RE-EXAMINATION OF THE METAMORPHIC EVOLUTION OF
THE LARSEMANN HILLS, EAST ANTARCTICA

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Abstract: A study of the high grade metamorphic rocks in the Larsemann Hills, East Antarctica, shows that besides the dominant lower pressure metamorphism in the area, earlier granulate facies relics are also present. The lower pressure metamorphism experienced four episodes (M21 to M24), and all can be discerned from both the felsic and mafic rocks. 207Pb/206Pb ages of zircons from a syenitic orthogneiss formed in M21 concentrate at c. 550 Ma. Temperature of the episodes and pressure of some have been estimated. A qualitative P-T path of the lower pressure metamorphism (M2) is sketched, M21 → M22 is decompressional heating, M22 → M23 isothermal decompression (ITD) and M23 → M24 isobaric cooling (IBC). A rare assemblage grandidierite–kornerupine–tourmaline is discovered in blue gneiss in Stornes Peninsula, and the borosilicate crystallization corresponded mainly with the M23 substage. The late greenschist facies overprint (M3) on high grade rocks is related to shear zones which may exist between the Bollingen Islands and the western Larsemann Hills. It is this presumable boundary which might be accompanied by the ultramafic rock of olivine-bearing mafic granulate that makes two structural domains, the west with early high pressure relics, while the east without, but with higher geothermal gradient than the west in later high grade metamorphism.

Key words: earlier granulate facies relics, incongruent melting, borosilicate, structural domains, Larsemann Hills

Introduction

Prydz Bay comprises one part of the generally recognized late Proterozoic metamorphic belt. The high grade metamorphism is ascribed to the c. 1000 Ma event (Stüwe et al., 1989; Stüwe and Powell, 1989a; Sheraton et al., 1984). This data is acquired by geological comparison with neighbouring Rayner Complex, Enderby Land (Black et al., 1987) and Rauer Group (Harley, 1988; Fitzsimons and Harley, 1991).

The Larsemann Hills are located in the centre of Prydz Bay (Fig. 1). The study of metamorphic geology on the outcrops (Stüwe and Powell, 1989a) shows that the high grade event is lower pressure granulate facies metamorphism occurring c. 1000 Ma. No earlier metamorphic relics have been found, and a model of perturbation in the asthenosphere is put forward.

According to our observation on petrographic and reaction features, earlier granulate facies relics have been discerned, and asymmetry exists on both sides of the Larsemann Hills. Furthermore, a c. 550 Ma age for the upper limit of the lower pressure granulate facies metamorphism is obtained.

Our study is based on detailed petrographic observation on the felsic and mafic rocks, with reference to some borosilicate minerals, being combined with appropriate pressure and temperature calculation.

Geological and Petrological Outline

The Larsemann Hills sequence consists essentially of four lithological units: (1) Layered paragneisses comprise a major part of the outcrops and can be subdivided into banded gneiss, migmatitic gneiss, leucogneiss and felsic gneiss: between them concordant or transitional contacts are observed (Fig. 2). Garnet, sillimanite, biotite, K-feldspar and cordierite occur regionally. Orthopyroxene poikiloblasts can be found locally. Garnet or orthopyroxene segregations are sometimes present. (2) Mt–Ilm–Sp–Sl–Crd gneiss or “blue gneiss” as defined by Stüwe et al. (1989), is heterogeneous in texture. Garnet is present or not. In the Stornes Peninsula a rare grandidierite–kornerupine–tourmaline assemblage of borosilicate is found. Cordierite or sillimanite segregations have also been found. (3) Syenitic orthogneiss or “pink granite” by Stüwe et al. (1989), crops out in northern Mirror Peninsula. Besides its dominant granitic feature, relic garnet and biotite (Fig. 3a) and biotite—sillimanite coronas around magnetite and/or ilmenite (Fig.
biotite inclusion association in K-feldspar (Fig. 3a). The biotite has the highest TiO$_2$ content (6.57 wt%) in the area (Table 2). Both present as inclusions, the cordierite is magnesio-richest (Mg/(Mg+Fe) = 0.84) (Table 2), while the plagioclase (An$_{13.7}$) is relatively albitic to the later ones (An$_{15.43}$).

In the mafic rocks the M$_1$ event is shown by the inclusion association orthopyroxene–clinopyroxene–plagioclase (Fig. 3e, g and i) in hornblende. The orthopyroxene is unique in that it has the highest Al$_2$O$_3$ content (up to 2.49 wt%) (Table 2) in mafic rocks in the area, and the plagioclase (An$_{40}$) is most albitic in the rock type.

The second metamorphic event (M$_2$)

Four substages can be discerned in this event of metamorphism through the assemblage evolution of both the felsic and mafic rocks.

The first substage (M$_{21}$) in the felsic rocks is characterized by the preferred orientation of biotite–sillimanite–feldspars–quartz ± ilmenite ± magnetite ± garnet. The syenitic orthogneiss is probably formed at this stage because the garnet of M$_2$ and M$_3$ (see next section) can be observed overgrown in part of the orthogneiss (Fig. 2).

In the mafic rocks the M$_{21}$ hornblende grains form a preferred orientation, as is the case for the Brattstrand Bluffs coastline (Fitzsimons and Harley, 1991). The assemblage at this stage is hornblende–plagioclase (An$_{70}$) ± magnetite ± ilmenite.

The second substage (M$_{22}$) is distinguished in the felsic rocks as the garnet–cordierite–sillimanite–K-feldspar–spinel assemblage, which is ascribed to incongruent fluid–absent melting reactions (Grant, 1985) as the following:

\begin{align*}
\text{biotite + sillimanite + quartz} & = \text{garnet + K-feldspar + melt} \quad (1) \\
\text{biotite + sillimanite + quartz} & = \text{garnet + cordierite + K-feldspar + melt} \quad (2)
\end{align*}

The anhedral garnet rich in the inclusion association biotite–sillimanite–quartz is formed through the above reactions. This garnet is generally fine grained and aligned along the foliation. The geologic characteristic of the Larsemann Hills is to some degree similar to that of the Brattstrand Bluffs coastline (Fitzsimons and Harley, 1991), and the garnet may equate here to the garnet I in the Brattstrand Bluffs area described by Fitzsimons and Harley (1991).

The spinel–cordierite symplectite and spinel–quartz intergrowths are generally considered to reflect reactions like:

\begin{align*}
\text{garnet + sillimanite} & = \text{cordierite + spinel} \quad (3) \\
\text{garnet + sillimanite} & = \text{spinel + quartz} \quad (4)
\end{align*}

which are indicative of decompression (Bohlen et al., 1986, Fig. 2; Clarke and Powell, 1991).

The spinel was considered as one phase of the peak assemblage (Stüwe and Powell, 1989a), while in the Rauer
### Table 1. Critical assemblages of the Larsemann Hills lithologies and their relationship with metamorphic events.

<table>
<thead>
<tr>
<th>Rock type</th>
<th>M1 assemblage</th>
<th>M2 assemblage</th>
<th>M3 assemblage</th>
<th>M4 assemblage</th>
<th>M5 assemblage</th>
<th>M6 assemblage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Blue gneiss</td>
<td>SIl - Br - Pl - Kfs</td>
<td>Br - Sil - Pl - Ilm</td>
<td>SiL - SpL - Crd</td>
<td>Gtd - Kms - GtCrd</td>
<td>Trm - QzS - Kfs</td>
<td>Chl - Ms</td>
</tr>
<tr>
<td></td>
<td>- Qx + Ilm + Mt</td>
<td>Mt - Or - Qse</td>
<td>SiL - Crd</td>
<td>Br - SiL</td>
<td>SIl - Crd</td>
<td>Crd - Cld</td>
</tr>
<tr>
<td></td>
<td>+ GtS + Trm</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Layered gneiss</td>
<td>GtS - Br - SIl - Kfs</td>
<td>Br - Kfs - Pl - SIl</td>
<td>GtS - SIl - Kfs - Crd</td>
<td>GtS - Kfs - Crd</td>
<td>Br - SIl</td>
<td>Chl - Ms - Ep</td>
</tr>
<tr>
<td></td>
<td>- Pt - QxS + Ilm</td>
<td>- QxS - Ilm + Gt</td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>Synthetic orthogneiss</td>
<td>GtS - Br - Pl - Kfs</td>
<td>Br - Kfs - Qxs - Pl</td>
<td>GtS - Kfs + SIl</td>
<td>GtS - Kfs - Crd</td>
<td>Br - M6 + SIl</td>
<td>Chl - Ms</td>
</tr>
<tr>
<td></td>
<td>- Ilm - Mt + Gt</td>
<td>- GtC + Sps + SPl</td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Mafic and Ultramafic</td>
<td>OpS - Cpx - Pla</td>
<td>Hbl - Plz - Ilm - Mt</td>
<td>OpS - Cpx + Pl2 + Hbl</td>
<td>OpS - Cpx - Pl1</td>
<td>Hbl - Pl</td>
<td>Tr - Cal - Chl</td>
</tr>
<tr>
<td>granulites</td>
<td></td>
<td></td>
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</tbody>
</table>

Mineral abbreviations are after Kretz (1983). The boldface refers to the dominant assemblage in the rock. The subscript of Pl refers to the anorhinite content of plagioclase from sample 2020.

### Table 2. Representative microprobe analyses of the minerals in the felsic and mafic rocks from the Larsemann Hills.

<table>
<thead>
<tr>
<th>Mineral</th>
<th>M1</th>
<th>M2</th>
<th>M3</th>
<th>M4</th>
<th>M5</th>
<th>M6</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO2</td>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>TiO2</td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>Al2O3</td>
<td></td>
<td></td>
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<tr>
<td>Cr2O3</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>FeO*</td>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>MgO</td>
<td></td>
<td></td>
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<tr>
<td>CaO</td>
<td></td>
<td></td>
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<td></td>
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<td></td>
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<tr>
<td>Total</td>
<td></td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>

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*Total Fe as FeO, Gtt(c) and Gtt(r) refer to the core and rim, respectively.
Fig. 3. Photomicrographs of the critical assemblages and textures in the Larsemann Hills. The width of the picture f, h is 1.0 mm; a, b, c, g 0.5 mm; d, e, i 0.25 mm. (a) relict garnet and biotite (M$^1$) in late K-feldspar; (b) spinel (M$^2$) along the biotite (M$^2$) cleavage, subidioblastic sillimanite (M$^2$) is included in spinel; (c) ilmenite is mantled by the sillimanite–biotite corona (M$^2$); (d) grandidierite included in cordierite between them is the quartz corona; (e) orthopyroxene and clinopyroxene (M$^1$) included in hornblende (M$^2$); (f) orthopyroxene (M$^2$) grown on the cleavage of plagioclase and the boundary between Pl and Hbl (M$^1$); (g) clinopyroxene (M$^1$) and hornblende (M$^2$) inclusions in orthopyroxene (M$^2$); (h) hornblende (M$^1$) inclusion in plagioclase (M$^2$); (i) orthopyroxene and plagioclase rim (M$^2$) around plagioclase core (M$^1$) and between the hornblende and clinopyroxene (M$^2$).
Group and Brattstrand Bluffs coastline it was considered a retrograde one (Harley, 1988; Fitzsimons and Harley, 1991). The former is preferred in the Larsemann Hills according to the next description (about $M_2^3$) and the spinel formation reaction is possible like this:

$$Fe^{2+}Fe^{3+} + (Mg,Fe)_6Al_2$$

$$Mt \quad Bt$$

$$= \{Fe^{2+},Mg\}Al_2 + (Mg,Fe)_6\{Al,Fe^{3+}\}_2$$

$$Spl \quad Bt$$

The substitutions involved in reaction (5) are $Fe^{2+}Mg_{1-}$, $Fe^{3+}Al_1$ between magnetite, biotite and spinel. When the reaction shifts from left to right, we can see the increase of both the $Fe^{2+}/(Mg + Fe^{3+})$ and the $Fe^{3+}$ content of biotite. Titanium occurs generally as a substitutional element in magnetite, sometimes ilmenite is present, therefore the biotite on the right side should have a higher Ti content. Calculation on thermodynamic data (Lin et al., 1985) shows that the high entropy side is on the right of the reaction equation. All these are concordant with the increase of temperature when the reaction occurs.

The transition of magnetite along the rim into spinel is the manifestation of the reaction. As the transition degree increases, we can find the sparsely distributed spinel dots along the magnetite rim, to the spinel corona, even the complete spinel (Fig. 3b). It is this corona of spinel around magnetite and/or ilmenite that results in the interpretation of a retrograde history by some petrologists. Furthermore, unlike the reactions (1) or (2), the reaction (5) cannot form garnet and the sillimanite is not a reactant. The euhedral sillimanite grains included in spinel (Fig. 3b) and the absence of garnet in some biotite–magnetite rich rocks (part of the blue gneiss) in Mirror Peninsula (Fig. 2) and Stornes Peninsula is quite consistent with this point.

As to the pressure change of reaction (5), we can find spinel along the fracture cleavage in the biotite (Fig. 3b), which may imply that decomposition occurred when the spinel was formed.

From the above discussion we can say, the reaction (5) may take place in the decompressional heating process.

Just like the $M_2^1$ biotite retained in $M_2^2$ in the felsic rocks, the $M_2^1$ hornblende can be preserved in $M_2^2$ in the mafic rocks, perhaps with some adjustment in composition (Table 2), and tend to be in equilibrium with the new grown $M_2^2$ association: orthopyroxene–clinopyroxene–hornblende–plagioclase (Fig. 3f, i). If the hornblende ($M_2^1$) is slightly transformed, earlier inclusion of $M_1$ can be preserved (Fig. 3c), and the orthopyroxene distributed along the cleavage of plagioclase and the boundary between plagioclase and hornblende (Fig. 3f), as is similar to the spinel growth along the biotite cleavage in felsic rocks (Fig. 3b), indicating presumably a decompressional process from $M_1^2$ to $M_2^2$. When the hornblende is nearly consumed, it may occur as inclusions in orthopyroxene and plagioclase ($An_{92}$) (Fig. 3g, h). Clinopyroxene of $M_1$ occasionally survived due to its relative immobility feature in the $M_2^2$ orthopyroxene (Fig. 3g).

Combining the assemblage features of $M_2^1$ and $M_2^2$ for both the felsic and mafic rocks, it can be seen that the preservation of much part of $M_2^1$ in $M_2^2$ is obvious, suggesting that it is a continuous process from $M_2^1$ to $M_2^2$. The third substage ($M_2^2$) is typical of the porphyroblast segregations, such as garnet, orthopyroxene, cordierite megacrystals mantled by essentially K-feldspar in the felsic rocks. According to Fitzsimons and Harley (1991), subidioblastic garnet (II) in the Brattstrand Bluffs coastline is inclusion-free other than quartz, as is the general case in the Larsemann Hills, but in a euhedral garnet segregation (c. 15 mm in diameter) collected from Friendship Peak in Mirror Peninsula a small amount of biotite, rutile, spinel, cordierite inclusions has been observed, indicating that the idiomorphic garnet in the Larsemann Hills formed after the assemblage ($M_2^2$) spinel–cordierite–garnet (I). The anhedral garnet (I) never contains spinel or cordierite. The formation mechanism of orthopyroxene segregation is similar to that of the garnet, through the following incongruent reactions (Stüwe and Powell, 1989b)

$$\text{biotite + cordierite + quartz} = \text{garnet + K-feldspar + H}_2\text{O}$$

$$\text{biotite + quartz} = \text{orthopyroxene + K-feldspar + H}_2\text{O}$$

The cordierite corona around spinel and the plagioclase corona around garnet might correspond with this period:

$$\text{garnet + sillimanite + quartz} = \text{plagioclase}$$

which indicates decompression from $M_2^2$ to $M_2^3$ (Essene, 1989, Fig. 2).

In mafic rocks the $M_3^1$ assemblage generally occurs as the orthopyroxene and plagioclase coronas on hornblende, and in the extreme case the coarse assemblage orthopyroxene–clinopyroxene–plagioclase ($An_{85}$) formed in the differentiation veins or lenses. The fourth substage ($M_2^4$) is characterized by reappearance of biotite as a stable phase in the felsic rocks. The sillimanite–biotite corona around the magnetite and ilmenite (Fig. 3c) and the green biotite adjacent to the garnet segregations indicate the introduction of a fluid and/or isobaric cooling towards the stable geotherm (Fitzsimons and Harley, 1991; Stüwe and Powell, 1989a).

In the mafic rocks, the substage is slightly manifested by the association of green hornblende–plagioclase around earlier pyroxenes or along the cleavage of earlier minerals, coinciding perhaps with the biotite–sillimanite corona in felsic gneisses.

From both the petrography and reaction textures in the felsic and mafic rocks, it is deduced that except the independent history of $M_1$, the $M_2^1$ to $M_2^4$ story is a continuous process.

**Borosilicate minerals in the Stornes Peninsula**

The borosilicate association grandidierite–kornerrupine–tourmaline has been observed and is confined to the blue gneiss. The grandidierite is identified by the optical properties (E. S. Grew, personal communication; Fig. 3d) and the
electron microprobe analysis (Table 2). It is the first occurrence of the mineral in Antarctica (E. S. Grew, personal communication). The kornerupine is identified by both composition (Table 2) and X-ray powder diffraction data: 2.996(77), 2.607(100), 2.094(45), 1.762(11). The "radial tourmaline" in the Stornes Peninsula described by Stüwe et al. (1989, Fig. 3c) may be the kornerupine.

Generally, the three borosilicate minerals are not in contact to each other (Lonker, 1988). The scatter of the minerals implies the presence of B₂O₃ potential gradients (Grew et al., 1990). In the Stornes Peninsula, however, we can find the contacts between the borosilicates: tourmaline has two forms, one included in kornerupine, the other replaces the latter, while granddiditerite is included in kornerupine. Combined with the petrographic feature of other minerals we have the following crystallization order:

\[ \text{Trm}_1 \rightarrow \text{Sil} \rightarrow \text{Bt} \rightarrow \text{Sil} \rightarrow \text{Spl} \rightarrow \text{Gdd} \rightarrow \text{Krn} \rightarrow \text{Crd} \rightarrow \text{Trm}_2 \rightarrow \text{Gbs} \rightarrow \text{M}_2^1 \rightarrow \text{M}_2^2 \rightarrow \text{M}_2^3 \rightarrow \text{M}_2^4 \rightarrow \text{M}_2^5 \rightarrow \text{M}_2^6 \]

Spinel and sillimanite inclusions occur in granddiditerite. This is consistent with the conclusion that spinel was formed earlier in the metamorphic history. The sequence Gdd \rightarrow Krn \rightarrow Trm corresponded with the diminishing ratio of B₂O₃/H₂O (Manning and Pichavant, 1983), and is consistent with the increasing fluid (water) activity after the peak metamorphism.

On the petrographic aspect, the coarse segregations of cordierite and garnet (M₂³) are quite similar in that both are mantled with K-feldspar, and the radial kornerupine is also closely related to the K-feldspar segregation, meanwhile most of the coarse minerals mentioned above occur heterogeneously and not foliation-bound, suggesting they formed in the same period, i.e. M₂³ substage. The tourmaline (II) and gibbsite in blue gneiss are presumably concordant with the biotite and sillimanite stage in felsic gneisses, both implying the increasing water activity and cooling from M₂³ to M₂⁶. The granddiditerite is closely related with anatexis in regional metamorphism (Lonker, 1988; Grew et al., 1990), which is not inconsistent with the above interpretation.

**Implication of the third metamorphic event (M₂³)**

The M₂³ is manifested by the greenschist facies overprint on high grade rocks (Table 1). The intensity of development of M₂³ in the Larsemann Hills appears to increase from Mirror Peninsula in the east to Stornes Peninsula in the west, coinciding with the case of deformation. It suggests that fracture or shear zones may exist west of Stornes Peninsula, as the greenschist facies overprint is associated with the development of mylonite zones (Fitzsimons and Harley, 1991). In fact, late shear zones have been discovered in the Bolingen Islands (Fig. 1), west of Larsemann Hills (Stüwe et al., 1989; Thost et al., 1988).

**Temperature and Pressure Conditions**

As garnet is absent from mafic granulites, many geobarometers cannot be used. The temperature estimate is based mainly on the biotite–garnet pair for the felsic rocks and the two pyroxene thermometry for mafic granulites (Table 3).

The M₁ biotite and garnet are separated and included in feldspar (Fig. 3a). A temperature of 888°C by biotite–garnet thermometer is got. This extraordinary high value may be resulted from the extremely high TiO₂ content in biotite. The two pyroxene pair gives 850°C (Table 3). Although the estimates are very equivocal, the high temperature feature of M₁ can be implied by some mineral characteristics such as the dark red biotite has the high TiO₂ content, and the highest Al₂O₃ content of orthopyroxene in the mafic rocks (Harley, 1988). From the same sample the orthopyroxenes of later stages give the Al₂O₃ content lower than 1.0 wt%.

The geobarometry of Grt–Silt–PI–Qtz assemblage (Essene, 1989) gives c. 9.0 kbar at the reference temperature 850°C for M₁. Although the geobarometry is very sensitive to the CaO content of garnet, the pressure estimate is very weakly defined, the chemistry of M₁ minerals such as the magnesiocumic cordierite in the felsic rocks and the most albitic plagioclase both in the mafic and felsic rocks is in sharp contrast with that of the lower pressure metamorphism (Holdaway and Lee, 1977; Spear et al., 1991), suggesting at least the medium pressure feature of M₁ event.

M₁ temperature (573 ± 20°C) is obtained through biotite–garnet pair in which the garnet rim is corroded into sillimanite, quartz and magnetite (M₂⁶), and then mantled by cordierite (M₂³).

M₂³ temperature is 777 ± 30°C by the Grt–Bt pair and 740–815°C by the Opx–Cpx pair (Table 3), while the garnet–hercynitic spinel geobarometry gives the pressure 4.3 kbar (Bohlen et al., 1986), which are consistent with that for assemblage involving garnet–sillimanite–spinel–cordierite–plagioclase–quartz: 750°C, 4.5 kbar (Stüwe and Powell, 1989a).

M₂⁶ is characterized by the occurrence of coarse euhedral garnet in which the composition zoning is detected, X₉₇ (= Mg/(Mg+Fe)) increases from core to rim. Pairing the garnet

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**Table 3. Thermometry of biotite–garnet pair in the felsic rocks and the clinopyroxene–orthopyroxene pair in the mafic rocks.**

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>M₁</th>
<th>M₂³</th>
<th>M₂⁶</th>
<th>M₃</th>
<th>M₄</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bt–Grt</td>
<td>22103</td>
<td>20602</td>
<td>22103</td>
<td>22103</td>
<td></td>
</tr>
<tr>
<td>Sample No.</td>
<td>20201</td>
<td>20201</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Opx–Cpx</td>
<td>850</td>
<td>740–815</td>
<td></td>
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<td></td>
</tr>
</tbody>
</table>

Notes: The estimates from biotite–garnet pair are after Ferry and Spear (1978), Ghent and Stout (1981) and Hodges and Spear (1982), respectively. The pyroxene thermometry is of Lindsley (1983).
with the matrix biotite, we have the temperature 733 → 764 → 772 (±30)°C from core to rim, suggesting the garnet crystallized in the slightly heating process (M2), although the absolute value of temperature is ambiguous. As to the grandidentite formed in this stage, the low temperature synthesis limit of Fe-bearing grandidentite (Fe/(Fe + Mg) = 0.1) is 700°C at 1.0 kbar fluid pressure at the QFM buffer (Olesch and Seifert, 1976), which is not in contradiction to the above estimate.

M24 temperature is estimated using the segregation garnet (rim)–green biotite (adjacent) pair. The value 634 ± 25°C may represent the closure temperature (Spear, 1989), and the real temperature may be higher than this.

As the garnet–biotite thermometer in the granulite facies and the two-pyroxene thermometer are not quite reliable, the above estimates are not unambiguous and only qualitative. Combined with the petrographic and textural constraints, we have the sketchy outline of P–T path of the M2 metamorphism (Fig. 4). The M1 is the independent earlier metamorphic record and not included in the path. M21–M22 is the decompressional heating section, M22–M23 isothermal decompression (ITD) and M23–M24 isobaric cooling (IBC).

Fig. 4. Sketch diagram of the P-T path from M21 to M24. The M22 location is from Stüwe and Powell (1989a).

Geochronology

It is generally considered that most part of Prydz Bay gneiss is asecribed to the c. 1000 Ma event through comparison with the Rayner Complex (Black et al., 1987; Stüwe and Powell, 1989a) and Rauer Group (Harley, 1988; Fitzsimons and Harley, 1991) from the latter the U-Pb data 1070–1030 Ma have been obtained.

In fact, the Rauer Group and the Larsemann Hills have quite different lithological characteristics. The former is of Archaean or basement feature while the latter Proterozoic or cover sequence feature (Fitzsimons and Harley, 1991). The metamorphic conditions are also different in the two areas. Therefore the Larsemann Hills may have experienced quite a different history from that of the Rauer Group. An independent age should be got from the Larsemann Hills.

As the Rb-Sr, Ar-Ar and K-Ar systems can be easily reset in later thermal events, the Pb-Pb single zircon stepwise evaporation technique is chosen (Zhao et al., 1992). The samples are collected from the syenitic orthogneiss in Mirror Peninsula. Although some relics can be preserved in the granite, it is essentially magmatic in crystallization, with subsequent M22–M24 overprints to some extent. Measurement shows that the substantial Pb loss of zircon occurs only in the rim. The cores yield stable 207Pb/206Pb ages of 547 ± 9 Ma and 556 ± 7 Ma, which correspond with the time of M21 metamorphism in the Larsemann Hills.

The Sm-Nd internal isochrons give 540 ± 75 (2σ) Ma in the mafic rock (whole rock–Opx–Hbl–Pl, formed in M22) and 497 ± 7 (2σ) Ma in the felsic rock (whole rock–Grt–Kfs, formed in M23) (Zhao, unpublished data). They are quite consistent with the 207Pb/206Pb age (M21) and the interpretation on the metemorphism evolution in the present study.

The 40Ar/39Ar plateau age of biotite concentrating at 494 Ma (Zhao et al., 1992) is considered to represent the cooling time of the biotite, constraining the upper limit of M24. Therefore the M23–M24 gap is only 53–61 m.y. In the short history from M21 to M24 is consistent with the incompleteness of the reactions in metamorphism and preservation of multi-stage textures in the metamorphic rocks.

Correlation with Other Outcrops in the Prydz Bay

Through geological (especially with respect to metamorphism) comparison between the different areas in Prydz Bay, several models have been put forward in interpreting the history of the region. The thermal regime of the crust in the Larsemann Hills was controlled by a perturbation in the asthenosphere, with magma invasion of the crust and some symmetry around the Larsemann Hills exists (Stüwe and Powell, 1989a) or the Larsemann Hills coincide with a synformal core, and the Brattstrand Bluffs coastline with an antiformal core (Fitzsimons and Harley, 1991). It seems that the thermal gradient of the high grade gneiss is lower in west of Larsemann Hills than that in the east except the Rauer Group (105–114°C/kbar: 7.0–9.0 kbar, 800–850°C). The Munro Kerr Mountains far west is 106°C/ kbar (7.0–8.0 kbar, 750–850°C) (Stüwe and Powell, 1989a), the Bologen Islands in the west 129°C/kbar (c. 6.0 kbar at 775°C) (Motoyoshi et al., 1991), while the Brattstrand Bluffs coastline in the east gives the condition 142°C/kbar (c. 6.0 kbar at 850°C) (Fitzsimons and Harley, 1991), the Larsemann Hills in itself is 167°C/kbar (4.5 kbar, 750°C) (Stüwe and Powell, 1989a).

On the other hand, the rock types on both sides of the Larsemann Hills appear different. The paragneisses in the west are dominant in felsic gneiss (Stüwe and Powell, 1989a), while in the east abundant in cordierite and ilmenite or cordierite–spinel– sillimanite gneiss (Fitzsimons and Harley, 1991). In addition, the ultramafic rock, namely, the olivine-bearing mafic granulite has been discovered in Bologen Islands (Motoyoshi et al., 1991), west of Larsemann Hills, and western Larsemann Hills, but not in the eastern Larsemann Hills or the Brattstrand Bluffs coastline in the east (Fitzsimons and Harley, 1991). While in the Sestrene Island, earlier high pressure relics (c. 10 kbar at 980°C) has been discovered in mafic granulites (Thost et al., 1991).

The above suggest that both sides of the Larsemann Hills
are presumably of different structural domains. The west underwent earlier high pressure metamorphism but lower in geothermal gradient in later high grade metamorphism. The east is essentially manifested by the late (c. 550 Ma) high grade event. It is deduced that the shear zones in the Bolingen Islands and (?) the gap area between the islands and the Larsemann Hills might represent the boundary between the two domains, together with the boundary is the presence of the ultramafic rock, i.e. the olivine-bearing mafic granulite. The ultramafics between the relative high pressure and lower pressure metamorphism areas in the Lützow–Holm Bay region may represent the position of a subduction zone (Hiroi et al., 1991).

Conclusion
On the basis of petrographic observation and pressure–temperature estimates, an earlier granulite facies event in the Larsemann Hills is discerned, which may be the high temperature, at least medium pressure type metamorphism.

The later metamorphism, M₂ to M₄, can be discriminated from both the felsic and mafic rocks. M₂ with amphibole facies assemblages comprises the dominant foliation, the temperature is 573 ± 20°C. The syenitic orthogneiss formed at this stage gives the zircon ²⁰⁶Pb/²⁰⁶Pb age c. 550 Ma.

Through incongruent melting in felsic rocks and dehydration reaction in mafic rocks, M₂ → M₄ assemblages give the condition 4.5 kbar at 750°C. The subsequent isothermal decompression was responsible for the presence of M₃ assemblage in which the unique porphyroblast segregations are formed, temperature 772 ± 30°C. Then in M₄ the geothermal gradient tended to become stable, and isobaric cooling took place; a temperature record for the period is 634 ± 25°C.

A rare assemblage granddierite–kornerupine–tourmaline is found in Stornes Peninsula. The crystallization of the borosilicate minerals is mainly in the M₃ substage. The M₃ is shown by greenschist facies overprint on high grade rocks and is related with shear zones that may exist between the Bolingen Islands and Larsemann Hills. It is this probable boundary which might be accompanied by the ultramafic rock of olivine-bearing mafic granulite that makes two structural domains, the west with early high pressure relics, while the east without, but with higher geothermal gradient than the former in later high grade metamorphism.

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