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Authors	Paulsen, Timothy S.; Encarnación, John; Grunow, Anne M.; Stump, Edmund; Pecha, Mark; Valencia, Victor A.
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Correlation and Late-Stage Deformation of Liv Group Volcanics in the Ross-Delamerian Orogen, Antarctica, from New U-Pb Ages

Timothy S. Paulsen,^{1,*} John Encarnación,² Anne M. Grunow,³ Edmund Stump,⁴
Mark Pecha,⁵ and Victor A. Valencia⁶

1. Department of Geology, University of Wisconsin, Oshkosh, 800 Algoma Boulevard, Oshkosh, Wisconsin 54901, USA; 2. Department of Earth and Atmospheric Sciences, St. Louis University, 3242 Lindell Boulevard, O'Neil Hall 205, St. Louis, Missouri 63108, USA; 3. Byrd Polar Research Center, Ohio State University, 108 Scott Hall, 1090 Carmack Road, Columbus, Ohio 43210, USA; 4. School of Earth and Space Exploration, Arizona State University, Tempe, Arizona 85287, USA; 5. Department of Geosciences, University of Arizona, Gould-Simpson Building, 1040 East Fourth Street, Tucson, Arizona 85721, USA; 6. School of the Environment, Washington State University, Pullman, Washington 99164-2812, USA

ABSTRACT

We present new zircon U-Pb isotopic data for volcanic rocks from deformed metavolcanic-sedimentary successions of the widespread Ross Supergroup in the Queen Maud Mountains, Antarctica. Zircon U-Pb analyses of Liv Group volcanic rocks thought to be Neoproterozoic in age instead have early Paleozoic ages. Zircon U-Pb analyses of five samples assigned to the Fairweather Formation have yielded 524 ± 9 to 514 ± 9 Ma (2σ) crystallization ages, whereas six samples assigned to the Taylor Formation have yielded 510 ± 12 to 490 ± 6 Ma (2σ) crystallization ages. Although these ages imply that the Fairweather Formation is generally older than the Taylor Formation, the age uncertainties show a 17-My overlap that is consistent with previous suggestions for temporal correlation of these formations. On a regional scale, Liv Group volcanism overlapped with the emplacement of ~ 535 – 490 Ma plutonic rocks associated with the early Paleozoic Queen Maud batholith as well as igneous rocks found elsewhere along the early Paleozoic Pacific-Gondwana margin. Collectively, these igneous rocks provide plausible zircon sources for similar age detrital zircon populations found in outboard siliciclastic rocks belonging to the Leverett, Taylor, Fairweather, Greenlee, and Starshot Formations of the Queen Maud Mountains. The youngest crystallization age yielded by the deformed Taylor Formation (~ 490 Ma) assumes regional significance because it represents the youngest volcanic rock yet identified within the Ross orogen in Antarctica and provides important new evidence for latest Cambrian or younger deformation, possibly associated with orogenic collapse during slab rollback at the terminal stages of the Ross orogeny.

Online enhancements: supplemental table.

Introduction

The late Neoproterozoic–early Paleozoic Ross-Delamerian orogenic belt forms part of a greater collage of igneous, sedimentary, and metamorphic rocks that collectively comprise a subduction-related accretionary belt along Gondwana's paleo-Pacific margin known as the Terra Australis orogen (fig. 1; Boger 2011). Rocks of the Ross-Delamerian orogenic belt are presently exposed unconformably beneath unmetamorphosed Devonian-Jurassic sedimentary rocks of

the Beacon Supergroup (Barrett 1991) in the Transantarctic Mountains, a major rift flank mountain range located on the margin of the Mesozoic-Cenozoic West Antarctic Rift system (Fitzgerald et al. 1986). Igneous, sedimentary, and metamorphic rocks of the Ross orogen are of widespread interest because of their significance for understanding the tectonic history of Gondwana's paleo-Pacific margin (Stump 1992; Encarnación and Grunow 1996; Grunow et al. 1996b; Goodge 1997; Goodge et al. 1993, 2002, 2004a, 2004b; Cawood 2005; Boger 2011) and the possibility that they may offer insight into the Cambrian explosion of life, since the sediment flux caused by erosion of

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* Author for correspondence; email: paulsen@uwosh.edu.

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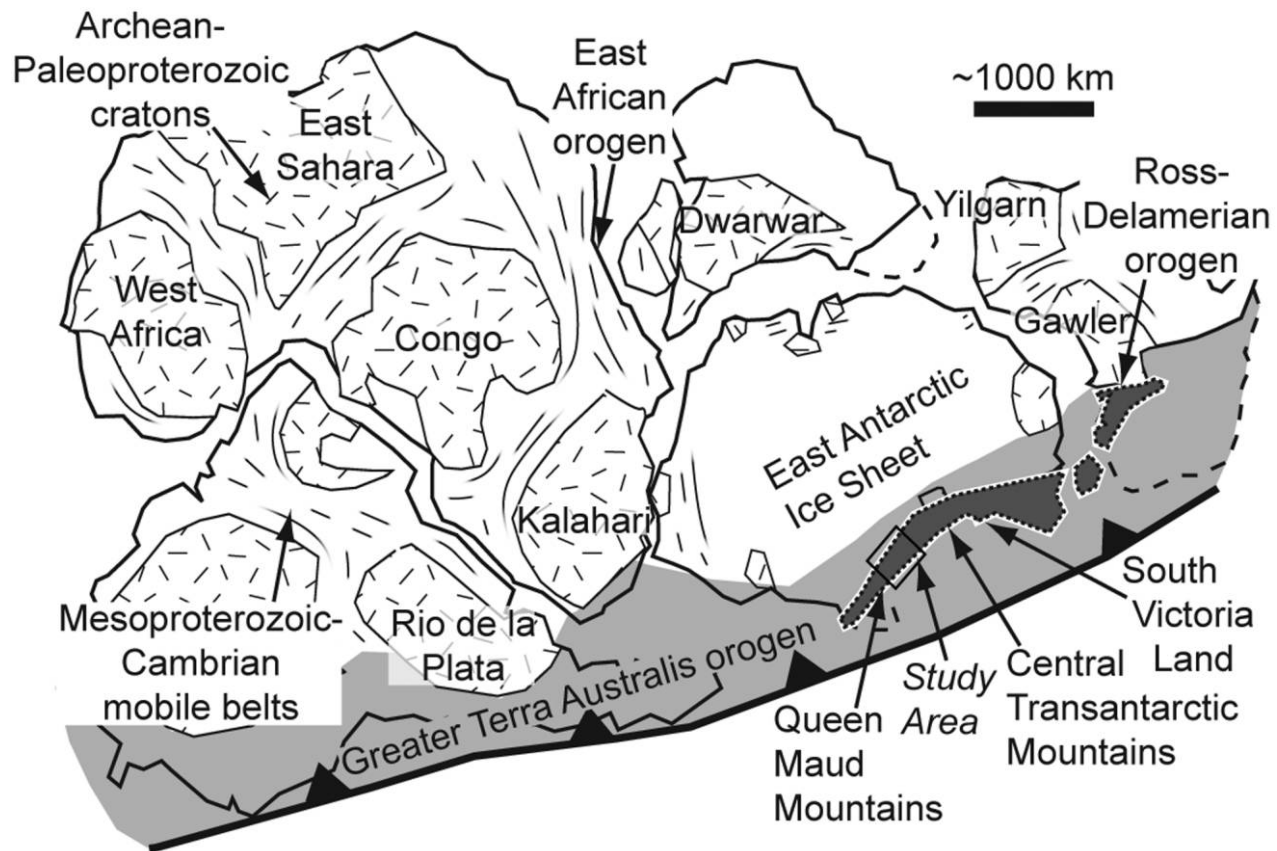


Figure 1. Gondwana reconstruction showing the Ross-Delamerian orogen within the greater latest Neoproterozoic to late Paleozoic Terra Australis orogen as well as the major Precambrian cratons and mobile belts of Gondwana. Figure modified from Paulsen et al. (2016b).

rocks of similar age is postulated to have seeded the oceans with the elements required for organisms to develop hard body parts (Squire et al. 2006). These rocks have been the subject of numerous field and petrographical studies (Stump 1995) as well as modern geochronologic and geochemical work (Encarnación and Grunow 1996; Goodge and Dallmeyer 1992, 1996; Goodge et al. 1993, 2001, 2002, 2004a, 2004b, 2012; Paulsen et al. 2013, 2015, 2016a, 2016b, 2017; Hagen-Peter et al. 2015, 2016; Hagen-Peter and Cottle 2016). Nevertheless, geochronologic studies have mostly been conducted on a reconnaissance basis throughout large sectors of the orogen, despite the fact that such data are important to help constrain the spatial-temporal patterns of magmatism related to the evolution of the Ross-Delamerian orogen (Borg and DePaolo 1991; Encarnación and Grunow 1996; Curtis et al. 2004; Squire and Wilson 2005; Goodge et al. 2012; Hagen-Peter et al. 2015; Hagen-Peter and Cottle 2016) and metamorphic-magmatic episodes associated with Gondwana assembly (Boger and Miller 2004; Cawood 2005; Boger 2011) and Iapetus rifting (Grunow et al. 1996b).

The intent of this article is to present new zircon U-Pb geochronologic data for volcanic rocks that outcrop along the Shackleton and Liv Glaciers of the Queen Maud Mountains to better constrain the timing, extent, and relationship of magmatism, deposition, and deformation in this remote sector of the Ross orogenic belt (fig. 1). Throughout this article, we use the 2015 International Chronostratigraphic Chart time scale (Cohen et al. 2013 [updated]).

Geology of the Queen Maud Mountains

Early work in the Queen Maud Mountains indicated that the Ross Supergroup contained at least two distinct tectonostratigraphic packages known as the Beardmore and Liv Groups (fig. 2; Stump 1995). The Beardmore Group was considered to have an older (Neoproterozoic) depositional age and includes greenschist facies siliciclastic rocks known as the La Gorce, Duncan, and Party Formations (Laird et al. 1971; Stump 1981, 1982, 1995). The Liv Group was considered to have a younger depositional age (Cambrian)

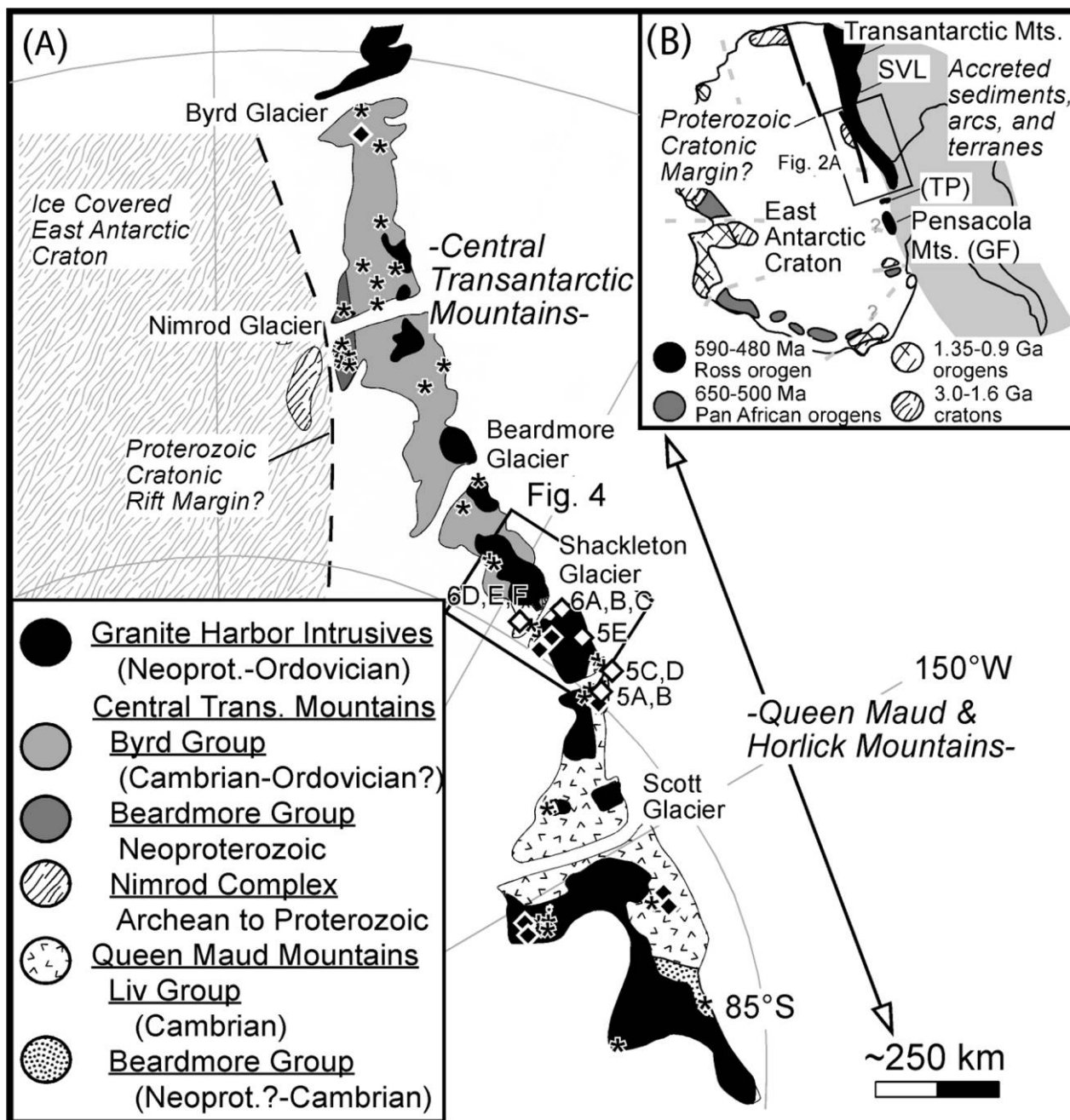


Figure 2. A, Simplified geologic map showing intrusive, sedimentary, and metamorphic basement rocks of the Ross orogen from the central Transantarctic Mountains to the south through the Queen Maud and Horlick Mountains. There are abundant volcanic rocks in early to middle Cambrian stratigraphic packages (i.e., the Liv Group) in the Queen Maud Mountains, whereas the early to middle Cambrian stratigraphic packages (i.e., the Byrd Group) in the central Transantarctic Mountains are practically devoid of volcanic rocks. The boundary separating the Liv Group from the Byrd Group occurs west of the Shackleton Glacier. Asterisks indicate approximate locations of detrital zircon samples analyzed by previous authors for the Queen Maud and central Transantarctic Mountains. Diamonds indicate approximate locations of volcanic samples analyzed in this article (white diamonds) and by previous authors (black diamonds) for the Queen Maud and central Transantarctic Mountains. Labels 5A-E and 6A-F correlate with localities of samples analyzed in this article and their age data shown in figures 5 and 6. B, Locality of the Transantarctic Mountains (approximately the black area of the Ross orogen) as well as the area of the figure with respect to major crustal age provinces and the trace of the possible Proterozoic rift margin of East Antarctica. Figure compiled from Stump (1982), Borg and DePaolo (1994), Goodge (2007), and Goodge and Finn (2010). GF = Gambacorta Formation; Mts. = mountains; Neoprot. = Neoproterozoic; SVL = south Victoria Land; TP = Thiel Mountains porphyry; Trans. = Transantarctic.

and contains volcanic and volcanoclastic rocks interbedded with carbonate and siliciclastic rocks (Rowell and Rees 1989) that include the Wyatt, Ackerman, Leverett, Taylor, Fairweather, and Greenlee Formations (Stump 1995; Rowell et al. 1997; Rowell and Rees 1989; Wareham et al. 2001). More recent paleontological and geochronologic studies (as in fig. 3) have yielded late Neoproterozoic to Cambrian ages for sedimentary and volcanic rocks of the Liv and Beardmore Groups, indicating that rocks belonging to both units are at least in part temporal correlatives to the Cambrian Byrd Group (Shackleton Limestone and Douglas and Starshot Formations) in the central Transantarctic Mountains (fig. 3).

Intrusions of the Queen Maud batholith (McGregor 1965; Murtaugh 1969; Borg et al. 1990; Stump 1995) invade Ediacaran to Cambrian metasedimentary and metavolcanic rocks of the Beardmore and Liv Groups in the Queen Maud Mountains. Batholith intrusions range from granite to gabbro in composition and have isotopic patterns indicative of derivation from mantle and Precambrian basement sources (Borg et al. 1990; Borg and DePaolo 1991, 1994). Early K-Ar and Rb-Sr geochronology suggested that some intrusions and volcanic country rocks of the Queen Maud batholith crystallized and cooled as early as the late Proterozoic (Faure et al. 1979; Felder and Faure 1979), but many of these older ages suffered from large errors, possible excess Ar, and open system behavior. More recent zircon U-Pb geochronology indicates that emplacement of intrusive and volcanic rocks primarily occurred after 535 Ma (Encarnación and Grunow 1996; Van Schmus et al. 1997; Paulsen et al. 2013), and ^{40}Ar - ^{39}Ar data indicate ~500–475 Ma metamorphic and igneous cooling (fig. 3; Grunow and Encarnación 2000a, 2000b; Paulsen et al. 2004, 2008, 2013).

Despite recent progress toward understanding the geochronology of tectonomagmatic events in the Queen Maud batholith, there are significant uncertainties about the timing of magmatism, deformation, and metamorphism in the broader Ross orogen because of the limited number of U-Pb age analyses over a relatively large region (Encarnación and Grunow 1996; Vogel et al. 2002). The paucity of age analyses is particularly problematic for understanding the correlation and tectonic significance of volcanic rocks hosted by the Queen Maud batholith. Although these volcanic rocks have been recognized since the early geologic exploration of the continental interior (McGregor 1965), their ages and correlation with each other and magmatic pulses along the Pacific-Gondwana margin (Encarnación and Grunow 1996; Goodge et al. 2012; Hagen-Peter and Cottle 2016; Hagen-Peter et al. 2016) remain poorly constrained.

The problems stemming from lack of age data are highlighted by U-Pb crystallization ages obtained from volcanic rocks of the Liv Group in the region of the Shackleton and Liv Glaciers (fig. 4). Van Schmus et al. (1997) reported 550 ± 15 and 516 ± 6 Ma zircon U-Pb crystallization ages for volcanic rocks of the Fairweather and Taylor Formations, respectively. The ~550 Ma crystallization age overlaps with the age of alkaline igneous rocks known from the south Victoria Land area of the Ross orogenic belt, such as the 551 ± 4 Ma age of syenite intrusions from the Skelton Glacier area (Rowell et al. 1993; Encarnación and Grunow 1996; Hagen-Peter and Cottle 2016) as well as the 545 ± 5 Ma intrusions from the central Transantarctic Mountains (Goodge et al. 1993, 2012). However, this ~550 Ma age for the Fairweather Formation apparently predates the oldest (~535 Ma) crystallization ages of rocks within the Queen Maud batholith (fig. 3). This older ~550 Ma age for the Fairweather Formation assumes regional tectonic significance in the Ross orogenic belt because it could represent the oldest volcanic rock recognized in the Ross-Delamerian orogenic belt outside of ~650 Ma rhyolite clasts dated in south Victoria Land as well as mafic volcanic rocks associated with a 668 Ma gabbro intrusion, which are postulated to be related to earlier late Neoproterozoic rifting (Goodge et al. 2002; Cooper et al. 2011). More recently, U-Pb age analyses of volcanic rocks of the Taylor Formation at Taylor Nunatak yielded a relatively more precise 505 ± 1.5 Ma crystallization age (Encarnación et al. 1999). This age overlaps crystallization ages of rocks within the Queen Maud batholith and depositional ages indicated by detrital zircon analyses of Liv Group sandstones (Paulsen et al. 2015, 2017), but it is the youngest volcanic rock recognized in the Ross-Delamerian orogenic belt outside of ~500 Ma volcanic rocks in the Thiel and Pensacola Mountains (the Thiel Mountains porphyry and the Gambacorta Formation in fig. 2b). This ~500 Ma age for the Taylor Formation is significantly younger than the 526 ± 2 to 516 ± 6 Ma crystallization ages (Encarnación and Grunow 1996; Van Schmus et al. 1997) yielded by other volcanic rocks of the Liv Group.

Apart from the three zircon ages from Liv Group samples mentioned above, there have been no other U-Pb age analyses of the volcanic successions of the Liv Group in the area from the Shackleton Glacier to the Liv Glacier. Here we provide new crystallization ages of 11 deformed volcanic rocks of the Fairweather, Taylor, and Greenlee Formations (fig. 4), including two samples acquired from the Polar Rock Repository (those with PRR numbers) and nine from other collections. The new ages establish important

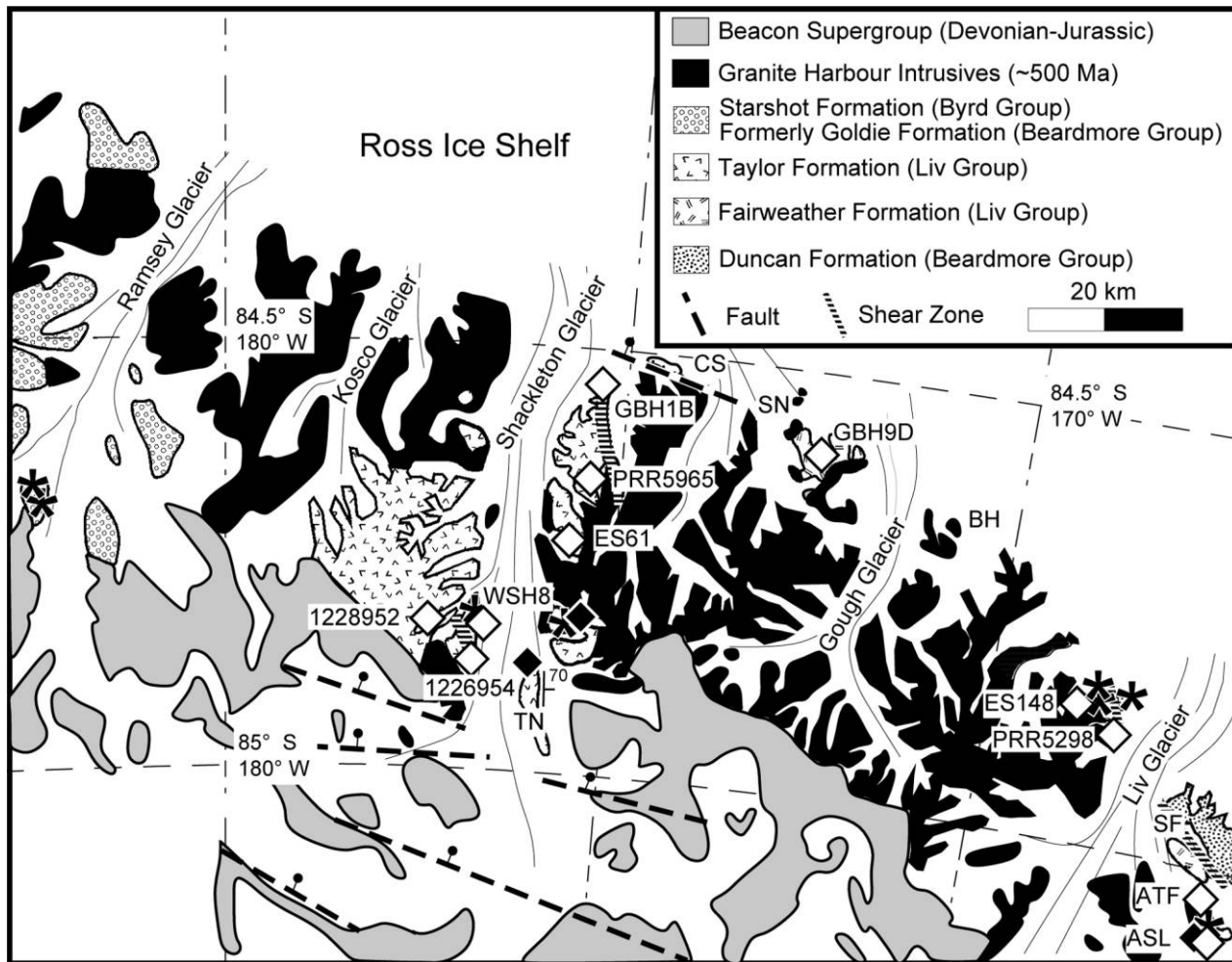


Figure 4. Simplified map showing the sample localities and geology in the Ramsey, Shackleton, and Liv Glacier areas (modified from Mirsky 1969). Diamonds show localities of volcanic rock samples analyzed herein (Fairweather Formation: ATF, ASL, PRR-5298, ES148, GBH9D; Taylor Formation: GBH1B, ES61, PRR-5965, 1226954, 1228952; Greenlee Formation: WSH8), and asterisks show locations of previous detrital zircon U-Pb age analyses reported in Paulsen et al. (2015, 2017). BH = Bravo Hills; CS = Cape Surprise; SN = Sage Nunataks; SF = Spillway Fault; TN = Taylor Nunatak.

correlations between major volcanic-volcaniclastic units in the area and provide limits to the timing of their deformation.

U-Pb Methods

Paulsen et al. (2008) describe the methods used for the separation and analysis of zircons from samples GBH9D and GBH1B. The methods used for the separation and analysis of zircons from the remaining nine samples are described below. Mineral concentrates were obtained from the samples by conventional mineral separation techniques. A split of these grains (generally 50–100 grains) was selected from the grains available and incorporated into a 1-inch epoxy mount together with fragments of Sri Lanka standard

zircon. The mounts were sanded to a depth of $\sim 20 \mu$, polished, imaged, and cleaned prior to isotopic analysis.

U-Pb geochronology of zircons was conducted by laser ablation multicollector inductively coupled plasma mass spectrometry at the Arizona Laser-Chron Center (Gehrels et al. 2008). The analyses involve ablation of zircon with a New Wave/Lambda Physik DUV193 excimer laser (operating at a wavelength of 193 nm) using a spot diameter of 35μ . The ablated material is carried in helium into the plasma source of a GV Instruments IsoProbe device, which is equipped with a flight tube of sufficient width that U, Th, and Pb isotopes are measured simultaneously. All measurements were made in static mode, using $10^{11}\text{-}\Omega$ Faraday detectors for ^{238}U , ^{232}Th , ^{208}Pb ,

and ^{206}Pb ; a $10^{12}\text{-}\Omega$ Faraday collector for ^{207}Pb ; and an ion-counting channel for ^{204}Pb . Ion yields are ~ 1.0 mV/ppm. Each analysis consisted of one 12-s integration on peaks with the laser off (for backgrounds), twelve 1-s integrations with the laser firing, and a 30-s delay to purge the previous sample and prepare for the next analysis. Ablation pits are ~ 12 μ in depth.

For each analysis, the errors in determining $^{206}\text{Pb}/^{238}\text{U}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ result in a measurement error of $\sim 1\%$ – 2% (at the 1σ level) in the $^{206}\text{Pb}/^{238}\text{U}$ age. The errors in measurement of $^{206}\text{Pb}/^{207}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ also result in $\sim 1\%$ – 2% uncertainty (at the 1σ level) in the $^{207}\text{Pb}/^{206}\text{Pb}$ age for grains that are older than 1000 Ma but are substantially larger for younger grains due to low intensity of the ^{207}Pb signal. Interpreted ages are based on $^{206}\text{Pb}/^{238}\text{U}$ for grains younger than 1000 Ma and on $^{207}\text{Pb}/^{206}\text{Pb}$ for grains older than 1000 Ma. This division at 1000 Ma results from the increasing uncertainty of $^{206}\text{Pb}/^{238}\text{U}$ ages and the decreasing uncertainty of $^{207}\text{Pb}/^{206}\text{Pb}$ ages as a function of age.

Common Pb correction is accomplished by using the measured ^{204}Pb and assuming an initial Pb composition from Stacey and Kramers (1975; with uncertainties of 1.0 for $^{206}\text{Pb}/^{204}\text{Pb}$ and 0.3 for $^{207}\text{Pb}/^{204}\text{Pb}$). Our measurement of ^{204}Pb is unaffected by the presence of ^{204}Hg because backgrounds are measured on peaks (thereby subtracting any background ^{204}Hg and ^{204}Pb) and because very little Hg is present in the argon gas.

Interelement fractionation of Pb/U is generally $\sim 20\%$, whereas apparent fractionation of Pb isotopes is generally $\sim 2\%$. In-run analysis (one or two every fifth measurement) of fragments of a large zircon crystal with a known age of 564.2 ± 3.2 Ma (2σ error; Gehrels et al. 2008) is used to correct for this fractionation. The uncertainty resulting from the calibration correction is generally 1% – 2% (2σ) for both $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ ages.

Concentrations of U and Th are calibrated relative to Sri Lanka zircon, which contains ~ 518 ppm of U and 68 ppm of Th.

The analytical data are reported in table S1 (available online). Uncertainties shown in this table are at the 1σ level and include only measurement errors. Analyses that are $>15\%$ discordant (by comparison of $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ages) or $>5\%$ reverse discordant (shown in italics in table S1) are not considered further.

The inferred age is determined from the $^{206}\text{Pb}/^{238}\text{U}$ age for analyses with a $^{206}\text{Pb}/^{238}\text{U}$ age <1000 Ma and from the $^{207}\text{Pb}/^{206}\text{Pb}$ age for analyses with a $^{206}\text{Pb}/^{238}\text{U}$ age >1000 Ma. The resulting interpreted ages are shown in figures 5 and 6 on Pb*/U concordia dia-

grams and weighted mean diagrams using the routines in Isoplot (Ludwig 2008). The weighted mean diagrams show the weighted mean (weighting according to the square of the internal uncertainties), the uncertainty of the weighted mean, the external (systematic) uncertainty that corresponds to the ages used, the final uncertainty of the age (determined by quadratic addition of the weighted mean and external uncertainties), and the MSWD of the data set. The systematic error includes contributions from the standard calibration, the age of the calibration standard, the composition of common Pb, and U decay constants and in this data set range from 1.2% to 2.3% (2σ).

U-Pb Results

The zircon grains yielded from all of these samples are subhedral to euhedral, with prismatic morphologies that display internal oscillatory, patchy, and unzoned domains, some of which surround xenocrystic cores. Ninety-nine percent ($n = 349$ of 351) of the U-Pb age analyses that meet acceptable concordance thresholds have U/Th ratios of <10 , suggesting that the zircons we analyzed primarily grew during igneous crystallization (Rubatto 2002; Hoskin and Schaltegger 2003), a result consistent with the presence of zircon grains with oscillatory zoned interiors and rational crystal faces (Corfu et al. 2003). The following presentation of the zircon U-Pb ages is separated by formation (the Fairweather Formation and the Taylor/Greenlee Formations).

Fairweather Formation. Seventy-four percent ($n = 153$ of 206) of the U-Pb age analyses of zircons from five samples of the Fairweather Formation (ATF, ASL, PRR-5298, ES148, and GBH9D) met acceptable concordance thresholds. The cumulative zircon age distribution from all of the samples ranges from 1555 Ma (Mesoproterozoic) to 423 Ma (Silurian).

Samples ATF and ASL were collected from localities to the east of the Liv Glacier at Mount Fairweather, where Stump (1985) reported interbedded volcanic and siliciclastic rocks belonging to the Fairweather Formation (fig. 4). Sample ATF is a metafelsite that comes from the 39-m-thick unit from the bottom of section 3 for the eastern ridge of Mount Fairweather. A weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 515 ± 9 Ma from a concordant set of analyses is considered to be the age of igneous crystallization for sample ATF (fig. 5A). One discordant zircon age analysis yields a 423 Ma age, which is younger than the rock's crystallization age and is presumably due

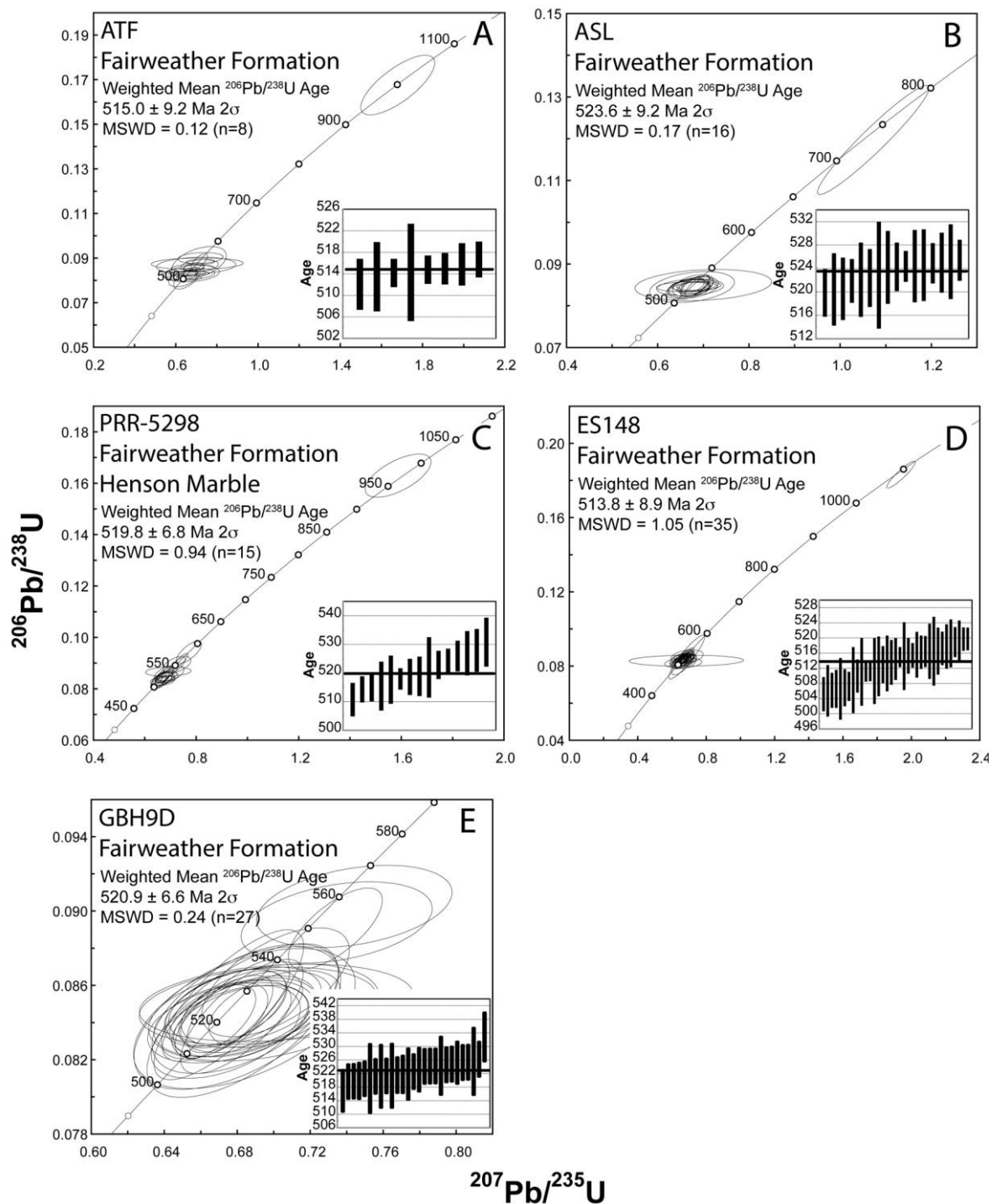


Figure 5. Concordia diagrams of zircon U-Pb data and weighted mean age plots of individual analyses for Fairweather Formation samples ATF (A), ASL (B), PRR-5298 (C), ES148 (D), and GBH9D (E). Error bars are 1σ from inferred U-Pb ages ($^{206}\text{Pb}/^{238}\text{U}$ age for samples younger than 1000 Ma and $^{207}\text{Pb}/^{206}\text{Pb}$ age for samples older than 1000 Ma). Data error ellipses are 2σ .

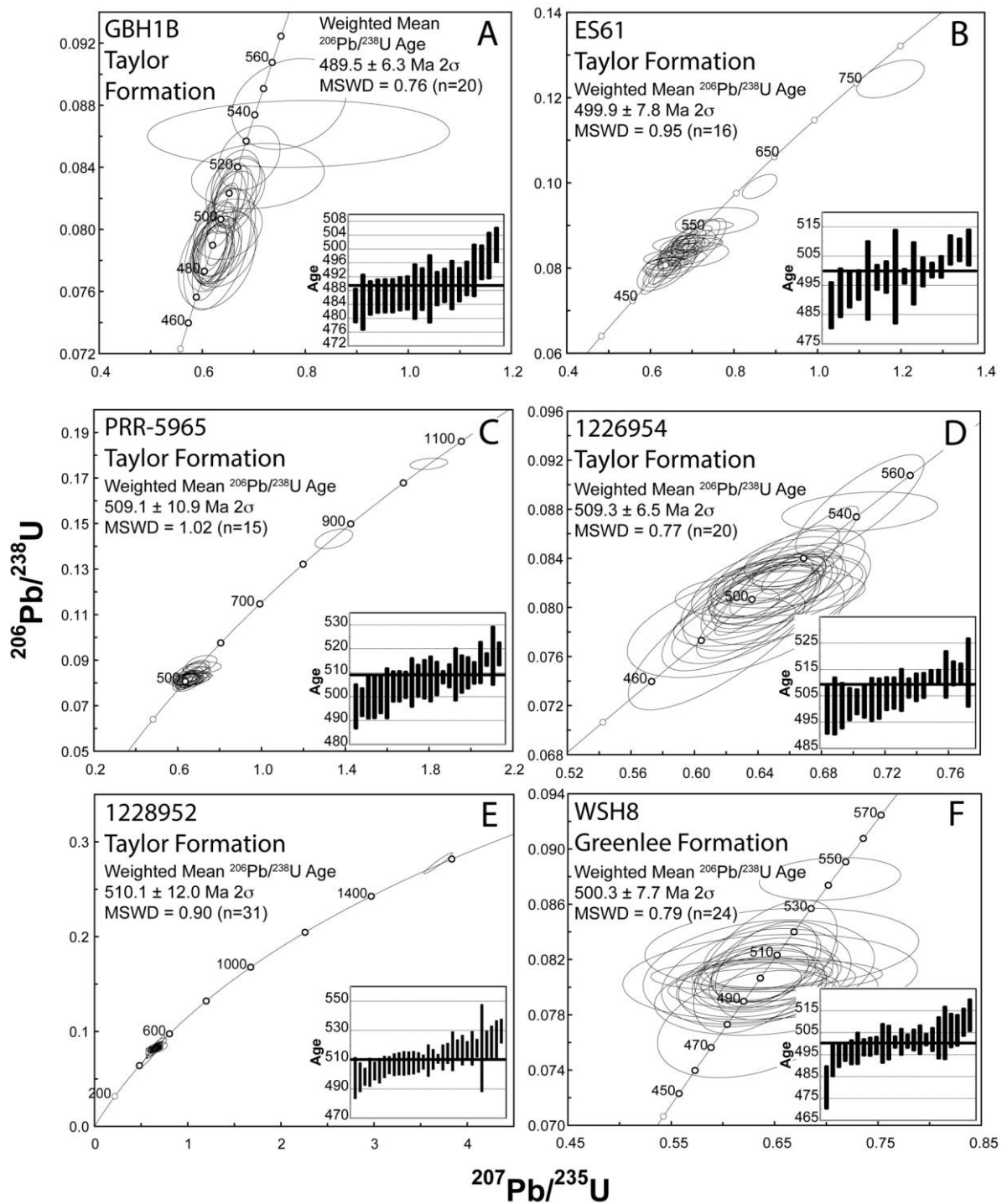


Figure 6. Concordia diagrams of zircon U-Pb data and weighted mean age plots of individual analyses for Taylor Formation samples GBH1B (A), ES61 (B), PRR-5965 (C), 1226954 (D), and 1228952 (E) and for Greenlee Formation sample WSH8 (F). Error bars are 1σ from inferred U-Pb ages ($^{206}\text{Pb}/^{238}\text{U}$ age for samples younger than 1000 Ma and $^{207}\text{Pb}/^{206}\text{Pb}$ age for samples older than 1000 Ma). Data error ellipses are 2σ .

to Pb loss. Sample ASL is a metafelsite that comes from a small spur that extends to the northeast from near the summit of Mount Fairweather, which is the same sample locality as the volcanic rock sample (ASJ) that yielded an upper intercept discordia age of 550 ± 15 Ma (Van Schmus et al. 1997). A weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 524 ± 9 Ma from a concordant set of analyses is considered to be the age of igneous crystallization for sample ASL (fig. 5B).

Samples PRR-5298 and ES148 were collected from localities to the west of the Liv Glacier at Mount Fairweather (fig. 4; Stump 1985; Grunow et al. 1996a). Sample PRR-5298 comes from a ~10-m-thick white micaceous rock with mafic spots that was described as a metamorphosed tuff within marble collected from the Henson Marble Member of the Fairweather Formation just above the level of the Liv Glacier at the Tusk. It likely has a tuff-volcaniclastic protolith, which is strongly supported by the presence of apparently unabraded, euhedral, oscillatory-zoned crystals that occur within the zircon split. A weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 520 ± 7 Ma from a concordant set of analyses is considered to be the age of igneous crystallization for sample PRR-5298 (fig. 5C). Sample ES148 is a metafelsite collected from interbedded volcanic and siliciclastic rocks of the Fairweather Formation at Mount Henson. A weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 514 ± 9 Ma from a concordant set of analyses is considered to be the age of igneous crystallization for sample ES148 (fig. 5D).

Sample GBH9D is a quartz-rich volcanic breccia collected from deformed metavolcanic rocks and marbles at a small nunatak located to the east of Polaris Peak (fig. 4). A weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 521 ± 7 Ma from a concordant set of analyses is considered to be the age of igneous crystallization for sample GBH9D (fig. 5E). Three concordant zircon age analyses yield 557–550 Ma ages, which presumably reflect inheritance.

The 524 ± 9 to 514 ± 9 Ma volcanic crystallization ages for our samples of the Fairweather Formation are consistent with an early Cambrian age (possibly late Terreneuvian or Series 2 from 521 to 509 Ma) suggested by a solitary probable archaeocyathan found at Mount Fairweather in the Henson Marble Member (Grunow et al. 1996a) as well as 512–504 Ma maximum depositional ages indicated by detrital zircon age populations for the Fairweather Formation (Paulsen et al. 2015, 2017).

Taylor/Greenlee Formations. Eighty percent ($n = 217$ of 272) of the U-Pb age analyses of zircons from the Taylor Formation (GBH1B, ES61, PRR-5965, 1226954, and 1228952) and the Greenlee Formation (WSH8) met acceptable concordance thresholds. The cumulative zircon age distribution from all of the

samples range from 1042 Ma (Mesoproterozoic) to 480 Ma (early Ordovician).

Samples GBH1B, ES61, and PRR-5965 were collected from localities to the east of lower stretches of the Shackleton Glacier (fig. 4). Sample GBH1B is a protomylonitic rhyolite porphyry collected from brightly colored pink, red, and green mylonitic volcanic rocks of the Taylor Formation at Nilsen Peak along the Massam Glacier near the Transantarctic Mountain front. A weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 490 ± 6 Ma from a concordant set of analyses is considered to be the age of igneous crystallization for sample GBH1B (fig. 6A). Samples ES61 and PRR-5965 were collected from interbedded volcanic and siliciclastic rocks of the Taylor Formation at Mount Wendland. Sample ES61 is a metafelsite with a definite welded shard structure with spherulites. A weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 500 ± 6 Ma from a concordant set of analyses is considered to be the age of igneous crystallization for sample ES61 (fig. 6B). Sample PRR-5965 is a dacite cobble collected from a sheared monomictic conglomerate, which likely represents locally eroded volcanic material from interbedded Taylor Formation volcanic rocks, considering its monomictic character. A weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 509 ± 11 Ma from a concordant set of analyses is considered to be the age of igneous crystallization for sample PRR-5965 (fig. 6C), which also represents a maximum depositional age for the Taylor Formation that is consistent with the 500 Ma crystallization age of sample ES61 from Mount Wendland.

Samples 1226954, 1228952, and WSH8 were collected from localities on the west flank of upper stretches of the Shackleton Glacier (fig. 4). Sample 1226954 is a protomylonitic felsite porphyry collected from brightly colored pink and green volcanic outcrops visible on the high slopes of Mount Greenlee at Taylor Nunatak. A weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 509 ± 6 Ma from a concordant set of analyses is considered to be the age of igneous crystallization for sample 1226954 (fig. 6D). Samples 1228952 and WSH8 were collected from interbedded volcanic and siliciclastic rocks of the Taylor Formation along the Shackleton Glacier at Mount Greenlee. Sample 1228952 is a metafelsite collected from southern exposures of the Mount Greenlee sequence to the north of granites that intrude the sequence. A weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 510 ± 12 Ma from a concordant set of analyses is considered to be the age of igneous crystallization for sample 1228952 (fig. 6E). Sample WSH8 is a massive to crudely layered plagioclase crystal tuff of the Greenlee Formation that alternates with black argillite and is overlain by a massive mafic unit. A weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 500 ± 8 Ma from a concordant set of analyses is

considered to be the age of igneous crystallization for sample WSH8 (fig. 6F).

The 510 ± 12 to 490 ± 6 Ma volcanic crystallization ages yielded by the Taylor and Greenlee Formations are consistent with U-Pb dates and paleontological (trilobite) data from felsic volcanic rocks at Lubbock Ridge (516 ± 6 Ma U-Pb crystallization age; Van Schmus et al. 1997) and carbonate units (Cambrian, Series 3) interbedded with volcanic rocks (505 ± 1.5 Ma U-Pb crystallization age) at Taylor Nunatak (Encarnación et al. 1999). The Taylor and Greenlee Formation crystallization ages are also consistent with 515 and 502 Ma maximum depositional ages indicated by detrital zircon age populations for the Taylor and Greenlee Formations at Lubbock Ridge and Mount Greenlee, respectively (Paulsen et al. 2015, 2017).

Discussion and Conclusions

The 524 ± 9 to 490 ± 6 Ma U-Pb crystallization ages reported in this article provide new constraints on the depositional ages of the volcanic and associated interbedded metasedimentary rocks that comprise the Fairweather and Taylor/Greenlee Formations of the Liv Group in the Shackleton to Liv Glacier area. These and the previously published Taylor Formation crystallization ages indicate that the Fairweather Formation (533–505 Ma) is likely to be early to middle Cambrian in age (Terreneuvian to early Series 3) and that the Taylor/Greenlee Formations (522–483 Ma) are middle to late Cambrian (Series 2 to Furongian) in age. These ages, taken together with the biostratigraphic and detrital mineral age data shown in figure 3, confirm the temporal correlation of portions of the Taylor, Greenlee, and Fairweather Formations, as previously suggested by Stump (1985) and Wade and Cathey (1986).

In other regions like the central Transantarctic Mountains, the depositional age of synorogenic sediments of the Ross orogen is bracketed by detrital mineral ages, metamorphic cooling ages, and cross-cutting igneous ages that allow Cambrian to earliest Ordovician deposition (Goodge et al. 2004a). By contrast, the 533–483 Ma ages (Cambrian, Terreneuvian to Series 2) from syndepositional volcanic units from which our samples were collected provide important new data that tightly constrain deposition in this sector of the Ross orogen to primarily lie within the Cambrian. On a regional scale, the 533–483 Ma volcanic ages overlap with biostratigraphic, U-Pb volcanic crystallization, and maximum depositional ages (based on detrital zircon U-Pb ages) for interbedded sedimentary and volcanic rocks found elsewhere in the Ross orogen (fig. 3) as well as volcanic

rock successions in Australia and New Zealand (Squire and Wilson 2005). The 533–483 Ma volcanic crystallization ages are significantly younger than the poorly defined discordia upper intercept age of 550 ± 15 Ma for a volcanic rock from the Fairweather Formation at Mount Fairweather (Van Schmus et al. 1997), eliminating these interbedded meta-volcanic and metasedimentary rocks as candidates for Precambrian cratonic crust and the oldest volcanic rocks exposed in the Ross-Delamerian orogenic belt. The 533–483 Ma crystallization ages therefore preclude correlation of our samples with ~670–650 Ma synrift sedimentary rocks of the Beardmore Group in the central Transantarctic Mountains and the Skelton Group of south Victoria Land (fig. 3; Goodge et al. 2002; Cooper et al. 2011).

The 533–505 Ma volcanic crystallization ages of Liv Group volcanic rocks indicate deposition concomitant with the younger emplacement of ~535–490 Ma plutonic rocks found in the Queen Maud batholith (Queen Maud Mountains) and other Cambrian igneous rocks along the early Paleozoic Pacific-Gondwana margin (fig. 3). Figure 7 shows a relative age probability diagram for all zircon U-Pb crystallization ages obtained for volcanic and plutonic rocks in this study and previous studies (Encarnación and Grunow 1996; Van Schmus et al. 1997; Vogel et al. 2002; Paulsen et al. 2008, 2013) for the Queen Maud batholith. The probability analysis shows a primary 521 Ma age probability peak for the crystallization ages of igneous rocks. The zircon U-Pb igneous crystallization ages reported herein provide important new evidence for an early Cambrian (Terreneuvian to Series 2) flare up of magmatism in the Queen Maud batholith sector of the Ross orogen, which overlaps in age with 565–492 Ma magmatism in south Victoria Land (Hagen-Peter et al. 2015; Hagen-Peter and Cottle 2016) and 545–485 Ma magmatism in the central Transantarctic Mountains (Goodge et al. 2012).

The new crystallization ages reported in this article add substantially to our understanding of the age of magmatism in the Queen Maud batholith, thereby warranting an examination of its relation to sandstone provenance in the region. Importantly, the probability age distribution and peaks yielded by zircon U-Pb igneous crystallization ages in the Queen Maud Mountains (fig. 7A) are similar to the youngest detrital zircon age probability peaks found in siliciclastic rocks from this area, suggesting that siliciclastic components of these successions mark the deposition of sedimentary material associated with, and derived from, arc-related rocks of the Ross orogen (Stump 1995; Encarnación and Grunow 1996; Myrow et al. 2002; Goodge et al. 2004b; Paulsen et al. 2015). The older 638, 608, and 563 Ma detrital zircon

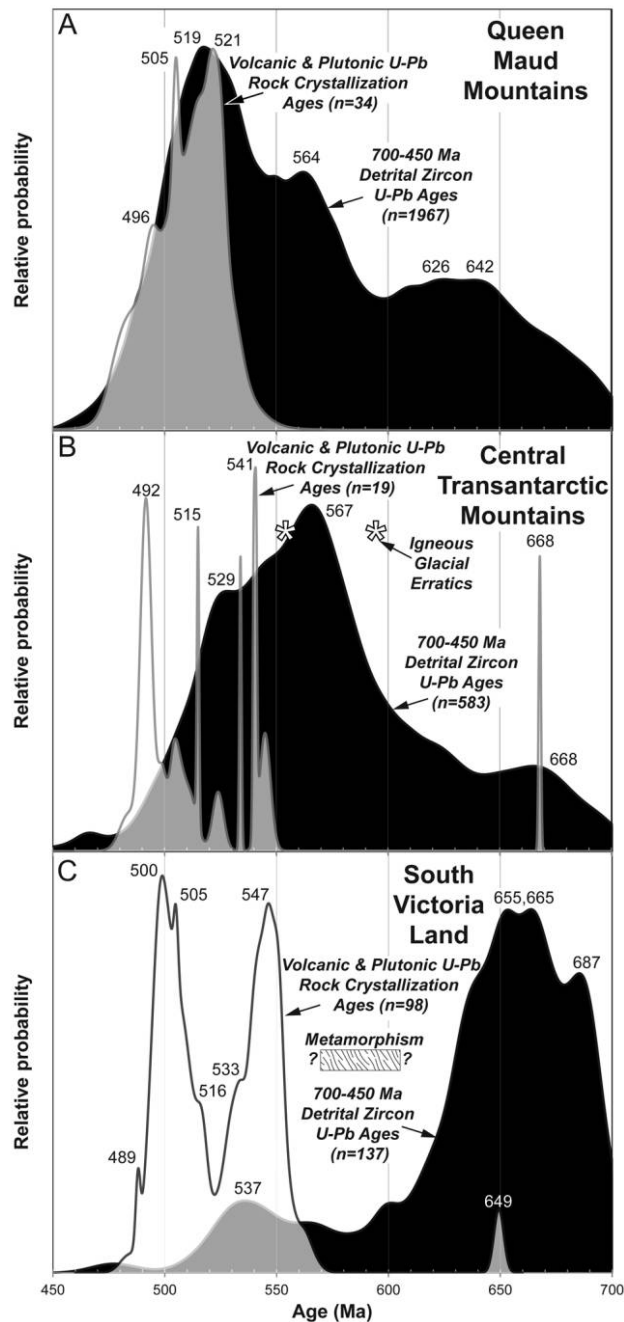


Figure 7. Stacked relative age probability diagrams (Ludwig 2008) for U-Pb igneous crystallization ages and detrital zircon ages falling within the 700–450 Ma time period from the Queen Maud Mountains (A), central Transantarctic Mountains (B), and south Victoria Land (C) sectors of the Ross orogen. These probability curves take each age and its normally distributed uncertainty (for measurement error only) and sum them into a single curve. Asterisks in B show ages of igneous glacial erratics reported in Goodge et al. (2012). Detrital zircon ages are compiled for the Liv and Beardmore Groups (A) from Stump et al. (2007) and Paulsen et al. (2015, 2017); for the Byrd Group (B) from Goodge et al. (2002, 2004b) and Paulsen et al. (2017); and for the Skelton

age peaks yielded by siliciclastic rocks of the Queen Maud Mountains might reflect erosion of unexposed igneous rocks along the Pacific-Gondwana margin (Goodge et al. 2004b) or the interior of the East Antarctic craton. For example, in the central Transantarctic Mountains, magmatic crystallization ages also overlap with <545 Ma detrital zircon age populations yielded by synorogenic Byrd Group sandstones (fig. 7B). Similar older detrital zircon ages have also been found in these sandstones (Goodge et al. 2004b), where glacial clasts with 590–554 Ma crystallization ages suggest the presence of older unexposed igneous rocks (Goodge et al. 2012). In south Victoria Land, ~560–488 Ma igneous rocks do not appear to be sources for the older detrital zircons that have been found to dominate Skelton Group sandstones, which were apparently deformed and metamorphosed prior to igneous intrusion (Hagen-Peter and Cottle 2016; Hagen-Peter et al. 2016). However, Skelton Group sediment appears to have been derived from older igneous rocks, of which local exposures occurring in south Victoria Land and the central Transantarctic Mountains have yielded 668–650 Ma ages. It is also possible that similar age populations are at least in part derived from long-distance transport from Dronning Maud Land (Jacobs 1999) or the East African orogen (Squire et al. 2006).

The 490 Ma (GBH1B) crystallization age for our deformed Taylor Formation volcanic sample from Nilsen Peak overlaps with similar age plutonic rocks found throughout the orogen, some of which in the Queen Maud Mountains are also deformed (Paulsen et al. 2008, 2013). This sample is also similar in age to 497–482 Ma U-Pb crystallization ages yielded by clasts of rhyolite porphyry and tuff found in the basal conglomerates belonging to the Devonian (post-Ross orogen) Taylor Group of the Beacon Supergroup in south Victoria Land (Wysoczanski et al. 2003). The 490 Ma (GBH1B) crystallization age assumes regional significance in the Ross orogenic belt because it represents the youngest in situ volcanic rock yet iden-

Group (C) from Stump et al. (2007), Cooper et al. (2011), and Paulsen et al. (2017). Volcanic and plutonic U-Pb crystallization ages are compiled from data presented in this article as well as Goodge et al. (1993, 2004b, 2012), Rowell et al. (1993), Hall et al. (1995), Encarnación and Grunow (1996), Cooper et al. (1997), Van Schmus et al. (1997), Encarnación et al. (1999), Cox et al. (2000), Cook and Craw (2001), Allibone and Wysoczanski (2002), Mellish et al. (2002), Read et al. (2002), Vogel et al. (2002), Stump et al. (2004, 2006), Wysoczanski and Allibone (2004), Cottle and Cooper (2006a, 2006b), Read (2010), Cooper et al. (2011), Paulsen et al. (2013, 2016c), Hagen-Peter et al. (2015), Martin et al. (2015), and Hagen-Peter and Cottle (2016).

tified within the Ross orogen and therefore provides important new evidence for Cambrian (Furongian) or younger deformation in this sector of the range.

The majority of $^{40}\text{Ar}/^{39}\text{Ar}$ mineral cooling ages for metamorphic and plutonic rocks of the Queen Maud batholith range from ~502 to 473 Ma and therefore broadly overlap with this phase of magmatism (fig. 3; Grunow and Encarnación 2000b; Paulsen et al. 2004, 2008). The timing of this magmatic and orogenic cooling phase shows remarkable overlap with the emplacement of undeformed intrusions, orogenic cooling, and exhumation that occurred after ~500 Ma along the length of the Ross orogen and its northern extension, the Delamerian orogen in Australia (Borg et al. 1990; Goodge et al. 1991, 2004a, 2012; Goodge and Dallmeyer 1992, 1996; Stump 1995; Encarnación and Grunow 1996; Turner et al. 1996; Foster and Gray 2000; Allibone and Wysoczanski 2002; Foster et al. 2005; Squire and Wilson 2005; Foden et al. 2006; Paulsen et al. 2007; Talarico et al. 2007; Rocchi et al. 2011; Hagen-Peter and Cottle 2016; Hagen-Peter et al. 2016). This orogenic phase might have been associated in part with a geodynamic change to extension along the Gondwana margin that was fostered by rollback of the subducting slab with possible influences from lithospheric delamination (Turner et al. 1996; Foster and Gray 2000; Wareham et al. 2001; Foden et al.

2006; Rocchi et al. 2011). Indeed, orogenic collapse provides an attractive tentative model for the development of shear zones observed to crosscut Taylor Formation volcanic rocks in the Nilsen Peak (sample GBH1B) and Mount Greenlee (samples 1226954 and 1228952) areas in the Shackleton Glacier region (Stump 1985; Paulsen et al. 2004), where high-angle dips and steeply plunging mineral stretching lineations in the volcanic rocks are consistent with the structural orientations expected during crustal extension.

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