Latest Neoproterozoic basin inversion of the Beardmore Group, central Transantarctic Mountains, Antarctica

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Abstract. Structural and age relationships in Beardmore Group rocks in the central Transantarctic Mountains of Antarctica indicate that they experienced a single deformation in latest Neoproterozoic to early Paleozoic time. New structural data contrast with earlier suggestions that Beardmore rocks record two orogenic deformations, one of the early Paleozoic Ross orogeny and a distinct earlier tectonic event of presumed Neoproterozoic age referred to as the Beardmore orogeny. In the Nimrod Glacier area, Beardmore metasedimentary rocks contain only a single set of geometrically related regional structures associated with the development of upright, large- and small-scale flexural-slip folds. Deformation of Beardmore strata involved west directed contraction of modest regional strain at relatively high crustal levels. Existing ages of detrital zircons from the Cobham and Goldie formations constrain Beardmore Group deposition to be younger than \( \approx 600 \) Ma. This is significantly younger than previous age estimates and suggests that Beardmore deposition may be closely linked to a latest Neoproterozoic East Antarctic rift margin. The lack of structural evidence for polyphase deformation and the relatively young depositional age for the Beardmore Group thus raises the question of a temporally and/or tectonically unique Beardmore orogeny. Here I suggest that Beardmore shortening may be related to tectonic inversion of East Antarctic marginal-basin strata because of localized compression during proto-Pacific seafloor spreading. Basin inversion is but one stage in a protracted Ross tectonic cycle of rifting, tectonic inversion, subduction initiation, and development of a mature convergent continental margin during latest Neoproterozoic and early Paleozoic time. The term "Beardmore orogeny" has little meaning as an event of orogenic status, and it should be abandoned. Recognition of this latest Neoproterozoic history reenforces the view that the broader Ross orogeny was not a single event but rather was a long-lived postrifting tectonic process along the East Antarctic margin of Gondwanaland.

Introduction

The Transantarctic Mountains constitute a major orogenic belt recognized for its early Paleozoic role in Gondwanaland amalgamation and its more recent tectonic history involving Jurassic to Neogene rifting and uplift. Traditionally, the pre-Devonian evolution of the Transantarctic Mountains is viewed in terms of three orogenic cycles: (1) the Precambrian Nimrod urgeny [Grindley and Laird, 1969], manifested in high-grade metamorphic rocks of the Nimrod Group; (2) the Neoproterozoic Beardmore orogeny [Grindley and McDougall, 1969], based on deformation in the Beardmore Group prior to deposition of Lower Cambrian sediments; and (3) the early Paleozoic Ross orogeny [Gunn and Warren, 1962], which imparted the principal pre-Jurassic structural fabric upon the present mountain belt. The two Precambrian urgenies were originally defined on the basis of geologic relations in the central Transantarctic Mountains, whereas early Paleozoic structures of the Ross orogen are recognized along the length of the mountain belt. The tectonic evolution of the Ross orogen is important in the broader context of global plate tectonics at the close of the Proterozoic because this belt contains rock assemblages that record Archean and Proterozoic craton growth, Rodinian supercontinent rifting, and plate-margin convergence related to Gondwanaland amalgamation.

Critical to our understanding of the Ross orogen is the relationship between crystalline basement and (meta)sedimentary supracrustal assemblages, which together show evidence of protracted Ross tectonism during the latest Neoproterozoic and earliest Paleozoic [Kleinschmidt and Tessensohn, 1987; Flöttemann und Kleinschmidt, 1991, Goodge et al., 1991, 1993b; Dallmeyer and Wright, 1992; Rowell et al., 1992, 1993; Goodge and Dallmeyer, 1996]. In the central Transantarctic Mountains, recent structural, petrologic, and thermochronologic data show that rocks of the Nimrod Group experienced major dynamothermal activity during the Ross orogeny that is not recorded in adjacent units [Goodge and Dallmeyer, 1992, 1996; Goodge et al., 1993a, b]. Although magmatic and depositional events in the Nimrod Group can be traced to \( \approx 3.0 \) Ga, there are few if any geologically recognizable events over this time period which could reasonably be assigned "orogenic" status. Because the principal metamorphic and structural features of the Nimrod Group are now recognized as a deep-crustal manifestation of the Ross orogeny, there is little discernible evidence for the Nimrod orogeny.

The younger orogenic record, however, remains obscured by uncertain depositional, age, and deformational relationships in widespread graywacke units commonly interpreted as Neoproterozoic in age. In most areas, these siliciclastic rocks show clear evidence of Ross tectonism, and several workers have reported evidence of an earlier deformation attributed to the Beardmore orogeny [Schmidt et al., 1965: Grindley and McDougall, 1969; Laird et al., 1971; Stump et al., 1991; Storey et al., 1992]. In the central Transantarctic Mountains, evidence for the Beardmore orogeny stems from apparently unconformable stratigraphic relations between Neoprotero-
zoic (Beardmore Group) and Lower Cambrian (Ryrd Group) sediments and from two reported deformations in the Neoproterozoic rocks [Grindley and Laird, 1969; Laird et al., 1971; Stump et al., 1991; Stump, 1995]. However, because the Ross and Beardmore structures are nearly coaxial and because of poor internal biostratigraphic and age control in the Beardmore Group, the relationship between the Beardmore and Byrd sedimentary units and the nature of the Beardmore orogeny remain uncertain. This uncertainty is compounded by reinterpretation in several locations of the Beardmore-Byrd unconformity as faults [Rowell et al., 1986; Rees et al., 1989]. In the Pensacola Mountains, metagraywackes once interpreted as Neoproterozoic in age (Patuxent Formation) [Schmidt et al., 1978] are now known to consist of two distinct sequences (pre-Middle Cambrian and earliest Ordovician) [Rowell et al., 1992, 1994; Millar and Storey, 1995]. Such age relationships demonstrate that at least some of the deformation previously held to be Neoproterozoic is necessarily post 500 Ma in age (i.e., Ross). These relationships therefore call into question the existence of a tectonically unique Beardmore orogeny, rather than a continuum of deformation operating within a common tectonic framework through the latest Neoproterozoic and early Paleozoic.

Evidence for the Beardmore orogeny should be present in the Nimrod Glacier area (Figure 1), where it was initially reported that the polydeformed Beardmore Group unconformably underlies Byrd Group sediments. In conjunction with recent work in the Nimrod Group, a comparative study of structural relations in the Beardmore Group was undertaken in the upper Nimrod Glacier area during the 1990-1991 austral summer. Aerial support provided by the German GANOVEX VII expedition permitted landings at key exposures of the Beardmore Group in the Cobham Range and at Kn-Tiki Nunatak. Although field examination of these exposures was brief (2 days), the new structural data presented here have an important bearing on tectonic relations between basement and supracrustal rocks in the region. From these structural data, combined with depositional age constraints, I present a new tectonic model that explains deformation of the Beardmore Group as an early phase of basin inversion associated with the Ross orogeny.

Geologic Setting

In the Nimrod Glacier area, Precambrian metamorphic rocks, Neoproterozoic to lower Paleozoic sedimentary rocks, and Cambrian-Ordovician granitoids are exposed beneath a flat-lying Devonian to Jurassic Gondwana sedimentary and igneous cover sequence (Figure 1) [Gunn and Walcott, 1962; Grindley et al., 1964, Grindley and McDougall, 1969; Laird et

Figure 1. Generalized geologic map of the central Transantarctic Mountains in the vicinity of Nimrod Glacier [after Grindley and Laird, 1969; Laird et al., 1971; Goodge et al., 1993a]. Inset shows location in Antarctica. Abbreviations are as follows: EA, Fast Antarctica; NG, Nimrod Glacier area; TM, Transantarctic Mountains; and WA, West Antarctica.
al., 1971; Rowell et al., 1988b; Boré et al., 1990]. Basement metamorphic rocks of the Precambrian Nimrod Group [Grindley et al., 1964] consist of strongly deformed high-grade gneiss and schist of Archean to Paleoproterozoic age that likely represent part of the East Antarctic craton [Gunnar, 1982; Boré et al., 1990; Goode et al., 1991; Walker and Goode, 1991; Bennett and Fanning, 1993]. Outboard of this basement complex to the Ross Sea side of the present Transantarctic Mountains lie two major supracrustal assemblages, the Beardmore and Byrd groups. Neoproterozoic(?), Beardmore Group strata [Grindley and Warren, 1964] are divided into a sequence of calc-schists (Cobham Formation) [Laird et al., 1971] conformably overlain by turbiditic meta-graywacke and slate (Goldie Formation) [Gunn and Walcott, 1962]. The Byrd Group [Laird, 1963] includes Lower Cambrian shallow-water carbonates (Shackleton Limestone) and unconformably overlying Middle to Upper Cambrian elastic units (Starshot and Douglas formations) [Laird, 1963; Skinner, 1964; Rowell et al., 1988b; Rowell and Rees, 1989; Rees and Rowell, 1991]. Cambrian to Ordovician granitoid plutons (Granite Harbour Intrusives) intrude all of the basement and supracrustal assemblages noted above, forming part of a regionally extensive syntectonic to posttectonic magmatic province [Gunnar and Warren, 1962; Gunnar and Mattinson, 1975; Gunnar, 1976; Boré et al., 1990; Goode et al., 1993b]. This paper focuses on the pre-Devonian nongranitic units.

Rocks of the Nimrod Group are exposed in the Miller and Geologists ranges (Figure 1). These rocks are unique in the Ross Sea sector of the Transantarctic Mountains in terms of lithology, age, metamorphism, and structural relations, although metamorphic counterparts may exist in northern Victoria Land [e.g., Kleinschmidt and Tessensohn, 1987; Goode and Dallmeyer, 1996]. Geologic and structural relations of the Nimrod Group are described in detail by others [Grindley et al., 1964; Gunnar, 1969; Grindley, 1972; Goode et al., 1990, 1991, 1992, 1993a; Peacock and Goode, 1995]. Upper amphibolite to lower granulitic facies Nimrod schists and gneisses include a variety of metasedimentary and metagneissic lithologies that host relict eclogites. These rocks are characterized by pervasive L-S ductile tectonite fabrics and four stylistically distinct types of folds [Goode et al., 1991, 1993a]. Regionally consistent mesoscopic and microscopic fabric relations indicate that the Nimrod tectonites and folds formed during SE directed, strike-parallel ductile shear. High-grade metamorphism and deformation occurred between about 540 and 520 Ma and are temporally associated with the Ross orogeny [Goode and Dallmeyer, 1992, 1996; Goode et al., 1993b]. Thus the dynamothermal events once viewed as a distinct Nimrod orogeny [Grindley and McDougall, 1969; Grindley, 1977; Gunnar and Faure, 1972] are now recognized as the response of deep-seated crystalline rocks within the East Antarctic craton to Ross-age tectonism [Goode et al., 1993a]. It remains unclear, however, how high-grade Nimrod deformation relates to nearly contemporaneous but kinematically different deformation in the supracrustal Beardmore and Byrd groups.

Beardmore Group strata in the Nimrod Glacier area include the Cobham Formation, a lower sequence of calcareous schist, impure marble and quartzite, and the Goldie Formation, a conformably overlying sequence of nonfossiliferous graywacke and slate. The Cobham Formation is exposed only on the west side of the Cobham Range and the west end of Kon-Tiki Nunatak (Figure 1). Goldie rocks are widely exposed from north of Nimrod Glacier to south of Beardmore Glacier, beyond which they correlate with similar siliciclastic rocks of the Duncan and La Gorce formations [Stump, 1981; Stump et al., 1986]. The conformable Cobham–Goldie sequence records inner continental-shelf sedimentation followed by deep water turbidite deposition across a subsiding rifted continental margin [Laird et al., 1971]. Matrix and grain compositions of the La Gorce Formation indicate a dominantly cratonic or crystalline source terrain [Stmit and Stump, 1986]. There is no biostratigraphic evidence constraining the depositional age of the Beardmore Group, but a Neoproterozoic age is indicated by a Sm-Nd isochron age of 762 Ma [Boré et al., 1990] from a gabbro associated with pillow basalts interlayered with Goldie turbidites. Single-grain Pb-evaporation ages for detrital zircons in Goldie quartzite include distinct populations at about 0.6–0.7, 1.0–1.2, 2.2–2.3, and 2.8 Ga [Walker and Goode, 1994; Walker, 1996], indicating latest Neoproterozoic deposition. Grains from one Goldie sample collected at Kon-Tiki Nunatak yielded ages as young as ~585 Ma [N.W. Walker, personal communication, 1994]. The morphology and isotopic behavior of the younger grains in particular indicate a young, volcano-plutonic source for at least part of the Goldie detritus. The range of detrital zircon ages corresponds well with known East Antarctic igneous and metamorphic events [Tingey, 1991], indicating a cratonal provenance; because of the distinctive younger population, a Laurentian source appears unlikely [Walker, 1996].

The Byrd Group is dominated by the thick, fossiliferous Lower Cambrian Shackleton Limestone, containing trilobite and archaeocyathid faunas of Atabanian, Botomian, and possibly Toyanian age [Laird and Waterhouse, 1962; Laird, 1963; Debenne and Kruse, 1986; Rowell et al., 1988a; Rowell and Rees, 1989; Palmer and Rowell, 1995]. The Shackleton primarily represents a broad subtidal shelf, but it also contains discontinuous carbonate patch reef, storm, and sandbar deposits [Rees et al., 1989]. Unconformably overlying the Shackleton Limestone are carbonate-clast conglomerates and siliciclastic arenites (Douglas Conglomerate) interpreted as synorogenic alluvial fan deposits [Rowell et al., 1988b; Rees and Rowell, 1991]. The Douglas Conglomerate is poorly dated, but stratigraphic relations indicate it is probably Middle to lower Upper Cambrian [Pantaija and Rees, 1991]. As noted earlier, the Goldie-Shackleton contact is controversial. Previously, it has been considered conformable [Laird, 1963], unconformable [Laird et al., 1971; Stump et al., 1991], and faulted [Rowell et al., 1986; Rees et al., 1989]. An unconformable relation between these units, if it exists, would provide conclusive evidence for a pre-Early Cambrian deformation of the Beardmore Group, but uncertainty in the nature of the contact leaves the existence of such a deformation in question.

Laird et al. [1971] and Stump et al. [1991] described structural data from the Beardmore Group in the Nimrod Glacier area that indicated two deformations, the younger of which corresponded to Ross events (Table 1). Both Beardmore and Byrd group rocks display contractional structures associated with Ross deformation, here referred to as D5, in the context of previous studies, which represents the primary supracrustal
Table 1. Summary of Structural Relations Reported for Beardmore Group Sedimentary Rocks in the Ninrod Glacier Area

<table>
<thead>
<tr>
<th>Laird et al. [1971]*</th>
<th>Stump et al. [1991]*</th>
<th>This Study</th>
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<tbody>
<tr>
<td>Late Ross deformation ($D_2$), producing open, upright folds ($F_2$) about horizontal north trending axes; associated with subvertical to steeply east dipping axial planes ($S_2$); affected Beardmore and Byrd groups</td>
<td>Late Ross deformation ($D_2$), producing upright west vergent folds ($F_2$) and steep NW striking cleavage ($S_2$); affected Beardmore and Byrd groups</td>
<td>Single Ross deformation ($D_3$), producing west directed flexural-slip folds ($F_1$) about gently south plunging axes, with steep east dipping cleavage ($S_1$); affected Beardmore Group; presumed affect on Byrd Group not observed</td>
</tr>
<tr>
<td>Early Beardmore (?) deformation ($D_1$), producing recumbent isoclinal folds ($F_1$) about subhorizontal axial planes ($S_1$); affected Beardmore Group only</td>
<td>Early Beardmore deformation ($D_1$), producing east vergent recumbent folds ($F_1$) and subvertical N-S cleavage ($S_1$); $F_1$ mesofolds locally recumbent; affected Beardmore Group only</td>
<td>No Beardmore deformation</td>
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* Abbreviations for deformations (D), folds (F), and cleavages (S) are designations given here to correlate inferred ages of structures.

deformation in the central Transantarctic Mountains [Gunn and Walcott, 1962; Grindley, 1963; Laird, 1963; Laird et al., 1971; Oliver, 1972; Gunner, 1976; Stump, 1981; Rowell et al., 1986; Rees et al., 1987; Stump et al., 1991; Stump, 1995]. In this region, $D_2$ formed subhorizontal folds about north to northwest trending axes with steep axial planes. An earlier phase of Beardmore deformation, here referred to as $D_1$, is reportedly marked by tight to isoclinal mesoscale folds and an outcrop-scale recumbent isoclinal in the Cobham Range [Laird et al., 1971], by refolded isoclinal folds in Goldie rocks at Kon-Tiki Nunatak [Stump et al., 1991], and by crosscutting cleavages and both east and west vergent mesofolds at Cotton Plateau [Edgerton, 1987; Stump et al., 1991; Stump, 1995]. However, the field observations and structural data presented here are incompatible with structural interpretations made by these earlier workers for the existence of $D_2$. Below I present new structural observations, and I discuss how these data are inconsistent with a distinct Beardmore orogenic deformation.

Beardmore Group Structural Relations

Cobham and Goldie strata were studied in the Cobham Range north of Ninrod Glacier (Figure 2) and at Kon-Tiki Nunatak in the middle of Ninrod Glacier (Figure 3). In the western Cobham Range, these units form an east dipping homocline. Farther east in the Holyoake Range (Figure 1), Goldie beds are folded with the Shackleton Limestone, although the contact between them is obscured and may be structural. At Kon-Tiki Nunatak, the Beardmore units form a large-scale, south plunging synform.

The Cobham Formation includes calcareous schist, phyllite, mica schist, impure quartzite, and minor calc-silicate and marble. Bedding on a scale of centimeters is well preserved, parallel to metamorphic foliation, and shows relict laminations and cross bedding. Following Laird et al. [1971], the contact between the Cobham and Goldie units lies above the uppermost gray calcareous schist. The Goldie Formation is characterized by alternating thin-bedded quartzofeldspathic meta-arenite (metagraywacke), quartzite, and black slate, with minor calcareous schist. Tabular-bedded meta-arenite layers show relict cross bedding and graded bedding indicative of deposition as turbidite flows.

Beardmore Group rocks in this area were metamorphosed during two low-grade events, $M_1$, and $M_2$, as indicated by textural and microfabric relations. $M_1$, a regional metamorphism in the biotite zone of the greenschist facies, formed a bedding-parallel foliation ($S_0$) shown by the gran-shape preferred orientation of muscovite and biotite (Figure 4). Layering is commonly cut by a discrete macroscopic cleavage ($S_1$) at moderate to high angles to $S_0$, in places manifested by crenulations and microfolds that form spaced crenulations (Figure 4). Although $S_1$ crosses $S_0$, $S_1$ could either be related to one protracted syn-$M_1$ deformation or to a younger, post-$M_1$ deformation event. Neoblastic muscovite along crenulation planes was observed in only one thin section, indicating that the deformation which formed $S_1$ probably occurred during the waning stages of $M_1$. In outcrop as well as in thin section, no textural evidence for more than one bedding-normal cleavage was observed.

A second metamorphism ($M_2$) is shown by sparse but areally widespread albite-epidote to hornblende hornfels-facies assemblages in the vicinity of Granite Harbour plutons, such as at Half Dome Nunatak (Figure 1). $M_2$ contact metamorphism produced spotted phyllites and hornfels, depending on host lithology, marked by coarse-grained poikiloblastic biotite, chloritoid, Ca-amphibole, and, locally, garnet. Randomly oriented poikiloblasts overgrowing $S_0$ and $S_1$ indicate that $M_2$ is strictly postkinematic (Figure 4).

Cobham Range

In the Cobham Range, relict bedding and compositional layering ($S_0$) dip gently to moderately east (Figure 5). Bedding laminations and cross bedding are locally preserved, although a layer-parallel mineral foliation is evident in thin section. A single cleavage ($S_1$) dips steeply east, at a higher angle than
bedding (Figures 5 and 6a), indicating that the strata here comprise one limb of an upright fold. Mesoscopic asymmetric folds with sharp to angular hinges are common ($F_1$, Figure 6b) and display east dipping, west verging axial planes. Exposure does not typically allow for unambiguous measurement of fold axes; to avoid dispersion of orientations, only geometrically well-expressed axes were measured. Down dip slickenside lineations ($L_{sd}$) contained within $S_0$ are...
Figure 4. Thin section photomicrograph of Goldie Formation phyllite from the Cobham Range. Light is plane-polarized; view is ~3 mm across. Early layer-parallel foliation (S₀), defined by grain-shape preferred orientation of fine-grained muscovite and biotite, is deformed by high-angle crenulations (S₁) parallel to regional macroscopic cleavage. Unstrained, coarse-grained biotite poikiloblast contains curved inclusion trails, showing that it overgrew foliation after formation of S₁.

Figure 5. Equal-area, lower hemisphere stereoplots of Beardmore Group structures from the Cobham Range and Kon-Tiki Nunatak. In the Cobham Range, bedding and layer-parallel foliation (S₀) dip uniformly east; average bedding is indicated by dashed great circle. Gently south plunging mesoscopic folds (F₁) are associated with a steep, east dipping axial-plane cleavage (S₁; average cleavage indicated by dotted great circle) and slickenside lineations (L₁cb) at a high angle to the fold axes. Bedding orientations on west limb of syncline define regional β axis shown by open circle. Asymmetric F₁ folds show a westward vergence (see Figure 6b). At Kon-Tiki Nunatak, bedding and foliation (S₀) are folded about a large-scale, south plunging synform (Figure 3) with β axis shown by open circle. Average orientations of bedding in each limb are shown by conjugate great circles in the diagram (dashed). Folded bedding surfaces show slickenside lineations (L₁cb) at a high angle to fold axes. Gently south plunging mesoscopic folds (F₁) are colinear with bedding-cleavage intersection lineations (L₁cb) and are associated with a steep, east dipping axial-plane cleavage (dotted; S₁). Note that measured F₁, measured L₁cb, β axis, and average planes of S₀ and S₁ are coincident.
approximately perpendicular to the average fold axis (Figure 5), indicating a flexural-slip fold mechanism. Sets of en echelon intralateral tension fractures dip moderately west in units with high relative competency (Figure 6c). I interpret the structurally coherent bedding-cleavage relations, the geometry of the asymmetric folds, axis-normal slickenside lineations, and the asymmetry of en echelon tension fractures as evidence that the Beardmore units in the Cobham Range were affected by a single west directed deformation.

As noted by Laird et al. [1971], the Cobham marls contain numerous asymmetric mesoscopic folds. These folds, such as shown in Figure 6b, contain shallow to moderately east dipping axial planes. Our field party observed no recumbent isoclinal folds of the type described by Laird et al. [1971, Figure 19]. An earlier field party did find one such fold hinge (E. Stump, personal communication, 1996), but its orientation and associated structures are uncertain. If these thinly bedded rocks were isoclinally folded by deformation, one would expect to observe stratal repetition, inverted bedding, inverted cleavage, refolded fold axes, or a combination of these. The structural data presented here are inconsistent with reversals of stratigraphic indicators, an early cleavage, or kinematic fabrics, and the structural elements observed are not those associated with the ductile flow required to form isolated large-scale, recumbent isoclinals as seen in some nappé terrains.

**Kon-Tiki Nunatak**

Most of Kon-Tiki Nunatak consists of interbedded Goldie metaturbidites and slates, underlain to the west by Cobham calcareous schists (Figure 3). Beardmore rocks are folded to form a single large-scale, south plunging syncline, defined by a great circle to upright bedding (Figure 5). The rocks display mesoscopic folds (F1) with gently south plunging axes. F1 folds are asymmetric, open to tight, and show opposite vergence on either side of the central synclinal axis (Figure 7). F1 fold axes are parallel to the β axis defined by the great circle to bedding and to bedding-cleavage intersection lineations (L1cb) that plunge uniformly south (Figure 5). A single axial-
Figure 7. Sketch from photograph mosaic of folds within Goldie and Cobham formations at Kon-Tiki Nunatak, showing detailed relation between locally tight folds and larger west vergent syncline. View is toward the southeast (note this is the opposite view shown in cross section of Figure 2 and sketch in Figure 6), with Nimrod Glacier in the foreground and Mt. Markham area of Queen Elizabeth Range in the background. Note the southeastward view is approximately same as that shown by Stump et al. [1991, Figure 3]; their figure is incorrectly labeled with east to the right and west to the left. Beardmore units include quartzite and calcareous quartzite (dark gray) and slate (light gray). Approximate Goldie-Cobham contact indicated. Height of main exposure is approximately 150 m. Inset drawings show detailed structural/fabric relations at locations indicated on main sketch (all views looking southeast or as otherwise indicated). (a) Angular relation between relict bedding in Goldie Formation (S0; dipping steeply right, or west) and axial-plane cleavage (S1; dipping steeply left, or east, parallel to pencil). View down intersection lineation (L1crb) to south-southeast. Pencil is 14 cm. (b) Parasitic asymmetric folds (F1) in Goldie Formation calc-schist on east side of synclinal axis. Axial-profile view looking south-southeast. Hammer is 33 cm. This fold occurs on east limb of large syncline, and steep east dipping axial-plane cleavage indicates westward overall vergence. Note cleavage refracted into fan shape through hinge of fold in thick layer (center left). (c) Open folds in Goldie Formation at west end of Kon-Tiki Nunatak, showing cleavage refraction through relict beds of different composition. Principal cleavage (S1) is nearly perpendicular to layering in thicker, more competent quartzites (gray) and dips at moderate angle to left (east) within schistose to phyllitic units (white). Quartzite layers also contain east dipping quartz veins and mineralized fractures; these fractures do not offset S1 and may have formed during the same deformation event. Axial-profile view down intersection lineation (L1crb) to south-southeast. Lens cap is 6 cm in diameter. (d) Detailed outcrop sketch showing parasitic isoclinal folds in hinge area of main syncline. Gray layers are prominent tabular-bedded quartzite. (e) Detailed outcrop sketch showing core of main Kon-Tiki syncline, marked by laterally continuous marker beds (light and dark gray). Note at least one outer layer is offset by low-angle reverse fault, and no isoclinal recurrent folds are apparent in western limb (right). (f) Detailed outcrop sketch showing uniformly east dipping layers in western limb of main syncline. No isoclinal recurrent folds are apparent, although sedimentary layers offset along a moderately east dipping reverse fault. Apparent offset in fault results from perspective view to near (left) and distant (right) cliff faces, separated by near-vertical line.
plane cleavage ($S_1$) dips steeply east (Figures 7a, 7b, and 7c), although individual outcrops display cleavage fans (Figure 7b) and cleavage refraction between layers of different composition (Figure 7c). Bedding surfaces contain a slickenside lineation ($L_{sk}$) that varies in plunge but is always at a high angle to mesoscopic fold axes, indicating that the folds formed predominantly by a flexural-slip mechanism.

Several tight to isoclinal folds are evident on the limbs of the large Kun-Tiki syncline when viewed to the southeast (Figure 7). East of the syncline axis, Goldie beds are involved in tight, upright structures, and folds show nearly vertical axial planes and an S-shaped geometry consistent with limb appression adjacent to the syncline hinge (Figure 7d). Smaller-scale, asymmetric folds (such as shown in Figure 7b) also show a consistent S-shaped geometry east of the syncline axis. Within the hinge of the main syncline, some beds are thicken by bedding-parallel reverse faults (Figure 7e), recording greater contraction in the hinge region. In the gently east dipping western limb of the Kun-Tiki syncline, subsidiary folds display a Z-shaped geometry (Figure 7c) and in places they are cut by high-angle reverse faults (Figure 7f). The large-scale folds and associated reverse faults indicate strong contraction in the core of the Kun-Tiki syncline, and their geometry is best explained by a single contractional deformation event. Although the existing data do not prove that the folds and faults are the same age, their geometry is consistent with such an interpretation.

An oblique aerial view of the folds in the syncline axis was also shown by Stump et al. [1991, Figure 3] (not this is not an axial profile view, as discussed below) and cited as evidence of polyphase folding, in which gently east dipping recumbent isoclines were taken to be refolded by steeper fold. From field observations and structural data, we found no recumbent isoclinal folds of the type previously described which predate formation of the main syncline. Rallic, the exposure on the northwest face of Kon-Tiki Nunatak consists of an open, asymmetric syncline showing opposing limb dips and opposite senses of fold vergence on either limb. In particular, folds within the western limb are not recumbent isoclines but are close, sharp-hinged parasitic folds that show a westward vergence. These folds, referred to as Beadmore-age F1, structures by Stump et al. [1991], do not have an eastward vergence. Therefore, in contrast to earlier interpretations, (1) both sets of folds on either side of the main syncline axis are sympathetic folds generated near the syncline hinge, and they do not require two episodes of deformation, and (2) the folds show opposite senses of vergence on either side of the hinge, as would be expected for a single large fold, precluding different vergence of polyphase structures. As in the Cobham Range, therefore, Beadmore rocks at Kon-Tiki Nunatak were affected by only a single contractional deformation and a late-stage static metamorphism.

**Evidence for the Beadmore Orogeny**

The main lines of evidence for a distinct orogenic event that predates the Ross orogeny are reports of a stratigraphic unconformity between the Goldie and Shackleton formations [Grindley and Laird, 1969; Laird et al., 1971] and polyphase deformation in the Goldie Formation [Laird et al., 1971; Stump et al., 1991; Stump, 1995]. Reconnaissance field study of the greater Nimrod Glacier region by Laird et al. [1971] was the first published account suggesting pre-Early Cambrian deformation of the Goldie rocks. Stump et al. [1991] focused their study in the Cotton Plateau area south of Nimrod Glacier (Figure 1), employing structural data reported by Edgerton [1987].

Stratigraphic evidence for the Goldie-Shackleton unconformity is not definitive, despite widespread acceptance of this major orogenic marker in the Transantarctic Mountains. Laird et al. [1971] described the Goldie Shackleton contact at five localities as an angular unconformity lying on "an unweathered surface." Among these, "breccias" of uncertain origin are described at two locations, an undulating surface overlying "drag-folded schist" is described at one location, at least one contact is sheared [Laird et al., 1971, Figure 6], three locations show no angular discordance, and one contact is reportedly "slightly brecciated and slickensided." These observations are more typical of tectonic rather than stratigraphic contacts, although Laird et al. [1971] did not specifically address the possibility that the contacts are structural. Reexamination of the Goldie-Shackleton contact at several localities north of Nimrod Glacier led Rowell et al. [1986] and Rees et al. [1989] to conclude that most if not all of them are faulted rather than depositional. Stump et al. [1991] affirmed the interpretation by Laird et al. [1971] that these units are unconformable at Cotton Plateau, but Stump [1992, p. 27] later equivocated on this point.

Polyphase deformation within the Goldie Formation is also open to question. In the Cobham Range, as described above, Cobham and Goldie strata dip homoclinally to the east. Across a synformal axis in the eastern flank of the range, these units dip steeply west. In their initial survey of the Cobham Range, Laird et al. [1971] reported common sub-horizontal isoclines in both formations. The axes of these folds plunge shallowly to the southeast and south, and they are reportedly associated with an axial plane schistosity. These mesofolds probably correspond to the F1, recognized here because of their geometric and stylistic similarities. As noted above, Laird et al. [1971, Figure 19] also reported the presence of a single recumbent isocline with notably greater amplitude within the Cobham Formation. Stump et al. [1991] described somewhat different fold orientations in the Cobham Range, with fold axes plunging in the "NE and SE quadrants," which they interpreted as having an easterly vergence. Both groups described gently north plunging folds (F2) with near-vertical axial planes and a locally displayed cleavage. Stump et al. [1991] interpreted the F2 folds as west vergent, but no crosscutting cleavages were documented.

Beadmore strata at Kon-Tiki Nunatak also reportedly experienced two generations of folding. Near the syncline axis in the central section of the nunatak, Stump et al. [1991] reported recumbent isoclinal folds (F1) with gently east dipping axial planes, refolded by tight upright folds (F2). No orientations were given, but the F1 folds are stated to have similar orientations as structures at Cotton Plateau, south of Nimrod Glacier. The apparent recumbent isoclinal nature of the folds illustrated by Stump et al. [1991, Figure 3] is not an axial profile view; therefore the true shape and orientation of these folds is not represented. At Cotton Plateau, Goldie strata are folded in a syncline that is disrupted by a shear zone oriented subparallel to the fold hinge [Edgerton, 1987; Stump et al.,]
1991]. Beds in the eastern limb of this syncline are steeply west dipping, and beds of unknown orientation in the western limb are described as recumbently folded. Along the northern edge of Cotton Plateau, Stump et al. [1991] reported a north striking, near-vertical cleavage \( (S_1) \) and a northwest striking, subvertical to steeply northeast dipping cleavage \( (S_2) \). \( S_2 \) is interpreted to be axial-planar to east vergent \( F_1 \) folds, whereas \( S_2 \) is axial-planar to west vergent \( F_2 \) folds. Stump [1995] stated that such crosscutting fold-cleavage relations are displayed along the western side of Cotton Plateau.

Despite its regional importance with respect to Gondwanawide tectonic events at the end of the Proterozoic, structural evidence for the Beadmore orogeny is contradictory and inconclusive. In particular, the structural relations described by Laird et al. [1971], Edgerton [1987], Stump et al. [1991], and Stump [1995] provide insufficient evidence of polyphase deformation in the Beadmore Group. Briefly noted are the following problems. First, folds designated as \( F_1 \) are described with axes that are not contained within axial planes, as they should be for cylindrical folds. Likewise, nonco planar \( F_2 \) fold axes and \( S_2 \) cleavage indicate geometric incompatibilities in folds designated as \( F_2 \). These geometric problems with the fold elements raise uncertainties about their interpretation as polyphase structures. Second, the early \( F_1 \) axes are described as uniform in orientation; they are not dispersed by \( F_2 \), as would be expected for nearly orthogonal successive folding events. Third, \( F_2 \) axes as measured and as derived geometrically from fold bedding and cleavage attitudes are not colinear, suggesting either that the criteria by which different fold generations were defined are invalid or that folds of a single generation were reoriented locally by high-strain structural features such as reported at Cotton Plateau [Stump, 1995]. Fourth, \( S_1 \) cleavage reported are both steeper and more uniform in orientation than \( S_2 \), not as would be expected if associated with early recumbent folds and then subsequently deformed. In fact, \( S_1 \) cleavage orientations were used elsewhere to define the orientation of \( F_2 \), illustrating the uncertainty about whether \( S_1 \) is more uniform or more variable than \( S_2 \). Perhaps the two cleavages described in the Cotton Plateau area represent strain heterogeneity associated with local deformation effects on a regional Ross cleavage. Last, as noted above, I cannot confirm the existence of an isolated recumbent isocl ine in the Cobham Range. If it does exist, as suggested by others, it is difficult to explain how an early recumbent isocl ine could form by deformation in a low-grade sequence of well-bedded strata and not display evidence of cleavage reorientation during \( F_2 \). Thus the Goldie structures described in earlier reports do not adequately document polyphase deformation of the Beadmore Group.

Part of the difficulty in distinguishing single versus multiple deformations may stem from uncertainty in the origin of the folds themselves. Cobham and Goldie strata represent calcareous muds and turbidites of the continental shelf to rise. It is possible that some of the "early" structures recognized by previous workers might be the result of soft-sediment movement in an unstable slope setting, as earlier cautioned by Grindley and Laird [1969]. They recognized only a single east dipping axial-plane cleavage related to Ross deformation in the Beadmore Group, but they noted NE vergent recumbent folds in the Cobham Range and at Cotton Plateau which they speculated may have formed by "gravity gliding." Soft-sediment slump structures with an easterly vergence and a lack of an associated cleavage might be expected in shallow-water carbonates and turbidite flows associated with an east facing continental rift margin. This mode of origin should be considered a plausible explanation for the early recumbent structures noted by Laird et al. [1971].

In summary, the structural relations reported here indicate (1) that the Beadmore Group deformation patterns are generally coherent and regionally distributed and (2) that they are most simply interpreted in terms of a single, chiefly contractional deformation. The occurrence of both isoclines and open, upright folds is compatible with a single, locally variable folding event, and this deformation produced only a single regional cleavage. Gunn and Walcott [1962], Oliver [1972], and Gunnar [1976] also observed evidence for only a single deformation in the Goldie Formation east of the study area and as far south as Beadmore Glacier.

These findings raise the larger question of whether or not there was a Neoproterozoic Beadmore orogeny as distinct from Ross events. The ages of detrital zircons extracted from Goldie and Cobham strata constrain the age of Beadmore Group deformation to be younger than \( -600 \) Ma, and posttectonic Granite Harbour plutons constrain deformation to be older than \( -500 \) Ma. That this time frame is shorter than previously recognized is significant. The structural and age relations discussed here point to deformation of Beadmore Group rocks during an early phase of Ross orogenesis. This conclusion supports the suggestion by Grindley and Laird [1969] that folds in the Beadmore Group "may have developed during an early phase of the Ross Orogeny before deposition of the Lower Cambrian Byrd Group." Although the duration of Beadmore Group deformation cannot presently be bracketed to any less than about 100 m.y., it is unlikely that two distinct orogenic cycles (in contrast with deformation pulses) occurred within this time period. Therefore I interpret deformation of the Beadmore rocks as indicative of protracted Ross orogenic activity, perhaps spanning 100 m.y., as also indicated by diachronous syntectonic to posttectonic Ross magmatism [Goodge et al., 1993b; Rowell et al., 1993; Encarnación and Gunnar, 1996]. On this basis, I recommend that the term "Beadmore orogeny" be abandoned and that the spatially variable and diachronous deformations observed at various crustal levels in the central Transantarctic Mountains (and perhaps the entire orogen) be included under a broadly defined Ross orogeny. As more age data become available, they appear to trace a prolonged continuum of tectonically related processes that are best described by an inclusive orogenic terminology, rather than by assignment of discrete deformation episodes to individual orogenic cycles.

The angular discordance between Beadmore and Byrd group rocks, if stratigraphic, does provide evidence that deformation of Beadmore rocks commenced prior to deposition of the Byrd Group. The Shackleton Limestone itself does not preserve regular cleavage patterns [Rees et al., 1989, Palmer and Rowell, 1995], yet some Byrd Group sedimentary rocks do show evidence of at least two "deformations" during Ross time. In the northern Holyoake Range, tilted Douglas Conglomerate lies with angular unconformity on Shackleton Limestone [Rowell et al., 1988b], implying two pulses of Middle and Late Cambrian deformation. Here the Douglas is dominated by coarse clasts of Shackleton Limestone, and it is
regarded as a synorogenic sediment. South of Byrd Glacier, the Douglas unit displays refolded folds recording three pulses of deformation [Rees et al., 1987], yet these deformation "events" are all regarded as phases of late Ross tectonism in the latest Ordovician. Thus lower Paleozoic units contain polyphase folds that are interpreted as the product of a single orogenic period rather than distinct tectonic events. Although I assert that the Beardmore Group shows evidence of only one deformation, some of the other reported fold complexities may have developed in the same way during one regional event.

If several generations of structures are recognized in lower Paleozoic units, why are these not all observed in the Beardmore rocks? It is possible that individual units may be partially detached from one another mechanically, that they were deformed at different crustal levels, or that they were deformed over a broad period of orogenic contemporaneity with deposition. Mechanical detachment is likely between basement and supracrustal units, and perhaps as well between preorogenic and synorogenic supracrustal sedimentary units such as those in the Beardmore and Byrd groups. Thus different units may record different parts of the orogenic history, and evidence of polyphase deformation in any one unit does not necessarily imply multiple orogenic cycles.

The Pensacola Mountains may show a similar tectonic history, where, for example, Rowell et al. [1992] recognized multiple Middle Cambrian to Early Ordovician deformations on the basis of stratigraphic and facies relationships. In that area, the Patuxent Formation, a graywacke sequence traditionally considered to be Neoproterozoic in age, is commonly correlated with other Paleozoic siliciclastic rocks such as the Beardmore Group. Rowell et al. [1992] inferred that most of the Patuxent graywackes are, in fact, a lower Paleozoic sequence deposited outboard of the Lower Cambrian Schneider Hills Limestone, both of which were deformed during the early Middle Cambrian, prior to deposition and deformation of a Middle Cambrian carbonate platform (Nelson Limestone). Although the eastern (inner) Patuxent sequence is only constrained to be pre-Middle Cambrian, recent data confirm that the western (outer) sections are Early to Middle Cambrian in age [Rowell et al., 1994; Millar and Storey, 1995; Van Schmus et al., 1995]. These new age data suggest that successive coaxial deformations are quite closely spaced in time and

Figure 8. Structural relations in the upper Nimrod Glacier area, summarized on stereonets for basement and supracrustal domains. Beacon Supergroup sedimentary rocks are omitted. Nimrod Group ductile L-S tectonites in the Geologists and Miller ranges show uniform orientation of elongation lineation (L_e) and foliation (S) deformed by gently NW plunging open folds [from Goodge et al., 1993a]. Structures in Beardmore Group in Cobham Range and Kon-Tiki Nunatak are from this paper (Figure 5). Beardmore Group cleavages (shown as synoptic S_i and S_j) at Cotton Plateau are from Stump et al. [1991]. Kinematic features in Nimrod Group metamorphic tectonites indicate top-to-the-southeast ductile translation at deep crustal levels [Goodge et al., 1992, 1993a], whereas anchizone to lower greenschist facies Beardmore Group structures are consistent with west directed shortening at high crustal levels.
may not represent different orogenic events [Rowell et al., 1992; Storey et al., 1992, 1996].

Neoproterozoic to Early Paleozoic Tectonics of the Ross Margin

Regional Structural Relations

Structural relations among basement and supracrustal units in the Nimrod Glacier region are summarized in Figure 8. Large and small scale structures in rocks of the Byrd and Beardmore groups indicate a general E-W contraction, and structural relations in the Beardmore Group indicate a westward vergence. Metamorphic mineral assemblages and textures in the Beardmore Group record regional greenschist facies metamorphism overprinted by an albite-epidote to hornblende hornfels facies contact event, indicating structural depths of <10 km during deformation. In contrast, the orientations of ductile tectonite fabrics in the Nimrod Group [Goode et al., 1993a] indicate a sinistral-oblique sense of basement shear (top-SE). Synkinematic metamorphic mineral assemblages indicate that ductile movement within the Nimrod Group occurred at depths of about 35-40 km, following a cycle of deep-crustal eclogite formation [Goode et al., 1992; Peacock and Goode, 1995]. Geochronologic data indicate some overlap in time between deformation of the Beardmore and Nimrod groups (about 585-500 Ma for Beardmore Group deformation and 540-520 Ma for peak-temperature Nimrod displacement), suggesting temporal and orogenic links between the kinematically distinct deformations (Figure 8).

In order to compare deformational events in the Beardmore and Nimrod groups, it is necessary to consider them at the appropriate relative crustal levels. The central Transantarctic Mountains are dissected by range-parallel Cenozoic normal faults with generally nonrotational motion, although some faults may be pre-Devonian in age. Individual fault systems have down-to-the-east (toward the Ross Ice Shelf) displacements of up to 5000 m [Barrett, 1965], and cumulative vertical displacements across the core of the range may be as much as 2000 m [Fitzgerald, 1992]. In the upper Nimrod Glacier area, the Kukri penaneal at the base of the Beacon Supergroup is offset in a down-to-the west sense along several faults (Figure 1), but its absence in the Miller Range suggests a down-to-the-east normal fault underlies Marsh Glacier. On the basis of inferred P-T conditions of metamorphism, Beardmore and Byrd group rocks restore to relatively high structural levels outboard of deeper-level Nimrod basement, and structures within them likely reflect their Neoproterozoic to Early Ordovician orientations except for vertical displacement. Given this assumption, the Beardmore structures indicate a dominantly west directed contractional deformation at higher crustal levels relative to nearly orocline-parallel movement within the deeper Nimrod Group.

Age Relations

Timing relationships with respect to regional events are summarized in Figure 9. Detrital zircon ages from stratigraphically lower portions of the Beardmore Group indicate that deposition was principally younger than 650 Ma and was in

![Figure 9. Time correlation diagram, showing ages of Neoproterozoic to early Paleozoic events in the central Transantarctic Mountains. Sources of age data are as follows: 1 from Borg et al. [1990]; 2 from Rowell et al. [1993]; 3 inferred from ages of detrital zircons in Beardmore Group with igneous morphology, see 4; 4 from Walker and Goode [1994], Walker [1996], and N. W. Walker (personal communication, 1994); 5 from Rowell and Rees [1989], Pannaja and Rees [1991], and Rees and Rowell [1991]; 6 from this paper; 7 from Goode et al. [1993b]; 8 from Adams et al. [1982], and Goode and Dallmeyer [1992, 1996]; 9 from Van Schmus et al. [1995]; 10 from Gunnar [1976], Adams et al. [1982], and Borg et al. [1990]. Timing of Laurentian rift-margin development is from Bond et al. [1984].]
part younger than 585 Ma [Walker and Goodge, 1994; Walker, 1996; N.W. Walker, personal communication, 1994]. An upper age limit on Beardmore deposition is unconstrained and could have persisted to the very latest Neoproterozoic at ~550 Ma. A Sm-Nd isochron age on a gabbronoritic unit within the Goldie Formation (762 Ma [Borg et al., 1990]) and a Nd mantle-separation age on pillow basalt from the Skeleton Group (800-700 Ma [Rowell et al., 1993]) suggest that deposition of the siliciclastic units could have spanned a considerable time period. The presence of young, pristine, apparent first-cycle igneous zircons among older age populations in the Beardmore Group (N.W. Walker, personal communication, 1995), suggests a mixed provenance that included the East Antarctic craton and an episodically active volcanic system, perhaps related to extensional rifting through the middle to late Neoproterozoic. This is consistent with bimodal basalt-rhyolite layers that occur sporadically through the Goldie Formation and correlative units farther south [Wade and Cathey, 1986; Stump, 1995; Van Schmus et al., 1995]. Beardmore Group deformation therefore appears to be post 585 Ma and could either predate or be synchronous with ductile Nimrod Group deformation. Lower Cambrian Byrd Group limestones represent the oldest sediments in the region that contain macrofossils, but they are not the first carbonates. Calcareous sections of the Cobham Formation record Neoproterozoic inner continental-shelf sedimentation [Laird et al., 1971], perhaps on a transient, unstable early platform associated with a subsiding rifted continental margin. Shelf sedimentation continued through the Early Cambrian, and these sedimentary rocks were probably deformed by the Middle or Late Cambrian. Discrete pulses of Byrd Group deformation may have persisted into the earliest Ordovician [Rees et al., 1987]. Nimrod Group tectonites record principal ductile shear between about 540 and 520 Ma, although this deep-crustal movement probably began somewhat earlier. Synkinematic to postkinematic mineral cooling ages from the Nimrod tectonites indicate a period of cooling through the Middle Ordovician, roughly synchronous with the main phase of postkinematic granite plutonism in the region. Although some age constraints are not tight, events recorded by each of these rocks units document a linked history of deposition, deformation, magmatism, and crustal cooling over a period of at least 100-150 m.y. I suggest that use of the term "Ross cycle" (similar to the broader term "Pannotia cycle" suggested by Stump [1987] to unify events along the southern margin of Gondwanaland) is appropriate to describe this multifaceted sequence of events.

An important consequence of recently discovered young detrital zircon ages in the Beardmore Group is that Beardmore deposition may be nearly coeval (within 40 m.y.) with that of the stratigraphically lowest unit in the Byrd Group, the Lower Cambrian Shackleton Limestone. Although some of the Beardmore sediments are undoubtedly older than the Shackleton Limestone, it is also possible that part were deposited at about the same time as the Shackleton in different depositional settings across a continental margin, as was recently postulated for a carbonate-siliciclastic sequence in the Putsacula Mountains [Rowell et al., 1992]. In contrast to the traditional view that the Beardmore Group represents a Proterozoic elastic association significantly predating development of the lower Paleozoic carbonate platform, the Goldie turbidites may represent latest Proterozoic to early Paleozoic deep-water deposition across the East Antarctic continental slope and rise outboard of a discontinuous and time-transgressive carbonate shelf. Calcareous units in the Cobham Formation indicate that at least an incipient carbonate shelf existed during the latest Neoproterozoic and perhaps was succeeded by the younger Shackleton Limestone. By considering the Beardmore and Byrd groups to be broadly synchronous, a Lower Cambrian deep water basinal sequence might be inferred for the central Transantarctic Mountains, as elsewhere [see Rowell and Rees, 1989; Rowell et al., 1992], and it explains a continental component in the isotopic character of Beardmore sediments [Borg et al., 1990; Borg and DePaolo, 1991]. This problem should be addressed by sedimentological analysis of the Beardmore Group; however, if it is true that the Beardmore and Byrd sequences are of similar age, then surely they were deformed during the same orogenic cycle.

**Tectonic Model for Beardmore Group Deformation**

What tectonic processes could lead to a contractional deformation of the Beardmore Group between about 600 and 500 Ma? The data presented here and regional geologic relationships are not easily explained in terms of collision, foreland thrust-belt deformation, or forearc convergence, all classical cases of shortening involving plate convergence. First, there are no crustal elements demonstrated to have collided with this margin of East Antarctica during the Neoproterozoic and early Paleozoic. Second, a foreland thrust belt is unlikely, because the modest observed strain patterns were probably not associated with great horizontal displacements and because one would expect older-over-younger, or at least deep water over shallow water facies, structural sequences to develop. Third, forearc contraction is also implausible in this case because one would expect intra- to outer-arc shortening to have a vergence toward the subducting plate, rather than toward the craton.

The structural data presented in this paper may be explained by a tectonic model in which Beardmore rocks were deformed by modest west vergent contraction during latest Proterozoic initiation of the Ross orogenic cycle. I propose a four-stage tectonic history for the East Antarctic margin, involving early rifting and passive-margin development, tectonic inversion of the continental margin sequences, initiation of a subduction boundary between East Antarctica and the proto-Pacific ocean, and, finally, development of a mature oblique-convergent plate margin (Figure 10).

### Stage 1: Rifting (About 700-600 Ma)

The first stage involved rifting of the Proterozoic Rodinian supercontinent, which included East Antarctica. The general configuration of the resulting East Antarctic rift margin (Figure 10a), inferred largely from sedimentological relations, was characterized by structural thinning of cratonal basement by down-to-the-east normal faults and a thin carbonate-shale platform deepening out to a continental slope-rise submarine fan sequence. At this time, carbonate deposition on a subsiding shelf and turbidite deposition across the outer rise are represented by the Cobham and Goldie formations, respectively. At this stage, East Antarctica was separated from
A. Rift stage (about 700-600 Ma)

B. Basin inversion stage (about 600-550 Ma)

C. Subduction initiation stage (about 550 Ma)

D. Convergent-margin stage (about 550-450 Ma)

Figure 10.
another continental plate, possibly Laurentia, by a narrow spreading ocean basin as the conjugate passive margins contracted thermally. Spreading was most likely perpendicular to the rift margin. The age of synrift to postrift Beardmore Group sediments suggests that rifting was underway by at least 650 Ma and that marginal-basin deposition may have continued for the next 50-100 m.y. Isotopic evidence from some mafic igneous units indicates that initial rifting may have occurred earlier (~750 Ga), but the stratigraphic relations of these rocks with the sedimentary sequences is somewhat uncertain [Borg et al., 1990; Rowell et al., 1993]. The close correspondence between detrital zircon ages from the Beardmore Group and known cratonic ages from East Antarctica (see summary by Tingey [1991]) suggests that the provenance for Beardmore detritus was confined to East Antarctica, with no contribution from Laurentia [Walker, 1996].

Stage 2: Basin Inversion (About 600-550 Ma)

Between about 600 and 550 Ma, Beardmore sediments experienced a modest contractional deformation characterized by upright, open folds with a cratonward vergence (Figure 10b). This early contractual phase may represent basin inversion resulting from compressional forces related to ocean spreading [e.g., Withjack et al., 1995]. The known outcrop and structural relations do not constitute proof of structural inversion, yet the plausibility of this scenario is discussed below. Such a mechanism accounts, however, for west vergent contraction by reactivation along earlier extensional structures, modest bulk crustal strain, and a short time span between rifting and major shortening during the peak of Ross orogenesis. Furthermore, the ultimate driving force of this marginal-basin inversion is oceanic spreading, not plate convergence. Although the present age constraints permit this deformation stage to be as early as ~585 Ma, I suggest it was relatively young (see Figure 9) in order to allow time for sediment accumulation and thermal subsidence of the continent-ocean rift transition.

Stage 3: Subduction Initiation (About 550 Ma)

As spreading continued in the proto-Pacific ocean, the global plate-kinematic balance led to the initiation of subduction along the East Antarctic continental margin (Figure 10c), a change that perhaps was triggered by events shaping Gondwanaland. Subduction may have been spawned by compressional stresses concentrated in the transitional crust of the craton margin during early stages in spreading [Dewey, 1988; Bott, 1992], although the plate may have broken further outboard in mechanically weaker oceanic lithosphere. Although early contraction in the previous stage was likely normal to the rift margin, subsequent subduction (freed of mechanical constraints imposed by plate coupling) may have initiated at an oblique angle following reorganized plate kinematics. During this time, there was continued carbonate deposition on what remained of the submergent part of the craton, and shortening continued in the inner parts of the marginal-basin sequence. Some early contractual faults may have become inactive during this stage.

Stage 4: Steady State Convergence (About 550-450 Ma)

Between 550 and 450 Ma, protracted subduction led to a steady state condition of plate convergence (Figure 10d). By this stage, the East Antarctic margin had undergone a complete transformation from rifting to subduction. The continental-margin system was characterized by synchronous continental-margin magmatism, continued contraction in the forearc and

Figure 10. Schematic model of Neoproterozoic to early Paleozoic tectonic evolution of the Ross orogen in the vicinity of the central Transantarctic Mountains. Approximate scale is as shown, with no vertical exaggeration; horizontal distances between cratonic margin and oceanic ridges are diagrammatic in Figures 10a and 10b. Lithospheric mantle is light gray. (a) Rift stage (about 700-600 Ma) during separation of East Antarctic and Laurentian cratons as a result of spreading normal to rift margin. East Antarctic margin is blanketed by rift-margin prism of Beardmore Group shelf carbonates and slope-rise submarine fans, with shelf facies progradational over slope-rise sequence. Interval of principal passive-margin sedimentation occurs between about 650 and 550 Ma. (b) Basin inversion stage (about 600-550 Ma), in which rift-margin normal faults are reactivated as contractual structures in response to compressional stresses localized across transition from continental to oceanic crust. Deformation of rift-margin sequence is moderate, consisting of upright, craton-vergent folds cored by reactivated reverse faults, principally affecting Beardmore Group. (c) Subduction initiation stage (~550 Ma), driven by global plate-kinematic balance. Global plate motions [see Dalziel, 1992] may have resulted in oblique subduction at an early stage. Note some early contractual faults become inactive at this stage, but inboard upper plate deformation continues. (d) Convergent-margin stage (about 550-450 Ma), consisting of oblique subduction of paleo-Pacific oceanic lithosphere beneath East Antarctic margin of fledgling Gondwanaland. Initial stages of oblique plate convergence lead to continued shortening in high-level supracrustal sequences, strike-slip faulting at high crustal levels, and sinistral orogen-parallel shear at deeper levels within an incipient magmatic arc (Nimrod Group shear zone). Note continued sedimentation across broken shelf through Middle Cambrian (~520 Ma) and development of a continental-margin magmatic system (dominant between about 540 and 490 Ma). As continental-margin upland is developed, erosion leads to formation of syntectonic Middle Cambrian to earliest Ordovician alluvial-fan and braid-plain deposits (upper Byrd Group). By the Late Ordovician, steady state subduction relieves compression on upper plate East Antarctic margin and may have led to intra-arc and forearc extension. Eventually, contractual deformation is transferred to the accretionary complex at the upper plate leading edge; an accretionary complex is absent in the present Transantarctic Mountains and may either be submerged beneath the Ross Sea or was removed by subsequent rifting to western Marie Byrd Land or New Zealand.
along the axis of the orogenic system, intra-arc and forearc
denudation, and sinistral arc-basement strike-slip shear. 
Orogen-parallel ductile deformation in the Nimrod Group and
some early plutonic rocks, representing deep crustal levels at
high temperatures, appears to record along-strike ductile flow
that is most easily explained by subduction with a sinistral
Continued shortening of the supracrustal units indicates that
sedimentary and basement units were mechanically detached
from one another during this period. Together, these two
different kinematic records, coupled with differences in
inferred crustal level, are similar to modern convergent-margin
systems involving oblique subduction and strain partitioning
between high-level sedimentary rocks deformed dominantly
by contraction in a forearc position and deeper-level intra-arc
strike-slip shear. Partitioned strain as a manifestation of
sinistral-oblique plate convergence during Ross time is
compatible with the predicted plate motions during early
spreading between East Gondwanaland and Laurentia derived
from paleomagnetic and paleogeographic studies [e.g.,
Dalziel, 1992]. This stage represents the culmination of Ross
orogenic activity, including deformation, metamorphism,
magmatism, and syntectonic sediment deposition. 
Ultimately, steady state convergence may have led to intra-arc and
forearc extension and a transfer of contractional deformation
to the leading edge of the East Antarctic plate, marking the end of
Ross orogenic activity in the early Paleozoic.

Implications of Structural Inversion

As outlined here, this model of basin inversion during early
stages in Ross activity explains: (1) the relatively modest
structural shortening in the Beardsmore Group; (2) the
cratonward vergence of deformation, presumably controlled by
reactivation along earlier extensional structures; (3) how the
early phase of contraction relates to early ocean-basin
spreading and ultimately links to Ross convergence; (4) isotopic variations in post-tectonic granites, which imply
eastward crustal thinning; and (5) the similar but somewhat
diachronous timing of supracrustal and crystalline basement
deformation. In more general terms, the complete evolution
from rifting to convergence as illustrated in Figure 10
provides a unifying framework with which to explain the
variations in tectonic setting of sediment deposition, the
different kinematics of deformation at different crustal levels
as a result of strain partitioning, and the transformation from
passive- to convergent-margin dynamics.

Many examples of tectonic inversion occur in settings
characterized by plate-margin convergence [e.g., Lowell.
1995]. However, Withjack et al. [1995] proposed a mechanism
of rift basin inversion following passive margin development
as a result of two principal factors: ridge-push forces and
continental resistance to plate motion. In the Fundy basin of
eastern Canada, Withjack et al. [1995] emphasized that struc-
tural inversion occurred in the absence of collision or
subduction, and it can only be attributed to the processes of
North Atlantic seafloor spreading. Age constraints are limited,
but structural inversion in the Fundy basin, from the time of
latest synrift sediment deposition to contraction, apparently occurred within a period of about 30-100 m.y.
Such a time span is of the same order of magnitude as that
given by present constraints on Beardsmore Group sediment-
ation and deformation. The Mesozoic evolution of the Fundy
basin could therefore provide a plate-tectonic analog for
deformation of the Beardsmore Group. Beardsmore sediments
most likely represent rift-margin turbidites deposited during
the latest Neoproterozoic to earliest Paleozoic. Deformation
probably occurred substantially after the main episode of
crustal rifting but perhaps as little as 50 m.y. after the end of
Beardsmore sedimentation, and it likely occurred during
development of a late Neoproterozoic Pacific ocean basin.
Using the Fundy basin evolution as an example, Beardsmore
Group deformation may have occurred during or shortly after
the rift-to-drift transition, as a result of mid-ocean ridge
boundary forces, rather than as a result of collisional or sub-
duction-related events. At the time of Fundy basin inversion
(about 170-120 Ma), the distance between the spreading
Atlantic ridge and the North American margin was as little as
500-600 km; this suggests that structural inversion of the
Beardsmore marginal-basin sediments could have occurred soon
after deposition at the edge of a relatively narrow ocean basin.

It has been suggested that Neoproterozoic rifting of the
Rodinian supercontinent resulted in the separation of East
Antarctica/Australia from Laurentia [Dalziel, 1991, 1992;
Moore, 1991; Hoffman, 1991]. One criticism of this concep-
tual model is that the subsequent Neoproterozoic to early
Paleozoic histories of the East Antarctic and western North
American margins appear to be quite different; the Pacific East
Antarctic margin between about 550 and 450 Ma was probably
one of plate convergence, whereas western North America
remained a passive margin through this period [Stewart, 1972;
Bond et al., 1984]. If these two plate margins were conjugate
to one another, why was one involved in plate consumption
so quickly while the other was continuing to drift after initial
rifting? In part, an answer may stem from one of the points
raised by Withjack et al. [1995] with respect to the Fundy
basin, that of the rifted continent’s resistance to plate motion.
In the Fundy case, resistance of North American continental
lithosphere to westward motion in the Mesozoic may have
contributed to the onset of compressive forces along the
continent-ocean transition. In this case, East Antarctic con-
tinental lithosphere presently comprising the Transantarctic
Mountains lay along the outer proto-Pacific edge of the
coalescing Gondwanaland supercontinent [e.g., Elliot, 1975;
Stump, 1987]. What developed initially as a Rodinian rift
margin in Proterozoic time [Hoffman, 1991; Dalziel, 1992]
behind it a massive amalgamating supercontinent (Figure
11). On the other side, Laurentia was drifting away from
Rodinia, was relatively isolated at least up to the latest
Proterozoic [Dalziel et al., 1994], and had few outer plate-
boundary forces acting upon it. The basin inversion model
therefore also explains why early contractional deformation of
the Beardsmore (and even lower Byrd?) supracrustal assem-
lages, believed to have occurred in the absence of subduction
or collision, was coincident with passive-margin subsidence
in western North America.

Storey et al. [1992] regarded this asymmetric development of
the two corresponding margins as a result of episodic
reversals in plate motions along the Antarctic margin.
Flöttmann et al. [1994] proposed tectonic inversion of
Neoproterozoic Adelaidean rocks in southeastern Australia
during the Delamerian orogeny (about 516-490 Ma). In their
view, basin reactivation was concomitant with subduction during Ross time. The basin inversion model outlined here differs in two regards: (1) it allows for marginal-basin contraction prior to the onset of transpressional convergence, and (2) it explains contraction as a result of the transition between rifting and convergence. This model successfully integrates structural, stratigraphic, and geochronologic data from several areas along the Transantarctic Mountains [Rowell et al., 1992; Storey et al., 1992, 1996] (see section on Neoproterozoic to Early Paleozoic Tectonics of the Ross Margin), and it is consistent with the independent conclusion that Beardmore Group deformation is but one manifestation of a protracted plate-tectonic continuum referred to as the Ross orogeny. In this view, the Ross orogeny is an expression of a coherent plate-tectonic cycle, including early rift-margin structural inversion and later plate-margin transpression.

A Final Note on the Beardmore Orogeny

The ideas presented in this paper represent a working, testable model for orogenesis in the Transantarctic Mountains; equally plausible models must explain the structural evidence for a single Beardmore deformation and the young apparent age of the Beardmore Group. Further study will help to resolve remaining stratigraphic, structural, and geochronological problems.

Regardless of the plausibility of the tectonic model outlined here, structural data indicate that deformation of the Beardmore Group is not easily distinguished from what has traditionally been referred to as the Ross orogeny. Grindley and Laird [1969] suggested nearly three decades ago that Beardmore Group deformation may have been linked to an early Ross phase. Recent reports from other areas suggest similar age and deformation patterns [Rowell et al., 1992; Storey et al., 1996]. In the sense that it is important to reserve the term orogeny to describe tectonic events that occur at distinctly different times and under different tectonic regimes, I caution that the so-called “Beardmore orogeny” may just be a deformational expression, perhaps at an incipient stage, of a broader cycle of Ross-age tectonic activity. It does not, on the evidence available, deserve separate orogenic status. As better age control becomes available for deformation of the Beardmore Group itself, its relationship to Ross events will be better understood. In the meantime, Beardmore Group deformation is best viewed in the context of linked events spanning the latest Neoproterozoic and early Paleozoic.

Conclusions

Geological field evidence and structural data reported in this paper indicate that Beardmore Group rocks in the central
Transantarctic Mountains experienced only a single, principally west vergent deformation. Contrary to earlier reports, no structural evidence for multiple deformations was found. U-Pb ages obtained from detrital zircons in the Goldie and Cobham formations indicate that the Beardmore Group is significantly younger than previously assumed (≤600 Ma). The detrital mineral ages suggest a mixture of Archean to Neoproterozoic sources, including a likely local volcanic source for the youngest population from the East Antarctic craton. The Beardmore Group may represent a deep water facies deposited at roughly the same time as the Byrd Group across a tectonically active East Antarctic rift margin, and it does not represent an allochthonous terrane of exotic origin.

In terms of postdepositional tectonic events, rocks of the Beardmore Group were not affected by a temporally and tectonically unique cycle of orogenesis, and I recommend that the term Beaddermore orogeny be abandoned. I speculate that Beardmore Group structural relations, when compared regionally with the Nimrod and Byrd groups, may be explained by contractional tectonic inversion along the newly created Neoproterozoic rift margin. A subsequent transition to left-oblique convergence of paleo-Pacific oceanic lithosphere beneath cratonic East Antarctica resulted in strain partitioning between deep-level strike slip and continued shallow-level contraction in the Beardmore and Byrd groups. Deformation of the Beardmore Group is therefore interpreted as an expression of marginal-basin inversion, following Rodinia rifting but prior to or during the onset of plate subduction leading up to the main phase of the Ross orogeny. The Ross orogenic cycle constitutes a broad and diachronous sequence of tectonic events during the latest Neoproterozoic to early Paleozoic manifested by different strain histories in rocks occupying different crustal levels. Discrete phases of Ross deformation are related to the tectonic transition from a rift margin to an oblique-convergent margin, and they are integrally related to the cycle of supercontinent breakup and reassembly.

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GOODGE: BEARDOORE INVERSION DURING ROSS OROGENESIS


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