip motion did not begin along the TMF until c. 30 Ma (Salvini et al. 1997). Thus, as a trigger for initiation of early Cenozoic uplift, transtension in the early Cenozoic may not apply. It is suggested here that a more logical trigger for the initiation of the uplift of the TAM is the propagation southward of seafloor spreading from the Adare Trough into the continental crust under the western Ross Sea. The initiation of the exhumation of the TAM in Victoria Land as determined by AFT profiles (c. 55 Ma) occurs about the same time as spreading jumped from east of the Inselin Bank to west of the Inselin Bank (53.5 Ma) (S. Cande, J. Stock pers comm. 1999). Thus, it is suggested that as continental extension propagated southward, the timing of the initiation of uplift of the TAM also became younger. The relative timing of these events appears to fit the kinematic regime along the front of the TAM, which while admittedly still poorly constrained, nevertheless is consistent with the following scenario. Seafloor spreading in the Adare Trough, initiated in the early Cenozoic, propagated southward into the Victoria Land Basin and along the front of the TAM. This continental extension, localised to basins along the front of the TAM, provided a mechanism for lithospheric decoupling of East Antarctica from the WARS and a trigger for the initiation of uplift of the TAM as a lithospheric flexure. Initial uplift of the TAM was thus accompanied largely by rift-perpendicular extension, consistent with a dominantly normal sense, shear kinematic regime in the TMF. This stress regime subsequently changed to dextral strike-slip at c. 30 Ma. At about the same time as this apparent stress change, seafloor spreading stopped in the Adare Trough (c. 28 Ma) (Cande et al. 1998; Cande 1999).

LANDSCAPE EVOLUTION AND EROSION RATES IN THE DRY VALLEYS BLOCK OF THE TRANSANTARCTIC MOUNTAINS

The rates of erosion and landscape evolution of the Dry Valleys block in SVL has been addressed via a variety of approaches: thermochronology, geomorphology, the

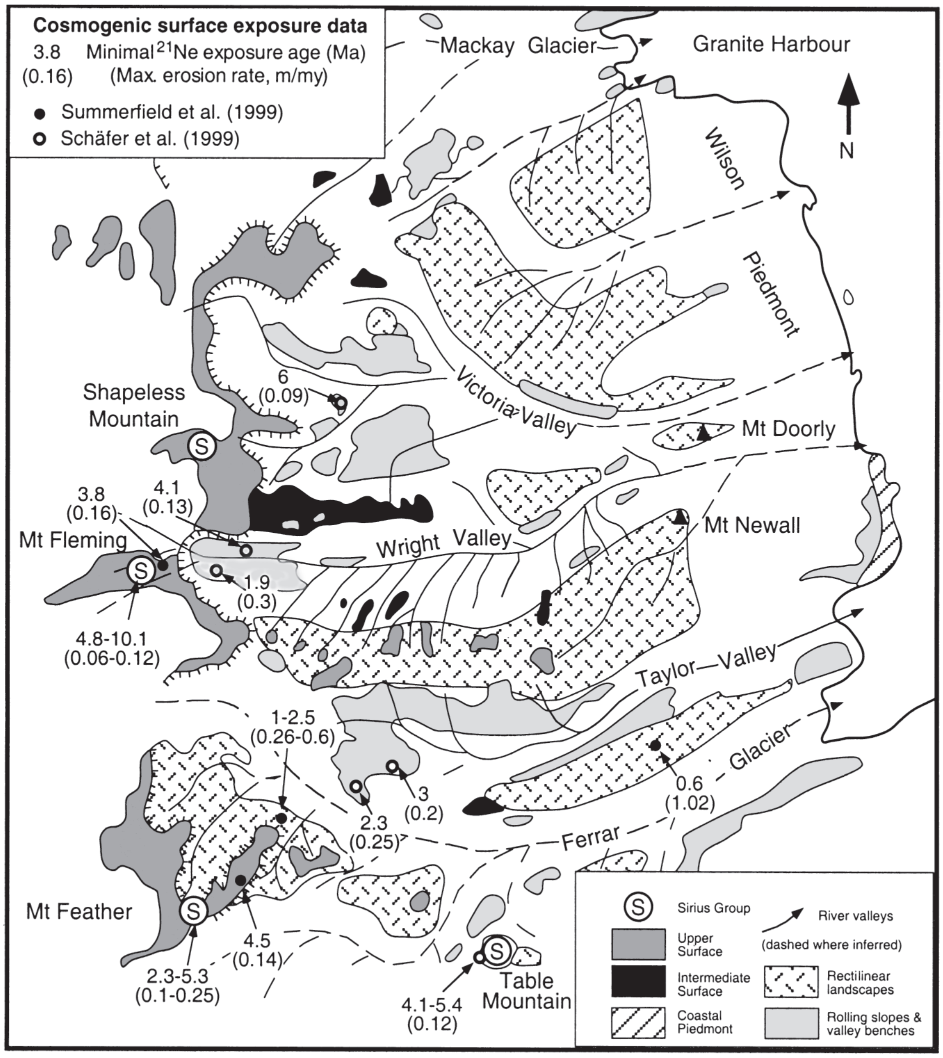


Fig. 8 Distribution of main landscape types together with the pattern of river valleys in the Dry Valleys block (modified from Sugden et al. 1995a) along with representative ^{21}Ne cosmogenic surface exposure ages (and erosion rates) from Summerfield et al. (1999) and Schäfer et al. (1999). Ages (and erosion rates) at each sample locality are either averaged from several samples where several ages are almost the same, or a range of ages or erosion rates is given. In general, exposure ages are older (with lower erosion rates) on the upper erosion surface of Sugden et al. (1995a) and younger on the rectilinear surfaces (with slightly higher erosion rates). The minimal ^{21}Ne exposure ages of Schäfer et al. (1999) on deposits of the Sirius Group indicate these deposits cannot be Pliocene in age.

presence and age of surficial ash deposits and Pliocene cinder cones, the preservation of Miocene ice, and surface exposure age dating. In the summary below, discussion of glacial deposits such as the Sirius Group and its role in constraining landscape evolution is avoided as much as possible, as the age of those deposits based on recycled diatoms is controversial and unsubstantiated (e.g., Harwood & Webb 1998; Miller & Mabin 1998; Stroeven et al. 1998).

The geomorphology of the Dry Valleys block is well presented in a paper by Sugden et al. (1995a). Landforms can be generally summarised as planation surfaces, rectilinear slopes of 26–36°, and valleys that were originally cut by streams and modified by rivers (Fig. 8). There are three main planation surfaces: (1) An upper surface with relief of 10–30 m contains inselbergs that are often flat topped and bounded by rectilinear slope. The upper surface has a prominent, steep, coastal-facing escarpment. (2) The intermediate surface contains inselbergs near this escarpment, which are erosional remnants of the upper surface, progressing into buttes with increasing distance from the escarpment. (3) The lower surface is mainly cut into granite nearer the coast. Sugden et al. (1995a) considered that the origin of the landforms was relict from an earlier fluvial environment but with subsequent minor glacial modification.

Constraining the age of the landforms in the Dry Valleys depends largely on the age of unweathered ash deposits that rest on desert pavement and in sand-wedge troughs under thin (1–2 cm) ventifact pavements. These ash deposits are most likely all from the adjacent Erebus Volcanic Province and give laser fusion $^{40}Ar/^{39}Ar$ ages on sanidine and volcanic glass of 4.33 ± 0.07 to 15.15 ± 0.02 Ma (Marchant et al. 1993, 1996). Ashes with ages of 7 and 11 Ma are also found on rectilinear slopes in Arena Valley, implying little slope development since emplacement of ash. The age of the ashes, their unweathered nature, and the surfaces they rest on indicate little landscape modification since they were deposited and the maintenance of a hyper-arid climate since at least 15 Ma (Marchant et al. 1993, 1996). In support of this interpretation, volcanic ash dated at 8.1 Ma has been found preserved in thin glacial till overlying glacier ice in Beacon Valley. In order for this ice to have remained undisturbed and to have avoided sublimation, stable cold polar conditions have existed since at least 8 Ma (Sugden et al. 1995b). Based upon the mutual truncation of rectilinear slopes truncating faults and TMF fault zones crosscutting valley benches, Sugden et al. (1995a) suggested that faulting across the TMF was coeval with downcutting to form the Dry Valleys. Thus, faulting across the TMF had ceased by the time that erosion to form the Dry Valleys was over at c. 15 Ma.

Cosmogenic surface age exposure dating is a relatively recent and powerful tool for constraining landscape evolution, especially in the Quaternary, sometimes back to Pliocene times, but rarely as far back as the Miocene. To determine a "surface age", zero erosion must be assumed since the exposure event. However, as this is not true in the majority of situations, a "surface exposure age" should be considered a minimal age only. In some circumstances, information about the rates of erosion can be obtained. In SVL, a number of studies have used exposure dating to "date" boulders on the Sirius Group, determine rates of erosion on that deposit, and to further constrain the glacial history of the region (Brook & Kurz 1993; Brook et al. 1993, 1995a,b; Ivy-Ochs et al. 1995; Bruno et al. 1997; Schäfer et al. 1999). Other studies have determined rates of erosion on the upper planation surface and the rectilinear slopes (Summerfield et al. 1999). Although exposure ages for the Sirius Group deposits have been found as old as 10 Ma, minimum ages of 2–6 Ma are more prevalent, but all data indicate low erosion rates of <0.25 m/m.y. Likewise, for the planation surfaces, erosion rates are low, <0.15 m/m.y., with typical minimal ^{21}Ne exposure ages of 4–5 Ma. Minimal ^{21}Ne ages for rectilinear slopes are younger, typically <2.5 Ma and with erosion rates of 1 m/m.y . The old minimum cosmogenic surface exposure ages on Sirius Group deposits indicate that it cannot be as young as late Pliocene as suggested by the age of recycled diatoms (e.g., Webb et al. 1984; McKelvey et al. 1991). It has been suggested that the apparent young appearance of the TAM suggests dramatic surface uplift since the Pliocene, with surface uplift rates in the order of 1 km/m.y. (Behrendt & Cooper 1991). However, Wilch et al. (1993a, b), by dating cinder cones, determined that surface uplift at the mouth of the Taylor Valley since c. 2.5 Ma was limited to <300 m. In addition, Bruno et al. (1997) found that the variation in production rate of cosmogenic nuclides with elevation within the TAM suggest minimal surface uplift in the last few million years.

Mean surface elevation and relief present in the TAM before the onset of early Cenozoic in the TAM is poorly constrained, thus making estimates of rock uplift or surface uplift since then also poorly constrained. Estimates of pre-early Cenozoic mean surface elevation have ranged from 500 m (Gleadow & Fitzgerald 1987; Fitzgerald 1992) to 1200 m (Sugden et al. 1995a). In Fig. 4, an estimate of 700 m (ten Brink et al. 1997) is used. Constraints on the amount of exhumation since the early Cenozoic are more reliable as they are based on AFTT and do not require constraining the early Cenozoic land surface elevation. Rates of exhumation in the Dry Valleys have been estimated from the AFT data, which indicate an average rate since the onset of early Cenozoic exhumation of c. 100 m/m.y. (e.g., Gleadow & Fitzgerald 1987). This average exhumation rate has been determined from sites such as Mt Doorly and Mt England just east of the inland edge of the TMF. Within this average rate, episodes of faster and slower exhumation, beyond either the precision of AFTT or younger than that time period covered by the AFT data, are to be expected. Rates of exhumation also vary across the TAM, decreasing inland as the overall amount of rock uplift decreases. In detail, exhumation rates also vary depending on the location and other factors such as the presence of valleys and ridges. The slope of the AFT profile under a break in slope gives an apparent exhumation rate, but only if advection of isotherms during exhumation has not modified the slope of the profile. The effects of advection cause one to initially underestimate the true exhumation rate soon after the initiation of exhumation (Peacock 1995) and then overestimate the true exhumation rate once steady state is reached (Stüwe et al. 1994; Mancktelow & Grasemann 1997). This is not normally a problem in the TAM as the "rapid exhumation" under the break in slope is actually quite slow, <300 m/m.y., so that advection has not modified the slope of the profile (Brown & Summerfield 1997). Thus, at Mt Barnes, the average rate of exhumation since 55 Ma is 82 m/m.y. (Fig. 9), but the measured rate from 55–42 Ma from least squares regression is c. 120 m/m.y. Modelling of confined FT lengths from profiles like Mt Doorly also indicate that the initial 10–15 m.y. of exhumation was likewise more rapid, up to twice as fast as the average (Fitzgerald 1992, 1994). However, beyond this, the AFTT gives little control on exhumation rate. Information on erosion rate using cosmogenic surface exposure ages only extends back to c. 10 Ma, and by inference from geomorphic arguments and the presence of ash deposits, to c. 15 Ma. Attempts have been made to extend the cosmogenically determined erosion rate back to 20 Ma by assuming "uplift rates" (Schäfer et al. 1999), but these assumed uplift rates are poorly constrained.

CONCLUSION

On a first-order scale, the landscape evolution of Antarctica has been controlled by tectonics associated with the breakup of Gondwana and the dispersal of continents as Antarctica moved to a polar location. Continental dispersal led to the development of the circum-polar current (e.g., Kennett 1978; Lawver et al. 1992), the establishment of continental glaciation with ice sheets growing to the continent margin in the early Oligocene, development of the East Antarctic Ice Sheet in the Miocene, and the present-day cold polar hyper-arid environment (e.g., Barrett 1996 and references therein). Along the TAM, the landscape evolution, especially in Cenozoic times, is a result of the interaction of tectonic and climatic influences. Tracking the timing and patterns of exhumation along the TAM using thermochronology, while still relatively rudimentary in scope and coverage, has allowed the tectonics of the region to be better delineated, especially the relationship of the exhumation history of the TAM with regional tectonic events. Exhumation events in the Early Cretaceous can be related to either initial stretching between Australia and Antarctica or at the other end of the TAM to tectonics probably involving the translation of West Antarctic microplates and early extension within the WARS, although these later events are still unclear. In addition, there are subtle Cretaceous exhumation patterns that may suggest possible extension within the Wilkes subglacial basin lying along the inland flank of the TAM, or early extension within the WARS. The Late Cretaceous exhumation events in the TAM are most likely related to the main phase of East–West Antarctic extension, possibly due to block tilting associated with extension along low-angle detachments (Fitzgerald & Baldwin 1997).

Whereas in the past, the relationship of early Cenozoic uplift of the TAM to regional tectonic events has been problematic, the discovery of Eocene–Oligocene seafloor spreading in the Adare Trough (Cande et al. 1998), and its possible extrapolation south into continental crust along the

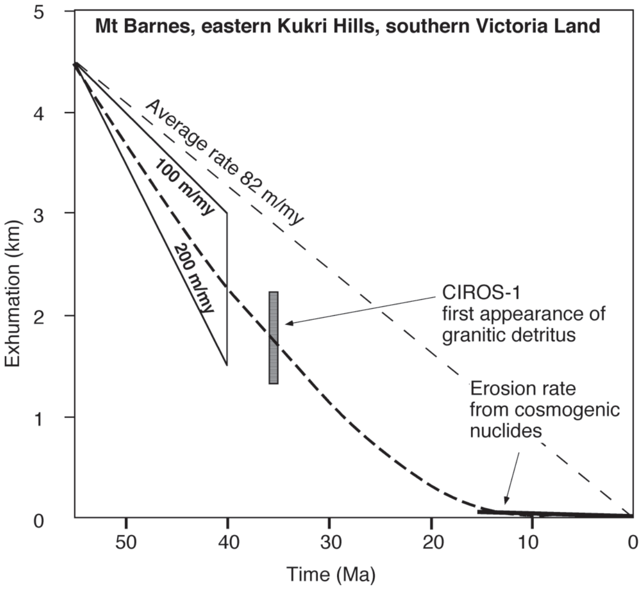


Fig. 9 Generalised exhumation curve for the Dry Valleys incorporating AFT data from Mt Barnes, CIROS-1 drillhole (noting the first appearance of granitic detritus; Barrett et al. 1989), and erosion rates determined from cosmogenic surface dating (e.g., Ivy-Ochs et al. 1995; Schäfer et al. 1999; Summerfield et al. 1999).

front of the TAM, provides a trigger for flexural uplift of the TAM. Subtle trends within the Cenozoic exhumation patterns along the TAM suggest that the initiation of uplift gets younger to the south, probably associated with the propagation of extension southward. Cenozoic exhumation patterns across the TAM suggest that the initiation of uplift gets younger inland, most likely due to escarpment retreat. Kinematics within the TMF are dominantly normal sense shear from the onset of uplift until c. 30 Ma. At c. 30 Ma, the stress regime changes to dextral strike-slip (Wilson 1995), which is about the same time as seafloor spreading (and hence the suggested continental extension to the south along the front of the TAM) ceases in the Adare Trough.

The average rate of exhumation since the early Cenozoic within the TAM, in the order of 100 m/m.y., is very slow compared to most tectonic regimes. However, the initial rate of exhumation was probably about twice that rate for the first 10–15 m.y. By c. 36 Ma, the TAM had undergone enough erosion to expose granitic basement, which was then eroded and deposited off Butter Point in SVL where it was cored in the CIROS-1 drillhole (Barrett et al. 1989). Between c. 40 and 15 Ma, we have little information about the exhumation history other than that inferred from offshore drillholes, but overall the rate of exhumation probably started to slow slightly as the climatic regime cooled and continental glaciation started to develop. However, at c. 15 Ma or earlier, the rate of erosion dropped dramatically as a hyper-arid cold polar environment developed. The onset of early Cenozoic uplift of the TAM (plus an associated drop in base-level in the adjacent Victoria Land Basin) was followed by escarpment retreat, formation of planation surfaces, and downcutting by fluvial systems (Sugden et al. 1995a). The present-day landscape has relict characteristics of this fluvial environment, with relatively minor glacial modification and extremely low erosion rates since c. 15 Ma. Thus, landscape development within the Dry Valleys as well as faulting across the TMF was largely complete by c. 15 Ma (Sugden et al. 1995a). However, tectonic activity is still ongoing within the TMF, as shown by the widespread McMurdo Group alkaline volcanism, c. 400 m of subsidence in the Miocene at the mouth of the Taylor Valley followed by c. 300 m of surface uplift in the Pliocene, and minor faulting in the Quaternary (Jones 1997).

ACKNOWLEDGMENTS

I thank the organising committee for inviting me to submit this synthesis. Thanks also to Peter Barrett, Terry Wilson, Christine Siddoway, Suzanne Baldwin, Scott Miller, Joann Stock, and Steve Cande for helpful discussions on various parts of this paper. Jamie Shulmeister is thanked for his review of this paper. Support is from the National Science Foundation Office of Polar Programs.

REFERENCES

Atwater, T. 1970: Implications of plate tectonics for the Cenozoic tectonic evolution of western North America. *Bulletin of the Geological Society of America 81*: 3513–3536.

Baldwin, S. L.; Siddoway, C.; Fitzgerald, P. G.; Fanning, C. M.; Luyendyk, B. P. 2001: Marie Byrd Land mylonites record onset of continental extension in West Antarctica. *Eos, Transactions, American Geophysical Union 82 (47)*: F1234.

Balestrieri, M. L.; Bigazzi, G.; Ghezzo, C.; Lombardo, B.1994: Fission track dating of apatites from the Granite Harbour Instrusive suite and uplift-denduation history of the Transantarctic Mountains in the area between the Mariner and David Glaciers (northern Victoria Land, Antarctica). *Terra Antartica 1:* 82–87.

Balestrieri, M. L.; Bigazzi, G.; Ghezzo, C. 1997: Uplift–denudation of the Transantarctic Mountains between the David and the Mariner Glaciers, Northern Victoria Land (Antarctica): constraints by apatite fission-track analysis. *In*: Ricci, C. A. *ed*. The Antarctic region: geological evolution and processes. Siena, Terra Antartica Publication. Pp. 547–554.

Barrett, P. J. 1965: Geology of the area between the Axel Heiberg and Shackleton Glaciers, Queen Maud Mountains, Antarctica. *New Zealand Journal of Geology and Geophysics 8*: 344–370.

Barrett, P. J. 1979: Proposed drilling in McMurdo Sound—1979. *Memoir of the National Institute of Polar Research, Special Issue 13*: 231–239.

Barrett, P. J. 1981: History of the Ross Sea region during the deposition of the Beacon Supergroup 400–180 million years ago. *Journal of the Royal Society of New Zealand 11*: 447–458.

Barrett, P. J. 1991: The Devonian to Triassic Beacon Supergroup of the Transantarctic Mountains and correlatives in other parts of Antarctica. *In*: Tingey, R. J. *ed*. The geology of Antarctica. Oxford Monographs on Geology and Geophysics. Oxford, Clarendon Press. Pp. 120–152.

Barrett, P. J. 1996: Antarctic paleoenvironment through Cenozoic times—a review. *Terra Antartica 3*: 103–119.

Barrett, P. J.; Elliot, D. H. 1973: Reconnaissance geologic map of the Buckley Island Quadrangle, Transantarctic Mountains, Antarctica. Reston, Va., United States Geological Survey.

Barrett, P. J.; Lindsay, J. F.; Gunner, J. 1970: Reconnaissance geologic map of the Mount Rabot Quadrangle, Transantarctic Mountains, Antarctica. Reston, Va., United States Geological Survey.

Barrett, P. J.; Hambrey, M. J.; Harwood, D. M.; Pyne, A. R.; Webb, P. N. 1989: Synthesis. *In*: Barrett, P. J. *ed*. Antarctic Cenozoic history from the CIROS-1 drillhole, McMurdo Sound. *Department of Scientific and Industrial Research Bulletin 245*: 241–252.

Barrett, P. J.; Henrys, S. A.; Bartek, L. R.; Brancolini, G.; Busetti, M.; Davey, F. J.; Hannah, M. J.; Pyne, A. R. 1995: Geology of the margin of the Victoria Land Basin off Cape Roberts, southwest Ross Sea. *In*: Cooper, A. K.; Barker, P. F.; Brancolini, G. *ed*. Geology and seismic stratigraphy of the Antarctic margin. *Antarctic Research Series Vol. 68*. Washington D.C., American Geophysical Union. Pp. 183–207.

Behrendt, J. C.; Cooper, A. K. 1991: Evidence of rapid Cenozoic uplift of the shoulder escarpment of the Cenozoic West Antarctic rift system and a speculation on possible climate forcing. *Geology 19*: 315–319.

Bell, R. E.; Blankenship, D. D.; Finn, C. A.; Morse, D. L.; Scambos, T. A.; Brozena, J. M.; Hodge, S. M. 1998: Influence of subglacial geology on the onset of a West Antarctic ice stream from aerogeophysical observations. *Nature 394*: 58–62.

Bentley, C. R. 1991: Configuration and structure of the subglacial crust. *In*: Tingey, R. J. *ed*. The geology of Antarctica. Oxford, Clarendon. Pp. 335–364.

Bott, M. H. P.; Stern, T. A. 1992: Finite element analysis of Transantarctic Mountain uplift and coeval subsidence in the Ross Embayment. *Tectonophysics 201*: 341–356.

Bradshaw, J. D. 1989: Cretaceous geotectonic patterns in the New Zealand region. *Tectonics 8*: 803–820.

Brook, E. J.; Kurz, M. D. 1993: Surface-exposure chronology using in situ cosmogenic ^{3}He in Antarctic quartz sandstone boulders. *Quaternary Research 39*: 1–10.

Brook, E. J.; Kurz, M. D.; Ackert, R. P.; Denton, G. H.; Brown, E. T.; Raisbeck, G.; Yiou, F. 1993: Chronology of Taylor Glacier advances in Arena Valley, Antarctica using in situ cosmogenic ^{3}He and ^{10}Be. *Quaternary Research 39*: 11–23.

Brook, E. J.; Brown, E. T.; Kurz, M. D.; Ackert, R. P.; Raisbeck, G. M.; Yiou, F. 1995a: Constraints on age, erosion, and uplift of Neogene glacial deposits in the Transantarctic Mountains determined from in situ cosmogenic ^{10}Be and ^{26}Al. *Geology 23*: 1063–1066.

Brook, E. J.; Kurz, M. D.; Ackert, R. P.; Raisbeck, G.; Yiou, F. 1995b: Cosmogenic nuclide exposure ages and glacial history of late Quaternary Ross Sea drift in McMurdo Sound, Antarctica. *Earth and Planetary Science Letters 131*: 41–56.

Brown, R. W. 1991: Backstacking apatite fission-track "stratigraphy": A method for resolving the erosional and isostatic rebound components of tectonic uplift histories. *Geology 19*: 74–77.

Brown, R. W.; Summerfield, M. A. 1997: Some uncertainties in the derivation of rates of denudation from thermochronologic data. *Earth Surface Processes and Landforms 22*: 239–248.

Bruno, L. A.; Baur, H.; Graf, T.; Schluchter, C.; Signer, P.; Wieler, R. 1997: Dating of Sirius Group tillites in the Antarctic Dry Valleys with cosmogenic ^{3}He and ^{21}Ne. *Earth and Planetary Science Letters 147*: 37–54.

Busetti, M.; Spadini, G.; Van der Wateren, F.; Cloetingh, S.; Zanolla, C. 1999: Kinematic modelling of the West Antarctic rift system, Ross Sea, Antarctica. *Global and Planetary Change 23*: 79–103.

Cande, S. 1999: Seafloor spreading constraints on Cenozoic motion between East and West Antarctica. *In*: Baldwin, S. L.; Lister, G. S. *ed*. Mid-Cretaceous to Recent plate boundary processes in the southwest Pacific. Penrose Conference, Arthur's Pass, South Island, New Zealand. P. 19.

Cande, S. C.; Mutter, J. C. 1982: A revised identification of the oldest sea-floor spreading anomalies between Australia and Antarctica. *Earth and Planetary Science Letters 58*: 151–160.

Cande, S.; Stock, J.; Mueller, D.; Ishihara, T.; Tikku, A. 1998: Seafloor spreading constraints on Cenozoic motion between East and West Antarctica. *EOS, Transactions of the American Geophysical Union 79*: F906.

Cape Roberts Science Team 2000: Studies from the Cape Roberts Project, Ross Sea Antarctica. Initial report on CRP-3. Siena, Italy, Terra Antartica Publication. 211 p.

Capponi, G.; Messiga, B.; Piccardo, G. B.; Scambelluri, M.; Traverso, G.; Vannucci, R. 1990: Metamorphic assemblages in layered amphibolites and micaschists from the Dessent Formation (Mountaineer Range, Antarctica). *Memorie della Societa Geologica Italiana 43*: 87–95.

Collinson, J. W. 1991: The palaeo-Pacific margin as seen from East Antarctica. *In*: Thomson, M. R. A.; Crame, J. A.; Thomson, J. W. *ed*. Geological evolution of Antarctica. Cambridge, Cambridge University Press. Pp. 199–204.

Coney, P. J. 1987: The regional tectonic setting and possible causes of Cenozoic extension in the North American Cordillera. *In*: Coward, M. P.; Dewey, J. F.; Hancock, P. L. *ed*. Continental extensional tectonics. Special Publication, Geological Society. Pp. 177–186.

Cooper, A. K.; Davey, F. J. 1985: Episodic rifting of the Phanerozoic rocks of the Victoria Land Basin, western Ross Sea, Antarctica. *Science 229*: 1085–1087.

Cooper, A. K.; Davey, F. J.; Behrendt, J. C. 1987: Seismic stratigraphy and structure of the Victoria Land Basin, western Ross Sea, Antarctica. *In*: Cooper, A. K.; Davey, F. J. *ed*. The Antarctic continental margin. Geology and geophysics of the Western Ross Sea. Vol. 5B. Earth Science Series. Houston, Texas, Circum-Pacific Council for Energy and Mineral Resources. Pp. 27–76.

Cooper, A. K.; Davey, F. J.; Hinz, K. 1991: Crustal extension and origin of sedimentary basins beneath the Ross Sea and Ross Ice Shelf, Antarctica. *In*: Thomson, M. R. A.; Crame, J. A.; Thomson, J. W. *ed*. Geological evolution of Antarctica. Cambridge, Cambridge University Press. Pp. 285–292.

Cox, K. G. 1988: The Karoo Province. *In*: MacDougall, J. D. *ed*. Continental flood basalts. Dordrecht, Kluwer Academic Publishers. Pp. 239–271.

Curtis, M. L.; Storey, B. C. 1996: A review of geological constraints on the pre-breakup position of the Ellsworth Mountains within Gondwana: implications for Weddell Sea evolution. *In*: Storey, B. C.; King, E. C.; Livermore, R. A. *ed*. Weddell Sea tectonics and Gondwana breakup. Special Publication, London, Geological Society. Pp. 11–30.

Dalziel, I. W. D. 1992: Antarctica: A tale of two supercontinents. *Annual Review of Earth and Planetary Sciences 20*: 501–526.

Dalziel, I. W. D.; Elliot, D. H. 1982: West Antarctica: problem child of Gondwana. *Tectonics 1*: 3–19.

Dalziel, I. W. D.; Lawver, L. A.; Murphy, J. B. 1999: Plume, orogenesis, supercontinental fragmentation and ice-sheets. *In*: Skinner, D. N. B. *ed*. 8th Symposium on Antarctic Earth Sciences, 5–9 July 1999, programme and abstracts. Wellington, New Zealand. Victoria University of Wellington. P. 78.

Damaske, D.; Behrendt, J.; McCafferty, A.; Saltus, R.; Meyer, U. 1994: Transfer faults in the western Ross Sea: new evidence from the McMurdo Sound/Ross Ice Shelf aeromagnetic survey (GANOVEX VI). *Antarctic Science 6*: 359–364.

Davey, F. J.; Brancolini, G. 1995: The Late Mesozoic and Cenozoic structural setting of the Ross Sea region. *In*: Geology and stratigraphy of the Antarctic margin. *Antarctic Research Series Vol. 68*. Washington D.C., American Geophysical Union. Pp. 167–182.

De Wit, M.; Jeffery, M.; Bergh, H.; Nicholatsen, L. 1988: Geological map of sectors of Gondwana. American Association of Petroleum Geologists and University of Witwatersrand.

Divenere, V. J.; Kent, D. V.; Dalziel, I. W. D. 1994: Mid-Cretaceous paleomagnetic results from Marie Byrd Land, West Antarctica: a test of post-100 Ma relative motion between East and West Antarctica. *Journal of Geophysical Research 99*: 15115–15139.

Divenere, V.; Kent, D. V.; Dalziel, I. W. D. 1995: Summary of paleomagnetic results from West Antarctica: implications for the tectonic evolution of the Pacific margin of Gondwana during the Mesozoic. *In*: Storey, B. C.; King, E. C.; Livermore, R. A. *ed*. Weddell Sea tectonics and Gondwana breakup. Special Publication, London, Geological Society. Pp. 31–43.

Divenere, V.; Kent, D. V.; Dalziel, I. W. D. 1996: Early Cretaceous paleomagnetic results from Marie Byrd Land, West Antarctica: implications for the Weddellia collage of crustal blocks. *Journal of Geophysical Research 100*: 8133–8152.

Elliot, D. H. 1975: Tectonics of Antarctica: a review. *American Journal of Science 275*: 45–106.

Elliot, D. H. 1992: Jurassic magmatism and tectonism associated with Gondwanaland breakup: an Antarctic perspective. *In*: Storey, B. C.; Alabaster, T.; Pankhurst, R. J. *ed*. Magmatism and the causes of continental breakup. Special Publication, London, Geological Society. Pp. 165–184.

Elliot, D. H.; Foland, K. A. 1986: K-Ar age determinations of the Kirkpatrick Basalt, Mesa Range. *In*: Stump, E. *ed*. Geological investigations in northern Victoria Land. *Antarctic Research Series Vol. 46*. Washington D.C., American Geophysical Union. Pp. 279–288.

England, P.; Molnar, P. 1990: Surface uplift, uplift of rocks, and exhumation of rocks. *Geology 18*: 1173–1177.

Findlay, R. H.; Skinner, D. N. B.; Craw, D. 1984: Lithostratigraphy and structure of the Koettlitz Group, McMurdo Sound, Antarctica. *New Zealand Journal of Geology and Geophysics 27*: 513–536.

Fitzgerald, P. G. 1992: The Transantarctic Mountains of southern Victoria Land: the application of apatite fission track analysis to a rift shoulder uplift. *Tectonics 11*: 634–662.

Fitzgerald, P. G. 1994: Thermochronologic constraints on post-Paleozoic tectonic evolution of the central Transantarctic Mountains, Antarctica. *Tectonics 13*: 818–836.

Fitzgerald, P. G. 1995: Cretaceous and Cenozoic exhumation of the Transantarctic Mountains: evidence from the Kukri Hills of southern Victoria Land compared to fission track data from gneiss at DSDP site 270. *In*: Ricci, C. A. *ed*. Seventh International Symposium on Antarctic Earth Sciences. Siena, Italy. P. 133.

Fitzgerald, P. G.; Baldwin, S. L. 1997: Detachment fault model for the evolution of the Ross Embayment. *In*: Ricci, C. A. *ed*. The Antarctic region: geological evolution and processes. Siena, Terra Antartica Publication. Pp. 555–564.

Fitzgerald, P. G.; Gleadow, A. J. W. 1988: Fission track geochronology, tectonics and structure of the Transantarctic Mountains in northern Victoria Land, Antarctica. *Isotope Geoscience 73*: 169–198.

Fitzgerald, P. G.; Gleadow, A. J. W. 1990: New approaches in fission track geochronology as a tectonic tool: examples from the Transantarctic Mountains. *Nuclear Tracks 17*: 351–357.

Fitzgerald, P. G.; Stump, E. 1997: Cretaceous and Cenozoic episodic denudation of the Transantarctic Mountains, Antarctica: new constraints from apatite fission track thermochronology in the Scott Glacier region. *Journal of Geophysical Research 102*: 7747–7765.

Fitzgerald, P. G.; Sandiford, M.; Barrett, P. J.; Gleadow, A. J. W. 1986: Asymmetric extension associated with uplift and subsidence of the Transantarctic Mountains and Ross Embayment. *Earth and Planetary Science Letters 81*: 67–78.

Fitzgerald, P. G.; Sorkhabi, R. B.; Redfield, T. F.; Stump, E. 1995: Uplift and denudation of the central Alaska Range: a case study in the use of apatite fission-track thermochronology to determine absolute uplift parameters. *Journal of Geophysical Research 100*: 20175–20191.

Fleming, T. H.; Elliot, D. H.; Foland, K. A.; Jones, L. M.; Bowman, J. R. 1993: Disturbance of Rb-Sr and K-Ar isotopic systems in the Kirkpatrick Basalt, north Victoria Land: implications for middle Cretaceous tectonism. *In*: Findlay, R. H.; Unrug, R.; Banks, M. R.; Veevers, J. J. *ed*. Gondwana eight: Assembly, evolution and dispersal. Rotterham, A. A. Balkema.

Ford, A. B.; Barrett, P. J. 1975: Basement rocks of the south-central Ross Sea, site 270, DSDP Leg 28. *In*: Hayes, D. E.; Frakes, L. A. *ed*. *Initial Reports of the Deep Sea Drilling Project 28*. Washington D.C., U.S. Government Printing Office. Pp. 861–868.

Gleadow, A. J. W.; Fitzgerald, P. G. 1987: Uplift history and structure of the Transantarctic Mountains: new evidence from fission track dating of basement apatites in the Dry Valleys area, southern Victoria Land. *Earth and Planetary Science Letters 82*: 1–14.

Gould, L. M. 1935: Structure of the Queen Maud Mountains, Antarctica. *Geological Society of America Bulletin 46*: 973–984.

Grunow, A. M.; Kent, D. V.; Dalziel, I. W. D. 1987: Evolution of the Weddell Sea Basin, new paleomagnetic constraints. *Earth and Planetary Science Letters 86*: 16–26.

Grunow, A. M.; Kent, D. V.; Dalziel, I. W. D. 1991: New paleomagnetic data from Thurston Island: implications for the tectonics of West Antarctica and the Weddell Sea. *Journal of Geophysical Research 96*: 17935–17954.

Gunn, B. M.; Warren, G. 1962: Geology of Victoria Land between the Mawson and Mulock Glaciers, Antarctica. *New Zealand Geological Survey Bulletin 71*. 157 p.

Gust, D. A.; Biddle, K. T.; Phelps, D. W.; Uliana, M. A. 1985: Associated Middle to Late Jurassic vulcanism and extension in southern South America. *Tectonophysics 116*: 223–253.

Harwood, D. M.; Webb, P. N. 1998: Glacial transport of diatoms in the Antarctic Sirius Group: Pliocene refrigerator. *GSA Today 8*: 1–8.

Heimann, A.; Fleming, T. H.; Elliot, D. H.; Foland, K. A. 1994: A short interval of Jurassic continental flood basalt volcanism in Antarctica as demonstrated by ^{40}Ar/^{39}Ar geochronology. *Earth and Planetary Science Letters 121*: 19–41.

Hergt, J. M.; Chappell, M. T.; McCullock, M.; MacDougall, I.; Chivas, A. R. 1989: Geochemical and isotopic constraints on the origin of the Jurassic dolerites of Tasmania. *Journal of Petrology 30*: 841–883.

Hooper, P. R.; Rehacek, J.; Duncam, R. A. 1993: The basalts of Lesotho, Karoo Province, Southern Africa (abstract). *EOS, Transactions of the American Geophysical Union 74*: 553.

Ivy-Ochs, S.; Schlüchter, C.; Kunik, P. W.; Dittrich-Hannen, B.; Beer, J. 1995: Minimum ^{10}Be exposure ages of the early Pliocene for the Table Mountain plateau and the Sirius Group at Mount Fleming, Dry Valleys, Antarctica. *Geology 23*: 1007–1010.

Jones, S. 1997: Late Quaternary faulting and neotectonics, south Victoria Land, Antarctica. *Journal of the Geological Society of London*: 645–652.

Kellogg, K. S.; Rowley, P. D. 1989: Structural geology and tectonics of the Orville Coast region, southern Antarctic Peninsula, Antarctica. U.S. Geological Survey. 25 p.

Kennett, J. P. 1978: The development of planktonic biogeography in the Southern Ocean during the Cenozoic. *Marine Micropaleontology 3*: 301–345.

Kreuzer, H.; Hohndorf, A.; Lenz, H.; Vetter, U.; Tessensohn, F.; Muller, P.; Jordan, H.; Harre, W.; Besang, C. 1981: K-Ar and Rb-Sr dating of igneous rocks from north Victoria Land. *Geologisches Jahrbuch B 41*: 267–273.

Lawver, L. A.; Gahagan, L. M. 1994: Constraints on the timing of extension in the Ross Sea region. *Terra Antartica 1*: 545–552.

Lawver, L. A.; Gahagan, L. M. 1995: Revised plate reconstructions and implications for extension in the Ross Sea region (abstract). 7th International Symposium on Antarctic Earth Science, Siena, Italy. P. 239.

Lawver, L. A.; Scotese, C. R. 1987: A revised reconstruction of Gondwanaland. *In*: MacKenzie, G. *ed*. Gondwana six: Structure, tectonics and geophysics. Washington D.C., American Geophysical Union. Pp. 17–23.

Lawver, L. A.; Royer, J.-Y.; Sandwell, D. A.; Scotese, C. T. 1991: Evolution of the Antarctic continental margins. *In*: Thomson, M. R. A.; Crame, J. A.; Thomson, J. W. *ed*. Geological evolution of Antarctica. Cambridge, Cambridge University Press. Pp. 533–539.

Lawver, L. A.; Gahagan, L. M.; Coffin, M. F. 1992: The Development of paleoseaways around Antarctica. *In*: The Antarctic paleoenvironment: a perspective on global change. *Antarctic Research Series Vol. 56*. Washington D.C., American Geophysical Union. Pp. 7–30.

Lawver, L. A.; Gahagan, L. M.; Dalziel, I. W. D. 1998: A tight fit—Early Mesozoic Gondwana: a plate tectonic perspective. *In*: Origin and evolution of continents. Memoirs of the National Institute of Polar Research, Special Issue. Tokyo, National Institute of Polar Research. Pp. 214–229.

LeMasurier, W. E.; Thomson, J. W. 1990: Volcanoes of the Antarctic plate and Southern Oceans. *Antarctic Research Series Vol. 48*. Washington D.C., American Geophysical Union.

Lister, G. S.; Etheridge, M. A.; Symonds, P. A. 1991: Detachment models for the formation of passive continental margins. *Tectonics 10*: 1038–1064.

Luyendyk, B.; Cisowski, S.; Smith, C.; Richard, S.; Kimbrough, D. 1996: Paleomagnetic study of the northern Ford Ranges, western Marie Byrd Land, West Antarctica: motion between West and East Antarctica. *Tectonics 15*: 122–141.

McGinnis, L. D.; Bowen, R. H.; Erickson, J. M.; Aldred, B. J.; Kreamer, J. L. 1985: East-West Antarctic boundary in McMurdo Sound. *Tectonophysics 14*: 341–356.

McKelvey, B. C.; Webb, P. N. 1962: Geological investigations in southern Victoria Land, Antarctica. Part 3: Geology of Wright Valley. *New Zealand Journal of Geology and Geophysics 5*: 143–162.

McKelvey, B. C.; Webb, P. N.; Harwood, D. M.; Mabin, M. C. G. 1991: The Dominion Range Sirius Group: a record of the late Pliocene-early Pliestocene Beardmore Glacier. *In*: Thomson, M. R. A.; Crame, J. A.; Thomson, J. W. *ed*. Geological evolution of Antarctica. Cambridge, Cambridge University Press. Pp. 675–682.

Mancktelow, N. S.; Grasemann, B. 1997: Time-dependent effects of heat advection and topography on cooling histories during erosion. *Tectonophysics 270*: 167–195.

Marchant, D. R.; Swisher, C. C.; Lux, D. R.; West, D. P.; Denton, G. H. 1993: Pliocene paleoclimate and East Antarctic Ice-Sheet history from surficial ash deposits. *Science 260*: 667–670.

Marchant, D. R.; Denton, G. H.; Swisher, C. C.; Potter, N. 1996: Late Cenozoic Antarctic paleoclimate reconstructed from volcanic ashes in the Dry Valleys region of southern Victoria Land. *Geological Society of America Bulletin 108*: 181–194.

Marshall, J. E. A. 1994: The Falkland Islands: a key element in Gondwana paleogeography. *Tectonics 13*: 449–514.

Mayes, C. L.; Lawver, L. A.; Sandwell, D. T. 1990: Tectonic history and new isochron chart of the South Pacific. *Journal of Geophysical Research 95*: 8543–8567.

Miller, M. F.; Mabin, M. C. G. 1998: Antarctic Neogene landscapes—in the refrigerator or in the deep freeze? *GSA Today 8*: 1–3.

Miller, S. R.; Fitzgerald, P. G.; Baldwin, S. L. 1998: Laterally progressive crustal cooling in the Transantarctic Mountains: fission-track evidence for ancient escarpment retreat. *EOS, Transactions of the American Geophysical Union 79*: F365.

Miller, S. R.; Fitzgerald, P. G.; Baldwin, S. L. 1999: Faulting, uplift, and denudation across the Transantarctic Mountain Front near Cape Surprise and Shackleton Glacier. *In*: Skinner, D. N. B. *ed*. 8th International Symposium on Antarctic Earth Sciences, 5–9 July 1999, programme and abstracts. Wellington, New Zealand, Victoria University of Wellington. P. 212.

Peacock, S. M. 1995: Calculating exhumation rates from thermochronologic data. *EOS, Transactions of the American Geophysical Union 76*: F569.

Peirce, H. W.; Damon, P. E.; Shafiqullah, M. 1979: An Oligocene(?) Colorado Plateau edge in Arizona. *Tectonophysics 61*: 1–24.

Redfield, T. F. 1994: The Transantarctic Mountains of northern Victoria Land: underplating, episodic uplift, and flexural suppression. PhD thesis, Tempe, Arizona State University.

Richard, S. M.; Smith, C. H.; Kimbrough, D. L.; Fitzgerald, P. G.; Luyendyk, B. P.; McWilliams, M. O. 1994: Cooling history of northern Ford Ranges, Marie Byrd Land, West Antarctica. *Tectonics 13*: 837–857.

Roland, N. W.; Tessensohn, F. 1987: Rennick faulting—an early phase of Ross Sea rifting. *Geologishes Jarhbuch B66*: 203–229.

Rowley, P. D.; Vennum, W. R.; Kellogg, K. S.; Laudon, T. S.; Carrara, P. E.; Boyles, J. M.; Thomson, M. R. A. 1983: Geology and plate tectonic setting of the Orville Coast and eastern Ellsworth Land, Antarctica. *In*: Oliver, R. L.; James, P. R.; Jago, J. B. *ed*. Antarctic earth science. Cambridge, Cambridge University Press. Pp. 245–250.

Salvini, F.; Brancolini, G.; Busetti, M.; Storti, F.; Mazzarini, F.; Coren, F. 1997: Cenozoic geodynamics of the Ross Sea region, Antarctica: crustal extension, intraplate strike-slip faulting, and tectonic inheritance. *Journal of Geophysical Research 102*: 24669–24696.

Schäfer, J. M.; Ivy-Ochs, S.; Wieler, R.; Leya, I.; Baur, H.; Denton, G. H.; Schlüchter, C. 1999: Cosmogenic noble gas studies in the oldest landscape on earth: surface exposure ages of the Dry Valleys, Antarctica. *Earth and Planetary Science Letters 167*: 215–226.

Schäfer, T.; Olesch, M. 1998: Multiple thermal evolution of Oates Land (Northern Victoria Land, Antarctica): evidence from fission track analysis. *In*: Van den Haute, P.; Corte, F. D. *ed*. Advances in fission-track geochronology. Dordrecht, Kluwer Academic Publishers. Pp. 241–253.

Schmidt, D. K.; Rowley, P. D. 1986: Continental rifting and transform faulting along the Jurassic Transantarctic rift, Antarctica. *Tectonics 5*: 279–291.

Schopf, J. M. 1969: Ellsworth Mountains: position in West Antarctica due to sea-floor spreading. *Science 164*: 63–66.

Spencer, J. E.; Reynolds, S. J. 1989: Middle Tertiary tectonics of Arizona and adjacent areas. *In*: Jenney, J. P.; Reynolds, S. J. *ed*. Geologic evolution of Arizona. Tucson, Arizona Geological Society Digest. Pp. 539–574.

Stagg, H. M. J.; Willcox, J. B. 1992: A case for Australia-Antarctica separation in the Neocomian (ca. 125 Ma). *Tectonophysics 210*: 21–32.

Stern, T.; ten Brink, U. S. 1989: Flexural uplift of the Transantarctic Mountains. *Journal of Geophysical Research 94*: 10315–10330.

Stock, J.; Molnar, P. 1987: Revised history of early Tertiary plate motion in the southwest Pacific. *Nature 325*: 495–499.

Storey, B. C. 1996: Microplates and mantle plumes in Antarctica. *Terra Antartica 3*: 91–102.

Storey, B. C.; Kyle, P. R. 1999: The Weddell Sea region: a plume impact site and source for Gondwana magmas during the initial stages of Gondwana breakup. *In*: Skinner, D. N. B. *ed*. 8th International Symposium on Antarctic Earth Sciences, 5–9 July 1999, programme and abstracts. Wellington, New Zealand, Victoria University of Wellington. P. 292.

Storey, B. C.; Hole, M. J.; Pankhurst, R. J.; Millar, I. L.; Vennum, W. R. 1988: Middle Jurassic within-plate granites in West Antarctica and their bearing on the breakup of Gondwanaland. *Journal of the Geological Society, London 145*: 999–1007.

Storey, B. C.; Vaughan, A. P. M.; Millar, I. L. 1996: Geodynamic evolution of the Antarctic Peninsula during Mesozoic times and its bearing on Weddell sea history. *In*: Storey, B. C.; King, E. C.; Livermore, R. A. *ed*. Weddell Sea tectonics and Gondwana breakup. Special Publication, London, Geological Society. Pp. 87–103.

Storey, B. C.; Dalziel, I. W. D.; Garrett, S. W.; Grunow, A. M.; Pankhurst, R. J.; Vennum, W. R. 1998: West Antarctica in Gondwanaland: crustal blocks, reconstruction and breakup processes. *Tectonophysics 155*: 381–390.

Stroeven, A. P.; Burkle, L. H.; Kleman, J.; Prentice, M. L. 1998: Atmospheric transport of diatoms in the Antarctic Sirius Group: Pliocene deep freeze. *GSA Today 8*: 1–8.

Stump, E. 1995: The Ross Orogen of the Transantarctic Mountains. New York, Cambridge University Press. 284 p.

Stüwe, K.; White, L.; Brown, R. 1994: The influence of eroding topography on steady-state isotherms: application to fission track analysis. *Earth and Planetary Science Letters 124*: 63–74.

Sugden, D. E.; Denton, D. H.; Marchant, D. R. 1995a: Landscape evolution of the Dry Valleys, Transantarctic Mountains: tectonic implications. *Journal of Geophysical Research 100*: 9949–9967.

Sugden, D. E.; Marchant, D. R. Jr; Potter, N. P.; Souchez, R. A.; Denton, G. H.; Swisher, C. C.; Tison, J.-L. 1995b: Preservation of Miocene glacier ice in East Antarctica. *Nature 376*: 412–414.

Summerfield, M. A. 1991: Global geomorphology: an introduction to the study of landforms. New York, John Wiley. 537 p.

Summerfield, M. A.; Stuart, F. M.; Cockburn, H. A. P.; Sugden, D. E.; Denton, G. H.; Dunai, T.; Marchant, D. R. 1999: Long-term rates of denudation in the Dry Valleys, Transantarctic Mountains, southern Victoria Land, Antarctica based on in situ-produced cosmogenic ^{21}Ne. *Geomorphology 27*: 113–129.

ten Brink, U. S.; Bannister, S.; Beaudoin, B. C.; Stern, T. A. 1993: Geophysical investigations of the tectonic boundary between East and West Antarctica. *Science 261*: 45–50.

ten Brink, U. S.; Hackney, R. I.; Bannister, S.; Stern, T. A.; Makovsky, Y. 1997: Uplift of the Transantarctic Mountains and the bedrock beneath the East Antarctic ice sheet. *Journal of Geophysical Research 102*: 27603–27621.

Tessensohn, F.; Wörner, G. 1991: The Ross Sea rift system, Antarctica: structure, evolution, and analogues. *In*: Thompson, M. R. A.; Crame, J. A.; Thomson, J. W. *ed*. Geological evolution of Antarctica. Cambridge, Cambridge University Press. Pp. 273–277.

Tulloch, A. J.; Kimbrough, D. L. 1989: The Paparoa metamorphic core complex, Westland-Nelson, New Zealand: Cretaceous extension associated with fragmentation of the Pacific margin of Gondwana. *Tectonics 8*: 1217–1234.

Turnbull, I. M.; Forsyth, P. J.; Mortimer, N. 1999: Field tests of Transantarctic Mountains uplift models, Cape Roberts hinterland. *In*: Skinner, D. N. B. *ed*. 8th International Symposium on Antarctic Earth Sciences, 5–9 July 1999, programme and abstracts. Wellington, New Zealand, Victoria University of Wellington. P. 306.

van der Beek, P.; Cloetingh, S.; Andriessen, P. 1994: Mechanisms of extensional basin formation and vertical motions at rift flanks: constraints from tectonic modeling and fission track thermochronology. *Earth and Planetary Science Letters 121*: 417–433.

Veevers, J. J.; Powell, C. M.; Rotts, S. R. 1990: Review of seafloor spreading around Australia. I, Synthesis of the patterns of spreading. *Australian Journal of Earth Science 38*: 373–389.

Warren, G. 1969: Sheet 14—Terra Nova Bay–McMurdo Sound area. Geologic map of Antarctica. *Antarctic Map Folio Series, Folio 12—Geology*. American Geographical Society.

Weaver, S. D.; Adams, C. J.; Pankhurst, R. J.; Gibson, I. L. 1992: Granites of Edward VII Peninsula, Marie Byrd Land: anorogenic magmatism related to Antarctic-New Zealand rifting. *Geological Society of America Special Paper 272*: 281–290.

Weaver, S. D.; Storey, B. C.; Pankhurst, R. J.; Mukasa, S. B.; Divenere, V. J.; Bradshaw, J. D. 1994: Antarctic-New Zealand rifting and Marie Byrd Land lithospheric magmatism linked to ridge subduction and mantle plume activity. *Geology 22*: 811–814.

Webb, P. N.; Harwood, D. M.; McKelvey, B. C.; Mercer, J. H.; Stott, L. D. 1984: Cenozoic marine sedimentation and ice volume variation on the East Antarctic craton. *Geology 12*: 287–291.

White, R. S.; MacKenzie, D. 1989: Magmatism at rift zones: the generation of volcanic continental margins and flood basalts. *Journal of Geophysical Research 94*: 7685–7729.

Wilch, T. I.; Denton, G. H.; Lux, D. R.; McIntosh, W. C. 1993a: Limited Pliocene glacier extent and surface uplift in middle Taylor Valley, Antarctica. *Geografiska Annaler 75A*: 331–351.

Wilch, T. I.; Lux, D. R.; Denton, G. H.; McIntosh, W. C. 1993b: Minimal Pliocene–Pleistocene uplift of the Dry Valleys sector of the Transantarctic Mountains: a key parameter in ice-sheet reconstructions. *Geology 21*: 841–844.

Wilson, T. J. 1992: Mesozoic and Cenozoic kinematic evolution of the Transantarctic Mountains. *In*: Yoshida, Y.; Kaminuma, K.; Shiraishi, K. *ed*. Recent progress in Antarctic Earth Science. Tokyo, Terra Scientific Publishing Company (TERRAPUB). Pp. 303–314.

Wilson, T. J. 1993: Jurassic faulting and magmatism in the Transantarctic Mountains: implications for Gondwana breakup. *In*: Findlay, R. H.; Banks, M. R.; Unrug, R.; Veevers, J. *ed*. Gondwana 8: Assembly, evolution, and dispersal. Rotterdam, Netherlands, A.A. Balkema. Pp. 563–572.

Wilson, T. J. 1995: Cenozoic transtension along the Transantarctic Mountains-West Antarctic rift boundary, southern Victoria Land, Antarctica. *Tectonics 14*: 531–545.

Wilson, T. J. 1999: Cenozoic structural segmentation of the Transantarctic Mountains rift flank in southern Victoria Land. *Global and Planetary Change 23*: 105–127.