Neoproterozoic-Cambrian basement-involved orogenesis within the Antarctic margin of Gondwana

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ABSTRACT

High-grade metamorphic tectonites of the Nimrod Group in the central Transantarctic Mountains compose a major ductile shear zone that formed within the paleo-Pacific margin of Gondwana. Despite demonstrated Precambrian protoliths, the timing of metamorphism and tectonite development has been poorly constrained. Igneous rocks of diverse compositions intrude the Nimrod tectonites. Four intrusive units with incipient to well-developed ductile fabrics yield U-Pb zircon ages of 541–521 Ma, and a nondeformed pegmatite has a U-Pb zircon age of ~515 Ma. These data show that early Paleozoic Ross magmatism was compositionally, texturally, and temporally more heterogeneous than previously recognized. Fabrics in the igneous rocks are concordant with those in their host tectonites, indicating that Nimrod tectonism was in part synchronous with plutonism. U-Pb ages of 525–522 Ma for metamorphic monazite from two pelitic tectonites support this interpretation. Thus, ductile deformation was in its peak to waning stages between about 540 and 520 Ma. This timing provides compelling evidence for transcurrent basement involvement in oblique plate convergence along the Neoproterozoic to Early Cambrian Antarctic margin of Gondwana.

INTRODUCTION

The Transantarctic Mountains represent a fundamental tectonic boundary that separates Precambrian cratonal basement and cover of East Antarctica from a complex mosaic of Precambrian to Mesozoic terranes in West Antarctica (Fig. 1). This ~3000-km-long belt is generally considered to be coincident with the paleo-Pacific margin of the East Antarctic craton. Prior to the breakup of Gondwana, this margin was the locus of a complex history of Proterozoic to early Paleozoic orogenesis, culminating in the Cambrian-Ordovician Ross orogeny, a pronounced tectonic event that was synchronous with Pan-African suturing of the Gondwana supercontinent (Kennedy, 1964; Stump, 1987) and that overprinted Neoproterozoic to early Paleozoic deformations. The tectonic expression of these events, including contraction and translation or transpression, varies along the orogen.

High-grade metamorphic rocks of the Nimrod Group in the central Transantarctic Mountains (Fig. 1) exhibit deformational features attributed to the Nimrod orogeny (Grindley, 1972). Structural studies have shown that these rocks were deformed within a wide ductile shear zone in response to orogen-parallel tectonic displacements (Goodge et al., 1991a; Hansen and Goodge, 1991). On geologic grounds this deformation is pre-Ordovician, but previous geochronologic data permitted the deformation to be as old as Mesoproterozoic (Grindley and McDougall, 1969; Goodge et al., 1991a). We present U-Pb ages of metamorphic monazite from Nimrod Group pelitic schists and igneous zircon from variably deformed igneous rocks that intrude Nimrod hosts which constrain Nimrod deformation to the latest Proterozoic and Early Cambrian.

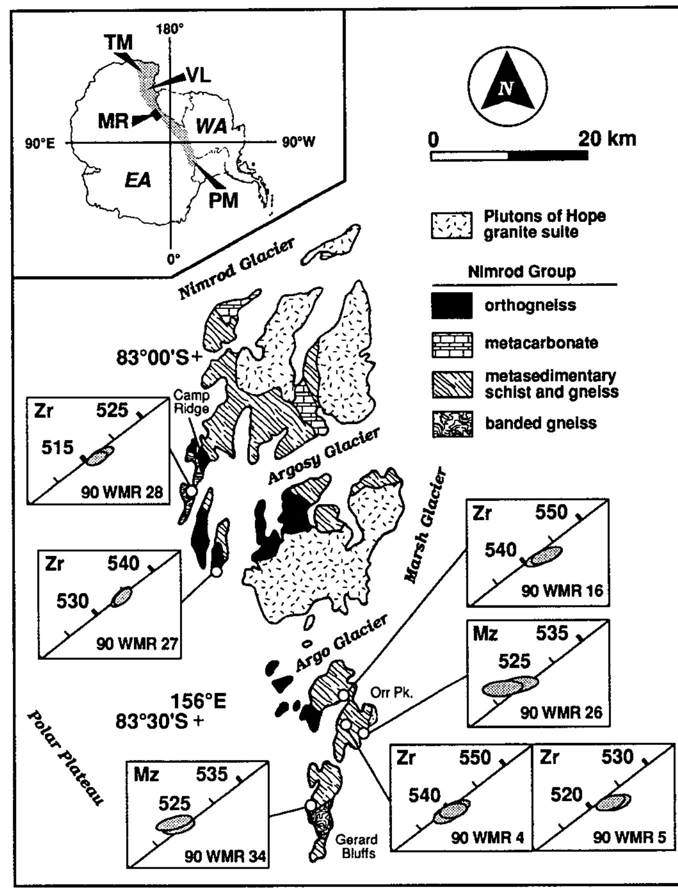

Figure 1. Geologic map of Miller Range (Grindley, 1972; Goodge et al., 1991a) showing location of U-Pb samples. Some igneous units are too small to display at this scale. Inset shows location of Miller Range (MR), Pensacola Mountains (PM), and Victoria Land (VL) within Transantarctic Mountains (TM). EA = East Antarctica; WA = West Antarctica. Zircon (Zr) and monazite (Mz) U-Pb data are shown on concordia plots (for each, abscissa is $^{206}Pb/^{238}U$, ordinate is $^{207}Pb/^{235}U$, and ages are in Ma).

GEOLOGIC SETTING

The Nimrod Group is an outlier of the East Antarctic craton exposed in the Miller and Geologists ranges of the central Transantarctic Mountains (Fig. 1). These rocks are separated from a lower grade supracrustal assemblage to the east by unknown structures concealed by glaciers. The supracrustal assemblage includes low-grade clastic and calcareous metasedimentary rocks of the Neoproterozoic Beardmore Group and lower Paleozoic carbonate and siliciclastic rocks of the Byrd Group (Lower to Middle Cambrian Shackleton Limestone and Douglas Conglomerate). Nondeformed plutons of the ~500 Ma Hope granite suite intrude all units.

The Nimrod Group includes pelitic schist, amphibolite, metacarbonate, quartzofeldspathic and mafic layered gneiss, migmatite, metaperidotite, relict eclogite, and orthogneiss (Grindley et al., 1964; Goodge et al., 1991a). Deformation during the Nimrod orogeny and synchronous northeast displacement along the Endurance thrust fault (Grindley, 1972) were reinterpreted by Goodge et al. (1991a) as a result of top-to-the-southeast ductile shear. Pervasive L-S (L = elongation foliation, S = foliation) tectonite fabrics formed in a southwest-dipping, sinistral ductile shear zone ~12–15 km thick (Hansen and Goodge, 1991), under metamorphic conditions of 650–750 °C and 8–14 kbar (Goodge et al., 1992).

U-Pb ages of detrital zircons demonstrate an Archean to Proterozoic provenance for clastic Nimrod protoliths (Gunner and Mattinson, 1975; Gunner, 1983; Walker and Goodge, 1991). A $^{207}Pb/^{206}Pb$ date of 1.72 Ga from an L-S tectonite orthogneiss (reported in Goodge et al., 1991a) provided a Paleoproterozoic maximum age for ductile tectonism. Amphibole K-Ar dates of 1.05–1.00 Ga were interpreted to record a Neoproterozoic orogenic event (Grindley and McDougall, 1969). However, amphibole and mica from Nimrod tectonites yielded $^{40}Ar/^{39}Ar$ dates of 525–485 Ma (Goodge and Dallmeyer, 1992), showing that the K-Ar ages are unreliable due to excess radiogenic argon. Rb-Sr and K-Ar dates of 520–450 Ma from Nimrod metamorphic rocks (McDougall and Grindley, 1965; Grindley and McDougall, 1969; Adams et al., 1982) probably record a thermal overprint due to regional Hope magmatism (Gunner, 1976; Pankhurst et al., 1988; Borg et al., 1990). Rb-Sr and U-Pb ages of 500–463 Ma from postkinematic plutons provide an upper limit on the age of regional deformation (McDougall and Grindley, 1965; Gunner and Mattinson, 1975; Gunner, 1976; J. M. Mattinson, 1991, personal commun.). Thus, existing age data only limit ductile Nimrod deformation to between about 1.7 and 0.5 Ga.

U-Pb GEOCHRONOMETRY

Nimrod tectonites in the Miller Range are intruded by a compositionally diverse suite of plutons, dikes, and sills (Goodge et al., 1991b) that are characterized by variable development of solid-state deformation fabrics. In order to determine the timing of tectonite development, we dated zircon from several igneous units of this suite and individual crystals of metamorphic monazite from two Nimrod pelitic tectonites.

U-Pb analytical procedures were similar to those described by Walker (1992). Analytical data are given in Table 1 (see Fig. 1). Some of the zircon data are slightly discordant; we attribute this to small amounts of inheritance rather than to Pb loss. Thus, we accept the U-Pb ages as the best estimate of the zircon ages. Three of four monazite analyses show slight reverse discordance; for these we interpret the $^{207}Pb^{*}/^{235}U$ (see Table 1) date as the monazite age.

Orthogneisses and Plutons

A coarse-grained orthoclase-augen biotite orthogneiss (90 WMR 4) has an L-S tectonite fabric that is kinematically consistent with that of host tectonites. Four zircon fractions yield concordant to slightly discordant ages of 540–538 Ma. These data indicate that the orthogneiss protolith was emplaced at about 540–538 Ma and was synchronously and/or subsequently deformed.

A moderately foliated garnet-biotite pegmatite (90 WMR 5) crosscuts the tectonite fabric of the orthogneiss described above. Porphyroclastic orthoclase shows deformation bands and microfaults, but matrix quartz shows only minor subgrain formation, indicating that weak solid-state deformation followed pegmatite solidification. Two zircon fractions yield U-Pb ages of 524–521 Ma.

A hornblende-biotite tonalite (90 WMR 16) with abundant mafic enclaves is generally poorly foliated but locally displays well-foliated margins concordant with Nimrod shear fabrics. This tonalite exhibits minor evidence of intracrystalline strain in feldspar, quartz, biotite, and hornblende. Two zircon fractions yield slightly discordant U-Pb ages of 543–541 Ma, interpreted as the time of late synkinematic deformation.

Two zircon fractions from a nonfoliated but weakly lineated hornblende-biotite granite (90 WMR 27) yield concordant U-Pb ages of 534 Ma. This granite contains orthoclase megacrysts surrounded by discontinuous mosaics of quartz showing subgrain formation, indicative of weak solid-state deformation. These features indicate that the granite was emplaced during the waning stages of deformation.

Two zircon fractions from a biotite granite pegmatite sill (90 WMR 28) that is concordant to layering in its host gneiss have slightly discordant U-Pb ages of 516–515 Ma. This sill exhibits a decussate, serrated contact with enclosing gneisses, and the two are separated by a prominent biotite selvedge. These features, coupled with the absence of solid-state deformation textures observed in other samples, indicate that this sill was generated as an in situ melt during the final stages of dynamothermal tectonism.

Pelitic Schist

U-Pb ages were obtained for individual monazite crystals from two samples of pelitic schist (90 WMR 26, 90 WMR 34) that contain

TABLE 1. ZIRCON AND MONAZITE DATA, MILLER RANGE, TRANSANTARCTIC MOUNTAINS

Sample	Fraction properties$^{\phi}$	Amount analyzed (mg)	Concentration+ (ppm)		Pb isotopic composition#			Age and uncertainty ** (Ma)		
			Pb*	U	208/206	207/206	206/204	206Pb*/238U	207Pb*/235U	207Pb*/206Pb*
90 WMR 4	Zr, nm 1°, -100 +75 μm, ab	1.6	93.6	1089	0.08806	0.05862	49 504	539.5 ±0.3	539.6 ±0.6	542.6 ±0.8
	Zr, nm 2°, -100 +75 μm, ab	1.7	49.6	574.2	0.09047	0.05855	59 172	540.2 ±0.3	540.4 ±0.5	541.3 ±0.7
	Zr, m 2°, -100 +75 μm, ab	1.6	96.7	1128	0.08882	0.05859	47 393	538.3 ±0.3	538.7 ±0.5	540.1 ±0.8
	Zr, m 2°, -75 +64 μm, ab	1.0	46.5	540.2	0.09324	0.06097	5 492	539.8 ±0.5	540.3 ±0.6	542.2 ±1.4
90 WMR 5	Zr, nm 3°, -75 +64 μm, ab	0.7	13.0	162.9	0.04827	0.05917	14 837	521.5 ±0.3	524.4 ±0.5	537.1 ±1.2
	Zr, nm 3°, -100 +75 μm, ab	0.1	27.7	344.8	0.07490	0.06735	1 555	521.0 ±0.6	523.1 ±0.9	529.8 ±2.8
90 WMR 16	Zr, m 2°, -100 +75 μm, ab	1.0	24.2	279.1	0.10177	0.06195	4 148	540.9 ±0.4	542.4 ±0.6	546.6 ±3.2
	Zr, m 3°, -100 +75 μm, ab	1.2	32.0	372.9	0.08823	0.06136	5 208	541.0 ±0.4	543.2 ±0.6	551.7 ±3.1
90 WMR 27	Zr, m 3°, -100 +75 μm, ab	1.2	20.6	233.1	0.13792	0.06027	6 906	534.1 ±0.9	534.3 ±1.0	535.6 ±1.8
	Zr, m 3°, -75 +64 μm, ab	1.0	29.5	333.4	0.13727	0.05891	17 225	534.0 ±0.3	533.8 ±0.5	533.6 ±0.9
90 WMR 28	Zr, m 7°, -100 +75 μm, ab	0.6	160	1927	0.11584	0.06146	3 876	515.1 ±0.3	516.0 ±0.4	521.6 ±2.3
	Zr, nm 7°, -75 +64 μm, ab	0.9	130	1543	0.14814	0.07317	1 071	515.7 ±0.7	516.8 ±0.7	521.8 ±5.6
90 WMR 26	Mz	~0.01	925	4097	2.039	0.11989	234.2	524.9 ±1.4	524.3 ±2.4	521.7 ±16.3
	Mz	~0.01	1309	4375	3.018	0.08904	461.8	523.6 ±1.0	521.3 ±1.9	512.6 ±8.9
90 WMR 34	Mz	~0.01	1403	5001	2.772	0.08169	610.9	524.6 ±1.1	524.9 ±1.8	525.2 ±7.3
	Mz	~0.01	1440	4987	2.875	0.07917	681.8	525.6 ±1.2	524.7 ±1.2	523.6 ±6.1

$^{\phi}$Zr = zircon, Mz = monazite, nm = nonmagnetic, m = magnetic, ab = abraded. Values in micrometers indicate size range of zircons prior to 24 to 48 h abrasion.

+Asterisk denotes radiogenic Pb corrected for common Pb in zircon and laboratory blank Pb. Isotopic composition of common Pb is estimated from Stacey and Kramers (1975) but an uncertainty of ±0.1 is assigned to the 207/204 ratio. Total procedural blanks are ~2 picograms for U and 8 - 38 picograms for Pb. Because a mixed U-Pb tracer was employed, uncertainty in the weights of monazite grains affect only concentration data, not calculated U-Pb ages.

#Measured isotopic ratios corrected for mass fractionation of ~ 0.11% per atomic mass unit based on replicate analyses of NIST SRM 981 and 982 and adjusted for small amount of ^{206}Pb in tracer.

**Decay constants: ^{238}U = 1.5513 E-10/yr ; ^{235}U = 9.8485 E-10/yr. Atom ratio $^{238}U/^{235}U$ = 137.88. Uncertainty in the calculated ages is stated at the two-sigma level and estimated from combined uncertainties in calibrations of mixed ^{205}Pb - ^{233}U - ^{235}U tracer, measurement of isotopic ratios of Pb and U, common and laboratory blank Pb isotopic ratios, Pb and U mass fractionation corrections, and reproducibility in measurement of NIST Pb and U standards.

the regional L-S tectonite fabrics. One sample (26) is a sillimanite-garnet-muscovite-biotite schist showing quartz-filled extensional fractures crossing aligned sillimanite bundles, and the other (sample 34) is a garnet-kyanite-biotite-muscovite schist containing tapered and bent kyanite crystals. We interpret the age data to indicate metamorphic growth of the monazites in the interval 525–522 Ma, and thus to directly record the age of Nimrod tectonism.

DISCUSSION

New field and geochronometric data from the Nimrod Group provide evidence for the following important conclusions. First, Cambrian-Ordovician plutonic rocks in the Nimrod Glacier area are compositionally, texturally, and temporally more heterogeneous than previously known. Our results, coupled with those of earlier workers, show that magmatism in the central Transantarctic Mountains occurred between ca. 540 and 500 Ma. Variably formed solid-state fabrics in igneous rocks from the Miller Range indicate a period of progressively dissipating tectonism during this period. Second, monazite U-Pb ages from Nimrod tectonites indicate high-temperature ductile deformation at ~524 Ma. The monazite ages are within the range of emplacement ages of the deformed igneous rocks, and they are slightly older than amphibole and mica $^{40}Ar/^{39}Ar$ cooling ages from Nimrod tectonites (Goodge and Dallmeyer, 1992). Third, these data demonstrate the involvement of Nimrod Group basement rocks during latest Proterozoic to early Paleozoic orogenesis. Fourth, it remains unknown when ductile deformation began and how long it lasted. All of the dated orthogneisses contain L-S tectonite fabrics that are weaker than their host gneisses, indicating that the *onset* of deformation preceded intrusion of the igneous bodies. In contrast, the 1.72 Ga Camp Ridge orthogneiss displays a well-developed S-C tectonite fabric concordant with its host (Goodge et al., 1991a), showing that this body was emplaced prior to deformation.

Our data thus provide evidence that Nimrod ductile deformation was as young as ca. 540–520 Ma. Correlating this deformation with regional tectonic events in the Transantarctic Mountains is hampered by two significant problems: (1) uncertainty regarding the precise age of the Precambrian-Cambrian time boundary and (2) regional spatial and temporal variation among deformation events that are traditionally viewed as an expression of Ross orogenesis. Recent evidence indicates that the Precambrian-Cambrian boundary may be 540 Ma or younger (e.g., Compston et al., 1992; Cooper et al., 1992). Our data indicate that Nimrod deformation occurred essentially at the Precambrian-Cambrian boundary, within present resolution. Because the Ross orogeny in the Nimrod Glacier area is classically defined by deformation of the Lower Cambrian Shackleton Limestone, placement of the Precambrian-Cambrian boundary is of fundamental importance in deducing whether the Shackleton and Nimrod deformations can be temporally correlated, or whether orogenesis was episodic during the late Precambrian and early Paleozoic. The Ross orogeny, however, can no longer be viewed as a single Middle to Late Cambrian deformation event. Studies along the Transantarctic belt between the Pensacola Mountains and Victoria Land (Fig. 1) emphasize the likelihood of multiple Neoproterozoic to Middle Cambrian deformation events (Schmidt et al., 1965; Rees et al., 1987; Rowell et al., 1991, 1992). In addition, the nature and age of Beardmore deformation (Laird et al., 1971; Stump et al., 1986) as a separate Neoproterozoic orogenic event are suspect (Rowell et al., 1986; Pankhurst et al., 1988). Thus, until more precise geochronometric data become available, the existing uncertainties in timing make it difficult to correlate specific events defined solely on the basis of either biostratigraphic or geochronometric data.

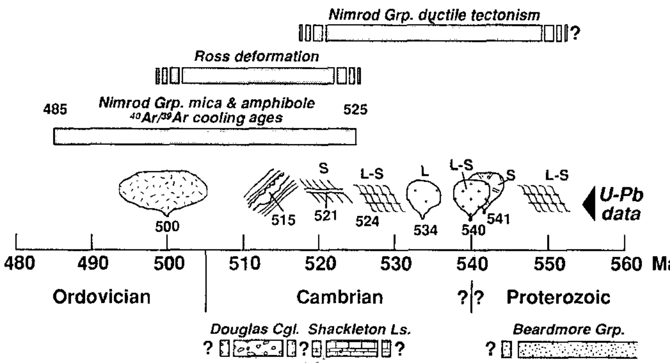

Figure 2. Time line showing correlation of geochronometrically and biostratigraphically dated tectonic events in Nimrod Glacier area. Ages are in Ma. All U-Pb data are from this paper, except those from undeformed Hope plutons; additional age constraints discussed in text. Sketches above time line show relation of fabrics in plutonic and schistose rocks (S = foliation; L = elongation lineation). Relative age of pre–540 Ma ductile deformation is inferred from geologic relations.

Despite these uncertainties, there appears to be temporal overlap between Nimrod ductile tectonism in the interval ca. 540–520 Ma and shortening of Byrd Group supracrustal rocks in the Middle to Late Cambrian (Fig. 2). This temporal overlap suggests a possible tectonic link, although no direct structural tie is permitted by present exposure. If deformation of the Shackleton and Nimrod rocks was cogenetic, what tectonic setting could simultaneously produce shallow-level contraction and deeper level orogen-parallel translation? We suggest that a transcurrent margin characterized by obliquely subducting paleo-Pacific(?) oceanic lithosphere beneath the East Antarctic craton may have been the dominant regime in this part of the orogen. Deformation of Nimrod basement occurred well after Neoproterozoic (~750 Ma) rifting along the East Antarctic margin inferred from sedimentological evidence. Whether this rifting event represents separation from Laurentia as postulated by Dalziel (1991) and Moores (1991) is uncertain: our data do not test the SWEAT hypothesis.

In the context of the latest Proterozoic to early Paleozoic margin of Antarctic Gondwana, many parts of the Transantarctic Mountains have been viewed in a contractional framework. Structural relations in northern Victoria Land (Fig. 1) indicate orogen-normal shortening across most of the belt (Gibson and Wright, 1985; Kleinschmidt and Tessensohn, 1987; Flöttmann and Kleinschmidt, 1991), although Ordovician(?) strike-slip motions have been suggested (Weaver et al., 1984; Bradshaw et al., 1985). Evidence for orogen-normal shortening of Neoproterozoic to lower Paleozoic supracrustal rocks continues through the Byrd-Nimrod-Beardmore sector of the central Transantarctic Mountains (Laird et al., 1971; Stump, 1981; Rees et al., 1987) to the Pensacola Mountains, where coaxial deformations were superimposed (Schmidt et al., 1965; Rowell et al., 1992). Shallow-level supracrustal deformation is thus of similar geometry along much of the orogen, indicating a broadly contractional convergent margin.

The constraints on Nimrod tectonism presented here provide perhaps the first clear evidence of orogen-parallel displacement associated with Cambrian tectonism postulated by others (Weaver et al., 1984; Rowell and Rees, 1990). The details of this Precambrian-Cambrian plate-margin history await further petrologic, structural, and geochronometric study, but orogenic patterns in the central Transantarctic Mountains differ from those in other parts of the orogen.

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