1 **Synchronous alkaline and subalkaline magmatism during the late Neoproterozoic–early**
2 **Paleozoic Ross orogeny, Antarctica: Insights into magmatic sources and processes**
3 **within a continental arc**

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11
12 **Abstract**

13 Extensive exposure of intrusive igneous rocks along the Ross orogen of Antarctica—an
14 ancient accretionary orogen on the margin of East Gondwana—provides an exceptional
15 opportunity to study continental arc magmatism. There is significant petrologic and
16 geochemical variability in igneous rocks within a ~500-km-long segment of the arc in
17 southern Victoria Land. The conspicuous occurrence of carbonatite and alkaline silicate
18 rocks (nepheline syenite, A-type granite, and alkaline mafic rocks) adjacent to large
19 complexes of subalkaline granitoids is not adequately explained by traditional models for
20 continental arc magmatism. Extensive geochemical analysis (>100 samples) and zircon U-Pb
21 geochronology ($n = 70$) confirms that alkaline and carbonatitic magmatism was partially
22 contemporaneous with the emplacement of large subduction-related igneous complexes in
23 adjacent areas. Major pulses of subalkaline magmatism were compositionally distinct and
24 occurred at different times along the arc. Large bodies of subalkaline orthogneiss and granite
25 (*sensu lato*) were emplaced over similar time intervals (ca. 25 Myr) to the north (ca. 515–492
26 Ma) and south (ca. 550–525 Ma) of the alkaline magmatic province, although the onset of
27 these major pulses of magmatism were offset by ca. 35 Myr. Alkaline and carbonatitic
28 magmatism spanned at least ca. 550–509 Ma, overlapping with voluminous subalkaline
29 magmatism in adjacent areas. The most primitive rocks from each area have similarly
30 enriched trace element compositions, indicating some common characteristics of the magma
31 sources along the arc. The samples from the older subalkaline complex have invariably low
32 Sr/Y ratios (<40), consistent with relatively shallow magma generation and differentiation.
33 The younger subalkaline complex and subalkaline rocks within the area of the alkaline
34 province extend to higher Sr/Y ratios (up to ~300), indicative of generation and
35 differentiation at deeper levels. The significant spatial and temporal diversity in magmatism
36 can be explained by a tectono-magmatic model that invokes alternating phases of extension
37 and contraction in the overriding plate. The low Sr/Y magmatism of the earlier subalkaline
38 complex and the largely contemporaneous alkaline and carbonatitic magmatism occurred
39 during a phase of extension in the overriding plate ca. 550–525 Ma. The differences in
40 composition of this early phase of magmatism can be explained by variable lithospheric
41 thickness along the margin, with the alkaline magmas derived from partial melts of thick
42 metasomatized lithosphere. The younger subalkaline complex was emplaced during a phase
43 of contraction in the overlying plate ca. 515–505 Ma. The latest phase of (post-orogenic)
44 magmatism occurred during extension in the overriding plate ca. 505–492 Ma.

Highlights

- Alkaline and subalkaline magmatism were contemporaneous in a continental arc.
- Large subalkaline igneous complexes were emplaced at different times.
- Alkaline magmas were sourced from metasomatized lithospheric mantle.
- Magmatic compositions record phases of shortening and extension in the upper plate.

Key words

Subduction-related magmatism; Alkaline magmatism; Zircon U-Pb geochronology; Gondwana; Ross orogen; Antarctica

1. Introduction

The processes of subduction and associated magmatism have significant influence on the geochemical evolution of the crust and mantle. Subduction recycles crust back into the mantle, re-enriching it in incompatible elements (Pearce, 1983; Pearce and Peate, 1995; Plank and Langmuir, 1993). Island arcs above subduction zones are important areas of crustal generation from mantle-derived melts (e.g., Gill, 1981). Continental arcs are settings of “distillation” of pre-existing continental crust (e.g., Ducea and Barton, 2007) and may also be important settings of crustal growth (e.g., Rogers and Hawkesworth, 1989). Classic models for the source of arc magmas invoke an H₂O-fluxed mantle wedge (Davies and Bickle, 1991; Gill, 1981) with variable contribution from the subducted slab and sediments. The magmas then evolve along calc-alkaline or tholeiitic trends with products ranging from basalt to rhyolite. As Ducea and Barton (2007) noted, such models are fundamentally based on island arcs. Processes are complicated in continental settings due to the ascent of magmas through a thick column of continental lithosphere, accompanied by stalling, fractionation, and assimilation. The result is compositional and isotopic variability that is not adequately explained by traditional models for subduction-related magmatism.

Extensive exposure of continental arc granitoids associated with the Neoproterozoic–

early Paleozoic Ross orogeny of Antarctica (Fig. 1) provides an exceptional opportunity to study variations in the character of magmatism along a major continental arc. Within a ~500 km-long segment of the arc in southern Victoria Land (sVL), igneous rocks display significant petrologic and geochemical variability, with the conspicuous occurrence of a suite of alkaline silicate rocks and carbonatites adjacent to large complexes of subalkaline, subduction-related granitoids. Zircon U-Pb data from ~70 samples from sVL provide the densest geochronologic dataset to-date for igneous rocks in the Ross orogen. These data shed light on the extreme variability in the composition of magmatism across southern Victoria Land and the spatial and temporal relationships between alkaline and subalkaline magmatism within the arc. Alkaline and carbonatitic igneous rocks are most commonly associated with intraplate settings (e.g., Fitton and Upton, 1987; Sørensen, 1974; Woolley, 1989) or post-orogenic magmatism (e.g., Bonin, 1990; Bonin et al., 1998), and the context of these rocks within the magmatic arc of the Ross orogen has been enigmatic. However, the petrologic, geochemical, and geochronologic associations of the igneous rocks in sVL can be interpreted in the context of alternating phases of extension and contraction in the overriding plate.

2. Geologic Background

The Ross orogen is a belt of deformed and metamorphosed sedimentary rocks and granitic (*sensu lato*) plutons exposed along the Transantarctic Mountains (Fig. 1). The orogen spans ~3500 km across Antarctica and extends into southeastern Australia and southwestern New Zealand (Foden et al., 2006). Deformation, metamorphism, and magmatism in the Ross orogen occurred in response to the subduction of paleo-Pacific oceanic lithosphere beneath the margin of the East Antarctic Craton during the late Neoproterozoic–early Paleozoic

(summarized in Stump, 1995). Igneous rocks exposed along the magmatic arc are collectively referred to as the Granite Harbour Intrusive Complex (Fig. 1; Gunn and Warren, 1962).

In the southern Victoria Land segment of the orogen—the largest ice-free area of Antarctica—metasedimentary rocks of the Skelton Group host plutonic rocks of the Granite Harbour Intrusive Complex (Fig. 1; Cox et al., 2012). The igneous and metamorphic basement is unconformably overlain by Devonian–Triassic siliciclastic sedimentary rocks of the Beacon Supergroup (summarized in Barrett, 1981), and both the basement and cover rocks were intruded by the regionally extensive Ferrar dolerite sills in the Jurassic (Encarnación et al., 1996; Heimann et al., 1994). The composition of the Granite Harbour Intrusives in sVL is highly variable, both in terms of petrology and geochemistry. In this study, we divide these igneous rocks into three groups based on their geographic distribution: 1) the Dry Valleys area, north of the Howchin Glacier, 2) the Koettlitz Glacier area, between the Howchin and Mulock Glaciers, and 3) the Byrd and Darwin Glacier areas, between the Mulock and Byrd Glaciers (Fig. 2). It should be noted that these broad areas do not perfectly delineate the differences in magmatic character across sVL, but generally envelop magmatic provinces of different composition and age.

2.1 Dry Valleys Area

The Dry Valleys (DV) area is defined here as the area between the Fry and Howchin Glaciers (Figs. 2 and A1). Areally extensive plutonic rocks in this area, hosted in migmatitic metasediments of the Skelton Group (Figs. 3 and A2), have been divided into three magmatic suites based on field characteristics (e.g., pluton elongation, contacts with country rock,

internal fabrics, etc.; Figs. 3 and A2), age, petrology, and geochemistry (Allibone et al., 1993a; b; Smillie, 1992). The DV1a, including the ~1000 km² Bonney Pluton, and DV1b suite plutons are typically elongate in the NW-SE direction and are interpreted as subduction-related granitoids emplaced in a regime of contraction across the arc (Allibone et al., 1993a; b; Smillie, 1992). The DV2 suite rocks have been interpreted as "post-tectonic" granitoids, emplaced at pressures as low as 1.2 kbar (~4 km) after the cessation of subduction (Allibone et al., 1993a; b; Smillie, 1992). Magmatism in this area was relatively short-lived, with a vast majority of the exposed plutonic rocks emplaced from 515–492 Ma (Hagen-Peter et al., 2015).

2.2 Koettlitz Glacier area

The Koettlitz Glacier (KG) area is defined here as the area between the Howchin and Mulock Glaciers (Figs. 2 and A3). The igneous rocks in the KG area vary widely in composition, including subalkaline rocks, as well as calcite-bearing nepheline syenites, carbonatites, alkaline gabbros, and A-type granites of the Koettlitz Glacier Alkaline Suite (Figs. 3, A4; Cooper et al., 1997; Hall et al., 1995; Read et al., 2002; Read, 2010; Rowell et al., 1993; Worley and Cooper, 1995; Worley et al., 1995). The igneous rocks are variably deformed, ranging from undeformed granitoids (e.g., Figs. 4A, 4C) to orthogneiss with strong fabrics defined by aligned micas and extensively recrystallized quartz (e.g., Figs. 3D, 4B). The intrusions in the KG area are relatively small (typically < 10 km²) compared to the large dominantly subalkaline DV plutons (e.g., ~1000 km² Bonney Pluton). Carbonatite dikes—generally small (~1-m-wide) and associated with nepheline syenites—have been identified in at least four locations across the KG area (Figs. 3F; A4e; Cooper et al., 1997;

Hall et al., 1995; Read, 2010; this study). The southern extent of the KG area is currently defined by an A-type granite that crops out on the north side of the Mulock Glacier (Fig. 2; Cottle and Cooper, 2006a). Published geochronology data suggest that magmatism in the KG area occurred ca. 551–517 Ma (Cooper et al., 1997; Cottle and Cooper, 2006a; Encarnación and Grunow, 1996; Hall et al., 1995; Mellish et al., 2002; Read et al., 2002; Rowell et al., 1993). In an unpublished PhD thesis, Read (2010) determined that magmatism started slightly earlier at ca. 557 Ma and persisted until ca. 511 Ma.

2.3 Byrd and Darwin Glacier Areas

The Byrd and Darwin Glaciers (BG) area is defined here as the area between the Mulock and Byrd Glaciers (Figs. 2, A5, A6). The Ross-age basement in the BG area consist of granitoids, mafic, and minor ultramafic rocks hosted in migmatitic metasedimentary rocks of the Horney Formation (Borg et al., 1989; Carosi et al., 2007; Stump et al., 2006). Large granitoid bodies are variably strained, from relatively undeformed with magmatic alignment of K-feldspars (Fig. 3J) to highly strained orthogneisses (Figs. 3H, 4E, 4H), and cut by deformed and undeformed leucocratic dikes as well as lamprophyres (Figs. A7c, A7g, A7h, A7i). The granitoids have been variably divided into different magmatic suites (Encarnación and Grunow, 1996; Simpson and Cooper, 2002), although the interpretations of BG area samples in this study are largely independent of this framework. Mafic plugs and small plutons, including the ~0.5 km² layered Fontaine Pluton (Cottle and Cooper, 2006b), are common (Figs. 3I, A7f, A7g). Sparse geochronology on igneous rocks in the BG area suggests that magmatism spanned ca. 547–507 Ma (Carosi et al., 2007; Cottle and Cooper, 2006b; Encarnación and Grunow, 1996; Stump et al., 2006).

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170 **3. Methods**

171 *3.1 Whole-Rock Geochemistry*

172 Whole-rock geochemistry data for samples with the prefix “P” were reported in Hagen-Peter
173 et al. (2015), and were acquired by the methods described therein. Whole-rock geochemical
174 analyses for samples with prefixes “RG” and “DG” were performed at ALS Minerals
175 (alsglobal.com; Reno, NV, USA). Rock chips were crushed and sieved to < 75 µm.
176 Approximately 0.2 g of the pulverized sample was added to 0.9 g of lithium-metaborate flux.
177 The mixture was fused at 1000°C and weighed for loss-on-ignition. The resulting glass bead
178 was dissolved in 100 mL of 4% HNO₃ plus 2% HCl. Major- and trace-elements were then
179 measured by solution inductively-coupled plasma atomic emission spectrometry (ICP-AES)
180 and inductively-coupled plasma mass spectrometry (ICP-MS), respectively. Standard
181 reference solutions, duplicate analyses, and blanks were included in the measurement
182 routines.

183

184 *3.2 Zircon U-Pb Geochronology*

185 Geochronology data for the DV area samples were previously reported in Hagen-Peter et al.
186 (2015). Data for the KG and BG areas were collected by the same methods, described here.
187 Zircon separation, imaging, and isotope analyses were done at University of California, Santa
188 Barbara. Zircons were separated from rock samples using standard crushing, sieving, density
189 separation, and magnetic separation techniques. Grains were hand-selected using a binocular
190 microscope—with care taken to select crack- and inclusion-free grains—and mounted in
191 epoxy resin. The disks were polished to expose median sections of the zircon grains and

carbon coated. Zircons were imaged using an FEI Quanta 400f secondary electron microscope (SEM) with a cathodoluminescence (CL) detector. The CL images were used to guide the placement of laser-ablation pits.

Zircons were ablated in a He environment using a Photon Machines 193 nm ArF excimer laser and HelEx cell. Laser energy of 3–4 mJ and 4 Hz repetition rate were used to ablate pits 24 or 31 μm in diameter. Ablation periods were 15–20 seconds (0.2 or 0.4 second integration periods), preceded by two “cleaning” shots (to removed common lead from the surface of the sample) and 12–15 second washout periods for the signal to return to baseline. The ablated sample was transported in a He+Ar mixture to a Nu Plasma multi-collector ICP-MS for the measurement of U-Th-Pb isotopes. Isotopes ^{238}U and ^{232}Th were measured with Faraday cups equipped with 10^{11} -ohm resistors, while ^{208}Pb , ^{207}Pb , ^{206}Pb , and $^{204}(\text{Pb}+\text{Hg})$ were measured on secondary electron multipliers. Data reduction and age calculations were done using Iolite version 2.31 (Paton et al., 2010; 2011) and Isoplot version 3.00 (Ludwig, 2003).

Every six to eight analyses were bracketed by analyses of reference zircons. The ‘91500’ reference zircon (1065.4 ± 0.3 Ma $^{207}\text{Pb}/^{206}\text{Pb}$ ID-TIMS age and 1062.4 ± 0.4 Ma $^{206}\text{Pb}/^{238}\text{U}$ ID-TIMS age; Wiedenbeck et al., 1995; 2004) was used as a primary reference to correct for mass bias, Pb/U and Pb/Th fractionation, and instrumental drift over the analytical sessions. Reference zircons ‘GJ-1’ (ID-TIMS $^{207}\text{Pb}/^{206}\text{Pb}$ age = 607.7 ± 0.67 Ma and $^{206}\text{Pb}/^{238}\text{U}$ age = 601.86 ± 0.37 Ma; Horstwood et al., 2016) and ‘Plešovice’ (337.13 ± 0.37 Ma $^{206}\text{Pb}/^{238}\text{U}$ ID-TIMS age Sláma et al., 2008) were used to assess data accuracy and external reproducibility. Throughout all analytical sessions, ‘GJ-1’ yielded a weighted-mean $^{206}\text{Pb}/^{238}\text{U}$ age of 602.4 ± 7.3 (2 s.d.; $n = 454$; MSWD = 1.1, and ‘Plešovice’ yielded a

weighted-mean $^{206}\text{Pb}/^{238}\text{U}$ age of 339.2 ± 4.2 (2 s.d.; $n = 313$; MSWD = 1.1).

4. Results

4.1 Whole-Rock Geochemistry

Major- and trace-element geochemistry for representative samples from the DV, KG, and BG areas are shown in Figures 5–7, and full results are provided in Table S1. Samples from the KG and BG areas span a range of SiO_2 content from ~45–78 wt% (not including carbonatites), whereas samples from the DV area span ~48–75 wt%, with a gap from ~60–69 wt% in our samples. On a total alkali versus silica (TAS) diagram, most DV and BG samples plot within the subalkaline field, with several low- SiO_2 samples slightly within the alkaline field. Samples from the KG area plot both within the subalkaline and alkaline fields, and several samples have very high alkali concentrations (e.g., $\text{Na}_2\text{O} + \text{K}_2\text{O} > 10$ wt%, Fig. 5A). On a modified alkali-lime index diagram (MALI; Frost et al., 2001), DV samples plot primarily in the calc-alkalic and alkali-calcic fields, BG samples plot primarily in the calcic and calc-alkalic fields, and KG area samples span the calcic through alkalic fields (Fig. 5B). Most samples from each area plot near the boundary of the metaluminous and peraluminous fields in terms of Shand's index, although several samples from the KG area extend far into the peraluminous and peralkaline fields (Fig. 5C). The samples from each area do not plot along simple trends on Harker diagrams, and there is generally more dispersion in major oxide and trace element concentrations at lower SiO_2 contents (Fig. 6). The dispersion is especially pronounced in KG area samples, with ranges of approximately 16, 18, 10, and 14 wt% for Al_2O_3 , CaO , Fe_2O_3 , and MgO , respectively, in samples with < 60 wt% SiO_2 . The arrays of major element concentrations (Al_2O_3 , CaO , Fe_2O_3 , and MgO shown in Fig. 6)

generally converge at high SiO₂ contents for samples from each area. Nepheline ± calcite syenites from the KG area have very low MgO contents and correspondingly low Mg# < 10 (molar Mg/[Mg+Fe]). Several KG area samples have very high CaO concentrations (> 16 wt%) coupled with high MgO and low Fe₂O₃ concentrations, and correspondingly high Mg# > 70. Samples from each area have widely variable trace element concentrations, with samples from the DV and KG areas much more strongly enriched in Sr, with high Sr/Y values in the “HiSY” range (Sr/Y > 40) of Tulloch and Kimbrough (2003).

On the discrimination diagram of Whalen et al. (1987), DV and BG samples plot within or near the boundary of the I- and S-type granite fields, whereas many KG area samples extend well into the A-type granite field (Fig. 7). On tectonic discrimination diagrams of Pearce et al. (1984), DV and BG samples plot primarily in the volcanic arc granite fields, whereas several KG area samples extend into the within-plate granite fields.

Normalized trace- and rare earth-element (REE) diagrams for the most-primitive silicate rocks (SiO₂ < 52 wt%; Mg# > 69) and carbonatites are shown in Figure 8. All of the primitive silicate samples show enriched trace element patterns with high concentrations of the large-ion lithophile elements (LILE) relative to the more-compatible high-field strength elements (HFSE) and heavy REE, the latter less than 10 times primitive mantle concentrations. The patterns typically show negative Nb, P, and Ti anomalies and positive K and Pb anomalies. The samples show enriched patterns on REE diagrams, with light REE concentrations up to ~100 times chondrite for the silicate rocks and ~700 times chondrite for the carbonatites, as well as normalized La/Yb ratios of 3.7–5.8 for the silicate rocks and 2.9–4.3 for the carbonatites. Some of the samples display moderately negative to moderately positive Eu anomalies (~0.45–1.63; $Eu/Eu^* = [(Eu_{norm}/(Sm_{norm} * Gd_{norm}))^{1/2}]$).

4.2 Zircon U-Th-Pb Geochronology

Representative CL images of zircons and concordia diagram examples are shown in Figures 9 and 10, respectively. A summary of the U-Pb geochronology data is provided in Table 1, and full results are provided in Table S2. Weighted-mean $^{206}\text{Pb}/^{238}\text{U}$ dates were calculated for most samples. Some samples have mean-square weighted deviates (MSWD) greater than expected for a statistically uniform normally-distributed population (Wendt and Carl, 1991). This excess dispersion in some samples may be due to a number of factors, including the presence of small amounts of common lead in the zircons when they crystallized and/or minor degrees of lead loss, though the exact cause of the excess dispersion is typically unresolvable in our data. We take the dates reported in Table 1 as the best estimates of the crystallization ages of the rocks in which the zircons were hosted. The uncertainties on the reported dates take into account both the internal uncertainty of the U-Pb isotope ratio measurements and the long-term reproducibility of secondary reference zircons, resulting in total uncertainties of approximately $\pm 6\text{--}7$ Ma (i.e., $\sim 1.2\%$) for most samples. Samples from the DV area range from ca. 515–492 Ma. Samples from the KG area range from ca. 565–499 Ma. Samples from the BG area range from ca. 550–503 Ma.

5. Discussion

5.1 Petrologic and Geochemical Variations in Magmatism

There is significant petrologic and geochemical variability in the igneous rocks from along a ~ 500 km segment of the Ross orogen in southern Victoria Land (sVL). Large complexes of orthogneiss, granitoids, and mafic rocks were emplaced in the DV and BG areas, whereas

relatively small bodies of alkaline silicate rocks and carbonatites were emplaced in the KG area. Although the igneous rocks from both the DV and BG areas are dominantly subalkaline and plot primarily in the “volcanic arc granite” fields on tectonic discrimination diagrams (Fig. 7), these complexes are compositionally distinct. In particular, granitoids from the BG area are generally more calcic and ferromagnesian, and less alkaline than those from the DV area. Some of the igneous rocks in the KG area have distinctly alkaline compositions and plot in the A-type granite and “within-plate granite” fields. These rock types, including carbonatites, nepheline syenites, alkaline gabbros, and A-type granites, are much more commonly associated within intraplate tectonic environments (e.g., Fitton and Upton, 1987; Sørensen, 1974; Woolley, 1989) or post-orogenic (e.g., Bonin, 1990; Bonin et al., 1998) magmatism than a continental arc setting.

The geochemical variability of the igneous rocks within each area also rules out a common genetic relationship among the samples within each complex. On Harker diagrams, the samples from each complex do not form coherent arrays, and there is significant dispersion in major- and trace-element concentrations at a given SiO₂ (Fig. 6) or MgO content. These geochemical features could not have developed through simple fractional crystallization/melting processes. Instead, they require variability in magma sources and/or magma evolution processes—for example, partial melting under different P-T conditions or different degrees of contamination.

Despite significant differences in major- and trace-element concentrations, the most primitive silicate rocks from each area have remarkably similar normalized trace- and rare earth-element patterns, with similar LILE/HFSE and light REE/heavy REE ratios, negative Nb, P, and Ti anomalies, and positive K and Pb anomalies (Fig. 8). These samples do show

evidence of some degree of differentiation—for example slight to moderate Eu anomalies (~0.45–1.63) and SiO₂ up to 52 wt%—but with relatively high Mg# (>69), as well as Cr and Ni concentrations (50–1680 and 8–194 ppm, respectively), indicating they are the closest representations of primitive magmas in this segment of the arc. These similarities suggest that there are some common features in the magma source(s) in each area, despite the gross petrologic and compositional range of the igneous rocks. The normalized trace element patterns of carbonatites are generally more enriched than the most primitive silicate rocks and have several distinct features, namely high concentrations of Th, U, Nb, and Ta, and extreme negative Ti anomalies (particularly in two samples; Fig. 8)

One prominent difference in trace element composition is the variable Sr/Y ratios of samples from each area. Samples from the DV and KG areas plot in both the “HiSY” (Sr/Y > 40) and “LoSY” (Sr/Y < 40; Tulloch and Kimbrough, 2003) fields on a Sr/Y versus SiO₂ diagram (Fig. 6), whereas BG samples from this study plot exclusively in the LoSY field. In the KG area samples, there appears to be a weak negative correlation between Sr/Y and SiO₂, but there is no obvious correlation in the DV and BG samples, suggesting that this ratio is not significantly affected by assimilation–fractional crystallization (AFC) processes. Therefore, the Sr/Y ratio of the samples is taken to reflect the trace element composition of the magma source, or the conditions of melting in the source region. Tulloch and Kimbrough (2003) interpreted HiSY igneous rocks in the Median Batholith of New Zealand and the Peninsular Range of North America to reflect high-pressure melting at the base of the crust during shallow-slab subduction and underthrusting of outboard arc segments. It is therefore inferred that the BG samples represent relatively shallow melting and differentiation (above the garnet stability field), whereas the HiSY DV and KG area samples record higher-pressure

melting conditions.

5.2 Transient Magmatism Along the Arc

The timing of magmatism was variable along the Ross orogen, and the large geochronologic dataset presented here reveals significant variability within a ~500 km-long segment in sVL (Fig. 11). In the BG area, the largest calcic and calc-alkalic plutons were emplaced over a period of ca. 550–525 Ma. Only two new ages from this study fall outside of this range (ca. 510 Ma and 502.8 ± 6.2 Ma), and these are from small leucocratic dikes that crosscut older diorite and granitic orthogneiss. Other age data for the BG area are sparse. Cottle and Cooper (2006b) determined an age of 546 ± 10 Ma for a composite mafic pluton at Fontaine Bluff (Fig. A5). Stump et al. (2006) reported ages of 546 ± 7 and 531 ± 8 for a strained granite and a diorite body, respectively, from near the Ramseier Glacier (Fig. A6). A.F. Cooper (pers. comm.) determined an age of 537 ± 4 Ma for a granodiorite in the Brown Hills area (Fig. A5). Encarnación and Grunow (1996) reported an age of 515 ± 8 Ma for a biotite-hornblende granite at Cooper Nunatak (Fig. A5). Based on a single titanite fraction measured by thermal-ionization mass spectrometry, this date may represent a crystallization age or the timing of closure to lead diffusion during cooling. Nonetheless, most of the large bodies of megacrystic granite, granitic orthogneiss, and diorite/gabbro were emplaced over ca. 25 million years, from 550–525 Ma.

The majority of magmatism in the DV area spanned a similar range of ca. 23 million years, but this major pulse of magmatism occurred ca. 515–492 Ma (Fig. 11; Hagen-Peter et al., 2015)—later than in the BG area. A significant volume (inferred from areal exposure) of primarily calc-alkalic plutonic rocks was emplaced during this period, including the ~1000

km² Bonney Pluton (Allibone and Wysoczanski, 2002; Cox et al., 2000; Encarnación and Grunow, 1996; Hagen-Peter et al., 2015). Martin et al. (2014) dated a ca. 545 Ma xenolith (hosted within Cenozoic Mt. Morning Volcanics) with geochemical affinity to DV granites, and Allibone and Wysoczanski (2002) determined an age of ca. 530 Ma for a small body of orthogneiss in the DV area. However, a dense geochronologic dataset indicates that most of the exposed DV granitoids were emplaced during the period ca. 515–492 Ma.

In contrast to the DV and BG areas, relatively small volumes of compositionally diverse igneous rocks were emplaced in the KG area over a longer period, ca. 565–500 Ma. The oldest sample (RG11-212; 564.7 ± 7.0 Ma) is a two-mica garnet-bearing orthogneiss. A subalkaline gabbro (RG11-146; 561.9 ± 6.8 Ma) is of similar age. Other alkaline mafic rocks, A-type granites, and syenites from this study span ca. 550–542 Ma, and a calcite-bearing nepheline syenite from the north side of the Walcott Glacier was emplaced at ca. 509 Ma. Previously published zircon U-Pb ages for carbonatites, nepheline syenites, A-type granites, and alkaline mafic rocks range from ca. 557–531 Ma (Cooper et al., 1997; Cottle and Cooper, 2006a; Hall et al., 1995; Mellish et al., 2002; Read, 2010). Subalkaline granitic orthogneiss, granitoids, and mafic rocks were emplaced in the KG area over the period of ca. 562–499 Ma, largely overlapping the period of alkaline magmatism. It is possible that magmatism in the KG area was episodic—with pulses of alkaline and subalkaline intrusions separated by millions of years. However, uncertainties of ± 6 –7 million years on weighted-mean $^{206}\text{Pb}/^{238}\text{U}$ dates—interpreted as emplacement ages—preclude distinguishing between episodic and continuous magmatism.

The 564.7 ± 7.0 Ma two-mica granite (RG11-212) from the KG area is the oldest exposed rock of the Granite Harbour Intrusive Complex (GHI) yet identified (Fig. 11) and

therefore places constraints on the initiation of Ross-age magmatism in this region. Goodge et al. (2012) identified peaks at 559 ± 5 Ma and 589 ± 5 Ma in a multi-modal zircon U-Pb age distribution from a granitic glacial clast from the central Transantarctic Mountains (cTAM), interpreting the older peak to represent the crystallization age of the rock. Other records of early magmatism during the Ross orogen include ca. 581 Ma and 595 Ma detrital zircon age peaks in sedimentary rocks from the Queen Maud Mountains (QMM; Stump et al., 2007) and sVL (Paulsen et al., 2011), respectively, and a ca. 584 Ma glacially transported igneous clast dredged off the Wilkes Land coast (Goodge and Fanning, 2010). However, the exact provenance of these detrital samples is unknown.

Based on the geochronology of exposed igneous rocks, the onset of magmatism appears to have been transient along the Ross orogen (Fig. 11). Magmatism commenced in the KG area ca. 565 Ma. The oldest exposed igneous rocks dated in the cTAM (Goodge et al., 2012), BG area, and northern Victoria Land (nVL; Black and Sheraton, 1990; Bomparola et al., 2006; Fioretti et al., 2005; Tiepolo and Tribuzio, 2008) are all ca. 550–545 Ma. Magmatism in the Queen Maud Mountains appears to have commenced at ca. 535 Ma (Paulsen et al., 2013), nearly contemporaneous with the emplacement of small bodies of orthogneiss in the DV area (Allibone and Wysoczanski, 2002). However, if the ca. 545 Ma xenolith with DV granite geochemical affinity (Martin et al., 2014) is taken into account, the early magmatic phase (ca. 550–545 Ma) is recorded in the cTAM, BG and DV areas, and nVL, and only slightly later in the Queen Maude Mountains (ca. 535 Ma; Paulsen et al., 2013). Even if magmatism *commenced* at approximately the same time (ca. 550 Ma, and slightly earlier in the KG area) along much of the arc, the major pulses of compositionally distinct magmatism clearly occurred at different times within the sVL segment.

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400 *5.3 Tectonic Controls on Magmatic Variability*

401 The significant variability in the timing and composition of magmatism across sVL most
402 likely reflects differences in both magma source *and* tectonic processes, both spatially (along
403 500 km of the sVL margin) and temporally (from ca. 565–490 Ma). Alkaline magmatism at
404 convergent margins is typically associated with post-orogenic (PO) stages. The distinction of
405 the alkaline PO magmatism from the preceding subduction- and/or collision-related
406 magmatism may be due to relatively high-temperature melting of lithospheric mantle that
407 was metasomatically enriched during subduction, followed by increasing primitive mantle
408 components (Bonin et al., 1998). However, there are an increasing number of accounts of
409 alkaline and carbonatitic magmatism within active subduction zone settings (e.g., Abratis and
410 Wörner, 2001; Carmichael et al., 1996; D’Orazio et al., 2007; Petrone et al. 2003).

411 Cooper et al. (1997), Cottle and Cooper (2006a), Mellish et al. (2002), and Read et al.
412 (2002) suggested that the alkaline magmatism in the KG area was confined to an extensional
413 or transtensional jog along the convergent margin and predated widespread calc-alkaline
414 magmatism in southern Victoria Land. The new geochronology and geochemistry data
415 presented in this study confirm that the alkaline and carbonatitic magmatism within the KG
416 area was long-lived (at least ca. 550–509 Ma) and was broadly contemporaneous with
417 subalkaline magmatism within the KG area, and with dominantly subalkaline magmatism in
418 the adjacent DV and BG areas. Formation of the alkaline magmas in an extensional jog
419 remains a viable hypothesis, although the occurrence of subalkaline magmatism within the
420 KG area and A-type granitoids in the vicinity of Darwin Glacier (Fig. A5; Foggydog suite of
421 Simpson and Cooper, 2002) preclude a clearly delineated extensional province with

compositionally restricted magmatism. The subalkaline magmatism in the BG is demonstrably subduction-related, but the invariably low Sr/Y of these rocks indicates melt generation and differentiation at relatively shallow depths (above the garnet-in and plagioclase-out reactions; Tulloch and Kimbrough, 2003), consistent with an extensional deformation regime in the overriding plate. The high Sr/Y of the younger DV subduction-related magmatism suggests that these melts were generated at deeper levels (below the garnet-in and plagioclase-out reactions; Tulloch and Kimbrough, 2003) during crustal thickening and/or underthrusting.

Our preferred model for the tectono-magmatic evolution of the Ross orogen in sVL (Fig. 12) unifies our observations, components of previously established models for the Ross orogeny, and well-established models for accretionary orogens in general. Collins (2002a; 2002b) recognized that long-lived accretionary orogens are characterized by “tectonic switching”, i.e., repeated alternation between contraction and extension of the overriding plate. Kemp et al. (2009) identified multiple phases of alternating extension and contraction within the Australian Tasmanides—the northward (in present day coordinates) extension of the Paleozoic active margin along East Antarctica—marked by secular trends in the Nd, Hf, and O isotope composition of igneous rocks and zircons. Johnson et al. (2016) interpreted one full sequence of extension-contraction-extension during the Delamerian phase of Tasmanide orogenesis, synchronous with the later stages (ca. 517–490 Ma) of the Ross orogeny in Antarctica.

The compositional evolution of igneous rocks in sVL seems to record one cycle of extension-contraction-extension of this segment of the Antarctic margin, following an earlier phase of pre-magmatic thickening and metamorphism in the upper plate ca. 590–570 Ma

(Hagen-Peter et al., 2016). The ca. 565–525 Ma igneous rocks exposed in sVL—mafic subalkaline and alkaline rocks in the KG area and low Sr/Y subduction-related rocks in the BG area—have compositions indicative of formation beneath/within an extending upper plate during subduction. We propose that the distinct compositions of the synchronous suites of rocks in these areas are due to differences in lithospheric thickness and the composition of the lithospheric mantle beneath the margin. In our model (Fig. 12A), the low Sr/Y subalkaline melts of the BG area were generated beneath relatively thin lithosphere, above the spinel lherzolite to garnet lherzolite transition at ~50–60-km depth (18–20 kbar at 1200°C; Klemme and O'Neill, 2000), and differentiated in thin crust above eclogite facies at ~30–40-km depth (also supported by the negative Eu anomalies in the primitive BG samples; Fig. 8). The alkaline silicate rocks and carbonatites were generated by low-degree partial melting within metasomatically enriched sub-continental lithospheric mantle (SCLM). The SCLM has been invoked as a potential source of alkaline silicate rocks and associated carbonatites in many localities (e.g., Arzamastsev et al., 2001; Bailey, 1987; Downes et al., 2005; Serri et al., 1993), and also as the source of late-stage lamprophyres and potassic granitoids in the Ross orogen (Rocchi et al., 2009). We also consider the SCLM to be the likely source of the alkaline and carbonatitic rocks in the KG area because: 1) The strongly enriched trace element compositions of the primitive silicate rocks and carbonatites are at odds with a depleted asthenospheric mantle source; 2) The very low degrees of partial melting required to extract highly alkaline and carbonatitic melts from primitive or depleted asthenospheric mantle are considered unlikely for a subduction zone setting; 3) The persistence of alkaline magmatism for ca. 50 million years requires a static source not in the convecting asthenospheric mantle (assuming a down-dip velocity of 20–40 km/Myr for

asthenosphere convecting through corner flow; Turner and Hawkesworth, 1997). Exposed rocks within the ca. 565–525 Ma age range are very sparse in the DV area, perhaps due to highly oblique subduction beneath a stepped margin (Fig.12). The step-like geometry of the margin in sVL had been previously recognized by several studies (e.g., Goodge and Dallmeyer, 1996; Stump, 1992; Stump et al., 2006), as has the inference of oblique subduction (Encarnación and Grunow, 1996; Goodge et al. 1993; Goodge and Dallmeyer, 1996; Stump et al., 2006).

The younger intrusions in sVL—the high Sr/Y granitoids in the DV and KG areas and small leucogranite dikes in the BG area—are consistent with derivation beneath/within a thickened upper plate. Thickening of the overriding plate during early magmatism in the DV area is also supported by Hf isotope data presented by Hagen-Peter et al. (2015). The onset of widespread magmatism within the DV area may be due to a reorientation of the subduction vector relative to the margin, i.e., the establishment of a subduction vector at a higher angle to the DV margin (Fig. 12B; Encarnación and Grunow, 1996; Stump et al., 2006). Contraction in the overriding plate and a reorientation in the subduction vector may have been due to the docking of the “Beardmore Terrane” to the south of the Byrd Glacier ca. 525 Ma (Stump et al., 2006). Although our model does not strictly rely on this hypothesized collisional event, the waning of voluminous magmatism in the BG area and alkaline magmatism in the KGAS (only one calcite-bearing nepheline syenite in the KG area is younger than 525 Ma), is consistent with such a model. The latest phase of magmatism in sVL (~505–492 Ma) is characterized by the emplacement of alkali calcic granitoids and mafic dikes, including lamprophyres, at shallow crustal levels (Allibone et al., 1993a; 1993b; Encarnación and Grunow, 1996; Hagen-Peter et al., 2015)—representing extensional post-

orogenic magmatism that is recorded at roughly the same time along the entire arc (Hagen-Peter et al., 2015, and references therein). At this time, either convergence between the Pacific Plate and Gondwana waned, or the accretionary margin moved outboard, perhaps due to the docking of the Beardmore Terrane and other continental fragments along the margin (e.g., Rocchi et al., 2011).

6. Conclusions

Igneous rocks with grossly similar petrographic and geochemical compositions occur along a ~500 km segment of the Ross orogen in southern Victoria Land. Large complexes of dominantly subalkaline granitoids in the Dry Valleys and Byrd Glacier area bound a suite of small-volume alkaline silicate rocks and carbonatites, more commonly associated world-wide with intraplate and continental-rift tectonic settings. The onset and duration of magmatism was transient along the arc—with relatively large subalkaline complexes emplaced in the DV and BG areas over a similar duration (~25 Myr) but during different periods—and new geochronology confirms that the alkaline and carbonatitic magmatism was at least partially contemporaneous with the large-scale subduction-related magmatism. Despite the significant petrologic and major-element differences between rocks, the most primitive samples from each area have broadly similar and strongly enriched trace element compositions, suggesting some commonalities in the magma sources for each region. The gross petrologic, geochemical, and geochronologic differences in the magmatism can be explained by alternating phases of extension and contraction in the overriding plate. Whereas many phases of “tectonic switching” are recorded in other accretionary orogens, only one phase of extension-contraction-extension seems to be recorded in the sVL segment of the Ross

orogen, perhaps due to docking of a microcontinent and waning of convergence along this segment of the Gondwana margin.

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Fig. 1. Simplified geologic map of metamorphic and igneous rocks of the Ross orogen, exposed in the Transantarctic Mountains (modified from Goodge, 2007). Igneous rocks of the Granite Harbour Intrusives (GHI) were emplaced into supracrustal metasedimentary rocks along the Antarctic margin. The only exposed contacts between GHI and Archean and Proterozoic basement rocks are in the central Transantarctic Mountains.

Fig. 2. Satellite image of southern Victoria Land showing the division of the DV, KG, and BG areas. Thick dashed lines delineate the boundaries between these areas but do not imply strict geologic or physiographic boundaries. The distribution of samples from this study is shown by stars of different colors. Multiple samples were collected at many of the sites, and so the actual number of samples exceeds the number of stars. The four outlined areas are enlarged in geologic maps in Figs. A1, A3, A5 and A6.

Fig. 3. Field photos of representative rock types from each area (Dry Valleys area: A–C; Koettlitz Glacier area: D–F; Darwin/Byrd Glacier area: G–L). (A) Migmatitic paragneiss of the Skelton Group pervasively intruded by leucogranite dikes and sills. (B) Augen gneiss fabric in a texturally-old component of the Bonney Pluton. (C) Porphyritic diorite enclaves within an undeformed granite. (D) Compositional layering in a biotite orthogneiss. (E) Tapered nepheline syenite dike cross-cutting marble. (F) Very coarse-grained calcite (Cal)-bearing nepheline (Ne) syenite dike cutting marble. (G) Approximately 100 meter-thick sill of granitic orthogneiss beneath layered gneisses of the Horney Formation. (H) Boudinaged pod of fine-grained hornblende diorite within grey orthogneiss. Both contain leucogranite dikes and sills. (I) Large pod of gabbroic-diorite cross-cut by a sub-vertical dike of granitic orthogneiss and networks of leucogranite dikes. (J) Megacrystic K-feldspar granite

intermingled with a coarse leucocratic granite with a layer of migmatite. (K) Interlayering of orthogneiss and migmatitic paragneiss. (L) Pyroxene-biotite enclaves within a coarse granite.

Fig. 4. Photomicrographs of igneous rocks from the Koettlitz and Darwin/Byrd Glacier areas. Both plane-polarized (left) and cross-polarized (right) photomicrographs are shown for each sample. (A; RG11-180) Calcite (Cal) intergrown with potassium feldspar (Kfs) and nepheline (Ne) in a Cal-bearing Ne syenite. (B; RG11-212) Sigmoidal quartz (Qtz) porphyroclasts in an extensively recrystallized biotite (Bt) + garnet + muscovite (Ms) orthogneiss. (C; RG11-46) Altered clinopyroxene (Cpx) and orthopyroxene (Opx) intergrown with plagioclase (Pl) and minor amphibole. (D; RG11-257) Igneous fluorite (Fl) in a leucogranite. (E; DG12-118) Granitic orthogneiss with feldspar (Fsp) augen, recrystallized Qtz bands, and aligned Bt. (F; DG12-152) Undeformed equigranular granite with zoned Pl. (G; DG12-121) Garnet (Gt)-bearing leucogranite. (H; DG12-230) Cpx partially replaced by an elongate amphibole (Am) grain in an orthogneiss. (I; DG12-117) Large plagioclase porphyroblast with inclusions of Cpx, Am, and Bt. (J; DG12-14) Crystallographically coherent Cpx crystal surrounding Bt in a gabbro (Ol = olivine). All scale bars are 1 mm unless otherwise noted.

Fig. 5. (A) Total alkalis versus silica (TAS) and (B) modified alkali-lime index (MALI; Frost et al., 2001) diagrams demonstrate the strong alkali enrichment in some rocks from the KGAS. Rocks from the DV and BG areas are dominantly subalkaline and calc-alkalic or calcic. (C) Discrimination of samples with >60 wt% SiO₂ based on Shand's index. Color-filled fields outline previously published data for the different areas from the following references: **DV:** Allibone et al., 1993b; Cox et al., 2000; Forsyth et al., 2002; Smillie, 1992; **KG:** Aslund, 1990; Cottle and Cooper, 2006a; Hall et al., 1995; Mellish et al., 2002; Read et al., 2002; Rowell et al., 1993; **BG:** Cottle and Cooper, 2006b; Simpson and Cooper, 2002.

Fig. 6. Geochemical discrimination diagrams for samples from the DV, KG, and BG areas. Igneous rocks from each area span the range from gabbro to granite (*sensu lato*). Significant dispersion of major- and trace-element concentrations at a given SiO₂ precludes simple fractional crystallization/melting cogenetic relationships for the igneous rocks in each area. Major oxide concentrations are in wt%, and trace element concentrations are in ppm.

Fig. 7. Tectonic discrimination diagrams for rocks with >57 wt% SiO₂. Igneous rocks from the KG area commonly fall within the A-type granite and within-plate granite fields. Most samples from the DV and BG areas fall within or near the boundaries of the I- and S-type granites and volcanic arc granite fields.

Fig. 8. Trace element and rare earth element (REE) patterns of the most primitive silicate rocks (SiO₂ <52 wt%; Mg# >69) and carbonatites normalized to primitive mantle and the chondritic uniform reservoir, respectively (Sun and McDonough, 1989). All of the primitive samples show strong enrichments in the large ion lithophile elements (LILE) relative to high field strength elements and light REE relative to heavy REE.

Fig. 9. Cathodoluminescence (CL) images of representative zircons from different areas. Circles show the locations of laser pits for U-Th-Pb isotope analyses. For scale, the laser pits are ~24 μm in diameter. The numbers next to the circles are the $^{206}\text{Pb}/^{238}\text{U}$ dates (in Ma) for the corresponding analysis. Propagated uncertainties on $^{206}\text{Pb}/^{238}\text{U}$ dates are typically ± 7 Myr (full results provided in Table S2).

Fig. 10. Examples of inverse concordia diagrams and $^{206}\text{Pb}/^{238}\text{U}$ date distributions for representative samples from each area. These examples demonstrate the effects of ancient (P68770, P68851, RG11-137, DG12-160) and “young” (P68851, P62293, RG11-85) inheritance, the presence of common Pb, and Pb loss. The effects of these processes on the U-Pb systematics are annotated on several of the diagrams. Blue ellipses represent analyses used in the calculation of weighted-mean $^{206}\text{Pb}/^{238}\text{U}$ dates. Red ellipses represent analyses omitted from the weighted-mean dates.

Fig. 11. Timeline summarizing the zircon U-Pb age data for the Granite Harbour Intrusives along the Ross orogen. Closed symbols are from this study. Open symbols and grey bars are previously published data from the following sources: **northern Victoria Land:** Black and Sheraton, 1990; Bomparola et al., 2006; Fioretti et al., 2005; Rocchi et al., 2015; Tiepolo and Tribuzio, 2008; **central Victoria Land:** Rocchi et al., 2009; **southern Victoria Land:** Allibone and Wysoczanski, 2002; Cooper et al., 1997; Cottle and Cooper, 2006a,b; Cox et al., 2000; Encarnación and Grunow, 1996; Hall et al., 1995; Martin et al., 2014; Mellish et al., 2002; Read et al., 2002; Read, 2010; Rowell et al., 1993; Simpson, 2002; **central Transantarctic Mountains:** summarized in Goodge et al., 2012; **Queen Maud Mountains:** Paulsen et al., 2013.

Fig. 12. Generalized model for the tectono-magmatic evolution of the Ross orogen in sVL shown in map view, perspective block diagrams, and arc-parallel cross sections (not-to-scale). The stages of the model are summarized in the bulleted lists and explained in detail in the text. A final stage of post-orogenic magmatism (ca. 505–492 Ma) is not depicted in the diagram.

Fig. A1. Simplified geologic map of the Granite Harbour Intrusives in the Dry Valleys area (modified from Cox et al., 2012). Metamorphic and cover units are removed to emphasize the distribution of igneous rocks. The colors correspond to different magmatic suites. The circles with letters show the locations of the field photographs shown in Figures 3 and A2.

Fig. A2. Additional field photos from the Dry Valleys area. (a) Massive equigranular granite of the Hedley Pluton with xenoliths of coarser, gneissic granite from the Bonney Pluton. (b) Undeformed equigranular granite from the DV2 suite. Photos provided by GNS Science (New Zealand).

Fig. A3. Geologic map of the KG area in the vicinity of the Walcott Glacier (modified from Cox et al., 2012). Alkaline mafic rocks, syenite, A-type granite, and carbonatite, as well as sub-alkaline orthogneiss and granite, were emplaced into metasedimentary rocks of the Skelton Group. Approximately 250 samples were collected in this area; the stars mark locations of samples for which geochemical, geochronologic, and/or isotopic data are presented. The circles with letters show the locations of the field photographs shown in Figures 3 and A4.

Fig. A4. Additional field photos from the Koettlitz Glacier area. (a) Gneissic banding in a granite cut by a several generations of leucocratic dikes. (b) Grey orthogneiss intruded by leucogranite dikes and sills. (c) Two-mica orthogneiss with strained Z-shaped garnets. (d) Coarse-grained gabbroic diorite cut by 2–3 cm-wide leucogranite sills and ~30 cm wide granite dike. A ~10 cm-wide layer of fine-grained diorite in between two leucogranite sills is apparently offset across the granite dike. (e) Very coarse calcite and amphibole grains in a carbonatite plug. (f) Gneissic A-type granite.

Fig. A5. Simplified geologic map of the Darwin Glacier area (modified from Carosi et al., 2007; Cottle, 2002; Simpson and Cooper, 2002). Some granitic and granodioritic rocks, including orthogneisses with variable fabric development are grouped together. Large areas of diorite and gabbroic diorite are mapped as separate, predominantly mafic units. Stars show the locations of samples with geochemistry, geochronologic, and/or isotopic data from this study. Multiple samples were collected in the vicinity of the stars on the map, but only a single star is plotted for clarity. Letters show the locations of corresponding field photographs shown in Figs. 3 and A7.

Fig. A6. Simplified geologic map of the Byrd Glacier area (modified from Carosi et al., 2007). All plutonic rocks, including megacrystic granites, orthogneisses, and mafic rocks are grouped together at the scale of this map. Stars show the locations of samples with geochemistry, geochronologic, and/or isotopic data from this study. Multiple samples were collected in the vicinity of the stars on the map, but only a single star is plotted for clarity. Letters show the locations of corresponding field photographs shown in Figures 3 and A8.

Fig. A7. Additional field photos from the Darwin Glacier area. (a) Contact between leucocratic and grey hornblende (Hbl) orthogneisses, both cross-cut by leucogranite dikes. (b) Centimeter- to meter-scale compositional zoning in an orthogneiss with feldspar augen. (c) Grey Hbl orthogneiss with abundant strained mafic enclaves in contact with a leucocratic orthogneiss cut by a domino-boudinaged leucogranite dike. (d) High density of strained mafic enclaves in coarse-grained granite. (e) Strained mafic enclaves showing stretching > flattening strain. (f) Variation of grain size within a gabbro. (g) Gabbro cross-cut by a ~1.5-meter-wide dike of biotite-garnet granite, in-turn cross-cut by a garnet-bearing pegmatitic

leucogranite. (e) Three parallel lamprophyre dikes cross-cutting a very-coarse-grained biotite orthogneiss. (i) Pyroxenite dike cross-cutting biotite orthogneiss.

Fig. A8. Additional field photos from the Byrd Glacier area. (a) Boudinaged lens of migmatitic paragneiss (Horney Formation) enveloped in granitic orthogneiss. (b) Large body of granitic orthogneiss intruded by a Ferrar dolerite sill. (c) Anastomosing dikes of leucogranite cross-cutting diorite. (d) Intermingling of coarse granite and diorite.

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Figure 1

Figure 2

Figure 3

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Figure 5

Figure 6

Figure 7

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Figure 9

Figure 10

Figure 11

Figure 12

A) ca. 550–525 Ma: upper-plate extension

- hinge rollback and extension
- generation of alkaline melts in SCLM beneath KG area
- shallow extraction and differentiation of magmas in BG area
- sparse magmatism in DV area

B) ca. 525–505 Ma: upper-plate thickening

- hinge advance and thickening
- reorientation of subduction vector
- mostly subalkaline magmatism KG area
- voluminous subalkaline magmatism in DV area
- sparse crustal magmatism in BG area

Fig. A1. Simplified geologic map of the Granite Harbour Intrusives in the Dry Valleys area (modified from Cox et al., 2012). Metamorphic and cover units are removed to emphasize the distribution of igneous rocks. The colors correspond to different magmatic suites. Circles with letters indicate the locations of field photos shown in Figures 3 (capital letters) and A2 (lower-case letters).

Fig. A2. Additional field photos from the Dry Valleys area. (a) Massive equigranular granite of the Hedley Pluton with xenoliths of coarser, gneissic granite from the Bonney Pluton. (b) Undeformed equigranular granite from the DV2 suite.

Fig. A3. Geologic map of the KGAS in the Walcott Glacier area (modified from Cox et al., 2012). Alkaline mafic rocks, syenite, A-type granite, and carbonatite, as well as sub-alkaline orthogneiss and granite, were emplaced into metasedimentary rocks of the Skelton Group. Approximately 250 samples were collected in this area; the stars mark locations of samples for which geochemical, geochronologic, and/or isotopic data are presented. The circles with letters show the locations of the field photographs in Figures 3 (capital letters) and A4 (lower-case letters).

Fig. A4. Additional field photos from the Koettlitz Glacier area. (a) Gneissic banding in a granite cut by a several generations of leucocratic dikes. (b) Grey orthogneiss intruded by leucogranite dikes and sills. (c) Two-mica orthogneiss with strained Z-shaped garnets. (d) Coarse-grained gabbroic diorite cut 2–3 cm-wide leucogranite sills and ~30 cm wide granite dike. A ~10 cm-wide layer of fine-grained diorite in between two leucogranite sills is apparently offset across the granite dike. (e) Very coarse calcite and amphibole grains in a carbonatite plug. (f) Gneissic A-type granite.

Fig. A5. Simplified geologic map of the Darwin Glacier area (modified from Carosi et al., 2007; Cottle, 2002; Simpson and Cooper, 2002). Granitic and granodiorite rocks, including orthogneisses with variable fabric development are grouped together. Large areas of diorite and gabbroic diorite are mapped as separate, predominantly-mafic units. Stars show the locations of samples with geochemistry, geochronologic, and/or isotopic data from this study. Multiple samples were collected in the vicinity of the stars on the map, but only a single star is plotted for clarity. Letters indicate the locations of field photographs shown in Figures 3 (capital letters) and A7 (lower-case letters).

Fig. A6. Simplified geologic map of the Byrd Glacier area (modified from Carosi et al., 2007). All plutonic rocks, including megacrystic granites, orthogneisses, and mafic rocks are grouped together at the scale of this map. Stars show the locations of samples with geochemistry, geochronologic, and/or isotopic data from this study. Multiple samples were collected in the vicinity of the stars on the map, but only a single star is plotted for clarity. Letters indicate the locations of field photographs shown in Figures 3 (capital letters) and A8 (lower-case letters).

Fig. A7. Additional field photos from the Darwin Glacier area. (a) Contact between leucocratic and grey hornblende (Hbl) orthogneisses, both cross-cut by leucogranite dikes. (b) Centimeter- to meter-scale compositional zoning in an orthogneiss with feldspar augen. (c) Grey Hbl orthogneiss with abundant strained mafic enclaves in contact with a leucocratic orthogneiss cut by a domino-boudinaged leucogranite dike. (d) High density of strained mafic enclaves in coarse-grained granite. (e) Strained mafic enclaves showing stretching > flattening strain. (f) Variation of grain size within a gabbro. (g) Gabbro cross-cut by a ~1.5-meter-wide dike of biotite-garnet granite, in-turn cross-cut by a garnet-bearing pegmatitic leucogranite. (e) Three parallel lamprophyre dikes cross-cutting a very-coarse-grained biotite orthogneiss. (i) Pyroxenite dike cross-cutting biotite orthogneiss.

Fig. A8. Additional field photos from the Byrd Glacier area. (a) Boudinaged lens of migmatitic paragneiss (Horney Formation) enveloped in granitic orthogneiss. (b) Large body of granitic orthogneiss intruded by a Ferrar dolerite sill. (c) Anastomosing dikes of leucogranite cross-cutting diorite. (d) Intermingling of coarse granite and diorite.

Table 1

Summary of samples and U-Pb results

Sample name	Rock description	Location		U-Pb age ^a (Ma)	± ^b (Ma)	<i>n</i>	MSWD ^c
		Lat (°S)	Long (°E)				
<i><u>Dry Valleys area</u></i>							
RG11-233	Hbl diorite	-78.073	163.895	492.2	6.1	27	1.9
P64422 ^d	Hbl diorite	-77.482	162.821	493.2	6.1	35	0.8
P68826	pink granite	-77.877	162.636	495.5	6.1	25	1.2
P62101	granite	-77.231	162.067	495.6	6.1	20	1.3
RG11-241	diorite	-78.161	163.647	495.9	6.1	23	1.5
P68775	granite	-76.943	162.161	495.9	6.2	30	2.0
P68845	Hbl granite	-76.953	162.247	496.1	6.1	26	1.2
P68844	Hbl granite	-76.952	162.267	496.8	6.1	28	1.0
P62293	megacrystic granite	-77.369	163.007	496.9	6.2	22	1.7
P68802	pink granite	-76.810	162.443	497.1	6.2	29	1.4
RG11-234	cpx-bt gabbro-diorite	-78.060	163.849	498.6	6.5	13	2.6
P68760	granite	-76.934	162.473	500.1	6.6	18	1.6
RG11-224	Bl diorite	-78.078	163.867	500.2	6.1	26	1.4
P49945	Bt granodiorite	-77.556	161.858	500.6	6.6	17	2.1
P64403	Hbl gabbro	-77.329	162.815	501.9	6.4	20	1.6
P68796	pink granite	-76.802	162.157	502.1	6.6	23	2.0
P68766	granite	-76.861	162.551	502.6	6.6	10	0.5
P68843	porphyritic granite	-76.950	162.292	504.2	6.4	19	0.9
P68770	weakly-foliated granite	-76.892	162.248	504.8	7.0	14	2.8
P68878	porphyritic diorite dike	-76.804	162.135	505	6.5	18	1.5
P68851	megacrystic Hbl granite	-76.903	162.196	505.5	6.5	16	1.5
P68782	weakly-foliated granite	-76.895	162.395	506.4	6.4	16	1.0
P68861	massive Hbl diorite	-76.884	162.485	507.8	6.4	20	1.5
P49944	Qtz monzodiorite	-77.569	161.628	508.9	6.4	23	1.3
P68860	foliated Hbl tonalite	-76.884	162.485	509.4	6.6	17	1.9
P62321	foliated diorite	-77.533	162.962	509.5	6.3	20	1.1
P68787	Hbl monzodiorite	-76.866	162.349	511	6.4	28	2.2
P68756	megacrystic granite	-76.944	162.335	514.5	6.7	10	0.7
P68758	granite	-76.939	162.396	515.3	6.4	11	1.0
P62324	foliated diorite	-77.535	162.900	515.4	6.6	28	1.3

Sample name	Rock description	Location		U-Pb age ^a (Ma)	$\pm^b$ (Ma)	<i>n</i>	MSWD ^c
		Lat (°S)	Long (°E)				

Koettiltz Glacier area

RG11-51	zoned dike edge	-78.256	162.880	499.4	6.4	16	1.9
RG11-44	gabbro/diorite	-78.255	162.877	500.3	6.2	20	2.5
RG11-222	felsic porphyry sill	-78.266	162.598	500.5	6.6	5	1.2
RG11-85	Ne + Cal syenite	-78.198	163.143	509.1	6.4	16	1.6
RG11-50	zoned dike core	-78.256	162.880	510.1	8.8	4	2.4
RG11-70	granite	-78.255	163.237	518.2	6.4	16	1.3
RG11-71	granite	-78.255	163.237	522.3	6.7	11	1.2
RG11-121	Bt granite	-78.255	163.942	528.4	6.8	13	1.2
RG11-160	gabbro/diorite	-78.273	162.711	541.7	6.7	24	2.3
RG11-141	Hbl granodiorite	-78.283	162.792	543.7	7.1	8	0.6
RG11-183	granitoid	-78.270	162.755	544.3	6.7	9	0.9
RG11-202	Qtz syenite	-78.300	162.739	549.5	7.1	15	1.3
RG11-117	Hbl diorite	-78.372	163.066	551	6.7	27	1.1
RG11-24	orthogneiss	-78.217	163.067	552	7.0	7	0.9
RG11-119	leucogranite dike	-78.372	163.066	553.2	8.1	7	2.3
RG11-146	gabbro	-78.284	162.795	561.9	6.8	24	1.4
RG11-212?-mica + Grt orthogneiss		-78.315	162.799	564.7	7.0	12	1.6
RG11-137	orthogneiss	-78.199	163.005	~540	—	—	—

Byrd Glacier area

DG12-112	Grt granite dike	-80.469	156.142	502.8	6.2	23	2.2
DG12-25	leucogranite dike	-80.452	156.321	~510	—	—	—
DG12-23	gabbroic diorite	-80.452	156.321	524.9	6.9	16	1.6
DG12-72	Bt granite dike	-80.442	156.239	529.9	6.9	9	0.7
DG12-210	leucogranite sill	-80.049	159.448	531.4	8.0	3	0.3
DG12-199	Hbl granodiorite	-80.449	156.730	531.7	8.6	4	1.3
DG12-77	massive granite	-80.442	156.244	535.5	7.4	3	0.9
DG12-117	gabbro	-79.874	159.040	537.4	6.6	27	1.1
DG12-118	orthogneiss	-79.874	159.040	539.2	6.7	27	1.6
DG12-13	megacrystic granitoid	-80.479	156.371	~540	6.5	—	—
DG12-182	Hbl orthogneiss	-79.923	158.325	540.6	6.9	26	3.1

Sample name	Rock description	Location		U-Pb age ^a (Ma)	$\pm^b$ (Ma)	<i>n</i>	MSWD ^c
		Lat (°S)	Long (°E)				
DG12-178	gabbro	-79.931	158.791	541.4	6.7	32	0.6
DG12-127	gabbro	-79.862	158.976	541.6	6.6	29	0.7
DG12-228	orthogneiss	-79.899	159.131	542.6	7.2	24	2.1
DG12-152	granitoid	-79.478	160.971	543.8	7.2	5	0.7
DG12-230	gabbroic diorite	-79.899	159.131	544.8	7.0	14	1.9
DG12-156	orthogneiss	-79.754	158.879	545.7	7.2	5	0.2
CM-1	orthogneiss	-79.575	160.150	546.0	6.9	12	0.7
DG12-160	orthogneiss	-79.922	158.694	547.9	6.7	23	0.8
DG12-257	gabbro	-79.862	158.984	548.3	7.0	10	0.7
DG12-145	orthogneiss	-79.185	160.406	549.7	7.5	5	1.0
DG12-132	orthogneiss	-79.787	158.536	~550	—	—	—

^a "U-Pb ages" with a corresponding uncertainty are weighted-mean $^{206}\text{Pb}/^{238}\text{U}$ mean ages. Ages preceded by "~" are estimated ages of samples for which mean ages were not calculated.

^b Uncertainties on mean ages are quadratic additions of the internal 2 S.E. and external uncertainty estimated from the reproducibility of the GJ-1 and Plešovice reference zircons.

^c MSWD—mean square of weighted deviates.

^d Samples with prefix "P" are housed in the New Zealand Institute of Geological and Nuclear Science (GNS) PETLAB database (www.pet.gns.cri.nz).