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Cenozoic transtension along the Transantarctic Mountains-West Antarctic rift boundary, southern Victoria Land, Antarctica

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Abstract. Brittle fault arrays mapped along the structural boundary between the Transantarctic Mountains and the West Antarctic rift system are oriented obliquely to the axis of the mountains and offshore rift basins. The north to northwest trending regional rift boundary is thus not controlled by continuous rift border faults. Instead, the rift margin trend must be imposed by inherited lithospheric weaknesses along the ancestral East Antarctic craton margin. Fault kinematic solutions indicate that a dextral transtensional regime characterized the rift boundary in the Cenozoic and that dominantly transcurrent motion occurred during the most recent faulting episode. The Transantarctic Mountains are considered to be a rift-flank uplift, yet no substantial isostatic uplift is expected in a transtensional setting, and the mechanism of large-magnitude Cenozoic uplift of the mountains remains problematical. Regional deformation patterns in Victoria Land and the Ross Sea can be explained by a transtensional model and are not compatible with large-magnitude crustal stretching within the West Antarctic rift system in the Cenozoic. The crustal thinning across the rift system more likely took place in the Mesozoic, when major West Antarctic crustal block motions occurred. The Cenozoic intracontinental deformation can be related to plate interaction resulting from the global Eocene plate reorganization, prior to the final separation between Antarctica and a narrow salient of the southeastern Australian margin. Displacement magnitude was probably minor, and thus early Tertiary east-west Antarctic motion is unlikely to account for discrepancies in global plate motion circuits.

Introduction

The Transantarctic Mountains (TM) form a lithospheric boundary that transects the Antarctic continent and separates cratonic East Antarctica from the mosaic of crustal blocks that compose West Antarctica (Figure 1) [*Dalziel and Elliot*, 1982]. The TM form the rim of the submarine Ross embayment, which is underlain by the Mesozoic-Cenozoic West Antarctic rift system. This association has led to the widely accepted model that the TM form an uplifted rift shoulder, possibly analogous to the Colorado Plateau-Basin and Range boundary or the flanks of the East African rift system [*Fitzgerald et al.*, 1986; *Tessensohn and Wörner*, 1991]. In comparison to other rift flanks, however, the TM have exceptional relief, with elevations typically above 2000 m and peaks reaching 4500 m and a more simple tilt block structure that is laterally continuous for over 2500 km.

The TM appear to have marked a fundamental structural boundary throughout the progressive fragmentation of the Gondwana supercontinent. Magmatism and extension were focused along this boundary during the initial, Jurassic breakup phase [*Elliot*, 1992; *Wilson*, 1993]. Relative motions between West Antarctic microplates and East Antarctica during subsequent, Jurassic, and Cretaceous progressive fragmentation of Gondwana was accommodated, at least in part, along portions of this intraplate boundary [*Schmidt and Rowley*, 1986; *Grunow et al.*, 1991; *Lawver and Gahagan*, 1991; *DiVenere et al.*, 1994]. This paper presents structural kinematic results that allow more detailed resolution of the Cenozoic motion history along the East-West Antarctica boundary and discusses the implications for intracontinental block displacements and Tertiary plate motion circuit models.

Uplift of the TM has been linked with mechanical and thermal processes associated with the West Antarctic rift system. Proposed models include development as the upper plate margin of an asymmetric extensional orogen [*Fitzgerald et al.*, 1986] and as a flexurally uplifted footwall of a half-graben basin [*Stern and ten Brink*, 1989]. The boundary constraints of these uplift models imply specific geometries and kinematics for the structural margin between the TM and adjacent Ross embayment rift basins. This study investigated the architecture and structural evolution of the boundary between the TM and the rift system and provides new information of use in evaluating TM uplift models and the relation of uplift to rift development in time and space.

Most rift studies have focused on rift basins and basin-margin structures. In Antarctica, however, only the TM rift margin is exposed above the covering ice sheet and thus is the only major structural boundary within the rift system that can be directly studied. I applied brittle structural analysis methods to this margin in a region in southern Victoria Land (Figure 1) where Cenozoic faulting was documented from fission track studies [*Fitzgerald*, 1987, 1992]. Whereas all preceding work had supposed that major faults were oriented parallel or orthogonal to the regional trend of the TM, I found an array of northeast faults, oblique to the north to northwest trending TM margin. This requires that the preexisting infrastructure, not range-parallel rift faults, controlled the regional trend of the TM chain. Fault kinematic solutions demonstrate Cenozoic dextral transtension oblique to the TM margin. This kinematic regime is discussed within the framework of regional structural patterns in the West Antarctic rift, postulated intracontinental

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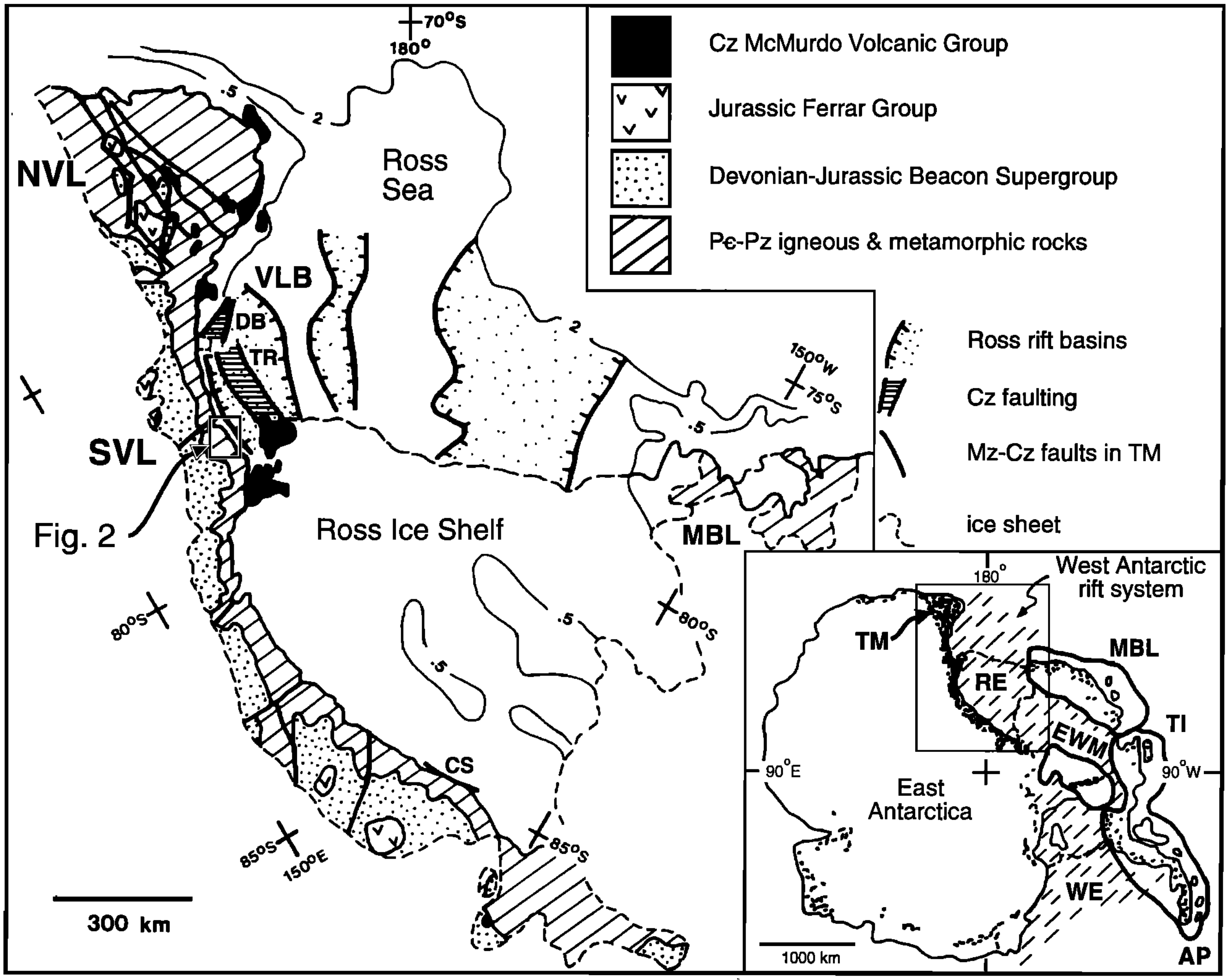

Figure 1. Tectonic setting of the Transantarctic Mountains. Inset shows the Transantarctic Mountains (TM) along the margin of cratonic East Antarctica; the crustal blocks that form West Antarctica (MBL, Marie Byrd Land; TI, Thurston Island; AP, Antarctic Peninsula; EWM, Ellsworth-Whitmore Mountains); the West Antarctic rift system (RE, Ross embayment; WE, Weddell embayment); and the location of the tectonic map. Note that in this sector, "West Antarctica" lies to the east of "East Antarctica". Map shows the main lithotectonic units of the TM and the principal structural elements of the TM and Ross Sea. NVL is northern Victoria Land; SVL, southern Victoria Land; VLB, Victoria Land Basin; DB, Drygalski basin; TR, Terror rift; MBL, Marie Byrd Land; and CS marks downfaulted Beacon strata in the central TM at Cape Surprise. Box marks the location of the study area in southern Victoria Land (Figure 2).

block motions within Antarctica, and plate boundary configurations along the Antarctic margins. Proposed Transantarctic Mountains uplift models are reexamined in light of transtensional rift kinematics.

TM Tectonic Setting

The present TM coincide with the edge of the East Antarctic craton. The East Antarctic margin had a protracted Neoproterozoic and Paleozoic history as a divergent and convergent plate boundary during development of the Ross orogen [*Borg and DePaolo*, 1991; *Dalziel*, 1992; *Stump*, 1992]. The Ross mountain belt was denuded to produce the regional Kukri erosion surface (Kukri Peneplain) and, beginning in the Devonian, elongate intracratonic or foreland basins parallel to the ancestral margin subsided and were filled by clastic rocks of the Devonian-Jurassic Beacon Supergroup [*Barrett*, 1991; *Collinson*, 1991].

Multiple rift phases occurred within Antarctica during Gondwana breakup and appear to have repeatedly reactivated the ancestral East Antarctic margin. The change from a compressional to an extensional setting is marked by syndepositional tectonism associated with Early-Middle Jurassic volcanogenic deposits that form the uppermost Beacon strata and the lower part of the Ferrar Group [*Elliot*, 1992]. Extrusion and intrusion of circa 175-180 Ma Kirkpatrick Basalt and Ferrar Dolerite occurred along the length of the East Antarctic margin [*Elliot*, 1992]. Crustal extension was coeval with emplacement of the Jurassic magmatic rocks in the TM [*Wilson*, 1993], and a subsiding volcano-tectonic rift zone developed along the present site of the TM and possibly extended over a portion of West Antarctica [*Schmidt and Rowley*, 1986; *Lawver and Gahagan*, 1991; *Elliot*, 1992; *Wilson*, 1993; *Grunow*, 1993]. The development of rift basins in the Ross embayment is considered to have involved two principal rift phases of presumed Cretaceous and Cenozoic age [*Cooper et al.*, 1987, 1991; *Tessensohn and Wörner*, 1991]. There is increasing evidence from fission track dating for an episodic TM uplift history, with initial Cretaceous uplift followed by the principal uplift phase beginning in the Eocene, ~55-60 Ma [*Stump and Fitzgerald*, 1992]. The Cenozoic phase in southern Victoria Land produced >5000 m of uplift [*Fitzgerald*, 1992]. Seismic reflection data show that Cenozoic faulting is localized within the westernmost Ross Sea adjacent to the TM in Victoria Land [*Cooper et al.*, 1987; *Della Vedova et al.*, 1992]. Although there is, as yet, no age constraint on the inception of this Cenozoic faulting, this spatial correlation has led to models linking TM uplift with West Antarctic rifting [*Fitzgerald et al.*, 1986; *Stern and ten Brink*, 1989]. New seismic reflection data from the central TM, in contrast, have revealed little evidence of Cenozoic faulting [ten Brink et al., 1993], in spite of the fact that the TM range there has elevations equivalent to Victoria Land.

TM Architecture

The generally undeformed character of the Paleozoic-Mesozoic Beacon and Ferrar sequences that form the peaks along the spine of the TM was recognized during the earliest exploration of Antarctica, and pioneering geologists proposed that the TM are a fault-bounded horst block upthrown relative to the Ross Sea depression and to bedrock underlying the polar plateau ice sheet [*David and Priestley*, 1914; *Gould*, 1935]. The gentle regional dip of Beacon strata toward the polar plateau margin of the range later suggested the alternative geometry of a tilt block bounded only on the Ross margin by a major normal fault zone [e.g., *McGregor*, 1965]. The TM range was subsequently described as a mosaic of tilted fault blocks with a horst and graben morphology [*Katz*, 1982]. Recent models are essentially unchanged, with the TM depicted as extensive linear to curvilinear, asymmetric fault blocks bounded by a continuous normal fault zone, called the Transantarctic Mountains Front (TMF), along the Ross embayment coastline [*Fitzgerald et al.*, 1986; *Stern and ten Brink*, 1989; *Tessensohn and Wörner*, 1991; *Le Masurier*, 1990; *Behrendt and Cooper*, 1991]. The TMF has been interpreted to accommodate the large differential relief between the mountains and the subsided crust of the adjacent Ross embayment; vertical displacements of the order of 15-20 km have been cited [*Tessensohn and Wörner*, 1991]. A steep Bouguer gravity anomaly along the TM coastline together with seismic evidence for thinned crust beneath the Ross embayment have been interpreted to reflect an abrupt change in crustal thickness, presumed to be related to displacement across the TMF [*Behrendt et al.*, 1991]. In southern Victoria Land the Victoria Land rift basin lies just offshore, and hence the TMF zone has been compared with the normal border fault systems that bound half-graben basins in the East African rift system [*Tessensohn and Wörner*, 1991].

The long-standing view of the TMF as a coastal fault zone with a north to northwest trend and large-scale, normal, down-to-the-east (i.e., down toward West Antarctica) displacement is based mainly on morphotectonic arguments. In southern Victoria Land, faults were presumed to control east facing physiographic scarps and the low-lying coastal piedmont glaciers [*Gunn and Warren*, 1962]. Mapping of TMF faults is hampered by coastal ice cover, the lack of post-Jurassic, pre-upper Cenozoic strata in the TM, and the restriction of the Devonian-Jurassic Beacon Supergroup strata to the higher, western portion of the range (Figure 1). Large-magnitude post-Jurassic faulting along the coast has thus only been demonstrated on outcrop at a single location in the central TM, where Beacon and Ferrar units have been downfaulted ~5 km [*Barrett*, 1965]. Major basin-bounding normal faults in the Ross Sea are thought to parallel the coastline [*Cooper et al.*, 1987], but, owing to the large distance between offshore seismic reflection profiles, fault trends can not be resolved in any detail.

Fitzgerald [1987, 1992] used fission track dating to examine the structure and uplift history of the coastal border of the TM in southern Victoria Land. Discrepancies between fission track ages at a given elevation along traverses approximately perpendicular to the mountain trend were ascribed to fault displacement between localities, with the age differential used to calculate the magnitude of offset. Faulting is inferred to postdate the inception of uplift at ~55 Ma. The fission track results indicate that TMF faulting extends ~20 km inland from the coastline and produced ~1200-2000 m of down-to-the-east displacement across that distance [*Fitzgerald*, 1992]. The fission track data alone do not constrain the exact position or orientation of the faults that accommodated this displacement, however.

Brittle Fault Study

Design, Location, and Methodology of Field Study

Southern Victoria Land (Figure 1) was selected for the first systematic regional study of brittle faults in the TM because of the presence of coastal outcrops within the TMF zone, where strains should be greatest, and the fission track constraints on the timing of faulting from the work of *Fitzgerald* [1987, 1992]. As in most of Antarctica, mesoscale structural analysis was mandated by the paucity of outcrops and because it is unlikely that major faults will be exposed owing to exploitation of the associated fractured zones by glaciers. A search for brittle mesoscale features was completed in bedrock outcrops along a 50-km longitudinal profile of the TMF along the McMurdo Sound coast, along a 25-km transect across the TMF along the south coast of Granite Harbour, and at selected localities an additional 20 km inland along New Glacier and Mackay Glacier (Figure 2). A transverse profile across the continuous exposure along the ridge extending eastward from Mount Doorly was also investigated, to relate fault patterns to fission track results [*Gleadow and Fitzgerald*, 1987; *Fitzgerald*, 1987, 1992]. All of the exposures along the transects consist of

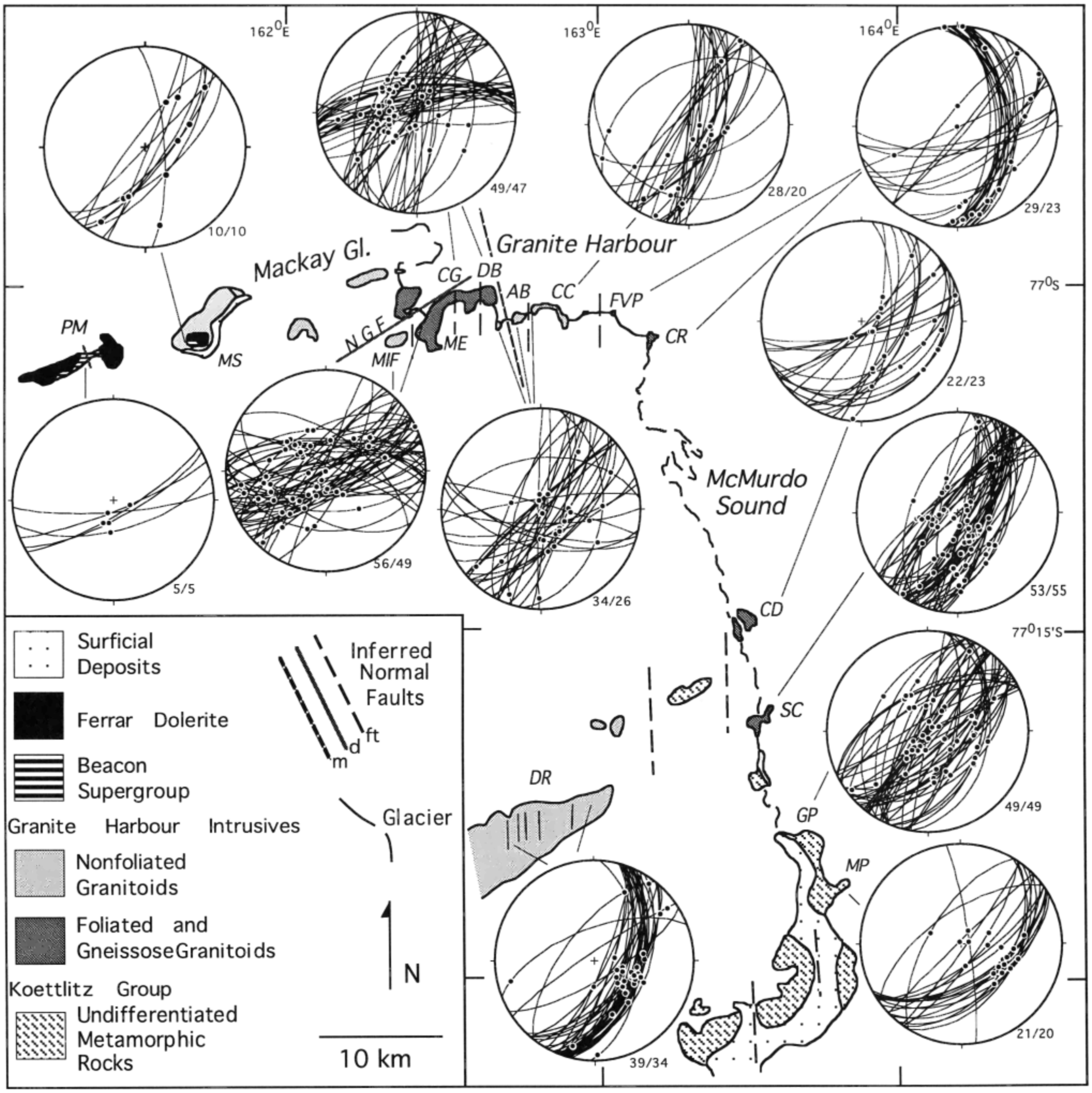

Figure 2. Geologic map of the study area in southern Victoria Land (location shown in Figure 1); faults as mapped by *Gunn and Warren* [1962] and *Fitzgerald* [1987, 1992] based on morphology (m), apparent offsets of Ferrar Dolerite (d), and offset fission track isochrons (ft). Fault planes and striae are shown on lower hemisphere, equal area projections (numbers denote number of faults/number of striae). PM is Pegtop Mountain; MS, Mount Suess; NGF, New Glacier Fault, inferred to underlie the New Glacier; MIF, Minnehaha Ice Falls; ME, Mount England; CG, Cape Geology; DB, Discovery Bluffs; AB, Avalanche Bay; CC, Couloir Cliffs; FVP, First View Point; CR, Cape Roberts; CD, Cape Dunlop; SC, Spike Cape; GP, Gneiss Point; MP, Marble Point; and DR, Doorly ridge.

Proterozoic-early Paleozoic igneous and metamorphic rocks with the exception of the Beacon Supergroup outcrop at Pegtop Mountain (Figure 2).

All striated surfaces discovered in the outcrops were measured. The outcrops, though limited in overall extent, have continuous exposure along a variety of directions, ensuring that fault surfaces of all orientations could be identified if present. Displacement sense and magnitude were recorded where markers were present; markers consisted of compositional layering in metasedimentary units, gneissic banding, pegmatitic dikes and veins, and mafic inclusions in granitoids. In the absence of markers, which was common in exposures of isotropic granitoids, displacement sense alone was determined from intersection relations of minor fractures with the fault surface [*Angelier et al.*, 1985; *Petit*, 1987], less commonly from accretionary fiber steps on fault surface veins, and, in rare cases, from compressive or tensile bridge structures along the fault trace [*Gamond*, 1987]. The quality of each displacement sense determination was rated as definite, for offset markers and some exceptionally developed fiber steps and reidel shear arrays; probable, for well-developed and consistent sense indicators; possible, for poorly defined or locally developed sense indicators; and inferred, where sense was based on comparison with nearby faults of similar attitude and slip direction. A systematic search for crosscutting relations between faults and superposition of striae sets on fault surfaces was carried out, but such features proved to be rare.

Fault Character

Outcrop-scale faults most typically are discrete, striated planes, commonly with a surface polish or a thin, mineralized veneer. Cataclastic zones from 2-20 cm thick, and rarely to 60-300 cm thick, are locally developed, particularly in homogeneous granitic lithologies. They consist of anastomosing microfault surfaces and local pockets or stringers of fine-grained breccia. Measured fault displacements range from 1.5 cm to 90 m, with most between 5 and 350 cm. The development of the cataclastic zones, particularly within major gullies, may indicate substantially larger displacements, but this could not be demonstrated owing to the lack of markers. Exposed fault traces ranged between 2 and >30 m; in some cases, cataclastic zones marked by linear, incised gullies could be traced for 200-750 m to marginal ice cover. Because all outcrops are circumscribed by ice, the trace lengths are minima.

Fault surface veins consist of either smooth cryptocrystalline fill or smooth to patchy fibrous coatings. The veins are composed of quartz, chlorite, and, less commonly, contain epidote. Patchy fibrous veins with subhorizontal fibers are particularly well developed at Cape Roberts. Both individual fault planes and cataclastic zones are marked by bright red alteration of the host rock. Alteration zones of similar appearance that occur around fractures in granitic basement rocks just to the south of the study area were investigated by *Craw and Findlay* [1984]. They found that the red color is due to hematite staining of feldspars and ascribed the alteration to either fluid flux generated by emplacement of Jurassic Ferrar Dolerite or to superposition of the Jurassic event on an older, Ordovician alteration event related to late stage granite emplacement. The fault surface veins noted in this study indicate that fluid flux was associated with Cenozoic faulting, and hence the alteration along fault surfaces may also have occurred at that time.

Fault Geometry, Distribution, and Age

Abrupt morphologic breaks, apparent offsets of Ferrar Dolerite sheets, and offsets in fission track isochrons suggested to earlier workers that major north-northwest and east-northeast normal faults were present in southern Victoria Land (Figure 2) [*Gunn and Warren*, 1962; *Fitzgerald*, 1987, 1992]. In this study, no faults of the coast-parallel, northwest orientation were found; instead, fault strikes range between north and east (Figures 2 and 3). Although there is a continuum of fault strikes within this range, three fault arrays can be distinguished based on geometric criteria and regional field relations (Figure 3). North-northeast faults (~020°) are differentiated here from a northeast fault set (~040°), although there is overlap between them. The distinction is important, however, because the 040° direction parallels a regional early Paleozoic dike swarm and a fault set presumed to be coeval [*Keiller*, 1991]. A third group of faults strikes east-northeast.

North-northeast faults (~020°, range 350-029°) are well developed in all coastal outcrops except Marble Point (Figure 2). At Granite Harbour the north-northeast faults are not developed to the west of Cape Geology-Discovery Bluffs outcrops, and they become dominant toward the east (Figure 2). This distribution coincides spatially with the limits of TMF faulting as recorded by fission track results [*Fitzgerald*, 1992]. The specific positions of the faults inferred by Fitzgerald are not further constrained by the present results, however, because observed variations in the intensity of fracture development could not be unambiguously related to fault zones, as opposed to dike zones or intrusive contacts. A single, uniform array of north-northeast faults and cataclastic fault zones is present at Doorly ridge (Figure 2); associated low-relief scarps are interpreted here as fault-line scarps, resulting from a greater resistance to erosion imparted by hydrothermal alteration of host rocks along the zones, rather than as neotectonic fault scarps as inferred by *Fitzgerald* [1992]. North-northeast fault planes have a uniform, moderate to steep easterly dip; striae are dominantly dip parallel with a slight oblique component (Figure 3), and displacements are dominantly normal-dextral. A subordinate low-rake striae population occurs on the moderately dipping fault planes (Figure 3); dextral faults of this type are the dominant fault set at Cape Roberts (Figure 2). Rare superposition of striae in coastal outcrops indicate that the shallow striae mark a younger, discrete slip event reactivating the older normal faults.

The northeast fault set (~040°, range 030-060°) is present at all localities, with the exception of Beacon exposures at Pegtop Mountain (Figure 2). Fault planes dip moderately southeast to moderately northwest, with subvertical faults abundant (Figure 3). Mutually crosscutting conjugate normal fault planes occur at Marble Point and Spike Cape. Striae on faults of this set have variable rakes from moderate northeast to moderate southwest (Figure 3); normal-dextral oblique slip on southeasterly dipping planes is dominant overall. Rare but consistent superposition of striae sets indicates that the subpopulation of low-rake striae postdate the moderate- to high-rake striae; the low-rake striae occur mainly on subvertical fault planes (Figure 3) indicating that only the steep planes were suitably oriented for reactivation. Paleozoic mafic dikes with trends between ~030 and 050°, parallel to the northeast fault set, are present at all basement rock localities in Granite Harbour [*Janosy*, 1991] and at Doorly ridge. The dominant morphologic grain in Granite Harbour outcrops, expressed as relatively deeply eroded, steep-sided, linear gullies, is controlled by northeast dikes,

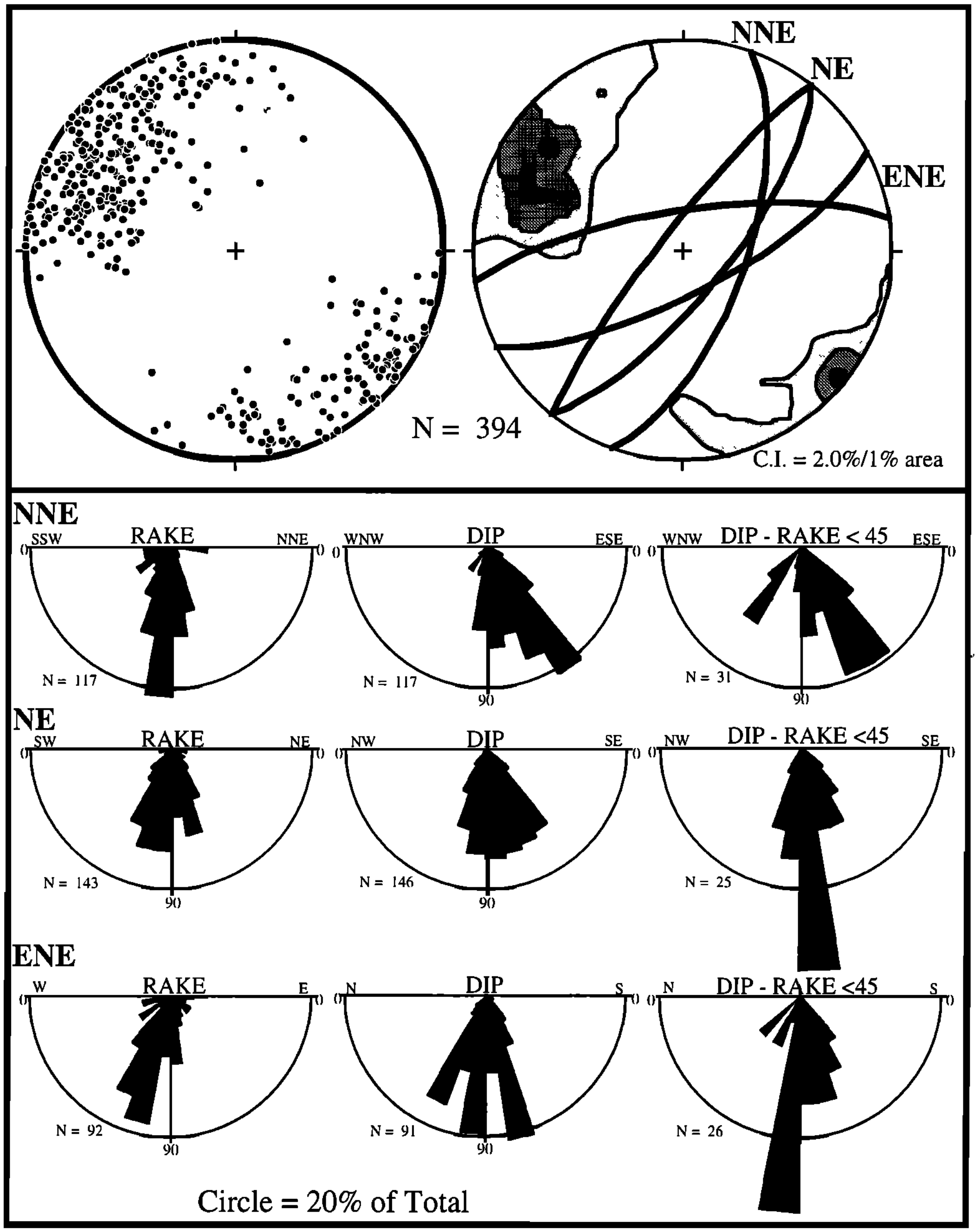

Figure 3. Geometry of faults in the study area. (top) Fault poles from the entire area on lower hemisphere, equal-area projections; the scatter plot shows that fault strikes range between north and east; the contour plot shows the three fault arrays differentiated for the region. (bottom) Histogram plots of striae rake, fault plane dip, and the dip of fault planes with low-rake striae for the north-northeast, northeast, and east-northeast fault arrays.

fractures, and faults. The Spike Cape peninsula (~030°), the northern coast of Gneiss Point, and the Marble Point peninsula (~048°) are also controlled by faults of this set (Figure 2).

The east-northeast fault array (061-110°) is restricted to Granite Harbour outcrops, with the exception of Cape Dunlop. East-northeast faults are particularly well developed between Minnehaha Ice Falls and Discovery Bluffs, supporting the suggestion that a major transverse normal fault occurs along the New Glacier, but the dominant fault orientation is more easterly than the ~055° trend inferred for the "New Glacier fault" (Figure 2) [*Gunn and Warren*, 1962; *Fitzgerald*, 1987, 1992]. At Mount England there are two distinct conjugate normal fault sets, locally forming small graben, with trends of ~063° and ~080°. At Pegtop Mountain there is a single, east-northeast fault zone (~070°) with ~90 m of down-to-the-south offset of Beacon and Ferrar units; the northwest fault inferred by *Fitzgerald* [1992] actually reflects an intrusive ramp in Ferrar Dolerite. Faults dip moderately north and south (Figure 3) and striae have dominantly moderate to steep westerly rakes (Figure 3). A subgroup of low-rake striae occur on subvertical fault planes (Figure 3). A single instance of shallow striae superposed on steep striae on an east-northeast fault was observed at Cape Dunlop.

Ages for the fault sets are only constrained by distribution in relation to offset fission track isochrons and geometric arguments. Along approximately east-west profiles along the south side of Granite Harbour and between the western edge of Doorly ridge and Spike Cape, down-to-the-east, vertical displacement of 1100-1200 m is indicated by offset Cenozoic fission track ages, with an additional 700-800 m displacement required if a 2-3° westward tilt of fault blocks is assumed [*Fitzgerald*, 1992]. Approximately half of this amount must be accommodated between Mount England and Avalanche Bay and in the equivalent transect along Doorly ridge; additional displacement of ~1000 m was determined for a covered fault just inland from Cape Dunlop and Spike Cape [*Fitzgerald*, 1992]. Given the magnitudes of these offsets, it is unlikely that brittle structures related to these displacements would be absent on outcrop. My fault data therefore indicate that Cenozoic displacement along the north-northeast and northeast fault arrays must have produced the observed offsets in the fission track ages. Normal slip on these fault arrays must thus postdate the onset of uplift at ~55-60 Ma [*Fitzgerald*, 1992], and the younger strike-slip motion documented by superposed striae must be a more recent Cenozoic event.

The distribution of the north-northeast fault set defined in this study corresponds closely with the zone of TMF faulting as defined by *Fitzgerald* [1987, 1992] based on offsets in fission track ages. At Doorly ridge, where faulting is indicated by both offset Ferrar dolerite and fission track ages [*Fitzgerald*, 1992], a single, homogeneous north-northeast fault array is developed. The north-northeast orientation of this fault set is oblique to both northwest trending ductile deformation fabrics in the basement rocks and to the ~040° trend of the regionally developed Paleozoic dike swarm, hence these faults are unlikely to represent reactivated, preexisting fracture planes. The homogeneous regional geometry of moderate eastward dip and approximately downdip striae is compatible with normal slip on newly formed fault planes. These factors suggest that the north-northeast fault set formed during Cenozoic uplift of the TM.

The timing of development and slip on the northeast fault set, of closely similar orientation, is more ambiguous. At Granite Harbour outcrops there is a clear association between mafic dikes and northeast faults, and oblique slip has locally occurred along reactivated dike margins. There is a regional parallelism between the fault set and the Paleozoic dike swarm as a whole, although there are no dikes exposed in coastal outcrops, where faults of the northeast set are abundant. Both the dominance of steeply dipping fault planes and the variability in striae rake are indicative of reactivation of preexisting surfaces. Taken together, the evidence suggests that northeast fracture planes formed in the Paleozoic were subsequently reactivated. Much of the motion is likely to be Cenozoic because these planes are close to the north-northeast trend of probable Cenozoic faults and thus were favorably oriented for reactivation.

East-northeast faults have been assigned a post-Jurassic age based on apparent offsets of Ferrar Dolerite across them [*Gunn and Warren*, 1962]; no Cenozoic offsets across the inferred faults can be demonstrated by fission track data, however [*Fitzgerald*, 1992]. An east-northeast fault set, developed synchronously with Jurassic Ferrar Dolerite emplacement, is developed within Beacon exposures on the polar plateau margin of the TM [*Wilson*, 1993]. The same Jurassic, syn-Ferrar relation is likely at Pegtop Mountain, but younger faulting can not be ruled out. The parallelism between the east-northeast faults in bedrock exposures and the Jurassic fault set developed in Beacon strata suggests they may be coeval. Striae rakes on the basement faults are predominantly oblique, whereas those on faults in Beacon units are approximately downdip. This may indicate that the striae on the basement faults are recording a separate, younger slip event. The most likely scenario is development of the east-northeast fault planes during Jurassic rifting, followed by reactivation during the Cenozoic morphologic uplift of the TM.

Kinematic Analysis

Kinematic contraction and extension axes were reconstructed from the fault slip data using the assumption that the principal axes are oriented at 45° to the fault normal and the slip vector within the movement plane [*Marrett and Allmendinger*, 1990]. About 80% of the measured faults were used in the kinematic analysis. At many localities a substantial number of faults with inferred displacement are included. In such cases, inferred displacements are based on definite determinations from offset markers applied within a group of faults having homogenous attitudes and slip directions (e.g., Doorly ridge, Figures 2 and 4), where it is unlikely that diametrically opposite displacements occurred. Average kinematic axes derived from subgroups divided according to the quality of slip sense determination are not significantly different for any locality. Inclusion of faults rated as possible or inferred results in larger data sets which better represent the overall fault populations within each subarea.

Results are presented as contour-plus-scatter plots of kinematic axes from individual localities. Such plots display the degree to which the axes cluster or disperse and hence the level of coherency of the average kinematic solutions. The average kinematic axes were determined using Bingham distribution statistics with a uniform weighting of the fault slip data based on the assumption that the fault kinematics are scale invariant [*Marrett and Allmendinger*, 1990]. Several factors precluded deriving average kinematic axes weighted according to the magnitude of fault displacement. Measured displacements are geographically biased based on exposed rock type and resultant presence/absence of markers. Since most faults are discrete surfaces and fault trace lengths are biased owing to ice cover, displacement estimates based on gouge thickness or fault size are not meaningful. Thus it is assumed here that the measured faults are representative of the entire fault population within the area and that the

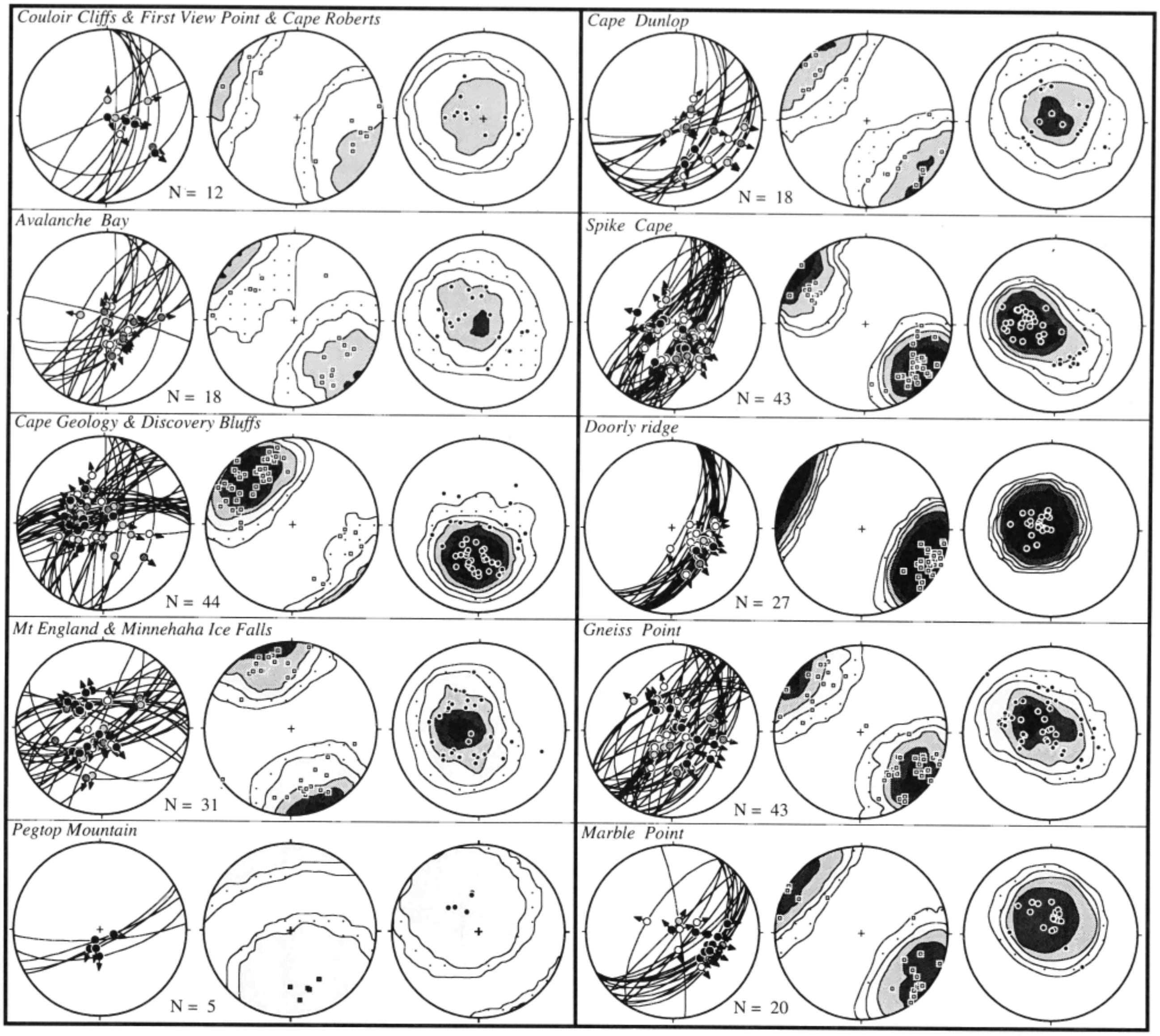

Figure 4. Fault kinematic data for high-rake fault population (striae rake ≥ 45°). Lower hemisphere, equal-area projections depict fault slip data, contoured extension axes (open squares), and contoured contraction axes (solid circles) for each locality. Great circles mark fault planes, small dots mark striae, and arrows indicate relative motion of hanging wall block. Type of circle fill indicates quality rating of displacement sense determination; solid is definite; dense stipple, probable; light stipple, possible; and white, inferred. Contouring is by the Kamb method with contours at 2, 4, 6, 8, 10, and 12 sigma.

average kinematic axes mark the principal contraction and extension directions within each crustal domain.

On the basis of the sparse but regionally consistent superposition of low-rake on high-rake striae on each of the regional fault arrays, the fault population is divided into two kinematic subgroups. Kinematic solutions for the large population having rakes ≥45° are presented in Figure 4 and for the much smaller population with rakes <45° in Figure 5. Because of the small numbers in the low-rake subgroup, populations are grouped from all localities in Granite Harbour and all localities along the McMurdo Sound coast, with Cape Roberts-First View Point, where low-angle slip is dominant, treated as an individual population. Solutions from low-rake fault populations at individual localities are closely similar to the results from the regional groupings in all cases.

The kinematic solutions for the high-rake fault populations (Figure 4) exhibit coherent clusters of kinematic axes for individual locations and are regionally homogeneous along both the longitudinal McMurdo Sound coast profile and the transverse Granite Harbour profile. Extension axes are the most tightly grouped, with average northwest-southeast trends and subhorizontal plunges. The majority of trends vary by only 10°, and all are within a 30° range, with plunges <15° except at two localities. The contraction axes are all close to vertical, with plunges >70°, with the

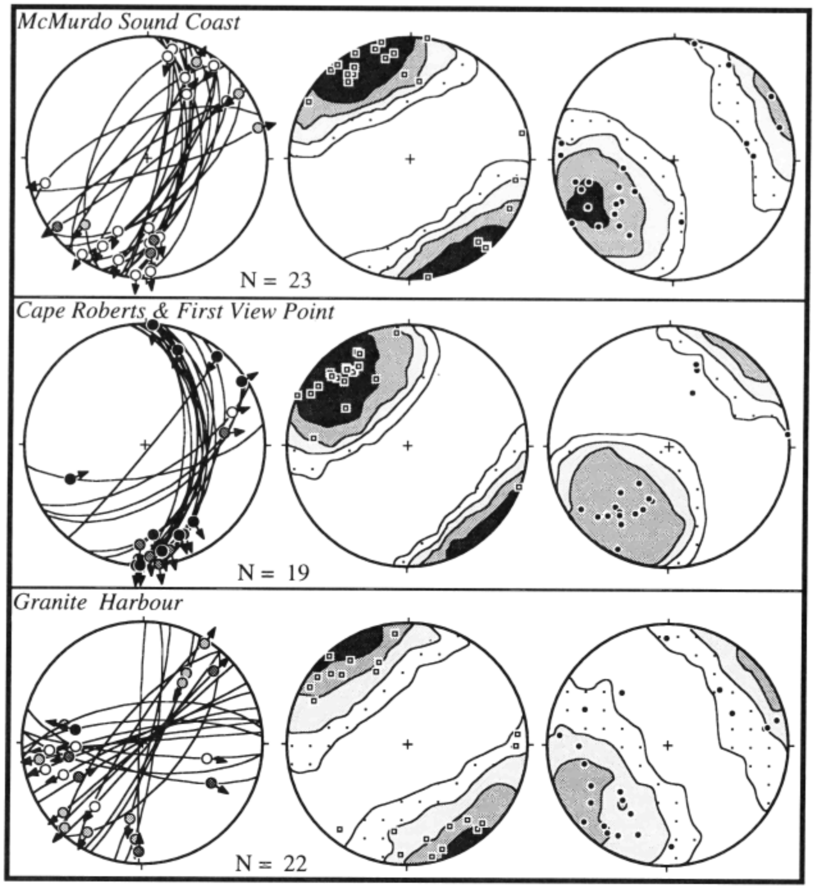

Figure 5. Fault kinematic data for low-rake fault population (striae rake <45°). Explanation is the same as Figure 4.

exception of two localities with 60° plunges. In general, the contraction axes are more dispersed, with an elongation toward a girdle distribution at some localities, showing the influence of the oblique-slip component on the fault arrays.

The low-rake fault populations also yield coherent and regionally uniform kinematic solutions (Figure 5). The extension directions are subhorizontal and trend northwest-southeast. The average extension directions for the McMurdo Sound coast and Granite Harbour low-rake groups are oriented 15-20° clockwise relative to the high-rake population directions. The Cape Roberts solution is closely similar to the high-rake results. The contraction axes for the low-rake populations have low plunges and are oriented southwest-northeast. Overall, the change between the high-rake and low-rake subpopulations represents a permutation in the average contraction direction from subvertical to subhorizontal, with the average extension direction undergoing a small clockwise rotation.

A particular concern in evaluating the kinematic results is whether the fault sample is truly representative. Is it possible that large-magnitude displacement has occurred on northwest trending, range-parallel faults now covered by ice? The consistency of the fault pattern and the total absence of northwest trending faults throughout the region argue against this. There is continuous outcrop along traverses where major offset is indicated by fission track data, such as at Doorly ridge and between Mount England and Avalanche Bay, hence to accommodate the displacement, both large- and small-magnitude offsets must have occurred on the northeast faults that were measured.

A second concern derives from the lack of good age control for faulting. In the absence of stratigraphic control on the ages of the distinct fault sets, is it justified to assume that motion along them was approximately coeval? The high degree of kinematic compatibility between the north-northeast, northeast, and east-northeast fault arrays is illustrated in Figure 6, showing average kinematic axes derived from each fault array compared with average axes derived from the total data sets. For high-rake faults the east-northeast fault array yields a more northerly extension axis. The geometric argument presented previously for a possible Jurassic age for these faults may support the interpretation that this marks a distinct, Jurassic extension direction. The difference in trend of ~25° is greater than the range of solutions from the individual localities, but since most faults of this set are concentrated in the western sector of the Granite Harbour traverse, it is not clear from the present data whether this change has any regional significance. For low-rake faults, there is a ~35° dispersion in the trends of the extension and contraction axes, but data are fewer and individual solutions less tightly constrained. For both the high- and low-rake solutions the overall kinematic

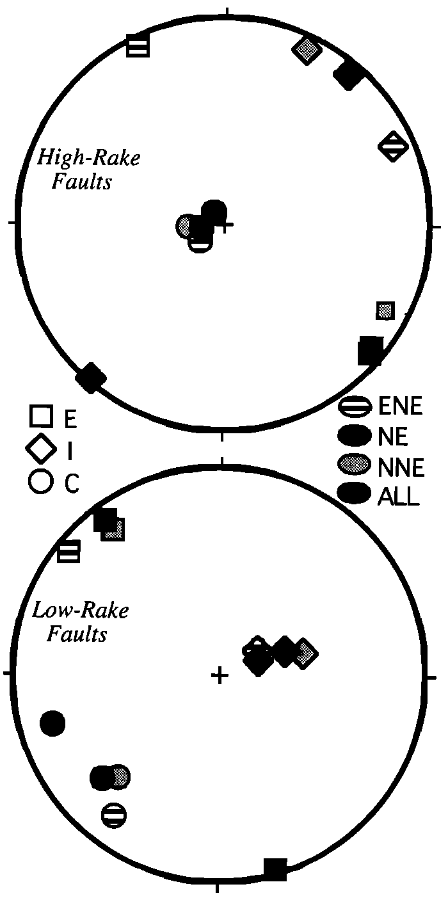

Figure 6. Comparison of kinematic axes derived from each fault array and the total fault population. E is extension axes; I, intermediate axes; and C, contraction axes.

compatibility between the fault sets, the clustering of kinematic axes at individual localities, and the regional consistency of the solutions all indicate either that slip on the fault sets was coeval or that any multiple events had nearly identical strain regimes. In the absence of direct evidence for a series of events I assume that the two-stage history of approximately dip-slip followed by approximately strike-slip faulting events is valid.

The stretching direction for extension was at a markedly oblique angle to the mountains and the presumed trend of the TMF during both phases of faulting. The orientation of the stretching direction with respect to the range indicates that a significant component of right-lateral translation must have accompanied divergence across the TMF. A clockwise rotation of the extension direction of ~14° occurred between the two kinematic phases. At the same time the average contraction direction changed from subvertical to subhorizontal, marking a permutation in the least and intermediate strain axes. Thus a change from an oblique extensional or transtensional regime to a dominantly transcurrent regime is indicated by the kinematic results.

Structural Model

The TM axis in southern Victoria Land trends ~350°, whereas the fault arrays documented here are all northeast and oblique to the axis of the mountain range. The trend of the mountains thus can not be imposed by regionally continuous faults along the TMF structural boundary, as has always been assumed. Instead, the TMF appears to be formed, at least along its exposed, coastal margin, by an en echelon array of northeast faults. This proposed en echelon architecture along a northwest trending, high-relief rift shoulder conflicts with traditional notions of morphotectonics. If no regional, range-bounding faults are present, what controls the regional architecture of the mountains? The preexisting infrastructure of the lithosphere is the most likely control on the trend of the rift-margin uplift and adjacent rift basin. During the Neoproterozoic through early Paleozoic history of the Ross orogen a series of lithospheric-scale structural boundaries were developed. The Neoproterozoic rifted margin and early Paleozoic active margin and associated suture zones are all believed to be approximately parallel to and close to the position of the present TM [e.g., *Borg and DePaolo*, 1991; *Dalziel*, 1992; *Stump*, 1992]. These features mark a zone or possibly a suite of subparallel zones that cuts the entire lithosphere and was thus likely to act as a first-order strain guide during the Mesozoic-Cenozoic rifting of the region.

Both the location and mechanics of rifting are known to be strongly controlled by weaknesses in the continental lithosphere [e.g., *Dunbar and Sawyer*, 1989]. Where a lithospheric weakness nucleates rifting, oblique, rather than orthogonal, rifting may be common because far-field stresses that drive extension are not likely to be perpendicular to the preexisting weak zone. There is, in fact, increasing evidence for oblique rifting in different sectors of the East African Rift, including unidirectional oblique extension [*Tiercelin et al.*, 1988] and rotation of the extension direction from orthogonal to oblique to the rift axis during ongoing rifting [*Strecker et al.*, 1990; *Ring et al.*, 1992]. Of particular note is the recent proposal that both fault arrays and the extension direction are oblique to the long axis of the rift basins and associated rift flanks in the western branch of the East African Rift [*Scott et al.*, 1992].

Widely accepted geometric models have been developed for orthogonal rifting. Either rectilinear arrays of rift-parallel normal faults and perpendicular, strike-slip transfer faults [e.g., *Gibbs*, 1984] or arcuate, rift-parallel, dip- to oblique-slip normal border fault systems and cross-rift oblique- to strike-slip accommodation zones [e.g., *Rosendahl*, 1987] are thought to be typical. The fault patterns documented in this study clearly do not conform with these models. Normal faults are oriented obliquely, rather than parallel, to the regional rift trends and are characterized by oblique-slip displacements. Faults are present that are approximately perpendicular to the trend of the TM (~070°), and these have been compared to "transfer faults" as defined in the orthogonal rifting model [*Fitzgerald*, 1992]. Rather than the strike-slip motion predicted by such a model, however, the transverse faults have dominantly normal-sinistral displacements and are not kinematically compatible with a simple transfer fault origin. No general geometric models for oblique rifting have been proposed, but some experimental and observational studies provide insight into fault geometries in such settings.

Analogue experiments have shown that oblique rifting is characterized by en echelon arrays of faults oriented obliquely to the rift zone and that strain is partitioned between arrays of normal, oblique-slip, and strike-slip faults in cases where the extension direction is oriented at a low angle to the rift axis [*Withjack and Jamison*, 1986; *Tron and Bruhn*, 1991]. There are some striking similarities between the TMF fault arrays and fault patterns produced in model experiments simulating divergent, right-lateral oblique rifting (Figure 7). The dominance of north-northeast to northeast normal faults together with the characteristic component of dextral oblique slip along them compares well with experiments where the displacement direction lies at 30 - 45° to the rift trend. Development of east-northeast faults with sinistral-oblique components matches the 30° case most closely. The second kinematic phase, marked by low-angle dextral slip on the northeast faults and low-angle sinistral slip on the east-northeast faults, equates with the case where displacement occurs at only 15° to the rift trend. The permutation from subvertical to subhorizontal greatest contractional strain, shown by the high-rake and low-rake TMF fault populations, occurs in the experiments where displacement angles are $\leq 30°$. The average extension axes determined from the TMF fault populations approximate the stretching vector in the *Tron and Bruhn* [1991] models for 15-30°, however, no faults with sinistral-oblique components were formed in those experiments.

In Baja California, where an intracontinental extensional transform zone developed prior to spreading in the Gulf of California, faults have oblique trends and a dominant normal-dextral oblique-slip motion pattern [*Angelier et al.*, 1981]. In the East African Rift, oblique extension has been accommodated by subperpendicular sets of oblique-slip faults, with one set subparallel to the extension direction and acting as a transfer fault array and a second group at high angles to the extension direction forming an en echelon border fault system [*Scott et al.*, 1992]. Where rotation of the extension direction has occurred, early rift-parallel faults and perpendicular transfer faults are reactivated as oblique-slip faults and younger crosscutting oblique-slip fault arrays are developed [*Strecker et al.*, 1990; *Ring et al.*, 1992].

The common features in these examples indicate that the key signatures of transtensional versus orthogonal rifting are the oblique trends of fault sets with respect to the rift axis and the oblique-slip motion along the faults. The fault trends in the East African case have been attributed to the influence of preexisting structural grain [*Scott et al.*, 1992]. The analogue experiments show, however, that faults are initiated at oblique angles to preexisting trends except in the orthogonal rifting case [*Withjack and Jamison*, 1986; *Tron and Bruhn*, 1991]. Along the TMF the oblique fault sets were either initiated during the Cenozoic rift phase or mark preexisting fractures that were oriented for easy normal-slip reactivation [*Etheridge*, 1986]. A notable feature of the

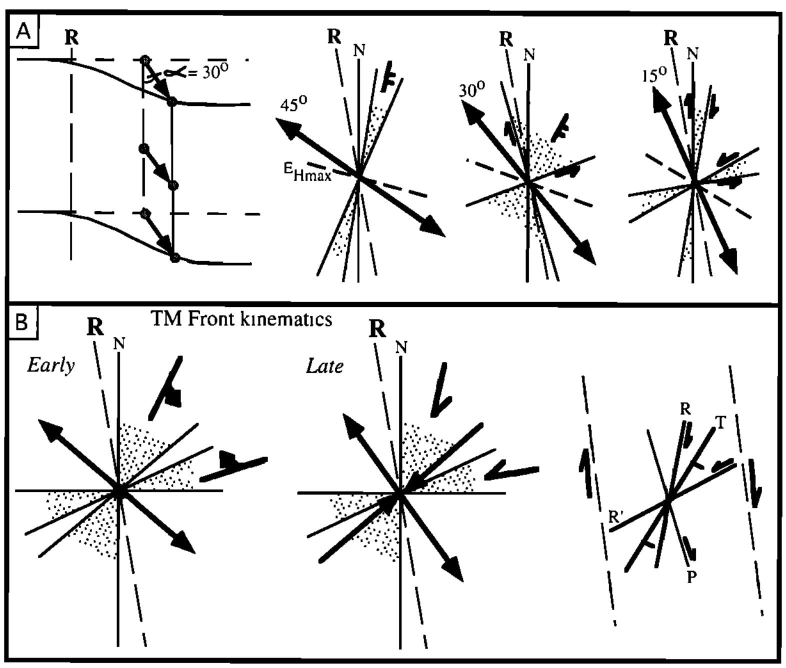

Figure 7. (a) Fault patterns formed in oblique rifting experiments of *Withjack and Jamison* [1986]. Model results are shown relative to the 350° regional trend of the TM for right-oblique displacement directions (bold arrows) relative to the rift trend (R) equal to 15°, 30°, and 45°. (b) Generalized, two-stage Cenozoic TM front fault kinematics derived from this study. For early stage, fault arrays have normal-oblique displacements, the extension direction (bold arrow) is subhorizontal and northwest-southeast, and the contraction direction is subvertical (solid circle); this matches the 30° case in Figure 7a most closely. For late stage, fault arrays have low-angle oblique to strike-slip displacements, the extension direction is subhorizontal and northwest-southeast, and the contraction direction is subhorizontal and southwest-northeast; this matches the 15° model in Figure 7a most closely. Reidel-type fracture arrays associated with right-lateral transcurrent strain regime are shown for comparison.

TMF case is the absence of faults parallel to the extension direction that would act as transfer structures. This is particularly surprising, given that there are regionally developed ductile structural fabrics in the basement rocks that trend ~320° [*Cox*, 1993; *Allibone et al.*, 1993], subparallel to the extension direction derived from the fault populations. This could possibly be because the variable dips of the refolded ductile fabrics precluded reactivation as strike-slip brittle faults during rifting.

The key difference between transtensional and transcurrent strain regimes also appears to be oblique-slip motions along faults. Although the fault sets along the TMF approximate the trends of shear and extension fractures expected to form along a right-lateral strike-slip fault zone (Figure 7), they have normal-oblique displacements, rather than the predicted dip- and strike-slip motions.

Discussion

TMF Faulting and Regional Deformation Patterns

Various structural features in the TM and Ross Sea are compatible with Cenozoic transtensional deformation. Seismic reflection data show that there was a change from an early, wide rift system with multiple basins to localized Cenozoic faulting within the "Terror rift" along the western margin of the Ross Sea (Figure 1) [*Cooper et al.*, 1987, 1991; *Tessensohn and Wörner*, 1991]. This narrowing of the zone of active deformation could be explained by a change from a rift to a dominantly transcurrent regime. The Cenozoic fault geometries within the Terror rift as interpreted from seismic data by *Cooper et al.* [1987] are, in some cases, characterized by upward splay patterns, surficial synforms and antiforms capping fault zones, and apparent normal and reverse offsets of reflectors that are all geometric elements that may be diagnostic of flower structures associated with transtensional or transpressional fault zones [e.g., *Harding*, 1985]. Oblique, northeast trending Cenozoic faults have been mapped in the Drygalski basin at the northern end of the Victoria Land Basin (Figure 1) and interpreted to reflect Neogene extension oblique to the TM margin [*Della Vedova et al.*, 1992]. Cenozoic brittle fault patterns and volcanic structures in the northern Victoria Land sector of the TM indicate northwest-southeast extension oblique to the TM trend [*Lanzafame and Villari*, 1991; *Salvini and Storti*, 1994; *Stackebrandt*, 1994]. The alignment of Cenozoic volcanic centers along structurally controlled glacial valleys transverse to the TM in southern Victoria Land may indicate that oblique stretching caused preexisting transverse structures to become "leaky".

At least for the Victoria Land-Ross Sea portion of the TM-West Antarctic rift system I believe that transtensional deformation provides a more consistent explanation for the regional deformation patterns than does orthogonal rifting. The transtensional model also makes testable predictions about regional structural patterns. In a dominantly transcurrent regime, localized zones of transtension and transpression would be expected at offsets and bends in regional fault systems, and these should be evident in both TM and Ross Sea structural patterns. The possible flower structure patterns along Cenozoic faults in the Ross Sea may be marking such zones, and this can be tested by high-resolution seismic reflection profiling. Detailed mapping of TM structures, particularly at major transverse offsets in the rift flank, may reveal transtensional or transpressional geometries. Block rotations around subvertical axes are also expected in a transtensional setting [e.g., *Tron and Bruhn*, 1991] and may be detectable in paleomagnetic data from the region.

Rift-Related Intracontinental Displacements

The Cenozoic fault arrays along the TMF in southern Victoria Land document a right-oblique transtensional kinematic regime. The superposed slip directions on the fault sets indicate that the transcurrent component became dominant in the late stage history. The subordinate component of Cenozoic crustal stretching perpendicular to the TMF indicated by these data is not compatible with the interpretation that the marked change in crustal thickness across the TMF is the result of Cenozoic rifting [e.g., *Behrendt et al.*, 1991]. This, in turn, implies that the rifting that produced the thin crust that characterizes the Ross embayment as a whole predates the Cenozoic rift phase. Because relative motion between the West Antarctic crustal blocks and the East Antarctic craton must have been produced by rift-related displacements, this has implications for both the timing and kinematics of intracontinental block motions within Antarcica.

Structural and morphologic basins in the Ross embayment are broadly parallel to the TM, and it has generally been assumed that extension was orthogonal to these features throughout the Mesozoic-Cenozoic development of the West Antarctic rift system. My structural data show that this was not the case in the Cenozoic. Other evidence indicates that Mesozoic rift episodes that affected the system also did not have simple, orthogonal kinematics. Marine geophysical data require both extension and dextral translation between East and West Antarctica in the Jurassic, presumably accommodated through the West Antarctic rift system [*Lawver and Gahagan*, 1991]. Large-scale rotations and translations of the Antarctic Peninsula, Thurston Island, and Ellsworth-Whitmore Mountains blocks in the late Jurassic and early Cretaceous are required by paleomagnetic data [*Grunow et al.*, 1991; *Grunow*, 1993] and must have been produced by motion along transpressional and transtensional structural boundaries within the rift system. Paleomagnetic data from the Marie Byrd Land block indicate post-100 Ma translation away from the East Antarctic craton [*DiVenere et al.*, 1994], and it is likely that this motion was complete prior to the split between New Zealand-Campbell Plateau and Marie Byrd Land at ~85 Ma because the Ross rift basins, which presumably accommodated the relative motion, are oriented at a high angle to and are truncated by the passive rifted margin produced during this breakup. Thus the major periods of block motion in West Antarctica occurred between the Jurassic and Late Cretaceous, and the multiple rifting events that induced these motions must have produced the crustal thinning presently observed in the Ross embayment.

Calculations of the magnitude of stretching within the West Antarctic rift system have been based on the assumption that stretching occurred perpendicular to the TM and subparallel basins [e.g., *Storey*, 1991; *Behrendt et al.*, 1991]. This type of calculation, with the further assumption that all of the stretching was coeval with TM uplift over the last ~55 Ma, was used to evaluate the role of Antarctic intraplate deformation in resolving discrepancies in plate motion circuit models [*Kamp and Fitzgerald*, 1987]. The complex, multistage kinematics of West Antarctic rifting, the argument that the major crustal stretching in the system was of Mesozoic age, and the narrow zone of dextral transtensional deformation postulated here for the Cenozoic rift phase, together appear to preclude the several hundred kilometers of early Tertiary extension between East and West Antarctica that would be required to reconcile global plate motion circuits [*Stock and Molnar*, 1987; *Acton and Gordon*, 1994]. The dextral transcurrent component along the TM-rift boundary documented here is of the opposite sense to the early Tertiary sinistral strike-slip motion of ~800 km calculated assuming fixed hotspots by *Acton and Gordon* [1994].

The magnitude of Cenozoic dextral translation is presently unconstrained, but no Cenozoic block motions have been detected paleomagnetically. The isolation of the Antarctic continent by breakup and seafloor spreading was essentially complete by ~85 Ma, with the exception of a junction between northern Victoria Land and a narrow salient projecting from the southeastern Australian margin, including Tasmania and the South Tasman Rise, which remained juxtaposed until ~30-40 Ma [*Lawver et al.*, 1992]. The time frame for dextral transtension is probably between this time and the onset of TM uplift at ~55-60 Ma. The latter event is approximately coeval with the Eocene global plate reorganization [*Cooper et al.*, 1991; *Stump and Fitzgerald*, 1992]. The episode of transtensional deformation along the TM-Ross Sea boundary was thus most likely the result of stresses generated by the redirected plate interactions prior to complete separation of the irregular Antarctica-Australia margin in the vicinity of the northwestern Ross Sea. The imposed far-field stresses were sufficient to produce reactivation of the fundamental lithospheric weakness along the ancestral East Antarctic margin.

Implications for TM Uplift Models

General rift-flank uplift models invoke heat transfer from the rift, dynamic support during active extension, and/or regional isostatic adjustments during extension [e.g., *Weissel and Karner*, 1989; *Braun and Beaumont*, 1989]. Combinations of thermal transfer and isostatic compensation for load changes due to normal faulting have been proposed to explain TM uplift [*Fitzgerald et al.*, 1986; *Stern and ten Brink*, 1989]. Uplift due either to thinning of the mantle lithosphere [*Fitzgerald et al.*, 1986] or to flexure in response to mechanical unloading of the footwall block [*Stern and ten Brink*, 1989] occurs as a result of mass redistribution perpendicular to the rift flank caused by lithosphere-scale normal faulting. In a transtensional regime, however, a markedly different mass redistribution would occur because loads are shifted axially along the rift. Unless the proportion of divergence and thus the component of orthogonal unloading is dominant, little isostatic uplift would occur in a transtensional setting. My fault data document a transtensional regime during Cenozoic uplift of the TM and, by analogy with experimental models, suggests that the transport direction during rifting was at angles of only 15-30° to the TM trend. Simple flexural uplift resulting from mechanical unloading alone is unlikely to have resulted from such oblique rift kinematics. Although some combination of mechanical and thermal processes acting on the margin of the obliquely extending rift must have produced the observed Cenozoic uplift, the nature of the uplift mechanism remains unresolved.

Conclusions

Brittle faults along the margin between the TM and the Ross Sea sector of the West Antarctic rift system have a consistent oblique orientation with respect to the regional mountain trend and record dextral transtension across the rift boundary. The rift margin appears to consist of an oblique, en echelon array of oblique-slip faults, rather than a regionally continuous, rift-parallel normal border fault system. The persistent north to northwest trend of the TM must therefore be inherited from the preexisting infrastructure along the margin of the East Antarctic craton. Increasing evidence of oblique extension in other continental rifts suggests that transtensional rifting may be typical where rifts nucleate along profound lithospheric weaknesses. Rift morphotectonics may thus be more indicative of the nature of the rift infrastructure than of rift kinematics. Models for rift-flank uplift that rely on orthogonal extension to mechanically unload the rift margin and induce isostatic uplift do not appear to be viable for obliquely extending rifts unless the component of divergence is more dominant than translation along the rift axis. Uplift mechanisms for such transtensional rift margins are not yet understood.

A zone of localized transtension along the TM-West Antarctic rift margin in Victoria Land is indicated by regional structural patterns. Large-scale Cenozoic crustal stretching within the West Antarctic rift system is not compatible with this kinematic regime. Major early Tertiary displacement between East and West Antarctica across an intraplate boundary within the TM-Ross embayment sector of the rift system is thus unlikely and can not be invoked to resolve discrepancies in global plate motion circuit models. The Cenozoic transtensional rift phase can be linked with plate interactions along the Antarctic-Australian margin and appears to mark an episode of intracontinental deformation related to the final decoupling of the Antarctic continent by a continuous spreading ridge system.

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