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Compressional and extensional tectonics in low–medium pressure granulites from the Larsemann Hills, East Antarctica

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Abstract – Meta-sediments in the Larsemann Hills that preserve a coherent stratigraphy, form a cover sequence deposited upon basement of mafic-felsic granulite. Their outcrop pattern defines a 10 kilometre wide east–west trending synclinal trough structure in which basement-cover contacts differ in the north and the south, suggesting tectonic interleaving during a prograde, Dp thickening event. Subsequent conditions reached low-medium pressure granulite grade, and structures can be divided into two groups, Ds and Dp, each defined by a unique lineation direction and shear sense. Ds structures which are associated with the dominant gneissic foliation in much of the Larsemann Hills, contain a moderately east-plunging lineation indicative of west-directed thrusting. Ds comprises a co-linear fold sequence that evolved from early intrafolial folds to late upright folds. D3 structures are associated with a high-strain zone, to the south of the Larsemann Hills, where S3 is the dominant gneissic layering and folds sequences resemble D2 folding. Outside the Ds high-strain zone occurs a low-strain D3 window, preserving low-strain D3 structures (minor shear bands and upright folds) that partly re-orient Ds structures. All structures are truncated by a series of planar pegmatites and parallel D4 mylonite zones, recording extensional dextral displacements.

D2 assemblages include coexisting garnet-orthopyroxene pairs recording peak conditions of ~ 7 kbar and ~ 780 °C. Subsequent retrograde decompression textures partly evolved during both D2 and D3 when conditions of ~ 4-5 kbar and ~ 750 °C were attained. This is followed by Ds shear zones which formed around 3 kbar and ~ 550 °C.

It is tempting to combine Ds, Ds structures in one tectonic cycle involving prograde thrusting and thickening followed by retrograde extension and uplift. The available geochronological data, however, present a number of interpretations. For example, Ds was possibly associated with a clockwise P–T path at medium pressures around ~ 1000 Ma, by correlation with similar structures developed in the Rauer Group, whilst D3 and D4 events occurred in response to extension and heating at low pressures at ~ 550 Ma, associated with the emplacement of numerous granitoid bodies. Thus, decompression textures typical for the Larsemann Hills granulites maybe the combined effect of two separate events.

1. Introduction

The Larsemann Hills, located on the Ingrid Christensen Coast of Prydz Bay, East Antarctica (Fig. 1), consist of a number of small ice-free peninsulas and off-shore islands representing some 60 km² of exposure. The area is dominated by migmatitic pelitic, psammitic and felsic paragneisses which are extensively intruded by peraluminous granitic and pegmatitic bodies (Stüwe, Braun & Peer, 1989; Dirks, Carson & Wilson, 1993). This region forms part of the extensive high-grade metamorphic complex of East Antarctica, which is generally interpreted to have evolved during one Neoproterozoic (~ 1000 Ma) granulite facies event (see, e.g. Tingey, 1981; Black et al. 1987; Stüwe, Braun & Peer, 1989). Within the Prydz Bay area, post-peak metamorphic textures, indicative of decompression, have been used to infer a clockwise P–T path for this event (see, e.g. Harley, 1988; Stüwe & Powell, 1989a; Nichols & Berry, 1991; Harley & Fitzsimons, 1991; Fitzsimons & Harley, 1992).

The contribution of the Pan-African event (~ 550 Ma) to the structural and metamorphic evolution of the Larsemann Hills has become increasingly recognized. Zhao et al. (1992) report Pan-African ages for syn-deformational intrusives in the Larsemann Hills, which led Ren et al. (1992) to conclude that the pervasive structural and P–T evolution of the area is of Pan-African age. Dirks, Carson & Wilson (1993) alternatively suggest that the observed decompression textures in the Larsemann Hills, rather than being the result of a continuous clockwise P–T evolution during
either a single 1000 Ma or 550 Ma event, may instead be the result of superposition of two unrelated metamorphic peaks (one at ~ 1000 Ma, post-dated by a second at lower pressure at ~ 550 Ma).

In this paper, we present stratigraphic, structural and kinematic data relevant to the Larsemann Hills, expanding on the earlier structural analyses of Stüwe, Braun & Peer (1989) and Dirks, Carson & Wilson (1993). Relevant micro-textures, recent geochronology and kinematics are integrated within this expanded structural framework. Discussion will focus on the relative effects of the Neoproterozoic (~ 1000 Ma) event and the Pan-African (~ 550 Ma) event in the Larsemann Hills in the light of this information.

2. Lithologies

The Larsemann Hills (including Steinnes Peninsula, Fig. 1) consist predominantly of upper amphibolite to granulite grade paragneisses intruded by numerous generations of pegmatitic and granitic bodies. The area forms part of a metasedimentary sequence which extends to at least the Bolingen Islands to the south (Dirks & Hand, 1995) and the Brattstrand Bluffs (Fitzsimons & Harley, 1991) and western Rauer Group to the north (Harley, 1987). Two primary associations can be identified: (i) a composite mafic-felsic orthogneiss, which forms the boundary of and is interleaved with a (ii) metasedimentary pile, which comprises the bulk of the outcrop in the Larsemann Hills (Figs 2, 3). Figure 4 shows a comparison of previous workers lithological descriptions with that presented here.

2.a. Basement

'Composite orthogneiss' crops out on Bartangen and Steinnes Peninsula and on northern offshore islands (e.g. Kolloy, northeast Vikoy and Fold islands). The composite orthogneiss largely consists of felsic gneiss with discontinuous pods, boudins and layers of mafic units that generally comprise around 10-20%, but locally up to 80-90% of the total rock volume.

The host orthogneiss is homogeneous, light-brown, medium-grained (0.5-2.0 mm) and composed of quartz, feldspar, biotite ± orthopyroxene in southern outcrops, but is more trondhjemitic on northern offshore islands. The mafic component is dominated by hornblende-orthopyroxene assemblages with variable amounts of clinopyroxene, biotite, plagioclase ± quartz that are commonly truncated by felsic leucosomes with minor orthopyroxene-biotite ± clinopyroxene. Rare calc-silicate boudin trains are present.
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Figure 2. Lithological map of the Larsemann Hills, with inferred stratigraphic column for meta-sedimentary units. Major intrusive lithologies are listed in order of degree of deformation. ‘m’ represents presence of metabasite bodies in relevant units.

Figure 3. Lithological map of Steinnes Peninsula. Major south-plunging F_{3}^{1st} structure is shown, the synonymous of F_{3} of Fitzsimons & Harley (1991). The dominant foliation present is S_{2}.
as narrow (1 m) layers within the composite orthogneiss (e.g. Bartrtangen).

2.b. Metasediments
The inferred cover sequence consists of alternating aluminous pelites that grade to more psammitic and felsic units. The constituent units are described in the order they occur, from the contact with the basement orthogneiss in the southeast Larsemann Hills (Fig. 2). As there are no sedimentary younging criteria observed within the sequence, the presented order should not be strictly interpreted as a progressively younging sequence. The description of lithologies and lithological stratigraphy presented here differs from that of previous workers (e.g. Sheraton, Black & McCulloch, 1984; Stüwe, Braun & Peer, 1989; Dirks, Carson & Wilson, 1993) and several additional lithological units are described.

2.b.1. Pelite 1
This unit constitutes around 10% of the outcrop in the Larsemann Hills, and is generally present as continuous layers of variable thickness (~ 10–400 m). No metabasite layers or pods are present within this unit. Pelite 1 is characterized by a wispy, 'rope-like' appearance produced by bundles of coarse-grained sillimanite, spinel and magnetite interleaved with generally pink K-feldspar-rich leucosomes (up to 10 mm wide) that locally account for 75–85% of the unit. Cordierite is common in the sillimanite–spinel domains, but restricted to the margins of the K-feldspar leucosomes. The nature of the contact with the mafic–felsic orthogneiss is obscured, however, the contact with the overlying composite pelite 2 is generally sharp.

2.b.2. Composite pelite 2
This composite unit, which comprises around 10% of exposure, consists of a number of rock types that are interbedded on a metric scale. It varies in thickness from narrow (20–50 m wide) occurrences at Bartrtangen, to larger bodies (up to 100 m) on central Stornes Peninsula and Sigdoy Island (Figs 1, 2). Composite pelite 2 comprises the following units:

(i) A discontinuous coarse-grained, cordierite-dominated, spinel–sillimanite–biotite–ilmenite–magnetite gneiss with highly variable amounts of coarse subhedral garnet, K-feldspar and quartz ('blue gneiss' of Stüwe, Braun & Peer, 1989). Cordierite forms an extensive multigrained matrix in which all other phases are present. Cordierite also forms inclusion-free coronas on garnet (Stüwe & Powell, 1989a). Massive pods of coarse-grained cordierite–biotite and euhedral spinel–magnetite are common.

(ii) Megacrystic orthopyroxene-bearing tonalite (+biotite and garnet) occurs as narrow layers and lenses up to 50 cm wide, and is comprised of equidimensional plagioclase–quartz–biotite matrix host large blasts of euhedral orthopyroxene (up to 30 mm diameter, see Stüwe & Powell, 1989b, fig. 1c). Orthopyroxene blasts may have a 'spongy' appearance due to numerous inclusions of quartz, and are variably mantled by coarse biotite.
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(iii) Sulphide-rich quartzite occurs as narrow (4–5 m), laterally discontinuous layers and forms characteristic dark yellow and iron-oxide stained gossan-like outcrops with coarse skeletal cubic limonite (after pyrite?) and numerous patches of secondary anhedral azurite and malachite. Associated coarse-grained blue quartzite layers and pods are also common.

(iv) A distinctive, homogeneous, medium grained (~1 mm), light grey, plagioclase–biotite–spinel±magnetite±quartz gneiss, the ‘granular-porphyroblastic-gneiss’ (GPG sub-unit), occurs as lensoid, discontinuous pods up to 200 m long and several tens of metres wide. This sub-unit characteristically hosts large (up to 30 mm diameter), euhedral Fe–Mg porphyroblasts that typically consist of a single unzoned almandine garnet or cordierite blast, which are mantled by coarse plagioclase (cf. Stüwe & Powell, 1989b). Inclusions within the blasts are generally coarse biotite, plagioclase feldspar, ilmenite and magnetite, which may or may not be aligned with the external foliation. Multi-minerallic aggregates are also present within this sub-unit. These are composed of a aggregate core with variable amounts of garnet–cordierite–kornerupine–spinel–biotite–grandidierite, again variably mantled by plagioclase (Carson, Hand & Dirks, 1995; Ren et al. 1992). Such blasts and multi-minerallic aggregates generally cross-cut the host biotite foliation, but locally they can be flattened within and tightly to isoclinally folded by that foliation.

2.b.3. Pelite

Pelite is similar to pelite, but it is garnet-bearing. It locally contains a very leucocratic quartz-bearing, K-feldspar–cordierite sub-unit (‘felsic cordierite gneiss’ = ‘white gneiss’, Stüwe, Braun & Peer, 1989) which is rarely garnetiferous with distinctive cordierite–quartz symplectites. This unit is particularly common on central and eastern Stornes Peninsula. Metabasite pods and layers consisting of predominantly hornblende–plagioclase–biotite±orthopyroxene are present, but are uncommon.

2.b.4. Semi-pelite

This unit accounts for up to 40% of the total outcrop in the Larsemann Hills, and up to 80% of outcrop on Broknès Peninsula (Fig. 2). Semi-pelite is a heterogeneous fine- to medium-grained garnet–sillimanite–spinel–plagioclase–quartz±biotite gneiss with a characteristic sugary texture and a yellow to pale weathering colouration. The modal amounts of spinel and sillimanite are highly variable and cordierite is uncommon. Magnetite forms coarse (20–30 mm diameter) euhedral segregations. Semi-pelite has a transitional contact with pelite, with garnet becoming more abundant at the expense of cordierite (Fig. 2). A common feature of this unit is the occurrence of discontinuous hornblende, orthopyroxene, clinopyroxene, biotite, plagioclase-bearing mafic lenses and boudin pods (~5% of the unit). This unit is equivalent to the ‘yellow gneiss’ of Stüwe, Braun & Peer (1989); ‘migmatitic paragneiss’ of Dirks, Carson & Wilson (1993) and ‘striped migmatite’ of Zhao et al. (1992).

2.b.5. Psammite

This unit accounts for ~40% of the total outcrop and up to 70% of outcrop on Stornes Peninsula (Fig. 2). It generally forms a homogeneous unit with equigranular granitic appearance, with quartz, plagioclase, K-feldspar and variable amounts of garnet and biotite. Sillimanite is rare and cordierite is absent. Small aluminous enclaves (~10–20 cm diameter) are composed of coarse-grained sillimanite–spinel±magnetite and minor cordierite can also be present. This unit is highly migmatized with up to three overprinting generations of garnet-bearing felsic leucosomes that cross-cut the primary compositional layering. Numerous metabasite pods and lenses are present (up to 5 m wide), dominated by hornblende–plagioclase assemblages with minor, variable, amounts of orthopyroxene, biotite and, occasionally, clinopyroxene. Narrow, laterally extensive layers (30–40 cm wide) of coarse-grained radiating sprays of kornerupine are present. Rare garnet, orthopyroxene, plagioclase-bearing metabasite assemblages are also present. Psammite, is gradational with semi-pelite with cordierite, spinel and sillimanite becoming gradually less common.

2.b.6. Psammite

This felsic unit constitutes around 10% of total outcrop and differs from psammite, in that it contains more felsic component relative to Fe–Mg phases and lacks small aluminous enclaves and mafic lenses. Sillimanite and cordierite are also absent. This unit is common along the contact with the basement orthogneiss on the northern offshore islands, but generally absent elsewhere.

2.c. Intrusives

The Larsemann Hills are intruded by various generations of granitic and pegmatitic bodies (Fig. 2). Pegmatites and migmatitic bodies have developed synchronously and, less commonly, post-dating regional high grade deformation. The intrusive bodies have been interpreted as locally derived melts based on S-type geochemical signatures (Stüwe, Braun & Peer, 1989; Zhao et al. 1992). Dirks, Carson & Wilson (1993) described a detailed history of pegmatite intrusions, and these will not be discussed further.
2.c.1. Zhong Shan gneiss

This massive leucocratic quartz-feldspathic gneiss 'painted rock' – Stüwe, Braun & Peer, 1989; 'granitic gneiss I' – Dirks, Carson & Wilson, 1993; 'gneissic leucogranite' – Zhao et al. (1992), crops out on northeastern Broknes Peninsula and preserves a clear foliation defined by stringers of garnet, rare biotite and a compositional banding defined by quartz-feldspar leucosomes. Hornblende–plagioclase-bearing metabasite units are present as isolocinally folded layers (Dirks, Carson & Wilson, 1993). On Steinnes Peninsula, this unit occurs as a strongly-foliated, narrow transposed layer (3–4 m wide). The primary nature of the Zhong Shan gneiss is unclear, but previous workers assumed an igneous origin based on its homogeneous nature (Stüwe, Braun & Peer, 1989; Dirks, Carson & Wilson, 1993), however, a sedimentary origin cannot be discounted. Zhao et al. (1992) determined a zircon \(^{207}\text{Pb}/^{206}\text{Pb}\) age of 940 ± 6 Ma, and a Sm–Nd three point isochron (garnet + Kspar + whole rock) age of 497 ± 7 Ma for this rock.

2.c.2. Progress granite

A relatively large body of this granite outcrops on Mirror Peninsula ('granitic gneiss V' – Dirks, Carson & Wilson, 1993; ‘pink granite’ – Stüwe, Braun & Peer, 1989; 'syenogranite' – Zhao et al. (1992), with other minor occurrences on Steinnes Peninsula (Fig. 3; 'granitic orthogneiss', GOG – Fitzsimons & Harley, 1991), Barrtangen, Vikoy Island and various islands to the north of Uspo Island (Fig. 2). It is comprised of medium- to coarse-grained K-feldspar and quartz with subordinate plagioclase. Biotite locally defines a weak to moderate foliation. Accessory phases include, in order of abundance, sub-anhedral garnet with rounded inclusions of quartz, magnetite, spinel, monazite and zircon. Zhao et al. (1992) determined a zircon \(^{207}\text{Pb}/^{206}\text{Pb}\) age of 547 ± 9 Ma for the crystallization age of the Progress Granite from Mirror Peninsula.

2.c.3. Grovness Enderbitite

A massive but well-foliated non-megacrystic orthopyroxene–tonalite body is exposed on southern Grovness Peninsula. It is dominated by medium-grained subhedral polygonal plagioclase and quartz with strongly corroded grains of orthopyroxene. Rare, strongly embayed, anhedral garnet, occurs both as isolated grains and in contact with orthopyroxene. Small euhedral magnetite grains (which occur on the rims of orthopyroxene grains), monazite, zircon and apatite are accessory phases. Subhedral biotite occurs as inclusions with orthopyroxene grains. An intrusive origin is preferred, based on the massive, extremely homogenous nature of this unit, but a sedimentary origin cannot be ruled out.

2.c.4. Dalkoy granite

The Dalkoy granite (Fitzsimons & Harley, 1991; ‘granitic gneiss VI’, Dirks, Carson & Wilson, 1993; ‘grey granite’, Stüwe, Braun & Peer, 1989) crops out on the northern portion of Dalkoy Island. It consists of medium-grained K-feldspar, quartz, plagioclase and biotite. K-feldspar and plagioclase are variably replaced by muscovite and sericite. Accessory phases include rounded magnetite grains, zircon, apatite and rare monazite. The unit contains a variably developed foliation defined by euhedral biotite and aligned quartz and feldspar. This foliation is axial planar to open folds defined by thin, discontinuous, quartzfeltdspathic leucosomes. The contact of the Dalkoy granite is concordant with the pervasive \(S_2\) foliation of the host pelitic gneiss. The Dalkoy granite has been correlated with the Landing Bluff granitoid suite to the south by Stüwe, Braun & Peer (1989) and Fitzsimons & Harley (1991), for which a 493 ± 17 Ma age has been determined (Rb–Sr whole rock; Tingey, 1981).

2.c.5. Storneskola megacrystic K-feldspar orthogneiss units

Two porphyritic K-feldspar orthogneiss units occur as small isolated exposures on Storneskola (Fig. 1) and Stornes Peninsula regions, and account for less than 5% of total outcrop. Augen K-feldspar orthogneiss, is comprised of K-feldspar, plagioclase, quartz, biotite and minor garnet, with large (20–30 mm diameter) Kfeldspar augen, which locally spectacularly define the ambient lineation (e.g. islands north of Storneskola). The augen K-feldspar orthogneiss, can occur in association with, and is clearly intruded by, the prophyroblastic K-feldspar orthogneiss, which consists of numerous large, euhedral, undeformed, K-feldspar porphyroblasts (20–70 mm long), within a finer ground-mass of K-feldspar, quartz, plagioclase, biotite (Fig. 5). Garnet occurs as subhedral grains within the ground-mass, and also occurs with coarse biotite as selvedges along the unit perimeter.

2.c.6. Late K-spar granitic dyke

A pink magnetite–sillimanite-bearing granitic dyke occurs on Broknes Peninsula and Sigdoy Island and has been described by Dirks, Carson & Wilson (1993) and Stüwe, Braun & Peer (1989). Zircon \(^{207}\text{Pb}/^{206}\text{Pb}\) ages of 556 ± 7 Ma (sample ZG20405) have been obtained (Zhao et al. 1992).

2.d. Basement and cover relationships

Based on the lithological similarity with mafic–felsic orthogneiss from the southeastern Rauer Group and Berrneset Peninsula (Bolingen Island, Dirks & Hand, 1994) for which Archaean and Palaeoproterozoic formation ages (respectively) have been suggested
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Figure 5. Augen K-feldspar orthogneiss (labelled 'a', foliation is $S_2$) intruded by the porphyroblastic K-feldspar orthogneiss, (labelled 'p'), coarse unoriented phenocrysts of K-feldspar arrowed, Storneskola. Lens cap ~ 50 mm diameter.

(\text{U-Pb} – Black & Sheraton, 1993; \text{Rb-Sr}, sample 81285291 – Sheraton, Black & McCulloch, 1984), we assume that the mafic–felsic composite orthogneiss represents an archaean basement complex to the metasedimentary sequence of the Larsemann Hills. The extensive metasedimentary sequence, for which a range of Mesoproterozoic ages have been determined (1300–1600 Ma \text{Rb-Sr} ages – Sheraton, Black & McCulloch, 1984) is considered to have been derived from, and deposited on, the Archaean orthogneiss basement. Similar assumptions on the nature of the relationship between composite orthogneiss and paragneiss sequences is also assumed in the Brattstrand Bluffs and associated outcrops by Fitzsimons & Harley (1991, 1992) and the Bolingen Islands (Dirks & Hand, 1995). The primary nature of the orthogneiss–paragneiss contact is difficult to determine due to subsequent deformation and transposition. The nature of both the basement orthogneiss and its relationship with metasediments differ, north to south, across the Larsemann Hills. On northern offshore islands, the mafic–felsic orthogneiss is more trondhjemitic, hosts a higher mafic component and is interleaved and infolded with psammite, associations that are not observed along the southern margin where mafic–felsic orthogneiss is in contact with pelite. This asymmetry from south to north may represent lateral sedimentary facies variation, but more likely represents structural disturbance.

2.d.l. Metabasite bodies

Discontinuous, concordant lenses and pods of metabasite, consisting of variable amounts of hornblende–plagioclase–biotite ± orthopyroxene ± clinopyroxene, occur within pelite, semi-pelite, psammite, and psammite, units. The primary nature of these metabasite units is problematical. Stüwe, Braun & Peer (1989) suggested that the bulk compositions of two pyroxene–plagioclase ‘rafts’ were consistent with derivation from meta-igneous precursors, and biotite-bearing, pyroxene-poor varieties have bulk compositions close to felsic orthogneiss, but assumed either a volcanic or intrusive origin. Dirks, Carson & Wilson (1993) assumed an intrusive origin for the metabasite lenses based on (i) the presence of such units within what they assumed to be felsic igneous bodies (their ‘granitic gneiss I and II’), and (ii) the correlation of the metabasites with the abundant Proterozoic tholeiitic dykes of the Archaean Vestfold Hills Complex based on a similar geochemical signature. Such geochemical criteria are, however, somewhat inconclusive and ambiguous and the assumption that the metabasite host lithology is of igneous origin is difficult to substantiate. A volcanogenic sedimentary origin for the metabasite units is considered more likely.

The occurrence and abundance of metabasite units is strongly stratigraphically controlled (Fig. 2). Metabasites within the sedimentary pile occur only within mid- to upper units in the described stratigraphy and with increasing abundance (Fig. 3). It is tempting to suggest that this indicates a younging direction, given that basic volcanism is common in the early stages of extensional basin development. If so, the sequence, as described, is inverted.

3. Structural relationships

The structure of the Larsemann Hills is affected by two major high-grade events $D_2$ and $D_3$, during which metamorphic conditions reached granulite facies grade. $D_4$ and $D_5$ are composite events, each consisting of a number of fold generations and foliation forming events each with identically and uniquely directed fold axes and mineral elongation lineations. $D_4$ is a major crustal thickening event which is coeval with the metamorphic peak. $D_5$ consists of a number of discrete high-grade sillimanite-bearing mylonites and $D_6$ is characterized by the development of high-grade shear systems with a normal sense of movement. $D_7$ consists of a number of discrete high-grade sillimanite-bearing mylonites and $D_8$ is characterized by the development of planar low-grade brittle features which accommodated minor offsets. The structural nomenclature described here differs significantly from that suggested by Stüwe, Braun & Peer (1989) and any similarity in numbering of deformational schemes is coincidental (Fig. 6). A summary of structural geometries described here is presented in Figure 7.

3.a. $D_1$ deformation

There is little information concerning the geometry and nature of $D_1$ deformation. We assign $D_1$ deformation to post-depositional, early prograde burial and crustal thickening, all of which is obscured by late prograde, peak and subsequent, retrograde metamorphism. Probably the best indication of the
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D1: Interleaving of basement and sedimentary cover, possibly due to thrusting resulting in burial

D2: WNW-directed thrusting resulting in uplift

D2 low strain zone:
(a) formation of S2 form surface and L2x
(b) F2m interfolial folds with strong asf // S2, L2x and L2a
(c) F2a, isoclinal folds, weak asf, L2a // L2x
(d) F2c, reclined chevron folds, no asf, L2x // L2a
(e) F2d open upright folds, no asf, L2x // L2a
(f) F2e open upright folds, no asf, L2x // L2a

D2 high strain zone:
(a) formation of S2a form surface with L2x, followed by 2-3 episodes of isoclinal folding and transpositions resulting in new form surface S2b which truncates S2a
(b) F2f open upright folds, no asf, L2a // L2a
(c) F2g open upright folds, no asf, L2x variable

D3: N-S extension resulting in limited uplift

D3 low strain zone:
(a) F3 open upright folds // L3x
(b) development of mostly sinistral shear bands associated with leucosome. Minor dextral shear bands also present. Intersection lineation of these sets of shears // with L3x

D3 high strain zone:
(a) formation of S3a form surface with L3x, followed by 2-3 episodes of isoclinal folding and transposition locally resulting in a new form surface S3b

D4: Dextral shear along N-NW trending mylonite zones.

D5: Brittle faults - associated with the formation of Lambert Graben?

D1 - D3 of Sims et al. (1994) are correlated to structures developed within the Archaean Craton of the Vestfold Hills.

Figure 6. Correlative framework comparing deformation schemes proposed for the southern Prydz Bay region, including the Larsemann Hills, Brattstrand Bluffs, the Bolingen Islands and the Rauer Group. * D1 and D2 events listed here represents very early events that are difficult to confidently correlate. † D1 - D3 of Sims et al. (1994) are correlated to structures developed within the Archaean Craton of the Vestfold Hills.

Figure 7. Summary of the structural and geological history of the Larsemann Hills. Possible scenarios for age of deformation events shown on the left. Abbreviations used: asf, axial surface foliation; //, parallel; L3x, fold hinge orientation of F3b-a.

nature of D1 deformation may be derived from the distribution pattern of lithological units across the Larsemann Hills (Fig. 2). Whereas to the south, the assumed lower unit of the stratigraphic sequence is in contact with the basement gneiss, to the north relationships are exactly reversed with basement in contact with some of the assumed highest stratigraphic units. The asymmetrical nature of the cover–basement contact can be well explained by tectonic interleaving and thrust–pile stacking during D1 thrusting. Evidence for pre-early syn-D2 event(s) is preserved within peak D2 garnets from the composite pelite2, pelite3 and
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3.b. D₂ thrusting event

The mappable form surface in the majority of the Larsemann Hills formed during D₃. It may consist of a number of truncating foliation generations that are locally complexly folded and has therefore been described for Broknes Peninsula as D₂₋₄ by Dirks, Carson & Wilson (1993). Almost all folds and multiple foliations within D₂ are co-linear, meaning that linear elements including F₁ fold hinges, intersection lineations between various S₂ foliation generations and mineral extension lineations preserved on S₂ surfaces (L₂x) are commonly parallel. The average orientation of this lineation direction changes from moderately plunging to east-northeast to east in the western Larsemann Hills to south-southeast on Broknes Peninsula, depending on the relative position of D₂ structures in large-scale D₃ folds (Figs 8, 9).

Regionally, two structural D₂ domains can be distinguished (Fig. 8). These are, (i) zones with a relatively planar, apparently simple gneissic foliation (e.g. Seal Cove and Small Peninsula), which truncate or transpose, (ii) domains with an earlier S₃ form surface that is multiply folded. These domains can be interpreted as high- and low-strain zones respectively. The planar S₂ high-strain zones remain active as ductile shear zones up to the later stages of D₂, whereas the 'low-strain' areas represent domains where S₂ ceased to develop relatively early on during D₂, and further strain was accommodated by progressive co-linear F₂ folding. Based on the co-linearity of linear elements in all D₃ domains, it is assumed they all formed during the one composite D₃ event, associated with a continuous kinematic framework, in this case northwest-directed thrusting (Dirks, Carson & Wilson, 1993, fig. 4e).

3.b.1. F₂ fold sequences

S₂ is folded by at least five progressively developed generations of F₂ folds (F₂₋₅), all of which are not semi-pelite, as microscopic, tightly to isoclinally folded, inclusion trails of fibrolitic sillimanite, which are commonly oblique to the pervasive S₂ foliation. Similar micro-textural relationships are described by Fitzsimons & Harley (1991) which are used as evidence of early deformation events (D₁₋³(F&H)) from the Brattstrand Bluffs to the northeast. (D₁₋³(F&H) refers to the deformation scheme proposed for the Brattstrand Bluffs by Fitzsimons & Harley, 1991.)
necessarily present in any one location. Fs2a folds have only been observed on outcrop-scale, and are rare isoclinal intrafolial folds with a moderate to strongly developed axial planar foliation (Ss2a) defined by biotite, sillimanite and compositional layering (Fig. 10a). Ss2a is only identifiable within the fold closure, but is otherwise indistinguishable from the Ss1 foliation it folds. The hinge region commonly has a characteristic 'feathered' or flame-like appearance as the folded lithological contact is irregular (Fig. 10a). Fs2b tight to isoclinal folds are similar to Fs2a folds, and are characterized by a weak axial surface foliation (Ss2b) generally defined by biotite and/or sillimanite, and a more regular fold surface. Fs2c folds are macroscopic (~ 200–400 m amplitude), open to tight, folds with inclined to recumbent axial surfaces and angular hinge regions giving them a chevron fold appearance. Axial surface foliations are absent or weak and only developed in biotite-rich units. Good examples of these folds can be seen on eastern Stornes Peninsula, Eliza Kate and Breadloaf islands (Fig. 2). Fs2c folds can only be found in the low-strain Ds domains. They may have been present in the high-strain zones as well, but in that case all evidence is completely obliterated due to intense transposition. In contrast the common Fs2d folds are present in all Ds domains. They vary in scale from 1–5 mm crenulations of biotite, to 5 m wide, upright to slightly reclined, open to closed cylindrical folds that contain no axial planar foliation. In outcrop Fs2d fold generations are co-linear with the ambient mineral extension direction (Ls2x) present on Ss (Fig. 10b), although variations in absolute orientation occur due to late D3 folding (Fig. 8). A final set of Fs2e folds are open to tight upright folds that differ from previous generations in that they can locally fold Ls2x.

Within the relatively planar high-strain zones, a number of fold types may be present. In general one or more generations of tightly to isoclinal folded, highly disharmonic, parallel folds may be distinguished, some of which preserve sheath fold geometries that are not everywhere absolutely co-linear with the ambient mineral extension lineation. These folds are subsequently overprinted by Fs2f type folds.

Although Ss, especially in the high-strain zones, may be equated with a non-coaxial flow foliation,
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Figure 10. (a) Typical F2a fold with ‘flamelike’ geometry and characteristic strong axial surface fabric, Stornes Peninsula. Hammer handle ~ 0.75 m. (b) Interference pattern of co-linear F2b folds viewed in a plane normal to the lineation, Broknes Peninsula. Pencil (arrowed) indicates direction of L2x and fold hinge.

kinematic indicators in S2 are infrequent (Fig. 11). They are restricted to uncommon mesoscopic fold asymmetries for folds that are not co-linear with L2x (fig. 4e of Dirks, Carson & Wilson, 1993), leucosome-filled shear bands and rotated boudins, all of which are consistent with west- to northwest-directed thrusting if the current orientation of the D2 high-strain zones is taken as a reference frame (Dirks, Carson & Wilson, 1993).

3.c. D3 extension event

A second major composite event, D3, overprints and locally transposes D2 structures. These are completely replaced by a new gneissic layering, S3, on the north and south sides of the Larsen-IV Hills (Fig. 9). Therefore, the larger part of the Larsen-IV Hills represents a window of low D3 strain in which D2 deformation is largely preserved.

D3 structures can be categorized in the same manner as D2 structures with (a) D3 high-strain zones where a number of truncating foliations, with the same mineral elongation lineation direction, are paralleled by various generations of F3 fold hinges, and (b) D3 low-strain domains where S3 constitutes the form surface and D3 effects are limited to later folds, shear bands, migmatite development and partial metamorphic re-equilibration. The average L3x orientation plunges 47° to 210° with all generations of F3 fold axes sub-parallel to this orientation in both high- and low-strain areas (Fig. 9). In regions transitional from S2 to S3-dominated domains, two lineations may be preserved on foliation surfaces, that is, in semi-pelite on southern Stornes Peninsula, L2x sillimanite-spinel lineations overprint an earlier east-plunging L2x lineation defined by sillimanite and aligned ‘trails’ of large (5–20 mm diameter), subhedral garnet.

3.c.1. F3 fold generations and low-strain structures

Fold sequences in D3 differ from D2 high-strain zones to low-strain D3 domains. In D3 high-strain zones the sequence of fold geometries are similar to those described for D2 high-strain zones. Early intrafolial folds (F3a) are refolded by up to two generations of tight to isoclinal folds (F3b_c) with generally weak to no axial planar foliations and a late generation of upright, closed to open F3d folds.

F3 folds in low-strain D3 domains (F3ls) are generally upright open to isoclinal folds with axial surfaces that dip around 80° towards 100–130° with fold hinges paralleling the regional L3x recorded from D3 high-strain zones. Fold amplitudes vary from several centimetres to several kilometres and two major F3ls folds can be recognized on Sigdoy Island and Steinsnes Peninsula (e.g. upright south plunging F3ls structure (fig. 3a of Fitzsimons & Harley, 1991). Large-scale F3ls folding is responsible for major re-orientation of D3 structural elements (Figs 8, 9). Along the axial surface trace of the major open F3ls fold that runs through Sigdoy Island, numerous small-scale parasitic folds are present (Fig. 12a). On Foxtrott Oskar Island and Vikoy Island interference of these small scale folds with upright F3ls folds produces type 1 fold interference geometries. The F3ls structure on Sigdoy Island is relatively open compared with the F3ls tight to isoclinal major and parasitic fold structures on Steinsnes Peninsula (Fig. 3), suggesting intensification of D3 strain towards the northeast. Similarly, Fitzsimons & Harley (1991) report D3(F&H) (D this paper) strain intensification northeast from Steinsnes Peninsula to the Bratstrand Bluffs with ‘increasing structural depth’ (e.g. figs 3b, 7 and 8 of Fitzsimons & Harley, 1991; fig. 2 of Fitzsimons & Harley, 1992). An alternative description, however, is a progressive transition from a D3ls which preserves upright F3ls structures folding an S2 foliation, intensifying to the northeast into D3 high-strain zone, dominated by S3 foliation (= S(F&H)), transposing D3ls structures.

Additional D3 structures that are present in low-strain domains are sets of asymmetrical crenulation bands or cleavage boudins (Platt & Vissers, 1980) that form discrete discontinuous ductile shear bands,
commonly with co-planar leucosome development (Fig. 12b). Within exposures west of the $F_3^{\text{hsz}}$ fold trace through Sigdoy Island such shear bands are limited to one set which is upright to steeply dipping (toward $\sim 130^\circ$) with generally an apparent sinistral movement sense in plan view. To the east of this fold trace (Broknes Peninsula), leucosome-bearing sinistral shear bands occur as overprinting sets displaying a progressive clockwise rotation from old to young ($= S_{8\text{a}, b, c}$, Dirks, Carson & Wilson, 1993). This suggests that the west limb of the major $F_3^{\text{hsz}}$ fold was stable in the $D_3$ stress field whilst the east limb rotated. This also indicates that the east-plunging direction of $L_3^{\text{x}}$ is likely to be closest to its original orientation. The upright $D_3$ shear bands and associated leucosomes are also developed in the axial surfaces of mesoscopic, parasitic $F_3^{\text{hsz}}$ folds on Foxtrot Oskar, Fold, and Sigody islands, where they are associated with offsets of up to 0.5 m in the hinge region (Fig. 12b). Within extreme southwest exposures (e.g. Storneskola), $D_3$ sinistral shear bands form conjugate sets with a set dipping moderately toward $200^\circ$ and with a dextral sense in plane view. The transport vector, where recorded, is sub-parallel to $L_3^{\text{x}}$ in $D_3^{\text{hsz}}$ domains which also parallels the intersection lineation of the conjugate $D_3^{\text{hsz}}$ shear bands (Fig. 9).

Kinematic indicators in $S_3$ high-strain domains consistently preserve a south-down normal movement sense with a minor dextral component (Fig. 11). These indicators include shear bands, back-rotated calc-silicate boudin blocks (Fig. 12c) and asymmetrical metabasite boudins in mafic-felsic orthogneiss (Barrtangen), and rare $\sigma$-blasts defined by cordierite mantles on garnet (Fig. 12d, southern Stornes Peninsula). The locations of major $D_3$ kinematic indicators are shown on Figure 9.

The composite $D_3$ event in the Larsemann Hills equates with $D_{3-4(P&H)}$ events reported from the
Brattstrand Bluffs coastline to the northeast. Both a tightly folded foliation, \( S_{3(\mathrm{F\&H})} \), and a flat-lying shear foliation, \( S_{4(\mathrm{F\&H})} \), preserves a predominantly south-plunging lineation (\( L_{3(\mathrm{F\&H})} \) and \( L_{4(\mathrm{F\&H})} \)), similar to the composite \( S_3 \) foliation and \( L_{3x} \) lineation presented here for the Larsemann Hills. \( F_{4(\mathrm{F\&H})} \) fold axes parallel this \( L_{4(\mathrm{F\&H})} \), similar to the co-linear nature of multiple \( F_3 \) fold generations developed in the southern Larsemann Hills.

3.d. \( D_4 \) deformation

\( D_4 \) deformation is confined to the development of up to 20 cm wide, upright amphibolite-grade mylonite zones that formed along planar north–south-trending garnet–sillimanite–spinel-bearing pegmatites (\( D_4 \), Dirks, Carson & Wilson, 1993; Fig. 13a). These mylonites are best developed on the western Broknes Peninsula and Sigdoy Island, and are relatively uncommon elsewhere. The movement sense is typically dextral, east-down (fig. 6 of Dirks, Carson & Wilson, 1993) along a 20–30°S pitching sillimanite lineation, and offsets are generally less than 20 m. Within pegmatites, \( S_4 \) is defined by sillimanite and quartz ribbons, whilst garnet and Zn-spinel (~ 6 wt % ZnO) are stable. K-feldspar porphyroclasts are semiplastically deformed and develop asymmetrical tails that are stretched along \( S_4 \). Planar north–south orientated fractures that contain garnet, cordierite, quartz, magnetite and randomly orientated biotite, occur both adjacent to, and distal from, \( D_4 \) mylonite zones, and are considered to have formed syn-\( D_4 \). No offset is associated with these fractures (Fig. 13b). The stability of garnet, sillimanite, Zn-spinel and K-feldspar in \( D_4 \) mylonites, and garnet, cordierite and biotite in planar fractures suggests at least amphibolite grade conditions during \( D_4 \). Similar planar pegmatites with sheared margins occur throughout the Prydz Bay region (see, e.g. Harley, 1987; Fitzsimons & Harley, 1991; Figure 6).

3.e. \( D_5 \) deformation

\( D_5 \) deformation is restricted to uncommon planar, sub-vertical brittle faults with variably developed sub-horizontal slickenlines and abundant epidote. These have been described as \( D_7 \) by Dirks, Carson & Wilson (1993).
4. Structural–metamorphic relationships

This section describes metamorphic textures in reference to the structural–kinematic framework and briefly outlines preliminary pressure–temperature estimates. The textures observed within metapelitic units are consistent with many of the textural features observed in metapelitic sequences of the Brattstrand Bluffs (Fitzsimons & Harley, 1991, 1992).

4.a. Domains with $S_2$ as dominant foliation

$S_2$ assemblages preserved within pelite, composite pelite, and pelite contain variable amounts of garnet, cordierite, quartz, plagioclase, K-feldspar, spinel, sillimanite and biotite. Two textural varieties of garnet are commonly present. Garnet, is strongly corroded and generally elongate in $S_2$ and $L_{2a}$, and characterized by abundant inclusions of tightly folded fibrolitic sillimanite and biotite. Inclusion-free moats separate sillimanite and biotite, within garnet, suggesting destabilization of early biotite–sillimanite to form garnet (Grant, 1985; cf. Brattstrand Bluffs, Fitzsimons & Harley, 1991, 1992) via the prograde reaction:

$$\text{biotite} + \text{sillimanite} + \text{quartz} = \text{garnet} + \text{K-feldspar} + \text{melt}, \quad (1)$$

early during $D_2$. Garnet and early foliation forming, spinel, are separated by cordierite moats which may form coarse-grained overgrowths aligned in $S_2$ or delicate symplectic intergrowths with plagioclase, quartz and spinel. Possible retrograde reactions include:

$$\text{garnet} + \text{spinel} = \text{cordierite} + \text{plagioclase} \quad (2)$$

and

$$\text{garnet} + \text{sillimanite} = \text{cordierite} + \text{quartz} + \text{spinel}. \quad (3)$$

The coarse-grained cordierite overgrowths consist of large (several mm) grains with relatively straight grain boundaries suggesting textural re-equilibration after formation of the coronas in response to syn-$D_2$ decompression. This type of texture is very common in the semi-pelite. In contrast the delicate symplectic intergrowths involving cordierite are in textural disequilibrium and must have formed relatively late. Although many of these are also aligned in $S_2$ or $L_{2a}$, this may have resulted from replacement of $D_2$ phases rather than syn-$D_2$ growth, and the origin of these symplectites including spinel is probably syn-$D_3$. This is suggested by the common occurrence of such symplectites in $D_3$ high-strain zones.

Although inclusion trails in garnet, suggest that sillimanite and biotite were not stable during peak-$D_2$ in quartz-bearing pelites (see above), matrix sillimanite and biotite in combination with a second generation garnet are very common in these assemblages. Garnet, occurs as inclusion-free subhedral grains, which may mantle garnet. The renewed stability of biotite generally and of the assemblage garnet, sillimanite, biotite, quartz in pelites probably occurred during $D_3$ as this assemblage is common in $D_3$ high-strain zones. It suggests hydration and/or cooling of the terrain which is consistent with the growth of various borosilicates at the onset of $D_3$, due to an influx of a fluid (possibly boron-enriched, Carson, Hand & Dirks, 1995), which is consistent with the stabilization of biotite in $S_4$ zones in the Brattstrand Bluffs (Fitzsimons & Harley, 1991, 1992). The presence of narrow coronas of cordierite and rarely biotite on spinel in the presence of quartz probably represents late syn- to post-$D_3$ cooling (see, e.g. Fitzsimons & Harley, 1992). This is also consistent with thin coronas of garnet on hornblende and orthopyroxene in metabasite lenses on Steinnes Peninsula.

Trails of garnet, orthopyroxene, plagioclase defining $S_4$ and $L_{2a}$ are common within the Grovness enderbite, and corroded grains of garnet are in contact with orthopyroxene. This is a feature that is also common in many of the metabasite layers that occur within part of the stratigraphy. Preliminary geothermobarometry relevant to orthopyroxene–garnet–plagioclase–quartz assemblages (Harley & Green, 1982; Newton & Perkins, 1982; Harley, 1984a, b; Sen
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& Bhattacharya, 1984) based on core compositions of coexisting orthopyroxene–garnet pairs yield peak conditions of ~ 7 kbar and ~ 780 °C. These peak $P$–$T$ conditions compare favourably with previously reported peak conditions reported in the nearby Bolingens Island (e.g. 6 kbar, ~ 760 °C, Motoyoshi, Thost & Hensen, 1991) and the Reinbolt Hills (7 kbar, ~ 800 °C, Nichols & Berry, 1991, fig. 1) to the south and from the Brattstrand Bluffs to the north (6 kbar, ~ 850 °C, Fitzsimons & Harley, 1991, 1992).

4.b. Domains with $S_3$ as dominant foliation

Relevant textures from metabasites with $D_3$ domains include the destabilization of garnet generating orthopyroxene and plagioclase symplectites which are enveloped by a biotite $S_3$ foliation. On Barrtangen such coronas contain fine platelets of symplectic biotite that are aligned parallel with the external $S_3$ foliation. Within basement composite orthogneiss on Fold Island, hornblende, plagioclase, orthopyroxene-bearing metabasite preserves orthopyroxene, plagioclase, biotite ± clinopyroxene symplectites that are strongly elongate in the regional $L_3$ direction. These textures may represent destabilization of garnet, now consumed, via the retrograde reactions:

$$\text{garnet} + \text{quartz} = \text{orthopyroxene} + \text{plagioclase} \quad (4)$$

and

$$\text{garnet} + \text{quartz} + \text{hornblende} = \text{orthopyroxene} + \text{plagioclase} + \text{H}_2\text{O}, \quad (5)$$

indicative of decompression (see, e.g. Harley, 1988; Nichols & Berry, 1991; Clarke & Powell, 1991). On Kolloy and northwest Vikoy islands, coarse-grained biotite selvedges commonly occur at the contact of felsic and mafic sub-units of the basement orthogneiss. The contact is folded by open to tight upright $F_{as}$ folds, and biotite in the selvedge develops a well-developed preferred orientation parallel to the axial plane of the $F_{as}$ folds. Within the biotite selvedge, coarse-grained garnet develops a plagioclase corona that overprints the biotite $S_{as}$ foliation, indicating post-$S_{as}$ garnet breakdown.

Within semi-pelite from the $D_3$ high-strain zone on southern Storness Peninsula, garnet, is locally overgrown by a mantle of inclusion-free garnet, and wrapped by sillimanite aligned in $L_{ax}$. Garnet and sillimanite are subsequently consumed by cordierite that is elongated in tails along $L_{ax}$. Within nearby pelite, trails of garnet, aligned in $L_{ax}$ are wrapped by sillimanite aligned in $L_{ax}$. Garnet is subsequently consumed by $L_{ax}$-parallel, delicate cordierite–spinel symplectites, via reaction (3).

Preliminary $P$–$T$ results using the quantitative average $P$–$T$ approach outlined by Powell & Holland (1994) give very similar results for $D_2$ and $D_3$ domains using the various matrix and corona assemblages. Mineral zoning in garnet, cordierite or spinel is minimal, as are most compositional variations in texturally different generations of these minerals (Carson, unpub. data). Pelite$_4$, composite pelite$_2$, pelite$_1$ and metabasite lenses from $D_4$ and $D_3$ domains record pressures around 4 to 5 kbar, over a range of temperatures between 650–800 °C.

$P$–$T$ estimates on the amphibolite-grade assemblages in $D_4$ mylonites using garnet–biotite geothermometry (Ganguly & Saxena, 1984; Bhattachary a et al. 1992) and average pressure–temperature calculations of Powell & Holland (1994) result in pressures of ~ 3.0 kbar at ~ 550–650 °C.

5. Structural–metamorphic correlations

The similarity of pervasive structural features, orientations and kinematics along the Prydz Bay coast, enables apparently simple correlations of structural and metamorphic events. These correlations are limited to the ‘Neoproterozoic’ terrains and cannot be made to the Archaean Vestfold Hills Block.

In the Bolingen Islands immediately southwest of the Larsemann Hills (Fig. 1), Dirks & Hand (1995) describe dominantly east–west-trending, south-dipping foliations that contain an early, east-plunging lineation related to thrusting in the south Boligen Islands and a later south-plunging lineation related to normal movements in the north Bolingen Islands. This latter zone is the direct westward extension of the $D_4$ high-strain zone in Barrtangen. Thus, $D_2$ and $D_3$ domains similar to those in the Larsemann Hills are readily apparent in the Bolingen Islands, which are also truncated by later pegmatites paralleled by amphibolite grade shear zones. Thost, Hensen & Motoyoshi (1991) report peak conditions of 10 kbar and 980 °C from basement gneiss in the southwest Bolingen Islands, using garnets which Dirks & Hand (1995) describe as being aligned in $L_{ax}$; Thost, Motoyoshi & Hensen (1988) and Thost, Hensen & Motoyoshi (1992) also report a second metamorphic overprint (syn-$D_4$ ?) of ~ 6 kbar and ~ 775 °C from the same area, which is comparable with $P$–$T$ results of ~ 6 kbar and ~ 760 °C obtained from calc-silicate units elsewhere in the Bolingen Islands (Motoyoshi, Thost & Hensen, 1991).

In the Brattstrand Bluffs some 30–35 km northeast of the Larsemann Hills (Fig. 1) a similar metasedimentary sequence is described by Fitzsimons & Harley (1991, 1992). They identified a shallowly south-dipping planar foliation ($S_{sfr\&h}$), truncating a tight to isoclinal folded $S_{stfr\&h}$ foliation which itself contains colinear intrafolial folds. Both foliations, $S_3$ and $S_{sfr\&h}$ preserve a south-plunging mineral extension lineation. These structures record a normal sense of movement (Fitzsimons & Harley, 1991, 1992; Dirks & Carson, unpub. data) and can be correlated with the progressively folded and truncated foliations in $D_3$ normal shear domains from the Larsemann Hills.
while earlier assemblages in low-strain zones and inclusions (called D₃ and D₂ by Fitzsimons & Harley, 1991) may represent D₄ features. Coarse-grained garnets, rapped in the south-plunging lineation (L₂ₙ and L₃ₜ₉ [P(EBH)]) can contain inclusions of prismatic sillimanite (texturally distinct from isoclinally folded fibrolite sillimanite trails) which are aligned sub-parallel to the regional L₂ₙ direction and may represent the preservation of D₃ features in the Brattstrand Bluffs.

Fitzsimons & Harley (1991, 1992) previously report peak P–T estimates of 6 kbar and 850 °C followed by decompression and cooling to 3–4 kbar and 770 °C, conditions that are near identical to those now identified in the Larsemann Hills. Rocks in the Brattstrand Bluffs are also truncated by a late set of north–south pegmatites with sheared margins (D₄ [F(BH)]).

The Rauer Group (Fig. 1) is dominated by a series of progressively overprinting co-linear folds and foliations recording a very constant fold axes and mineral lineation plunging moderately to the southeast (see, e.g. Harley, 1987; Sims et al., 1994), and is associated with thrust kinematics (Dirks et al. 1993; Sims et al. 1994). Most of this deformation may be equivalent to the composite D₃ event recorded from the Larsemann Hills. It is associated with decompression of ~ 2–4 kbar from peak conditions of 7–9 kbar at temperatures in excess of 850 °C (Harley, 1988; Harley & Fitzsimons, 1991), dated at 1000–1050 Ma (zircon U–Pb, Kinny, Black & Sheraton, 1993). Although a new zircon population with an age of ~ 500 Ma is common in a large variety of lithologies in the Rauer Group (Kinney, Black & Sheraton, 1993), the only events definitely linked to the Pan-African are north–south trending pegmatite dykes (500 ± 12 Ma, Kinny, Black & Sheraton, 1993), with amphibolite-grade shear zones at their margins.

The Landing Bluff granitoid units (Fig. 1) contain mafic layers and biotite-rich schlieren that preserve one generation of open upright folds, with fold hinges that are co-linear with the regional D₃ lineation, similar to that recorded from the Larsemann Hills, Brattstrand Bluffs (Fitzsimons & Harley, 1991) and the Bolingen Islands (Dirks & Hand, 1994). These granitoid units are widely accepted as Pan-African post-orogenic intrusive bodies (e.g. Stüwe, Braun & Peer, 1989; Sheraton, Black & McCulloch, 1984; Fitzsimons & Harley, 1991), for which a 493 ± 17 Ma Rb–Sr age (whole rock) has been obtained (Tingey, 1981). Given that the granitoids preserve one generation of F₃ folding, however, suggests that these granitoid bodies intruded late syn-D₃. Additionally, moderately west-dipping planar coarse-grained pegmatites cross-cut these granitoids, again similar to those reported elsewhere in Prydz Bay.

Using the orientation of the dominant linear features and kinematic indicators in high-grade foliations, it is simple to correlate D₂, D₃ and D₄ events along the length of Prydz Bay. Figure 6 summarizes the structural and geochronological correlations in southeastern Prydz Bay.

6. Timing of the structural events
Most of the pervasive deformational and metamorphic features of eastern Prydz Bay outcrops, south of the Archaean craton of the Vestfold Hills, have been previously assumed to have evolved during a tectonic event that occurred ~ 1000 Ma ago (see, e.g. Fitzsimons & Harley, 1991, 1992; Harley et al., 1992; Harley & Fitzsimons, 1991; Thost, Hensen & Motoyoshi, 1991; Sheraton, Black & McCulloch, 1984; Stüwe, Braun & Peer, 1989; Stüwe & Powell, 1989a). While syn-deformational orthogneiss units from the Rauer Group strongly indicate a significant ~ 1000 Ma tectono-thermal event in that area (Kinny, Black & Sheraton, 1993), the effects of the Pan-African orogeny in the Prydz Bay area are previously thought to be limited to low-grade discrete shear zones (e.g. Kinny, Black & Sheraton, 1993; Harley, 1987; Fitzsimons & Harley, 1991) and various post-orogenic intrusive bodies (e.g. Tingey, 1981). Recent geochronology in the Larsemann Hills, however, suggests a major metamorphic imprint of Pan-African age (~ 550 Ma).

Sm–Nd isochron ages from mafic gneiss and Zhong Shan Gneiss near Zhong Shan station (Fig. 2) yield Pan-African ages (whole rock–orthopyroxene–hornblende–plagioclase, 540 ± 75 Ma; whole rock–garnet–K-feldspar, 497 ± 7 Ma, respectively, Zhao et al. 1993), Step-wise evaporation ²⁰⁶Pb/²⁰⁷Pb techniques applied to zircon separates (Kober, 1986, 1987) from the Progress granite on Mirror Peninsula truncates all F₂ fold generations and contains a variable though generally weak biotite foliation that is generally discordant with the pervasive S₂ foliation in the host gneiss. The granitic foliation can be discordant by up to 30° with the external S₂ gneissic foliation. The granite is elongated in the D₃ high-strain zone, however, it is not certain that this is a D₃ deformational effect. At Barrtangen, however, the Progress granite is isoclinally folded and strongly transposed with the pervasive S₂ high-strain zone foliation and K-feldspar and biotite are aligned in L₃ₙ (Fig. 14), whilst on Steiness Peninsula, this unit occurs in the core of the major F₁ₕₙ structure and is folded by that structure. These observations suggest a relative intrusive age for the Progress granite to syn-S₂–pre-S₂.

When considering the above ages for these intrusive units and their relevance for tectonic interpretations, it is important to consider whether they represent real crystallization ages. Zhao et al. (1992) argue that the
Zhong Shan gneiss and argue that it is related to
morphism is difficult to constrain. It is possible that
event, implying that the Larsemann Hills region was
Group. This suggests that the Zhong Shan orthogneiss
evidence for at least D2 being of ~ 1000 Ma age, is the
correlation of D2 with pervasive deformation within

206Pb/207Pb zircon ages obtained from the Progress
granite represent a primary crystallization age because
the reported zircon populations are homogeneous and
display clear age plateaus that suggest no serious
radiogenic Pb loss. An alternative interpretation is
that the 206Pb/207Pb system suffered very serious
radiogenic Pb loss, resulting in effective homogeniza-
tion of the zircon’s radiogenic Pb profile during a Pan-
African thermal event. The ages may, therefore,
simply represent Pan-African reset ages. Sm–Nd ages
obtained from various lithologies (Zhao et al., 1993),
may also represent Pan-African reset ages, given that
recent studies on closure temperatures for Sm–Nd
systems can be as low as 600 °C (see, e.g. Mezger,
Essene & Halliday, 1992). A number geochrono-
logical–structural scenarios are therefore possible
(Fig. 7).

Assuming that the 206Pb/207Pb zircon ages obtained
from the Progress Granite (Zhao et al. 1992) represent
a primary crystallization age, the D4 is constrained to
a structural episode that occurred ~ 550 Ma and D3
to an episode younger or equal to 550 Ma. Zhao et al.
(1993) obtained a 206Pb/207Pb apparent age of
~ 960 ± 6 Ma from four zircon separates from the
Zhong Shan gneiss and argue that it is related to
the syn-D3 ~ 1000 Ma felsic intrusives in the Rauer
Group. This suggests that the Zhong Shan orthogneiss
intruded during a Neoproterozoic tectonothermal
event, implying that the Larsemann Hills region was
subsequently reworked during the Pan-African event,
and the area was subjected to at least two major
tectonothermal events.

If the Pan-African ages simply represent reset ages,
then the real age of the major deformational events in
the Larsemann Hills, D3–D4, and associated met-
armorphism is difficult to constrain. It is possible that
D2–D3 represent the ~ 1000 Ma metamorphic event
widely reported elsewhere in the East Antarctic Shield
(e.g. Sheraton, Black & McCulloch, 1984). Supportive
evidence for at least D3 being of ~ 1000 Ma age, is the
correlation of D3 with pervasive deformation within
the Rauer Group. As discussed above, the Rauer
Group is dominated by thrust kinematics associated
with south–southwest-plunging fold axes and per-
vasive lineation (D4, Sims et al., 1994; D3–4, Harley,
1987), that are co-linear with D4 structures in the
Larsemann Hills. Syn-D4 felsic intrusives from the
Rauer Group (Kinny, Black & Sheraton, 1993)
indicate a Neoproterozoic age for pervasive D4
deformation. If the regional correlation between D3 in
the Larsemann Hills and the Rauer Group is valid,
then this strongly suggests D3 in the Larsemann Hills
is also Neoproterozoic (~ 1000 Ma) in age. Support-
ning this interpretation, the Zhong Shan gneiss,
which intruded pre- to early syn-S2 and records all
episodes of D3 deformation is dated at ~ 960 Ma
(206Pb/207Pb zircon, Zhao et al., 1993). An additional
age constraint comes from Sm–Nd studies of peak
metamorphic garnet in metabasites on Søstrene
Island (~ 25 km west of the Larsemann Hills). These
garnets are aligned in a lineation associated with
thrusting that parallels L2x (Dirks & Hand, 1994) and
yields two point isochron ages of ~ 988 Ma (Hensen,
pers. comm., 1994), further supporting a ~ 1000 Ma
origin for D3.

An additional possibility is that the extensional D3
event is of Pan-African age (~ 550 Ma). The wide-
spread occurrence of ‘post-orogenic’ voluminous
granitoid bodies in the southern Prydz Bay area, often
interpreted as Pan-African intrusives, and the common
accepted view that many of the Pan-African ages
obtained for this region represent a period of extensive
resetting of isotopic systems, suggest a major regional
thermal perturbation at ~ 550 Ma. It is unlikely that
such a widespread high-grade thermal event associated
with extensive felsic magmatic activity was unac-
companied by significant deformation (see, e.g. Sandi-
ford et al. 1991), and given that the last major high-
grade event in the region is D3 suggests that this event
may be the structural manifestation of the Pan-
African thermal anomaly. The syn-D3 nature of Landing Bluff granitoid bodies also supports a Pan-
African age for D3.

The age of D4 is well constrained by the age of the
planar pegmatites which are deformed by D4 exten-
sional mylonites. Similar planar pegmatites from the
Rauer Group constrained to the Pan-African (500 ±
12 Ma, U–Pb SHRIMP techniques, Kinny, Black &
Sheraton, 1993), provide a maximum age for D4.
These structures may represent the waning effects of
the D4 high-grade extension event.

7. Conclusions

The meta-sedimentary sequence of the Larsemann
Hills preserves a workable stratigraphy that can be
recognized throughout the area. Metabasite inter-
calations restricted to specific lithological units suggest
a metavolcanic derivation, possibly in a mid-Pro-
terozoic extensional basin associated with volcanics. The opening of this basin may have coincided with an extensional event and the emplacement of voluminous dyke generations in the Vestfold Hills Archaean Block at 1370 and 1250 Ma (Dirks et al. 1993; Hoek, Dirks & Passchier, 1992; Lanyon, Black & Seitz, 1993), which is consistent with both Sm–Nd CHUR model ages of 1300–1600 Ma for most of these sediments in the Prydz Bay area (Sheraton, Black & McCulloch, 1984).

The high-grade structures in the Larsemann Hills can be subdivided in two groups, D2 and D3, each defined by a unique lineation direction and shear sense (Figs 3, 8). Each event is composite in nature and comprises high- and low-strain domains characterized by similar fold sequences. An important implication of this is that the gneissic layering in different parts of the Larsemann Hills did not necessarily develop all at the same time. S2 dominated zones may have developed 500 Ma later than S3 dominated zones, even though S2 and S3 appear identical in outcrop except for the mineral lineation direction they contain. The structural geometry of the Larsemann Hills is dominated by a major D3 high-strain zone to the south, and a zone of diffuse D3 reactivation to the north and with a low D3 strain pocket in between, where S2 structures are dominant. Assemblages in garnet–orthopyroxene-bearing tonalitic rocks record peak P–T conditions of ~7 kbar and ~780 °C early during D2 followed by decompression and cooling to ~4.5 kbar at ~700 °C with progressive D3 thrusting. During D2 normal shearing, peak conditions may have been as high as 800 °C at ~4.5 kbar resulting in a second generation of re-equilibration textures and symplectites on remnant D2 assemblages, mainly in aluminous pelites. During D3 the growth of hydrous phases, especially biotite is important, as is the emplacement of a number of porphyritic granites.

To explain the Proterozoic decompression textures and the thermal regime prevailing in Prydz Bay, several workers have suggested tectonic models involving crustal thickening, delamination of the lower crust and subsequent extensional collapse (e.g. Fitzsimons & Harley, 1991; Thost, Hensen & Motoyoshi, 1991; Nichols & Berry, 1991) even though structural–kinematic data to prove or disprove these models are generally lacking. It is almost inevitably assumed that this event occurred ~1000 Ma ago, whilst the 550 Ma events are equated with a passive thermal overprint and limited deformation (e.g. Stüwe & Powell, 1989a; Fitzsimons & Harley, 1991). Although our structural–kinematic data support a continuous D3 thickening, D3 extension thinning model, resulting in decompression textures, problems exist if one tries to match the structures with available age data. One possibility explored here is that D2 and D3 are separated by ~500 Ma. In this case, the superposition of two independent P–T loops, one at ~1000 Ma and a later one at slightly lower pressures at ~550 Ma, may give the appearance of a smooth retrograde P–T path (Dirks, Carson & Wilson, 1993). In fact it appears that most uplift in the Larsemann Hills occurred during D2 and that decompression is a result of thrust-related uplift. D3 may have been purely extensional in nature affecting a long since stabilized normal thickness crust.

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