



Examining the Nuna-Rodinia supercontinent cycle in East Antarctica and testing correlations to the southwest United States: evidence from the Read Group (Shackleton Range) and the Nimrod Complex

Joe H. Metcalfe^{a,b,*}, Teal R. Riley^a, Michael J. Flowerdew^c, Nicholas J. Gardiner^b, Sebastian Fischer^b, Ian L. Millar^d, Joshua M. Garber^b, Maurice Brodbeck^e, David van Acken^e, Cristina Talavera^f, Heejin Jeon^g, Martin J. Whitehouse^g

^a British Antarctic Survey, High Cross, Madingley Road, Cambridge CB3 0ET, UK

^b School of Earth & Environmental Sciences, University of St Andrews, Bute Building, Queen's Terrace, St Andrews KY16 9TS, UK

^c CASP, Madingley Rise, Madingley Road, Cambridge CB3 0UD, UK

^d British Geological Survey, Keyworth, Nottingham NG12 5GG, UK

^e UCD School of Earth Sciences, University College Dublin, Belfield, Dublin 4, Ireland

^f School of Geosciences, University of Edinburgh, Edinburgh, UK

^g Swedish Museum of Natural History, Stockholm, Sweden

ARTICLE INFO

Keywords:

Geochronology

Pb isotopes

Hf isotopes

East Antarctica

Supercontinent

Nuna

Polar Rock Repository

ABSTRACT

The Read Group of the southern Shackleton Range and the Nimrod Complex of the central Transantarctic Mountains (cTAMs) form two cratonic blocks with unknown spatial extent and ambiguous crustal affinities. They may form an extension of the Mawson Continent, which is generally thought to adjoin the Gawler Craton of South Australia with the Adélie Craton of East Antarctica. Additionally, variations of the original SWEAT (SouthWest United States – East Antarctica) hypothesis for the Nuna-Rodinia supercontinent cycle envisages that the Mojave and Yavapai orogenic provinces of the American Southwest continue into East Antarctica, either via the Shackleton Range or beneath the Ross-Delamerian orogen in the cTAMs.

New U-Pb zircon geochronology and whole-rock geochemical analyses of *meta*-igneous suites from the Read Group indicate arc-related magmatic pulses at c. 1850–1825, 1810–1790, and 1780–1770 Ma. The extensive metasedimentary and metavolcanic paragneiss sequences of the Read Group are defined by detrital zircons with primary age peaks at c. 1830 Ma and secondary age peaks between 2700–1900 Ma. The paragneisses are cut by 1810–1770 Ma granitoid intrusions, demonstrating that their formation was prior to c. 1810 Ma. K-feldspar Pb isotope compositions from the Read Group ($\mu = 10.0$ – 10.1) are different to the Nimrod Complex ($\mu = 8.1$ – 9.8), suggesting that the Read Group and the Nimrod Complex are not contiguous and are therefore unlikely to form a coherent cratonic block with the Gawler-Adélie cratons. By contrast, K-feldspar Pb and zircon Lu-Hf isotope compositions from the Read Group are an excellent match for the Mojave and Yavapai provinces. The Mojave-Yavapai provinces also have coeval magmatic pulses and matching detrital zircon age spectra to the Read Group. A tectonic piercing point between the Shackleton Range and Mojave province is supported by the new data and is consistent with the Laurentian affinity of the Mesoproterozoic Coats Land block to the north, raising the possibility that both Coats Land and the Shackleton Range are remnants of the Great Proterozoic Accretionary Orogen, which remained attached to Laurentia throughout the Nuna-Rodinia supercontinent cycle until the break-up of Rodinia at c. 750 Ma.

1. Introduction

Palaeogeographical reconstructions for the Proterozoic were initially

modelled under the assumption that pre-Gondwanan East Antarctica originated and has remained a single contiguous crustal block since c. 1000 Ma (e.g., Craddock 1972; Elliot 1975; Borg et al. 1990). These

* Corresponding author at: British Antarctic Survey, High Cross, Madingley Road, Cambridge CB3 0ET, UK.

E-mail address: jostca79@bas.ac.uk (J.H. Metcalfe).

<https://doi.org/10.1016/j.precamres.2026.108235>

Received 31 March 2026; Received in revised form 24 July 2026; Accepted 27 July 2026

Available online 21 August 2026

0301-9268/© 2026 The Author(s). Published by Elsevier B.V. This is an open access article under the CC BY license (<http://creativecommons.org/licenses/by/4.0/>).

models were underpinned by the correlation of several Mesoproterozoic “Grenville” orogenic belts into a circum-orogen interpreted to wrap the so-called ‘East Antarctic craton’ against peripheral cratonic fragments with African (Grunehogna; Halpern 1970; Barton et al. 1987; Marshall et al. 2010) and Indian (Napier; Grew 1982; Motoyoshi 1999; Ishizuka 2008; Król et al. 2023) affinities (e.g., Yoshida 1995). However, post-millennial geological investigations on coastal exposures of Archaean-Proterozoic terranes, coupled with advancements in sub-glacial geophysics, have revealed that the Mesoproterozoic orogenies are crisscrossed and overprinted by late Neoproterozoic-Cambrian “Pan-African” orogenic belts (Fitzsimons 2000a, Fitzsimons 2000b; Harley 2003; Zhao et al. 2003; Jacobs & Thomas 2004; Fig. 1), meaning that East Antarctica did not assemble into its present-day configuration until Gondwana assembly (c. 500 Ma). Despite these advancements, concealment of the continental interior by the East Antarctic Ice Sheet has precluded the emergence of a preferred crustal architecture and various competing models remain distributed throughout the literature. For example, it is still unclear whether any of these Pan-African and/or Grenville orogenic belts continue beneath the ice (e.g., Fitzsimons 2003; Goodge et al. 2010, 2024) and, by implication, whether or not they stitch together a number of Archaean-Palaeoproterozoic basement terranes which face the ice sheet interior. Understanding the temporal and spatial limits of these basement terranes is therefore critical to refining

the formation of not only East Antarctica as a coherent continent, but also its role in the tectonic assembly of the ancient supercontinents Nuna, Rodinia and Gondwana throughout the Proterozoic.

The Palaeoproterozoic Shackleton Range of East Antarctica is an example where crustal affinity remains ambiguous. Detailed field and analytical investigations from the EuroShack expedition in 1994/95 (Tessensohn & Thomson 1999) built on earlier geological mapping and sampling in the mid- to late- 20th century, but it has attracted relatively little attention since. Geochronological and geochemical data from the Read Group within the southern Shackleton Range (Pankhurst et al. 1983; Will et al. 2009, 2010) show that this region has avoided isotopic resetting beyond the Palaeoproterozoic and is therefore regarded as cratonic, yet it is unknown how far south the Read Group may continue, and whether it is contiguous or not with any other basement terranes in East Antarctica. The geographically closest potential link to the Read Group is the Mesoarchaeoan-Palaeoproterozoic Nimrod Complex of the Miller and Geologists Range – the only exposed crystalline basement of the Transantarctic Mountains, some 1800 km of ice sheet cover away (Fig. 1). The Nimrod Complex preserves evidence for tectono-magmatic events at c. 3100–2900, c. 2500 Ma, c. 1720 Ma and c. 530 Ma (Goodge & Fanning 1999, 2016; Goodge et al. 2001). Such age patterns closely resemble the proposed ‘Mawson Continent’ (Fanning et al. 1995, 1996; Payne et al. 2009), which originally was thought to encompass just the

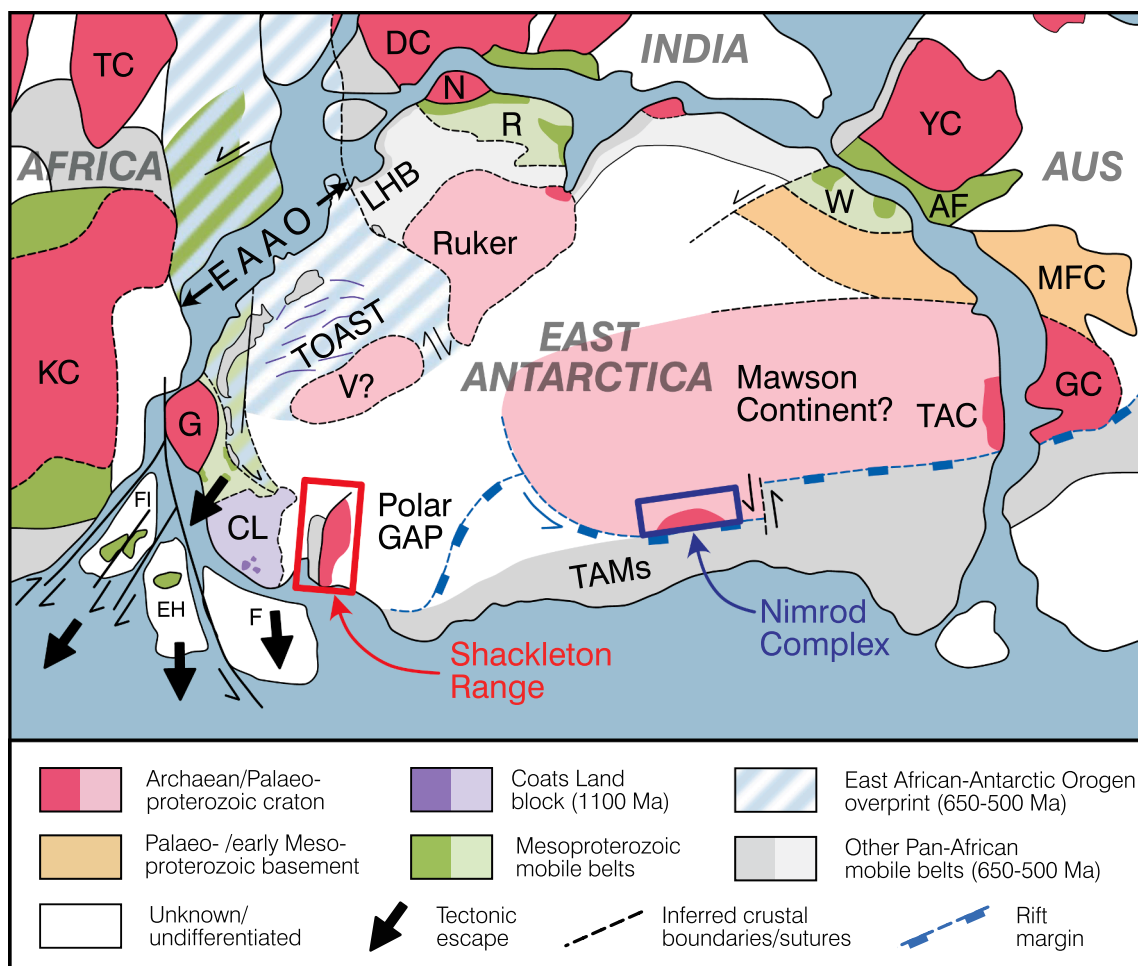


Fig. 1. Sketch tectonic map of East Antarctica, Australia (AUS), southern Africa, and India after Gondwana assembly at c. 500 Ma, adapted from Jacobs & Thomas (2004) and (Fitzsimons, 2000a). AF – Albany-Fraser Orogen, CL – Coats Land, DC – Dhawar Craton, EAAO – East African Antarctic Orogen, EH – Ellsworth-Haag, F – Filchner block, FI – Falklands Islands, G – Grunehogna Craton, GC – Gawler Craton, KC – Kalahari Craton, MFC – Madura-Forrest-Coompana provinces, N – Napier Craton, NC – Nimrod Complex, R – Rayner Complex, TAC – Terre Adélie Craton, TAMS – Transantarctic Mountains, TC – Tanzania Craton, TOAST – Tonian Oceanic Arc Super Terrane, V – Valkyrie Craton, W – Wilkes, Y – Yilgarn Craton. Solid colours refer to exposed tectonic domains, paler shades infer/speculate on their possible sub-glacial extent.

Gawler Craton of South Australia and its inferred counterparts along the East Antarctic coastlines of Terre Adélie and King George V Land (Fanning et al. 1995, 1996; Peucat et al. 1999; Fig. 1). Despite differing interpretations of geophysical models (e.g., Stål et al. 2019; Jordan et al. 2022), the Nimrod Complex has generally been considered an extension of the Mawson Continent (e.g., Goodge & Fanning 2016). However, any further continuation beyond the Nimrod Complex (e.g., towards the Shackleton Range) is inhibited by a discontinuity in the Mawson-like aeromagnetic signature near the Pensacola Mountains (Jordan et al. 2022). The low magnetic ‘Pensacola Embayment’ occupies the South Pole region, truncating the ‘Southern Mawson Province’ which instead shows high amplitude positive magnetic anomalies and low Bouguer gravity anomalies typical of thick (>40 km) continental lithosphere (Jordan et al. 2022; Fig. 2). The Shackleton Range lies on the opposite side of the Pensacola Embayment from the Nimrod Complex, within the sparsely covered and magnetically complex PolarGAP province (Forsberg et al. 2018; Jordan et al. 2022; Fig. 2). Therefore, whilst the Shackleton Range may still form a rifted fragment or later addition to the Mawson Continent, its incorporation into a coherent cratonic core within East Antarctica is considered unlikely, and its crustal affinity may be aligned elsewhere.

To the north of the Shackleton Range is the magnetically defined Coats Land block (Figs. 1 & 2), which includes very sparsely exposed c. 1110 Ma felsic igneous intrusives (Gose et al. 1997) that have been correlated with Midcontinent Rift large igneous provinces (LIPs) in west Texas and the Lake Superior region (Loewy et al. 2011). Little to no exposure between the Coats Land nunataks and the Shackleton Range

means their tectonic relationship remains unclear, however their proximity and the Laurentian affinity of Coats Land ponders whether this link could be extended to the Shackleton Range. Indeed, the original SWEAT hypothesis (SouthWest United States – East Antarctica) of Moores (1991) noted the overlapping age (c. 1800–1600 Ma) of the Palaeoproterozoic Mojave-Yavapai-Mazatzal orogenic belts of southwestern United States (SWUS) with the southern Shackleton Range and the two regions would have a broadly concordant structural trend when rotated into their respective Nuna orientations (Dalziel 2024). However, similar so-called “piercing points” to the Mojave promontory have been championed for the Nimrod Complex (e.g., Goodge et al. 2001, 2008, Goodge & Fanning 2016) and the Gawler Craton (Karlstrom et al. 2001; Betts et al. 2008, 2011, 2016), highlighting the need for further investigation into the southern Shackleton Range to restore its position in global plate tectonic reconstructions throughout the Proterozoic.

Here, we provide a comprehensive set of new U-Pb zircon geochronology, K-feldspar Pb isotope, zircon Lu-Hf isotope and whole-rock geochemical data from a series of orthogneisses, paragneisses and granitoids from the Read Group of the Shackleton Range. We also present comparative data from Nimrod Complex and discuss the closest crustal affinities of both terranes in the context of palaeocontinent reconfigurations during the Nuna-Rodinia cycle.

2. Geological setting of the Shackleton Range

The Shackleton Range (80°07′–80°50′S, 19°–31°W) of East Antarctica (Fig. 3) is a mountain block situated within southern Coats Land, inland

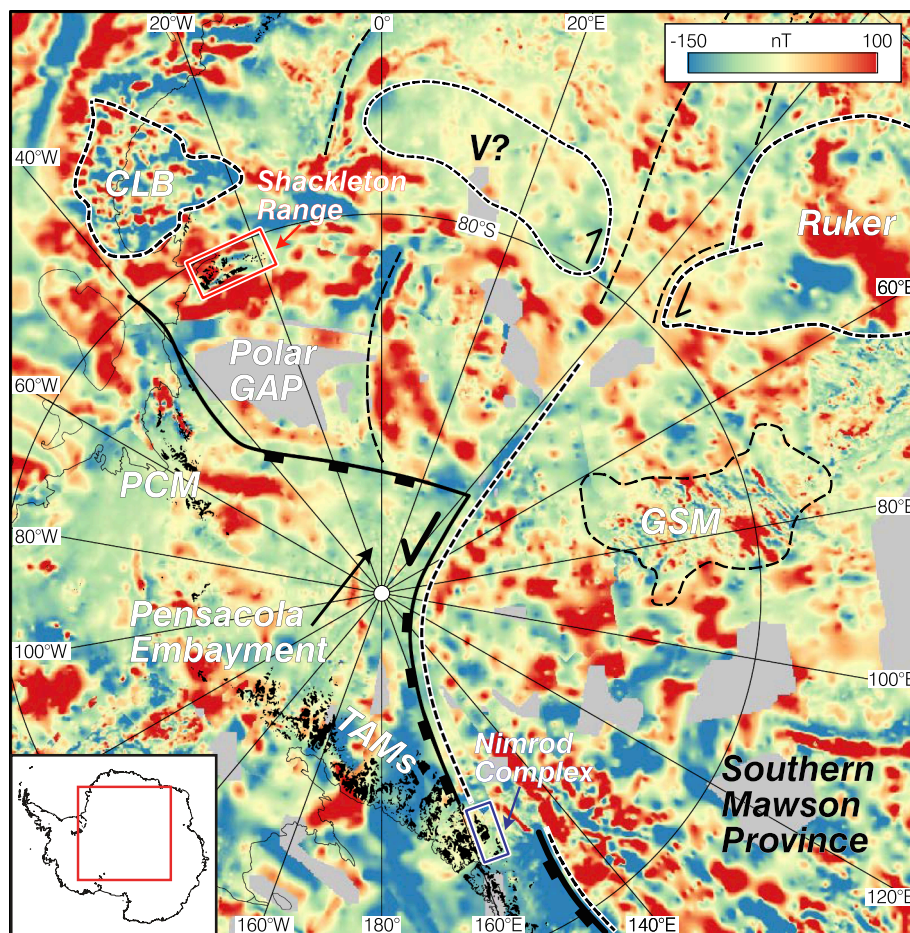


Fig. 2. Inset of ADMAP2 overlain by possible cratonic, orogenic and rifted margin boundaries, after Mieth et al. (2014) and Jordan et al. (2022). Red and blue boxes shows the locations of the Shackleton Range and Nimrod Complex respectively, the bold black line and ticks indicate the possible rifted margin and Pensacola Embayment into the Southern Mawson Province. Regions of grey are without available data. CLB – Coats Land Block, GSM – Gamburtsev Subglacial Mountains, PCM – Pensacola Mountains, TAMS – Transantarctic Mountains, V – Valkyrie Craton.

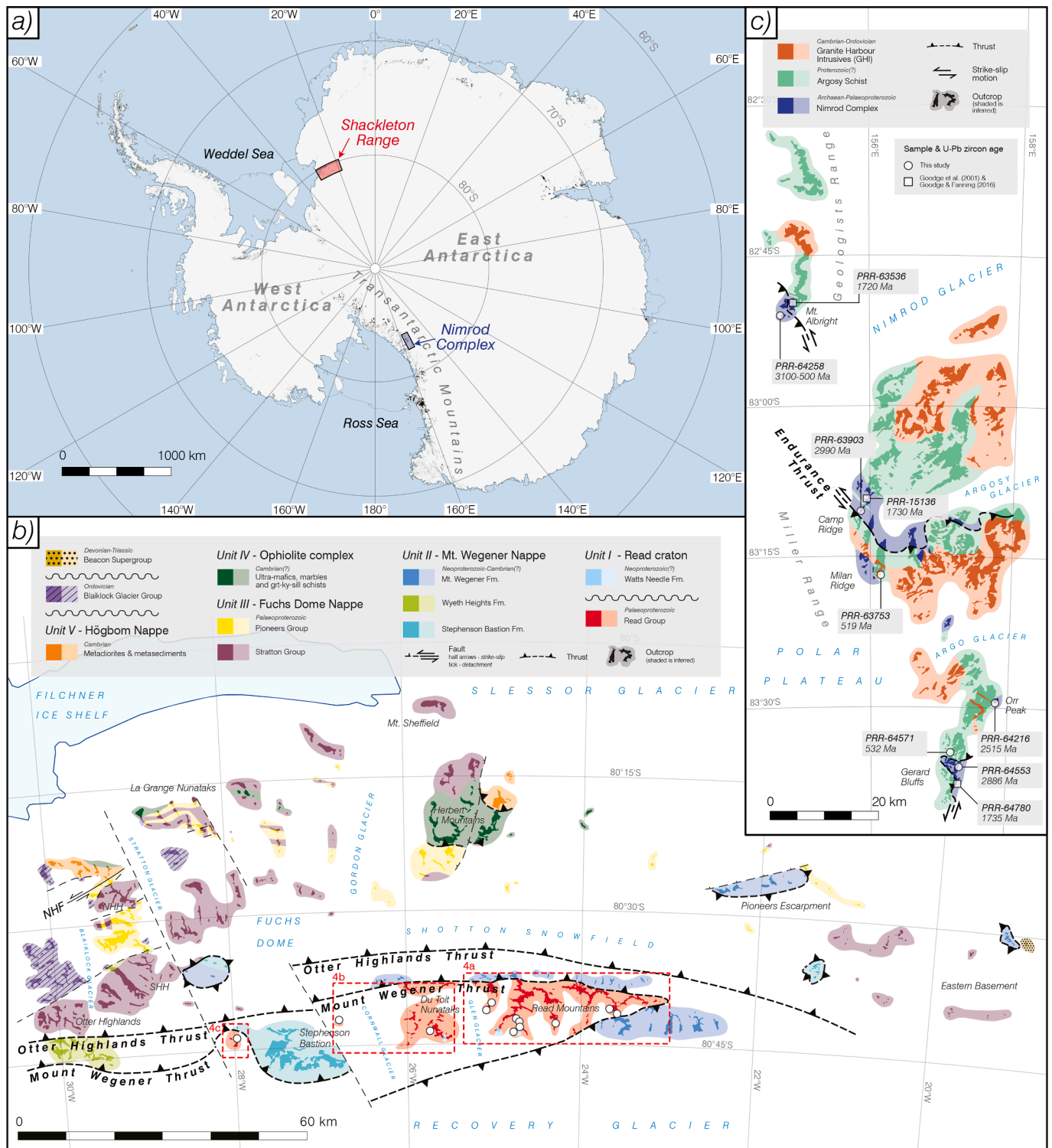


Fig. 3. a) Location map of the Shackleton Range and Nimrod Complex within Antarctica, b) Geological map of the Shackleton Range, after Clarkson et al. (1995) and Kleinschmidt (2021). Dashed red boxes show the approximate limits of Fig. 4(a-c), c) Geological map of the Miller and Geologists ranges after Goodge & Fanning (2016), with sample numbers, locations and U-Pb zircon ages determined by either this study (circles) or by Goodge et al. (2001) and Goodge & Fanning (2016) (squares). Solid colours refer to exposed rock, paler shades infer their minimal sub-glacial extent.

from the Filchner Ice Shelf and the Weddell Sea. Sandwiched between the Slessor and Recovery glaciers, it likely forms a normal fault-bounded horst structure (Clarkson 1972; Skidmore & Clarkson 1972), extending roughly 240 km eastward before disappearing beneath the East Antarctic Ice Sheet. Its dominant east-west structural trend suggests it is distinct from the mostly north-south trending Transantarctic Mountains

(TAMs) (Clarkson et al. 1995; Kleinschmidt 2021; Figs. 1-3) which extend only to the Pensacola Mountains some 250 km to the south. The TAMs straddle the palaeo-Pacific margin of East Antarctica and record the Pan-African or “Ross” accretion (615–470 Ma; Stump 1995; Encarnación & Grunow 1996; Storey et al. 1996; Myrow et al. 2002; Goodge 2020) of various crustal fragments which now comprise much of

the present-day West Antarctica.

The overall structure of the Shackleton Range is a series of five E-W trending nappes or units assigned by Tessensohn et al. (1999) and revised by Kleinschmidt (2021), each of which have been variably reworked by Ross orogenesis. Autochthonous crystalline basement rocks of the southernmost Read Group (Unit I) and the overriding, low-grade metasedimentary Mount Wegener nappe (Unit II) show dominantly southward-directed thrust transport, the latter bounded by the E-W trending Mount Wegener and Otter Highlands thrusts (Kleinschmidt & Buggisch 1994; Fig. 3). To the north are the Fuchs Dome (Unit III) and Högbom (Unit V) nappes, which were originally mapped together as the Stratton and Pioneers groups of Tessensohn & Thomson (1990) and Clarkson et al. (1995). The Stratton and Pioneers groups broadly correspond to re-mobilised crystalline basement and its cover of now metasedimentary and metavolcanic sequences, respectively (Clarkson et al. 1995). They show mostly Palaeoproterozoic protolith ages and Ross-age isotopic resetting, although the Högbom Nappe appears to show only Ross protolith ages (c. 530 Ma) and thus warrants a separate tectonic unit (Zeh et al. 1999; Will et al. 2009, 2010). Both units differ from the southern Shackleton Range by recording dominantly westward-directed transport, best exhibited by blastomylonitic gneisses at Mt. Sheffield and several sinistral strike-slip faults (e.g., Northern Haskard Fault, NHF; Fig. 3) in the Haskard Highlands (Marsh 1983; Zeh et al. 1999). Slices of eclogite-facies mafic to ultramafic meta-igneous rocks in these regions (Schmädicke & Will 2006; Romer et al. 2009) may represent relics of an ophiolite complex (Unit IV) or hyper-extended mantle interposed between units III and V. In the Haskard Highlands, the northern units are unconformably overlain by Ordovician red-bed siliciclastics of the Blaiklock Glacier Group which represent molasse deposits to the Ross orogen (Clarkson 1972; Clarkson & Wyeth 1983;

Buggisch et al. 1999; Fig. 3).

Most workers originally proposed a sinistral transpressive collision of the Kalahari Craton and East Antarctica during the closure of the Mozambique Ocean as the most likely tectonic scenario for the formation of the Shackleton Range (Tessensohn et al. 1999; Buggisch & Kleinschmidt 2007). In this case, the Read Group, Fuchs Dome nappe, and Högbom nappe basement units represent fundamentally different terranes with the eclogite-facies ophiolite relics interposed between them possibly representing a Pan-African collisional suture between East and West Gondwana. However, U-Pb geochronology and isotopic geochemistry by Will et al. (2009, 2010) revealed similarities between the northern and southern basement. These authors proposed that the Read Group and much of the Fuchs Dome nappe as far north as the Northern Haskard Highlands and the La Grange Nunataks belong to a single Palaeoproterozoic *Southern Terrane* as part of a wider tripartite sub-division of the Shackleton Range based on differences in age and isotopic composition. A Pan-African *Northern Terrane*, broadly equivalent to the Högbom nappe, and a 'Grenville' *Eastern Terrane* were also determined, although the latter is poorly constrained and marred by confusion on the locality/identification of collected samples (see Kleinschmidt 2021). From here, we focus only on the Palaeoproterozoic Read Group and below summarise its existing field, geochronological, and geochemical characteristics.

2.1. The Read Group

The Read Group (Tessensohn & Thomson 1990; Buggisch et al. 1994; Clarkson et al. 1995) crops out within the Read Mountains and Du Toit Nunataks and defines the southern belt of the range (Figs. 3 & 4). It forms an east–west trending anticlinal window of high-grade

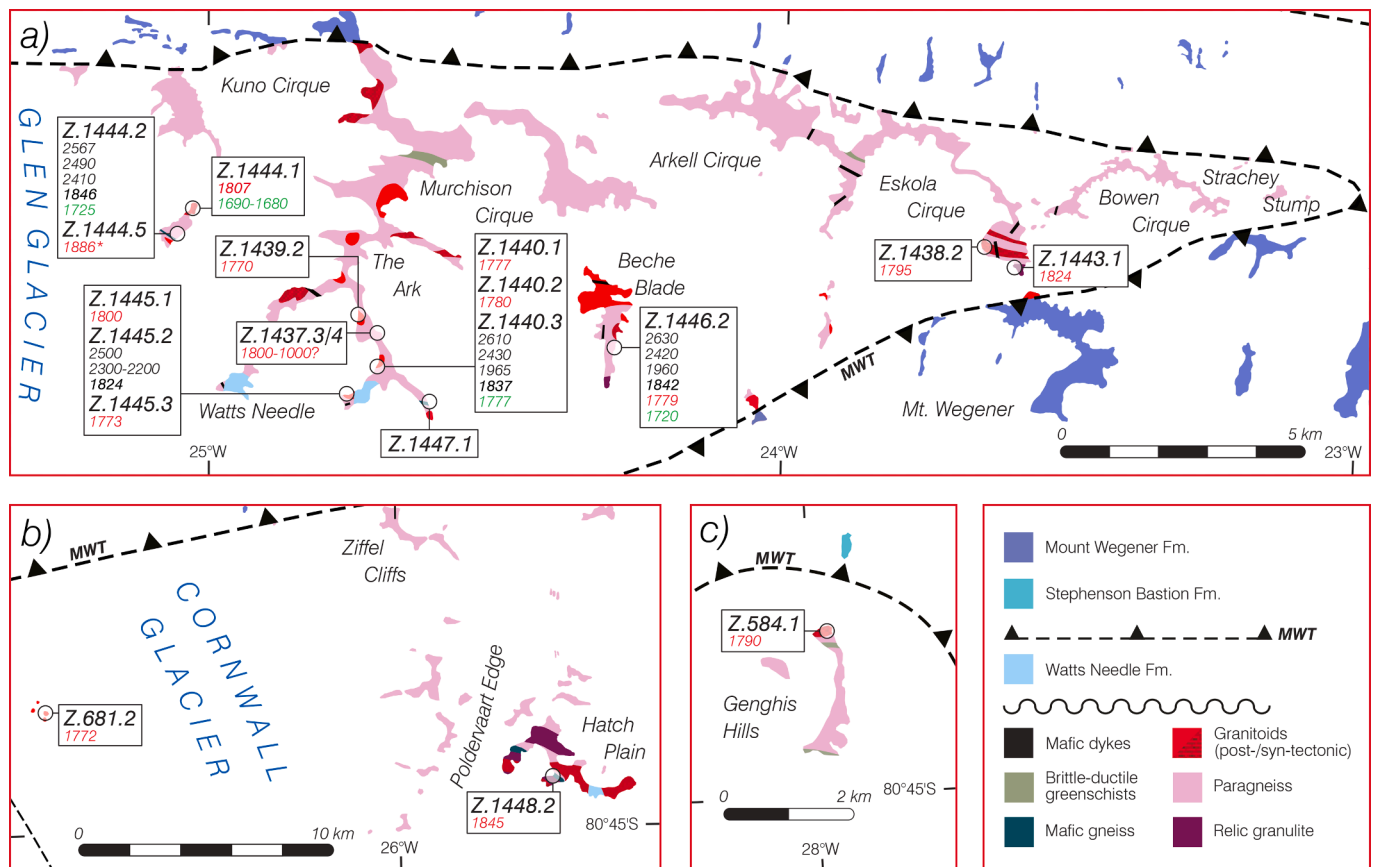


Fig. 4. Sketch geological map of Read Group exposures at a) Read Mountains, b) Du Toit Nunataks and Cornwall Glacier, c) Genghis Hills. White circles show sample locations, boxes show sample number (bold) and key U-Pb zircon ages from this study (red: magmatic, green: metamorphic, black: detrital/inherited). Maps modified and inferred from Clarkson et al. (1995), Schubert & Olesch (1995).

metamorphic units, surrounded by low-grade Cambrian shales and siliciclastics of the Mount Wegener Formation (Buggisch et al. 1994; Buggisch & Henjes-Kunst 1999) which were overthrust during the Ross orogeny (c. 615–470 Ma; Goodge 2020). Both the Read Group and Mount Wegener Formation pinch out to the east, although the latter may continue westward, forming a continuous nappe with the lithologically similar Stephenson Bastion and Wyeth Heights formations (Buggisch et al. 1994; Clarkson et al. 1995). Read Group exposures in this western region are rare and are only tentatively ascribed close to Stephenson Bastion at Genghis Hills (Buggisch & Kleinschmidt 1999) (Fig. 4c), likely because this region was downfaulted along NNW-SSE trending detachments related to Mesozoic rifting in the Weddell Sea (Krohne et al. 2018). The Read Group is thought to be autochthonous, given it has largely escaped metamorphic overprinting by Pan-African age orogenesis and is unconformably overlain by shallow-marine sediments of the Mesoproterozoic Watts Needle Formation (Marsh 1983; Clarkson et al. 1995; Jacobs et al. 2026). This, and the fact it faces the interior of the East Antarctic shield, has led to the Read Group's incorporation as part of a composite cratonic core concealed beneath much of the East Antarctic Ice Sheet.

2.1.1. Lithological subdivisions and structural evolution

The complex of igneous and metamorphic suites which comprise the Read Group are summarised here and are largely drawn from the works of Talarico & Kroner (1999), which we refer to for a more detailed overview. Relic granulite units, mapped as enderbites or charnockites by previous workers (e.g., Schubert & Olesch 1995), occur sporadically on the western spur of Hatch Plain and the ridge between Eskola Cirque and Bowen Cirque (Fig. 4 & Fig. 5a–e), and have also been reported from the southern tip of Beche Blade (Talarico & Kroner 1999). These units are defined by medium-grained granoblastic felsic to intermediate gneisses, often inter-banded with metre-scale, retrogressed mafic gneisses (Fig. 5b–c). The gneisses carry mm-scale garnet and orthopyroxene porphyroblasts both extensively replaced by chlorite and epidote (Fig. 5d–e), particularly within the Eskola Cirque region, and display highly lobate grain boundaries in thin section, all indicative of at least granulite-facies metamorphism (M1) which was subsequently retrogressed to amphibolite (M2, M3) and then greenschist facies conditions (M4) (Table 1). Thermobarometric studies on the Hatch Plain enderbites confirm metamorphic conditions reached up to 830 °C/~8 kbar (M1 in Table 1) (Schubert & Will 1994; Schubert & Olesch 1995; Talarico & Kroner 1999).

Compositionally diverse sequences of amphibolite-facies (690 °C/~5–6 kbar; Schubert & Olesch 1995; M2, M3 in Table 1) paragneisses form the most common rock types within the Read Group (Clarkson et al. 1995; Talarico & Kroner 1999; Fig. 6a–c). They are interpreted as paragneiss sequences given their fine-medium grained pelitic or quartzitic matrix and interlayering quartzitic and calcareous metasediments (Fig. 6a). Biotite-hornblende varieties form much of the southeast-trending ridge immediately south of The Ark, where they are sub-vertically interlayered with mafic gneisses, quartzites, and calc-silicates (Fig. 6a–c). These may represent volcanoclastic packages given the abundance of amphibole and comparatively low quartz. Talarico & Kroner (1999) report cordierite- and sillimanite-bearing paragneisses further north towards the peneplain between Glen Glacier and Murchison Cirque, and eastward to Beche Blade and the dividing ridge between Arkell Cirque and Eskola Cirque. For simplicity and their interlayered association within the field (Clarkson et al. 1995; Talarico & Kroner 1999), we have grouped these units together and interpret them as a continuous sequence of medium- to high-grade metasedimentary rocks. Most paragneiss sequences are migmatized, stromatic to nebulitic quartz-plagioclase leucosomes a common feature throughout. Several generations of leucosomes formed (Talarico & Kroner 1999), with early garnet-bearing and tightly folded varieties cutting an earlier fine-grained mylonitic matrix.

The gneissic suites are intruded by a series of metaluminous to

peraluminous, garnet-bearing granitic to gabbroic bodies (Clarkson et al. 1995). Some of these bodies, such as those at Eskola Cirque, are bound by steeply N-dipping shear zones and are arranged into lenticular sheets and boudins (Fig. 7a). Internally, they show a mylonitic matrix wrapping megacrystic feldspar augen (Fig. 7b–c), suggesting that they were emplaced syn- or pre-tectonically. Other intrusive suites, for example south of The Ark, are absent of such deformational fabrics from outcrop to thin section scale (Fig. 7d–f) and are instead more indicative of magmatic flow fabrics, with feldspar megacrysts randomly oriented and appearing to have overgrown boundaries with enclosed dioritic xenoliths (Fig. 7d). The plutonic rocks at The Ark may be post-tectonic as a result, representing a later pulse of magmatism possibly associated with several generations of undeformed leucocratic sheeted dykes. The dykes dip gently to the west and are particularly pervasive at The Ark (Talarico & Kroner 1999), where they have clearly been emplaced under brittle conditions, facilitating minor m-scale reverse displacements (Fig. 6a). Like their host rocks, they are consistently garnetiferous and contain coarse-grained orthoclase and microcline which enclose numerous μm -scale spherical to lobate quartz inclusions (Fig. 6b & d). Sporadically, up to 10 m thick lower-greenschist facies shear zones crosscut the gneisses and granitoids and show a fracture cleavage with a top-to-the-W shear sense (Clarkson et al. 1995; Talarico & Kroner 1999). Such fracture zones usually occur close to the Mt. Wegener thrust-nappe according to Talarico & Kroner (1999), who suggest they may relate to partial reworking of the Read Group during the Ross-age nappe tectonics (M5 in Table 1). Finally, localised N-S trending metabasaltic or doleritic dykes crosscut all other lithologies and are inferred as late Proterozoic in age (Clarkson 1981).

2.1.2. Previous geochronological, geochemical and isotopic constraints

The most reliable ages from the Read Group to date come from Hatch Plain in the Du Toit Nunataks, west of the Read Mountains (Figs. 3–4). A garnetiferous metagranite from here was dated at 1850 ± 13 Ma (U-Pb zircon, LA-SF-ICP-MS) and is interpreted as the igneous protolith age for the relic granulite gneisses (Will et al. 2009), whilst whole-rock Rb-Sr analysis on an intrusive granite in this region returned an isochron age of 1763 ± 21 Ma (Pankhurst et al. 1983). These ages might constrain the upper- and lower-limits of M1 respectively. *In-situ* Th-U-Pb monazite dating of cordierite-sillimanite-garnet paragneisses produced the best estimates for the timing of subsequent M3 metamorphism within the Read Group, yielding ages of c. 1670 Ma (Will et al. 2009). A series of whole-rock K-Ar and Rb-Sr ages from across the Read Mountains mostly scatter between 1670 and 1540 Ma (Pankhurst et al. 1983; Buggisch et al. 1994 – and references therein) and probably represent isotopic closure of these systems during post-metamorphic cooling (M4). Ross-age (615–470 Ma) isotopic disturbances of these accessory phases are seemingly absent in the Read Group.

Trace element compositions of mafic gneisses in the Read Group fit best within tholeiitic-MORB fields and suggest they probably formed within a back-arc environment, sourced from relatively juvenile melts with little crustal input (Will et al. 2010). Positive ϵNd and ϵHf values and calculated depleted mantle model ages between 2200–2050 Ma for these rocks and their felsic counterparts are consistent with a juvenile crustal origin (Will et al. 2010). Whole-rock and K-feldspar Pb isotopic data are few in number from the Read Group but suggest a variable but radiogenic Pb mantle source for these rocks, plotting on or above typical crustal growth curve models (e.g., Stacey & Kramers 1975) (Will et al. 2010; Flowerdew et al. 2012).

3. Analytical methods

All Read Group samples analysed were acquired from the British Antarctic Survey (BAS) archives in Cambridge and Keyworth, UK. Twenty samples collected predominantly during the 1994/95 Euro-Shack expedition (two samples collected by P. D. Clarkson in 1968–70 field seasons) were selected for analytical work and included a consort of

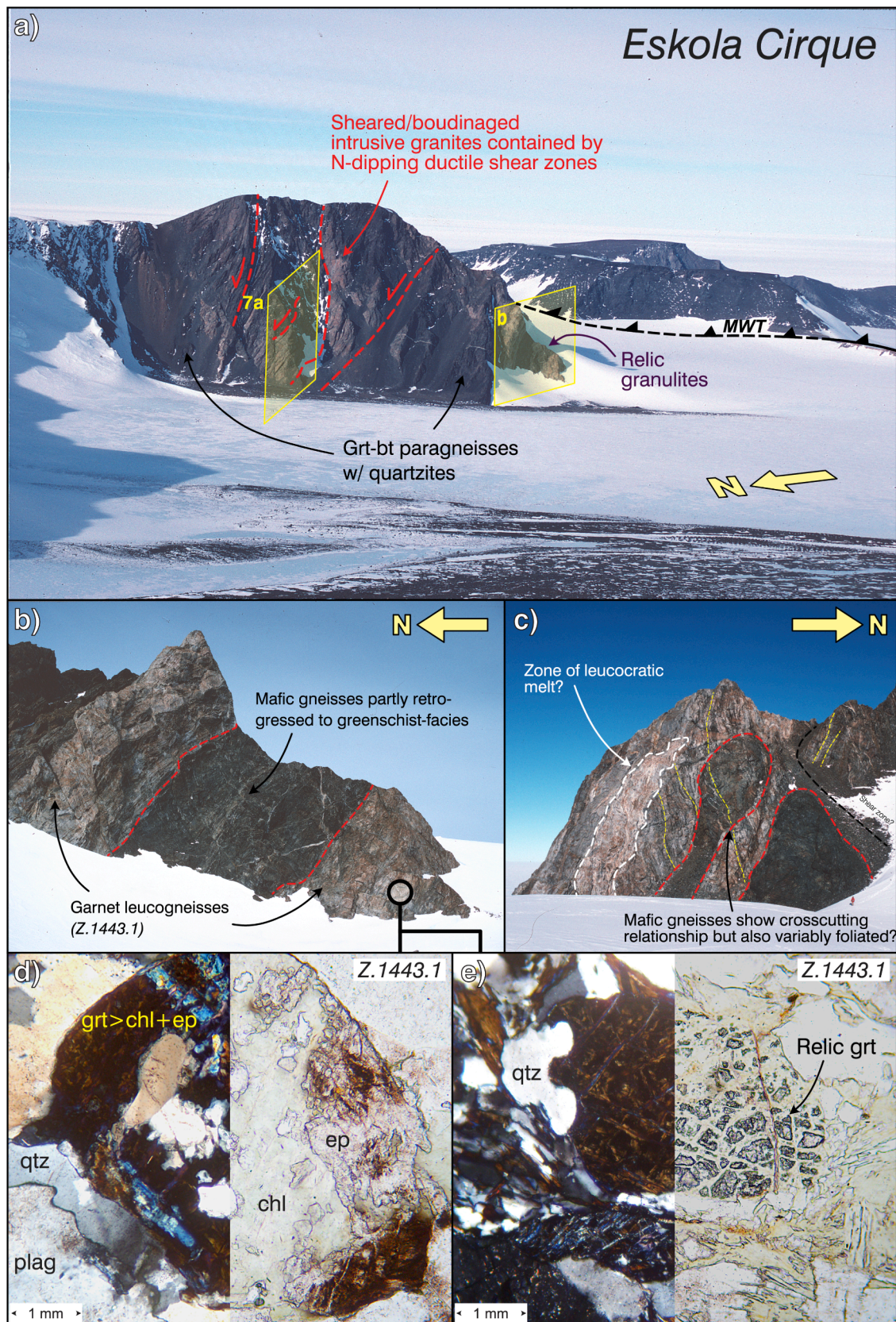


Fig. 5. a-c) Photographs of the eastern ridge of Eskola Cirque. Yellow inset boxes show the frame and perspective of (b) and Fig. 7a, (c) is the reverse perspective view of (b). d-e) XPL/PPL photomicrographs of retrogressed/pseudomorphed garnet from the relic granulites (Z.1443.1) at Eskola Cirque, their approximate location shown in (b). Mineral abbreviations: chl – chlorite, ep – epidote, grt – garnet, plag – plagioclase, qtz – quartz.

Table 1

Metamorphic and structural evolution of the Read Group, partly synthesized by Clarkson et al. (1995) and Talarico & Kroner (1999). Mineral abbreviations: ab – albite, act – actinolite, bt – biotite, chl – chlorite, cpx – clinopyroxene, crd – cordierite, ep – epidote, grt – garnet, hbl – hornblende, Kfsp – K-feldspar, opx – orthopyroxene, plag – plagioclase, qtz – quartz.

Stage	Metamorphic conditions	Stable parageneses	Structural deformation/intrusion	Age
M1	Granulite-facies. 760–830 °C/8 kbar.	opx-cpx-qtz-grt-bt-plag-Kfsp	Isoclinal folding along NE-SW axes. Magmatic underplating?	1850–1825 Ma?
M2 (?)	Upper amphibolite-facies(?)	Bt-hbl-grt-plag-qtz	Migmatization and ductile shearing. Syn- to post-tectonic granitoid intrusion?	1810–1790 Ma?
M3	Upper amphibolite-facies. 630–700 °C/5 kbar.	bt-hbl-grt-sill-crd-Kfsp-plag-qtz	Isoclinal folding along NW-SE axes, plunging E. Migmatization and ductile shearing.	1690–1670 Ma
M4	Upper-greenschist-facies. 500–550 °C/4 kbar.	act-ep-ab-chl	Static overprint associated with post-metamorphic cooling/uplift?	1600–1500 Ma?
M5 (?)	Greenschist-facies		m-dm thick E-W trending brittle-ductile shear zones.	c. 500 Ma?

felsic and mafic orthogneisses, paragneisses, and intrusive granitoids sourced across an east–west transect between Eskola Cirque and Genghis Hills (Fig. 3b & 4, Table 2). Eight samples collected by J. W. Goodge and one sample collected by S. G. Borg are from the Miller and Geologists Range in the cTAMs (Fig. 3) and were loaned from the Polar Rock Repository. These samples were selected to provide an additional comparison to its potential correlatives in the Read Group and included eight felsic to intermediate orthogneisses belonging to the Nimrod Complex (Goodge & Fanning 2016), and a granodiorite from the Cambrian-Ordovician Granite Harbour Intrusive Suite (GHI; Gunn & Warren 1962) (Fig. 3c). Sample selection was partly made according to those previously dated via U-Pb zircon geochronology by Goodge et al. (2001) and Goodge & Fanning (2016). Samples which did not have a published age were dated by SIMS U-Pb zircon geochronology at the NordSIMS facility, Stockholm. The age data acquired from these samples agree well with previous geochronology studies from the Nimrod Complex and GHI suite (e.g., Goodge et al. 2001, Goodge & Fanning, 1999, 2016) and were obtained to assist in the interrogation of our K-feldspar Pb and zircon Lu-Hf isotope datasets. They are thus not presented or discussed in detail in the text but can be found in the [supplementary information](#) and are summarised in Table 2.

3.1. Whole-rock geochemistry

Samples were first crushed and then milled into a fine powder using an agate TEMA mill at the British Antarctic Survey in Cambridge. Whole rock compositions were then analysed using an energy-dispersive X-ray fluorescence (ED-XRF) spectrometer (SPECTROTOM XEPOS HE) at the School of Earth & Environmental Sciences, University of St Andrews. The samples were analysed for major elements on glass beads, prepared by mixing 0.5 g of sample powder with 5 g of Li-borate flux (Li-metaborate and Li-tetraborate in an 80:20 mix), melting in a Pt-crucible in a muffle furnace at 1050 °C, and casting and cooling in a Pt-mould. For trace element analysis, samples were prepared into pressed powder pellets by mixing 4 g of fine sample powder and 1 g of CEREOX binding wax and pressing in a stainless-steel die at 10 t pressure. The measured data were interpreted using the SPECTRO XLABPRO software, using a fundamental parameter-calibration made with > 30 international certified reference materials (CRM; largely silicates) for major

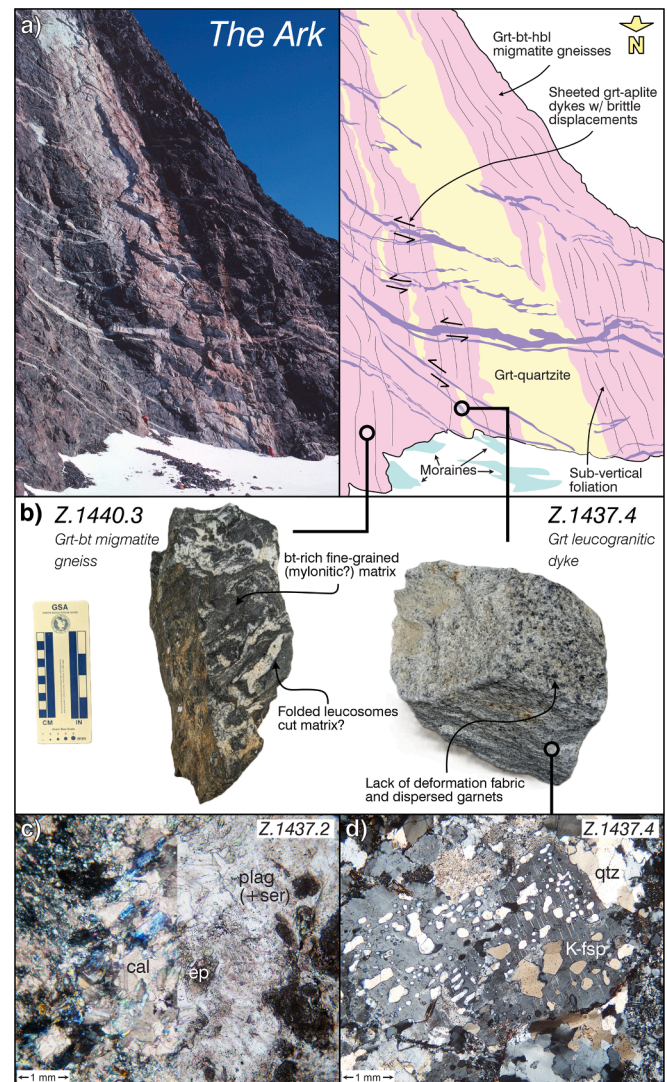


Fig. 6. a) Photograph and sketch of the cliff face section southeast of The Ark. Sub-vertical biotite gneisses interbanded with quartzites and calc-silicates. Both are cut by leucogranitic dykes emplaced under small-scale brittle displacements, b) Hand samples of Z.1440.3 and Z.1437.3 and their approximate locations in (a), c) PPL/XPL photomicrograph of a calc-silicate segregation (Z.1437.2), d) PPL/XPL photomicrograph of a leucogranite dyke (Z.1437.4). Mineral abbreviations: cal – calcite, ep – epidote, K-fsp – K-feldspar, plag – plagioclase, qtz – quartz.

elements. The trace element calibration uses slightly fewer CRM, as values for measured elements are not available for all CRM. Loss-on-ignition (LOI) was determined by weighing ~ 1 g of sample powder before and after heating overnight at 1000 °C.

3.2. U-Pb zircon geochronology

Laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) and secondary ionization mass spectrometry (SIMS) techniques were both employed to carry out U-Pb zircon geochronology across three separate laboratories (University of St Andrews, Swedish Museum of Natural History, University of Edinburgh). Below we summarise the analytical procedures for both techniques. Full U-Pb datasets and zircon images (BSE/CL) are provided in the [supplementary information](#).

LA-ICP-MS analyses were undertaken at the St Andrews Geochronology and Geochemistry facility (StAGE) at the University of St

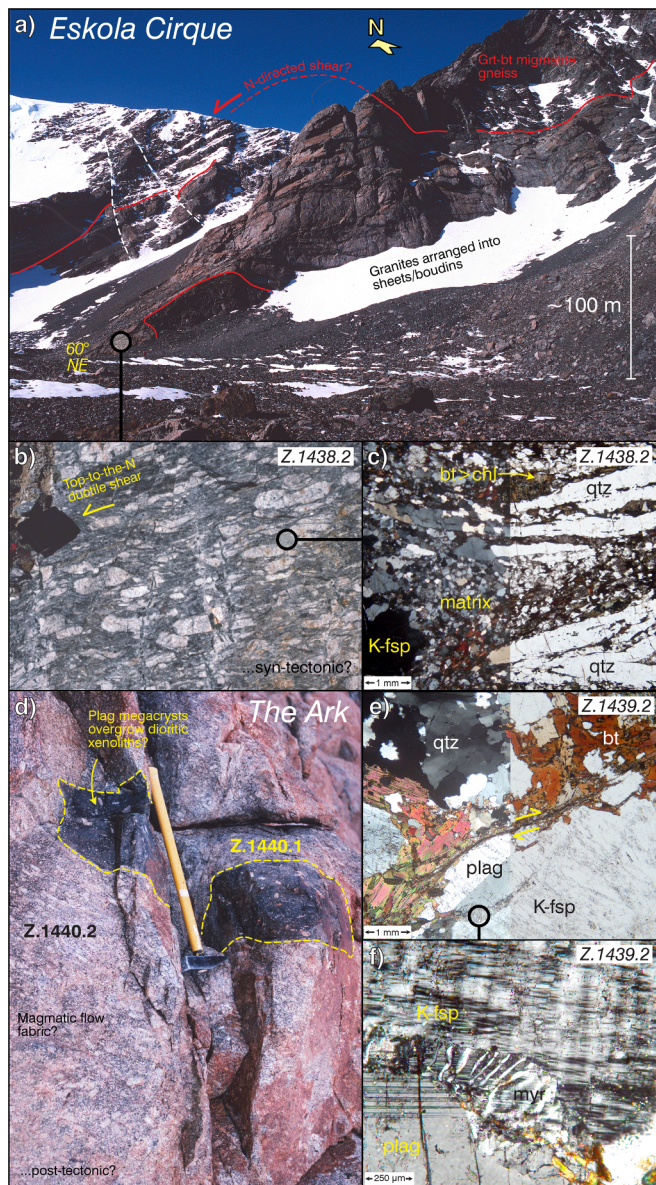


Fig. 7. a–b) Photographs of mylonitic granite sheets along the eastern ridge of Eskola Cirque, the location of (a) is shown in Fig. 5a, c) XPL/PPL photomicrograph of a mylonitic granodiorite (Z.1438.2) from Eskola Cirque, d) Photograph of a porphyritic granitoid (Z.1440.2) and dioritic enclaves (Z.1440.1) at The Ark, e–f) XPL/PPL photomicrograph of an analogous porphyritic granitoid (Z.1439.2) from The Ark. Mineral abbreviations: bt – biotite, chl – chlorite, myr – myrmekite.

Andrews, UK using a New Wave UP213 aperture imaged 213 nm Nd:YAG solid-state laser coupled to an Agilent 8800 Triple Quadrupole ICP-MS operated in single-quad mode. The laser was operated at 45 % output and a frequency of 10 Hz, producing an estimated fluence of 2 to 2.5 J cm⁻². Helium was used as carrier gas at a flow rate of 775 ml/min. 10 ml/min of N₂ and 850 ml/min of Ar were subsequently mixed into the carrier gas, before flowing through a signal smoothing device (“squid”) and then entering the plasma. Additional spot analyses for samples Z.681.2, Z.1440.3, Z.1444.2, and Z.1446.2 were undertaken using a RESOLUTION-SE 193 nm excimer laser ablation system with a Laurus Technic S155 Sample Cell connected to the same ICP-MS set up. The RESOLUTION laser was operated at 8 Hz frequency and a fluence of 5 J cm⁻² with He and sample Ar flow rates of ~ 400 mL/min and ~ 1100 mL/min, respectively. An analytical spot size of 24 or 25 µm was utilised across all analytical runs for standard and unknown analyses. 91500

zircon (1065.4 ± 0.3 Ma; Wiedenbeck et al. 1995) was used as the primary reference material to correct for instrumental drift, mass bias, and laser-induced elemental fractionation. Plešovice (337.13 ± 0.37 Ma; Sláma et al. 2008), OGC (3465.4 ± 0.6 Ma; Stern et al. 2009), and Maniitsoq (3008.70 ± 0.72 Ma; Marsh et al. 2019) zircons were used as secondary reference materials, their respective ages were reproduced within error across all analytical runs. For trace elements, NIST SRM 612 (Pearce et al. 1997) was used as the primary reference material and ²⁹Si was used as an internal standard, assuming a stoichiometric Si content in zircon of 15 wt%. Data reduction of the raw isotopic and elemental values was performed using Iolite 4 (Paton et al. 2011). All age and concordia calculations were done using IsoplotR (Ludwig 2012; Vermeesch 2018) and do not incorporate the long-term uncertainty (~2 %) of the StAGE LA-ICP-MS setup.

SIMS analyses were undertaken at both the NordSIMS facility within the Swedish Museum of Natural History in Stockholm and the Ion Microprobe Facility (IMF) housed at the University of Edinburgh, performed by a CAMECA ims1280 and ims1270 ion microprobe, respectively. At the NordSIMS facility, a focused ¹⁶O₂ primary beam was initiated by a Hyperion II radio-frequency plasma iron source and tuned to a current of 10 nA and a spot size of 15 µm. The methodology otherwise mirrored that of Whitehouse & Kamber (2005), employing a single-ion counting electron multiplier (EM) to measure and amplify the positive secondary ions at a mass resolution (M/ΔM) of c. 5000 and using the in-house developed software for data reduction. The analytical procedures at the IMF were similar to those described above with some minor differences. The ¹⁶O₂ primary beam was operated at ~ 12 kV and a current of 4.9–4.7 nA, producing a spot size of 18 µm. A higher mass resolution of c. 6000 was preferred to avoid Hf-based interferences. At both laboratories, 91500 zircon was used as the primary reference material and measured for its ²⁰⁶Pb/²³⁸U¹⁶O and ²³⁸U¹⁶O₂/²³⁸U¹⁶O compositions to calibrate U–Th–Pb determinations of unknowns (Jeon & Whitehouse 2015). M257 (561 ± 0.3 Ma; Nasdala et al. 2008) or Qinghu (159.5 ± 0.2 Ma; Li et al. 2013) zircons and OGC zircon were analysed as secondary reference materials at the NordSIMS and IMF facilities, respectively. For unknowns with ²⁰⁴Pb determination above limits of detection, common-Pb correction was applied to U–Pb isotopic ratios and ages, assuming the present-day terrestrial ²⁰⁷Pb/²⁰⁶Pb composition of 0.836 estimated by Stacey & Kramers (1975).

3.3. K-feldspar Pb isotopic analysis

Grain mounts of the light fraction retained following zircon separation were imaged by a Zeiss EVO MA 15 SEM housed at CASP and equipped with Oxford Instruments Ultim Max 100 EDS and Unity BSE/EDS detectors to determine major element composition and to locate fresh inclusion-free K-feldspar. In-situ Pb isotopic analyses of K-feldspar were carried out at the National Centre for Isotope Geochemistry, UCD School of Earth Sciences, University College Dublin. Analyses of samples Z.1437.3, Z.1439.2, Z.1444.1, and Z.1448.2 were carried out using a Thermo Fisher Scientific Neptune MC-ICP-MS, coupled with a Teledyne Cetac Analyte G2 193 nm excimer laser ablation system, and closely followed the procedure outlined by Tyrrell et al. (2012). ²⁰²Hg was monitored to correct for isobaric interference of ²⁰⁴Hg on ²⁰⁴Pb and ²⁰³Tl and ²⁰⁵Tl were measured and employed during standard-sample bracketing to correct for instrumental mass bias. Feldspars were ablated along a line at a speed of 2 µm/s with ablation spot sizes of 88 × 88 µm using a laser energy of 3.4 J/cm² with 20 Hz frequency. NIST612 glass served as the primary reference material, whilst accuracy and precision of the analyses were monitored through repeat analyses of Shap K-feldspar (see supplementary information), which produced values within uncertainty of those reported by Liebmman et al. (2023) and Tyrrell et al. (2006). Data were reduced offline using in-house developed software. All remaining samples were analysed with the Teledyne Cetac Analyte G2 193 nm excimer laser attached to a Thermo Fisher Scientific Neoma MC-ICP-MS. The method was similar to that

Table 2

All Read Group, Nimrod Complex and GHI suite samples analysed for whole-rock geochemistry, K-feldspar Pb isotopes, zircon Lu-Hf isotopes and U-Pb zircon geochronology in this study. Their preferred ages and geological meaning are stated.

Sample	Locality	Lithology	WR XRF	K-feldspar Pb (model μ, κ)	Zircon $\epsilon_{\text{Hf}}(t)$	U (ppm)	Th/U	Preferred age(s) $\pm 2\sigma$ (Ma) a: concordia, b: $^{207}\text{Pb}/^{206}\text{Pb}$ weighted mean, c: discordia upper-/lower-intercept, d: kernel density estimator age peaks	Interpretation
READ GROUP									
<i>Relic granulites</i>									
Z.1443.1	Eskola Cirque (–80.67556, –23.81111)	Charnockitic orthogneiss	X			430–4130	0.02–0.36	$^c1824 \pm 23$ ($n = 10$, MSWD = 3.8)	Magmatism
Z.1448.2	Hatch Plain (–80.73778, –25.7025)	Charnockitic orthogneiss	X	10.2, 3.6	+7.5 to +12.7	350–2810	0.36–1.86	$^b1845 \pm 9$ ($n = 35$, MSWD = 0.68)	Magmatism
<i>Mafic gneisses</i>									
Z.1444.3 (M2)	Glen Glacier (–80.68611, –25.11111)	Amphibolite	X						
Z.1447.1 (M2)	Watts Needle (–80.73528, –24.68028)	Amphibolite	X						
Z.1448.3 (M1)	Hatch Plain (–80.73778, –25.7025)	Garnet- amphibolite	X						
<i>Paragneisses</i>									
Z.1440.3	The Ark (–80.7225, –24.74611)	Migmatitic garnet gneiss	X			210–1140	<0.01–0.36	$^d1777, 1837, 1967, 2431, 2611$	Detrital, M2(?) metamorphism
Z.1444.2	Glen Glacier (–80.68611, –25.11111)	Migmatitic garnet gneiss	X			180–1340	0.35–0.42	$^d1724, 1846, 1950, 2200, 2400–2600$	Detrital, M3 metamorphism
Z.1445.2	The Ark (–80.72472, –24.78222)	Migmatitic garnet gneiss	X			90–2170	0.42–0.49	$^d1850, 2506$	Detrital
Z.1446.2	Beche Blade (–80.70167, –24.33306)	Hornblende- biotite gneiss	X			90–8400	0.49–1.86	$^d1718, 1779, 1842, 1963, 2417, 2627$	Detrital, M2(?), M3 metamorphism
<i>Syn-tectonic granitoids</i>									
Z.584.1	Genghis Hills (–80.74583, –28.02361)	Gneissic granodiorite		9.9, 3.9	–6.6 to –3.0	40–150	0.25–1.43	$^b1790 \pm 5$ ($n = 22$, MSWD = 1)	Magmatism
Z.1438.2	Eskola Cirque (–80.69, –23.695)	Mylonitic granodiorite		9.9, 3.9	–1.7 to +2.8	520–1470	0.03–0.46	$^a1795 \pm 5$ ($n = 5$, MSWD = 0.3)	Magmatism
Z.1444.1	Glen Glacier (–80.68611, –25.11111)	Gneissic granodiorite		10.2, 3.8	+1.4 to +5.7	220–710	1: 0.35–0.90 2: 0.20–0.45	1: $^a1807 \pm 16$ ($n = 10$, MSWD = 1.9) 2: $^a1693 \pm 25$ ($n = 2$, MSWD = 0.1)	1: Magmatism 2: M3 metamorphism?
<i>Post-tectonic granitoids</i>									
Z.681.2	W of Du Toit Nunataks (–80.71806, –26.80278)	Granoblastic leucogranite		10.0, 3.7	–5.6 to +1.6	90–3310	0.03–0.50	$^b1772 \pm 4$ ($n = 43$, MSWD = 1.9)	Magmatism
Z.1439.2	The Ark (–80.70917, –24.77028)	Porphyritic quartz monzonite	X	10.2, 3.7		340–1680	0.07–0.84	$^b1770 \pm 7$ ($n = 3$, MSWD = 1)	Magmatism
Z.1440.1/.2	The Ark (–80.7225, –24.74611)	Porphyritic granite (hosting microdiorite)				350–12750	0.07–1.6	$^c1778 \pm 11$ ($n = 19$, MSWD = 1.3)	Magmatism
Z.1445.1	The Ark (–80.72472, –24.78222)	Granoblastic granite				690–8950	0.05–1.69	1803.2 ± 6.8 (Single grain $^{207}\text{Pb}/^{206}\text{Pb}$ age)	?
<i>Leucocratic dykes</i>									
Z.1437.3/.4	The Ark (–80.71556, –24.75583)	Garnet leucogranite	X	10.5, 3.5		3170–16170	0.03–0.47	$^c1800–1000?$	?
Z.1444.5	Glen Glacier (–80.68611, –25.11111)	Garnet aplite				2480–10280	0.06–0.39	$^c1886 \pm 63$ ($n = 8$, MSWD = 0.99)	Xenocrystic?
Z.1445.3	The Ark (–80.72472, –24.78222)	Garnet aplite	X			640–2390	0.04–0.53	$^a1773 \pm 23$ ($n = 2$, MSWD = 0.11)	Magmatism
NIMROD COMPLEX									
PRR-64258	Mt. Albright (–82.833, 157.842)	Layered orthogneiss				30–6120	0.07–0.85	1: 3100? 2: 530?	1: Magmatism? 2: Ross Pb-loss?
PRR-64553	Gerard Bluffs (–83.583, 157.37)	Layered orthogneiss		8.1, 12.5	–0.7 to +1.5	170–2000	0.07–0.13	1: $^c2886 \pm 6$ ($n = 7$, MSWD = 1.5) 2: $^c537 \pm 180$ ($n = 7$, MSWD = 1.5)	1: Magmatism 2: Ross Pb-loss
PRR-64216	Orr Peak (–83.484, 157.90)	Hornblende- biotite orthogneiss		9.5, 3.4	+0.1 to +4.8	150–640	0.32–0.50	1: $^c2515 \pm 11$ ($n = 12$, MSWD = 2.2)	1: Magmatism 2: Ross Pb-loss

(continued on next page)

Table 2 (continued)

Sample	Locality	Lithology	WR XRF	K-feldspar Pb (model μ, κ)	Zircon $\epsilon_{\text{Hf}}(\text{t})$	U (ppm)	Th/U	Preferred age(s) $\pm 2\sigma$ (Ma) a: concordia, b: $^{207}\text{Pb}/^{206}\text{Pb}$ weighted mean, c: discordia upper-/lower-intercept, d: kernel density estimator age peaks	Interpretation
PRR-64571	Gerard Bluffs (–83.575, 157.20)	Biotite-hornblende-garnet orthogneiss		9.8, 3.9	–7.9 to –3.8	140–1210	0.07–0.46	2: $^{587} \pm 229$ ($n = 12$, MSWD = 2.2) $^{532} \pm 5$ ($n = 8$, MSWD = 2.8)	Magmatism
PRR-63903	Camp Ridge (–83.187667, 155.9302)	Layered orthogneiss		9.1, 4.7		70–850	0.02–0.49	a1 : c. 2990 a2 : c. 1730–1700	1: Magmatism 2: Nimrod Pb-loss
PRR-63536 (03-403A; Goodge & Fanning 2016)	Mt. Albright (–82.82645, 154.958633)	Hornblende-biotite orthogneiss		9.0, 4.4				$^{b1720} \pm 9$ ($n = 12$, MSWD = 0.9) (see Goodge & Fanning 2016)	Magmatism
PRR-64780 (99-300; Goodge & Fanning 2016)	Gerard Bluffs (–83.6327, 157.2237)	Hornblende-biotite layered orthogneiss		10.7, 4.4				$^{c1735} \pm 9$ ($n = 13$, MSWD = 1.2) (see Goodge & Fanning 2016)	Magmatism
PRR-15136 (M-28; Goodge et al. 2001)	Camp Ridge (–83.1494, 156.0114)	Orthogneiss		9.0, 4.7				$^{b1730} \pm 14$ (see Goodge et al. 2001)	Magmatism
GHI SUITE									
PRR-63753	Milan Ridge (–83.274483, 156.307183)	Granodiorite		9.5, 4.1		1120–6360	0.03–0.65	$^{a519} \pm 4$ ($n = 6$, MSWD = 2.3)	Magmatism

described above for the Neptune, but with $10^{13}\Omega$ amplifiers installed on the 202 and 204 masses to improve accuracy of the ^{202}Hg correction and measurement of the ^{204}Pb peak. Data were reduced offline using the ‘Pb Isotopes DRS’ data reduction scheme in Iolite 4 (Paton et al., 2011). Pb–Pb plots and calculation of model source indices (μ , ω , κ) and model ages were completed using PbIso (Armistead et al. 2024).

3.4. Zircon Lu–Hf isotopic analysis

Lu–Hf isotopes in zircon were measured by LA-MC-ICPMS at the St Andrews Geochronology (StAGE) laboratory at the University of St Andrews (UK) using an Applied Spectra RESOLUTION-SE 193 nm excimer laser ablation system coupled to a Nu Plasma MC-ICP-MS. All spots were measured with a 50 μm spot either overlapping or adjacent to previously analysed U–Pb isotopic spots, a fluence of 6.5 J/cm^2 , a repetition rate of 10–12 Hz, and an ablation period of 50 s, leading to pit depths on the order of 40–50 μm . The laser was fired thrice at the same spot size before each shot to remove surface contamination; this material was allowed to wash out for 30 s prior to ablation, and each ablation was followed by a 40 s delay prior to the next cleaning shot. The ablated aerosol was transported to the ICPMS with a combination of Ar (nebulizer Ar gas flow ~ 0.95 L/min) and He (~ 300 mL/min). A suite of primary and secondary zircon standards were run every 9 unknowns. 91500 zircon ($^{176}\text{Lu}/^{177}\text{Hf} = 0.000311 \pm 0.000136$; Blichert-Toft 2008) was used as the primary standard for normalizing final $^{176}\text{Lu}/^{177}\text{Hf}$ ratios, whilst Plešovice ($^{176}\text{Hf}/^{177}\text{Hf} = 0.282482$; Sláma et al. 2008) was used as the primary standard for normalizing final $^{176}\text{Hf}/^{177}\text{Hf}$ ratios. 91500, Mud Tank, Grey Hill and Temora2 zircons were used as secondary standards. Data were collected across three sessions within the same overall run. In the second session, the measured $^{178}\text{Hf}/^{177}\text{Hf}$ ratio for several of the standards was sometimes above the recommended range of acceptable values (1.46688–1.46746; Spencer et al. 2020), although only one unknown sample was affected. However, the analyses with elevated $^{178}\text{Hf}/^{177}\text{Hf}$ ratios have $^{176}\text{Hf}/^{177}\text{Hf}$ values within analytical uncertainty of the sample mean, and therefore they have not been excluded (see the supplementary information for a full outline of this method and the obtained values of the secondary standards). Data reduction followed the methods detailed in Granseth et al. (2021) as implemented in the ‘Hf Isotopes XBeta’ data reduction scheme for Iolite 4 (Paton et al., 2011).

$\epsilon_{\text{Hf}}(\text{t})$ values for all unknowns were calculated using the ^{176}Lu decay constant of Söderlund et al. (2004) and CHUR parameters of Bouvier et al. (2008) and plotted using the HafAn plugin for GCDKit 6.3.0 (Janoušek et al. 2006; Janoušek 2024).

4. Results

4.1. Whole-rock geochemistry and inferred tectonic settings

Thirteen samples encompassing the major different lithotectonic units were chosen for whole-rock major and trace element geochemical analyses (Fig. 8a–i). Two relic granulites (Z.1443.1, Eskola Cirque; Z.1448.2, Hatch Plain) are metaluminous to peraluminous and show relatively high SiO_2 content (>75 wt%), with both belonging to the subalkaline series and granite fields (Fig. 8a–c). Z.1448.2 is significantly more K_2O -rich (4.46 wt%) than Z.1443.1 (0.63 wt%), whereas Z.1443.1 is more Mg-rich ($\text{Mg}\# = 38.7$) than Z.1448.2 ($\text{Mg}\# = 14.2$). Typical trace element discrimination plots for inferring the tectonic settings of granites (e.g., Fig. 8e: Rb vs. Nb + Y or Nb vs. Y; Pearce et al. (1984)) show Z.1443.1 within the volcanic arc field and Z.1448.2 along the volcanic arc/within-plate granites boundary.

Three mafic gneisses from between Hatch Plain and Watts Needle are variably associated with the M1 relic granulites (Z.1448.3) or the M2 paragneiss sequences (Z.1444.3, Z.1447.1). Together, they conform to subalkaline gabbro to gabbroic diorite compositions (Fig. 8a). All three gneisses also fall within the tholeiitic series and correspond to relatively low TiO_2 contents of between 0.8–1.8 wt%, consistent with igneous rather than sedimentary protoliths. The Ti/Y vs. Zr/Y diagram (Fig. 8g) is useful in distinguishing plate margin basalts from those formed in an intra-plate setting (Pearce & Gale 1977). The lower Ti/Y (190–310) and Zr/Y (2–5) ratios show that all three mafic gneisses are consistent with a plate margin origin in this classification. Combined with their reduced Nb/Y ratios (0.2–0.6), such values are fitting of the Iceland mantle array, plotting either within or close to E-MORB fields (Fig. 8h). Fig. 8h also includes whole-rock data from Will et al. (2010), who analysed five mafic gneisses from the Read Mountains also with E-MORB type compositions and elevated Rb and Ba concentrations which mirror those of this study (4–50 times MORB compositions). Will et al. (2010) proposed a back-arc setting for the protolith of the mafic gneisses and suggested

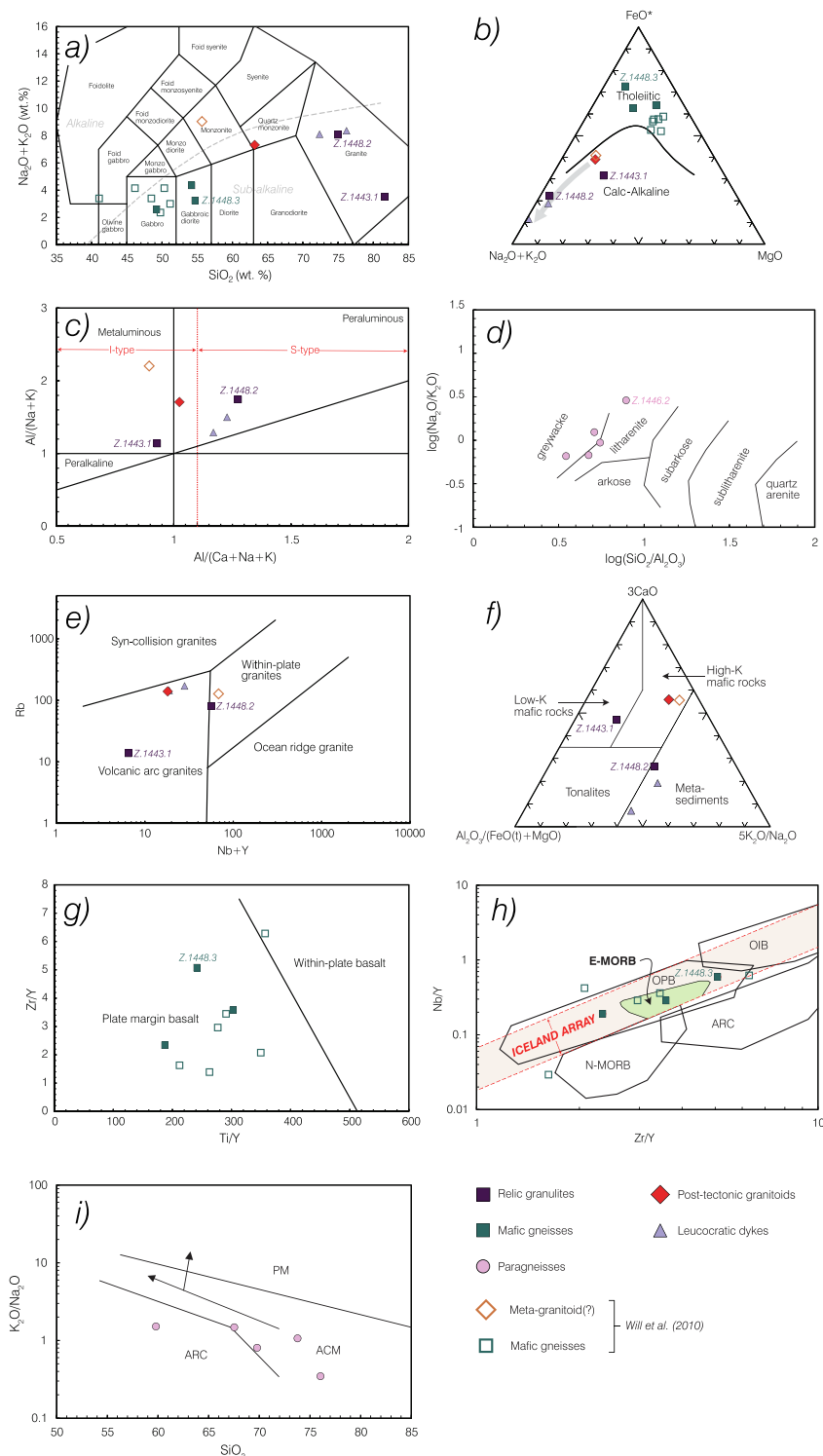


Fig. 8. Major and trace element classification and tectonic setting discrimination diagrams for Read Group samples from this study and Will et al. (2010). a) SiO_2 vs. $\text{Na}_2\text{O} + \text{K}_2\text{O}$; Middlemost (1994) b) FeO^* vs. MgO vs. $\text{Na}_2\text{O} + \text{K}_2\text{O}$; Irvine & Baragar (1971), c) $\text{Al}/(\text{Ca} + \text{Na} + \text{K})$ vs. $\text{Al}/(\text{Na} + \text{K})$; Maniar & Piccoli (1989), d) $\log(\text{SiO}_2/\text{Al}_2\text{O}_3)$ vs. $\log(\text{Na}_2\text{O}/\text{K}_2\text{O})$; Pettijohn et al. (1987), e) $\text{Nb} + \text{Y}$ vs. Rb ; Pearce et al. (1984), f) 3CaO vs. $5\text{K}_2\text{O}/\text{Na}_2\text{O}$ vs. $\text{Al}_2\text{O}_3/(\text{FeO}(\text{t}) + \text{MgO})$; Laurent et al. (2014), g) Ti/Y vs. Zr/Y ; Pearce & Gale (1977), h) Nb/Y vs. Zr/Y ; Fitton et al. (1997), i) SiO_2 vs. $\text{K}_2\text{O}/\text{Na}_2\text{O}$; Roser & Korsch (1988). The mafic gneisses include both the M1 and M2 sub-groupings referred to in the text and Z.1448.3 (M1) is labelled to distinguish it from the two M2 mafic gneisses.

spikes in concentration of the fluid-mobile elements can be explained by subsequent incorporation of these rocks into a volcanic arc setting.

Five paragneisses from Glen Glacier, The Ark and Beche Blade show variable SiO_2 content of between 59–76 wt%. Most samples are Al_2O_3 -rich (13–17 wt%), although Z.1446.2 from Beche Blade has an Al_2O_3 content of 9.7 wt% and is also reduced in K_2O whilst more concentrated

in Na_2O relative to the other paragneisses. On a geochemical classification of metasedimentary rocks from Pettijohn et al. (1987) (Fig. 8d), Z.1446.2 indicates a litharenite protolith whilst the remaining samples cluster closer to greywacke compositions. Tectonic setting discrimination diagrams such as SiO_2 vs. $\text{K}_2\text{O}/\text{Na}_2\text{O}$ (Fig. 8i; Roser & Korsch 1988) suggest an active continental margin to arc setting for the sedimentary

protoliths of these rocks.

All three intrusive felsic rocks from The Ark in the central Read Mountains are undeformed and interpreted as post-tectonic. They conform to subalkaline and peraluminous compositions (Fig. 8a–c), although there are some differences between the porphyritic granitoid (Z.1439.2) and the undeformed leucocratic dykes (Z.1437.4 & Z.1445.3). Z.1439.2 falls along the quartz monzonite-granodiorite fields with elevated ferromagnesian major oxide content (i.e., FeO, MgO, MnO, TiO₂), whilst the post-kinematic dykes are more granitic (SiO₂: 71–76 wt%) and plot within the S-type field on the alumina saturation index diagram (Fig. 8c). Z.1439.2 is significantly enriched in large-ion lithophile elements (LILE) such as Ba (>2000 ppm) and Sr (>600 ppm) and also compatible elements such as Ni (>200 ppm), all typical for late Archaean to early Proterozoic sanukitoids which are thought to result from interaction between recycled crustal material with peridotitic mantle (Martin et al. 2009). The leucocratic dykes are slightly less enriched in these elements and are closer to typical tonalite-trondhjemite-granodiorite values, similar to the relic granulites. The tertiary diagram from Laurent et al. (2014) (Fig. 8f) attempts to determine the likely source components for granitoids. Z.1439.2 sits within the high-K mafic rocks field whilst the two leucocratic dykes indicate melting from a metasedimentary protolith, the obvious candidate being the paragneisses in which they intrude. All the intrusive rocks fall within the volcanic arc granite field in the Nb + Y vs. Rb and Y vs. Nb tectonic setting discrimination diagrams (Fig. 8e).

4.2. U-Pb zircon geochronology

Here we report U-Pb zircon geochronology data for the respective lithotectonic units of the Read Group (see also Table 2). All single grain ages and kernel density estimator age peaks reported refer to their ²⁰⁷Pb/²⁰⁶Pb age. BSE and/or CL images of zircons for each sample and the analytical spot positions can be found within the [supplementary information](#).

4.2.1. Relic granulites

Z.1443.1 and Z.1448.2 are both medium-grained charnockitic orthogneisses belonging to the suite of relic granulites preserved within the Read Group. Z.1443.1 was sampled from Eskola Cirque in the eastern Read Mountains and yields elliptical, dark brown and CL-dark zircons ranging from 50 to 250 µm in length and 1:1 or 2:1 in aspect ratio. Thirty LA-ICP-MS analyses (StAGE, St Andrews) and ten SIMS analyses (NordSIMS, Stockholm) were undertaken. Only one LA-ICP-MS analysis is concordant at 1816 ± 43 Ma (²⁰⁷Pb/²⁰⁶Pb age = 1831 Ma), whilst the remaining data are discordant and do not form a single isochron regression. Most zircons from Z.1443.1 are variably metamict, often severely fragmented, and were likely susceptible to the effects of common-Pb and Pb-loss. The smaller analytical spot size (c. 15 µm) and depth of SIMS can assist in analysing only preserved domains. Nine SIMS analyses produce an isochron regression with an upper-intercept age of 1823 ± 26 Ma (2σ, n = 9, MSWD = 4.2) and therefore close to the single concordant analysis from the LA-ICP-MS session. Together, they form an upper- and lower-intercept age of 1824 ± 23 Ma and 381 ± 30 Ma (2σ, n = 10, MSWD = 3.8) (Fig. 9a), respectively, the former interpreted as the most accurate time estimate for igneous crystallisation of the gneiss protolith.

Z.1448.2 is from Hatch Plain in the Du Toit Nunataks, west of the Read Mountains. Zircons from Z.1448.2 are subhedral prismatic, dark brown and range between 50 to 250 µm in length and 2:1 or 3:1 in aspect ratio. Often, they retain xenocrystic cores truncated by thin (<25 µm) rims. Sixty-five LA-ICP-MS analyses were undertaken across fifty-five cores and ten rims. Concordant core analyses produce a concordia age of 1848 ± 11 Ma (2σ, n = 26, MSWD = 2.4) (Fig. 9b), whilst those between -2 and 7 % discordance produce a similar ²⁰⁷Pb/²⁰⁶Pb weighted mean age of 1845 ± 9 Ma (2σ, n = 35, MSWD = 0.7) (Fig. 9b), the preferred age for crystallisation of the igneous protolith. Rims and

overgrowths are variably discordant and show lower Th/U (0.02–0.30) than cores (>0.30). An isochron regression including rims and overgrowths only produces upper and lower intercept ages of 1765 ± 169 Ma and 177 ± 110 Ma (2σ, n = 10, MSWD = 22), respectively. By including core analyses, the intercept ages shift to 1842 ± 17 Ma and 211 ± 30 Ma (2σ, n = 59, MSWD = 4.4) (Fig. 9b).

4.2.2. Paragneisses

The four paragneiss samples (Z.1440.3, Z.1444.2, Z.1445.2 & Z.1446.2) collected between Glen Glacier, The Ark and Beche Blade all show variable zircon grain morphologies (rounded to sub-prismatic) and similar multi-modal U-Pb age distributions in concordia space, consistent with a sedimentary origin. An adaptive kernel density estimator plot (KDE) in IsoplotR (Vermeesch 2018) for a total of 170 analyses across all samples returns a primary ²⁰⁷Pb/²⁰⁶Pb age peak of c. 1818 Ma, and a secondary peak at c. 2433 Ma (Fig. 9c). When using a discordance filter of between -2 and 7 %, the age distribution (n = 70) shifts to a primary peak at c. 1827 Ma and two secondary peaks at c. 2435 and c. 2600 Ma. Fifteen concordant analyses (~21 % total population) younger than 1810 Ma are recorded. Given the paragneisses are crosscut by the intrusive granitoids and dykes (1810–1770 Ma), zircons of this age likely represent isotopic resetting either during syn- to post-tectonic granitoid intrusion and/or M2 metamorphism. Excluding these analyses does not appear to shift the age distribution significantly (i.e., less than 1 myr). The older age peaks often correspond to rounded/sub-angular zircons with bright CL characteristics, whilst the c. 1850–1830 Ma population tend to be sub-prismatic and CL-dark, resembling those from the relic granulites (e.g., Z.1448.2).

4.2.3. Syn-tectonic granitoids

Z.584.1 is a well-banded gneissic granodiorite from Genghis Hills – a north–south trending ridge west of the Read Mountains and Stephenson Bastion. It contains pale-brown and CL-dark sub-prismatic zircons, 100–300 µm in length and 2:1 to 4:1 in aspect ratio. Some grains are mantled by thin (<25 µm) CL-bright rims. Twenty-five SIMS analyses targeted on fifteen cores and ten rims were carried out at the NordSIMS facility. Analyses between -2 and 7 % discordance yield a ²⁰⁷Pb/²⁰⁶Pb weighted mean age of 1790 ± 5 Ma (2σ, n = 22, MSWD = 1) (Fig. 9d). Concordia ages for cores (1788 ± 5 Ma) are slightly older than rims (1785 ± 11 Ma), but they are within uncertainty.

Z.1438.2 is a mylonitic granodiorite from Eskola Cirque in the eastern Read Mountains. Z.1438.2 contains dark brown, rounded-prismatic zircons ranging from 50 to 300 µm long and 1:1 and 3:1 in aspect ratio. Coarser grains occasionally retain core-rim structures and all are non-luminescent in CL. Thirteen SIMS analyses targeted on eleven cores and two rims were undertaken at the IMF in Edinburgh. An isochron regression including both cores and rims produces an upper intercept of 1795 ± 10 Ma and a lower intercept of 332 ± 156 Ma (2σ, n = 11, MSWD = 2.2) (Fig. 9e). This excludes one older concordant core (1904 ± 14 Ma) and one significantly discordant analysis. Five concordant cores produce an indistinguishable concordia age of 1795 ± 5 Ma (2σ, n = 5, MSWD = 0.3). Most grains yield low Th/U ratios of less than 0.15.

Z.1444.1 is a gneissic granodiorite from near Glen Glacier. It contains brown, mostly sub-prismatic zircons 100–200 µm in length and 2:1 or 3:1 in aspect ratio. Some stubby grains roughly 50–100 µm in diameter also occur. Twenty LA-ICP-MS analyses across both populations were undertaken, with ten concordant analyses producing a concordia age of 1807 ± 16 Ma (2σ, n = 10, MSWD = 1.9) (Fig. 9f). Two concordant stubby grains have a much younger age of 1693 ± 25 Ma (2σ, n = 2, MSWD = 0.1) and one additional stubby grain is slightly older at 1755 ± 44 Ma. These 1807 and 1693 Ma age signatures generally correlate with higher (0.35–0.90) and lower (0.20–0.40) Th/U, respectively, which is consistent with their difference in morphology and may imply they represent successive magmatic and metamorphic episodes.

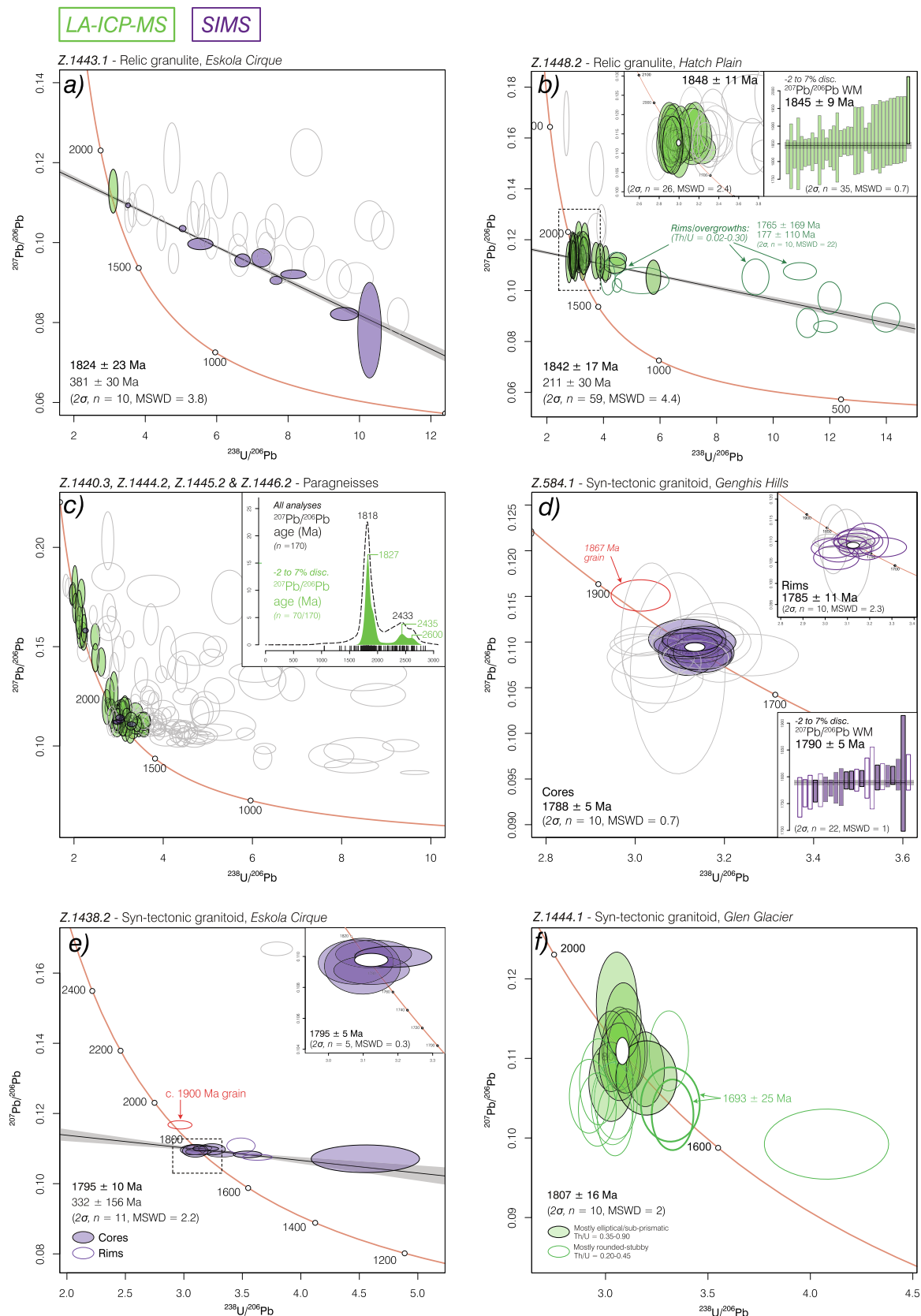


Fig. 9. Tera-Wasserburg U-Pb concordia diagrams and interpreted age calculations for Read Group samples. a-b) Relic granulites, c) Paragneisses, d-f) Syn-tectonic granitoids, g-j) Post-tectonic granitoids, k-l) Leucocratic dykes. Z.1437.3 & Z.1437.4 did not produce any well constrained or geological meaningful data and are not included but can be viewed in the [supplementary information](#). Green and purple ellipses refer to LA-ICP-MS and SIMS analyses respectively.

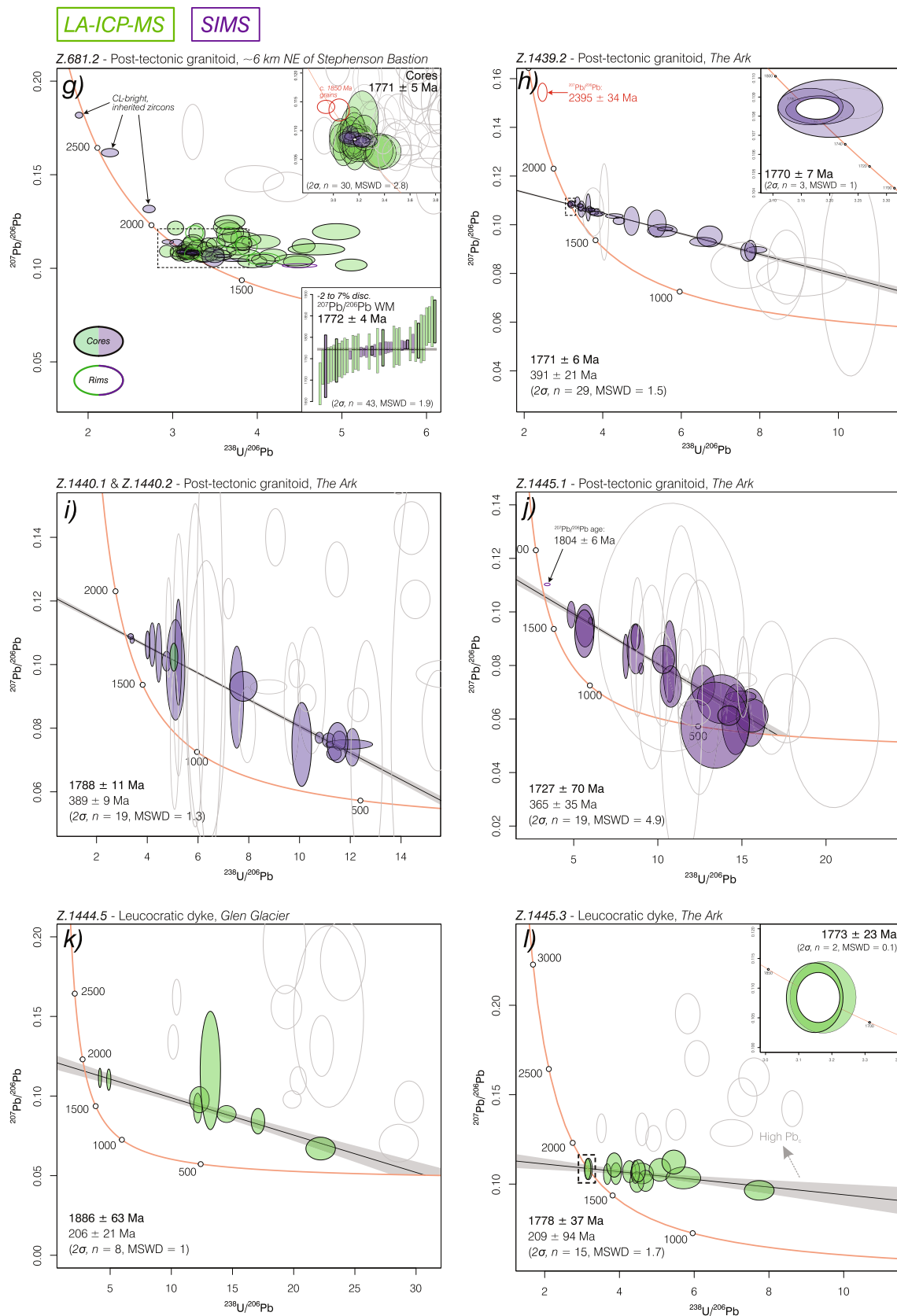


Fig. 9. (continued).

4.2.4. Post-tectonic granitoids

Z.681.2 is a granoblastic leucogranite from a group of small nunataks west of the Du Toit Nunataks. It contains brown, prismatic to sub-

prismatic zircons 100–150 μm in length and with 2:1 aspect ratio. Most grains form subtle core-rim structures and are non-luminescent in CL except for a minority of bright xenocrystic cores. Fifty-nine LA-ICP-

MS analyses and twenty-seven SIMS analyses were carried out at the StAGE and NordSIMS facilities, respectively. Cores and rims within ± 2 and 7 % discordance produce a $^{207}\text{Pb}/^{206}\text{Pb}$ weighted mean age of 1772 ± 4 Ma (2σ , $n = 43$, MSWD = 1.9) (Fig. 9g) and generally low Th/U (<0.15), indicating they are cogenetic. Cores alone return an indistinguishable concordia age of 1771 ± 5 Ma (2σ , $n = 30$, MSWD = 2.8) (Fig. 9g). Additionally, two older cores are concordant at 1862 ± 11 Ma (2σ , $n = 2$, MSWD = 0.027), whilst three marginally discordant CL-bright cores show $^{207}\text{Pb}/^{206}\text{Pb}$ ages between 2670–2120 Ma and elevated Th/U of 0.3–0.9.

Z.1439.2 is a porphyritic quartz monzonite from The Ark in the western Read Mountains. Zircons from Z.1439.2 are brown and CL-dark, long-prismatic or sometimes needle-shaped, and up to 250 μm in length and 4:1 in aspect ratio. Thirty-seven SIMS analyses were carried out at the NordSIMS facility. Three analyses were concordant and yielded a concordia age of 1770 ± 7 Ma (2σ , $n = 3$, MSWD = 1) (Fig. 9h). The remaining data for Z.1439.2 produced a discordia array with a very similar upper-intercept age of 1771 ± 6 Ma and a lower-intercept of 391 ± 21 Ma (2σ , $n = 29$, MSWD = 1.5) (Fig. 9h). Seven analyses were rejected due their large error resulting from common-Pb correction. One additional analysis was not included in the calculation given it is close to concordia at c. 2400 Ma and thus represents an inherited grain.

Z.1440.1 is a microdiorite enclave hosted within a porphyritic granite (Z.1440.2) from The Ark. Zircons from both samples are brown and CL-dark, with sub-prismatic or acicular form, and ranging between 50–350 μm in length and 2:1 to 5:1 in aspect ratio. Fourteen LA-ICP-MS analyses and forty SIMS analyses were carried out at the StAGE and NordSIMS facilities, respectively. All data from both samples plot discordantly and most are skewed towards high common-Pb, but the best-preserved grains form an isochron regression with upper- and lower-intercept U-Pb discordia ages of 1788 ± 11 Ma and 389 ± 9 Ma (2σ , $n = 19$, MSWD = 1.3) (Fig. 9i), respectively. Thirty-five analyses were excluded from age calculations due to high common-Pb or large error ellipses, both of which were generally associated with severely fragmented grains or unintended spot placement over cracks or inclusions.

Z.1445.1 is a granoblastic granite from the southwestern ridge of The Ark. It contains brown and CL-dark zircons, sub-prismatic to acicular in form, 50–250 μm in length and 2:1 to 4:1 in aspect ratio. Thirty-eight SIMS analyses were undertaken at the NordSIMS facility. All data are either discordant or show significant Pb loss effects towards c. 360 Ma, likely a factor of their extremely high U content (up to 9000 ppm). Analyses from the most well-preserved grains produce upper and lower intercept ages of 1727 ± 70 Ma and 365 ± 35 Ma (2σ , $n = 19$, MSWD = 4.9) (Fig. 9j), respectively, whilst one marginally discordant analysis (92 % concordant) returns a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1803 ± 7 Ma. However, the grain in question is heavily altered and partially cracked, and thus it is uncertain if either of these ages have any geological meaning.

4.2.5. Post-tectonic leucocratic dykes

Z.1437.3 and Z.1437.4 are compositionally identical leucocratic dykes sampled from the series of sheeted dykes on the southeastern ridge of The Ark. Both samples include pale and CL-dark zircons, with acicular form ranging from 50–400 μm in length and 2:1 to 5:1 in aspect ratio. Many grains are internally mosaiced, which may represent relic zircon from intense hydrothermal alteration. Twenty-five and thirty-one LA-ICP-MS analyses were carried out for Z.1437.3 and Z.1437.4, respectively. All data are discordant and bimodally skewed towards $(^{207}\text{Pb}/^{206}\text{Pb})_c$ ratios of c. 0.19 or 0.59. Forcing a linear regression through the least affected analyses produced upper-intercept ages ranging between 1800 and 1000 Ma with exceptionally high MSWD (>36), meaning these data cannot be reliably interpreted for geological meaning. Most analyses from both samples tend to show Th/U between 0.04 and 0.15, supporting a hydrothermal/metasomatic origin. A Tera-Wasserburg plot of these data is not included in Fig. 9 but can be viewed in the [supplementary information](#).

Z.1444.5 is a garnetiferous aplite dyke from a northeast-southwest ridge next to Glen Glacier. Zircons from Z.1444.5 are pale and CL-dark, with prismatic form and between 50–300 μm in length and 2:1 to 4:1 aspect ratio. Akin to those from Z.1437, the grains show signs of intense hydrothermal alteration internally. Twenty-three LA-ICP-MS analyses were undertaken, with eight of these analyses producing upper-intercept and lower-intercept ages of 1886 ± 63 Ma and 206 ± 21 Ma (2σ , $n = 8$, MSWD = 0.99) (Fig. 9k), respectively. Given Z.1444.5 cuts its host paragneisses (Z.1444.2), which were likely deposited at c. 1810 Ma, these zircons may be xenocrystic, or the upper intercept age is over-estimated – possibly due to common-Pb and Pb-loss resulting from fluid-assisted emplacement. This, and the fact that the aplite dykes are relatively undeformed, suggests that Z.1444.5 was probably emplaced much later and likely after the intrusion of the gneissic granites found close by (Z.1444.1; c. 1810 Ma).

Z.1445.3 is a garnetiferous aplite dyke from the southwestern ridge of The Ark. It is lithologically equivalent to Z.1444.5 and crosscuts its host paragneisses (Z.1445.2). Z.1445.3 contains brown and CL-dark zircons with sub-prismatic to rounded form, ranging from 50–250 μm in length and 1:1 to 4:1 in aspect ratio. Grains are variably fractured but otherwise structurally homogeneous and do not appear metasomatized. Twenty-seven LA-ICP-MS analyses were undertaken, with two concordant analyses returning a concordia age of 1773 ± 23 Ma (2σ , $n = 2$, MSWD = 0.11) (Fig. 9l). One analysis on a core returns a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2110 ± 174 Ma and is thus probably xenocrystic. Eleven analyses with high common-Pb were rejected, whilst the remaining discordant data produce an indistinguishable upper-intercept of 1778 ± 37 Ma and a lower-intercept of 209 ± 94 Ma. (2σ , $n = 15$, MSWD = 1.7) (Fig. 9l).

4.3. K-feldspar Pb and zircon Lu-Hf isotopic analysis

4.3.1. Read Group

Seven samples analysed for U-Pb zircon geochronology were further selected for Pb isotopic analysis of K-feldspar. One relic granulite (Z.1448.2), three syn-tectonic granitoids (Z.584.1, Z.1438.2, Z.1444.1), two post-tectonic granitoids (Z.681.2 & Z.1439.2), and one leucocratic dyke (Z.1437.3) were investigated to compare differences in Pb isotopic character across the major lithological suites within the Read Group. Feldspars from these samples show $^{206}\text{Pb}/^{204}\text{Pb}$ values between 15.8–16.3, $^{207}\text{Pb}/^{204}\text{Pb}$ of 15.3–15.5, and $^{208}\text{Pb}/^{204}\text{Pb}$ of 35.3–36.1 (Fig. 10a–b). Using the U-Pb zircon magmatic crystallisation ages determined for these samples above, these Pb isotope compositions yield model source μ ($^{238}\text{U}/^{204}\text{Pb}$) and κ ($^{232}\text{Th}/^{238}\text{U}$) values of 9.8–10.6 and 3.4–4.1 respectively and model ages between 1700–1400 Ma. The five syn- to post-tectonic granitoids dated between 1810–1770 Ma overlap in uranium and thorogenic Pb-Pb space but have lower $^{206}\text{Pb}/^{204}\text{Pb}$ (15.7–16.1) than the 1845 Ma relic granulite (Z.1448.2; 16.1–16.3) (Fig. 10). One exception is the gneissic granodiorite from Genghis Hills (Z.584.1), which shows a slightly more radiogenic $^{206}\text{Pb}/^{204}\text{Pb}$ composition (up to 16.3) as well as higher $^{208}\text{Pb}/^{204}\text{Pb}$ (up to 36.1), producing a distinct linear array in Pb-Pb space straddling just beneath the Pb evolution curve. Feldspars from the leucocratic dyke (Z.1437.3) cluster at higher $^{206}\text{Pb}/^{204}\text{Pb}$ (~ 16.2) and $^{207}\text{Pb}/^{204}\text{Pb}$ (~ 15.5) than the other granitoid suites, instead overlapping somewhat with the 1845 Ma relic granulite (Z.1448.2) (Fig. 10).

Five of the above samples (Z.681.2, Z.584.1, Z.1438.2, Z.1444.1, Z.1448.2) were additionally selected for Lu-Hf isotopic analysis of zircon and these data are presented fully in the [supplementary information](#). The 1845 Ma relic granulite (Z.1448.2) shows $\epsilon\text{Hf}(t)$ values between +7.5 and +12.7 and single-stage depleted mantle model ages which overlap the crystallisation age (1880–1690 Ma) (Fig. 10c). By contrast, the 1810–1790 Ma syn-tectonic granitoids exhibit lower $\epsilon\text{Hf}(t)$ values between -1.7 and +5.7 (Z.1444.1, Z.1438.2) or between -6.6 and -3.0 (Z.584.1). One 1772 Ma post-tectonic granitoid (Z.681.2) overlaps with the syn-tectonic granitoids, showing $\epsilon\text{Hf}(t)$ values between -5.6 and +1.6. Such initial Hf compositions correspond to single-stage depleted

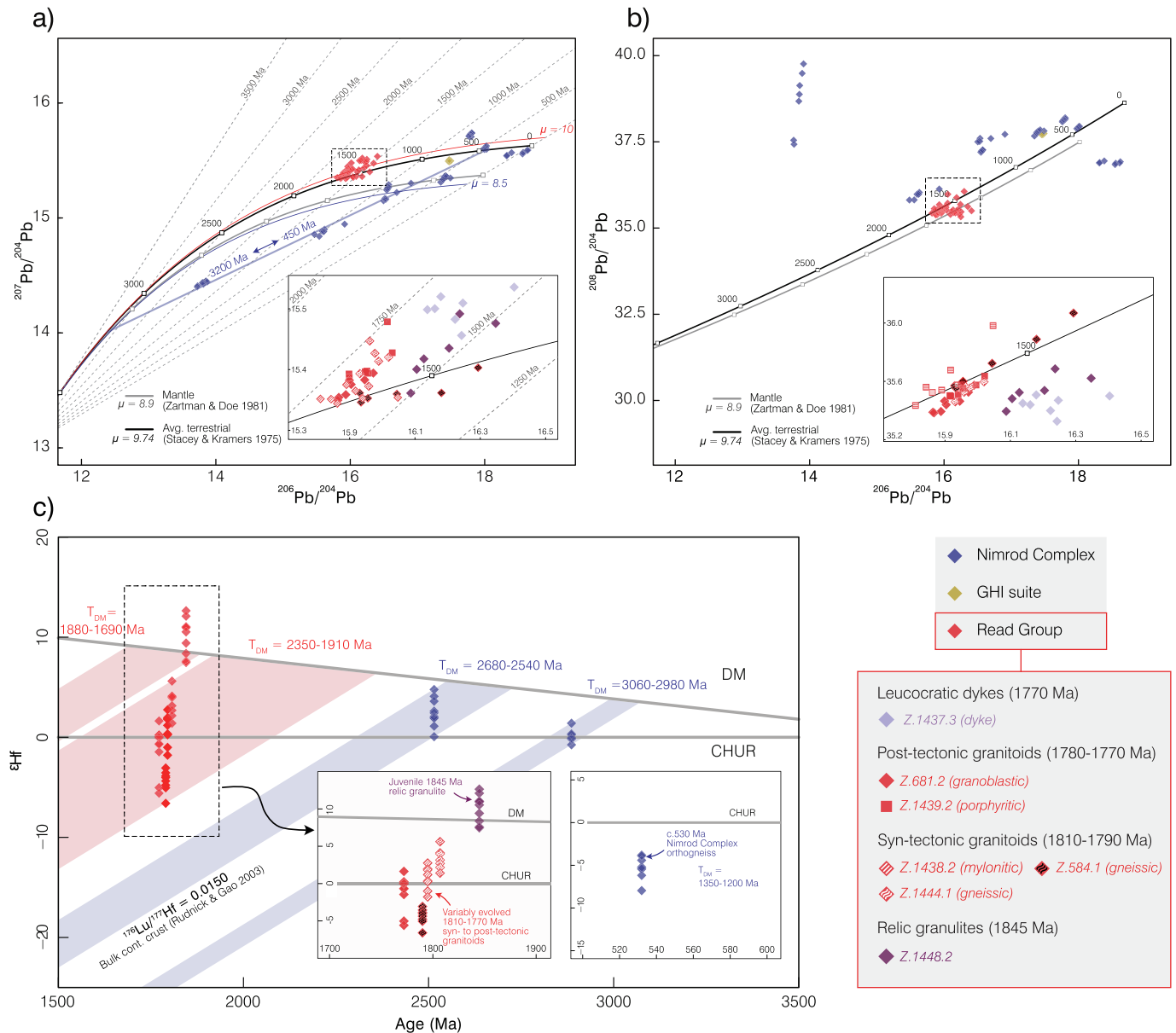


Fig. 10. Uranogenic (a) and thorogenic (b) Pb-Pb diagrams for Read Group, Nimrod Complex and GHI suite K-feldspar samples. The black curve represents the average terrestrial Pb isotopic composition modelled by Stacey & Kramers (1975). The grey curve represents the mantle (Zartman & Doe 1981). The red and blue lines are growth curves with μ of 10 and 8.5 respectively. The grey dashed lines represent palaeoisochrons. c) U-Pb zircon crystallisation ages vs. initial ϵ_{Hf} values for zircon grains analysed from the Read Group and Nimrod Complex. The shaded red and blue regions are the margins of single-stage Hf isotope growth lines for the Read Group and Nimrod Complex respectively, using a $^{176}\text{Lu}/^{177}\text{Hf}$ ratio of 0.0150 (Bulk continental crust; Rudnick & Gao 2003). Their corresponding single-stage model ages are labelled accordingly. Two inset plots are presented: the left-hand inset is a blown-up of the Read Group analyses, showing the inter-sample variation. The right-hand inset shows the 532 Ma orthogneiss (PRR-64571) from the Nimrod Complex. Abbreviations: CHUR – Chondritic Uniform Reservoir, DM – Depleted Mantle.

mantle model ages between 2350 and 1910 Ma (Fig. 10c).

4.3.2. Nimrod Complex & GHI suite

Feldspars from the Nimrod Complex orthogneisses show a wide spread of $^{206}\text{Pb}/^{204}\text{Pb}$ (13.7–18.6), $^{207}\text{Pb}/^{204}\text{Pb}$ (14.4–15.7) and $^{208}\text{Pb}/^{204}\text{Pb}$ (35.8–39.8) compositions, all but one of which plot on or below the Stacey & Kramers (1975) growth curve and mostly on or below the mantle growth curve (Zartman & Doe 1981) (Fig. 10a-b). Protolith and metamorphic ages for these samples (c. 3000–2900, 2515, 1730, or 530 Ma) are known from their U-Pb zircon ages determined by this study (see supplementary information) or by Goodge et al. (2001) and Goodge & Fanning (2016). Using these ages, model source μ and κ values for these feldspars mostly range between 8.1–9.8 and 3.4–12.5,

respectively. In uranium space, up to six samples appear to form an isochron which intercepts the growth curves at c. 3200 and c. 650–440 Ma (Fig. 10a). Two orthogneisses from Orr Peak (PRR-64216; 2515 Ma) and Gerard Bluffs in the Miller Range (PRR-64780; 1735 Ma) do not lie on this isochron. PRR-64216 (Orr Peak) shows higher $^{206}\text{Pb}/^{204}\text{Pb}$ (18.3–18.7) and $^{207}\text{Pb}/^{204}\text{Pb}$ (15.5–15.6) but remains below the Stacey & Kramers (1975) growth curve ($\mu = 9.5$), whilst PRR-64780 (Gerard Bluffs) is significantly more enriched ($\mu = 10.7$). In the thorogenic plot, samples neither cluster nor correlate and are widely scattered towards elevated $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ (Fig. 10b). A felsic orthogneiss from Gerard Bluffs (PRR-64553; 2886 Ma) shows a near vertical linear array with increasingly high $^{208}\text{Pb}/^{204}\text{Pb}$ (37.5–39.8) against consistently low $^{206}\text{Pb}/^{204}\text{Pb}$ (~14.4). A GHI suite granodiorite from Milan

Ridge (PRR-63753; 519 Ma) has Pb isotopic ratios ($^{206}\text{Pb}/^{204}\text{Pb} = 17.5$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.5$; $^{208}\text{Pb}/^{204}\text{Pb} = 37.7$) which yield model source μ and κ values of 9.5 and 4.1 respectively, plotting just below the [Stacey & Kramers \(1975\)](#) growth curve and roughly along the c. 3200 Ma Nimrod Pb-Pb isochron.

Three samples from the Nimrod Complex (PRR-64216, PRR-64553, PRR-64571) were additionally selected for Lu-Hf isotopic analysis of zircon and these data are presented fully in the [supplementary information](#). The c. 2886 Ma felsic orthogneiss from Gerard Bluffs shows $\epsilon\text{Hf}(t)$ values mostly between -0.7 and +1.5 and single-stage depleted mantle model ages between 3053–2976 Ma, whilst the c. 2515 Ma orthogneiss from Orr Peak shows marginally more positive values between +0.1 and +4.8 and model ages of 2712–2542 Ma ([Fig. 10c](#)). A c. 532 Ma orthogneiss (PRR-64571) from Gerard Bluffs showed more negative $\epsilon\text{Hf}(t)$ values between -7.9 and -3.8 and model ages of 1347–1193 Ma ([Fig. 10c](#)).

5. Interpretation of the Read Group

5.1. 1850 – 1825 Ma relic granulites

The zircon $^{207}\text{Pb}/^{206}\text{Pb}$ weighted mean age of 1845 ± 9 Ma for Z.1448.2 ([Fig. 9b](#)) from Hatch Plain in the Du Toit Nunataks closely matches the 1850 ± 13 Ma magmatic protolith age determined by [Will et al. \(2009\)](#) for a nearby orthogneiss. The lesser-constrained age of 1823 ± 17 Ma for Z.1443.1 from Eskola Cirque ([Fig. 9a](#)), some ~ 40 km to the east, may represent a successive magmatic pulse, or that it has simply been skewed to a younger age because of data imprecision. Both samples belong to those mapped as relic granulites preserved across the Read Mountains, although the high-K/shoshonitic geochemistry of Z.1448.2 differs from the more calc-alkaline Z.1443.1 and is typical for post-collisional or within-plate granites. Such a trend might predict Z.1448.2 to have formed later in the orogenic cycle (i.e., post-accretion of the island arc) but given Z.1448.2 is older or at least of broadly equivalent age to Z.1443.1, it is more likely that their difference in geochemistry was primarily driven by difference in composition of their igneous protoliths' parent magmas, with Z.1448.2 sourced from pelitic metasediments (S-type) and Z.1443.1 from low-K mafic rocks (I-type). Isotopically depleted, positive $\epsilon\text{Hf}(1845 \text{ Ma})$ (+7.5 to +12.7) of zircons from Z.1448.2 ([Fig. 10c](#)) show that the relic granulites were likely generated as juvenile melts with relatively little crustal input. The K-feldspar Pb isotopic data from Z.1448.2 contradict this in that they are enriched relative to the average terrestrial Pb growth curve of the upper crust ([Stacey & Kramers 1975](#)). Emplacement of the igneous protoliths between 1850 and 1825 Ma provides an upper-age constraint for the timing of M1 granulite-facies metamorphism within the Read Group and subsequent emplacement of the syn- to post-tectonic granitoids at c. 1810–1770 Ma sets the lower age limit for this event. Discordant zircon rim and overgrowth analyses from Z.1448.2 are consistent with a metamorphic origin (low Th/U; 0.02–0.30) and define an upper-intercept age (1765 ± 169 Ma) which overlaps the interval of granitoid emplacement, and a lower-intercept age indicative of Pb-loss during the late Triassic (177 ± 110 Ma). P-T estimates on the relic granulite parageneses concur at around $830^\circ\text{C}/\sim 8$ kbar ([Schubert & Will 1994](#); [Schubert & Olesch 1995](#); [Talarico & Kroner 1999](#)). The relatively low-pressure could be explained by magmatic underplating or possibly a result of thermal resetting during the interval and transition from M1 to M2 as alluded to by [Schubert & Olesch \(1995\)](#) and supported by the inferred ages of zircon rims in Z.1448.2 (c. 1765 Ma). The lower-intercept ages for Z.1443.1 (c. 381 Ma) and Z.1448.2 (c. 211–177 Ma) are replicated in the syn- and post-tectonic granitoids and their significance is discussed in section 5.3.

5.2. c. 1810 Ma paragneisses

The primary detrital zircon $^{207}\text{Pb}/^{206}\text{Pb}$ age peak (c. 1827 Ma;

[Fig. 9c](#)) of the four paragneisses overlaps with the protolith emplacement age of the relic granulites (c. 1850–1825 Ma). Coupled with their similarly CL-dark, high U zircons (c. 400–4000 ppm), which would be unlikely to survive long distance transport, this suggests a proximal source component for the paragneiss precursors. Secondary peaks at c. 2435 and c. 2600 Ma show the contribution of older sources. Units of early Palaeoproterozoic age in the Shackleton Range are known only from the Mathys Gneiss (c. 2310–2470 Ma; [Pankhurst et al. 1983](#); [Brommer et al. 1999](#); [Riley et al. 2020](#)) in the La Grange Nunataks, but no *in situ* crust > 2500 Ma has yet been identified. This may indicate that crust of this age is more widespread within the Shackleton Range than currently known, allowing for a close association of the Read Group with the Fuchs Dome nappe during the mid to late Palaeoproterozoic as has been previously suggested (e.g., [Will et al. 2009, 2010](#)). The older components could also indicate that *in situ* > 2500 Ma crust exists south of the Read Mountains where the craton is extrapolated to continue into the PolarGAP province ([Fig. 2](#); [Jordan et al. 2022](#)) or may simply be derived from previously contiguous terranes (i.e., Mawson Continent, Laurentia).

Pinning a maximum deposition age for the sedimentary protoliths is complicated by the emplacement of syn- to post-tectonic granitoids during 1810–1770 Ma. Grains younger than 1810 Ma account for 21 % of the total population and must represent metamorphic or magmatic resetting given the clear intrusive relationships preserved between the granitoids and their host migmatite gneisses. Deposition of the paragneiss precursors must have concluded fairly rapidly as a result, lasting perhaps no longer than ~ 15 myr between the emplacement of the youngest relic granulite precursor (c. 1824 Ma) and emplacement of the first plutonic suites (c. 1807 Ma). The youngest concordant analysis across the paragneisses is c. 1733 Ma in age, though otherwise little evidence for late Palaeoproterozoic metamorphism is evident within the zircon populations. This is in stark contrast to two paragneisses from the Fuchs Dome nappe in the Eastern Basement and Pioneers Escarpment analysed by [Will et al. \(2009\)](#), where grains less than < 1810 Ma in age account for 62 % of the total population. The youngest zircon from these eastern samples has an age of 1620 Ma and they show a primary age peak of 1702 Ma. The authors attributed these younger ages to metamorphic growth and resetting of zircon during the high-T/low-P upper-amphibolite facies metamorphic episode which affected the entire Shackleton Range (M3 in the Read Group; [Table 1](#)). One of the samples showed a second primary peak at 1767 Ma, which was deemed meaningless by [Will et al. \(2009\)](#), however this closely matches the age of the post-tectonic granitoids within the Read Mountains. The age spectra older than 1810 Ma are generally similar to the Read Mountains paragneisses, suggesting they may have shared the same or similar detrital source.

5.3. 1810–1770 Ma syn- to post-tectonic granitoids and leucocratic dykes

The granitoids scattered across the Read Group show calc-alkaline and I-type geochemistry typical of volcanic island arc settings and can be grouped into a 1810–1790 Ma syn-tectonic suite and a 1780–1770 Ma post-tectonic suite. A metamorphic or thermal resetting event between or during granitoid emplacement is hinted at by a minority of 1780–1730 Ma metamorphic and magmatic zircons hosted within the paragneisses. Further, the low Th/U ratios (<0.15) and matching ages of both cores and rims from samples Z.1438.2 (1795 Ma, syn-tectonic; [Fig. 9f](#)) and Z.681.2 (1772 Ma, post-tectonic; [Fig. 9e](#)) differ greatly from the other granitoids which mostly range between values of 0.2–1.0. In particular, the fine-grained mylonitic matrix of Z.1438.2 ([Fig. 7b-c](#)) is evidence it was affected by intense shearing, most likely during or very rapidly after its emplacement. These successive magmatic pulses and their corresponding differences in age and emplacement style indicate an episode of north-directed ductile shearing and metamorphism affected the Read Mountains either during or just after the initial pulse (c. 1800 Ma?) but before the second pulse of undeformed granitoids at

1770 Ma. Significant reworking of the granulites post-emplacement is ruled out due to the absence of overprinting deformational fabrics or stable metamorphic mineral parageneses. This may be highlighted in the Pb isotope composition of K-feldspars, which show slightly enriched values along a geochron close to the age of their emplacement (1750–1650 Ma) and paradoxically older than those from the relic granulite from the Du Toit Nunataks which have been reworked during M2 (Z.1448.2, 1600–1500 Ma; Fig. 10). Zircon Lu-Hf compositions across both granulite suites yield $\epsilon_{\text{Hf}}(t)$ values between -6.6 and +5.7 and are consistent with their magma sources incorporating more crustal material than the juvenile relic granulites ($\epsilon_{\text{Hf}}(1845 \text{ Ma})$: +7.5 to +12.7) which preceded them (Fig. 10c).

As with the post-tectonic granulites, the leucocratic dyke (Z.1437.3) from The Ark shows enrichment of LREEs and fluid-mobile elements such as Ba and Sr and are completely undeformed internally. However, they exhibit a slightly more radiogenic Pb signature than the other granulites, instead overlapping somewhat with the relic granulite, despite the best estimate intrusion age for the dykes (Z.1445.3, 1773 Ma) matching emplacement of the post-tectonic granulites. The combined geochronological, geochemical and isotopic datasets indicate that dyke intrusion closely followed or overlapped emplacement of the post-tectonic granulites and was strongly fluid-assisted. Such fluids could have been derived and released from the second plutonic pulse and induced partial melting of the surrounding gneiss country rock, potentially explaining their similar Pb isotopic composition. The dominantly brittle environment of dyke emplacement (Fig. 6a) indicates that these waning stages of magmatism within the Read Group coincided with a rapid period of uplift toward the upper-crust and above the brittle-ductile transition zone.

Five of the granulites and leucocratic dykes, plus both of the relic granulites (Z.1443.1 & Z.1448.2), show lower-intercept ages of c. 390–380 Ma or c. 200 Ma (Fig. 9). It is unclear what, if any, geological significance is recorded in these lower-intercept ages. They may reflect one or a combination of: (1) slow cooling following Pan-African events (e.g., Möller et al. 2000) or (2) cooling events recorded by apatite and zircon fission tracks during the Devonian, Late Carboniferous and Late Triassic from the Weddell Sea (T. R. Riley, unpublished data), western Dronning Maud Land (Jacobs & Lisker 1999) and the Shackleton Range (Krohne et al. 2018), respectively.

5.4. 1720–1670 Ma upper-amphibolite facies metamorphism

Late Palaeoproterozoic metamorphism is well documented across most regions of the Shackleton Range. This event was first reported in the Read Group by Will et al. (2009) through electron microprobe Th-U-Pb monazite ages of c. 1670 Ma for two sillimanite-cordierite bearing paragneisses from the Du Toit Nunataks. A single grain U-Pb zircon concordia age from these lithologies predates the monazite ages at $1695 \pm 12 \text{ Ma}$, and together they were interpreted to represent the timing of a high-P/low-T upper-amphibolite facies ($690\text{--}730^\circ\text{C}/\sim 5\text{--}6 \text{ kbar}$; Schubert & Will 1994; Schubert & Olesch 1995) metamorphic episode which affected the entire Read Group and its crystalline basement counterparts in the northern Shackleton Range. However, we have demonstrated that the post-tectonic granulites and the several generations of leucocratic dykes of the Read Mountains intruded at c. 1770 Ma during a period of rapid uplift towards the upper crust. Their undeformed nature and crosscutting of earlier migmatitic and metamorphic fabrics prove that they did not undergo significant deformation post-emplacement. Further, resetting of the U-Pb zircon system at this interval was not found to be widespread in our dataset and, where present, appears restricted to the western Read Mountains and Du Toit Nunataks. The $1695 \pm 12 \text{ Ma}$ age determined for the paragneiss at Hatch Plain (Will et al. 2009) matches populations of stubby, metamorphic zircons and/or rims of 1730–1680 Ma age detected within the syn-tectonic granulite (Z.1444.1) and paragneiss (Z.1444.2) near Glen Glacier. Given the monazite ages from the Du Toit Nunataks and abundance of c. 1700 Ma

metamorphism recorded in the northern and eastern Shackleton Range, it is probable that metamorphic conditions during M3 were simply reduced in the central and eastern Read Mountains and predominantly insufficient to stimulate new zircon growth or reset its U-Pb system. We therefore envisage the Read Mountains to have occupied a higher crustal level than the Du Toit Nunataks during M3 metamorphism, perhaps since being downfaulted during Mesozoic rifting in the Weddell Sea. This would explain the apparently contradictory K-feldspar Pb data, where the c. 1845 Ma Du Toit Nunataks relic granulite appears younger than that of the 1810–1770 Ma granulites from the central Read Mountains, presumably because the Pb isotopic fingerprint of the former was re-equilibrated during M3 whilst the Read Mountains were unaffected.

6. Discussion

6.1. Tectono-metamorphic evolution of the Read Group

Our new extensive dataset allows us to critically assess previous interpretations of the evolution of the Read Group (e.g., Schubert & Will 1994; Schubert & Olesch 1995; Talarico & Kroner 1999; Will et al. 2009, 2010) and propose a new tectonic model (Fig. 11). We suggest that the igneous precursors of the relic granulites and their associated M1 mafic gneisses were emplaced within juvenile island arc and back-arc settings between 1850–1825 Ma (Fig. 11). Low- to intermediate-pressure granulite facies metamorphism and associated M1/D1 deformation of the juvenile melts is interpreted to have resulted from continued magma underplating of the island arc and subduction-related compression. Litharenitic and volcanoclastic sedimentary protoliths of the paragneiss sequences were deposited within a relatively narrow back-arc basin; the sediment accumulated 1850–1825 Ma juvenile magmatic material and minor components of 2700–1900 Ma age crust from a late-Archaeon to early Palaeoproterozoic craton. The older, >1900 Ma components may be preserved as relic slices within the Fuchs Dome nappe to the north and/or the unexposed PolarGAP province directly south.

Alternatively, the paragneiss precursors could have been deposited within the forearc, with sedimentation ceasing by c. 1810 Ma due to accretion of the forearc basin onto the matured island arc. M2/D2 metamorphism and migmatization of the metasediments was followed by emplacement of the syn- to post-tectonic granulites and associated M2 mafic gneisses, potentially driven by dehydration melting of the subducting slab (Fig. 11). The first pulse of syn-tectonic granulites between 1810–1790 Ma coincided with localized normal-sense ductile shearing associated with rapid uplift and isobaric cooling of the arc. Continued episodic magmatism manifested in the c. 1780–1770 Ma post-tectonic granulites, which fractionated towards the waning stages of their emplacement, generating the undeformed leucocratic dykes, the brittle emplacement of the latter documenting continued exhumation above the brittle-ductile transition zone. The cessation of magmatism after c. 1770 Ma is likely to be a result of subduction termination or potentially, slab break-off (Fig. 11). The identification of rare metamorphic zircons and monazite growth, or resetting, between 1720–1670 Ma in the metamorphic and plutonic suites at the Du Toit Nunataks testify to a secondary metamorphic event (M3) in this region, contemporaneous with widespread c. 1700–1650 Ma signatures detected across the Eastern Basement, Pioneers Escarpment and Haskard Highlands of the Fuchs Dome nappe. We interpret this metamorphic event to be related to arc-continent collision that led to crustal thickening and telescoping of the Read arc within c. 2500 Ma crust analogous to the Fuchs Dome nappe (e.g., PolarGAP province?), explaining the typical high-T/low-P amphibolite facies metamorphism recorded across the Shackleton Range at this time. The central and eastern Read Mountains may not have experienced this event as indicated by the unaffected 1770 Ma post-tectonic granulites and dykes, suggesting this region may have occupied a higher crustal level during M3 (Fig. 11) prior to being tectonically juxtaposed across an inferred detachment in

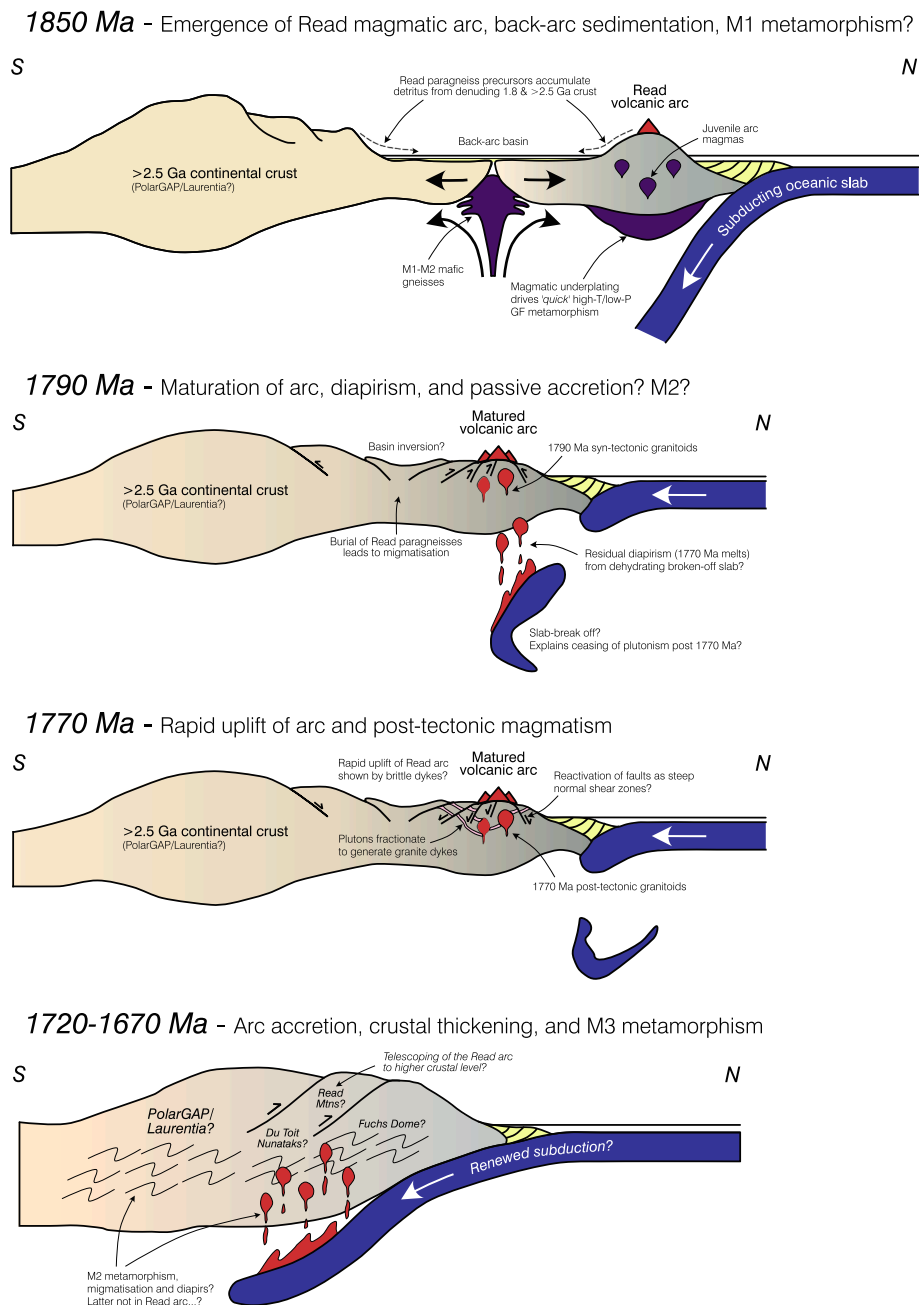


Fig. 11. Cartoon tectonic model proposed for the evolution of the Read Group. 1850 Ma) Juvenile arc magmas form from dehydration melting of a subducting oceanic slab. Successive magmatic underplating of the emerging Read volcanic arc drives high-T/low-P granulite facies metamorphism (M1) of the arc magmas to form the relic granulites. Subsequently, back-arc spreading develops the sequence of M1- and M2-related mafic gneisses and forms a sedimentary basin in which the Read paragneiss precursors are deposited, accumulating detritus from the Read volcanic arc (1850 Ma) and an older crustal block (>2500 Ma; PolarGAP province-Laurentian shield?). 1790 Ma) Maturation and accretion of the Read volcanic arc manifests in the emplacement of the syn-tectonic granitoids, basin inversion of the back-arc sedimentary basin and M2 metamorphism/migmatization of the relic granulites and paragneisses. 1770 Ma) Slab-break off ceases diapirism after emplacement of the 1770 Ma post-tectonic granitoids, which fractionate to generate the leucocratic dykes. This coincided with rapid uplift of the Read volcanic arc to the brittle-ductile transition zone. 1720–1670 Ma) Renewed subduction drives crustal thickening and associated M3 metamorphism across the Read Group and the entire Shackleton Range. Telescoping of the Read arc within the older PolarGAP-Laurentia(?) crustal block may explain the position of the c. 2500 Ma Fuchs Dome nappe to the north of the Read arc, whilst a lower crustal level of the Read Mountains relative to the Du Toit Nunataks may be reflected in the scarcity of M3 metamorphism in the former. 'N' and 'S' orientations refer to present-day orientation between the Read Mountains and the PolarGAP province.

the vicinity of Glen Glacier (Figs. 3 & 4). Although no field evidence can be cited for a NNW-SSE trending fault in this region, broadly parallel sub-vertical detachments have been mapped at Stephenson Bastion and Genghis Hills, which have offset Read Group exposures and the Mount Wegener and Otter Highlands thrusts to the north (Figs. 3 & 4; Clarkson et al. 1995).

6.2. Comparisons with Australo-Antarctic (Mawson) and Laurentian provinces

Whilst contesting models dispute their precise geographical fit (e.g., SWEAT, AUS-WUS, AUS-MEX; Moores 1991; Brookfield 1993; Karlstrom et al. 1999, 2001; Wingate et al. 2002), it is widely accepted that Australo-Antarctica (Mawson) and Laurentia became juxtaposed during

the assembly of Nuna (c. 1600 Ma), before rifting during break-up at c. 1300 Ma, and then re-amalgamating to form the northern part of Rodinia (c. 900 Ma) (Li et al. 2023; Cao et al. 2024). The Nuna-Rodinia supercontinent cycle oversaw a long-lived period of crustal growth (1800–1000 Ma) along an Andean-type active margin colloquially referred to as the Great Proterozoic Accretionary Orogen (GPAO; Condie 2013). The Mojave, Yavapai, and Mazatzal arcs (1850–1600 Ma) of southwestern United States (SWUS) represent one end of the GPAO, from where it is traced into the North American continental shield and southern Greenland before potentially extending as far as NW Scotland and Fennoscandia (Condie 2013). However, the SWUS provinces are truncated on their western flank by the Neoproterozoic rifted margin of the North American Cordillera and may have once continued into a region close to the Shackleton Range (Moore 1991; Dalziel 2024). Considering its possible ties as well with the Nimrod Complex and the Gawler-Adélie cratons, the Shackleton Range thus lies at a geological crossroads between the Mawson and SWUS provinces and may form a continuation of either or both terranes. Thus, with the caveat that the Read Group and the Nimrod Complex represent comparatively small windows located on the boundaries of their respective crustal domains, the comprehensive suite of ages and isotopic data we have presented from both regions nonetheless allow for the most robust test yet of their closest crustal affinities and are discussed in detail below.

6.2.1. Age correlations

Equally compelling age correlations to the Read Group can be made across the Mawson and SWUS provinces (Fig. 12). For example, the 1850–1825 Ma relic granulites match well with both the 1840 Ma Elves Chasm Gneiss (Babcock et al. 1979; Hawkins et al. 1996; Holland et al. 2015) in the Mojave Province and the 1860–1850 Ma Donington Suite granulitoids from the Gawler Craton (Mortimer et al. 1988; Schaefer 1998;

Reid et al. 2008; Belousova et al. 2009). Further, the Donington Suite granulitoids shows gneissic and migmatitic fabrics associated with granulite facies metamorphism (~750 °C/6 kbar; Reid et al. 2008). Although no granulite facies conditions have been reported from the Elves Chasm Gneiss, matrix monazites have been dated at 1674 Ma (Williams & Jercinovic 2002) and are a close match with monazite ages from the Du Toit Nunataks (1667–1672 Ma; Will et al. 2009). Younger calc-alkaline plutons of similar age to the Read Group granulitoids are also widely scattered across the Gawler and Mojave terranes. In Mojavia, they range in age from 1790–1690 Ma (Holland et al. 2015), becoming progressively younger and more juvenile towards the Mojave-Yavapai transition zone. The 1790–1770 Ma Tip Top and Wertigo granites (McAvaney et al. 2016; Fig. 12) in the Gawler Craton also show S-type and LREE-enriched geochemical patterns and are analogous to the post-tectonic granulitoids of the Read Group. Although *in-situ* crust of 1850–1770 Ma age is not known from the Nimrod Complex or the Adélie Craton of East Antarctica, glacial igneous clasts from the cTAMs dated between 1880–1790 Ma (Goodge et al. 2017) point to its presence beneath the East Antarctic Ice Sheet (i.e., Nimrod Igneous Province; Goodge & Finn 2010; Fig. 12) and are a potential link with the Read Group. Finally, 1690–1670 Ma M3 metamorphism reported from the Du Toit Nunataks and sparingly from the Read Mountains is contemporaneous with pervasive high-grade metamorphism, transpressional shear and the emplacement of syn- to post-tectonic magmatic suites within the Gawler Craton, collectively known as the *Kimban orogeny* (1730–1690 Ma; Daly et al. 1998; Vassallo & Wilson 2002; Dutch et al. 2010). Such an orogenic interval is also preserved within the Adélie Craton, the Nimrod Complex (*Nimrod orogeny*), and, crucially, the Mojave and Yavapai provinces (1720–1690 Ma *Yavapai orogeny*) (Fig. 12).

Most authors envisage the American Southwest to have developed through the stitching of multiple arc terranes to explain the Nd, Pb and

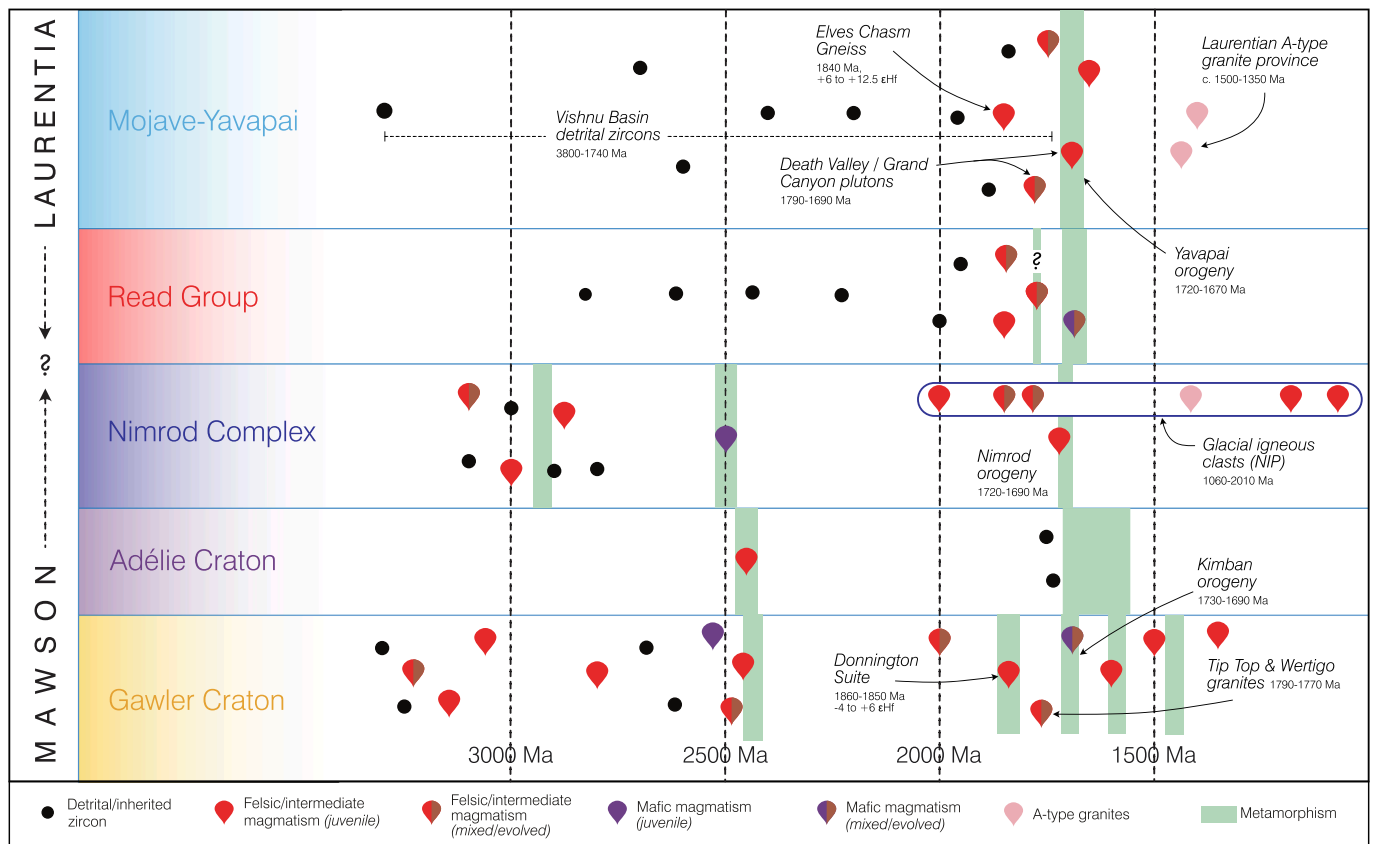


Fig. 12. Magmatic and metamorphic episodes of the Read Group in comparison with those documented from the Mojave-Yavapai provinces, the Nimrod Complex, the Adélie Craton, and the Gawler Craton.

Sr isotopic discontinuities between the Mojave, Yavapai and Mazatzal provinces (Zartman 1974; Bennett & DePaolo 1987; Wooden & Mueller 1988; Wooden & Miller 1990). Crust older than 1850 Ma is apparently absent from the Mojave Province, but its presence is implied from its isotopically evolved nature and 3800–1740 Ma age detritus accumulated within the Vishnu Basin metasediments of the Grand Canyon and Death Valley (Barth et al. 2000, 2009; Shufeldt et al. 2010; Strickland

et al. 2013; Holland et al. 2018; Hillenbrand et al. 2025; Figs. 12 & 13). Holland et al. (2018) suggested a Mawson Continent source for the Vishnu metasediments and thus a close link between Mojavia and East Antarctica early-on in the assembly of Nuna, whereas Hillenbrand et al. (2025) used multi-dimension scaling (MDS) analysis to suggest that the Trans-Hudson and Penokean orogens likely accounted for ~ 60–80 % of zircon detritus with Mojave Province metasediments, thus arguing for a

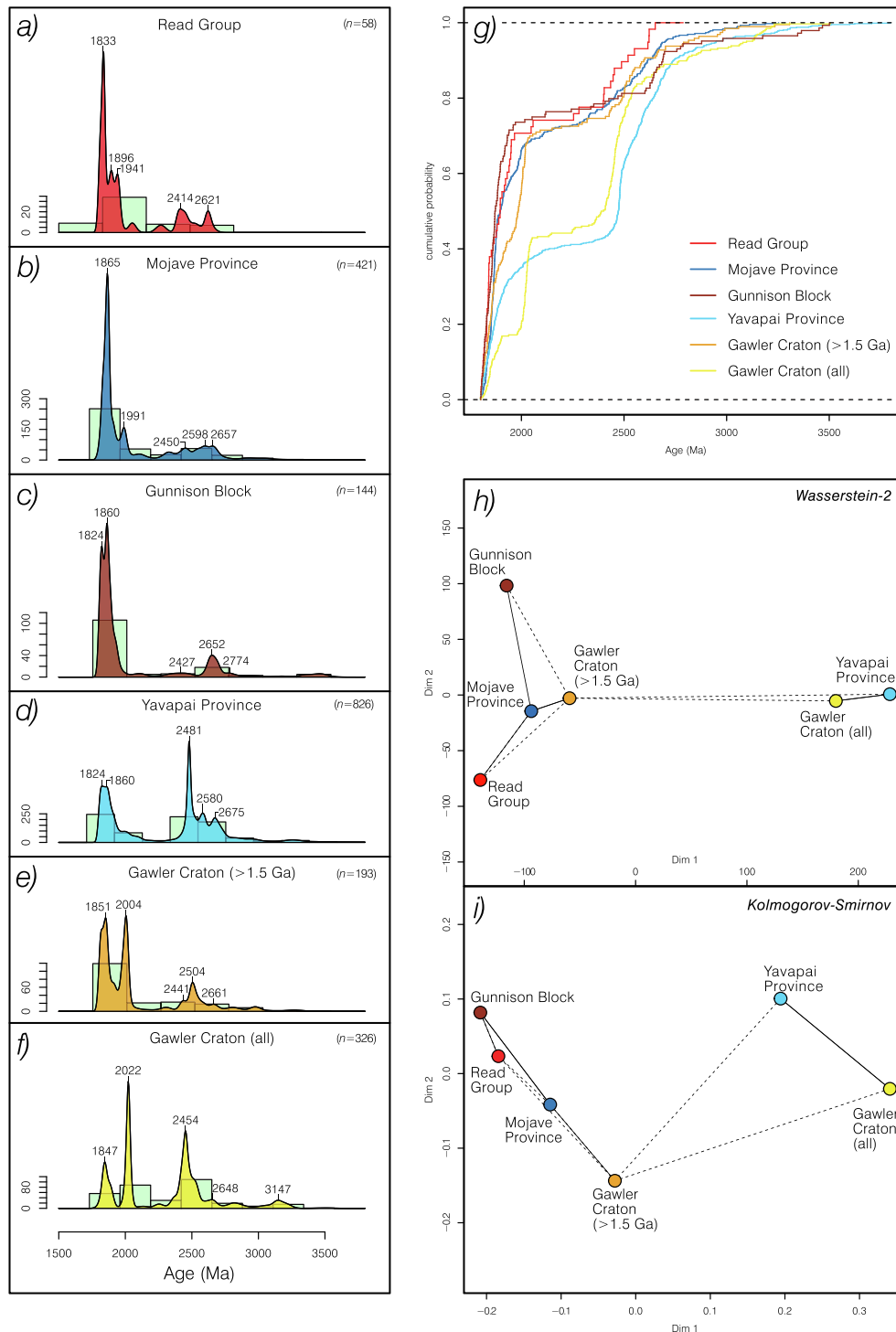


Fig. 13. a-f) Kernel density estimator (KDE) plots of U-Pb detrital zircon ages from metasedimentary rocks of the a) Read Group, b) Mojave Province (Death Valley), c) Gunnison Block, d) Yavapai Province (Grand Canyon “Vishnu Schists”), e) Gawler Craton (>1.5 Ga minimum deposition age), f) Gawler Craton (all). g) Cumulative age distribution (CAD) plot of all comparative terranes, h) Multi-scale dimension (MDS) plots of all comparative terranes. Solid lines indicate nearest neighbours and dashed lines indicate second nearest neighbours. Data from the Read Group is taken from this study. Data from all comparative provinces was sourced from Hillenbrand et al. (2024) and references therein. An adaptive kernel bandwidth was used for all groups.

dominantly Laurentian rather than East Antarctic source. Comparing our Read Group detrital zircon data against the same dataset shows it has greatest resemblance to the Death Valley metasediments of the Mojave Province and those from the Gunnison Block – an isotopically evolved crustal block within the Yavapai Province of northern Colorado (Hillenbrand et al. 2024). In kernel density estimator (KDE) plots, including only zircon $^{207}\text{Pb}/^{206}\text{Pb}$ ages > 1800 Ma, all three provinces produce primary age peaks at c. 1860–1830 Ma and less prominent secondary peaks at c. 2000–1900, c. 2450–2415, or c. 2650–2600 Ma (Fig. 13a-c). By contrast, the Vishnu Schists from Grand Canyon

(Yavapai Province) shift to a bi-modal age profile with primary and secondary peaks of c. 2480 and c. 1860–1825 Ma respectively (Fig. 13d). Palaeoproterozoic metasediments from the Gawler Craton yield equally prominent peaks at c. 1850 and c. 2000 Ma and a secondary peak at c. 2500 Ma (Fig. 13e-f). Thus, although these terranes all share similar detrital age peaks, their relative prominence is different and this is interrogated well in the cumulative age distribution (CAD) and MDS plots (Fig. 13h-g) where the Read, Gunnison, Mojave and Gawler (>1.5 Ga) terranes show close affinity whilst the Yavapai metasediments (Vishnu Schists) are markedly different and likely accumulated from a

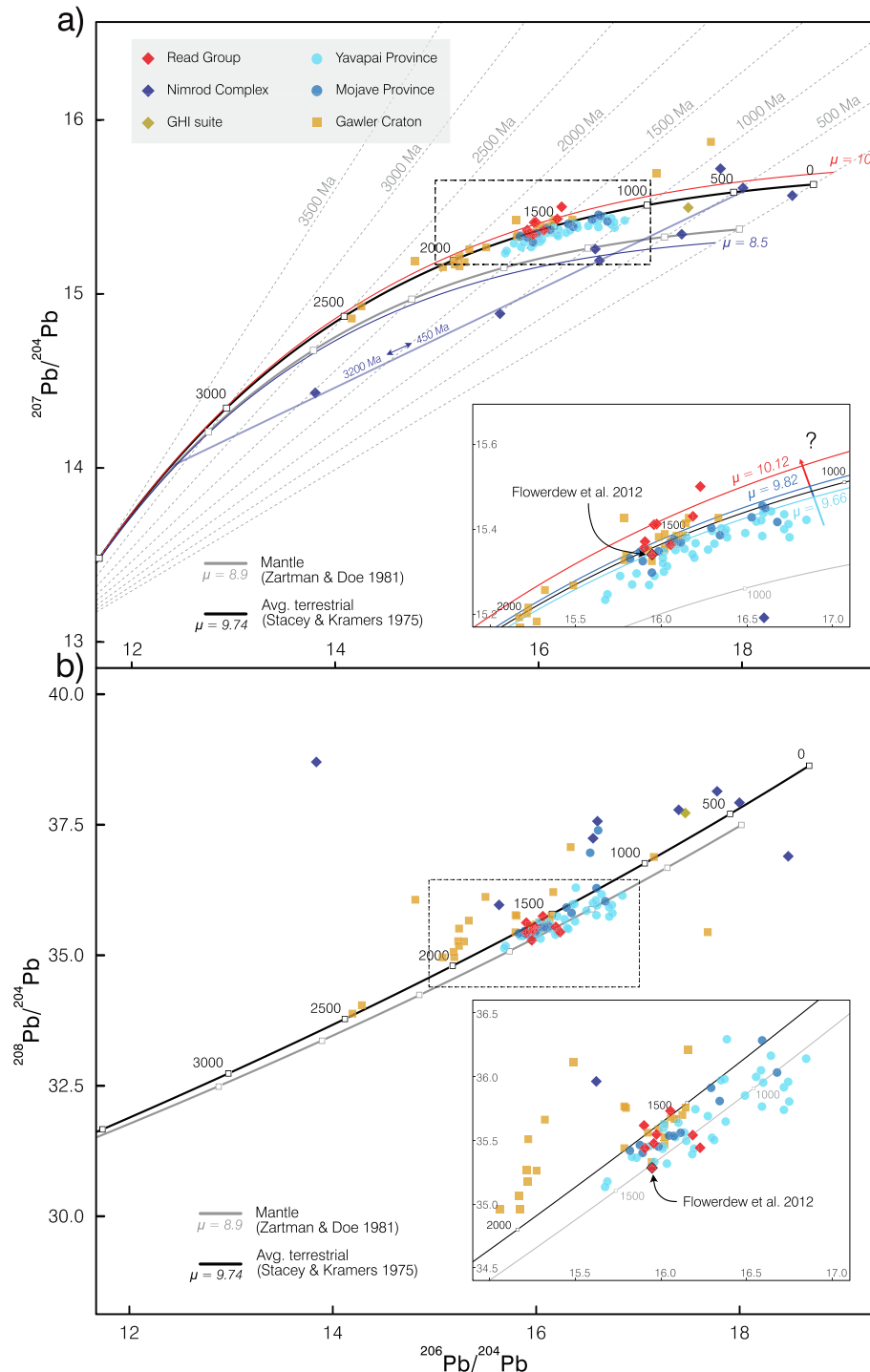


Fig. 14. a) Uranogenic, and b) thorogenic, Pb-Pb diagrams of Read Group, Nimrod Complex and GHI suite K-feldspar Pb analyses compared with K-feldspar Pb data from the Gawler Craton (Chapman et al. 2021; Liebmann et al. 2024) and the Mojave and Yavapai provinces (Bouchet et al. 2014; Hillenbrand 2025).

different or additional source (e.g., Wyoming Craton; Spencer et al. 2024). Nearest neighbour lines on the Wasserstein-2 and Kolmogorov-Smirnov MDS plots indicate that the Read Group matches best with the Mojave or Gunnison terranes (Fig. 13g).

6.2.2. Comparison of K-feldspar Pb and zircon Lu-Hf isotopic signatures

Correlations based on age data alone are limited in their scope, but the addition of K-feldspar Pb and zircon Lu-Hf isotopic data provides an additional parameter that allows more robust comparisons between potentially contiguous sectors. Comparable Pb and Hf isotope datasets are readily available from the Gawler Craton (Reid & Payne 2017; Chapman et al. 2021; Liebmann et al. 2024), and the American Southwest essentially played birthplace to isotopic mapping on the regional scale (e.g., Zartman 1974; Bennett & DePaolo 1987; Wooden & Mueller 1988; Wooden & Miller 1990; Wooden et al. 2013; Holland et al. 2015,

2018, 2020; Mako et al. 2015).

Figs. 14 and 15 compares our new data against these external K-feldspar Pb and zircon Lu-Hf isotope datasets, respectively. Despite a wider spread for *Kimban*-age and older (1.67–3.25 Ga) samples, several Gawler Craton Pb-Pb analyses overlap well with the Read Group in uranogenic and thorogenic space (Fig. 14a-b), yielding mostly similar $^{206}\text{Pb}/^{204}\text{Pb}$ (15.7–16.3), $^{207}\text{Pb}/^{204}\text{Pb}$ (15.3–15.5), and $^{208}\text{Pb}/^{204}\text{Pb}$ (34.8–36.8) as well as matching average model source μ and κ values of 10.0–10.1 and 3.7–4.0 respectively. The Mojave and Yavapai provinces overall demonstrate similar Pb-Pb compositions but with marginally higher $^{206}\text{Pb}/^{204}\text{Pb}$ (15.6–16.8) and lower $^{207}\text{Pb}/^{204}\text{Pb}$ (15.2–15.4) (Fig. 14), shifting their average model source μ and κ values to 9.7–9.8 and 3.6–3.8 respectively, straddling just below or along the Stacey & Kramers (1975) growth curve. Feldspars from the Nimrod Complex show significantly more variable $^{206}\text{Pb}/^{204}\text{Pb}$ (13.8–18.5), $^{207}\text{Pb}/^{204}\text{Pb}$

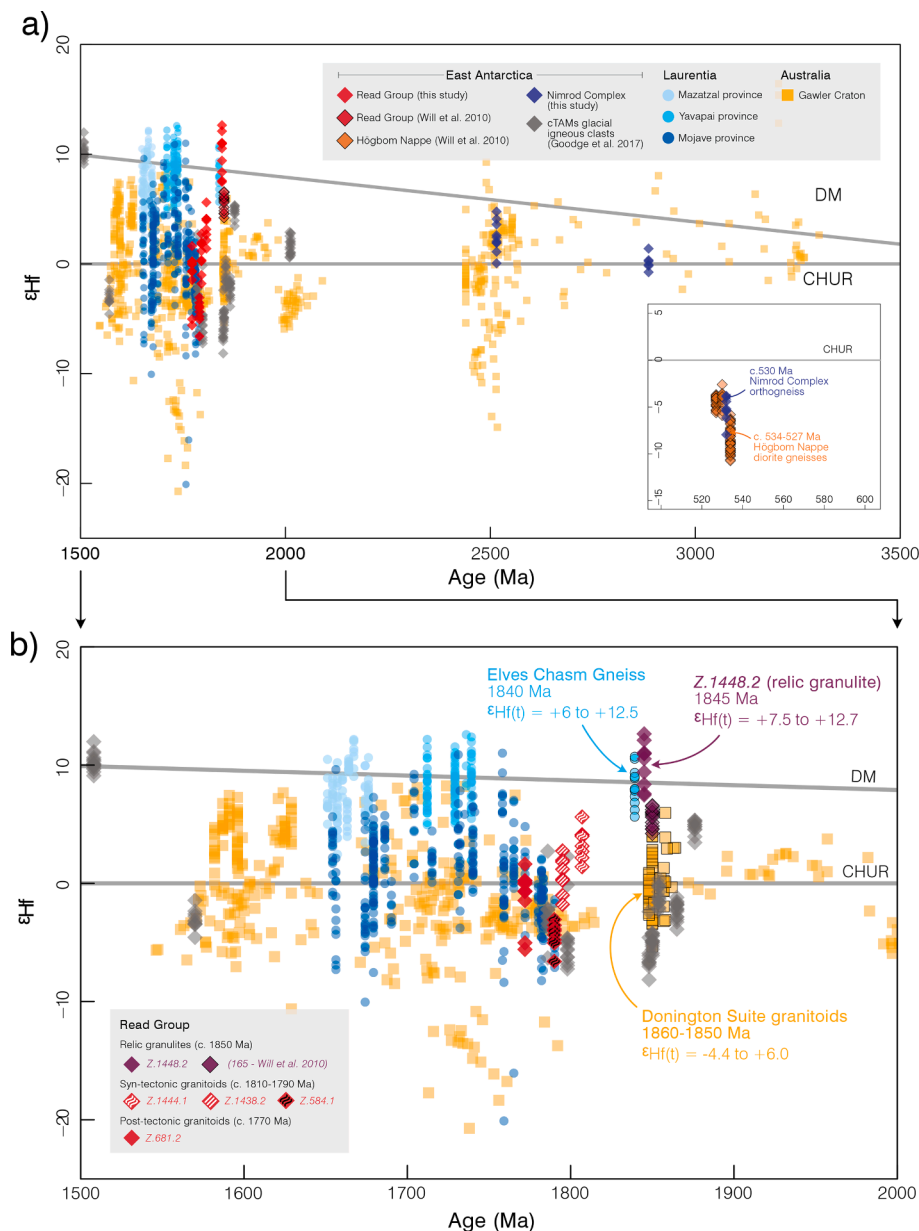


Fig. 15. a) U-Pb zircon crystallisation ages (3500–1500 Ma) vs. initial ϵHf values for zircon grains analysed from the Read Group and Nimrod Complex compared against published Hf isotope data from the Shackleton Range (Read Group, Högbom nappe; Will et al. 2010), cTAMs glacial igneous clasts (Goode et al. 2008; 2017), the Mojave, Yavapai and Mazatzal provinces (Wooden et al. 2013; Holland et al. 2015, 2018, 2020; Mako et al. 2015), and the Gawler Craton (Reid & Payne 2017 and references therein), (b) is a blown-up version of (a) with a constrained age range between 2000 and 1500 Ma, to demonstrate inter-sample variations of the Read Group and comparisons with the Elves Chasm Gneiss (blue circles with black outline) and the Donington Suite granitoids (orange squares with black outline).

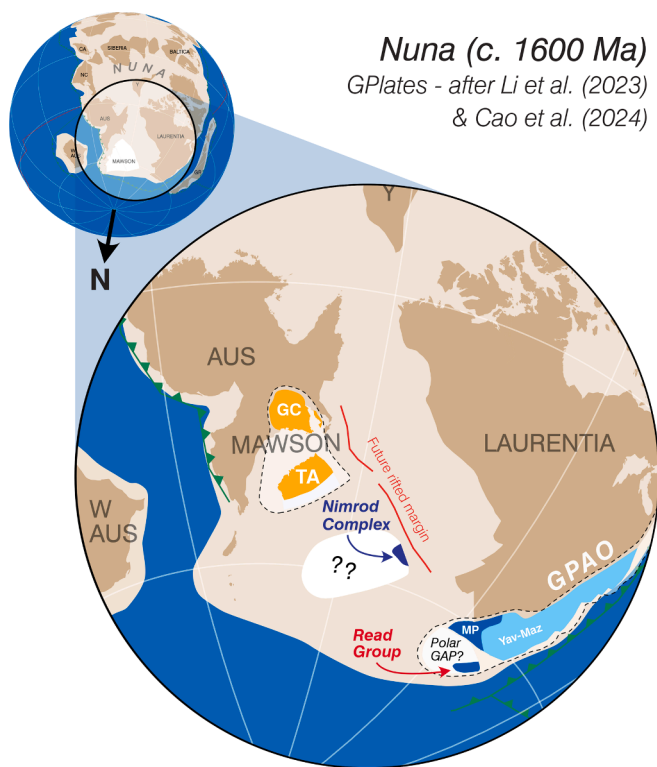


Fig. 16. Palaeogeographic reconstruction of the Nuna supercontinent (inset, top left) at c. 1600 Ma. The possible continuation of the Great Proterozoic Accretionary Orogen into the PolarGAP province of East Antarctica is shown, adjoining the Read Group with the Mojave Province. The Mawson Continent is shown to include only the Gawler and Adélie cratons, the Nimrod Complex (NC) may represent a separate crustal block with unknown spatial extent. Made and manually edited from GPlates software after the 1.8 Ga – present template from Li et al. (2023) and Cao et al. (2024). Abbreviations: AUS – Australia, CA – Cathaysia, GC – Gawler Craton, GR – Granite Rhyolite Province, MP – Mojave Province, NC – North China, TA – Terre Adélie Craton, W AUS – Western Australia, Y – Yangtze, Yav-Maz – Yavapai and Mazatzal provinces.

(14.4–15.7) and $^{208}\text{Pb}/^{204}\text{Pb}$ (35.9–38.7) than the Read and Gawler terranes (Fig. 14). Most samples plot well below the Stacey & Kramers (1975) growth curve in uranogenic space with mantle-like average model source μ values of 8.1–9.8 and form a linear array in uranogenic space which intersects various Pb growth curves at c. 3200 Ma and c. 650–440 Ma (Fig. 14a), consistent with Pb mobilisation during the Ross orogeny.

None of the Pb–Pb compositions from the Nimrod Complex overlap with any of the Read, Gawler or SWUS provinces. Contiguous cratonic terranes of shared crustal origin and evolution would generally be expected to share similar Pb isotopic fingerprints, however our new data do not support cratonic continuity between the Nimrod Complex and the Read Group or Gawler Craton. Therefore, a continuation of a coherent cratonic block beyond the Adélie coastline and into the interior of East Antarctica towards the Nimrod Complex, and beyond to the Shackleton Range, is not supported by the K-feldspar Pb data. Mostly positive zircon $\epsilon\text{Hf}(t)$ (–0.7 to +4.8), coupled with model ages which overlap their known crystallisation ages, support a juvenile, mantle-like source for the Archaean rocks of the Nimrod Complex (Fig. 10c). Although some of these initial Hf compositions overlap similar age rocks from the Gawler Craton (Reid & Payne 2017; Fig. 15), the uniquely juvenile Pb isotopic fingerprints suggest the Nimrod Complex may represent an isotopically distinct cratonic block of unknown extent beneath the ice.

An alternative tectonic configuration for the Read Group may be an affinity with the Mojave and Yavapai provinces. Despite their subtle difference in Pb isotopic character as outlined above, a more evolved

and radiogenic isotopic signature for the Read Group compared to the Mojave Province might be expected given the overall trend of increasingly evolved and older crust along a NW-transect across the SWUS accretionary orogens (Zartman 1974; Bennett & DePaolo 1987; Wooden & Mueller 1988; Wooden & Miller 1990). This fits with a model source μ increasing from 9.66 for the Yavapai Province, to 9.82 in the Mojave Province, and 10.12 for the Read Group (Fig. 14a), but also the wide range of $\epsilon\text{Hf}(t)$ values (–6.6 to +5.7) from the suite of syn- to late-tectonic granitoids of the Read Group which are most consistent with age-equivalent rocks from the Mojave province and, notably, several glacial igneous clasts from the cTAMs (Goodge et al. 2017) (Fig. 15). The strongly juvenile $\epsilon\text{Hf}(t)$ values (+7.5 to +12.7) from the 1845 Ma relic granulite (Z.1448.2) are perhaps most significant with respect to linking terranes. Will et al. (2010) reported lower $\epsilon\text{Hf}(t)$ values of +4.2 to +6.6 for an c. 1850 Ma orthogneiss from the Read Group, which they linked to the Donington Suite granitoids in the Gawler Craton ($\epsilon\text{Hf}(t) = -4$ to +6; Schaefer 1998; Reid et al. 2008; Belousova et al. 2009). However, such values also marginally overlap with those from the c. 1840 Ma Elves Chasm Gneiss (+6 to +12.5) in the Upper Granite Gorge, Grand Canyon (Hawkins et al. 1996; Holland et al. 2015; Fig. 15). The sample we have analysed as part of this study (Z.1448.2) yields indistinguishable values to those from obtained for the Elves Chasm Gneiss and do not overlap with the Donington Suite granitoids (Fig. 15).

The Mojave Province and its equivalents in the Yavapai Province (Gunnison Block; Hillenbrand et al. 2024) also provide perhaps the best match for U–Pb detrital zircon ages and show overlapping magmatic and metamorphic episodes with the Read Group (Figs. 12 & 13). Moreover, a Mojave rather than Mawson affinity satisfies the incoherent aeromagnetic anomalies of the PolarGAP province with that of the Southern Mawson Province and aligns with the newly discovered Pensacola Embayment along the Neoproterozoic rifted margin of the TAMs (Jordan et al. 2022) (Fig. 2), which Dalziel (2024) proposed may have once been occupied by the Mojave promontory. In this plate reconfiguration, the NE–SW trending SWUS accretionary orogens would continue into a region close to the Read Mountains and imply the continuation of the Great Proterozoic Accretionary Orogen into East Antarctica (Fig. 16). Links between the American Southwest and East Antarctica are not unfamiliar. Most recently, detrital U–Pb and Lu–Hf zircon data from the Watts Needle and Stephenson Bastion formations, which overlie the Read Group, indicate a link with the Mesoproterozoic A-type granite province of southwest Laurentia (Jacobs et al. 2026). Whilst *in-situ* analogues to the Laurentian A-type granites remain elusive in East Antarctica, they are predicted to continue somewhere beneath the ice by c. 1.47–1.41 Ga A-type rapakivi granite moraine clasts from the cTAMs (Goodge et al. 2008, 2017; Fig. 12). Furthermore, the enigmatic c. 1100 Ma Coats Land block north of the Shackleton Range shares similarities in lithology, age and Pb isotopic character with LIPs of the Laurentian Midcontinent rift inboard of the Grenville orogen (Loewy et al. 2011). This raises the possibility that both Coats Land and the Shackleton Range are remnants of Laurentia which remained attached throughout the Nuna–Rodinia supercontinent cycle until Rodinia break-up at c. 750 Ma (Li et al. 2023; Cao et al. 2024), with the Pensacola Embayment to the south potentially the preservation of this separation.

7. Conclusions

- Newly defined magmatic pulses within the Read Group at c. 1850–1825 Ma, 1810–1790 Ma and 1780–1770 Ma are linked to a juvenile arc terrane which developed on a continental margin setting. These pulses constrain the timing and duration of granulite (M1; 1825–1810 Ma) and amphibolite facies (M2; 1790–1770 Ma) metamorphism, which was likely driven by a combination of magmatic underplating of the Read arc and its subsequent accretion to an older crustal block (c. 2500 Ma – PolarGAP?).
- Extensive sequences of amphibolite-facies paragneisses within the Read Group were deposited by c. 1810 Ma, possibly within a back-

arc setting. They contain a prominent age signature which matches the M1 relic granulites (1850–1825 Ma) but also late-Neoproterozoic to early-Palaeoproterozoic (2700–1900 Ma) detritus which is absent from the Read Group. Detritus of these ages may be partly derived from c. 2500 Ma crust in the Fuchs Dome nappe which would imply that these terranes were already proximal prior to Pan-African reworking and potentially juxtaposed during renewed accretion and crustal thickening associated with the widespread M3 metamorphism (1720–1670 Ma) across the Shackleton Range.

- Although the age pattern and K-feldspar Pb isotope signature of the Read Group correlates well with both Australian (Gawler) and SWUS (Mojave-Yavapai) provinces, the uniquely juvenile Pb isotope signature of the Nimrod Complex ($\mu = 8.1\text{--}9.8$) strongly indicates that a coherent cratonic link between the Gawler-Adélie cratons and the Shackleton Range does not exist beneath the ice. Instead, the Read, Nimrod and Gawler-Adélie cratons likely represent isotopically distinct cratonic blocks which are potentially welded together by orogenic belts of unknown age.
- The Read Group U-Pb zircon age, zircon Lu-Hf and K-feldspar Pb isotope signature strongly overlaps that of metagneous suites from the Mojave and Yavapai provinces (e.g., Elves Chasm Gneiss) of southwestern United States. This similarity implies a shared crustal evolution for the two regions, and that they were once contiguous. This is further supported by the similarity of detrital zircon U-Pb age spectra obtained from the Read Group paragneisses with those from the Death Valley metasediments within the Mojave Province. The slightly more radiogenic Pb isotopic signature obtained from the Read Group ($\mu = 10.0\text{--}10.1$) compared with the Mojave Province ($\mu = 9.8$) may reflect a continuation into East Antarctica of the trend of the SWUS provinces transect of increasingly older and isotopically evolved crust.
- A Laurentian crustal affinity for the Shackleton Range aligns with recent detrital zircon Lu-Hf and U-Pb data from the sedimentary cover of the Read Group. It is also consistent with links between the Coats Land block with LIPs of the Laurentian Midcontinent Rift. Potentially, both Coats Land and the Shackleton Range are remnants of Laurentia which remained attached throughout the Nuna-Rodinia supercontinent cycle until Rodinia break-up at c. 750 Ma.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Acknowledgements

This study was partly funded by the IAPETUS2 Doctoral Training Partnership scheme through the Natural Environmental Research Council (grant NE/S007431/1). Materials prefixed with 'PRR-' are based on services provided by the Polar Rock Repository with support from the National Science Foundation, under Award OPP-2436582. Peter D. Clarkson collected two of the samples used in this study and is thanked for sharing his experiences in the field area. Ian Fitzsimons and Joachim Jacobs are thanked for their positive and constructive reviews which improved the manuscript considerably. The authors also thank Mark Evans at the British Antarctic Survey, Diana Sahy at the British Geological Survey, Stuart Nightingale at CASP, and Kerstin Lindén and Andreas Karlsson at the Swedish Museum of Natural History for sample preparation support. This is NordSIMS contribution number 828.

Appendix A. Supplementary data

Supplementary data to this article include full zircon U-Pb and Lu-Hf datasets and associated BSE/CL zircon images, full K-feldspar Pb-Pb datasets and associated PbIsO inputs and outputs, and full whole-rock

geochemistry datasets. The supplementary data can be found online at <https://doi.org/10.1016/j.precamres.2026.108235>.

References

- Armistead, S.E., Eglinton, B.M., Pehrsson, S.J., 2024. PbIso: an R package and web app for calculating and plotting Pb isotope data. *Can. J. Earth Sci.* 61 (1), 1–15. <https://doi.org/10.1139/cjes-2023-0029>.
- Babcock, R.S., Brown, E.H., Clark, M.D., Livingston, D.E., 1979. Geology of the older Precambrian rocks of the Grand Canyon Part II. the Zoroaster Plutonic complex and related rocks. *Precamb. Res.* 8 (3–4), 243–275. [https://doi.org/10.1016/0301-9268\(79\)90031-7](https://doi.org/10.1016/0301-9268(79)90031-7).
- Barth, A.P., Wooden, J.L., Coleman, D.S., Fanning, C.M., 2000. Geochronology of the Proterozoic basement of southwesternmost North America, and the origin and evolution of the Mojave crustal province. *Tectonics* 19 (4), 616–629. <https://doi.org/10.1029/1999TC001145>.
- Barth, A.P., Wooden, J.L., Coleman, D.S., Vogel, M.B., 2009. Assembling and disassembling California: a zircon and monazite geochronologic framework for Proterozoic crustal evolution in southern California. *J. Geol.* 117 (3), 221–239. <https://doi.org/10.1086/597515>.
- Barton, J.M., Klemd, R., Allsopp, H.L., Auret, S.H., Copperthwaite, Y.E., 1987. The geology and geochronology of the Annandagstoppane granite, Western Dronning Maud Land, Antarctica. *Contrib. Miner. Petrol.* 97, 488–496. <https://doi.org/10.1007/BF00375326>.
- Belousova, E.A., Reid, A.J., Griffin, W.L., O'Reilly, S.Y., 2009. Rejuvenation vs. recycling of Archean crust in the Gawler Craton, South Australia: evidence from U–Pb and Hf isotopes in detrital zircon. *Lithos* 113 (3–4), 570–582. <https://doi.org/10.1016/j.lithos.2009.06.028>.
- Bennett, V.C., DePaolo, D.J., 1987. Proterozoic crustal history of the western United States as determined by neodymium isotopic mapping. *GSA Bull.* 99 (5), 674–685. [https://doi.org/10.1130/0016-7606\(1987\)99<674:PCHOTW>2.0.CO;2](https://doi.org/10.1130/0016-7606(1987)99<674:PCHOTW>2.0.CO;2).
- Betts, P.G., Giles, D., Aitken, A., 2011. Palaeoproterozoic accretion processes of Australia and comparisons with Laurentia. *Int. Geol. Rev.* 53 (11–12), 1357–1376. <https://doi.org/10.1080/00206814.2010.527646>.
- Betts, P.G., Armit, R.J., Stewart, J., Aitken, A.R.A., Ailleres, L., Donchak, P., Hutton, L., Withnall, I., Giles, D., 2016. Australia and Nuna. *Geol. Soc. Lond. Spec. Publ.* 424 (1), 47–81. <https://doi.org/10.1144/SP424>.
- Betts, P.G., Giles, D., Schaefer, B.F., 2008. Comparing 1800–1600 Ma accretionary and basin processes in Australia and Laurentia: possible geographic connections in Columbia. *Precamb. Res.* 166 (1–4), 81–92. <https://doi.org/10.1016/j.precamres.2007.03.007>.
- Blichert-Toft, J., 2008. The Hf isotopic composition of zircon reference material 91500. *Chem. Geol.* 253, 252–257. <https://doi.org/10.1016/j.chemgeo.2008.05.014>.
- Borg, S.G., Depaolo, D.J., Smith, B.M., 1990. Isotopic structure and tectonics of the central Transantarctic Mountains. *J. Geophys. Res. Solid Earth* 95 (B5), 6647–6667. <https://doi.org/10.1029/JB095iB05p06647>.
- Bouchet, R.A., Blichert-Toft, J., Reid, M.R., Levander, A., Albarède, F., 2014. Similarities between the Th/U map of the western US crystalline basement and the seismic properties of the underlying lithosphere. *Earth Planet. Sci. Lett.* 391, 243–254. <https://doi.org/10.1016/j.epsl.2014.02.004>.
- Bouvier, A., Vervoort, J.D., Patchett, P.J., 2008. The Lu–Hf and Sm–Nd isotopic composition of CHUR: Constraints from unequilibrated chondrites and implications for the bulk composition of terrestrial planets. *Earth Planet. Sci. Lett.* 273, 48–57.
- Brookfield, M.E., 1993. Neoproterozoic Laurentia–Australia fit. *Geology* 21, 683–686. [https://doi.org/10.1130/0091-7613\(1993\)021<0683:NLAF>2.3.CO;2](https://doi.org/10.1130/0091-7613(1993)021<0683:NLAF>2.3.CO;2).
- Brommer, A., Millar, I.L., Zeh, A., 1999. Geochronology, structural geology and petrology of the northwestern Lagrange Nunataks, Shackleton Range. *Antarctica. Terra Antarctica* 6 (3), 269–278.
- Buggisch, W., Bachtadse, V., Henjes-Kunst, F., 1999. Lithostratigraphy, facies, geochronology and palaeomagnetic data from the Blaiklock Glacier Group, Shackleton Range. *Antarctica. Terra Antarctica* 6 (3), 229–239.
- Buggisch, W., Henjes-Kunst, F., 1999. Stratigraphy, facies, and provenance analyses of the lower Cambrian Mount Wegener Formation of the Shackleton Range. *Antarctica. Terra Antarctica* 6 (3), 211–228.
- Buggisch, W., Kleinschmidt, G., 1999. New evidence for nappe tectonics in the southern Shackleton Range. *Antarctica. Terra Antarctica* 6 (3), 203–210.
- Buggisch, W. & Kleinschmidt, G. (2007). The Pan-African nappe tectonics in the Shackleton Range. In: Cooper, A. K., Barrett, P. J., Storey, B., Stump, E. & Wise, W. and the 10th ISAES editorial team (eds.) Antarctica: a keystone in a changing world. Proceedings of the 10th ISAES, USGS Open-File Report, 104. 10.3133/ofr20071047SRP58.
- Buggisch, W., Kleinschmidt, G., Höhndorf, A., Pohl, J., 1994. Stratigraphy and facies of sediments and low-grade metasediments in the Shackleton Range. *Antarctica. Polarforschung* 63 (1), 9–32.
- Cao, X., Collins, A.S., Pisarevsky, S., Flament, N., Li, S., Hasterok, D., Müller, R.D., 2024. Earth's tectonic and plate boundary evolution over 1.8 billion years. *Geosci. Front.* 15 (6), 101922. <https://doi.org/10.1016/j.gsf.2024.101922>.
- Chapman, N.D., Mefre, S.J., Maas, R., Ehrig, K., 2021. Pb-isotope ratios and the petrogenesis of the Tunkillia Suite, Gawler Craton. *Aust. J. Earth Sci.* 68 (8), 1071–1091. <https://doi.org/10.1080/08120099.2021.1915868>.
- Clarkson, P.D., 1972. Geology of the Shackleton Range: a preliminary report. *British Antarctic Survey Bulletin* 31, 1–15.
- Clarkson, P.D., 1981. Geology of the Shackleton Range. IV, the dolerite dykes. *British Antarctic Survey Bulletin* 53, 201–212.

- Clarkson, P. D., Tessensohn, F., Thomson, M. R. A., Belyatsky, B., Braun, H.-M., Buggisch, W., Grikurov, G. E., Hofmann, J., Hohndorf, A., Kamenev, E. N., Kameneva, G. J., Kleinschmidt, G., Kreuzer, H., Leat, P. T., Olesch, M., Paech, H.-J., Pankhurst, R. J., Peters, M., Roland, N. W., Schubert, W., Soloviev, I. A., Spaeth, G., Techmer, K. S., Thomson, J. W., Weber, B., Weber, K. & King, S. (1995). Geological map of the Shackleton Range, Antarctica. BAS GEOMAP Series, Sheet 4, 1:250 000, with supplementary text. Cambridge, British Antarctic Survey.
- Clarkson, P.D., Wyeth, R.B., 1983. Geology of the Shackleton Range. III: the Blaiklock Glacier Group. *Bulletin-British Antarctic Survey* 52, 233–244.
- Condie, K.C., 2013. Preservation and recycling of crust during accretionary and collisional phases of Proterozoic orogens: a bumpy road from Nuna to Rodinia. *Geosciences* 3 (2), 240–261. <https://doi.org/10.3390/geosciences3020240>.
- Craddock, C. (1972). Antarctic Tectonics. In: Adie, R. J. (ed.). Antarctic geology and geophysics. Symposium on Antarctic Geology and Solid Earth Geophysics, Oslo, 6–15 August 1970 (Vol. 1). Universitetsforlaget, Oslo, 449–445.
- Daly, S.J., Fanning, C.M., Fairclough, M.C., 1998. Tectonic evolution and exploration potential of the Gawler Craton, South Australia. *AGSO Journal of Australian Geology & Geophysics* 17, 145–168.
- Dalziel, I.W., 2024. Rodinia palaeogeography: Laurentia as the geological 'Key'. *Geol. Soc. Lond. Spec. Publ.* 542 (1), 51–85. <https://doi.org/10.1144/SP542-2022-343>.
- Dutch, R.A., Hand, M., Kelsey, D.E., 2010. Unravelling the tectonothermal evolution of reworked Archean granulite facies metapelites using in situ geochronology: an example from the Gawler Craton, Australia. *Journal of Metamorphic Geology* 28 (3), 293–316. <https://doi.org/10.1080/08120090802266550>.
- Elliott, D.H., 1975. Tectonics of Antarctica: a review. *Am. J. Sci.* 275, 45–106.
- Encarnación, J., Grunow, A., 1996. Changing magmatic and tectonic styles along the paleo-Pacific margin of Gondwana and the onset of early Paleozoic magmatism in Antarctica. *Tectonics* 15 (6), 1325–1341. <https://doi.org/10.1029/96TC01484>.
- Fanning, C.M., Daly, S.J., Bennett, V.C., Menot, R.P., Peucat, J.J., Oliver, R.L., Monnier, O., 1995. The 'Mawson Block': once contiguous Archean to Proterozoic crust in the East Antarctic Shield and the Gawler Craton, Australia. In: Ricci, A.C. (Ed.), 7th International Symposium on Antarctic Earth Sciences, 124. University of Siena, Siena.
- Fanning, C.M., Moore, D.H., Bennett, V.C., Daly, S.J., 1996. The 'Mawson Continent': Archean to Proterozoic crust in the East Antarctica Shield and Gawler Craton, Australia. a cornerstone in Rodinia and Gondwanaland. *Geological Society of Australia Abstracts* 41, 135.
- Fitton, J.G., Saunders, A.D., Norry, M.J., Hardarson, B.S., Taylor, R.N., 1997. Thermal and chemical structure of the Iceland plume. *Earth Planet. Sci. Lett.* 153 (3–4), 197–208. [https://doi.org/10.1016/S0012-821X\(97\)00170-2](https://doi.org/10.1016/S0012-821X(97)00170-2).
- Fitzsimons, I.C., 2000a. A review of tectonic events in the East Antarctic Shield and their implications for Gondwana and earlier supercontinents. *J. Afr. Earth Sc.* 31 (1), 3–23. [https://doi.org/10.1016/S0899-5362\(00\)00069-5](https://doi.org/10.1016/S0899-5362(00)00069-5).
- Fitzsimons, I.C., 2000b. Grenville-age basement provinces in East Antarctica: evidence for three separate collisional orogens. *Geology* 28 (10), 879–882. [https://doi.org/10.1130/0091-7613\(2000\)28<879:GBPIEA>2.0.CO;2](https://doi.org/10.1130/0091-7613(2000)28<879:GBPIEA>2.0.CO;2).
- Fitzsimons, I.C., 2003. Proterozoic basement provinces of southern and southwestern Australia, and their correlation with Antarctica. *Geol. Soc. Lond. Spec. Publ.* 206, 93–130. <https://doi.org/10.1144/GSL.SP.2003.206.01.07>.
- Flowerdew, M.J., Tyrrell, S., Riley, T.R., Whitehouse, M.J., Mulvaney, R., Leat, P.T., Marshall, H.R., 2012. Distinguishing East and West Antarctic sediment sources using the Pb isotope composition of detrital K-feldspar. *Chem. Geol.* 292, 88–102. <https://doi.org/10.1016/j.chemgeo.2011.11.006>.
- Forsberg, R., Olesen, A.V., Ferraccioli, F., Jordan, T.A., Matsuoka, K., Zakrajsek, A., Ghidella, M., Greenbaum, J.S., 2018. Exploring the Recovery Lakes region and interior Dronning Maud Land, East Antarctica, with airborne gravity, magnetic and radar measurements. *Geol. Soc. Lond. Spec. Publ.* 461 (1), 23–34. <https://doi.org/10.1144/SP461.17>.
- Goodge, J.W., 2020. Geological and tectonic evolution of the Transantarctic Mountains, from ancient craton to recent enigma. *Gondw. Res.* 80, 50–122. <https://doi.org/10.1016/j.gr.2019.11.001>.
- Goodge, J.W., Fanning, C.M., Bennett, V.C., 2001. U–Pb evidence of ~ 1.7 Ga crustal tectonism during the Nimrod Orogeny in the Transantarctic Mountains, Antarctica: implications for Proterozoic plate reconstructions. *Precamb. Res.* 112 (3–4), 261–288. [https://doi.org/10.1016/S0301-9268\(01\)00193-0](https://doi.org/10.1016/S0301-9268(01)00193-0).
- Goodge, J. W. & Fanning, C. M. (1999). 2.5 by of punctuated Earth history as recorded in a single rock. *Geology*, 27(11), 1007–1010. 10.1130/0091-7613(1999)027<1007:BYOPEH>2.3.CO;2.
- Goodge, J.W., Fanning, C.M., 2016. Mesoarchean and Paleoproterozoic history of the Nimrod complex, central Transantarctic Mountains, Antarctica: Stratigraphic revisions and relation to the Mawson continent in East Gondwana. *Precamb. Res.* 285, 242–271. <https://doi.org/10.1016/j.precambres.2016.09.001>.
- Goodge, J.W., Fanning, C.M., Brecke, D.M., Licht, K.J., Palmer, E.F., 2010. Continuation of the Laurentian Grenville province across the Ross Sea margin of East Antarctica. *J. Geol.* 118 (6), 601–619. <https://doi.org/10.1086/656385>.
- Goodge, J.W., Fanning, C.M., Fisher, C.M., Vervoort, J.D., 2017. Proterozoic crustal evolution of central East Antarctica: Age and isotopic evidence from glacial igneous clasts, and links with Australia and Laurentia. *Precamb. Res.* 299, 151–176. <https://doi.org/10.1016/j.precambres.2017.07.026>.
- Goodge, J.W., Fanning, C.M., Fisher, C.M., Vervoort, J.D., 2024. Crustal architecture of the paleo-Pacific rift margin of East Antarctica: evidence from U–Pb ages and O–Hf isotope compositions of Ross Orogen granitoids. *Geochem. Geophys. Geosyst.* 25 (4), e2024GC011435. <https://doi.org/10.1029/2024GC011435>.
- Goodge, J.W., Finn, C.A., 2010. Glimpses of East Antarctica: Aeromagnetic and satellite magnetic view from the central Transantarctic Mountains of East Antarctica. *J. Geophys. Res. Solid Earth* 115 (B9). <https://doi.org/10.1029/2009JB006890>.
- Goodge, J.W., Vervoort, J.D., Fanning, C.M., Brecke, D.M., Farmer, G.L., Williams, I.S., Myrow, P.M., DePaolo, D.J., 2008. A positive test of East Antarctica–Laurentia juxtaposition within the Rodinia supercontinent. *Science* 321 (5886), 235–240. <https://doi.org/10.1126/science.1159189>.
- Gose, W.A., Helper, M.A., Connelly, J.N., Hutson, F.E., Dalziel, I.W., 1997. Paleomagnetic data and U–Pb isotopic age determinations from Coats Land, Antarctica: Implications for late Proterozoic plate reconstructions. *J. Geophys. Res. Solid Earth* 102 (B4), 7887–7902. <https://doi.org/10.1029/96JB03595>.
- Granseth, A., Slagstad, T., Roberts, N.M.W., Hagen-Peter, G., Kirkland, C.L., Møkkelgjerd, S.H.H., Röhr, T.S., Coint, N., Sørensen, B.E., 2021. Multi-isotope tracing of the 1.3–0.9 Ga evolution of Fennoscandia; crustal growth during the Sveconorwegian orogeny. *Gondw. Res.* 91, 31–39. <https://doi.org/10.1016/j.gr.2020.10.019>.
- Grew, E. S. (1982). The Antarctic margin. *The Ocean Basins and Margins: The Indian Ocean*, 697–755.
- Gunn, B.M., Warren, G., 1962. Geology of Victoria Land between the Mawson and Mulock Glaciers, Antarctica. *New Zealand Geological Survey Bulletin* 71.
- Halpern, M., 1970. Rubidium-strontium date of possibly 3 billion years for a granitic rock from Antarctica. *Science* 169 (3949), 977–978. <https://doi.org/10.1126/science.169.3949.977>.
- Harley, S., 2003. Archean–Cambrian crustal development of East Antarctica: Metamorphic characteristics and tectonic implications. *Geol. Soc. Lond. Spec. Publ.* 206 (1), 203–230. <https://doi.org/10.1144/gsl.sp.2003.206.01.11>.
- Hawkins, D.P., Bowring, S.A., Ilg, B.R., Karlstrom, K.E., Williams, M.L., 1996. U–Pb geochronologic constraints on the Paleoproterozoic crustal evolution of the Upper Granite Gorge, Grand Canyon, Arizona. *GSA Bulletin* 108 (9), 1167–1181. [https://doi.org/10.1130/0016-7606\(1996\)108<1167:UPGCOT>2.3.CO;2](https://doi.org/10.1130/0016-7606(1996)108<1167:UPGCOT>2.3.CO;2).
- Hillenbrand, I.W., 2025. Lead Isotopes Constrain Precambrian Crustal Architecture, thermal history and Lithospheric Foundering in Laurentia. *Terra Nova* 37 (2), 65–76. <https://doi.org/10.1111/ter.12748>.
- Hillenbrand, I.W., Gilmer, A.K., Premo, W.R., Williams, M.L., Jercinovic, M.J., 2025. Re-evaluating the tectonic affinity of Proterozoic crustal provinces in the Southwest USA: Detrital zircon evidence for a Laurentian source for the Yavapai and Mojave Provinces. *GSA Bull.* 137 (7–8), 2965–2981. <https://doi.org/10.1130/B37882.1>.
- Hillenbrand, I.W., Gilmer, A.K., Williams, M.L., Karlstrom, K.E., Souders, A.K., Jercinovic, M.J., Vazquez, J., Premo, W.R., 2024. Heterogeneous multi-stage accretionary orogenesis—Evidence from the Gunnison block in the Yavapai Province, southwest USA. *Precamb. Res.* 401, 107256. <https://doi.org/10.1016/j.precambres.2023.107256>.
- Holland, M.E., Grambling, T.A., Karlstrom, K.E., Jones III, J.V., Nagotko, K.N., Daniel, C. G., 2020. Geochronologic and Hf-isotope framework of Proterozoic rocks from central New Mexico, USA: Formation of the Mazatzal crustal province in an extended continental margin arc. *Precamb. Res.* 347, 105820. <https://doi.org/10.1016/j.precambres.2020.105820>.
- Holland, M.E., Karlstrom, K.E., Doe, M.F., Gehrels, G.E., Pecha, M., Shufeldt, O.P., Begg, G., Griffin, W.L., Belousova, E., 2015. An imbricate midcrustal suture zone: the Mojave–Yavapai province boundary in Grand Canyon, Arizona. *GSA Bulletin* 127 (9–10), 1391–1410. <https://doi.org/10.1016/j.B31232.1>.
- Holland, M.E., Karlstrom, K.E., Gehrels, G., Shufeldt, O.P., Begg, G., Griffin, W., Belousova, E., 2018. The Paleoproterozoic Vishnu basin in southwestern Laurentia: Implications for supercontinent reconstructions, crustal growth, and the origin of the Mojave crustal province. *Precamb. Res.* 308, 1–17. <https://doi.org/10.1016/j.precambres.2018.02.001>.
- Irvine, T.N., Baragar, W.R.A.F., 1971. A guide to the chemical classification of the common volcanic rocks. *Can. J. Earth Sci.* 8 (5), 523–548. <https://doi.org/10.1139/e71-055>.
- Ishizuka, H., 2008. Protoliths of the Napier complex in Enderby Land, East Antarctica; an overview and implication for crustal formation of Archean continents. *J. Mineral. Petrol. Sci.* 103 (4), 218–225. <https://doi.org/10.2465/jmps.080328>.
- Jeon, H., Whitehouse, M.J., 2015. A critical evaluation of U–Pb calibration schemes used in SIMS zircon geochronology. *Geostand. Geoanal. Res.* 39 (4), 443–452. <https://doi.org/10.1111/j.1751-908X.2014.00325.x>.
- Jacobs, J., Läufer, A., Wang, C.C., Fossheim, Y., Jacques, G., Pedersen, L.E.R., 2026. Detrital U–Pb and Lu–Hf zircon signatures from the Shackleton Range, East Antarctica: Implications for supercontinent assembly in Nuna, Rodinia, and Gondwana. *Precamb. Res.* 444, 108189. <https://doi.org/10.1016/j.precambres.2026.108189>.
- Jacobs, J., Lisker, F., 1999. Post Permian tectono-thermal evolution of western Dronning Maud Land, East Antarctica: an apatite fission-track approach. *Antarct. Sci.* 11 (4), 451–460. <https://doi.org/10.1017/S0954102099000589>.
- Jacobs, J., Thomas, R.J., 2004. Himalayan-type indenter-escape tectonics model for the southern part of the late Neoproterozoic–early Paleozoic East African–Antarctic orogen. *Geology* 32 (8), 721–724. <https://doi.org/10.1130/G20516.1>.
- Janoušek, V., 2024. HafAn: a R-language script aiding interpretation of the Hf isotopic data. *J. Geosci.* 69 (3), 151–160. <https://doi.org/10.3190/jgeosci.391>.
- Janoušek, V., Bowes, D.R., Rogers, G., Farrow, C.M., Jahn, B., 2006. Interpretation of whole-rock geochemical data in igneous geochemistry: introducing Geochemical Data Toolkit (GCDkit). *J. Petrol.* 47 (6), 1255–1259. <https://doi.org/10.1093/ptrology/egl009>.
- Jordan, T.A., Ferraccioli, F., Forsberg, R., 2022. An embayment in the East Antarctic basement constrains the shape of the Rodinian continental margin. *Commun. Earth Environ.* 3 (1), 52. <https://doi.org/10.1038/s43247-022-00375-z>.
- Karlstrom, K.E., Åhäll, K.I., Harlan, S.S., Williams, M.L., McLelland, J., Geissman, J.W., 2001. Long-lived (1.8–1.0 Ga) convergent orogen in southern Laurentia, its extensions to Australia and Baltica, and implications for refining Rodinia. *Precamb. Res.* 111 (1–4), 5–30. [https://doi.org/10.1016/S0301-9268\(01\)00154-1](https://doi.org/10.1016/S0301-9268(01)00154-1).

- Karlstrom, K.E., Harlan, S.S., Williams, M.L., McLelland, J., Geissman, J.W., Åhäll, K.I., 1999. Refining Rodinia: Geologic evidence for the Australia–western US connection in the Proterozoic. *GSA Today* 9 (10), 1–7. <https://doi.org/10.1130/GSAT-1999-10-01-science>.
- Kleinschmidt, G., Buggisch, W., 1994. Plate tectonic implications of the structure of the Shackleton Range. *Antarctica. Polarforschung* 63 (1), 57–62.
- Kleinschmidt, G., 2021. The Shackleton Range and its surroundings. In: Kleinschmidt, G. (Ed.), *The Geology of the Antarctic Continent*. Borntraeger Science Publishers, pp. 218–253.
- Krohne, N., Lisker, F., Kleinschmidt, G., Kluegel, A., Lauer, A., Estrada, S., Spiegel, C., 2018. The Shackleton Range (East Antarctica): an alien block at the rim of Gondwana? *Geol. Mag.* 155 (4), 841–864. <https://doi.org/10.1017/S0016756816001011>.
- Król, P., Kusiak, M.A., Dunkley, D.J., Whitehouse, M.J., Wilde, S.A., Augland, L.E., 2023. The record of geological processes in zircon from polymetamorphosed orthogneisses from the Napier Mountains, Napier complex, East Antarctica. *J. Mineral. Petrol. Sci.* 118, S012. <https://doi.org/10.2465/jmps.230419>.
- Laurent, O., Martin, H., Moyen, J.F., Doucelance, R., 2014. The diversity and evolution of late-Archean granulites: evidence for the onset of “modern-style” plate tectonics between 3.0 and 2.5 Ga. *Lithos* 205, 208–235. <https://doi.org/10.1016/j.lithos.2014.06.012>.
- Li, X., Tang, G., Gong, B., Yang, Y., Hou, K., Hu, Z., Li, Q., Liu, Y., Li, W., 2013. Qinghu zircon: a working reference for microbeam analysis of U–Pb age and Hf and O isotopes. *Chin. Sci. Bull.* 58 (36), 4647–4654. <https://doi.org/10.1007/s11434-013-5932-x>.
- Li, Z.X., Liu, Y., Ernst, R., 2023. A dynamic 2000–540 Ma Earth history: from cratonic amalgamation to the age of supercontinent cycle. *Earth Sci. Rev.* 238, 104336. <https://doi.org/10.1016/j.earscirev.2023.104336>.
- Liebmann, J., Ware, B., Hartnady, M.I., Kirkland, C.L., Timms, N.E., Evans, N.J., 2023. Albany K-Feldspar: a New Pb Isotope Reference Material. *Geostand. Geoanal. Res.* 47 (3), 637–655. <https://doi.org/10.1111/grg.12487>.
- Liebmann, J., Ware, B., Mole, D.R., Kirkland, C.L., Fraser, G., Waltenberg, K., Bodorkos, S., Huston, D.L., Evans, N.J., McDonald, B.J., Rankenburg, K., Datta, P., Tessalina, S., 2024. A crustal Pb isotope map of southeastern Australia. *Sci. Data* 11 (1), 1222. <https://doi.org/10.1038/s41597-024-03996-5>.
- Loewy, S.L., Dalziel, I.W.D., Pisarevsky, S., Connelly, J.N., Tait, J., Hanson, R.E., Bullen, D., 2011. Coats Land crustal block, East Antarctica: a tectonic tracer for Laurentia? *Geology* 39 (9), 859–862. <https://doi.org/10.1130/G32029.1>.
- Ludwig, K.R., 2012. *User's manual for Isoplot 3.75: a geochronological toolkit for Microsoft Excel*. Berkeley Geochronology Center special. Publication 5, 75.
- Mako, C.A., Williams, M.L., Karlstrom, K.E., Doe, M.F., Powicki, D., Holland, M.E., Gehrels, G., Pecha, M., 2015. Polyphase Proterozoic deformation in the four Peaks area, central Arizona, and relevance for the Mazatzal orogeny. *Geosphere* 11 (6), 1975–1995. <https://doi.org/10.1130/GES01196.1>.
- Maniar, P.D., Piccoli, P.M., 1989. Tectonic discrimination of granulites. *Geol. Soc. Am. Bull.* 101 (5), 635–643. [https://doi.org/10.1130/0016-7606\(1989\)101<0635:TDQG>2.3.CO;2](https://doi.org/10.1130/0016-7606(1989)101<0635:TDQG>2.3.CO;2).
- Marsh, J., Jørgensen, T., Petrus, J., Hamilton, M., Mole, D., 2019. U–Pb, trace element, and hafnium isotope composition of the Maniitsoq zircon: a potential new Archean zircon reference material. *Twenty-Ninth Annual Goldschmidt Conference* 2161.
- Marsh, P.D., 1983. The stratigraphy and structure of the metamorphic rocks of the Haskard Highlands and Otter Highlands of the Shackleton Range. *British Antarctic Survey Bulletin* 60, 23–43.
- Marschall, H.R., Hawkesworth, C.J., Storey, C.D., Dhuime, B., Leat, P.T., Meyer, H.P., Tamm-Buckle, S., 2010. The Anandagstoppane Granite, East Antarctica: evidence for Archean intracrustal recycling in the Kaapvaal–Grunehogna Craton from zircon O and Hf isotopes. *J. Petrol.* 51 (11), 2277–2301. <https://doi.org/10.1093/petrology/egq057>.
- Martin, H., Moyen, J.F., Rapp, R., 2009. The sanukitoid series: magmatism at the Archean–Proterozoic transition. *Earth Environ. Sci. Trans. R. Soc. Edinb.* 100 (1–2), 15–33. <https://doi.org/10.1017/S1755691009016120>.
- McAvaney, S.O., Wade, C.E., Jagodzinski, E.A., 2016. Tip top and Wertigo granites, c. 1775 Ma magmatism on northeastern Eyre Peninsula. *MESA Journal* 80 (1), 63–76.
- Middlemost, E.A., 1994. Naming materials in the magma/igneous rock system. *Earth Sci. Rev.* 37 (3–4), 215–224. [https://doi.org/10.1016/0012-8252\(94\)90029-9](https://doi.org/10.1016/0012-8252(94)90029-9).
- Mieth, M., Jacobs, J., Ruppel, A., Damaske, D., Läufer, A., Jokat, W., 2014. New detailed aeromagnetic and geological data of eastern Dronning Maud Land: Implications for refining the tectonic and structural framework of Sør Rondane, East Antarctica. *Precamb. Res.* 245, 174–185. <https://doi.org/10.1016/j.precamres.2014.02.009>.
- Möller, A., Mezger, K., Schenk, V., 2000. U–Pb dating of metamorphic minerals: Pan-African metamorphism and prolonged slow cooling of high pressure granulites in Tanzania. *East Africa. Precambrian Research* 104 (3–4), 123–146. [https://doi.org/10.1016/S0301-9268\(00\)00086-3](https://doi.org/10.1016/S0301-9268(00)00086-3).
- Moore, E.M., 1991. Southwest US–East Antarctic (SWEAT) connection: a hypothesis. *Geology* 19 (5), 425–428. [https://doi.org/10.1130/0091-7613\(1991\)019<0425:SUSEAS>2.3.CO;2](https://doi.org/10.1130/0091-7613(1991)019<0425:SUSEAS>2.3.CO;2).
- Mortimer, G.E., Cooper, J.A., Oliver, R.L., 1988. The geochemical evolution of Proterozoic granulites near Port Lincoln in the Gawler orogenic domain of South Australia. *Precamb. Res.* 40, 387–406. [https://doi.org/10.1016/0301-9268\(88\)90077-0](https://doi.org/10.1016/0301-9268(88)90077-0).
- Motoyoshi, Y., 1999. Ultra-High Temperature (UHT) Metamorphism of the Napier complex, East Antarctica. *Korean Journal of Polar Research* 10 (2), 85–90.
- Myrow, P.M., Pope, M.C., Goodge, J.W., Fischer, W., Palmer, A.R., 2002. Depositional history of pre-Devonian strata and timing of Ross orogenic tectonism in the central Transantarctic Mountains. *Antarctica. Geological Society of America Bulletin* 114 (9), 1070–1088. [https://doi.org/10.1130/0016-7606\(2002\)114<1070:DHOPDS>2.0.CO;2](https://doi.org/10.1130/0016-7606(2002)114<1070:DHOPDS>2.0.CO;2).
- Nasdala, L., Hofmeister, W., Norberg, N., Martinson, J.M., Corfu, F., Dörr, W., Kamo, S.L., Kennedy, A.K., Kronz, A., Reiners, P.W., Frei, D., Kosler, J., Wan, Y., Götz, J., Häger, T., Kröner, A., Valley, J.W., 2008. Zircon M257—a homogeneous natural reference material for the ion microprobe U–Pb analysis of zircon. *Geostand. Geoanal. Res.* 32 (3), 247–265. <https://doi.org/10.1111/j.1751-908X.2008.00914.x>.
- Pankhurst, R.J., Marsh, P.D., Clarkson, P.D., Oliver, R.L., James, P.R., Jago, J.B., 1983. A geochronological investigation of the Shackleton Range. *Antarctic Earth Science* 176–182.
- Paton, C., Hellstrom, J., Paul, B., Woodhead, J., Hergt, J., 2011. Iolite: Freeware for the visualisation and processing of mass spectrometric data. *J. Anal. At. Spectrom.* 26 (12), 2508–2518. <https://doi.org/10.1039/C1JA10172B>.
- Payne, J.L., Hand, M., Barovich, K.M., Reid, A., Evans, D.A., 2009. Correlations and reconstruction models for the 2500–1500 Ma evolution of the Mawson Continent. *Geol. Soc. Lond. Spec. Publ.* 323 (1), 319–355. <https://doi.org/10.1144/SP323.16>.
- Pearce, J.A., Gale, G.H., 1977. Identification of ore-deposition environment from trace-element geochemistry of associated igneous host rocks. *Geol. Soc. Lond. Spec. Publ.* 7 (1), 14–24. <https://doi.org/10.1144/GSL.SP.1977.007.01.03>.
- Pearce, J.A., Lippard, S.J., Roberts, S., 1984. Characteristics and tectonic significance of supra-subduction zone ophiolites. *Geol. Soc. Lond. Spec. Publ.* 16 (1), 77–94. <https://doi.org/10.1144/GSL.SP.1984.016.01.06>.
- Pearce, N.J., Perkins, W.T., Westgate, J.A., Gorton, M.P., Jackson, S.E., Neal, C.R., Chenery, S.P., 1997. A compilation of new and published major and trace element data for NIST SRM 610 and NIST SRM 612 glass reference materials. *Geostand. Newsl.* 21 (1), 115–144. <https://doi.org/10.1111/j.1751-908X.1997.tb00538.x>.
- Pettijohn, F.J., Potter, P.E., Siever, R., 1987. *Sand and sandstone*. Springer, New York.
- Peucat, J.J., Ménot, R.P., Monnier, O., Fanning, C.M., 1999. The Terre Adélie basement in the East-Antarctica Shield: geological and isotopic evidence for a major 1.7 Ga thermal event; comparison with the Gawler Craton in South Australia. *Precamb. Res.* 94 (3–4), 205–224. [https://doi.org/10.1016/S0301-9268\(98\)00119-3](https://doi.org/10.1016/S0301-9268(98)00119-3).
- Reid, A., Hand, M., Jagodzinski, E., Kelsey, D., Pearson, N., 2008. Paleoproterozoic orogenesis in the southeastern Gawler craton, South Australia. *Aust. J. Earth Sci.* 55 (4), 449–471. <https://doi.org/10.1080/08120090801888594>.
- Reid, A.J., Payne, J.L., 2017. Magmatic zircon Lu–Hf isotopic record of juvenile addition and crustal reworking in the Gawler Craton, Australia. *Lithos* 292, 294–306. <https://doi.org/10.1016/j.lithos.2017.08.010>.
- Riley, T.R., Flowerdew, M.J., Pankhurst, R.J., Millar, I.L., Whitehouse, M.J., 2020. U–Pb zircon geochronology from Haag Nunataks, Coats Land and Shackleton Range (Antarctica): constraining the extent of juvenile late Mesoproterozoic arc terranes. *Precamb. Res.* 340, 105646. <https://doi.org/10.1016/j.precamres.2020.105646>.
- Romer, T., Mezger, K., Schmädicke, E., 2009. Pan-African eclogite facies metamorphism of ultramafic rocks in the Shackleton Range, Antarctica. *Journal of Metamorphic Geology* 27 (5), 335–347. <https://doi.org/10.1111/j.1525-1314.2009.00820.x>.
- Roser, B.P., Korsch, R.J., 1988. Provenance signatures of sandstone–mudstone suites determined using discriminant function analysis of major-element data. *Chem. Geol.* 67 (1–2), 119–139. [https://doi.org/10.1016/0009-2541\(88\)90010-1](https://doi.org/10.1016/0009-2541(88)90010-1).
- Rudnick, R., Gao, S., 2003. Composition of the Continental Crust. *Treatise Geochem* 3:1–64. <https://doi.org/10.1016/B0-08-043751-6/03016-4>.
- Schaefer, B.F., 1998. *Insights into Proterozoic tectonics from the southern Eyre Peninsula, South Australia*. University of Adelaide, Adelaide. Unpublished PhD thesis.
- Schmädicke, E., Will, T.M., 2006. First evidence of eclogite facies metamorphism in the Shackleton Range, Antarctica: Trace of a suture between East and West Gondwana? *Geology* 34 (3), 133–136. <https://doi.org/10.1130/G22170.1>.
- Schubert, W., Olesch, M., 1995. Cordierite in the Shackleton Range, Antarctica: first recorded occurrence. *Polarforschung* 63 (2/3), 153–164.
- Schubert, W., Will, T., 1994. Granulite-facies rocks of the Shackleton Range, Antarctica. Conditions of formation and preliminary petrogenetic implications. *Chem. Erde* 54, 355–371.
- Shufeldt, O.P., Karlstrom, K.E., Gehrels, G.E., Howard, K.E., 2010. Archean detrital zircons in the Proterozoic Vishnu Schist of the Grand Canyon, Arizona: Implications for crustal architecture and Nuna supercontinent reconstructions. *Geology* 38 (12), 1349–1352. <https://doi.org/10.1130/G30417.1>.
- Skidmore, M.J., Clarkson, P.D., 1972. Physiography and glacial geomorphology of the Shackleton Range. *British Antarctic Survey Bulletin* 30, 69–80.
- Sláma, J., Kosler, J., Condon, D.J., Crowley, J.L., Gerdes, A., Hanchar, J.M., Horstwood, M.S.A., Morris, G.A., Nasdala, L., Norberg, N., Schaltegger, U., Schoene, B., Turbett, M.N., Whitehouse, M.J., 2008. Plešovice zircon—a new natural reference material for U–Pb and Hf isotopic microanalysis. *Chem. Geol.* 249 (1–2), 1–35. <https://doi.org/10.1016/j.chemgeo.2007.11.005>.
- Söderlund, U., Patchett, P.J., Vervoort, J.D., Isachsen, C.E., 2004. The ¹⁷⁶Lu decay constant determined by Lu–Hf and U–Pb isotope systematics of Precambrian mafic intrusions. *Earth Planet. Sci. Lett.* 219, 311–324. [https://doi.org/10.1016/S0012-821X\(04\)00012-3](https://doi.org/10.1016/S0012-821X(04)00012-3).
- Spencer, J.E., Doe, M.F., Domanik, K.J., 2024. Abundant 2480 Ma Detrital Zircons in ~1740 Ma Vishnu Schist in Northwestern Arizona: Derivation from the Matichewan Large Igneous Province? *Am. J. Sci.* 324, 6. <https://doi.org/10.2475/001c.115633>.
- Spencer, C.J., Kirkland, C.L., Roberts, N.M.W., Evans, N.J., Liebmann, J., 2020. Strategies towards robust interpretations of in situ zircon Lu–Hf isotope analyses. *Geosci. Front.* 11, 843–853. <https://doi.org/10.1016/j.gsf.2019.09.004>.
- Stacey, J.S., Kramers, J.D., 1975. Approximation of terrestrial lead isotope evolution by a two-stage model. *Earth Planet. Sci. Lett.* 26 (2), 207–221. [https://doi.org/10.1016/0012-821X\(75\)90088-6](https://doi.org/10.1016/0012-821X(75)90088-6).

- Stål, T., Reading, A.M., Halpin, J.A., Whittaker, J.M., 2019. A multivariate approach for mapping lithospheric domain boundaries in East Antarctica. *Geophys. Res. Lett.* 46 (17–18), 10404–10416. <https://doi.org/10.1029/2019GL083453>.
- Stern, R.A., Bodorkos, S., Kamo, S.L., Hickman, A.H., Corfu, F., 2009. Measurement of SIMS instrumental mass fractionation of Pb isotopes during zircon dating. *Geostand. Geoanal. Res.* 33 (2), 145–168. <https://doi.org/10.1111/j.1751-908X.2009.00023.x>.
- Storey, B.C., Macdonald, D.I., Dalziel, I.W., Isbell, J.L., Millar, I.L., 1996. Early Paleozoic sedimentation, magmatism, and deformation in the Pensacola Mountains, Antarctica: the significance of the Ross orogeny. *Geol. Soc. Am. Bull.* 108 (6), 685–707. [https://doi.org/10.1130/0016-7606\(1996\)108<0685:EPSMAD>2.3.CO;2](https://doi.org/10.1130/0016-7606(1996)108<0685:EPSMAD>2.3.CO;2).
- Strickland, A., Wooden, J.L., Mattinson, C.G., Ushikubo, T., Miller, D.M., Valley, J.W., 2013. Proterozoic evolution of the Mojave crustal province as preserved in the Ivanpah Mountains, southeastern California. *Precamb. Res.* 224, 222–241. <https://doi.org/10.1016/j.precamres.2012.09.006>.
- Stump, E., 1995. *The Ross Orogen of the Transantarctic Mountains*. Cambridge University Press.
- Talarico, F.M., Kroner, U., 1999. Geology and tectono-metamorphic evolution of the Read Group, Shackleton Range: a part of the East Antarctic Craton. *Terra Antarctica* 6 (3–4), 183–202.
- Tessensohn, F., Kleinschmidt, G., Talarico, F.M., Buggish, W., Brommer, A., Henjes-Kunst, F., Kroner, U., Millar, I.L., Zeh, A., 1999. Ross-age amalgamation of east and west Gondwana: evidence from the Shackleton Range, east Antarctica. *Terra Antarctica* 6 (3–4), 317–325.
- Tessensohn, F., Thomson, M.R.A., 1990. European geological initiative for the Shackleton Range. *Antarct. Sci.* 2 (3), 265–266. <https://doi.org/10.1017/S0954102090000359>.
- Tessensohn, F., Thomson, M.R.A., 1999. The Euroshack project: a brief outline. *Terra Antarctica* 6 (3), 175–182.
- Tyrrell, S., Haughton, P.D.W., Daly, J.S., Kokfelt, T.F., Gagnevin, D., 2006. The use of the common Pb isotope composition of detrital K-feldspar grains as a provenance tool and its application to Upper Carboniferous paleodrainage, northern England. *J. Sediment. Res.* 76 (2), 324–345. <https://doi.org/10.2110/jsr.2006.023>.
- Tyrrell, S., Haughton, P. D., Daly, J. S. & Shannon, P. M. (2012). The Pb isotopic composition of detrital K-feldspar: A tool for constraining provenance, sedimentary processes and paleodrainage. 10.3749/9780921294825.ch11.
- Vassallo, J.J., Wilson, C.J.L., 2002. Palaeoproterozoic regional-scale non-coaxial deformation: an example from eastern Eyre Peninsula, South Australia. *J. Struct. Geol.* 24 (1), 1–24. [https://doi.org/10.1016/S0191-8141\(01\)00043-8](https://doi.org/10.1016/S0191-8141(01)00043-8).
- Vermeesch, P., 2018. IsoplotR: a free and open toolbox for geochronology. *Geosci. Front.* 9 (5), 1479–1493.
- Whitehouse, M.J., Kamber, B.S., 2005. Assigning dates to thin gneissic veins in high-grade metamorphic terranes: a cautionary tale from Akilia, southwest Greenland. *J. Petrol.* 46 (2), 291–318. <https://doi.org/10.1093/petrology/egh075>.
- Wiedenbeck, M.A.P.C., Alle, P., Corfu, F.Y., Griffin, W.L., Meier, M., Oberli, F.V., Von Quadt, A., Roddick, J.C., Spiegel, W., 1995. Three natural zircon standards for U-Th-Pb, Lu-Hf, trace element and REE analyses. *Geostand. Newslett.* 19 (1), 1–23. <https://doi.org/10.1111/j.1751-908X.1995.tb00147.x>.
- Will, T.M., Frimmel, H.E., Zeh, A., Le Roux, P., Schmädicke, E., 2010. Geochemical and isotopic constraints on the tectonic and crustal evolution of the Shackleton Range, East Antarctica, and correlation with other Gondwana crustal segments. *Precamb. Res.* 180 (1–2), 85–112. <https://doi.org/10.1016/j.precamres.2010.03.005>.
- Will, T.M., Zeh, A., Gerdes, A., Frimmel, H.E., Millar, I.L., Schmädicke, E., 2009. Palaeoproterozoic to Palaeozoic magmatic and metamorphic events in the Shackleton Range, East Antarctica: constraints from zircon and monazite dating, and implications for the amalgamation of Gondwana. *Precamb. Res.* 172 (1–2), 25–45. <https://doi.org/10.1016/j.precamres.2009.03.008>.
- Williams, M.L., Jercinovic, M.J., 2002. Microprobe monazite geochronology: putting absolute time into microstructural analysis. *J. Struct. Geol.* 24 (6–7), 1013–1028. [https://doi.org/10.1016/S0191-8141\(01\)00088-8](https://doi.org/10.1016/S0191-8141(01)00088-8).
- Wingate, M.T., Pisarevsky, S.A., Evans, D.A., 2002. Rodinia connections between Australia and Laurentia: no SWEAT, no AUSWUS? *Terra Nova* 14 (2), 121–128. <https://doi.org/10.1046/j.1365-3121.2002.00401.x>.
- Wooden, J.L., Barth, A.P., Mueller, P.A., 2013. Crustal growth and tectonic evolution of the Mojave crustal province: Insights from hafnium isotope systematics in zircons. *Lithosphere* 5 (1), 17–28. <https://doi.org/10.1130/L218.1>.
- Wooden, J.L., Miller, D.M., 1990. Chronologic and isotopic framework for Early Proterozoic crustal evolution in the eastern Mojave Desert region, SE California. *J. Geophys. Res. Solid Earth* 95 (B12), 20133–20146. <https://doi.org/10.1029/JB095iB12p20133>.
- Wooden, J.L., Mueller, P.A., 1988. Pb, Sr, and Nd isotopic compositions of a suite of late Archean, igneous rocks, eastern Beartooth Mountains: implications for crust-mantle evolution. *Earth Planet. Sci. Lett.* 87 (1–2), 59–72. [https://doi.org/10.1016/0012-821X\(88\)90064-7](https://doi.org/10.1016/0012-821X(88)90064-7).
- Yoshida, M. (1995). Assembly of East Gondwanaland during the Mesoproterozoic and its rejuvenation during the Pan-African period. In: Yoshida, M. & Santosh, M. (eds.) *India and Antarctica During the Precambrian*: Memoir Geological Society of India, Bangalore, 22–45.
- Zartman, R.E., 1974. Lead isotopic provinces in the Cordillera of the western United States and their geologic significance. *Econ. Geol.* 69 (6), 792–805. <https://doi.org/10.2113/gsecongeo.69.6.792>.
- Zartman, R.E., Doe, B.R., 1981. Plumbotectonics—the model. *Tectonophysics* 75 (1–2), 135–162. [https://doi.org/10.1016/0040-1951\(81\)90213-4](https://doi.org/10.1016/0040-1951(81)90213-4).
- Zeh, A., Millar, I.L., Kroner, U., Gorz, I., 1999. The structural and metamorphic evolution of the northern Haskard Highlands, Shackleton Range, Antarctica. *Terra Antarctica* 6 (3), 249–268.
- Zhao, Y., Liu, X. H., Liu, X. C. & Song, B. (2003). Pan-African events in Prydz Bay, east Antarctica, and their implications for east Gondwana tectonics. 10.1144/GSL.SP.2003.206.01.12.