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GEOPHYSICAL STUDIES OF THE WEST ANTARCTIC RIFT SYSTEM

J. C. Behrendt,[1] W. E. LeMasurier,[2] A. K. Cooper,[3] F. Tessensohn,[4] A. Tréhu,[5] and D. Damaske[4]

Abstract. The West Antarctic rift system extends over a 3000 x 750 km, largely ice covered area from the Ross Sea to the base of the Antarctic Peninsula, comparable in area to the Basin and Range and the East African rift system. A spectacular rift shoulder scarp along which peaks reach 4-5 km maximum elevation marks one flank and extends from northern Victoria Land-Queen Maud Mountains to the Ellsworth-Whitmore-Horlick Mountains. The rift shoulder has maximum present physiographic relief of 5 km in the Ross Embayment and 7 km in the Ellsworth Mountains-Byrd Subglacial Basin area. The Transantarctic Mountains part of the rift shoulder (and probably the entire shoulder) has been interpreted as 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. The rift system is characterized by bimodal alkaline volcanic rocks ranging from at least Oligocene to the present. These are exposed asymmetrically along the rift flanks and at the south end of the Antarctic Peninsula. The trend of the Jurassic tholeiites (Ferrar dolerites, Kirkpatric basalts) marking the Jurassic Transantarctic rift is coincident with exposures of 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 continuously (including the Dufek intrusion) along the lower- elevation (1-2 km) section of Transantarctic Mountains to the Weddell Sea. Widely spaced aeromagnetic profiles in West Antarctica indicate the absence of Cenozoic volcanic rocks in the ice covered part of the Whitmore-Ellsworth-Mountain block and suggest their widespread occurrence beneath the western part of the ice sheet overlying the Byrd Subglacial Basin. A German Federal Institute for Geosciences and Natural Resources (BGR)-U.S. Geological Survey (USGS) aeromagnetic survey over the Ross Sea continental shelf indicates rift fabric and suggests numerous submarine volcanoes along discrete NNW trending zones. A Bouguer anomaly range of approximately 200 (+50 to -150) mGal having 4-7 mGal/km gradients where measured in places marks the rift shoulder from northern Victoria Land possibly to the Ellsworth Mountains (where data are too sparse to determine maximum amplitude and gradient). The steepest gravity gradients across the rift shoulder require high density (mafic or ultramafic?) rock within the crust as well as at least 12 km of thinner crust beneath the West Antarctic rift system in contrast to East Antarctica. Sparse land seismic data reported along the rift shoulder, where velocities are greater than 7 km/s, and marine data indicating velocities above 7 km/s beneath the Ross Sea continental shelf support this interpretation. The maximum Bouguer gravity range in the Pensacola Mountains area of the Transantarctic Mountains is only about 130 mGal with a maximum 2 mGal/km gradient, which can be explained solely by 8 km of crustal thickening. Large offset seismic profiles over the Ross Sea shelf collected by the German Antarctic North Victoria Land Expedition V (GANOVEX V) combined with earlier USGS and other results indicate 17-21 km thickness for the crust beneath the Ross Sea shelf which we interpret as evidence of extended rifted continental crust. A regional positive Bouguer anomaly (0 to +50 mGal), the width of the rift, extends from the Ross Sea continental shelf throughout the Ross Embayment and Byrd Subglacial Basin area of the West Antarctic rift system and indicates that the Moho is approximately 20 km deep tied to the seismic results (probably coincident with the top of the asthenosphere) rather than the 30 km reported in earlier interpretations. The interpretation of horst and graben structures in the Ross Sea, made from marine seismic reflection data, probably can be extended throughout the rift (i.e., the Ross Ice shelf and the Byrd Subglacial Basin areas). The near absence of earthquakes in the West Antarctic rift system probably results from a combination of primarily sparse seismograph coverage and, secondarily, suppression of earthquakes by the ice sheet (e.g., Johnston, 1987) and very high seismicity shortly after deglaciation in the Ross Embayment followed by abnormally low seismicity at present (e.g., Muir Wood, 1989). The evidence of high temperatures at shallow depth beneath the Ross Sea continental shelf and adjacent Transantarctic Mountains is supportive of thermal uplift of the mountains associated with lateral heat conduction from the rift and can possibly also explain the volcanism, rifting, and high elevation of the entire rift shoulder to the Ellsworth- Horlick- Whitmore Mountains. We infer that the Gondwana breakup and the West Antarctic rift are part of a continuously propagating rift that started in the Jurassic when Africa separated from East Antarctica (including the failed Jurassic Transantarctic rift). Rifting proceeded clockwise around East Antarctica to the separation of New Zealand and the Campbell Plateau about 85-95 Ma and has continued (with a spreading center jump) to its present location in the Ross Embayment and West Antarctica. The Cenozoic activity of the West Antarctic rift system appears to be continuous in time with rifting in the same area that began only in the late Mesozoic. Although the mechanism for rifting is not completely explained, we suggest a combination of the flexural rigidity model (Stern and ten Brink, 1989)

[1]U.S. Geological Survey, Denver, Colorado.
[2]Department of Geology, University of Colorado, Denver.
[3]U.S. Geological Survey, Menlo Park, California.
[4]Federal Institute of Geosciences and Natural Resources, Hanover.
[5]College of Oceanography, Oregon State University, Corvallis.

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proposed for the Ross Embayment and the thermal plume or hot spot concepts. The propagating rift may have been "captured" by the thermal plume.

INTRODUCTION

This paper is an effort to integrate the geophysical research over the West Antarctic rift system over the past three decades, including new data in the Ross Sea area (reviewed in an expanded paper where errors and accuracies are discussed and additional references provided [Behrendt et al., 1991b], within the concepts of continental rifting developed for other areas during the past decade. The relative paucity of data in Antarctica requires us to follow an approach very similar to planetologists, that is, use analogs and be willing to speculate. We have done this here and hope we have distinguished the facts from the interpretation.

One of the world's largest and least known active rift systems extends through the interior of West Antarctica (Figures 1 and 2). The largely aseismic 3000 km x 750 km area of Cenozoic volcanism and Late Cretaceous and Cenozoic extension called the West Antarctic rift system (see names discussion by Behrendt et al. [1991b]) underlies the Ross Embayment and Byrd Subglacial Basin and is comparable in area to the Basin and Range or the East African rift (Figure 2; see also LeMasurier [1990] and Tessensohn and Wörner, [1991]).

Middle Jurassic tholeiites associated with the Jurassic Transantarctic rift [Schmidt and Rowley, 1986] coincide with exposures of Cenozoic bimodal alkaline volcanic rocks along the flank of the West Antarctic rift system along the Transantarctic Mountain front from the Ross Sea to the Horlick Mountains, where the two rift systems diverge. The Jurassic rift passes between the Whitmore and Horlick Mountains and extends into Dronning Maud Land to the Weddell Sea (Figure 1) in a region of older lower topography (1-2 km elevation) than that of the section of the Transantarctic Mountains bordering the Ross Embayment (over 4 km high) (Figure 3). In contrast, the West Antarctic rift (Figures 1 and 2) curves north from the Horlick Mountains and passes west of the Ellsworth-Whitmore Mountains scarp. Rift-associated late Cenozoic bimodal alkalic volcanic rocks are also exposed along the Amundsen- Bellingshausen Sea coast and at the south end of the Antarctic Peninsula (Figure 2). Plate reconstructions and motions [e.g., Bradshaw, 1989; Stock, 1989; Lawver et al., 1991] suggest that West Antarctica reached its present general configuration about 100 Ma. Similarities [Storey et al., 1989] between crustal blocks [e.g. Dalziel and Elliot, 1982] indicate that West Antarctica is not composed of far-traveled exotic terranes.

Both geological and geophysical evidence suggests that, in addition to the Cenozoic and Jurassic magmatic episodes, late Mesozoic-early Cenozoic crustal extension, associated with right-lateral strike-slip displacement (e.g., [Ford, 1972; Storey and Nell, 1988; Kellogg and Rowley, 1989] in the southern Antarctic Peninsula- Ellsworth Mountains area and [Stock, 1989] in the Campbell Plateau area) may have accompanied or immediately followed the separation of the New Zealand-Campbell Plateau block from Marie Byrd Land about 85 Ma [Jankowski et al., 1983; Cooper et al., 1987a; Bradshaw, 1989]. Although no exposed magmatic rocks of late Mesozoic age are known in the Ross Embayment-Byrd Subglacial Basin, major extension probably occurred there during this time [Cooper et al. 1987a; Stock, 1989] and continued (episodically?) in the Cenozoic; it is possible that late Mesozoic volcanic rocks are buried beneath the ice or in the lower part of rift basins beneath the Ross Embayment and Byrd Subglacial Basin.

TOPOGRAPHY

The Cenozoic West Antarctic rift shoulder is marked by spectacular nearly continuous mountain escarpments of pre-Cenozoic rocks that extend along the Transantarctic Mountains from northern Victoria Land to the Ellsworth-Whitmore-Horlick Mountain front (Figures 2 and 3). Relief ranges from 5 km from summits of the highest peaks (4 km) in the Transantarctic Mountains to the continental shelf at the edge of the Ross Sea (Figures 1 and 2), to 7 km in the Ellsworth Mountains from the summit of Mount Vinson, about 5 km above sea level, to the greatest depths of about 2 1/2 km in the Byrd Subglacial Basin and the Bentley subglacial trench [Drewry, 1983]. Subglacial erosion and graben formation have occurred simultaneously since glaciation about Oligocene time [LeMasurier and Thomson, 1990] in the Byrd Subglacial Basin (2 to 2.5 km depths), and both have contributed to the present relief, although extension and graben formation may be as old as late Mesozoic. The "sedimentary" basin fill within the Byrd Subglacial Basin consists largely of moving ice, in contrast to the sedimentary rock in basins in the Ross Embayment, which accounts for the lower-relief topography of the continental shelf there (Figure 2). The rugged graben and horst topography of the floor of the Byrd Subglacial Basin revealed by radar ice sounding [Drewry, 1983] is probably what we would expect for the Ross Embayment floor had the grounded ice sheet covered it since Oligocene time.

The Transantarctic Mountains-Ellsworth-Whitmore-Horlick flank is a well-defined rift shoulder (Figures 1-3) and where bordering the best studied Ross Embayment is represented by mountains that (1) have been built in a tensional field, (2) show a lack of folding, and (3) can be accounted for by the elastic flexure of two cantilevered plates separated by a stress-free edge [Stern and ten Brink, 1989]. In contrast, the opposing Amundsen-Bellingshausen flank [Behrendt et al., 1991b] with exposed outcrops in Marie Byrd Land, lacks a well-defined shoulder, resulting in overall asymmetrical topography. It is marked by graben and horst topography [LeMasurier and Thomson, 1990] and by pre-Cenozoic basement elevations now no more than 2500 m; post erosion surface subsidence of at least 3 km accompanied rifting [LeMasurier, 1990]. We interpret the main cause of uplift along the Cenozoic West Antarctic rift shoulder to be late Cenozoic tectonism associated with rifting although isostatic adjustment accompanying erosion contributes (e.g. [Tingey, 1985; England and Molnar, 1990]). The abrupt drop (Figures 2 and 3) at the north end of the Ellsworth Mountains separates these mountains from the southern Antarctic Peninsula and was interpreted as a major tectonic break by Behrendt [1964b] and as a rift by Doake et al. [1983] and Maslanyj and Storey [1990]. In

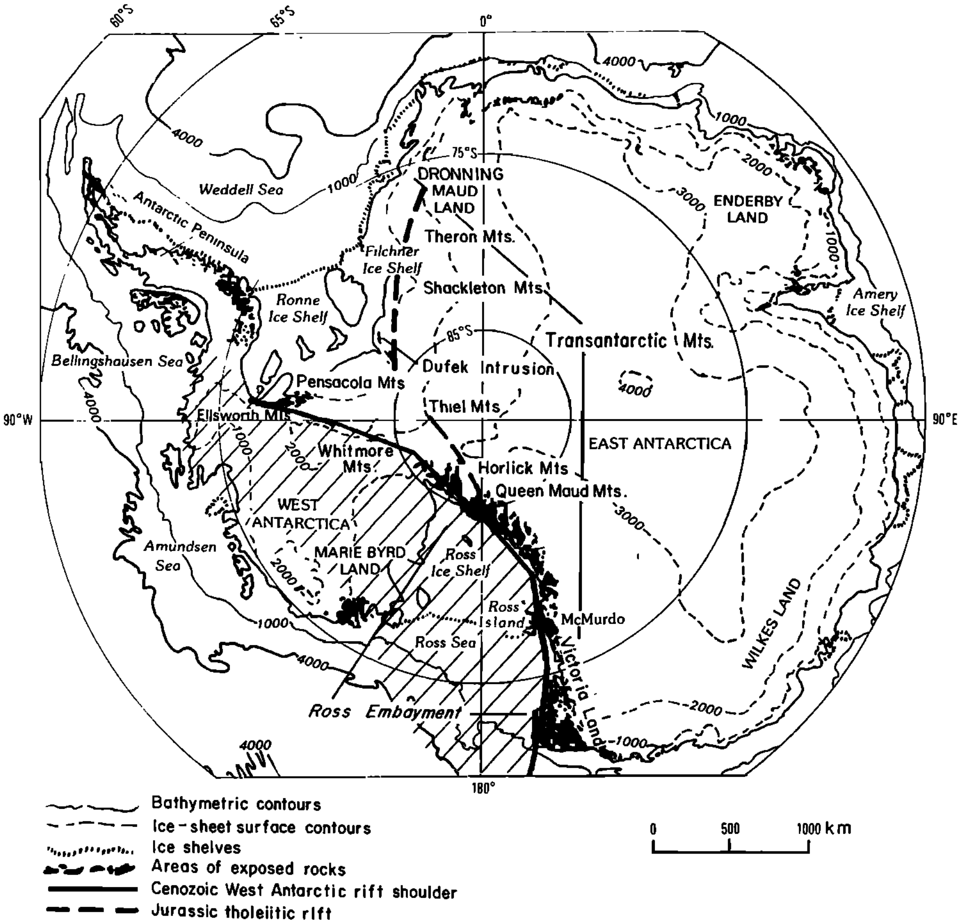

Fig. 1. Index map of Antarctica showing some of the features discussed in the text. The Transantarctic Mountains extend across the continent from Victoria Land near the Ross Sea to the Theron Mountains near the Weddell Sea and comprise the ranges shown by the heavy dashed line as well as those bordering the Ross Embayment. Diagonal lines show approximate location of West Antarctic rift system. Heavy line is approximate rift shoulder. To conform with convention and other publications, maps of Figures 1, 2, 4, and 8 (covering all or large regions of Antarctica) have grid north (parallel to 0° Meridian) at top; the larger scale maps (Figures 6, and 10 and Plate 1) have normal convention. All maps use polar stereographic projection.

contrast, elevations along the Transantarctic Mountains are lower toward the Weddell Sea (not a part of the Cenozoic rift shoulder; see dashed line profile in Figures 1 and 3). The maximum elevations range from about 2 km in the Thiel and Pensacola Mountains (including the Dufek intrusion) and drop to about 1 1/2 km in the Shackleton Mountains and 1 km in the Theron Mountains.

Fitzgerald [1989], using fission track dates, interpreted 5-6 km of uplift in the Transantarctic Mountains of southern Victoria Land beginning about 60 Ma and continuing at an average calculated rate of about 100 m/m.y. since that time [Fitzgerald, 1989; Tingey, 1985]. Behrendt and Cooper [1991] inferred 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 (e.g. ~1 km/m.y.), than the mean rate. Data from other similar tectonically active rift areas in the world suggest that uplift rates as great as 1-2 km/m.y. are not uncommon [Behrendt and Cooper, 1991]. 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 [1989] that the Transantarctic Mountains had reached half their present elevation by early Oligocene time. Uplift history along the entire 3000 km rift shoulder was episodic and complex with transverse faulting and differential uplift [Behrendt and Cooper, 1991].

Possibly the asymmetry in uplift characteristics between the rift shoulder and the Amundsen- Bellingshausen flank is caused by the very rigid lithosphere underlying East Antarctica [Stern and ten Brink, 1989] contrasted with

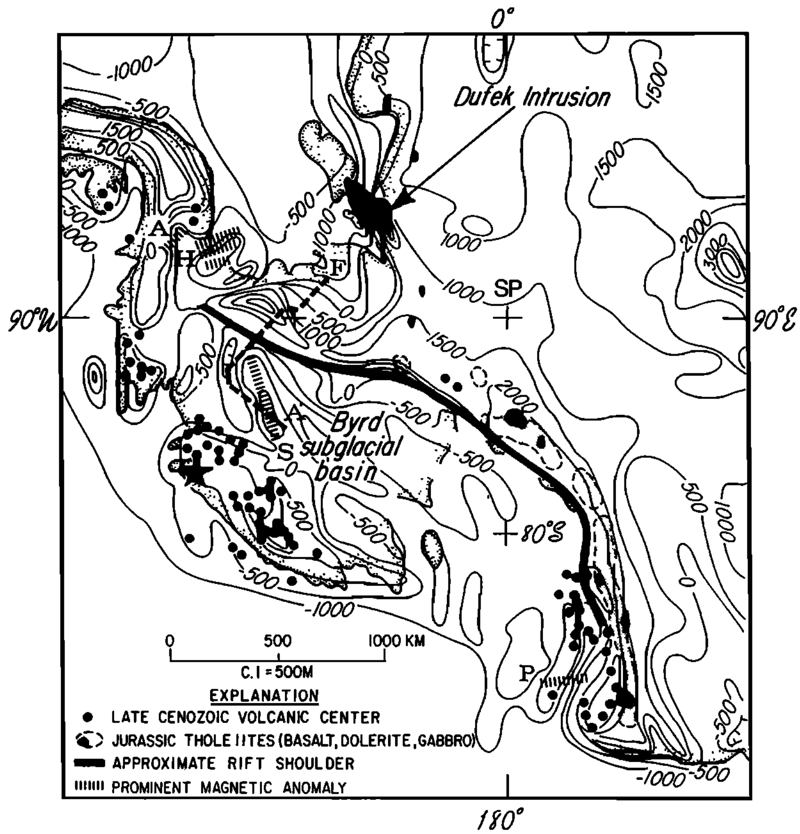

Fig. 2. Generalized isostatically adjusted bedrock elevation map after ice removal assuming sufficient time. Contour interval 500 m (modified from Drewry [1983]). All known high-amplitude linear (greater than 200 km long) magnetic anomalies greater than 1000 nT amplitude are indicated. However, coverage is sufficient not to have missed any only in areas of the detailed aeromagnetic survey (e.g., Plate 1). P, Polar 3 anomaly (Plate 1), S, Sinuous Ridge anomaly [Jankowski et al., 1983]. H, Haag Nunataks anomalies [Garrett et al., 1987]; A, Antarctic Peninsula Traverse anomaly [Behrendt, 1964b]. Approximate location of late Cenozoic volcanic centers from Gonzales-Ferran [1982] and LeMasurier and Thomson [1990]. Locations of Jurassic tholeiites (basalt, dolorite, and gabbro) from Craddock [1969]. Volcanic Toney Mountain is indicated by star; see Figure 8. The area beneath the rift zone (elevations below sea level) is probably underlain by crust about 20 km thick based on interpretation of Bouguer anomaly map (Figure 8), Bouguer anomaly interpretation and seismic determinations of Moho depth beneath Ross Sea Shelf (Figure 10). Location of profile A'-F (Figure 5) is indicated. Grid north is at top.

(probably less rigid) oceanic lithosphere offshore. Geophysical studies indicate multiple periods of subsidence in grabens in Ross Sea basins from as early as late Mesozoic [Cooper et al., 1987a; 1991], which is probably characteristic of the ice covered low areas in the Ross Embayment-Byrd Subglacial Basin.

CENOZOIC VOLCANIC ROCKS

The late Cenozoic highly alkaline, bimodal character of the volcanic rocks exposed along the rift flanks suggested even to early workers [Hamilton and Boudette, 1962] that volcanism in this region was rift related. However, the highly asymmetric distribution (Figure 2) of volcanic fields and the ice cover obscured both the geometry and scale of the rift system. Quite likely the bulk of the Cenozoic volcanic activity occurred beneath the ice and sea covered low elevation parts of the rift proper [Behrendt and Wold, 1963; Jankowski et al., 1983; Behrendt et al., 1991b]) rather than along its flanks.

The volcanic exposures (Figure 2) are arrayed roughly parallel to the deep subglacial topographic trough that forms the axial part of the rift. Basaltic rocks make up between 70% and 90% of the total volume of Cenozoic volcanic rock [LeMasurier, 1990; Kyle, 1990] in the Marie Byrd Land and Ross Sea provinces. In contrast, volcanoes at the Bellingshausen Sea and Southern Antarctic Peninsula (Figures 1 and 2) end of the rift system are relatively small, widely scattered, and entirely basaltic. The basaltic rocks in general produce the high amplitude magnetic anomalies discussed below.

Where complete data sets are available, they reveal especially narrow ranges in La/Nb (0.6-1.4), $^{87}Sr/^{86}Sr$ (0.70258-0.70387), and $^{143}Nd/^{144}Nd$ (0.51285-0.51307), indicating that basalts throughout the rift system were derived from a depleted mantle source region (i.e., at least

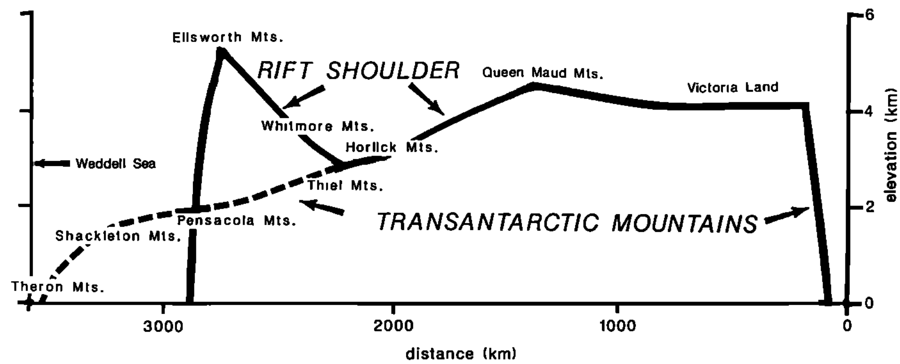

Fig. 3. Generalized topographic profile along the crest of the highest peaks parallel to and along the Cenozoic West Antarctic rift shoulder (Figures 1 and 2) (solid line) from the north coast of Victoria land at right to the Ellsworth Mountains at left compared with lower topography (dashed line) of highest peaks in the Transantarctic Mountains from the Horlick Mountains to the Weddell Sea. Low areas within the mountains probably are the result of glacial and fluvial erosion, differential uplift and transverse down faulting and are not shown in these profiles.

100 km deep within the asthenosphere) with no perceptible contribution from lithospheric mantle or crustal sources [Stuckless and Ericksen, 1976; Futa and LeMasurier, 1983; Wörner et al., 1989; Hole and LeMasurier, 1990]. Rift-related basalts from the Ross Sea region, Marie Byrd Land, and western Ellsworth Land all have K/Ba ratios <50, whereas Antarctic Peninsula basalts have K/Ba ratios >50 [Hole and LeMasurier, 1990]. Furthermore, the geochemical characteristics of the rift-related basalts are indistinguishable from those of oceanic island basalts, and the Antarctic Peninsula basalts differ from both in the above LIL-element characteristics. These distinctions become especially significant in helping to define the effects of mantle plume activity in the West Antarctic rift system, as discussed in a later section.

In general terms, the picture that seems to be emerging for rift-related magmatic activity in this region involves (1) the generation of alkaline basaltic magma from mantle sources, largely within the asthenosphere; (2) eruption of a large proportion of this material without major modification; and (3) differentiation of other batches of basaltic magma in crustal reservoirs by a complex variety of fractional crystallization processes.

Drill cores from McMurdo Sound record volcanic activity of Oligocene or older age (CIROS [Barrett, 1989]) and early Cenozoic (MSSTS, [Gamble et al., 1986]); the oldest exposed alkaline rocks along the Transantarctic Mountains front are 18-25 Ma [Schmidt-Thomé et al., 1990]. In Marie Byrd Land, basaltic volcanism seems to have begun around 25-30 Ma [LeMasurier, 1990] and continued with few interruptions. In the Bellingshausen volcanic province [Rowley et al., 1989] the oldest basalts are 7-10 Ma in Ellsworth Land. Although Mount Erebus (Ross Island) and Mount Melbourne [Wörner and Viereck, 1989] are the only demonstrably active volcanoes in the rift system, 15 others are suspected of Holocene activity [LeMasurier, 1990]. The distribution of ages suggests therefore that the rift system has propagated gradually toward the Bellingshausen Sea-Antarctic Peninsula, while remaining active at the Ross Sea end and that no part of the rift system is inactive.

AEROMAGNETIC SURVEYS

Aeromagnetic survey flights in the late 1950s and early 1960s [Behrendt and Wold, 1963; Behrendt, 1964a; Behrendt et al., 1991b] are still the only data available in much of West Antarctica. In 1978, aeromagnetic surveys were combined with radar ice sounding [Behrendt et al., 1980; Jankowski et al., 1983; Drewry, 1983] (Figure 4) over part of the West Antarctic rift system and the Middle Jurassic Dufek intrusion. In this paper we combine these data as published by Drewry [1983] with an earlier compilation [Behrendt 1964a] to produce the statistical distribution map shown in Figure 4.

Evidence for Cenozoic Volcanic Rocks Beneath Ice

The obvious correlation of high-amplitude short-wavelength magnetic anomalies with outcrops of late Cenozoic volcanic rocks [Behrendt et al., 1991b] in the Marie Byrd Land area (Amundsen-Bellingshausen flank) (Figures 2 and 4) has long been noted [Behrendt and Wold, 1963; Behrendt, 1964a]. The prominent (>1000 nT) linear magnetic "Sinuous Ridge" anomaly was interpreted as caused by volcanic rocks of unknown age [Jankowski et al., 1983] (Figure 2). We interpret most anomalies in the area covered within the closed "2" contour of Figure 4 over the ice-covered area of Marie Byrd Land (Byrd Subglacial Basin) where depths to sources are <1 km below the base of the ice [Behrendt and Wold, 1963; Jankowski et al., 1983] to be the result of late Cenozoic volcanism similar to that of the volcanic ranges of Marie Byrd Land (Figure 2), although some are probably caused by older shallow sources.

The character of the magnetic field changes abruptly, as marked by the north trending part of the ">2" contour in Figure 4a, about halfway between the outcropping volcanoes of Marie Byrd Land and the Ellsworth Mountains (Figures 4 and 5, a representative profile). This "break" first described by Behrendt and Wold [1963], is quite apparent in other profiles [e.g., Behrendt, 1964a; Drewry, 1983; Behrendt et al., 1991b]. Seismic

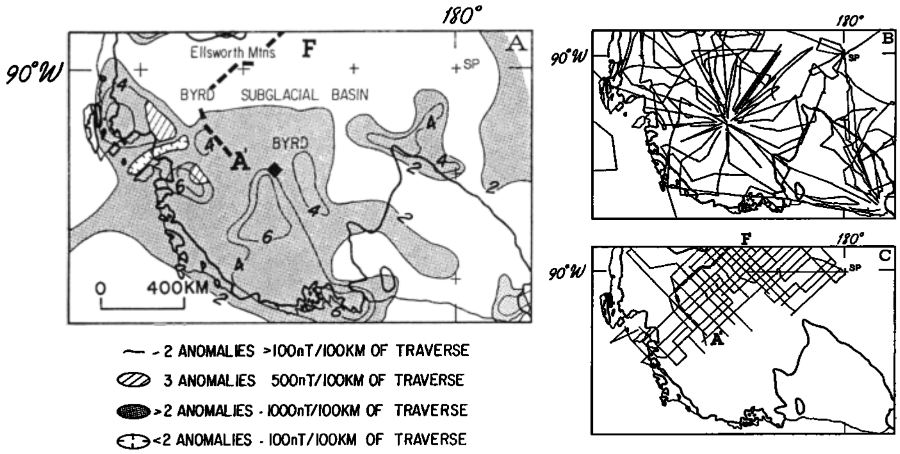

Fig. 4. (a) Frequency of occurrence of short wavelength high amplitude anomalies. Contours show number of anomalies greater than 100 nT/100 km of flight line averaged for a 1° of latitude grid (111 x 111 km). Modified from Behrendt [1964a] using additional data collected in 1978 [Jankowski et al., 1983; Drewry, 1983]. The location of profile A'-F (Figure 5) is indicated. The magnetic "break" is defined by the north trending >2 contour in the Byrd Subglacial Basin. Byrd Station is 80°S, 120°W.
(b) Locations of flight lines used by Behrendt (1964a).
(c) Location of flight lines collected in 1978 at approximately 100 km spacing combined with radar ice sounding [Jankowski et al., 1983; Drewry, 1983]. Location of profile A'-F (Figure 5) is indicated by heavy dashed line. Grid north is at top of a) b) and c).

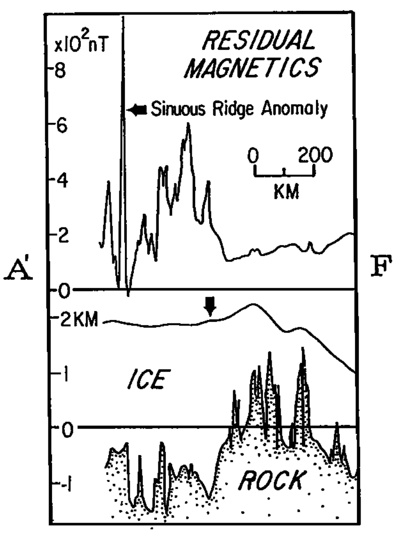

Fig. 5. Radar ice sounding and aeromagnetic profile across inferred extended crust of Byrd Subglacial Basin to near Ellsworth Mountains [from Drewry, 1983]. Vertical exaggeration of topography is 200:1. The arrow marks the interpreted rift shoulder and also the "break" in magnetic anomalies. The magnetic "break" (also Figure 4) separates a region of numerous high-amplitude anomalies, interpreted as caused mostly by late Cenozoic volcanic rocks as sources (90% $\leq$ 1 km below bedrock surface) contrasted with nonmagnetic rock (probably metasedimentary because of exposures in the Ellsworth Mountains). The ~1000-nT Sinuous Ridge anomaly is located in Figure 2. The location of the profile is indicated in Figure 4.

refraction data, discussed later, support the shallow magnetic depth and volcanic source interpretation. The absence of shallow magnetic sources on the Ellsworth Mountains side of the "break" in Figure 4 is good evidence for the lack of Cenozoic volcanic rocks beneath the ice sheet there. The area towards the Ellsworth Mountains from the magnetic "break" in Marie Byrd Land mostly contains nonmagnetic rock as thick as 5 km, with individual depth determinations as great as 7-11 km [Jankowski et al., 1983]. Behrendt and Wold [1963] showed a broad regional negative magnetic anomaly over the Byrd Subglacial Basin which was generally correlated to topography. Possibly this is the result of thermally demagnetized rock caused by regionally high heat flow in the West Antarctic rift.

A profile across part of the West Antarctic rift and rift shoulder (Figure 5) illustrates the magnetic "break" between the region of inferred shallow Cenozoic volcanic rock and the region underlain by thick nonmagnetic rocks which probably extends into the sedimentary Ellsworth Mountains. Figure 6 illustrates that the magnetic "break" appears to approximately mark the inferred edge of the rift shoulder (arrow) at least on this profile.

Behrendt [1964a, b]; Behrendt et al. [1974]; Behrendt et al. [1980, 1991a, b]; Dalziel et al. [1987]; Garrett et al. [1987; 1988], and Bosum et al. [1989] discuss magnetic anomalies over the Transantarctic Mountains (including the Dufek intrusion), Whitmore Mountains area, and southern Antarctic Peninsula. The interpreted intrusive and extrusive magnetic rocks inferred to cause the anomalies range from Precambrian (Haag Nunataks, Figure 2) [Garrett et al., 1987; Maslanyj and Storey, 1990] to Middle Jurassic [e.g., Behrendt et al., 1980] in both the southern Antarctic Peninsula and Transantarctic

Mountains. We cannot differentiate on the basis of magnetic data alone between pre-Cenozoic igneous rocks and Cenozoic volcanic rocks. There are, however, late Cenozoic volcanic outcrops in the southern Antarctic Peninsula (Figure 2). Note the position of several linear anomalies each (about 1000 nT in intensity) indicated in Figure 2, similar to the "Sinuous Ridge" anomaly [Jankowski et al., 1983] in the Byrd subglacial Basin, which we suggest was caused by the rift-related Cenozoic(?) volcanism [Behrendt et al., 1991b]. Maslanyj and Storey [1990] reported recent British Antarctic Survey (BAS) data (generally consistent with Figure 4, although not included) over the Ellsworth-Whitmore Mountains, southern Antarctic Peninsula, and Weddell Embayment, mostly out of the area of this paper.

ROSS SEA SHELF-NORTHERN VICTORIA LAND AEROMAGNETIC SURVEY

Widely spaced magnetic profiles (Figure 4) over the Ross Sea Shelf and adjacent part of the Transantarctic Mountains (Figure 6) [Behrendt, 1964a; Behrendt et al., 1987] show numerous high-amplitude (300-2000 nT), short-wavelength anomalies over marine and ice-covered areas. Although Jurassic bodies several kilometers thick such as the Dufek intrusion [Behrendt et al., 1974, 1980] or the Kirkpatrick basalts [Behrendt et al. 1991b] produce high-amplitude anomalies, Ferrar-dolerite sills, (although quite magnetic) because they are essentially thin sheets, produce only small anomalies (< 100 nT) at their edges [Pederson et al., 1981; Behrendt et al., 1991b]. We infer that the high-amplitude, shallow source, short-wavelength anomalies over the Ross Sea Shelf (Figure 5) and grounded ice at its inland end are predominantly caused by late Cenozoic volcanic rocks (having substantial vertical thickness) [Behrendt, 1964a; Robinson, 1964; Pederson et al., 1981; Behrendt et al., 1991b].

A 4.4 km x 22 km spaced aeromagnetic survey was flown over northern Victoria Land and the southwestern Ross Sea [Bosum et al., 1989; Behrendt et al., 1991a, b]. These data [BGR-USGS, 1987], contoured at a 5-nT interval (Plate 1), illustrate features which include (1) the numerous small circular anomalies interpreted as submarine, late Cenozoic volcanic edifices and subvolcanic intrusions; (2) the 1700 nT, 200 km long, Polar 3 anomaly (centered about 74°S 168°E); and (3) the 80 to -100 nT anomaly over the Victoria Land basin.

Ten anomalies having short (1-5 km) wavelengths and high (100 to more than 1000 nT) amplitudes were found along seismic reflection profiles using magnetic gradiometer data (e.g., Figure 7), and are attributed to late Cenozoic volcanic flows and subvolcanic intrusions

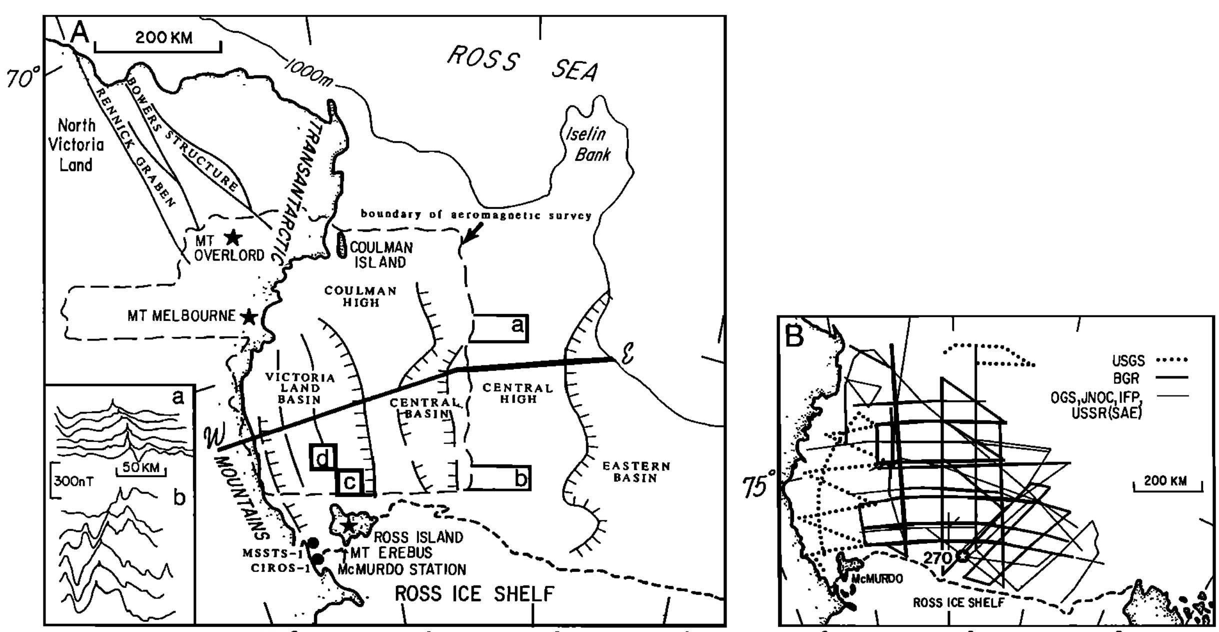

Fig. 6. (a) Index map of structures and geographic locations discussed in the Ross Sea shelf- North Victoria Land area. Active volcanoes Mt. Erebus and Mt. Melbourne and late Cenozoic volcano Mt. Overlord (Pleistocene?) are indicated. Boxes a and b are location of stacked magnetic profiles a and b at lower left of this figure. (Tics indicate anomaly correlations between profiles). (Structures from Davey et al. [1982, 1983] and Cooper et al. [1987a]). Locations of anomalies C and D [Behrendt et al., 1987] in Figure 8, MSSTS-1 and CIROS 1 core holes, and approximate location of cartoon (Figure 10) profile W-E are indicated. 1000 m bathymetric contour is shown. (b) Locations of seismic reflection profiles over the Ross Sea shelf. The USGS and BGR profiles (also indicated in Plate 1) are used in interpretations in this paper. Other profiles: Institute Francaies du Petrole (IFP), France, Japanese National Oil Company (JNOC), Japan Soviet Antarctic Expedition (SAE), USSR; Oservetoreo Geofisico Sperimentale (OGS), Italy. DSDP core hole 270 is indicated.

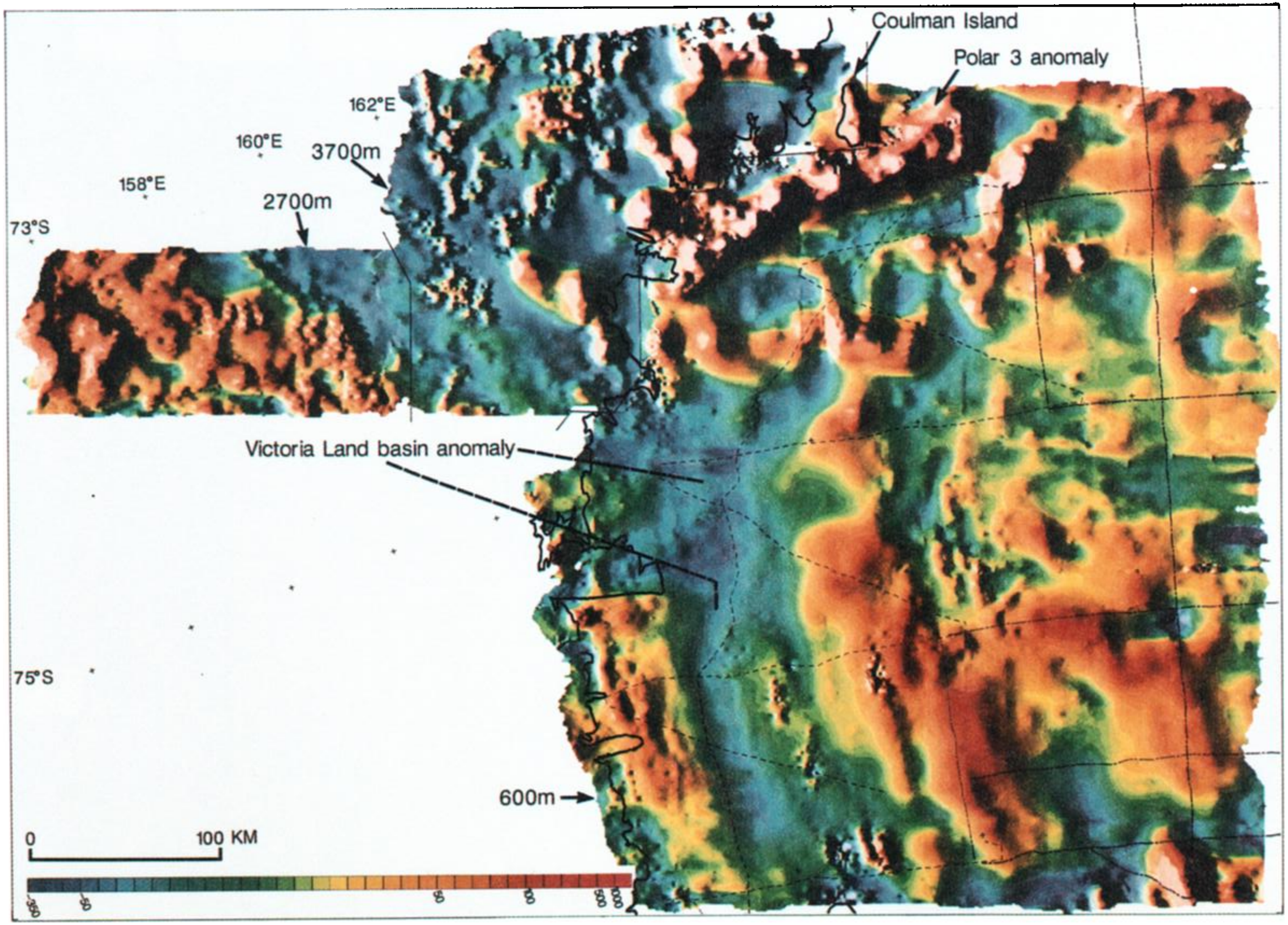

Plate 1. Color shaded relief (apparent sun angle 15° from west) aeromagnetic map of western Ross Sea shelf (right half of figure) and part of northern Victoria Land (top-central and left part of figure). Location of survey is shown in Figure 6. Areas flown at different elevations are separated by heavy solid line segments: 600 m Ross Sea, 3700 m northern Victoria Land, 2700 m East Antarctic ice sheet at west (far left). Flight lines were spaced 4.4 km (E-W) (some at 8.8 km) by 22 km (N-S). This map was compiled using an 800 x 800 m grid. Locations of marine seismic reflection profiles collected by USGS *S.P. Lee* in 1984 (short dash lines) and BGR *Explora* in 1980 (dash-dot lines) are indicated. The north tip of Ross Island is present at the south edge of the map; the coastline is indicated by the solid line. Contour interval is 5 nT between +50 and -50 nT and compressed as indicated in color bar beyond this range. The Polar 3 anomaly probably caused by late Cenozoic volcanic rock and a subvolcanic (mafic?) intrusion. The -80 to -100 nT north trending anomaly 80 to 100 km wide at the southwest corner of the map overlies the Victoria Land Basin. The elongate western area of magnetic coverage (2700 m) overlies the East Antarctic ice sheet. The major northwest trending break in magnetic character at the right of this area probably marks the east edge of the buried Precambrian shield.

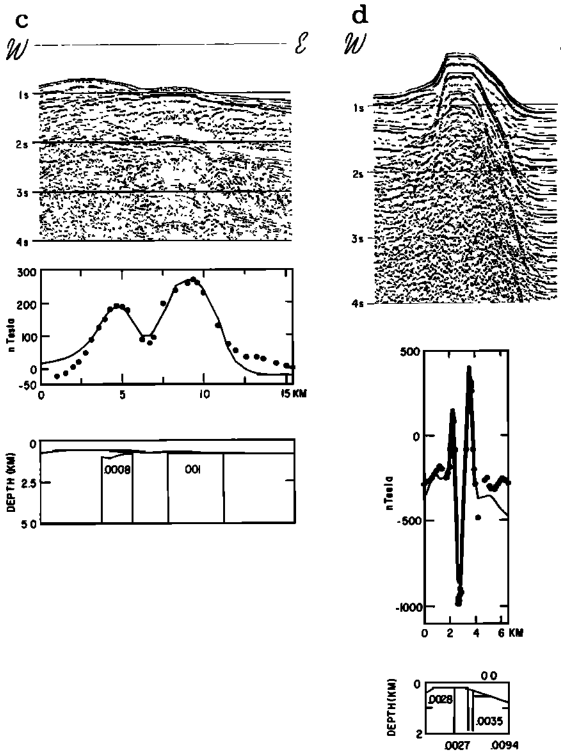

Fig. 7. Seismic reflection profiles (time in seconds indicated) compared with observed (dots) and theoretical (solid line) marine magnetic profiles from Behrendt et al. [1987]. Locations are shown in Figure 6. Apparent susceptibilities of theoretical models shown in cgs units. Profile C (anomaly C [Behrendt et al., 1987]): Sources interpreted as late Cenozoic subvolcanic intrusions. The smooth submarine topography across anomaly C probably indicates that former volcanic edifices were removed by glacial erosion; therefore the intrusion is younger than the Neogene sedimentary rock but older than latest deglaciation. Profile D (anomaly D [Behrendt et al., 1987]): The prominent submarine volcano is probably Holocene; i.e., the edifice was extruded on the seafloor since deglaciation.

[Behrendt et al., 1987] that penetrate a thick (up to 14 km) sedimentary section. Based on theoretical magnetic models compared with seismic reflection profiles (Figure 7) for several of the very short wavelength (1-3 km) anomalies along ship tracks, we infer that causative bodies for many of the about 100 small anomalies (Plate 1) defined by the aeromagnetic survey [Behrendt et al., 1991a] penetrate from the basement through the sedimentary section. Depth estimates indicate that the tops of the sources are essentially at the sea floor and probably would have topographic expression like those in Figure 7 were seismic or bathymetric profiles measured over them. These penetrating bodies have been defined by available seismic reflection data only in the Victoria Land basin area.

The small magnetic anomalies do not occur randomly over the marine area but are concentrated along linear zones mostly parallel to the Victoria Land basin (Figure 6 and Plate 1). We interpret this lineation as evidence of rift fabric. Note the north-northwest "grain" over the Ross Sea part of the survey. The zone of lineations of north-northwest trending anomalies observed in Plate 1 can be extended nearly to the edge of the Eastern Basin (Figure 6A) where additional east-trending profiles are shown.

The Polar 3 anomaly has sharp peaks similar to those over late Cenozoic volcanic rocks exposed on Coulman Island (Figure 6 and Plate 1) and thus we interpret the source as an intrusive and extrusive complex [see Bosum et al., 1989; Behrendt et al., 1991b]. Possibly the Polar 3 anomaly is related to transfer faults tectonically [Wörner et al., 1989; Behrendt et al., 1991b]. The Polar 3 anomaly is similar in amplitude and wave-length ($\geq$1000 nT, 200+km) to other anomalies indicated in Figure 2 such as the "Sinuous Ridge" (Figures 2, 4, and 5) [Jankowski et al., 1983], in the ice-covered areas of the West Antarctic rift system in Marie Byrd Land.

The source of the -80 to -100 nT anomaly over the Victoria Land basin [Behrendt et al., 1991b], inferred to be filled with a 14-km-thick sedimentary sequence [Cooper et al., 1987a], has been interpreted in two ways. One model fit to a profile across this anomaly suggests a deep (about 12-14 km) magnetic basement overlain by nonmagnetic sedimentary rock [Behrendt et al., 1987, 1991a]. A second model fit to the same profile is interpreted to represent a shallow Curie isotherm resulting from late Cenozoic rifting and a thinner (6-8 km) sedimentary section overlying rock possibly consisting of thermally demagnetized volcanic flows [Behrendt et al., 1991a, b] caused by rift related high heat flow.

BOUGUER GRAVITY ANOMALIES

Figure 8 is a compilation of Bouguer anomalies from all available sources [Behrendt et al., 1991b].

Thinned Crust Beneath Ross Embayment-Byrd Subglacial Basin

One of the largest-magnitude Bouguer anomalies in the world was recognized as early as 1960 [Bentley et al., 1960] as marking what we now interpret as the West Antarctic rift shoulder (Figure 8). Bouguer anomalies increase by about 200 mGal from about -150 mGal over the Transantarctic Mountains to about +50 mGal over the inferred extended crust beneath the Ross Ice shelf and Byrd Subglacial Basin. From magnetic depth estimates (discussed above) and limited seismic refraction data (next section) we know that normal basement density rocks are probably quite shallow (< 1 km) beneath the ice in the Byrd Subglacial Basin. Because there are no gravity data on the outcrops of the Ellsworth Mountains, we cannot determine a maximum range and gradient across this part of the rift shoulder.

Bouguer anomalies tied to seismic determinations of 17- to 21-km Moho depth beneath the Ross Sea Continental shelf (discussed below) provide the primary evidence for extension of the crust throughout the ice-covered Ross Embayment Byrd Subglacial Basin area of the West Antarctic rift system. Early crustal thickness maps based on Bouguer anomalies showed approximately 30-km-deep

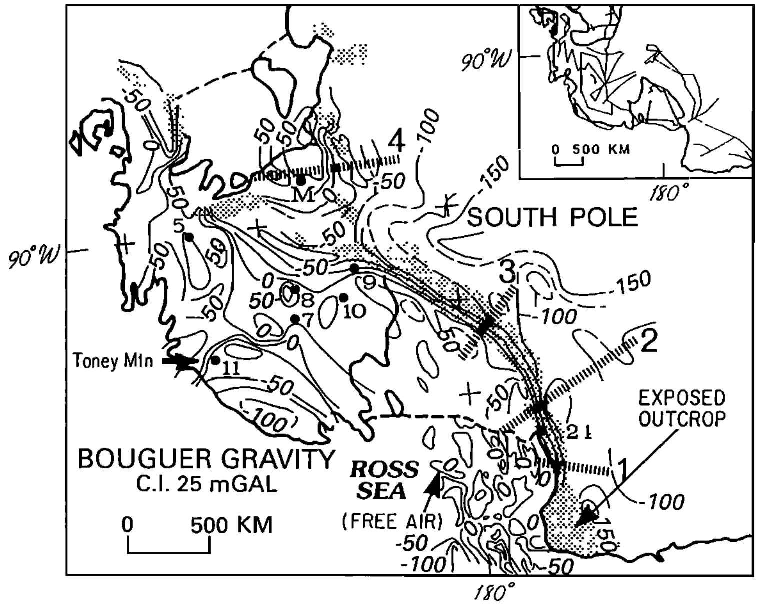

Fig. 8. Bouguer anomaly contour map for West Antarctica compiled from data collected at seismic reflection stations (about 30-40 km spaced) where ice thickness measurements were made by oversnow traverse parties led from 1956 to 1964 by J. C. Behrendt, C. R. Bentley, F. K. Chang, A. P. Crary, M. Hochstein, E. S. Robinson, E. Thiel, and F. Van der Hoeven. The inset map shows locations of these oversnow traverses. Additional data: Ross Ice shelf from the map of Robertson et al. [1982]. The free air anomaly map over the Ross Sea shelf is from Davey and Cooper [1987]. The numbered bands indicate areas where data density allowed reasonably accurate calculation of gradients: 1, Duerbaum et al. [1989] 2, Robinson [1964]; Smithson [1972]; Robinson and Splettstoesser [1984]; 3, Robinson [1964]; Robinson and Splettstoesser [1984]; 4, Behrendt et al. [1974]. Seismic velocity columns (Figure 11), 11 (at Toney Mountain), 7 (Byrd Station), 8 (Whitmore Mountains), 5 (Ellsworth Mountains) 24 (Ferrar dolerite near McMurdo) 10 and 9 (Horlick Mountains) from Bentley and Clough [1972] are indicated. M is Moho reflection (work of M. Hochstein, 1963-1964, as discussed by Bentley [1973])

Moho over the area of the West Antarctic rift (Byrd Subglacial Basin-Ross Embayment) in contrast with the approximately 40-km-deep Moho across the rift shoulder [e.g., Bentley et al., 1960], calculated solely from the difference in Bouguer anomaly. In this paper we interpret the data to indicate that the base of the crust beneath the rift (Byrd Subglacial Basin-Ross Embayment) is probably closer to 20 km depth, typical of rift stage crust [e.g., Klitgord et al., 1988], by adjusting the Bentley et al. [1960] regional interpretation made with many fewer gravity data than Figure 8 to the seismic Moho determinations beneath the Ross Sea continental shelf.

Rift Shoulder Anomaly

Many authors [e.g., Robinson, 1964; Smithson, 1972; Behrendt et al., 1974; Robinson and Splettstoesser, 1984; Davey and Cooper, 1987; Stern and ten Brink, 1989] have computed theoretical models and explained the great change in gravity across the front of the Transantarctic Mountains as a result primarily of an abrupt 10- to 20-km change in crustal thickness. If we assume, as a first approximation an infinite slab having a density contrast of 0.4 g/cm^3 at the crust-mantle boundary to account for the entire 200-mGal change, then the change in crustal thickness, using 41.85 mGal/km for a density contrast of 1.0 g/cm^3, is 12 km. Were the 200 mGal difference to be caused solely by a difference of crustal thickness as great as 20 km, a density contrast of 2.4 g/cm^3 would be implied. There is a significant difference between the maximum 200-mGal change across the front of the Transantarctic Mountains near McMurdo (profile 2, Figure 8) and the maximum 130-mGal anomaly range at the front of the Transantarctic Mountains in the Pensacola Mountains area (profile 4, Figure 8) [Behrendt et al., 1974].

Bouguer Anomaly Gradient Across Transantarctic Mountain Front

We examined the steepest gravity gradients measured across outcrops of Paleozoic or older sedimentary or crystalline rock at the front of the Transantarctic Mountains. In the McMurdo area the Bouguer anomaly gradient ranges from 4 to 7 mGal/km and, based on fewer data, appears to be as steep as 4 mGal/km in the Beardmore Glacier area [Robinson, 1964; Robinson and Splettstoesser, 1984]. Two-dimensional gravity models have been computed across the front of the Transantarctic Mountains in the Beardmore Glacier and McMurdo areas of the Ross Embayment; two in the McMurdo area are shown in Figures 9b and 9c taken from along profile 2,

Figure 8. A common feature of these models is a steep or near vertical 10- to 20-km step in the Moho, interpreted generally as a fault extending to the surface at the front of the Transantarctic Mountains, and a high-density mass (e.g., Figure 9c) within the interpreted extended crust at this boundary.

In contrast, the lower 2-mGal/km gradient observed across the 130-mGal gravity change in the Pensacola Mountains section of the Transantarctic Mountains was fit with an approximate 15°-20° dip on the assumed Moho [Behrendt et al., 1974] (Figure 9a) tied to a 24-km seismic reflection depth to Moho (Figure 8). See additional discussion by Behrendt et al. [1991b].

The Bouguer values across the Amundsen-Bellingshausen flank of the West Antarctic rift system decrease to about -100 mGal from the +50-mGal Bouguer anomaly over the Byrd Subglacial Basin. The gradient does not appear as steep over this flank of the rift nor the crust as thick beneath it as across the Ellsworth-Whitmore side of the rift, which again demonstrates the asymmetry of the West Antarctic rift system.

Ross Sea Basins

Davey and Cooper [1987] required high density basin fill (2.7 g/cm^3) below 5-7 km to fit the gravity anomaly over the 14-km-thick section in the Victoria Land Basin (Figures 6, 10 and 11). This suggests that a substantial part of the deeper section could be high-density basalt flows contrasted to shallower lower density sedimentary rock (supported by seismic results discussed below), possibly similar to those observed in the Midcontinent Rift system of North America [Behrendt et al., 1990]. The high-density mass beneath the other Ross Sea basins was also shown by Hayes and Davey [1975].

The Central Basin is marked by a positive gravity anomaly which seismic interpretations [Tréhu et al., 1989 and manuscript in preparation] of line 2 (Figure 10) indicate can be explained by high density basin fill (basalt?) or more likely by a high-density, high-velocity rift cushion at the base of the crust.

SEISMIC INVESTIGATIONS

Large Offset Surveys: Ice Sheet

Refraction results obtained for the upper crust collected on the oversnow traverses at about 33 sites in the area of Figure 2 have been summarized [Bentley and Clough, 1972; Rooney et al., 1987]. All data were collected on the ice sheet, which has a seismic velocity of 3.9 km/s; therefore evidence of lower-velocity sedimentary rock would not have been observed as a first arrival although it was inferred indirectly in several cases [Bentley and Clough, 1972; Behrendt et al., 1974; Rooney et al., 1987]. Behrendt et al. [1991b] reviewed these results in some detail; we only summarize them here.

The high velocities (> 6.7 km/s) within the shallow crust (Table 1) are similar to a few measured in McMurdo Sound [McGinnis et al., 1985; Kim et al., 1986] and beneath the Ross Sea shelf [Cooper et al., 1987b; Tréhu et al., 1989] discussed below, which we interpret to be associated with mafic magma intruded during rifting.

However, because the Horlick Mountains are located where the Jurassic Transantarctic rift and the Cenozoic West Antarctic rift shoulder are coincident, the high seismic velocities could be the result of either Jurassic or Cenozoic intrusions or both. However, column 8 (Figure 8, Table 1) overlies the interpreted West Antarctic rift well away from the possible rocks of the Jurassic

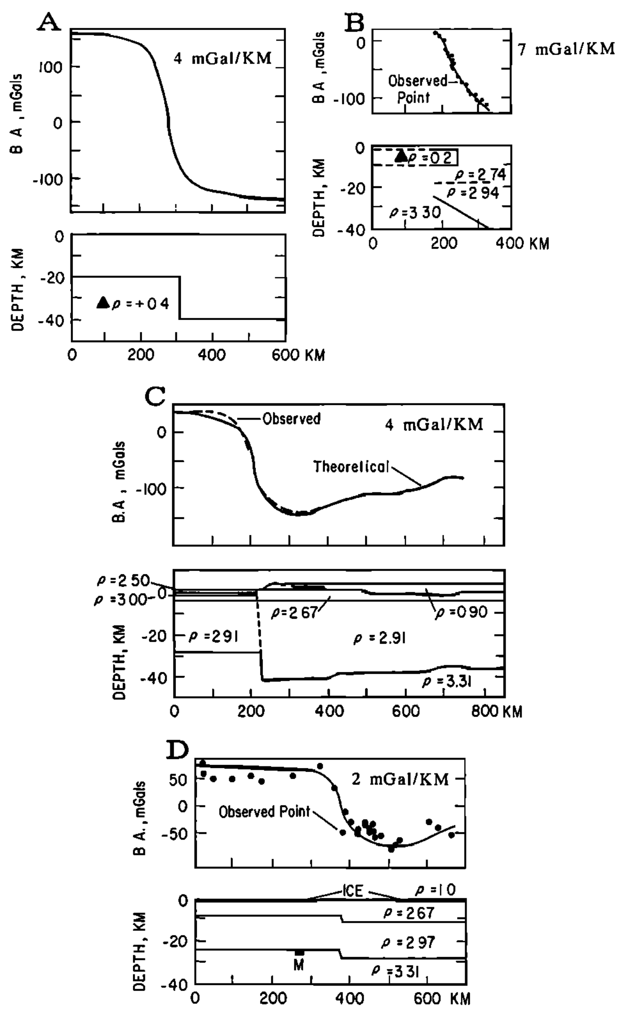

Fig. 9. Theoretical and observed Bouguer anomaly profiles for several models, maximum gradients are indicated. (A) Model for an assumed crustal thickness change of 20 km from 20 to 40 km with a simple density contrast of 0.4 g/cm^3 across an assumed Moho. The maximum gradient is 4 mGal/km, and the total anomaly range is about 330 mGal (contrasted with actual observed maximums of about 200 mGal). This type of model cannot explain the observed data in profiles B, C and D here and in Figure 8 but is required to produce the steep gravity gradient solely by a change in crustal thickness. (B) profile of Smithson [1972] approximately along 2 in Figure 8; (C) profile of Robinson [1964]; and Robinson and Splettstoesser [1984] approximately along 2 in Figure 9; (D) Behrendt et al. [1974], approximately along 4 in Figure 8 across the Pensacola Mountains. M is Moho reflector (work of M. Hochstein, 1964 as discussed by Bentley [1973]).

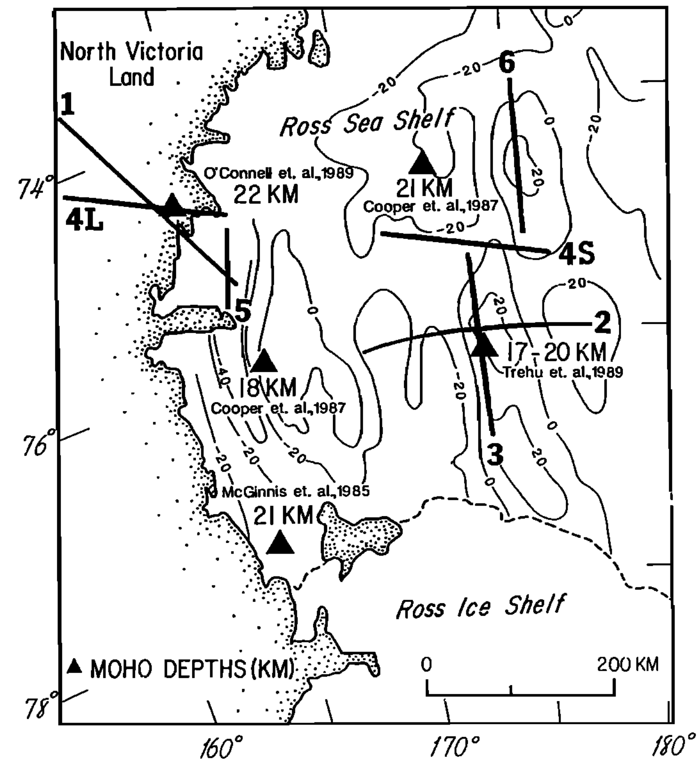

Fig. 10. Locations of large-offset seismic experiments along profiles in Ross Sea shelf-Victoria Land area. Profiles 1-6 are from the 1988-89 GANOVEX V expedition. Profile 3 is along the strike of the Central Basin (Figure 6). Generalized free air gravity contours at a 20 mGal interval are from Davey and Cooper [1987].

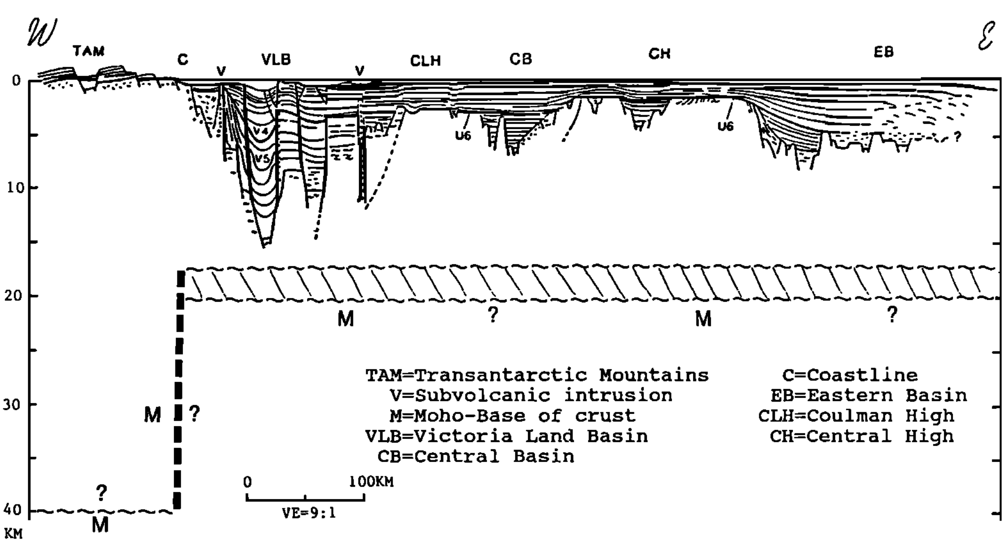

Fig. 11. Cartoon illustration of geologic structure across the Ross Sea shelf modified from seismic reflection interpretation of Cooper et al. [1990] for upper crust based on preliminary seismic crustal thickness determinations from GANOVEX V (Figure 11). Unconformity U6 [Hinz and Block, 1983] is correlated with that between acoustic units v4 and v5 [Cooper et al., 1987a]. The Moho is shown as a horizontal band from 18 to 20 km to indicate uncertainty in its depth. The approximate location of this profile is indicated by W-E in Figure 6. The vertical dashed line indicates that Moho probably changes thickness by up to 20 km at rift shoulder (Transantarctic Mountains front), however, the dip of the Moho in the transition zone is unknown, but is probably not vertical. Note that the old crust has been nearly rifted through beneath the Victoria Land basin.

Table 1. Seismic Refraction Velocity Columns From Bentley and Clough [1972]

Column	Remarks
11	Toney Mountain, 4-6 km/s andesite basalt flows over 6.1 km/s metamorphic basement
7	Byrd Station, 4.3 km/s, magnetic volcanic rock over 5.9 km/s basement
10	Near rift shoulder, 6.7 km/s about 2 km below sea level
9	Near Horlick Mountains, near rift shoulder, 7.0 km/s ~0.5 km below sea level
8	Over rift, 5.8 km/s basement over ~7 km/s refraction about 4 km below sea level
24	Ferrar dolerite sill in Transantarctic Mountains, 6.8 km/s

Locations on Figure 8.

Transantarctic rift, suggesting that Cenozoic mid and lower crustal mafic intrusions may be the source of the 7 km/s velocity there. In contrast to these high shallow or midcrustal velocities is the comparatively low 6 km/s mean (subice) crustal velocity measured by M. Hochstein [Bentley, 1973; Behrendt et al., 1974].

Large Offset Surveys: Ross Sea

McGinnis et al. [1985], using land techniques on sea ice, report a 7.2- km/s refractor in the lower crust beneath McMurdo Sound which is reasonable for stretched underplated and intruded lower crust. Furlong and Fountain [1986] discuss examples of such crust having seismic velocities of 7.0-7.8 km/s. Kim et al. [1986] show bands of reflections and diffractions observed in McMurdo Sound as late as about 7.8 s beneath the Victoria Land Basin (Figure 6), which are quite similar in appearance to reflections from probable underplated material reported for other rifts [e.g., Behrendt et al., 1990].

Velocities were reported from sonobuoys for the sedimentary rocks in the three basins [Davey et al., 1983], beneath the Ross Sea Shelf (Figures 6 and 11). These ranged from 1.7 to 4 km/s in the upper few kilometers of the Victoria Land Basin but increased to as great as 5.6 km/s in the lower 6-8 km of this 14-km thick basin [Cooper et al., 1987a]. Because these velocities are significantly higher than the 4.4 km/s measured for the Devonian to Jurassic Beacon Supergroup (e.g., Column 10, Table 1, Figure 8) and the 4.6 km/s measured for Cenozoic volcanic rocks at Toney Mountain (Figure 8), they probably are rift-related volcanic rocks of pre-early Oligocene [Behrendt et al., 1991b] age.

The highest crustal velocities measured in the western Ross Sea [Cooper et al., 1987b] are 7.3 km/s at 8.3 km depth beneath the east flank of the Victoria Land Basin and 7.4 km/s at 3.0 km depth, 7.4 km/s at 3.5 km depth, and 6.9 km/s at 2.8 km depth beneath the Coulman high (Figure 6). Cooper et al. [1987b] reported a velocity of 6.5 km/s as acoustic basement from within the central region of the Victoria Land Basin. We infer that velocities of 6.9-7.4 km/s probably indicate mafic or ultramafic intrusions typical of underplated crust [e.g., Furlong and Fountain, 1986] at quite shallow depths beneath the Ross Sea which is unusual. These rocks may be the source of the high densities required within the crust to fit the steep observed gravity gradients (Figure 9) and total Bouguer anomaly range and are additional evidence of rifted extended crust. Tréhu et al. [1989 and manuscript in preparation, 1991] observed a high-velocity (>7.2 km/s) lower crustal transition zone (rift cushion) beneath the Central Basin.

Moho Determinations: Ross Sea Shelf

Cooper et al. [1987b] report deep reflections at two sites (Figure 10) which we interpret as Moho because they have depths of 18 and 21 km. BGR and USGS (German Antarctic North Victoria Land Expedition V, 1988-1989) conducted a large-offset seismic experiment on land and at sea using a 43.5 L airgun source (Figure 10). From these results Tréhu et al. [1989 and manuscript in preparation] reported Moho depths of 17-21 km (typical of extended rifted crust [Klitgord et al., 1988]) beneath the Central basin and Coulman High, and O'Connell et al. [1989] reported a Moho depth of 22 km on land near the coast. A similar depth to the Moho of 21 km beneath McMurdo Sound was reported by McGinnis et al. [1985]. All of these Moho depths are consistent with each other and approximately consistent with the earlier gravity models of Davey and Cooper [1987], which suggested attenuated (rifted) crust beneath the basins. In the Bouguer anomaly section we concluded, by comparison with these measured depths to the Moho beneath the Ross Sea shelf, that the Moho depths throughout the topographically low area of the rift (Ross Embayment-Byrd Subglacial Basin) are more likely closer to 20 km than the 30 km as previously reported [e.g., Bentley et al., 1960].

Marine Multichannel Seismic Reflection Surveys

Since 1980, approximately 35,000 km of marine common-depth-point reflection profiles have been acquired in the Ross Sea (Figure 6); the USGS and BGR data have been described by Hinz and Block [1983], Cooper et al. [1987a], and Cooper et al. [1991]. These authors identify up to seven acoustic units and unconformities within the reflection section, which they interpret to be of late Mesozoic and younger age. Acoustic basement may be composed of metasedimentary and/or igneous units of Precambrian to late Mesozoic age. A schematic interpretation of these data, which Behrendt et al. [1991b] discuss at greater length, along one profile across the three basins in the Ross Sea is shown in Figure 11.

Significance of Ross Sea Reflection Results to Study of the West Antarctic Rift System

1. These seismic reflection data (tied to the sparse drill holes) across the part of the rift underlying the Ross Sea shelf provide the only age control for the ice-covered and sea-covered parts of the rift beneath the Ross Embayment and Byrd Subglacial Basin.

2. Rifting in the Ross Sea, and likely throughout the West Antarctic rift system, has been episodic; the majority of basement downfaulting and brittle crustal deformation occurred prior to development of unconformity U6 (Eocene to Oligocene time), and probably in Late Mesozoic time [Cooper et al., 1991].

3. The interpretation of horst and graben structures in the Ross Sea (Figure 11), probably can be extended beneath the Ross Ice Shelf and the Byrd Subglacial Basin.

4. The orientation of the structural basins underlying the Ross Sea shelf defined by seismic reflection results are subparallel to the "grain" of the West Antarctic rift system (e.g., the Transantarctic Mountains shoulder and Figure 2 and Plate 1).

5. Together with the aeromagnetic data (Plate 1) the seismic reflection data provide the best available geophysical evidence for concealed submarine or subglacial Late Cenozoic volcanic structures in the West Antarctic rift system.

DISCUSSION

The West Antarctic rift system seems unusual compared with other rift systems: (1) Even after allowing for isostatic adjustment after ice removal, the elevation is anomalously low throughout most of the area (Figure 2). (2) Although earthquakes in rifts generally appear to be small in magnitude [Strecker and Bosworth, 1991] Antarctica is nearly aseismic [Kaminuma, 1982; Adams et al., 1985]. However, considering the following explanations, it is more similar to other rifts than first appears.

Anomalous Low Elevations

Various lines of evidence [Behrendt et al., 1991b] support a thin, hot, very weak [Stern and ten Brink, 1989], dense extended lithosphere which is subsiding rapidly. By comparison with the Bouguer anomaly map (Figure 8) we infer that similar crust and lithosphere is characteristic of the entire West Antarctic rift system. The Byrd Subglacial Basin area of the rift is filled with about 3 km of ice at about 1.5 km snow surface elevation. If the ice were compressed to low density Neogene sedimentary rock, the surface elevation would be about sea level for the same mass load; this elevation should be compared with other rifts rather than the approximate mean 1.5 km depth of the base of the ice because the ice is isostatically compensated [Bentley, 1983]. Other areas with consistently low elevation during active rifting include the North Sea and the Midcontinent Rift system of North America.

Low Seismicity

Although reasonably high quality seismographs have operated continuously since the International Geophysical Year (1957-1958), Antarctica appears largely aseismic [Adams et al., 1985]. Bentley [1983] considers it likely that three >4 magnitude earthquakes were associated with fault zones that control the location of glaciers (e.g., the Rennick graben, Figure 6) [Roland and Tessensohn, 1987]. Low seismicity associated with the West Antarctic rift system [Behrendt et al., 1991b] is primarily the result of sparse seismograph coverage and secondarily the result of possible suppression by ice [e.g., Johnston, 1987] or very high seismicity in the Ross Embayment immediately after deglaciation followed by low seismicity at present [e.g., Muir Wood, 1989]. It probably would not be possible to detect the Rio Grande rift in the western United States solely from seismograph coverage typical of that in Antarctica.

Mechanisms for Rifting

Active extension in the West Antarctic rift system since at least early Oligocene [Barrett, 1989] is implied by the volcanic activity (Figure 2) and rapid uplift of the rift shoulder [Behrendt and Cooper, 1991]. Extension and basin downfaulting was probably even greater in late Mesozoic time [Fitzgerald et al., 1986; Cooper et al., 1987a; Behrendt et al., 1991b]. The New Zealand-Campbell plateau block broke away from Marie Byrd Land (about 80 Ma) [Bradshaw, 1989] when the major Ross Sea shelf grabens were likely downfaulted in the late Mesozoic. Uplift of the Transantarctic Mountains commenced about 60 Ma [Fitzgerald, 1989].

Because Antarctica is essentially surrounded by mid-ocean ridges, the only apparent source for stress is ridge push [Okal, 1981], yet it seems unlikely that this could be the source of the extension. Smith and Drewry [1984] proposed that the rise of the Transantarctic Mountains is a delayed effect caused by overriding by East Antarctica of a hot spot that formed under West Antarctica in Late Cretaceous time, resulting in phase changes in the upper mantle leading to uplift. This model does not address the asymmetry [Fitzgerald et al., 1986] in the Ross Embayment. However, we suggest that the asymmetry is the result of different flexural rigidity of the lithosphere bordering the rift on the rift shoulder side from that of the Amundsen- Bellingshausen flank along the lines of the discussion by Stern and ten Brink [1989].

Fitzgerald et al. [1986] proposed a simple shear model for uplift of the Transantarctic Mountains and extension observed in the basins beneath the Ross Sea shelf and estimated about 200 km (25-30%) of extension assuming an average of 25-30 km crustal thickness. The large-offset seismic results obtained over the Ross Sea shelf (Figure 10) indicate that the crust beneath the Ross Sea shelf may be only 17-21 km thick requiring even greater extension. If, for example, a 40-km-thick crust were stretched to 20 km across the presently 750-km-wide rift during late Mesozoic-Cenozoic time, about 350-400 km total extension is implied. If a significant thickness of volcanic flows were extruded and underplating occurred, this of course would require still greater extension for a given present crustal thickness [e.g., Behrendt et al., 1990] but would probably be localized rather than uniformly distributed.

Stern and ten Brink [1989], reject the simple shear model for faulting at the Transantarctic Mountains front, because this mechanism could not provide the observed magnitude (5 km) of uplift of the Transantarctic Mountains. Instead they propose a model in which two lithospheric plates of vastly different effective thermal ages are juxtaposed. They estimate flexural rigidities of 1×10^{25} and 4×10^{22} Nm for the two plates, East Antarctica and the extended crust underlying the Ross Embayment, respectively. Their corresponding lithospheric elastic thicknesses are 115 ± 10 km and 19 ± 5 km respectively. The 19-km value is essentially the same as the crustal thickness measured beneath the Ross Sea continental shelf (Figures 10 and 11) and suggests that the shallow asthenosphere is essentially at the base of the crust. In the model these plates are separated at the mountain front by a stress free boundary. Uplift of the rigid cantilevered beam is driven by heating at the free edge. Their model implies effective thermal ages for East Antarctica of about 500-600 Ma and the Ross Embayment of 25 Ma; the latter we find quite appropriate for the late Cenozoic volcanism observed considering the uncertainties.

Stern and ten Brink [1989] suggest that about 70 m.y. is required for an average density contrast of 1.5% associated with thermal conduction to penetrate 50 km horizontally beneath the Transantarctic Mountains to provide the necessary thermal uplift. This is within the 90-65 Ma proposed by Smith and Drewry [1984] for the initiation of hot spot overriding and is consistent with Fitzgerald's [1989] interpretation of start of uplift of the Transantarctic Mountains at 60 Ma. Stern and ten Brink [1989] note that it could take several tens of millions of years of thermal conduction to weaken the edge of the East Antarctic plate and start the uplift process. The evidence for rapid uplift [Behrendt and Cooper, 1991] suggests that once started it proceeds rapidly similar to other rift shoulders, which could account for the high Neogene uplift rates proposed [Behrendt and Cooper, 1991; Behrendt et al., 1991b]. The Smith and Drewry [1984] model has an additional attraction in that it suggests that the later volcanism observed in Victoria Land (and offshore according to the magnetic and seismic data discussed in this paper) would have started about 20 m.y. ago. They point out that this volcanism should be increasing in intensity, reaching its maximum development today, again consistent with the 25 Ma thermal age calculated for the Ross Embayment [Stern and ten Brink, 1989]. However, the Oligocene (or older) dates for Cenozoic volcanic rock recovered from the CIROS hole [Barrett, 1989] indicates that volcanism started at least by that time.

The Stern and ten Brink [1989] and Smith and Drewry [1984] discussions of heat flow are consistent with the geophysical data discussed in this paper: (1) evidence of extension seen throughout the area of the Ross Sea aeromagnetic survey (Plate 1), and seismic surveys (Figures 10 and 11); (2) widespread submarine and subglacial volcanism (Figure 4 and Plate 1) inferred throughout the West Antarctic rift system, and (3) evidence of extension throughout the rift interpreted from the Bouguer anomaly data combined with the seismic depths to Moho beneath the Ross Sea shelf (Figures 8 and 10). Very high heat flow based on xenolith thermobarometry [Berg et al., 1989], a 40^{o}C/km gradient in the CIROS hole [Barrett, 1989] and the sparse [Blackman et al., 1987] heat flow data are consistent with the magnetic evidence of shallow volcanism (Plate 1 and Figure 7) in the Victoria Land Basin area. High heat flow values of 1.7 HFU at 80^{o}S, 120^{o}W and 1.9 HFU at 82^{o} 53'S, 136^{o} 40'W [Alley and Bentley, 1988] from ice covered areas of the rift are supportive of active rifting there.

Possibly the mafic extrusion, intrusion and underplating which we interpret from the magnetic data, Bouguer anomalies and high seismic velocities result from a thermal plume mechanism [LeMasurier and Rex, 1989] such as that discussed by White and McKenzie [1989]. The rift flank uplift and subsidence seem closely related to magmatic activity. The plume mechanism proposed for this region rests on (1) the association of contemporaneous magmatism, centrifugal migration of felsic activity [LeMasurier and Rex, 1989], and a stationary plate environment, (2) isotopic data which suggest an asthenospheric source, and (3) the fact that Cenozoic West Antarctic rift basalts are geochemically indistinguishable from oceanic island basalts that are widely accepted as products of plume magmatism [Futa and LeMasurier, 1983; LeMasurier and Rex, 1989; Hole and LeMasurier, 1990] (but are clearly distinguishable from Antarctic Peninsula postsubduction basalts). However, had any substantial doming resulted from such a thermal plume [e.g., White and McKenzie, 1989], very substantial volcanic flows (flood basalts?) resulting from decompression melting [Futa and LeMasurier, 1983; LeMasurier, and Rex, 1989; Schmidt-Thome et al., 1990; Mueller et al., 1990; Behrendt et al., 1991b] would be expected beneath the Byrd Subglacial Basin and Ross Embayment overlying the interpreted extended crust. The sparse geophysical evidence available (discussed here) provides partial support for their existence. The White and McKenzie [1989] model implies a plume head having a diameter on the order of 2000 km comparable to the West Antarctic rift system.

The 5 km of relief across the Transantarctic Mountains front in the North Victoria Land-South Victoria Land area continues from 4 to 7 km relief along the entire rift shoulder (Figures 2 and 3) and likely is related to the thermal structure extending throughout the West Antarctic rift system. Possibly the Stern and ten Brink [1989] model could be applied to this scarp. Perhaps a plume mechanism partly explains the curvature of the rift shoulder (Figures 1 and 2). Whatever the mechanism, we interpret that the high uplift of the rift shoulder and the Amundsen-Bellingshausen flank resulted from lateral thermal heating from the upwarped shallow asthenosphere beneath the Byrd Subglacial Basin and Ross Embayment.

We inferred that shallow to lower crustal intrusions are the sources of the positive Bouguer anomaly over the rift, steep gradients along the rift shoulder, and high seismic velocities (6.7-7.4 km/s) at shallow crustal depth. If underplating and lower crustal intrusions are present, they may account for a few deep reflections reported from beneath McMurdo Sound and the Ross Sea shelf [McGinnis et al., 1985; Kim et al., 1986; Cooper et al., 1987b; Tréhu et al., 1989] similar to those interpreted elsewhere than Antarctica [e.g., Behrendt et al., 1990]. Supportive evidence is provided by reported gabbro norite cumulates from xenoliths collected from volcanic rock near McMurdo [Berg et al., 1989] and Mount Melbourne [Wörner and Viereck, 1989] (Figure 6). The Polar 3, Sinuous Ridge and other >1000-nT magnetic anomalies (Figure 2) are evidence for great accumulations of mafic (or ultramafic?) intrusive (and extrusive) magmatic rocks.

Progression of Rifting From Jurassic to Present

LeMasurier [1990] noted above that the age of the oldest Cenozoic volcanism is about Oligocene (or earlier) in the Ross Embayment and progressively decreases towards the end of the rift near the Antarctic Peninsula, where it is Pliocene in age. The age of earliest rifting in the Ross Sea basins is not known but is probably late Mesozoic. We suggest that the Gondwana breakup and the West Antarctic rift system are part of a continuously operating single system. There is a progression, as also discussed by Lawver et al. [1991], in rifting and separation around East Antarctica from Jurassic (179-162 Ma) for Africa [e.g., Ford and Kistler, 1980; White and McKenzie, 1989] to Cretaceous (about 130 Ma) for greater India [Johnson et al., 1976] and about 110-90 Ma for Australia [Cande and Mutter, 1982] to Late Cretaceous-early Cenozoic (about 95 to 85 Ma [Bradshaw, 1989]) for New Zealand and the Campbell Plateau from Marie Byrd Land. We propose that this rifting has propagated into West Antarctica (with a spreading center jump) to its present location in the Ross Embayment [Davey, 1981] and West Antarctica. This idea is generally supported by the structures between the Ross Sea and New Zealand revealed in the Geosat data [Sandwell and McAdoo, 1988] and the suggestions of thermal activity beginning in West Antarctica in possibly Late Cretaceous time [Smith and Drewry, 1984]. Possibly the spreading center was "captured" by the plume head (as suggested for other areas by Kent [1991]. Weinstein and Olson [1989] noted that plumes are entrained into upwelling flow beneath spreading centers.

Most extension in the rift in West Antarctica is probably late Mesozoic in age, but a significant part of the total could have occurred during the Cenozoic [Stock, 1989; J.M. Stock, personal communication, 1989]. The major part of the rift process, including all the dated alkali volcanism, took place after the separation of Antarctica from Australia and New Zealand. As it is a process restricted to the Antarctica plate with no apparent link to the surrounding plate tectonic features, the cause for the late Cenozoic volcanism and rifting will have to be assumed in mantle processes under the stationary Antarctic plate.

We realize that because of the sparse data base, we have been forced to use analogs with other areas and a lot of speculation. We hope, however, that these ideas may stimulate acquisition of badly needed geophysical data and encourage others to interpret their work throughout the area within the context of these ideas.

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J.C. Behrendt, U.S. Geological Survey, MS 964, Box 25046, Federal Center, Denver, CO 80225.
A.K. Cooper, U.S. Geological Survey, MS 999, 345 Middlefield Road, Menlo Park, CA 94035.
D. Damaske and F. Tessensohn, Federal Institute of Geosciences and Natural Resources, PO 510153, D-3000, Hanover 51, Federal Republic of Germany.
W.E. LeMasurier, Department of Geology, 1200 Larimer St., University of Colorado, Denver, CO 80204
A. Tréhu, College of Oceanography, Oregon State University, Corvallis, OR 97331.

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