TECTONICS, VOL. 12, NO. 6, PAGES 1460-1478, DECEMBER 1993

KINEMATIC EVOLUTION OF THE MILLER RANGE SHEAR ZONE, CENTRAL TRANSANTARCTIC MOUNTAINS, ANTARCTICA, AND IMPLICATIONS FOR NEOPROTEROZOIC TO EARLY PALEOZOIC TECTONICS OF THE EAST ANTARCTIC MARGIN OF GONDWANA

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Abstract. High-grade ductile tectonites of the Precambrian Nimrod Group in the central Transantarctic Mountains form the Miller Range shear zone (MRSZ). With no exposed boundaries, this zone has a minimum structural thickness of 12-15 km. Shear-sense indicators record consistent top-to-the-SE, or left-lateral, shear within the NW striking, moderately SW dipping zone. Cylindrical folds with axes normal to elongation lineation (L_e) are kinematically consistent with other shear indicators. They may represent early stages in the development of subordinate noncylindrical sheath folds, which indicate locally high bulk ductile strain and a moderate strain gradient. Pervasive, open to tight cylindrical folds with axes parallel to L_e formed during shear and may reflect a component of constrictional strain. Quartz *c* axis fabrics from micaceous quartzites show asymmetric single girdles evident of dominantly rhombohedral slip, with limited basal-plane slip, affirming both the consistency of shear sense and high-grade syn-kinematic conditions. Deformation did not persist during subamphibolite facies cooling, as shown by (1) a lack of basal-plane slip in ductilely deformed quartz, (2) a lack of quartz subgrains and grain shape-preferred orientation, and (3) the presence of oriented muscovite "fish" included within polygonal quartz grains, which show that quartz grain boundaries migrated and annealed under static conditions following ductile shear. From the uniform L_e orientation and consistent shear sense, we interpret that ductile deformation resulted from a single, kinematically simple, left-lateral (top-to-the-SE) shear event. Together, the scale, high total strains ($\gamma \geq 5$), fabric uniformity, and the widespread presence of asymmetric microstructures formed at high temperatures, all indicate that strain rates within the MRSZ were high and that it represents a major crustal structure. Orogen-parallel displacements within this zone during the latest Neoproterozoic to Early Cambrian were at a high angle to penecontemporaneous orogen-normal contraction in outboard supracrustal rocks, suggesting that the Neoproterozoic to early Paleozoic plate margin of Antarctica was characterized by left-oblique convergence in which strain within the orogen was partitioned into deep-level strike slip and shallow-level contraction.

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Paper number 93TC02192.
0278-7407/93/93TC02192$10.00

INTRODUCTION

The Transantarctic Mountains constitute a major orogenic belt in Antarctica, extending for ~3000 km between the present-day Weddell and Ross seas. Although the modern physiography of the Transantarctic Mountains is dominantly a manifestation of Cenozoic rift tectonics [Stern and ten Brink, 1989], its fundamental underlying tectonic architecture is largely a result of its Neoproterozoic to early Paleozoic history. The Neoproterozoic-early Paleozoic tectonic evolution of this orogen primarily involved interactions between the stabilized East Antarctic craton and younger tectonic elements along the Antarctic margin of Gondwana, and a better understanding of this tectonic history is relevant to evaluating plate reconstructions that have been proposed as models for supercontinental assembly, fragmentation, and reassembly [Dalziel, 1992].

The Transantarctic Mountains are noted for multiple Neoproterozoic to early Paleozoic tectonic events related to rifting and convergence, as are counterparts in Africa and Australia [Stump, 1987]. The central part of the orogen in the Nimrod Glacier area, however, is unique in exposing a Precambrian metamorphic terrain of the East Antarctic craton that lies inboard of younger, deformed sedimentary sequences (Figure 1). Previously, three orogenic events were recognized on the basis of geologic relations, the Precambrian Nimrod, Neoproterozoic(?) Beardmore, and Cambro-Ordovician Ross orogenies, although distinguishing these events temporally is a formidable problem because of poor age constraints, discontinuous exposures, and ambiguous structural relations. In the traditional view, the Nimrod deformation affected high-grade metamorphic rocks of the Nimrod Group [Grindley and McDougall, 1969; Grindley, 1972], whereas supracrustal rocks deposited along the outer margin facing the modern Ross Sea (Beardmore and Byrd groups) were only affected by the later Beardmore and Ross contractional deformations [Grindley and Laird, 1969; Laird et al., 1971; Stump, 1981; Stump et al., 1986, 1991].

Recent studies, however, provide an emerging picture of punctuated Antarctic-margin orogenesis during the Neoproterozoic to Early Ordovician that is associated with margin-parallel displacements. First, stratigraphic, structural, and geochronologic data show evidence of multiple tectonic events during the latest Proterozoic and early Paleozoic that do not fit into a simple Nimrod-Beardmore-Ross orogenic succession [Rees et al., 1987; Rowell and Rees, 1989; Rowell et al., 1991, 1992; Goodge et al., 1993]. At least four discrete tectonic events that occurred over a time span of ~80 m.y., straddling the Precambrian-Cambrian boundary, are now recognized in basement and supracrustal rocks of the central Transantarctic Mountains [Rowell et al., 1988, 1991, 1992; Goodge et al., 1993]. Second, stratigraphic, structural and isotopic evidence indicates that orogenic development involved a component of strike-slip deformation in addition to a clear record of regional contraction [Rowell and Rees, 1990; Goodge et al., 1991a]. Thus the concept of a narrowly defined event such as the Ross orogeny, regarded as the most pronounced

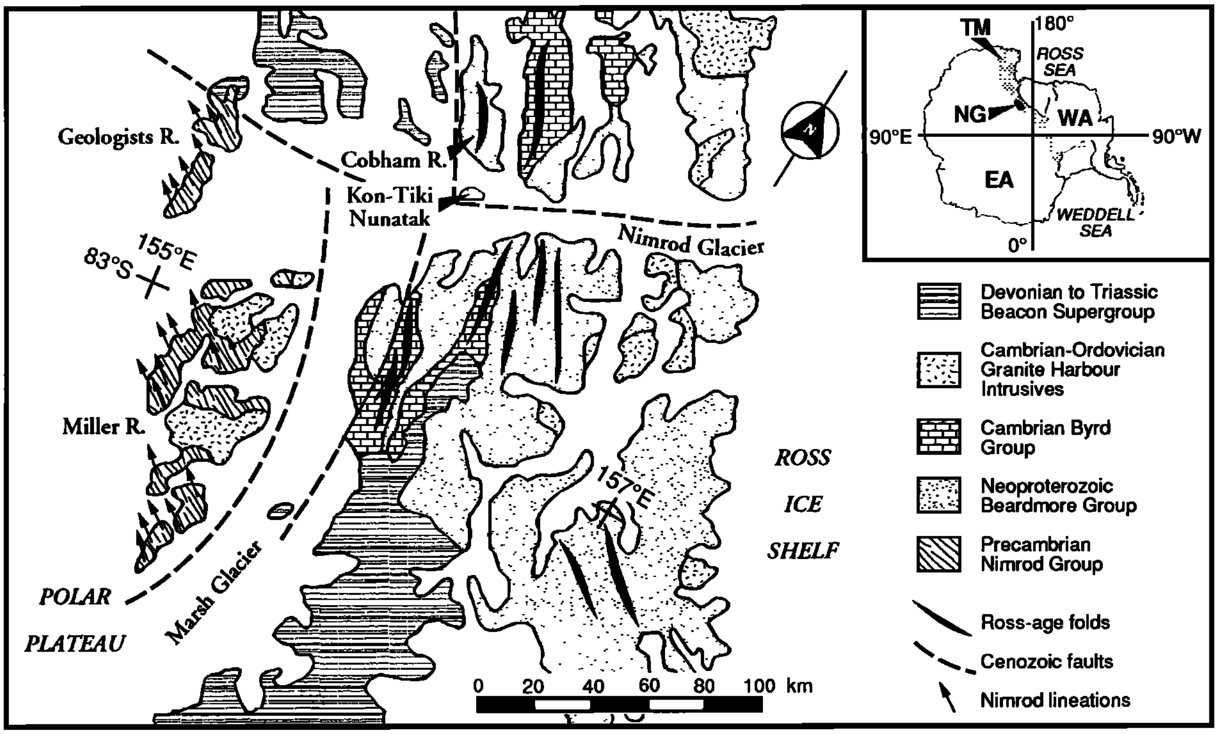

Fig. 1. Generalized geologic map of the central Transantarctic Mountains in the vicinity of Nimrod Glacier [from Grindley and Laird, 1969]. Arrows show trends of representative Nimrod Group mylonitic lineations. Inset shows location in Antarctica; EA denotes East Antarctica; NG denotes Nimrod Group; TM denotes Transantarctic Mountains; and WA denotes West Antarctica.

along the entire margin, should perhaps be abandoned in favor of a diachronous series of tectonic pulses that may have included nonorthogonal displacements.

In this paper we discuss the structural and kinematic history of L-S (L = elongation lineation; S = foliation) tectonites from the Precambrian Nimrod Group, exposed in the Miller and Geologists ranges of the central Transantarctic Mountains (Figure 1). We document field relations, ductile shear fabrics, the relation of these fabrics to fold development, and kinematic criteria to show that Nimrod tectonism is best explained by a single tectonic event involving synmetamorphic, penetrative, and progressive ductile flow within the 12 to 15-km-wide Miller Range shear zone. This ductile shear zone represents a large crustal structure with primarily top-to-the-SE, or left-lateral, displacement that was accompanied by minor NE-SW constriction. Deformation occurred under high-T conditions within the middle to lower crust, probably at relatively high strain rates. These interpretations are in contrast to those which explain Nimrod tectonism by NE directed shortening [Grindley, 1972]. They also modify the notion of a discrete décollement, including both the Endurance thrust of Grindley [1972] and the ~1-km-wide Endurance shear zone of Goodge et al. [1991a]. Our structural data, in conjunction with geochronologic data [Walker and Goodge, 1991; Goodge and Dallmeyer, 1992; Goodge et al., 1993] and structural relations within outboard supracrustal rocks, provide the framework for a tectonic model involving Neoproterozoic to early Paleozoic oblique plate interactions and strain partitioning at different crustal levels manifested by synchronous orogen-parallel and orogen-normal displacements.

TECTONIC SETTING

Rocks of the Precambrian Nimrod Group comprise an isolated high-grade metamorphic complex in the central Transantarctic Mountains, and they have no proven counterpart in the Ross Sea sector of the Transantarctic Mountains (Figure 1). In the Nimrod Glacier region, exposures of metamorphic, plutonic and sedimentary units are limited, due to extensive cover by ice, and by generally flat-lying, Devonian-Jurassic Gondwana overlap strata. Precambrian metamorphic rocks and Neoproterozoic to lower Paleozoic sedimentary units have long been known to record regional tectonism, reflected by contrasting lithologic assemblages, abrupt transitions in metamorphic grade, magmatism, and large-scale fold-belt structures [Gunn and Walcott, 1962; Grindley et al., 1964; Grindley and McDougall, 1969; Gunner, 1969; Laird et al., 1971; Grindley, 1972; Rowell et al., 1988; Goodge et al., 1991a, b].

The major pre-Devonian lithotectonic units of the Nimrod Glacier area include (Figure 1): (1) the Nimrod Group [Grindley and Warren, 1964], a lithologically diverse complex of strongly deformed high-grade gneisses and schists of Precambrian protolith age [Gunner and Mattinson, 1975; Gunner, 1983; Goodge et al., 1990, 1991a; Walker and Goodge, 1991]; (2) the Neoproterozoic(?) Beardmore Group, a thin sequence of metacarbonate and quartzite (Cobham Formation of Laird et al. [1971]) conformably overlain by low-grade, unfossiliferous turbiditic greywacke and slate (Goldie Formation of Gunn and Walcott [1962]); (3) the Byrd Group [Laird, 1963], Lower Cambrian shallow-water carbonates (Shackleton Limestone) [Laird and Waterhouse, 1962; Rowell and Rees, 1989; Rees et al., 1989] unconformably overlain by a sequence of Middle to Upper Cambrian clastic units (Starshot, Dick, and Douglas formations) [Skinner, 1964; Rowell and Rees, 1989, 1990]; and (4) the 550-480 Ma Hope granite suite, part of the regionally extensive Granite Harbour Intrusive Complex [Gunn and Warren, 1962; Gunner, 1976; Faure et al., 1979; Borg, 1983; Borg et al., 1990; Goodge et al., 1993]. In terms of lithology, age, metamorphism and structural relations, the Nimrod Group is unique within the Transantarctic Mountains, although a metamorphic terrain with similar lithologies exists in northern Victoria Land [e.g., Grew et al., 1984; Kleinschmidt and Tessensohn, 1987; Kleinschmidt et al., 1987]. Rocks of the Nimrod and Beardmore groups are not in exposed contact, yet Grindley and Laird [1969] inferred that Beardmore strata unconformably overlie the Nimrod Group, whereas Stump et al. [1991] suggested a correlation between the Cobham Formation and part of the Nimrod Group. An unconformable relation between Goldie clastic rocks and overlying Shackleton Limestone [Laird, 1963; Laird et al., 1971; Laird and Bradshaw, 1982; Stump et al., 1991] indicates that the Goldie is older than Early Cambrian. A gabbroic unit associated with pillow basalt in the Goldie strata yielded a Sm-Nd isochron age of 762 Ma [Borg et al., 1990], indicating that the Goldie rocks are probably Neoproterozoic in age. This age assignment for the Goldie assumes, however, that the dated mafic rocks are volcanic in origin.

Among the proposed Neoproterozoic to early Paleozoic deformation events in the central Transantarctic Mountains, the most conspicuous is the early Middle Cambrian to Early Ordovician Ross orogeny. This regional contractional deformation is recognized by open, upright folds and a steep east dipping (toward Ross Ice Shelf) cleavage in Beardmore and Byrd supracrustal rocks [Laird et al., 1971; Rowell et al., 1986; Rees et al., 1989; Stump et al., 1991; Goodge et al., 1991b], and it culminated in regional Cambro-Ordovician magmatism [McDougall and Grindley, 1965; Gunner and Mattinson, 1975; Gunner, 1976; Borg et al., 1990]. Recent studies suggest that the so-called Ross event, as defined by deformation of supracrustal sedimentary rocks, was punctuated, involving as many as three discrete pulses of activity during and after Early to Middle Cambrian deposition of Byrd Group sediments [Rowell and Rees, 1989; Rowell et al., 1991]. The age of the broadly defined Ross deformation is constrained by deformed Lower to Middle Cambrian sedimentary rocks [Rowell et al., 1992] and the emplacement of ~500 Ma posttectonic plutons [Gunner, 1976; also J. Mattinson, personal communication]. The Beardmore orogeny was originally defined on the basis of an angular unconformity between sedimentary rocks of the Beardmore and Byrd groups [Grindley and McDougall, 1969; Laird et al., 1971; Stump et al., 1991], although some of these boundaries are now recognized to be faults [Rowell et al., 1986]. Beneath this alleged unconformity where it was identified south of Nimrod Glacier [Laird et al., 1971; Stump et al., 1991], close to tight folds in Beardmore strata are nearly coaxial with Ross contractional structures. Stump et al. [1986] suggested that Neoproterozoic intrusive rocks cut deformed Goldie-equivalent strata, but reanalysis indicates that these intrusives are probably Late Cambrian to Ordovician in age (502 ± 5 Ma, Rb-Sr) [Pankhurst et al., 1988]. Therefore as presently known, deformation of Goldie rocks is only constrained to be post-762 Ma and pre-Early Ordovician. Geologic and geochronologic data have thus cast doubt on the occurrence of a definitive Neoproterozoic Beardmore deformation.

Early studies of the Nimrod Group indicated that these rocks were complexly deformed during a high-grade metamorphic event (Nimrod orogeny of Grindley and Laird [1969]) that is not expressed in younger sedimentary units [Grindley and Warren, 1964; Grindley and McDougall, 1969; Grindley, 1972]. K-Ar metamorphic mineral ages indicated this deformation may be as old as Neoproterozoic in age [Grindley and McDougall, 1969]. Goodge and Dallmeyer [1992] showed that incorporation of extraneous argon was the likely cause of the Neoproterozoic K-Ar ages, and they provided evidence of Nimrod cooling below ~500°C between 525-480 Ma. A ~1.7 Ga orthogneiss that crosscuts Nimrod metasedimentary rocks indicates a period of cryptic Paleoproterozoic tectonism [Goodge et al., 1991a], but the most severe Nimrod deformation, the subject of this paper, was in progress by about 540 Ma [Goodge et al., 1993].

The relationships between these orogenic events remain enigmatic. Doubt about the Goldie-Shackleton unconformity raises the possibility that the Ross and Beardmore events are the same, or perhaps closely related in time. Temporal overlap of Byrd and Nimrod group deformations, although of different structural styles and crustal levels, suggests cogenetic tectonism. Thus evidence is mounting for protracted but punctuated Neoproterozoic to early Paleozoic tectonism.

In addition to timing, the plate tectonic settings for these orogenic episodes are uncertain. The structural styles associated with the latest Proterozoic to early Paleozoic deformation events are compatible with an overall convergent plate-margin regime, perhaps involving subduction and/or arc accretion. Many workers have suggested that the early Paleozoic Antarctic plate margin was the site of convergence involving subduction of oceanic lithosphere beneath the East Antarctic craton [Elliot, 1975; Bradshaw et al., 1985; Gibson and Wright, 1985; Borg et al., 1987, 1990; Kleinschmidt and Tessensohn, 1987]. Structural relations are an important way to establish the relative timing of tectonic events, and they also provide a measure of geometric and kinematic compatibility among events. Here, we document the structural and kinematic evolution of Nimrod Group rocks that must be established in order to develop plausible larger-scale tectonic models.

MILLER RANGE SHEAR ZONE

The Nimrod Group, exposed in the Miller and Geologists ranges (Figure 1 and Plate 1), is a metamorphic basement complex composed of schist, metacarbonate, orthogneiss, and compositionally layered gneiss [Goodge et al., 1990, 1991a, b]. Grindley et al. [1964] gave similar map units formational status, but we defined a greater variety of units on the basis of rock composition and cross-cutting relations. We emphasize, however, that several of the metamorphic map units shown in Plate 1 are recognized on the basis of their overall lithologic character, and they may be quite heterogeneous on a smaller scale. Major lithologies among these map units include pelitic schist, micaceous quartzite, amphibolite, banded quartzofeldspathic to mafic gneiss, homogeneous (garnet-) biotite-hornblende gneiss, granitic to gabbroic orthogneiss, migmatite, calc-silicate gneiss, and marble. Structurally enclosed within Nimrod rocks, particularly along the boundary separating quartzofeldspathic gneiss from schists, are discrete tectonic blocks (0.5-50 m) of mafic and ultramafic composition (Plate 1). Many mafic blocks contain relict eclogite facies mineral assemblages [Goodge et al., 1992], as noted earlier by Grindley [1972], but metamorphic grade in all other rock types ranges from upper amphibolite to lower granulite facies [Goodge et al., 1992].

Grindley [1972] first documented structural relations of Nimrod Group rocks in the Miller Range. He described five phases of deformation, the first three of which he ascribed to the Nimrod orogeny, and the later two phases of faulting and flexure he interpreted as minor events during regional early Paleozoic tectonism. According to Grindley [1972], structures formed during high-grade metamorphism include both NW and NE trending tight to isoclinal folds. Two fold generations (F_1 and F_2) with NW trending axes were interpreted as the result of NE directed shortening along a structure he termed the Endurance thrust, placing banded gneisses upon the other units of the Nimrod Group. Grindley's third phase of Nimrod folding (F_3), marked by NE trending fold axes and no apparent axial plane schistosity, was a product of late-stage SE directed movement following the main-phase Endurance displacements.

In contrast to Grindley's findings, field study by Goodge et al. [1991a] showed that (1) the Endurance thrust is a distributed ductile shear zone with a structural width of ~1 km, rather than a discrete thrust surface (referred to by Goodge et al. [1991a] as the Endurance shear zone); and (2) kinematic features within this zone record principal top-to-the-SE, or left-lateral, displacement.

Building on this earlier work, detailed field and petrofabric study shows that rocks of the Nimrod Group throughout the Miller and Geologists ranges display well-formed ductile L-S tectonite fabrics characterized by composition layer-parallel shear foliation (S) and mineral elongation lineation (L_e) (Plate 1). S generally dips moderately to the SW, and L_e plunges gently to the NW and SE (see fabric diagrams, Plate 1). Ductile L-S tectonite fabrics are pervasive in the Nimrod Group, and we will here refer to these rocks generally as Nimrod tectonites; tectonite exposure and dip constrain a minimum structural thickness of 12-15 km before the fabrics disappear beneath snow and ice. The Nimrod tectonites formed during noncoaxial ductile shear, as developed below, and, collectively, they comprise the Miller Range shear zone (MRSZ), a term we introduce in order to convey the large region over which these rocks are exposed. We document relationships below which show that tectonites of the Endurance shear zone record extremely high strain, and we thus retain use of this separate term to delineate a zone of relatively higher strain within the MRSZ.

Nimrod ductile tectonism occurred under moderate-P, high-T conditions ($P \geq 8$ kbar; $T \approx 700°C$) in the upper amphibolite to lower granulite facies, as shown by synkinematic kyanite + garnet + muscovite + biotite + quartz in pelites, hornblende + plagioclase ± garnet ± clinopyroxene ± clinozoisite in mafic rocks, and by thermobarometry [Goodge et al., 1992]. Late synkinematic growth of sillimanite after kyanite reflects waning deformation along a decompression path. A variety of granitic to pegmatitic dikes and sills crosscut the Nimrod tectonites; these intrusive bodies display variable L-S fabric formation, and they are planar to boudinaged or folded. Formation of some pegmatitic melts may have been related to local in situ partial melting, consistent with the high-grade metamorphic conditions that accompanied shear deformation and outcrop evidence such as mafic mineral selvedges and serrate contacts. These features indicate that dike and sill intrusion occurred at several stages during ductile deformation. The great abundance of these synkinematic dikes and sills may have thermally enhanced ductility in some areas.

Tectonite Fabrics

Nimrod Group tectonites contain monoclinic to triclinic fabrics composed of tectonite foliation (S), elongation lineation (L_e), fabric asymmetries, and mesoscopic folds. S is composed of phyllosilicate and amphibole preferred orientation in schists and gneisses, and lenticular feldspar porphyroclasts in orthogneisses. L_e is formed by rodded quartz, crenulated mica, and elongate or aligned feldspar, amphibole, sillimanite or kyanite, depending on host lithology. L_e is contained within S (*XY* plane), commonly coincident with compositional layering. The plane normal to S and parallel to L_e (*XZ* plane) commonly contains microstructural asymmetries that reflect ductile shear vorticity, and it is therefore called the motion plane [Arthaud, 1969]. The plane normal to both S and the motion plane (*YZ* plane) may contain a lineation, a symmetric planar fabric, or an axial-profile view of folds.

Fabric asymmetry within the motion plane records the direction of tectonic transport during ductile deformation [e.g., Eisbacher, 1970; Berthé et al., 1979; Simpson and Schmid, 1983; Hanmer and Passchier, 1991]. Motion-plane asymmetries in the Nimrod tectonites are observed at the macroscopic, mesoscopic, and microscopic scales. Shear-sense indicators useful for kinematic interpretation of Nimrod tectonites include mesoscopic to microscopic asymmetric folds with axes normal to L_e; rotated clasts and winged inclusions [Hanmer and Passchier, 1991]; stretched, imbricated, and locally folded pull-aparts [Hanmer, 1986]; foliation "fish" [Hanmer, 1986]; types I and II S-C fabrics [Lister and Snoke, 1984]; shear bands [Platt and Vissers, 1980], developed mostly in paragneiss tectonites; σ and δ feldspar megacrysts [Passchier and Simpson, 1986], developed in orthogneiss tectonites; mica, amphibole, and kyanite "fish"; mineral grain shape-preferred orientations; and quartz *c* axis fabrics. In all cases, these asymmetries record top-to-the-SE, or left-lateral, shear parallel to L_e (mean N46°W).

Nimrod tectonites may be described generally as ductile L-S tectonites, but ultramylonites occur within the Endurance shear zone. Asymmetric motion-plane structures are nearly ubiquitous at the mesoscopic scale, and we interpreted them with confidence in the field (Plate 1). The types of mesoscopic structures displayed by the Nimrod tectonites, described below, are typical of those seen in other large-scale shear zones [e.g., Davidson, 1984]. Although tectonite fabrics displaying interpretable kinematics occur throughout the Nimrod complex, fabric intensity varies greatly at the outcrop and larger scales. This fabric heterogeneity may be the result of variations in either strain intensity or lithologic rheology, and we are unable to differentiate these mechanisms. However, the ubiquitous presence of tectonite fabrics indicates that the entire Nimrod complex forms a single large-scale shear zone, the locus of which may be represented by the Endurance shear zone, but the boundaries of which are not exposed.

Asymmetric and imbricated pull-aparts, including both boudinage and pinch-and-swell structures [Hanmer, 1986], occur throughout the MRSZ. These structures are locally associated with SE vergent folds, and themselves record SE directed displacement. Pull-aparts are preserved in four stages of formation: (1) pull-apart boudinage of a rigid layer; (2) rotation of pull-aparts into an en echelon series, indicating the vorticity of noncoaxial shear (Figure 2a); (3) imbrication and stacking of tabular detached boudins; and (4) deformation of the adjacent ductile layer, and locally the boudins themselves, into asymmetric SE vergent folds. Pull-aparts observed within the MRSZ range in size from 1-10 cm rotated amphibolite inclusions (Figure 2b) to relict eclogitic blocks ranging from 1-50 m in size (Figure 2c). Foliation "fish" [Hanmer, 1986], also common within the MRSZ (Figure 2d), allow confident kinematic interpretation in the field.

Types I and II S-C fabrics [Lister and Snoke, 1984] in granodiorite orthogneiss, and paragneiss and schist, respectively, consistently record top-to-the-SE displacement. Kinematically compatible shear bands [Platt and Vissers, 1980] are most common in paragneiss, yet they are developed locally in orthogneiss as well. Type I S-C fabrics, documented in the Camp Ridge orthogneiss of the Endurance shear zone [Goodge et al., 1991a, Figure 3], are also observed in other orthogneiss units. Large orthogneiss bodies of this type have semitabular shapes that appear to parallel the regional metamorphic foliation (Plate 1). Some orthogneiss units, however, display less strongly formed tectonite fabrics characterized chiefly by linear elements. The presence of S-C fabrics in orthogneiss units indicates that their parent igneous bodies intruded prior to, or during, Nimrod tectonism. Although some orthogneiss bodies in the Miller and Geologists ranges appear petrographically and texturally similar, they are not all of the same age. The Camp Ridge granodiorite orthogneiss that contains strong type I S-C fabrics within the Endurance shear zone yielded a zircon $^{207}Pb/^{206}Pb$ date of ~1.7 Ga [Goodge et al., 1991a], whereas a granodiorite orthogneiss with similar igneous protolith characteristics and tectonite fabrics in the southeastern Miller Range near Orr Peak yielded a zircon U-Pb date of ~540 Ma [Goodge et al., 1993].

Orthogneiss tectonites of the MRSZ also display feldspar σ porphyroclasts [Passchier and Simpson, 1986] that record SE directed shear. Within the Endurance shear zone both δ and σ porphyroclasts record sinistral, or SE-vergent, vorticity. In order for δ porphyroclasts to form, the ratio of recrystallization rate to deformation rate must be low and shear strain values must be high ($\gamma > 5$-10) [Passchier and Simpson, 1986]. Because mineral recrystallization rates were probably relatively high in this case, given the high-temperature synkinematic conditions [Goodge et al., 1992], the presence of δ porphyroclasts indicates relatively high strain rates during deformation within at least the Endurance shear zone.

The motion-plane views of these tectonites are relatively simple, comprised of foliations with or without folds and microstructures indicating consistent shear sense. In contrast, the plane perpendicular to foliation and lineation may display fabric disruptions on the scale of a hand sample to a that of a cliff face. These heterogeneities take many forms: (1) foliation phacoids, in which foliation and compositional layering are truncated along discrete shear surfaces resulting in lensoidal domains (Figure 2e); (2) discrete blocks, which themselves are foliated and locally boudinaged, enclosed in a second tectonite foliation that wraps completely or partially around the foliated clast (Figure 2f); (3) foliation that is folded around lineation-parallel folds, described below, or foliation and compositional layering that are boudinaged, broken, folded, stacked, or warped, often pinching off along layer strike. One layer may display folds or pinching and swelling along its length, whereas an adjacent layer is boudinaged, broken, and stacked. In each case, brittle failure is within certain layers and comprises part of a larger regime of ductile failure. These fabric heterogeneities provide evidence that local brittle failure accompanied dominant ductile tectonism.

Folds

Three types of folds occur within the MRSZ (Figures 3 and 4): (1) pervasive open to tight, cylindrical folds of compositional layering with axes parallel to L_e, present at the map to hand-sample scale; (2) SE vergent cylindrical folds with axes normal to L_e, observed at the macroscopic to microscopic scale; and (3) subordinate noncylindrical sheath folds elongated parallel to L_e, found at the mesoscopic scale. We refer to these folds as types 1, 2, and 3, respectively.

Map-scale type 1 folds deform compositional layering as open, gently NW plunging to subhorizontal folds in the Geologists and northern Miller ranges, and open to close, gently NW plunging folds in the southern Miller Range (Plate 1). These folds show no preferred vergence or asymmetry; they verge NE in the Geologists and northern Miller ranges, and SW in the southern Miller Range (Figures 3a and 3b). Locally, outcrop and hand-sample scale type 1 folds are extremely complex with a large variation in axial plane orientation (Figure 3c), yet due to the cylindrical nature of these folds the associated L_e-parallel view in each case is structurally simple (Figure 3d). Mesoscopic type 1 folds display open to tight interlimb angles. In some cases, well-developed L_e-parallel rods or mullions are exposed on fold-limb surfaces. Rootless type 1 folds, refolded by other folds of type 1 geometry, occur locally. Type 1 folds always deform compositional layering, which in many cases is parallel to tectonite foliation (S).

Type 1 folds display three different relationships with S as observed in axial profile.

1. Open folds deform coplanar compositional layering and S. A series of such folds resembles a corrugated surface.
2. Folds deform compositional layering, locally with coplanar S, but an axial planar tectonite foliation is also developed. These relations occur typically in tight to isoclinal examples.
3. Tight to isoclinal folds of compositional layering, contained within phacoidal domains, are truncated by discrete shear zones that are asymptotic to their limbs.

Type 2 folds are common within the MRSZ, but they are present chiefly within a 5-km-thick zone centered approximately on the Endurance shear zone near Camp Ridge and rarely in other areas (Plate 1). Type 2 folds are cylindrical, tight to isoclinal folds with axes normal, or at a high angle, to L_e. S is coplanar with type 2 axial planes. These structures, which fold compositional layering and, locally, tectonite foliation, display consistent SE vergence with axial planes at a shallow angle to S (Figure 3e). Type 2 folds are observed at outcrop, hand-sample, and microscopic scales. In outcrop type 2 folds are most commonly associated with strongly contrasting compositional layers [e.g., Goodge et al., 1991a; Figure 3], and they are locally associated with pull-apart or boudinage structures. The consistent SE vergent asymmetry viewed normal to L_e indicates that these folds formed during

Geologic Map of the Miller and Geologists Ranges, Central Transantarctic Mountains, Antarctica

by J. W. Goodge, V. L. Hansen, S. M. Peacock, M. J. Roberts, B. K. Smith, and N. W. Walker

Mapping completed 1985-91

Geology also modified from Barrett et al. [1970] and Grindley [1972]

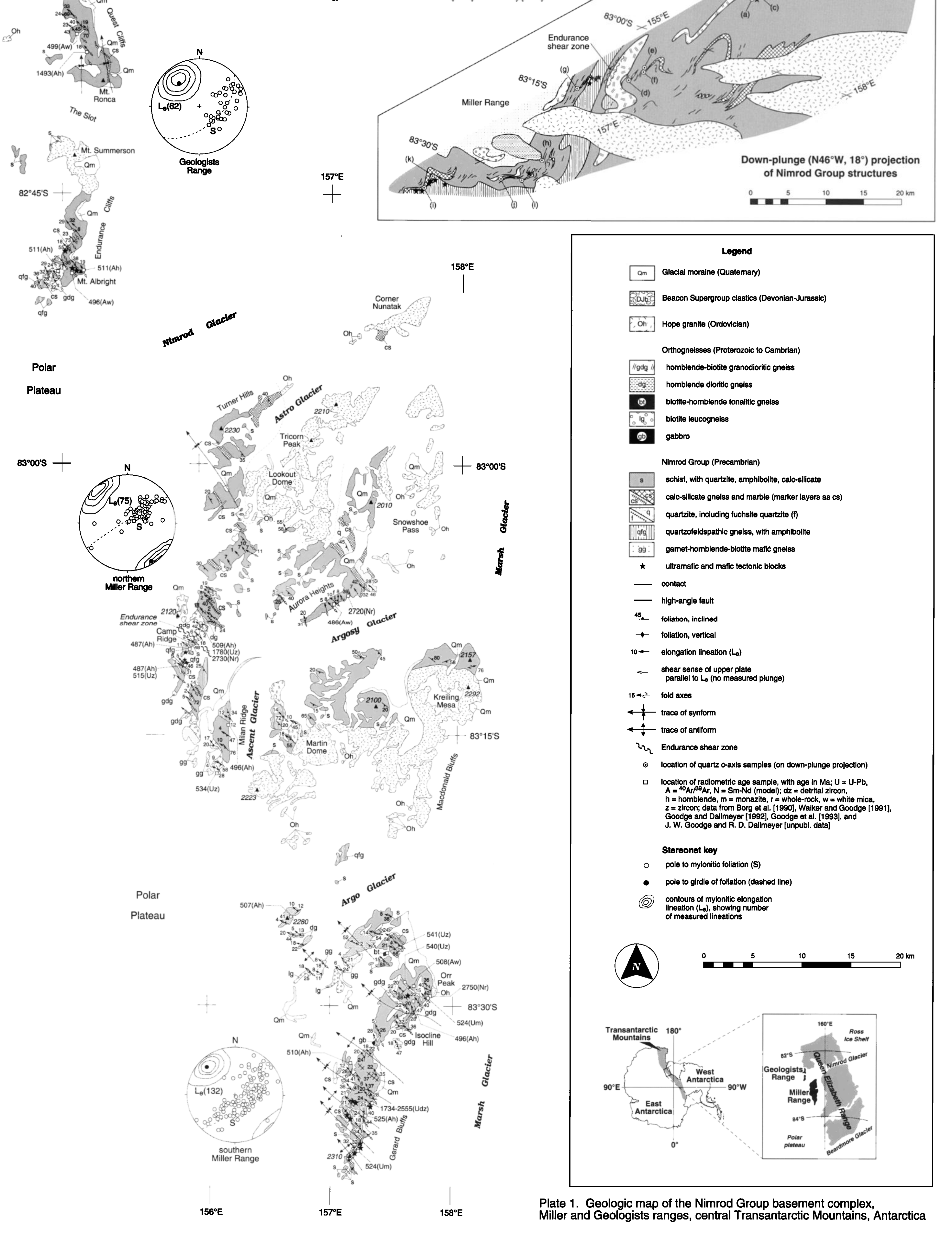

Plate 1. Geologic map of the Nimrod Group basement complex, Miller and Geologists ranges, central Transantarctic Mountains, Antarctica

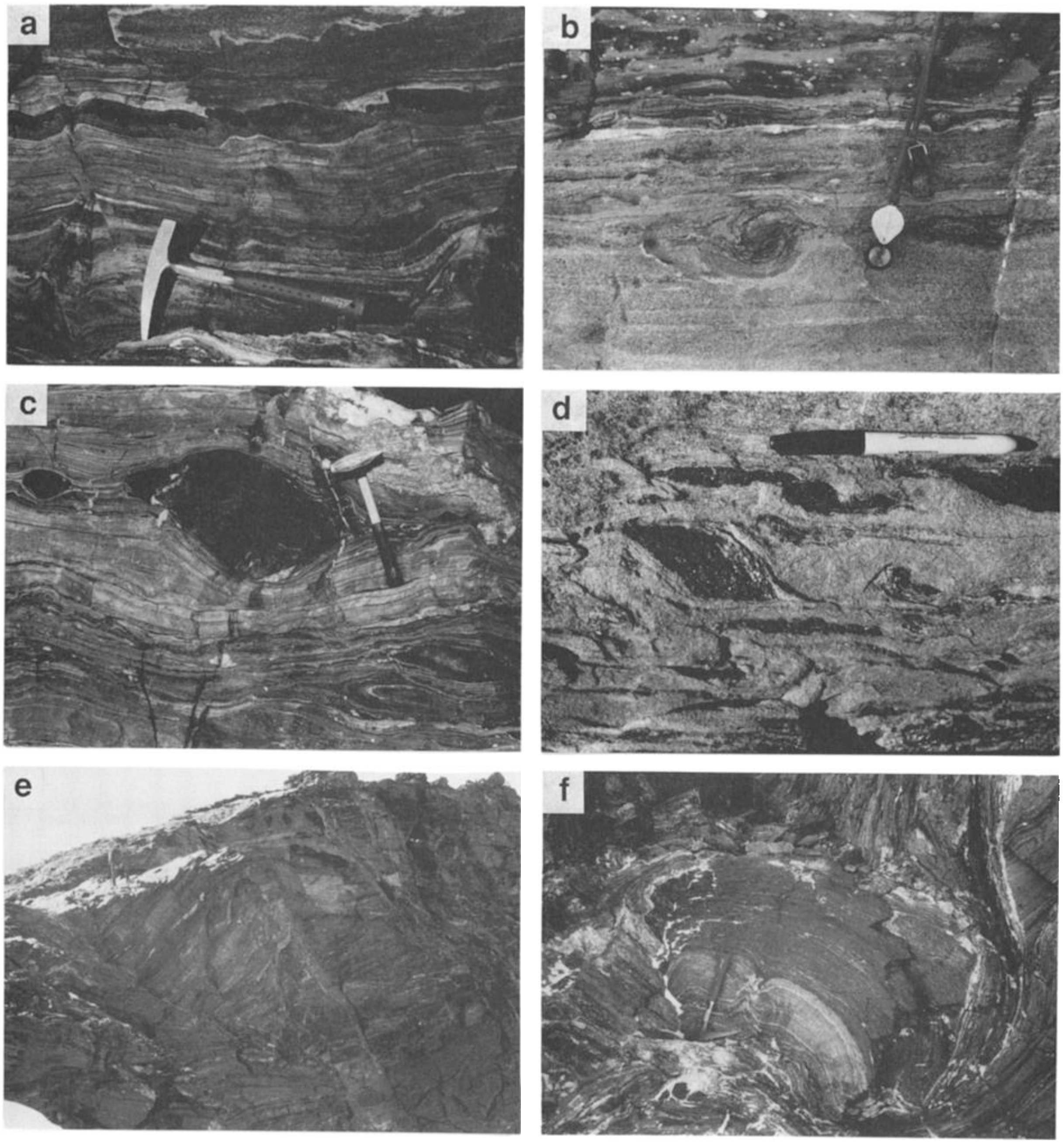

Fig. 2. Photographs of asymmetric motion-plane structures or fabrics that are used for determining shear sense within the MRSZ tectonites. All photographs viewed to SW, unless noted. (a) Tabular mafic-layer boudins showing tilt to form en echelon series resulting from progressive ductile shear after initial separation. Hammer is 33 cm long. (b) Detached tectonic inclusion of amphibolite in calc-silicate gneiss, showing sinistral rotation. Inclusion is 4 cm in diameter. (c) Mafic boudins within layered tectonites, showing detached and tailed form. These boudins contain relict eclogitic mineral assemblages. Hammer is 33 cm long. (d) Asymmetric foliation "fish" of amphibolite in light-colored calc-silicate host. Note planar tail truncations parallel to S. Pen is 14 cm long. (e) Outcrop of foliation phacoid (center) in which foliation and compositional layering are truncated along discrete shear surfaces at top and bottom. Person standing in upper left corner. View is to NW down plunge of L_e. (f) Outcrop showing elliptical block (center) that is foliated and boudinaged (light-colored layer), and which is enclosed by a second tectonite foliation that wraps around it. Hammer is 33 cm long. View is to NW down plunge of L_e.

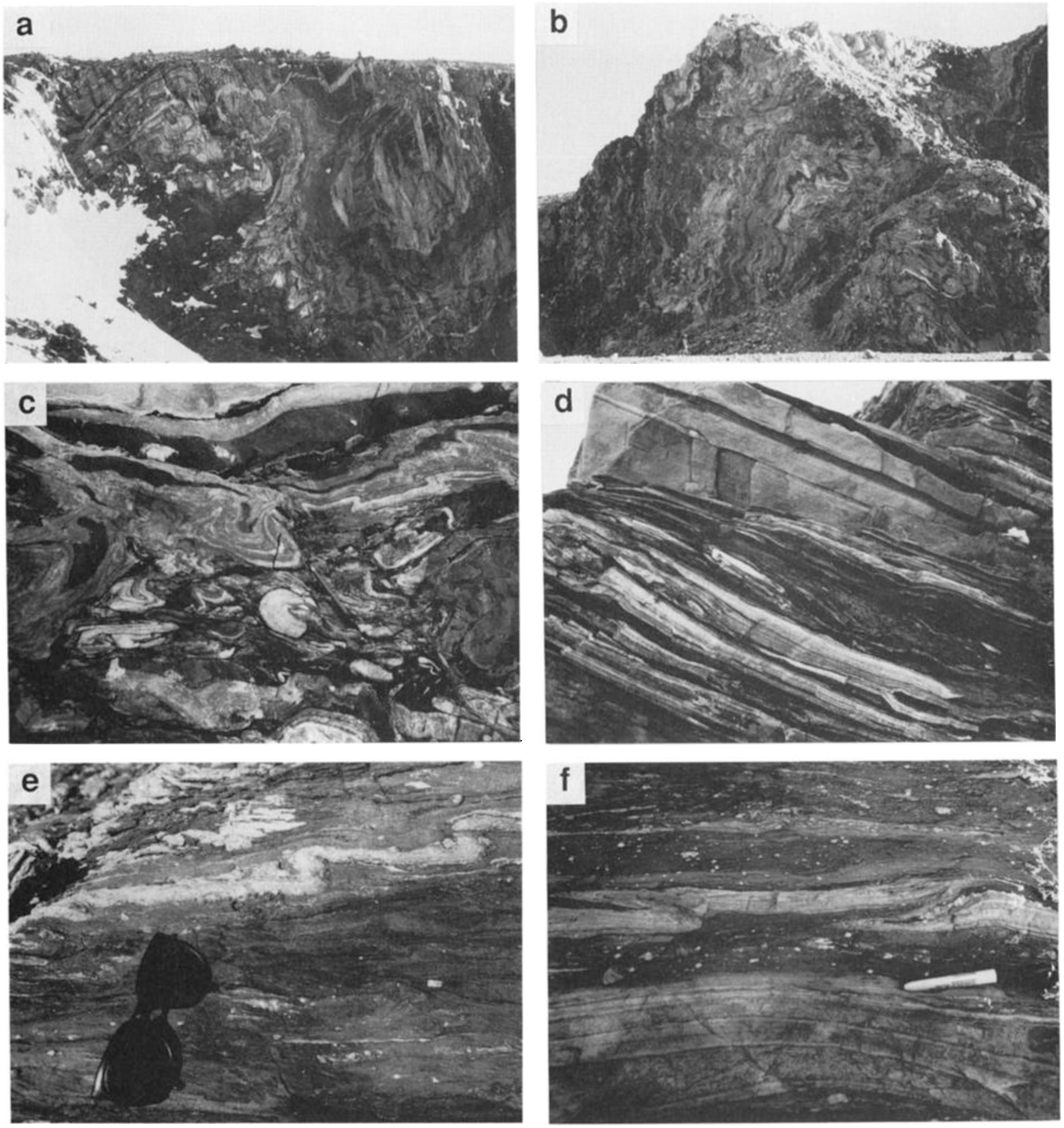

Fig. 3. Axial-profile views of folds in Nimrod tectonites. Type 1 folds shown in Figures 3a-3c, looking down the plunge of L_e to the NW. (a) Type 1 folds exposed along Endurance Cliffs in the Geologists Range, showing steep to moderately dipping axial planes and generally NE fold vergence. Cliff face is about 300 m high. (b) Type 1 folds exposed along Gerard Bluffs in the southern Miller Range. Folds in this cliff face are subordinate to larger NW plunging folds shown on Plate 1. Note steep to moderately dipping axial planes and SW fold vergence. Outcrop is about 200 m high. (c) Type 1 mesofolds in layered gneiss, showing variably oriented axial planes and no preferred fold vergence. Note structural complexity and layer discontinuity due to boudinage. Outcrop height is about 1 m. (d) Motion-plane view (looking SW) of outcrop shown in Figure 3c at same scale. Note extreme planar aspect to layering, general structural simplicity, and left-lateral (top-to-the-SE) fabric asymmetry in central dark layer. Outcrop height is about 1 m. (e) Motion-plane axial-profile view of type 2 mesofolds in layered schistose tectonites (looking SW). Axial planes are nearly parallel to S in tectonites, and folds show SE vergence. Sunglasses are 14 cm long. (f) Type 3 noncylindrical sheath folds viewed down tectonite L_e (to NW). Note near complete closure of fold limbs, tight to isoclinal interlimb angle, and foliation-parallel axial plane. Other type 3 folds show complete limb closure and SE vergent asymmetry.

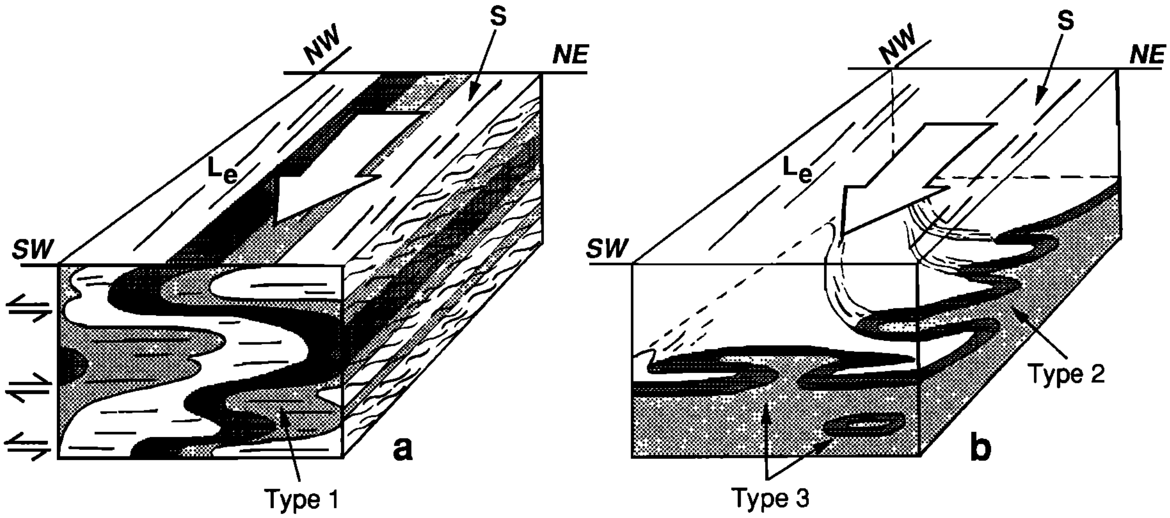

Fig. 4. Block diagrams summarizing different mesoscopic fold types relative to geometry of L-S tectonite fabrics. Open arrows show sense of ductile shear. (a) Open type 1 cylindrical folds with axes parallel to elongation lineation (L_e) that signify minor L_e-normal strain during dominant L_e-parallel flow. (b) Asymmetric type 2 cylindrical folds with axes normal to L_e, which may evolve to noncylindrical type 3 sheath folds with noses parallel to L_e at high strains.

noncoaxial ductile shear, and their asymmetry records shear-induced vorticity. Some folds of type 2 style are refolded by relatively younger type 2 folds within the Endurance shear zone. We interpret that the refolded type 2 folds reflect particularly high shear strains resulting from progressive SE directed flow.

Type 3 folds occur locally within the Endurance shear zone, and they are most common at the boundary between compositionally distinct units such as felsite and biotite gneiss, or marble and mica schist (Figure 3f). They are noncylindrical sheath folds [Quinquis et al., 1978; Cobbold and Quinquis, 1980], with axial planes parallel to S, isoclinal interlimb angles, and asymmetric profiles in sections parallel to L_e. Rarely observed profiles normal to L_e show limb closure; where exposure permits a three-dimensional view, L_e-parallel profiles record top-to-the-SE (left-lateral) ductile flow. These folds transpose compositional layering and S, and they probably record extreme ductile shear strain within the Endurance shear zone ($\gamma > 10$) [Cobbold and Quinquis, 1980].

Fold timing. Deformation of Nimrod metamorphic rocks produced a complicated set of structures, including not only the shear fabrics and folds described above, but also geometrically complex re-folded and/or rootless folds observed locally. These latter structures may have formed either prior to or during progressive ductile shear deformation, but we are unable to document an unambiguous record of preshear deformation. Likewise, some type 1 folds with tight interlimb angles could predate ductile shearing with their present geometry due to structural reorientation in the MRSZ. However, because of the nature of progressive ductile deformation, we cannot prove either that they are the product of an earlier tectonic event or that their present orientation within the MRSZ has any significance with respect to prior displacements.

Despite these ambiguities, the three types of folds described above are each interpreted to have formed during Nimrod ductile tectonism. Types 2 and 3 folds are themselves in kinematic agreement, and they are consistent with top-to-the-SE (left-lateral) displacement parallel to L_e. In addition, these folds display SE vergent tectonite fabrics on each limb as viewed within the motion plane. Therefore we interpret that these two fold types formed during ductile shear deformation. Folds such as these are common in shear zones [e.g., Cobbold and Quinquis, 1980; Hanmer and Passchier, 1991]. The proximity of types 2 and 3 folds to the Endurance shear zone provides evidence of relatively higher strain within this subzone of the MRSZ.

Type 1 folds, which fold compositional layering and, locally, S, show no preferred overall asymmetry and their fold axes consistently parallel L_e. L_e-parallel folds are common in large-scale shear zones [Christie, 1963; Eisbacher, 1970; Bell, 1978; Mattauer et al., 1981; Bell and Hammond, 1984; Sylvester and Janecky, 1988; Hansen, 1989], particularly those characterized by large bulk shear strains. If all type 1 folds had formed by rotation of preexisting folds during shear [e.g., Bryant and Reed, 1969; Bell, 1978], fold interlimb angles would be expected to tighten during rotation, resulting in tight to isoclinal colinear folds [e.g., Escher and Watterson, 1974; Bell, 1978; Skjernaa, 1980; Jamieson, 1987]. Most type 1 folds, however, exhibit open to close fold profiles; therefore these type 1 folds likely do not predate ductile shear deformation. Because S is locally deformed by type 1 folds, some must have formed after at least the earliest stages of noncoaxial shear. If the type 1 folds had formed after ductile shear ceased, opposing shear sense should be preserved on opposite fold limbs. However, opposite limbs in all cases that we examined record the same top-to-the-SE shear. Therefore type 1 folds must have formed during Nimrod tectonism with their axes initially parallel to L_e. Axes of type 1 folds, like L_e, thus approximately represent the maximum elongation axis of finite strain.

Type 1 folds correspond geometrically and descriptively to

Grindley's [1972] early phase (F_1 and F_2) folds, which he attributed to NE displacement along the Endurance thrust. From the distribution of these folds throughout the MRSZ, well away from the Endurance thrust as mapped by Grindley [1972], and the general lack of systematic vergence, we interpret that the type 1 folds did not form in response to major NE directed displacement. Rather, they formed as a result of regional NE-SW constriction perpendicular to the direction of penetrative translational shear, perhaps accommodated by movement along minor shears normal to L_e (Figure 4a). Furthermore, it is important to emphasize that type 1 fold axial profiles are viewed within the plane normal to tectonite L_e and not within the motion plane; hence these folds do not record shear sense.

Further evidence for establishing the nature and timing of fold formation comes from examination of widespread pegmatitic dikes and sills. These intrusive rocks, typically containing quartz + K-feldspar + biotite ± garnet ± tourmaline, are common within orthogneisses, schists, and gneisses. They are particularly abundant near zones of migmatite and relatively isotropic granite, but they are also common as swarms within the metamorphic tectonites, both concordant and discordant to S. Dikes and sills range in size but most commonly are <1 m wide. Pegmatitic dikes and sills show a textural range from unmodified igneous textures to strong L-S tectonite fabrics concordant to wall-rock fabrics. In some locations, dikes show successive stages of folding and boudinage, in which early formed dikes that have been folded and dismembered are crosscut by younger, generally planar dikes that are subparallel to the youngest shear foliations. The folded dikes are geometrically and stylistically similar to the type 1 folds observed in layered tectonites, as shown by fold-axis parallelism with L_e, a lack of asymmetry, and open interlimb angles. We consider all of these intrusive types to be generally synkinematic, although some of the most strongly deformed examples could predate ductile deformation by a considerable period. These relationships are evidence of multiple injections of pegmatitic igneous melts during folding and ductile shear deformation within the MRSZ.

Microstructural Fabrics

Microscopic examination of Nimrod tectonite fabrics confirms the left-lateral displacement sense interpreted in the field throughout the MRSZ. The form and orientation of mica, kyanite, and amphibole "fish" in type II S-C mylonitic quartzite, pelite, and amphibolite provide evidence of preannealing, noncoaxial, left-lateral deformation.

Intragranular and intergranular muscovite crystals in (garnet-)mica quartzite are needle-shaped with their long axes parallel to L_e, or parallel to the interpreted direction of maximum finite strain (X; Figures 5a and 5b). Despite the extreme aspect ratio (e.g., x:y:z = 30:3:1) of their grain shapes, locally preserved populations of asymmetric mica "fish" record consistent left-lateral shear sense. Preferred mineral grain shape is also shown by garnet observed in garnet-biotite quartzite (Figure 5c). In these rocks, rounded garnets show aspect ratios of 1.5:1:1 to 3:1:1. Within the motion plane, the long axes of the garnets consistently plunge NW relative to S, and they record a top-to-the-SE direction of shear. Asymmetric biotite pressure shadows on garnet reflect similar shear sense in these quartzites, as do quartz c axis orientations (see below).

Types I and II S-C fabrics are interpreted with confidence in thin section (Figures 5d to 5f). Kyanite "fish" in pelitic tectonites commonly show tapered ends that terminate along C planes, and locally they are bent, show undulose extinction, and are segmented by cross-grain fractures (Figure 5d). These textures indicate prekinematic to synkinematic growth of kyanite. Sillimanite needles or bundles are bent into parallelism with C planes (Figure 5e). Sillimanite generally lacks undulatory extinction and broken fibers, indicating that sillimanite growth was synkinematic to postkinematic. Sillimanite was never observed overprinting asymmetric fabric elements; however, sillimanite commonly replaces kyanite. Thus sillimanite growth was probably late synkinematic. Small and large garnets alike exhibit ellipsoidal shapes with long axes inclined, together with pressure shadows, in the direction of shear as in Figures 5c and 5e. Kyanite and muscovite "fish," and type II S-C fabrics in garnet-kyanite-muscovite pelitic schist, provide evidence that high-grade metamorphic conditions accompanied the strong noncoaxial ductile deformation (Figures 5d to 5f). The synkinematic assemblage kyanite + muscovite may exist at T = 700°C only at moderate to high P; hence muscovite remained stable even at the high temperatures of deformation that at lower P might induce reaction to anhydrous phases. Despite well-preserved shear-sense indicators, all samples observed in thin section lack strong quartz grain shape-preferred orientation. Quartz, chiefly preserved as equidimensional polygonal grains with straight to coarsely serrated grain boundaries, generally displays flat-field extinction (e.g., Figures 5a and 5b).

Quartz c Axis Fabrics

We analyzed quartz c axis fabrics in 12 specimens of quartzite selected from different structural levels throughout the MRSZ (Plate 1 and Figure 6). Samples were cut in two perpendicular sections, parallel to the motion plane and normal to S and L_e (the XZ and YZ planes, respectively). Quartz c axis orientations were measured optically using the universal stage [Turner and Weiss, 1963]. Data measured in the YZ plane were rotated into the XZ plane and are shown together in lower-hemisphere projections with the XZ plane as the perimeter great circle (Figure 6).

MRSZ c axis fabrics indicate that L_e is indeed an elongation lineation (i.e., that it is parallel to X) because the fabric girdle for each sample intersects S normal to L_e [Behrmann and Platt, 1982]. Furthermore, quartzite c axis fabrics show well-defined asymmetric single girdles which confirm that noncoaxial ductile flow was in a top-to-the-SE, or left-lateral, direction [e.g., Lister and Williams, 1979; Schmid and Casey, 1986].

In addition to kinematic information, quartz c axis fabrics also provide a means to qualitatively evaluate relative thermal conditions and finite strain within the MRSZ. Quartz c axis patterns are a function of intracrystalline slip, temperature, water content, strain rate, recrystallization rate, and deformation path [Tullis et al., 1973; Nicolas and Poirer, 1976; Lister and Hobbs, 1980; Hobbs, 1985; Law, 1990]. Asymmetric single-girdle fabrics can form as the result of an increasing noncoaxial component of strain, or as a result of increasing strain during simple shear [Schmid and Casey, 1986]. Within the MRSZ, the only indication of prominently higher relative strain is within the Endurance shear zone as marked by the presence of sheath folds, ultramylonite, and extremely attenuated asymmetric fold profiles. Elsewhere, S

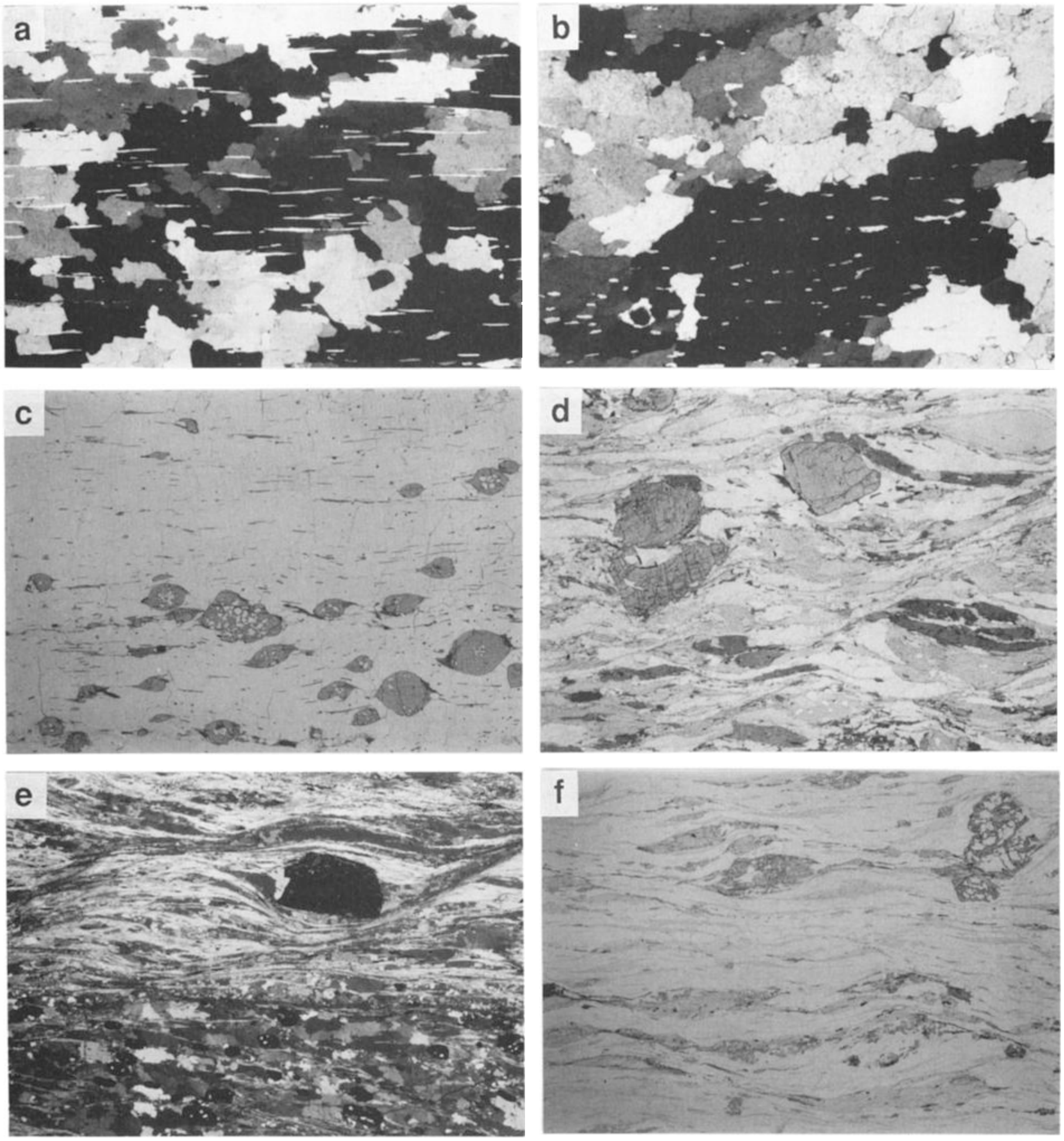

Fig. 5. Photographs of Nimrod tectonite microstructures. All photographs are parallel to the motion plane (*XZ*) unless noted. (a) (Garnet-) muscovite quartzite (89GGR14D; crossed nicols; 8 mm wide), with muscovite grain shape showing extreme aspect ratio (up to 30:1) with long axis parallel to S, and intragranular nature of muscovite in quartz. Quartz exhibits flat-field extinction and relatively straight, locally polygonal, grain boundaries. Rare very fine-grained muscovite "fish" preserved in lower right corner show dextral (top-SE) shear as viewed. Quartz *c* axis fabric diagram for this sample shown in Figure 6a. (b) View of quartzite shown in Figure 5a looking down L_e in section perpendicular to motion-plane (*XY*). Crossed-nicols; 5 mm wide. Note smaller aspect ratio of muscovite (2:1 to 5:1) than that shown in motion-plane section. (c) Garnet-biotite quartzite with asymmetric biotite pressure shadows on ellipsoidal garnets. Plane-polarized light; 10 mm wide. Garnets have long axes inclined relative to S in the direction of shear (dextral as viewed), and they are circular in the section normal to the motion plane and L_e. Quartz grains are equidimensional polygons with flat-field extinction in crossed-nicols view. Biotite is preserved in both intragranular and intergranular positions. (d) Garnet-kyanite-muscovite schist showing left-lateral type II S-C fabrics and kyanite crystals (dark gray) that are tapered, bent, and segmented. Plane-polarized light; 16 mm wide. (e) Garnet-sillimanite-muscovite schist. Sinistral shear sense indicators include grain shape-preferred orientation of ellipsoidal garnets (black), garnet pressure shadows of quartz, and type II S-C fabrics, deforming sillimanite. Crossed nicols view; 16 mm wide. (f) Garnet-kyanite-muscovite schist showing kyanite and muscovite "fish" and type II S-C fabrics. Shear sense is dextral as viewed. Plane-polarized light; 16 mm wide.

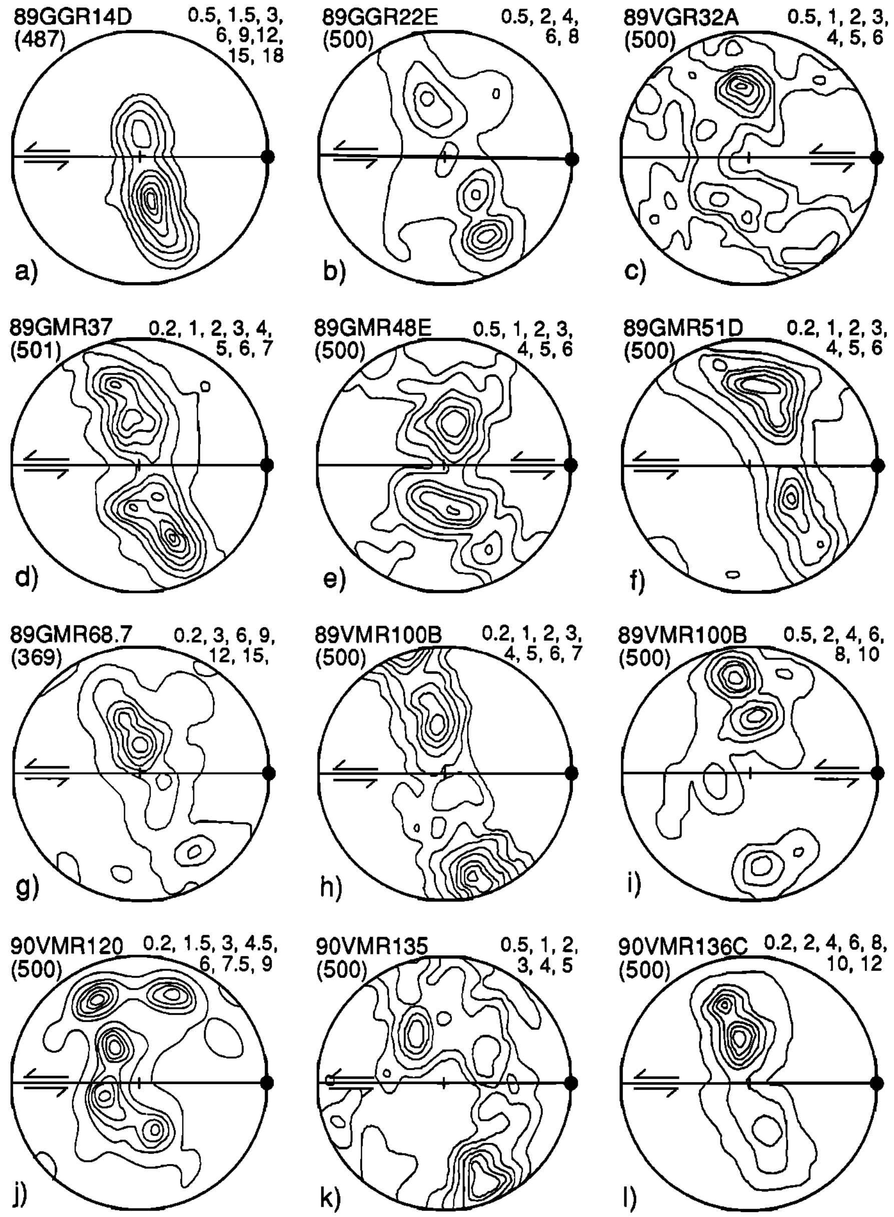

Fig. 6. Lower-hemisphere, equal-area stereonet projections of quartz *c* axis fabrics from selected quartzite tectonites sampled at different structural levels in the Geologists and Miller ranges (locations shown on Plate 1). Each diagram shows combined *c* axes measured from both the motion plane (*XZ*), and the plane normal to the motion plane and L_e (*YZ*). Number of total *c* axes measured in parentheses. Density contours indicated by the numbers to the upper right of each plot, in percent. Fabric diagrams shown in a structural reference frame with respect to S (horizontal line) and L_e (dot; downplunge direction). Paired arrows indicate shear sense determined from the inclination of the *c* axis girdle to S. Diagrams plotted using a spherical Gaussian function.

and L_e appear equally well developed, and on this basis, we interpret no obvious variation in strain intensity within the bulk of the MRSZ. However, several quartzite samples display equally well-developed single-girdle *c* axis fabrics despite their relative proximity to, or distance from, the Endurance shear zone (Figures 6a, 6b, 6d, 6f, 6g, 6h, 6i and 6k). By comparison with simulated fabric diagrams [Jessel and Lister, 1990], we interpret these single-girdle fabrics as the result of high-strain simple shear ($\gamma \geq 5$) under medium- to high-temperature conditions. Medium- to high-temperature simple shear is consistent with the simultaneous development of muscovite and kyanite "fish," and with estimates of temperature during Nimrod tectonite formation [Goodge et al., 1992].

The activity of different slip systems in quartz during ductile tectonism is principally related to temperature, strain rate, and water content [Hobbs, 1985]. The dominant single-girdle *c* axis patterns in MRSZ quartzites are evidence of principal deformation by rhombohedral and/or prismatic slip [Schmid and Casey, 1986]. Many MRSZ quartz *c* axis fabrics also show a lack of poles along or near the great circle at a small angle to Z (Figures 6a, 6b, 6d, 6f, 6g, 6j and 6l). An absence of poles of this orientation indicates a lack of basal-plane slip [Schmid and Casey, 1986], which occurs principally at moderate to low temperatures of deformation [Tullis, 1977; Jessel and Lister, 1990]. The lack of evident basal-plane slip is consistent with the interpretation of high-T deformation, but it also implies that Nimrod tectonites did not experience intracrystalline strain at lower subpeak temperatures (i.e., during thermal retrogression). This provides indirect, yet important, evidence that crystal-plastic flow ended while the Nimrod tectonites were still at high temperatures. These relative constraints on the kinematic and cooling histories of the Nimrod tectonites are valuable for deciphering possible tectonic paths of deformation, as discussed below.

Quartz within each of the measured samples occurs as polygonal grains that lack grain shape-preferred orientation and show chiefly flat-field extinction (e.g., Figures 5a and b), and intragranular muscovite "fish" are common. However, muscovite grains probably pinned quartz grain boundaries during deformation, and they were most likely intergranular during ductile flow. The preserved quartz textures may be the result of movement of grain dislocations, and grain and subgrain boundaries, resulting in annealing of petrographic textures. Annealing probably postdated ductile deformation because quartz grains lack grain shape-preferred orientation. Therefore we suggest that quartz *c* axis fabrics record the ductile shear, whereas quartz petrographic textures resulted from postkinematic thermal annealing. This interpretation would explain the polygonal quartz grain boundaries, flat-field extinction, a lack of grain shape-preferred orientation, and only minor subgrain formation, as well as intragranular muscovite, and it is consistent with the lack of evidence for intracrystalline slip along quartz basal planes. Thus cooling appears to have followed, rather than accompanied, deformation.

Temporal Relations

The absolute timing of ductile deformation is constrained by mineral geochronometer data (available age data from Nimrod units are shown in Plate 1). Goodge et al. [1993] described several Nimrod orthogneiss units that exhibit variably formed ductile deformation fabrics, ranging from orthogneiss bodies displaying well-formed L-S fabrics kinematically consistent with their host tectonites, to weakly deformed orthogneiss bodies expressing only linear or planar fabrics, to pegmatite sills containing no solid-state ductile deformation fabrics. U-Pb age data obtained from igneous zircons in these rocks indicate a period of progressively waning ductile deformation between about 540 and 520 Ma. Kyanite-zone pelitic schists containing well-formed L-S fabrics contain metamorphic monazites that yielded U-Pb ages of ~524 Ma. These age data are consistent with petrologic and geothermometric constraints indicating that Nimrod ductile deformation occurred at high temperatures (~700°C). The $^{40}Ar/^{39}Ar$ cooling ages of 525-487 Ma that were obtained from hornblende and muscovite [Goodge and Dallmeyer, 1992], earlier interpreted as evidence of magmatic heating by posttectonic plutons, probably record cooling following high-temperature ductile deformation and syntectonic to posttectonic magmatism. Thus ductile deformation was diminishing during the interval between about 540 to 520 Ma, although when this deformation began is unknown. Deformation in this interval during the Early Cambrian (based on new criteria for the Precambrian-Cambrian boundary) [Compston et al., 1992; Cooper et al., 1992] is broadly synchronous with contractional deformation of Byrd Group supracrustal rocks to the east. Tectonic implications of this temporal overlap are discussed below.

Strain Rate

A uniform L_e orientation and widespread, consistent shear sense indicate that ductile deformation in the MRSZ resulted from a single, kinematically simple, top-to-the-SE shear event. General shear strains within the MRSZ were probably moderate to high ($\gamma \geq 5$), and locally very high ($\gamma \geq 10$), as indicated from structures such as type 3 sheath folds, extreme fold attenuation, ultramylonite, δ porphyroclasts, and quartz *c* axis fabrics. High relative strain rates may also be qualitatively inferred. MRSZ fabrics comprise a thick zone with great geometric uniformity and consistency of asymmetric microstructures formed during high-temperature ductile deformation, in places associated with synkinematic magmatism. Synkinematic mineral assemblages indicate that upper amphibolite to granulite facies conditions (T ≈ 700°C) accompanied ductile deformation [Goodge et al., 1992]. The preservation of well-developed microstructural ductile shear fabrics requires that the ratio of strain rate to mineral recrystallization rate be high, where recrystallization rate is primarily a function of temperature. Nimrod tectonites formed at temperatures sufficient to promote rapid mineral recrystallization. In order to preserve a microstructural record of noncoaxial shear, high-T shear zones must undergo strain at a faster rate than their lower-T counterparts; if not, such zones would likely not preserve asymmetric microstructures. Thus the well-preserved, high-T microstructures described here are compatible with high strain rates.

DISCUSSION

To summarize structural relations within the MRSZ, Nimrod Group tectonites were ductilely deformed by dominantly noncoaxial shear within a broad zone (12-15 km wide) under moderate-P, high-T conditions in the upper amphibolite to lower granulite facies. Although the boundaries of this zone are not observed, its exposed structural width is comparable with that of the Grenville Front Tectonic Zone [Davidson, 1984] and the Great Slave Lake Shear Zone [Hanmer, 1988]. L-S tectonite fabrics are pervasive, and only minor strain gradients are apparent. Relative displacement across the zone was left-lateral, or top-to-the-SE, accompanied by local perpendicular constrictional strain (Figure 7). On the basis of well-formed microstructures and quartz *c* axis fabrics, we estimate that total strains were high, generally of the order of $\gamma \geq 5$, with $\gamma \geq 10$ locally within the Endurance shear zone, as shown by ultramylonite, sheath folds, and δ porphyroclasts. High total strains, fabric uniformity, consistent shear sense, and high synkinematic temperatures collectively argue for high strain rates within the MRSZ. Microstructural and quartz crystallographic fabrics, with little evidence of quartz basal-plane slip, indicate that crystal-plastic flow within the MRSZ

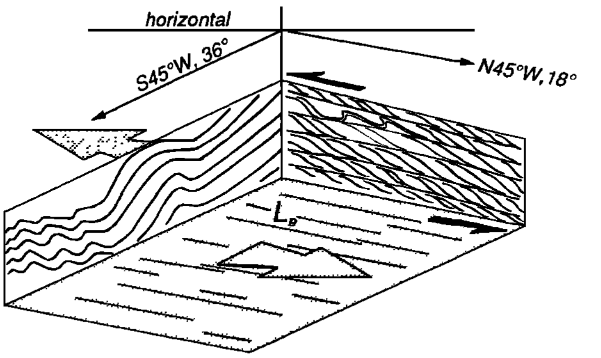

Fig. 7. Composite block diagram of Nimrod structural elements. View is to SW, looking in dip direction of S (broad block faces on top and bottom). L_e is consistently oriented SE-NW, and face oriented N45°W is the motion plane. Simplified downplunge projection shown on Plate 1 is viewed looking at the face oriented S45°W. Shear sense is consistently top-to-the-SE (black arrows on motion plane), giving a general direction of ductile flow shown by large gray arrows. Also shown are types 1 and 2 folds.

did not take place during progressive cooling; rather, the tectonites cooled to temperatures representative of the greenschist facies only after bulk ductile displacement within the zone had ceased. This cooling pattern is verified by thermochronologic data, which indicate that high-temperature Nimrod ductile tectonism waned between about 540 to 520 Ma [Goodge et al., 1993] but that postkinematic cooling to >350°C occurred by ~485 Ma [Goodge and Dallmeyer, 1992].

The large scale, high strain, and inferred deep-crustal conditions of the MRSZ indicate that it represents a major crustal structure. This zone may have overprinted a preexisting crustal boundary, as expressed by the presence of mafic and ultramafic tectonic blocks distributed within it. Mafic blocks with relict eclogite facies parageneses [Goodge et al., 1992] and ultramafic bodies containing high-temperature anthophyllite-talc assemblages [Goodge et al., 1990] represent entrained lower-crustal or upper-mantle fragments within the Nimrod Group. High-pressure metamorphism of these blocks (12-25 kbar) [Goodge et al., 1992] probably predates ductile flow within the MRSZ and may signify an earlier collisional event at substantially deeper crustal levels. These blocks may have initially been tectonically incorporated in their hosts within either a Franciscan-type subduction zone setting or along an Alpine-type collisional suture belt. Whatever their origin, the high-pressure tectonic blocks may mark an ancient crustal-penetrating suture that was reactivated as an intraplate structure by the MRSZ. The relative timing between such events is completely unconstrained at present. It is possible that an event which led to entrainment of mafic and ultramafic rocks is substantially older than ductile deformation within the MRSZ, or these events may reflect different stages between early collision and subsequent shear. Despite remaining ambiguities, the scale and geometry of the MRSZ in its larger structural context are well illustrated in the downplunge projection (mean L_e) shown on Plate 1. This projection shows several key features of the MRSZ:

1. Lithologic layering and foliation of MRSZ tectonites have an apparent dip to the SW (toward the polar plateau). Although major displacements within the MRSZ were normal to the downplunge projection, a constrictional component of strain is apparent from the large-scale type 1 folds.

2. Mafic and ultramafic tectonic blocks (marked by stars) are concentrated along the boundary between schist and layered gneiss units, which in the western Miller Range is generally coincident with the Endurance shear zone. The blocks are out of petrologic equilibrium with their host tectonites, within which they are boudinaged as relatively rigid bodies. The presence of these "exotic" blocks suggests that the Nimrod Group tectonites may mark a fundamental tectonic boundary, although the origin of this boundary may predate formation of the MRSZ.

3. Pretectonic to syntectonic plutonic bodies, generally small and conformable with tectonite layering, and posttectonic plutons, exhibiting subhorizontal lenticular shapes, are readily distinguished. Older plutonic bodies are commonly distended and folded (many occur as layer-parallel bodies within gneissic units, too small to show at the scale of Plate 1), whereas posttectonic bodies lacking deformation features clearly crosscut all metamorphic units as their ascent through the crust was arrested. The downplunge projection thus not only offers a revealing view of MRSZ structures, but it provides a window into the middle to lower crust of the Antarctic plate-boundary regime during the Neoproterozoic-Cambrian transition which contains elements of a collisional or subduction-related magmatic-arc system.

Isotopic compositions of posttectonic plutons reinforce the concept that the MRSZ may have reactivated a crustal-scale boundary, in that these plutons are isotopically distinct from the Nimrod tectonites they intrude [Borg et al., 1990]. These ~500 Ma plutons have average ε_{Nd} values of -11 and yield Sm-Nd depleted-mantle model ages (T_{Dm}) of about 2.0 Ga, considerably different from any Nimrod lithologies (ε_{Nd} = -21 to -24 and $T_{Dm} \approx 2.7$ Ga). This contrast in isotopic composition appears to preclude the presence of significant Nimrod Group melt components in the ~500 Ma granites. Conversely, a ~1.7 Ga orthogneiss near Camp Ridge (Plate 1) has a Sm-Nd model age similar to other Nimrod units (T_{Dm} = 2.73 Ga) [Borg et al., 1990]. The regional dip of MRSZ tectonites to the SW, the absence of ductile tectonites in any of the outboard sedimentary assemblages, and the inferred presence of isotopically distinct source materials beneath the Nimrod tectonites all suggest that the MRSZ represents a major crustal structure dipping toward the craton that juxtaposes reworked cratonic basement materials and outboard supracrustal assemblages. It must be emphasized, however, that no constituents of known supracrustal assemblages (Beardmore and Byrd groups) are present within the MRSZ, as recognized either on lithologic, metamorphic, structural, or isotopic grounds. The geometry and strike-slip kinematics of the MRSZ may therefore reflect accommodation of principally margin-parallel motion along an older lithospheric boundary.

In order to understand the tectonic regime in which the MRSZ formed, it is necessary to recall deformation patterns in the outlying supracrustal rocks. Regionally, rocks of the Byrd and Beardmore groups are affected by open to tight folds that parallel the present Transantarctic Mountains [Grindley and Laird, 1969; Oliver, 1972; Stump, 1981; Stump et al., 1986; Rees et al., 1987], although the specific ages of these structures are uncertain in most cases. In the vicinity of Nimrod glacier, Byrd Group rocks are deformed by upright NNW-NW trending folds with subhorizontal axes [Grindley and Laird, 1969; Laird et al., 1971; Rees et al., 1987], the hallmark signature of contractional Ross deformation in this area. In more detail, however, Byrd Group rocks appear to be affected by multiple deformations [Rees et al., 1987, 1989; Rowell et al., 1992], including several generations of tilting, folding, and thrust-fault displacement. Beardmore Group rocks are deformed by broad folds that parallel regional Ross structural trends [Grindley and Laird, 1969; Laird et al., 1971; Stump et al., 1991]; however, they reportedly underlie an unconformity at the base of Lower Cambrian Byrd Group strata, indicating an earlier (Beardmore) deformation of pre-Early Cambrian age [Laird et al., 1971; Stump et al., 1991]. Structural data from Beardmore rocks in the Cobham Range and at Kon-Tiki Nunatak (Figure 1) [Goodge et al., 1991b] indicate that deformation there was west vergent, resulting in large-scale folds, flexural-slip mesofolds, en echelon tension gashes, and a steep east dipping cleavage. These relations, and similar observations reported by others [Laird et al., 1971; Edgerton, 1987; Stump et al., 1991], suggest that the latest major supracrustal deformation involved west directed shortening. Inadequate exposure, poor biostratigraphic control, and difficult access make it impossible at this time to resolve the vergence of other latest Proterozoic to early Paleozoic deformations within the supracrustal sequences; until further study is completed, it may only be stated with assurance that these rocks were involved in orogen-normal contraction. The Early Cambrian along-strike displacements we have documented for the MRSZ are therefore at a high angle to the principal directions of latest Proterozoic to Early Ordovician orogen-normal shortening recorded in outboard supracrustal rocks.

Based on structural relations of the MRSZ and the supracrustal rocks described above, we propose a tectonic model that may serve to explain the relation between generally synchronous but stylistically distinct deformations. Because of many uncertainties, several assumptions must first be stated. First, because crystalline metamorphic rocks of the Nimrod Group are nowhere in exposed contact with the outboard supracrustal sequences, the plausibility of our tectonic model rests on the temporal overlap between ductile MRSZ deformation and supracrustal shortening [Goodge et al., 1993], rather than on direct structural observation. Second, the Transantarctic Mountains have undergone only minor nonvertical tectonic modification since the Devonian; the modern Transantarctic Mountains front is the result of differential uplift along orogen-parallel post-Triassic normal faults, as shown by subhorizontal Beacon Group strata overlying the Kukri peneplain [Grindley and Laird, 1969; Laird et al., 1971]. From this, we infer that the orientations of tectonite fabrics within the MRSZ, which are as young as early Middle Cambrian in age, are not greatly different from when they formed. Third, the general tectonic setting for these deformations is one of a convergent plate margin in which oceanic lithosphere of the proto-Pacific ocean was subducted beneath the East Antarctic continent [Elliot, 1975; Borg et al., 1987, 1990]. It is possible, however, that such a margin did not have a simple configuration as in the case of the modern Andean margin of South America but perhaps was more complicated, involving fringing volcanic arcs and multiple subduction zones as along the modern Pacific-Eurasian plate boundary.

Given these relations, we argue that the late Neoproterozoic to early Paleozoic tectonic setting of the central Transantarctic Mountains was one of left-oblique subduction beneath the East Antarctic craton (Figure 8). In this model, strain partitioning accommodated contemporaneous deformation of basement and supracrustal sequences [e.g., Fitch, 1972; Jarrard, 1986], in which strike-parallel ductile deformation within the MRSZ at deep crustal levels occurred at the same time as high-level shortening of the supracrustal Beardmore and Byrd sequences. Ductile tectonites within the MRSZ represent the diffuse equivalents of discrete, higher level strike-slip faults within the upper plate, perhaps transecting the magmatic arc represented by syntectonic to posttectonic Granite Harbour plutons. The orientation of MRSZ tectonites suggests that the strike-slip faults are listric toward the craton. Orogen-parallel displacements within the MRSZ are consistent with a lack of pronounced metamorphic gradients across the zone, as would be expected in the case of a large-scale, purely dip-slip collisional suture. Shortening within the supracrustal sequences probably occurred within a forearc setting, where both west and east vergent thrust displacements are possible. An active forearc environment with syntectonic conglomeratic and siliciclastic sediment deposition is a plausible setting for at least part of the Byrd Group. The outer accretionary prism of such a convergent-margin system has not been identified, either because it has yet to be recognized or because of subsequent truncation and translation. An obliquely subducting plate margin in which upper-plate transcurrent faults pass downward into distributed zones of ductile flow could help to explain isotopic patterns observed in the ~500 Ma posttectonic plutons [Borg et al., 1990]. If the MRSZ represents a large-magnitude, strike-slip structural boundary within the upper plate, it may be possible that granitic magmas generated by subduction would carry an isotopic signature of more outboard, partly subducted materials. Granitic rocks in this case were apparently derived from younger materials that may presently underlie the Nimrod Group.

An obliquely subducting margin requires a low angle of subduction (≤25°) in order to generate the plate coupling forces that lead to development of strain partitioning within the upper plate [Jarrard, 1986]. In addition, generation of granitoid magmas along this margin during and after orogenic deformation suggests an arc-trench distance >200 km, based on simple geometric construction; distances of this magnitude or greater are observed in modern margins characterized by oblique subduction (e.g., Sumatran margin of Eurasia). Those parts of the orogen exposed in the present-day Transantarctic Mountains, although complicated by younger extension, are greater than 200 km in width; thus the breadth of this orogen is compatible with generally shallow subduction and magma generation. Ductile tectonite fabrics within the MRSZ reflect a clear component of shear subparallel to the orogen that is nearly orthogonal to the direction of maximum shortening in supracrustal rocks. Furthermore, this ductile deformation occurred at significant crustal depths (≥26 km). If the depth of Nimrod deformation was actually below that at which strain is effectively partitioned, depending on relative plate velocities, temperature, and geometry, it is possible that the SE directed (mean 134° in present-day coordinates) displacements recorded by ductile strain parameters in Nimrod tectonites in fact preserve the true plate convergence direction, rather than a partitioned component. Whichever case is more likely, there remains plausible evidence for partitioned deformation at different crustal levels within the upper plate of an oblique

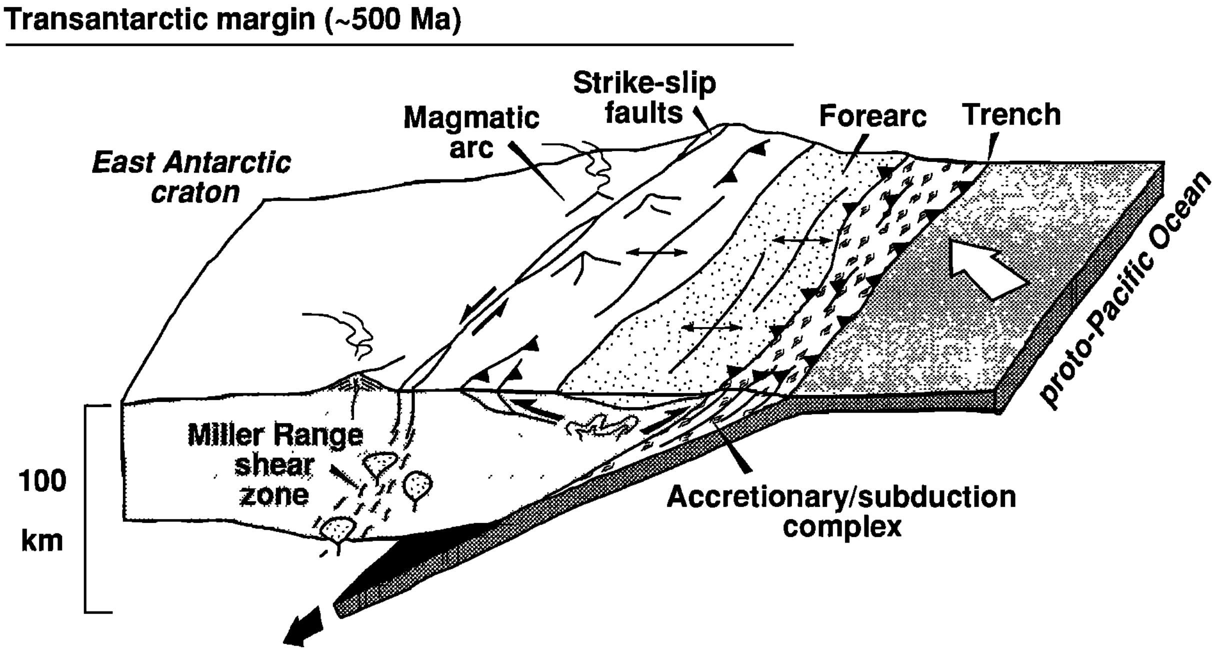

Fig. 8. Schematic diagram of Transantarctic Mountains plate-boundary regime at ~500 Ma. In this model, relative plate motions involving left-oblique subduction between the East Antarctic craton and proto-Pacific oceanic lithosphere is partitioned into both contractional and strike-slip components of deformation. Inboard strike-slip faults that become shallow with depth away from the subducting plate diffuse into broad zones of ductile shear within the root zone of a magmatic arc represented at depth by the Granite Harbour intrusives. The Miller Range shear zone formed as a result of strike-parallel ductile displacements at depths >25 km. Deformation in the forearc region occurred by punctuated contraction of supracrustal sedimentary rocks, including possible Neoproterozoic deformation of rift-facies Beardmore Group clastics and Middle Cambrian to Early Ordovician fold and thrust deformation of pretectonic to syntectonic Byrd Group sediments. Rocks of an accretionary or subduction complex are not preserved.

convergent plate boundary.

Rowell and Rees [1990] and Rowell et al. [1992] used sedimentological evidence to argue that parautochthonous Byrd Group sediments accumulated along an active Early Cambrian plate margin of East Antarctica. Carbonate units contain features indicative of platform deposition, but siliciclastic and conglomeratic facies may have been deposited in structurally controlled basins distributed along a transform (or transpressive) margin. Rowell et al. [1992] defined two Cambrian sedimentary belts ("marginal cratonic belt" and "Queen Maud terrane") that are subparallel to the present trend of the Transantarctic Mountains and that coincide with crustal age provinces defined by Borg et al. [1990]. Independent evidence that Early Cambrian deposition of the marginal cratonic belt occurred in a tectonically active strike-slip or transpressive environment is thus supportive evidence that deformation in the Transantarctic Mountains occurred in an oblique-subduction setting. Although Weaver et al. [1984] postulated early Paleozoic strike-slip displacements for northern Victoria Land, the kinematics of ductile deformation within the MRSZ provide the first evidence for a relative sense of motion along this plate boundary, that of left slip. Finally, left-oblique motion along this margin is compatible with, but not confirmed by, a Cambrian plate reconstruction [Dalziel, 1992] in which Neoproterozoic rift separation of East Antarctica from another continent (perhaps Laurentia) initiated proto-Pacific ocean basin spreading along an axis at an angle to the East Antarctic margin that may have led to oblique subduction.

Acknowledgments. This study was supported by a grant from the U.S. National Science Foundation Division of Polar Programs (DPP88-16807). Mike Roberts provided outstanding logistical and geological assistance in the field. We thank Rick Law for discussion of quartz *c* axis patterns. Simon Hanmer and Peg Rees provided constructive reviews.

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(Received January 13, 1993;
revised July 29, 1993,
accepted August 5, 1993.)