CONTRIBUTIONS TO ANTARCTIC RESEARCH II
ANTARCTIC RESEARCH SERIES, VOLUME 53, PAGES 67-112

THE WEST ANTARCTIC RIFT SYSTEM: A REVIEW OF GEOPHYSICAL INVESTIGATIONS

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The West Antarctic rift system extends over a 3000 × 750 km, largely ice-covered area from the Ross Sea to the Bellingshausen Sea, comparable in area to the Basin and Range and the East African rift systems. A spectacular rift-shoulder scarp along which peaks reach 4 to 5 km maximum elevation 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. Various lines of evidence, no one of which is definitive, lead us to conclude that the Transantarctic Mountains part of the rift shoulder (and possibly the entire shoulder) has been rising since about 60 Ma at episodic rates of the order of 1 km/m.y. most recently since mid-Pliocene time rather than continuously at the mean rate of 100 m/m.y. We suggest (Behrendt and Cooper, 1991) a possible synergistic relation between episodic tectonism in the West Antarctic rift system and the waxing and waning of the Antarctic ice sheet (Webb, 1990) approximately coincident in time with rifting, mountain uplift, and volcanism since Oligocene (or earlier) time. The West Antarctic rift system is characterized by bimodal alkaline volcanic rocks ranging from about Oligocene or earlier to the present. These are exposed along the rift shoulder part of the Transantarctic Mountains and on the lower Amundsen-Bellingshausen flank in Marie Byrd Land and at the south end of the Antarctic Peninsula. The trend of the Jurassic tholeiites (Ferrar dolerites and Kirkpatrick basalts) marking the Jurassic Transantarctic rift crop out coincidently 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 continuously (including the Dufek intrusion) along the lower-elevation (1 to 2 km) section of the 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 Mountains block and suggest their widespread occurrence beneath the western part of the ice sheet overlying the Byrd Subglacial Basin. A BGR-USGS aeromagnetic survey over the Ross Sea continental shelf indicates rift fabric and suggests numerous submarine volcanoes along discrete NNW trending zones. Large offset seismic profiles over the Ross Sea shelf collected by the German Antarctic Northern Victoria Land expedition V (GANOVEX V), combined with earlier USGS and other results, indicate 17–21 km thickness for the crust in the Ross Sea, which we interpret as evidence of extended continental crust. A regional positive gravity anomaly extends from the Ross Sea continental shelf throughout the subglacial area (Byrd Subglacial Basin) of the West Antarctic rift system and indicates that the crust is approximately 20 km thick rather than the 30 km reported in earlier interpretations. An approximately 200 (+50 to −150) mGal Bouguer anomaly having 4–7 mGal/km gradients where measured in places extends across the rift shoulder from northern Victoria Land possibly to the Ellsworth Mountains (where data are too sparse to determine maximum amplitude and gradient). In contrast, the maximum Bouguer gravity range is only about 130 mGal with a maximum 2 mGal/km gradient in the Pensacola Mountains area of the Transantarctic Mountains, which is easily explained by 24-km-deep Moho deepening about 8 km at a dip of 15°–20° beneath the mountains. The steepest gravity gradients across the rift shoulder require high-density (mafic or ultramafic?) rock within rifted 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 lower crustal velocities about 7 km/s beneath the Ross Sea continental shelf support this interpretation. The near absence of earthquakes in the West Antarctic rift system probably results from a combination of primarily

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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). Therefore we do not consider the low seismicity significant evidence against the presence of an active rift. We suggest that the origin of the West Antarctic rift system is due to a continuation of Gondwana breakup that started in the Jurassic when Africa rifted from East Antarctica (including the failed Transantarctic rift), 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 Byrd Subglacial Basin–Ross Embayment was probably extended greatly during late Mesozoic time, but a significant part of the total could have occurred during Cenozoic time. The Cenozoic activity of the West Antarctic rift system appears to be continuous in time with rifting in the same area that began in the late Mesozoic. The major part of the rifting, including extrusion of all of the dated alkali volcanism, took place after the separation of Antarctica from Australia and New Zealand.

INTRODUCTION

In this paper we review the geophysical research over the West Antarctic rift system over the past three decades and present new compilations including new data in the Ross Sea area. The paper is an effort to integrate these Antarctic results within the concepts of continental rifting developed for other areas during the past decade. The Gondwana breakup and the West Antarctic rift system are part of a continuously operating single system. The rifting has probably progressed clockwise around Antarctica from separation of Africa in the Jurassic to extension of continental crust in the Ross Embayment–Byrd Subglacial Basin from late Mesozoic through Cenozoic time.

One of the world's largest and least known active rift systems extends approximately 3000 km through the interior of West Antarctica, roughly parallel to the Pacific continental margin (Figure 1). First named the Cenozoic West Antarctic rift system by *LeMasurier* [1978], this largely aseismic 3000 km × 750 km area of late Cenozoic volcanism and Late Cretaceous and Cenozoic extension has also been referred to as the Ross Sea rift system [*Tessensohn and Wörner*, 1991] and the Transantarctic rift system [*Tessensohn*, 1979; *Stock*, 1989*b*] and the West Antarctic rift system by *LeMasurier* [1990*a*] and *Behrendt and Cooper* [1991]. Further complicating the nomenclature, *Schmidt and Rowley* [1986] proposed the Jurassic Transantarctic rift, to refer to an older rift system based on exposed Jurassic tholeiites along the Transantarctic Mountains extending from the Ross Sea to the Weddell Sea. Indeed, K. S. Kellogg (personal communication, 1990) has raised the question as to existence of a Jurassic rift. Kellogg noted that the only evidence for Jurassic rifting is extensive tholeiitic magmatism and the coeval timing of the initial breakup of Gondwanaland; no actual rift features (e.g., normal faults or rift basins) of Jurassic age are known. However, *Schmidt and Rowley*'s [1986] interpretation is that these features are buried beneath the ice. The Early Jurassic tholeiitic magmatic rocks (basalts, dolerite sills, and the Dufek intrusion) were intruded and extruded as part of initiation of Gondwana rifting from Africa about 184–154 Ma [*Ford and Kistler*, 1980; *Schmidt and Rowley*, 1986; *White and McKenzie*, 1989]. Tholeiitic magmatism, which generally has a shallow upper mantle origin, may only represent a failed early stage of rifting along the Ferrar-Kirkpatrick-Dufek exposures associated with the breakup of Gondwana [*Ford and Kistler*, 1980]. *Ford and Kistler* [1980] noted that Jurassic rifting along the Transantarctic Mountains ceased long before it ever reached the aulacogen stage of development.

Because the extension and rifting which we are primarily concerned with here probably began in the Late Cretaceous and continued throughout the Cenozoic, in this paper we use the term "West Antarctic rift system" to define a 3000-km-long sub-sea level bedrock valley (Figure 2) (see also *Tessensohn and Wörner* [1991]) that arcs through West Antarctica curving away from the Transantarctic Mountains between the Ross Sea and the base of the Antarctic Peninsula (Figures 1 and 2).

Recent seismic studies [*McGinnis et al.*, 1985; *Cooper et al.*, 1987*b*, 1990; *Trehu et al.*, 1989; *O'Connell et al.*, 1989] show that the crust is thinner beneath the Ross Sea part of the rift system, and older interpretations of gravity [*Bentley et al.*, 1960; *Woollard*, 1962] over the topographic trough of the Byrd Subglacial Basin reinterpreted below here indicate crustal thinning which we infer to result from late Mesozoic-Cenozoic extension. Additional gravity, aeromagnetic, and seismic studies have defined a variety of structures and anomaly patterns that seem to be characteristic of rift systems elsewhere in the world. These geophysical characteristics of the rift system are the main focus of this paper.

Because exposures in magmatic rocks correlated with the Jurassic Transantarctic rift partly overlap the West Antarctic rift system, it is important to recognize and separate the characteristics of each. The Middle Jurassic tholeiites coincide with the flank of the West Antarctic rift system along the Transantarctic Mountain front from the Pacific Ocean to within 500 km of the south pole. In the area of the Horlick Mountains, the two rift systems diverge. The Jurassic rift passes between the Whitmore and Horlick Mountains and extends into Queen Maud Land to the Weddell Sea (Figure 1) in a region of lower topography (1 to 2 km elevation) than that of the section of the Transantarctic Mountains bordering the Ross Embayment (over 4 km high) (Figure 3). In contrast,

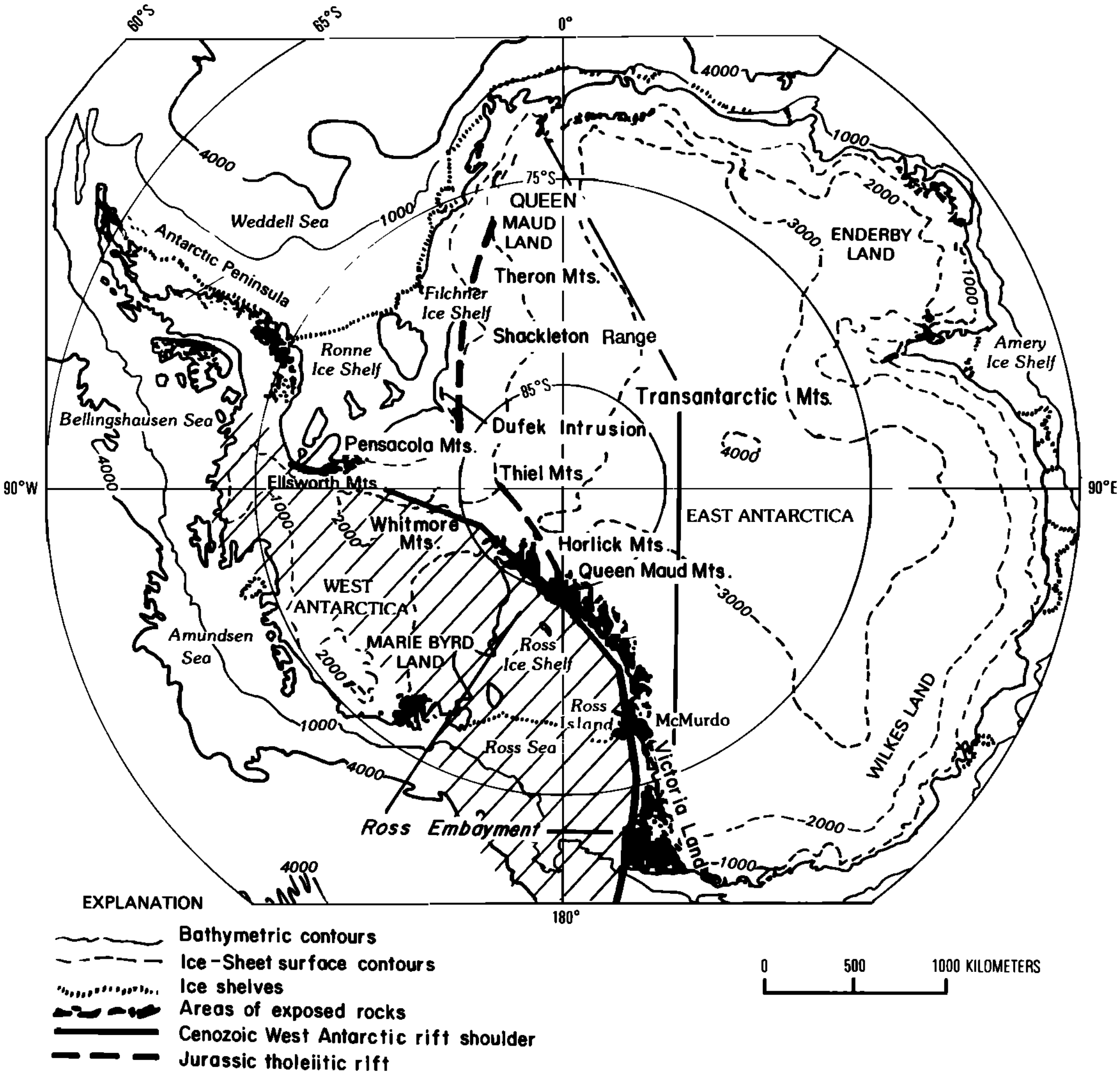


Fig. 1. Index map of Antarctica showing some of the features discussed in the text. The Transantarctic Mountains extend across the continent from Victoria Land near the Ross Sea to the Theron Mountains near the Weddell Sea and comprise the ranges shown by the heavy dashed line, as well as those bordering the Ross Embayment. The shaded area depicts the approximate location of the West Antarctic rift system. The heavy line illustrates the approximate location of the rift shoulder. To conform with convention and other publications, Figures 1, 2, 5, 12, and 13 (covering all or large regions of Antarctica) have grid north (parallel to 0° meridian) at the top. All other larger-scale maps in this paper conform to normal convention. All maps use polar stereographic projection.

the bimodal alkalic West Antarctic rift system continues beyond the Horlick Mountains and passes west of the Ellsworth-Whitmore Mountains (as indicated by aeromagnetic data discussed below) extending toward the Bellingshausen Sea. Associated late Cenozoic bimodal alkalic volcanic rocks, whose magmatic origins are inferred to be at least 100 km deep, are exposed along the west coast, in Marie Byrd Land and at the south end of the Antarctic Peninsula (Figure 2).

The distribution of igneous rocks associated with the two rifts suggests that the interior of West Antarctica has been the site of at least two episodes of lithospheric extension during the past 200 m.y., but that these episodes have only partially coincided geographically (i.e., along the Transantarctic Mountains bordering the Ross Embayment). Plate reconstructions and motions [*Norton and Sclater*, 1979; *Lawver and Scotese*, 1987; *Bradshaw*, 1989; *Stock*, 1989*a*, *b*; *Lawver et al.*, 1991] suggest that West Antarctica reached its present general configuration about 100 Ma. Similarities between crustal blocks indicate that West Antarctica is not composed of far traveled exotic terranes [*Storey et al.*, 1989].

Both geological and geophysical evidence suggests that, in addition to the late Cenozoic and Jurassic

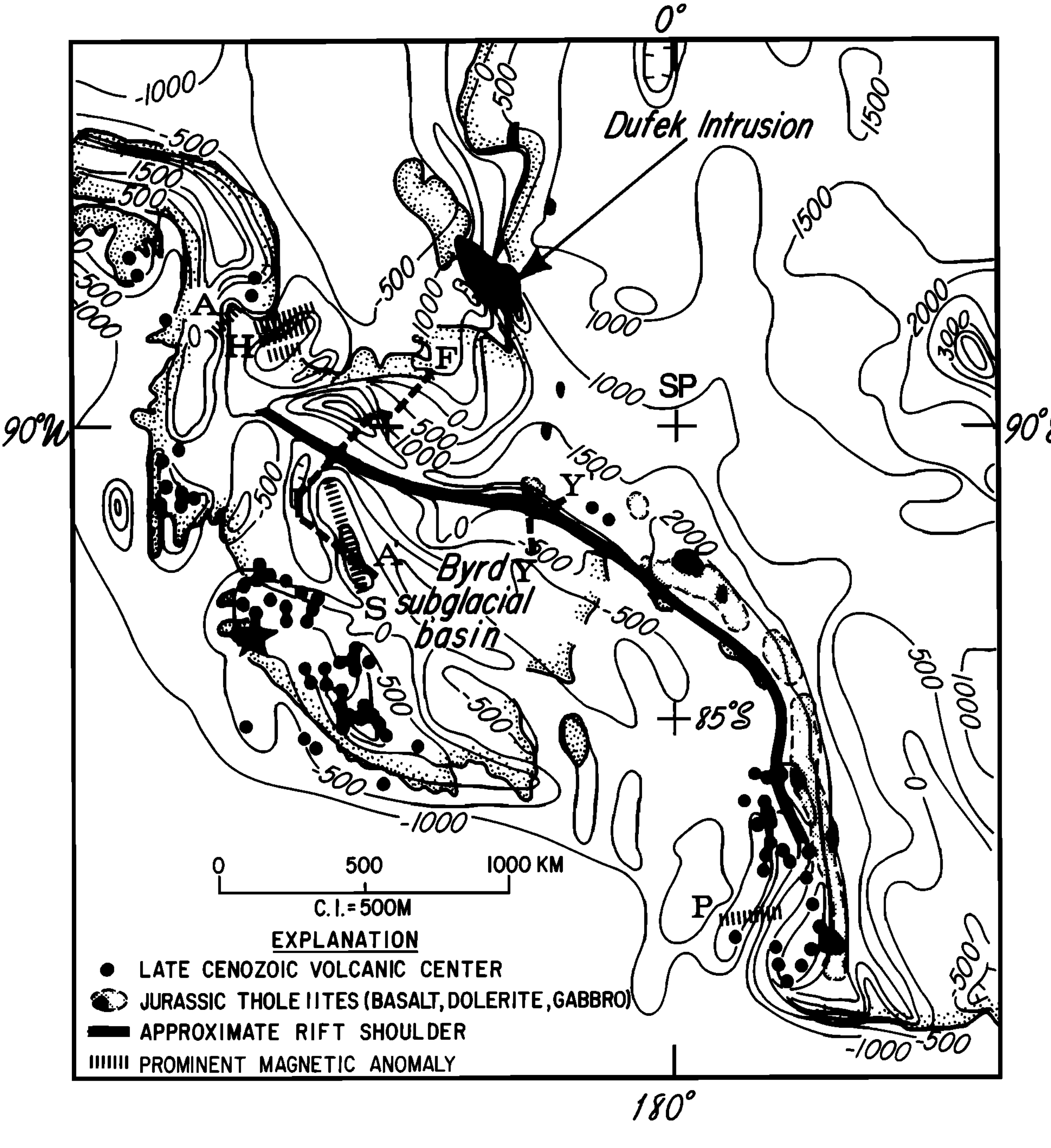


Fig. 2. Generalized isostatically adjusted bedrock elevation map after ice removal, assuming sufficient time for recovery. The contour interval is 500 m (modified from *Drewry* [1983]). All known high-amplitude linear (greater than 200 km long) magnetic anomalies greater than 1000 nT amplitude are indicated. However, coverage is only sufficient so as not to miss any magnetic anomalies in areas of the detailed aeromagnetic survey (e.g., Plates 1 and 2 and Figures 8 and 9). P, Polar 3 anomaly (Plates 1 and 2 and Figure 8); S, Sinuous Ridge anomaly [*Jankowski et al.*, 1983]; H, Haag Nunataks anomalies [*Garrett et al.*, 1987]; A, Antarctic Peninsula Traverse anomaly [*Behrendt*, 1964*b*]. The locations of late Cenozoic volcanic centers are from *González-Ferrán* [1982] and *LeMasurier and Thomson* [1990]. Locations of Jurassic tholeiites (basalt, dolerite, and gabbro) are from *Craddock et al.* [1969]. The volcanic Toney Mountain is indicated by a star (see also Figures 13 and 16). The area beneath the rift zone (elevations below sea level) is probably underlain by crust about 20 km thick, on the basis of interpretation of a Bouguer anomaly map (Figure 13), *Woollard*'s [1962] Bouguer anomaly interpretation, and seismic determinations of Moho depth beneath Ross Sea shelf (Figure 15). Locations of profiles are indicated for Figure 4 (Y—Y′) and Figure 6 (A′—F). Grid north is at the top.

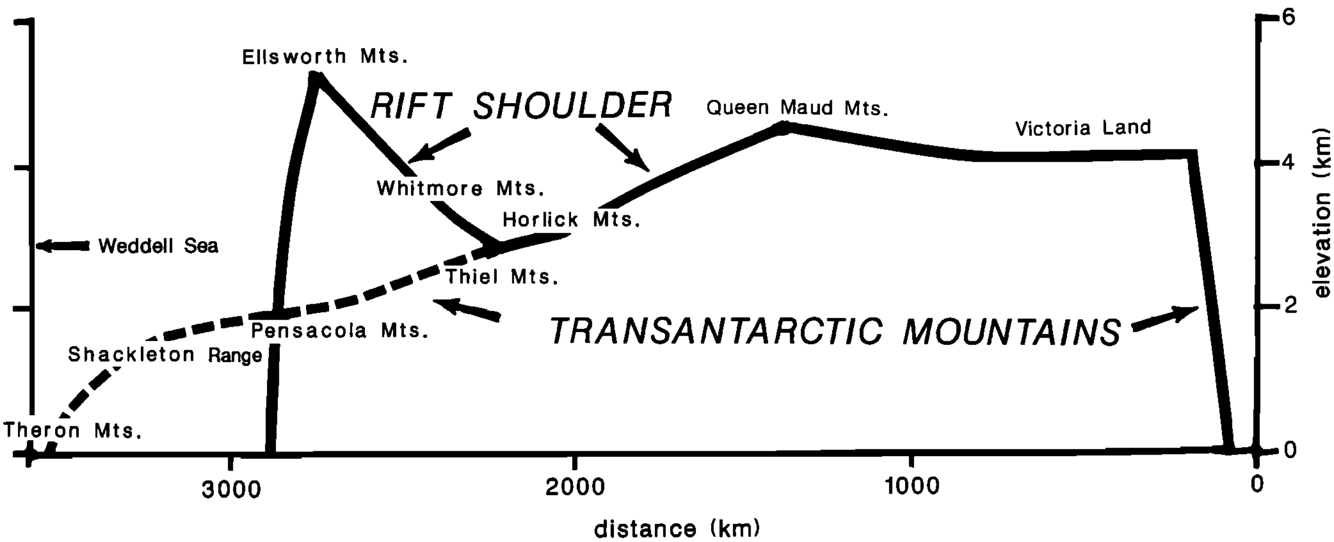


Fig. 3. Generalized topographic profile along the crest of the highest peaks parallel to and along the Cenozoic West Antarctic rift shoulder (Figures 1 and 2) (solid line) from the north coast of Victoria Land at right to the Ellsworth Mountains at left, compared with lower topography (dashed line) of 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 fluvial erosion, differential uplift, and transverse downfaulting and are not shown in these profiles.

magmatic episodes, late Mesozoic–early Cenozoic crustal extension, associated with right-lateral strike-slip displacement [*Ford*, 1972; *Storey and Nell*, 1988; *Kellogg and Rowley*, 1989], in the southern Antarctic Peninsula–Ellsworth Mountains area and [*Stock*, 1989*b*] 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 [*Cooper et al.*, 1982; *Jankowski et al.*, 1983; *Cooper et al.*, 1987*a*; *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.*, 1987*a*; *Stock*, 1989*a*, *b*] and continued (episodically?) in the Cenozoic. Possibly late Mesozoic volcanic rocks are buried beneath the ice or in the lower part of sedimentary basins.

The discussion of topography and the brief description of late Cenozoic volcanic rocks which follow are intended to provide an overall view of the rift. The rapid topographic uplift is discussed by *Behrendt and Cooper* [1991]. Further descriptions of the volcanic rocks are given by *LeMasurier and Rex* [1989] and *LeMasurier and Thomson* [1990].

TOPOGRAPHY

The Cenozoic West Antarctic rift shoulder, 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), has been a subject of geologic speculation since it was first described in southern Victoria Land by *Debenham* [1913]. Relief ranges from 5 km for summits of the highest peaks in the Transantarctic Mountains to the continental shelf at the edge of the Ross Sea (Figures 1 and 2), in the Ellsworth Mountains to 7 km from the summit of Mount Vinson, about 5 km above sea level, to the greatest depths of about 2.5 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 Rex*, 1983] 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-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. As discussed by *Stern and ten Brink* [1989], the part of rift shoulder 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. In contrast, the opposing (Pacific) flank of the West Antarctic rift system, the Amundsen Sea–Bellingshausen Sea–Ellsworth Land flank (referred to hereafter as the Amundsen-Bellingshausen flank), is a highland region also with exposed outcrops in Marie Byrd Land but lacks a well-defined shoulder, resulting in overall asymmetrical topography. Rather than a single, simple linear escarpment, the Amundsen-Bellingshausen flank is marked by graben-and-horst topography [*LeMasurier and Rex*, 1983] and by pre-Cenozoic basement elevations now no more than 2500 m. The Amundsen-Bellingshausen flank is marked by much lower maximum uplifted topography than the rift shoulder, with the present highest elevations being only about 2.5 to 3 km above sea level in the exposed Marie Byrd Land basement rocks.

We examined the amount of uplift and subsidence in the West Antarctic rift system on the basis of extreme elevations (Figure 3). The steep exposed scarp that exists almost continuously along the Transantarctic Mountains from northern Victoria Land to the Horlick Mountains to the Ellsworth Mountains is interpreted to be the expression of a major normal or extensional fault that defines the boundary between the West Antarctic rift system and the rift shoulder, with a likely strike-slip component [*Ford*, 1972; *Storey and Nell*, 1988; *Kellogg and Rowley*, 1989]. Figure 4 shows a subglacial view of the steep scarp and is suggestive of a very youthful topography [*Behrendt and Cooper*, 1991].

The range in extreme topography seen in Figure 3 along the rift shoulder is probably due to erosion as well as uplift and downfaulting. Based on youthful morphology, age of volcanism, tilted Beacon Supergroup rock, elevation, and maximum relief, we interpret the main cause of uplift along the Cenozoic West Antarctic rift shoulder to be late Cenozoic tectonism associated with rifting. The Cenozoic uplift can be modeled for a cantilevered lithospheric plate heated at the free edge [*Stern and ten Brink*, 1989].

Maximum elevations (Figure 3) vary from about 4 km in Victoria Land and the Queen Maud Mountains, dropping to about 3 km in the Horlick and Whitmore Mountains, but rise to 5 km in the Ellsworth Mountains (Figure 1). The abrupt drop 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* [1964*b*] and as a rift by *Doake et al.* [1983] and *Maslanyj and Storey* [1990]. The drop in elevation in the Horlick and Whitmore Mountains may have resulted from either a greater rate of erosion or differential uplift. In contrast, elevations along the Transantarctic Mountains are lower toward the Weddell Sea (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 1.5 km in the Shackleton Range and 1 km in the Theron Mountains. The low scarp bordering these ranges defines the boundary of the 1.7-km-deep Thiel-Crary trough beneath the Filchner Ice Shelf [*Behrendt et al.*, 1974]. We emphasize that the lower topography at the Weddell Sea end of the Transantarctic Mountain chain is not a part of the late Cenozoic rift. The age of uplift in this area is not known but appears to be significantly older than that of the West Antarctic rift system, partly based on the fact that the Cenozoic sedimentary rocks within the Thiel-Crary trough are tectonically undeformed [*Haugland*, 1982].

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 although 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. We interpret that the main cause of uplift, along the rift shoulder (Figures 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] refer to possible separate histories of individual fault blocks. They note that uneroded subaerial volcanic cones dated 4.2 and 3.5 Ma in Wright Valley (and other evidence) indicate that tectonic uplift is limited to about 300 m since the last glacial overriding. *Webb and Wrenn* [1982], *Wrenn and Webb* [1982], and *Ishman and Webb* [1988] also show evidence that indicates about 500 m of uplift occurred in the Taylor Valley area starting about 3 Ma. Although the maximum elevation in the Transantarctic Mountains in the McMurdo area (e.g., Mount Lister) is 4 km, about 20 km to the north, maximum elevations on ridges bordering the Taylor and Wright valleys are only about 2 km, suggesting to us differential uplift on transverse faults along the axis of the mountains in this area. *Wrenn and Webb* [1982] discussed differential movement in more detail in the dry valleys area. Transverse faulting along rift shoulders is common and is apparent [*Cooper et al.*, 1991] along the Transantarctic Mountains bordering the Ross Embayment. Transverse vertical faulting resulting in differential uplift is apparent between McMurdo and the Shackleton Glacier (about 85°S, 178°W, not shown) [*Katz*, 1982; D. Elliott, personal communication, 1990]. Aerial photography taken at an altitude of 12.5-km elevation along the Transantarctic Mountains north of McMurdo suggests faults transverse to the strike of the Transantarctic Mountains here. We interpret this as evidence of differential uplift (or subsidence) possibly along unrecognized transverse faults.

Fitzgerald [1989], using fission track dates from the relatively low dry valley area north of Mount

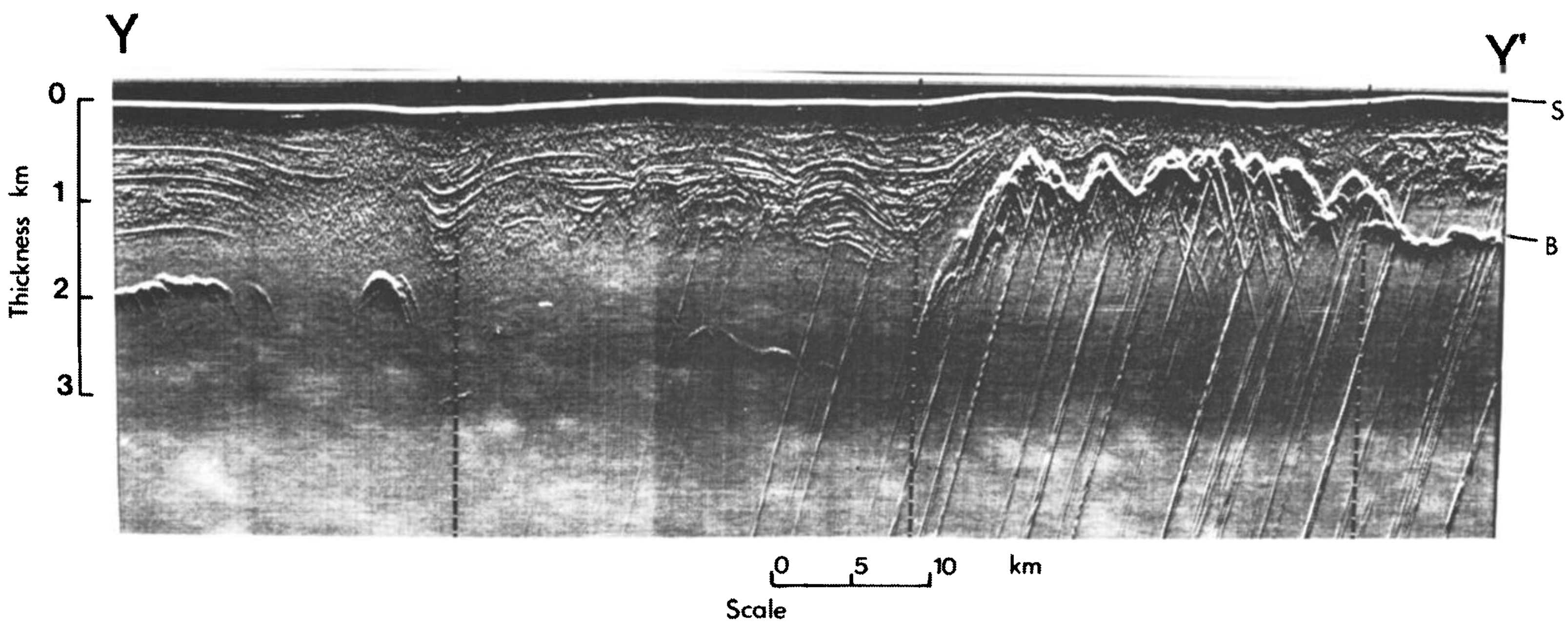


Fig. 4. Example of a radar ice-sounding profile Y—Y′ across the ice-covered part of the rift shoulder, from *Jankowski et al.* [1983]. The location is indicated in Figure 2. S, snow surface reflection; B, bedrock reflection.

Lister, interpreted 5 to 6 km of uplift in the Transantarctic Mountains of southern Victoria Land. Uplift began about 60 Ma and 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]. The extreme amount of uplift which *Fitzgerald* [1989] inferred in northern Victoria Land is 10 km. 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, which we interpret as evidence for episodic uplift. Therefore, we infer (as did *Behrendt and Cooper* [1991]) 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. Data from other similar tectonically active rift areas in the world suggest that uplift rates as great as 1 to 2 km/m.y. are not uncommon as discussed below, and thus we believe these high rates are reasonable for the rift shoulder.

Using an Occam's Razor or "principle of least astonishment" approach, various lines of evidence, none of which are independently conclusive, led *Behrendt and Cooper* [1991] to interpret that the Transantarctic Mountains part of the rift shoulder (and probably the entire shoulder) has been rising at a rate of approximately 1 km/m.y. during the latest episode, probably since early or middle Pliocene time. The following is evidence [*Behrendt and Cooper*, 1991] for this order of magnitude of uplift, no single item of which is definitive or which 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 to 5 km, Figure 2) of any suspected rift mountains on Earth.

2. The 5–7 km relief on the shoulder (Figure 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 about 10-m displacement (discussed in a later section) 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.*, 1991].

5. Unconformities in the Victoria Land basin [*Cooper et al.*, 1987*a*] in conjunction with the 100 m/m.y. average uplift rate since 60 Ma [*Fitzgerald*, 1989] require episodic uplift. We do not know of any active rift shoulder 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., *Höfle*, 1989; *McKelvey et al.*, 1991]. Although these striae possibly indicate higher ice sheet elevations in the past [*Höfle*, 1989; *Prentice et al.*, 1986], we think that the striae are more likely due to tectonic uplift of the peaks.

7. *Nothofagus* (Pliocene beech leaf) occurs in situ in Pliocene moraines at high elevations in the Transantarctic Mountains [*Webb and Harwood*, 1987]. This presence is consistent [*Mercer*, 1987] with, but does not necessarily require, a 1 km/m.y. uplift rate (Harwood et al., unpublished data).

8. Late Pliocene marine microfossils in the Sirius Formation 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].

9. Neotectonic topography of other rifts has uplift rates that are similar or higher (e.g., East Africa, 2 and 8 km/m.y. [see *Ebinger*, 1989]; western United States (Wasatch Mountains), 2 to 2.5 km/m.y. [see *Naeser et al.*, 1983]; and other areas [see *Ruddiman et al.*, 1989; *Parry and Bruhn*, 1987; *Lajoie*, 1986]).

Behrendt and Cooper [1991] considered age limits on the latest tectonic episode. The present Victoria Land–Ellsworth Mountains scarp could have been built in 4 to 5 m.y., assuming that all uplift was accommodated at about 1 km/m.y. We know however [e.g., *Fitzgerald*, 1989; *Barrett*, 1989] that uplift started much earlier. The age of the uplift of the Pensacola-Shackleton-Theron part of the Transantarctic Mountains is unknown, but the faulting of the Transantarctic Mountains from which 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 (Figure 3) is equal to that of the entire rift shoulder (from Victoria Land to the Ellsworth Mountains) prior to the latest (i.e., since Pliocene) episode of uplift, only 2 to 3 m.y. are needed for the rift shoulder to reach the present highest elevation (4 to 5 km). Estimates from *Harwood* [1986], *McKelvey et al.* [1991], *Webb and Harwood* [1987], and Harwood et al. (unpublished data) 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* between 2 and 3 Ma (Harwood et al., unpublished data).

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 300-m rift shoulder was episodic and complex with transverse faulting and differential uplift.

Amundsen-Bellingshausen Flank

The Amundsen Sea flank of the rift system is marked by lower maximum elevations, compared to the Transantarctic-Horlick-Ellsworth Mountains flank, for uplifted pre-Cenozoic basement. The magnitude and timing of the vertical displacements are also better constrained. The most conspicuous topographic features are the high volcanic peaks, which occur both in isolated and rectilinear chains, and the flat-topped coastal nunataks, which are distributed in a manner that simulates basin-and-range topography [*LeMasurier and Rex*, 1983]. The key structural datum in this region is a late Cretaceous–early Tertiary erosion surface of very low relief that bevels pre-Cenozoic granitic and metamorphic basement rocks throughout the mountainous uplifted area of Marie Byrd Land and Ellsworth Land [*Craddock et al.*, 1964; *LeMasurier and Rex*, 1982]. The erosion surface is exposed at different levels in different nunataks, at elevations ranging from 500 m to 2700 m, and is typically overlain by a thin sequence of basaltic flows and hydroclastic deposits. Dating of these volcanic rocks has revealed a systematic pattern of increasing age on successively higher remnants of the erosion surface. The displacements of the erosion surface seem to be the result of block faulting, as there is no evidence anywhere for multiple erosional surfaces, and the difference in elevation of the surface from one nunatak to another is more than 40 times the measurable erosional relief on the surface. The systematic pattern of increasing volcanic ages with higher elevations has been interpreted to represent sequential uplift of individual fault blocks, where the initiation of uplift was accompanied by volcanic eruptions [*LeMasurier and Rex*, 1983]. These relationships seem, at the very least, to represent 2.2 km of differential vertical uplift within the past 25 m.y. (i.e., 120 m/m.y.). Arguments for a higher episodic uplift rate for the rift shoulder probably apply here as well.

The regional extent of the erosion surface provides a basis for inferring subsidence when it is considered together with geophysical and chronologic data. The seismic profile at late Miocene (10.1 Ma [*LeMasurier*, 1990*a*]) Toney Mountain (75°45′S, 117°00′W) (Figure 2) described in a later section provides evidence that magnetic anomaly–producing rocks with seismic velocities similar to the volcanic rocks lie 3 km below sea level. From this we infer that the volcanic section extends to 3 km below sea level where it is in contact with basement rock inferred from seismic velocities. The erosion surface interpreted at depth is exposed only 80 km north of Toney Mountain. Hence, assuming that the erosion surface formed at or above sea level, it appears that there has been at least 3 km of posterosion surface subsidence beneath Toney Mountain.

The most significant aspect of topography of the exposed rock area in Marie Byrd Land is that it appears to record up to 2.7 km of horst uplift and up to 3 km of graben subsidence and is suggestive of an area within a rift rather than a rift shoulder. In contrast, the available data suggest that Cenozoic deformation on the Transantarctic Mountains flank was dominantly, perhaps exclusively, uplift and was largely subsidence associated with extension along the axial part of the rift system. Possibly this 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 (probably less rigid) oceanic lithosphere offshore. Geophysical studies indicate multiple periods of subsidence in grabens in Ross Sea basins since as early as late Mesozoic [*Cooper et al.*, 1987*a*], which is probably characteristic of the ice-covered low areas in the Ross Embayment–Byrd Subglacial Basin.

CENOZOIC VOLCANIC ROCKS

The large late Cenozoic volcanoes along the west coast of the Ross Sea and those along the Pacific coast of Marie Byrd Land [*LeMasurier*, 1990*a*, *b*] are, along with the rift shoulder, the most conspicuous features of the West Antarctic rift system. The highly alkaline, bimodal character of the volcanic rocks suggested even to early workers [*Hamilton and Boudette*, 1962] that volcanism in this region was rift related. However, the highly asymmetric distribution of volcanic fields and the ice cover obscured both the geometry and scale of the rift system. Quite likely the bulk of the volcanic activity, however, occurred beneath the ice and sea-covered low-elevation parts of the rift proper rather than along its flanks. Using gravity, seismic and magnetic patterns, and the reconnaissance mapping of subglacial topography by radar–echo-sounding techniques (Figure 2) [*Drewry*, 1983], it is possible to present an integrated view of the distribution of exposed volcanic rocks in relation to the still somewhat tentative boundaries of the rift system. Chronologic patterns of volcanism and the specific geochemical characteristics of the rocks add still more to the overall picture of the structure and evolution of the West Antarctic rift system. *Hole and LeMasurier* [1990] reported a roughly threefold increase in the age range of volcanic activity in the West Antarctic rift system compared with late Cenozoic alkaline basalts along the Antarctic Peninsula. They also estimated 3 orders of magnitude increase in the volume of eruptive products in Marie Byrd Land, which they considered evidence of a mantle plume.

Distribution

Cenozoic volcanic rocks are exposed discontinuously and asymmetrically along both flanks of the rift system from the Ross Sea to the Bellingshausen Sea (Figure 2) and southern Antarctic Peninsula. It can be seen from Figure 2 that these exposures are arrayed roughly parallel to the deep subglacial topographic trough that forms the axial part of the rift. The possible occurrence of subglacial volcanoes in this axial region can only be inferred from geophysical studies, and the least ambiguous possibilities are those along the floor of the Ross Sea, described below. The largest volcanoes and the only known occurrences of felsic rock related to the West Antarctic rift system are found along the Ross Sea coastal segment of the Transantarctic Mountains (between about 71° and 79°S) and along the Amundsen Sea coast of Marie Byrd Land between about 110° and 140°W (Figures 1 and 2). The major volcanoes of these two regions are typically shieldlike in form and structure (although Mount Erebus and Mount Melbourne are stratovolcanoes), display between 1 and 3 km of constructional relief and are between 10 and 50 km in base diameter. Basaltic rocks occur in these shield volcanoes, in thick laval (~2 km) and hydroclastic sequences of undetermined structure, and also in numerous minor occurrences of flows, cinder cones, and tuff cones. Felsic rocks (trachyte, phonolite, pantellerite, and comendite) occur either as summit sections of basalt shield volcanoes or as the major component of felsic shield volcanoes. The latter contain very small proportions of pryoclastic rock compared to typical parts of the world [*LeMasurier*, 1990*a*]. Intermediate rocks (benmoreite and mugearite) are usually associated with the felsic rocks in the West Antarctic rift system and are relatively uncommon. In Marie Byrd Land it has been estimated that basaltic rocks make up between 70% and 90% of the total volume of Cenozoic volcanic rock [*LeMasurier*, 1990*a*, *b*]; the Ross Sea provinces are similar in these characteristics [*Kyle*, 1990]. The basaltic rocks in general produce the high-amplitude magnetic anomalies discussed below.

In contrast, volcanoes at the Bellingshausen Sea end of the rift system are relatively small, widely scattered, and entirely basaltic. This area includes all of the volcanoes in Ellsworth Land and the southern end of the Antarctic Peninsula (Figure 1). These generalizations are not evident from Figure 2, which shows only schematic representations of known Cenozoic volcanic localities, irrespective of their size or composition. The relevance of these distribution patterns to geophysical interpretations is that one can be sure that Cenozoic volcanic rocks are a dominant rock type between about 71° and 79°S, along the west coast of the Ross Sea, and between 110° and 140°W, along the Pacific coast of Marie Byrd Land, and thus they provide "ground truth" for comparison with areas dominated by igneous rocks that are not related to the rift system.

Geochemical Characteristics

Volcanic rocks associated with the West Antarctic rift system are an alkaline bimodal assemblage, all of which belong to the sodic series of alkaline rocks [*LeBas et al.*, 1986]. They are similar in many ways to rocks in the eastern rift of the East African rift system, but the highly potassic rocks and carbonatites of the western rift in East Africa have not been found anywhere in the West Antarctic rift system.

The basaltic rocks are remarkably similar to one another throughout the rift system and indeed throughout West Antarctica as a whole [*Hole and LeMasurier*, 1990]. They are typically nepheline-normative and olivine-phyric and contain spinel lherzolite and other mantle-derived nodules in a few widely scattered localities (e.g., transitional alkali basalts and basinites in the Mount Melbourne volcanic field [*Wörner et al.*, 1989]).

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., within the asthenosphere) with no preceptible contribution from lithospheric mantle or crustal sources [*Stuckless and Ericksen*, 1976; *Futa and LeMasurier*, 1983; *Wörner et al.*, 1989; *Hole and LeMasurier*, 1990]. Some heterogeneity within this source region must be at least partly responsible for variations in the Zr/Nb and other high field strength element ratios among closely associated basalt localities [*Hole*, 1988; *LeMasurier*, 1990*a*, *b*], and variable but low degrees of partial melting are believed to be responsible for other geochemical variations, most notably in Nb/Y and Ta/Yb ratios [*Hole*, 1988].

None of the variations described above serves to distinguish basalts related to the West Antarctic rift system from non-rift-related postsubduction basalts from the Antarctic Peninsula. For this purpose, the best discriminators are relative and absolute abundances of K, Ba, and U, and a variety of ratios involving these elements. Rift-related basalts from the Ross Sea region, Marie Byrd Land, and western Ellsworth Land all have K/Ba ratios of <50, whereas Antarctic Peninsula basalts have K/Ba ratios of >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 large-ion lithophile (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 described in a later section.

Felsic rock types associated with the West Antarctic rift system include phonolite, trachyte, and several varieties of rhyolite, all of which are commonly peralkaline. However, the large felsic volcanoes tend to be either predominantly phonolitic (e.g., Mount Erebus and Mount Sidley), predominantly trachytic (Mount Melbourne and Mount Berlin), or predominantly rhyolitic (e.g., Mount Andrus). Isotopic and other geochemical studies all have concluded that the felsic and intermediate rocks evolved mainly by fractional crystallization of a basaltic parent and that crustal contamination and anatexis were of minimal significance [*Goldich et al.*, 1975; *Stuckless and Ericksen*, 1976; *Kyle*, 1981; *Wörner et al.*, 1989]. Evidently the conditions in this rift system were not suitable for the production of large-volume felsic ignimbrites like those found in the Rio Grande and East African rift systems.

In Marie Byrd Land, felsic volcanic activity has migrated centrifugally outward along rectilinear ranges from the center of the province toward its perimeter during the past 18 m.y., while basaltic activity has shown no chronologic pattern of spatial distribution. This has been interpreted as a manifestation of crustal doming along the flanks of the rift system, related in turn to the rise of an asthenospheric diapir or plume [*LeMasurier and Rex*, 1989].

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.

Chronology

Cenozoic volcanism associated with rifting seems to have begun roughly in late Eocene to middle Oligocene time in the Marie Byrd Land and Ross Embayment areas but did not begin until late Pliocene time at the Bellingshausen Sea end of the rift. Drill cores (Cenozoic Investigations of the Western Ross Sea (CIROS)) from McMurdo Sound record volcanic activity of Oligocene or older age [*George*, 1989], and 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*a*, *b*] and continued with few interruptions; the oldest felsic volcano formed around 18–20 Ma [*LeMasurier*, 1990*b*]. In the Bellingshausen volcanic province [*Rowley et al.*, 1990] the oldest basalts are 7–10 Ma in Ellsworth Land and about 7 Ma [*LeMasurier*, 1990*a*, *b*] in northern Alexander Island. 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*a*]. 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. This progression, if true, is significant and important to consider in any general model which might account for the tectonic activity of the West Antarctic rift system.

We do not, of course, know the age of volcanism beneath the Ross Sea shelf, Ross Ice Shelf, and Byrd Subglacial Basin inferred from the geophysical data discussed in the magnetics section. These geophysically interpreted volcanic rocks are substantially more extensive in area than the rocks exposed in uplifted areas (Figure 2).

GEOPHYSICAL INVESTIGATIONS IN WEST ANTARCTICA

We review the geophysical studies of the West Antarctic rift system, beginning with aeromagnetic data because it provides the widest coverage of the area, then discuss the available gravity profiles (also over a wide area), and finally turn to the generally sparse oversnow seismic investigations. In the Ross Sea, marine seismic studies are extensive.

AEROMAGNETIC SURVEYS

We discuss first the results from widely spaced flight lines magnetic (Figures 5 and 6) and oversnow traverses throughout the region. We then concentrate on our recent BGR-USGS investigations in the Ross Sea–northern Victoria land area (Figures 7–10 and Plates 1 and 2). We refer to recent geophysical surveys of the southern Antarctic Peninsula–Ronne Ice Shelf area by the British Antarctic Survey (BAS), as appropriate.

High-amplitude linear magnetic anomalies are commonly produced by exposed and unexposed volcanic rocks that are found in rift zones. In Antarctica, prominent magnetic anomalies [*Behrendt and Wold*, 1963; *Behrendt*, 1964*a*, *b*; *Behrendt and Bentley*, 1968; *Robinson*, 1964*b*; *Pedersen et al.*, 1981; *Jankowski et al.*, 1983; *Bosum et al.*, 1989; *Behrendt et al.*, 1991] mark both Cenozoic volcanic rocks in Marie Byrd Land and the Transantarctic Mountains and Early Jurassic mafic magmatic rocks in the Transantarctic Mountains (e.g., Figure 5 and Plate 1). In the Ross Sea area, our recent aeromagnetic survey (Figures 7–9 and Plates 1 and 2) provides much finer resolution of the magnetic signatures of these somewhat contrasting magnetic rock types than the earlier widely spaced surveys. For example, the exposed Ferrar dolerite sills produce low-amplitude anomalies compared with the Jurassic Kirkpatrick basalts and late Cenozoic volcanoes (Plate 1); therefore, we would not expect significant amplitude anomalies from sills buried beneath the ice sheet or continental shelf.

Aeromagnetic survey flights in the late 1950s and early 1960s [*Behrendt and Wold*, 1963; *Behrendt*, 1964*a*], which are still the only data available in much of West Antarctica, were made with poor navigational accuracy that produced horizontal errors distant from control points as great as 50 km. Although the actual magnetic data were measured to 1 or 2 nT, the widely spaced lines and lack of control for secular variation away from base stations made it impossible to produce reasonable contour maps of total intensity. The profiles were interpreted using mostly two-dimensional methods [e.g., *Behrendt and Wold*, 1963]. An aeromagnetic survey of Ross Island near McMurdo made by E. Thiel in 1958 with a 100-nT contour interval [*Robinson*, 1964*b*] showed prominent anomalies (e.g., 2000 nT over active Mount Erebus) produced by the late Cenozoic volcanic rocks that

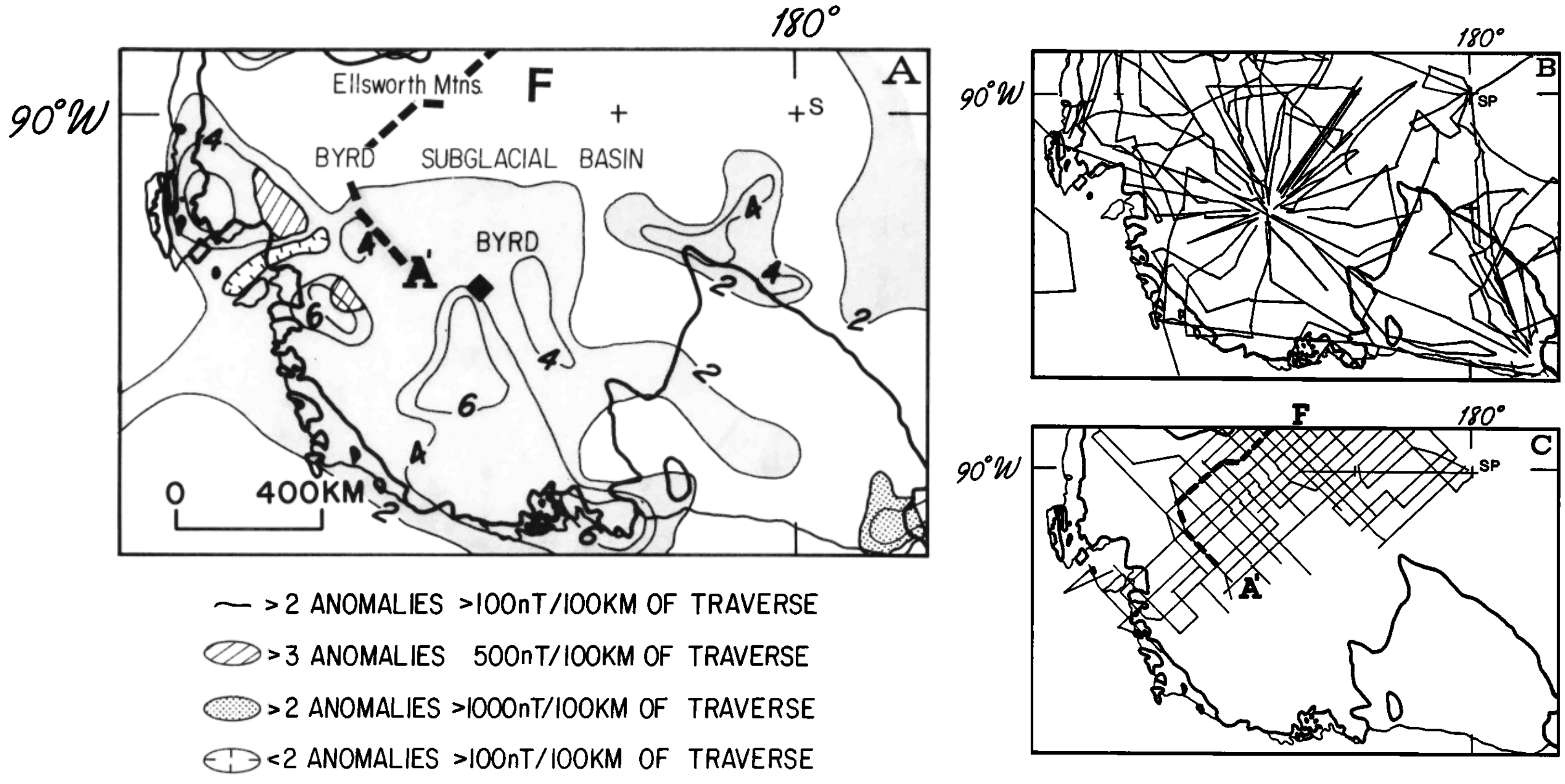


Fig. 5. (*a*) Frequency of occurrence of short-wavelength, high-amplitude anomalies. Contours show the number of anomalies greater than 100 nT per 100 km of flight line averaged for 1° of latitude grid (111 × 111 km). Modified from *Behrendt* [1968], using additional data collected in 1978 [*Jankowski et al.*, 1983; *Drewry*, 1983]. The location of profile A′—F (Figure 6) is indicated. The magnetic "break" is defined by ">2" contour in the Byrd Subglacial Basin. (*b*) Locations of flight lines used by *Behrendt* [1968]. (*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 6 and Plate 2) is indicated by a heavy line. Grid north is at the top of Figures 5*a*–5*c*.

comprise Ross Island. *Pedersen et al.* [1981] repeated this survey and published a 200-nT contour map. Although the anomalies were published along profiles [*Behrendt and Wold,* 1963; *Ostenso and Thiel,* 1964; *Behrendt,* 1964*b*], a statistical method [*Behrendt,* 1964*a*; *Behrendt and Bentley,* 1968] allowed the display of general magnetic character over Antarctica, where profiles were variably spaced.

In 1978, inertial navigation (with an accuracy to about 5 km [*Jankowski,* 1981]) was used for aeromagnetic surveys combined with radar ice sounding [*Behrendt et al.,* 1980, 1981; *Jankowski et al.,* 1983; *Drewry,* 1983] (Figures 4 and 5) over part of the West Antarctic rift system and the Middle Jurassic Dufek intrusion. The flight lines were too widely spaced and were not controlled for time variation of the magnetic field in order to contour the data, so the anomalies were shown along profiles. In this paper, we combine these data as published by *Drewry* [1983] with the earlier compilation [*Behrendt* 1964*a*, 1968] to produce the map shown in Figure 5. We did not have available the recently published BAS [*Maslanyj and Storey,* 1990] data, but in the area of Figure 5 the results would not be changed had they been included.

Evidence for Cenozoic Volcanic Rocks Beneath Ice

The obvious correlation of high-amplitude short-wavelength magnetic anomalies with outcrops of late Cenozoic volcanic rocks in the Marie Byrd Land area (Amundsen-Bellingshausen flank) (Figures 2 and 5) was noted earlier [*Behrendt and Wold,* 1963; *Behrendt,* 1964*a*]. Numerous magnetic anomalies, which we infer are caused by subglacial late Cenozoic volcanic rocks, extend over a broad area of Marie Byrd Land and the Byrd Subglacial Basin far from volcanic outcrops. The prominent (>1000 nT) linear magnetic "Sinuous Ridge" was interpreted as caused by volcanic rocks of unknown age [*Jankowski et al.,* 1983] (Figure 2). Possibly the Sinuous Ridge anomaly in the deepest part of the Byrd Subglacial Basin represents a change from alkaline to tholeiitic magmatism and the start of formation of an oceanic spreading center in the late Cretaceous or Cenozoic, although there is no gravity or other supporting geophysical evidence for this. We interpret most anomalies in the area covered within the closed "2" contour of Figure 5 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 certainly caused by older shallow sources. Locations of these ranges are shown by *LeMasurier and Rex* [1989]. Specifically Mount Takahe, Kohler Mountains, Hudson Mountains, Crary Mountains, Mount Toney, and Mount Sidley have associated high (100–500 nT) short-wavelength anomalies correlated with ridges and peaks [*Behrendt and Wold,* 1963; *Ostenso and Thiel,* 1964]. In contrast, the basement granitic and metamorphic rocks exposed in Marie Byrd Land have few or no magnetic anomalies where crossed by the profiles shown in Figure 5 [*Behrendt,* 1964*a*; *Ostenso and Thiel,* 1964], except for the anomalies over metamorphic basement within the "6 anomalies >100 nT per 100 km of traverse" contour at the bottom of Figure 5*a*.

Magnetic Boundary in Byrd Subglacial Basin

The character of the magnetic field changes abruptly, as marked by the north trending part of the ">2" contour in Figure 5*a*, about half way between the outcropping volcanoes of Marie Byrd Land and the Ellsworth Mountains (Figures 5 and 6, a representative profile). This "break" was first noted on an oversnow traverse by *Bentley and Ostenso* [1961] and described by *Behrendt and Wold* [1963] and is quite

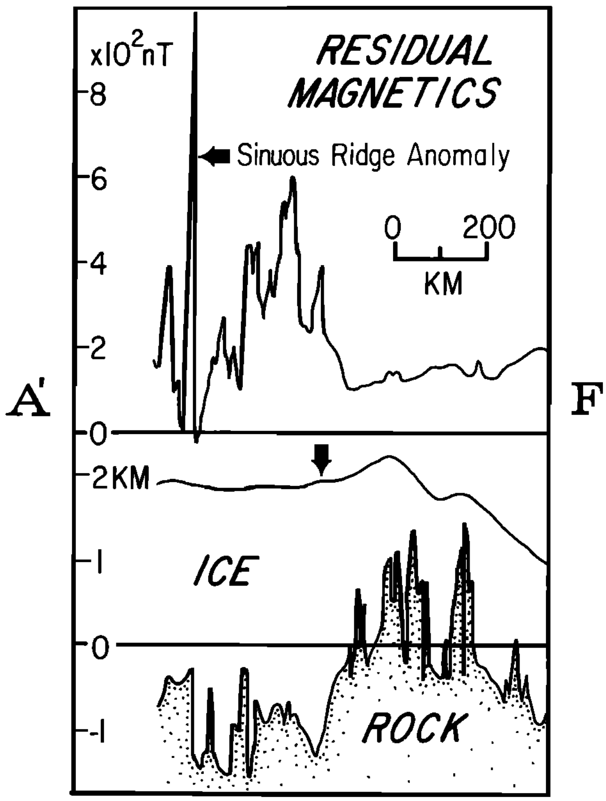


Fig. 6. Radar ice-sounding and aeromagnetic profile across the extended crust of Byrd Subglacial Basin and near the Ellsworth Mountains, from *Drewry* [1983]. Vertical exaggeration of the topography is 200:1. The arrows mark the interpreted rift shoulder and also the "break" in magnetic anomalies. The magnetic break (also Figure 5) separates a region of numerous high-amplitude anomalies, interpreted as caused mostly by upper Cenozoic volcanic rocks as sources (90% ≤ 1 km below bedrock surface), contrasted with nonmagnetic rock (possibly granitic but most likely metasedimentary because of exposures in the Ellsworth Mountains). The ~1000-nT anomaly is the Sinuous Ridge anomaly (Figures 2 and 5). The location of the profile is indicated in Figure 5.

apparent in other profiles [e.g., *Drewry*, 1983]. Although some of the shallow sources of these anomalies may be pre-Cenozoic mafic magnetic rock close to the base of the ice as well as Cenozoic volcanic rocks, the absence of shallow magnetic sources on the Ellsworth Mountains side of the "break" depicted in Figure 5 is good evidence for the lack of Cenozoic volcanic rocks beneath the ice sheet there.

Interpretation of magnetic depths along the magnetic-radar-ice-sounding profiles in Figure 5*c* combined with the earlier determined depths [*Behrendt and Wold*, 1963] were used to compile maps [*Jankowski et al.*, 1983] showing depth to magnetic basement and overlying thickness of nonmagnetic rock. On these maps (not shown), most depth to magnetic basement is less than 1 km in the area of steep short-wavelength anomalies within the "2" contour of Figure 5, including the Byrd Subglacial Basin. Seismic refraction data, discussed later, support the magnetic depth and volcanic source interpretation. The area toward 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].

A profile across part of the West Antarctic rift and rift shoulder (Figure 6) illustrates the magnetic break between the region of shallow, strongly magnetic rocks (magnetic basement close to the bedrock surface) interpreted as caused by late Cenozoic volcanic rock and the region underlain by thick nonmagnetic rocks which probably extends into the Ellsworth Mountains. The rugged subglacial topography is also continuous with that of the precipitous Ellsworth Mountains. Figure 6 illustrates the fact that the magnetic break (Figure 5) which we infer to mark the limit of late Cenozoic volcanic rocks in this area may also approximately mark the inferred edge of the rift shoulder (arrow). However, in the Ross Sea area and the Amundsen-Bellingshausen flanks of the West Antarctic rift system the magnetic Cenozoic rocks extend into the rift shoulders, so we cannot interpret the magnetic pattern to mark the rift shoulder where buried.

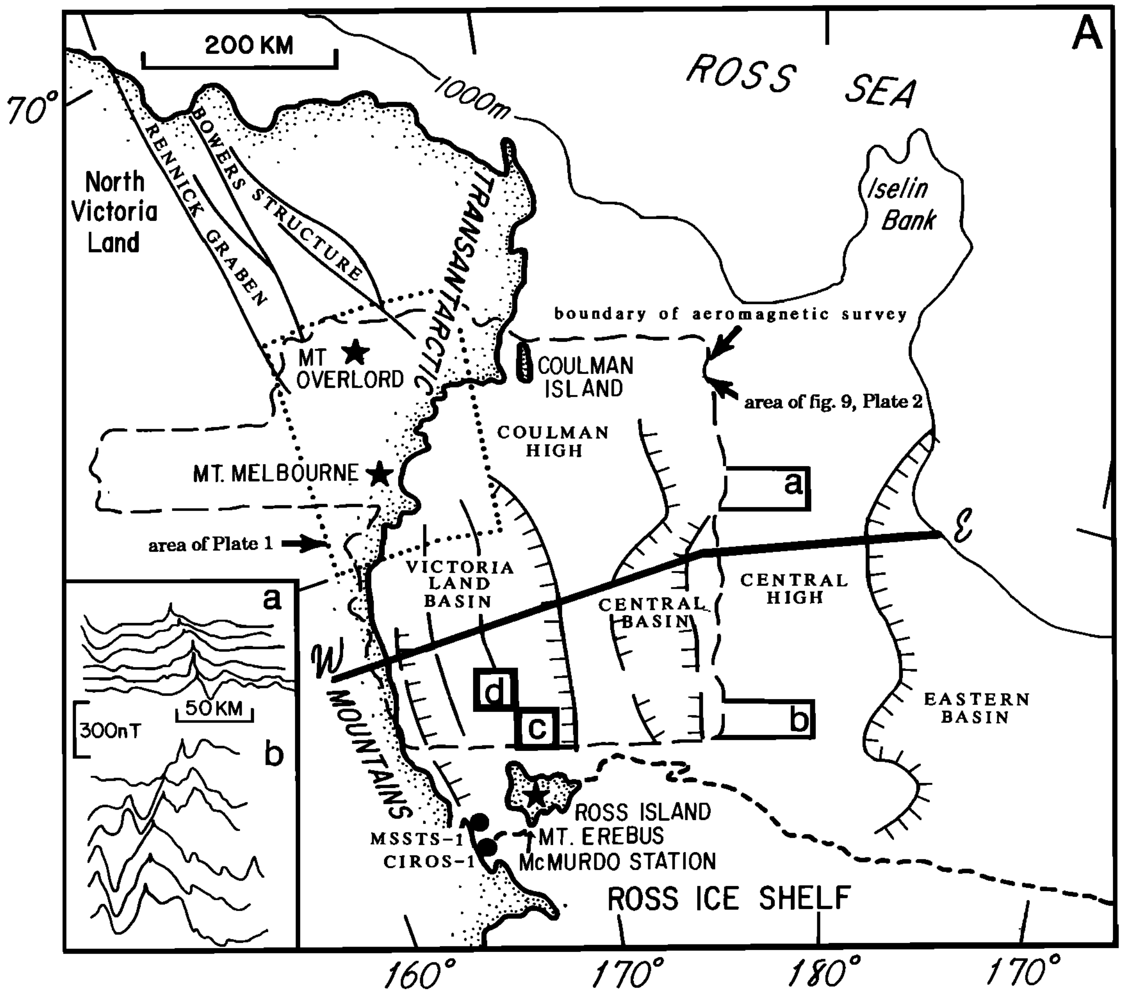
In contrast to the aeromagnetic anomalies over Marie Byrd Land (including the Byrd Subglacial Basin), which we correlate with late Cenozoic volcanic rocks, other high-amplitude, short-wavelength anomalies elsewhere over ice-covered areas are more problematic to interpret. *Behrendt* [1964*a*, *b*], *Behrendt et al.* [1974, 1980, 1981, 1990], *Garrett et al.* [1987, 1988], *Dalziel et al.* [1987], 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 [*Garrett et al.*, 1987; *Maslanyj and Storey*, 1990] to Middle Jurassic [e.g., *Behrendt et al.*, 1981] in both the southern Antarctic Peninsula and Transantarctic Mountains. We cannot differentiate between pre-Cenozoic igneous rocks and late Cenozoic volcanic rocks on the basis of magnetic data alone. We point out, however, that there are late Cenozoic volcanic outcrops in the southern Antarctic Peninsula (Figure 2) and 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 is caused by the rift-related Cenozoic(?) volcanism.

Anomalies in Haag Nunataks–Southern Antarctic Peninsula Area

Garrett et al. [1987] showed three linear anomalies (H in Figure 2, which designates location near Haag Nunataks) which are subparallel to and south of a similar 1300-nT anomaly (A in Figure 2) [*Behrendt*, 1964*b*]. *Garrett et al.* [1987] suggested that the high uplift of the northern Ellsworth Mountains (Sentinel Range) was a Cenozoic horst (which fits the interpretations of this paper) but interpret the sources of the linear anomalies (H in Figure 2) between the Ellsworth Mountains and the southern Antarctic Peninsula as being Precambrian in age because magnetic rock of that age is exposed at Mount Haag. They suggest Cenozoic rifting possibly separated the southern Antarctic Peninsula from the Ellsworth Mountains, thus extending the interpretation of this topographic tectonic break [*Behrendt*, 1964*b*; *Doake et al.*, 1983]. Considering that late Cenozoic volcanic rock [*Rowley et al.*, 1990] crops out about 50–100 km northeast of Mount Haag in the Merrick Mountains (not labeled) (75°06′S, 72°04′W), we suggest that Cenozoic mafic (or ultramafic?) magmatic rocks are possible sources for some of these high-amplitude linear anomalies (A, S, and H in Figure 2). *Maslanyj and Storey* [1990] reported recent BAS survey data over the Ellsworth–Whitmore Mountains, southern Antarctic Peninsula, and Weddell Embayment, mostly out of the area of this paper. Their results, however, did improve the definition of the Haag Nunataks and Antarctic Peninsula traverse anomalies, indicating greater complexity than the simple linear trends we indicate at the scale of Figure 2. Their data over the Ellsworth Mountain area are generally consistent with the smooth field indicated in Figure 5.

ROSS SEA SHELF–NORTHERN VICTORIA LAND AEROMAGNETIC SURVEY

Figure 7*a* shows an index map of this well-studied area. Widely spaced magnetic profiles over the Ross Sea shelf and adjacent part of the Transantarctic Mountains [*Behrendt*, 1964*a*, 1968; *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, 1981] or the Kirkpatrick basalts in the



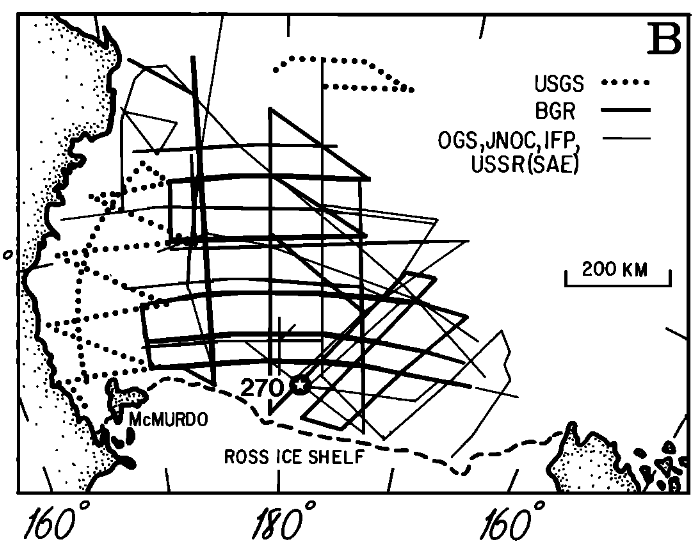


Fig. 7. (*a*) Index map of structures and geographic locations in the Ross Sea shelf–northern Victoria Land area. Active volcanoes Mount Erebus and Mount Melbourne and late Cenozoic volcano Mount Overlord (Pleistocene?) are indicated. Boxes a and b indicate the locations of the stacked magnetic profiles a and b shown in the lower left of this figure. (Ticks indicate anomaly correlations between profiles.) (Structures are from *Davey et al.* [1982, 1983] and *Cooper et al.* [1987*a*].) Locations of profiles C and D in Figure 10 and the MSSTS-1 and CIROS-1 core holes and the approximate location of the cartoon in Figure 17, profile W—E, are indicated. A 1000-m bathymetric contour is shown. (*b*) Locations of seismic reflection profiles over the Ross Sea shelf. The profiles from the U.S. Geological Survey and Bundesanstalt für Geowissenschaften und Rohstoffe (USGS and BGR) (also indicated in Plate 2) are used in interpretations in this paper. Other profiles include the Institut Français du Pétrole (IFP), France; Japanese National Oil Company (JNOC), Japan [*Sato et al.*, 1984]; Soviet Antarctic Expedition (SAE), USSR; and Oservetoreo Geofisico Sperimentale (OGS), Italy. Deep Sea Drilling Project core hole 270 is also indicated. Maps in Figures 7, 8, 9, and 15 and Plates 1 and 2 (all covering the Ross Sea–northern Victoria Land area) have normal map conventions for true north, approximately toward the top of each map.

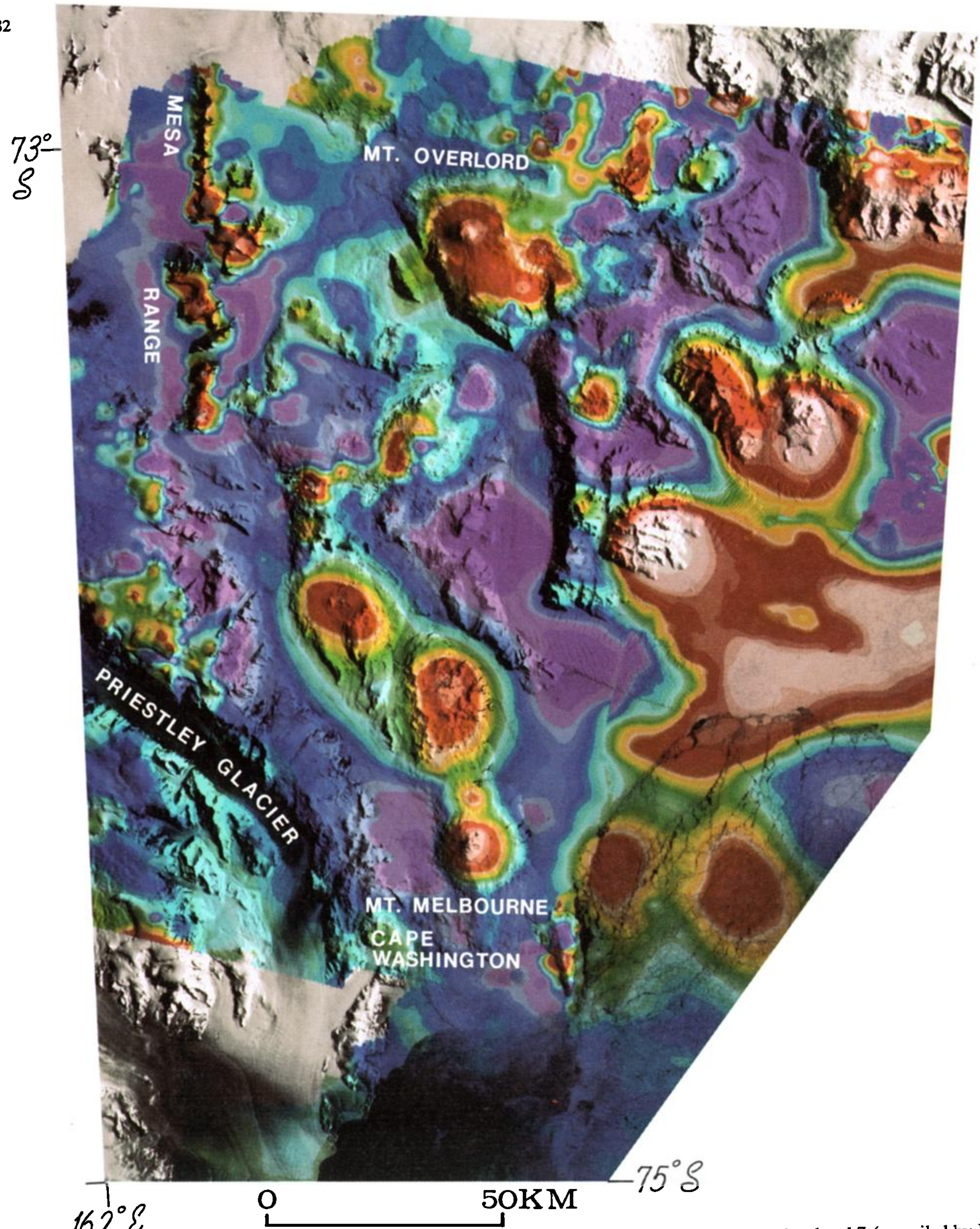


Plate 1. Aeromagnetic survey over part of northern Victoria Land, superimposed on enhanced Landsat band 7 (compiled by B. Lucchitta, U.S. Geological Survey). The area is shown in Figure 7. The contour interval is 5 nT except for the highest amplitude, as in Plate 2. Active volcano Mount Melbourne and late Cenozoic volcano Mount Overlord are indicated. Exposures of Jurassic Kirkpatrick basalts in the Mesa Range have several hundred nanotesla associated anomalies. At the north edge of Priestley Glacier, Ferrar dolerites are exposed in cliff faces [*Roland et al.*, 1989], but only a few tens of nanotesla anomalies are produced by the flat-lying sills. The small circular high-amplitude anomalies just north of the edge of Priestley Glacier are probably caused by small exposures of Kirkpatrick basalt (N. Roland, personal communication, 1989). Cape Washington consists of late Cenozoic volcanic rocks [*Wörner and Viereck*, 1989] and has an associated magnetic anomaly. The two approximately 10-km-wide circular anomalies 20 to 60 km northwest of Mount Melbourne must be caused by sources (intrusions?) buried beneath the nonmagnetic exposed rock [*Bosum et al.*, 1989].

Mesa Range (Plate 1) produce high-amplitude anomalies, we speculate 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) based on comparison of these anomalies to similar anomalies over young volcanic exposures in other areas [*Behrendt*, 1964*a*; *Robinson*, 1964*a*; *Pederson et al.*, 1981]. The possibility that these magnetic anomalies might be caused by Jurassic Ferrar-type dolerite sills is rejected because, although these bodies are quite magnetic, the sills are essentially thin sheets, produce only small anomalies and only at their edges, and do not produce anomalies greater than about 50–100 nT at the flight elevations (Plate 1) (also, for discussion of models for sills, see *Pederson et al.* [1981]). Therefore it is understandable that the numerous profiles crossing the Transantarctic Mountains (Figure 5) directly over dolerite sills [*Behrendt*, 1964*a*] did not record significant high-amplitude anomalies.

A 4.4 km × 20 km spaced aeromagnetic survey was flown over northern Victoria Land and the southwestern Ross Sea [*Bosum et al.*, 1989; *Behrendt et al.*, 1991]. These data [*BGR-USGS*, 1987], contoured at a 5-nT interval, provide substantially more information

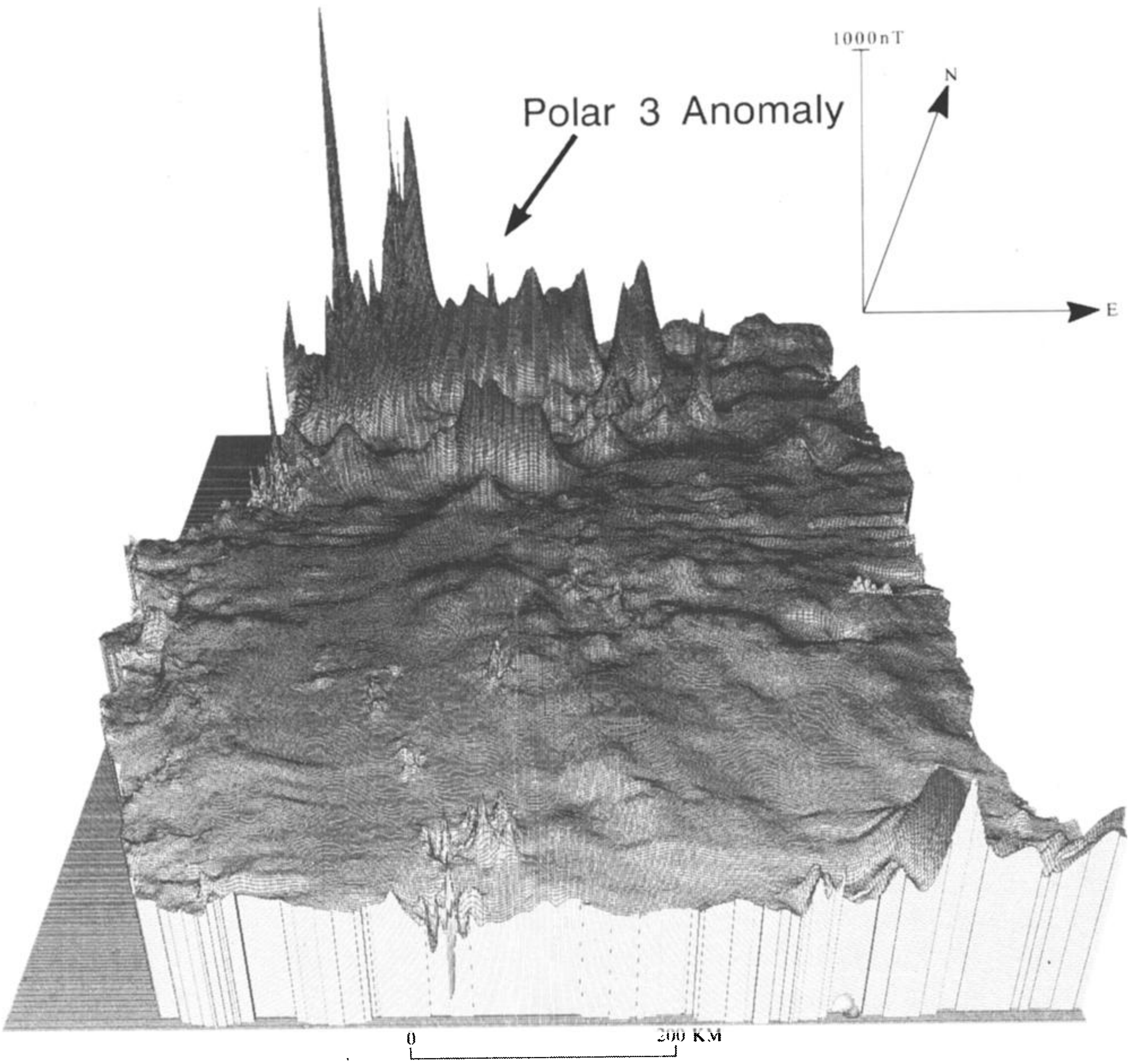


Fig. 8. A three-dimensional perspective of the total magnetic intensity map of the marine area east of the coast depicted in Plate 2, viewed from the south. Amplitudes are projected at the vertical scale shown. Geologically significant anomalies indicated here and in Plate 2 range over 3 orders of magnitude in amplitude and wavelength (compiled by D. Damaske).

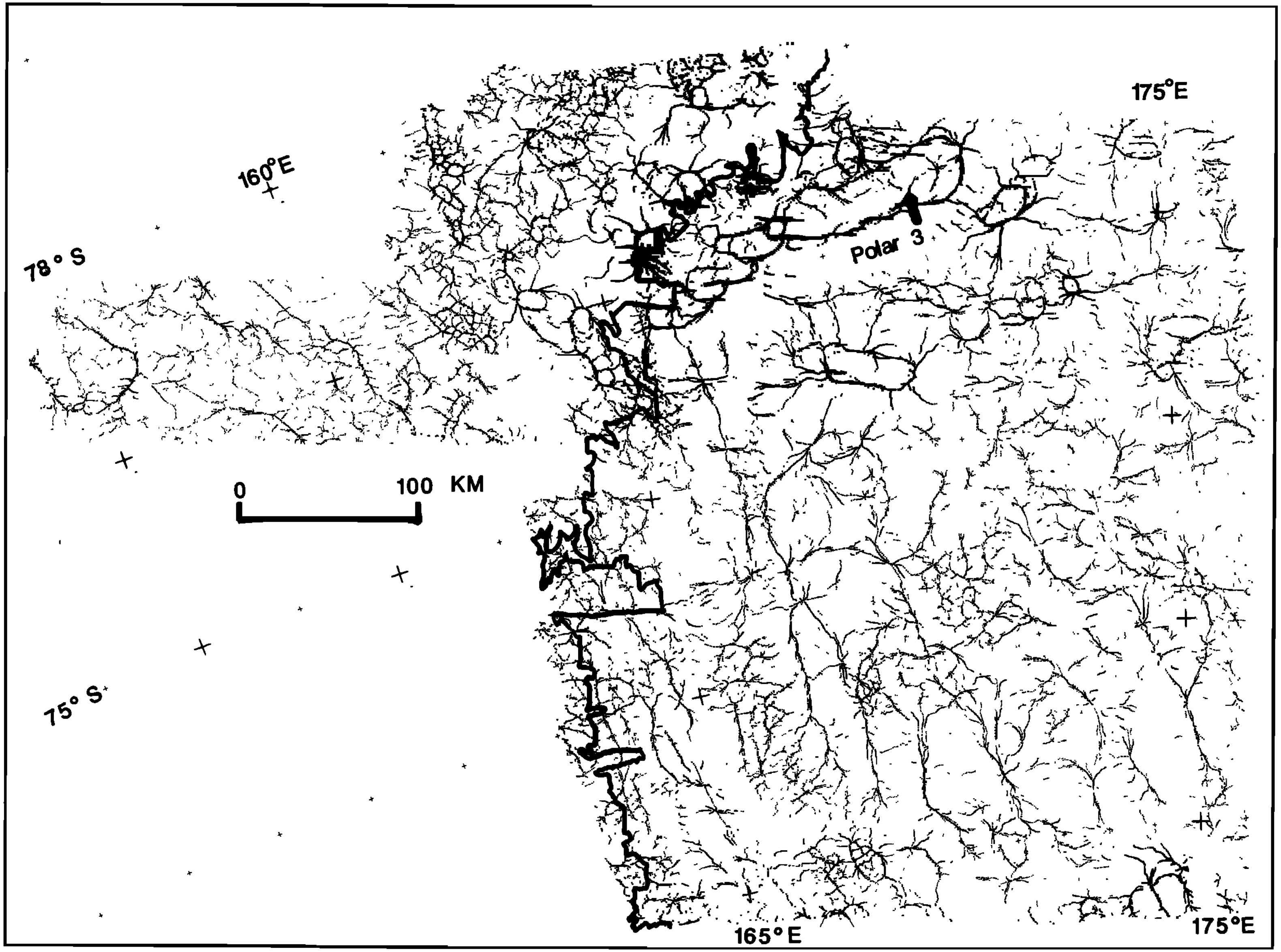


Fig. 9. Local maxima of the horizontal gradient of pseudogravity calculated from the aeromagnetic data found in Plate 2 (for explanations of methods, see *Cordell and Grauch* [1985] and *Blakely and Simpson* [1986]). This calculation procedure (which has nothing to do with gravity) normalizes the anomaly amplitudes and areally defines the edges of the bodies producing the magnetic anomalies. The lines drawn on this figure are the loci of inflection points and mark the edges of bodies (compiled by R. Saltus).

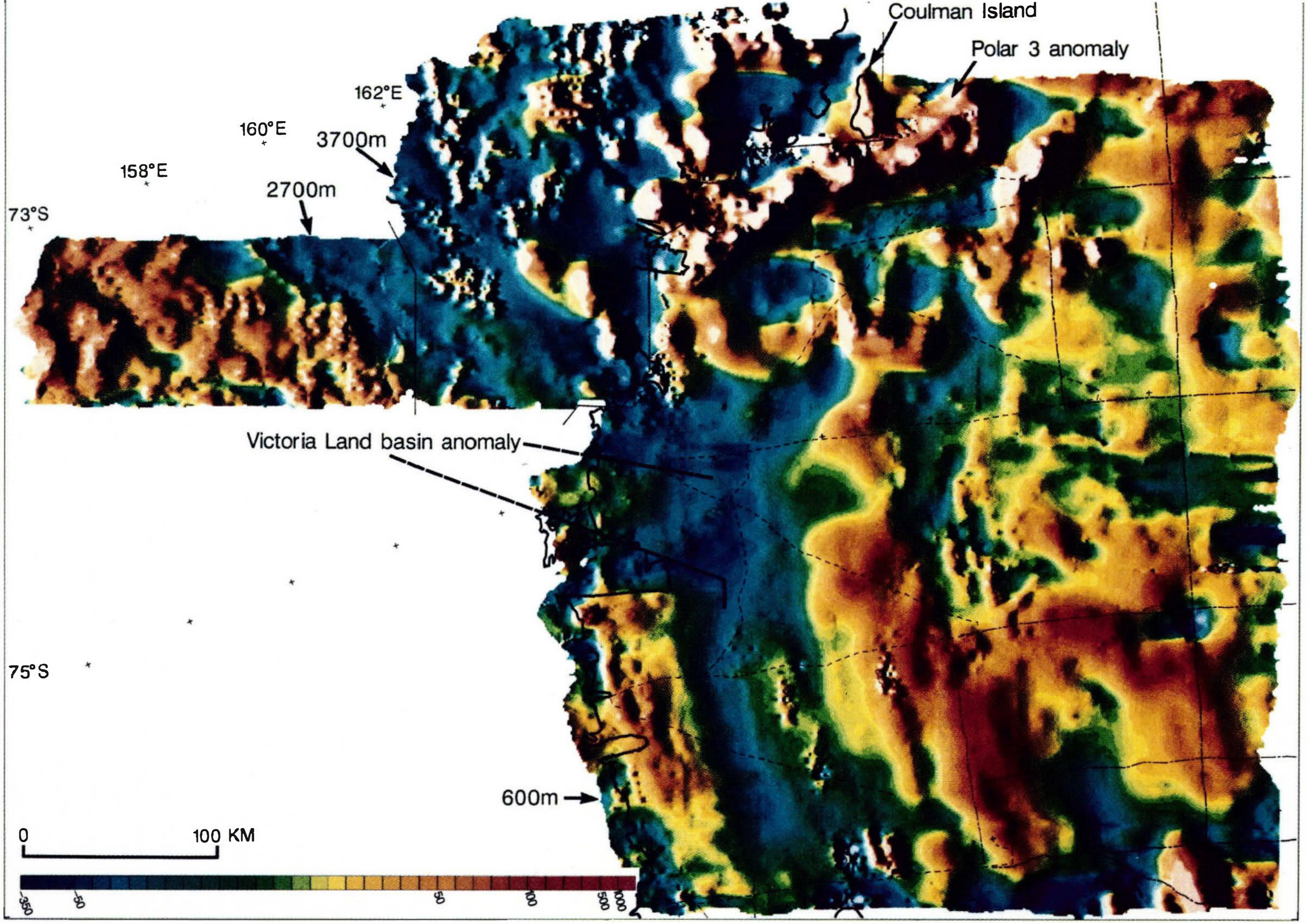


Plate 2. Color-shaded relief (apparent sun angle 15° from west) and aeromagnetic map of western Ross Sea shelf (right half of figure) and part of northern Victoria Land (top central and left part of figure). Location of the survey is shown in Figure 7. Areas flown at different elevations are separated 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 (EW) (some at 8.8 km) by 22 km (NS). This map was compiled using an 800 × 800 m grid. Locations of marine seismic reflection profiles collected by the U.S. Geological Survey's R/V *S.P. Lee* in 1984 (short dash lines) and the *Explora* of the Bundesanstalt für Geowissenschaften und Rorstoffe 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. The contour interval is 5 nT between +50 and −50 nT and compressed, as indicated in color bar beyond this range. The Polar 3 anomaly was probably caused by upper 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.

per unit area than that known for most of the West Antarctic rift system. Geologically significant individual anomalies defined by this survey range in amplitude from 10–20 nT up to 1700 nT and in length from 1–2 km to about 200 km. Figures 7–9 and Plates 1 and 2 illustrate these 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.

Figures 8 and 9 and Plate 2 show different presentations of the data set from this survey and thus should be examined together. For example, the shaded relief (Plate 2) accentuates rift fabric, but short-wavelength anomalies are hard to see at this small scale. The 5-nT contour interval present over most of the map defines well the longer-wavelength low-amplitude anomalies; however, anomalies greater than 100 and 500 nT tend to become obscured in the red shades. In contrast, Figure 8 provides the best subjective illustration of amplitude and gradient (which suggests depth to source) for all the anomalies in the Ross Sea area. However, the locations of anomalies are distorted by the three-dimensional presentation. Figure 9 provides the most information on location of edges of anomaly-producing bodies irrespective of depth. This type of presentation is somewhat analogous to a seismic epicenter map in which all data are projected to the surface irrespective of depth. In general, shapes of short-wavelength circular anomalies are not well resolved in Plate 2 and Figure 9 because of the limitations resulting from wide (4.4 km) flight lines compared with the smallest anomaly width of 1 to 2 km, which would only be crossed by one flight line, even considering the 700-m flight elevation and the 500–1400 m deep continental shelf.

Anomalies Caused by Late Cenozoic Volcanoes

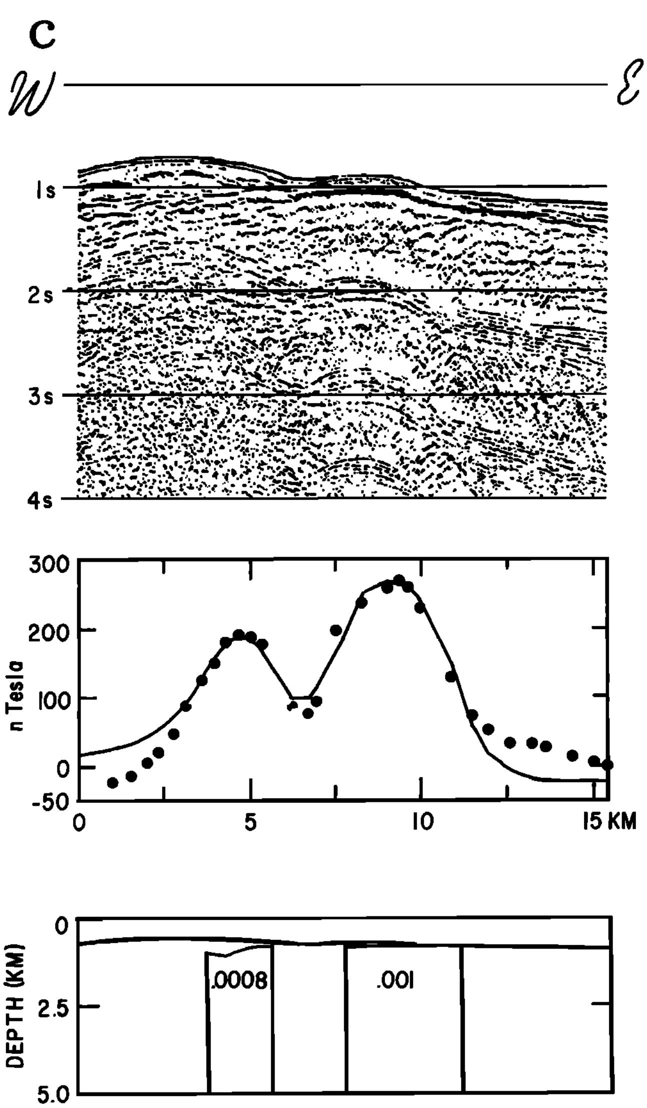
The steeper gradients of circular anomalies caused by volcanoes like Mount Overlord and Mount Melbourne (Plates 1 and 2) can be compared with those caused by similar but lower-gradient circular sources within the lower Paleozoic basement that can be seen both onshore and offshore (Plate 1). Cenozoic intrusions within competent crystalline basement rock probably could produce these circular anomalies (Plates 1 and 2) which have diameters of >10 km. Similar volumes of the same magma penetrating the thick sedimentary rock within the Victoria Land basin would produce sills with low-amplitude anomalies and the smaller-diameter (1 to 2 km) circular anomalies such as those observed over vertical bodies which underlie submarine volcanoes (e.g., Figure 10). However, older intrusions could produce sources of these ~10-km-diameter circular anomalies, some of which are not exposed (Plate 1), over outcropping basement.

Ten small anomalies having short (1–5 km) wavelengths and high (a few tens to more than 1000 nT) amplitudes were found along seismic reflection profiles using magnetic gradiometer data (e.g., Figure 10). They are attributed to late Cenozoic subvolcanic intrusions [*Behrendt et al.*, 1987] that penetrate a thick (up to 14 km) sedimentary section. The aeromagnetic survey defines many more (about 100) similar type small anomalies (Figure 8 and Plate 2) which may also have an analogous cause. Based on theoretical magnetic models compared with seismic reflection profiles for several of the very short wavelength (1–3 km) anomalies along ship tracks, we infer that causative bodies for many of these 100 small anomalies penetrate from the basement through the sedimentary section. Depth estimates indicate that the tops of the sources are essentially at the seafloor and probably would have topographic expression like those in Figure 10 were seismic or bathymetric profiles measured over them. However, only in areas of the Ross Sea floor underlain by late Cenozoic sedimentary rock [*Behrendt et al.*, 1987] can we infer that the ages of the (shallow) magmatic rock are late Cenozoic or Holocene (Figures 10*c* and 10*d*, respectively). These penetrating bodies have been defined by available seismic reflection data only in the Victoria Land basin area.

Magnetic Evidence of Rift Fabric Beneath Ross Sea Shelf

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 7 and Plate 2). We interpret this lineation as evidence of rift fabric. Note the north-northwest "grain" over the Ross Sea part of the survey, which can be better defined by locating the edges of anomaly-producing structures (Figure 9). Probably many of the north-northwest trending magnetic lineations are caused by faults in the magnetic basement, some of which are discernible on reflection profiles [*Cooper et al.*, 1987*a*]. This fabric can be seen across the survey area over the Victoria Land basin, Coulman high, Central basin (Figure 9 and Plate 2), and Central high (Figure 7). The short-wavelength (1 to 2 km) anomalies discussed above occur along some of these linear trends. This linear fabric (Figure 9 and Plate 2) could be, however, either Cenozoic or late Mesozoic in age, based on interpretation [*Cooper et al.*, 1987*a*] of several episodes of rifting.

The zone of lineations of north-northwest trending anomalies observed in Plate 2 can be extended nearly to the edge of the Eastern basin (Figure 7*a*) where additional east trending profiles are shown. We interpret the 1–3 km wavelength anomalies observed on the stacked profiles as shallow source (essentially at the seafloor) features, similar to anomalies to the west discussed above, on the basis of the depth



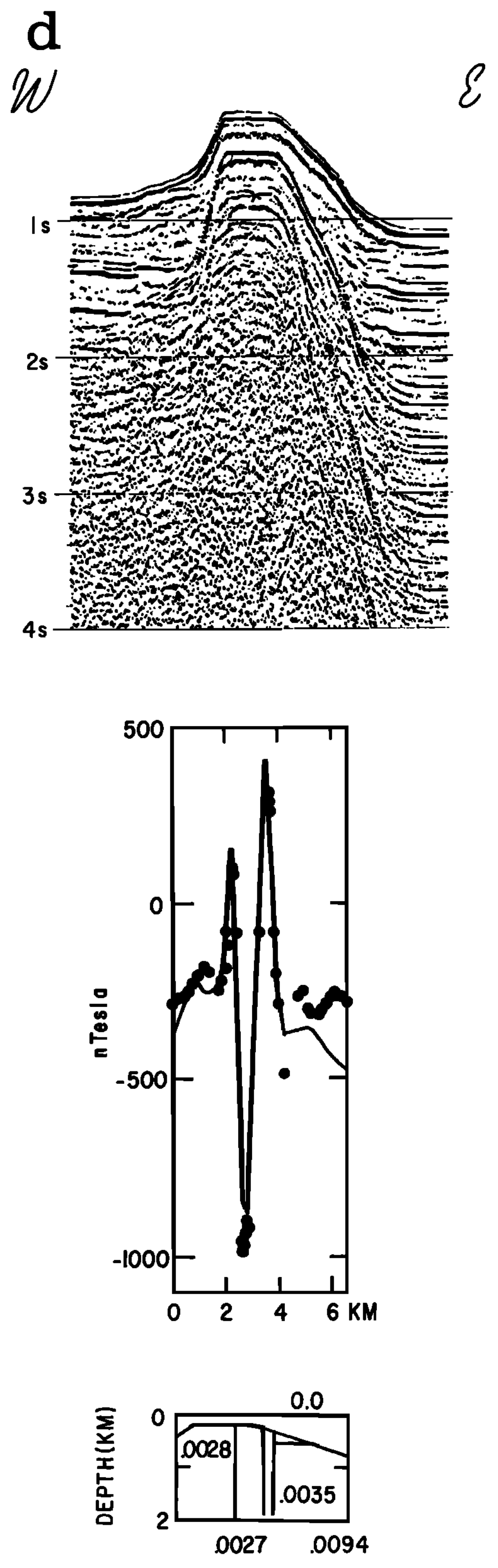


Fig. 10. (Left) Anomaly c [*Behrendt et al.*, 1987]. Seismic reflection profiles (time in seconds indicated) compared with observed (dots) and theoretical (solid line) marine magnetic profiles from *Behrendt et al.* [1987]. Apparent susceptibilities of theoretical models are shown in cgs units. Sources are interpreted to be upper 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 the latest deglaciation. (Right) Anomaly d [*Behrendt et al.*, 1987]. The prominent submarine volcano is probably Holocene; i.e., the edifice was extruded on the seafloor since deglaciation. The locations of anomalies c and d are shown in Figure 7.

estimates (horizontal extent of maximum gradient); but because these profiles cross the Central high only, we cannot say much about their age. A broad (about 40 km wide) north-northeast trending anomaly indicated by tick marks in box b of Figure 7*a* and having several sharp peaks exceeds 300 nT and indicates a shallow source.

Polar 3 Anomaly

The Polar 3 anomaly has sharp peaks (Plate 2 and Figure 8) similar to those over late Cenozoic volcanic rocks exposed on Coulman Island (Figure 7 and Plate 2), and thus we interpret the Polar 3 anomaly as a magmatic complex. Although generally linear, the curved edges of individual circular volcanic-intrusive centers can also be seen in Figure 9. *Bosum et al.* [1989] interpret serpentinized ultramafic rocks as one possible source. Probably the source consists of both volcanic deposits and subvolcanic intrusions. Transfer faults have been suggested to occur along the Transantarctic Mountains [*Fitzgerald*, 1989] and possibly the Polar 3 anomaly is related to these faults tectonically [*Wörner et al.*, 1989; *Cooper et al.*, 1991] because it has a similar trend. The Polar 3 anomaly is similar in amplitude and wavelength to other anomalies indicated in Figure 2 such as the Sinuous Ridge (Figures 2, 5, and 6) [*Jankowski et al.*, 1983], adjacent to the subglacial trench in the ice-covered areas of the West Antarctic rift system in Marie Byrd Land.

Other shallow-source magnetic anomalies occur over the Coulman high between the Victoria Land basin and Central basin (about 75°45′S, 170°E) (Figure 7 and Plate 2); these may be due to volcanoes and associated subvolcanic intrusions. Seismic velocities as high as 6.9–7.4 km/s are reported [*Cooper et al.*, 1987*b*] at shallow depths in this area as discussed below, which implies high-density mafic intrusions. We suggest the source of the 500-nT anomaly at the southeast corner of Plate 2 and Figure 8 is late Cenozoic volcanic rock.

Victoria Land Basin Anomaly

The source of the −80- to −100-nT anomaly over the Victoria Land basin, inferred to be filled with a 14-km-thick sedimentary sequence [*Cooper et al.*, 1987*a*], 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, 1991]. 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.*, 1991]. High heat flow in the Victoria Land basin area is suggested by sparse and inconclusive measurements [*Blackman et al.*, 1987], active volcanism, tectonic activity (Figure 11), thermobarometry [*Berg and Herz*, 1986], a 40°C/km thermal gradient [*White*, 1989], and high seismic velocity at shallow depth within the reflector sequence interpreted as evidence of volcanic layers (see below). These features lend support to the shallow Curie isotherm interpretation.

GRAVITY SURVEYS

Although gravity data were routinely collected on the (1956–1966) oversnow traverse program (1955–1964) throughout the West Antarctic rift system, seismic reflection determinations of ice thickness measurements at only every eighth or tenth station (about 30–40 km) and poor elevation control (absolute errors about ±50 m, corresponding to ±15 mGal error in free-air anomaly [*Bentley*, 1964]) limit their usefulness beyond ice thickness determination

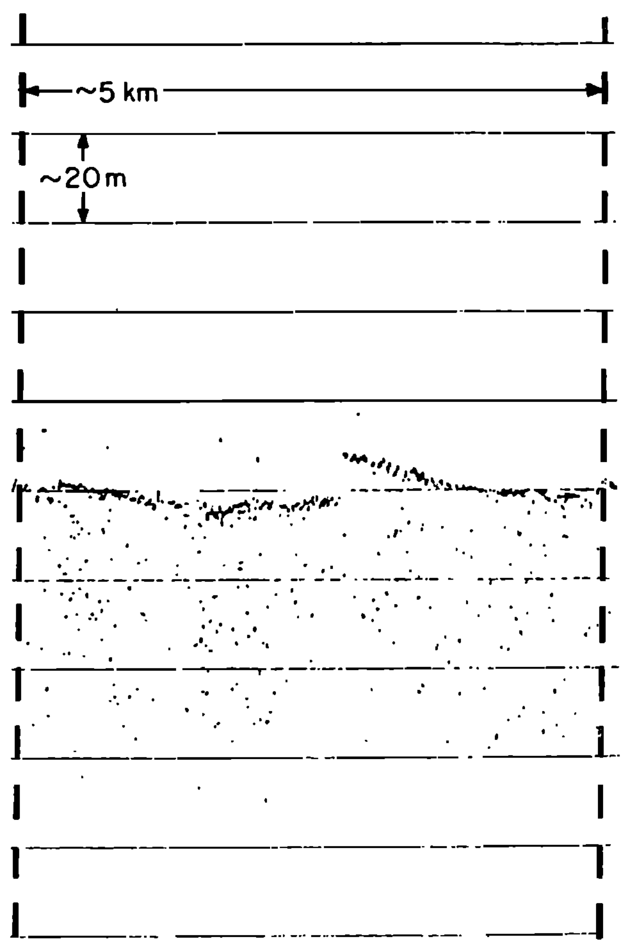


Fig. 11. Example of a 3.5-kHz bathymetric profile in the Victoria Land basin (see Figure 9), collected by R/V *S.P. Lee*, interpreted as a fault scarp cutting the sea bottom. Because this area was covered by grounded moving ice possibly as recently as about 7000 years B.P., the scarp must be Holocene in age.

between reflection soundings. In the McMurdo, northern Victoria Land, and Pensacola Mountains areas of the Transantarctic Mountains and in the Ross Sea and Ross Ice Shelf areas more accurate data have been gradually obtained but coverage is still very sparse. In this section we briefly discuss free-air anomalies over Antarctica (Figure 12) and the available Bouguer anomaly data for West Antarctica (Figure 13).

Free-Air Anomalies

Bentley [1968] compiled free-air gravity data (Figure 12) over 2° × 2° (222 km × 222 km) mean squares in Antarctica. He noted a mean free-air value near 0 for all of West Antarctica, implying isostatic balance on a regional basis for the entire area. Bentley discussed the −20 to −50 mGal "Transantarctic gravity anomaly" parallel to the Transantarctic

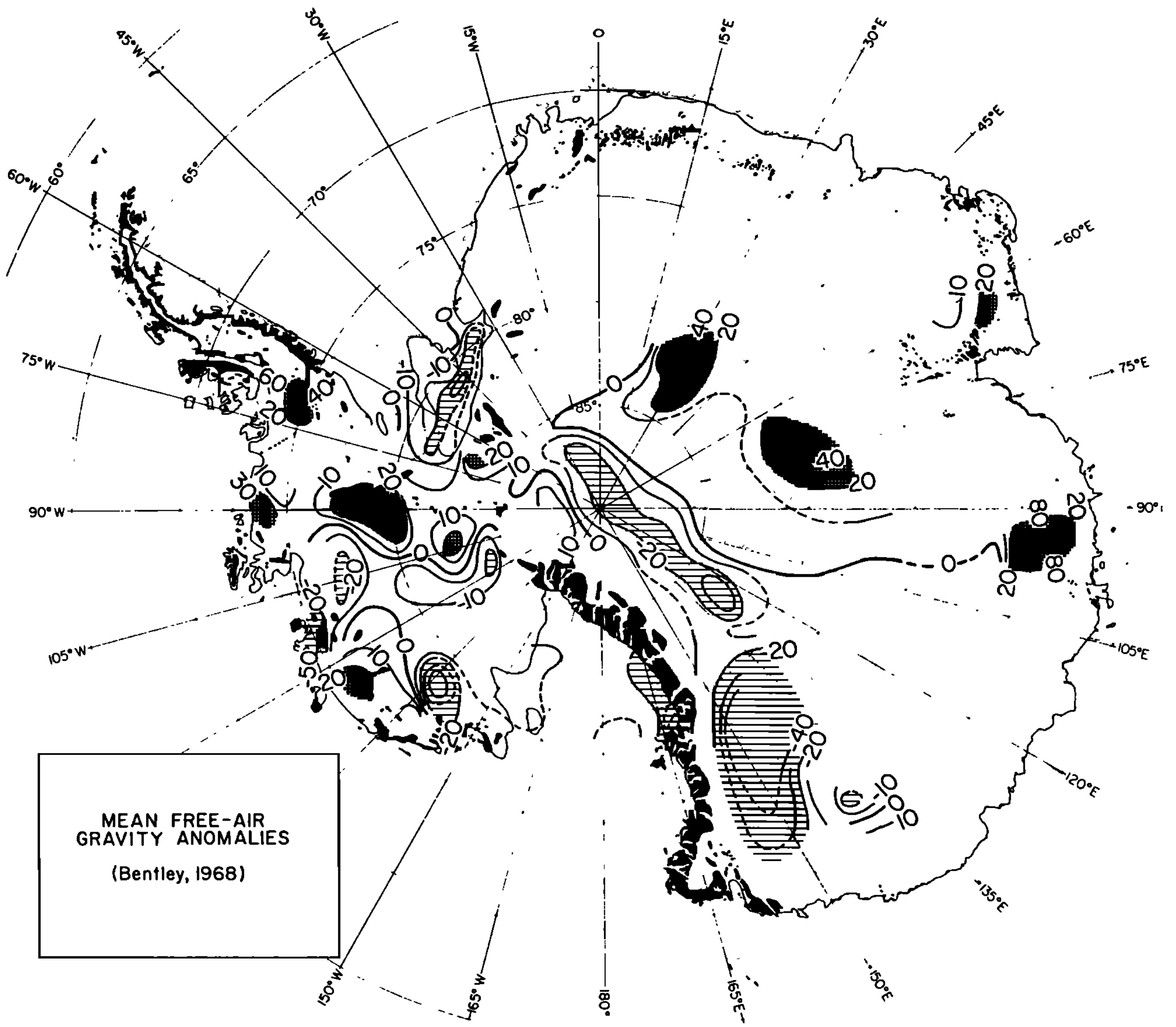


Fig. 12. Free-air gravity anomalies averaged over 2° × 2° squares from *Bentley* [1983], simplified from *Bentley* [1968]. Anomalies greater than 20 mGal in amplitude are indicated by horizontal hachures for negative anomalies and by cross-hatching for positive anomalies.

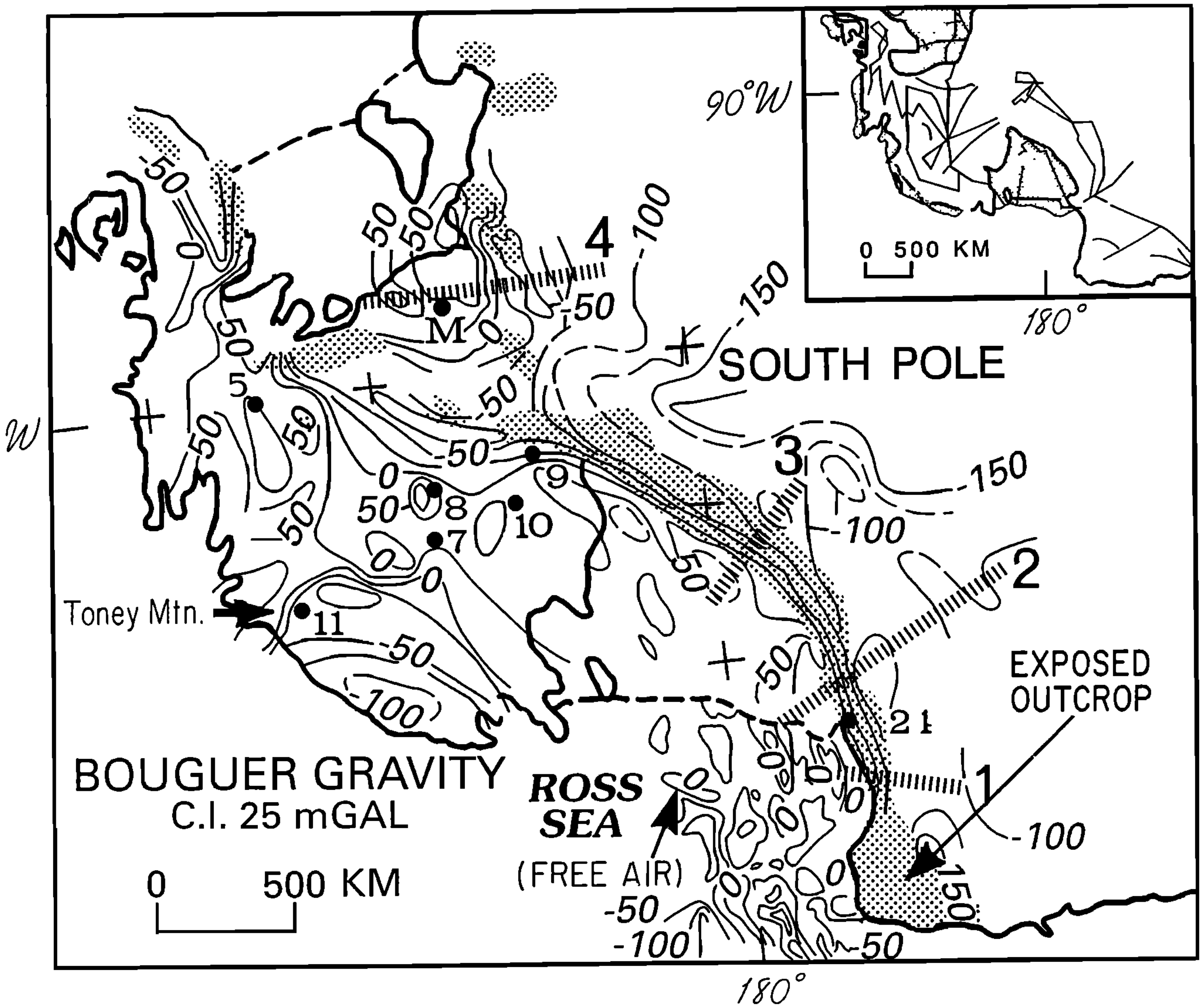


Fig. 13. Bouguer anomaly contour map for West Antarctica compiled from data collected at seismic reflection stations (about 30–40 km spaced) where ice thickness measurements were made by oversnow traverse parties led from 1956 to 1964 by J. C. Behrendt, C. R. Bentley, F. K. Chang, A. P. Crary, M. Hochstein, 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 heavy hachures indicate areas where data density allowed reasonably accurate calculation of gradients: 1 [*Duerbaum et al.*, 1989], 2 [*Robinson*, 1964*a*; *Smithson*, 1972; *Robinson and Splettstoesser*, 1984], *3* [*Robinson*, 1964*a*; *Robinson and Splettstoesser*, 1984], and 4 [*Behrendt et al.*, 1974]. Seismic velocity columns (Figure 16) from *Bentley and Clough* [1972] are indicated: 11 (at Toney Mountain), 7 (Byrd Station), 8 (Whitmore Mountains), 5 (Ellsworth Mountains), 24 (Ferrar dolerite near McMurdo), and 10 and 9 (Horlick Mountains). M is Moho reflection from M. Hochstein, 1963–1964 (cited by *Bentley* [1973]).

Mountains on the East Antarctic side which extends from northern Victoria Land nearly to the Pensacola Mountains. *Bentley* [1983] attributes the source of this anomaly to large glacio-isostatic imbalance, to abnormally thick or low-density crust, or to a deep-seated cause in the upper mantle. Although low-density sedimentary rock in the Wilkes basin (Figure 2) could partly explain the gravity low, the interpretation of a thick (>600 km) lithosphere in East Antarctica [*Stern and ten Brink*, 1989], resulting in a flexurally controlled basin which is regionally compensated, is largely the cause. Based on flexural rigidity and gravity models, the front of the Transantarctic Mountains in the McMurdo area is postulated to have a much higher rigidity (10^{25} N m) for the East Antarctic "cantilevered lithospheric plate" than for the less rigid (4×10^{22} N m), stretched, hot lithospheric plate underlying the Ross Embayment [*Stern and ten Brink*, 1989]. Thus, the Transantarctic (free-air) negative gravity anomaly (Figure 4) is evidence for regional compensation of the load of the Transantarctic Mountains as modeled by *Stern and ten Brink* [1989].

Free-air anomalies are always locally positive in mountains because the compensating mass, irrespective of thickness or rigidity of the lithosphere, is at a greater distance below the gravity station than the rugged mountain topography, so positive free-air anomalies over mountains in Figure 12 are expected. This is apparent in the data from the oversnow traverses near the Ellsworth Mountains and the southern Antarctic Peninsula [*Behrendt*, 1964*b*; *Behrendt et al.*, 1974]. A more recent paper illustrating this point [*Simpson et al.*, 1986] discusses isostatic anomalies in the United States and notes that "large amplitude anomalies can be produced by crustal bodies in complete local isostatic equilibrium." The map in Figure 12 is biased by the high topography, which therefore produces positive mean free-air anomalies over the 2° squares in the southern Antarctic Peninsula [*Behrendt*, 1964*b*], Pensacola and Ellsworth Mountains area [*Behrendt et al.*, 1974].

Bouguer Anomalies

The regional Bouguer anomaly map (Figure 13) of the West Antarctic rift system and Transantarctic Mountains area is more useful for our purposes than the free-air anomaly map. Figure 13 is a compilation of Bouguer anomalies from various sources [*Robinson*, 1964*a*; *Behrendt*, 1964*b*; *Behrendt et al.*, 1974; *Robertson et al.*, 1982; *Robinson and Splettstoesser*, 1984]. Data from 19 U.S. Geological Survey gravity stations acquired in the Beardmore Glacier area (collected by M. Hower and G. Perasso in 1985–1986) were used. We also included all the original 1957–1964 U.S. oversnow traverse data (used by *Bentley* [1968], who averaged the data over 2° × 2° squares and therefore smoothed the gradients over the rift flanks).

In using the Bouguer anomaly map, errors due to uncertainties in elevation (as great as ±50 m for the oversnow traverse data) and terrain effects must be kept in mind. The data in the mountains have the best elevation control but are subject to the greatest terrain effects. For example, the approximately 400 stations in the Pensacola Mountains area [*Behrendt et al.*, 1974] are estimated to have ±25 m absolute elevation error, corresponding to about ±5 mGal Bouguer anomaly error for stations on bedrock. Terrain corrections in the mountains are the most difficult to evaluate and could be as great as several tens of milligals. However, for most stations, terrain effects probably are 10 mGal or less [*Behrendt et al.*, 1974]. Terrain corrections for subglacial bedrock topography are not possible for most of the data used to compile the Bouguer anomaly map because of the scale and contour interval available for most maps (1/250,000 scale with 200-m contour intervals). Terrain corrections over the ice sheet are not possible with existing subglacial bedrock elevation data but could be as great as ±20 mGal over rugged subice topography buried by 1 km of ice [*Behrendt*, 1964*b*]. Most of the ice over the area depicted in Figure 13 is about 3 km thick, so probably the terrain errors are mostly less than 10 mGal here as well.

Considering all of the possible errors, we contoured the Bouguer anomaly data at a conservative 25-mGal interval and will only discuss the large regional anomalies ranging from 100 to 200 mGal in magnitude. Although the lack of data in the Marie Byrd Land–Ellsworth-Whitmore Mountains area is apparent, the map is nevertheless informative regarding the tectonics of the West Antarctic rift system.

Bouguer anomalies provide the primary evidence for the thinning of the crust throughout the ice-covered Byrd Subglacial Basin area of the West Antarctic rift system. Early crustal thickness maps based on Bouguer anomalies showed approximately 30-km-thick crust over the area of the West Antarctica rift in contrast with the approximately 40-km-thick crust in East Antarctica [*Bentley et al.*, 1960; *Woollard*, 1962; *Groushinsky and Sazhina*, 1982], calculated from the difference in Bouguer anomaly between these two areas. However, these regional studies were based on comparisons with other parts of the world and contributed information similar to that obtained by averaging regional elevations. There are no seismic refraction or reflection measurements of Moho depth anywhere in the ice-covered Byrd Subglacial Basin area. In this paper we interpret the data to indicate that the crust beneath the rift (Byrd Subglacial Basin–Ross Embayment) is probably closer to 20 km thick by adjusting the *Bentley et al.* [1960] and *Woollard* [1962] regional interpretation tied to the seismic determinations of 17–21 km depth to the Moho beneath the Ross Sea continental shelf. (See the seismic results below.) This value is typical for rift stage crust [e.g., *Klitgord et al.*, 1988].

Rift Shoulder Anomaly

One of the largest-magnitude Bouguer anomalies in the world was recognized as early as 1962 [*Woollard*, 1962] as marking the transition from West Antarctica to East Antarctica in the Ross Embayment (Figure 13). 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. We did not compute Bouguer anomalies over the Ross Sea shelf but include the free-air anomaly map of *Davey and Cooper* [1987] for this area. For a mean water depth of 500 m, a Bouguer correction of +34 mGal would need to be added to the regional free-air anomaly field over the Ross Sea shelf. Because water depths vary from 400 to 1400 m (e.g., Bouguer corrections of 28 to 96 mGal) over the shelf, caution should be used in using the free-air map. *Davey and Cooper* [1987] used bathymetry and seismic reflection data along specific free-air anomaly profiles in their interpretations.

Many authors [*Robinson*, 1964*a*; *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 between West Antarctica and East Antarctica as a result primarily of an abrupt change in crustal thickness across the front of the Transantarctic Mountains. There are significant regional differences, however, as revealed by these models along the Transantarctic Mountains.

If we examine the 200-mGal change at the front of the Transantarctic Mountains near McMurdo and 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 change, then the change in crustal thickness, using 41.85 mGal/km for a density contrast of 1.0 g/cm^3, $\Delta h = 200/(0.4 \times 41.85) = 12$ 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 0.24 g/cm^3 would be implied.

This value is consistent with those calculated by *Robinson* [1964*a*] and *Robinson and Splettstoesser* [1984]. There is a significant difference between the maximum 200-mGal change across the front of the Transantarctic Mountains near McMurdo (profile 2 in Figure 13) and the maximum 130-mGal anomaly range at the front of the Transantarctic Mountains in the Pensacola Mountains area.

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*a*; *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 14*b* and 14*c* along profile 2 in Figure 13. A common feature of these models (e.g., Figure 14*c*) 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.

In contrast, the steepest gradient in the Pensacola Mountains area at the Filchner Ice Shelf is only about 2 mGal/km (Figure 14*d*) [*Behrendt et al.*, 1974]. 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]. The model is tied to a 24-km depth to the Moho on the West Antarctica side of the Pensacola Mountains determined by wide-angle seismic reflection (Figure 14). In this Pensacola Mountains profile, a crustal thickness increase of only about 8 km was required across the front of the Transantarctic Mountains to account for the observed gravity difference. One explanation for this difference is that the West Antarctic rift system trend, which is superimposed in the Ross Embayment area on the Jurassic Transantarctic rift, diverges at the inland end of the Ross Ice Shelf and continues along the Ellsworth-Whitmore Mountains trend.

Another possibly more likely explanation for the lower 2-mGal/km Bouguer anomaly gradient in the Pensacola Mountains is the deliberate omission of the Dufek intrusion in the location of the profile selected for modeling [*Behrendt et al.*, 1974]. Although it was not possible to measure a comparable gravity gradient for stations on the exposures of the Dufek intrusion because of variable erosion and terrain effects, *Behrendt et al.* [1974] calculated about 85-mGal maximum gravity effect of the thickest section of gabbro. For a density contrast of 0.27–0.33 g/cm^3, 85 mGal corresponds to 8.8 to 6.2 km total thickness of the Jurassic tholeiitic intrusion. We suggest that possibly the exposed Dufek intrusion is an example of underplated and shallow- to lower-crustal intrusions which we infer to exist elsewhere along the Transantarctic Mountain front, where gradients of 4–7 mGal are observed. If the Dufek intrusion was buried by several kilometers of sedimentary rock, the high gradient would result.

We examined maximum possible gravity gradients in the Ross Embayment area with a simple extreme model (Figure 14*a*), again assuming a density differential of 0.4 g/cm^3 across the Moho to explain the gravity difference. Various depths to the Moho determined by seismic investigations are reported beneath the Ross Sea shelf: at McMurdo Sound (21 km) [*McGinnis et al.*, 1985], at Central basin (17–21 km) [*Trehu et al.*, 1989], and near the coast beneath the Transantarctic Mountains at about 74°34′S (22 km) [*O'Connell et al.*, 1989] (Figure 15). If we assume a 20-km Moho depth on the Ross Sea side of the Transantarctic Mountains with a vertical step to 40 km at the front of the mountains, we obtain a gradient of 4 mGal/km significantly lower than the maximum of 7 mGal/km in the McMurdo area reported by *Smithson* [1972] (Figure 14*c*). This model also produces an unreasonable maximum difference of 334 mGal for an infinite slab compared with about 200 mGal observed. Therefore we conclude (as did *Smithson* [1972]) that a significant excess mass, such as a high-density intrusion like the Dufek intrusion, exists at a shallow depth within the crust on the West Antarctic side of the Transantarctic Mountains. Because of the steep gradient observed, contrast in mantle density as great as 0.4 g/km^3 could not account for the anomaly. *Robinson and Splettstoesser* [1984] (Figure 14*b*) also required a high-density (3.0 g/cm^3) layer within the crust to fit the observed gradient in the McMurdo area. These results are consistent with gravity models for marine data in the Ross Sea [*Davey and Cooper*, 1987], although such models are

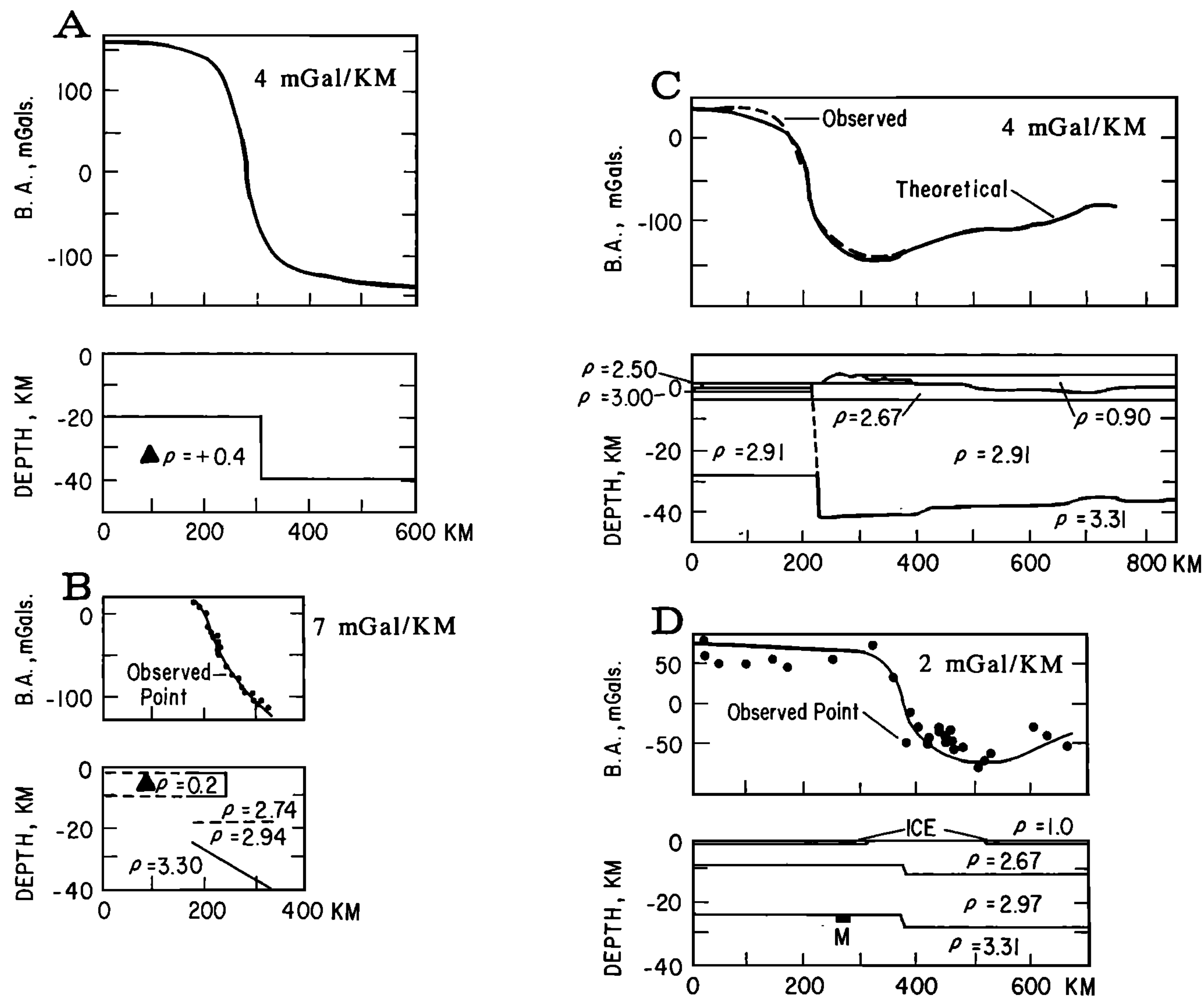


Fig. 14. Theoretical and observed Bouguer anomaly profiles for several models with maximum gradients indicated. (*a*) Model for an assumed crustal thickness change of 20 km from 20 to 40 km with a simple density contrast of 0.4 g/cm^3 across an assumed Moho. The maximum gradient is 4 mGal/km and the total anomaly range is about 330 mGal. This type of model cannot explain the observed data in the profiles shown in Figures 14*b*–14*d* and Figure 13 but is required to produce the steep gravity gradient solely by a change in crustal thickness. (*b*) Model approximately along 2 in Figure 13 [*Smithson*, 1972]. (*c*) Also approximately along 2 in Figure 13 [*Robinson*, 1964*a*, *b*; *Robinson and Splettstoesser*, 1984]. (*d*) Approximately along 4 in Figure 13 across the Pensacola Mountains [*Behrendt et al.*, 1974]. M is Moho reflector from M. Hochstein, 1964 (cited by *Bentley* [1973]).

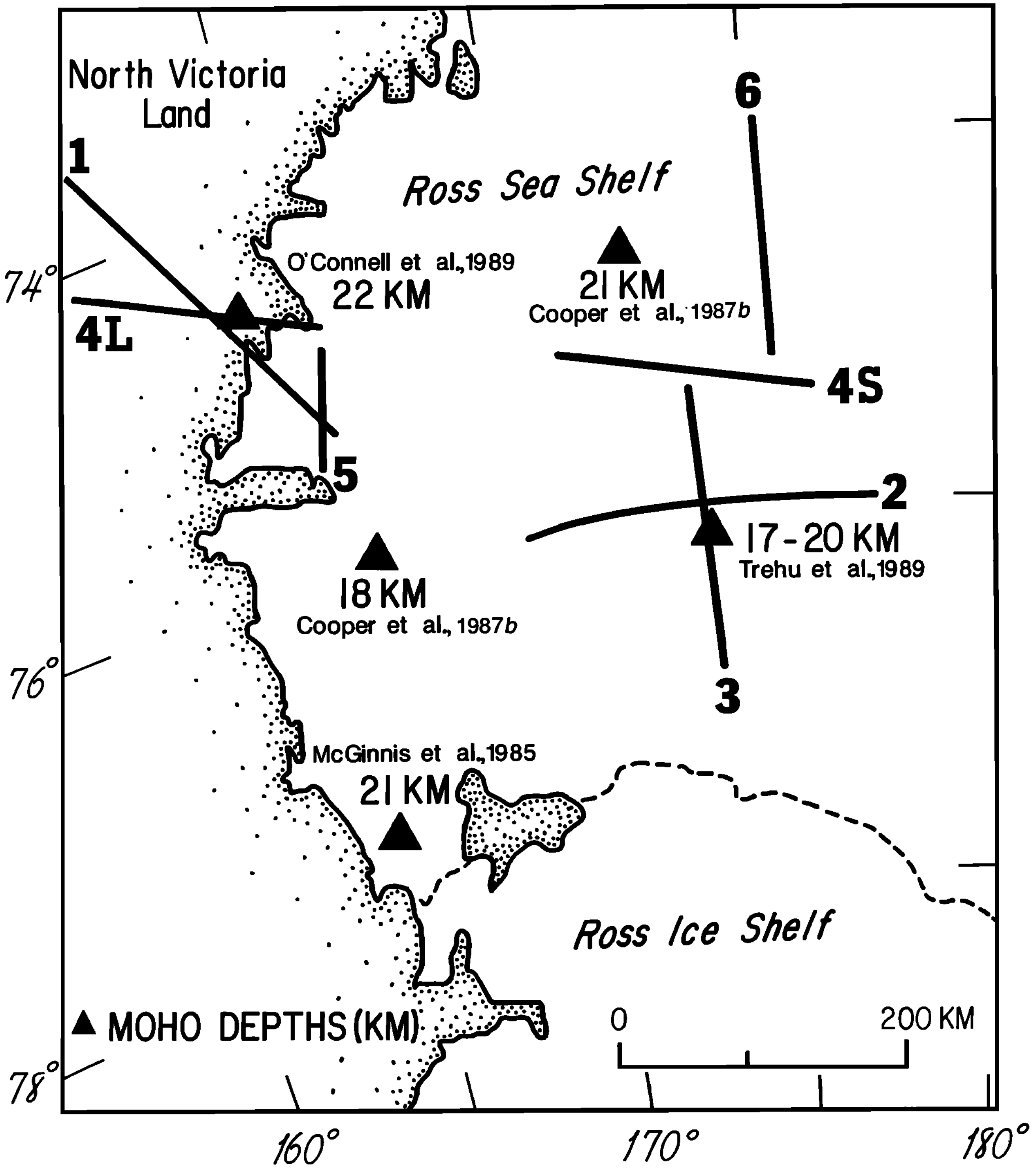


Fig. 15. Locations of large offset seismic experiments along profiles in the Ross Sea shelf–northern Victoria Land area. Profiles 1–6 are from the 1988–1989 GANOVEX V expedition. Profile 3 is along the strike of the Central basin (see Figure 7).

complicated by the low-density Cenozoic sedimentary rocks filling the rift basins beneath the Ross Sea shelf. The marine profiles, of course, did not cross the Transantarctic Mountains rift shoulder but did match part of the steep gradient. We discuss seismic evidence for high-density rock at the East Antarctica–West Antarctica transition in the next section.

We can conclude from the sparse Bouguer anomaly data (Figure 13) across the Transantarctic Mountains that the significant difference in the observed Bouguer anomaly range (about 200 mGal in the Ross Embayment and 130 mGal in the Pensacola Mountains–Filchner Ice Shelf area) and the difference in gradient (about 4–7 mGal/km and 2 mGal/km, respectively) indicate that greater crustal thinning and a notable excess mass are present at this boundary beneath the Ross Embayment. We interpret the excess mass to be associated with shallow mafic intrusions within the attenuated crust in the West Antarctic rift system.

Victoria Land Basin

Important questions remain about the possible presence of underplated or shallow-crustal high-density intrusions suggested by the 4- to 7-mGal/km gradients or thick sections of basalt within the rift basins. The high-density basin fill (2.7 g/cm^3) used below 5–7 km to fit the gravity anomaly over the 14-km-thick section in the Victoria Land basin [*Davey and Cooper*, 1987; *Cooper et al.*, 1987*a*] suggests that a substantial part of the section could be high-density basalt flows contrasted to lower-density sedimentary rock, possibly similar to those observed in the Midcontinent rift system of North America [*Behrendt et al.*, 1990]. The high-density mass beneath the Ross Sea basins was also shown by *Hayes and Davey* [1975] and inferred in the Ross Embayment [*Cooper et al.*, 1991]. We consider this supportive of the unwarped Curie isotherm hypothesis for the negative magnetic anomaly discussed previously; i.e., high-density basalt could have a low magnetization because of shallow high temperatures. The high seismic velocities, 5.6 km/s at 6.8 km depth within parts of the basin [*Cooper et al.*, 1987*b*], also support the basalt interpretation as discussed below.

Marie Byrd Land Area

The regional Bouguer anomaly gradient associated with the front of the Transantarctic Mountains in the Ross Embayment continues along the West Antarctic rift system shoulder to the Ellsworth Mountains (Figure 13). The Bouguer anomalies range from a minimum observed value of about –75 mGal in the Whitmore Mountains (not contoured because the data are too sparse) to about +50 mGal over the Byrd Subglacial Basin. We know from the combination of magnetic depth estimates discussed above and the limited seismic refraction data discussed in the next section 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 maximum range and gradient there.

The only places along the West Antarctic rift system rift shoulder where gradients equal to or steeper than 4 mGal/km have been measured are where the Cenozoic and Jurassic magmatic rock exposures are approximately coincident. Along the Ellsworth-Whitmore-Horlick Mountains section of the Cenozoic rift shoulder no exposures of either Jurassic tholeiites or late Cenozoic volcanics are known.

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 as across the Ellsworth-Whitmore side of the rift, which again may demonstrate the asymmetry of the West Antarctic rift system, although caution should be taken when inferring gradients in areas of widely spaced data. The 150-mGal difference corresponds to 9-km difference in depth to the Moho rather than the 12 km calculated above for the 200-mGal difference across the rift shoulder, assuming the same 0.4 mGal/km across the crust-mantle boundary only.

Southern Antarctic Peninsula–Ellsworth Mountains Area

Steep gradients of 100 mGal or greater exist in the southern Antarctic Peninsula–Ronne Ice Shelf area (Figure 13). The maximum range is from about –70 to +60 mGal [*Behrendt*, 1964*b*]. Minimum values are probably not as negative over the narrow southern Antarctic Peninsula because of the geometric effect of the high topography on the gravity field as pointed out for the two isostatic models calculated by *Behrendt* [1964*b*]. As was discussed in the magnetics section, we are uncertain where the West Antarctic rift system terminates toward the Ronne Ice Shelf. *Garrett et al.* [1987] and *Kellogg and Rowley* [1989] discussed evidence for Cenozoic extension between the Antarctic Peninsula and the Ellsworth Mountains and right-lateral strike-slip faulting (the Weddell fault system) along the east side of the Antarctic Peninsula. The Ellsworth fault system [*Kellogg and Rowley*, 1989] lies within the complex topography separating the Antarctic Peninsula and the Ellsworth Mountains and forms a boundary with the area of extended crust suggested by gravity data (Figure 13) and the linear prominent (>1000 nT) magnetic anomalies (Figure 2).

SEISMIC INVESTIGATIONS

Large Offset Surveys—Ice Sheet

Multichannel seismic reflection data (analog) collected on the oversnow traverses in the area shown by the Bouguer anomaly data in Figure 13 were primarily made to determine ice thickness; few reflections from deeper than the ice bed were reported. The only velocity control, used to interpret vertical incidence reflection profiles, is from large offset (reflection and refraction) surveys. Refraction results obtained for the upper crust 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].

Marie Byrd Land–Horlick Mountains Area

Figure 16 shows a seismic velocity column across the West Antarctic rift system; the locations of stations (11, 7, 10, and 9) along the profile are indicated in Figure 13. The 4.5-km/s velocity measured at Toney Mountain can unambiguously be associated with andesitic-basaltic flows exposed nearby (Figure 2). In the same profile the 6.1-km/s layer is interpreted as high-grade metamorphic(?) basement. In the Byrd Station profile the 4.3-km/s velocity is also correlated with magnetic volcanic rock [*Behrendt and Wold*, 1963], on the basis of the magnetic interpretations discussed previously. The distinct velocity change [*Bentley and Clough*, 1972] on either side of the axis of the Byrd Subglacial Basin (compare columns 7 and 10 in Figure 16) can be seen. The break in the magnetic anomaly pattern in this area (Figures 5 and 6) discussed in the magnetics section also occurs between these two profiles. Seismic refraction data [*Bentley and Ostenso*, 1961; *Bentley and Clough*, 1972] show about 1 km of 5.2–5.3 km/s velocity overlying 6.3-km/s and 6.1-km/s rock, respectively, in the area interpreted as thick nonmagnetic rock (east of break in Figures 5 and 6). These velocities could indicate granitic terrane as originally suggested [*Bentley and Ostenso*, 1961], on the basis of the nonmagnetic nature of the subice rock; however, as one of these refraction determinations (profile 5 in Figure 13) is close to the Ellsworth Mountains, we still subscribe to the interpretation of *Behrendt and Wold* [1963] that the velocities correspond to the thick metasedimentary rock section (lower Paleozoic) exposed in the Ellsworth Mountains.

Profile 10 in Figure 16 near the Horlick Mountains is on the Cenozoic West Antarctic rift shoulder, as defined in this paper. The 6.7-km/s and 7.0-km/s velocities are more typical of lower crust [*Bentley and Clough*, 1972] and are similar to others reported in the area away from profile C—C′ in Figure 16. These high velocities are also similar to those measured in McMurdo Sound [*McGinnis et al.*, 1985; *Kim et al.*, 1986] and beneath the Ross Sea shelf [*Cooper et al.*, 1987*b*; *Trehu et al.*, 1989]. We interpret these high velocities, typical of lower crustal rocks, to be associated with mafic magma intruded during rifting. Because the Horlick Mountains are located where the Jurassic Transantarctic rift and Cenozoic West Antarctic rift shoulder are coincident, however, the high seismic velocities could be the result of either Early Jurassic or Cenozoic intrusions or both. A velocity of 6.8 km/s (profile 24 in Figure 13) [*Bentley and Clough*, 1972; *Robinson and Splettstoesser*, 1984] was reported for the Jurassic Ferrar dolerite near McMurdo, whereas a velocity of 4–5 km/s is typical of a Cenozoic basalt (as measured, for example, at Toney Mountain [see *Bentley and Clough*, 1972]). However, profile 8 in Figure 13 near the Whitmore

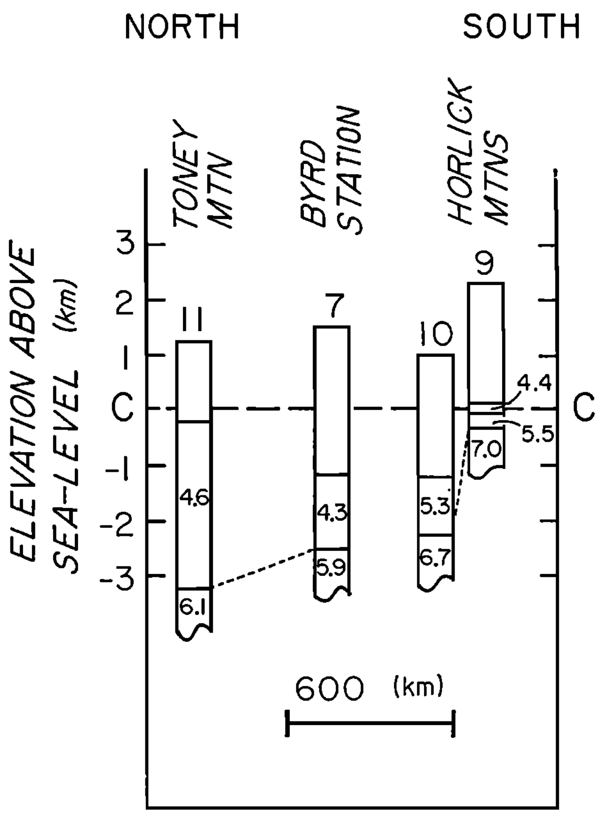


Fig. 16. Seismic velocity columns 11 to 9 along the section (see Figure 13), measured from the 1956–1962 oversnow traverses (section C—C′), from *Bentley and Clough* [1972]. Velocities are shown in kilometers per second. Correlations between columns are shown by dashed lines. The blank upper part of each column corresponds to ice $V_p = 3.9$ km/s. We interpret the velocities as follows. The 4.6-km/s velocity at profile 11 (at Toney Mountain) is correlated with exposed volcanic rock. The 4.3-km/s velocity at Bryd Station (profile 7) is also correlated with volcanic rock because of shallow sources of magnetic anomalies in the area. At profile 9 (near the Horlick Mountains) (see Figure 2), the 4.4-km/s velocity layer is probably Beacon Supergroup Paleozoic-Mesozoic sedimentary rock; the 6.7-km/s velocity at profile 10 and 7.0-km/s velocity at profile 9 are interpreted as mafic or ultramafic intrusive rock. The 5.9–6.1-km/s velocities at profiles 7 and 11 are probably crystalline granitic basement, as are possibly the lower 5.3-km/s and 5.5-km/s velocities at profiles 10 and 9.

Mountains (about 100 km on the rift side) exhibits an approximate 7 km/s velocity about 4 km below sea level underlying 5.8-km/s basement [*Bentley and Clough*, 1972]. This area overlies the interpreted rift well away from the magmatic rocks of the Jurassic Transantarctic rift system, 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 [*Behrendt et al.*, 1974], interpreted as evidence of a thick section of low-density nonmagnetic sedimentary rock at the Moho reflection location (Figure 13) northwest of the Pensacola Mountains. No high-density mass was required within the crust there to fit a model to the observed Bouguer anomaly profile (Figure 14*d*).

McMurdo Sound

McGinnis et al. [1985] and *Kim et al.* [1986] discussed deep reflection and refraction profiles in McMurdo Sound that involved using land techniques on sea ice. *McGinnis et al.* [1985] observed a 6.5-km/s refractor as shallow as 5.4 km beneath a 5.0-km/s refractor on a north-south profile. On another east-west profile they report a 7.2-km/s refractor in the lower crust. The velocity below the Mohorovic discontinuity, which occurs at 21 km depth, is 8.2 km/s (Figure 15). Their 12-fold reflection profile shows bands of reflections in the lower crust similar to those interpreted elsewhere as caused by underplated basalt associated with rifting [e.g., *Behrendt et al.*, 1988, 1989]. A lower crustal velocity of 7.2 km/s 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. *McGinnis et al.* [1985] assume a 2.94-g/cm^3 density for the 6.5-km/s layer in a gravity model that is seaward of the steepest part of the gradient measured by *Smithson* [1972]. Therefore we consider this density and velocity consistent with our previous discussion of gravity data. *Kim et al.* [1986] show bands of reflections and diffractions as late as about 7.8 s beneath the Victoria Land basin and suggest these are sideswipe, but in appearance they are quite similar to reflections from probable underplated material reported for other rifts [e.g., *Behrendt et al.*, 1988, 1990].

Marine Large Offset Surveys—Ross Sea

In striking contrast in number to the sparse large offset seismic surveys done over ice sheets (only about 30–40 profiles on "land" in 33 years), marine data are several orders of magnitude more efficient to collect than land data. For example, the R/V *S.P. Lee* obtained 39 sonobuoy profiles over the Ross Sea shelf in a few weeks in 1984. Interpretations generally were made only to midcrustal depth, but some resulted in probable Moho reflections. *Cooper et al.* [1987*b*] discussed these results, and *Davey et al.* [1983] discussed 98 older, shallow-penetration sonobuoy profiles.

Velocities reported for the sedimentary rocks in the three basins beneath the Ross Sea shelf (Figures 7 and 17) 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.*, 1987*a*, *b*]. Because these velocities are significantly higher than the 4.4 km/s measured for the Devonian to Jurassic Beacon Supergroup rocks exposed nearby in the Transantarctic Mountains and buried beneath 1–2 km of ice elsewhere in the Transantarctic Mountains (e.g., profile 10 in Figure 16), and considering the 4.6 km/s measured for Cenozoic volcanic rocks at Toney Mountain (Figure 16), the rocks are probably in fact rift-related volcanic rocks of uncertain age. Whether or not these layered [*Cooper et al.*, 1987*a*] rocks (in the deepest Victoria Land basin) are Cenozoic, they must predate the overlying sedimentary rocks which *Cooper et al.* [1987*a*] interpret to be as old as early Oligocene, from correlations with the CIROS-1 drill site (Figure 7) [*Barrett*, 1989] on the flank of the Victoria Land basin. Ages of the deep, high-velocity layered reflectors probably are from Late Mesozoic to Paleogene [*Cooper et al.*, 1987*a*]. Another hypothesis to explain the high velocities would be diagenetically altered or metamorphosed sedimentary rock. We think this is unlikely considering they are probably not older than the Devonian-Jurassic Beacon Supergroup rocks which have lower velocity (4.4 km/s) where measured [*Bentley and Clough*, 1972].

The highest crustal velocities measured in the Ross Sea [*Cooper et al.*, 1987*b*] 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 beneath the Coulman high (Figure 7). *Cooper et al.* [1987*b*] reported a velocity of 6.5 km/s from within the central region of the Victoria Land basin at a depth of 9.5 km as acoustic basement on the reflection profiles. Although *Cooper et al.* [1987*b*] interpreted rocks with velocities greater than 6.3 km/s as igneous-metamorphic basement, here we refine this to 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. As mentioned in the gravity section, these rocks may be the source of the high densities required within the crust to fit the observed steep gravity gradients and a high range of total Bouguer gravity values and are additional evidence of rifted extended crust.

Moho Determinations—Ross Sea Shelf

Cooper et al. [1987*b*] report deep reflections at two sites which we interpret as the Moho because they have depths similar to other Moho determinations discussed below. The corresponding depths to Moho

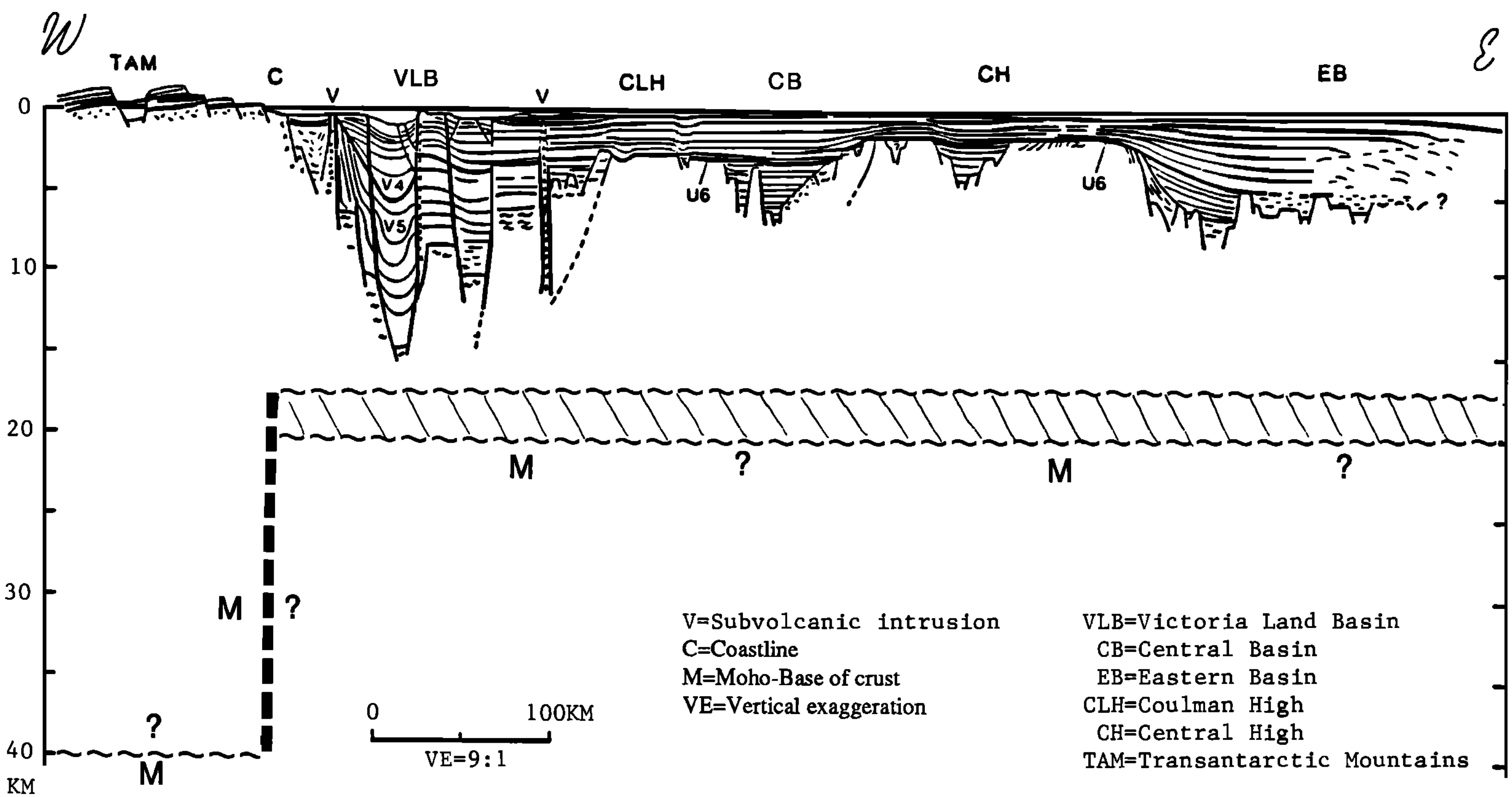


Fig. 17. Cartoon illustration of the geologic structure across the Ross Sea shelf, modified from the seismic reflection interpretation of *Cooper et al.* [1990] for the upper crust and based on preliminary seismic crustal thickness determinations from GANOVEX V (see Figure 15). Unconformity U6 [*Hinz and Block*, 1983] is correlated with that between acoustic units v4 and v5 [*Cooper et al.*, 1987*a*]. 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 7. The vertical dashed line indicates that the Moho probably changes thickness by up to 20 km at the rift shoulder (Transantarctic Mountains front); however, the dip of the Moho in the transition zone is unknown but is probably not vertical or even necessarily steep. Note that the old crust has been nearly rifted through beneath the Victoria Land basin.

are 18 km beneath the central Victoria Land basin and 21 km beneath the Coulman high (Figure 15).

BGR and USGS (GANOVEX V, 1984–1985) conducted a large offset seismic experiment on land and at sea using a 43.5-L air gun source (Figure 15). From these results *Trehu et al.* [1989] reported Moho depths of 17–21 km beneath the Central basin with a high-velocity lower crustal transition layer, 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.

The 17–21 km depths to Moho beneath the Ross Sea shelf are typical of extended and rifted continental crust, as described and defined by *Klitgord et al.* [1988] beneath the east coast area of North America, and the greater than 7 km/s velocities reported in the lower crust in the Ross Embayment area are typical of transitional crust. The very shallow high velocities (6.9–7.4 km/s) beneath the continental shelf are unusual.

Crustal Thickness Beneath the Ross Embayment–Byrd Subglacial Basin

As discussed previously, the Bouguer anomaly map (Figure 13) indicates approximately the same range (0 to +50 mGal) throughout the interpreted extended crust beneath the West Antarctic rift system from the Ross Embayment to the Byrd Subglacial Basin. Therefore, by comparison with measured depths to the Moho beneath the Ross Sea shelf, we conclude that the Moho depths throughout this topographically low area of the rift are more likely closer to 20 km than to 30 km, as previously reported [e.g., *Bentley et al.*, 1960; *Woollard*, 1962]. The earlier interpretations did recognize the continuity of what we now interpret as extended rifted crust, from the Ross Ice Shelf through the Byrd Subglacial Basin to the Bellingshausen Sea. There is no evidence for oceanic crust beneath the Byrd Subglacial Basin.

The 24-km depth to Moho obtained in 1963 from large offset seismic results by M. Hochstein (Figures 13 and 14*d*) [*Behrendt et al.*, 1974] northwest of the Pensacola Mountains appears more reasonable for the low topography in that area of West Antarctica than was realized at the time. The question remains whether the possible extended crust (Figure 14*d*) bordering the Pensacola Mountains in the Weddell Sea–Filchner Ice Shelf is the result of Jurassic Transantarctic rifting or Cenozoic rifting. The absence of any magmatic rocks younger than Jurassic suggests the former, but the faulted uplift of the Transantarctic Mountains in this area requires tectonic activity postdating the Jurassic Dufek intrusion (but possibly as old as Jurassic).

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 7); data from U.S. and German research have been described by *Hinz and Block* [1983], *Hinz and Kristofferson* [1987], *Cooper et al.* [1987*a*], 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 ages. Acoustic basement may be composed of metasedimentary and/or igneous units of Precambrian to late Mesozoic age. A schematic interpretation of these data along one profile across the three basins in the Ross Sea is shown in Figure 17.

The only control for age and lithology of interpreted reflectors offshore is provided from sparse drilling in the central (Deep Sea Drilling Project (DSDP)) and western (CIROS and McMurdo Sound Sedimentary and Tectonic Study (MSSTS)) Ross Sea (Figure 7) and is inferred from recycled and glacial-erratic materials found within the Ross Sea region (see summaries by *Hayes and Frakes* [1975], *Webb* [1981], *Truswell* [1983], *Barrett* [1986], *Davey* [1987], and *Barrett et al.* [1987, 1991]). Glacial-marine sedimentary rocks of early Oligocene and younger age comprise the upper part of the thick offshore sedimentary sections. Unconformities occur throughout the sedimentary section, with major events in the late Oligocene and late Miocene which we correlate with suggested episodic uplift of the Transantarctic Mountains. Mesozoic and lower Paleogene rocks have not been sampled, but *Cooper et al.* [1987*a*] suspected the presence of these within their inferred deeper sedimentary section on the basis of recycled microfossils in drill cores, coastal glacial erratics, and seafloor cores in the eastern Ross Sea. However, on the basis of large offset seismic data discussed above, we speculate that these rocks may also comprise layered volcanic sequences. Basement rocks have been recovered only at DSDP site 270 (Figure 7*b*) on a basement high and are composed of early Paleozoic(?) gneiss [*Hayes and Davey*, 1975].

The uppermost part of the sedimentary section is cut regionally by the Ross Sea unconformity, which lies 2 to 42 m below the seafloor [*Karl et al.*, 1987] and spans the interval 14.7 to 4.0 Ma [*Savage and Ciesielski*, 1983]. Another major unconformity, which cuts the sedimentary sections and tops of basement ridges throughout the Ross Sea, lies at 3 to 6 km depth. *Hinz and Block* [1983] name this unconformity U6 in the Eastern basin and Central basin (Figure 17), and *Cooper et al.* [1987*a*] refer to it as the unconformity between acoustic units V4 and V5 in the Victoria Land basin. Reflection units below unconformity U6 are isolated within early rift stage grabens [*Cooper et al.*, 1991], whereas units above U6 cover the entire region. The early rift grabens probably extend south throughout the West Antarctic rift system.

Rocks have not been sampled from below the unconformity U6, with the exception of early Paleozoic(?) basement rock at DSDP site 270 (Figure 7*b*). Glacial sedimentary units have been sampled only above U6; similar units have been sampled only above basement highs. U6 is late Oligocene in age at DSDP site 270 on the Central high [*Hinz and Block*, 1983], but rocks from a unit above unconformity U6 are of early Oligocene age at core hole CIROS-1 (Figure 7*a*) on the uplifted flank of the Victoria Land basin [*Cooper et al.*, 1990]. Unconformity U6 therefore appears time-transgressive and is probably Eocene to Oligocene in age. Drilling data are needed to control the ages on these strata so that the timing on the stages of rifting and associated crustal extension can be determined.

Victoria Land Basin

The Victoria Land basin (Figures 7 and 17) is characterized by a thick (up to 14 km) layered section, presumed to be sedimentary by *Cooper et al.* [1987*a*] and considered by us to be partly volcanic. Extensive basement faulting (early rift) occurred in the Victoria Land basin and was followed by faulting and deformation of the sedimentary section in the Terror rift (late rift). In contrast, large offset (1 to 2 km) faults along the east flank of the early-rift graben do not disrupt the nearly flat-lying sedimentary strata that fill the graben (Figure 17), indicating that downfaulting of the basement occurred mostly prior to, rather than during, deposition of the overlying sedimentary/volcanic section. Strata above and below U6 are disrupted in the Terror rift by large normal faults that extend from the basement to the seafloor (Figure 17) [*Cooper et al.*, 1987*a*]. These faults and nearby subvolcanic intrusions (e.g., Figure 10) are thought to be late Paleogene and Neogene (e.g., late rift) features that were emplaced after deposition of most of the sedimentary section. Faulting along the west side of the Victoria Land basin has occurred in part during the Eocene and younger uplift of the Transantarctic Mountains and has resulted in at least three major angular unconformities [*Cooper et al.*, 1987*a*]. Their existence is further evidence for episodic uplift and extension.

Eastern and Central Basins

Seismic reflection data from the Eastern and Central basins (Figures 7 and 17) show that the maximum sedimentary thickness is 6–8 km [*Hinz and Block*, 1983]. Basement rocks beneath the center and flanks of these basins appear highly eroded on reflection interpretations. Deformation of the sedimentary section is generally minor throughout the basins, and reflectors are disrupted only locally within intrabasement grabens and over basement structures. The Central basin overlies extended crust, as is evident from large offset seismic data where the Moho is 17–19 km deep [*Trehu et al.*, 1989]. The short-wavelength magnetic anomalies correlated with shallow volcanic sources (Plate 2) that are present in the general area of the Central basin have not been crossed by seismic reflection profiles. Therefore the interpretive cartoon of Figure 17 does not indicate penetrative structures such as that labeled V in the Victoria Land basin.

Significance of Reflection Results to Study of West Antarctic Rift System

The seismic reflection results summarized in Figures 7 and 17 and discussed briefly here are relevant to the subject of the West Antarctic rift system specifically, as follows.

1. These 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. Most of the 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.*, 1990, 1991].
3. The interpretation of horst and graben structures in the Ross Sea (Figure 17) probably extends 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 parallel to the "grain" of the West Antarctic rift system (e.g., the Transantarctic Mountains shoulder, Plate 2, and Figures 2 and 9).
5. Together with the aeromagnetic data (Plate 2), 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 early reconnaissance geophysical studies of *Bentley* [1964], *Behrendt* [1964*a*], *Bentley and Clough* [1972], and *Bentley and Robertson* [1982] encompassed all of the interior of West Antarctica, and these data provide a provisional basis for inferring boundaries and basic characteristics of the entire rift system in West Antarctica. The gravity data suggesting crustal thinning beneath the West Antarctic rift (Byrd Subglacial Basin), which show similarities to those in the East African and Rio Grande rift systems, provide the basis for the suggestion that alkaline volcanism in Marie Byrd Land was related to the same major intracontinental rift system as the volcanoes in the western Ross Embayment [*LeMasurier*, 1978; *LeMasurier and Rex*, 1982, 1989].

Many of the recently published discussions of rifting cited in this paper [*Cooper et al.*, 1987*a*; *Stern and ten Brink*, 1989; *Tessensohn and Wörner*, 1991] have concentrated on specific areas (e.g., the Ross Embayment, western Ross Sea, and Transantarctic Mountains bordering it) and most of the data collected since 1950 and discussed here are also concentrated in the Ross Embayment area. However, we attempt to place our geophysical review in the larger context of the West Antarctic rift system using the earlier results. This world class rift system appears unusual compared with other rift systems in at least two respects: (1) even after allowing for isostatic adjustment after ice removal, the elevation is anomalously low (as is the North Sea rift system) throughout most of the area (Figure 2) compared to other rifts; and (2) Antarctica is nearly aseismic [*Kaminuma*, 1982; *Adams et al.*, 1985], whereas earthquakes are typically associated with active rifting. Although we cannot totally explain these observations, we offer the following suggestions.

Anomalous Low Elevations

In the rift model proposed by *Stern and ten Brink* [1989], they required a very low rigidity (4×10^{22} N m) for the lithosphere beneath the Ross Embayment, which is approximately equal in thickness to the crust (about 20 km) beneath the Ross Sea shelf (Figure 17), as interpreted from large-offset seismic studies. *Wörner and Viereck* [1989] in their discussions on Mount Melbourne (located in Figure 7 and Plate 1) noted evidence for relative and absolute subsidence of this volcano, in contrast to the high uplift rate of the Transantarctic Mountains. Although the Neogene sedimentary section in the Victoria Land basin is relatively thin [*Cooper et al.*, 1987*a*], the deep topographic depression of the continental shelf over the basin (Figure 17) suggests subsidence faster than sedimentary deposition and ice erosion. The high seismic velocity (7+ km/s) in both the upper and lower crust beneath the Ross Sea, as discussed above, can be interpreted as evidence for dense mafic underplating and crustal intrusion and would also contribute to rapid subsidence. All of these lines of evidence support a thin, hot, very weak, dense extended lithosphere which is subsiding rapidly. By comparison with the Bouguer anomaly map (Figure 13), we infer that similar crust and lithosphere is characteristic of the entire West Antarctic rift system and accounts for the anomalously low elevation. 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 so that its density equaled that of 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]. Another area with consistently low elevation during active rifting is the Midcontinent rift system of North America [*Behrendt et al.*, 1988].

Low Seismicity

Although reasonably high quality seismographs have operated continuously since the International Geophysical Year (1956–1957), the only unambiguous earthquake, M_b 4.5 [*Adams et al.*, 1985], occurred in East Antarctica in 1982. *Bentley* [1983, 1991] discussed seismicity and rejected the argument of ice quakes for three >4 magnitude events next to major coastal outlet glaciers. He considers it more likely that these earthquakes were associated with fault zones that control the location of the glaciers. One of these earthquakes occurred in the Rennick graben (Figure 7), which *Roland and Tessensohn* [1987] interpreted as an early onshore branch of the rift system. We find it intuitively difficult to accept that stress buildup in ice alone could produce a >4 magnitude earthquake.

Johnston [1987] provided a reasonable explanation for the anomalously low seismicity of Antarctica and Greenland, on the basis of earthquake suppression by large continental ice sheets. *Johnston* [1987] proposed two theoretical explanations based on (1) the effect of static overburden (preferred) and (2) control of crustal pore pressure by insulation from a significant amount of meteoric water. The first explanation appears to require a compressive stress regime [*Johnston*, 1987], which is likely true for the continent as a whole because Antarctica is surrounded by spreading centers. However, the evidence for active extension associated with the West Antarctic rift system casts doubt on its applicability here. The second explanation, based on suppression of pore pressure by ice, would not seem to apply to the seawater-covered area of the Ross Embayment, although it might well apply in the thick ice-covered area of the Byrd Subglacial Basin.

Microseismicity is associated with active volcanoes such as Mount Erebus on Ross Island, and seismicity has been associated with a volcanic eruption on Deception Island [*Kaminuma*, 1982] at the north end of the Antarctic Peninsula (far from the West Antarctic rift system). Microearthquakes and "ultramicroearthquakes" of probable tectonic origins were reported to occur every 2 days [*Kaminuma*, 1982], measured by a station in the Transantarctic Mountains about 120 km from Mount Erebus (and thus too far to detect its activity). From stations on Ross Island *Rowe and Kienle* [1986] also reported tectonic microearthquakes which they associated with rifting. We associate these with faulting in the Victoria Land basin (Terror rift), as is suggested by the occurrence of Holocene fault scarps (such as shown in Figure 11). Greater frequency of microseismicity would probably be detected elsewhere within the West Antarctic rift system were more seismographs deployed. *Berg et al.* [1989] (from xenolith data) infer at present an anomalously high temperature at shallow depth. Possibly the very low rigidity of the extended crust of the Ross Embayment [*Stern and ten Brink*, 1989] and very high heat flow likely [*Berg et* *al.*, 1989] in the area (including the Transantarctic Mountains) have led to rapid creep along faults, resulting in movement with few earthquakes. Therefore we do not consider the low seismicity significant evidence against the presence of an active rift; for example, the Rio Grande rift in the western United States would probably not be detectable solely from sparse seismograph coverage typical of that in Antarctica.

A fault with 300 m vertical offset cutting a Pliocene moraine in the Beardmore Glacier area [*McKelvey et al.*, 1991], numerous other faults cutting this moraine, and fault scarps cutting the seafloor (e.g., Figure 11) over the Victoria Land basin imply large earthquakes in the past, although one could consider aseismic creep similar to that reported for a fault scarp in Death Valley [*Sylvester and Bies*, 1986]. We infer that fault scarps, such as the one shown in Figure 11, must have formed after the latest deglaciation, or else they would have been eroded by overriding ice. *Muir-Wood* [1989] discussed evidence for earthquakes and rapid stress release shortly after deglaciation in Fennoscandia. He interpreted 5 to 6 orders of magnitude higher earthquake frequency followed by an anomalously seismically quiet period at present. Possibly *Muir-Wood*'s [1989] results for Fennoscandia provide some explanation for the fault scarps we see in reflection profiles and the absence of earthquakes now in the active West Antarctic rift system, i.e., high seismicity immediately following deglaciation and low seismicity at present.

Okal [1981] concluded that although the Antarctic continent is essentially aseismic, the Antarctic plate has three times the area of the continent alone and is not aseismic. The total seismic energy released from the plate during 1925–1980 is comparable to that of the African and Nazca plates, which have similar tectonics. Okal concluded that the ring of spreading ridges does transmit stress, which he intrepreted as clearly showing the importance of ridge push as a driving mechanism for plate movement. Ridge push would result in a horizontal compressive stress regime in Antarctica favoring *Johnston*'s [1987] static overburden hypothesis for lack of seismicity but does not explain the extension in the West Antarctic rift system. The stress situation (i.e., a ring of spreading ridges around Antarctica) is similar in the African plate, where the East African rift developed in spite of this fact. The West Antarctic rift system is probably an analog. We do not mean to imply that a ring of spreading centers caused either rift.

Zoback et al. [1989], in discussing global patterns of stress and constraints on intraplate deformation, noted that whereas asthenospheric drag is a major source of stress in a number of plates, ridge push and forces due to plate collision and subduction cannot be ruled out. Unfortunately, data for the Antarctic plate are too sparse to make any attempt to examine possible correlation of shear stress and plate velocity. For both the Antarctic and Africa plates absolute motion is very slow [*Okal*, 1981], so asthenospheric drag would likely be very low in both cases.

Uplift of Rift Shoulder

Active extension in the West Antarctic rift system at least since early Oligocene [*Barrett*, 1989] is implied by the volcanic activity (Figure 2) and uplift of the rift shoulder. Late Mesozoic–early Cenozoic extension and basin downfaulting were probably even greater than in late Cenozoic time [*Cooper et al.*, 1982; *Cooper et al.*, 1987*a*; *Fitzgerald et al.*, 1986]. 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]. However, a growing body of evidence [*Behrendt and Cooper*, 1991] summarized in a previous section indicates rapid episodic uplift of the rift shoulder in the late Cenozoic and Holocene. We interpret this uplift along the Transantarctic Mountains bordering the Ross Embayment to mark the Cenozoic West Antarctic rift shoulder extending to the Ellsworth Mountains (but not to the Pensacola-Shackleton-Theron part of the Transantarctic Mountains) (Figure 3). If a maximum uplift of 10 km in northern Victoria Land [*Fitzgerald*, 1989] area is correct, then as much as 6 km of erosion has occurred in this area, resulting in the maximum altitude of about 4 km observed now. However, there is no direct evidence that the late Mesozoic and Cenozoic sedimentary rocks in basins beneath the adjacent continental shelf [*Cooper et al.*, 1987*a*] were ever deposited above the section exposed in the present rift shoulder ranges.

Mechanisms for Rifting

Because Antarctica is essentially surrounded by mid-ocean ridges, the only apparent source for stress to drive any tectonism is ridge push [*Okal*, 1981], yet it seems unlikely that this could be the source of the extension. Therefore another mechanism seems to be required. *Smith and Drewry* [1984] proposed that the rise of the Transantarctic Mountains is a delayed effect caused by overriding by East Antarctica of anomalously hot asthenosphere (hot spot) that formed under West Antarctica in Late Cretaceous time. In this model, the resultant increase in heat flow would result in phase changes in the upper mantle leading to uplift. *Fitzgerald et al.* [1986] pointed out that this model does not address the asymmetry (i.e., the high rift shoulder scarp on only one side) of the Transantarctic Mountains which they noted in the Ross Embayment.

In turn, *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. They estimated about 200 km (25–30%) of extension, and their model required simple shear in the rift basins as well as along a master décollement extending from the Central high to beneath the Transantarctic Mountains. In their estimate [*Fitzgerald et al.*, 1986], they assumed an average of 25–30 km crustal thickness for the attenuated crust beneath the Ross Sea shelf. The large offset seismic results obtained over the Ross Sea shelf (Figure 15) suggest that the crust beneath the shelf may be only 17–21 km thick, requiring even greater extension than the 200 km proposed by *Fitzgerald et al.* [1986]. The Byrd Subglacial Basin–Ross Embayment was probably extended greatly during late Mesozoic time, but a significant part of the total could have occurred during Cenozoic time, and would fall within constraints from plate reconstruction models [*Stock*, 1989*a*; J. M. Stock, personal communication, 1989]. If, for example, a 40-km-thick crust were stretched to 20 km across the rift now presently approximately 750 km wide, 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 would of course require still greater extension for a given present crustal thickness [e.g., *Behrendt et al.*, 1988, 1990] but would probably be localized (e.g., Figure 11) rather than uniformly distributed across the 750-km-wide rift system.

Stern and ten Brink [1989], although not disputing the asymmetry for the tectonic development of the Transantarctic Mountains, reject the simple shear model for faulting at the mountain front. They argue that this mechanism could not provide the observed magnitude (5 km) of uplift of the Transantarctic Mountains in the McMurdo area which they investigated. Instead they propose a model in which two lithospheric plates of vastly different effective thermal ages are juxtaposed and no shear stresses are transmitted across the boundary. They estimate flexural rigidities of 1×10^{25} and 4×10^{22} N m for the two plates, East Antarctica and the extended crust underlying the Ross Embayment, respectively. 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. Stern and ten Brink point out that their determinations are consistent with effective thermal ages for East Antarctica of about 600 Ma and the Ross Embayment of 25 Ma, which we find quite appropriate for the late Cenozoic volcanism observed.

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, thus providing the necessary thermal uplift. They note that this is within the 90–65 Ma proposed by *Smith and Drewry* [1984] for the initiation of hot spot overriding. *Stern and ten Brink*'s [1989] proposed 70 m.y. also is consistent with *Fitzgerald*'s [1989] interpretation of the start of uplift of the Transantarctic Mountains at 60 Ma. They 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 discussed above suggests that once it has started, it proceeds rapidly similar to other rift shoulders. This could account for the high Neogene uplift rates proposed here and by *Behrendt and Cooper* [1991]. The *Smith and Drewry* [1984] model has an additional attraction in that it suggests that the later volcanism observed in northern 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, which is again consistent with the 25-Ma thermal age calculated for the Ross Embayment [*Stern and ten Brink*, 1989] and the rapid uplift discussed above. However, the early Oligocene (or older) date 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 several of the observations from the geophysical data discussed in this paper, as follows: (1) evidence of extension seen throughout the area of the Ross Sea aeromagnetic survey (Figure 9 and Plate 2) and seismic surveys (Figure 17), (2) widespread submarine and subglacial volcanism 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. Very high heat flow based on xenolith thermobarometry [*Berg et al.*, 1989], a 40°C/km gradient in the CIROS hole [*White*, 1989], and the sparse *Blackman et al.* [1987] heat flow data are consistent with the presence of shallow volcanism inferred from short-wavelength magnetic anomalies (Plates 1 and 2 and Figures 8 and 10) in the Victoria Land basin area. High temperature in the upper crust also is consistent with the negative magnetic anomaly over the Victoria Land basin, interpreted as possible evidence of thermal demagnetization [*Behrendt et al.*, 1987, 1991] of layered volcanic rocks which we infer to be present in the deepest 8 to 9 km of the basin. Possibly the extensive magmatic crustal intrusion and underplating which we interpret from the Bouguer anomalies result from a thermal plume mechanism [*LeMasurier and Rex*, 1989] such as that discussed by *White and McKenzie* [1989]. Marie Byrd Land (the Amundsen-Bellingshausen 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, and a stationary plate environment; (2) isotopic data which suggest an asthenospheric source; and (3) the fact that Marie Byrd Land 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].

Rifting in West Antarctica

We think it is likely that the arguments presented above for the Ross Embayment apply as well to the ice-covered areas of Marie Byrd Land and its exposed volcanic rocks and to the adjacent rift shoulder defined by the Transantarctic Mountains–Whitmore-Ellsworth Mountains escarpment. Unfortunately, the geophysical data are very sparse throughout this area.

There are two lines of reasoning which suggest that the Marie Byrd Land plume model may be extended to the late Cenozoic volcanic and tectonic activity of the entire rift system. First, *White and McKenzie* [1989] have shown that it is geophysically and geologically reasonable, in several rift localities in the world, to visualize a plume head having a diameter on the order of 2000 km. Second, *Hole and LeMasurier* [1990] have shown that there is a strong geochemical coherence to all the rift-related basalts in West Antarctica, from western Ellsworth Land through Marie Byrd Land to the western Ross Embayment. These basalts are all indistinguishable from plume-related oceanic island basalts but are clearly distinguishable from Antarctic Peninsula postsubduction basalts that are not related to the West Antarctic rift system.

The early-rift grabens of the Ross Sea (Figures 7 and 17) are separated by broad, eroded basement ridges; the area has probable crustal thicknesses of 17–21 km. One ridge (Central high) continues north and west into the south Pacific Ocean as Iselin Bank. Early-rift grabens of the Eastern and Central basins can be traced from near the continental shelf edge (~1000-m contour, illustrated in Figure 7) to the edge of the Ross Ice Shelf. These grabens can probably be correlated with similar structures beneath the Campbell Plateau [e.g., *Bradshaw*, 1989]. The Victoria Land basin, however, terminates at its north end against the Transantarctic Mountains (Figure 7).

The Polar 3 anomaly may indicate a transfer fault if a basin similar to the Victoria Land basin were present with a north trend extending north at the northeast end of the Polar 3 anomaly. The negative magnetic anomaly (Plate 2) suggests that there could also be a continuation of the Victoria Land basin in northern Victoria Land (for example, Rennick graben) [see *Roland and Tessensohn*, 1987]. Early-rift grabens beneath the Ross Sea continental shelf probably continue south beneath the Ross Ice Shelf and throughout the ice-covered area of the West Antarctic rift system [*LeMasurier and Rex*, 1983; *Jankowski et al.*, 1983; *Cooper et al.*, 1991]. *LeMasurier and Rex* [1983] pointed out that the ice in Marie Byrd Land–Byrd Subglacial Basin is in fact much of the "sedimentary rock" filling the basins there. The subice topography contours in Figure 2, shown in sufficient detail by *Drewry* [1983], illustrate that the seafloor contours are much smoother beneath the Ross Sea shelf and Ross Ice Shelf (probably because of the sediment filling of basins) than the subice topography of the Byrd Subglacial Basin underlying the moving ground ice there.

The 5 km of relief across the Transantarctic Mountains front in the northern Victoria Land–southern Victoria Land area continues from 4 to 7 km relief along the entire rift shoulder (Figures 2 and 3). This relief 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 the scarp marking the interpreted rift shoulder along the Horlick-Whitmore-Ellsworth Mountains trend. Whatever the mechanism, we interpret the high uplift of the rift shoulder and the Amundsen-Bellingshausen flank as resulting from lateral thermal heating from the upwarped shallow asthenosphere beneath the Byrd Subglacial Basin and Ross Embayment.

In our discussion above of gravity and the sparse seismic refraction data in the West Antarctic rift system, we inferred that shallow- to lower-crustal intrusions are the sources of steep gravity gradients and high seismic velocities. If underplating and lower-crustal intrusions are present, they may account for a few deep reflections from beneath McMurdo Sound and the Ross Sea shelf [*McGinnis et al.*, 1985; *Kim et al.*, 1986; *Cooper et al.*, 1987*b*; *Trehu et al.*, 1989] as has been interpreted for rift systems elsewhere than Antarctica [e.g., *Behrendt et al.*, 1988, 1990]. The measured velocities interpreted as evidence for shallow- and lower-crustal intrusions (and/or underplating) range from about 6.7 to 7.4 km/s, as discussed in the seismic section. *Berg et al.* [1989] reported gabbro norite cumulates from xenoliths collected from volcanic rocks in the McMurdo area, and *Wörner and Viereck* [1989] reported similar rocks with xenoliths near Mount Melbourne. Both finds support the presence of mafic intrusions. The Polar 3 and other >1000-nT magnetic anomalies (Figure 2) are evidence for mafic (or ultramafic?) magmatic rocks. The only measurement of *Pn* (Moho) is 8.2 km/s in the McMurdo Sound area [*McGinnis et al.*, 1985], a velocity which seems high for an area of active rifting.

The late Cenozoic alkaline volcanic rocks exposed at sparse locations (Figure 2) along the West Antarctic rift system and in the CIROS core hole provide the only age control for Cenozoic magmatism (early Oligocene or earlier to present) associated with rifting. The magmas probably resulted from decompression melting of upwarped asthenosphere beneath the attenuated rifted crust [*Futa and LeMasurier*, 1983; *LeMasurier and Rex*, 1989; *Schmidt-Thomé et al.*, 1990; *Mueller et al.*, 1990]. On the basis of magnetic data, we interpret these volcanic rocks to occur over a much broader area (Figures 2, 4, and 5 and Plate 1) of the ice-covered West Antarctic rift system than their areas of outcrop. We interpret the most extensive area of rift-related volcanism beneath the ice-covered area of Marie Byrd Land to be that marked by the abundance of short-wavelength anomalies (shallow source) discussed in the magnetics section (Figure 5).

We realize that extending the tectonic interpretation for Cenozoic rifting and crustal thinning developed for the Ross Sea–northern Victoria Land area throughout the West Antarctic rift system is speculative. 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 this model.

Progression of Rifting From Jurassic to Present

LeMasurier [1990*a*, *b*] noted above that the age of the oldest Cenozoic volcanism is about early Oligocene (or earlier) in the Ross Embayment and progressively decreases toward 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 noted 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 like the peeling of an orange, extending 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 Late Cretaceous time [*Smith and Drewry*, 1984].

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. Much of the rift process, including all the dated alkaline volcanism, took place after the separation of Antarctica from Australia and New Zealand. As it is a process restricted to the Antarctic plate with no apparent link to the surrounding plate tectonic features, the cause for the rifting will have to be assumed to lie in mantle processes under the stationary Antarctic plate.

Additional Studies

Specifically, some of the additional studies necessary in the West Antarctic rift area are as follows:

1. Aeromagnetic surveys similar to the survey shown in Plate 1 combined with radar ice-sounding data over grounded ice, along transect zones crossing the rift.
2. Gravity traverses (airborne or surface) with ice thickness determinations over ice-covered areas and on outcrops across the steep Bouguer anomaly gradient (as shown in Figure 13) over the rift shoulder adjacent to the Horlick-Whitmore-Ellsworth Mountains area to determine if the 4–7 mGal/km gradient present along parts of the Ross Embayment is continuous to this area. This is important to help resolve the ambiguity whether the source of the steep gradient is associated with early Jurassic or Cenozoic rifting.
3. Large offset seismic surveys over the rift shoulder and extended crust in the thick ice area of Marie Byrd Land (Byrd Subglacial Basin and Ross Embayment) to measure crustal thickness and determine the areal extent of the ~7-km/s refractors.
4. Deep crustal (20-s recording time) multichannel (vertical incidence) reflection profiles in the Ross Embayment–Marie Byrd Land area. A deep profile could be done economically using marine techniques over the Ross Sea shelf but would be much more expensive in the ice-covered areas.
5. Heat flow measurements beneath the continental shelf and the floating and grounded ice sheet. If this is not possible, electrical measurements could provide indirect thermal structure determination.
6. In addition to the geophysical surveys, drill holes are needed in the basin and platform areas beneath the Ross Sea shelf (Figure 7), particularly to put constraints on the ages of the interpreted unconformities (Figure 17). This information is needed to better define the timing of the West Antarctic rift system, including the early rift stages (which are probably as old as late Mesozoic). If ever feasible, drilling is needed through the ice filling the Byrd Subglacial Basin and into the rock (including rift basins) beneath with shallow or preferably deep penetration.

SUMMARY

In this paper, we have attempted to review all of the geophysical results and better define the little known West Antarctic rift system, one of the great rift systems of the world. The major characteristics are as follows:

1. The West Antarctic rift system covers an area about 3000 by 750 km through West Antarctica. It is comparable in size to the Basin and Range in North America or the East African rift system, as pointed out by *LeMasurier* [1978, 1990*a*] and *Tessensohn and Wörner* [1991].
2. Bimodal alkaline volcanic rocks typical of other rifts, erupted from the Oligocene (or earlier) to the present, crop out throughout the West Antarctic rift system, and are probably widely present beneath water- and ice-covered areas. Late Cenozoic magmatic rocks are exposed (1) along the front of the Transantarctic Mountains from northern Victoria Land, being geographically coincident with Jurassic tholeiites (Ferrar dolerites and Kirkpatrick basalts), to about the Horlick Mountains but do not exist any further along the Transantarctic Mountains and (2) throughout the mountains of Marie Byrd Land to the southern Antarctic Peninsula. Magnetic data suggest the presence of Cenozoic volcanic rocks beneath the western part of the Byrd Subglacial Basin and indicate their absence beneath the eastern part of the basin (closest to the Ellsworth-Whitmore Mountains).

Magnetic data also suggest Cenozoic volcanic rocks are present on the Ross Sea shelf and possibly beneath the Ross Ice Shelf.

3. The rift system is marked by a spectacular physiographic scarp coincident with the Transantarctic Mountains in Victoria Land (about 5 km relief) through to the Ellsworth-Whitmore-Horlick Mountains where relief is as great as 7 km. The Amundsen-Bellingshausen flank of the rift system is lower and more internally fragmented by horst and graben structures, while the axial part or "rift valley" comprising early rift grabens is buried beneath the Ross Sea continental shelf, the Ross Ice Shelf, and the West Antarctic ice sheet. Rifting probably started in the late Mesozoic and has continued through the Cenozoic. However, the oldest known associated magmatic rocks are only about Oligocene or earlier.

4. There is a distinct difference between the elevations of mountains (rift shoulder) flanking the West Antarctic rift which were uplifted during late Cenozoic extension and elevations of those ranges in the Transantarctic Mountains beyond the Horlick Mountains toward the Weddell Sea not affected by late Cenozoic extension. The maximum elevations along the rift shoulder are about 4 km in Victoria Land, 3 km in the Horlick-Whitmore Mountains, and 5 km in the Ellsworth Mountains. In contrast, the maximum elevation in the Thiel and Pensacola Mountains is 2 km decreasing toward the Weddell Sea to $1\frac{1}{2}$ and 1 km in the Shackleton and Theron ranges, respectively. Uplift and faulting in this low area of the Transantarctic Mountains probably is substantially older than that which occurred along the highest parts of the Cenozoic West Antarctic rift shoulder.

5. Although not definitive, a growing body of physiographic, geophysical, glacial geomorphological, and paleontological evidence from the Transantarctic Mountains along the Ross Embayment leads us to the conclusion that the rift shoulder has been rising since about 60 Ma at episodic uplift rates of the order of 1 km/m.y. most recently since the mid-Pliocene rather than continuously at the mean rate of 100 m/m.y. This uplift rate could account for the present 4–5 km highest elevation since 2–5 Ma, but evidence also indicates earlier episodes of uplift and erosion. We suggest a possible climate forcing effect on the advance of the Antarctic ice sheet.

6. Although high-amplitude (>100 nT) short-wavelength magnetic anomalies are caused by both the late Cenozoic alkaline volcanic rocks and the Jurassic volcanic rocks (but not the Ferrar dolerite sills), magnetic data are useful in places in approximately delineating the extent of Cenozoic magmatism beneath the ice. Over the Ross Sea continental shelf, a detailed aeromagnetic survey reveals the presence of about 100 inferred shallow-source submarine volcanic edifices (or subvolcanic intrusions) along linear trends interpreted as rift fabric. Conspicuous shallow-source magnetic anomalies on widely spaced profiles over the ice-covered areas of Marie Byrd Land (correlated with late Cenozoic volcanoes in outcrop areas) suggest the presence of Cenozoic volcanism throughout a large area there. The absence of such shallow-source anomalies east of a prominent north trending break in the high-amplitude anomaly pattern over the Byrd Subglacial Basin indicates the absence of Cenozoic volcanism to the east of this line (i.e., toward the Ellsworth Mountains). The smooth magnetic field is probably caused by the thick nonmagnetic (lower Paleozoic) metasedimentary rock cropping out in the Ellsworth Mountains.

Several linear (several hundred kilometers long) 1000+ nT anomalies (e.g., the Polar 3 anomaly at the front of the Transantarctic Mountains in Victoria Land) exist throughout the West Antarctic rift. We suggest this is evidence for possible oblique slip within this extensional system. We speculate that the "Sinuous Ridge anomaly" in the deepest part of the Byrd Subglacial Basin represents a change from alkaline to tholeiitic magmatism and the start of formation of an oceanic spreading center.

7. One of the long-recognized great gravity anomalies of the world marks the transition from West to East Antarctica across the rift shoulder in the Ross Embayment part of the West Antarctic rift system. Compared with East Antarctica, Bouguer anomalies increase by as much as 200 mGal over the attenuated rifted crust underlying the Ross Sea shelf, Ross Ice Shelf, and ice-covered part of Marie Byrd Land (Byrd Subglacial Basin). Although data are sparse, it is apparent that this anomaly continues along the Ellsworth-Whitmore part of the rift shoulder. In contrast, the difference in Bouguer gravity across the West Antarctica–East Antarctica boundary is lower (only about 130 mGal maximum range) in the Pensacola Mountains–Filchner Ice Shelf area outside the Cenozoic rift area.

Although primarily the result of crustal thinning through rifting beneath the Ross Embayment, the steep gradients (4–7 mGal/km) measured at a few places (Beardmore Glacier, McMurdo, and northern Victoria Land areas) require high-density rock within the rifted crust, which we suggest is evidence of shallow- or lower-crustal intrusion of mafic or ultramafic rock associated with rifting. In contrast, the gravity gradient across the front of the Transantarctic Mountains in the Pensacola Mountains is only about 2 mGal/km, which is easily explained by a 15°–20° dip on the Moho. The thinned crust on the West Antarctic side is probably due to rifting, but whether this occurred in the Jurassic (post–Dufek intrusion) or later is unknown.

8. At several places close to the Horlick Mountains and the Whitmore Mountains part of the rift shoulder, and in the Ross Embayment–Ross Sea area, sparse seismic refraction data have shown evidence of 6.8–7.3 km/s velocity layers at relatively shallow depth within the crust. We consider this supportive of the gravity interpretation of high-density mafic (or ultramafic?) intrusions within the rifted crust. We cannot say whether these inferred intrusions resulted from Cenozoic or Mesozoic magmatic activity, but

they are likely to have been emplaced during major stages of extension, crustal thinning, and graben formation. The presence of similar high (6.9–7.4 km/s) velocities in the lower crust beneath the extended Ross Sea shelf suggests mafic (or ultramafic?) underplating there and likely elsewhere beneath the rift in ice-covered areas. In the 14-km-deep Victoria Land basin, high seismic velocities at shallow depths are interpreted as evidence of layered volcanic rather than sedimentary rocks, consistent with evidence of magmatic activity associated with rifting.

9. Reflections from Moho at several locations over the Ross Sea shelf and one at the northern Victoria Land mountain front show Moho depths from 17 to 22 km, consistent with our interpretation of crustal thinning caused by rifting. An uppermost mantle velocity of 8.2 km/s in McMurdo Sound is the only determination in the West Antarctic rift system area. The generally constant level in the Bouguer gravity map from the Ross Sea shelf through the Ross Embayment–Byrd Subglacial Basin to the Bellingshausen Sea indicates that the ~20-km-thick rift stage crust extends through this area as well and is consistent with the presence of early-rift grabens (see characteristic 3 above).

10. A combination of active volcanism, rapid topographic uplift at the rift shoulder, xenolith thermobarometry, interpretation of possible upwarp of Curie isotherm beneath the Victoria Land basin, a few heat flow measurements, and a high geothermal gradient (40°C/km, *White,* 1989) are all indicative of high heat flow in the Ross Embayment, characteristic of other young rift systems. The evidence of high temperatures at shallow depth beneath the Ross Sea continental shelf and adjacent Transantarctic Mountains is supportive of the model of thermal uplift associated with lateral heat conduction from the rift [*Stern and ten Brink,* 1989]. Possibly in combination with hot spot–thermal plume models this model can also explain the volcanism, rifting, and high elevation of the rift shoulder (Ellsworth Mountains) in greater Marie Byrd Land (including the Byrd Subglacial Basin).

We propose that the active West Antarctic rift system is a continuation of rifting that began in the Jurassic at the Weddell Sea end of the Transantarctic Mountains with separation of Antarctica from Africa and has continued clockwise around the present continental margin (with a spreading center jump) to its present location in the Ross Embayment and the Byrd Subglacial Basin. The rift system is active now and may have been so since the late Mesozoic.

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. If, as an extreme case, a 40-km-thick crust thinned to 20 km across the present 750-km-wide rift during late Mesozoic-Cenozoic time, a maximum of about 350-400 km total extension is implied. Significant underplating would require still greater extension. The major part of the rift process, including all the dated alkaline volcanism, took place after the separation of Antarctica from Australia and New Zealand. As it is a process restricted to the Antarctic plate with no apparent link to the surrounding plate tectonic features, the cause for the rifting must lie in mantle processes under the stationary Antarctic plate.

Acknowledgments. We thank Charles Bentley for assistance in obtaining some of the early gravity data (from the 1950s and 1960s). Baerbel Lucchitta of the U.S. Geological Survey (USGS), Flagstaff, Arizona, provided the aeromagnetic-enhanced Landsat images used. Discussions with our colleagues—particularly Peter Barker, Peter Barrett, Charles Bentley, David Harwood, Karl Kellogg, Philip Kyle, Lisa Morgan, Dwight Schmidt, Hans Schouten, Brian Storey, Gerhard Wörner, and Mary Lou Zoback—were helpful. David Harwood, P. Mueller, M. Schmidt-Thomé, and Cynthia Ebinger furnished unpublished manuscripts. Karl Kellogg, John Grow, Larry Lawver, Kevin Furlong, and others reviewed the manuscript of this paper. The National Science Foundation (NSF) and the Alfred Wegner Institute in Germany provided logistic support for the Bundesanstalt für Geowissenschaften und Rohstoffe–USGS Antarctic surveys. W.E.L. gratefully acknowledges the support of NSF through grant DPP 8020836 and the hospitality of the British Antarctic Survey during the preparation of this paper.

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(Received August 22, 1990;
accepted April 8, 1991)