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BORON AND BERYLLIUM MINERALS IN GRANULITE-FACIES PEGMATITES AND IMPLICATIONS OF BERYLLIUM PEGMATITES FOR THE ORIGIN AND EVOLUTION OF THE ARCHEAN NAPIER COMPLEX OF EAST ANTARCTICA

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Abstract: ČERNÝ included in his "abyssal class" those pegmatites originating by partial melting under granulite-facies conditions. Enrichment of abyssal pegmatites in rare elements such as Li, Be, B, Rb, and Cs is exceptional. Consequently, the presence of B minerals in pegmatites in granulite-facies rocks at Cape Andrahomana, SE Madagascar; in Rogaland, SW Norway; at Homagama, Sri Lanka; and in the Larsemann Hills, East Antarctica, have considerable geochemical interest. The host rocks are Al-rich paragneisses containing cordierite, almandine-pyrope, and sillimanite. The suite of B minerals is distinct from that in pegmatites associated with amphibolite-facies terrains in that tourmaline is not always present; instead, three phases related to sillimanite (grandidierite, werdingite, and boralsilite), dumortierite and prismatine are characteristic. Boron enrichment results from remobilization of B present in the host rock. Hence, the B pegmatites have retained the B signature characteristic of pelitic rocks, which is typically lost during granulite-facies metamorphism.

Abyssal pegmatites containing Be minerals are found in the Archean Napier Complex at two localities in Casey Bay, Enderby Land, East Antarctica. The Be silicates surinamite and beryllian sapphirine, and the Be oxides chrysoberyl and musgravite are present instead of beryl. The host rocks are Mg-Al-rich gneisses containing sapphirine. The Be pegmatites were emplaced on the prograde path of increasing pressure and temperature and recrystallized under ultra-high-temperature conditions during a single deformational-metamorphic episode at 2475 ± 25 Ma. Be could have originated in the Mg-Al-rich host rocks inasmuch as Be concentrations are reported from Mg-Al-rich rocks elsewhere in the Napier Complex and at other localities worldwide. This scenario for the Late Archean event does not require the prolonged (400 Ma) residence at depth proposed by other investigators; it is consistent with HENSEN's and MOTOYOSHI's model of insertion of anorthositic rocks above the rocks presently exposed, which resulted in increased load as well as heat.

key words: pegmatite, boron, beryllium, Antarctica, Archean/Archaean

1. Introduction

ČERNÝ (1982a, b, 1992) included in his "abyssal class" those pegmatites originating by

partial melting under granulite-facies conditions (see also GINZBURG *et al.*, 1979). Abyssal pegmatites are rarely mineralized, particularly in beryllium or boron, which is not surprising given the low concentrations of these elements in granulite-facies complexes. In unmetamorphosed pelitic sediments, the metamorphosed equivalents of which commonly host pegmatites, beryllium is more sparse than boron (*e.g.*, 1-18 ppm Be, 30-250 ppm B, HÖRMANN, 1969; LEEMAN and SISSON, 1996). However, in granulite-facies rocks B and Be contents are probably comparable, *i.e.*, B contents rarely exceed 5 ppm irrespective of protolith (LEEMAN and SISSON, 1996). Thus, the formation of pegmatites containing B or Be minerals (to be referred herein as "boron or B pegmatites" and "beryllium or Be pegmatites", respectively) must involve special processes. I have worked on six examples of B and Be abyssal pegmatites (Table 1) that meet the following criteria, two of which distinguish them from pegmatites intruded after the granulite-facies metamorphism:

1) Discordant contacts with country rocks or K-feldspar-rich composition, *i.e.*, evidence that the pegmatites are melts and not merely recrystallized feldspathic rocks that have not been melted,
2) High-temperature mineralogy, *e.g.*, Al_2SiO_5 + K-feldspar (Kfs) in aluminous compositions or orthopyroxene (Opx) in intermediate compositions. The suite of Be and B minerals in the high-temperature pegmatites is also diagnostic (Table 2),
3) Minimal retrogression of their granulite-facies host rocks.

Two important features of the B and Be pegmatites are

1) Host rocks are Al-rich paragneisses or metapelites containing cordierite, almandine-pyrope, sillimanite, and/or sapphirine,
2) No obvious plutonic source (see also GINZBURG *et al.*, 1979; ČERNÝ, 1982a).

To date, no abyssal pegmatite has been found to contain *both* B minerals and Be minerals. However, in B pegmatites, some minerals have elevated Be contents; and, to a lesser extent, a few minerals contain B in a Be pegmatite. Nonetheless, simultaneous enrichment in Be and B is possible. A specimen of the Homagama B pegmatite is

Table 1. Pegmatites reviewed in this study.

Locality	Main sources
Boron pegmatites	
Cap Andrahomana, SE Madagascar	LACROIX (1904, 1922); BEHIER (1961); GREW *et al.* (1998c)
Almgjotheii in the contact aureole of the Rogaland Complex, SW Norway	HUIJSMANS *et al.* (1982); GREW *et al.* (1998a, c)
Homagama, in the Wanni Complex, Sri Lanka	HIROI *et al.* (1990); GREW *et al.* (1995)
Larsemann Hills, Prydz Bay, East Antarctica	GREW *et al.* (1998a)
Beryllium pegmatites	
"Zircon Point", Casey Bay, Enderby Land, East Antarctica (Fig. 2)	GREW (1981)
"Christmas Point", Casey Bay, Enderby Land, East Antarctica (Fig. 2)	GREW (1981)

Note: BEHIER (1961) also reported a Gdd-bearing pegmatite from Fort-Dauphin, SE Madagascar.

modestly enriched in both B (30 ppm) and Be (44 ppm)(GREW *et al.*, 1995). The absence of a discrete Be phase in this pegmatite could be due to the incorporation of Be in cordierite, which is a relatively abundant accessory phase and contains 0.08 wt% BeO.

In the present paper, I will review my work on the six pegmatites listed in Table 1. Inasmuch as the work on B pegmatites has been published recently in detail (1995-1998), my review of these pegmatites will be brief and will stress the importance of the B pegmatites as geochemical signatures of their host rocks. I will also note similarities and differences between the B and Be pegmatites. On the other hand, my primary work on the Be pegmatites was published in 1981 and subsequent studies have caused me to substantially revise my original interpretations. Consequently, the present paper will focus on the Be pegmatites. Moreover, the metamorphic evolution of the host rocks of the two Be pegmatites is controversial. I will contend herein that when studied in conjunction with the granulite-facies host rocks, the Be pegmatites (plus barren pegmatites of the same generation) can provide important information on crustal evolution and geochemistry not obtainable from study of the host rocks alone.

2. Boron Pegmatites

Characteristic features of the B pegmatites at the four localities listed in Table 1 are their relatively small size and generally discordant relationship to the host rocks, their distinctive Al-rich mineral assemblages and dominance of K-feldspar over plagioclase (Tables 2, 3) and their close association with Al-rich host rocks (metapelites) containing cordierite, almandine-pyrope and sillimanite. The pelitic protoliths of the host rocks presumably contained about 100 ppm B; even more B was possibly present in the protoliths to the highly graphitic gneisses at Almgjotheii and to the gneisses containing abundant prismatine (=B-rich kornerupine), grandidierite and tourmaline in the Larsemann Hills (*e.g.*, CARSON *et al.*, 1995). Boron enrichment in the pegmatites can be explained by remobilization of B present in the host rocks, incorporation of the remobilized boron in partial melts, and crystallization of the melts as pegmatite bodies with borosilicates. As a result, a small proportion of the B lost from the metapelites during granulite-facies metamorphism was "captured" by the pegmatites, *i.e.*, the B pegmatites have retained the B signature characteristic of pelitic rocks, whereas the pelitic host rocks have lost this signature as a result of granulite-facies metamorphism.

The boron pegmatites have several features in common with the Be pegmatites described in the next section. The pegmatites typically form small, isolated, discordant bodies a few centimeters thick and about one meter in extent (*e.g.*, GREW *et al.*, 1995, Fig. 1) and generally can be tied to a structural and plutonic event. However, the pegmatites at Andrahomana form a string of lenses up to 1 m thick and extend several hundreds of meters (H. BESAIRIE, cited in BEHIER, 1961). Although the Larsemann Hills pegmatite is mostly layer-parallel and only locally discordant, the dominance of K-feldspar in this pegmatite distinguishes it from plagioclase-dominated leucosomes in the area and suggests that it is an anatectic granitic melt extracted from the leucosomes as described by CARSON *et al.* (1997).

The suite of boron minerals in the abyssal pegmatites differs markedly from that in granitic pegmatites associated with amphibolite-facies rocks in that tourmaline is much less

Table 2. Minerals of boron and beryllium in high-temperature rocks and their pegmatites.

Mineral	Formula	Ideal or max. BeO in wt%
Minerals with essential boron ; some also with non-essential beryllium		
Tourmaline group (Tur)	$(Na,Ca)(Mg,Fe)_3Al_6Si_6O_{27}(OH)_3(OH,F)$	0.10
Dumortierite group (Dum)	$(Al,\square,Ti,Mg)Al_6BSi_3O_{16}(O,OH)_2$	—
Kornerupine group (Krn)	$(\square,Fe,Mg)(Mg,Fe,Al)_9(Si,Al,B,Be)_5(O,OH,F)_{22}$	0.67
"Sillimanite family"		
Grandidierite (Gdd)	$(Mg,Fe)Al_3BSiO_9$	<0.01
Werdingite (Wrd)	$(Mg,Fe)_2Al_{12}(Al,Fe)_2Si_4B_2(B,Al)_2O_{37}$	0.55
Boralsilite (Bor)	$Al_{16}B_6Si_2O_{37}$	0.25
Minerals with essential beryllium		
Surinamite (Sur)	$(Mg,Fe)_3Al_4BeSi_3O_{16}$	4.0
Chrysoberyl (Cb)	Al_2BeO_4	20
Musgravite-9R (Mgr)	$(Mg,Fe,Zn)_2Al_6BeO_{12}$	5.7
Taaffeite-4H	$(Mg,Fe)_3Al_8BeO_{16}$	4.4
Other minerals with non-essential beryllium in amounts ≥ 0.1 wt% BeO		
Sapphirine (Spr)	$Mg_{2.73}Fe_{1.00}Al_{7.14}Be_{0.72}B_{0.01}Si_{2.40}O_{20}$	2.51
Cordierite (Crd)	$Na_x(Mg,Fe)_2(Al_{4-x}Be_x)Si_5O_{18}.nH_2O$ with $0<x<0.48$	1.94
Sillimanite (Sil)	Al_2SiO_5	0.12

Sources of data: Černý and Povondra (1966); Deer *et al.* (1986); Henry and Dutrow (1996); Moore and Araki (1983); Nuber and Schmetzer (1983); Schmetzer (1983): Grew (1981); Grew *et al.* (1995, 1998a, b, c); Grew, Marquez and Wiedenbeck (unpublished ion probe data).

Table 3. Representative mineral assemblages in the pegmatites and conditions estimated for their crystallization.

Locality	Minerals	Temperature, pressure
	Boron pegmatites, minor Be in Wrd, Bor, Sil, Prs	
Andrahomana	Qtz+Kfs+Sil+Grt+Gdd+Wrd; 2nd And, Bt	~600-700°C, ~3-4 kbar
Almgjotheii	Qtz+Kfs+Sil+Grt+Gdd+Dum; Wrd+Bor; 2nd Tur, Crn	*ditto*
Homagama	Qtz+Prs+Crd; Qtz+Kfs+And+Tur+Crd+Pl; ±Sil	$\geq$540°C, $\leq$3 kbar
Larsemann	Qtz+Kfs+Bor+Tur	~750°C, 4-5 kbar
	Beryllium pegmatites, trace B in Spr, Sil, Sur	
"Zircon Point"	Sil+Grt+Sur+Mgr*+Be Spr+Bt; Sur+Cb+Bt	$\geq$820°C, $\geq$10 kbar
"Christmas Pt."	Qtz+Sil+Grt+Sur+Bt+Opx; 2nd Crd	*ditto*

Note: And-andalusite, Bt-biotite, Crn-corundum, Grt-garnet; Kfs-K-feldspar, Opx-orthopyroxene; Prs-prismatine (B-rich Krn), Qtz-quartz (other abbreviations in Table 2). *Originally reported as taaffeite (Grew, 1981), but better described as musgravite (Schmetzer, 1983). Sources: B pegmatites-Grew *et al.* (1995, 1998a, c); Be pegmatites-Grew (1981 and unpublished data).

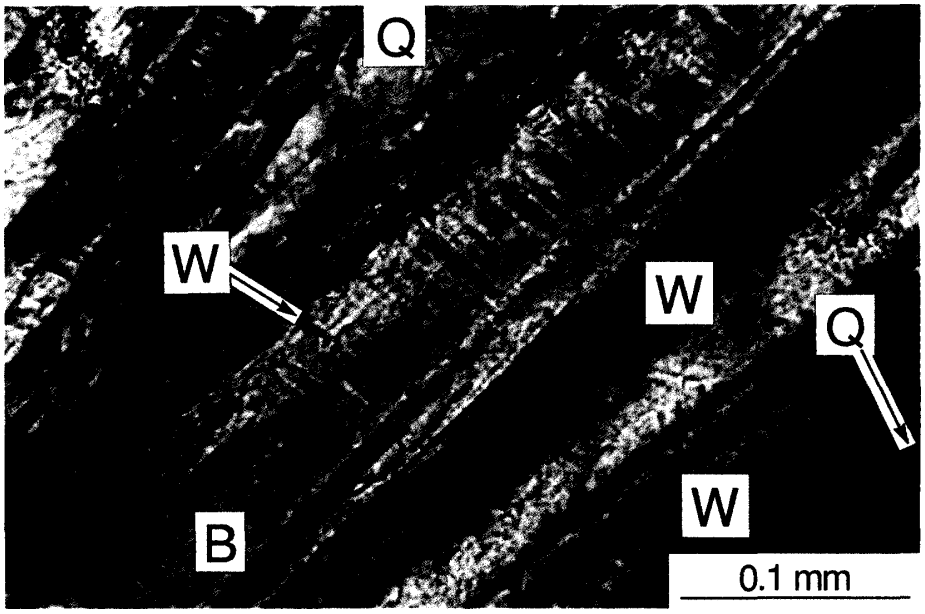

Fig. 1. Photomicrograph of ladder-like (scalariform) intergrowth of boralsilite (B, purplish) and werdingite (W, bluish-green and yellow). The adjacent prism is werdingite. Q-quartz. Crossed nicols. Sample HE138B2, Almgjotheii, Norway (a small-scale view of the intergrowth in plane light appears in Fig. 3 of GREW et al., 1998c).

important or even absent (Table 2). More prominent are minerals belonging to a family of boroaluminosilicates, sillimanite and andalusite that are related structurally to one another (PEACOR *et al.*, in press.). This "sillimanite family" includes the new mineral boralsilite, which resembles sillimanite optically and could be easily mistaken for it. The close relationship among the "sillimanite family" minerals is illustrated by their commonly forming intergrowths of parallel prisms (GREW *et al.*, 1998a, c) and a distinctive ladder-like or scalariform intergrowth between werdingite and boralsilite (Fig. 1). The peraluminous composition of the pegmatites is evidenced by the presence of the Al_2O_3-saturating assemblage Sil (or And)+Qtz (abbreviations in Tables 2 and 3) in most of the pegmatites and the Al-rich compositions of the borosilicates. Despite the aluminous composition, primary muscovite is absent. Andalusite is characteristic of three of the four pegmatites, and it appears to have been stable with K-feldspar at some stage during evolution of the pegmatites, implying that water activities were low.

No Be minerals have been reported in these pegmatites. However, pegmatitic werdingite and sillimanite contain the maximum BeO contents reported in these minerals (Table 2); pegmatitic prismatine from Homagama contains an above-average 0.14–0.18 wt% BeO (GREW *et al.*, 1995).

In contrast to the Be pegmatites, the B pegmatites formed late in the metamorphic evolution of their respective terrains. The temperatures estimated for crystallization (Table 3) are below the temperatures inferred for peak conditions for granulite-facies metamorphism of the host rocks, in some cases lower by 100°C (Almgjotheii, Andrahomana) to by more than 200°C (Homagama). Nonetheless, the crystallization temperatures are higher than those estimated for the solidi of several rare-element (Li or B) pegmatites, 475°C (LONDON, 1992). Pressures of crystallization are inferred not to have exceeded 5 kbar.

Many of the boron minerals appear to have crystallized from the pegmatite magma. In summary, the four B pegmatites evolved in a very different way from the two Be pegmatites reviewed herein although both types are inferred to have formed by anatexis of their aluminous host rocks.

3. Beryllium Pegmatites

3.1. Description and conditions of formation

Archean pegmatites with a granulite-facies mineralogy are rare in the Napier Complex except in a few areas such as eastern Casey Bay (*e.g.*, GREW and MANTON, 1979; SANDIFORD and WILSON, 1984, 1986; SHERATON *et al.*, 1987; this paper, Fig. 2 from GREW, 1981). Be minerals have been reported in only two pegmatites, one each at "Christmas Point" and "Zircon Point". The Be pegmatites form pods up to 1 m long and are clearly discordant. At "Christmas Point", the Be pegmatite is one of an *en echelon* set of barren pegmatites of the same generation. The Be-bearing and associated "barren" pegmatites appear to be filling pressure shadows of boudins of more competent rock pulled apart during the main deformation D_2 (SANDIFORD and WILSON, 1983, 1984). Host rocks are quartz granulites containing sapphirine, orthopyroxene, sillimanite and garnet, *i.e.*, Mg-Al-rich rocks typical of the Napier Complex.

The suite of beryllium minerals in these two pegmatites differs markedly from the suite typical of granitic pegmatites in that beryl is absent (Tables 2, 3). The most prominent Be mineral is surinamite, a beryllosilicate related to sapphirine; it is readily recognizable for its distinctive pleochroism (Fig. 3). Sapphirine associated with surinamite is beryllian. Like the B pegmatites, the mineral assemblages are peraluminous, and the "Christmas Point" assemblage is saturated in Al_2O_3 (Sil+Qtz). A small amount of boron is present in sillimanite, sapphirine and surinamite from "Zircon Point" (Table 4). These minerals incorporate substantial boron in B-rich metamorphic parageneses, *e.g.*, assemblages of sapphirine and sillimanite with borosilicates.

Textural relations suggest that cordierite ("Christmas Point"), kyanite ("Zircon Point") and some biotite in the pegmatites appeared after crystallization of the Be minerals, sillimanite and orthopyroxene, and most likely formed by retrogression during a later

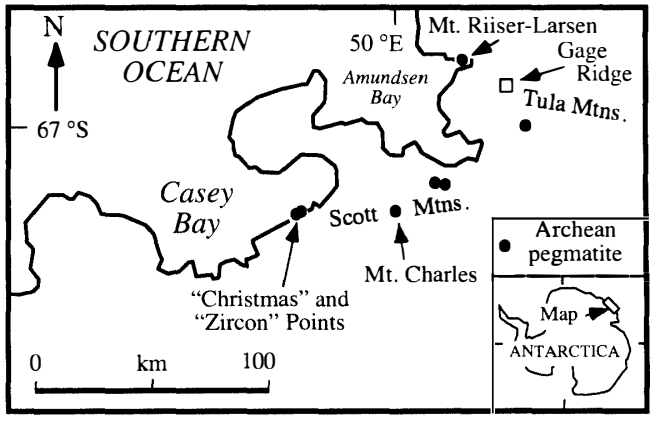

Fig. 2. Sketch map of western Enderby Land showing location of Late Archean pegmatites.

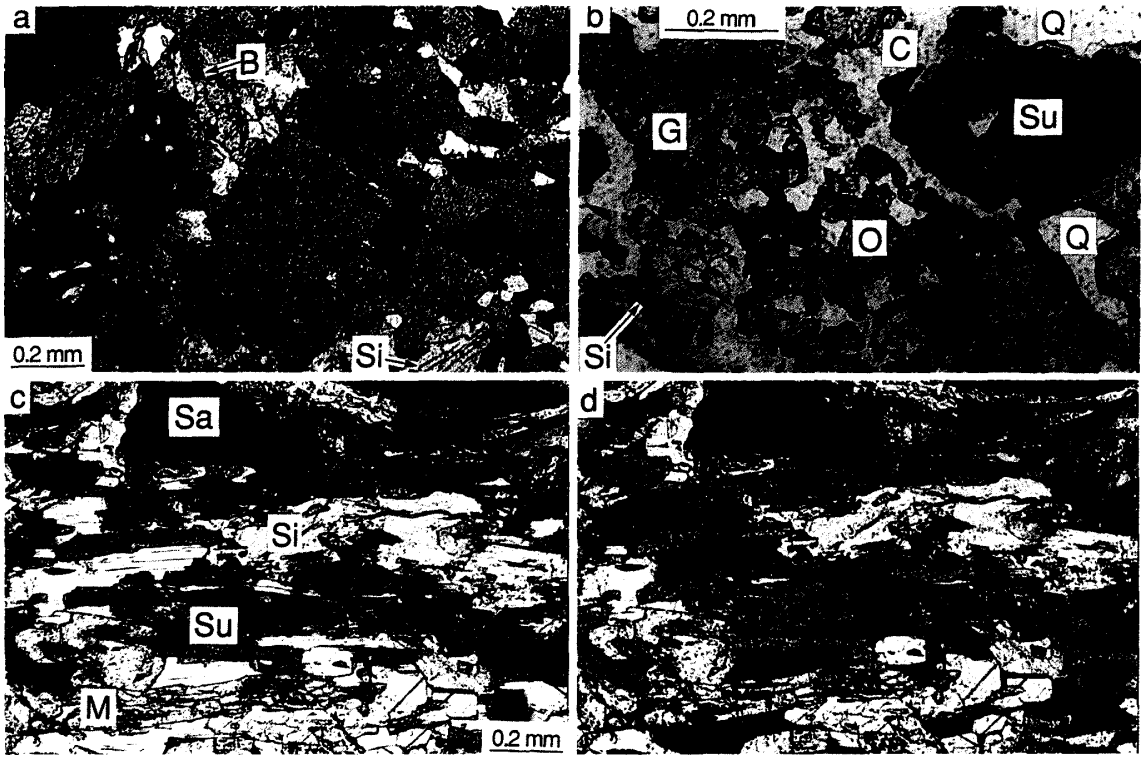

Fig. 3. Photomicrographs of surinamite from Casey Bay, Enderby Land, East Antarctica. a) Surinamite aggregate (purple, blue and green hues) with minor sillimanite (Si, cleavage), cordierite + quartz (low relief, some stained brown), and biotite (B, brown). Plane light. Sample 2292B, "Christmas Point", b) Surinamite (Su), garnet (G, high relief), orthopyroxene (O, brown), and sillimanite (Si, cleavage and colorless, one grain), which are isolated by moats of cordierite (C, low relief). Q-quartz. Plane light. Sample 2292B. c) and d) Foliated pegmatite with surinamite (Su, greenish-blue to purple pleochroism), sapphirine (Sa, blue to purplish blue pleochroism), biotite (colorless to brown pleochroism), sillimanite (Si, colorless, moderate relief), and musgravite (M, green tint, high relief). Photomicrographs were taken in plane light at right angles to one another. Sample 2234L/3, "Zircon Point".

Table 4. Boron contents of minerals in the Napier Complex Be pegmatites (in wt% B_2O_3).

Mineral	"Zircon Point"*	"Christmas Point"*	Maximum reported†
Sillimanite	0.06	0.002	0.66
Sapphirine (Ca-poor)	0.05	—	0.85
Surinamite	0.17	0.004	3.17

Sources; *GREW, MARQUEZ and WIEDENBECK (unpublished ion probe data), †RAMESH KUMAR *et al.* (1995); GREW (1996), GREW *et al.* (1998c).

(post-Archean) event in the amphibolite-facies (*e.g.*, GREW, 1981; SANDIFORD and WILSON, 1983).

In my original study of the pegmatites, I suggested direct crystallization of the Be minerals from a pegmatitic magma (GREW, 1981). However, this interpretation seems implausible in light of more recent studies, and I no longer accept it. FRANZ and MORTEANI (1984) and DE ROEVER and VRÁNA (1985) suggested that chrysoberyl, sur-

Fig. 4. View looking down at a Late Archean pegmatite on "Christmas Point" containing coarse, brown sillimanite prisms (Sil). Red color is from K-feldspar. Host rock is gray.

inamite, and associated minerals in the Napier Complex pegmatites formed by recrystallization of a lower-T assemblage with beryl or beryllian cordierite under granulite-facies conditions, *i.e.*, the pegmatites have been metamorphosed under high P-T conditions following emplacement. Both pairs of authors cited analogies with other occurrences, respectively, the formation of chrysoberyl from beryl and K-feldspar in metapegmatites and the formation of surinamite from metamorphic Be-bearing cordierite during a superimposed, higher pressure event (or phase) in the upper amphibolite or granulite facies. In addition, DE ROEVER and VRÁNA (1985) noted the contrast in the Napier Complex pegmatites between coarse-grained quartz, K-feldspar and sillimanite in grains several centimeters across of probable magmatic origin (Fig. 4) with the medium-grained Be minerals of possible metamorphic origin (Fig. 3). This distinction is supported by the foliation in the "Zircon Point" pegmatite (Fig. 3c, d). A third scenario not considered in the literature is that the coarse sillimanite prisms are xenocrysts and that the medium-grained Be-Al-rich assemblages are xenoliths that either reacted with Be-bearing pegmatitic melt or were reworked by residual Be-rich solutions from the melt (Simon HARLEY, personal communication, 1998). This scenario is difficult to distinguish from DE ROEVER and VRÁNA's (1985) on textural grounds alone. Nonetheless, the abundance of coarse sillimanite in the pegmatites and its absence in the host rocks suggest that the coarse sillimanite is more likely magmatic. Moreover, the proportion of ferromagnesian minerals

in the medium-grained aggregates is far greater than in most host rocks, which are predominantly quartz. Feldspar, although not rare in the host rocks, is absent from the aggregates.

Given the available evidence, my preferred scenario is that the coarse-grained minerals are magmatic and that the medium-grained Be-rich mineral aggregates are derived from the prograde metamorphism of *pegmatitic* beryllian cordierite; that is, these pegmatites *have been metamorphosed*. The presently observed aggregates could not have formed from the isochemical breakdown of beryllian cordierite. In order to balance reactions relating cordierite breakdown and the presently observed assemblages, substantial quartz must have been lost to the matrix of the aggregates, together with Na_2O and H_2O. For example, possible reactions for breakdown of cordierite with 0.5 Be per formula unit, the maximum reported in nature (Table 2) and close to the maximum synthesized ($x=0.65$, POVONDRA and LANGER, 1971), yield assemblages similar to those in the "Christmas Point" pegmatites (combining Fe with Mg):

$$4Crd(Na_{0.5}Mg_2Be_{0.5}Al_{3.5}Si_5O_{18}.nH_2O) \rightarrow 2Sur + 2Opx + 3Sil + 9Qtz + Na_2O + 4nH_2O.$$
$$12Crd(Na_{0.5}Mg_2Be_{0.5}Al_{3.5}Si_5O_{18}.nH_2O) \rightarrow 6Sur + 2Grt + 7Sil + 29Qtz + 3Na_2O + 12nH_2O$$

The "Zircon Point" assemblage requires even greater silica loss; possibly chrysoberyl was a reactant in this pegmatite. If the starting material had been typical host gneiss instead of cordierite, even more quartz and alkalis would have to have been driven out (and BeO added) to give the observed assemblages at either "Christmas Point" or "Zircon Point".

Further evidence for metamorphism of the pegmatite is a corona of orthopyroxene + plagioclase symplectite around hornblende (Fig. 5), which I collected from a pegmatite on

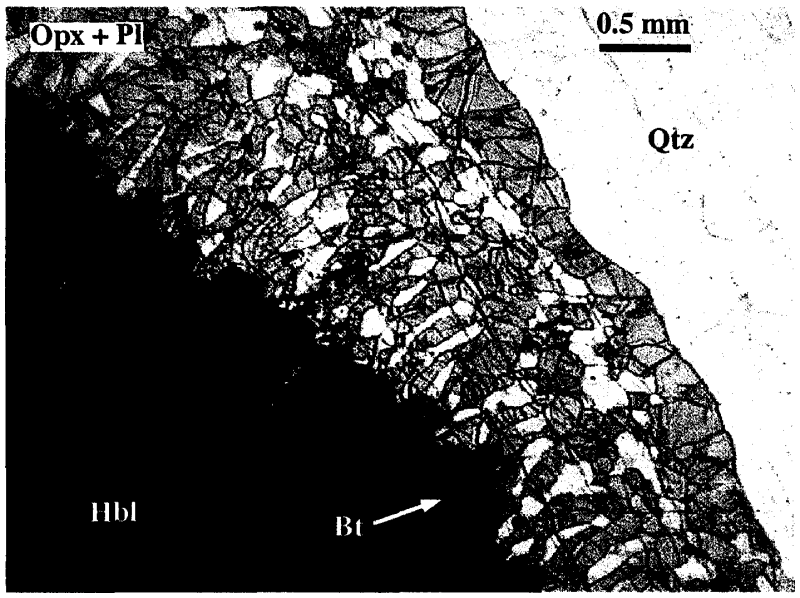

Fig. 5. Photomicrograph of a pegmatite from Mt. Charles showing embayment of a hornblende megacryst (Hbl) by orthopyroxene + plagioclase symplectite (Opx+Pl). Qtz-quartz. Small flakes of biotite (e.g., Bt) are present along margin of the hornblende (these flakes are oriented so as to appear equally dark as Hbl). Plane light. Sample 2098E; this pegmatite was dated by GREW and MANTON (1979, no. 2098A).

Mt. Charles in the Scott Mountains (Fig. 2). Although the corona admits to several interpretations, my preferred interpretation is that orthopyroxene+plagioclase formed from the prograde breakdown of hornblende. That the corona formed during decompression is implausible given the proposed counter-clockwise *P-T* path inferred for the Napier Complex (see below).

The Be mineral assemblages crystallized at higher temperatures and pressures than the B mineral assemblages (Table 3). The temperature estimate is based on the Qtz+Sil+Opx (+Sur) assemblage found locally in the "Christmas Point" pegmatite (Fig. 3b). HARLEY *et al.* (1990) calculated $T \geq 820$°C, $P \geq 10$ kbar for Qtz+Sil+Opx. Equally important is the presence of Opx+Sil+Qtz in the host rocks at "Christmas Point" where it formed from the reaction of sapphirine+quartz (GREW, 1982, Fig. 5). M. SANDIFORD (in GREW *et al.*, 1982) reported that sapphirine occurs with quartz in a pegmatite on McIntyre Island, 2 km north of "Zircon Point". In the absence of details on textural relations and composition of the sapphirine, it would be premature to cite this as evidence for Spr+Qtz stability in the pegmatites. Nonetheless, it is safe to conclude that the pegmatites experienced ultra-high metamorphic temperature conditions comparable to those affecting the host rocks.

This conclusion is consistent with available data on beryllium minerals. In contrast to most minerals containing essential beryllium (*e.g.*, BARTON, 1986), surinamite is restricted to high-temperatures. HÖLSCHER *et al.* (1986) synthesized the Mg-Al surinamite end member from unseeded gel at 775–1000°C and 9–20 kbar and from seeding experiments found a preliminary stability field to lie at $T \geq 650$°C and $P \geq 4$ kbar.

I previously rejected a non-magmatic origin for the Be as there was no obvious mechanism for transporting and concentrating Be under granulite-facies conditions (GREW, 1981). However, in view of the high mobility of trace elements reported by DEPAOLO *et al.* (1982), presence of Be enrichments in Napier Complex rocks at Gage Ridge (Table 5), and absence of visible connections between the Be pegmatites and plutonic rocks (*cf.*, ČERNÝ, 1982b), I later suggested that the Be originated in the country rocks and was transported and concentrated by fluids (GREW, 1984). However, given the dry nature of granulite-facies metamorphism, it seems more likely that anatexis played a more important role than simple fluid transport.

In contrast to B, Be is a rare constituent of sediments. Is it possible that the precursors to the Napier complex contained sufficient Be to be a source of Be for the pegmatites? Mg-Al-rich rocks compositionally similar to the Napier Complex host rocks are reported to contain Be-bearing minerals at other localities (Table 5). Mg-Al-rich rocks are characterized by high Mg/Fe ratios compared to typical pelitic sediments and by the presence of magnesian phases such as talc, sapphirine, and cordierite, *e.g.* kyanite+talc "white-schist" from Tajikistan and Zambia (GREW *et al.*, 1998b) and cordierite-sillimanite-spinel-sapphirine gneisses from the Eastern Ghats, India (RAMESH KUMAR *et al.*, 1995). The Napier Complex itself could have a significant Be signature. Research to date has turned up three occurrences of Be concentrations in the Napier Complex, two in Mg-Al-rich rocks (Gage Ridge and Mt. Riiser-Larsen, Table 5). Other potential carriers of Be in the Napier Complex are metapelitic rocks analogous to the surinamite-bearing mesoperthite gneisses from Surinam (type locality) and eastern Zambia (DE ROEVER *et al.*, 1976; DE RORVER and VRÁNA, 1985). Consequently, there is reason to suggest that Be concentrations are

Table 5. Mineral and whole rock beryllium signatures of non-metasomatic origin.

Locality	Mineral Host−BeO	Whole Rock Be	Source
	Mg-Al-rich rocks−Amphibolite-facies		
South Australia	Taaffeite−4.4 wt%	5.3 ppm	TEALE (1980)
Central Alps	Cordierite−0.8 wt%	—	ARMBRUSTER & IROUSCHEK (1983)*
Central Zambia	Kornerupine−0.67 wt%	—	GREW *et al.* (1998b)
SW Pamir, Tajikistan	Kornerupine−0.29 wt%	—	GREW *et al.* (1998b)
	Mg-Al-rich rocks−Granulite facies		
Strangways Range, Aust.	Surinamite	—	WOODFORD and WILSON (1976)
Gage Ridge, Enderby Land	Sapphirine (?)	10–23 ppm†	SHERATON *et al.* (1982)
Mt. Riiser-Larsen, *ditto*	Sapphirine−~0.1 wt%	—	CHRISTY (1989)
Eastern Ghats, India	Surinamite−4.2–4.8 wt%	—	RAMESH KUMAR *et al.* (1995)
	Pelitic and quartzofeldspathic rocks−Upper amphibolite and granulite facies		
Bakhuis Mtns., Surinam	Surinamite	—	DE ROEVER *et al.* (1976)
Chimwala area, E. Zambia	Surinamite	—	DE ROEVER and VRÁNA (1985)
Gage Ridge, Enderby Land	Sapphirine−1.7 wt%	5 ppm	GREW & SHERATON (unpubl. data)

Note: *See also IROUSCHEK-ZUMTHOR and ARMBRUSTER (1985). †Compare with average of 2.4±1.5 ppm Be for 21 granulite-facies pelites from Prydz Bay, Antarctica (SHERATON *et al.*, 1984).

present in the Mg-Al-rich and metapelitic rocks of the Napier Complex underlying western Casey Bay and that these could have been a source of Be for the pegmatites.

Mg-Al-rich rocks are commonly enriched in B, *e.g.*, the kornerupine-tourmaline-bearing whiteschists from Tajikistan and Zambia listed in Table 5. The B contents of the minerals in the "Zircon Point" pegmatite could be a faint B signature that survived the ultra-high-temperature metamorphism of an originally B-bearing rock. B is probably more easily mobilized than Be, and thus less likely to remain unless concentrations in the protolith were substantial.

3.2. The significance of the Be pegmatites in interpreting Napier Complex evolution

When studied in connection with their host rocks, granulite-facies pegmatites provide information not always available from study of the host metamorphic rocks by themselves. In the case of the Napier Complex, this information is critical in deducing which of two fundamentally different scenarios proposed for the evolution of the Napier Complex best fits available data. The pegmatites do not provide any information on the pre-Late Archean history of the Napier Complex, which extends back to about 3800 Ma (*e.g.*, SOBOTOVICH *et al.*, 1976; DEPAOLO *et al.*, 1982; BLACK *et al.*, 1986; SHERATON *et al.*, 1987; HARLEY and BLACK, 1997). The present discussion focuses on events in the Late Archean.

I propose herein Scenario 1 (Fig. 6), which is consistent with available information on the relative and absolute ages, structural relations (*e.g.*, SANDIFORD and WILSON, 1983, 1984), and metamorphism of the generation of granulite-facies pegmatites, including the Be

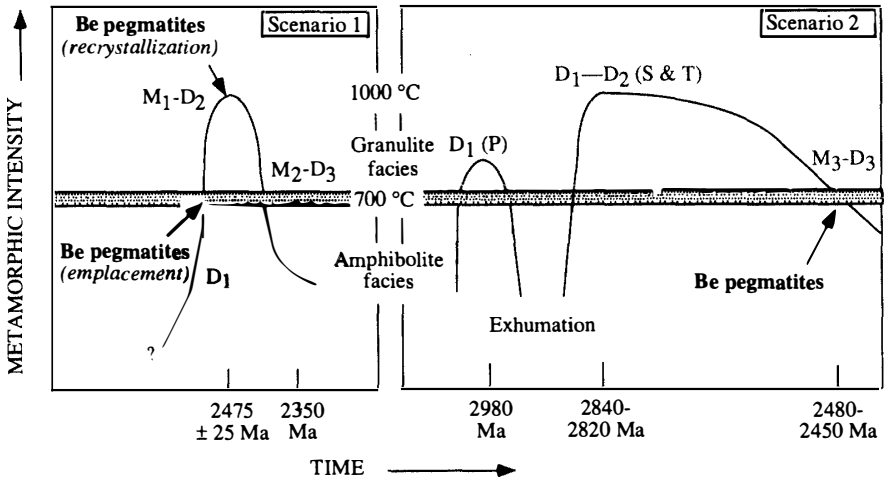

Fig. 6. Schematic diagram illustrating two scenarios proposed for the middle and late Archean evolution of the Napier Complex. Scenario 1 is based on the interpretations of GREW *and* MANTON *(1979),* GREW *et al. (1982),* DEPAOLO *et al. (1982),* SANDIFORD *and* WILSON *(1983, 1984) and* SANDIFORD *(1985). Scenario 2 was proposed by* HARLEY *and* BLACK *(1997) and is modified from the scenario summarized in* BLACK *and* JAMES *(1983) and* SHERATON *et al. (1987). P-Proclamation Island (northern Enderby Land), T-Tula Mountains, S-Scott Mountains.*

pegmatites. Scenario 2 is based on the extensive geochronological work by L.P. BLACK in collaboration with structural geologist P.R. JAMES and petrologist S.L. HARLEY (*e.g.*, BLACK and JAMES, 1983; BLACK *et al.*, 1983b; SHERATON *et al.*, 1987), and is the generally accepted interpretation. It has been recently updated and revised by HARLEY and BLACK (1997). Nonetheless, it is not supported by available information on the granulite-facies pegmatites.

Both groups of investigators agree on two observations. One is that the granulite-facies pegmatites were emplaced about 2500 Ma ago, *i.e.*, these pegmatites are Late Archean in age (GREW and MANTON, 1979; GREW *et al.*, 1982; BLACK *et al.*, 1983b). The second is that isotopic systems in the Napier Complex metamorphic rocks were highly disturbed and extensively reset at that time, although evidence of pre-Late Archean activity does remain. HARLEY and BLACK (1997, shown in Fig. 7) now give the age of the Late Archean event more precisely as 2480–2450 Ma. Sm-Nd data have confirmed the marked effect of the Late Archean event on isotopic systems (TAINOSHO *et al.*, 1994, 1997; OWADA *et al.*, 1994, 1995; OSANAI *et al.*, 1995). Also, structural relations suggest that the pegmatites are anatectic melts associated with the Late Archean event, a point that most investigators seem to agree on.

The disagreement between the two groups of investigators lies in relating the pegmatite emplacement to structural and metamorphic events. For the former, it is a question of assigning a given structural feature to one of the three deformational events in the Archean and earliest Proterozoic. BLACK *et al.* (1983b) dated two pegmatites in western Casey Bay as Late Archean; neither of which was reported to contain Be minerals. The one from McIntyre Island is described as a "sweat" pegmatite in the hinge of an F_3 fold (p. 298), but no details were given on the structural relations of the "old" folded sillimanite pegmatite from "Christmas Point" (p. 285). The latter pegmatite belongs to the same generation as

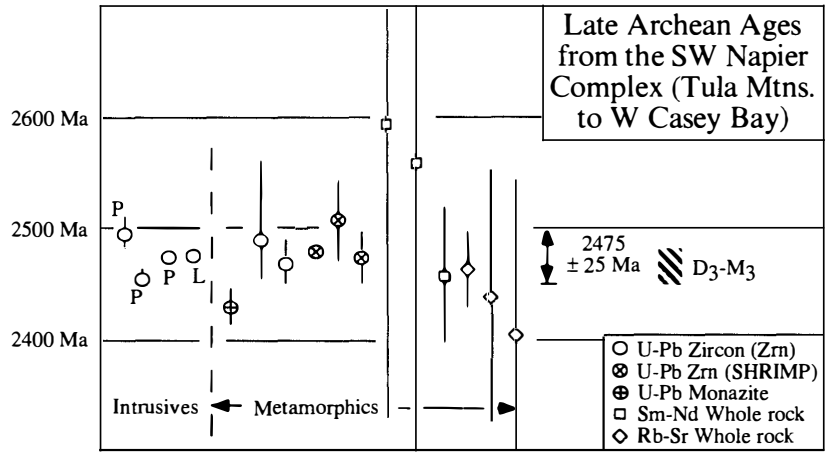

Fig. 7. Summary of isotopic ages between 2400 and 2600 Ma for intrusive rocks and gneisses in the southwestern part of the Napier Complex. The vertical lines indicate the stated uncertainties. Zircon and monazite U-Pb ages were determined on concordia diagrams; whole rock ages are isochrons. For the intrusive rocks, P-pegmatites from western Casey Bay, L-leuconorite from Mt. Hardy. The hachured block is the range of ages inferred for the D_3-M_3 event (HARLEY and BLACK, 1997). The data are taken from JAMES and BLACK (1981), GREW et al. (1982, recalculated by W.I. MANTON, personal communication, 1998, to 2496 + 16/−13 Ma for the upper intercept and 1135 ± 190 Ma for the lower intercept, MSWD = 0.60), BLACK et al. (1983a, b, 1984, 1986), SHERATON et al. (1987), OWADA et al. (1994), TAINOSHO et al. (1994, 1997), and HARLEY and BLACK (1997).

the Be pegmatite and the pegmatite illustrated in Fig. 4 at "Christmas Point". SANDIFORD and WILSON (1983, 1984) reported that pegmatites with a high-temperature mineralogy are commonly developed in inter-boudin spaces associated with D_2 boudinage in western Casey Bay, but they found no pegmatites were associated with D_3 boudinage, which shows minimal boudin separation. SANDIFORD and WILSON (1984) also reported pegmatite veins in conjugate sets related to F_2 folds, and thus presumed these pegmatites to be D_2 anatectic melts (also SANDIFORD and WILSON, 1986). In summary, SANDIFORD and WILSON (1984, 1986) assigned the pegmatites to the main deformational event, D_2; they found no pegmatites associated with D_3. At several localities in the Scott and Tula Mountains, I observed pegmatite pods up to one meter thick and several meters long; some contain coarse sillimanite prisms. These pegmatites lie at a high angle to compositional layering and are restricted to certain layers; that is, the pods appeared to be filling inter-boudin spaces (GREW, unpublished field notes). Thus, SANDIFORD and WILSON's assignment is the more compelling for me.

In the case of metamorphism, the question is whether the ultra-high-temperature assemblage is close in age to pegmatite emplacement (Scenario 1) or significantly older (Scenario 2). The presence of the ultra-high temperature assemblage Qtz + Opx + Sil in the pegmatites is not consistent with emplacement during M_3, for which BLACK *et al.* (1983a, b) and HARLEY (1983) estimated $T = 650$–720°C and $P = 5$–8 kbar. Simon HARLEY (personal communication, 1998) has suggested a xenolithic origin for the ultra-high-temperature assemblage Opx + Sil + Qtz (and Spr + Qtz, if this could be confirmed), thereby resolving the discrepancy, *i.e.*, the ultra-high-temperature assemblages are relict and could

be much older than the pegmatites. This scenario presupposes that the Be-rich assemblages are lower grade because Be enrichment must have originated in the pegmatites. Available data on the stability of Be minerals point to the surinamite-bearing assemblages being high temperature. I prefer the simpler scenario whereby the orthopyroxene, sillimanite, quartz, *and* surinamite crystallized together from pre-existing pegmatitic minerals under ultra-high-temperature conditions. The breakdown of pegmatitic beryllian cordierite to surinamite or musgravite + sapphirine occurred under the same ultra-high temperature conditions experienced by the host rocks during the peak M_1 event.

HENSEN and MOTOYOSHI (1992) noted that prograde metamorphism of pre-existing pegmatites fits their scenario for the metamorphic evolution of the Napier Complex. This prograde recrystallization under increasing T and P conditions inferred for the pegmatites mirrors the breakdown of cordierite to Spr + Opx + Qtz reported in Mg-Al-rich rocks at Mt. Riiser-Larsen and other localities in an area over 2000 km^2 around Amundsen Bay (SHERATON *et al.*, 1987; MOTOYOSHI and HENSEN, 1989; HENSEN and MOTOYOSHI, 1992). Breakdown of cordierite to surinamite in the pegmatites and to sapphirine in the host rocks requires an increase in pressure as well as temperature. HENSEN and MOTOYOSHI (1992) also reported the breakdown of Spr + Kfs + Qtz to Osumilite + Grt + Sil with a further increase of pressure at very high temperature, but whether this event affected the Be pegmatites is not known.

The two scenarios not only differ in the age assigned to the ultra-high temperature event, but also in the nature of the prograde path of the ultra-high temperature event. On the basis of new age data, HARLEY and BLACK (1997) proposed that a period of exhumation intervened between two high-grade events early in the history of the Napier Complex (Fig. 6). According to this reinterpretation, the apparent prograde character of the Crd→Spr + Qtz + Opx reaction described by MOTOYOSHI and HENSEN (1989) resulted from superposition of a 2840 Ma ultra-high temperature event on the lower-temperature 2980 Ma granulite-facies event. According to Scenario 1, the reaction Crd→Spr + Qtz + Opx could be prograde because cordierite formation is presumed to be coeval with pegmatite emplacement, and was part of the same cycle as the very-high-temperature event. The *P-T* path overall for the ultra-high temperature metamorphism would be counter-clockwise, as proposed by HENSEN and MOTOYOSHI (1992).

The Late Archean event appears to have lasted as long as 50 Ma. HARLEY and BLACK (1997, p. 88) concluded that zircon growth and resetting of isotopic systems during their D_3-M_3 event occurred over a "relatively protracted" time of 25–30 Ma. Figure 7 shows that allowing for uncertainties, almost all the age data plot within a 50 Ma interval (2450–2500 Ma). The relatively large uncertainties for the data on the metamorphic rocks preclude fixing the age of the metamorphism more precisely than 2475 ± 25 Ma. The differences in the ages obtained on individual pegmatites exceed the measured uncertainties, and thus HARLEY and BLACK (1997) inferred several episodes of pegmatite emplacement during their D_3-M_3 event. This interpretation presumes that the uncertainties near ±10 Ma and less are a true measure of the accuracy of the pegmatitic U-Pb zircon ages. Despite the precision on the pegmatite ages, the interval of time between intrusion and recrystallization of the pegmatites implicit in Scenario 1 (Fig. 6) was too short to be resolved by the geochronological data on the metamorphic rocks.

4. Conclusions

Pegmatites in granulite-facies terrains are geochemical indicators of host-rock composition. For example, B is a characteristic trace constituent of pelitic sediments. Boron pegmatites retain this B signature, which has been lost from the host cordierite-garnet gneisses as a result of granulite-facies metamorphism. Be is potentially another such geochemical tracer. The protoliths of the Napier Complex Mg-Al-rich rocks have been a subject of considerable debate (*e.g.* SHERATON, 1980; SHERATON *et al.*, 1982); the effects of ultra-high-temperature metamorphism have complicated the task of deducing these protoliths. Is Be a characteristic trace element of these rocks? At several localities worldwide, Be-bearing minerals are found in amphibolite-facies Mg-Al-rich rocks in which there is no evidence for metasomatic introduction of Be (Table 5). A Be signature could result from a distinctive depositional environment and thus be a diagnostic geochemical feature. Confirmation that relatively high Be contents characterize the Napier Complex Mg-Al-rich rocks could simplify deducing which of several protoliths are precursors for these rocks.

The Late Archean pegmatites, including those enriched in Be, provide several major constraints on the relative and absolute timing of deformational and metamorphic events in the Napier Complex that are not obvious in study of the host rocks by themselves. According to Scenario 2, the Napier Complex resided at depth under high T conditions for nearly 400 Ma (Fig. 6). Such a prolonged residence at depth is not typical for high-T terrains, *e.g.*, Limpopo Belt (KAMBER *et al.*, 1996), and is not required by Scenario 1. Pegmatite emplacement and the very-high-temperature metamorphism are part of a single, possibly protracted event. The counter-clockwise P-T path with increasing P as well as T on the prograde path during this event constrains possible tectonic models for the extreme heating of the Napier Complex. Relationships deduced from the pegmatites are consistent with HENSEN and MOTOYOSHI's (1992) model of insertion of gabbroic or anorthositic rocks above the rocks presently exposed, which resulted in increased load as well as heat. No direct evidence for such an intrusive has been reported to date, although small bodies of premetamorphic anorthositic rocks have been found in the Napier Complex, *e.g.*, leuconorite from Mount Hardy dated at $2474+6/-3$ Ma (JAMES and BLACK, 1981; SHERATON *et al.*, 1987; HARLEY and BLACK, 1997). This leuconorite is coeval with one of the pegmatites (Fig. 7). It is tempting to suggest that this leuconorite is comagmatic with an overlying anorthosite subsequently eroded away and that the pegmatites are a consequence of heating underneath such an anorthosite.

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