

Late Archaean granulite facies metamorphism in the Vestfold Hills, East Antarctica

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Abstract

Granulites of the Vestfold Hills record a pulsed end-Archaean to early Palaeoproterozoic M1–M2 evolution that is distinct from other Archaean areas in East Antarctica and cratonic domains placed adjacent to East Antarctica in Gondwana reconstructions. Pressure and temperature conditions of the end-Archaean to earliest Palaeoproterozoic (2501–2496 Ma) M1 granulite facies metamorphism in the Vestfold Hills have been constrained from mineral assemblages and thermobarometry of Fe-rich paragneisses. Reintegrated compositions of exsolved subcalcic clinopyroxenes and pigeonites in a metaironstone yield temperatures of 895 ± 35 °C, whilst reintegrated compositions of perthitic feldspars in semipelitic paragneisses give minimum estimates of 860 ± 30 °C. These results rule out the extreme ultrahigh temperature (UHT) conditions previously proposed for M1 in the Vestfold Hills. Pressures of metamorphism during M1 are estimated as 8.1 ± 0.9 kb at 850 ± 40 °C from hercynite + sillimanite + almandine + corundum and retrieved Fe–Mg–Al relations in orthopyroxene coexisting with garnet. A second metamorphic event, M2, occurred at 600–660 °C and 6–8 kb based on thermometry of recrystallised pyroxene neoblasts and thermobarometry applied to M2 garnet–quartz symplectites formed on orthopyroxene and garnet. The intervening emplacement of the magmatic Crooked Lake Gneiss Group precursors occurred at similar or shallower pressures prior to D2–M2, an event that caused tectonic interleaving and reactivation of the Vestfold Hills basement at mid-crustal depths in the earliest Palaeoproterozoic, prior to its unroofing to shallower levels (3–5 kb) by 2470 Ma. The lack of correlative Archaean histories in areas that were formerly adjacent in Gondwanan reconstructions is consistent with the Vestfold Hills region either being exotic to the East Antarctic Shield until the final (Neoproterozoic to Cambrian) amalgamation of Gondwana, or being accreted to part of East Antarctica in a Proterozoic event distinct from the Rayner–Eastern Ghats tectonism that united much of India with Antarctica at 1000–900 Ma.

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1. Introduction

Reconstruction of East Gondwana and its Precambrian precursors, and correlation of former East Gondwanan

and Rodinian crustal domains now dispersed between India, Asia, Australia and Antarctica, is dependent upon a thorough knowledge of the tectonothermal histories of those fragments that remain preserved in the present-day continents (Tingey, 1991; Fitzsimons, 2003). The Vestfold Hills form the only exposed portion of a late Archaean to Palaeoproterozoic (2520–2475 Ma) craton fragment that occurs in the East Antarctic Shield (Fig. 1) within the region that formerly was sited opposite east and

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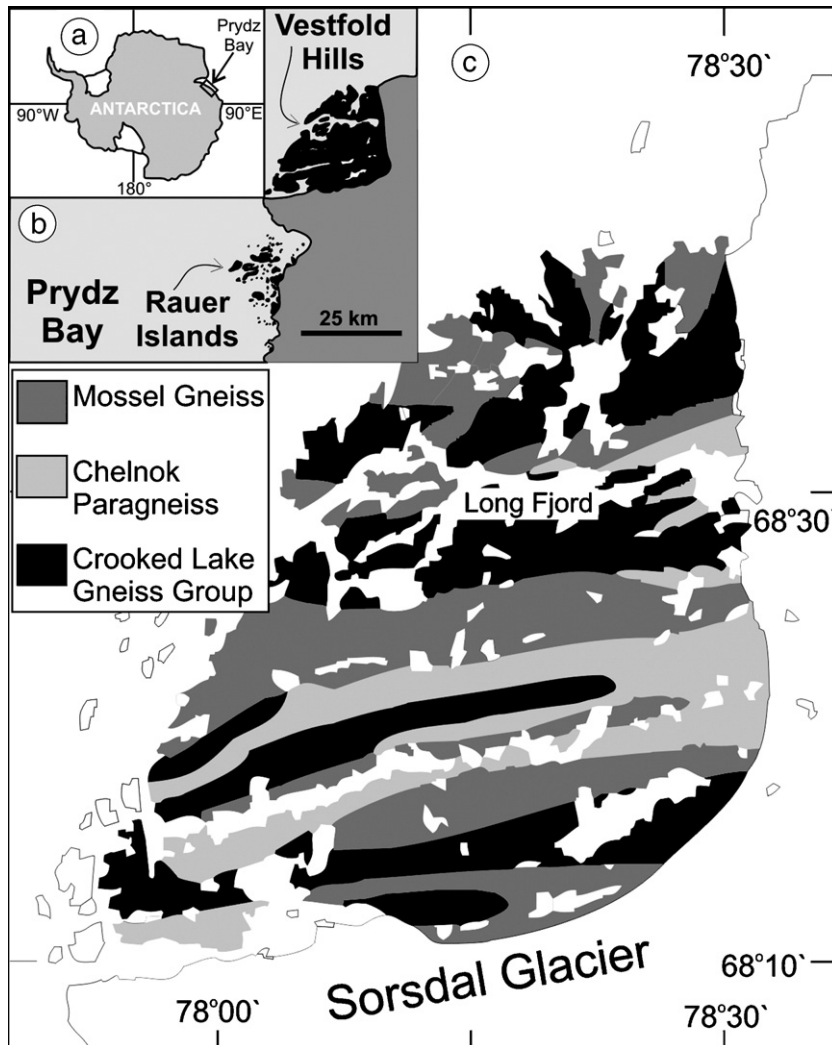


Fig. 1. a) Position of the Vestfold Hills in Antarctica. b) Position of the Vestfold Hills and Rauer Islands in the Prydz Bay. c) Simplified geology of the Vestfold Hills.

northeastern India according to recent Gondwana reconstructions (e.g. Dasgupta and Sengupta, 2003). Given this, and the possibilities for alternative correlations of the Vestfold Hills with a number of Archaean areas in northern India and Bangladesh (e.g. Fitzsimons, 2003), it is important that its Pressure–Temperature–time (P – T – t) history and geological evolution is well characterised and distinguished from those of neighbouring terranes in Antarctica.

Within Antarctica the craton that includes the Vestfold Hills is separated from other Archaean areas such as the Napier Complex of Enderby Land (Tingey, 1991; Harley and Black, 1997) and the Ruker Complex of the Southern Prince Charles Mountains (Mikhalsky et al., 2001; Boger et al., 2006) by Neoproterozoic and Cambrian high-grade

belts that form much of Prydz Bay and MacRobertson Land (Fitzsimons, 2000a,b; Boger et al., 2000, 2001; Harley, 2003). Previous geochemical, isotopic and geochronological studies demonstrate that the Archaean basement of the Vestfold Hills is distinct from the Napier Complex and Ruker Terrane (Black et al., 1991; Snape and Harley, 1996; Snape et al., 1997). Adjacent to the Vestfold Hills is the complex poly-deformed and poly-metamorphosed Rauer Terrane (Harley, 2003), which includes both Archaean and Mesoproterozoic crustal components, deformed under high-grade conditions in events at 1030–990 Ma and 530–510 Ma. Although the Archaean component of this terrane was initially thought to be a reworked equivalent of the Vestfold Hills basement (Collerson and Sheraton, 1986a, b), isotope geochemistry

and geochronology (Kinny et al., 1993; Harley et al., 1998) convincingly prove that this is not the case, and that the Vestfold Hills Archaean basement forms a unique cratonic fragment with its own P – T record and geological history (Oliver et al., 1982; Collerson et al., 1983; Harley and Hensen, 1990; Passchier et al., 1991; Tingey, 1991; Dirks et al., 1993; Harley and Fitzsimons, 1995; Snape and Harley, 1996).

Mineral assemblages in the gneisses of the Vestfold Hills have been used previously to estimate the pressure–temperature (P – T) conditions of metamorphism (Collerson and Sheraton, 1986a; Harley, 1993) (Table 1). However, these existing estimates are in conflict as to the peak conditions attained, and are based on lithologies that potentially record different tectonothermal events. Collerson and Sheraton (1986a) estimated P – T conditions in the range 8–10 kb and 900–1050 °C for the first recorded high-grade metamorphism (M1) using two-pyroxene thermometry and garnet–orthopyroxene barometry, applied to mafic and tonalitic orthogneisses and rare sub-aluminous and Fe-rich paragneisses. In contrast, Harley (1993) obtained peak M1 conditions of 830–880 °C at P < 8.5 kb based on the mineral assemblages sapphirine + spinel + orthopyroxene and sapphirine + cordierite + sillimanite + corundum + feldspar which occur in a compositionally distinct suite of Mg–Al rich granulites (*Taynaya Paragneiss*). As the latter rocks occur as boudinaged lenses that contain fabrics which are truncated by the main foliation or gneissic layering present in adjacent and enclosing orthogneisses, it is not clear whether the P – T conditions obtained from them by Harley (1993) are representative of the main metamorphism that affected the dominant orthogneisses and Fe-rich paragneisses of the Vestfold Hills. There are also contrasting estimates for the M2 event. Whereas Collerson and Sheraton (1986a) estimated granulite facies P – T conditions in the ranges 6–8 kb and 750–850 °C, Snape (1997) suggested that M2 took place under upper amphibolite-facies conditions on the basis of the observation of syntectonic (syn-D2) amphibole and plagioclase in dioritic and mafic *Crooked Lake Gneiss* lithologies.

In this work a suite of metapelites and calcsilicates from the relatively abundant Fe-rich paragneiss unit, the *Chelnok Paragneiss*, have been investigated in order to better constrain the P – T conditions of peak metamorphism, provide direct comparisons with data from the *Taynaya Paragneiss* in the northern Vestfold Hills, and improve the constraints on the post-peak P – T evolution of the terrane. These results are used to discuss the implications of the P – T record for the timing of exhumation and relationship of the basement rocks to adjacent regions in East Antarctica and former Gondwana.

Table 1

Timing and P – T conditions of events in the Vestfold Hills, based on previous work

Age	Event	Previous P and T estimates
Phanerozoic?	D7 brittle/ductile deformation events	
Ca. 1000–500 Ma	?Late alkaline dykes Amphibolite-facies metamorphism in the SW dykes/D6–M6	602–660 °C at 6.5–7.5 kb amphibolite-facies
1241 Ma	Intrusion of group 4 Fe-rich tholeiite	
?	Lamprophyre dyke intrusion	
1380 Ma	Intrusion of group 3 Fe-rich tholeiite dykes	4–5 kb
?	Lamprophyre dyke intrusion	
	D5 brittle deformation/pseudotachylite	≤4–5 kb
	D4, M4 ductile mylonite zones	Amphibolite-facies
1754 Ma	Intrusion of group 2 Fe-rich tholeiite dykes	
2240 Ma	Intrusion of group 2 (?) High-Mg tholeiite dykes	2–3 kb ^{###}
?	Intrusion of group 1 Fe-rich tholeiite dykes	
?	(Low-Ti and PM subgroup) Intrusion of group 1 High-Mg tholeiite dykes	
2472–2240 Ma	Intrusion of group 1 High-Mg tholeiite dykes (High-Ti subgroup)	3–5 kb (or shallower) ^{##}
2472 Ma	Intrusion of the <i>Snezhnyy Quartz Diorite</i> dykes**	
pre-2472 Ma	D3 deformation	
2475 Ma	D2 deformation and M2 metamorphism	750–850 °C at 6–8 kb* or upper amphibolite-facies ^c
2501 to 2475 Ma	Emplacement of <i>Crooked Lake Gneiss Group</i>	
2496 Ma	D1 deformation and M1 metamorphism	1000–1100 °C at 8–10 kb* or 830–880 °C at 3.5–8.5 kb [#]
2526 to 2501 Ma	Emplacement of the <i>Mossel Gneiss</i>	
Pre-2501 Ma	Deposition of the <i>Taynaya Paragneiss</i>	
Pre-2526 Ma	Deposition of the <i>Chelnok Paragneiss</i>	
ca. 2800 Ma	Inherited zircons in orthogneisses	

*Collerson and Sheraton (1986a); [#]Harley (1993); Snape (1997);

**These 2477 ± 5 Ma dykes are related to the Crooked Lake magmatism (Black et al., 1991) but have been included by Snape et al. (2001) in the *Vestfold Dyke Swarm*.

^{###}Seitz (1994); ^{####}Harley and Christy (1995).

2. Background geology

The Vestfold Hills basement comprises a high-grade terrane composed of dominant orthogneisses and less abundant paragneisses. Rocks of the Vestfold Hills are classified into five units according to their composition and to their relative age with respect to two main metamorphic-deformation events, D1–M1 and D2–M2. These units are the *Taynaya Paragneiss*, *Chelnok Paragneiss*, *Tryne Mafic Gneiss*, *Mossel Gneiss* and *Crooked Lake Gneiss Group* (Collerson et al., 1983; Snape et al., 1997, 2001). The oldest recognised lithologies include the metasedimentary *Taynaya Paragneiss* and *Chelnok Paragneiss* and the metabasic suite of the *Tryne Mafic Gneiss*, all of which pre-date the earliest recognised tectonothermal event, D1–M1. These rock units were intruded by pre-D1 tonalites and granodiorites that were later deformed and are now grouped into the *Mossel Gneiss* unit (Collerson et al., 1983). D1 deformation, which occurred under granulite facies conditions, is responsible for the upright to steeply-plunging isoclinal folding of these lithologies about steep axial surfaces now generally oriented broadly east–west (Oliver et al., 1982; Collerson et al., 1983; Collerson and Sheraton, 1986a; Snape and Harley, 1996).

Later granitic to gabbroic intrusives that collectively comprise the *Crooked Lake Gneiss Group* were emplaced into the deformed (D1–M1) basement consisting of the paragneisses, mafic gneiss and *Mossel Gneiss* prior to a second tectonothermal event, D2–M2 (Snape et al., 2001). Syn-deformational (syn-D2) magmatism is minor. All the units noted above were deformed in kilometre-scale, steeply-plunging open to isoclinal folds and sub-vertical coaxial high strain zones in D2, which was accompanied by M2 metamorphism (the D2–M2 event).

The principal magmatic accretion events forming the protoliths to the *Mossel Gneiss* and *Crooked Lake Gneiss Group*, and most of the high-grade tectonothermal evolution of the Vestfold Hills, occurred within the 50 million year time interval between 2520 and 2470 Ma (Black et al., 1991; Snape et al., 1997, 2001). D1–M1 and D2–M2 are further constrained by zircon U–Pb dating of bracketing magmatic rocks and rare syn-deformational leucosomes to have occurred at 2496 ± 1 Ma and 2475 ± 1 Ma respectively (Snape et al., 1997). A later deformation episode, D3, is manifested as E-trending open warping of D2 structures. These F3 open warps deform syn-D2 axial planar leucosomes that are dated at ca. 2475 Ma (Snape et al., 1997). D3 had no associated high-grade metamorphism, and will not be considered further here.

The post-D3 Proterozoic evolution of the Vestfold Hills is dominated by the intrusion of mafic dykes of the

Vestfold Dyke Swarms (Collerson and Sheraton, 1986b; Kuehner and Green, 1991; Snape et al., 2001) and the formation of localised shear-zones, mylonites, cataclastites and pseudotachylites (Passchier et al., 1991; Hoek et al., 1992). The mafic dykes intruded in several distinct pulses between 2240 Ma and 1241 Ma (Black et al., 1991; Lanyon et al., 1993), into basement that had already been cooled and partially exhumed to depths of 8–11 km by 2240 Ma (Seitz, 1994; Harley and Christy, 1995). In the Northern part of the Vestfold Hills the dykes are undeformed and display chilled margins and baked contacts (Collerson and Sheraton, 1986b; Passchier et al., 1991). Most of the dykes are unmetamorphosed with the exception of a small area in the Southwest of the Vestfold Hills. In this area some dykes are deformed, with xenomorphic garnet locally developed along dyke margins and in fracture zones within and across the dykes. Such features have been interpreted as the result of fluid access during a Neoproterozoic or younger (1000 to 500 Ma) overprint at P – T conditions of 5.6–7.5 kb and 600–660 °C (Kuehner and Green, 1991).

3. Criteria for selection of Chelnok Paragneiss samples

The mafic, tonalitic and felsic orthogneisses that dominate the Vestfold Hills are characterised by garnet-absent assemblages such as orthopyroxene + clinopyroxene + plagioclase + ilmenite + pargasite, clinopyroxene + plagioclase + ilmenite + pargasite + quartz and orthopyroxene + plagioclase/K-feldspar + ilmenite + pargasite + quartz (Collerson and Sheraton, 1986a; Snape, 1997). These assemblages provide only very broad constraints on the conditions of M1 and M2, and are not in general amenable to quantitative P – T calculations even with recent advances in THERMOCALC (Holland and Powell, 1998). Temperature-sensitive mineral assemblages such as sapphirine + orthopyroxene + spinel and cordierite + corundum + sillimanite + spinel occur in the *Taynaya Paragneiss* in the vicinity of Tryne Fjord and Taynaya Bay (Harley, 1993). However, P – T sensitive mineral assemblages of more regional extent are restricted to calcsilicate gneisses, garnet-bearing dioritic gneisses, metaironstones and aluminous metapelites that occur as rare boudins, lenses and horizons within the mainly semipelitic to quartzitic *Chelnok Paragneiss*.

Six samples of *Chelnok Paragneiss* from the Long Fjord region of the northern Vestfold Hills (Fig. 1) have been selected for detailed analysis in this study on the basis of petrographic observations on the full suite of

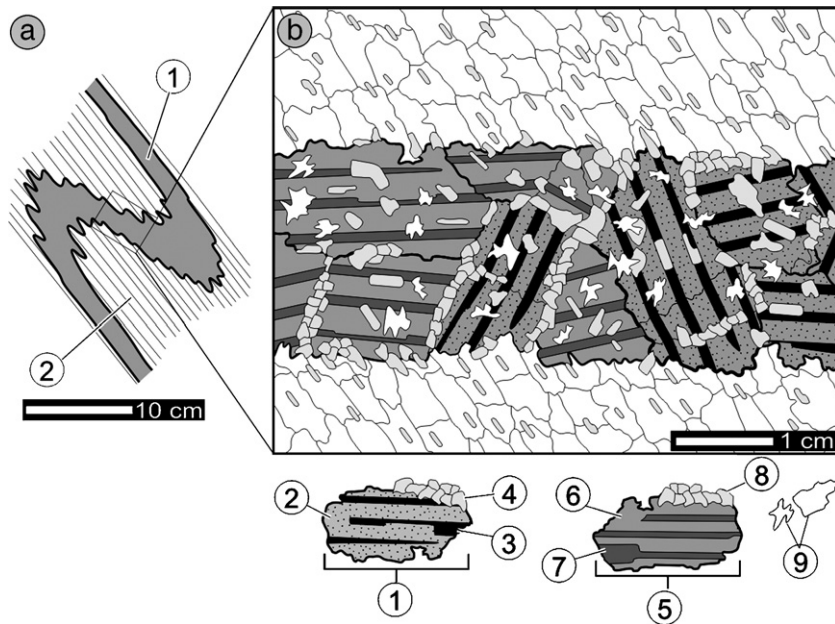


Fig. 2. a) Structural setting of meta-ironstone sample VH543. The pyroxene-rich band (grey layer) occurs within a quartz-rich and pyroxene-poor band, both of which are folded about a tight F1 fold that is intrafolial to the larger-scale gneissic layering. Continuous lines are S1 axial planar foliations to this isolated F1 fold. Scale bar is 20 cm. b) Microstructure of the pyroxene-rich layer in VH543. The pyroxene-rich layer is composed of: initial coarse grained *pigeonite* (1, Opx_1) now exsolved into orthopyroxene (2, Opx_{2a}) and clinopyroxene (3, Cpx_{2a}) lamellae. The exsolution lamellae are weakly deformed and replaced by more polygonal to lobate neoblastic grains of orthopyroxene and clinopyroxene (4, Opx_{3a} and Cpx_{3a}). Initial coarse *Fe-augite* grains (5, Cpx_1) are exsolved into clinopyroxene (6, Cpx_{2b}) and orthopyroxene (7, Opx_{2b}) lamellae. These exsolution lamellae are weakly deformed and replaced by polygonal to lobate neoblasts of orthopyroxene and clinopyroxene (8, Opx_{3b} and Cpx_{3b}). Within the quartz-rich layers (top and bottom) an axial plane foliation is defined by the dimensional preferred orientation of quartz (9) and neoblastic pyroxenes (Opx_{3a} , Opx_{3b} , Cpx_{3a} and Cpx_{3b}) (the scale-bar is 5 mm).

paragneisses collected by I. Snape in 1992–93 and 1993–94 (40 samples). These samples contain P – T sensitive mineral assemblages that, though rare, can be found in *Chelnok Paragneiss* units throughout the Vestfold Hills. The samples have been selectively collected from the *Chelnok Paragneiss* units that are located in the same broad region of the Vestfold Hills as the *Taynaya Paragneiss* studied by Harley (1993), in order to enable comparison of the P – T data. Moreover, based on the geological mapping of Snape et al. (2001) the *Chelnok Paragneiss* units from which the selected samples have been collected are located in regions of relatively low D2 strain, in which F1 folds and S1 fabrics can be recognised without being extensively overprinted on the metre to microscopic scale by D2 structures and S2 fabrics.

The following gneisses have been selected on the basis of the described criteria: a meta-ironstone (VH543); two garnet + orthopyroxene + plagioclase gneisses (VH536A and VH546C); two garnet + orthopyroxene + plagioclase + quartz + exsolved alkali-feldspar gneisses (VH207B and VH208); and a garnet-rich metapelite containing spinel and corundum (VH522). The garnet-bearing gneisses (VH

546A, VH446C, VH207B, VH208 and VH552) were all collected from typical structural domains in which F1 folds and S1 fabrics are well developed within *Chelnok Paragneiss*, locally discordant to S2 developed in adjacent and enclosing *Crooked Lake Gneiss Group* felsic orthogneisses. The meta-ironstone VH543 was collected from within a structural domain in which the meso-scale (tens of metres to decimetre) effects of F2 coaxial refolding are minor, so that D1–M1 structures are well preserved.

4. Petrography

4.1. Meta-ironstone VH543

Compositional layering on a 0.2–1-cm scale is defined by alternating quartz-rich and pyroxene-rich layers. This layering is folded about F1 close to isoclinal folds, associated with the development of an axial planar foliation, S1. This fabric is developed in the quartz rich layers, and is defined by the dimensional preferred orientations of both fine-grained pyroxenes (<2 mm) and larger deformed quartz grains (0.5–1 cm length) that preserve sutured grain boundaries (Fig. 2).

Within the mafic pyroxene-rich layers the pyroxene occurs in two different textural settings:

1) As coarse (1–10 mm), randomly oriented grains (Opx_1 and Cpx_1) that preserve subhedral to prismatic shapes. These coarse pyroxenes are inverted and exsolved (Figs. 2b and 3a and b) into 1–2 μm clinopyroxene (Cpx_{2a} and Cpx_{2b}) and orthopyroxene (Opx_{2a} and Opx_{2b}) lamellae along (001). Point counting of images of the exsolved pyroxenes indicate that Cpx_{2a} lamellae constitute 18–22% of the Opx_1 grains whilst Opx_{2b} lamellae constitute 23–27% of the Cpx_1 grains. A second generation of finer (<0.5 μm) exsolution lamellae intersect with a 45° angle the pre-existing lamellae and either dislocate them or kink them through an average of 10° (Fig. 3c). Both generations of exsolution lamellae, whose orientations are variable as they are controlled by the random orientations of their host grains, are undeformed by and hence *post-date* the main D1 event as manifested in the S1 axial planar foliation present in the quartz-rich layers. The lamellae are weakly deformed by a post-S1 event.

2) As finer grained (<2 mm) pyroxenes (Cpx_{3a} , Cpx_{3b} , Opx_{3a} and Opx_{3b}) that replace the inverted coarse pyroxenes along their grain boundaries and in microfractures. These finer recrystallised pyroxenes have a polygonal to lobate shapes, are either strain-free or only weakly deformed, and may show fine (<0.5 μm) exsolution lamellae.

4.2. Garnet–orthopyroxene–plagioclase–biotite gneiss VH546A and VH546C

This gneiss is composed of plagioclase (35%), orthopyroxene (30%), garnet (20%), biotite (10%), quartz and ilmenite (5%). Initial orthopyroxene (Opx_1), garnet (Grt_1) and plagioclase (Pl_{1a}) with minor amount of biotite (Bt_1) and ilmenite constitute an inequigranular granuloblastic assemblage of 0.5–1 cm grainsize (Fig. 4). Quartz is not part of the initial assemblage as it occurs only in retrograde symplectites involving garnet. Initial orthopyroxene (Opx_1) occurs both as isolated xenoblastic crystals and as groups of grains showing mutual sutured contacts. Opx_1 has an internal foliation defined by elongated plagioclase and biotite grains that have a different orientation with respect to orthopyroxene cleavages. Fine (1–2 mm) polygonal grains of plagioclase (Pl_{1b}) mantle coarse euhedral to sub-euhedral grains (Pl_{1a}). Biotite is a minor constituent of the rock and occurs both as inclusions in coarse pyroxenes and garnets as well as in the matrix. An early generation of garnet (Grt_1) forms isolated coarse grains within the plagioclase matrix and displays a slightly curved internal foliation defined by the alignment and preferred dimensional orientation of early

biotite and plagioclase. Finer (20–200 μm), more granuloblastic and inclusion-free neoblastic minerals (Opx_2 , Pl_2 and Bt_2) replace coarse minerals along 1–3 mm width high-temperature micro-shear zones (Figs. 4 and 5a) inferred to reflect the superposition of D2–M2 on the pre-existing D1–M1 assemblages.

A later generation of finer (100–300 μm) garnet (Grt_2) occurs as curvilinear symplectites together with worm-like

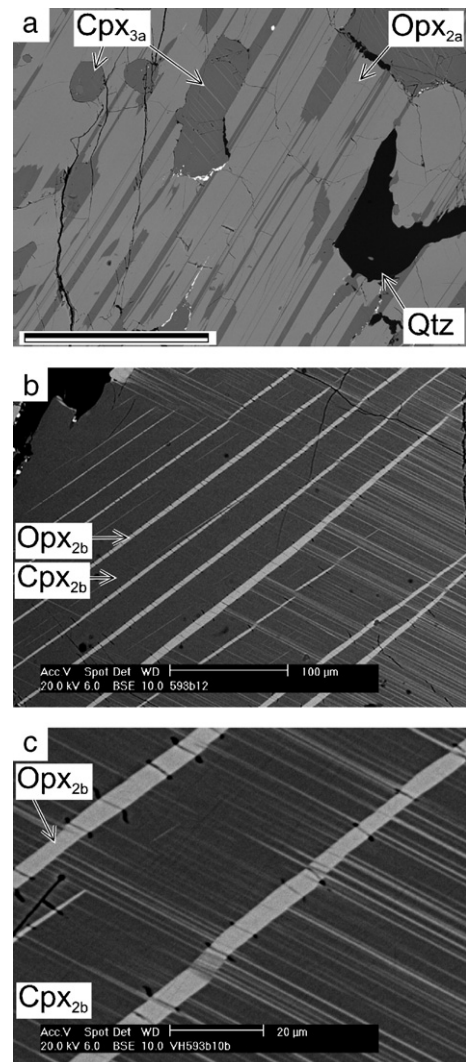


Fig. 3. a) Original orthopyroxene grain (Opx_1) now exsolved into large orthopyroxene_{2a} areas (Opx_{2a}) (light part) hosting smaller clinopyroxene_{2b} lamellae (Cpx_{2a}) (dark part). b) Original clinopyroxene₁ grain (Cpx_1) now exsolved into large clinopyroxene_{2b} areas (Cpx_{2b}) (dark part) hosting smaller NE–SW trending orthopyroxene_{2b} lamellae (Opx_{2b}) (light part). Note the second generation of finer NW–SE trending exsolution lamellae at high angle with the pre-existing ones. c) Higher magnification of lower-right part of Fig. 4b, showing the second generation of exsolution lamellae intersecting and dislocating the pre-existing ones (Opx_{2b} and Cpx_{2b}). Images are BSEI (Back Scattered Electron Images).

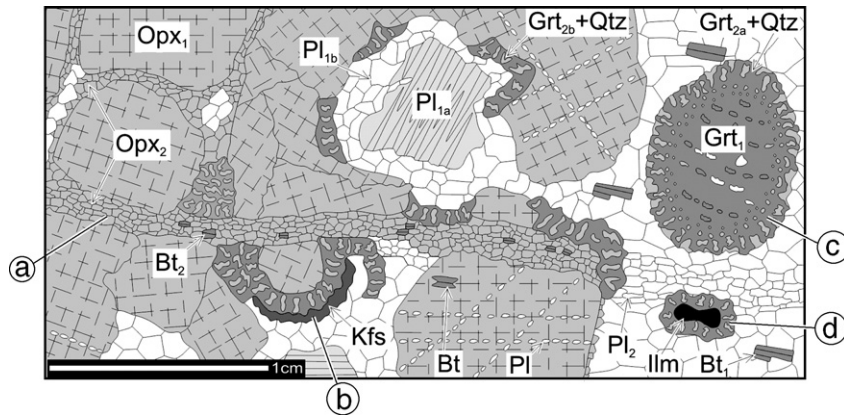


Fig. 4. Sketch of the microstructural setting of the garnet–orthopyroxene–plagioclase gneiss. Letters also refers to photos in Fig. 5. a) Fine grain and inclusion-free orthopyroxene₂ (Opx₂), biotite₂ (Bt₂) and plagioclase₂ (Pl₂) replace a coarse orthopyroxene₁ (Opx₁) with plagioclase inclusions (Pl) along a D2 micro-shear-zone. b) Composite corona of garnet_{2b} (Grt_{2b})–quartz (Qtz) symplectites and irregular K-feldspar (Kfs) rim formed between coarse orthopyroxene (Opx₁) and plagioclase (Pl_{1b}). Note the rational contact between garnet_{2b} and K-feldspar while plagioclase appears resorbed at the contact with K-feldspar. c) Garnet_{2a}–quartz symplectites between garnet₁ and plagioclase_{1b}. d) Garnet_{2b}–quartz symplectites between ilmenite (Ilm) and plagioclase_{1b}.

to blebby quartz at the contacts between both coarse and finer orthopyroxene and plagioclase (Fig. 5b) and therefore post-dates the formation of the micro-shear zones associated with D2–M2. Garnet_{2b}–quartz symplec-

tites also occur between garnet₁ and plagioclase (Fig. 5c) and around isolated ilmenite grains within the surrounding plagioclase (Fig. 5d). K-feldspar occurs both as an irregular corona between plagioclase and garnet₂–quartz

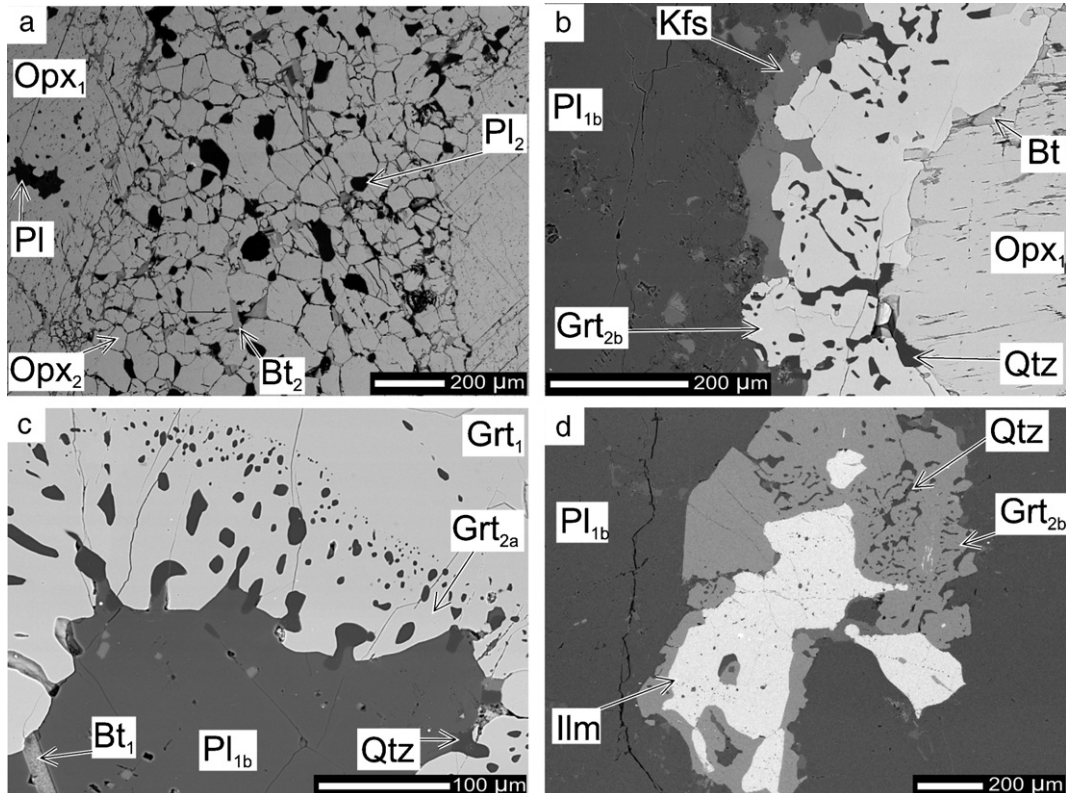


Fig. 5. BSEI microphotographs corresponding to lettered subdomains a to d highlighted in Fig. 4.

symplectites (Fig. 5b) around orthopyroxene and biotite, or as isolated patches inside plagioclase near contacts with the symplectites. Garnet_{2b}–quartz symplectites have rational contacts with the rimming K-feldspar whereas plagioclase is strongly resorbed at the K-feldspar interface, indicating that K-feldspar is a product of the reaction that formed garnet_{2b} + quartz rather than being a later phase replacing the symplectites. The formation of the composite garnet–quartz–K-feldspar coronas between plagioclase and orthopyroxene are consistent with the generalised reaction:



progressing from left to right with reaction. The role of K-feldspar is uncertain but could reflect the presence of saline fluid (Newton, 1986; Franz and Harlov, 1998; Harlov et al., 1998) or alternatively the breakdown of biotite during the corona-forming reaction, in which case some of the observed ilmenite may have formed along with the K-feldspar. During progress of reaction (1) garnet would form by preferentially consuming the ferrosilite component of orthopyroxene and anorthite component of plagioclase. As a consequence orthopyroxene would be expected to show a core to rim relative increase in enstatite content, garnet_{2b} would be expected to be grossular- and almandine-rich, and the remaining plagioclase enriched in its albite and K-feldspar components. However, as the amount of K-feldspar component that can be accommodated in the plagioclase solid solution is likely to be low at the reaction-forming temperature K-feldspar will appear as a separate phase.

4.3. Garnet–orthopyroxene–plagioclase–K-feldspar–quartz gneisses VH207B and VH208

These gneisses are composed of biotite (30%), orthopyroxene (30%), garnet (20%), plagioclase (10%) mesoperthitic feldspar (10%), quartz and ilmenite (10%). They are typified by polygonal granoblastic textures with grain sizes of 0.2–2 mm (Fig. 6). They differ from the gneisses described in the previous section (VH546A and VH546C) in the presence of prograde granoblastic quartz and mesoperthitic feldspar, and a higher abundance of biotite. Alkali feldspars, which occur as exsolved grains showing both lenticular and lamellar mesoperthite, are surrounded by plagioclase moats.

4.4. Corundum–spinel metapelite VH552

This quartz-deficient metapelite is composed of garnet (60%), biotite (20%), corundum (15%), spinel and il-

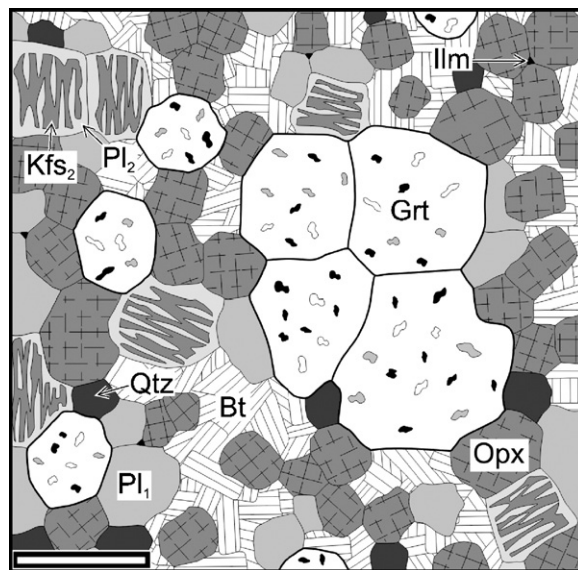


Fig. 6. Sketch of the microstructural setting of the garnet–orthopyroxene–plagioclase–K-feldspar–quartz. Granoblastic assemblage of garnet, orthopyroxene, plagioclase, K-feldspar and quartz. Scale-bar is 2 mm.

menite (5%). Garnet occurs as 1–2-centimetre grains set in a finer 1–2-millimetre-grain matrix. Coarse garnet contains inclusions of spinel, biotite, ilmenite and rare corundum. Biotite and spinel inclusions occur as resorbed grains usually reduced to amoeboid-shaped grains. Ilmenite within the garnet occurs as rounded or needle-like inclusions arranged in different sets of intersecting trails. Corundum inclusions occur as polygonal–skeletal grains adjacent to resorbed spinel and biotite. The matrix is composed of weakly foliated biotite flakes together with prismatic–idioblastic to sub-idioblastic grains of corundum. Green spinel is rare and is rimmed by corundum. A potential generalised reaction involving the formation of garnet and corundum, at the expense of spinel and sillimanite in this rock is:



This reaction, which progresses from left to right to produce a final assemblage of spinel + garnet + corundum on exhaustion of sillimanite, can be used to provide a limiting condition for the *P–T* stability of garnet + corundum in this instance.

5. Mineral chemistry

Minerals were analysed on a Cameca Camebax electron microprobe at the Grant Institute of Earth Science, University of Edinburgh, under operating conditions of 20 kV and 10 nA beam current. Metal, metal oxide and

Table 2
Mineral compositions of pyroxenes

Phase									Pyroxene								
Thin section	VH543								VH546A			VH546C		VH207B		VH208	
	Opx ₁				Cpx ₁				Opx ₁		Opx ₂	Opx ₁					
Position	Opx _{2a}	Cpx _{2a}	Opx _{3a}	Cpx _{3a}	Opx _{2b}	Cpx _{2b}	Opx _{3b}	Cpx _{3b}	Core	Rim		Core	Rim	Core	Rim	Core	Rim
SiO ₂	48.18	46.96	46.76	48.94	46.44	48.00	45.63	47.85	49.92	50.99	49.7	50.1	50.3	49.05	50.38	48.98	50.14
TiO ₂	0.17	0.07	0.08	0.17	0.04	0.08	0.01	0.07	0.01	0.04	0.03	0.05	0.02	0.08	0.08	0.13	0.07
Al ₂ O ₃	1.54	0.58	0.84	1.51	0.92	1.40	0.85	1.3	0.93	0.74	0.73	1.14	0.94	4.23	2.51	4.1	2.59
Cr ₂ O ₃	0.02	0.00	0.00	0.02	0.01	0.01		0.02	0.03	0.08	0.01	0.11	0.12	0	0.01	0.01	0
FeO	22.91	45.01	44.39	22.07	46.56	26.23	47.83	25.37	32.84	31.4	32.12	32.58	31.47	27.11	26.95	27.22	26.91
MnO	0.14	0.32	0.30	0.14	0.3	0.17	0.33	0.15	0.3	0.25	0.28	0.3	0.24	0.42	0.4	0.44	0.36
MgO	5.64	6.38	6.53	5.63	4.22	3.62	4.06	3.78	15.17	16.27	15.33	15.21	15.86	18.46	19.33	18.26	19.27
CaO	20.67	0.63	0.85	21.08	1.39	20.10	0.57	20.78	0.26	0.36	0.38	0.32	0.38	0.04	0.04	0.05	0.05
Na ₂ O	0.22	0.02	0.03	0.20	0.03	0.19	0.04	0.15	0.02	0.01	0.01	0.04	0.02	0.01	0.01	0.01	0.01
K ₂ O	0	0	0.01	0.01	0.01	0.01	0	0	0.01	0.01	0	0.01	0.01	0	0.01	0.01	0.01
Total	99.47	99.96	99.78	99.78	99.93	99.82	99.32	99.47	99.49	100.15	98.59	99.86	99.36	99.41	99.73	99.2	99.41
Si	1.935	1.964	1.96	1.95	1.968	1.948	1.959	1.946	1.963	1.978	1.969	1.962	1.97	1.874	1.916	1.878	1.913
Ti	0.005	0.001	0.002	0.005	0.001	0.003	0	0.002	0	0.001	0.001	0.001	0.001	0.19	0.113	0.185	0.116
Al	0.073	0.039	0.042	0.071	0.046	0.067	0.043	0.063	0.043	0.034	0.034	0.052	0.043	0.002	0.002	0.004	0.002
Cr	0	0	0	0.001	0	0	0	0.001	0.001	0.003	0	0.003	0.004	0	0	0	0
Fe	0.77	1.694	1.556	0.722	1.65	0.89	1.718	0.863	1.08	1.019	1.064	1.067	1.031	0.866	0.857	0.872	0.858
Mn	0.005	0.012	0.011	0.005	0.011	0.006	0.012	0.005	0.01	0.008	0.009	0.01	0.008	0.014	0.013	0.014	0.012
Mg	0.338	0.263	0.408	0.335	0.267	0.219	0.26	0.229	0.889	0.941	0.905	0.888	0.926	1.051	1.096	1.043	1.096
Ca	0.89	0.042	0.038	0.9	0.063	0.874	0.026	0.905	0.011	0.015	0.016	0.013	0.016	0.002	0.002	0.002	0.002
Na	0.017	0.001	0.002	0.016	0.003	0.015	0.003	0.012	0.002	0.001	0.001	0.003	0.002	0.001	0.001	0.001	0.001
K	0	0	0	0	0.001	0.001	0	0	0	0	0	0	0	0	0.001	0	0
Total	4.033	4.016	4.018	4.004	4.010	4.023	4.021	4.026	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000
X_{Mg}	0.30	0.20	0.21	0.31	0.14	0.2	0.13	0.21	0.45	0.48	0.46	0.45	0.47	0.548	0.561	0.545	0.56
Fe ³⁺	0.063	0.032	0.036	0.034	0.007	0.046	0.042	0.052	0.03	0.006	0.027	0.021	0.014	0.058	0.052	0.052	0.056
Wo	42	2	2	44	3	42	1	43	1	1	1.000	1	1	0	0	0	0
En	19	20	21	18	14	12	13	13	45	48	46	45	47	54	56	54	56
Fs	39	78	77	38	83	46	86	44	54	51	53	54	52	46	44	46	44
X_{Al}^*									0.021*	0.0145*	0.016*	0.023*	0.0185*				
X_{Al}														0.064	0.028	0.063	0.028

In VH543 the structural formula is calculated on the basis of 8 cations, and Fe³⁺ and members calculated according to Lindsley and Andersen (1983). In VH207B and VH208 the structural formula is calculated on the basis of 8 cations and Fe³⁺ then calculated from charge deficiency.

$$X_{\text{Al}}^* = [\text{Al} - 2\text{Ti} - \text{Cr}]/2. \quad X_{\text{Al}} = [\text{Al} - 2\text{Ti} - \text{Cr} - \text{Fe}^{3+}]/2.$$

natural mineral standards were used and data reduced using on-line PAP correction procedures. Representative compositions of clinopyroxene, garnet, biotite, feldspars, corundum and spinel are presented as [Tables 2–7](#), available as electronic appendices.

5.1. Clinopyroxene

The original pre-exsolution pyroxene compositions in meta-ironstone VH543 have been restored (reintegrated) by rastering sets of adjacent $25 \mu^2$ areas along line traverses perpendicular to the exsolved lamellae and integrating the measured compositions. A minimum of twenty lamellae were intersected along each traverse. Six clinopyroxene and five orthopyroxene grains were analysed and the integrated analyses for each averaged, yielding typical uncertainties in Wo, En and Fs components of less than ± 1 mol%. Following this procedure, the reintegrated compositions of Cpx₁ in meta-ironstone VH543 indicate that these pyroxenes were Fe-augites with Wo_{32–34}En_{12–13}Fs₅₄ and X_{Mg} 0.18–0.19 ([Fig. 7](#) and [Table 2](#)). Exsolved lamellae of Cpx_{2b} are Wo_{42–45}En_{12–13}Fs_{43–46} and X_{Mg} 0.20–0.22 while Opx_{2b} are Wo_{2–3}En_{13–14}Fs_{83–85} X_{Mg} 0.13–0.14. Neoblastic polygonal Opx_{3b} grains are Wo_{1–2}En_{13–14}Fs_{84–85} and X_{Mg}

0.13–0.14 while Cpx_{3b} are Wo_{43–45}En_{12–14}Fs_{42–44} and X_{Mg} 0.21–0.23.

5.2. Orthopyroxene

5.2.1. Meta-ironstone VH543

Restored (reintegrated) compositions of Opx₁ indicate that these pyroxenes were pigeonites with Wo_{10–11}En₂₀Fs_{69–70} and X_{Mg} 0.22. Exsolved lamellae of Cpx_{2a} are Wo_{41–45}En_{18–19}Fs_{37–41} and X_{Mg} 0.29–0.32 while Opx_{2a} are Wo_{1–4}En_{20–21}Fs_{76–79} and X_{Mg} 0.20–0.21 ([Fig. 7](#) and [Table 2](#)). Neoblastic polygonal Opx_{3a} grains are Wo_{1–2}En_{20–21}Fs_{77–79} and X_{Mg} 0.20–0.21 while Cpx_{3a} are Wo_{44–45}En₁₈Fs_{37–38} and X_{Mg} 0.31–0.32. The second generation of exsolution lamellae which intersect the coarser ones in Cpx₁ and Opx₁ ([Fig. 7](#)) as well as the exsolution lamellae in the neoblastic polygonal grains (Opx_{3a} and Cpx_{3a}) are too fine to be analysed.

5.2.2. Garnet–orthopyroxene–plagioclase gneiss VH546A and VH546C

Orthopyroxene is Wo₁En_{44–50}Fs_{40–55} with X_{Mg} in the range 0.45–0.50 ([Fig. 8](#) and [Table 2](#)). Coarse Opx₁ have X_{Mg} values of 0.45–0.50 with a core to rim increase of

Table 3
Garnet analyses

Phase				Garnet					
	Thin section	VH546A		VH546C	VH207B	VH208		VH552	
Position		Grt ₁ core	Grt _{2a} rim	Grt _{2b} rim	Core	Rim	Core	Rim	
SiO ₂		37.61	36.58	37.29	37.48	37.47	37.24	36.91	37.17
Al ₂ O ₃		20.89	20.87	20.61	20.68	20.43	20.94	20.68	21.1
Cr ₂ O ₃		0.03	0.07	0.52	0.37	0	0.02	0.02	0
FeO		31.84	30.98	34.05	32.55	32.83	33.62	32.87	34.84
MnO		0.63	0.74	0.87	0.87	1.69	1.78	1.93	0.04
MgO		5.89	5.3	3.46	3.72	5.91	5.58	5.68	6.11
CaO		2.85	4.29	4.01	4.83	0.76	0.71	1.04	0.62
Total		99.76	98.86	100.81	100.5	99.1	99.92	99.17	99.89
Si		2.972	2.918	2.961	2.974	2.996	2.959	2.952	2.949
Ti		0.001	0.002	0.000	0	0.002	0.002	0.002	0.001
Al		1.944	1.962	1.929	1.933	1.925	1.962	1.949	1.970
Cr		0.002	0.005	0.033	0.023	0.000	0.001	0.001	0.000
Fe		2.104	2.067	2.261	2.16	2.195	2.234	2.198	2.290
Mn		0.042	0.05	0.058	0.058	0.114	0.120	0.131	0.002
Mg		0.694	0.63	0.410	0.44	0.704	0.661	0.677	0.723
Ca		0.241	0.367	0.348	0.411	0.065	0.061	0.089	0.053
Total		8.000	8.001	8.000	7.999	8.000	8.000	8.000	7.988
X_{Mg}		0.248	0.23	0.153	0.17	0.243	0.228	0.236	0.239
Fe ³⁺		0.108	0.193	0.116	0.103	0.081	0.114	0.141	0.13
Alm		68	66	74	71	71	73	71	74
Sps		1	2	2	2	4	4	4	0
Pyr		23	20	13	14	23	21	22	24
Grs		8	12	11	13	2	2	3	2

Structural formulae calculated on the basis of 8 cations. Fe³⁺ is calculated from charge deficiency.

Table 4
Biotite analyses

Phase	Biotite							
Thin section	VH546A			VH546C		VH207B	VH208	VH552
Position	Inclusion	Bt ₁	Bt ₂	Inclusion	Bt ₁			
SiO ₂	37.03	35.38	35.09	36.19	36.64	37.1	36.61	34.75
TiO ₂	6.28	5.09	4.59	6.3	3.67	4.32	4.65	4.18
Al ₂ O ₃	14.69	14.02	13.39	14.31	14.34	13.85	14.37	17.18
Cr ₂ O ₃	0.24	0.08	0.26	0.21	0.21	0.01	0.03	0.01
FeO	15.41	19.27	22.14	15.76	17.85	14.73	15.08	16.06
MnO	0.03	0.02	0.05	0.02	0.04	0.02	0.05	0.01
MgO	13.09	11.06	9.59	12.72	12.6	14.98	14.18	13.13
CaO	0.09	0.04	0.01	0.05	0.06	0.03	0	0.01
Na ₂ O	0.07	0.06	0.05	0.05	0.06	0.16	0.13	0.24
K ₂ O	8.53	9.21	8.73	8.82	9.24	9.78	9.84	9.35
Total	95.46	94.23	93.9	94.43	94.71	94.98	94.94	94.92
Si	5.512	5.487	5.532	5.479	5.592	5.580	5.522	5.258
Al ^{IV}	2.488	2.513	2.468	2.521	2.408	2.420	2.478	2.742
Al ^{VI}	0.090	0.049	0.020	0.033	0.171	0.035	0.077	0.322
Ti	0.703	0.593	0.545	0.718	0.421	0.488	0.527	0.475
Cr	0.029	0.009	0.032	0.025	0.026	0.001	0.004	0.001
Fe	1.918	2.499	2.919	1.996	2.278	1.853	1.903	2.033
Mn	0.004	0.002	0.006	0.003	0.005	0.003	0.006	0.001
Mg	2.905	2.556	2.255	2.872	2.867	3.359	3.188	2.962
Ca	0.015	0.007	0.002	0.008	0.010	0.004	0.000	0.001
Na	0.019	0.017	0.015	0.016	0.016	0.046	0.038	0.072
K	1.619	1.822	1.756	1.704	1.799	1.877	1.894	1.804
Total	15.302	15.554	15.550	15.375	15.593	15.666	15.637	15.670
X _{Mg}	0.6	0.51	0.44	0.59	0.56	0.64	0.63	0.59

Structural formula calculated on the basis of 23 oxygens.

0.05 units while finer Opx₂ have X_{Mg} of 0.46 (Fig. 8). The increase of X_{Mg} is in agreement with the progress of reaction (1) defined above. Al₂O₃ contents are low, in the range of 0.71–1.14 wt.% for Opx₁, with a rimwards decrease of 0.04 units. Opx₂ has an Al₂O₃ content of 0.71–0.76 wt.% that is lower than the cores of Opx₁. The Fe³⁺ contents of orthopyroxenes were estimated from charge balance after normalisation to 8 cations. The resultant estimate of 0.01–0.05 c.p.f.u. (cations per formula unit) is of similar magnitude to the measured Al, which is 0.03–0.05 c.p.f.u. The reliable estimation of Fe³⁺ is important, especially in the Al-poor pyroxenes considered here, since Fe³⁺ contributes to the calculation of the Al-tschermaks component ($X_{Al}=[Al-2Ti-Cr-Fe^{3+}]/2$) which is used in Al-based thermobarometers (e.g. Harley and Green, 1982; Harley, 1984b). In the sample considered here the calculated Fe³⁺ content is very high with respect to Al, yielding in most cases zero or negative Al-tschermak values. However, the estimated Fe³⁺ is within the range of analytical uncertainty and therefore in the thermobarometric calculations the Al-tschermaks component has been computed considering all the iron as ferrous ($X_{Al}=[Al-2Ti-Cr]/2$).

5.2.3. Garnet–orthopyroxene–plagioclase–K-feldspar–quartz gneiss VH207B and VH208

Othopyroxene in VH207B and VH208 is Wo₀ En_{54–56}Fs_{44–46} and is more Mg- and Al-rich than those in VH546A and VH546C. The Fe^{tot} content of 0.886–0.849 is up to 0.2 c.p.f.u. lower than in the orthopyroxenes of samples VH546A and VH546C (Fig. 8 and Table 2). The X_{Mg} content is in the range 0.561–0.541, and shows a core to rim increase antithetic to an observed rimwards decrease in X_{Al} (0.064 to 0.023).

5.3. Garnet

5.3.1. Garnet–orthopyroxene–plagioclase gneiss VH546A and VH546C

Garnet is Alm_{66–73}Grs_{7–15}Pyr_{12–23}Sps_{1–2} (Fig. 9 and Table 3) with $X_{Mg}=0.15–0.25$. Elemental maps of coarse garnet₁ rimmed by symplectitic garnet_{2a} are complex, displaying composite element zoning (Figs. 10, 11 and 12). Isolated and irregularly shaped domains in the garnet cores are Alm_{68–69}Grs_{7–8}Pyr_{22–23}Sps_{1–2} and preserve lower-Ca, higher-Mg, higher-Fe and higher- X_{Mg} (=0.25) contents with respect to the surrounding garnet grain areas, whilst

Table 5
Plagioclase analyses

Phase	Plagioclase															
Thin section	VH546A					VH546C				Thin section	VH207B			VH208		
Position	Inclusion	Pl _{1a} core	Pl _{1a} rim	Pl _{1b}	Pl ₂	Inclusion	Pl _{1a} core	Pl _{1a} rim	Pl _{1b}	Position	Pl ₁ core	Pl ₁ rim	Pl ₂	Pl ₁ core	Pl ₁ rim	Pl ₂
SiO ₂	58.86	59.59	59.42	59.42	60.51	60.58	59.95	59.87	60.36	SiO ₂	63.62	64.39	65.59	62.77	63.81	63.94
TiO ₂	0	0	0	0	0	0	0	0	0	TiO ₂	0	0	0	0	0	0
Al ₂ O ₃	24.85	24.63	24.66	24.71	24.54	24.32	25.15	24.69	24.42	Al ₂ O ₃	22.26	21.89	20.99	22.65	21.63	18.4
FeO	0.86	0.04	0.04	0.59	0.1	0.03	0.06	0.04	0.04	FeO	0.11	0.2	0.04	0.1	0.31	0.11
MgO	0.05	0.02	0.01	0.02	0.01	0.01	0.02	0.01	0.01	MgO	0.02	0.02	0.02	0.03	0	0.03
CaO	7.72	7.19	7.1	6.76	6.59	6.46	7.3	6.89	6.89	CaO	3.63	3.19	0.03	0.04	0.01	0.54
Na ₂ O	7.08	7.27	7.44	7.29	7.67	7.75	6.95	7.47	7.5	BaO	0.02	0.02	2.01	3.78	2.64	0.02
K ₂ O	0.27	0.41	0.35	0.51	0.22	0.21	0.4	0.28	0.39	Na ₂ O	9.54	9.77	10	9.47	10.01	1.25
Total	99.69	99.13	99.02	99.3	99.63	99.35	99.83	99.25	99.62	K ₂ O	0.24	0.23	1.02	0.21	0.17	14.9
										Total	99.44	99.71	99.7	99.07	98.65	99.19
Si	2.649	2.682	2.678	2.675	2.702	2.711	2.676	2.687	2.7	Si	2.828	2.851	2.901	2.804	2.854	2.983
Al	1.318	1.306	1.309	1.311	1.291	1.283	1.323	1.306	1.287	Al	1.166	1.142	1.094	1.192	1.14	1.012
Mg	0.003	0.001	0.001	0.001	0.001	0.001	0.001	0.001	0	Fe	0.004	0.007	0.001	0.004	0.012	0.004
Fe	0.032	0.001	0.001	0.022	0.004	0.001	0.002	0.001	0.002	Ti	0	0	0	0	0	0
Na	0.618	0.634	0.65	0.636	0.664	0.673	0.601	0.65	0.65	Mg	0.001	0.001	0.001	0.001	0.005	0.002
Ca	0.372	0.347	0.343	0.326	0.315	0.309	0.349	0.331	0.33	Na	0.822	0.838	0.858	0.82	0.868	0.113
K	0.016	0.023	0.02	0.029	0.012	0.012	0.023	0.016	0.022	Ca	0.173	0.151	0.095	0.181	0.127	0.001
Total	5.01	4.994	5.002	5.002	4.99	4.99	4.975	4.993	4.993	K	0.014	0.013	0.057	0.012	0.01	0.887
Ab	61	63	64	64	67	68	62	65	65	Ba	0	0	0.001	0.001	0	0.01
An	37	35	34	33	32	31	36	33	33	Total	5.008	5.003	5.01	5.016	5.016	5.012
Or	2	2	2	3	1	1	2	2	2	Ab	81	84	85	81	86	11
										An	17	15	9	18	13	0
										Or	1	1	6	1	1	88
										Cs	0	0	0	0	0	1

Structural formulae calculated on the basis of 8 cations. P: polygonal. EL: exsolution lamellae. I: inclusion.

Table 6
K-feldspar analyses

Phase	K-feldspar			
Thin section	VH546A	VH546C	VH207B	VH208
	Kfs	Kfs	Kfs ₂	Kfs ₂
Position	C	C	EL	EL
SiO ₂	63.72	64.34	64.37	66.67
TiO ₂	0.03	0	0	0
Al ₂ O ₃	18.4	18.22	18.52	20.38
FeO	0.12	0.16	0.13	0.04
MgO	0.01	0.01	0.03	0.02
BaO	1.28	0.45	0.35	0.04
CaO	0.05	0.05	0.05	1.21
Na ₂ O	0.69	0.93	1.53	10.59
K ₂ O	15.52	15.34	14.34	0.73
Total	99.82	99.5	99.32	99.68
Si	2.976	2.993	2.986	2.939
Al	1.013	0.999	1.012	1.059
Fe	0.005	0.006	0.005	0.002
Ti	0.001	0	0	0
Mg	0.001	0.001	0.002	0.002
Na	0.063	0.084	0.138	0.905
Ca	0.003	0.003	0.003	0.057
K	0.925	0.91	0.849	0.041
Ba	0.023	0.008	0.006	0.001
Total	5.01	5.004	5.001	5.006
Ab	6	8	14	90
An	1	0	0	6
Or	91	91	85	4
Cs	2	1	1	0

Structural formulae calculated on the basis of 8 cations. EL: exsolution lamellae. C: coronite.

Mn shows no zoning (Figs. 10 and 11). Classic retrograde zoning (increase in Fe/Mg) occurs towards the grain boundaries, along micro-fractures and along inclusion trails. Coarse garnet₁ is rimmed by a symplectitic garnet_{2a}, which seals the garnet₁ micro-fractures and inclusion trails. These rims and narrow zones of garnet_{2a} are Alm_{65–68}Grs_{10–15}Pyr_{19–21}Sps_{1–2}, with X_{Mg} of 0.21–0.24. Grossular contents of garnets in the symplectites are zoned parallel to the quartz radial inclusions (Fig. 12), the difference in grossular content between high-Ca and low-Ca areas being in the range 4–5 mol%. High-Ca areas in the symplectites have also higher Alm (by 3 mol%) and lower X_{Mg} (by 0.02 units). The radial Ca zoning of the symplectitic garnet_{2a} contrasts with the zoning of coarse garnet, which is both concentric and oriented along the internal fractures and inclusion trails. Symplectitic garnet_{2a} is separated from the coarse garnet₁ by a continuous zone of Ca-rich (Figs. 11 and 12) garnet₁, which is locally replaced along the microfractures and inclusion trails by low-Ca areas of garnet_{2a}. This observation indicates that garnet_{2a}–quartz symplectites post-date the formation of the retrograde Ca-rich zone on garnet₁, which is itself inter-

preted to be retrograde as it is characterised by an increase in Ca and Fe coupled with a decrease in Mg.

The garnet zoning relationships described above are consistent with a four-stage model of garnet formation as showed in Fig. 13. An initial garnet₁ (Fig. 13a) is fractured (Fig. 13b). During and/or after fracturing this garnet₁ is first replaced by higher-Ca garnet along the grain boundaries and micro-fractures (Fig. 13c) and later replaced by garnet_{2a} (Fig. 13d) which seals the micro-fractures. The initial fracturing of M1 garnet is inferred from this textural progression to have occurred in D2, so that garnet_{2a} formed in M2.

Synplectitic garnet_{2b}, which forms coronas around pyroxenes has the composition Alm_{66–73}Grs_{9–15}Pyr_{12–18}Sps₂, with X_{Mg} of 0.14–0.21. This garnet is more Al-rich and Pyr-poor (i.e. lower X_{Mg}) than garnet_{2a} (Fig. 9).

5.3.2. Garnet–orthopyroxene–plagioclase–K-feldspar–quartz gneiss VH207B and VH208

Garnet is Alm_{71–73}Grs_{2–3}Pyr_{20–23}Sps₄ with X_{Mg} of 0.22–0.25 (Fig. 9 and Table 3) and it is more Alm- and

Table 7
Corundum and spinel analyses

Phase	Corundum	Spinel
Thin section	VH552	VH552
SiO ₂	0.01	0.02
Al ₂ O ₃	99.25	57.16
Cr ₂ O ₃	0.01	0.02
FeO	0.69	32.2
MnO	0.01	0.01
MgO	0.02	6.9
CaO	0.02	0
Total	100.04	96.37
Si	0	0.001
Ti	0	0.001
Al	1.99	1.931
Cr	0	0
Fe	0.01	0.77
Mn	0	0.001
Mg	0	0.295
Ca	0	0
Total	2.003	3.000
		Fe ₃₊
		X_{Mg}
		X_{hc}
		X_{gal}
		X_{spl}
		X_{chr}
		X_{mag}
		X_{ulv}

Cations of corundum calculated on the basis of 3 oxygens. Cations of spinel calculated on the basis of 4 oxygens. Fe³⁺ estimated using the site balance calculation $\text{Fe}^{3+} = 2 - (\text{Al} + \text{Cr} + \text{Ti})$, where Ti, Al, Cr Fe^{tot}, Mn and Mg have been normalised to a total of 3. hc: hercynite. gal: galaxite. spl: spinel. chr: chromite. ulv: ulvospinel.

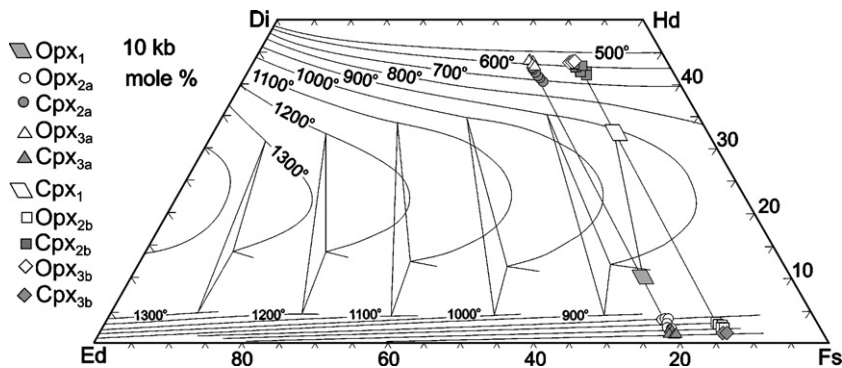


Fig. 7. Pyroxene quadrilateral plot showing the chemical compositions of different generations of pyroxenes in meta-ironstone VH543.

Sps-rich that Garnet₁ in VH546a–c, and has lower Grs. Rimward zoning features a decrease of 0.03 units in X_{Mg} , 1–2 mol% Pyr and 1–2 mol% in Alm.

5.3.3. Corundum–spinel metapelite VH552

Garnet in the corundum-bearing metapelite is almandine-rich (Table 3), with Alm_{70–75}Pyr_{23–27}Grs_{1–2} with X_{Mg} of 0.23–0.29.

5.4. Plagioclase and alkali-feldspars

5.4.1. Garnet–orthopyroxene–plagioclase gneiss VH546A and VH546C

Plagioclase in VH546A and VH546C is Ab_{59–69}An_{28–39}Or_{1–4} (Table 4). Coarse plagioclase_{1a} (Pl_{1a}) is An_{33–39} with a core to rim decrease of An of 6 units, while both plagioclase_{1b} (Pl_{1b}) and plagioclase₂ (Pl₂) are less anorthitic with An_{32–34}. Plagioclase inclusions are An_{28–37}. Rimming alkali-feldspar in VH546A and

VH546C is Ab_{5–10}An_{0–1}Or_{88–93}Cs_{1–2}, with a BaO content in the range of 0.42–1.28 wt.%.

5.4.2. Garnet–orthopyroxene–plagioclase–K-feldspar–quartz gneiss VH207B and VH208

Polygonal Pl₁ is Ab_{80–86}An_{13–19}Or₁Cs₀ with a rimwards decrease in An of up to 6 units (Table 4). The exsolved feldspars are Ab_{84–91}An_{9–12}Or_{1–6}Cs₀ (Pl₂) and Ab_{11–17}An_{0–1}Or_{82–89}Cs_{0–1} (K-f₂) in VH207B, and Ab_{87–91}An_{6–12}Or_{1–5}Cs₀ (Pl₂) and Ab_{10–17}An₀Or_{81–88}Cs_{0–2} (K-f₂) in VH208 (Fig. 14, Tables 5 and 6). The original pre-exsolution feldspar compositions have been restored (reintegrated) by rastering a continuous set of adjacent 25 μ^2 areas along a line traverse perpendicular to the exsolved lamellae and integrating the measured compositions. A minimum of twenty lamellae were intersected along each traverse and a minimum of 30 analysis performed. Six grains were analysed for each specimen and the integrated analyses averaged, yielding typical

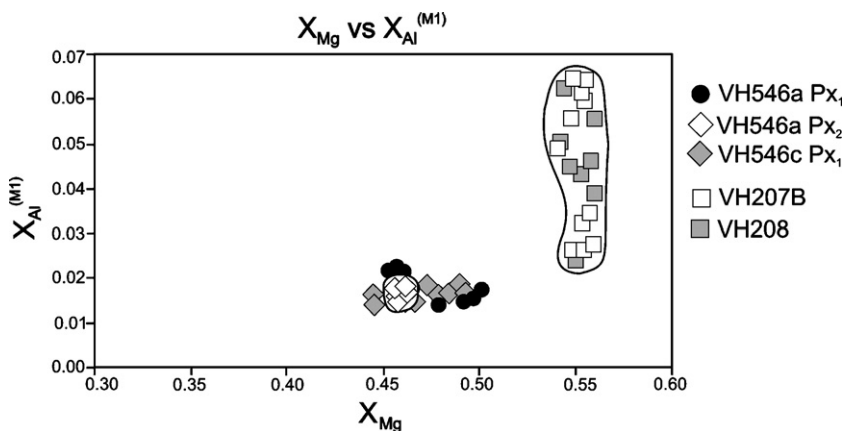


Fig. 8. Orthopyroxene chemical compositions, expressed in terms of X_{Al} and X_{Mg} . Cores of orthopyroxene₁ range up to high X_{Al} in VH207B and VH208 but show little variation in X_{Mg} , whereas in VH546a and VH546c variation only occurs in X_{Mg} .

uncertainties in each feldspar component (Ab, An, Or + Cs) of less than ± 2 mol%. In VH207B the reintegrated K-feldspar (K-f₁) is Ab₄₈An₃Or + Cs₄₉. In VH208 the reintegrated K-feldspar (K-f₁) is more albite-rich than in VH207B with a composition of Ab₇₀An₄Or + Cs₂₆.

5.5. Biotite

Biotite in the garnet–orthopyroxene–plagioclase–gneiss (VH546A and VH546C) spans the Mg-biotite and Fe-biotite field of Foster (1960) and Guidotti et al. (1975) (Table 5), with X_{Mg} in the range 0.43–0.63, Ti contents of 0.34–0.72 p.f.u. and Al^{VI} contents of 0.02–0.28. These biotites generally show little chemical va-

riation irrespective of their different microstructural settings, with the only distinct features being the higher Ti and X_{Mg} preserved in biotite inclusions compared with biotites in the matrix (Bt₁) and in the shear zones (Bt₂).

Biotite in the garnet–orthopyroxene–plagioclase–K-feldspar–quartz gneisses (VH207B and VH208) is Mg-biotite (Table 5) and has a X_{Mg} in the range 0.62–0.65, higher than biotites in VH546A and VH546C. Al^{VI} and Ti contents are respectively 0.01–0.12 and 0.49–0.54. Biotite in the corundum–spinel pelite VH552 lies between the Mg-biotite and Fe-biotite fields of Foster (1960) and Guidotti et al. (1975) (Table 5), with X_{Mg} 0.42–0.63, Al^{VI} 0–0.78 and Ti 0.14–0.66 c.p.f.u.

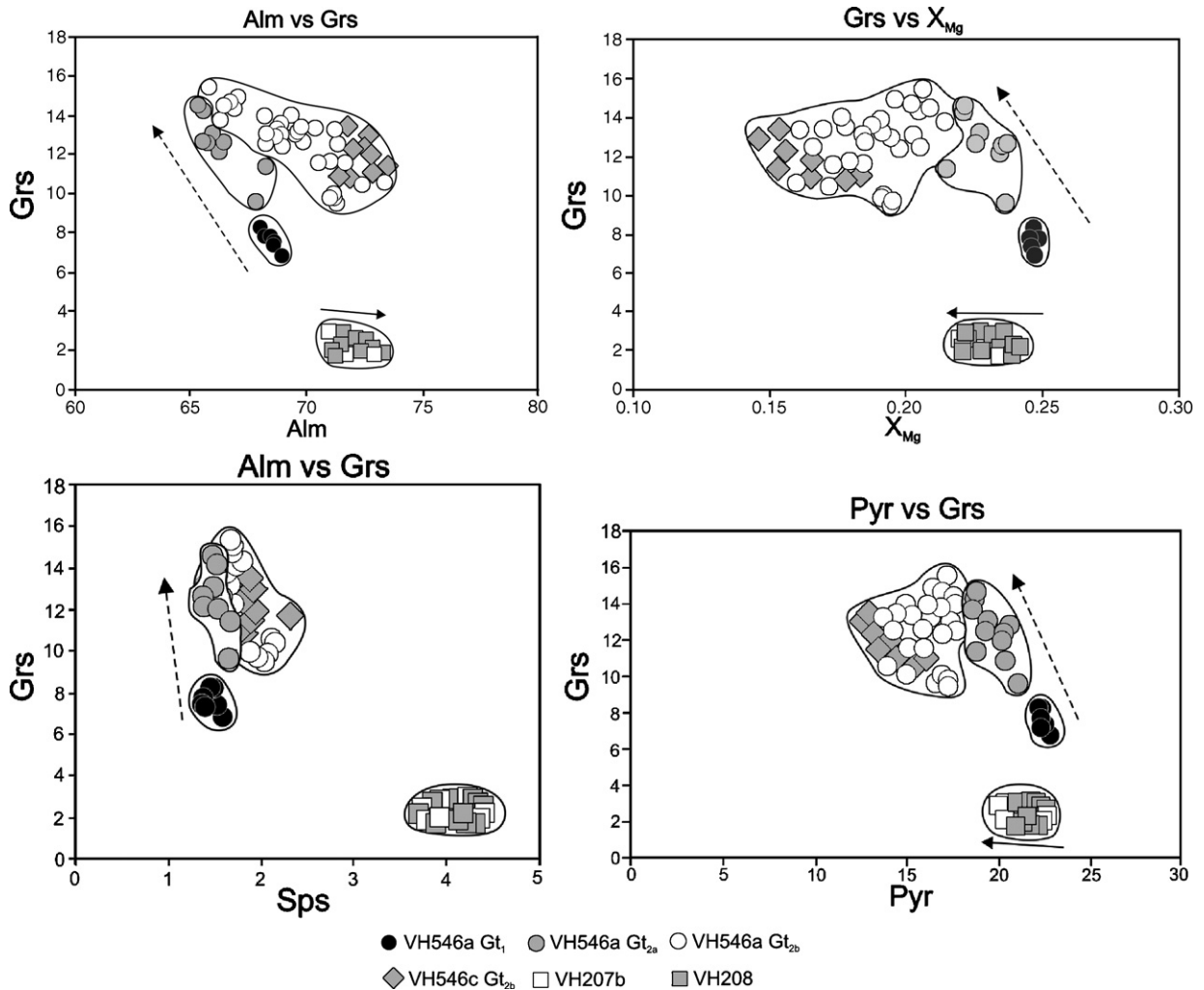


Fig. 9. Chemical compositions of garnets, expressed in terms of the molecular proportions of the main components (Alm, Grs, Sps, Pyr) and X_{Mg} . Only the compositions of the lower-Gr_s areas of garnet₁ are shown. Broken arrows indicate the compositional trend between garnet_{1a} and garnet_{2a} in VH546a. Note the decreases in X_{Mg} , of pyrope and almandine and increase in grossular from Gt₁ to Gt_{2a}, and the higher almandine, lower pyrope and lower X_{Mg} content of garnet_{2b} with respect to garnet_{2a}. Continuous arrows depict core to rim compositional changes in polygonal garnets in VH207B and VH208; these exhibit negligible zoning in grossular.

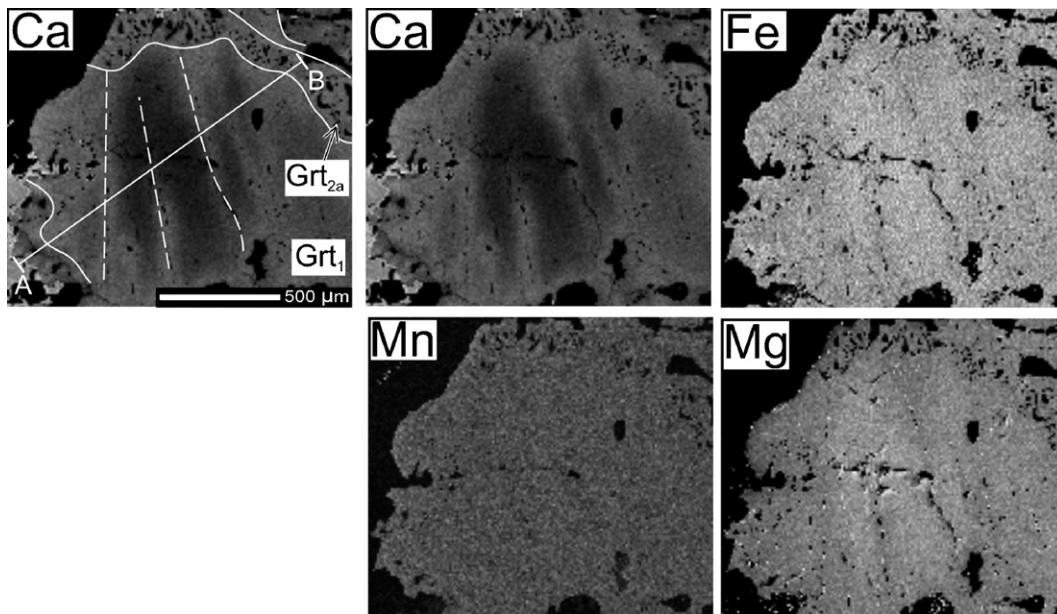


Fig. 10. X-ray element map of coarse garnet₁ rimmed by garnet_{2a}. Darker areas signify lower concentrations of the mapped element. Broken lines are micro-fractures while the continuous line separates garnet₁ from the garnet_{2a}+quartz symplectites.

5.6. Corundum

Corundum in VH552 has a FeO content of 0.5–0.9 wt.% (Table 7).

5.7. Spinel

Spinel in VH552 is Hc_{67–71}Spl_{27–30}Mag_{2–3} (Table 7), manganese and chromium contents are negligible and X_{Mg} is in the range of 0.25–0.28.

5.8. Ilmenite

Ilmenite has a X_{Mg} of 0.04–0.07, a MgO content of 0.06 wt.% and MnO of 0.01 wt.%. Chemical profiles across ilmenite grains rimmed by garnet₂–quartz symplectites in VH546C show no Fe, Mg or Mn zoning.

6. Thermobarometry

6.1. Temperatures of D1–M1

Conventional thermometers that rely on Fe–Mg exchange equilibria have been demonstrated in a number of studies to seriously underestimate peak metamorphic temperatures in granulites (Harley, 1989, 1998; Frost and Chacko, 1989; Pattison and Begin, 1994; Fitzsimons and Harley, 1994). Hence, in this study peak temperatures for the D1–M1 metamorphism in the Vestfold Hills have been estimated instead using the reintegrated compositions of exsolved pyroxenes in VH543, and exsolved alkali feldspars in VH207B and VH208.

Application of the graphical two-pyroxene thermometer (Lindsley and Andersen, 1983) to the reintegrated compositions of pre-exsolution pigeonite and Fe-augite in the meta-ironstone VH543 (Fig. 7) yields a temperature estimate of 895 ± 35 °C. As the preserved exsolution lamellae in these pyroxenes, though weakly deformed, are not overprinted by the S1 axial planar

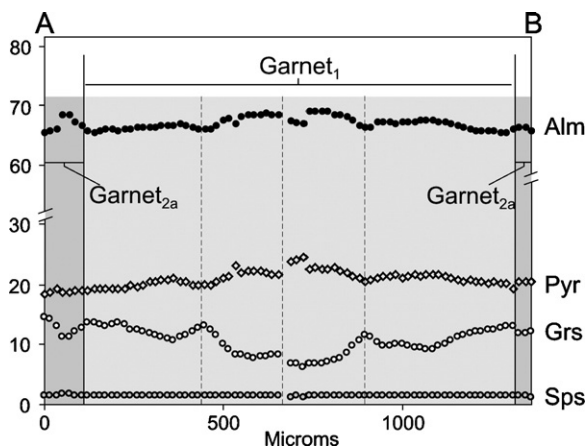


Fig. 11. Compositional profile across the garnet illustrated in Fig. 10. Broken lines are fractures while continuous lines are the boundaries between garnet₁ and garnet_{2a}.

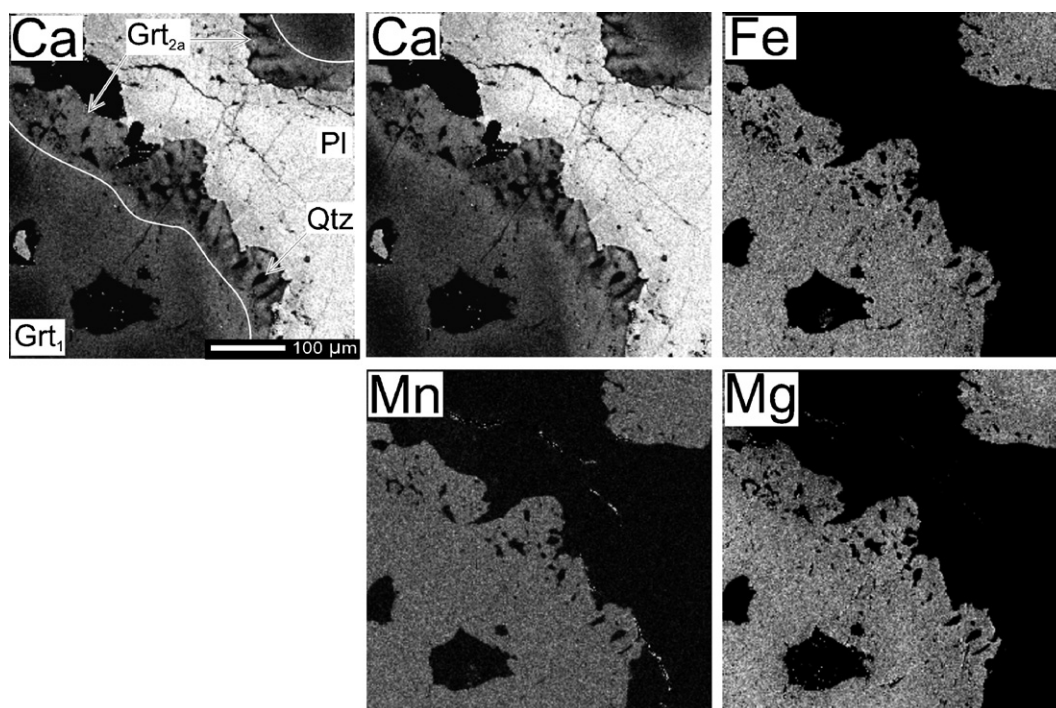


Fig. 12. Higher magnification of a contact between garnet₁ and garnet_{2b}. Dark areas indicate lower concentrations. Note the radial Ca-zoning of the garnet in the garnet_{2a}+quartz symplectites and the lighter (higher-Ca) rim separating them from the internal part of garnet₁.

fabric it is considered that exsolution of these lamellae post-dated S1, and hence that the reintegrated pyroxenes were stable under peak D1–M1 conditions. Late- to post-D1 recrystallisation of pyroxene produced the observed neoblast clusters and trails, which affect both host pyroxene and lamellae on their terminations at grain edges. These neoblasts may have formed at temperatures in the range 600–700 °C according to the isotherms of [Lindsley and Andersen \(1983\)](#), although these estimates must be regarded as minima given the propensity for the pyroxenes to re-equilibrate down-temperature.

Three alternative descriptions of the temperature-composition relations in ternary feldspars have been

applied to calculate minimum temperatures of D1–M1 based on reintegrated grain compositions in VH207B and VH208. The errors quoted below for these alternative temperature estimates include the calibration uncertainties and the uncertainties arising from the typical spread in Ab:Or:An generated from area analysis and reintegration of exsolved lamellae. The thermometer of [Elkins and Grove \(1990\)](#) yields the highest temperature estimates, at 835 ± 60 °C for VH208 and 900 ± 30 °C for VH207B (Fig. 14). The [Fuhrman and Lindsley \(1988\)](#) thermometer yields temperatures of 855 ± 50 °C for VH208 and 870 ± 40 °C for VH207B. Lastly, the [Lindsley and Nekvasil \(1989\)](#) thermometer yields estimates of 815 ± 50 °C (VH208) and 850 ± 40 °C (VH207B). In all

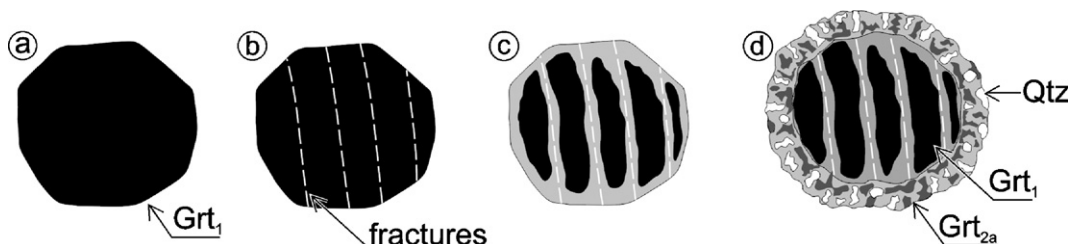


Fig. 13. Four-step model for the formation of the garnet-zoning pattern. An initial garnet₁ with a low-Ca content (a) is fractured (b) then replaced by a Ca-rich garnet both along its boundaries and micro-fractures (c). A later radial zoned garnet_{2a} with symplectitic quartz replaces the coarse garnet₁ along grain boundaries and micro-fractures sealing them.

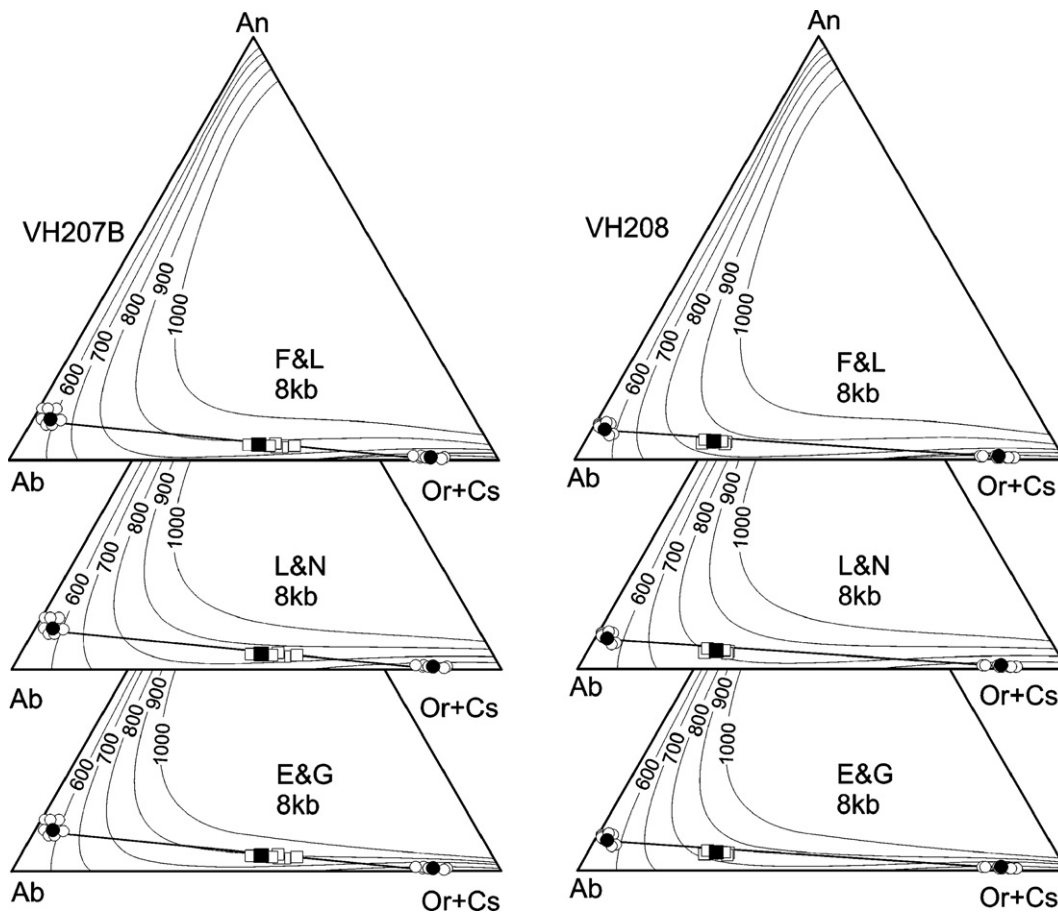


Fig. 14. Or–Ab–An ternary plot of the compositions of feldspars. Unfilled circles are the analysed compositions of the exsolution lamellae, averaged to yield the filled circles. Unfilled squares are the reintegrated compositions of the initial feldspars, reconstructed by rastering a line traverse perpendicular to the feldspar lamellae. Filled squares are the averaged compositions of the initial feldspars based on integrating several (5 to 10) such rastered traverses. Isotherms are calculated using the solid solution models of: F and L) Fuhrman and Lindsley (1988); L and N): Lindsley and Nekvasil (1989); E and G) Elkins and Grove (1990).

cases the Or-rich ternary feldspar in VH207B provides the highest minimum-temperature conditions for D1 and hence suggests peak M1 metamorphism at $T > 850 \pm 40$ °C based on the most conservative calibration, that of Lindsley and Nekvasil (1989). This minimum temperature is consistent with those calculated using reintegrated pyroxenes in the meta-ironstone (VH543) and is, moreover, also consistent with temperatures estimated by Harley (1993) from the *Taynaya Paragneiss* in the Long Peninsula region (830–880 °C).

6.2. Pressure conditions of D1–M1 and application of convergence–retrieval methods

Calculation of the peak conditions of the main metamorphic event (D1–M1) was also attempted using the composition of the least retrogressed parts of garnet₁ to-

gether with the most Al-rich pyroxene₁ core in the garnet–orthopyroxene–plagioclase gneisses VH5208 and VH207B (Fig. 15), based on the logic that this pyroxene may record a composition closest to that present at the metamorphic peak condition (e.g. Harley, 1989). In this case the temperatures were estimated using garnet–orthopyroxene Fe–Mg exchange thermometry (Harley, 1984a; Lee and Ganguly, 1988; Ganguly et al., 1996) and orthopyroxene Al-thermometry (Harley and Green, 1982; Aranovich and Berman, 1997). Pressures were estimated using formulations of the Fe-system garnet–orthopyroxene–plagioclase–quartz barometer (GAFS: Bohlen et al., 1983; Essene, 1989; Moecher et al., 1988), the Mg-system garnet–orthopyroxene–plagioclase–quartz barometer (GAES: Perkins and Newton, 1981; Newton and Perkins, 1982; Essene, 1989; Eckert et al., 1991), and the Al- in orthopyroxene formulation of Harley (1984b). The results of these calculations are summarised in Table 7 and Fig. 15.

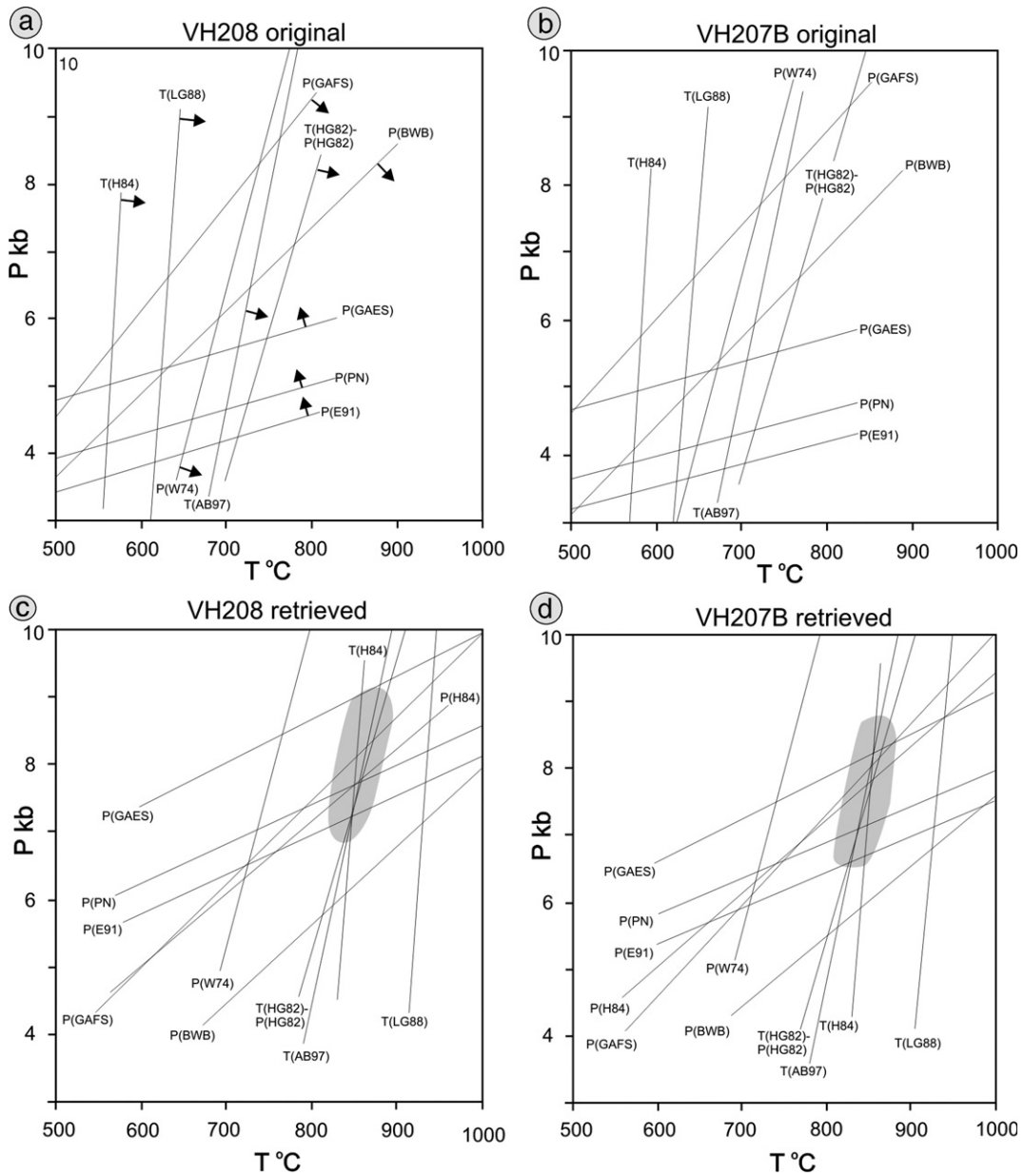


Fig. 15. Temperature and pressure estimates from samples VH207B and VH208. a–b) Estimates calculated using the “original” chemical composition of the less retrogressed cores of the garnet and orthopyroxene. These compositions are uncorrected for the effects of any retrograde Fe–Mg exchange. Arrows indicate the directions of movement of the thermobarometers that results from correction of mineral compositions for retrograde Fe–Mg exchange. c–d) Estimates calculated using the “retrieved” chemical compositions of the less retrogressed cores of the garnet and orthopyroxene. These compositions are corrected for the retrograde Fe–Mg exchange. Shaded area is the preferred P – T range, as discussed in the text. Thermometers: $T(H84)$: [Harley \(1984a\)](#); $T(LG88)$: [Lee and Ganguly \(1988\)](#); $T(G96)$: [Ganguly et al. \(1996\)](#); $T(HG82)$: barometer of [Harley and Green \(1982\)](#) recast as a thermometer for the given nominal pressure; $T(AB97)$: [Aranovich and Berman \(1997\)](#). Garnet–orthopyroxene–plagioclase–quartz barometers: $P(PN)$: [Perkins and Newton \(1981\)](#); $P(GAES)$: [Moecher et al. \(1988\)](#), and [Essene \(1989\)](#); $P(BWB)$: [Bohlen et al. \(1983\)](#); $P(GAFS)$: [Moecher et al. \(1988\)](#) and [Essene \(1989\)](#); $P(E91)$: [Eckert et al. \(1991\)](#). Grt–Opx–Al barometers: $P(W71)$: [Wood \(1974\)](#); $P(HG82)$: [Harley and Green \(1982\)](#); $P(H84)$: [Harley \(1984b\)](#). Typical uncertainties on each thermometer calibration/method are ± 50 °C. Typical uncertainties on the garnet–orthopyroxene–plagioclase–quartz barometers are ± 1 kb, whilst those on the Grt–Opx–Al barometers are greater than ± 1 kb.

Pressures estimated using the core compositions, uncorrected for the later effects of any Fe–Mg exchange, range widely between 5 and 15 kb at nominal temperatures of 850 °C, depending on the calibrations used (Figs. 15 and 16; Table 8). Such extreme scatter in part reflects the high iron content of the samples, which is outside the calibration ranges of most of the thermobarometers, but also is indicative of intense retrograde Fe–Mg cation exchange (e.g. Frost and Chacko, 1989; Fitzsimons and Harley, 1994). Notable features of the P – T data (Figs. 15 and 16; Table 8) are that the Grt–Opx Fe–Mg temperatures are very low compared with other methods (<630 °C c.f. 700–750 °C) and that the pressures calculated using the Harley (1984b) barometer are also anomalously low. These features are symptomatic of the effects of late Fe–Mg exchange, as recognised and described by Fitzsimons and Harley (1994) and Pattison and Begin (1994), and can be corrected for using their retrieval methods, providing that an independent and robust estimate of temperature (or pressure) is available.

In order to improve the pressure estimates, the temperatures retrieved independently from pyroxene and feldspar thermometry have been assumed to represent the peak or near-peak conditions under which garnet–orthopyroxene Al equilibrium was attained but following which both phases re-equilibrated in their Fe–Mg compositions. The X_{Mg} values of each phase have then been recalculated so that the garnet–orthopyroxene Fe–Mg thermometers (Harley, 1984a; Lee and Ganguly, 1988; Ganguly et al., 1996) retrieve temperatures within the range indicated by the pyroxene and feldspar thermometers (830–910 °C). These adjusted garnet and orthopyroxene X_{Mg} values have then been combined with the original Al content of orthopyroxene and further revised iteratively so that the Harley (1984a,b) Fe–Mg thermometer, Harley and Green (1982) Mg–Al thermobarometer and Aranovich and Berman (1997) Fe–Al thermometer yield convergent temperature estimates. The resultant garnet and orthopyroxene X_{Mg} values, along with the original Al content of orthopyroxene, have then been inputted into the garnet–orthopyroxene–plagioclase–quartz barometer formulations and revised pressures estimated. These ‘retrieved’ pressures at the independently-constrained temperatures are depicted for VH208 and VH207B on Fig. 15c and d respectively. This methodology is a variant of that employed by Fitzsimons and Harley (1994), incorporating the additional step of recalculating the pressures using garnet–orthopyroxene–plagioclase–quartz barometer formulations and then comparing the extent to which these converge once the assumed effects of post-peak Fe–Mg exchange are corrected.

Arrows in Fig. 15a indicate the P – T directions in which the thermobarometers are shifted after mineral compositions are corrected for the effect of retrograde exchange. As expected from previous studies (e.g. Fitzsimons and Harley, 1994) the Al-thermometers of Harley and Green (1982) and Aranovich and Berman (1997) both translate up-temperature to be consistent with the Fe–Mg thermometry, and with these adjustments the agreement between different versions of the garnet–orthopyroxene–plagioclase–quartz barometers (and the method of Harley (1984b)) is improved. Translation of the GAFS barometer (Moecher et al., 1988; Essene, 1989) to lower pressures, coupled with translation of the GAES barometer (Moecher et al., 1988; Essene, 1989) to higher pressures results in their overlap for samples VH208 and VH207B, within typical uncertainties of 1 kb, at the temperatures defined for convergence of the thermometers. Given the Fe-rich character of the garnet–orthopyroxene assemblages and the greater degree of sensitivity of the GAES barometer to the mineral composition adjustments, our preferred pressure estimate corresponding to the peak D1–M1 temperatures is that based on the GAFS barometer (Moecher et al., 1988; Essene, 1989). At temperatures of 850 ± 40 °C the retrieved pressure estimates based on this strategy (with 1σ errors) are 8.0 ± 1.2 kb for VH208 and 7.8 ± 1.2 kb for VH207B (Fig. 15c and d).

The retrieval methods based on Mg–Fe–Al in orthopyroxene coexisting with garnet do not result in convergent P – T estimates for VH546A and VH546C (Fig. 16a and b), samples in which the Al– in orthopyroxene thermometers yield lower temperatures than the Fe–Mg exchange methods. However, in the Fe- and Ca-rich sample VH546A the GAFS and GAES barometers are robust to Fe–Mg adjustments, and the converged pressure estimate of 8.4 ± 1.2 kb (850 ± 40 °C) based on these geobarometers is considered a reasonable pressure estimate for D1–M1, consistent with VH208 and VH207B. Considering all three samples as sharing the same D1–M1 P – T conditions leads to a statistically averaged estimate of 8.1 ± 0.9 kb for M1 metamorphism in the Vestfold Hills, at 850 ± 40 °C. This pressure estimate lies at the high end of the range defined by Harley (1993) from the *Taynaya Paragneiss* (<8.5 kb), but is consistent with those earlier estimates.

The occurrence of sillimanite as the stable aluminosilicate polymorph in the Chelnok and Taynaya paragneisses provides only a weak constraint on pressures, at $<11 \pm 0.8$ kb for 850 ± 40 °C. However, calculations on the assemblage spinel + sillimanite + garnet + corundum in the Fe_2O_3 – Al_2O_3 – SiO_2 (FAS) system using the THERMOCALC internally consistent thermodynamic data set of Holland and Powell (1998) allows a tighter pressure

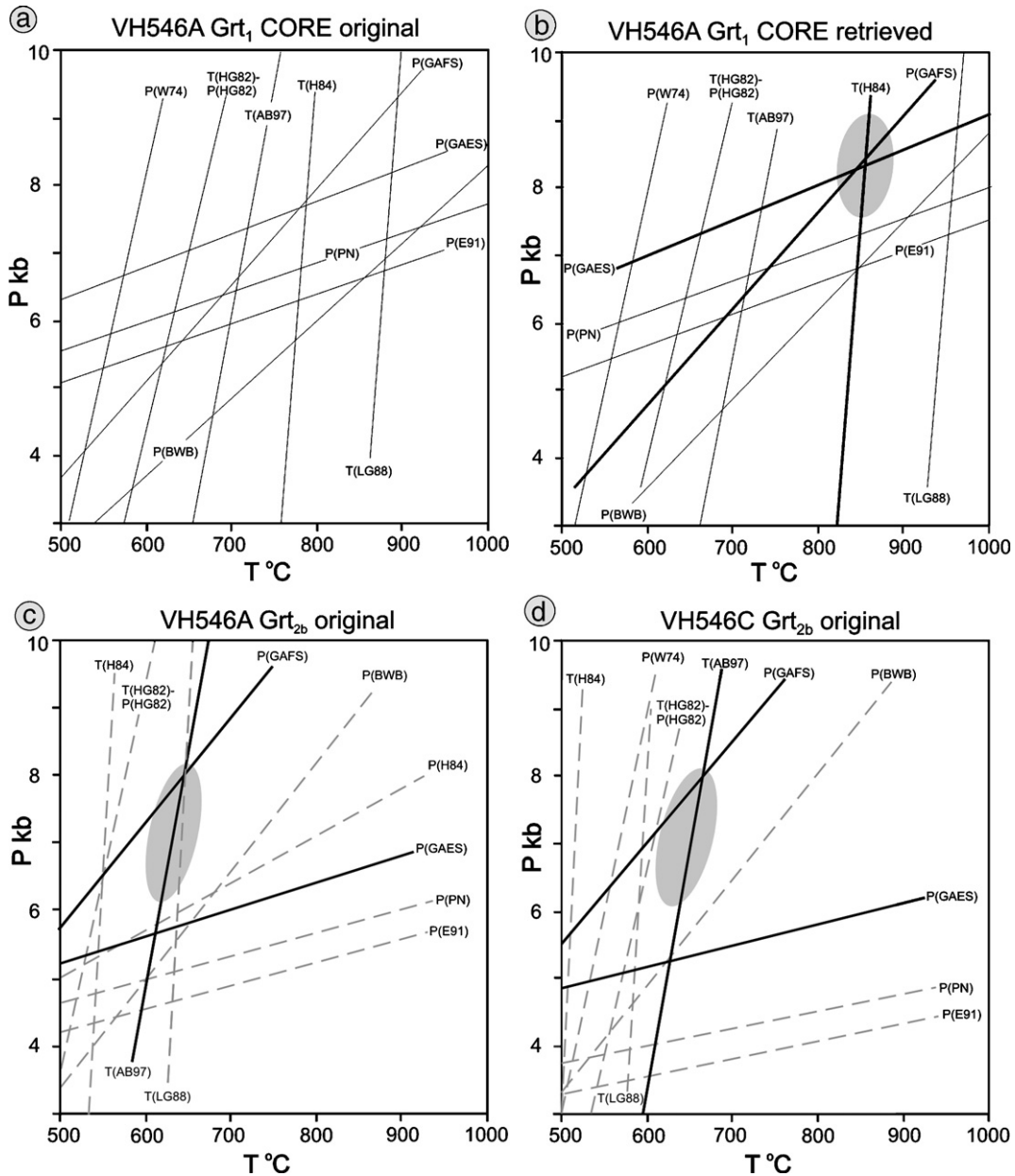


Fig. 16. Temperature and pressure estimates from samples VH546A and VH546C. a–b) Estimates using the compositions of less retrogressed cores of garnet₁ and orthopyroxene₁. c–d) Estimates for the formation of the secondary garnet_{2b}–quartz symplectites between orthopyroxene and plagioclase. “original” are estimates calculated using the mineral chemical compositions uncorrected for retrograde Fe–Mg exchange; “retrieved” are estimates calculated using the chemical compositions corrected for retrograde Fe–Mg exchange as described in the text. Thermometers and barometers are as listed in the caption to Fig. 15. Typical uncertainties on each thermometer calibration/method are ± 50 °C. Typical uncertainties on the garnet–orthopyroxene–plagioclase–quartz barometers are ± 1 kb, whilst those on the Grt–Opx–Al barometers are greater than ± 1 kb.

constraint to be applied for the aluminous metapelite in the *Chelnok Paragneiss*. In this case the reaction:



has been modelled for the mineral compositions in VH552 using activity–composition relations of [Holland and](#)

[Powell \(1998\)](#). Based on these data, the peak garnet + spinel + corundum assemblage is calculated to have formed at 8.6 ± 1.4 kb for temperatures of 850 ± 40 °C ([Table 8](#)), consistent with the results summarised above for the garnet–orthopyroxene gneisses.

In summary, geothermobarometry, THERMOCALC calculations and retrieval–convergence methods applied

to the *Chelnok Paragneiss* indicate peak D1–M1 metamorphism at P – T conditions of 8.1 ± 0.9 kb and 850 ± 40 °C.

6.3. Post-peak garnet formation and retrograde P – T conditions

Metamorphic conditions associated with post-peak garnet growth in VH546A and VH546C have been estimated using the composition of symplectitic garnet_{2b} together with the composition of the adjacent orthopyroxene rims (Table 8). This procedure yields a large spread in calculated P – T conditions, dependent both on method (e.g. Fe–Mg exchange thermometry) and calibration (e.g. Harley (1984a) versus Lee and Ganguly (1988)). Extensive Fe–Mg re-equilibration is indicated by the low garnet–orthopyroxene Fe–Mg exchange temperatures, which lie in the range 510–640 °C. Al³⁺ in orthopyroxene thermometers (Harley and Green, 1982; Aranovich and Berman, 1997) also yield low temperature estimates, 540–650 °C, which are likely to have been influenced by late Fe–Mg exchange. The Al³⁺ in orthopyroxene temperature estimates also are subject to large calibrational uncertainties in this case because of the low Al contents ($X_{\text{Al}} = 0.019$ – 0.014) of the analysed orthopyroxenes. Given these caveats, the estimate of 600–650 °C obtained from the Fe-system Aranovich and Berman (1997) thermometer is used here as the preferred indicator of the temperatures of formation of Fe-rich garnet_{2b} as it involves less extrapolation in compositional space, and hence is subject to less error arising from uncertainties in the activity–composition relations, than the thermometers based on activity–composition relations grounded in Mg-system experiments (e.g. Harley and Green, 1982).

Pressures of garnet corona and symplectite formation were estimated using garnet–orthopyroxene–plagioclase–quartz barometers calibrated in the Fe-system (GAFS: Bohlen et al., 1983; Moecher et al., 1988; Essene, 1989) and the Mg-system (GAES: Perkins and Newton, 1981; Essene, 1989; Eckert et al., 1991). The results of these calculations are summarised in Table 8 and depicted in Fig. 16c–d. As before, there is a large spread in pressures calculated at 600–650 °C, from 8 to 3.4 kb. This spread is calibration and formulation dependent. The Fe-system barometers (P_{GAFS} and P_{BWB}) yield higher pressures (5–8 kb) than the Mg-system barometers P_{PN} (Perkins and Newton, 1981) and P_{E91} (Eckert et al., 1991) (3.5–5.5 kb). The GAES barometer of Moecher et al. (1988) lies between the extremes and yields similar pressures to P_{BWB} (Bohlen et al., 1983). Correction of the mineral compositions for Fe–Mg exchange generally has the effect of increasing P_{PN} , P_{GAES} and P_{E91} and of

Table 8

a) Temperature estimates. b) Pressure estimates

Sample		Composition					Fe–Mg exchange				
		Garnet–orthopyroxene–plagioclase–quartz gneiss					P-reference				
		$X_{\text{Mg-grt}}$	X_{grs}	$X_{\text{Mg-px}}$	X_{Al}		$T(\text{H84})^a$	$T(\text{LG88})^a$	$T(\text{G96})^a$	$T(\text{HG82})^a$	$T(\text{AB97})^a$
VH207B core	Original	0.246	0.022	0.548	0.064	7.8	591	649	650	794	747
VH207B core	Retrieved	0.256	0.022	0.429	0.064	7.8	851	931	930	854	848
VH208 core	Original	0.236	0.03	0.545	0.063	7.7	578	638	638	792	747
VH208 core	Retrieved	0.246	0.03	0.417	0.063	7.7	851	935	934	857	854
VH546A Grt ₁ core	Original	0.248	0.079	0.454	0.023*	8	789	886	885	668	727
VH546A Grt ₁ core	Retrieved	0.258	0.079	0.441	0.023*	8	851	955	953	674	736
VH546C Grt _{2b}	Original	0.169	0.136	0.48	0.014*	8	556	646	646	576	665
VH546A Grt _{2b}	Original	0.153	0.115	0.473	0.019*	8	520	601	601	625	644

B

Pressure

Garnet–orthopyroxene–plagioclase–quartz gneiss

Sample		Composition					<i>T</i> -reference	Grt–Opx–Plag–Qz					Grt–Opx–Al		
		$X_{\text{Mg-grt}}$	X_{grs}	$X_{\text{Mg-opx}}$	X_{Al}	X_{An}		<i>P</i> (PN) ^b	<i>P</i> (GAES) ^b	<i>P</i> (BWB) ^b	<i>P</i> (GAFs) ^b	<i>P</i> (E91) ^b	<i>P</i> (W74) ^c	<i>P</i> (HG82) ^c	<i>P</i> (H84) ^c
VH207B core	Original	0.246	0.022	0.548	0.064	0.17	850	4.8	5.9	7.7	9.5	4.4	13.9	10.2	−0.2
VH207B core	Retrieved	0.256	0.022	0.429	0.064	0.17	850	7.2	8.2	6	8	6.7	12.8	7.6	7.8
VH208 core	Original	0.236	0.03	0.545	0.063	0.19	850	5.2	6.1	8	10.1	4.8	13.7	10.2	−1
VH208 core	Retrieved	0.246	0.03	0.417	0.063	0.19	850	7.7	9	6.2	8.1	7.2	12.6	7.4	7.7
VH546A Grt ₁ core	Original	0.248	0.079	0.454	0.023*	0.39	850	7.1	8	6.6	8.7	6.6	22.7	17.7	16.8
VH546A Grt ₁ core	Retrieved	0.258	0.079	0.441	0.023*	0.39	850	7.3	8.4	6.9	8.3	6.8	22.6	17.4	17.7
VH546C Grt _{2b}	Original	0.169	0.136	0.48	0.014*	0.34	650	5.2	5.8	5.7	8	4.7	14.3	12.2	6.1
VH546A Grt _{2b}	Original	0.153	0.115	0.473	0.019*	0.34	670	4.2	5.2	5.9	8	3.7	13.1	10.5	1.4

Corundum–spinel metapelite

Sample	Composition								<i>T</i> -reference	<i>P</i> (TC)
	Garnet				Spinel					
VH552	<i>X</i> _{Mg}	<i>X</i> _{grs}	<i>X</i> _{alm}	<i>X</i> _{prp}	<i>X</i> _{Mg}	<i>X</i> _{hrc}	<i>X</i> _{spl}	<i>X</i> _{mag}		
	0.24	0.02	0.75	0.23	0.26	0.71	0.27	0.02	850	8.6

“Original” is based the mineral compositions uncorrected for Fe–Mg exchange.

“Retrieved” is the mineral composition corrected for the effects of any Fe–Mg exchange.

“*P*-reference” is the nominal pressure at which thermometers are calculated.

“*T*-reference” is the nominal temperature at which the barometers are calculated.

P(TC): THERMOCALC (Holland and Powell, 1998).

$X_{\text{Al}} = [\text{Al} - 2\text{Ti} - \text{Cr} - \text{Fe}^{3+}] / 2$. * X_{Al} calculated considering all iron as ferrous * $X_{\text{Al}} = [\text{Al} - 2\text{Ti} - \text{Cr}] / 2$.

^a Thermometers: *T*(H84): Harley (1984a); *T*(LG88): Lee and Ganguly (1988); *T*(G96): Ganguly et al. (1996); *T*(HG82): barometer of Harley and Green (1982) recast as a thermometer for the given nominal pressure; *T*(AB97): Aranovich and Berman (1997).

^b Grt–Opx–Pl–Qtz barometers: *P*(PN): Perkins and Newton (1981); *P*(GAES): Moecher et al. (1988), and Essene (1989) Mg-system barometer; *P*(BWB) Bohlen et al. (1983); *P*(GAFS): Moecher et al. (1988) and Essene (1989) Fe-system barometer; *P*(E91): Eckert et al. (1991).

^c Grt–Opx–Al barometer: *P*(W71): Wood (1974); *P*(HG82): Harley and Green (1982); *P*(H84): Harley (1984b).

decreasing P_{GAFS} and P_{BWB} (Fitzsimons and Harley, 1994), leading to convergence at retrieved pressures P greater than initial P_{GAES} but close to initial P_{GAFS} (e.g. VH207B, VH208). On this basis, allowing for only a minor amount of Fe–Mg readjustment, and using the Fe-system GAFS calibration as previously applied to the peak assemblages on the basis that it involves less extrapolation in activity–composition space, pressures of garnet corona formation are placed between 7 ± 1 kb at 600 – 650 °C. Post-peak and M2 garnet corona formation is inferred to have occurred at pressures, and crustal depths, similar to those under which peak D1–M1 metamorphism occurred.

7. Discussion

7.1. Metamorphic history and P – T path of the Vestfold Hills

The results of our thermobarometry on the Chelnok Paragneiss are combined in Fig. 17 to produce a summary P – T path for the Vestfold Block over the time interval 2501 – 2496 Ma (D1–M1) to 2475 Ma (post D2–M2). The P – T conditions of D2–M2 are not tightly constrained by the data obtained here, as the only S2 fabrics recognised are the 600 – 650 °C pyroxene recrystallisation seams in VH546A that are then overprinted by the garnet coronas formed at 6 – 8 kb at 600 – 650 °C. Thermobarometric

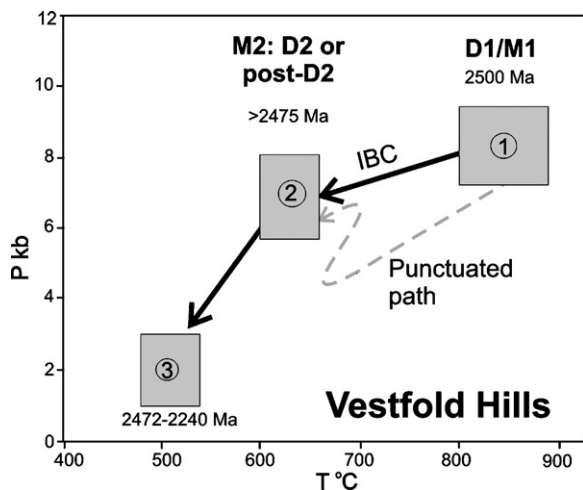


Fig. 17. P – T path for the Vestfold Hills. 1) P – T box for peak M1 conditions. 2) P – T box for post-M1 to M2 conditions corresponding to the formation of retrograde garnet₂–quartz symplectites in VH546A and VH546C. 3) P – T conditions at or after 2475 Ma. Continuous black line is a hypothetical P – T path in which the transition from D1–M1 conditions to post-M1 and M2 is accompanied only by cooling. Broken grey line is an alternative P – T path which allows for proposed thickening associated with magmatic underplating and intrusion of the post-M1 but largely pre-D2 Crooked Lake Gneiss Group.

analysis of Crooked Lake Gneisses Group samples is required in order to improve these estimates of the P – T conditions of D2–M2.

The post-peak P – T path traverses from 8.1 ± 0.9 kb and 850 ± 40 °C at ca. 2500 Ma to 7.0 ± 1.0 kb and 650 – 600 °C by 2477 Ma. Hence, the end-Archaeon P – T path for the Vestfold Hills is potentially characterised by an initial phase of slow near-isobaric cooling (IBC), through 300 – 200 °C at pressures of between 8 and 6 kb. The pressure and temperature uncertainties on this path lead to a dP/dT gradient of the simplest possible path of 1 – 7 bars/°C, with a dT/dt of 10 ± 3 °C/Ma. Such a simple IBC path also would imply that the Crooked Lake Gneiss Group magmatic protoliths were emplaced into Mossel Gneiss and Chelnok Paragneiss basement at relatively deep-crustal levels (6 – 8 kb).

This simple IBC post-D1 P – T path is unlikely to represent the actual end-Archaeon metamorphic history of the Vestfold Block. This P – T history is more likely to have been punctuated, with magmatism and then D2–M2 causing a P – T excursion prior to final cooling. Firstly, intrusion of the voluminous gabbroic to granitic Crooked lake Gneiss Group protoliths between 2500 and 2475 Ma is likely to have caused heating and some magmatic thickening, depending on the depths of intrusion. Secondly, D2–M2, which is inferred to have progressed at temperatures similar to or greater than those obtained from the coronas, involved regional-scale refolding and interleaving of pre-D2 lithologies and the Crooked Lake Gneisses Group (Oliver et al., 1982; Snape and Harley, 1996; Snape et al., 1997), almost undoubtedly leading to thickening, and hence a pressure as well as temperature excursion in the post-D1 to M2 P – T path. These effects are indicated by the alternative P – T path shown on Fig. 18 as a dashed grey line. In this case the emplacement depths of the Crooked Lake Gneiss Group precursors may be significantly less than those implied by the simple IBC path model. Hence, the alternative P – T paths proposed here and constrained at their high- and lower- T ends by this study could in principle be tested using barometry on magmatic mineral assemblages where these are still preserved in the Crooked Lake Gneisses Group rocks.

Evidence of brittle structures and localised contact alteration on 2475 Ma cross-cutting late- and post-Crooked Lake Gneisses Group dykes (Black et al., 1991; Snape et al., 1997) together with crystallisation depth estimates from the earliest suites of mafic dykes (Seitz, 1994) indicate that the Vestfold Hills gneisses were at shallower crustal pressures (3 – 5 kb) at or soon after 2475 Ma. Thus, it appears that rapid exhumation of the Vestfold Hills basement gneisses occurred following

D2–M2, consistent with the proposals of Seitz (1994) and Harley and Christy (1995).

7.2. The place of the Vestfold Hills in former Gondwana

The end-Archaeon to earliest Palaeoproterozoic P – T , deformational and magmatic history of the Vestfold Hills now documented through this and previous studies (Black et al., 1991; Harley, 1993; Snape et al., 1997, 2001) is distinct from that preserved in the adjacent Rauer Islands (Harley and Fitzsimons, 1995; Harley et al., 1995). The principal Archaeon magmatic and deformation events in the Rauer Islands are at least 2840 Ma in age (Kinny et al., 1993; Harley et al., 1998), significantly older than those recorded in the Vestfold Hills. The Vestfold history is also distinct from other Archaeon terranes in East Antarctica (Fitzsimons, 2000b; Harley, 2003; Fig. 18). In particular,

past correlations with granulites of the Napier Complex (e.g. Collerson and Sheraton, 1986a) are not tenable. Although the Napier Complex preserves a strong late-Archaeon to early Palaeoproterozoic isotopic record in its zircon and monazite U–Pb age data (e.g. Asami et al., 1998; Grew et al., 2001; Carson et al., 2002; Kelly and Harley, 2005; Hokada et al., 2005), in detail the age spectra differ, the UHT event in the Napier Complex precedes 2510 Ma (Kelly and Harley, 2005) and no ca. 2500–2475 Ma magmatism is recorded in the UHT region.

The unique end-Archaeon to earliest Palaeoproterozoic P – T , deformational and magmatic history of the Vestfold Hills allows it to be exotic to East Antarctica prior to its amalgamation with the adjacent Rauer terrane, as exposed in the Rauer Islands (Fig. 1b). The latter area records not only Archaeon crustal accretion and metamorphism but also latest Mesoproterozoic to earliest Neoproterozoic magmatism,

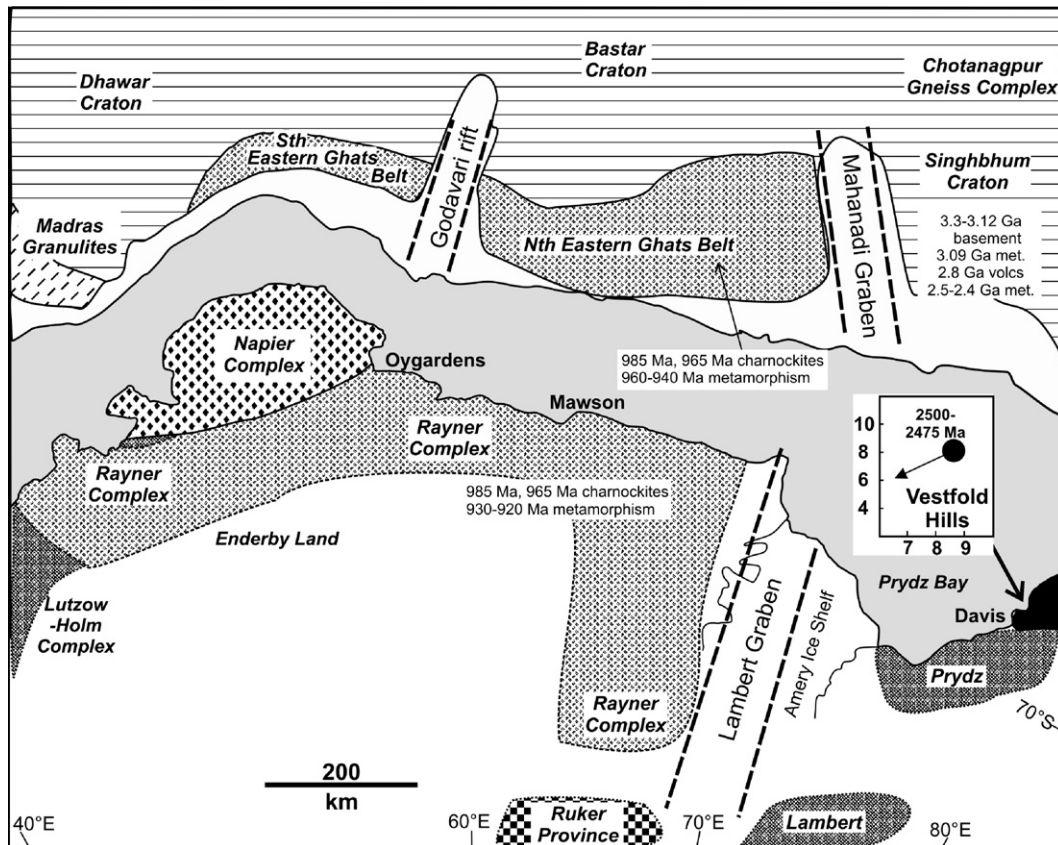


Fig. 18. Close-fit Gondwana reconstruction of East Antarctica and India in the sector (current longitudes) 40 °E to 80 °E, modified from Harley (2003) incorporating evidence from Acharyya (2001), Boger et al. (2006), Misra and Johnson (2005) and this work. The principal tectonic units, provinces and complexes are defined on the basis of their age–event histories. Archaeon blocks or cratons include the Dharwar, Bastar and Singhbhum in India and the Napier Complex, Ruker Province and Vestfold Block (black ornament) in East Antarctica. Each of these, including the Vestfold Block and adjacent Rauer Terrane, have distinct Archaeon crustal histories. The Eastern Ghats Belt (India) and Rayner Complex (Antarctica) are correlated based on their similar earliest Neoproterozoic crustal histories. Intense and extensive high-grade tectonothermal events affect older basement(s) and Neoproterozoic cover sequences in the Prydz Belt, the Lambert Province and the Lutzow–Holm Complex.

deformation and metamorphism (1030–990 Ma: Kinny et al., 1993). As these events appear to be absent in the Vestfold Hills, it is likely that its amalgamation with the Rauer terrane occurred in the Neoproterozoic, after ca. 990 Ma but prior to or during the Cambrian (530–500 Ma) Prydz tectonothermal event that affected the Rauer Islands and the Prydz Bay region to the SW (Zhao et al., 1992, 1995; Hensen and Zhou, 1995, 1997; Carson et al., 1996; Boger et al., 2001; Fitzsimons, 2003; Harley, 2003).

Taking this further, if the Archaean basement of the Vestfold Hills did not form part of the East Antarctic Shield prior to the Neoproterozoic then with what areas in other former Gondwana continents does it potentially correlate?

The most obvious potential candidates for extension of the Vestfold Hills Block into other parts of former Gondwana are the Archaean cratons that occur bounding the Eastern Ghats Belt of eastern India (Fig. 18). The Eastern Ghats are widely held to correlate with the Rayner Complex or Province of East Antarctica (Fitzsimons, 2000a; Harley, 2003; Rickers et al., 2001), recording c. 1000 Ma to 920 Ma arc accretion followed by the collision of India with a craton now underlying East Antarctica (Harley, 2003). To the present north of the Eastern Ghats, and opposite or close to the Vestfold Hills region in generally accepted Gondwana reconstructions (Fig. 18), are the Archaean Singhbhum Craton and the high-grade Chotanagpur Gneissic Complex (Acharyya, 2001; Mukhopadhyay, 2001; Misra and Johnson, 2005). However, examination of the geochronology and age-event records of these regions demonstrates conclusively that neither can be correlated with the Vestfold Hills Block, at least on the basis of current evidence. The Chotanagpur Gneissic Complex is dominated by high grade events at c. 1650–1500 Ma with overprinting in the Central Indian Tectonic Zone at c. 1000 Ma (Acharyya, 2001). The Singhbhum Craton is characterised by the presence of mid-Archaean (c. 3300–3100 Ma) amphibolites and tonalitic gneisses that were metamorphosed at c. 3090 Ma and exposed prior to the extrusion of volcanics at c. 2800 Ma. Although a tectonothermal event is suggested in the early Palaeoproterozoic (c. 2500–2400 Ma), this is not a period of major new crust formation (Mukhopadhyay, 2001; Misra and Johnson, 2005). In short, there is no current evidence for extension of the Vestfold Hills Block into eastern India and no requirement for it to have been part of a greater Indian craton lying to the south of the Central Indian Tectonic Zone (Acharyya, 2001) prior to 1000 Ma.

Fitzsimons (2000a,b) has recently documented evidence for the existence of a Cambrian-aged orogenic belt, the Prydz–Denman–Darling Orogen, that can be inferred to lie to the west, south and east of the Vestfold Hills in Antarctica (Fig. 18) and which joins onto the western

margin of Australia. Johnson (2003) has described Pan-African aged metasediments within the Indian crust underthrusting the Himalaya, and has proposed on the basis of reconstructions of East Gondwana that this crust may correlate at least in part with the Prydz–Denman–Darling Orogen. If this connection is valid, then this Cambrian orogenic belt would cut or terminate the Central Indian Tectonic Zone to the (present) north and east of the Singhbhum Craton, allowing Vestfold Block type Archaean basement to occur beneath the present day Ganges Basin and Bangladesh.

8. Conclusions

The peak P – T conditions calculated in this study are in accord with and improve on those estimated by Harley (1993) using sapphirine-bearing mineral assemblages from the distinct *Taynaya Paragneiss* unit of the Vestfold Hills. However, there is no support for the ultrahigh temperature metamorphic conditions proposed earlier by Collerson and Sheraton (1986a), and there is no good evidence for peak metamorphic temperatures in excess of ca. 900 °C. Instead, the peak ca. 2500 Ma D1–M1 metamorphism in the Vestfold Hills occurred at 8.1 ± 0.7 kb and 850 ± 40 °C. This was followed either by cooling or by cooling with minor decompression until the production of post-D2, M2 garnet corona assemblages at ca. 2475 Ma and 7 ± 1 kb, 650–600 °C. The intervening emplacement of the *Crooked lake Gneiss Group* magmatic rocks occurred at pressures that are as yet unconstrained by geobarometric data.

The distinctive geological and metamorphic P – T evolution of the Vestfold Hills at 2500–2575 Ma, together with its lack of further tectonothermal activity until at least the Neoproterozoic, suggests that this area may be a fragment of a more extensive Archaean high-grade terrane that was exotic to East Antarctica, only amalgamated onto the latter after 990 Ma, and most likely in the Cambrian. If this is the case, then other fragments of this former Archaean terrane may now occur beneath the present eastern end of the Himalayas, and areas such as Bangladesh, to the south of the Himalayas. This speculation remains to be tested through petrological and geochronological studies of further xenoliths from granitoids exposed in Himalayan nappes (Johnson, 2003) and of drill core material from the sub-Gangetic basement in north-east India and Bangladesh.

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References

- Acharyya, S.K., 2001. The nature of the Mesoproterozoic Central Indian Tectonic Zone with exhumed and reworked older granulite. *Gondwana Res.* 4, 197–214.
- Aranovich, L.Y., Berman, R.G., 1997. A new garnet–orthopyroxene thermometer based on reversed Al_2O_3 solubility in FeO – Al_2O_3 – SiO_2 orthopyroxene. *Am. Mineral.* 82, 345–353.
- Asami, M., Suzuki, K., Grew, E.S., Adachi, M., 1998. Chime ages for granulites from the Napier Complex, East Antarctica. *Polar Geosci.* 11, 172–199.
- Black, L.P., Kinny, P.D., Sheraton, J.W., Delor, C.P., 1991. Rapid production and evolution of late Archaean felsic crust in the Vestfold Block of East Antarctica. *Precambrian Res.* 50, 283–310.
- Boger, S.D., Carson, C.J., Wilson, C.J.L., Fanning, C.M., 2000. Neoproterozoic deformation in the Radok Lake region of the northern Prince Charles Mountains, East Antarctica: evidence for a single protracted orogenic event. *Precambrian Res.* 104, 1–24.
- Boger, S.D., Wilson, C.J.L., Fanning, C.M., 2001. Early Palaeozoic tectonism within the East Antarctic craton: the final suture between east and west Gondwana? *Geology* 29, 463–466.
- Boger, S.D., Wilson, C.J.L., Fanning, C.M., 2006. An Archaean province in the southern Prince Charles Mountains, East Antarctica: U–Pb zircon evidence for c. 3170 Ma granite plutonism and c. 2780 Ma partial melting and orogenesis. *Precambrian Res.* 145, 207–228.
- Bohlen, S.R., Wall, V.J., Boettcher, A.L., 1983. Geobarometry in granulites. In: Saxena, S.K. (Ed.), *Kinetics and Equilibrium in Mineral Reactions*. Springer-Verlag, New York, pp. 141–171.
- Carson, C.J., Fanning, C.M., Wilson, C.J.L., 1996. Timing of the Progress Granite, Larsemann Hills, evidence for early Palaeozoic orogenesis within the East Antarctic Shield and implications for Gondwana assembly. *Aust. J. Earth Sci.* 43, 539–553.
- Carson, C.J., Ague, J.J., Coath, C.D., 2002. U–Pb geochronology from Tonagh Island, East Antarctica: implications for the timing of ultra-high temperature metamorphism of the Napier Complex. *Precambrian Res.* 116, 237–263.
- Collerson, K.D., Sheraton, J.W., 1986a. Bedrock geology and crustal evolution of the Vestfold Hills. In: Pickard, J. (Ed.), *The Antarctic Oasis: Terrestrial Environments and History of the Vestfold Hills*. Academic Press, Sydney, pp. 21–62.
- Collerson, K.D., Sheraton, J.W., 1986b. Age and geochemical characteristics of mafic dyke swarms in the Archaean Vestfold Block, Antarctica: inference about Proterozoic dyke emplacement in Gondwana. *J. Petrol.* 27, 853–886.
- Collerson, K.D., Reid, E., Millar, D., McCulloch, M.T., 1983. Lithological and Sr–Nd isotopic relationships in the Vestfold Block: implications for Archaean and Proterozoic crustal evolution in the East Antarctic Shield. In: Oliver, R.L., James, P.R., Jago, J.B. (Eds.), *Antarctic Earth Science*. Australian Academy of Science, Canberra, pp. 77–84.
- Dasgupta, S., Sengupta, P., 2003. Indo-Antarctic correlation: a perspective from the Eastern Ghats Belt. In: Yoshida, M., Windley, B.W., Dasgupta, S., Powell, C.McA. (Eds.), *Proterozoic East Gondwana: Supercontinent Assembly and BreakupSpec.* Publ. Geol. Soc. Lond., vol. 206, pp. 131–144.
- Dirks, P.H.G.M., Hoek, J.D., Wilson, C.J.L., Sims, J.R., 1993. The Proterozoic deformation of the Vestfold Hills basement complex, East Antarctica; implications for the tectonic evolution of adjacent granulite belts. *Precambrian Res.* 65, 277–295.
- Eckert, J.O., Newton, R.C., Kleppa, O.J., 1991. The H of reaction and recalibration of garnet–pyroxene–plagioclase–quartz geobarometers in the CMAS system by solution calorimetry. *Am. Mineral.* 76, 148–160.
- Elkins, L.T., Grove, T.L., 1990. Ternary feldspar experiments and thermodynamic models. *Am. Mineral.* 75, 544–559.
- Essene, E.J., 1989. The current status of thermometry in metamorphic rocks. In: Daly, J.S., Cliff, R.A., Yardley, B.W.D. (Eds.), *Evolution of Metamorphic BeltsSpec.* Publ. Geol. Soc. Lond., vol. 43, pp. 1–44.
- Fitzsimons, I.C.W., 2000a. Grenville-age basement provinces in East Antarctica: evidence for three separate collisional orogens. *Geology* 28, 879–882.
- Fitzsimons, I.C.W., 2000b. A review of tectonic events in the East Antarctic Shield and their implications for Gondwana and earlier supercontinents. *J. Afr. Earth Sci.* 31, 3–23.
- Fitzsimons, I.C.W., 2003. Proterozoic basement provinces of southern and south-western Australia, and their correlation with Antarctica and India. In: Yoshida, M., Windley, B.W., Dasgupta, S., Powell, C.McA. (Eds.), *Proterozoic East Gondwana: Supercontinent Assembly and BreakupSpec.* Publ. Geol. Soc. Lond., vol. 206, pp. 93–130.
- Fitzsimons, I.C.W., Harley, S.L., 1994. The influence of retrograde cation exchange on granulite P – T estimates and a convergence technique for the recovery of peak metamorphic conditions. *J. Petrol.* 35, 543–576.
- Foster, M.D., 1960. Interpretation of composition of trioctahedral micas. *US Geol. Surv. Prof. Pap.* 345b, 11–46.
- Franz, L., Harlov, D.E., 1998. High-grade K-feldspar veining in granulite from the Ivrea–Verbano Zone, northern Italy: fluid flow in the lower crust and implications for granulite facies genesis. *J. Geol.* 106, 455–472.
- Frost, B.R., Chacko, T., 1989. The granulite uncertainty principle: limitations on thermobarometry in granulites. *J. Geol.* 97, 435–450.
- Fuhrman, M.L., Lindsley, D.H., 1988. Ternary-feldspar modelling and thermometry. *Am. Mineral.* 73, 201–215.
- Ganguly, J., Cheng, W., Tirone, M., 1996. Thermodynamics of aluminosilicate garnet solid solution; new experimental data, an optimised model, and thermometry applications. *Contrib. Mineral. Petrol.* 126, 137–151.
- Grew, E.S., Suzuki, K., Asami, M., 2001. CHIME ages of xenotime, monazite and zircon from beryllium pegmatites in the Napier Complex, Khmara Bay, Enderby Land, East Antarctica. *Polar Geosci.* 14, 99–118.
- Guidotti, C.V., Cheney, J.T., Conatore, P.D., 1975. Interrelationships between Mg/Fe and octahedral Al content in biotite. *Am. Mineral.* 60, 849–853.

- Harley, S.L., 1984a. An experimental study of the partitioning of Fe and Mg between garnet and orthopyroxene. *Contrib. Mineral. Petrol.* 86, 359–373.
- Harley, S.L., 1984b. The solubility of Alumina in orthopyroxene coexisting with garnet in $\text{FeO-MgO-Al}_2\text{O}_3\text{-SiO}_2$ and $\text{CaO-FeO-MgO-Al}_2\text{O}_3\text{-SiO}_2$. *J. Petrol.* 25, 665–996.
- Harley, S.L., 1989. The origins of granulites: a metamorphic perspective. *Geol. Mag.* 126, 215–247.
- Harley, S.L., 1993. Sapphirine granulites from the Vestfold Hills, East Antarctica: geochemical and metamorphic evolution. *Antarct. Sci.* 5, 349–402.
- Harley, S.L., 1998. On the occurrence and characterisation of ultrahigh-temperature crustal metamorphism. In: Treloar, P.J., O'Brien, P.J. (Eds.), *What Drives Metamorphism and Metamorphic Reactions?* Spec. Publ. Geol. Soc. Lond., vol. 138, pp. 81–107.
- Harley, S.L., 2003. Archaean–Cambrian development of East Antarctica: metamorphic characteristics and tectonic implications. In: Yoshida, M., Windley, B.F. (Eds.), *Proterozoic East Gondwana: Supercontinent Assembly and Breakup*. Spec. Publ. Geol. Soc. Lond., vol. 206, pp. 203–230.
- Harley, S.L., Black, L.P., 1997. A revised Archaean chronology for the Napier Complex, Enderby Land, from SHRIMP ion-microprobe studies. *Antarct. Sci.* 9, 74–91.
- Harley, S.L., Christy, A.G., 1995. Titanium-bearing sapphirine in a partially melted aluminous granulite xenolith, Vestfold Hills, Antarctica: geological and mineralogical implications. *Eur. J. Mineral.* 7, 637–653.
- Harley, S.L., Fitzsimons, I.C.W., 1995. High-grade metamorphism and deformation in the Prydz Bay region, East Antarctica: terranes, events and regional correlations. In: Yoshida, M., Santosh, M. (Eds.), *India and Antarctica During the Precambrian*. Mem. Geol. Soc. India, vol. 34, pp. 73–100.
- Harley, S.L., Green, D.H., 1982. Garnet–orthopyroxene barometry for granulites and peridotites. *Nature* 300, 697–701.
- Harley, S.L., Hensen, B.J., 1990. Archaean and Proterozoic high-grade terranes of East Antarctica (40–80 °E): a case study of diversity in granulite facies metamorphism. In: Ashworth, J.R., Brown, M. (Eds.), *High-Temperature Metamorphism and Crustal Anatexis*. Unwin Hyman, London, pp. 320–370.
- Harley, S.L., Snape, I., Fitzsimons, I.C.W., 1995. Regional correlations and terrane assembly in East Prydz Bay: evidence from the Rauer Group and Vestfold Hills. *Terra Antarctica* 2, 49–60.
- Harley, S.L., Snape, I., Black, L.P., 1998. The early evolution of a layered metagneous complex in the Rauer Group, East Antarctica: evidence for a distinct Archaean terrane. *Precambrian Res.* 89, 175–205.
- Harlov, D.E., Hansen, E.C., Bigler, C., 1998. Petrologic evidence for K-feldspar metasomatism in granulite facies rocks. *Chem. Geol.* 151, 373–386.
- Hensen, B.J., Zhou, B., 1995. A Pan-African granulite facies metamorphic episode in Prydz Bay, Antarctica: evidence from Sm–Nd garnet dating. *Aust. J. Earth Sci.* 42, 249–258.
- Hensen, B.J., Zhou, B., 1997. East Gondwana amalgamation by Pan-African collision? Evidence from Prydz Bay, Eastern Antarctica. In: Ricci, C.A. (Ed.), *The Antarctic Region: Geological Evolution and Processes*. Terra Antarctica Publication, Siena, pp. 115–119.
- Hoek, J.D., Dirks, P.G.H.M., Passchier, C.W., 1992. A late-Proterozoic extensional–compressional tectonic cycle in East Antarctica. In: Yoshida, Y., Kaminuma, K., Shiraishi, K. (Eds.), *Recent Progress in Antarctic Earth Science*. Terrapub, Tokyo, pp. 137–143.
- Hokada, T., Misawa, K., Yokoyama, K., Shiraishi, K., Yamaguchi, A., in press. SHRIMP and electron microprobe chronology of UHT metamorphism in the Napier Complex, East Antarctica: evidenced by zircon growth at > 1000 °C. *Contrib. Mineral. Petrol.*
- Holland, T.J.B., Powell, R., 1998. An internally consistent thermodynamic data set for phases of petrological interest. *J. Metamorph. Geol.* 16, 309–343.
- Johnson, M.R.W., 2003. Insight into the nature of Indian crust underthrusting the Himalaya. *Terra Nova* 15, 46–51.
- Kelly, N.M., Harley, S.L., 2005. An integrated microtextural and chemical approach to zircon geochronology: refining the Archaean history of the Napier Complex, East Antarctica. *Contrib. Mineral. Petrol.* 149, 57–84.
- Kinny, P.D., Black, L.P., Sheraton, J.W., 1993. Zircon ages and distribution of Archaean and Proterozoic rocks in the Rauer Islands. *Antarct. Sci.* 5, 193–206.
- Kuehner, S.M., Green, D.H., 1991. Uplift history of the East Antarctic Shield: constraints imposed by high-pressure experimental studies of Proterozoic mafic dykes. In: Thomson, M.R.A., Crame, J.A., Thomson, J.W. (Eds.), *Geological Evolution of Antarctica*. Cambridge University Press, Cambridge, pp. 1–6.
- Lanyon, R., Black, L.P., Seitz, H.M., 1993. U–Pb zircon dating of mafic dykes and its application to the Proterozoic geological history of the Vestfold Hills, East Antarctica. *Contrib. Mineral. Petrol.* 115, 184–203.
- Lee, H.Y., Ganguly, J., 1988. Equilibrium compositions of coexisting garnet and orthopyroxene: experimental determinations in the system $\text{FeO-MgO-Al}_2\text{O}_3\text{-SiO}_2$, and applications. *J. Petrol.* 29, 93–114.
- Lindsley, D.H., Andersen, D.J., 1983. A two-pyroxene thermometer. *Proceedings of the Thirteenth Lunar and Planetary Science Conference*, part 2. *J. Geophys. Res.* 88, A887–A906.
- Lindsley, D.H., Nekvasil, H., 1989. A ternary feldspar model for all reasons. *Eos* 70, 506.
- Mikhalsky, E.V., Sheraton, J.W., Laiba, A.A., Tingey, R.J., Thost, D.E., Kamenev, E.N., Fedorov, L.V., 2001. *Geology of the Prince Charles Mountains, Antarctica*. Geoscience Australia Bulletin. Australian Geological Survey Organisation, Canberra. 210 pp.
- Misra, S., Johnson, P.T., 2005. Geochronological constraints on the evolution of the Singhbhum Mobile Belt and associated basic volcanics of the Eastern Indian Shield. *Gondwana Res.* 8, 129–142.
- Moecher, D.P., Essene, E.J., Anovitz, L.M., 1988. Calculation and application of clinopyroxene–garnet–plagioclase–quartz geobarometers. *Contrib. Mineral. Petrol.* 100, 92–106.
- Mukhopadhyay, D., 2001. The Archaean nucleus of Singhbhum: the present state of knowledge. *Gondwana Res.* 4, 307–318.
- Newton, R.C., 1986. Fluids of granulite facies metamorphism. In: Walter, J.V., Wood, B.J. (Eds.), *Fluid–Rock Interactions During Metamorphism*. Springer-Verlag, Berlin, pp. 36–59.
- Newton, R.C., Perkins, D., 1982. Thermodynamic calibration of geobarometers for charnokites and basic granulites based on the assemblages garnet–plagioclase–orthopyroxene (clinopyroxene)–quartz with applications to high grade metamorphism. *Am. Mineral.* 67, 203–222.
- Oliver, R.L., James, P.R., Collerson, K.D., Ryan, A.B., 1982. Precambrian geologic relationships in the Vestfold Hills, Antarctica. In: Craddock, C. (Ed.), *Antarctic Geoscience*. University of Wisconsin Press, Madison, pp. 435–444.
- Passchier, C.W., Bekendam, R.F., Hoek, J.D., Dirks, P.G.H., de Boorder, H., 1991. Proterozoic geological evolution of the northern Vestfold Hills, Antarctica. *Geol. Mag.* 128, 307–318.
- Pattison, D.R.M., Begin, N.J., 1994. Zoning patterns in orthopyroxene and garnet in granulites: implications for geothermometry. *J. Metamorph. Geol.* 12, 410–487.

- Perkins, D., Newton, R.C., 1981. Chamokite geobarometers based on coexisting garnet–pyroxene–plagioclase–quartz. *Nature* 292, 144–146.
- Rickers, K., Mezger, K., Raith, M.M., 2001. Evolution of the continental crust in the Proterozoic Eastern Ghats Belt, India and new constraints for Rodinia reconstruction: implications from Sm–Nd, Rb–Sr and Pb–Pb isotopes. *Precambrian Res.* 112, 183–210.
- Seitz, H.-M., 1994. Estimation of emplacement pressure for 2350 Ma high-Mg tholeiite dykes, Vestfold Hills, Antarctica. *Eur. J. Mineral.* 6, 195–208.
- Snape, I., 1997. Geological and geochemical anatomy of Archaean gneisses from the Vestfold Hills and Rauer Group, East Antarctica. Unpubl. Ph.D. thesis, University of Edinburgh.
- Snape, I., Harley, S.L., 1996. Magmatic history and the high-grade geological evolution of the Vestfold Hills, East Antarctica. *Terra Antarctica* 3, 23–38.
- Snape, I., Black, L.P., Harley, S.L., 1997. Refinement of the timing of magmatism, high-grade metamorphism and deformation in the Vestfold Hills, East Antarctica, from new SHRIMP U–Pb zircon geochronology. In: Ricci, C.A. (Ed.), *The Antarctic Region: Geological Evolution and Processes*. Terra Antarctica Publication, Siena, pp. 139–148.
- Snape, I., Seitz, H.-M., Harley, S.L., Bain, J.H.C., 2001. Geology of the northern Vestfold Hills, East Antarctica (1:30,000 scale map). Australian Geological Survey Organisation, Canberra.
- Tingey, R.J., 1991. The regional geology of Archaean and Proterozoic rocks in Antarctica. In: Tingey, R.J. (Ed.), *The Geology of Antarctica*. Oxford University Press, Oxford, pp. 1–73.
- Wood, B.J., 1974. The solubility of alumina in orthopyroxene coexisting with garnet. *Contrib. Mineral. Petrol.* 46, 1–15.
- Zhao, Y., Song, B., Wang, Y., Ren, L., Li, J., Chen, T., 1992. Geochronology of the late granite in the Larsemann Hills, East Antarctica. In: Yoshida, Y., Kaminuma, K., Shiraishi, K. (Eds.), *Recent Progress in Antarctic Earth Science*. Terrapub, Tokyo, pp. 155–161.
- Zhao, Y., Liu, X., Song, B., Zhang, Z., Li, J., Yao, Y., Wang, Y., 1995. Constraints on the stratigraphic age of metasedimentary rocks from the Larsemann Hills, East Antarctica: possible implications for Neoproterozoic tectonics. *Precambrian Res.* 75, 175–188.