JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 95, NO. B5, PAGES 6647–6667, MAY 10, 1990

Isotopic Structure and Tectonics of the Central Transantarctic Mountains

SCOTT G. BORG, DONALD J. DEPAOLO, AND BRIAN M. SMITH

Berkeley Center for Isotope Geochemistry, Department of Geology and Geophysics, University of California, and Earth Science Division, Lawrence Berkeley Laboratory, Berkeley, California

Regional patterns of Nd, Sr, and O isotopic ratios of ~500 Ma granitic rocks are used to identify the ages and areal extents of three crustal provinces in the central Transantarctic Mountains. One of the provinces is the edge of the East Antarctic Craton, which isotopic analyses show is composed of Archean rocks thrust over Proterozoic rocks. The other two provinces compose the Beardmore microcontinent, which we deduce was allochthonous to East Antarctica and was emplaced in late Precambrian or early Paleozoic time. Evidence for a former ocean basin between the Beardmore microcontinent and East Antarctica is provided by basalt and gabbro of mid-ocean ridge character, dated by Sm-Nd at ~760 Ma, associated with marine sediments now located at the suture. The granitic rocks formed over a westward-dipping subduction zone that was active at ~500 Ma. The East Antarctic Craton is exposed in the Miller Range, which is a tectonic composite of reworked Archean and early Proterozoic material containing ~500 Ma peraluminous granites with model ages (T_{DM}) of 2.0 Ga, high $\delta^{18}O$ (+11 to +12‰) and high initial $^{87}Sr/^{86}Sr$ (0.7324 to 0.7417). East of the Marsh Glacier the granitic rocks are metaluminous to weakly peraluminous with model ages of 1.3 to 1.8 Ga, high $\delta^{18}O$ (+9 to +13‰) and lower $^{87}Sr/^{86}Sr$ (0.7068 to 0.7191). East of the Shackleton Glacier, gabbro, tonalite, diorite, and granodiorite have low $\delta^{18}O$ (+6 to +7‰), low initial $^{87}Sr/^{86}Sr$ (0.7045 to 0.7059) and high ε_{Nd} (+0.4 to +1.7). These isotopic provinces correspond to differences in age and composition of the middle and lower crust at the time of formation of the granitic magmas. The boundaries of the isotopic provinces also correspond to discontinuities in provenance, lithology, structural style, and grade of metamorphism of prebatholithic metasedimentary rocks. The isotopic data indicate that the granitic magmas were formed mostly by crustal anatexis in the areas west of the Shackleton Glacier. This is typical of early Paleozoic granitic batholiths elsewhere in the world and has led to speculation that subduction was not involved in granitic magmatism at that time of earth history. However, the granitic rocks located west of the Shackleton Glacier, by virtue of their mantle-like isotopic compositions and their association with metavolcanic rocks, appear to be subduction-related. The tectonic history deduced for the Gondwana margin, as represented in the central Transantarctic Mountains, began with deposition of sediments on an Atlantic-type rifted margin at ~760 Ma. The Beardmore microcontinent was most likely accreted in association with folding of the clastic sedimentary rocks before middle Early Cambrian time (550 Ma). Carbonate sedimentation and volcanism along the eastern margin of the Beardmore microcontinent commenced in Cambrian time. Folding and metamorphism of all older units occurred in late Cambrian time followed by emplacement of granitic rocks at ~500 Ma.

INTRODUCTION

The history of plate motions on the earth is relatively well known for the past 200 million years [*Ziegler et al.*, 1983; *Scotese et al.*, 1988], but for earlier times in earth history there is considerable uncertainty because of the absence of ocean floor magnetic anomaly patterns to guide reconstructions. Paleozoic and earlier plate motions can be inferred from paleomagnetic and paleontological data [*Ziegler et al.*, 1979; *Scotese et al.*, 1979], but these approaches leave some uncertainty in the reconstructions.

Isotopic studies can contribute to paleo-plate tectonic interpretations because they can be used to map large scale basement provinces and structure; data that help to constrain past continental reconstructions, to identify suspect cratons [*Bennett and DePaolo*, 1987; *Borg et al.*, 1987*b*], and to assess the uniformity of past tectonism and magmatism on the earth. Although tectonic and magmatic processes in the past may have been similar to those of the present, questions remain regarding the plate tectonic processes that operate during the formation of continents. The relative roles of continental margin versus island arc magmatism and the relationship of granitic batholiths to subduction remain particularly poorly understood.

In this paper we present the results of a regional, reconnaissance isotopic (Nd, Sr, and O), and petrological study of granitic and metamorphic rocks of the Transantarctic Mountains. Our regional geochemical mapping approach can be used to establish large-scale structural, tectonic, and petrogenetic characteristics of subcontinental-scale areas. It is particularly well suited to the study of basement tectonics in Antarctica, where exposure is limited and access is difficult (Figure 1) [*Borg et al.*, 1984, 1987*a*; *Borg and Stump*, 1987; *Borg and DePaolo*, 1987].

CRUSTAL STRUCTURE AND ISOTOPIC PATTERNS IN GRANITIC ROCKS

Studies of late Paleozoic and younger granitic provinces have resulted in the development of schemes relating the chemical compositions of granitic rocks to tectonic setting [e.g., *Pitcher*, 1982, 1987; *Pearce et al.*, 1984; *Harris et al.*, 1986]. The tectonic setting and pre-batholithic crustal structure can also be deduced from Nd, Sr, O, and Pb isotopic studies [*Kistler and Peterman*, 1978; *DePaolo*, 1981; *Farmer and DePaolo*, 1983; *Taylor*, 1986]. The development of the use of Nd and Sr isotopic compositions of granitic rocks to

Paper number 89JB03214.
0148-0227/90/89JB-03214$05.00

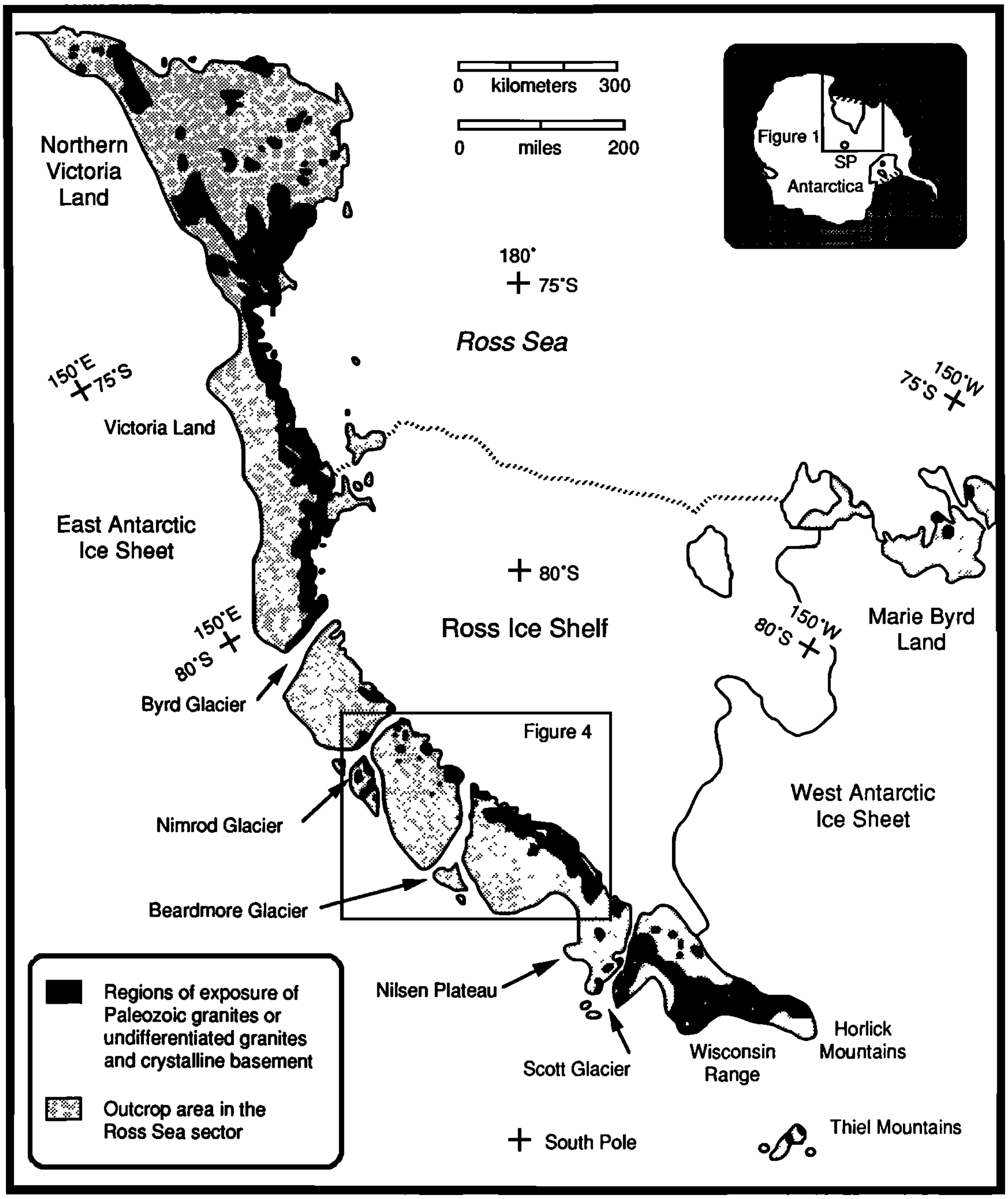

Fig. 1. Location diagram of the Transantarctic Mountains and the study area (Figure 4).

problems of crustal structure is given in papers by *Farmer and DePaolo* [1983, 1984] and *Bennett and DePaolo* [1987]. We give a summary here, which integrates the Nd and Sr isotopic patterns with those for O isotopes [*Taylor*, 1986; *Walawender et al.*, 1989].

Figure 2 shows the isotopic values for Nd, Sr, and O expected for granitoids emplaced into four different types of crust. Type O crust is ocean floor with normal, relatively thin sedimentary cover. Type A crust is composed of thick sedimentary accumulations on oceanic crust, as might occur in continental slope and rise environments. Type B crust is thinned Precambrian basement overlain by a thick section of continental margin sediments, a crustal type also found on rifted continental margins [*Grow et al.*, 1979]. Type C crust is continental craton of normal thickness.

The Nd, Sr, and O isotopic compositions of continental margin granitic rocks are a function of at least three parameters: (1) crustal thickness, (2) crustal age, and (3) magma flux from the mantle [*Farmer and DePaolo*, 1983, 1984]. For Sr, there is a fourth parameter, the Rb/Sr ratio of the crust. The oxygen isotopic composition is additionally dependent on the fraction of sedimentary or metasedimentary rocks in the crust. The isotopic compositions of granitoids derived from each type of crust are distinct. Furthermore, the ε_{Nd} value of granitoids in crust types B and C are dependent on the age of the specific crust involved. For example, Meso-

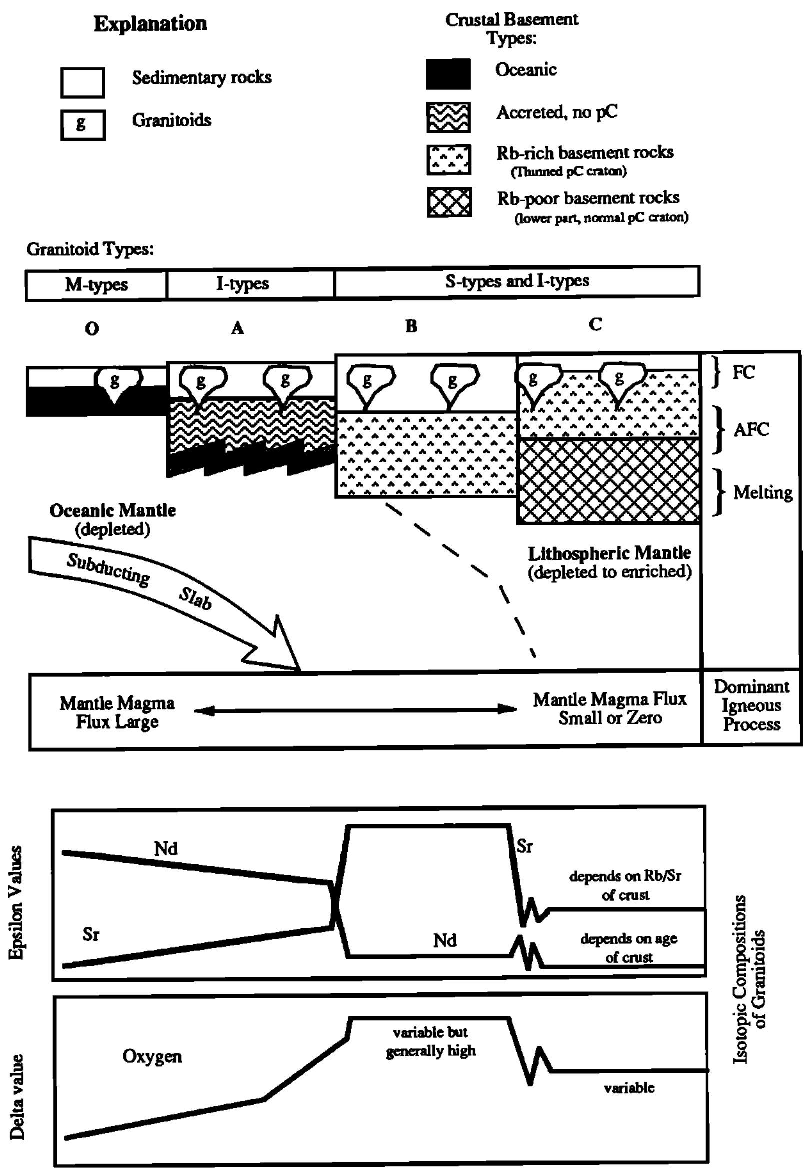

Fig. 2. Model relationship between the isotopic signature of granitoids and crustal structure. FC is the fractional crystallization. AFC is the assimilation-fractional crystallization. Crust types O, A, B, and C are discussed in text.

zoic peraluminous granitoids from the western U.S. that are emplaced into 1.8 Ga crust typically have ε_{Nd} values of -12 ± 2, whereas those emplaced into 2.2 Ga crust typically have values of -18 ± 2. The oxygen isotope patterns inferred for the different types of crust are derived from the work of *Solomon and Taylor* [1981], *Taylor* [1986], and *Bennett et al.* [1987]. In general, continental cratonic rocks have relatively high $\delta^{18}O$ values compared to the mantle, and they are particularly high in regions where the crust has a large metasedimentary component.

In type A crust (and to some degree in type B crust), the ε_{Nd}, ε_{Sr}, and $\delta^{18}O$ values shift regularly with distance inland from the subduction zone [*Farmer and DePaolo*, 1983; *Taylor and Silver*, 1978; *DePaolo*, 1981]. In these terranes, the magmas appear to be mixtures of crustal and mantle materials and the amount of the crustal component increases inland in a regular fashion. This has been interpreted to reflect systematic shifts in crustal thickness, magma flux, and geothermal gradient normal to the trend of the subduction zone.

The schematic crustal section shown in Figure 2 represents the structure expected on an Atlantic-type continental margin which has subsequently become a convergent margin. However, it is not universal and other possible continental margin configurations would affect the petrogenesis of granitic rocks and modify the isotopic patterns to some degree. For example, if the main magmatic arc associated with the subduction zone is in type B or C crust it would be possible to generate mixed (i.e. mantle-crust) magmas rather than purely crustal anatectic melts. While this appears to be the case in the Andes [*Thorpe and Francis*, 1979; *Hawkesworth et al.*, 1982], it has generally been found that in type B or C crust the amount of mantle magma added to the crust is small [*Farmer and DePaolo*, 1984]. In areas where mixed magmas occur in

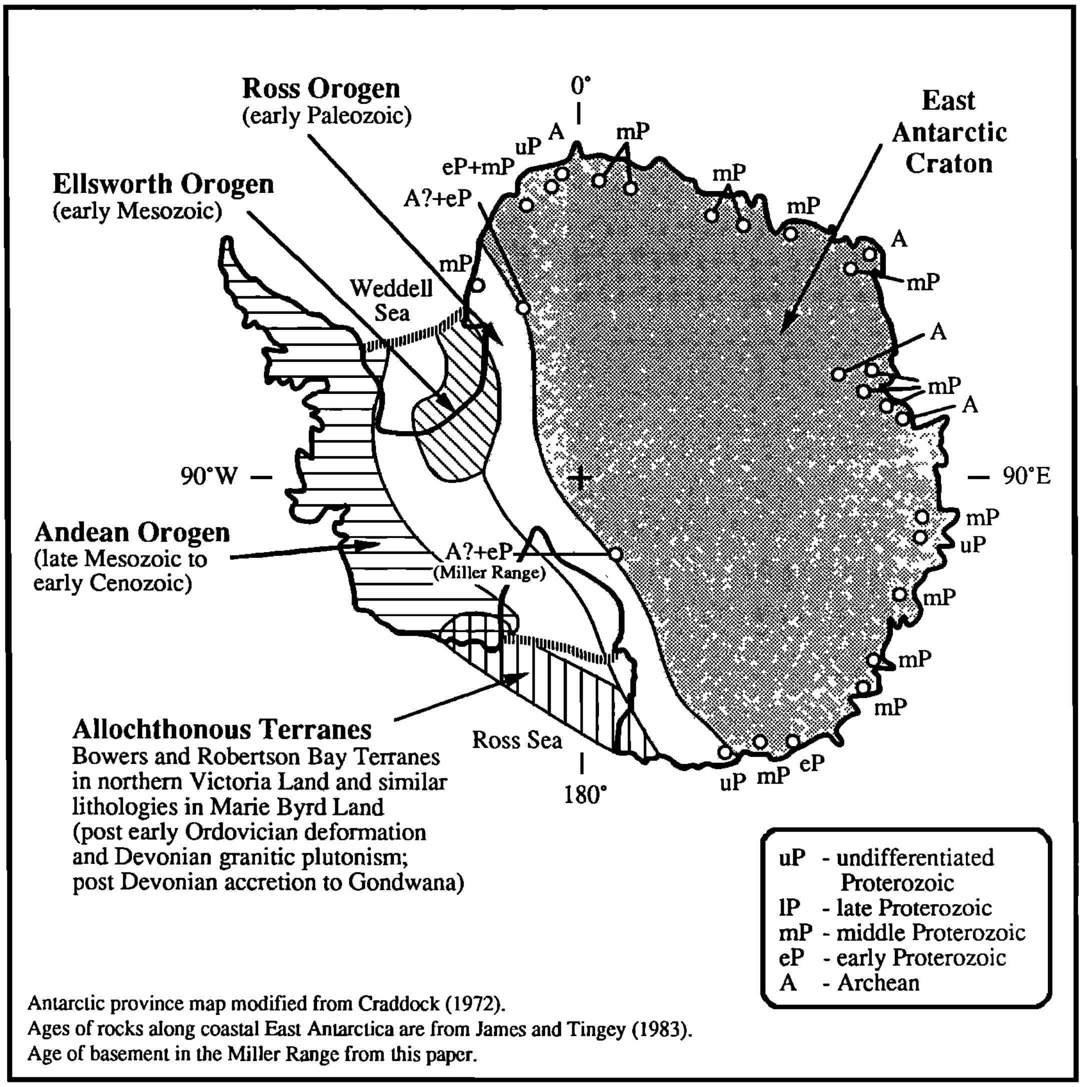

Fig. 3. Tectonic sketch map of Antarctica showing ages around the margin of the East Antarctic Craton. The tectonic evolution of the Ross Orogen is discussed in this paper. Regions to the left of the Ross Orogen on the diagram are composed of younger rocks or were accreted/emplaced after the early Paleozoic.

type B crust, there might be gradients in isotopic composition such as those shown for type A crust (Figure 2).

GEOLOGY OF THE CENTRAL TRANSANTARCTIC MOUNTAINS

The Transantarctic Mountains span the continent from northern Victoria Land to the Weddell Sea (Figure 1). The range consists of Precambrian to early Paleozoic metamorphic and intrusive basement rocks overlain by flat-lying, primarily continentally-derived sediments of Devonian to Triassic age (Beacon Supergroup). Diabase sills and basalt flows were emplaced during the Jurassic (Ferrar Group). The metasedimentary, metavolcanic, and granitic rocks of the pre-Devonian basement define the Ross Orogen (Figure 3), which is thought to have developed along the margin of the East Antarctic Craton during late Precambrian to early Paleozoic time [*Laird et al.*, 1971; *Elliot*, 1975; *Stump*, 1976; *Gunner*, 1976; *Faure et al.*, 1979; *Adams et al.*, 1982; *Borg*, 1983, 1984; *Borg et al.*, 1986*b*, 1987*b*]. Granitic rocks associated with the Ross Orogeny have been called the Granite Harbour Intrusive Complex [*Gunn and Warren*, 1962]. Granitic rocks assigned to this formation from localities throughout the Transantarctic Mountains have yielded K-Ar and Rb-Sr whole rock dates ranging from 460 to 570 Ma. Archean and Proterozoic igneous and metamorphic rocks are known from outcrops around the margin of East Antarctica, but have not previously been unequivocally identified in the Transantarctic Mountains (Figure 3).

The results reported here suggest that use of the term "Granite Harbour Intrusive Complex" for early Paleozoic granitic rocks throughout the Transantarctic Mountains should be reevaluated because (1) there may be more than one tectonic event in the interval between 570 and 460 Ma and (2) the tectonic setting and petrogenesis of the granites at the type locality (Granite Harbour in Victoria Land) may not be directly applicable to the ~500 Ma granites in the Transantarctic Mountains south of the Byrd Glacier. In this paper we avoid use of the formation name.

The central Transantarctic Mountains extend from the Byrd Glacier southward to the Scott Glacier (Figure 1). The geology of this area is described in a series of publications based on reconnaissance mapping [e.g., *Grindley*, 1963, 1981; *Grindley and Warren*, 1964; *Grindley et al.*, 1964; *Faure et al.*, 1968; *Mirsky*, 1969; *Grindley and Laird*, 1969; *McGregor and Wade*, 1969; *Laird et al.*, 1971; *Stump*, 1976; *Wade and Cathey*, 1986; *Stump et al.*, 1988]. The Precambrian and Paleozoic stratigraphy is summarized by *Stump* [1982].

The oldest rocks in the central Transantarctic Mountains comprise the Nimrod Group [*Grindley et al.*, 1964], which is exposed in the Miller Range (Figure 4) and immediately to the north in the Geologists Range. These rocks are thought to be part of the East Antarctic Craton and include marble, micaceous quartzite, garnet-mica schist, quartzo-feldspathic gneiss, and amphibolite which have been complexly deformed [*Grindley*, 1972]. *Gunner and Faure* [1972] report a Rb-Sr whole rock isochron age of 1984 ± 77 Ma for metasediments of the Nimrod Group and interpret this as a minimum age of sedimentation. Recent work suggests that the Miller Formation, previously considered an amphibolite grade portion of the Nimrod Group [cf. *Gunn and Walcott*, 1962; *Grindley et al.*, 1964], is distinct from the lower grade metamorphic rocks of the Nimrod Group (hereinafter referred to as Argosy formation) and that the high grade rocks were thrust eastward over the Argosy formation after emplacement of the Camp Ridge granodiorite (see discussion below) and before emplacement of granites at ~500 Ma [*Borg et al.*, 1986*a*; *Goodge and Borg*, 1987].

The next younger sequence, the Beardmore Group, consists of late Precambrian sedimentary rocks whose relation to the Nimrod Group is equivocal. The Cobham Formation in the upper Nimrod Glacier area consists of shallow water clastic and carbonate sediments [*Laird et al.*, 1971]. The Goldie, Duncan, and La Gorce Formations, exposed from the Byrd Glacier to the Horlick Mountains, are dominated by graywacke-shale sequences that are interpreted to be turbidite deposits [*Gunn and Walcott*, 1962; *McGregor*, 1965; *Wade et al.*, 1965*a*, *b*; *Minshew*, 1967; *Laird et al.*, 1971; *Smit and Stump*, 1986]. In the study area, the Goldie Formation conformably overlies the Cobham Formation [*Laird et al.*, 1971] and *Stump et al.* [1988] correlate the Cobham Formation and the lower Goldie Formation with part of the Nimrod Group (Argosy formation as used here). *Elliot* [1975] suggested that the Beardmore Group was deposited along a passive continental margin. In this scenario, the Cobham Formation and the Argosy formation would represent initial deposits on a subsiding Atlantic-type continental margin immediately preceding deposition of the turbidites. *Faure et al.* [1979] argued that the Beardmore Group was deposited prior to 810 Ma.

The Beardmore Group was folded in latest Precambrian time, an event referred to as the Beardmore Orogeny [*Grindley and McDougall*, 1969; *Stump*, 1981; *Stump et al.*, 1988]. The Beardmore Orogeny must have occurred after deposition of the Goldie Formation and before deposition of the Lower Cambrian Shackleton Limestone of the Byrd Group (~800 to ~550 Ma).

There is conflicting evidence about the existence and extent of magmatism associated with the Beardmore Orogeny. *Faure et al.* [1979] suggest that granitic plutons in the Wisconsin Range and the Nilsen Plateau (southeast of the study area, Figure 1) formed during this event. However, *Pankhurst et al.* [1988] believe that the Beardmore Orogeny did not include magmatic activity in regions southeast of the study area. Northwest of the study area, the Carlyon Granodiorite is a possible Beardmore-related intrusion because its whole rock Rb-Sr isochron age is ~568 Ma [*Felder and Faure*, 1980; *Skinner*, 1983]. Within the study area there is no evidence of granitic magmatism related to the Beardmore deformational event.

Sedimentary rocks of the Byrd Group unconformably overlie the Beardmore Group in the upper Nimrod Glacier area (Figure 4). In the Shackleton Glacier area, the Liv Group is inferred to unconformably overlie the Beardmore Group. The Byrd Group consists of thick limestone and conglomeratic sedimentary rocks and the Liv Group consists of interbedded limestone, silicic volcanic rocks, and volcaniclastic rocks. Fossils in the Shackleton Limestone (Byrd Group) [*Laird and Waterhouse*, 1962] and the Taylor Formation (Liv Group) [*Yochelson and Stump*, 1977] indicate an early Cambrian age (~550 Ma). *Stump* [1976] proposed a model of deposition on a subsiding shelf behind a volcanic arc on the edge of the East Antarctic Craton, although the tectonic environment in which these rocks were deposited remains controversial [see *Laird and Bradshaw*, 1982].

All of the above-mentioned units were folded (or refolded) and metamorphosed during the Ross Orogeny [*Gunn and Warren*, 1962]. Granitic plutons were emplaced at ~500 Ma. We have not found any regional deformational fabrics in granitic plutons intruding rocks initially deformed in the Ross event.

AGE AND LITHOLOGY OF GRANITIC ROCKS

Because of the paucity of detailed geochronological work on plutonic rocks in the study area (Figure 4), we have evaluated the age of each of the sampled plutons based on field relations and available radiometric data. U-Pb analyses of some samples (J. Mattinson, unpublished data, 1988) indicate crystallization ages ranging from 520 to 500 Ma. With the exception of the Camp Ridge granodiorite in the Miller Range, intrusion of all granites in the region closely followed the Ross Orogeny.

The available U-Pb age data are for quartz-bearing plutons and might not apply to mafic plutons. Mafic plutons could be unrecognized parts of the pre-batholithic basement or coarse-grained equivalents of the Jurassic Ferrar Group. To evaluate these possibilities, the following criteria have been used: (1) Where plutons intrude the Byrd or Liv Groups, association with an event earlier than the Cambro-Ordovician Ross Orogeny can be eliminated and the plutons cannot be part of the pre-batholithic basement. The additional observation that the plutonic rocks under consideration are undeformed is taken as further evidence that they postdate the Ross Orogeny, the most recent major deformation documented. (2) Where intruded by quartz-bearing plutons of the 500 Ma suite we can eliminate the possibility that the mafic plutons are part of the Ferrar Group.

The ages of two undeformed, relatively mafic plutons were also determined by the Rb-Sr method (Table 3). The diorite of the Campbell Hills (85 DCT 6) yielded a whole-rock-plagioclase date of 470 Ma. This rock is intruded by pegmatites related to a granite dated at ~500 Ma (J. Mattinson, unpublished data, 1988). A diorite from the north side of Ramsey Glacier (85 BCT 167), which intruded sediments believed to be part of the Beardmore Group, yielded a Rb-Sr whole-rock-plagioclase date of 460 Ma. These ages are

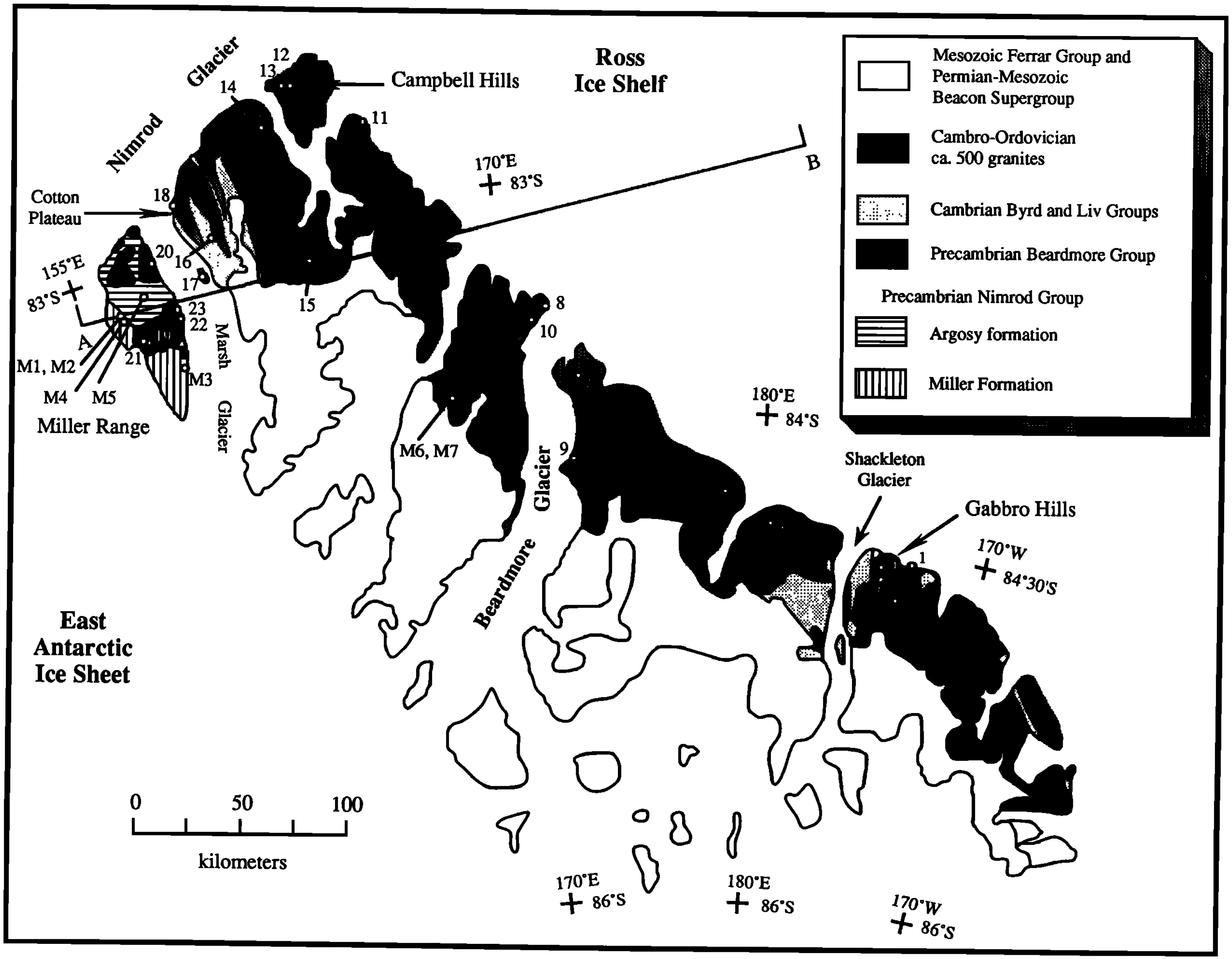

Fig. 4. Geologic sketch map of the central Transantarctic Mountains from the Nimrod Glacier to the Shackleton Glacier. Geology compiled from *Grindley and Laird* [1969], *Gunner* [1976], and our mapping.

probably slightly young, as is a common feature of Rb-Sr mineral ages in deep-seated plutonic rocks. A small amount of reequilibration may also have occurred in response to intrusion of the Jurassic Ferrar Group.

Within the study area, the 500 Ma intrusive rocks are strongly discordant epizonal plutons ranging in composition from gabbro to granite. The rocks are calc-alkaline according to the classification scheme of *Irvine and Baragar* [1971]. Lithology and chemical type are related to geographic position. Whereas granites in the Miller Range, which range from 63 to 72 wt % SiO_2, are uniformly peraluminous, east of the Marsh Glacier (Figure 4) the intrusive rocks are metaluminous to weakly peraluminous granodiorite and monzogranite. Gabbro, diorite, tonalite, and granodiorite are most common in the southeastern part of the region, southeast of the Shackleton Glacier. Exceptions are quartz diorite (85 BCT 3) immediately east of the Marsh Glacier and diorite (85 DCT 6) in the Campbell Hills. Mafic and felsic plutons were observed in close proximity in only two areas and in both cases the mafic plutons are older. In the Campbell Hills, diorite is intruded by monzogranite and in the Gabbro Hills, gabbro and diorite are intruded by tonalite and granodiorite.

ANALYTICAL PROCEDURES

Sm-Nd and Rb-Sr isotopic and concentration measurements were made on samples of granitic rocks and on samples of pre-batholithic metamorphic rocks. Table 1 contains isotopic data for whole-rock samples of metamorphic basement. Data for whole-rock powders of basalt and associated gabbro from the Cotton Plateau and for plagioclase and amphibole+clinopyroxene separates from the gabbro are shown in Table 2. Table 3 contains data for 23 whole-rock granitic samples and 2 plagioclase separates. Four of the metamorphic samples and six of the granites are powders of rocks analyzed by *Gunner* [1976].

Nd and Sr isotopic compositions are expressed as normalized ratios in epsilon notation following *DePaolo and Wasserburg* [1976, 1977]. Model reservoir compositions and decay constants are given in table footnotes and are identical to those used by *Farmer and DePaolo* [1983]. The isotopic composition of O is also expressed as a normalized ratio, in delta notation, and is calculated relative to V-SMOW [*O'Neil*, 1986].

Most of the granite samples show only minor sericitization of plagioclase and chloritization of biotite, but two samples exhibit evidence of more extensive hydrothermal alteration. Sample 85 BCT 21 shows extensive sericitization of plagioclase and chloritization of biotite. Sample 85 DCT 6 has moderately to extremely sericitized plagioclase and severe alteration of clinopyroxene to amphibole. Despite the alteration, the igneous textures of both samples are well preserved.

Samples prepared for analysis averaged 30 kg in mass and were broken into fist size pieces either at the sample locality or under clean conditions in the laboratory. Weathered portions were discarded and hammer impact sites were brushed off. About 15–20 kg of rock chunks were crushed to gravel (maximum dimension ~5 mm) using a jaw crusher with steel plates. Contamination was minimized by disassembly and thorough cleaning of the jaw crusher between samples and by discarding the first 1/2 kg of sample crushed. The gravels were then split using a Jones Splitter and ~0.5 kg splits were pulverized in a tungsten-carbide ring and puck mill to produce whole-rock powders. Preparation of powders from *Gunner* [1976] are described by *Gunner* [1971]. Powder of the basalt from the Cotton Plateau was provided by E. Stump.

Sample dissolution and chemical separation procedures for Rb, Sr, Sm, and Nd are given by *Papanastassiou et al.* [1977] and *DePaolo* [1978]. Procedures for Sr and Nd isotopic analyses and concentration measurements by isotope dilution techniques are given by *DePaolo* [1981]. Average total chemistry blank levels were negligible for Nd (125 pg) and Sm (18 pg). Rb and Sr total chemistry blanks were significantly different for the two HCl columns used for the chemical separations. For one column, blanks were 240 pg for Rb and 190 pg for Sr and are negligible. For the other column, blanks were 6.9 ng for Rb and 0.75 ng for Sr and are negligible only for samples with concentrations greater than ~50 ppm. Because of this situation, the former column was used for all samples with low Rb or Sr concentrations.

Whole-rock and mineral O-isotopic data for the granitic rocks are listed in Table 4. Quartz was selected for O-isotopic analysis because it is particularly resistant to diffusional oxygen exchange during prolonged cooling (B. M. Smith and B. J. Giletti, Oxygen isotope evidence for high-temperature cooling of porphyritic felsites from the Isle of Skye, N. W. Scotland, submitted to *Earth and Planetary Science Letters*, 1989.) (hereinafter referred to as B. M. Smith and B. J. Giletti, submitted manuscript, 1989). Three samples did not contain sufficient quartz to separate easily, so plagioclase separates or splits of whole-rock powder were analyzed. Mineral separates were hand-picked under a binocular microscope from 20–60 mesh fractions of the gravel splits. To facilitate comparisons between different analyzed materials, the data were recalculated to whole-rock values using the diffusional closure temperatures from B. M. Smith and B. J. Giletti (submitted manuscript, 1989), fractionation factors from *Bottinga and Javoy* [1975], and petrographic estimates of modal mineralogy. Mineral separates were not available for the six samples from *Gunner* [1976].

Oxygen was liberated by reaction with BrF_5 at 600°C [*Clayton and Mayeda*, 1963] and was converted to CO_2 over a resistance heated carbon rod. O-isotope ratios were determined on a Finnigan-MAT 251 mass spectrometer and are considered accurate and precise to ±0.2‰. The average deviation from the mean for 3 duplicate analyses is 0.07‰. Two analyses of NBS-28 quartz standard during the course of this study yielded a value of +9.62 ± 0.07‰, similar to values reported by most other stable isotope laboratories.

RESULTS

Age and Isotopic Character of the Metamorphic Basement

From data on metamorphic rocks (Tables 1 and 2), we have identified three distinct components of the pre-batholithic crystalline basement: (1) the Miller Range metamorphic rocks, (2) the Goldie Formation of the Beardmore Group, and (3) basalt and gabbro associated with the Goldie Formation at the Cotton Plateau. In the Miller Range, which contains rocks of the East Antarctic Craton, Sm-Nd model ages of 2.7 to 2.9 Ga were determined for both the Miller Formation and the Argosy formation (Table 1). The Camp

TABLE 1. Rb-Sr and Sm-Nd Concentrations and Isotopic Ratios for Prebatholithic Metamorphic Rocks

Map No. Locality	Sample	[Rb]	[Sr]	[Sm]	[Nd]	$^{87}Rb/^{86}Sr$	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd(0)$	$\varepsilon_{Nd}(0)$	$\varepsilon_{Nd}(500)$	Nd Model Age T_{DM}	$^{87}Sr/^{86}Sr(0)$	$^{87}Sr/^{86}Sr(500)$	$\varepsilon_{Sr}(500)$
							Nimrod Group							
Miller Fm														
M1 Camp Ridge	519 MR	59.62	71.63	1.484	8.818	2.4259(±11)	0.10183(±4)	0.510294(±22)	−30.13	−24.09	2.76	0.77806(±3)	0.76078	807.8
M2 Camp Ridge	517 MR	134.4	476.0	6.917	38.02	0.8187(±3)	0.11006(±8)	0.510363(±12)	−28.78	−23.26	2.88	0.72712(±3)	0.72129	246.9
M3 S of Orr Peak	86 BMR M37			11.95	70.24		0.10289(±50)	0.510318(±15)	−29.66	−23.68	2.75			
Argosy Fm														
M4 Aurora Heights	86 BMR M24			5.549	31.74		0.10575(±7)	0.510395(±19)	−28.15	−22.36	2.72			
Camp Ridge GD ($T_c \approx 1.7$ Ga)														
M5 N Camp Ridge	86 BMR M28			7.53	38.72		0.11760(±50)	0.510504(±24)	−26.03	−21.00	2.73			
							Beardmore Group							
Goldie Formation (eastern-type)														
M6 Hampton Ridge	A-2	235.7	89.26	7.086	36.32	7.6906(±40)	0.11800(±9)	0.511250(±24)	−11.45	−6.42	1.68	0.77249(±4)	0.71769	195.8
M7 Hampton Ridge	A-3	124.8	157.6	5.970	29.88	2.2998(±10)	0.12085(±8)	0.511323(±25)	−10.02	−5.18	1.61	0.73626(±3)	0.71987	226.8

See text for analytical procedures. Concentrations are in ppm. Mass spectrometric procedures are given by *DePaolo* [1981]. $^{143}Nd/^{144}Nd$ normalized to $^{146}Nd/^{142}Nd = 0.63613$. $^{87}Sr/^{86}Sr$ normalized to $^{86}Sr/^{88}Sr = 0.1194$. Epsilon notation follows *DePaolo and Wasserburg* [1979]. ε_{Nd} is calculated with respect to a chondritic reservoir with present $^{143}Nd/^{144}Nd = 0.511836$ and $^{147}Sm/^{144}Nd = 0.1967$. $\lambda_{Sm} = 6.54 \times 10^{-12}$ yr^{-1}. ε_{Sr} is calculated with respect to a uniform reservoir with present $^{87}Sr/^{86}Sr = 0.7045$ and $^{87}Rb/^{86}Sr = 0.0827$. $\lambda_{Rb} = 1.42 \times 10^{-11}$ yr^{-1}. $\varepsilon_{Nd}(500)$, $\varepsilon_{Sr}(500)$, and $^{87}Sr/^{86}Sr(500)$ are the compositions calculated at 500 Ma, whereas $^{143}Nd/^{144}Nd(0)$, $\varepsilon_{Nd}(0)$, and $^{87}Sr/^{86}Sr(0)$ are the measured compositions. Model ages are calculated relative to a depleted mantle model (ε_{Nd} $[DM] = 8.6 - 1.91T$; T = age in Ga). Thus, the model age is found by solving for T_{DM} in the following equation (modified from *Farmer and DePaolo* [1983]): $8.6 - 1.91T_{DM} = \varepsilon_{Nd}(0) - f_{Sm/Nd}Q_{Nd}T_{DM}$ where $f_{Sm/Nd} = (^{147}Sm/^{144}Nd_{sample})/(^{147}Sm/^{144}Nd_{CHUR}) - 1$, and $Q_{Nd} = 25.13$. Because sample 86 BMR M28 is a granite, its model age was calculated from the initial composition ($\varepsilon_{Nd} = 8.79$) and crystallization age ($T_c = 1.7$ Ga) using an Sm/Nd ratio which is dependent on crustal age. The relation used is: $f_{Sm/Nd} = -0.25 - 0.08T_{DM}$, and the model age equation becomes: $8.6 - 1.91T_{DM} = \varepsilon_{Nd} - (-0.25 - 0.08T_{DM})Q_{Nd}(T_{DM} - T_c)$. This is a quadratic with one reasonable root. See *DePaolo* [1988] and D. J. DePaolo et al. (The continental crust age distribution: Methods of determining mantle separation ages from Sm-Nd isotopic data and application to the southwestern United States, submitted to *Journal of Geophysical Research*, 1990) (hereinafter referred to as D. J. DePaolo et al., submitted manuscript, 1990). The tabulated uncertainty in the isotopic ratios is the 2-sigma error from the ratio measurement and pertains to the last digit or digits of the respective ratio. Uncertainty in the parent-daughter ratios was estimated from measured concentrations and respective 2-sigma uncertainties in the following equation: ([±2-sigma parent] / [parent] + [±2-sigma daughter] / [daughter]) / (1-[±2-sigma daughter] / [daughter]) * [parent isotope] / [daughter isotope]. Samples 517 MR and 519 MR [from *Gunner*, 1976] were probably collected below the Endurance Thrust and thus they would be part of the Argosy Formation as defined by *Grindley* [1972]. However, on the basis of metamorphic grade and arguements in the text, they are here assigned to the Miller Formation. Samples 519MR, 517MR, A-2, and A-3 are described by *Gunner* [1976]; powders were provided by Gunter Faure.

TABLE 2. Sm-Nd Isotopic Data for the Cotton Plateau Gabbro and Basalt

Rock	Sample	[Sm]	[Nd]	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd(0)$	$\varepsilon_{Nd}(0)$	$\varepsilon_{Nd}(760)$
Gabbro	85 BCT 114 wr	1.4075	4.4419	0.19168(±12)	0.512143(±24)	6.00	6.50
	85 BCT 114 cpx+	2.0888	6.0270	0.20964(±16)	0.512251(±21)	8.11	6.86
	85 BCT 114 plagioclase	0.8141	2.9859	0.16494(±10)	0.512029(±18)	3.77	6.88
Basalt	GAU wr	3.9148	13.7839	0.17180(±11)	0.512060(±16)	4.38	6.81

See text for analytical procedures. Concentrations are in ppm. Abbreviations: wr = whole rock, cpx+ = clinopyroxene + amphibole, plag = plagioclase. Mass spectrometric procedures are given by *DePaolo* [1981]. $^{143}Nd/^{144}Nd$ normalized to $^{146}Nd/^{142}Nd = 0.63613$. $^{87}Sr/^{86}Sr$ normalized to $^{86}Sr/^{88}Sr = 0.1194$. Mineral separations by heavy liquid and magnetic methods. Sample 85 BCT 114 was collected by E. Stump for S. Borg and D. DePaolo. A powder split of sample GAU was provided by E. Stump. Epsilon notation follows *DePaolo and Wasserburg* [1979]. ε_{Nd} is calculated with respect to a chondritic reservoir with present $^{143}Nd/^{144}Nd = 0.511836$ and $^{147}Sm/^{144}Nd = 0.1967$. $\lambda_{Sm} = 6.54 \times 10^{-12}$ yr^{-1}. $\varepsilon_{Nd}(760)$ is the calculated composition at 760 Ma whereas $\varepsilon_{Nd}(0)$ is the measured composition. Numbers in parentheses are $\pm 2\sigma$ analytical uncertainty and correspond to the last two digits of the preceeding ratio. Uncertainty in the parent-daughter ratios was estimated from measured concentrations and their respective uncertainties (2σ). See Table 1 equation in footnote.

Ridge granodiorite, which has a crystallization age of ~1.7 Ga based on Pb-Pb ages (V. Bennett, unpublished data, 1988), also has a Sm-Nd model age of 2.7 Ga (Table 1). This granodiorite intrudes folded, amphibolite-grade metamorphic rocks which we assign to the Miller Formation and therefore provides a minimum age for the folding. The Camp Ridge granodiorite contains penetrative shear fabrics that indicate east-directed overthrusting and, thus, provides a maximum age of the eastward thrusting identified in the Miller Range by *Grindley* [1972] and reinterpreted by *Goodge and Borg* [1987]. The Goldie Formation at Hampton Ridge yields Sm-Nd model ages of ~1.6–1.7 Ga (Table 1), indicating that it could not have been derived by erosion of rocks such as those exposed in the Miller Range. This is one of the observations that suggest that the Beardmore Group was deposited far from the East Antarctic Craton, and that both it and the associated basement rocks are allochthonous.

At the Cotton Plateau, an aphanitic pillow basalt was found interbedded with the Goldie Formation; gabbro associated with the basalt was also sampled (*Stump et al.* [1988]; E. Stump, personal communication, 1985). In thin section, the gabbro is a cumulate rock containing plagioclase and clinopyroxene with hypidiomorphic granular texture and a wide range of grain size. Due to alteration, the plagioclase contains some epidote and the clinopyroxene has alteration rims of amphibole. The sample is moderately friable and some calcite is present on fractures, grain boundaries, and in interstices. Based on normative mineralogy, these rocks are olivine tholeiites. They have about 50 wt % SiO_2 with K_2O contents of 0.15 and 0.11 wt %, and are thus low-K tholeiites (E. Stump, unpublished data, 1987; Borg, unpublished data, 1987). These rocks have the petrological characteristics of ocean floor basalts and suggest that the Goldie Formation in the Cotton Plateau area was deposited on oceanic crust.

Because of the potential for determining the depositional age of the Goldie Formation at the Cotton Plateau, we dated the gabbro and basalt by the Sm-Nd method. Low loss on ignition (0.58 wt %, E. Stump, unpublished data, 1987) indicates that the basalt sample is relatively fresh. Whole-rock splits of each sample were used along with plagioclase and clinopyroxene from the gabbro. It was not possible to eliminate all alteration products from the mineral separates. The data (Table 2) have been plotted on an isochron diagram in Figure 5 and isochrons have been calculated using the method of *York* [1969]. Using all four points, the calculated age is 746 ± 82 Ma (2σ). Excluding the whole-rock analysis of the gabbro, the three remaining points yield an age of 762 ± 24 Ma (2σ). The calculated error is low because of the co-linearity of the points. A more conservative error estimate can be made by using the 1σ analytical uncertainty of the end points. This calculation yields an age uncertainty of +60 and −71 Ma. Omitting the whole-rock gabbro from the isochron may be reasonable because of the secondary calcite present and we regard the latter date as a better estimate of the age.

Regardless of which isochron is used, the initial ε_{Nd} value for the gabbro and basalt is +6.8. Mid-ocean ridge basalt is estimated to have had an ε_{Nd} value of +8 ± 2 at 760 Ma, so the initial value of the gabbro is within the range expected for late Proterozoic MORB. The high ε_{Nd} value does not require the basalt to have been formed in normal oceanic crust. In fact, the Sm/Nd ratio of the basalt is lower than would be expected for "normal" MORB, but is in the range observed for so-called E-type MORB. The isotopic composition and Sm/Nd ratio would also allow the basalt to have formed in a back-arc or island-arc setting. However, the high ε_{Nd} value indicates that it is unlikely that the basalt formed on a continent; an oceanic setting is probable.

Nd, Sr, and O Isotopic Compositions of the Granitoids

The data for the granitic rocks are presented in Tables 3 and 4 and sample locations are shown on Figure 4. Initial Nd and Sr isotopic compositions (denoted ε_{Nd} and ε_{Sr}) were calculated from measured values using an age of 500 Ma. The uncertainty in ε_{Nd} for a ±20 m.y. age uncertainty is negligible. The uncertainty in ε_{Sr} for a ±20 m.y. age uncertainty is dependent on the measured Rb/Sr ratio. For most samples this age uncertainty leads to an uncertainty in initial ε_{Sr} of <15 ε-units. For four of the samples the uncertainty in initial ε_{Sr} is >15 but <25 ε-units. For three samples, all from the Campbell Hills area, Rb/Sr is high enough to generate an uncertainty of >50 ε-units for ±20 m.y. For these three samples, an age of 500 Ma has been used but error bars for ε_{Sr} calculated for ±10 m.y. are shown on the diagrams. This age uncertainty was used because one of the samples in question has been dated at 505 ± 10 Ma by U-Pb isotopes in zircon (J. Mattinson, unpublished data, 1988).

The granite rocks have a wide range of isotopic compositions; ε_{Nd} varies from +2 to −12, ε_{Sr} varies from +8 to +537, and $\delta^{18}O_{wr}$ varies from +6.0 to +12.7‰. In ε_{Nd}-ε_{Sr} space (Figure 6), the data form a curved, concave-upward array typical of calc-alkaline granitic provinces (e.g., Figure

TABLE 3. Rb-Sr and Sm-Nd Concentrations

Map No.	Locality	Sample	[Rb]	[Sr]	[Sm]	[Nd]	$^{87}Rb/^{86}Sr$
							Gabbro
1	Sage Nunataks	86 BGH 9	72.65	224.06	6.316	25.348	0.9384(±4)
2	Amphibole Peak	86 BGH 20	5.810	329.23	3.528	14.020	0.05108(±2)
3	Olds Peak	86 BGH 25	83.91	181.18	5.899	25.657	1.3404(±7)
4	Longhorn Spurs	86 BGH 31	90.30	148.57	8.078	36.841	1.7610(±9)
							Beardmore Glacier Area
5	Woodall Peak	85 BCT 167	28.56	263.25	5.797	27.887	0.3140(±2)
		plagioclase	19.13	446.17			0.1241(±1)
6	O'Leary Peak	85 BCT 171	139.54	119.49	7.790	37.699	3.3888(±11)
7	Mt. Harcourt	647 MH	249.12	162.97	10.661	50.435	4.4239(±16)
8	Mt. Hope	593 MH	196.45	269.79	13.424	74.762	2.1073(±5)
9	Beetle Spur	672 MH	233.40	399.63	8.284	46.267	1.6903(±10)
10	Granite Pillars	85 BCT 44	206.06	214.37	11.553	64.987	2.7879(±17)
11	Cape Goldie	85 BCT 39	312.03	70.71	2.177	8.814	12.8860(±34)
12	Campbell Hills	85 DCT 6	93.91	437.89	7.333	38.156	0.6207(±3)
		plagioclase	45.25	965.16			0.1357(±1)
13	Campbell Hills	85BCT 31	431.68	79.49	11.164	62.298	15.8947(±79)
14	N of Mt. Heiser	85 BCT 21	398.66	85.93	12.774	75.033	13.5656(±77)
15	Mt. Macbain	85 BCT 116	150.29	391.79	3.388	18.888	1.1102(±5)
16	Bartrum Plateau	85 BCT 3	101.21	261.67	4.478	21.758	1.1206(±3)
17	Moody Nunatak	85 BCT 8	194.82	208.00	9.178	48.375	2.7183(±7)
18	Panorama Point	85 BCT 1	147.35	188.11	8.500	42.606	2.2670(±9)
							Miller
19	SE Martin Dome	85 BCT 80	308.75	216.38	5.108	27.745	4.1522(±19)
20	E Snowshoe Pass	85 BCT 93	233.25	147.72	6.466	32.960	4.6002(±15)
21	Hockey Cirque	72 MR	357.06	164.26	4.556	23.284	6.2909(±34)
22	Macdonald Bluffs	297 MR	158.67	392.40	9.280	57.211	1.1703(±4)
23	Dike Cirque	315 MR	184.49	203.74	6.201	33.500	2.6208(±9)

See text for analytical procedures. Concentrations are in ppm. Mass spectrometric procedures are given by *DePaolo* [1981]. $^{143}Nd/^{144}Nd$ normalized to $^{146}Nd/^{142}Nd = 0.63613$. $^{87}Sr/^{86}Sr$ normalized to $^{86}Sr/^{88}Sr = 0.1194$. Epsilon notation follows *DePaolo and Wasserburg* [1979]. ε_{Nd} is calculated with respect to a chondritic reservoir with present $^{143}Nd/^{144}Nd = 0.511836$ and $^{147}Sm/^{144}Nd = 0.1967$. $\lambda_{Sm} = 6.54 \times 10^{-12}$ yr^{-1}. ε_{Sr} is calculated with respect to a uniform reservoir with present $^{87}Sr/^{86}Sr = 0.7045$ and $^{87}Rb/^{86}Sr = 0.0827$. $\lambda_{Rb} = 1.42 \times 10^{-11}$ yr^{-1}. ε_{Nd}, ε_{Sr}, and $^{87}Sr/^{86}Sr(i)$ are the calculated initial compositions, whereas $^{143}Nd/^{144}Nd(0)$, $\varepsilon_{Nd}(0)$, and $^{87}Sr/^{86}Sr(0)$ are the measured compositions. Initial compositions are calculated at 500 Ma (see text). Model ages are calculated relative to a depleted mantle model (ε_{Nd} $[DM] = 8.6 - 1.91T$; T = age in Ga). Thus, in general the model age is found by solving for T_{DM} in the following equation (modified from *Farmer and DePaolo*, [1983]): $8.6 - 1.91T_{DM} = \varepsilon_{Nd}(0) - f_{Sm/Nd}Q_{Nd}T_{DM}$ where $f_{Sm/Nd} = (^{147}Sm/^{144}Nd_{sample})/(^{147}Sm/^{144}Nd_{CHUR}) - 1$, and $Q_{Nd} = 25.13$. However, because these are granitic rocks, the model ages were calculated from the initial composition (ε_{Nd}) and crystallization age (T_c) using an Sm/Nd ratio which is dependent on crustal age. The relationship used is: $f_{Sm/Nd} = -0.25 - 0.08T_{DM}$, and the model age equation becomes: $8.6 - 1.91T_{DM} = \varepsilon_{Nd} - (-0.25 - 0.08T_{DM})Q_{Nd}(T_{DM} - T_c)$. This is a quadratic with one reasonable root. See *DePaolo* [1988] and D. J. DePaolo et al. (submitted manuscript, 1990). The tabulated uncertainty in the isotopic ratios is the 2-sigma error from the ratio measurement and pertains to the last digit or digits of the respective ratio. Uncertainty in the parent-daughter ratios was estimated from measured concentrations and respective 2-sigma uncertainties. See equation in Table 1 footnote. The initial Sr-isotopic compositions of samples 85 BCT 21, 85 BCT 31, and 85 BCT 39 should be used with caution because of high Rb/Sr and the age uncertainty. Samples with suffix MH or MR are described by *Gunner* [1976]; powders were provided by the Gunter Faure.

10 of *Liew and McCulloch* [1985]). The data can be divided into three groups: (1) Samples with $\varepsilon_{Nd} > 0$ and $\varepsilon_{Sr} < 30$ form the first group and all are from the Gabbro Hills area, on the eastern side of the range. (2) Samples with ε_{Nd} between -1 and -8, and ε_{Sr} between $+40$ and $+220$ form the second group. These samples are from the region west of the Gabbro Hills and east of the Marsh Glacier. (3) The third group is composed of samples with ε_{Nd} between -10 and -12, and ε_{Sr} between $+400$ and $+540$. These samples are all from the Miller Range, in the western part of the study area.

In Figure 7, isotopic compositions have been projected onto section line A-B of Figure 4, which is perpendicular to the structural trends associated with the Ross Orogeny. This line represents a structural cross-section from the cratonic metamorphic rocks (Miller Formation and Argosy formation) of the Miller Range through the metamorphic rocks (Beardmore, Byrd, and Liv Groups) of the Ross Orogen. The ε_{Nd} values decrease from east to west with a break at the Marsh Glacier, the 50 km point along section A-B (Figure 7*a*). In the Miller Range (west side), ε_{Nd} clusters tightly between -10 and -12 (see also Figure 6), whereas immediately east, ε_{Nd} is -5 and -8. From the west side of the Marsh Glacier to the 300 km mark, ε_{Nd} is scattered but generally increases from -6 ± 1 to -2 ± 1 (Beardmore Glacier area from Figure 6). Sample 85 BCT 1 ($\varepsilon_{Nd} = -8.2$) is from the boundary between the Miller Range and Beardmore Glacier areas. It is listed last in the Beardmore Glacier area granites (Tables 3 and 4) and is noted with an arrow on the diagrams. East of 300 km, ε_{Nd} is between $+0.4$ and $+1.7$ (Gabbro Hills area from Figure 6). The ε_{Sr} data define a westward increasing pattern (Figure 7*b*) with breaks similar to those shown by the Nd data. Figure 7*c*, a plot of $\delta^{18}O$ along section A-B, shows a pattern of westward increasing $\delta^{18}O_{wr}$. The shifts in the Nd and Sr isotopic compositions are not entirely correlated with the $\delta^{18}O$ trend, but all samples collected east of the 300 km mark have $\delta^{18}O_{wr}$ <7.3‰, whereas west of the 300 km mark all samples have $\delta^{18}O_{wr}$ >8.8‰. In the central part of the section, between and Isotopic Ratios for the ~500 Ma Granites

$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd(0)$	$\varepsilon_{Nd}(0)$	ε_{Nd}	Nd Model Age T_{DM}	$^{87}Sr/^{86}Sr(0)$	$^{87}Sr/^{86}Sr_{(i)}$	ε_{Sr}
Hills Area							
0.15072(±11)	0.511708(±24)	−2.50	0.44	1.18	0.712577(±29)	0.70589	28.1
0.15222(±6)	0.511778(±25)	−1.13	1.72	1.07	0.704823(±24)	0.70446	7.8
0.13907(±7)	0.511678(±20)	−3.08	0.61	1.17	0.714931(±31)	0.70538	20.9
0.13264(±10)	0.511711(±22)	−2.44	1.66	1.08	0.717437(±21)	0.70489	13.9
(*Marsh to Shackleton Glaciers Area*)							
0.12574(±8)	0.511432(±22)	−7.89	−3.36	1.48	0.714231(±30)	0.71199	114.8
					0.712999(±25)		
0.12500(±5)	0.511522(±10)	−6.13	−1.55	1.34	0.733162(±27)	0.70902	72.5
0.12787(±9)	0.511464(±24)	−7.27	−2.87	1.45	0.742207(±33)	0.71068	96.2
0.10861(±7)	0.511407(±23)	−8.38	−2.75	1.44	0.723389(±36)	0.70837	63.4
0.10831(±6)	0.511302(±21)	−10.43	−4.78	1.59	0.722735(±32)	0.71069	96.3
0.10753(±5)	0.511388(±17)	−8.75	−3.05	1.46	0.727868(±31)	0.70800	58.1
0.14941(±10)	0.511372(±20)	−9.07	−6.05	1.68	0.798658(±26)	0.70684	41.6
0.11625(±8)	0.511278(±22)	−10.90	−5.76	1.66	0.714765(±27)	0.71034	91.4
					0.711536(±30)		
0.10840(±7)	0.511201(±25)	−12.41	−6.76	1.73	0.822098(±28)	0.70884	70.1
0.10298(±9)	0.511184(±23)	−12.74	−6.75	1.73	0.811917(±30)	0.71526	161.2
0.10851(±6)	0.511415(±17)	−8.23	−2.59	1.42	0.715664(±27)	0.70775	54.6
0.12450(±7)	0.511352(±20)	−9.46	−4.84	1.59	0.717344(±32)	0.70936	77.4
0.11476(±6)	0.511304(±17)	−10.39	−5.16	1.62	0.735055(±31)	0.71569	167.3
0.12068(±9)	0.511169(±24)	−13.03	−8.18	1.83	0.735210(±29)	0.71906	215.2
Range Area							
0.11137(±6)	0.511022(±18)	−15.90	−10.46	1.99	0.762329(±28)	0.73274	409.6
0.11868(±9)	0.511045(±18)	−15.45	−10.47	1.99	0.774517(±25)	0.74174	537.4
0.11836(±6)	0.510996(±25)	−16.41	−11.41	2.05	0.778957(±34)	0.73413	429.3
0.09812(±7)	0.510975(±21)	−16.82	−10.53	1.99	0.741393(±34)	0.73305	414.0
0.11197(±5)	0.510955(±18)	−17.21	−11.81	2.07	0.751101(±33)	0.73243	405.1

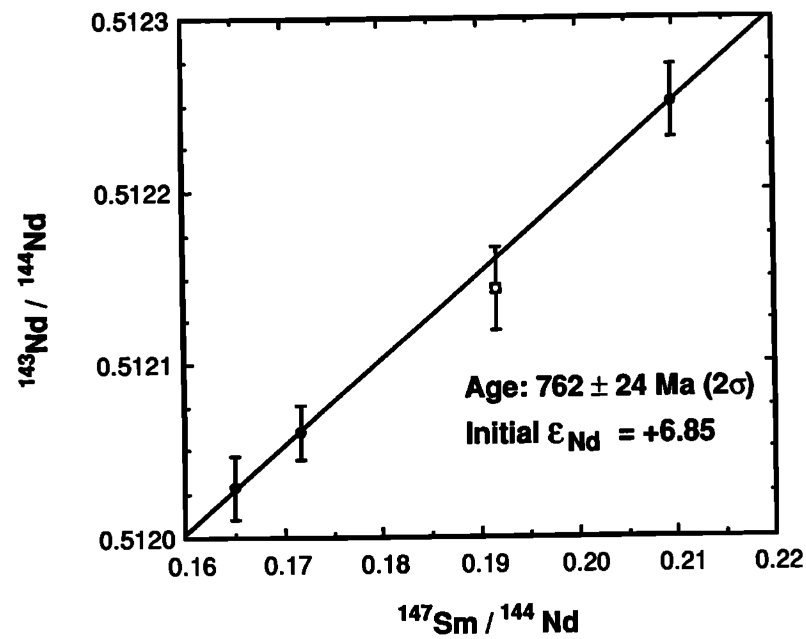

Fig. 5. Sm-Nd isochron diagram for the basalt and gabbro at Cotton Plateau. The age of 762 ± 24 Ma (2σ) with initial $\varepsilon_{Nd} = +6.85$ is the best estimate of the crystallization age and initial ε_{Nd} of these rocks. It is based on minerals separated from the gabbro and the whole-rock basalt and is calculated by the method of *York* [1969]. Exclusion of the whole-rock gabbro (open square) is warranted because of secondary calcite and related weathering products. However, using all four points only the age uncertainty is appreciably changed: an age of 746 ± 82 Ma (2σ) with initial $\varepsilon_{Nd} = +6.76$ is calculated. Error bars on the diagram are $\pm 2\sigma$ analytical uncertainty. Because the analytical uncertainty of each data point is large relative to the scatter about the line a more conservative estimate of the age uncertainty may be obtained by using the 1σ errors of the two extreme points on the isochron. This results in age uncertainty of +60 and −71 Ma. See text for additional discussion.

TABLE 4. Oxygen Isotopic Compositions for the ~500 Ma Granites

Map No.	Locality	Sample	Material	$\delta^{18}O$ %* Measured	Corrections %o†	$\delta^{18}O$ % Whole Rock
			Gabbro Hills Area			
1	Sage Nunataks	86 BGH 9	quartz	9.16	−2.0	7.16
2	Amphibole Peak	86 BGH 20	plagioclase	7.52	−0.3	7.22
3	Olds Peak	86 BGH 25	quartz	8.00(±7)	−2.0	6.00
4	Longhorn Spurs	86 BGH 31	quartz	8.58(±7)	−2.0	6.58
		Beardmore Glacier Area (Marsh to Shackleton Glaciers Area)				
5	Woodall Peak	85 BCT 167	whole rock	9.75	0.0	9.75
6	O'Leary Peak	85 BCT 171	quartz	10.67	−1.8	8.87
10	Granite Pillars	85 BCT 44	quartz	12.35	−1.8	10.55
11	Cape Goldie	85 BCT 39	quartz	13.63	−1.5	12.13
13	Campbell Hills	85 BCT 31	quartz	14.16	−1.5	12.66
14	N of Mt Heiser	85 BCT 21	quartz	14.20	−1.5	12.70
15	Mt Macbain	85 BCT 116	quartz	12.51	−1.8	10.71
16	Bartrum Plateau	85 BCT 3	plagioclase	9.60	−0.3	9.30
17	Moody Nunatak	85 BCT 8	quartz	11.76	−1.8	9.96
18	Panorama Point	85 BCT 1	quartz	12.58	−1.5	11.08
			Miller Range Area			
19	SE Martin Dome	85 BCT 80	quartz	13.24	−1.5	11.74
20	E Snowshoe Pass	85 BCT 93	quartz	12.53(±5)	−1.5	11.03

*Uncertainty in measured values ≈ 0.2‰. Tabulated uncertainities are 1/2 the difference between replicate measurements and refer to the last decimal place.

†Fractionation corrections: (See text also). Plagioclase-whole rock in a gabbro or diorite was estimated to be 0.3‰. Quartz-whole rock in a tonalite was estimated to be 2.0‰. Quartz-whole rock in a granodiorite was estimated to be 1.8‰. Quartz-whole rock in a granite was estimated to be 1.5‰.

the 50 and 300 km marks, $\delta^{18}O$ is particularly high and variable. In plots of $\delta^{18}O$ versus ε_{Nd} and ε_{Sr} (Figures 8 and 9), the same three groupings are evident with the same geographic relations shown in Figures 7.

STRUCTURAL AND PETROGENETIC IMPLICATIONS

The isotopic data on the metamorphic rocks are useful to interpretations of regional geologic development. For in-

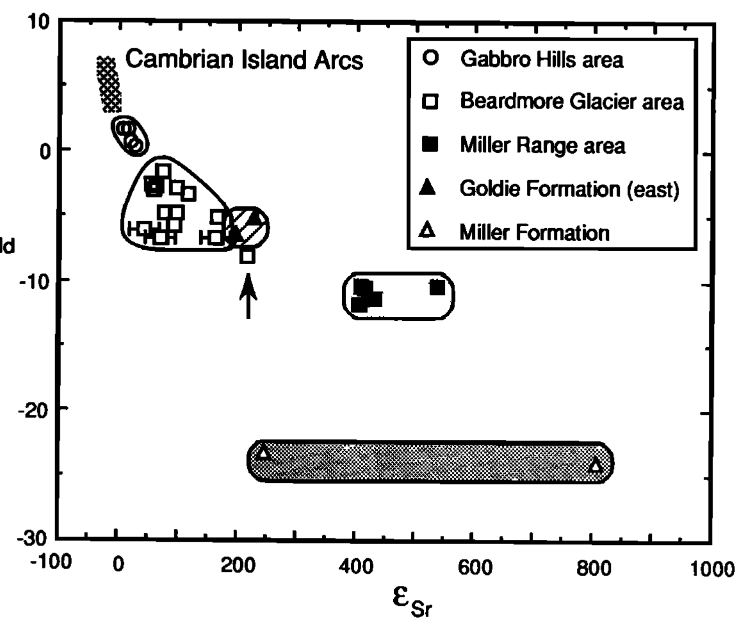

Fig. 6. ε_{Nd} versus ε_{Sr} for the ~500 Ma granites and pre-batholithic metamorphic rocks. Initial Nd and Sr isotopic compositions of the ~500 Ma granites (circles and squares). Data for the metamorphic basement (Goldie and Miller Formations) are calculated at 500 Ma. The granite data define a general array between depleted mantle (as represented by the field estimated for Cambrian island arcs) and crustal rocks which is typical of calc-alkaline batholiths. However, the data also form three clusters correlated with geography as shown. The sample noted by the arrow is from the boundary between the Beardmore Glacier area and the Miller Range area. Archean material of the Nimrod Group is not a component of the ~500 Ma granites. Based on this diagram, metasedimentary rocks of the Goldie Formation may have been a component in some Beardmore Glacier area samples. However, the granites with isotopic compositions consistent with complete melting of the Goldie Formation are not as aluminous as one might expect for minimum melts from this metasedimentary unit.

stance, the Nd isotopic data indicate that our samples of the Goldie Formation are not likely to be correlative with the Argosy formation [cf. *Stump et al.*, 1987] or any other rocks in the Miller Range. *Stump et al.* [1988] correlated Argosy Formation with Cobham Formation and lowermost Goldie Formation in the upper Nimrod Glacier area (Figure 4). In this region the Goldie Formation is lithologically diverse, containing quartzite and fine conglomerate [*Laird et al.*, 1971] and diamictite beds and basaltic pillow lava [*Stump et al.*, 1988] in addition to turbidite beds. It is also structurally complex [*Stump et al.*, 1989]. Recumbent folds with nearly horizontal axial planes are apparently not uncommon [e.g., *Laird et al.*, 1971, Figure 19; *Edgerton*, 1987, photos 13, 14]. In the more easterly part of the range, where our samples were taken, the Goldie Formation is monotonous graywacke and shale with open, upright folds [*Gunn and Walcott*, 1962; *Gunner*, 1976, Figures 5, 6]. Consequently, we suspect that the Goldie Formation at the Cotton Plateau is different from the Goldie Formation in the eastern part of the range and that the Cotton Plateau rocks should be assigned to the Argosy formation for consistency.

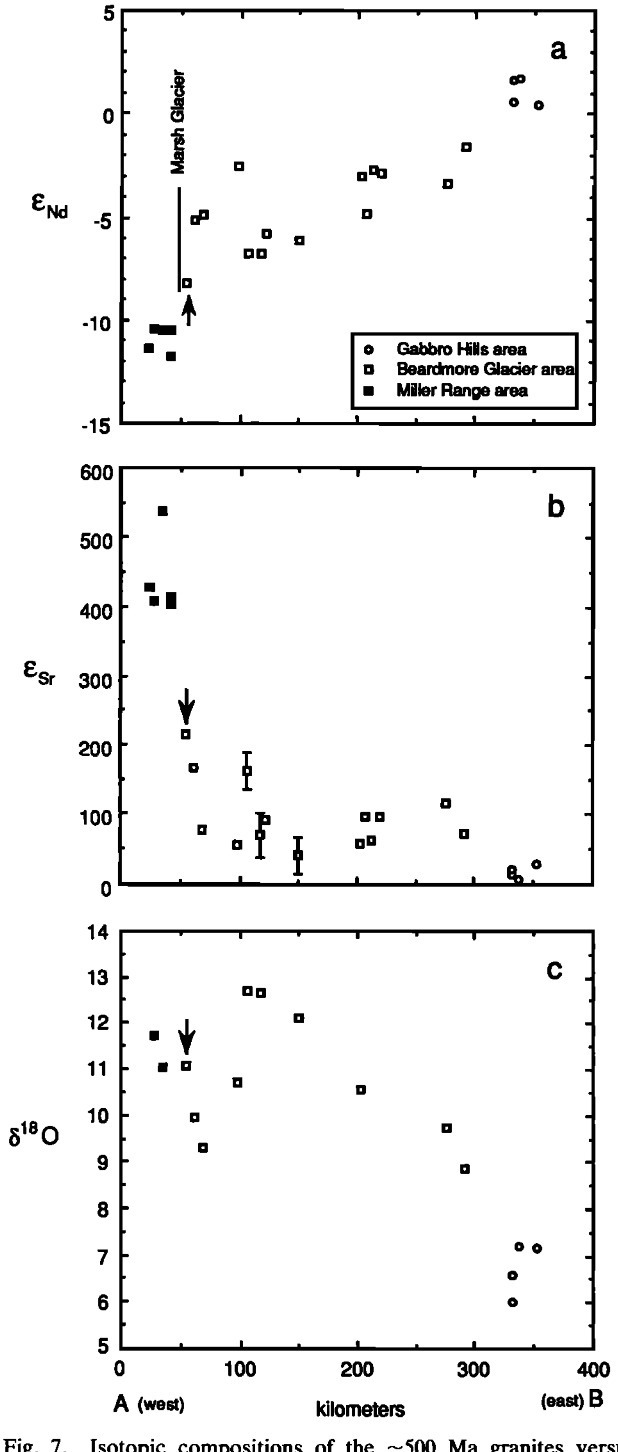

Fig. 7. Isotopic compositions of the ~500 Ma granites versus distance along cross-section A-B. (*a*) ε_{Nd}, (*b*) ε_{Sr}, and (*c*) $\delta^{18}O$ versus distance along cross-section A-B. This diagram shows the relationship between geographic position and isotopic composition for each of the granite groups. The arrow notes sample 85 BCT 1 which lies along the boundary between the Miller Range area and the Beardmore Glacier area. Error bars are shown on initial Sr-isotopic compositions of samples with high Rb/Sr and represent age uncertainty of ±10 m.y. Similar age uncertainty for all other samples results in error bars smaller than the size of the symbols.

Field relations of the Camp Ridge granodiorite (~1.7 Ga) in the Miller Range provide evidence for pre-mid Proterozoic deformations [cf *Stump et al.*, 1989]. The Archean model age of the Camp Ridge granodiorite further indicates that at ~1.7 Ga, the lower crust beneath the Miller Formation was composed of Archean rocks. This contrasts with the interpretation based on the 500 Ma granites that the lower crust has an age of ~2.0 Ga. This is explained by eastward thrusting of the Miller Formation and the Camp Ridge granodiorite to their present position overlying younger crustal rocks [*Goodge and Borg*, 1987]. The time of this thrusting is constrained only to between ~1.7 and ~0.5 Ga. The thrusting could be related to east vergent structures in the Goldie Formation at the Cotton Plateau (and to structures in the Argosy formation) attributed to the Beardmore Orogeny by *Edgerton* [1987] and *Stump et al.* [1989]. In this case, the age of thrusting would be confined to the time between ~750 Ma and ~550 Ma.

Based on the model described earlier in the paper (Figure 2), we interpret the isotopic signatures of the granites in the Miller Range and the Beardmore Glacier area as being formed by melting of Precambrian crystalline lower crust. However, it is known that, in general, continental cratons may be underlain by mantle lithosphere that has isotopic ratios of Sr and Nd similar to those of Precambrian crustal rocks. This is inferred to be the case for the central Transantarctic Mountains based on the isotopic compositions of the Mesozoic Ferrar diabases and Kirkpatrick basalts [e.g., *Elliot et al.*, 1989], although it is not known whether this special "enriched" mantle lithosphere material was present at the time of formation of the granites in the early Paleozoic. Based solely on the Sr and Nd data we can not rule out the possibility that the parental magmas of the granites in the Miller Range and Beardmore Glacier area were derived from subcontinental mantle lithosphere. However, the oxygen isotopic data argue strongly against this possibility; there are no known magmas derived from mantle sources, and no known mantle samples, that have $\delta^{18}O$ values higher than about +8.5, and even values this high are rare.

For the Miller Range granites, the peraluminous character of all samples strongly supports the interpretation of crustal derivation and also indicates a large metasedimentary component in the lower crust. Furthermore, the tight clustering of Sr, Nd, and O isotopic compositions allows the additional interpretation that the lower crust in the Miller Range was relatively homogeneous. Sm-Nd model ages calculated using

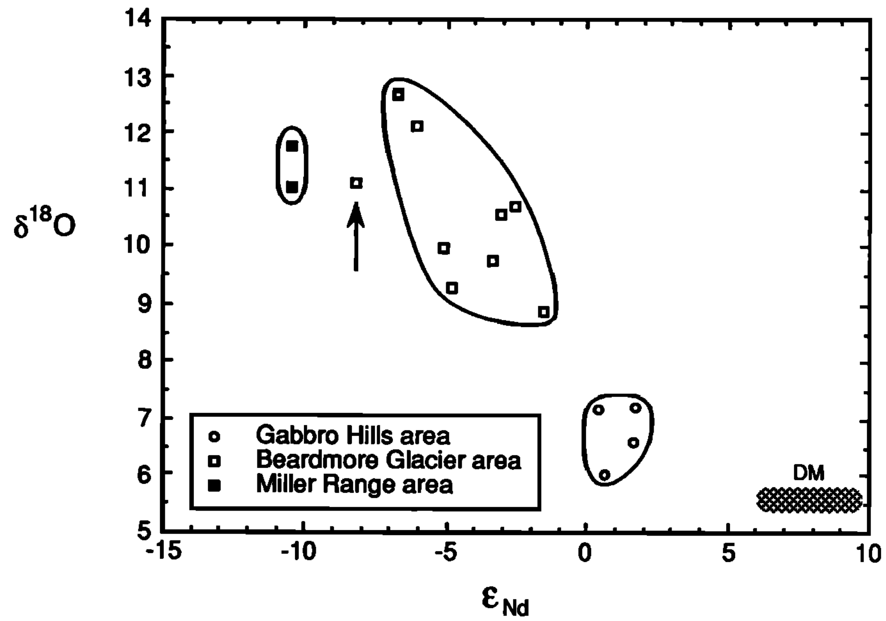

Fig. 8. $\delta^{18}O$ versus ε_{Nd} for the ~500 Ma granites.

initial ε_{Nd} values and average crustal $^{147}Sm/^{144}Nd$ values [*DePaolo*, 1988] are ~2.0 Ga. Therefore, the Miller Range area corresponds to Type-C crust, or Type-B crust far from a subduction zone (Figure 2), but with a thrust sheet of reworked Archean material in the upper crust rather than sedimentary cover.

For the Beardmore Glacier area granites, the interpretation of crustal derivation is supported by the presence of some slightly peraluminous granites. Furthermore, the lithologic variation from monzogranite to diorite with no definitive correlation between lithology and isotopic composition (low, but variable, ε_{Nd} values and high, but variable, ε_{Sr} and $\delta^{18}O$ values) allows the interpretation that the lower crust in this region was, lithologically, relatively heterogeneous. Also, although the Goldie Formation cannot be ruled out as a magma source based only on the isotopic compositions, the fact that both diorites and granites in the region have essentially the same isotopic compositions argues against the Goldie as a primary source because the Goldie Formation could produce only peraluminous melts. The ε_{Nd} values of granites in the Beardmore Glacier area correspond to Sm-Nd model ages of 1.3–1.8 Ga. The high $\delta^{18}O$ values indicate that the Precambrian basement here contains a substantial component of metasedimentary rock while the lack of strongly

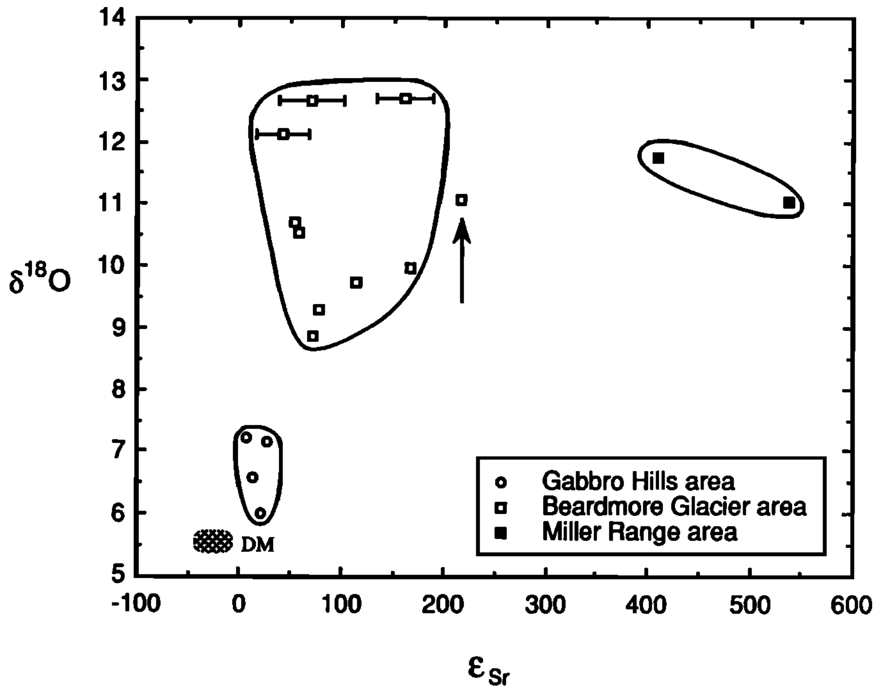

Fig. 9. $\delta^{18}O$ versus ε_{Sr} for the ~500 Ma granites.

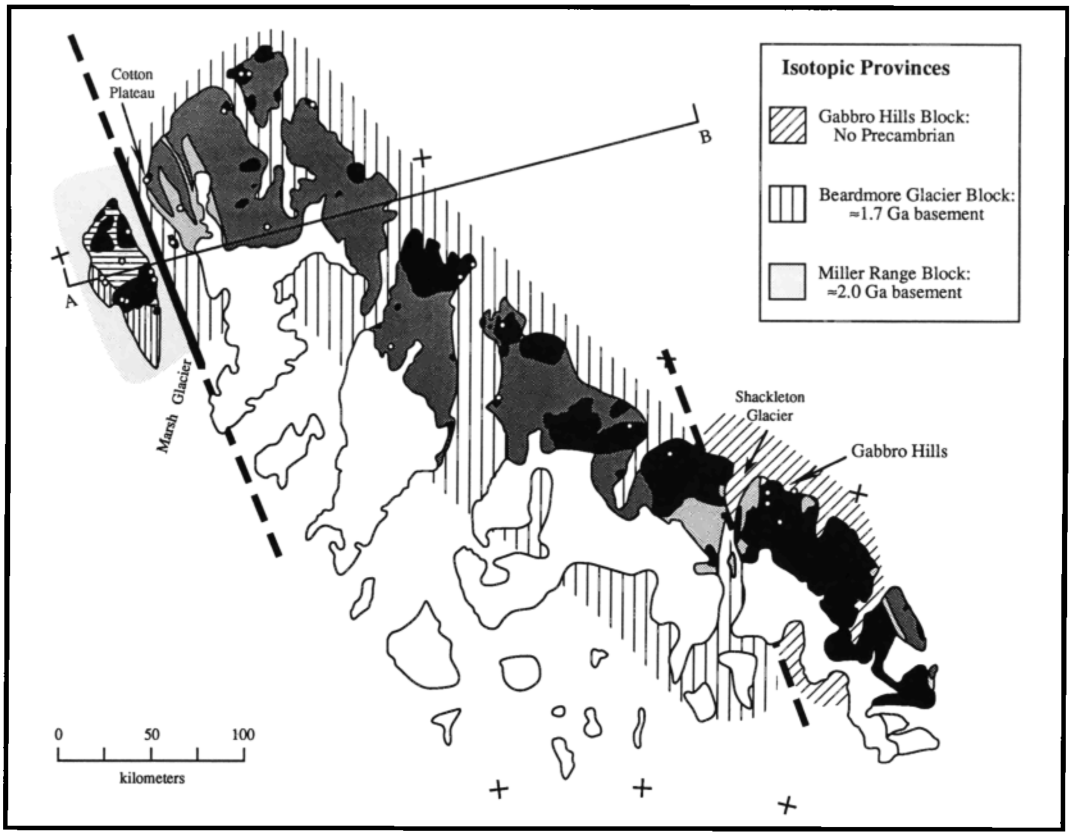

Fig. 10*a*. Isotopic provinces of the central Transantarctic Mountains. The area covered is the same as Figure 4.

peraluminous granites and the presence of metaluminous granites suggests that pelitic rocks were not abundant. We interpret the Beardmore Glacier area as being Type B continental crust (Figure 2).

Model ages of granites in the Beardmore Glacier area are distinct from those of the granites in the Miller Range, implying that the Marsh Glacier lies astride a crustal age province boundary. The west-to-east trends toward slightly higher ε_{Nd} values and slightly lower $\delta^{18}O$ values within the Beardmore Glacier area suggests that in the easternmost part of this region the granitic magmas may have included minor components of high-ε_{Nd} mantle-derived magma.

The positive ε_{Nd} values, low ε_{Sr} values, and low $\delta^{18}O$ values of the granitic rocks of the Gabbro Hills area indicate that the magmas were derived mainly by differentiation of mantle-derived magmas with small amounts of assimilated continental material, probably in the form of sedimentary rocks. A depleted mantle source is indicated by the high ε_{Nd} values and this means that the enriched mantle inferred from the Ferrar magmas was either not yet formed or absent in this region. Had enriched mantle been the source of magma at ~500 Ma we would expect ε_{Nd} values no higher than about −2 based on Ferrar compositions summarized by *Elliot et al.* [1989] and extrapolated from 180 Ma to ~500 Ma using estimated $^{147}Sm/^{144}Nd$ ratios as enriched as 0.13. We infer that the area we have sampled east of the Shackleton Glacier does not contain Precambrian crystalline basement, and that this area is Type-A crust (Figure 2).

The crustal structure in the central Transantarctic Mountains at the time of granite emplacement (~500 Ma), as inferred from the isotopic data, is shown on Figure 10 (10*a*-map view; 10*b*-composition section). The lower crustal provinces defined by the granites are referred to as the Miller Range Block (MRB), the Beardmore Glacier Block (BGB), and the Gabbro Hills Block (GHB).

Discussion

Allochthonous Nature of the Beardmore Microcontinent

The Marsh Glacier marks the eastern boundary of the MRB on the margin of the East Antarctic Craton and is inferred to mark the location of a suture between the MRB and BGB (Figure 10). The BGB and the GHB, which we refer to together as the Beardmore microcontinent, are allochthonous and possibly far-travelled terranes. This inference is born out by (1) the discontinuity in isotopic compositions of the ~500 Ma granites at the Marsh Glacier, which indicates a sharp boundary in the lower crust between two different continental blocks, (2) the Nd isotopic data on the Goldie Formation (of the type east of the Cotton Plateau), which preclude derivation from crustal sources on the East Antarctic Craton as represented by the rocks exposed in the Miller Range, and (3) the interpretation that the gabbro and basalt associated with the Cotton Plateau-type Goldie Formation (adjacent to the Marsh Glacier) were formed in an oceanic setting at ~760 Ma.

Within this framework, the metamorphic basement rocks correspond well to the lower crustal provinces or to the suture between the MRB and BGB. Specifically, the Miller Formation characterizes the East Antarctic Craton (MRB),

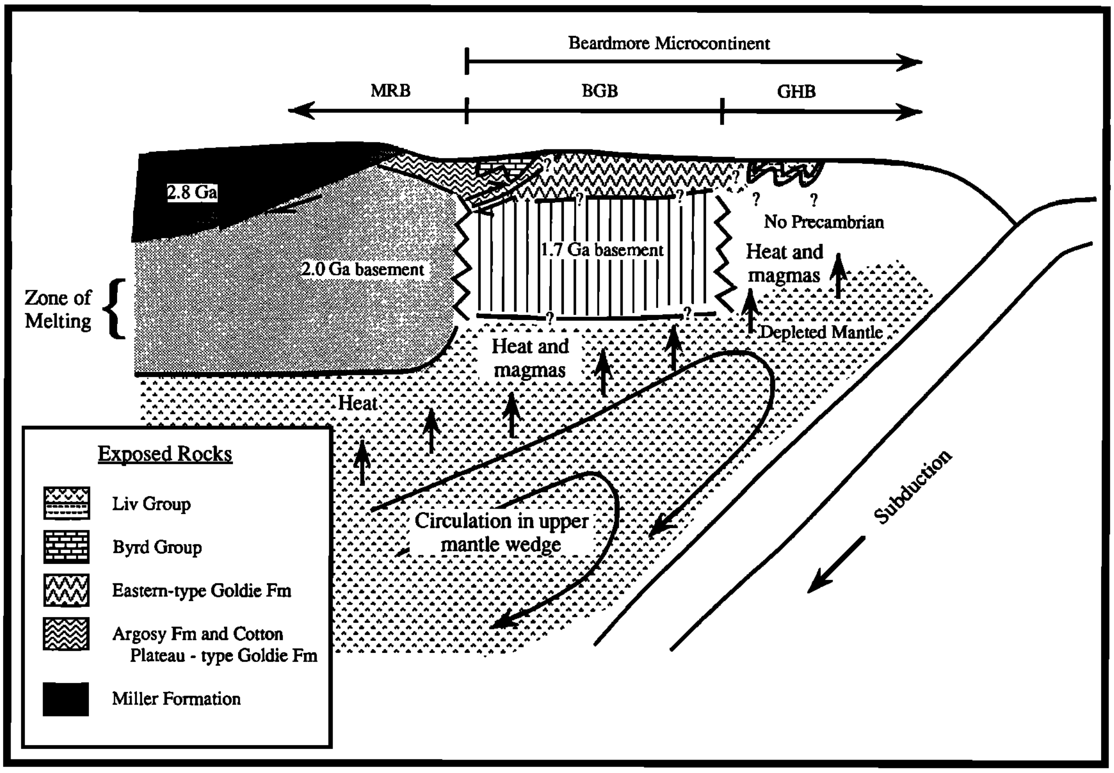

Fig. 10*b*. Composite cross-sectional sketch illustrating the crustal structure and tectonic setting inferred for the central Transantarctic Mountains after folding of the Ross Orogeny and during emplacement of the ~500 Ma granites.

whereas the eastern-type Goldie Formation along with the Liv Group characterize the Beardmore microcontinent (BGB and GHB). The Cotton Plateau-type Goldie Formation (including the basalt and gabbro), the Cobham Formation, and the Argosy formation characterize the boundary between the craton and the microcontinent and represent the material occluded between these blocks during accretion. Because of present proximity, the Byrd Group appears to be associated with the suture terrane although it may represent a depositional setting straddling the suture between the craton and the microcontinent. It is noteworthy that the Byrd Group has not been found more than about 40 km east of the suture (Figure 4). Whether or not this formation was deposited as an overlap assemblage is important as a possible constraint to timing of accretion of the Beardmore microcontinent.

A thrust fault, active between ~760 and ~500 Ma, placing Cotton Plateau-type Goldie Formation and associated rocks over the boundary is inferred from the interpretation that these rocks were deposited on oceanic crust at about 760 Ma, and that the granite (85 BCT 1) that intruded them at ~500 Ma was derived from continental basement. This inference is consistent with the existence of east-vergent early structures in the Goldie Formation at Cotton Plateau [*Edgerton*, 1987; *Stump et al.*, 1989]. Thus, the oceanic basin which must have existed at ~760 Ma along the margin of the East Antarctic Craton was closed by accretion of the Beardmore microcontinent prior to emplacement of the granites at ~500 Ma.

Accretion of the Beardmore microcontinent to the East Antarctic Craton is constrained to be after deposition of the (Cotton Plateau-type) Goldie Formation at ~760 Ma and earlier than the granitic magmatism at ~500 Ma. However, if the Byrd Group was deposited as an overlap assemblage on both the Cotton Plateau-type Goldie Formation and the eastern-type Goldie Formation, then accretion occurred prior to ~550 Ma. Because there are apparently no east vergent structures in the Byrd Group it is not likely that accretion post-dated deposition of the Byrd Group.

One problem with the interpretation of the Beardmore microcontinent as allochthonous is that there have been no igneous rocks identified that might correspond to the subduction that must have occurred to close the ocean basin between the Beardmore microcontinent and the East Antarctic Craton. This is not an insurmountable problem because the convergence may have been highly oblique, or the subduction may have been at a very low angle so that no, or very little, magmatism was produced at the margin. On the other hand, the Miller Range is a relatively small area and may not be fully representative. We have recently identified several deformed plutons north of the Byrd Glacier which may be correlative with the basement rocks of the Miller Range. One of these is the Carlyon Granodiorite, which gave an Rb-Sr age of 568 Ma [*Felder*, 1980; *Felder and Faure*, 1980]. This granite may have been associated with accretion of the Beardmore microcontinent and with more work, it may be possible to document magmatism associated with Beardmore accretion. A tectonic model for the development of the region, showing a logical sequence of events starting at ~750 Ma and ending with the granitic magmatism at ~500 Ma, is shown in Figure 11.

Subduction and Early Paleozoic Granitic Magmatism

The isotopic compositions of the ~500 Ma plutonic rocks in the Gabbro Hills region indicate a large component of

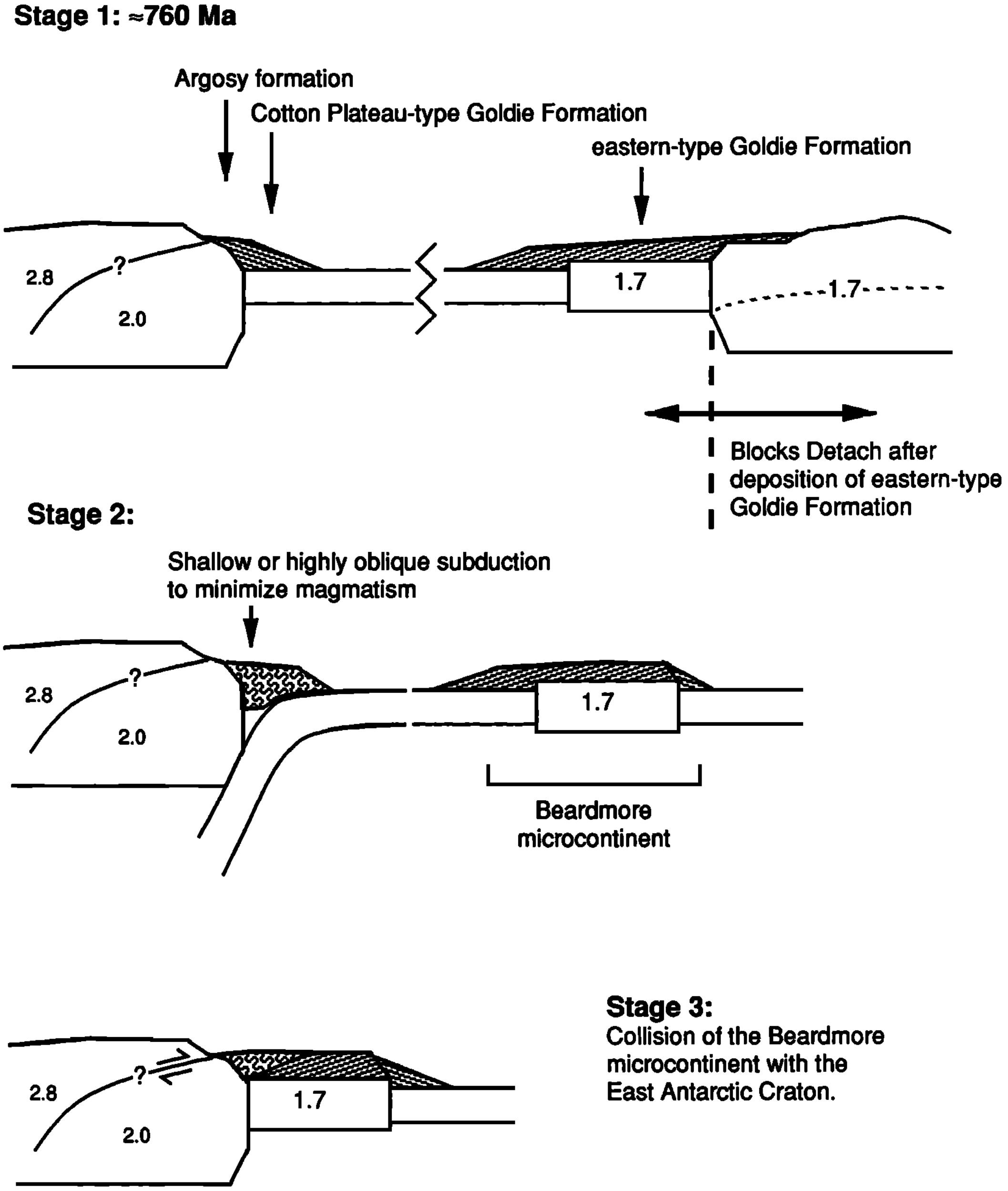

Fig. 11. Composite cross-sectional sketches illustrating a tectonic model for the development of the central Transantarctic Mountains from ~760 Ma through emplacement of the ~500 Ma granites.

magma derived from depleted, oceanic-type mantle sources. We interpret this as evidence of proximity to a westward dipping subduction zone which was responsible for generating the granitic magmas both in this region and in the terranes to the west. This inference is supported by the observation that the early Paleozoic metamorphic country rocks in the Gabbro Hills area are largely volcanic and volcaniclastic (the Liv Group) and that Precambrian continental basement is absent.

The crustal provinces are depicted in association with a convergent continental margin in our model for granitic petrogenesis (Figure 10*b*). Oceanic lithosphere was subducted to the west beneath a collage of provinces which made up the early Paleozoic margin of the East Antarctic Craton. Subduction-related mantle-derived magmas were parental to granitic rocks in the Gabbro Hills area and in the eastern part of the Beardmore Glacier block. Farther inland there is little evidence of mantle-derived magma. Granitic magmas were generated by lower-crustal melting caused by heat supplied by advecting material in the upper mantle wedge.

The evidence for a subduction zone setting for the granitic rocks of the Transantarctic Mountains is sparse. Nevertheless, there is sufficient similarity between the observed isotopic and geochemical patterns and those of Mesozoic granitic batholiths in western North America [*Kistler and Peterman*, 1973; *Taylor and Silver*, 1978; *DePaolo*, 1981; *Farmer and DePaolo*, 1983] to warrant the conclusion that a subduction zone was involved. However, the bulk of the granitic rocks in the Transantarctic Mountains, including those described by *Borg et al.* [1987*b*] in the Wilson Terrane of Northern Victoria Land, appear to be the products of crustal anatexis with only small amounts of magma derived

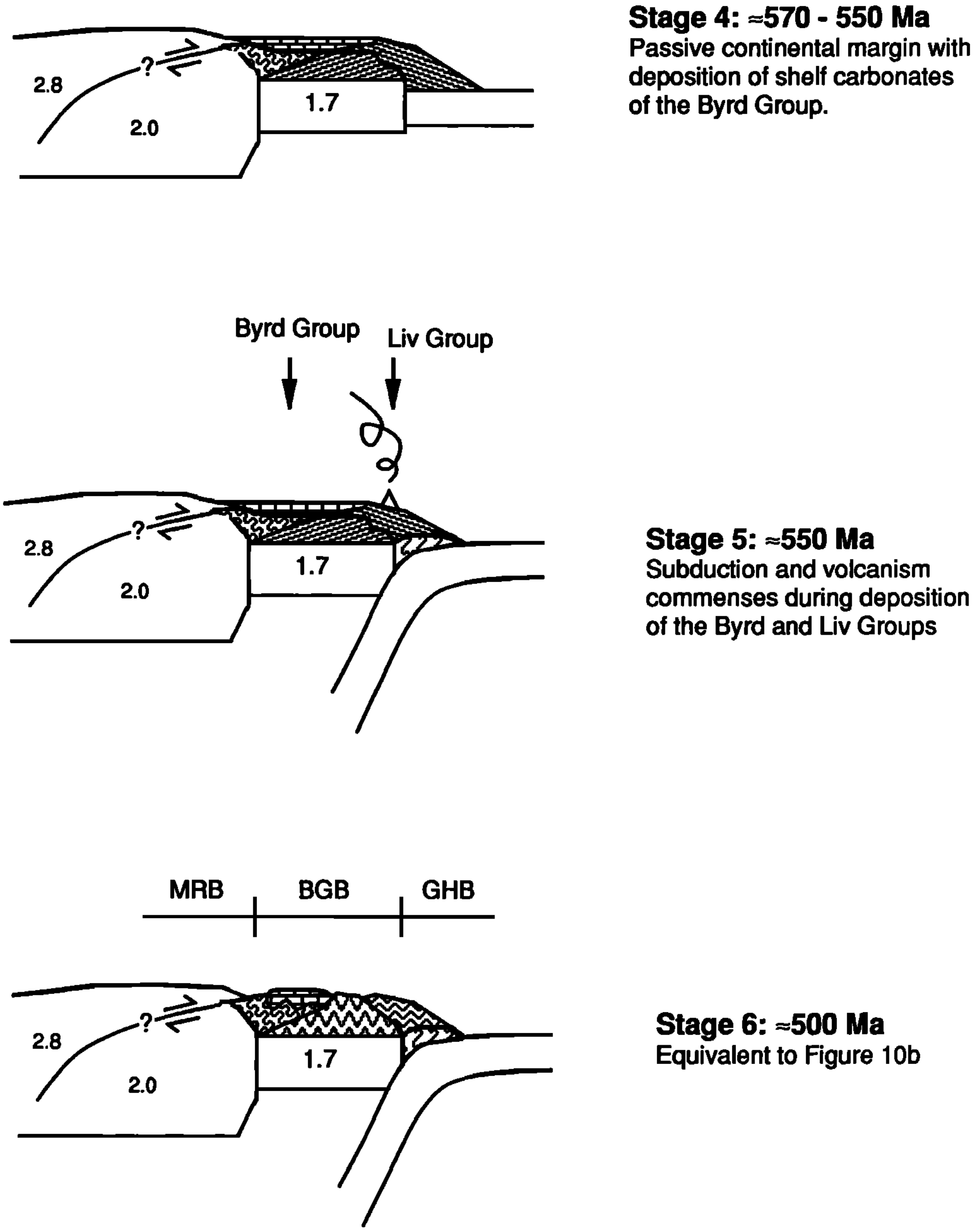

Fig. 11. (continued)

from the mantle. This aspect of early Paleozoic occurrences of granitic magmatism worldwide is noteworthy [*Pitcher*, 1987]. If the existing data are not biased, this characteristic distinguishes early Paleozoic batholiths from Mesozoic circum-Pacific batholiths. This suggests that the subduction process that characterized the closure of early Paleozoic oceans was different from that of the Mesozoic in that the former generally failed to produce island arc type crust. This conclusion, however, could be premature to the extent that the vast majority of Paleozoic batholiths occur in Asia, and they are not well studied. Indeed, recent work in the Canadian Cordillera suggests crustal growth by production of island arcs from the mantle throughout the Phanerozoic [*Samson et al.*, 1989]. Preservational bias is another problem because it is possible that island arc type crust is more easily destroyed by later tectonic reworking.

The Precambrian of the East Antarctic Craton

The isotopic data on the rocks of the Miller Range allow some inferences to be made about the structure and Precambrian development of this part of the Antarctic craton. The existence of a 1.7 Ga granodiorite intruded into and derived from Archean crust suggests that a 1.7 Ga orogenic belt was superimposed on the edge of an Archean craton. This indicates that the Antarctic craton west of the Miller Range under the polar ice cap is likely to be of Archean age. It is also consistent with the 2.0 Ga model ages of the granitic rocks of the Miller Range. These model ages, which exceed the crystallization age of the Camp Ridge granodiorite by 0.3 Ga, are characteristic of Proterozoic crustal belts that formed at the fringes of Archean cratons [*Bennett and DePaolo*, 1987]. The relatively old model ages result from incorporation of sedimentary material derived from the Archean craton into the crust formed in the 1.7 Ga orogenic event. This same sediment component helps account for the high $\delta^{18}O$ values of the Paleozoic granites formed by anatexis of the cratonic rocks [cf. *Bennett et al.*, 1987].

CONCLUSIONS

Isotopic data on ~500 Ma granitic rocks have allowed three basement age provinces to be identified in the central Transantarctic Mountains. Integration of these data with geological and isotopic data on the country rocks leads to a revised model for the late Precambrian and early Paleozoic tectonics of this portion of the Gondwana margin.

The oldest of the three age provinces is exposed in the Miller Range, on the inland side of the Transantarctic Mountains. This range is shown by Sm-Nd isotopic measurements to contain Archean and Proterozoic rocks that are part of the East Antarctic Craton. Metasedimentary rocks in this area have an Archean provenance age. The Archean gneisses are present only in the uppermost crust, mostly above a major low-angle shear zone [*Goodge and Borg*, 1987]. The lower crust has a Sm-Nd model age of ~2.0 Ga as indicated by the isotopic composition of 500 Ma peraluminous granites which pierce the veneer of Archean rocks.

To the east of the Miller Range, the Nd isotopic compositions of the ~500 Ma granitic rocks, coupled with the O-isotopic evidence for crustal derivation, indicate that the crust has a model age of about 1.6 ± 0.2 Ga. The boundary between this age province and the older one in the Miller Range lies beneath the Marsh Glacier. This boundary is inferred to be a suture because of the interpretation that the late Precambrian sedimentary rocks now found at the boundary were deposited on oceanic crust at about 760 Ma. This inference is supported by the interpretation that the metasedimentary rocks east of the Cotton Plateau have a provenance age of 1.6–1.7 Ga, and could not have been derived from the East Antarctic Craton in this area. The terrane east of the Marsh Glacier and extending to the Gabbro Hills, which we call the Beardmore microcontinent, is therefore allochthonous to the East Antarctic Craton and possibly far-travelled.

The rocks of the Gabbro Hills, east of the Shackleton Glacier and some 300 km east of the Marsh Glacier, define the third basement province. This region is composed of ~500 Ma granodiorite, diorite, tonalite, and gabbro which have mantle-like isotopic characteristics and intrude volcanic and volcaniclastic country rocks. This area contains no Precambrian basement and is interpreted as a continental margin arc assemblage formed at the eastern margin of the Beardmore microcontinent. The presence of this terrane east of the older crustal basement blocks suggests that a westward-dipping subduction zone was responsible for the granitic magmatism throughout the central Transantarctic Mountains at ~500 Ma. The Gabbro Hills area is one of very few examples of early Paleozoic granitic batholiths that have an island-arc geochemical character.

The results of this study put new constraints on the timing of deformational events in the Transantarctic Mountains and suggest that significant revisions to the stratigraphy may be required. The deformation of the Miller Formation in the Miller Range is now known to have occurred before emplacement of a 1.7 Ga granodiorite. This deformation is thus not related to late Precambrian–early Paleozoic tectonic events. Eastward thrusting of the Miller Formation is constrained to the interval between 1.7 Ga and 500 Ma and this thrusting may be related to accretion of the Beardmore microcontinent. The Beardmore microcontinent was accreted to east Antarctica before emplacement of the granites at ~500 Ma, but it may have preceded deposition of the Byrd Group at ~550 Ma. Accretion occurred after 760 Ma, the age of the oceanic-type basalt and gabbro of the Cotton Plateau. This accretion event may have caused the folding of the Argosy formation (in the Miller Range) and the Goldie Formation (in the Cotton Plateau) and thus would constitute what has heretofore been referred to as the Beardmore Orogeny. The similar tectonic histories of the Argosy Formation of the Miller Range and the "Goldie Formation" of the Cotton Plateau suggest that they may be correlative [*Stump et al.*, 1988] but isotopic data suggests these are distinct from the Goldie Formation east of the Cotton Plateau. Reconsideration of the terms "Goldie Formation" and "Beardmore Group" is clearly warranted.

Acknowledgments. Funding for this research was provided by National Science Foundation Grants DPP83-16807 and DPP86-14649. Special thanks to John Goodge, James Mattinson, Brad K. Smith, and Victoria Bennett for collaboration on portions of the field work, and to Victoria Bennett, James Mattinson, and Edmund Stump for unpublished data. Logistic support for this research was provided by the National Science Foundation (Division of Polar Programs, Polar Operations Section), the U.S. Navy (Naval Support Force Antarctica and Squadron VXE-6), and Antarctic Services Inc.

REFERENCES

Adams, C. J. D., J. E. Gabites, and G.W. Grindley, Orogenic history of the Central Transantarctic Mountains: New K-Ar age data on the Precambrian-Lower Paleozoic basement, in *Antarctic Geoscience*, edited by C. Craddock, pp. 817–826, University of Wisconsin Press, Madison, 1982.

Bennett, V. C., and D. J. DePaolo, Proterozoic crustal history of the western United States as determined by neodymium isotopic mapping, *Geol. Soc. Am. Bull.*, *99(5)*, 674–685, 1987.

Bennett, V. C., D. J. DePaolo, and B. M. Smith, Neodymium and oxygen isotopic evidence for the origin of middle Proterozoic anorogenic granites of the western United States, *Geol. Soc. Am. Abstr. Program.* *19(7)*, 586, 1987.

Borg, S. G., Petrology and geochemistry of the Queen Maud Batholith, central Transantarctic Mountains, with implications for the Ross Orogeny, in *Antarctic Earth Science*, edited by R. L. Oliver, P. R. James, and J. B. Jago, p. 165–169, Australian Academy of Sci., Canberra, 1983.

Borg, S. G., Granitoids of northern Victoria Land, Antarctica, Ph.D. thesis, 355 pp., 1 plate, Arizona State Univ., Tempe, 1984.

Borg, S. G., and D. J. DePaolo, Nd isotopic evidence for ≈1.8 Ga crust in the northeastern Robertson Bay Terrane, northern Victoria Land, and implications for tectonic models (abstract), paper presented at 5th International Symposium Antarctic Earth Science, Sci. Comm. on Antarct. Res., Cambridge, U. K., August 1987.

Borg, S. G., and E. Stump, Paleozoic magmatism and associated tectonic problems of northern Victoria Land, Antarctica, in *Gondwana Six: Structure, Tectonics, and Geophysics*, edited by G. D. McKenzie, pp. 67–75, *Geophys. Monogr. Ser.*, *40*, AGU, Washington, D. C., 1987.

Borg, S. G., B. W. Chappell, M. T. McCulloch, E. Stump, D. Wyborn, and J. R. Holloway, Compositional polarity of granitoids with implications to regional geology, northern Victoria Land, Antarctica (abstract), *Geol. Soc. Am. Abstr. Prog.*, *16*, 450, 1984.

Borg, S. G., J. W. Goodge, D. J. DePaolo, and J. Mattinson, Field studies of granites and metamorphic rocks: Central Transantarctic Mountains, Antarctica, *Antarctic J. U.S.*, *21(5)*, 43–45, 1986*a*.

Borg, S. G., E. Stump, and J. R. Holloway, Granitoids of northern Victoria Land, Antarctica: A reconnaissance study of field relations, petrography, and geochemistry, in *Geological Investiga-*

tions in Northern Victoria Land, Antarctica, Antarctic Res. Ser., vol. 46, edited by E. Stump, pp. 115–188, AGU, Washington, D. C., 1986*b*.

Borg, S. G., J. W. Goodge, V. C. Bennett, D. J. DePaolo, and B. K. Smith, Geochemistry of granites and metamorphic rocks, Central Transantarctic Mountains, *Antarctic J. U.S.*, *22(5)*, 21–23, 1987*a*.

Borg, S. G., E. Stump, B. W. Chappell, M. T. McCulloch, D. Wyborn, R. L. Armstrong, and J. R. Holloway, Granitoids of northern Victoria Land, Antarctica: Implications of chemical and isotopic variations to regional crustal structure and tectonics, *Am. J. Sci.*, *287*, 127–169, 1987*b*.

Bottinga, Y., and M. Javoy, Oxygen isotope partitioning among the minerals in igneous and metamorphic rocks, *Rev. Geophys. Space Phys.*, *13*, 401–418, 1975.

Clayton, R. N., and T. K. Mayeda, The use of bromine pentafluoride in the extraction of oxygen from oxides and silicates for isotopic analysis, *Geochim. Cosmochim. Acta*, *27*, 43–52, 1963.

Craddock, C., Antarctic tectonics, in *Antarctic Geology and Geophysics* edited by R. J. Adie, pp. 449–455, Universitetsforlaget, Oslo, 1972.

DePaolo, D. J., Study of magma sources, mantle structure and the differentiation of the earth from variations of $^{143}Nd/^{144}Nd$ in igneous rocks, Ph.D. thesis, 360 pp., Calif. Inst. of Technol., Pasadena, 1978.

DePaolo, D. J., A neodymium and strontium isotopic study of Mesozoic calc-alkaline granitic batholiths of the Sierra Nevada and Peninsular Ranges, California, *J. Geophys. Res.*, *86*, 10,470–10,488, 1981.

DePaolo, D. J., Age dependence of the composition of continental crust: Evidence from Nd isotopic variations in granitic rocks, *Earth Planet. Sci. Lett.*, *90*, 263–271, 1988.

DePaolo, D. J., and G. J. Wasserburg, Inferences about magma sources and mantle structure from variations of $^{143}Nd/^{144}Nd$, *Geophys. Res. Lett.*, *3*(12), 743–746, 1976.

DePaolo, D. J., and G. J. Wasserburg, The sources of island arcs as indicated by Nd and Sr isotopic studies, *Geophys. Res. Lett.*, *4*(10), 465–468, 1977.

DePaolo, D. J., and G. J. Wasserburg, Petrogenetic mixing models and Nd-Sr isotopic patterns, *Geochim. Cosmochim. Acta*, *43*, 615–627, 1979.

Edgerton, D. G., Kinematic orientations of Precambrian Beardmore and Cambro-Ordovician Ross orogenies, Goldie Formation, Nimrod Glacier area, Antarctica, M. S. thesis, 135 pp., 1 map plate, Arizona State Univ., Tempe, 1987.

Elliot, D. H., Tectonics of Antarctica: A review, *Am. J. Sci.*, *275-A*, 45–106, 1975.

Elliot, D. H., P. R. Kyle, and R. J. Pankhurst, Jurassic basic magmatism and crustal anatexis in Antarctica, *28th Int. Geol. Congress Abstr.*, *1*, 447, 1989.

Farmer, G. L., and D. J. DePaolo, Origin of Mesozoic and Tertiary granite in the western United States and implications for Pre-Mesozoic crustal structure, 1, Nd and Sr isotopic studies in the geocline of the Northern Great Basin, *J. Geophys. Res.*, *88*, 3379–3401, 1983.

Farmer, G. L., and D. J. DePaolo, Origin of Mesozoic and Tertiary granite in the western United States and implications for Pre-Mesozoic crustal structure, 2, Nd and Sr isotopic studies of unmineralized and Cu- and Mo-mineralized granite in the Precambrian Craton, *J. Geophys. Res.*, *89(B12)*, 10,141–10,160, 1984.

Faure, G., J. G. Murtaugh, and R. J. E. Montigny, The geology and geochronology of the basement complex of the central Transantarctic Mountains, *Can. J. Earth Sci.*, *5*, 555–560, 1968.

Faure, G., R. Eastin, P. T. Ray, D. McLelland, and C. Shultz, Geochronology of igneous and metamorphic rocks, central Transantarctic Mountains, in *Fourth International Gondwana Symposium Papers*, vol. 1, edited by B. Laskar and C. S. Rao, pp. 805–813, Hindustan, Delhi, India, 1979.

Felder, R. P., Geochronology of the Brown Hills, Transantarctic Mountains, M.S. thesis, 110 pp., Ohio State Univ., Columbus, 1980.

Felder, R. P., and G. Faure, Rubidium-strontium age determination of part of the basement complex of the Brown Hills, central Transantarctic Mountains. *Antarctic J. U. S.*, *15*(5), 16–17, 1980.

Goodge, J. W., and S. G. Borg, Metamorphism and crustal structure in the Miller Range, central Transantarctic Mountains (abstract), paper presented at 5th International Symposium Antarctic Earth Science, Sci. Comm. on Antarct. Res., Cambridge, U. K., August 1987.

Grindley, G. W., The geology of the Queen Alexandra Range, Beardmore Glacier, Ross Dependency, Antarctica, *N. Z. J. Geol. Geophys.*, *6*, 307–347, 1963.

Grindley, G. W., Polyphase deformation of the Precambrian Nimrod Group, central Transantarctic Mountains, in *Antarctic Geology and Geophysics*, edited by R. J. Adie, pp. 313–318, Universitetsforlaget, Oslo, 1972.

Grindley, G. W., Precambrian rocks of the Ross Sea region, *R. Soc. N. Z. J.*, *11*, 411–423, 1981.

Grindley, G. W., and M. G. Laird, Geology of the Shackleton Coast, scale 1:1,000,000, *Antarct. Map Folio Ser.*, folio 12, plate XV, edited by V. C. Bushnell, Am. Geogr. Soc., New York, 1969.

Grindley, G. W., and I. McDougall, Age and correlation of the Nimrod Group and other Precambrian rock units in the central Transantarctic Mountains, Antarctica, *N.Z. J. Geol. Geophys.*, *12*, 391–411, 1969.

Grindley, G. W., and G. Warren, Stratigraphic nomenclature and correlation in the western Ross Sea region, in *Antarctic Geology*, edited by R. J. Adie, pp. 314–333, North-Holland, Amsterdam, 1964.

Grindley, G. W., V. R. McGregor, and R. I. Walcott, Outline of the geology of the Nimrod-Beardmore-Axel Heiberg Glaciers region, Ross Dependency, in *Antarctic Geology*, edited by R. J. Adie, pp. 206–219, North Holland, Amsterdam, 1964.

Grow, J. A., R. E. Mattick, and J. S. Schlee, Multichannel seismic depth sections and interval velocities over outer continental shelf and upper continental slope between Cape Hatteras and Cape Cod, in *Geological and Geophysical Investigations of Continental Margins*, edited by J. S. Watkins, L. Montadert, and P. W. Dickerson, pp. 65–83, American Association of Petroleum Geologists, Tulsa, Okla., 1979.

Gunn, G. M., and R. I. Walcott, The geology of the Mt. Markham region, Ross Dependency, Antarctica, *N. Z. J. Geol. Geophys.*, *5*, 407–426, 1962.

Gunn, G. M., and G. Warren, Geology of Victoria Land between the Mawson and Mullock Glaciers, Antarctica, *N. Z. Geol. Survey Bull.*, *71*, 157 pp., 1962.

Gunner, J., Age and origin of the Nimrod Group and of the Granite Harbour Intrusives, Beardmore Glacier region, Antarctica, Ph.D. thesis, 231 pp., Ohio State Univ., Columbus, 1971.

Gunner, J., Isotopic and geochemical studies of the pre-Devonian basement complex, Beardmore Glacier region, Antarctica, *Rep. 41*, 126 pp., Inst. of Polar Stud., Ohio State Univ., Columbus, 1976.

Gunner, J., and G. Faure, Rb-Sr geochronology of the Nimrod Group, central Transantarctic Mountains, in *Antarctic Geology and Geophysics*, edited by R. J. Adie, pp. 305–312, Universitetsforlaget, Oslo, 1972.

Harris, N. B. W., J. A. Pearce, and A. G. Tindle, Geochemical characteristics of collision-zone magmatism, *Geol. Soc. London Spec. Pub.*, *19*, 67–81, 1986.

Hawkesworth, C. J., M. Hammill, A. R. Gledhill, P. van Calsteren, and G. Rogers, Isotope and trace element evidence for late-stage intra-crustal melting in the High Andes, *Earth Planet. Sci. Lett.*, *58*, 240–254, 1982.

Irvine, T. N., and W. R. A. Baragar, A guide to the chemical classification of the common volcanic rocks, *Can. J. Earth Sci.*, *8*, 523–548, 1971.

James, P. R., and R. J. Tingey, The Precambrian evolution of the East Antarctic metamorphic shield—A review, in *Antarctic Earth Science* edited by R. L. Oliver, P. R. James, and J. B. Jago, pp. 5–10, Australian Academy of Science, Canberra, 1983.

Kistler, R. W., and Z. E. Peterman, Variations in Sr, Rb, K, Na, and initial $_{87}Sr/_{86}Sr$ in Mesozoic granitic rocks and intruded wall rocks in central California, *Geol. Soc. Am. Bull.*, *84*, 3489–3512, 1973.

Kistler, R. W., and Z. E. Peterman, Reconstruction of crustal blocks of California on the basis of initial strontium isotopic compositions of Mesozoic granitic rocks, *U. S. Geol. Surv. Prof. Pap.*, *1071*, 17 pp., 1978.

Laird, M. G., and J. D. Bradshaw, Uppermost Proterozoic and Lower Paleozoic geology of the Transantarctic Mountains (review paper), in *Antarctic Geoscience*, edited by C. Craddock, pp. 525–534, University of Wisconsin Press, Madison, 1982.

Laird, M. G., and J. B. Waterhouse, Archeocyathine limestones of Antarctica, *Nature*, *194*(4831), 861, 1962.
Laird, M. G., G. D. Mansergh, and J. M. A. Chappell, Geology of the central Nimrod Glacier area, Antarctica, *N. Z. J. Geol. Geophys.*, *14*, 427–468, 1971.
Liew, T. C., and M. T. McCulloch, Genesis of granitoid batholiths of Peninsular Malaysia and implications for models of crustal evolution: Evidence from a Nd-Sr isotopic and U-Pb zircon study, *Geochim. Cosmochim. Acta*, *49*, 587–600, 1985.
McGregor, V. R., Geology of the area between the Axel Heiberg and Shackleton Glaciers, Queen Maud Range, Antarctica: Part 1, Basement complex, structure, and glacial geology, *N. Z. J. Geol. Geophys.*, *8*, 314–343, 1965.
McGregor, V. R., and F. A. Wade, Geology of the western Queen Maud Mountains, scale 1:1,000,000, *Antarct. Map Folio Ser.*, folio 12, plate XVI, edited by V. C. Bushnell, Am. Geogr. Soc., New York, 1969.
Minshew, V. H., Geology of the Scott Glacier and Wisconsin Range areas, central Transantarctic Mountains, Antarctica, Ph.D. thesis, 268 pp., Ohio State Univ., Columbus, 1967.
Mirsky, A., Geology of the Ohio Range-Liv Glacier area, scale 1:1,000,000, *Antarct. Map Folio Ser.*, folio 12, plate XVI, edited by V. C. Bushnell, Am. Geogr. Soc., New York, 1969.
O'Neil, J. R., Appendix: Terminology and standards, in *Rev. Mineral. 16*, 561–570, 1986.
Pankhurst, R. J., B. C. Storey, I. L. Millar, D. I. M. Macdonald, and W. R. Vennum, Cambrian-Ordovician magmatism in the Thiel Mountains, Transantarctic Mountains, and implications for the Beardmore orogeny, *Geology*, *16*(3), 246–249, 1988.
Papanastassiou, D. A., D. J. DePaolo, and G. J. Wasserburg, Rb-Sr and Sm-Nd chronology and genealogy of mare basalts from the Sea of Tranquility, *Proc. Lunar and Planet. Sci. Conf. 8th*, 1639–1672, 1977.
Pearce, J. A., N. B. W. Harris, and A. G. Tindle, Trace element distribution diagrams for the tectonic interpretation of granitic rocks, *J. Petrol.*, *25*, 956–983, 1984.
Pitcher, W. S., Granite type and tectonic environment, in *Mountains Building Processes*, edited by K. Hsü, pp. 19–40, Academic, San Diego, Calif., 1982.
Pitcher, W. S., Granites and yet more granites forty years on, *Geol. Rundshau*, *76*(1), 51–79, 1987.
Samson, S. D., W. C. McClelland, P. J. Patchett, G. E. Gehrels, and R. G. Anderson, Evidence from neodymium isotopes for mantle contributions to Phanerozoic crustal genesis in the Canadian Cordillera, *Nature*, *337*, 705–709, 1989.
Scotese, C. R., R. K. Bambach, C. Barton, R. Van der Voo, and A. M. Ziegler, Paleozoic base maps, *J. Geol.*, *87*, 217–277, 1979.
Scotese, C. R., L. M. Gahagan, and R. L. Larson, Plate tectonic reconstructions of the Cretaceous and Cenozoic basins, *Tectonophysics*, *155*(1–4), 27–48, 1988.
Skinner, D. N. B., The granites and two orogenies of southern Victoria Land, in *Antarctic Earth Science*, edited by R. L. Oliver, P. R. James, and J. B. Jago, pp. 160–163, Australian Academy of Science, Canberra, 1983.
Smit, J. H., and E. Stump, Sedimentology of the La Gorce Formation, La Gorce Mountains, Antarctica, *J. Sediment. Petrol.*, *56*(5), 663–668, 1986.
Smith, B. M., and B. J. Giletti, Oxygen isotope evidence for high-temperature cooling of porphyritic felsites from the Isle of Skye, N. W. Scotland, in press, 1989.
Soloman, G. C., and H. P. Taylor, The geographic distribution of $\delta^{18}O$ values in Mesozoic and early Cenozoic granitic rocks of the southwestern North American cordillera (abstract), *Geol. Soc. Am. Abstr. Programs*, *13*, 558, 1981.
Stump E., On the Late Precambrian-Early Paleozoic metavolcanic and metasedimentary rocks of the Queen Maud Mountains, Antarctica, and a comparison with rocks of similar age from southern Africa, Ph.D. thesis, 212 pp., Ohio State Univ., Columbus, 1976.
Stump, E., Observations on the Ross Orogen, Antarctica, in *Gondwana Five*, edited by M. M. Cresswell and P. Vella, pp. 205–208, Balkena, Rotterdam, 1981.
Stump, E., The Ross Supergroup in the Queen Maud Mountains, in *Antarctic Geoscience*, edited by C. Craddock, pp. 565–570, University of Wisconsin Press, Madison, 1982.
Stump, E., R. J. Korsch, and D. G. Edgerton, The myth of the Nimrod and Beardmore orogenies (abstract), paper presented at 5th International Symposium Antarctic Earth Science, Sci. Comm. on Antarct. Res., Cambridge, United Kingdom, 1987.
Stump, E., J. M. G. Miller, R. J. Korsch, and D. G. Edgerton, Diamictite from Nimrod Glacier area, Antarctica: possible Proterozoic glaciation on the seventh continent, *Geology*, *16*(3), 225–228, 1988.
Stump, E., R. J. Korsch, and D. G. Edgerton, The myth of the Nimrod and Beardmore orogenies, in *Geological Evolution of Antarctica*, edited by M. R. A. Thomson, J. A. Crame, and J. W. Thomson, Cambridge University Press, Cambridge, UK, in press, 1990.
Taylor, H. P., Jr., Igneous Rocks: II. Isotopic case studies of circumpacific magmatism, *Rev. Mineral.*, *16*, 273–317, 1986.
Taylor, H. P., Jr., and L. T. Silver, Oxygen isotope relationships in plutonic igneous rocks of the Peninsular Ranges batholith, southern and Baja California, *U. S. Geol. Surv. Open File Rep. 78–701*, 423–426, 1978.
Thorpe, R. S., and P. W. Francis, Petrogenetic relationships of volcanic and intrusive rocks of the Andes, in *Origin of Granitic Batholiths: Geochemical Evidence*, edited by M. P. Atherton and J. Tarney, pp. 65–75, Shiva, Kent, United Kingdom, 1979.
Wade, F. A., and C. A. Cathey, Geology of the basement complex, western Queen Maud Mountains, Antarctica, in *Geology of the Central Transantarctic Mountains*, *Antarct. Res. Ser.*, vol. 36, pp. 429–453, edited by M. D. Turner and J. Splettstoesser, AGU, Washington, D. C., 1986.
Wade, F. A., V. L. Yeats, J. R. Everett, D. W. Greenlee, K. E. LePrade, and J. C. Shenke, Geology of the central portion of the Queen Maud Range, Transantarctic Mountains, *Science*, *150*, 1808–1809, 1965*a*.
Wade, F. A., V. L. Yeats, J. R. Everett, D. W. Greenlee, K. E. LePrade, and J. C. Shenke, The Geology of the central Queen Maud Range, Transantarctic Mountains, Antarctica, *Res. Rep. Ser.*, *Antarct. Ser. 65-1*, 54 pp., Texas Tech. College, Lubbock, 1965*b*.
Walawender, M. J., R. G. Gastil, J. P. Clinkenbeard, W. V. McCormick, R. G. Eastman, R. S. Wernicke, M. S. Wardlow, and B. M. Smith, Origin and evolution of the zoned LaPosta-type plutons, eastern Peninsular Ranges Batholith, southern and Baja California, Nature and Origin of Cordilleran Magmatism, edited by J. L. Anderson, *Mem. Geol. Soc. Am.*, 1990.
Yochelson, E. L., and E. Stump, Discovery of Early Cambrian fossils at Taylor Nunatak, Antarctica, *J. Paleontol.*, *51*, 872–875, 1977.
York, D., Least squares fitting of a straight line with correlated errors, *Earth Planet. Sci. Lett.*, *5*, 320–324, 1969.
Ziegler, A. M., C. R. Scotese, and S. F. Barrett, Mesozoic and Cenozoic paleogeographic maps: in *Tidal Friction and the Earth's Rotation II*, edited by P. Brosche and J. SünderMann, pp. 240–252, Springer-Verlag, New York, 1983.
Ziegler, A. M., C. R. Scotese, W. S. McKerrow, M. E. Johnson, and R. K. Bambach, Paleozoic paleogeography, *Ann. Rev. Earth Planet. Sci.*, *7*, 473–502, 1979.

S. Borg, D. DePaolo, and B. Smith, Department of Geology and Geophysics, Berkeley Center for Isotope Geochemistry, University of California at Berkeley, Berkeley, CA 94720.

(Received May 1, 1989;
revised October 16, 1989;
accepted October 16, 1989.)