Evidence of rapid Cenozoic uplift of the shoulder escarpment of the Cenozoic West Antarctic rift system and a speculation on possible climate forcing

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ABSTRACT

The Cenozoic West Antarctic rift system, characterized by Cenozoic bimodal alkalic volcanic rocks, extends over a largely ice-covered area, from the Ross Sea nearly to the Bellingshausen Sea. It is bounded on one side by a spectacular 4- to 5-km-high rift-shoulder scarp (maximum bedrock relief 5 to 7 km) from northern Victoria Land–Queen Maud Mountains to the Ellsworth-Whitmore-Horlick Mountains. Jurassic tholeiites crop out with the late Cenozoic volcanic rocks along the section of the Transantarctic Mountains from northern Victoria Land to the Horlick Mountains. The Cenozoic rift shoulder diverges here from the Jurassic tholeiite trend, and the tholeiites are exposed discontinuously along the lower elevation (1–2 km) section of the Transantarctic Mountains to the Weddell Sea.

Various lines of evidence, no one of which is independently conclusive, lead us (as others have also suggested) to interpret the following. The Transantarctic Mountains part of the rift shoulder (and probably the entire shoulder) has been rising since about 60 Ma, at episodic rates of ~1 km/m.y., most recently since mid-Pliocene time, rather than continuously at the mean rate of 100 m/m.y. Uplift rates vary along the scarp, which is cut by transverse faults. We speculate that this uplift may have climatically forced the advance of the Antarctic ice sheet since the most recent warm period. We suggest a possible synergistic relation between episodic tectonism, mountain uplift, and volcanism in the Cenozoic West Antarctic rift system and waxing and waning of the Antarctic ice sheet beginning about earliest Oligocene time.

INTRODUCTION

The asymmetric West Antarctic rift system (LeMasurier, 1978) extends over a 3000 × 750 km, largely ice-covered area (Figs. 1 and 2), comparable in area to the Basin and Range province and the East African rift systems (Tessensohn and Woerner, 1991). The rift boundary is defined by a spectacular Cenozoic rift-shoulder scarp that extends from northern Victoria Land to the Queen Maud Mountains to the Horlick-Whitmore-Ellsworth Mountains, along which peaks reach maximum elevations of 4–5 km (Figs. 2 and 3). Maximum bedrock relief is 5 km in the Ross embayment and 7 km in the Ellsworth Mountains–Byrd subglacial basin area. The rift is characterized by Cenozoic bimodal alkalic volcanic rocks (LeMasurier, 1990; Kyle, 1990), typical of continental rifts, that range in age from Oligocene or older (George, 1989) to Holocene. In contrast, tholeiites (Ferrar dolerites, Kirkpatrick basalts) marking the Jurassic Transantarctic rift (Schmidt and Rowley, 1986) crop out with late Cenozoic volcanic rocks along the section of the Transantarctic Mountains from Victoria Land to the Horlick Mountains (Craddock et al., 1969). The Cenozoic rift shoulder diverges from the tholeiite trend in the Horlick Mountains (Figs. 1 and 2). The Jurassic tholeiites are exposed (including the Dufek intrusion) along the Transantarctic Mountains to the Weddell Sea, whereas the late Cenozoic volcanic rocks are exposed in Marie Byrd Land (Figs. 1 and 2) to the southern Antarctic Peninsula, but not in the Ellsworth-Whitmore Mountains area.

GEOPHYSICAL EVIDENCE

Widely spaced aeromagnetic profiles throughout West Antarctica indicate the probable absence of Cenozoic volcanic rocks in the ice-covered part of the Ellsworth-Whitmore Mountains area and suggest their widespread occurrence beneath the western Byrd subglacial basin (Behrendt, 1964; Jankowski et al., 1983; Behrendt and Cooper, 1990). An aeromagnetic survey over the western Ross Sea continental shelf indicates rift fabric and suggests numerous submarine volcanoes along discrete north-northwest–trending zones that are defined by late Mesozoic–Cenozoic rifting (Behrendt et al., 1987, 1991).

Large-offset seismic profiles over the Ross Sea shelf (McGinnis et al., 1985; Cooper et al., 1987b; Trehu et al., 1989) indicate that the crust beneath the Ross Sea shelf is 17–21 km thick; we interpret this as evidence for crustal extension. A regional positive Bouguer gravity anomaly extends from the Ross Sea continental shelf throughout the Ross Ice Shelf–Byrd subglacial basin area of the West Antarctic rift system; on the basis of the large-offset seismic and gravity results (Behrendt et al., unpublished), the extended crust beneath the Ross Ice Shelf–Byrd subglacial basin is also about 20 km thick, rather than 30 km as interpreted in earlier (Woollard, 1962) gravity studies.

The near absence of earthquakes in the Cenozoic West Antarctic rift system probably results from a combination of sparse seismograph coverage (the Rio Grande rift because of its low seismicity would not be detectable solely from similar seismograph spacing), suppression of earthquakes by the grounded ice sheet (e.g., Johnston, 1987), and high seismicity immediately after deglaciation in the Ross embayment followed by abnormally low seismicity at present (e.g., the interpretation for Fennoscandia by Muir Wood, 1989). We do not consider the low seismicity as significant evidence against the presence of an active rift.

We suggest that the Cenozoic West Antarctic rift system is a part of the rifting that started during the Jurassic when Africa rifted from East Antarctica. Rifting proceeded clockwise (Lawver et al., 1991) around East Antarctica to New Zealand and the Campbell Plateau, which separated from Marie Byrd Land at ca. 85–95 Ma. Rifting has since continued (with perhaps a spreading-center jump) to its present location in the Ross embayment and West Antarctica. The Byrd subglacial basin–Ross embayment was probably extended greatly during late Mesozoic and Cenozoic time, as allowed by plate-reconstruction models (Stock, 1989; J. M. Stock, 1989, personal commun.). If a 40-km-thick crust was extended to 20 km across the 750-km-wide rift, total extension of ~350–400 km is implied.

UPLIFT OF RIFT SHOULDER

The steep scarp marking the rift shoulder suggests a youthful topography. Although present rates of weathering may be low in the absence of water and dry-based glacial erosion may be minimal, repeated periods of Cenozoic glaciation and deglaciation (Webb, 1990) would probably have been accompanied by very high rates of fluvial and wet-based glacial erosion. The rift scarp is interpreted to be the expression of a major normal or extensional fault zone (with a likely strike-slip component: Ford, 1972; Storey and Nell, 1988; Kellogg and Rowley, 1989). We interpret that the main cause of uplift, along the rift shoulder (Figs. 1 and 2), is late Cenozoic tectonism, probably as modeled by Stern and ten Brink (1989), for a flexed continental lithospheric plate heated at the free edge (mountain front).

The variation of the high topography seen in Figure 3 along the rift shoulder is partly caused by erosion and by differential uplift on transverse faults. Denton et al. (1984) referred to possible separate histories of individual fault blocks. They noted that uneroded subaerial volcanic cones dated as 4.2 and 3.5 Ma in Wright Valley (and other evidence) indicate that tectonic uplift is limited to ~300 m since the last glacial overriding. Webb and Wrenn (1982), Wrenn and Webb (1982), and Ishman and Webb (1988) also showed evidence that indicates ~500 m of uplift occurred in the Taylor Valley area starting ca. 3 Ma. Although the maximum elevation in the Transantarctic Mountains in the McMurdo area (e.g., Mount Lister) is 4 km, ~20 km to the north, maximum elevations on ridges bordering the Taylor and Wright valleys are only ~2 km, suggesting to us differential uplift along the axis of the mountains in this area. Transverse faulting along rift shoulders is common and is apparent (Cooper et al., 1991) along the Transantarctic Mountains bordering the Ross embayment.

Figure 3 shows various maximum elevations: ~4 km in Victoria Land, ~4.5 km in the Queen Maud Mountains (Fig. 1), ~3 km in the Horlick and Whitmore mountains (perhaps resulting from a greater rate of erosion or differential uplift), and ~5 km in the Ellsworth Mountains. In contrast, elevations along the Transantarctic Mountains are much lower toward the Weddell Sea (dashed-line profile of Fig. 3), ranging from ~2 km in the Thiel and Pensacola mountains (Dufek intrusion) to ~1.5 km in the Shackleton Mountains and 1 km in the Theron Mountains. The solid-line elevations (Fig. 3) parallel the trend of the overdeepened subglacial basins (Fig. 2), the geophysical anomaly patterns, and the Cenozoic volcanic fields that characterize the Cenozoic West Antarctic Rift System (Behrendt

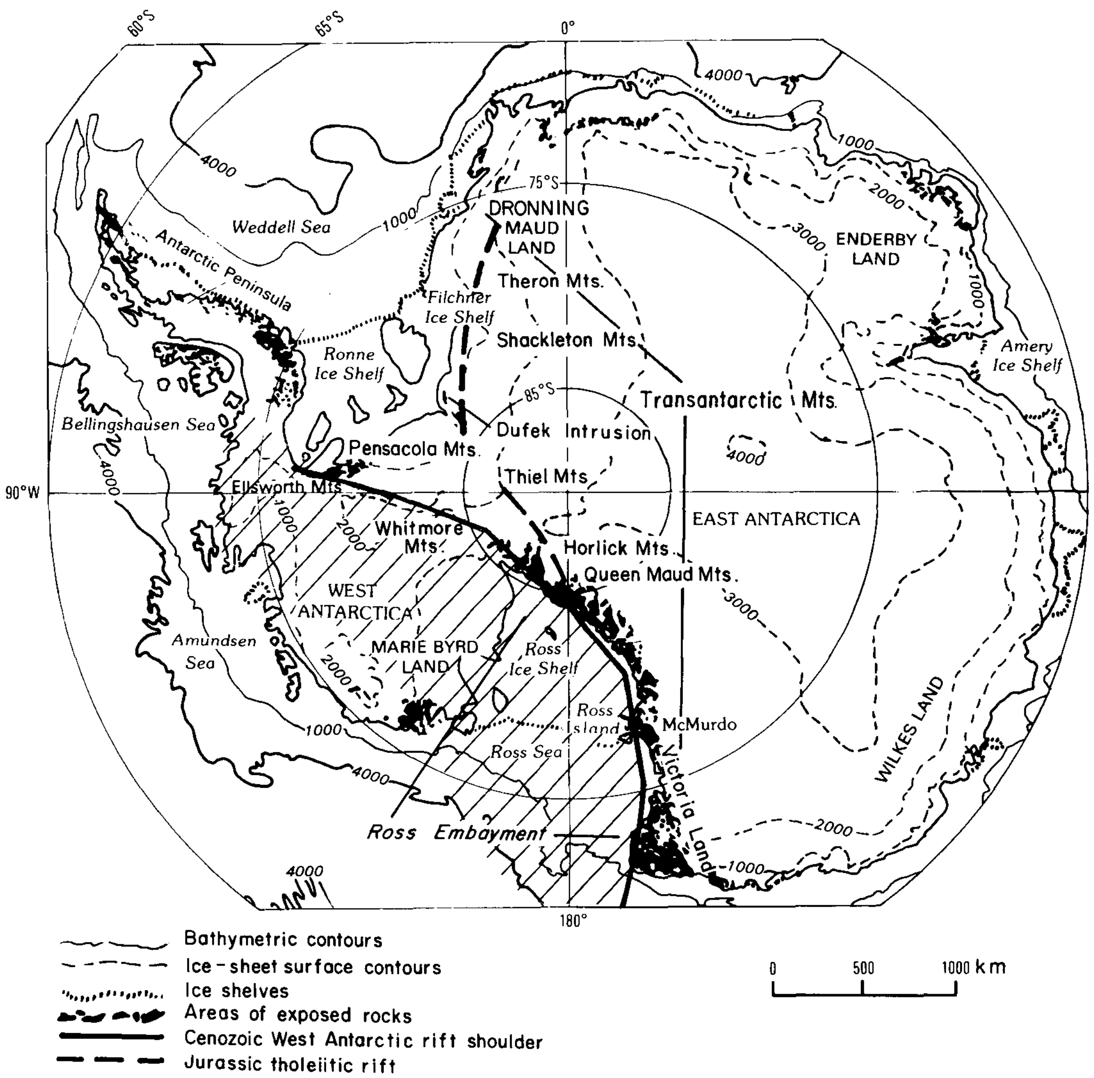

Figure 1. Index map of Antarctica showing some features discussed in text. Diagonal-rule area is approximate location of Cenozoic West Antarctic rift system. Heavy line is approximate rift shoulder; dashed line extends along lower elevations of Transantarctic Mountains from Horlick to Theron mountains.

et al., 1991, unpublished). Dashed-line elevations (Fig. 3) diverge and follow exposures of Jurassic tholeiitic rocks (Fig. 2) toward the Weddell Sea; they imply older uplift.

Fitzgerald (1989), using fission-track dates from the relatively low Dry Valley area north of Mount Lister, interpreted 5–6 km of uplift in the Transantarctic Mountains of southern Victoria Land. Uplift began about 60 Ma and has continued at an average calculated rate of about 100 m/m.y. since that time. These estimates are similar to those summarized by Tingey (1985). Seismic-reflection data from the Ross Sea adjacent to the Transantarctic Mountains show several distinct angular unconformities, along the western edge of the Victoria Land Basin, that have been interpreted by Cooper et al. (1987a) as evidence for episodic uplift. Therefore, we infer that the uplift of the Transantarctic Mountains and entire rift shoulder was episodic and has probably been an order of magnitude faster, at times including the present, than the mean rate.

Using an Occam's razor or "principle of least astonishment" approach, various lines of evidence, none of which is independently conclusive, lead us to interpret that the Transantarctic Mountains part of the rift shoulder (and probably the entire shoulder) has been rising at a rate of ~1 km/m.y. during the latest episode, probably since early or middle Pliocene time. The following is evidence for this order of magnitude of uplift, no single item of which is either definitive or cannot be explained by other mechanisms, as several of the authors cited have done.

1. The youthful-appearing rift shoulder has about the highest elevation (4–5 km, Fig. 2) of any suspected rift mountains on Earth.

2. The relief of 5–7 km on the shoulder (Fig. 2) is the greatest of any rift mountains, even if the rift fill (ice) were compressed to the density of sedimentary rock and that elevation contrasted with shoulder elevations.

3. Holocene fault scarps with ~10 m displacement (Behrendt, 1991) are interpreted from high-resolution marine (Ross Sea) reflection profiles.

4. Normal faults displace late Pliocene moraines, with as much as 300 m vertical displacement along the Transantarctic Mountains bordering the Ross embayment (McKelvey et al., 1990).

5. Unconformities in the Victoria Land Basin (Cooper et al., 1987a), in conjunction with the

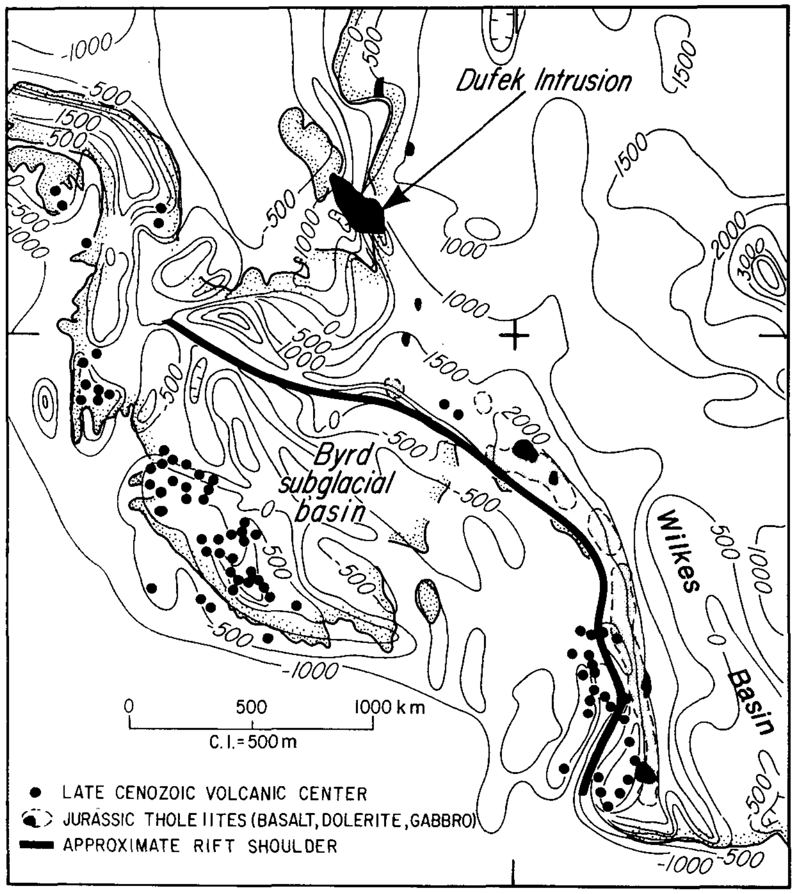

Figure 2. Generalized, isostatically adjusted, bedrock elevation map after ice removal, assuming sufficient time. Contour interval is 500 m (modified from Drewry, 1983). Location of late Cenozoic volcanic centers from Gonzales-Ferran (1982) and LeMasurier (1990). Locations of Jurassic tholeiites (basalt, dolorite, and gabbro) from Craddock et al. (1969). Area beneath rift zone (elevations below sea level) is probably underlain by ~20-km-thick crust. Wilkes basin contains thick section (>8 km) of sedimentary rock, as estimated from magnetic profiles.

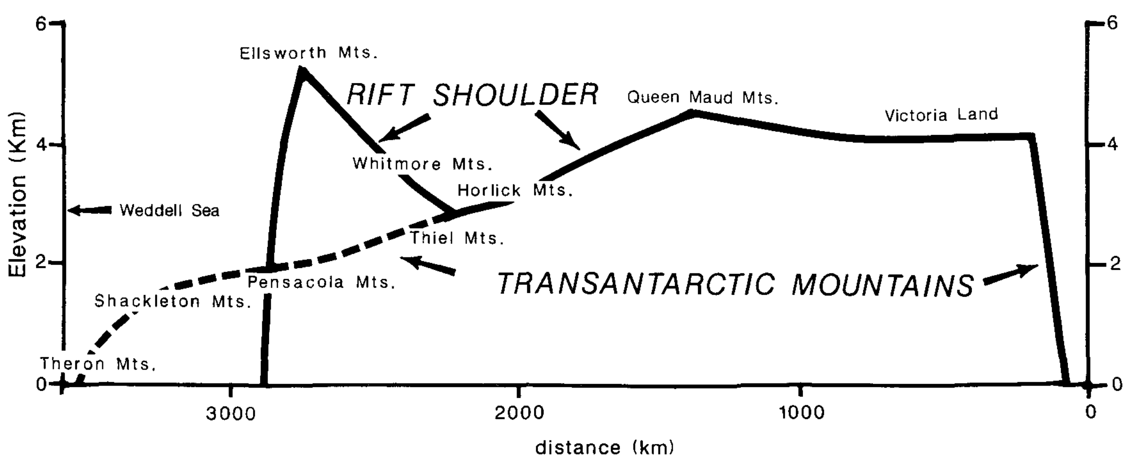

Figure 3. Generalized topographic profile along crest of highest peaks along interpreted rift shoulder (Figs. 1 and 2) (solid line) from north coast of Victoria Land at right to Ellsworth Mountains at left compared with lower topography (dashed line) of highest peaks in Transantarctic Mountains from Thiel Mountains to Weddell Sea coast at left. Interpreted Cenozoic West Antarctic rift shoulder has topography higher than any other rift. Low areas within mountains probably are results of glacial or fluvial erosion and differential uplift and are not shown in this profile.

average uplift rate of 100 m/m.y. since 60 Ma (Fitzgerald, 1989), require episodic uplift. We do not know of any active rift with a constant uplift rate for 60 m.y.

6. Glacial striae are found on peaks in the Transantarctic Mountains 600–1000 m above present ice levels (e.g., Hoefle, 1989; McKelvey et al., 1991). Although these striae possibly indicate higher ice-sheet elevations in the past (Hoefle, 1989; Prentice et al., 1986), we think that the striae more likely are due to tectonic uplift of the peaks, in agreement with McKelvey et al. (1991), Webb et al. (1986), and Harwood (1986). On the basis of glaciologic criteria, the ice sheet is unlikely to have been this much higher at some of the locations where these striae were observed. For example, both maximum and minimum reconstructions of grounded ice in the Ross embayment (Denton et al., 1989) indicate maximum ice-sheet elevations much lower than the highest striae.

7. *Nothofagus* (Pliocene beech leaf) is present in situ in moraines at high elevations in the Transantarctic Mountains (Webb and Harwood, 1987). This presence is consistent (Mercer, 1987) with, but does not necessarily require, an uplift rate of 1 km/m.y. because, as noted by Harwood et al. (unpublished), the mountains were significantly uplifted in the past 2–3 m.y., or Pliocene climate was considerably warmer than present, leading to extremely high tree-line limits, or the Pliocene Antarctic *Nothofagus* could endure colder conditions than its closest living relative.

8. Late Pliocene marine microfossils in the Sirius Group have been glacially transported from the Wilkes basin by the East Antarctic ice sheet to elevations between 1750 and 2500 m in the Transantarctic Mountains. These microfossil locations can be more easily explained if the mountains were much lower at 3 Ma than at present (Webb et al., 1984). Mercer (1978) suggested >1000 m of uplift since the Sirius Group was deposited (now known to be 2.5 Ma, D. Harwood, personal commun., 1990).

9. Neotectonic topography of other rifts has uplift rates that are similar or higher (e.g., East Africa, 2 and 8 km/m.y.—Ebinger [1989]; western United States, Wasatch Mountains, 2–2.5 km/m.y.—Naeser et al. [1983]; other areas—Ruddiman et al. [1989]; Parry and Bruhn [1987], Lajoie [1986]). Different tectonic settings have rates that are up to an order of magnitude greater (e.g., New Zealand, 10 km/m.y.—Wellman [1979]).

We can consider age limits on the latest tectonic episode. The present Victoria Land–Ellsworth Mountains scarp could have been built in 4–5 m.y., assuming that all uplift was accommodated at ~1 km/m.y. We know, however (e.g., Fitzgerald, 1989; Barrett et al., 1989), that uplift started much earlier. We do not know the time of the uplift of the Pensacola-Shackleton-Theron part of the Transantarctic Mountains; we know only that the faulting of the Transantarctic Mountains front that offsets the Jurassic Dufek intrusion is younger than its emplacement (Behrendt et al., 1974). Therefore, alternatively assuming that the 2 km elevation of the Pensacola Mountains (Fig. 3) is equal to that of the entire rift shoulder (from Victoria Land to the Ellsworth Mountains) prior to the latest (i.e., since the Pliocene) episode of uplift, only 2–3 m.y. are needed for the rift shoulder to reach the present highest elevation (4–5 km). Estimates from Harwood (1986), Harwood et al. (unpublished), McKelvey et al. (1991), and Webb and Harwood (1987) provide considerably better timing of this uplift in the Queen Maud Mountains area, where the start of uplift must postdate the deposition of the Sirius Group and the Pliocene *Nothofagus*, which occurred between 2 and 3 Ma (Harwood et al., unpublished).

Earlier episodes of uplift and erosion also occurred, as suggested by the maximum uplift of 10 km in northern Victoria Land (Fitzgerald, 1989) and interpretations by Barrett et al. (1989) that the Transantarctic Mountains had reached half their present elevation by early Oligocene time. Uplift along the entire 3000 km rift shoulder was episodic, and the motions were complex, involving transverse faulting and differential uplift.

LATE CENOZOIC CLIMATIC IMPLICATIONS

Researchers have long noted (e.g., Wright, 1923) a possible association of uplift of the Transantarctic Mountains and glaciation as referenced in Tingey (1985). It is interesting that our estimated limits for the start of the latest episode of uplift (2–5 Ma) approximately coincides with the start of the cold period at 2.5 Ma that changed Antarctic glaciation from temperate to polar (Harwood and Webb, 1990; Harwood et al., unpublished) and killed the *Nothofagus*. Whereas Stern and ten Brink (1989) and Webb (1990) noted that "overall the present Antarctic ice sheet appears to have had only a small effect on uplift of the Transantarctic Mountains" (~100 m), we suggest, as have Harwood (1986) and Webb and Harwood (1987), that the converse is not necessarily true. The postulated rapid uplift of ~1 km/m.y. since early or middle Pliocene time may in fact have led to the most recent advance of the East Antarctic ice sheet through climatic deterioration by a forcing mechanism similar to that proposed for the Northern Hemisphere (winter cooling and increased precipitation caused by comparable mountain uplift culminating in the Pliocene-Pleistocene ice ages) by Ruddiman and Kutzbach (1989), although their model required plateau uplift. Additionally, damming of ice-sheet flow by rapid uplift could also be a major factor in growth of the present East Antarctic ice sheet (Webb and Harwood, 1987; Harwood, 1986) and in changing the thermal character of the ice sheet from temperate to polar (Harwood and Webb, 1990; Harwood et al., unpublished). We suggest the possibility of a synergistic relation between episodic Cenozoic tectonism, mountain uplift, and volcanism in the West Antarctic rift system and waxing and waning of the Antarctic ice sheet (Webb, 1990) beginning about earliest Oligocene time.

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