Tectonic Model for Development of the Byrd Glacier Discontinuity and Surrounding Regions of the Transantarctic Mountains during the Neoproterozoic – Early Paleozoic

Edmund Stump[1] · Brian Gootee[1] · Franco Talarico[2]

[1] Department of Geological Sciences, Arizona State University, Tempe, Arizona AZ 85287-1404, USA

[2] Dipartimento di Scienze della Terra, Università di Siena, Via del Laterino 8, 53100 Siena, Italy

Abstract. The Byrd Glacier discontinuity is a major tectonic boundary crossing the Ross Orogen, with crystalline rocks to the north and primarily sedimentary rocks to the south. Most models for the tectonic development of the Ross Orogen in the central Transantarctic Mountains consist of two-dimensional transects across the belt, but do not address the major longitudinal contrast at Byrd Glacier. This paper presents a tectonic model centering on the Byrd Glacier discontinuity. Rifting in the Neoproterozoic produced a crustal promontory in the craton margin to the north of Byrd Glacier. Oblique convergence of a terrane (Beardmore microcontinent) during the latest Neoproterozoic and Early Cambrian was accompanied by subduction along the craton margin of East Antarctica. New data presented herein in support of this hypothesis are U-Pb dates of 545.7 ±6.8 Ma and 531.0 ±7.5 Ma on plutonic rocks from the Britannia Range, directly north of Byrd Glacier. After docking of the terrane, subduction stepped out, and Byrd Group was deposited during the Atdabanian-Botomian across the inner margin of the terrane. Beginning in the upper Botomian, reactivation of the sutured boundaries of the terrane resulted in an outpouring of clastic sediment and folding and faulting of the Byrd Group.

Introduction

By the time of publication of the Antarctic Map Folio Series (Anonymous 1969–1970), most of the Transantarctic Mountains (TAM) had been geologically mapped, and a general model had emerged for the formation of the late Proterozoic – early Paleozoic Ross orogenic belt. Following roughly along the trend of the present day TAM (Fig. 4.3-1), a suite of sediments, and in places volcanics (Ross Supergroup), accumulated in a continental margin setting from late Proterozoic to Middle Cambrian time, followed by deformation and metamorphism during the Ross Orogeny, and intrusion of magmas of batholithic proportions, which are found throughout most segments of the TAM. This understanding led eventually to the general consensus that subduction was active outboard of the present day TAM in a Cambro-Ordovician timeframe.

The central Transantarctic Mountains (CTM) figured significantly in interpretations of the evolution of the Ross Orogen. The craton along which the sediments of the Ross Orogen were deposited is exposed in the Miller and Geologists Ranges as outcroppings of amphibolite-grade metamorphics (Nimrod Group) (Grindley 1972; Grindley et al. 1964). The initial passive margin deposits (Beardmore Group) were thought to have accumulated during the Neoproterozoic, throughout a broad region from north

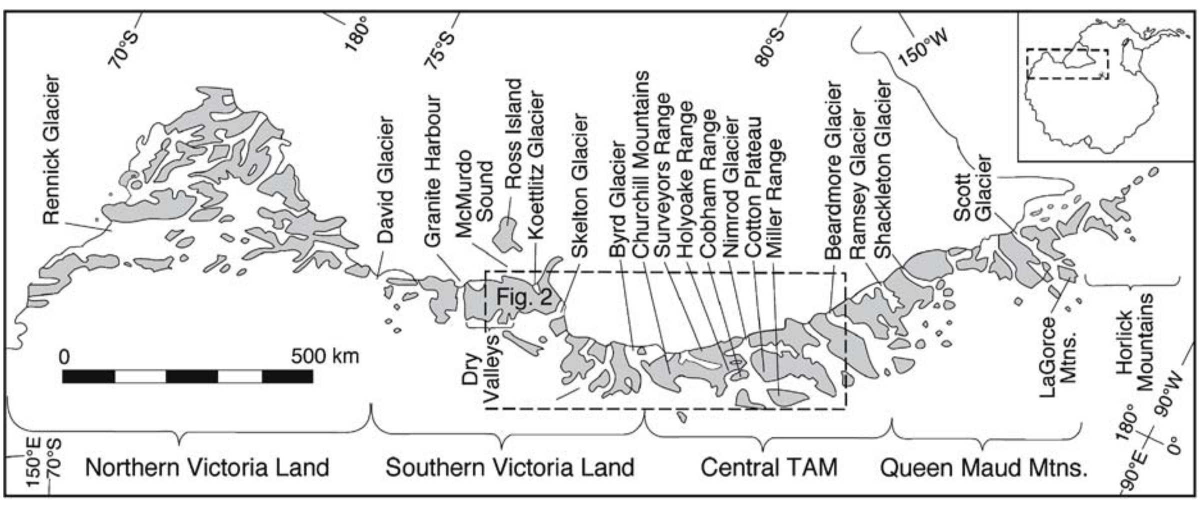

Fig. 4.3-1. Location map of Transantarctic Mountains

From: Fütterer DK, Damaske D, Kleinschmidt G, Miller H, Tessensohn F (eds) (2006) Antarctica: Contributions to global earth sciences. Springer-Verlag, Berlin Heidelberg New York, pp 181–190

of Nimrod Glacier to south of Beardmore Glacier (Gunn and Walcott 1962). Beardmore Group includes Cobham Formation, characterized by greenschist-grade schists, calc-schists, and marbles, conformably overlain by Goldie Formation, a sequence of metamorphosed metagraywacke and argillite (Laird et al. 1971). Folding of the Beardmore Group, recognized at the time by discordant contacts with overlying Lower Cambrian Shackleton Limestone, was named the Beardmore Orogeny (Grindley and McDougall 1969). The Cambrian Byrd Group, composed of Shackleton Limestone and overlying clastic units of the Starshot Formation, Dick Formation, and Douglas Conglomerate (Laird 1963; Skinner 1964), was deformed during the Ross Orogeny sometime during the Middle to Late Cambrian, and then intruded by the Cambro-Ordovician Granite Harbour Intrusives (Gunn and Warren 1962).

The ensuing decades have seen significant additions to and reinterpretations of the bedrock history of the CTM. Borg et al. (1990) presented an important model for the tectonic development of the region. Based on a study of Nd, Sr, and O isotopes in Granite Harbour Intrusives of the CTM, they recognized a major, lower-crustal province boundary beneath Marsh Glacier, with granitic rocks intruding Nimrod Group to the west having model ages (T_{DM}) of ~2.0 Ga and granitic rocks to the east having model ages of ~1.7 Ga. Borg et al. (1990) developed a tectonic model in which a terrane, which they called the "Beardmore microcontinent", converged obliquely on the craton, with Goldie Formation accumulating in the basin between. Suturing of the microcontinent in Neoproterozoic time was accompanied by deformation and erosion of the Goldie Formation, and subsequent deposition of Shackleton Limestone. Further subduction outboard of the Beardmore microcontinent led to volcanism, deformation, and intrusion associated with the Ross Orogeny.

The SWEAT hypothesis, that Laurentia had separated from Antarctica and Australia, provided the rifted cratonic margin along which the Ross Supergroup would be deposited (Moores 1991). It also provided Borg et al. (1990) an opposing margin from which their Beardmore microcontinent was produced.

"Nimrod Orogeny" was the name given to the episode of deformation and metamorphism that affected the Nimrod Group (Grindley and Laird 1969). Dating by Rb-Sr whole-rock analysis estimated this to have occurred during the Paleoproterozoic (Gunner and Faure 1972). Grindley (1972) had mapped a major structure (Endurance thrust) displacing units of the Nimrod Group. Goodge et al. (1991) found that the thrust is a distributed shear zone formed under amphibolite-facies conditions, which dips SW with top-to-the-SSE sense of shear. Dating a succession of intrusions both affected by and postdating the shearing showed that deformation began before 541 Ma and had waned by 520 Ma (Goodge et al. 1993b). This put in doubt the age of the Nimrod Group, until a more recent study using U-Pb, single-grain zircon analysis showed that deep-crustal metamorphism and magmatism (Nimrod orogeny) occurred 1 730–1 720 Ma, following initial crustal magmagenesis at ca. 3 100–3 000 Ma (Goodge and Fanning 1999; Goodge et al. 2001). The data from the Endurance shear zone were instrumental in development of a model of oblique subduction along the Pacific margin of Antarctic during the Early Cambrian (Goodge et al. 1993a).

The extent of Beardmore Group outcrop, as originally mapped, was considerably reduced when it was found that the age of detrital zircons in a number of samples was as young as Middle(?) Cambrian (Goodge et al. 2002). These sedimentary rocks are now assigned to the Starshot Formation (Myrow et al. 2002a). However, detrital zircons in the suite from Cobham and Goldie formations at Cobham Range and Cotton Plateau are no younger than Grenville age (Goodge et al. 2002). Gabbros and pillow basalts interbedded with Goldie Formation at Cotton Plateau, possibly associated with rifting, have been U-Pb zircon dated at 668 ±1 Ma (Goodge et al. 2002). Thus, the original recognition of Neoproterozoic sedimentation remains, albeit in a restricted area adjacent to the craton margin.

An important detail in the deformational history of Byrd Group was added when Rowell et al. (1988) discovered that Douglas Conglomerate overlies folded Shackleton Limestone unconformably in the northern Holyoake Range. Then Myrow et al. (2002a) showed that the carbonate to clastic transition within the Byrd Group is conformable in the southern Holyoake Range. More recently, we have mapped similar conformable and unconformable relationships in the area to the south of Byrd Glacier (Stump et al. 2004).

Archaeocyathid and trilobite dating of the Shackleton Limestone has shown it to have been deposited during the Atdabanian, Botomian, and possibly Toyonian stages of the Early Cambrian (Hill 1964; Debrenne and Kruse 1986; Palmer and Rowell 1995). Recently, Myrow et al. (2002a) have also found Botomian trilobites in the lower portion of the Starshot Formation where it conformably overlies Shackleton Limestone, constraining the age of the onset of clastic deposition in the CTM.

Since the earliest days of interpreting the evolution of the Ross Orogen, resolving the timing of isotopically dated volcanic and plutonic events with fossil-dated sedimentary episodes has been a fundamental problem. In recent years, with refinements of the Cambrian interval of the Geological Timescale, and with precise U-Pb dates, a clearer picture has emerged (Tucker and McKerrow 1995; Landing et al. 1998; Davidek et al. 1998; Bowring and Erwin 1998). By the timescale of Bowring and Erwin

(1998) the Atdabanian began about 525 Ma and the Botomian/Toyonian ended by about 510 Ma, with the Middle-Upper Cambrian boundary at 500 Ma and the Cambrian-Ordovician boundary at 490 Ma.

The geological patterns that can be followed for several hundred kilometers along the CTM, abruptly end at Byrd Glacier, where folded and thrust-faulted limestones and conglomerates to the south face amphibolite-grade metamorphics and plutonics to the north. In fact, the geology throughout southern Victoria Land (SVL) is distinctly different in detail from the geology of the CTM. In the region from the Dry Valleys to Skelton Glacier, Grindley and Warren (1964) subdivided the metamorphic units into greenschist-grade Skelton Group and amphibolite-grade Koettlitz Group. Recently Cook and Craw (2001, 2002) have lumped all the metamorphic rocks as Skelton Group following the original designation of Gunn and Warren (1962), owing to structural complexity, which has precluded a comprehensive stratigraphic interpretation.

The fundamental, stratigraphic characteristics of the Byrd Group are initial deposition of a thick sequence of nearly pure carbonate sediments followed abruptly by a thick sequence of clastic sediments. This pattern is not seen in any of the occurrences of Skelton Group. In the Skelton Glacier area, the "Marble unit" of Cook and Craw (2002) (Anthill Limestone of Gunn and Warren (1962) and Skinner (1982)) contains centimeter to meter wide layers of quartzite and argillite. Near the base of the marble to the north of Cocks Glacier, polymict, matrix-supported metaconglomerate occurs. The marble overlies a thick succession of clastics, mainly quartzite (Rowell et al. 1993). Above the Marble unit is the "Cocks unit" of Cook and Craw (2002) (Cocks Formation of Skinner (1982)), a schistose metasedimentary unit containing thin pillow basalts, and with conglomeratic lenses near its base. The units of the Skelton Group in the Skelton Glacier area were deformed prior to intrusion of two plutons, each dated at 551 Ma (Rowell et al. 1993; Encarnación and Grunow 1996). This relationship itself precludes correlation of Skelton Group (in the Skelton Glacier area) with Byrd Group.

Although marbles occur at places to the north of Skelton Glacier, again they do not show the thickness and purity of the Shackleton Limestone. For example, the Salmon Marble of Gunn and Warren (1962) and Findlay et al. (1984) in the area between Koettlitz and Blue Glaciers contains thin layers of quartzite, biotite schist, and calc-schist. The earliest plutonism into Skelton Group at the head of Koettlitz Glacier was at 540–530 Ma (Hall et al. 1995; Mellish et al. 2002; Read et al. 2002), again earlier than deposition of Shackleton Limestone.

In the Britannia Range, directly north of Byrd Glacier, poorly studied metasedimentary rocks are assigned to the Horney Formation (Borg et al. 1989). Based on the report of Borg et al. (1989) and our own reconnaissance observations, there is no marble within the Horney Formation, which is primarily amphibolite-grade micaceous schist and gneiss, with minor calc-silicate layers and pods. Again, these rocks are not a match for the Byrd Group. Until now the oldest reported plutonism in this area is from the Brown Hills, to the north of the Britannia Range, where the Carlyon Granodiorite has been dated at 515 ±8 Ma (Encarnación and Grunow 1996).

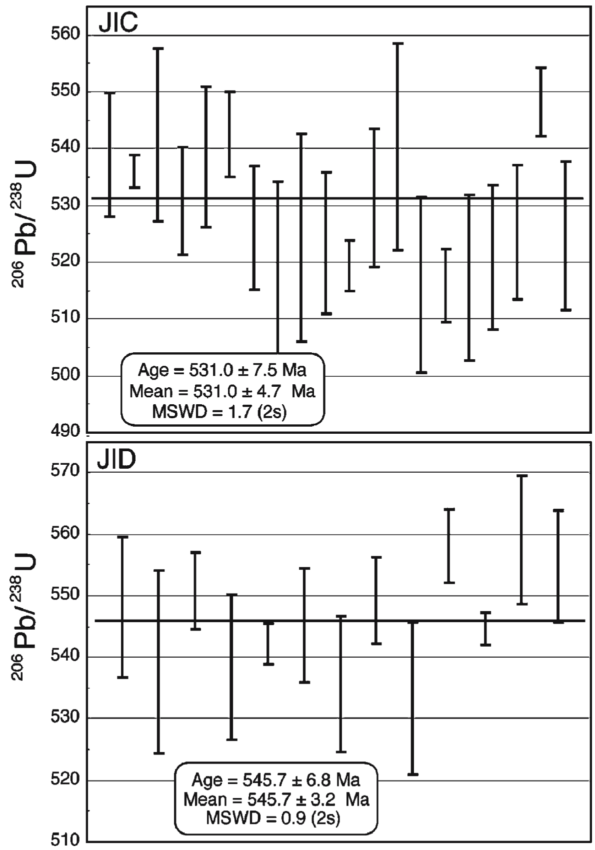

Fig. 4.3-2. Plots of $^{206}Pb/^{238}U$ ages from samples *JIC* and *JID* collected on north side of Byrd Glacier adjacent to Ramseier Glacier

Herein we report two new U-Pb dates on plutonic rocks from the western end of the Britannia Range, collected on the buttress to the south of Ramseier Glacier (Fig. 4.3-2, Table 4.3-1). Analyses were made on single grains of zircon using a laser-ablation ICPMS. The older of these intrusions has an age of 545.7 ±6.8 Ma. It is a granite with quartz, plagioclase, microcline, and minor biotite, showing signs of strain and alteration. The quartz is in part polygonized and has undulose extinction. Traces of epidote are associated with the plagioclase, and some of the biotite is altered to chlorite. The younger intrusion, with an age of 531.0 ±7.5 Ma is a diorite with approximately equal proportions of pristine hornblende, biotite, and plagioclase.

Table 4.3-1. U-Pb (zircon) geochronologic analyses by Laser-Ablation Multicollector ICP Mass Spectrometery

Sample-grain	U (ppm)	$^{206}Pb/^{204}Pb$	U/Th	Isotopic ratios							Apparent ages (Ma)					
				$^{207}Pb^*/^{235}U$	± (%)	$^{206}Pb^*/^{238}U$	± (%)	Error corr.	$^{206}Pb/^{207}Pb$	± (%)	$^{206}Pb^*/^{238}U$	± (Ma)	$^{207}Pb^*/^{235}U$	± (Ma)	$^{206}Pb^*/^{207}Pb^*$	± (Ma)
JIC-1	184	3154	2.3	0.67521	4.3	0.08715	2.1	0.49	17.797	3.8	538.7	10.9	524	18	460	84
JIC-2	198	1904	4.0	0.73406	4.5	0.08665	0.6	0.13	16.276	4.5	535.7	2.9	559	19	655	96
JIC-3	210	4513	2.7	0.69199	8.3	0.08773	2.9	0.35	17.480	7.8	542.1	15.2	534	35	500	172
JIC-4	195	2535	2.6	0.69402	5.6	0.08578	1.9	0.34	17.041	5.2	530.5	9.5	535	23	555	114
JIC-5	737	10102	1.3	0.69329	2.9	0.08710	2.4	0.83	17.321	1.6	538.3	12.4	535	12	520	36
JIC-6	320	3799	2.7	0.70028	5.2	0.08775	1.4	0.28	17.278	5.0	542.2	7.5	539	22	525	109
JIC-7	272	2560	2.5	0.65669	6.1	0.08499	2.2	0.36	17.844	5.7	525.8	10.9	513	24	454	125
JIC-8	324	6319	2.1	0.69441	5.9	0.08368	3.2	0.54	16.614	5.0	518.0	16.0	535	25	610	107
JIC-9	1213	5517	1.2	0.66185	4.0	0.08470	3.6	0.92	17.646	1.5	524.1	18.3	516	16	479	34
JIC-10	328	5813	2.1	0.69790	3.6	0.08455	2.5	0.68	16.704	2.6	523.2	12.4	538	15	599	57
JIC-11	222	2451	3.6	0.66531	5.6	0.08387	0.9	0.16	17.381	5.5	519.2	4.4	518	23	512	121
JIC-12	313	7709	1.6	0.67800	3.2	0.08587	2.4	0.74	17.463	2.2	531.1	12.1	526	13	502	47
JIC-13	305	6514	1.6	0.72004	6.7	0.08737	3.5	0.52	16.731	5.7	540.0	18.2	551	29	595	124
JIC-14	267	5530	1.9	0.70525	4.7	0.08329	3.1	0.66	16.283	3.6	515.7	15.4	542	20	654	76
JIC-15	629	7315	1.7	0.66092	2.4	0.08328	1.3	0.55	17.374	2.0	515.7	6.5	515	10	513	43
JIC-16	316	3835	2.5	0.69336	4.2	0.08353	2.9	0.70	16.610	3.0	517.1	14.6	535	18	611	65
JIC-17	354	4226	1.9	0.64985	4.2	0.08412	2.5	0.60	17.847	3.4	520.7	12.7	508	17	454	75
JIC-18	315	4428	3.0	0.68450	8.8	0.08487	2.3	0.27	17.096	8.5	525.1	11.8	530	36	548	186
JIC-19	791	4158	3.3	0.68962	1.9	0.08872	1.2	0.63	17.738	1.5	548.0	6.1	533	8	467	32
JIC-20	630	6492	1.8	0.67676	3.0	0.08475	2.6	0.86	17.267	1.6	524.4	13.1	525	12	527	34
JID-1	1520	2853	6.5	0.71037	4.1	0.08871	2.2	0.52	17.218	3.5	547.9	11.4	545	18	533	77
JID-2	2545	5380	11.1	0.68846	3.5	0.08722	2.9	0.83	17.468	1.9	539.1	14.8	532	14	501	43
JID-3	2488	2613	6.7	0.69960	4.0	0.08915	1.2	0.29	17.570	3.8	550.5	6.2	539	17	488	84
JID-4	6229	3304	5.8	0.67808	3.0	0.08706	2.3	0.77	17.703	1.9	538.1	11.8	526	12	472	42
JID-5	8788	4530	16.0	0.68849	1.2	0.08772	0.6	0.54	17.567	1.0	542.0	3.3	532	5	489	22
JID-6	5884	6617	17.0	0.68582	1.8	0.08822	1.8	0.99	17.735	0.2	545.0	9.2	530	7	468	5
JID-7	4641	9724	19.3	0.67840	2.5	0.08660	2.2	0.85	17.601	1.4	535.4	11.0	526	10	484	30
JID-8	1875	7820	6.7	0.72404	1.7	0.08890	1.3	0.80	16.929	1.0	549.0	7.0	553	7	570	22
JID-9	2696	3563	8.6	0.67881	2.7	0.08622	2.4	0.92	17.513	1.1	533.1	12.4	526	11	495	23
JID-10	1882	1148	8.0	0.73474	11.3	0.09039	1.1	0.10	16.963	11.2	557.9	5.9	559	49	565	245
JID-11	3217	10907	10.3	0.69404	2.3	0.08813	0.5	0.22	17.507	2.3	544.4	2.7	535	10	496	50
JID-12	4242	11191	12.9	0.70864	2.2	0.09056	1.9	0.87	17.619	1.1	558.8	10.4	544	9	482	24
JID-13	4116	7793	18.9	0.70931	2.0	0.08982	1.7	0.84	17.460	1.1	554.5	9.1	544	9	502	24

All errors are reported at the 1–σ level and incorporate only uncertainties from measurement of isotopic ratios. U concentration and U/Th have uncertainty of ~25%. Decay constants: $^{235}U = 9.8485 \times 10^{-10}$, $^{238}U = 1.55125 \times 10^{-10}$, $^{238}U/^{235}U = 137.88$. Isotope ratios are corrected for Pb/U fractionation by comparison with standard zircon with an age of 564 ±4 Ma. Initial Pb composition interpreted from Stacey and Kramers (1975), with uncertainties of 1.0 for $^{206}Pb/^{204}Pb$ and 0.3 for $^{207}Pb/^{204}Pb$.

Tectonic Model

Models that have evolved for the tectonic development of the CTM generally have taken a view in cross section transverse to the orogen (e.g., Borg et al. 1990; Goodge 1997). Such a view is successful for interpretation of the broad region of the CTM; however, it does not take into account the abrupt discontinuity at Byrd Glacier and the geological differences in southern Victoria Land. Our current research is aimed at understanding the tectonic discontinuity followed by Byrd Glacier. Research to the south of Byrd Glacier was begun during the 2000/2001 field season, with results presented elsewhere (Gootee 2002; Stump et al. 2002b; Stump et al. 2004). The following model may serve as a working hypothesis at this stage of our research. It draws on prior research and models of the CTM, in particular Borg et al. (1990) and Goodge et al. (1993a), but goes beyond these in its focus on the Byrd Glacier discontinuity.

Although components of Borg et al.'s (1990) model of the Beardmore microcontinent have been shown to be in error, we feel that the general model of a terrane colliding with the Antarctic continental margin is viable, and so we retain their nomenclature. Their model was based on three sets of isotopic data: (1) the difference in Sm-Nd model ages of Granite Harbour Intrusives across Marsh Glacier (T_{DM} ~ 2.0 Ga in the Miller Range to the west and T_{DM} ~ 1.7 Ga to the east), (2) the presence of basalt and gabbro with oceanic affinity associated with Goldie Formation at Cotton Plateau, adjacent to Marsh Glacier, and (3) T_{DM} of ~1.7 Ga in "Goldie Formation" from outcrops to the east of Cotton Plateau. The first and second of these data sets must be included in any model of the region, the third set not. The T_{DM} data from "Goldie Formation" were emphasized by Borg et al. (1990) to suggest the allochthonous nature of the terrane, with no apparent source on the Antarctic craton to provide detritus of that model age. However, further isotopic investigations on the Granite Harbour Intrusives showed that a crustal province with Sm-Nd model ages in the range 1.6–1.9 Ma underlies most of the Transantarctic Mountains from Shackleton Glacier to northern Victoria Land (Borg and DePaolo 1991; Borg and DePaolo 1994). Furthermore, it has been shown that the "Goldie Formation" rocks on which the analyses were performed belong to the Starshot Formation, and were deposited from local sources along with the associated Douglas Conglomerate in the Early(–Middle) Cambrian (Goodge et al. 2002; Myrow et al. 2002a,b).

Our model encompasses the CTM and part of SVL from Beardmore Glacier to the Dry Valleys. The first panel of Fig. 4.3-3 shows the general geology of this portion of the TAM. The boundaries of the major units on this map are used as geographical reference in the various stages of the model.

Pre-550 Ma

We begin in the Neoproterozoic prior to 550 Ma. The craton had formed in the Archean and Paleoproterozoic. Rifting occurred in the Neoproterozoic, perhaps around 670 Ma (Goodge et al. 2002), with the conjugate continent drifting away. In the area of Byrd Glacier a major offset in the cratonic margin occurred during the rifting. This was first suggested by Stump (1992). Goodge and Dallmeyer (1996) added that this step-like geometry was possibly due to transform faulting in the vicinity of Byrd Glacier, but we postulate that separation occurred along both Byrd Glacier and the craton margin to the west, analogous to the opening of the Red Sea/Gulf of Aden. The Beardmore microcontinent separated during the rifting and lay somewhere to the southeast in present coordinates. It is possible, but not required, that the Beardmore microcontinent was the block that rifted from the reentrant at Byrd Glacier. The rocks immediately to the north of Byrd Glacier and throughout most of Victoria Land are of the same basement-age province as the Beardmore microcontinent (Borg and DePaolo 1994).

Deposition of Beardmore Group occurred in the region between the Beardmore microcontinent and mainland Antarctica. Geochemical and isotopic composition of the gabbro associated with Goldie Formation at Cotton Plateau indicates an oceanic rather than continental setting of magmatism (Borg et al. 1990). The question of how far removed from its present day location this terrane sat is speculative, but the ocean between the terrane and the continent may have been no wider than the Red Sea today. We position the terrane in the second panel of Fig. 4.3-3 in a fairly proximal location.

In SVL clastics, carbonates, and volcanics (Skelton Group) were being deposited during the Neoproterozoic (Skinner 1982; Cook and Craw 2002).

~550 Ma, Ross Orogeny Commences

At some time in the latter part of the Neoproterozoic, subduction commenced along this margin of Antarctica. Indications of convergence are folding of Skelton Group crosscut by plutons dated at 551 ±4 Ma (Rowell et al. 1993; Encarnación and Grunow 1996). An even earlier age for the onset of subduction is suggested by detrital zircons of magmatic origin in Starshot Formation dated at 580–560 Ma (Goodge et al. 2002). The Beardmore microcontinent converged on the continent from the SSE. We postulate that this convergence was oblique in the sector of the CTM (following the direction indicated in the Endurance shear zone (Goodge et al. 1993)), but more orthogonal in the area north of Byrd Glacier due to the geometry established during rifting. If the term Ross orogeny is to be applied to the full cycle of deformation and plutonism affecting the Ross Supergroup, then it began at approximately this time.

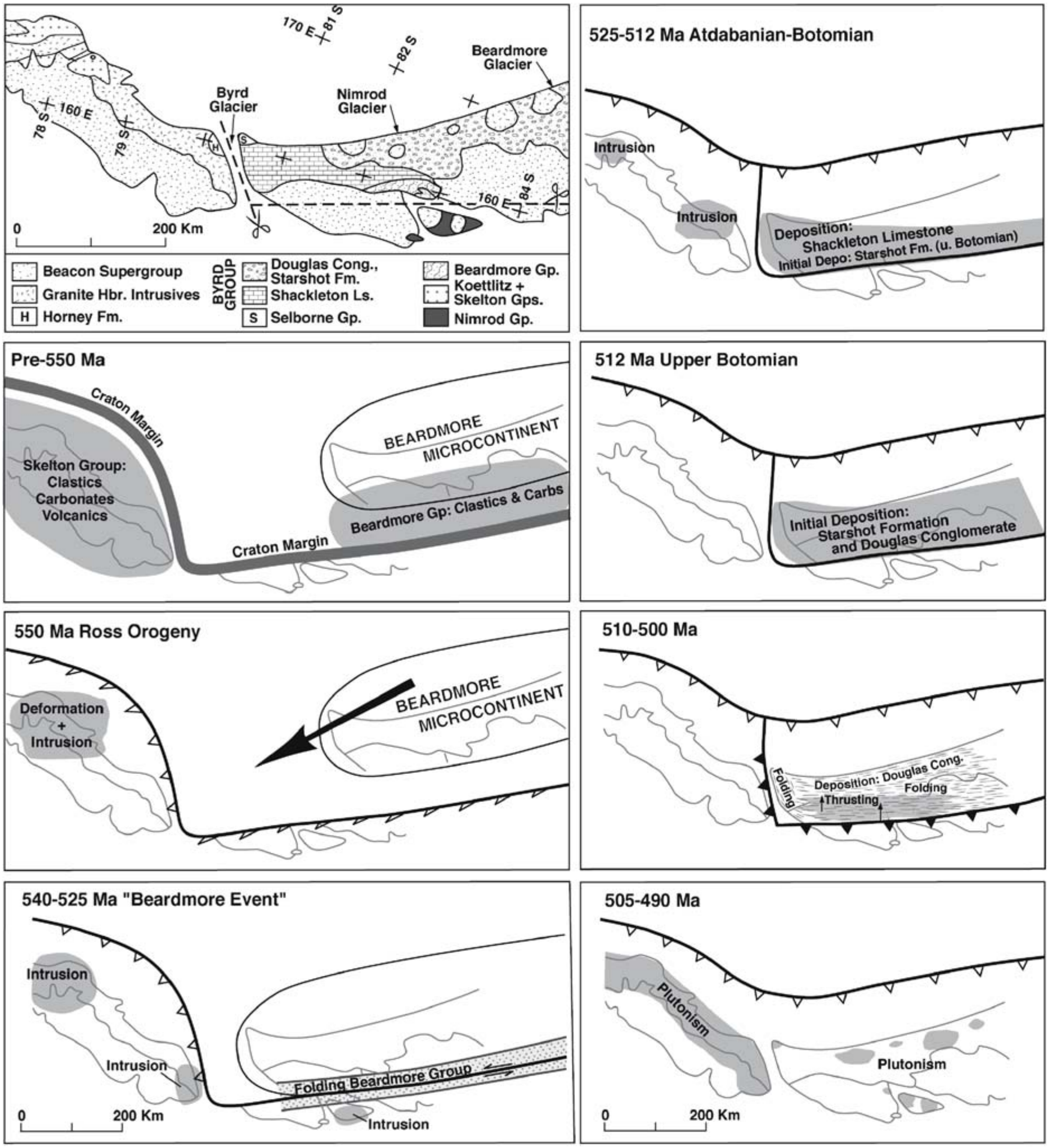

Fig. 4.3-3. Base geological map and tectonic model, southern Victoria Land and central Transantarctic Mountains. First panel is geological basemap. *Lines* between geological units are carried through the model in subsequent panels. The *dashed lines* in first panel represent sutured boundaries of the Beardmore Microcontinent, and the cuts that were made in developing the model. See text for discussion of individual panels of the model

540–525, "Beardmore Event"

By the beginning of the Cambrian the Beardmore microcontinent had begun to collide with the mainland, as indicated by plutonics in the Endurance shear zone dated at 540 Ma (Goodge et al. 1993b). How much earlier this shearing had commenced is unconstrained, but it could have occurred as early as 550 Ma, while plutons were intruding in SVL. Plutonism in the Britannia Range at 545–530 Ma indicates that subduction was continuing in the vicinity of Byrd Glacier during this interval. Beardmore Group was folded in a narrow zone adjacent to the craton, which includes Goldie Formation exposed at Cotton Plateau and

La Gorce Formation in the La Gorce Mountains, 600 km to the south (Stump et al. 2002a). Constraints on folding of La Gorce Formation are detrital zircons as young as 550 Ma (supplied from the magmatism in SVL?) (Vogel et al. 2001), and intrusion of porphyry (Wyatt Formation) dated at 525 ±3 Ma, which cross cuts the folds (Stump et al. 1986; Encarnación and Grunow 1996). A second episode of deformation in the La Gorce Mountains caused high-angle reverse faulting and widespread development of associated cleavage (Stump et al. 1986). At Cotton Plateau, folded Goldie Formation is overlain by basal Shackleton Limestone. Whether this contact is an unconformity or a fault remains controversial (Laird et al. 1971; Myrow et al. 2002a,b). Stump et al. (2002a) argued for two episodes of deformation at Cotton Plateau, the first a folding event recorded only in Goldie Formation, and the second causing folding and major offset of Goldie Formation and Shackleton Limestone along a high-angle shear zone. This is the scenario followed in our model. But even if the unconformity at Cotton Plateau is discounted, folding in the La Gorce Mountains occurred in the same timeframe as movement on the Endurance shear zone.

Identification of the Beardmore orogeny was based primarily on the deformation of Beardmore Group prior to deposition of Shackleton Limestone, but was generalized to encompass any orogenic activity in the Neoproterozoic (Grindley and McDougall 1969). Constraints from detrital zircons in La Gorce Formation now suggest that this folding episode was an Early Cambrian event (perhaps beginning within the last few million years of the Proterozoic) completed by 525 Ma, the beginning of the Atdabanian, which is the age of cross-cutting Wyatt Formation in the La Gorce Mountains and the oldest fossils in the Shackleton Limestone (Palmer and Rowell 1995).

In SVL a series of minor plutons span most of the time period 550–505 Ma when the main episode of Granite Harbour Intrusives began, suggesting subduction to the north of the Byrd Glacier discontinuity throughout that time (Rowell et al. 1993; Cooper et al. 1997; Cox et al. 2000; Allibone and Wysoczanski 2002; Read et al. 2002).

525–512 Ma, Atdabanian-Upper Botomian

We postulate that following docking of the Beardmore microcontinent, the subduction zone stepped out to the outboard side of the terrane. During the Atdabanian-Botomian there is ample evidence that subduction was occurring both south and north of the CTM. In the Queen Maud Mountains, on the outboard side of the docked terrane, volcanism of the Liv Group spanned the time 525–505 Ma (Encarnación and Grunow 1996; Van Schmus et al. 1997; Encarnación et al. 1999; Wareham et al. 2001). In the Darwin Glacier area, the Cooper granodiorite was intruded at 515 ±8 Ma (Simpson and Cooper 2002), and in the Dry Valleys area DV1b suite plutons were intruded at 531 ±10 Ma (Dun Pluton) and 516 ±10 Ma (Calkin Pluton) (Allibone and Wysoczanski 2002). It seems likely that the subduction indicated in the Queen Maud Mountains would have connected along the outboard side of the terrane to SVL.

By the beginning of the Atdabanian (525 Ma), erosion had occurred across the accreted, inboard side of the Beardmore microcontinent and deposition of an east-facing carbonate bank (Shackleton Limestone) had commenced (Rees et al. 1989; Gootee and Stump 2006 this vol.). More than 2 000 meters of carbonate deposition continued through most of the Botomian under quiescent conditions (Stump et al. 2004). That this carbonate basin existed without tectonic effects through this period may be owing to its location at the postulated reentrant to the previously rifted margin.

512 Ma

Before the end of the Botomian, clastic deposition conformably blanketed the carbonate platform, from eroding highlands in the region. In the Holyoake Range, Starshot Formation contains upper Botomian trilobites and is conformably interbedded with Douglas Conglomerate (Myrow et al. 2002a). In the area immediately to the south of Byrd Glacier, clastic and carbonate units interfinger conformably, and the interval is marked by pillow basalts (Stump et al. 2004). A tuff in Shackleton Limestone 50 meters beneath the first, overlying clastic unit has been U-Pb zircon dated as 512 ±5 Ma (Stump et al. 2004). Paleocurrent indicators signal a western source in the Holyoake and Surveyors Ranges (Myrow et al. 2002a,b). A source to the north of Byrd Glacier may also have been supplying sediment to the south at this time.

510–500 Ma

Folding and thrust faulting of Byrd Group began during the cycle of deposition of Starshot Formation and Douglas Conglomerate. Shackleton Limestone was eroded, providing a majority of the clasts for the Douglas, which is seen in unconformable contact with Shackleton at two localities (Rowell et al. 1988; Stump et al. 2004). Throughout the Holyoake Range and Churchill Mountains, the trends of folding and thrust faulting are primarily oriented N-S (Laird et al. 1971; Myrow et al. 2002a,b), whereas in the area immediately south of Byrd Glacier fold trends are ENE parallel to the glacier (Skinner 1964; Stump et al. 2004). We suggest that this pattern reflects a reactivation of the western and northern, sutured boundaries between the Beardmore microcontinent and the cratonic margin at the promontory along Byrd Glacier.

505–490 Ma

Although magmatism began by 550 Ma, as discussed above, it was volumetrically minor. Then between 505–490 Ma voluminous magmatism intruded large regions throughout the TAM (Borg 1983; Borg et al. 1987; Allibone et al. 1993a,b; Allibone and Wysoczanski 2002). The magmatism overlapped the latter stages of deformation and continued post-tectonically. To the south of Byrd Glacier a post-deformation pluton and dikes have been U-Pb dated at 492 ±2 Ma (Stump et al. 2002b). The magmatic flare-up was short-lived, ~15 million years, and was comparable to episodes recorded in the Sierra Nevada batholith of North America, postulated to have been caused by lithospheric-scale underthrusting of the magmatic arc (Ducea 2001). Late-stage dikes and minor plutons were intruded after the flare-up in southern Victoria Land for another 10–15 million years (Allibone and Wysoczanski 2002) and then Ross orogenic activity ended. Erosion cut deeply into the mountain belt and subsequent tectonic activity shifted outward along the continent.

Analytical Methods

Zircons from two samples were analyzed in the Radiogenic Isotope Geochemistry Laboratory of the Department of Geosciences, University of Arizona, with a Micromass Isoprobe multicollector ICPMS equipped with nine faraday collectors, an axial Daly detector, and four ion-counting channels. The Isoprobe is equipped with a New Wave DUV 193 laser ablation system with an emission wavelength of 193 nm. The analyses were conducted on 35–50 micron spots with an output energy of ~40 mJ and a repetition rate of 8 Hz. Each analysis consisted of one 20-s integration on the backgrounds (on peaks with no laser firing) and twelve one-second integrations on peaks with the laser firing. The depth of each ablation pit is ~12 microns. The collector configuration allows simultaneous measurement of ^{204}Pb in a secondary electron multiplier while ^{206}Pb, ^{207}Pb, ^{208}Pb, ^{232}Th, and ^{238}U are measured with Faraday detectors. All analyses were conducted in static mode.

Correction for common Pb was done by measuring ^{206}Pb/^{204}Pb, with the composition of common Pb from Stacey and Kramers (1975) and uncertainties of 1.0 for ^{206}Pb/^{204}Pb and 0.3 for ^{207}Pb/^{204}Pb. Fractionation of ^{206}Pb/^{238}U and ^{206}Pb/^{207}Pb during ablation was monitored by analyzing fragments of a large concordant zircon crystal that has a known (ID-TIMS) age of 564 ±4 Ma (2σ) (G. E. Gehrels, unpublished data). Typically this reference zircon was analyzed once for every four unknowns. The uncertainty arising from this calibration correction, combined with the uncertainty from decay constants and common Pb composition, contributed ~1% systematic error to the ^{206}Pb/^{238}U and ^{206}Pb/^{207}Pb ages (2-sigma level).

The reported ages are based on ^{206}Pb/^{238}U ratios because the errors of the ^{207}Pb/^{235}U and ^{206}Pb/^{207}Pb ratios are significantly greater (Table 4.3-1). This is due in large part to the low intensity (commonly ~1 mv) of the ^{207}Pb signal from these young grains. For each sample, the ^{206}Pb/^{238}U ages are shown on an age plot separate age plot (using plotting program of Ludwig 2001). The final age calculations are based on the weighted mean of the cluster of ^{206}Pb/^{238}U ages, with the error expressed both as the uncertainty of this mean and as the error of the age. The age error is based on the quadratic sum of the weighted mean error and the systematic error. Both are expressed at the 2σ level.

Acknowledgment

We thank Alex Pullen and George Gehrels for the U-Pb analyses. Funding sources include NSF grant OPP-9909463 (E. S.) and PNRA Research Project 4.13 (F. T.).

References

Allibone AH, Wysoczanski R (2002) Initiation of magmatism during the Cambrian-Ordovician Ross orogeny in southern Victoria Land, Antarctica. Geol Soc Amer Bull 114:1007–1018

Allibone AH, Cox SC, Graham IJ, Smillie RW, Johnstone RD, Ellery SG, Palmer K (1993a) Granitoids of the Dry Valleys area, southern Victoria Land, Antarctica: plutons, field relationships, and isotopic dating. New Zealand J Geol Geophys 36:281–297

Allibone AH, Cox, SC, Smillie RW (1993b) Granitoids of the Dry Valleys area, southern Victoria Land: geochemistry and evolution along the early Paleozoic Antarctic craton margin. New Zealand J Geol Geophys 36:299–316

Anonymous (1969–1970) Geologic Maps of Antarctica. American Geographical Society, New York

Borg SG (1983) Petrology and geochemistry of the Queen Maud Batholith, central Transantarctic Mountains, with implications for the Ross Orogeny. In: Oliver RL, James PR, Jago JB (eds) Antarctic earth science. Austral Acad Sci, Canberra, pp 165–169

Borg SG, DePaolo DJ (1991) A tectonic model of the Antarctic Gondwana margin with implications for southeastern Australia: isotopic and geochemical evidence. Tectonophysics 196:339–358

Borg SG, DePaolo DJ (1994) Laurentia, Australia, and Antarctica as a Late Proterozoic supercontinent: constraints from isotopic mapping. Geology 22:307–310

Borg SG, Stump E, Chappell BW, McCulloch MT, Wyborn D, Armstrong RL, Holloway JR (1987) Granitoids of northern Victoria Land, Antarctica: implications of chemical and isotopic variations to regional crustal structure and tectonics. Amer J Sci 287:127–169

Borg SG, DePaolo DJ, Wendlandt ED, Drake TG (1989) Studies of granites and metamorphic rocks, Byrd Glacier area. Antarct J US 24:19–21

Borg SG, DePaolo DJ, Smith BM (1990) Isotopic structure and tectonics of the central Transantarctic Mountains. J Geophys Res 95:6647–6667

Bowring SA, Erwin DH (1998) A new look at evolutionary rates in deep time: uniting paleontology and high-precision geochronology. GSA Today 8:1–8

Cook YA, Craw D (2001) Amalgamation of disparate crustal fragments in the Walcott Bay – Foster Glacier area, South Victoria Land, Antarctica. New Zealand J Geol Geophys 44:403–416

Cook YA, Craw D (2002) Neoproterozoic structural slices in the Ross orogen, Skelton Glacier area, south Victoria Land. New Zealand J Geol Geophys 45:133–143

Cooper AF, Worley BA, Armstrong RA, Price RC (1997) Synorogenic alkaline and carbonatitic magmatism in the Transantarctic Mountains of South Victoria Land, Antarctica. In: Ricci CA (ed) The Antarctic region: geological evolution and processes. Terra Antartica Publication, Siena, pp 245–252

Cox SC, Parkinson DL, Allibone AH, Cooper AF (2000) Isotopic character of Cambro-Ordovician plutonism, southern Victoria Land, Antarctica. New Zealand J Geol Geophys 434:501–520

Davidek K, Landing E, Bowring SA, Westrop SR, Rushton AWA, Fortey RA, Adrian JM (1998) New uppermost Cambrian U-Pb date from Avalonian Wales and age of the Cambrian-Ordovician boundary. Geol Mag 135:305–309

Debrenne F, Kruse PD (1986) Shackleton Limestone archaeocyaths. Alcheringa 10:235–278

Ducea M (2001) The California arc: thick granitic batholiths, eclogitic residues, lithospheric-scale thrusting, and magmatic flare-ups. GSA Today 11: 4–10

Encarnación JP, Grunow AM (1996) Changing magmatic and tectonic styles along the paleo-Pacific margin of Gondwana and the onset of early Paleozoic magmatism in Antarctica. Tectonics 15:1325–1341

Encarnación J, Rowell AJ, Grunow AM (1999) A U-Pb age for the Cambrian Taylor Formation, Antarctica: implications for the Cambrian time scale. J Geol 107:497–504

Findlay RH, Skinner DNB, Craw D (1984) Lithostratigraphy and structure of the Koettlitz Group, McMurdo Sound, Antarctica. New Zealand J Geol Geophys 27:513–536

Goodge JW (1997) Latest Neoproterozoic basin inversion of the Beardmore Group, central Transantarctic Mountains, Antarctica. Tectonics 16:682–701

Goodge JW, Dallmeyer RD (1996) Contrasting thermal evolution within the Ross Orogen, Antarctica Evidence from mineral 40Ar/39Ar Ages. J Geol 104:435–458

Goodge JW, Fanning CM (1999) 2.5 b.y. of punctuated Earth history as recorded in a single rock. Geology 27:1007–1010

Goodge JW, Borg SG, Smith BK, Bennett VC (1991) Tectonic significance of Proterozoic ductile shortening and translation along the Antarctic margin of Gondwana. Earth Planet Sci Letters 102:58–70

Goodge JW, Hansen VL, Peacock SM, Smith BK, Walker NW (1993a) Kinematic evolution of the Miller Range shear zone, central Transantarctic Mountains, Antarctica, and implications for Neoproterozoic to early Paleozoic tectonics of the East Antarctic margin of Gondwana. Tectonics 12:1460–1478

Goodge JW, Walker NW, Hansen VL (1993b) Neoproterozoic-Cambrian basement-involved orogenesis within the Antarctic margin of Gondwana. Geol 21:37–40

Goodge JW, Fanning CM, Bennett VC (2001) U-Pb evidence of ~1.7 Ga crustal tectonism during the Nimrod Orogeny in the Transantarctic Mountains, Antarctica: implications for Proterozoic plate reconstructions. Precambrian Res 112:261–288

Goodge JW, Myrow P, Williams IS, Bowring SA (2002) Age and provenance of the Beardmore Group, Antarctica: constraints on Rodinia supercontinent breakup. J Geol 110:393–406

Gootee B (2002) Geology of the Cambrian Byrd Group, Byrd Glacier area, Antarctica. MSc Thesis, Arizona State University, Tempe

Gootee B, Stump E (2006) Depositional environments of the Byrd Group, Byrd Glacier area: a Cambrian record of sedimentation, tectonism, and magmatism. In: Fütterer DK, Damaske D, Kleinschmidt G, Miller H, Tessensohn F (eds) Antarctica – Contributions to global earth sciences. Springer, Berlin Heidelberg New York, pp 191–194

Grindley GW (1972) Polyphase deformation of the Precambrian Nimrod Group, Central Transantarctic Mountains. In: Adie RJ (ed) Antarctic geology and geophysics. Universitetsforlaget, Oslo, pp 313–318

Grindley GW, Laird MG (1969) Geology of the Shackleton Coast. Antarctic Map Folio Series Folio 12, XIV

Grindley GW, McDougall I (1969) Age and correlation of the Nimrod group and other Precambrian rock units in the Central Transantarctic Mountains, Antarctica. New Zealand J Geol Geophys 12:391–411

Grindley GW, Warren G (1964) Stratigraphic nomenclature and correlation in the western Ross sea region. In: Adie RJ (ed) Antarctic Geology. North-Holland, Amsterdam, pp 314–333

Grindley GW, McGregor VR, Walcott RI (1964) Outline of the geology of the Nimrod-Beardmore-Axel Heiberg glaciers region, Ross Dependency. In: Adie RJ (ed) Antarctic Geology. North-Holland, Amsterdam, pp 206–219

Gunn BM, Walcott RI (1962) The geology of the Mt Markham region, Ross Dependency, Antarctica. New Zealand J Geol Geophys 5:407–426

Gunn BM, Warren G (1962) Geology of Victoria Land between the Mawson and Mulock glaciers, Antarctica. New Zealand Geol Surv Bull 7:1–157

Gunner JD, Faure G (1972) Rubidium-strontium geochronology of the Nimrod Group, central Transantarctic Mountains. In: Adie RJ (ed) Antarctic geology and geophysics. Universitetsforlaget, Oslo, pp 305–311

Hall CE, Cooper AF, Parkinson DL (1995) Early Cambrian carbonatite in Antarctica. J Geol Soc London 152:721–728

Hill D (1964) Archaeocyatha from the Shackleton limestone of the Ross system, Nimrod glacier area, Antarctica. Trans Royal Soc New Zealand Geol 2:137–146

Laird MG (1963) Geomorphology and stratigraphy of the Nimrod glacier-Beaumont bay region, southern Victoria Land, Antarctica. New Zealand J Geol Geophys 6:465–484

Laird MG, Mansergh GD, Chappell JMA (1971) Geology of the central Nimrod Glacier area, Antarctica. New Zealand J Geol Geophys 14:427–468

Landing E, Bowring SA, Davidek KL, Westrop SR, Geyer G, Heldmaier W (1998) Duration of the Early Cambrian: U-Pb ages of volcanic ashes from Avalon and Gondwana. Canadian J Earth Sci 35:329–338

Ludwig KJ (2001) Isoplot/Ex (rev. 2.49). Berkeley Geochronology Center Spec Publ 1a:1–56

Mellish SD, Cooper AF, Walker NW (2002) Panorama Pluton: a composite gabbro-monzodiorite early Ross Orogeny intrusion in southern Victoria Land, Antarctica. In: Gamble JA, Skinner DNB, Henrys S (eds) Antarctica at the close of a millennium. Royal Soc New Zealand Bull 35, Wellington, pp 129–141

Moores EM (1991) Southwest U.S.-East Antarctic (SWEAT) connection: a hypothesis. Geology 19:425–428

Myrow PM, Pope MC, Goodge JW, Fischer W, Palmer AR (2002a) Depositional history of pre-Devonian strata and timing of Ross orogenic tectonism in the central Transantarctic Mountains, Antarctica. Geol Soc Amer Bull 114:1070–1088

Myrow PM, Fischer W, Goodge JW (2002b) Wave-modified turbidites: Combined-flow shoreline and shelf deposits, Cambrian, Antarctica. J Sed Res 72:641–656

Palmer AR, Rowell AJ (1995) Early Cambrian trilobites from the Shackleton Limestone of the central Transantarctic Mountains. Paleontol Soc Mem 45:1–28

Read SE, Cooper AF, Walker NW (2002) Geochemistry and U-Pb geochronology of the Neoproterozoic-Cambrian Koettlitz Glacier alkaline province, Royal Society Range, Transantarctic Mountains, Antarctica. In: Gamble JA, Skinner DNB, Henrys S (eds) Antarctica at the close of a millennium. Royal Soc New Zealand Bull 35, Wellington, pp 143–151

Rees MN, Pratt BR, Rowell AJ (1989) Early Cambrian reefs, reef complexes, and associated lithofacies of the Shackleton Limestone, Transantarctic Mountains. Sedimentology 36:341–361

Rowell AJ, Rees MN, Cooper RA, Pratt BR (1988) Early Paleozoic history of the central Transantarctic Mountains: evidence from the Holyoake Range, Antarctica. New Zealand J Geol Geophys 31:397–404

Rowell AJ, Rees MN, Duebendorfer EM, Wallin ET, Van Schmus WR, Smith EI (1993) An active Neoproterozoic margin: evidence from the Skelton Glacier area, Transantarctic Mountains. J Geol Soc London 150:677–682

Simpson AL, Cooper AF (2002) Geochemistry of the Darwin Glacier region granitoids, southern Victoria Land. Antarctic Sci 14: 425–426

Skinner DNB (1964) A summary of the geology between Byrd and Starshot Glaciers, south Victoria Land. In: Adie RJ (ed) Antarctic geology. North-Holland, Amsterdam, pp 284–292

Skinner DNB (1982) Stratigraphy and structure of low-grade metasedimentary rocks of the Skelton Group, southern Victoria Land: does Teall Greywacke really exist? In: Craddock C (ed) Antarctic geoscience. University of Wisconsin Press, Madison, pp 555–563

Stacey JS, Kramers JD (1975) Approximation of terrestrial lead isotope evolution by a two-stage model. Earth Planet Sci Letters 26:207–221

Stump E (1992) The Ross Orogen of the Transantarctic Mountains in light of the Laurentia-Gondwana split. GSA Today 2:25–31

Stump E, Smit JH, Self S (1986) Timing of events during the late Proterozoic Beardmore Orogeny, Antarctica: geological evidence from the La Gorce Mountains. Geol Soc Amer Bull 97:953–965

Stump E, Edgerton DG, Korsch RJ (2002a) Geological relationships at Cotton Plateau, Nimrod Glacier area, bearing on the tectonic development of the Ross orogen, Transantarctic Mountains, Antarctica. Terra Antartica 9:3–18

Stump E, Foland KA, Van Schmus WR, Brand PK, Dewane TJ, Gootee BF, Talarico F (2002b) Geochronology of deformation, intrusion, and cooling during the Ross orogeny, Byrd Glacier area, Antarctica. Geol Soc Amer, Abstracts Progr 34:560–561

Stump E, Gootee BF, Talarico F, Van Schmus WR, Brand PK, Foland KA, Fanning CM (2004) Correlation of Byrd and Selborne Groups, with implications for the Byrd Glacier discontinuity, central Transantarctic Mountains. New Zealand J Geol Geophys 47:157–171

Tucker RD, McKerrow WS (1995) Early Paleozoic chronology: a review in light of new U-Pb zircon ages from Newfoundland and Britain. Canadian J Earth Sci 32:368–379

Van Schmus WR, McKenna LW, Gonzales DA, Fetter AH, Rowell AJ (1997) U-Pb geochronology of parts of the Pensacola, Thiel, and Queen Maud Mountains, Antarctica. In: Ricci CA (ed) The Antarctic region: geological evolution and processes. Terra Antartica Publication, Siena, pp 187–200

Vogel MB, Wooden JL, Stump E, McWilliams MO (2001) Detrital zircon provenance of Neoproterozoic to Cambrian successions: resolving continental configurations. Gondwana Res 4:809

Wareham CD, Stump E, Storey BC, Millar IL, Riley TR (2001) Petrogenesis of the Cambrian Liv Group, a bimodal volcanic rock suite from the Ross orogen, Transantarctic Mountains. Geol Soc Amer Bull 113:360–372