Canadian Mineralogist
Vol. 30, pp. 835-875 (1992)

THE GHOST LAKE BATHOLITH, SUPERIOR PROVINCE OF NORTHWESTERN ONTARIO: A FERTILE, S-TYPE, PERALUMINOUS GRANITE – RARE-ELEMENT PEGMATITE SYSTEM

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ABSTRACT

The Late Archean, 2685 Ma Ghost Lake batholith, in the Dryden area of northwestern Ontario, is a late-tectonic, subsolvus, peraluminous, S-type, collisional granite in the Superior Province that features an unambiguous genetic linkage with a proximal, zoned cluster of rare-element pegmatites (Mavis Lake Group). The batholith consists of eight internal units (GLB–1 to –8) that range from chemically primitive biotite and cordierite–biotite granites (GLB–1 and –2) to a highly evolved, tourmaline–muscovite-bearing pegmatitic granite facies (GLB–4 to –8) locally containing endogenous beryl and cassiterite. Salient petrochemical attributes include A/CNK_{mol} between 1.194 and 1.480, high SiO_2 (70.5–77.4%), Al_2O_3 (14.2–19.9%), low CaO, MgO and Fe, and, with increasing fractionation, enrichment of Na_2O, B, F, Be, Cs, Ga and Li, and depletion of K_2O, Ba, total *REE*, *LREE*, Sr and Zr. Strong fractionation is also revealed by Al/Ga (1835–7020), Ba/Rb (<0.0009–9), Ca/Sr (6–383), K/Ba (15–6560), K/Cs (650–29,590), K/Rb (44–387), Mg/Li (4–200), Na_2O/K_2O (0.15–16), Ce/Yb_N (2.6–247), Eu/Eu^* (<0.06–1.98) and total *REE* (1.8–57× chondrite). Genesis of the peraluminous, S-type granite – rare-element pegmatite association in the Dryden area is explained by a complex interplay of petrogenetic processes that involves five stages. Rare elements and B were previously concentrated in wackes and mudstone deposited in large basins; these rocks underwent step-wise dehydration reactions involving muscovite and biotite, under fluid-absent conditions, and successively released these elements to anatectic melts. Rare elements and volatiles were progressively concentrated *via* crystal–melt fractionation, the Harker trends of which were obscured by two stages of extraction of residual melt and by episodic, subsolidus redistribution *via* base-cation leaching. The late magmatic history of the batholith is marked by widespread exsolution of a volatile-rich phase, dispersion of a rare-element-F–B-rich fluid along shear zones and ensuing emigration of rare-element-rich melt–fluid systems upward from the cupola, which led to the regionally zoned Mavis Lake Pegmatite Group.

Keywords: Ghost Lake batholith, Superior Province, peraluminous granite, rare-element-enriched granitic pegmatite, Dryden, Ontario.

SOMMAIRE

Le batholite de Ghost Lake, situé près de Dryden, dans le nord-ouest de l'Ontario, d'âge archéen (2685 Ma), contient un granite tardi-tectonique, subsolvus, hyperalumineux, de type S, et lié à un régime de collision dans la province du Supérieur. Il démontre un lien génétique non ambigu avec un essaim proximal et zoné de pegmatites à éléments rares (Groupe de Mavis Lake). Nous reconnaissons huit unités internes (GLB–1 à –8) allant de granites chimiquement primitifs à biotite et biotite + cordiérite (GLB–1 et –2) jusqu'à un facies de granite pegmatitique très évolué, à tourmaline + muscovite et, localement, béryl endogène et cassitérite (GLB–4 à –8). Parmi les caractéristiques pétrochimiques importantes, notons une valeur de A/CNK_{mol} entre 1.194 et 1.480, une teneur en SiO_2 et Al_2O_3 élevée (70.5–77.4% et 14.2–19.9%, respectivement), une faible teneur en CaO, MgO et Fe, avec un enrichissement en Na_2O, B, F, Be, Cs, Ga et Li, et un appauvrissement en K_2O, Ba, terres rares, surtout les terres rares légères, Sr et Zr, avec progression du fractionnement. Le caractère poussé du fractionnement est mis en évidence par les rapports Al/Ga (1835–7020), Ba/Rb (<0.0009–9), Ca/Sr (6–383), K/Ba (15–6560), K/Cs (650–29,590), K/Rb (44–387), Mg/Li (4–200), Na_2O/K_2O (0.15–16), Ce/Yb_N (2.6–247), Eu/Eu^* (<0.06–1.98), et la concentration des terres rares (1.8–57 × les teneurs chondritiques). L'origine des granites hyperalumineux de type S et des pegmatites à éléments rares de cette région serait due à un concours complexe de processus pétrogénétiques, en cinq étapes. Les éléments rares et le bore ont été concentrés dans des grès et des argiles déposés dans de vastes bassins. Ces roches ont subi un métamorphisme progressif et, en particulier, une déshydratation de la muscovite et de la biotite en l'absence d'une phase fluide; ces éléments ont donc été incorporés directement dans un bain

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fondu anatectique. Les éléments rares et les composants volatils ont ensuite progressivement été concentrés par fractionnement de cristaux primaires. Les effets d'un tel fractionnment, normalement visibles sur des diagrammes de Harker, ont été rendus obscurs par deux stades d'extraction d'un magma évolué et par des épisodes de redistribution subsolidus par lessivage des espèces cationiques. L'évolution tardi-magmatique du batholite est marquée par l'exsolution répandue d'une phase volatile, la dispersion de cette phase fluide riche en éléments rares, fluor et bore le long de zones de cisaillement, et l'émigration de systèmes magmatiques enrichis en éléments rares et en phase fluide vers le haut, à partir de la coupole que manifeste l'essaim zoné de pegmatites du Groupe de Mavis Lake.

(Traduit par la Rédaction)

Mots-clés: batholite de Ghost Lake, province du Supérieur, granite hyperalumineux, pegmatite granitique à éléments rares, Dryden, Ontario.

Introduction

One of the important goals in the study of rare-element granitic pegmatites is to attempt to identify the source(s) of the lithophile metal enrichment. In many pegmatite districts, it is difficult or impossible to identify the source materials and initial petrogenetic processes responsible for the anomalous concentration of rare elements in the derivative granitic pegmatites, as the pegmatite-forming melts have become isolated from their now unexposed parental fertile granite (*e.g.*, Bernic Lake pegmatite group: Černý *et al.* 1981). The process of physical separation of such melts also increases the probability of obscuring the chemical and isotopic legacy of the source region owing to interaction with extraneous crustal materials during the ascent of the magma systems, thus engendering "hybrid" chemical signatures. Furthermore, in those localities where rare-element granitic pegmatites can be confidently linked genetically to a parental granite (*e.g.*, Greer Lake Pegmatite Group: Černý *et al.* 1981; Black Hills Pegmatite District: Shearer *et al.* 1987), it is common to find only the chemically most fractionated end-members in the source pluton, the pegmatitic granite facies of Černý & Meintzer (1988).

In rare cases, however, a fertile pegmatitic granite facies may exist within a granitic intrusive complex that also contains consanguineous, chemically and mineralogically more primitive peraluminous granitic rocks comparable with those derived in autochthonous peraluminous anatectic complexes (*e.g.*, Peña Negra anatexites: Bea 1991). Such a relationship provides an

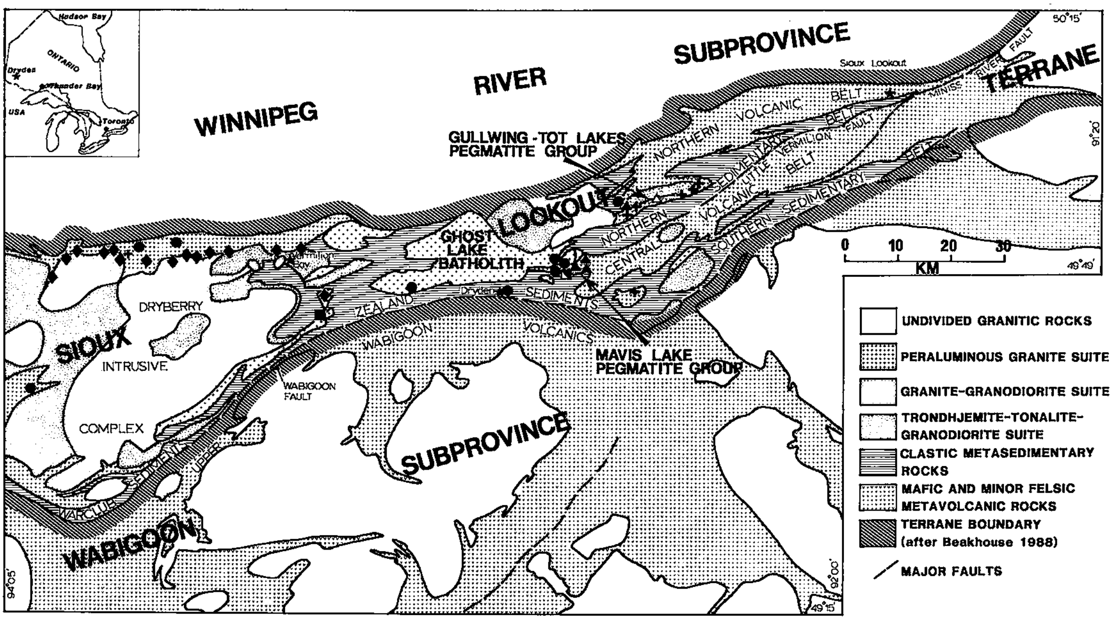

FIG. 1. Regional geological setting of the Ghost Lake batholith and lithophile element mineralization of the Sioux Lookout Terrane. Compiled from Breaks *et al.* (1978), Trowell *et al.* (1980), Sanborn-Barrie (1987), Beakhouse (1988) and Breaks (1989). Symbols for types of mineralization: filled circle: beryl type of rare-element pegmatite, filled triangle: albite–spodumene-type of rare-element pegmatite, open triangle: complex spodumene type of rare-element pegmatite, half-filled square: scheelite-bearing tourmalinite, filled square: cassiterite-bearing pegmatites, +: molybdenite-bearing pegmatites, filled diamond: uranium–thorium-rich pegmatites.

opportunity to infer the source regime and subsequent mechanisms of concentration of the various rare elements, both within the parent and its derivative rare-element pegmatites. The Late Archean Ghost Lake batholith in the Superior Province of northwestern Ontario is such an intrusive complex (Fig. 1).

The purpose of this paper is to document the geological, petrographic and chemical attributes of this peraluminous, S-type fertile granite and to propose a model to account for its derivation and that of the proximal aureole of rare-element pegmatites. A more detailed consideration of the geology, mineralogy and chemistry of the associated rare-element pegmatites will appear in a forthcoming paper (Breaks & Moore, in prep.). This present discourse is, in part, supported by a chemical data-base comprised of results of approximately 800 partial and complete bulk-rock and mineral analyses (Breaks 1989, unpubl. data).

Previous and Present Geological Investigations

Rare-element granitic pegmatites have been known in the northwest Superior Province since the early 1940s (Satterly 1941, p. 55). Spodumene-bearing pegmatites, however, were not discovered until the mid-1950s (Mulligan 1965, p. 63–65). Early geological investigations were undertaken by Moorhouse (1939), Burwash (1939), Satterly (1941) and Harding (1950). More recent geological work in the region has been carried out by Breaks *et al.* (1978), Page & Christie (1980), Trowell *et al.* (1980), Ucakuwun (1981), Sanborn-Barrie (1987), and Beakhouse (1988).

The current study involved a multi-year project marked by a progressive decrease in scale, from 1:63,360 regional reconnaissance coverage of 44,000 km^2 of the English River and Winnipeg River subprovinces (Breaks *et al.* 1978) to 1:60 mapping of individual bodies of rare-element granitic pegmatite (Breaks 1989).

Analytical Techniques

All analyses were undertaken at the Geoscience Laboratory Section of the Ontario Geological Survey. Concentrations of major and some trace elements were determined with a Philips PW 1400 X-ray-fluorescence spectrometer on fused and pressed-powder disks, respectively, with the following estimates of precision at the mid-range value: SiO_2 (0.8%), Al_2O_3 (0.3%), Fe_2O_3 (0.2%), MgO (0.3%), CaO (0.15%), Na_2O (0.5%), K_2O (0.15%), TiO_2 (0.12%), P_2O_5 (0.05%), MnO (0.015%). Proportions of other components were determined by a variety of methods: titrimetric, FeO (0.2%), infrared absorption, CO_2 (0.06%) and S (0.02%), and combustion/infrared absorption, H_2O^+ (0.2%) and H_2O^- (0.1%). Analytical methods, limits of detection and estimates of precision of the trace elements can be obtained from Depository of Unpublished Data, CISTI, National Research Council of Canada, Ottawa, Ontario K1A 0S2. Analyses for the rare-earth elements, Y and Cs, Hf and Ta (selected samples) were performed by Inductively Coupled Plasma Mass Spectrometry. Details of this technique and an estimate of precision are given elsewhere (Riddle *et al.* 1988, Doherty 1989).

Regional Geological Setting

Rare-element granitic pegmatites in the study area were initially described by Mulligan (1965), who coined the term *Dryden Pegmatite Field*. Two areally distinct clusters, some 10 km apart (Fig. 1), were named the Mavis Lake Pegmatite Group and the Gullwing Lake – Tot Lake Pegmatite Group (Breaks 1983, 1989). Because the Gullwing Lake – Tot Lake Pegmatite Group, like pegmatites to the west of Vermilion Bay (Fig. 1), are relatively distal to the Ghost Lake batholith, they will not be further described in this paper (see Breaks 1982, 1989). Regionally, the Ghost Lake batholith and this field are situated in the Winnipeg River – Wabigoon subprovincial boundary-zone, which has been recently defined by Beakhouse (1988, 1989) as the Sioux Lookout Terrane (Fig. 1). This 250 by 15–40 km zone is characterized by: (a) inverted stratigraphy and out-of-sequence thrust-stacking of metavolcanic and metasedimentary assemblages having an age between 2733±1 and 2706±2 Ma (Davis *et al.* 1988, Davis 1990, Blackburn *et al.* 1991), (b) a wide range in metamorphic grade, (c) zones of migmatized metasedimentary rocks, (d) a 150-km-long belt of peraluminous granite plutons, of which the Ghost Lake batholith is the largest, and (e) a distinctive metallogeny relative to the bounding Winnipeg River and Wabigoon subprovinces (Beakhouse 1988), characterized by enrichment in lithophile elements which, in addition to the rare elements (Li, Rb, Cs, Be, Nb, Ta and Ga), includes U, Th, Mo, W, and Sn.

Structural Setting of the Ghost Lake Batholith and Rare-Element Granitic Pegmatites

The Ghost Lake batholith is broadly concordant to the east-striking foliations within its supracrustal and granitic host-rocks. Deflections of this fabric to northerly and northeasterly strikes are, however, evident in migmatized metasedimentary rocks (Zealand unit in Fig. 1) along the western periphery of the batholith. Internally, the batholith is weakly foliated, as defined by preferred orientation of phyllosilicates, tabular enclaves of metawacke and metapelite, and segregations of biotite – cordierite – sillimanite. The intensity of foliation and the content of enclaves within unit GLB–1, described below, progressively diminish eastward; beyond Highway 601, north of Dryden, granitic rocks composing the eastern lobe of the batholith are either massive or exhibit magmatic layering. External contacts of the batholith in this region interdigitate with the host mafic metavolcanic rocks. This fact, coupled with a mild buckling of related pegmatitic granite dykes, establish that emplacement of the eastern part of the batholith was late relative to deformation in the country rocks.

Structural development of the supracrustal host-rocks

Four events of ductile deformation (D_1, D_2 and D_3) were recognized in rocks of the Sioux Lookout Terrane in the Dryden – Sandybeach Lake area (Blackburn *et al.* 1982, 1985, 1991, Breaks 1989, Chorlton 1990).

Critical field relations exposed along Highway 17 in Sanford Township afford insight into the relative timing between incipient migmatization of metasedimentary protoliths, subsequent ductile deformation, and emplacement of the chemically most primitive part of the Ghost Lake batholith. At the Eggli sheep farm in the village of Minnitaki, recumbent Z-folds of possible D_1 chronology involving peraluminous granite metatects occur in metasedimentary migmatite cross-cut by cordierite–biotite granite (GLB–1).

Emplacement of the chemically most primitive part of the batholith (GLB–1), however, can be constrained between D_1 and D_2. The latter deformation is marked by tight to isoclinal folds of Z-asymmetry that commonly exhibit a well-developed axial planar foliation and steeply plunging, southwest-trending axes. This planar fabric overprints isoclinally folded cordierite–biotite granite dykes petrographically identical to unit GLB–1 that were initially oriented discordantly to the S_0/S_1 fabric of the host metawacke–metapelite.

Major faults

A number of regionally significant faults slice through supracrustal assemblages and granitic masses proximal to the Ghost Lake batholith (Fig. 1). Structures of this kind have been viewed with importance by many investigators of pegmatites as possible conduits of water-saturated granitic magma (*e.g.*, Beus 1962, 1968, Černý 1989a). Two important faults, the Wabigoon Fault and the Little Vermilion Fault, lie near the Ghost Lake batholith and are interpreted to have had an early component of thrust movement and a later period of dextral transcurrent motion (Blackburn 1979). Evidence for structural and stratigraphic facing proximal to the Little Vermilion Fault reveals that assemblages of the Central Volcanic Belt (2733±1 Ma) were positioned over younger (<2698 Ma) clastic metasedimentary rocks of the Abram Group, possibly owing to thrusting (Davis *et al.* 1988, Blackburn *et al.* 1991). An analogous event marked by the Wabigoon Fault was hypothesized to have overstacked the Upper Wabigoon Volcanic Suite (2743±1 to 2735 $^{+4}_{-3}$ Ma: Davis *et al.* 1988) upon rocks of the 2714 Ma Warclub Group, which is correlative with the Zealand metasedimentary unit (Davis & Edwards 1982, Blackburn *et al.* 1991). The latest effects of faulting involved brittle deformation, which generated pseudotachylite in garnet-zone metawacke and caused local dextral offset of apophyses of beryl–tourmaline–muscovite granitic pegmatite related to the batholith, as at Dryden.

Rare-element granitic pegmatites

The Mavis Lake Pegmatite Group is mainly hosted in mafic metavolcanic rocks. They are confined to the northern limb of the west-plunging, D_3 Thunder Lake Syncline, where they strike parallel to the foliation of their host and exhibit local effects of late-tectonic deformation: mildly strained contacts, internal ductile shearing in quartz-rich zones, pull-apart structures ("healed fracture structure" of Jahns 1953, p. 577–578) in tourmaline and spodumene, and buckling of pegmatite dykes near the batholith contact. However, emplacement of some spodumene-bearing pegmatites, particularly those classified as of the albite type (Černý 1989a; see below), outlasted the late-tectonic deformation, as testified by their sharply discordant, planar contacts and absence of signs of ductile deformation.

Regional Metamorphic Setting of the Ghost Lake Batholith and Rare-Element Granitic Pegmatites

The southern margin of the Sioux Lookout Terrane in the Dryden area is the locus of a steep (50°C/km) metamorphic gradient that increases northward from the low-grade Wabigoon Subprovince and was inferred from comparison of data on mineral assemblages with experimentally determined conditions for relevant reactions (Breaks 1989). An intact facies-series in metapelite and metawacke of the Zealand Group is disposed parallel to the southern margins of the Sioux Lookout Terrane (Figs. 1, 2). It is of the low pressure – high temperature Abukuma style of regional metamorphism widespread in the northwestern Superior Province (Ayres 1978, Ermanovics & Froese 1978, Thurston & Breaks 1978). Metamorphic conditions, as estimated from data on mineral assemblages (Breaks *et al.* 1978, Bartlett 1978) and thermobarometric work (Campion *et al.* 1986), indicate a temperature range of 550 to 750°C at a constant pressure of about 0.4 GPa.

The pattern of metamorphic zones adjacent to the Ghost Lake batholith is notably asymmetrical (Fig. 2). High-grade migmatitic clastic metasedimentary rocks are situated only beyond the western to northwestern periphery of the batholith. In contrast, host rocks to the east and most of the southern contact belong to the medium-grade sillimanite–muscovite zone. A further notable feature is the slight discordance between the southern contact of the batholith and this metamorphic zone by post-metamorphic emplacement of the batholith. No contact-metamorphic aureole is associated with the Ghost Lake batholith, in contrast to Archean peralu-

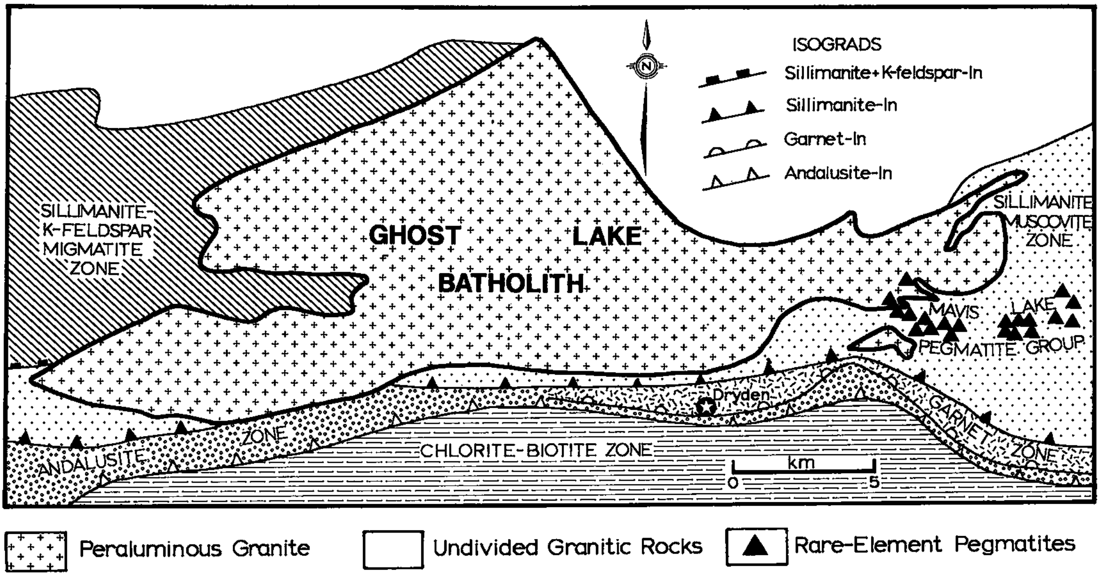

FIG. 2. Distribution of regional metamorphic zones and isograds in relation to the Ghost Lake batholith and rare-element pegmatites of the Mavis Lake Group.

minous granite plutons in the Yellowknife pegmatite district (Kretz *et al.* 1981, 1989) or in some Hercynian terranes (*e.g.*, Holtz & Barbey 1991). The Mavis Lake Pegmatite Group is situated within the sillimanite–muscovite zone, which is apparently an uncommon metamorphic zonal setting (Černý & Meintzer 1988). Elsewhere, this is exemplified by the Yellowknife Pegmatite District and part of the Black Hills Pegmatite District (Redden & Norton 1975, Shearer *et al.* 1987, and Helms & Labotka 1991).

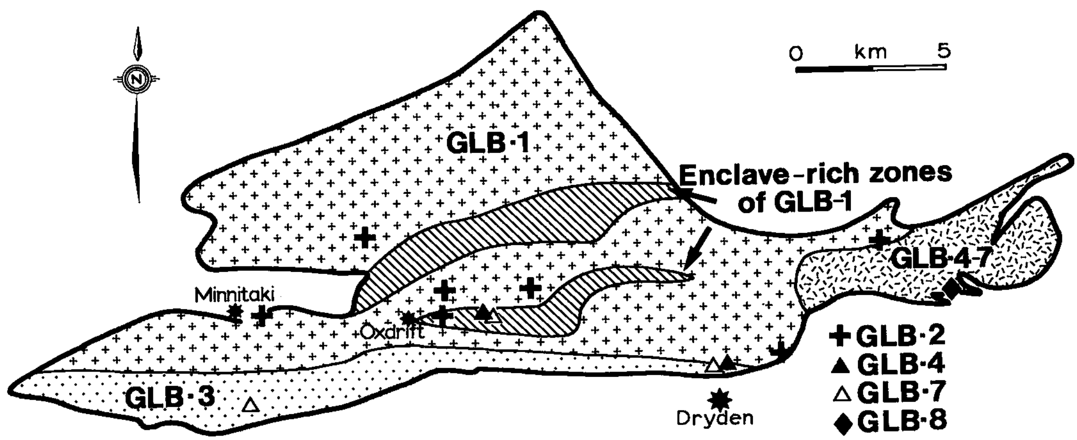

FIG. 3. Distribution of internal granitic units in the Ghost Lake batholith.

TABLE 1. SUMMARY OF FIELD AND PETROGRAPHIC FEATURES OF INTRUSIVE UNITS OF THE GHOST LAKE BATHOLITH AND RELATED METASOMATIC ROCKS

Intrusive Unit	Rock Type(s)	Petrographic Features	Varietal and Accessory[1] Mineralogy	Field Relations
GLB-1	Granite, rare granodiorite, quartz syenite and leucocratic quartz diorite. Small percent of biotite-cordierite pegmatitic leucogranite and biotite potassic pegmatite segregations.	General weakly foliated to massive. Also compositionally layered. Hypidiomorphic-granular, inequigranular textures prevalent.	Biotite, cordierite (apatite, zircon, monazite). Secondary muscovite and fibrolite are common. Rare garnet.	Shallow contact with host metatexite metasedimentary migmatite which is injected by apophyses of cordierite-biotite granite. Emplacement of GLB-1 therefore post-dates early migmatization and recumbent folding of granite metatects.
GLB-2	Granite to granodiorite	Massive and occasionally weakly foliated; fine- to medium-grained, allotriomorphic-granular. Overprinted by muscovite porphyroblasts.	Biotite, muscovite	Occurs as small masses and dykes cross-cutting GLB-1. GLB-2 intruded by veins of apatite-biotite potassic pegmatite inferred to be deep-level equivalents of GLB-7.
GLB-3: First-Stage Fertile Granite	Granite which locally grades into segregations of biotite-muscovite pegmatite leucogranite and potassic pegmatite. Quartz monzonite is rare and possibly of cumulate origin.	Weakly foliated to massive. Occasionally lineated. Medium-grained, equigranular, allotriomorphic-granular. Foliation overprinted by muscovite porphyroblasts and quartz-fibrolite-muscovite nodules.	Biotite, muscovite (garnet)	No observed intrusive relations with GLB-1 and -2. Cut by dykes of beryl - bearing GLB-7.
GLB-4 to -8: Second-Stage Fertile Granite				
GLB-4	Pegmatitic leucogranite	Massive and locally layered with sodic aplite. Characterized by blocky to irregular megacrysts of graphic K-Feldspar-quartz and plumose intergrowths of muscovite-quartz. Rare unit in Ghost Lake batholith.	Muscovite, biotite (garnet, tourmaline)	Rare unit in GLB and virtually confined to eastern batholith lobe. Dykes of similar rock which cut GLB-3 occur elsewhere as near Dryden.
GLB-5	Fine-grained leucogranite	Massive and locally layered with potassic pegmatite (GLB-6 and -7) or rarely with orbicular aplite. Fine- to medium-grained, allotriomorphic-granular.	Tourmaline (muscovite, garnet)	Intruded by dykes of GLB-7 and -8.
GLB-6	Muscovite-rich potassic pegmatite	Massive and locally layered with GLB-5 and rare orbicular aplite; coarse-grained, inequigranular, hypidiomorphic-granular.	Muscovite (tourmaline, biotite, garnet)	Age relations with GLB-5 are equivocal; this unit is layered with and intrusive in GLB-5.
GLB-7	Potassic pegmatite	Massive and locally layered with sodic aplite; coarse-grained, inequigranular, hypidiomorphic-granular.	Muscovite, tourmaline (garnet, biotite, beryl)	Dykes and irregular masses cut GLB-5.
GLB-8	Potassic pegmatite	Massive, coarse-grained, equigranular, hypidiomorphic-granular.	Toumaline, muscovite	Youngest unit of GLB. Occurs as 3-8 cm wide dykes which cut GLB-7 and layered aplite.
Metasomatic units				
Fibrolite-Rich Veins	Quartz-muscovite-biotite-fibrolite rock	Commonly folded and lineated. Fine- to medium-grained veins typically <3 cm thick.	Biotite, muscovite (apatite, dumortierite)	Developed repeatedly during migmatization of host metasedimentary rocks and crystallization of intrusive units GLB-1, -2, and -3.
Synmagmatic Contact Metasomatic Zone	Altered mafic metavolcanic rocks	Schistose, lineated mafic rocks; fine- to coarse-grained, granoblastic porphyroblastic	Dravite, holmquistite, muscovite, phlogopite, hornblende, actinolite (anthophyllite, chlorite, pyrite, zircon, rutile)	Zone is cut by GLB-7 and -8, folded to planar, beryl albitite dykes, and small exocontact masses of beryl pegmatitic granite.

1. accessory minerals (<1 percent) placed in parentheses

PETROGRAPHY OF THE GHOST LAKE BATHOLITH

The Ghost Lake batholith contains eight internal units (Fig. 3) divided on the basis of petrography and field relations (Table 1).

GLB–1

Most of the batholith consists of GLB–1, a white weathering, massive to modestly foliated, biotite and cordierite–biotite granite that is typically inequigranular

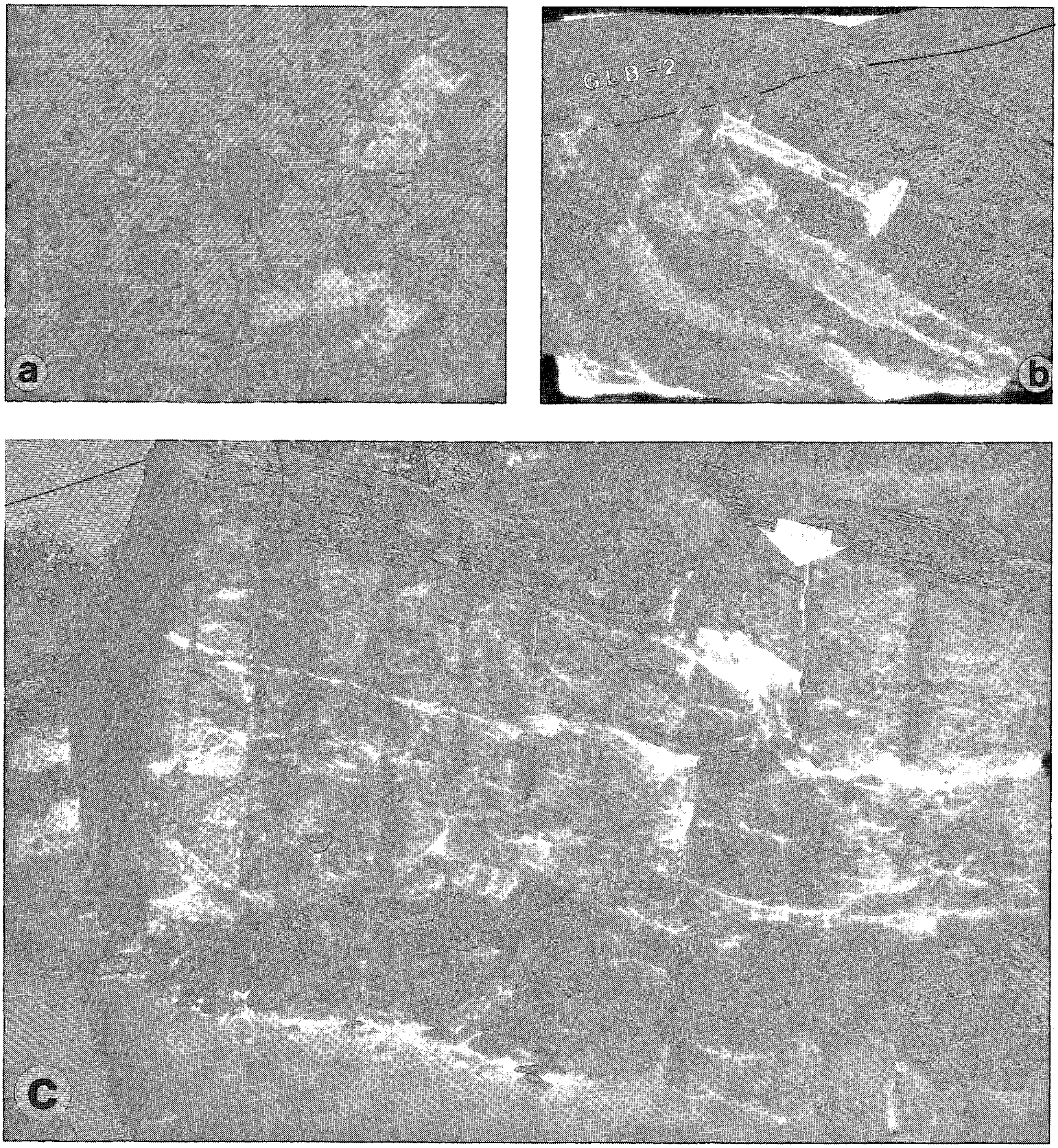

FIG. 4. Western part of the Ghost Lake batholith. a. Inequigranular granite, typical of GLB–1, with rectangular phenocrysts of cordierite. Barney Haukness farm in Eton Township. Coin diameter: 1.7 cm. b. Typical exposure of metasedimentary-enclave-rich part of GLB–1, the foliation of which is cut by a discordant dyke of GLB–2. Highway 17 road-cut at Oxdrift. c. Contact between cordierite–biotite granite associated with pegmatitic leucogranite and flatly disposed migmatitic metasedimentary rocks (metatexite). Arrow points to D_1 isoclinal fold in leucosome. Location near the Aubrey Creek bridge on Highway 17.

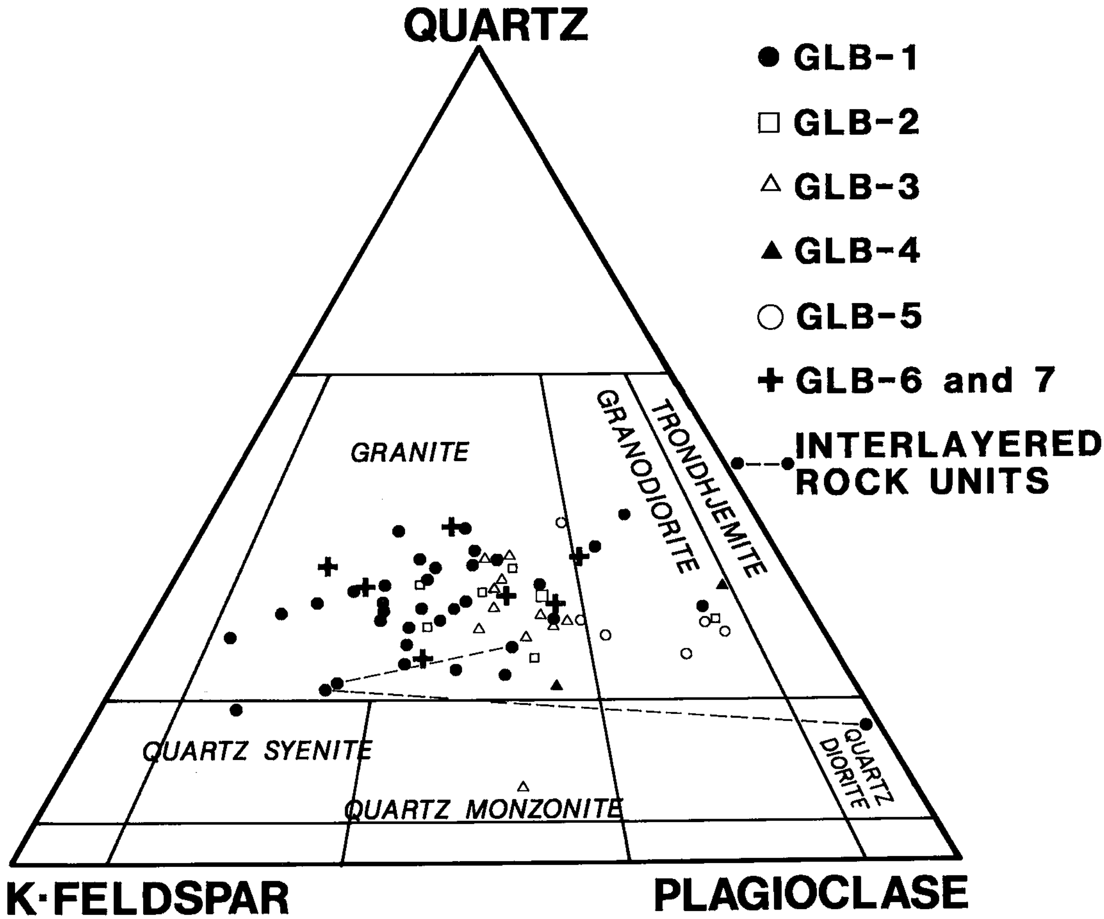

FIG. 5. Modal variation of quartz, plagioclase and K-feldspar in granitic units of the Ghost Lake batholith.

and coarse grained (Figs. 4a, 5). Granodiorite, quartz syenite and leucocratic quartz diorite are rare; the latter occurs as 0.3-m-wide injections into, and border zones around, rare mafic inclusions (*e.g.*, Oxdrift: Fig. 3) that cryptically grade into medium-grained layers of apatite–biotite granite. The margins of such hornblende–plagioclase mafic blocks contain a 1- to 3-cm-wide biotite-rich selvedge that was either induced by metasomatic interchange with a volatile-rich granitic melt precursor to GLB–1 or by subsolidus processes.

Two elongate zones of biotite granite occur near the central part of GLB–1 (Fig. 3) and contain abundant enclaves of migmatized metawacke and metapelite (Fig. 4b) identical to high-grade rock types found within the adjacent Zealand metasedimentary suite (Fig. 1). Intrusive contacts of GLB–1 with these host rocks are exposed only rarely and are generally shallowly dipping (Fig. 4c).

Pegmatite segregations occur sporadically throughout GLB–1 and most commonly consist of biotite and cordierite–biotite pegmatitic leucogranite. These contain irregular to euhedral, graphic microcline perthite (bulk composition $Or_{74}Ab_{26}$) – quartz megacrysts up to 25 by 60 cm within a plagioclase (An_{10-25}) – quartz – biotite matrix. Less common are apatite–biotite potassic pegmatite pods that grade into a smoky, quartz-rich core up to 0.4 by 1.5 m, into which project blocky perthitic microcline of bulk composition $Or_{75}Ab_{25}$.

Cordierite [Mg/(Mg+Fe) = 0.41] locally constitutes up to 5% of GLB–1 and occurs in elongate and square megacrysts (Fig. 4b) up to 1 by 3 cm and 2 cm across, respectively. It also commonly forms roughly circular intergrowths up to 1.5 m diameter with a subequal amount of quartz. These are similar to graphic intergrowths in peraluminous granites of the English River Subprovince (Breaks 1991) and in younger terranes

(West Moravia: Černý & Povondra 1967, Appalachian Piedmont: Speer 1981, French Massif Central: F. Holtz, pers. comm., 1992). Cordierite is variably replaced by a complex symplectite of chlorite, muscovite, andalusite and "fibrolite".

Accessory minerals in GLB–1 include garnet, apatite, zircon, monazite, sillimanite and rare sagenitic rutile in biotite. Opaque iron oxides are typically absent. Muscovite is largely of secondary origin after microcline, plagioclase, biotite and cordierite.

GLB–2

Unit GLB–2 comprises rare, narrow (<5 m), light grey, weakly deformed dykes of biotite granite and granodiorite mainly confined to the western part of GLB–1 (Fig. 4b) and adjacent metasedimentary migmatitic host-rocks (Fig. 3). These dykes, distinguished from GLB–1 by a finer grain-size and light grey plagioclase (An_{18-26}), are typically flat-lying, weakly foliated and may exhibit open Z-folds. The foliation is commonly interrupted by muscovite porphyroblasts. Temporal relations with GLB–3 are not known; however, it is possible that these dykes were emplaced after GLB–3 because deformed "fibrolite"-rich veins (see below), which occur in GLB–1 and –3, are cross-cut by GLB–2. Because GLB–1 and –2 represent the chemically least evolved parts of the batholith (see Petrochemistry, below), they will, in some of the discussion to follow, be referred to as relatively "primitive".

GLB–3

This unit is confined to a 25 by 0.5–2 km strip

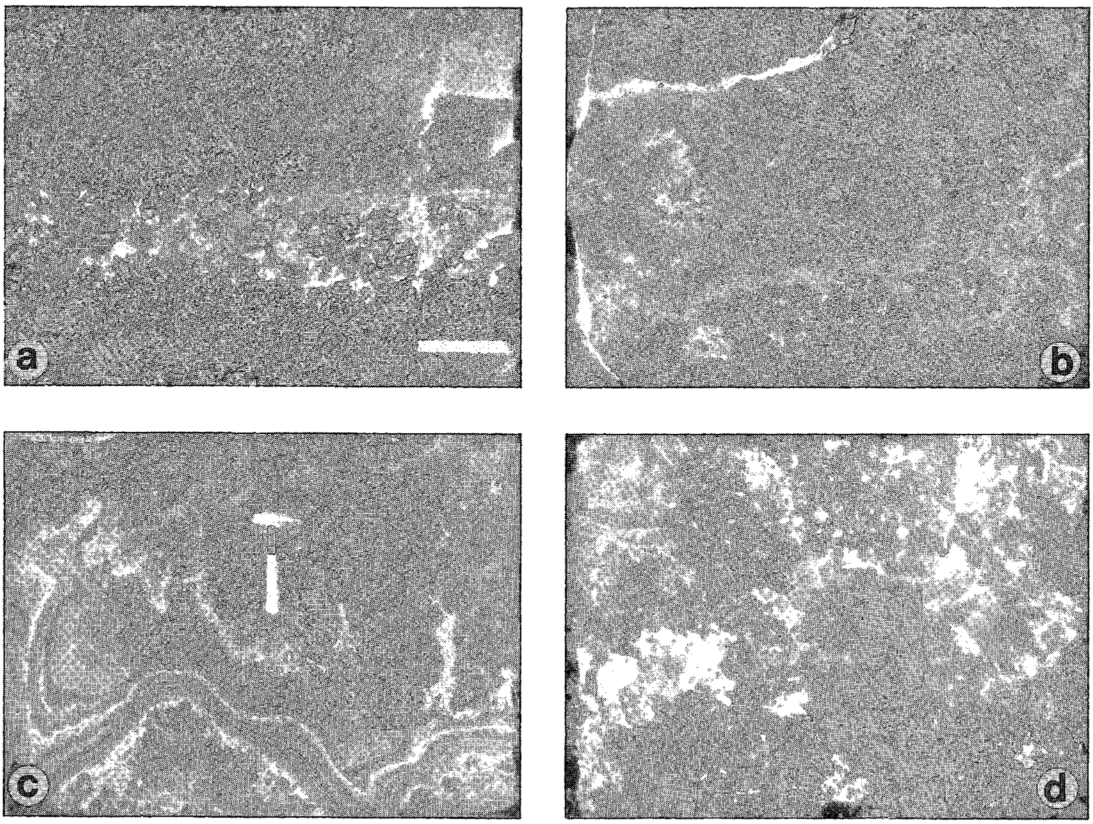

FIG. 6. Fertile granite and early metasomatic units. a. Biotite–muscovite granite typical of unit GLB–3, with gradational pods of potassic pegmatite. Bar: 10 cm. b. Layer of fine-grained leucogranite (GLB–5) truncated by muscovite-rich potassic pegmatite (GLB–6) at the Contact Beryl Occurrence. The contact has served as a substrate for the directional growth of muscovite in GLB–6. Coin diameter: 2.3 cm. c. Potassic pegmatite (GLB–7) featuring elongate megacrysts of microcline oriented normal to convoluted layering with sodic aplite at the Contact Beryl Occurrence. d. Blocky megacryst of microcline that projects into a quartz-rich segregation in beryl – garnet – muscovite – tourmaline potassic pegmatite at the Dryden Airport exposure. Coin diameter: 2.3 cm.

contiguous to the southern contact of the batholith west of Dryden (Fig. 3). It mainly consists of white, massive and weakly foliated, medium-grained biotite and muscovite–biotite granite (Fig. 5) containing a generally more sodic plagioclase (An_{3-17}). It may be distinguished from GLB–1 to the north by its finer grain-size, allotriomorphic-granular texture, absence of cordierite, and a relative scarcity of metawacke and metapelite enclaves and mafic schlieren. Segregations of pegmatitic leucogranite and potassic pegmatite with megacrysts of white to orange intermediate microcline perthite (obliquity = 0.84) of bulk composition $Or_{73}Ab_{27}$ are ubiquitous in GLB–3 (Fig. 6a) and are likely consanguineous by virtue of the gradational grain-size into the host and a common suite of accessory minerals (biotite, garnet and sillimanite).

Near the contact with andalusite-zone metasedimentary rocks, the foliation and a local lineation defined by biotite are more strongly developed. This zone also is the locus of muscovite porphyroblasts and sillimanite – muscovite – quartz nodules that overprint these structures. Narrow veins (1–10 mm) rich in "fibrolite" cross-cut the foliation and pegmatitic segregations in GLB–3 and may have formed contemporaneously with the nodules.

Age relations of GLB–1 and –2 with GLB–3 are uncertain. Dykes and irregular masses of GLB–4 and –7 sharply intrude foliated GLB–3 in several locations. Furthermore, it can be established that GLB–3 preceded the generation of the rare-element mineralization, as dykes of tapiolite–beryl potassic pegmatite sharply cross-cut GLB–3 near Minnitaki (Fig. 4). In view of the chemically more fractionated nature of GLB–3 relative to GLB–1 (see below) and the age relationships, we conclude that GLB–3 was emplaced during an intermediate stage, between GLB–1 and the main mass of pegmatitic granite (GLB–4 to –8). The portions of the batholith that contain GLB–3 and GLB–4 to –8 both have chemical features diagnostic of fertile granites (Černý & Meintzer 1988) and will be subsequently be designated as fertile granites of stage 1 and stage 2. We acknowledge that GLB–3 could have been derived from a source region and time distinct from GLB–1 and –2, with the two masses subsequently and fortuitously amalgamating into an ostensibly comagmatic, discrete pluton. However, the observed chemical trends (*e.g.*, Figs. 13–16, below) are characterized by continua displaying no significant gaps in the data, from which we infer consanguinity for the various internal units of the batholith.

GLB–4 to –8

The eastern lobe of the batholith (Fig. 3) consists predominantly of a complex of pegmatitic granite units, here divided according to the classification of Černý & Meintzer (1988): pegmatitic leucogranite (GLB–4), fine-grained leucogranite (GLB–5) and several types of potassic pegmatite (GLB–6, –7 and –8). Units GLB–4, –6 and –7 may be interlayered with or truncated by subordinate sodic aplite.

The eastern lobe is characterized by the virtual absence of metasedimentary enclaves, mafic segregations and a striking change in accessory minerals relative to GLB–1. The pegmatitic granites contain a conspicuous abundance of tourmaline and books of primary muscovite, coupled with a rarity of biotite and an absence of zircon, apatite and monazite. Plagioclase (An_{3-15}), as a discrete phase, is typically albite. Furthermore, the appearance of beryl and cassiterite, and modest to extreme enrichment in rare alkalis in muscovite and K-feldspar (Cs up to 1000 ppm in blocky microcline: Breaks 1989), indicate that these rocks can be classified as fertile pegmatitic granites as defined by Černý & Meintzer (1988).

Unit *GLB–4*, the least abundant, is chiefly distinguished by euhedral to irregular graphic megacrysts of perthitic microcline and quartz set within a matrix of albite, quartz, muscovite and minor biotite, garnet and tourmaline. Plumose aggregates of muscovite and quartz are commonplace and do not occur in the remaining units.

Unit *GLB–5* is widespread in the Dryden airport area southward to the batholith contact (Fig. 3). It is massive and commonly layered with potassic pegmatite, widely equigranular, and dominantly granodioritic in composition (Fig. 5). It may occur as layers 0.5 to 2 m in thickness that are locally laminated and that alternate with potassic pegmatite (GLB–6 and –7), and rare aplite containing tourmaline – muscovite – albite – quartz orbicules. It is also cross-cut by dykes and small masses of later potassic pegmatite (GLB–6 and 8: Fig. 6b). Tourmaline, commonly 10–15%, and lesser muscovite and garnet also are present.

Unit *GLB–6*, a muscovite-rich potassic pegmatite, contains homogeneous to locally graphic K-feldspar megacrysts from 2 to 16 cm in length embedded in a matrix with abundant muscovite (up to 35% of matrix) that is commonly intergrown with biotite in bladed aggregates, and lesser quartz, albite and tourmaline. Its distribution is restricted to outcrops at the Dryden Airport and the Contact Beryl Occurrence (described below). It may be a variant of GLB–7, as intrusive relations between the two were not encountered. Age relations with GLB–5 are equivocal, as it is both interlayered and intrusive into this unit.

GLB–7 occurs as 0.5- to 2-m-thick units repeatedly interlayered with sodic aplite (Fig. 6c), or more commonly, as dykes intruding GLB–5. It is characterized by blocky microcline projecting into quartz-rich internal pockets (Fig. 6d), by the abundance of coarse tourmaline, and by the presence of rare-element minerals such as beryl.

GLB–8 is the youngest internal unit in the batholith and is found only at the Contact Beryl Occurrence (Fig.

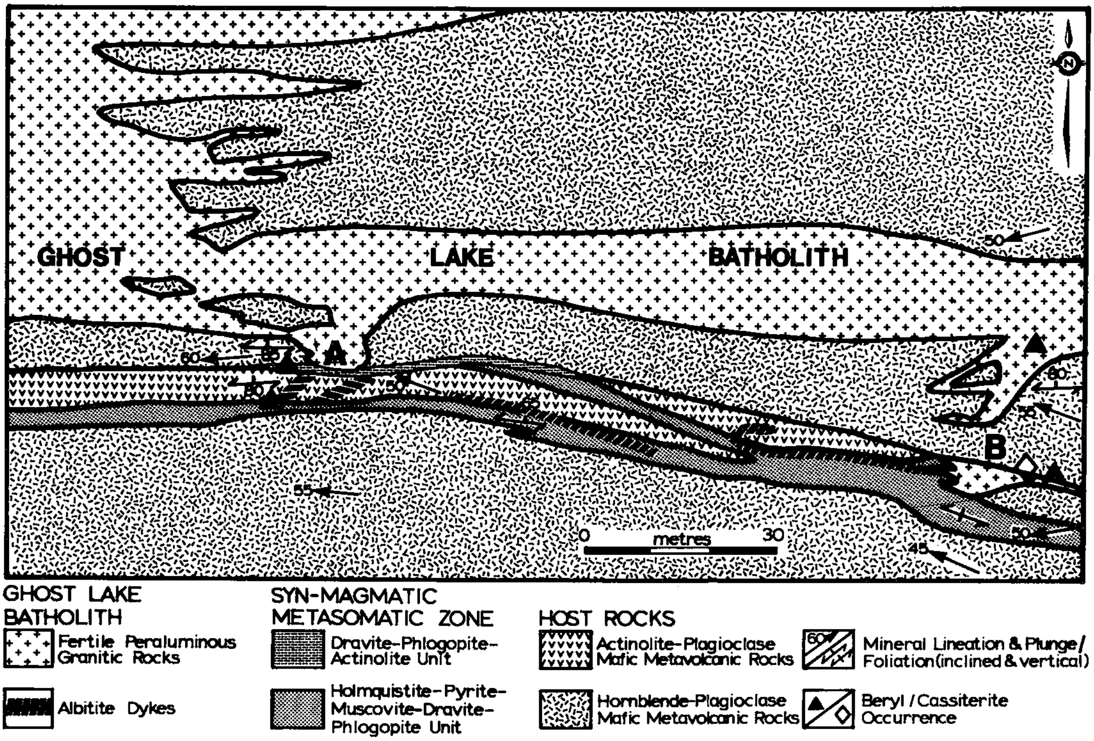

FIG. 7. Detailed geology of the contact zone between the Ghost Lake batholith and mafic metavolcanic host-rocks at the Contact Beryl Occurrence.

7) as a 3- to 8-cm-wide dyke cross-cutting GLB–7 and layered sodic aplite.

Zealand stock

This 0.8 by 2.7 km ovoid pluton (Fig. 9, below), first mapped by Satterly (1941), contains a suite of intrusive units petrographically similar to GLB–5 to –8. Thus, it is interpreted as an apophysis of nearby fertile granites of the Ghost Lake batholith, an inference also supported by similarities in indices of chemical fractionation (Breaks 1989).

METASOMATIC ROCKS

Several types of metasomatic rocks are associated with many internal units of the Ghost Lake batholith, and also occur in migmatized metasedimentary rocks and mafic metavolcanic host-rocks.

"Fibrolite"-rich veins

Thin sheets (<3 cm) of grey to porcellaneous white, "fibrolite" – biotite – muscovite containing minor dumortierite (identified by X-ray diffraction) or apatite or both are found within numerous exposures of enclave-rich GLB–1 (Fig. 8a, below) and adjacent migmatitic host-rocks and also occur in GLB–3. Field relations show that these tightly folded and highly lineated veins comprise at least three generations. They are not only associated with flat-lying shear zones along the base of recumbent folds in migmatitic metasedimentary rocks near the batholith contact, as at Minnitaki (Fig. 3), but are also found as earlier veins deformed by such folds. Nearby apatite–biotite granite (GLB–1) also contains contorted veins of apatite – muscovite – biotite – "fibrolite" that could be coeval with those controlled by the shear zones. The youngest veins cross-cut weakly foliated, muscovite–biotite granite bearing sillimanite–quartz nodules and related pods of potassic pegmatite of unit GLB–3 in Aubrey Township (49°47′08″N, 93°05′00″W). Dykes of beryl – tourmaline – biotite – muscovite potassic pegmatite (GLB–7) at the same exposure cross-cut GLB–3, but are not transected by the "fibrolite"-rich veins.

In summary, the "fibrolite"-rich veins affected several units of the Ghost Lake batholith including the first-stage fertile granite (GLB–3), as well as the mig-

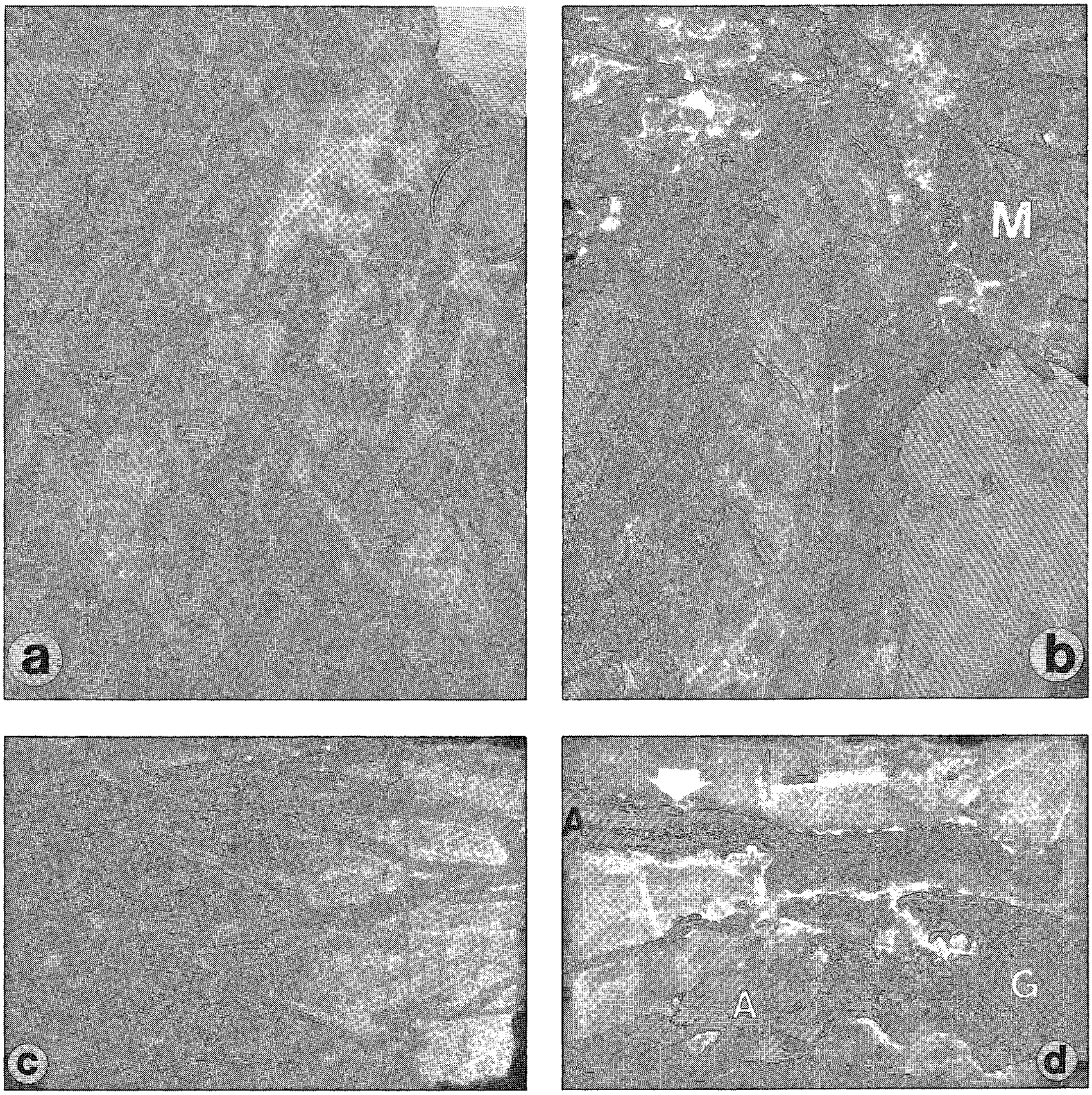

FIG. 8. Metasomatic rocks associated with Ghost Lake batholith. a. Deformed network of anastomosing veins of muscovite – biotite – "fibrolite" (dark area) hosted in GLB–1 at Oxdrift on Highway 17. Section was cut normal to the vein lineation. Coin diameter: 1.7 cm. b. Dravite – phlogopite – actinolite subunit of the zone of synmagmatic alteration, marked by hammer, which is intruded by potassic pegmatite of the Ghost Lake batholith (lower right). Host mafic metavolcanic rocks are denoted by "M". c. Net-vein system composed of holmquistite – biotite – plagioclase contained within actinolite-bearing mafic metavolcanic rocks situated at the periphery of the zone of synmagmatic metasomatism (*i.e.*, immediately adjacent to Fig. 8a). Coin diameter: 2.6 cm. d. Transition of cassiterite – albite – beryl greisen (G) into two dykes of cassiterite – beryl – tourmaline albitite (A) located at the periphery of the exocontact mass of pegmatitic granite (locality B in Fig. 7). Note the discontinuous tourmaline-rich margins, one of which is marked by the arrow.

matitic metasedimentary rocks structurally overlying GLB–1. A process such as subsolidus leaching spurred by a flux of mildly acidic fluid (Korikovskiy 1965, Losert 1968, Vernon 1979) may have been involved in the genesis and distribution of rare elements and associated volatiles in these rocks (see below). Similar secondary units occur in other peraluminous granite complexes, such as the South Mountain batholith, where Corey (1988) and Corey & Chatterjee (1990) have termed this "high-alumina hydrothermal alteration".

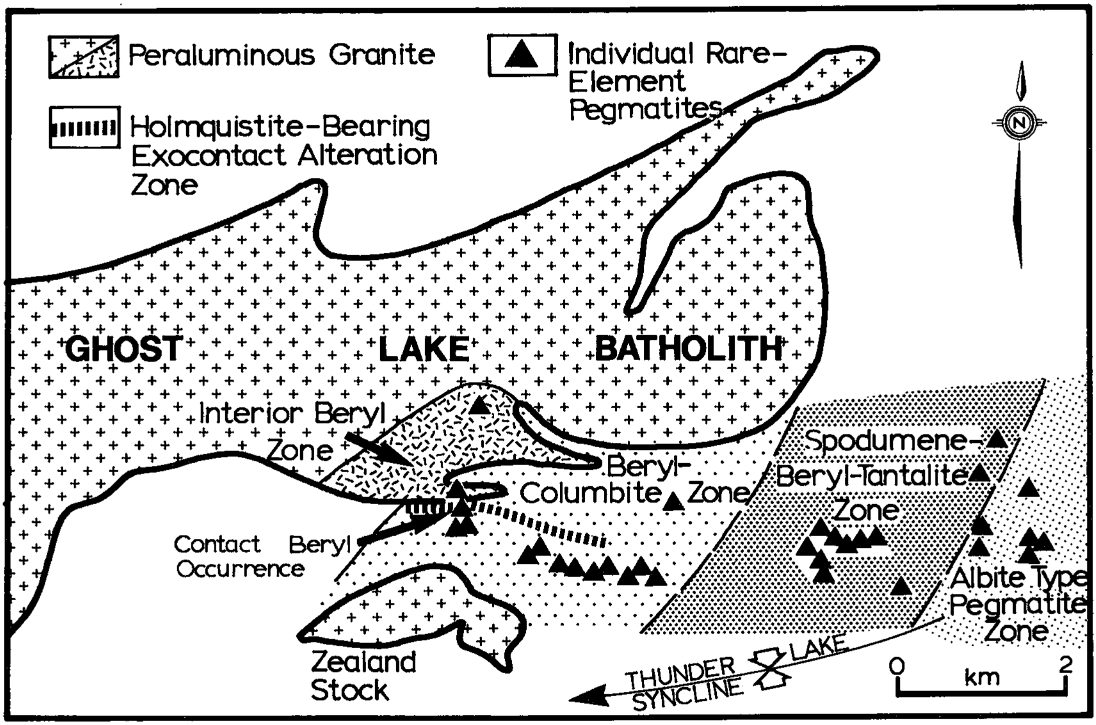

FIG. 9. Distribution of rare-element-enriched zones of the Mavis Lake Group in relation to the Ghost Lake batholith.

Synmagmatic endo- and exocontact metasomatic rocks

At the Contact Beryl Occurrence (Figs. 7, 8b, c, d, 9), there is a vein system consisting of several subunits situated at the contact between the Interior Beryl Zone of the Ghost Lake batholith and the host mafic metavolcanic rocks. This system is characterized by holmquistite, a Li-bearing amphibole normally restricted within 20 m of Li-rich pegmatites (Černý *et al.* 1981, London 1986b), but here occurring 3.5 km from the nearest known body of spodumene pegmatite of the Mavis Lake Group. Holmquistite occurs in the main subunit of the vein system (holmquistite – pyrite – muscovite – dravite – phlogopite rock in Fig. 7), a 2–10-m wide body that has been traced within a shear zone for at least 2.5 km to the east and also occurs in local net-veins (Fig. 8c). The second subunit, a dravite – phlogopite – actinolite rock, is locally cut by GLB–7 (locality A in Figs. 7 and 8b), whereas the gradational main subunit in turn is transected at locality B in Figure 7 by exocontact pegmatitic granite units correlative with GLB–5 and –6. That this exomorphic metasomatism is synmagmatic is supported by the presence of small inclusions containing holmquistite and anthophyllite in primary muscovite of GLB–7, which apparently originated from the main subunit of the vein system.

These synmagmatic units have a chemical signature enriched in Li–Rb–B–F–Cs–Sn (see Petrochemistry, below) that is similar to that of the late-magmatic alteration described in the next section. Such metasomatic rocks have not been widely recognized in the Superior Province and best compare with the holmquistite-bearing mafic metasomatic rocks in Russia (Gordiyenko *et al.* 1975).

A second episode of alteration is also established by field relations at the Contact Beryl Occurrence. This is recognized at nearby, small exocontact masses of garnet – muscovite – tourmaline pegmatitic granite (locality B in Fig. 7), which cross-cut the synmagmatic metasomatic zone, particularly by an albitite apophysis that extends 30 m west of this mass. These apophyses consist of cassiterite – beryl – tourmaline albitite and a subordinate muscovite-rich rock that possibly corresponds to the mica-rich class of greisens (Kuhne *et al.* 1972). The latter occur as peripheral pods up to 0.4 by 4.5 m defined by a general depletion in quartz, commencing from unaltered, layered muscovite-rich potassic pegmatite and gradational into an irregular zone of green muscovite – beryl – cassiterite – chrysoberyl greisen, which in turn changes into a peripheral zone of tourmaline – beryl – cassiterite albitite. The albitites may be similar to the type-IIA episyenites associated with Hercynian uranium-bearing peraluminous granites of western Europe (Cathelineau 1986, 1987). There are, however, little of the original magmatic textures remain in the study area; we are not aware of documentation concerning an association of type-IIA episyenites with greisen.

An alternative means of generating the peripheral albite-rich units was advanced in the classic study of London (1986a). He concluded that rare-element albitites can be primary units and conceivably resulted from the destabilization of a hydrous, alkali borosilicate fluid triggered by tourmaline crystallization. Only the bodies of tourmaline – cassiterite – beryl albitite (Figs. 7, 8d), described above, appear to be reasonably explicable by this mechanism. These masses, zoned by tourmaline concentrated along its margins, could have originated by the former fluid becoming depleted in boron, leaving the residue to crystallize as an inner albitite core enriched in rare elements and Sn. The beryl – tourmaline – quartz veins interconnected with the albitite masses would therefore logically represent crystallization of a liberated aqueous fluid. Sporadic miarolitic cavities containing euhedral beryl testify to local saturation in volatiles, which may relate to a decrease in H_2O solubility due to

TABLE 2. REPRESENTATIVE CHEMICAL COMPOSITIONS OF INTERNAL UNITS OF THE GHOST LAKE BATHOLITH AND LEUCOSOME FROM ADJACENT MIGMATIZED METASEDIMENTARY ROCKS

GLB Unit	1	1	1	1	1	2	3	3	4	5	6	7	8	Aplite Layer	Fibrolite Vein	Normal Zealand Leucosome	Fertile Zealand Leucosome
Sample #	C15-5	C20-3	D14-6B	D18-2A	E17-8B	E19-5	F9-2	F10-1	E29-1	E31-1	88-29	88-32	88-35	88-34		88-50	88-52
SiO_2	76.4	73.3	75.0	75.3	71.2	71.9	74.5	74.7	74.8	75.6	75.8	73.0	73.6	70.0	48.1	73.2	76.3
Al_2O_3	14.9	15.4	14.7	14.9	16.9	16.2	15.3	15.2	15.0	14.8	14.2	15.4	15.8	16.5	30.0	15.5	17.9
FeO	0.16	0.73	0.56	0.51	0.73	1.05	0.48	0.48	0.40	0.18	0.47	0.13	0.33	1.93	4.92	0.27	1.13
Fe_2O_3	0.00	0.00	0.03	0.03	0.28	0.32	0.00	0.05	0.00	0.53	0.46	0.12	0.63	0.35	1.93	0.12	0.21
MgO	0.04	0.32	0.13	0.14	0.25	0.49	0.05	0.06	0.05	0.03	0.08	0.15	0.09	0.05	2.28	0.12	0.48
CaO	1.94	0.85	0.83	0.81	0.92	2.32	0.81	0.71	0.52	0.39	0.13	0.13	0.20	0.35	0.16	2.45	1.10
Na_2O	3.73	1.98	2.13	1.74	1.72	3.95	3.60	4.12	4.97	4.64	3.70	3.44	4.02	7.67	0.37	3.91	1.76
K_2O	2.80	6.06	5.90	6.30	6.35	2.40	4.45	3.87	3.62	3.41	3.43	6.90	4.12	0.47	5.63	2.56	0.28
TiO_2	0.24	0.35	0.22	0.37	0.36	0.47	0.20	0.17	0.11	0.23	0.01	0.02	0.01	0.02	0.81	0.01	0.01
P_2O_5	0.12	0.11	0.11	0.12	0.11	0.10	0.09	0.14	0.16	0.11	0.05	0.10	0.09	0.08	0.07	0.19	0.13
MnO	0.01	0.01	0.01	0.00	0.01	0.01	0.02	0.02	0.01	0.04	0.05	0.01	0.02	1.22	0.07	0.01	0.13
CO_2	0.08	0.22	0.22	0.12	0.19	0.38	0.06	0.15	0.27	0.14	0.27	0.11	0.15	0.13	0.17	0.08	0.09
S	0.01	0.01	0.01	0.01		0.01			0.01	0.01	0.01	0.01	0.01		0.01	0.01	0.02
B_2O_3										0.74	0.15	0.07	0.36	0.36			0.07
F	0.02	0.03	0.03	0.02	0.04	0.05	0.03	0.03	0.03	0.04	0.04		0.03	0.03	0.08		
H_2O+	0.18	0.19	0.39	0.26	0.48	0.18	0.29	0.16	0.27	0.14	0.53	0.29	0.50	0.31	4.12	0.14	0.49
H_2O-	0.10	0.09	0.12	0.05	0.07	0.06	0.05	0.05	0.06	0.05	0.05	0.11		0.12	0.32		0.10
Total	100.73	99.65	100.39	100.68	99.61	99.89	99.36	99.91	100.28	101.08	99.43	99.99	99.96	99.59	99.04	98.57	100.20
Li	4	10	4	7	9	22	44	52	25	16	118	54	50	54	30	6	62
Be	<1	<1	<1	<1	<1	1	2	4	2	3	4	2	4	5	3	<1	99
Rb	80	190	140	150	140	80	270	280	300	390	652	1150	621	214	319	34	7
Sr	165	265	230	190	365	835	35	35	12	12	12	17	12	7	13	516	127
Cs		1.7	1.7		2.8	2.6	7.7			32	15	22	16	7	37	0.91	5.1
Ba	300	1020	1160	1070	1460	1210	240	90	20	20	<10	<10	13	<10	52	1790	211
Pb	32	37	37	49	42	14	35	25	13	17	<10	<10	<10	<10	73	45	12
La	7.4	11	40	13	42	31	9.6	6.2	1	9.6	0.46	0.39	0.69	0.75	175	9.2	2.1
Ce	13	21	83	25	81	59	18	13	2	18	0.93	0.29	1.5	2.3	385	16	4.1
Pr	1.3	2.3	9.7	3	8.9	6.3	1.9	1.5	0.23	1.4	0.09	<0.05	0.17	0.33	39	1.8	0.47
Nd	4.3	8.2	31	9.3	32	24	7.2	4.9	0.77	7.2	0.38	0.13	0.67	1.4	135	6.7	2.1
Sm	1.1	1.7	6.3	2.1	6.8	3.6	1.9	1.5	0.32	2.6	0.25	<0.05	0.39	1.1	34	1.6	0.47
Eu	0.64	0.80	0.89	0.87	0.95	1	0.22	0.12	<0.05	<0.05	<0.05	<0.05	<0.05	<0.05	0.49	1.5	0.48
Gd	0.94	1.5	4	1.3	4.8	1.9	1.7	1.3	0.27	2.5	0.20	<0.05	0.28	1.4	28	1.4	0.46
Tb	0.13	0.21	0.47	0.16	0.54	0.19	0.28	0.23	0.05	0.55	<0.05	<0.05	0.05	0.5	3.8	0.19	0.07
Dy	0.51	0.98	2	0.49	2	0.78	1.5	1.1	0.19	2.7	0.24	0.05	0.27	3.2	18	1.2	0.61
Ho	0.07	0.17	0.28	0.06	0.27	0.10	0.24	0.16	<0.05	0.41	<0.05	<0.05	<0.05	0.46	2.7	0.26	0.13
Er	0.14	0.43	0.58	0.10	0.44	0.26	0.59	0.38	0.07	1	<0.05	<0.05	0.06	1.1	5.4	0.67	0.50
Tm	<0.05	<0.05	0.08	<0.05	<0.05	<0.05	0.07	<0.05	<0.05	0.18	<0.05	<0.05	<0.05	0.22	0.59	0.09	0.08
Yb	0.12	0.29	0.52	0.10	0.26	0.19	0.49	0.31	0.06	1.3	0.07	<0.05	0.09	1.6	3.1	0.6	0.78
Lu	<0.05	<0.05	0.09	<0.05	<0.05	<0.05	0.06	<0.05	<0.05	0.19	<0.05	<0.05	<0.05	0.18	0.39	0.09	0.12
REE_T	<29.8	<48.7	178.9	<71.8	<179.8	<128.4	43.8	<30.8	<6.2	34.6	<2.9	<1.36	<4.37	<14.6	830	41.3	12.5
Y	2.1	4.4	7	1.6	6.9	3.2	8.5	5	1	8.5	0.99	0.30	1.1	17	66	5.9	3.8
Th	<10	20	30	20	30	20	20	10	20	20	10	<10	<10	<10	108	<10	<10
U	6	2	3	280	6	1	5	2	<1	7							
Zr	35	65	155	60	70	105	25	25	5	40	12	7	17	93	180	61	43
Hf		2.3	4.6		1.9	2.8	2.1			2.2	0.89	0.36			8	0.82	3.1
Sn	<0.8	<0.8	<0.8		2.1	0.9	4.8	6.3	3.8	4.9	130	13	66	4	5.5	1	2
Nb	<1	<1	<1	<1	1	1	9	3	5	1	70	9	28	14	55	<5	<5
Ta		0.26	0.27		0.39	0.13	1.9			2.6	11	2.4			2.5	0.1	1.8
Cr	19	<5	<5	<5	<5	<5	<5	<5	<5	<5	<10	<10	<10	<10	15	11	15
Ni	<5	5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	8	<5	5
V	7	11	6	7	13	23	4	2	3	3	<5	<5	<5	<5	18	6	11
Sc	2	2	3	2	3	3	2	2	2	2	3	<2	<2	<2	22	<2	3
Co	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	15	11	<5
Ga	14	14	13	13	17	18	25	26	12	33	41	28	33	36	91	18	34
Zn	26	22	18	16	27	40	28	32	42	23	41	18	72	29	188	8	21
B	12	10	13	9	14	5	130	12	<5	2300	480	210	1100	1100	39	7	210
F	200	300	260	180	440	500	310	310	290	410	380	60	300	300	840	<50	50
Cl	<30	59	259	532	75	<30	33	45	33	60	112	44	37	37	37		37
								Ratios									
A/CNKmol	1.175	1.357	1.291	1.338	1.480	1.218	1.255	1.242	1.152	1.232	1.417	1.154	1.383	1.200	4.296	1.136	3.447
Al/Ga	5260	5830	5990	6080	5270	4770	3240	3100	6630	2380	1836	2915	2540	2430	1750	4565	2790
Ba/Rb	3.8	6.3	8.3	7.1	10.4	15	0.9	0.3	0.07	0.05	<0.015	<0.01	0.02	0.23	0.16	53	30
K/Ba	78	49	42	49	36	17	154	357	1500	1415	>2850	>5730	2630	325	900	12	11
K/Cs		29590	28810		18825	7660	4800			885	1900	2605	2140	650	1050	23350	456
K/Rb	290	265	350	349	376	249	137	115	100	73	44	50	55	75	146	625	332
Mg/Li	53	192	195	120	167	134	7	7	13	11	4	17	11	6	455	120	46
Nb/Ta		<3.8	<3.7		2.6	7.7	4.7			0.4	6.4	3.8			22	<50	<2.8
Rb/Cs		112	82		50	31	35			12	44	52	39	9	8.6	37	1.4
Rb/Sr	0.5	0.7	0.6	0.8	0.4	0.1	8	8	25	32	54	68	52	10	25	0.07	0.06
Th/U	<1.7		10	0.07	5	12	4		>10	3							
Zr/Hf		28	34		37	38	12	5		18	14	19			23	74	14
$(Ce/Yb)_N$	26	18	39	11	75	75	9		8	2.1	3.1	>1.4	4.0	0.4	30	6.5	1.3
Eu/Eu*	1.95	1.54	0.55	1.62	0.51	1.2	0.30	0.26	<0.52	<0.06	<0.68		<0.47	0.12	0.05	3.1	3.2

the removal of boron *via* tourmaline crystallization, as inferred by London (1986a, 1990).

Petrography of the Rare-Element Granitic Pegmatites

The Mavis Lake Group consists of an east-trending, 8 km by 0.8 to 1.5 km concentration of bodies of granitic pegmatite and related metasomatic zones. It exhibits a classic regional zonation of pegmatite types (Cameron *et al.* 1949, Heinrich 1953) with increasing distance east of the parent batholith (Fig. 9), as defined by systematic changes in mineralogy, chemical association, and extent of postmagmatic replacement.

The *Interior Beryl Zone* comprises a 1.5 by 3.5 km area of pegmatitic granite within the Ghost Lake batholith (Fig. 9); it is marked by sporadic green beryl in potassic pegmatite dykes and masses (unit GLB–7).

In the adjacent mafic metavolcanic country-rocks, the *Beryl–Columbite Zone* constitutes the first grouping in the exocontact. This rare-element mineralization is contained in muscovite–tourmaline potassic pegmatites, as at the Taylor No. 1 and 2 Pegmatites or in locally albitic units at the *Contact Beryl Occurrence* (Fig. 9, Table 2). The former exhibit local emerald adjacent to phlogopite-rich selvedges in contact with metaultramafic host-rocks. This setting resembles that which led to the appearance of emerald at the Gravlotte mine in the Republic of South Africa, developed near an Archean albite-rich pegmatite (Robb & Robb 1985). Interspersed throughout the beryl–columbite zone are low concentrations of scheelite (<0.1 wt.% WO_3) in tourmalinite sheets, calc-silicate skarns and pillow selvedges.

The *Spodumene – Beryl – Tantalite Zone* is defined by the initial appearance of spodumene in pegmatites of the albite–spodumene type (Černý 1989a), about 3.5 km from the batholith contact. Here, swarms of tabular pegmatite dykes, up to 10 m in thickness and 280 m in length, generally strike parallel to the foliation in the host. Internal zoning is indistinct to absent and is best exemplified at the Fairservice No. 1 Pegmatite (Breaks 1989), which contains three gradational zones of increasing content of quartz: 1) potassic pegmatite, which contains minor interstitial spodumene and quartz, 2) spodumene–quartz-rich pegmatite, and 3) a discontinuous quartz-rich core zone with minor spodumene, blocky microcline and beryl.

The distal zone of the Mavis Lake Group is composed of extensively replaced, albite-rich pegmatites considered to be representative of the *Albite type* of Černý (1989a). These pegmatites form thin (<1 m) sheets rich in albite. They contain sporadic fine-grained aggregates of green muscovite and albite (interpreted as secondary after spodumene), tantalite, white beryl and rare, green tourmaline (tsilaisite).

Petrochemistry

Representative results of major, minor and trace element analyses of all units of the Ghost Lake Batholith and leucosome from the adjacent migmatized metasedimentary rocks are presented in Table 3. Compositions not tabulated in this paper but nevertheless utilized in the various diagrams are available from the Depository of Unpublished Data.

Ghost Lake batholith: major elements

Variation of Al_2O_3, CaO, total iron as FeO, K_2O, MnO, Na_2O, P_2O_5, and TiO_2 and selected trace elements Rb, Sn, Sr, Zr, and 14 *REE* against SiO_2 is portrayed in Figures 10, 11 and 12. These Harker plots reveal vaguely linear arrays of data points of decreasing Al_2O_3 and flat trends for CaO (Fig. 10), P_2O_5 and total iron as FeO (Fig. 11) with increasing SiO_2. The remaining elements mostly exhibit overlapping fields of concentration that are ireegular but do show some consistent trends from the most primitive to the most fractionated units of the Ghost Lake batholith.

The leucocratic character of the Ghost Lake batholith is reflected in the very low contents of FeO, Fe_2O_3, MgO and TiO_2. Total iron expressed as FeO rarely exceeds 1% and reveals a dominance of Fe^{2+} over Fe^{3+}. There is generally less than 0.5% MgO, which exhibits a dispersed variation with SiO_2 (Fig. 11). Unit GLB–1 has the widest variation of MgO (0.04–0.91%), and samples with the highest contents usually contain cordierite. In most GLB–3 and GLB–4 to –8 samples, MgO is extremely depleted and typically less than 0.15% (Fig. 11), which compares with leucogranites lacking cordierite, such as some of those in the Himalayas (*e.g.*, Castelli & Lombardo 1988). TiO_2 (0.02–0.51%) also exhibits a dispersed variation (Fig. 11), with the highest levels in GLB–2 and some GLB–3 samples, and relatively low contents (<0.2%) in GLB–4 to –8. TiO_2 contents exceed MgO in most samples of GLB–1, –2 and –3. MnO, not shown in the Harker plots, is largely below 0.02%; however, enrichment in many GLB–4 to –6 units (0.04–0.15%) is notable. One high value, 1.22% MnO, was found in layered aplite (Table 2) associated with GLB–7.

Data for K_2O and Na_2O and trace elements Rb, Sr and Zr can also be divided into similarly overlapping fields for GLB–1, –2, –3 and a composite field for the later, more fractionated units (GLB–4 to –8). There is a crude inverse correlation of K_2O or Sr fields of distribution with those for Na_2O ot Rb.

Chemical evolution of the Ghost Lake batholith, as revealed by increase in the Na_2O/K_2O ratio coupled with distinctive low CaO contents (Fig. 13), departs significantly from the calc-alkaline trend of Nockolds & Allen (1953) and the trondhjemitic trend of Barker & Arth (1976). The path of fractionation, marked by a curvilinear array of data points oriented subparallel to the

TABLE 3. VARIATION IN CONCENTRATION OF SELECTED TRACE ELEMENTS WITH METAMORPHIC GRADE IN CLASTIC METASEDIMENTARY ROCKS IN TWO RARE-ELEMENT PEGMATITE FIELDS OF THE SUPERIOR PROVINCE

I. DRYDEN AREA

	Li		Cs		Rb		B		F			
Metamorphic Zone	mean	range	mean	range	mean	range	mean	range	mean	range	N	N for Rb
					METAPELITES							
Chlorite-biotite	50	16-114	<57	<10-150	120	—	61	33-98	477	370-600	4	1
Andalusite	63	20-210	<10	<10-31	114	70-160	65	5-230	515	460-740	25	8
Sillimanite-Muscovite	170	37-590	<21	<10-90	158	120-260	124	5-560	516	370-720	16	5
Sillimanite-K-feldspar	47	27-76	<11	<10-20	110	90-130	<8	<5-16	540	410-630	11	2
Enclaves in Ghost Lake batholith	46	29-67	<10	<10-10	—	—	<5	—	513	340-650	3	—
					METAWACKES							
Chlorite-biotite	28	29-34	<10	<10-11	90	—	12	5-17	277	200-400	6	1
Andalusite	67	16-235	<17	<10-100	75	60-90	21	<5-120	413	320-830	14	2
Sillimanite-muscovite	110	24-355	<16	<10-50	—	—	41	<5-98	634	300-2140	9	—
Sillimanite-K-feldspar	38	11-94	<10	<10-20	—	—	<5	<5-8	540	310-1000	12	—
Enclaves in Ghost Lake batholith	36	26-48	<10	<10-20	—	—	<7.5	<5-24	569	440-770	10	—

II. WESTERN LAKE ST. JOSEPH AREA

	Li		Cs		Rb		
Metamorphic Grade	mean	range	mean	range	mean	range	N
					METAPELITES		
Medium	66	30-165	11	9-15	91	70-120	7
High	25	10-60	9	4-21	62	4-110	17
Granulite	14	5-20	<6	<2-7	61	40-80	9
					METAWACKES		
							9
Medium	45	10-185	11	3-30	68	20-160	39
High	11	5-30	6	1-9	53	10-110	9
Granulite	10	6-25	<5	<5-25	42	20-60	

N = number of samples

Na_2O–K_2O sideline, indicates the relative accumulation of Na_2O over K_2O in the two stages of fertile granite compared to GLB–1 and –2. As many compositions compare with the low-Ca granite of Turekian & Wedepohl (1961), this locus has been named the *low-calcium granite trend* (Breaks 1989). A similar magmatic trend has also been recognized in the various internal units of the Prosperous Granite in the Yellowknife Pegmatite District (P. Černý, pers. comm., 1990) and may characterize some rocks of the South Mountain batholith, Nova Scotia (Corey & Chatterjee 1990).

The well-defined inverse correlation between Na_2O and K_2O over a weight ratio range of 0.1 to 16.3 is shown in Figure 14; this trend is quite similar to those of other peraluminous granite suites such as the Manaslu pluton of Nepal (Le Fort 1981) and metaluminous suites as the Nelspruit Granite of South Africa (McCarthy & Robb 1977). Most internal units of the Ghost Lake batholith fall within discrete, partly overlapping fields, suggesting the following trend of Na-enrichment that correlates with generally decreasing K/Rb: GLB–1–2–(3–4)–(5–6) (Table 3). The wide range in Na_2O/K_2O of the batholith also distinctly contrasts with the limited field of distribution of the S-type granites of the Lachlan Fold Belt of Australia (Fig. 14). Chappell & White (1982, p. 92) indicated that low levels of Na, Ca and Sr are diagnostic of the Australian peraluminous S-type granites. It thus seems apparent that Archean S-type granites, at least in the study area and in the English River Subprovince 60 km to the north (Breaks 1991, unpubl. data) initially had some compositions similar to their Australian counterparts (*i.e.*, GLB–1 and –2), but in contrast are fractionated toward high levels of Na. Therefore, comparisons of Archean S-type granite suites with average S-type granites of the Lachlan Fold Belt, such as that proposed by Whalen *et al.* (1987), must be treated with caution.

All samples of the Ghost Lake batholith are peraluminous, as revealed by the aluminum saturation index of Zen (1986), in the range of 1.152 to 1.606, substantial levels of normative corundum (4 to 7%), which are comparable to those of the strongly peraluminous East Kemptville leucogranite (Kontak 1990), and absence of normative diopside (Chappell & White 1982).

Ghost Lake batholith: trace elements

Significant petrochemical variation in the Ghost Lake batholith is revealed by enrichment trends in B, Be, Cs, Ga, Li, Nb, Rb, Sn and Ta, and depletion in Sr, Ba, Zr, total *REE* and *LREE* with progression from GLB–1 to GLB-3 and thence to GLB–4 to –8 (Table 2).

High-field-strength elements

The high-field-strength elements Ga, Nb, Hf, Sn, Ta and Zr have limited ranges of concentration. These elements exhibit distinct trends of enrichment in the batholith, with the exception of Zr, which displays a converse trend. Maximum values for Ga (39 ppm), Nb (70 ppm), Sn (130 ppm) and Ta (17 ppm) are associated with pegmatitic granites in the eastern lobe, whereas the highest mean and single Zr values respectively occur in GLB–2 (mean 158 ppm) and GLB–1 (360 ppm). This progressive depletion of Zr denotes early fractional crystallization of zircon in GLB–1 and –2 and consequent removal of much of the Zr from the residual melts succeeding these units.

Data on Hf are more sparse, but reveal a limited overall range in concentration of 0.2–5.8 ppm that indicates little variation amongst the internal granitic units of the batholith, except for very low abundances found in GLB–6 and –7 (0.2–0.89 ppm). Zr/Hf ratios (12–37), which overlap the range for granitic rocks in general (30–40: Černý *et al.* 1985) are typically higher in the chemically more primitive units (GLB–1 and –2: 20–37) and diminish in the two fertile granite units (12–22). This trend compares well with that in other suites of fertile granite (*e.g.*, southeastern Manitoba: Černý *et al.* 1985).

Ta and Nb display an uncertain behavior in the batholith. In GLB–1 and –2, Nb, at a level mainly less than 3 ppm (Table 2), is considerably depleted relative to its average in the crust (11 ppm: Taylor & McLennan 1985, p. 67). A more limited Ta data-set (mean 0.45 ppm, range 0.13–0.91 ppm), however, reveals appreciable fractionation, as the Nb/Ta ratios are mostly less than 3.8 (Table 2) and considerably lower than that in the average crust (11: Taylor & McLennan 1985, p. 67). Higher contents of Ta occur in the fertile granite units GLB–3 (mean 2.5, range 0.49–4.9 ppm) and GLB–5 to –7 (mean 3.9 ppm, range 1.0–11 ppm: Table 2). However, this increase is sympathetic with that of Nb in the fertile granite units (1–70 ppm), such that Nb/Ta is constant. Values of the Nb/Ta ratio for GLB–3 (mean <2.7, range <2–4.7) largely overlap with those for GLB–4 to –7 (mean 5.4, range 0.4–15).

Concentrations of Ga in the Ghost Lake batholith have a limited overall range, 4–38 ppm. However, mean Ga contents exhibit a well-defined, progressive enrichment in the observed intrusive sequence: GLB–1 (14.5 ppm), GLB–2 (17.5 ppm), GLB–3 (21.2 ppm) and GLB–4 to –8 (27.8 ppm). This correlates with a drop in mean Al/Ga ratio, from the chemically primitive parts of the batholith (6125) to the fertile granite stages (GLB–3: 3712, GLB–4: –7: 2670), the latter of which lie in the range 1400–3500 for fertile granites in the Canadian Shield (Černý *et al.* 1985). More extreme buildups of Ga, up to 142 ppm, characterize the metasomatic units at the Contact Beryl Occurrence (unpubl. data), restite selvedges of "fertile" migmatites described below (50 ppm), and restite entrained in GLB–1 (48 ppm) and GLB–3 (67 ppm).

Tin exhibits a slight tendency for increased concentration in GLB–3 (mean 6.0 ppm, range 3.6–10.2 ppm) and most of GLB–4 to –8 (mean 4.5 ppm, range 2.7–7.1 ppm) relative to the chemically more primitive GLB–1 (mean <1.6 ppm, range <0.8–3.5 ppm). However, substantially higher Sn values occur in pegmatitic granites at the Contact Beryl Occurrence (mean 38 ppm, range 4–130 ppm), where cassiterite has been documented. Similarly, high trace levels of Sn also occur in exocontact pegmatite dykes adjacent to the southeastern contact of the batholith (mean 116 ppm, range 3.2–445 ppm). The Ghost Lake batholith can thus locally be classed as a stanniferous granite (*i.e.*, one with greater than 10 ppm Sn, as defined by Flinter 1971).

Li, Cs and Be

Lithium is generally enriched in GLB–3 (mean 68 ppm, range 15–132 ppm) and in the eastern lobe of the batholith (mean 56 ppm, range 16–198 ppm) relative to GLB–1 (mean 12 ppm; range 3–30 ppm). However, local concentrations of Li occur in cordierite–quartz segregations in GLB–1 (108–223 ppm) which, along with Be (35–51 ppm), can be incorporated into the cordierite structure (0.067–1.77% BeO: Černý & Po-

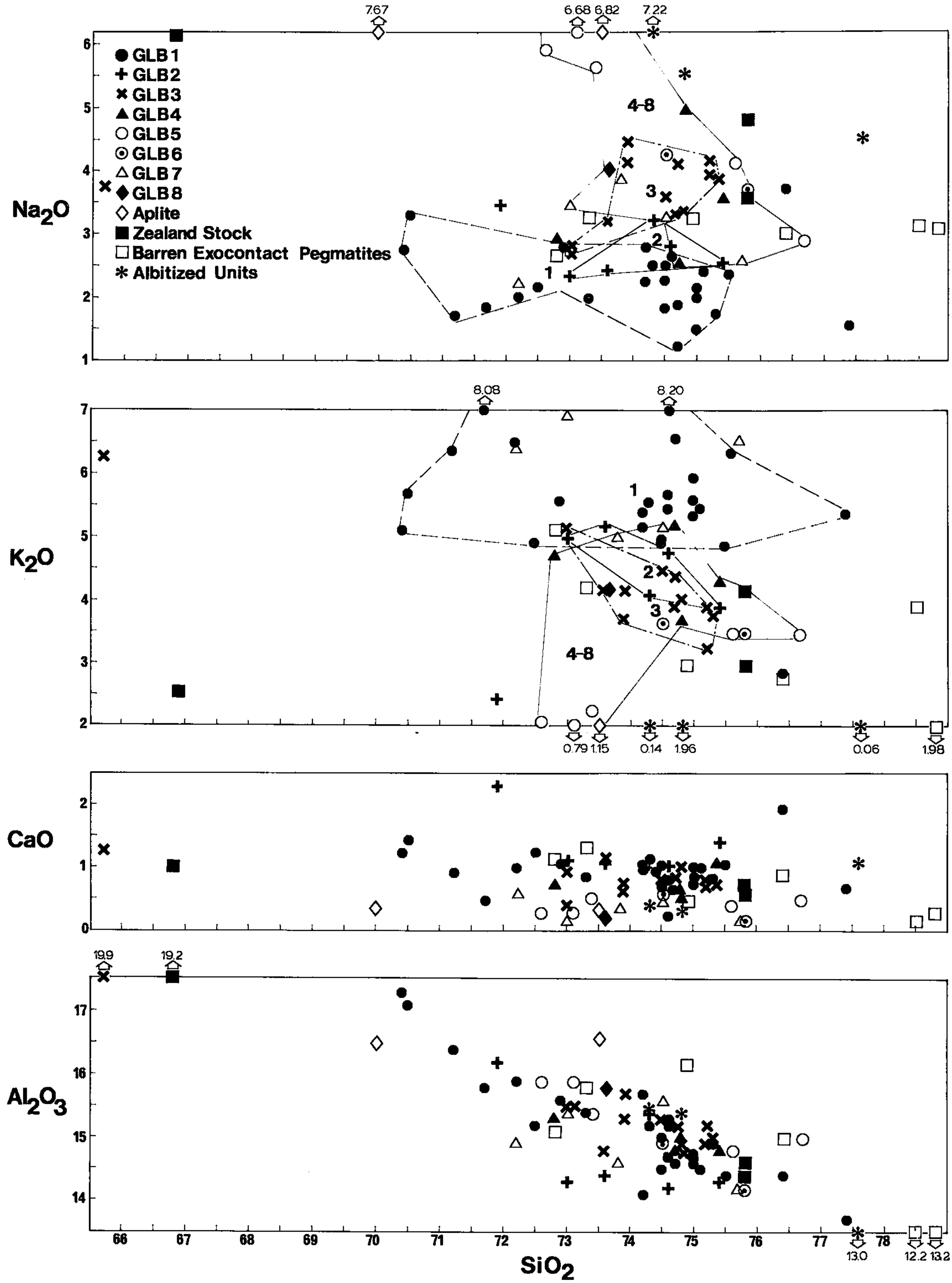

FIG. 10. Harker plots of Al_2O_3, CaO, K_2O and Na_2O in internal units of the Ghost Lake batholith, the Zealand stock and related, barren, exocontact pegmatites.

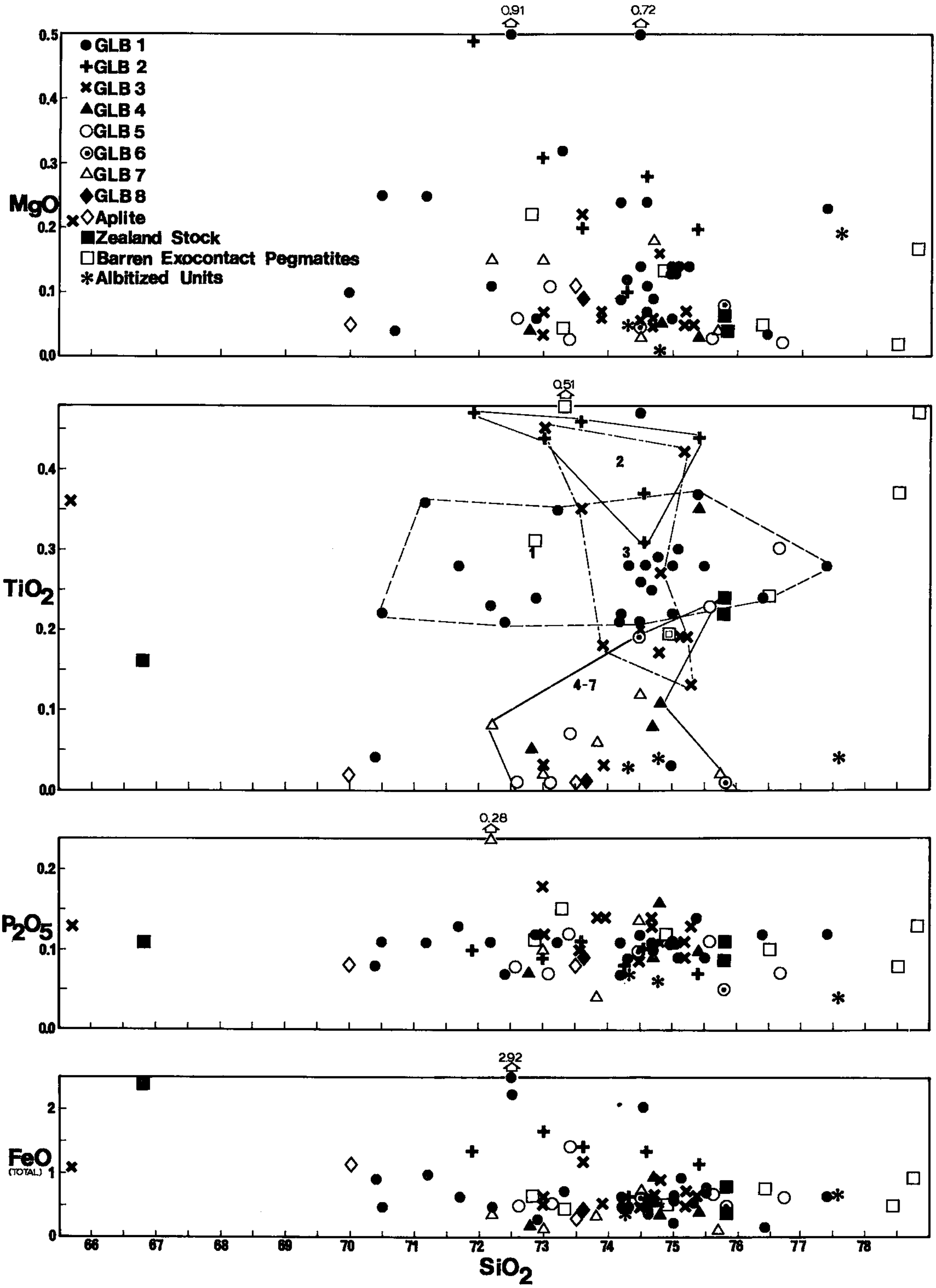

FIG. 11. Harker plots of total Fe as FeO, MgO, P_2O_5 and TiO_2 in internal units of the Ghost Lake batholith, the Zealand stock and related, barren, exocontact pegmatites.

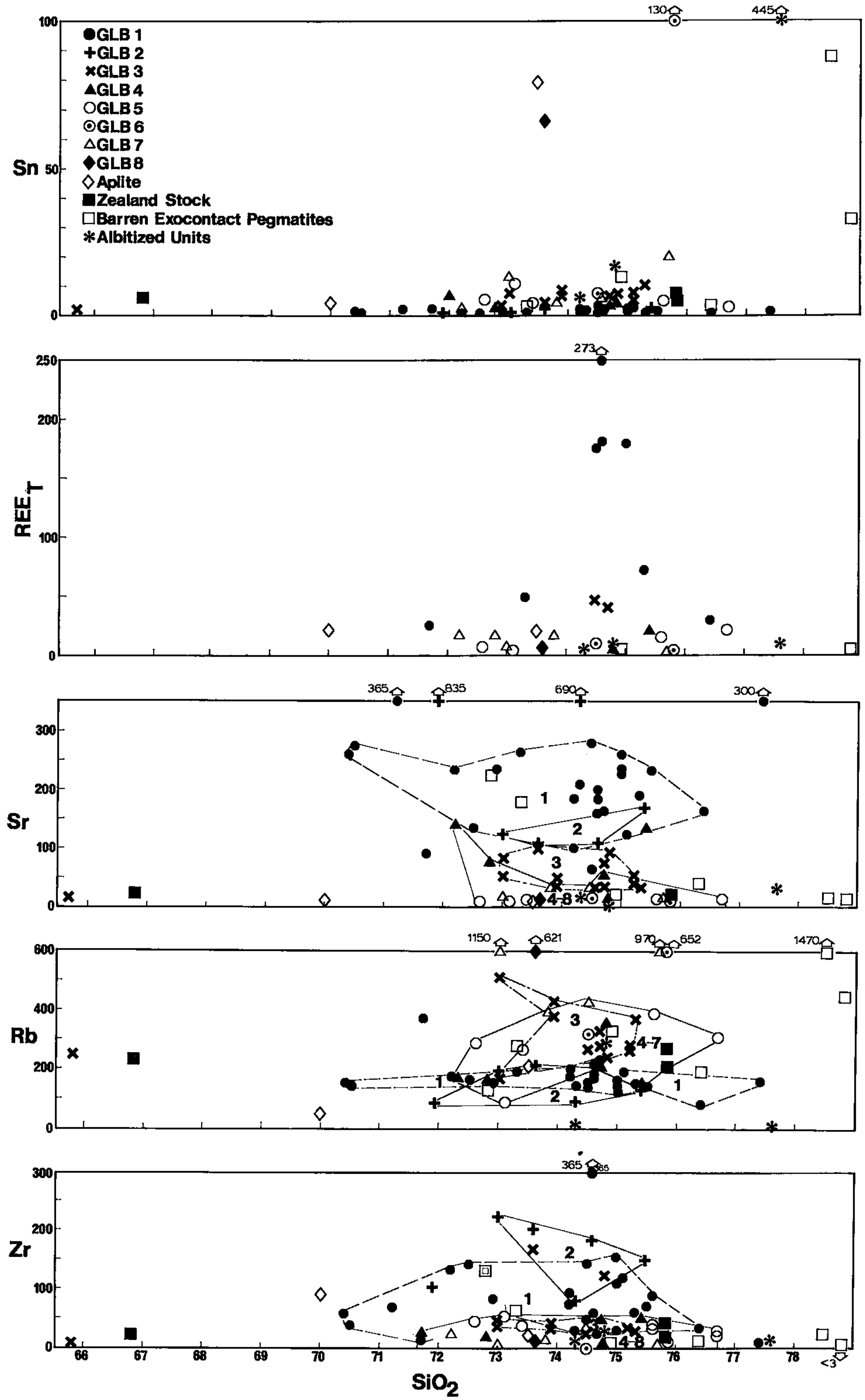

FIG. 12. Harker plots of Sn, total *REE*, Sr, Rb and Zr in internal units of the Ghost Lake batholith, the Zealand stock and related, barren, exocontact pegmatites.

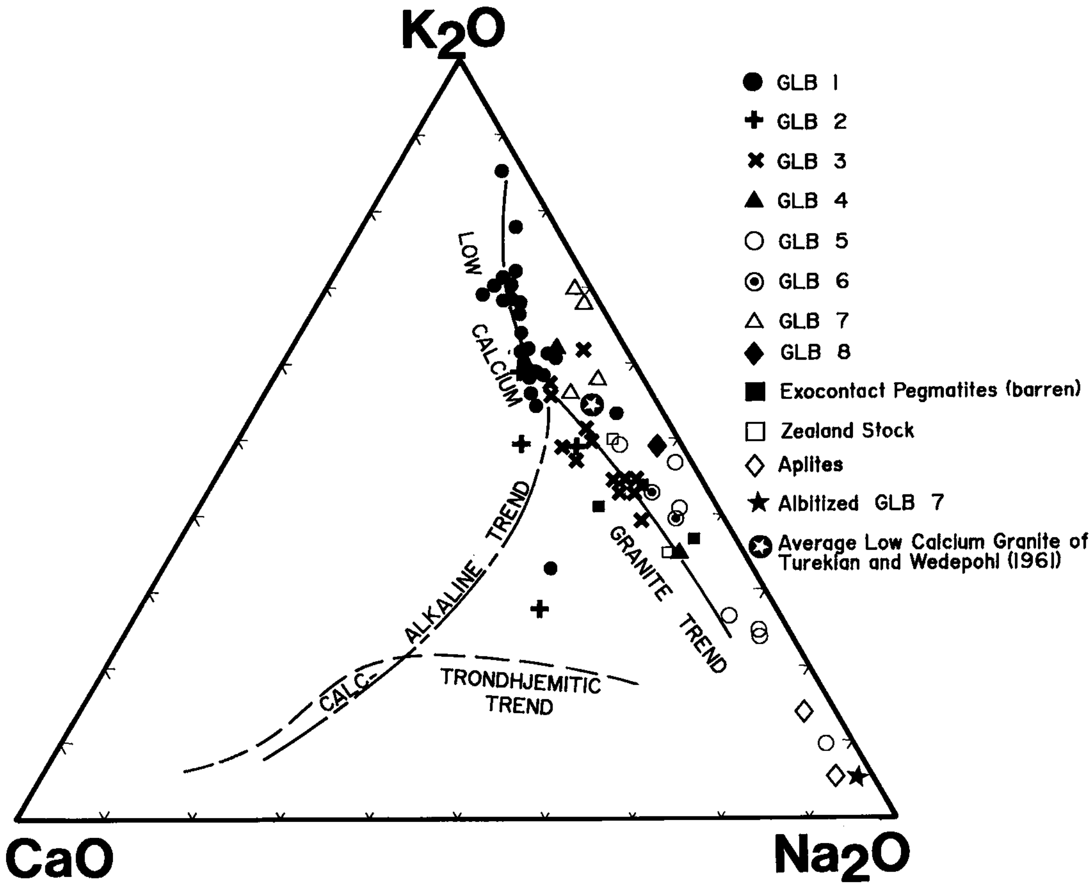

FIG. 13. Variation (wt.%) of CaO, Na_2O and K_2O in internal units of the Ghost Lake batholith, related, barren, exocontact pegmatites and the Zealand stock. Calc-alkaline and trondhjemitic trends respectively after Nockolds & Allen (1953) and Barker & Arth (1976).

vondra 1967). Stavrov (1978) indicated that cordierite can contain up to 2000 ppm Li.

Substantially higher values of Li are found in the dravite – phlogopite – actinolite (1270–1820 ppm), holmquistite-bearing (1020–2450 ppm) and beryl–muscovite greisen (mean 329, range 303–355 ppm) units of the synmagmatic metasomatic system and in late magmatic biotite-rich selvedges adjacent to GLB–7 (mean 1583 ppm, range 1400–1860 ppm) and GLB–1 (130 ppm).

Mg/Li relations for units of the Ghost Lake batholith, supracrustal host-rocks and rare-element pegmatites are depicted in Figure 15. The batholith occupies a broad field that defines a general trend of significantly increasing Li against slightly diminishing Mg. Unit GLB–1 exhibits the highest Mg/Li values (mean 120, range 27–540); these are considerably lower than the value for the average crust (2446: Taylor & McLennan 1985, p. 67). Units of fertile pegmatitic granite from the eastern lobe and unit GLB–3 have the lowest Mg/Li ratios, generally between 4 and 20.

No primary unit of the batholith or rare-element pegmatite falls within the range 200–1200 ppm Li. This hiatus, which corresponds to the *lithium gap* first mentioned by Norton (1973), possibly represents excess lithium in the pegmatite-forming melt that produced the units of stage-2 fertile granite. Consequently, this gap is interpreted as Li that was subsequently evacuated *via* the contact-metasomatic process. Li concentrations (mean 1340 ppm, range 178–2450 ppm, unpubl. results of 14 analyses) do span this gap, albeit shifted to higher Mg; they characterize the holmquistite – actinolite – phlogopite – dravite metasomatic rocks (Fig. 15). These rocks may represent the "missing link" between the

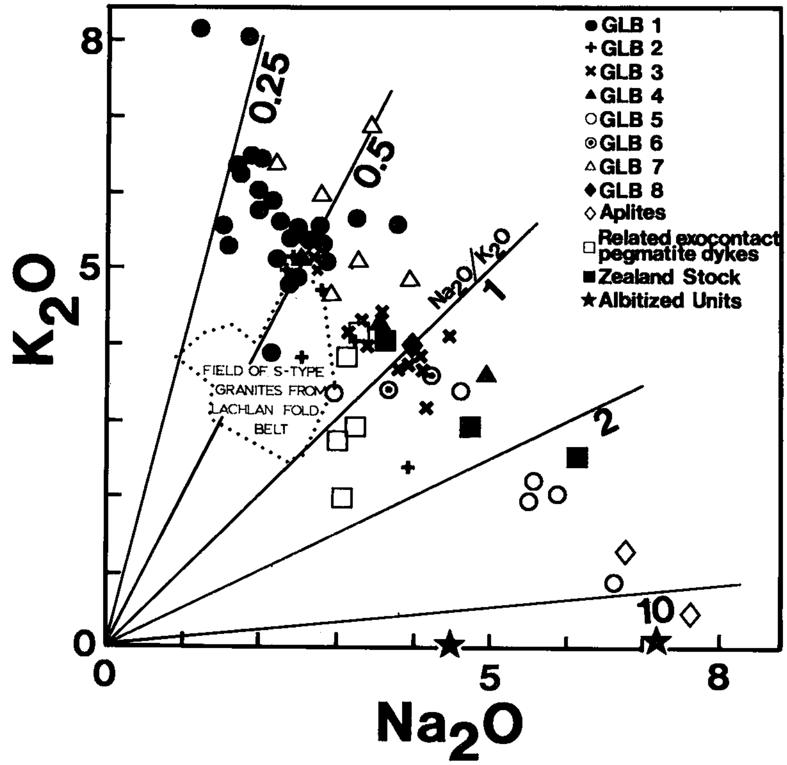

FIG. 14. K_2O–Na_2O variation in internal units of the Ghost Lake batholith, the Zealand stock, and related, barren, exocontact pegmatites. Dotted field encloses results of 316 analyses of S-type granites from the Lachlan Fold Belt of Australia (Chappell & White 1982).

levels of Li levels fixed in the parent Ghost Lake batholith and minimum levels of Li neccessary to crystallize the progenitor albite–spodumene pegmatites of the Mavis Lake Group. It should be noted that the Li gap does not include the Taylor beryl pegmatites, which have Li contents comparable to nearby units GLB–5 to –8. These pegmatites, however, may represent modified compositions, as they are typified by higher-than-normal Mg contents. This anomalous chemistry may possibly relate to metasomatic transfer *via* interaction of pegmatitic fluid with its metaultramafic host. It is plausible that these pegmatites endured considerable exomorphic loss of Li, possibly as a lithium borosilicate component (London 1986a), as its metasomatic selvedges contain abundant holmquistite and phlogopite (bulk Li = 2300 ppm).

The distribution of cesium is more imperfectly known, as most rocks fall below the 3 ppm analytical sensitivity limit save for recent results obtained by ICP–MS (0.05 ppm detection limit). Local anomalous concentrations occur within the Ghost Lake batholith, several of which fall into the category of extreme enrichment (>15 ppm) of Černý & Meintzer (1988). These are mainly pegmatitic granite units at the Contact Beryl Occurrence (mean 16 ppm, range 3–44 ppm) and parts of GLB–3 (mean 20 ppm, range 10–30 ppm). Greatly elevated levels of Cs characterize the synmagmatic metasomatic rocks: dravite – phlogopite – actinolite unit (mean 674 ppm, range 349–915 ppm), beryl–muscovite greisen (mean 189, range 183–195 ppm) and the holmquistite-bearing unit (mean 245 ppm, range 170–332 ppm). Late magmatic biotite-rich metasomatic selvedges contain even higher levels of Cs (mean 1583 ppm, range 1400–1860 ppm).

The distribution of beryllium also is incompletely established. Most samples of GLB–1 and –2 contain less than 1 ppm. Elevated concentrations occur sporadically in GLB–3, which has a range of <1–24 ppm. Pegmatitic granites of the eastern lobe have an overall range of 2–8 ppm, with no obvious trends toward increased abundance in units GLB–4 to –8. However, spatially related contact-metasomatic units and albitite dykes typically display modest to appreciable enrichment of Be (overall range: 2–1700 ppm Be).

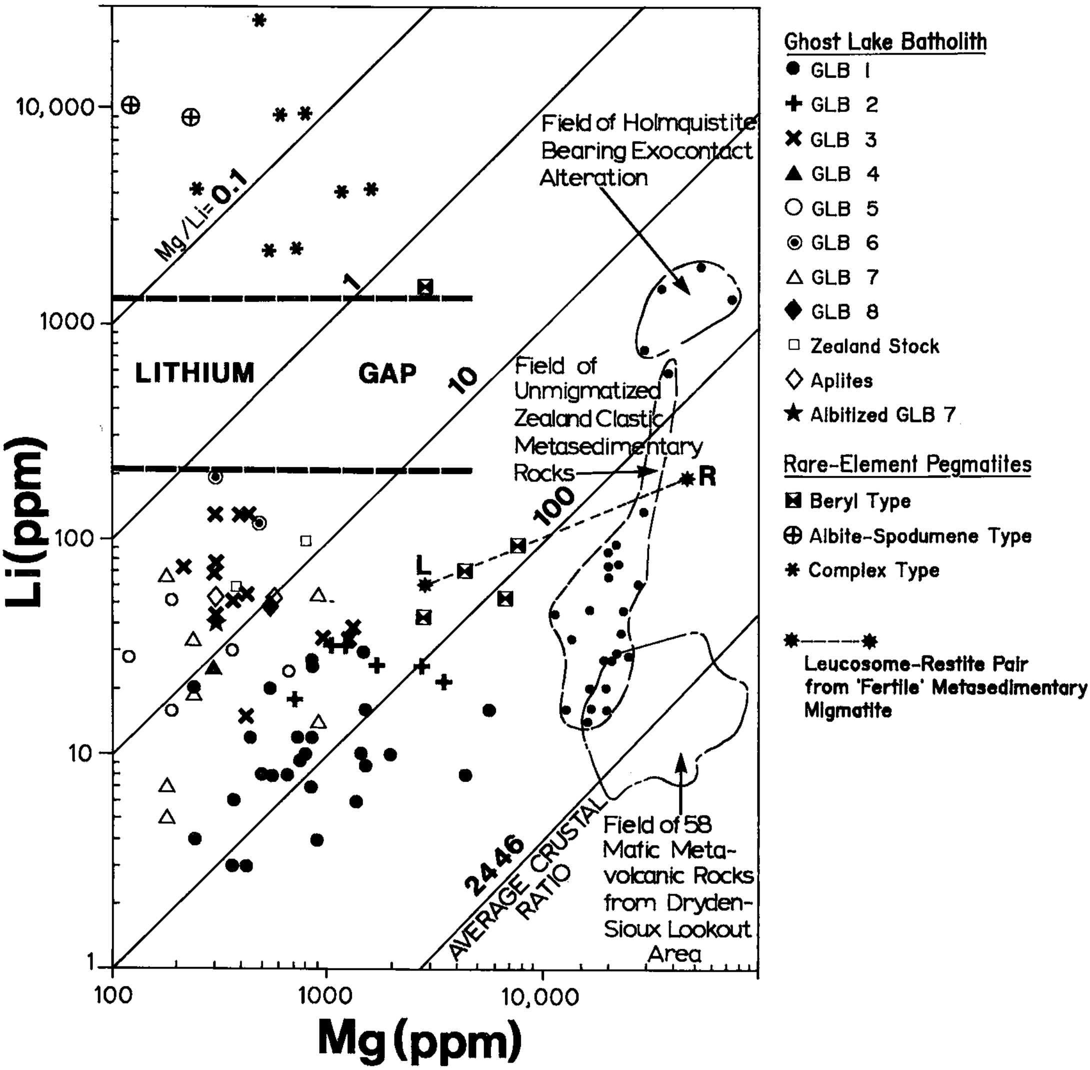

FIG. 15. Variation of Mg *versus* Li in samples of the Ghost Lake batholith and related rocks, supracrustal host-rocks, rare-element-enriched contact-metasomatic rocks, and rare-element pegmatites of the Dryden Field.

Ba, Rb and Sr

These three elements exhibit covariations that are typical of the influence during fractionation of feldspar and biotite (Tindle & Pearce 1981, Brown *et al.* 1981). Vectors for Rayleigh crystal fractionation suggest the importance of K-feldspar in controlling the distribution of Ba and Sr with respect to Rb. These elements exhibit trends of inverse concentration, reflected by rapidly decreasing Ba and Sr as a function of modestly rising levels of Rb. Most units of the Ghost Lake batholith yield an approximately linear distribution in a log-log plot (Fig. 16), which mark a progression from relatively poorly fractionated units GLB–1 and –2, marked by modest to very high levels of Ba (370–1840 ppm) and Sr (160–365 ppm) and relatively low amounts of Rb (80–230 ppm), to the chemically most evolved units (GLB–4 to –8: 270–1150 ppm Rb, 5–20 ppm Sr and –55 ppm Ba). Rocks of GLB–3 have overall values intermediate between GLB–1 to –2 and GLB–4 to –8. Such trends compare favorably with the model of perfect fractional crystallization of granitic melts under the influence of an intercumulus melt (McCarthy & Hasty 1976). However, it must be emphasized that subsolidus

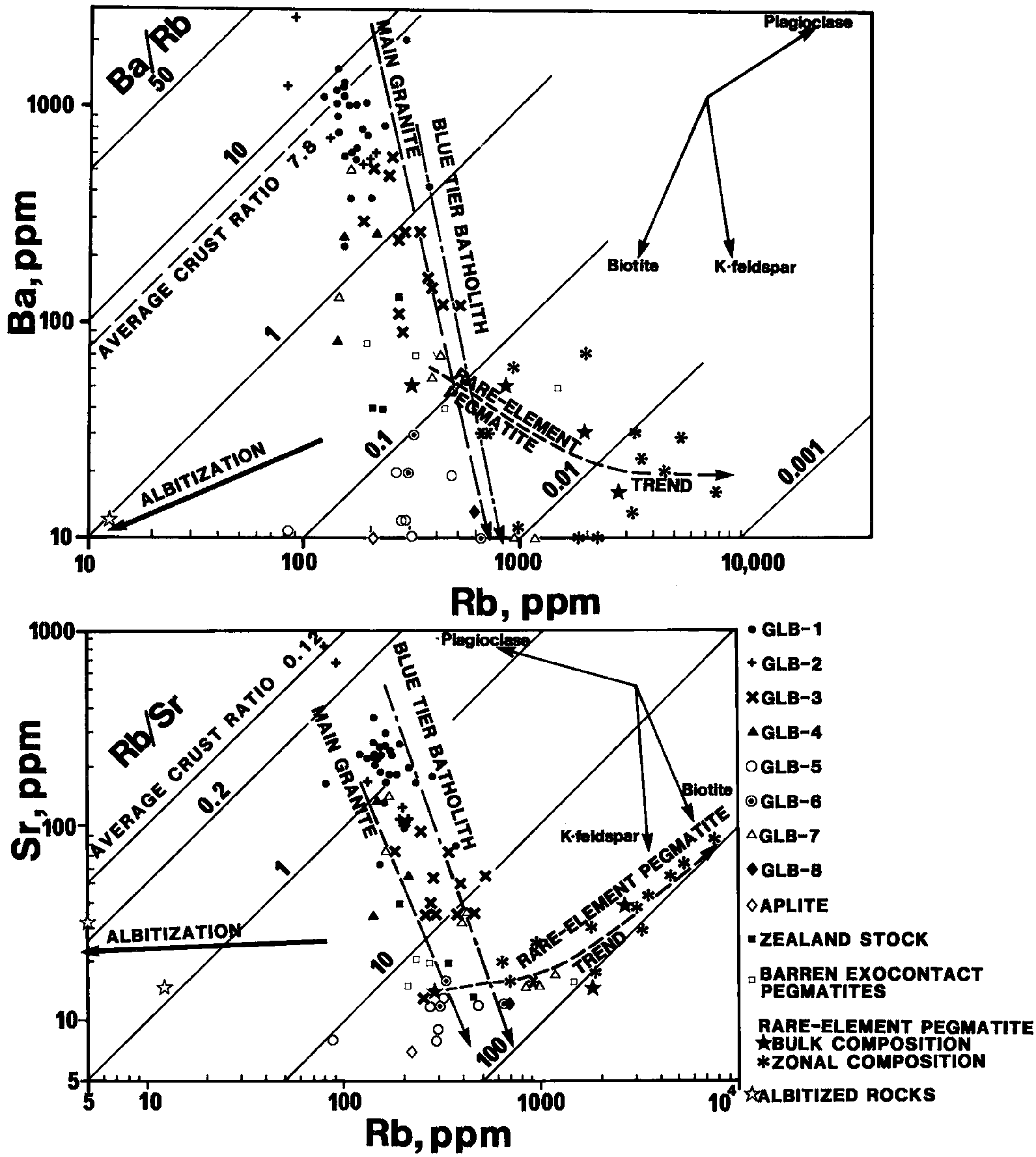

FIG. 16. Ba–Rb and Rb–Sr variation diagrams for the Ghost Lake batholith, related barren exocontact pegmatites, the Zealand stock and rare-element pegmatites. Comparative trends are from the Blue Tier Batholith of Tasmania (Groves & McCarthy 1978) and the Main Granite of the Bushveld Igneous Complex (McCarthy & Hasty 1976). The vectors reveal, in direction of decreasing proportion of remaining melt, the compositional change of the melt phase in response to Rayleigh fractionation of plagioclase, K-feldspar or biotite (after Tindle & Pearce 1981). Average crustal values after Taylor & McLennan 1985, p. 67).

mobilization of Ba, Rb and Sr probably occurred in all units of the batholith as suggested, for example, by the "fibrolite"-rich veins, muscovite porphyroblasts and muscovite – sillimanite – quartz nodules. Such mobilization may be reflected in the modest scattering of Ba, Rb and Sr in Figure 16. Fractionation trends in the Ghost Lake batholith parallel, for example, the composite field of the Main Granite – Bobbejaankop Granite – Lease

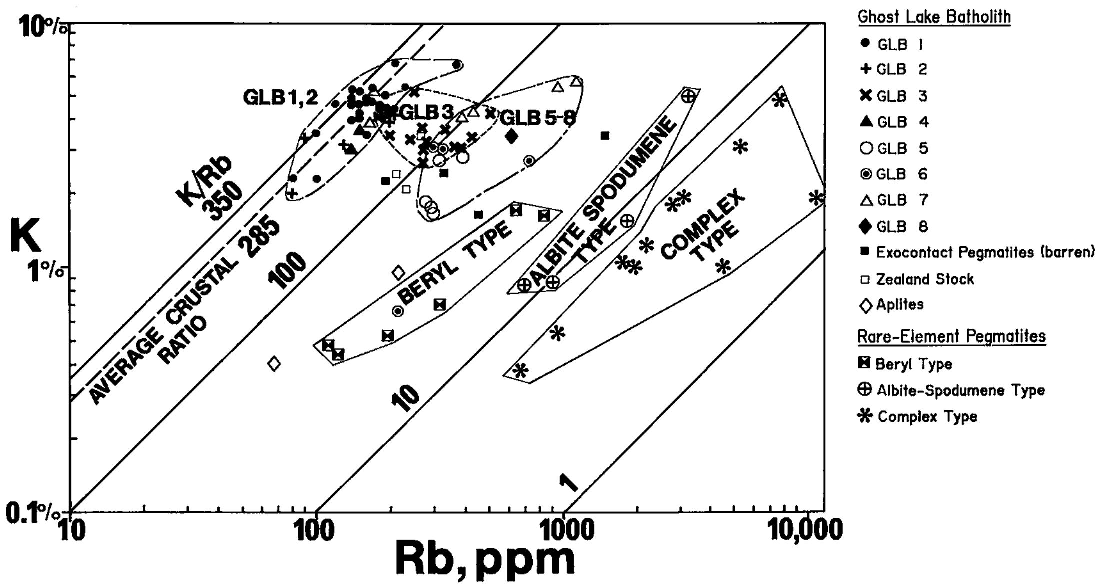

FIG. 17. Variation in K *versus* Rb for the Ghost Lake batholith, related barren, exocontact pegmatites, the Zealand stock and rare-element pegmatites. Average crustal ratio after Taylor & McLennan (1985, p. 67).

Granite of the Bushveld Igneous Complex (McCarthy & Hasty 1976, Kleeman & Twist 1989) and the Blue Tier Batholith of Tasmania (Groves & McCarthy 1978), both of which contain syngenetic cassiterite mineralization in late-stage units. The extremely low contents of Ba and Sr (<50 ppm) at high Rb concentrations (>200 ppm) that typify rocks formed by crystallization of residual liquids in these two tin-bearing complexes also are characteristic of most units of pegmatitic granite within the eastern lobe of the Ghost Lake batholith.

Notable departures from the linear Ba–Rb and Rb–Sr trends are produced by high levels of Rb in the chemically most evolved units of potassic pegmatite of the batholith (Contact Beryl Occurrence: 652–1150 ppm) and exocontact rare-element pegmatites of the Mavis Lake Group (700–3300 ppm), and lower levels of Rb in fine-grained leucogranites (87–390 ppm), primary aplites (52–214 ppm) and in rocks that have endured postmagmatic albitization (12–293 ppm). This implies, in the case of the potassic pegmatites within the eastern batholith lobe, attainment of such compositions by processes that may approximate incremental equilibrium crystallization, as modeled by McCarthy & Hasty (1976). It is also notable that rare-element pegmatite compositions lie on the same trend as the chemically most evolved units of the Ghost Lake batholith, which suggests a genetic link. The trend of rising Sr with Rb for the rare-element pegmatites (Fig. 16) is possibly due to progressively increasing levels of radiogenic ^{87}Sr produced from ^{87}Rb (Clark & Černý 1987), although one cannot dismiss introduction of Sr into pegmatite-forming melts by interaction with mafic metavolcanic host-rocks.

K–Rb relationships

Variation of K *versus* Rb for the Ghost Lake batholith units and spatially related rare-element pegmatites is displayed in Figure 17. Overall, the K/Rb ratio exhibits a wide range in the batholith, *i.e.*, from 387 to 45. Exclusive of exocontact units (barren pegmatites and the Zealand stock), data on the batholith fall into three distinct fields in Figure 17: GLB–1–2, GLB–3, and a composite of the pegmatitic granite units (GLB–4 to –8).

Chemically primitive rocks (GLB–1 and –2) have the highest K/Rb ratio (several exceed 349) and exhibit little change in K/Rb against increasing K (2.00–6.81 wt.%), which defines an elongate field essentially parallel to the average crustal value (Fig. 17). The GLB–1 and –2 field corresponds to the normal trend of Shaw (1968), and exhibits values comparable to the lower part of the range for peraluminous granites associated with low-pressure granulite domains in the nearby English River Subprovince (mean of 31 analyses: 464, range 244–818: unpubl. data).

The GLB–3 field is indicative of the pegmatitic-hydrothermal trend of Shaw (1968) which, according to Černý *et al.* (1985, p. 385), generally "marks a turning point in K/Rb evolution" of fertile granites and is characterized by a sharp decrease in the ratio in products of late- and postmagmatic crystallization. The field comprising the second stage of fertile granites (GLB–4 to –8) overlaps with the GLB–3 field; however, it shows a definite trend toward higher Rb contents (maximum: 1150 ppm). Several data points for pegmatitic leucogranite and potassic pegmatite data points do not fall within this ovoid field and represent compositions of units that lie outside the eastern lobe, *i.e.*, associated with GLB–1 and –2.

This field of positive slope may be called the "fertile pegmatitic granite trend"; it is similar in orientation to the fields defined by three types of rare-element pegmatites (Fig. 17). These isolated, *en échelon* fields progressively decrease in K/Rb values in the series beryl type – albite type – spodumene type – complex type. The beryl type, spatially nearest to the most fractionated part of the Ghost Lake batholith, lies closest to the parent rocks in terms of K/Rb and Rb contents. Potassium contents, however, trend to lower values, possibly lower owing to loss to phlogopite-rich exomorphic selvedges *via* a fluid phase.

Rare-earth elements

Behavior of the *REE* in the Ghost Lake batholith, associated metasomatic systems and the adjacent rare-element pegmatites is complex but similar to that in other suites of fertile granite (Černý & Meintzer 1988). Total concentrations of the *REE* exhibit no systematic variation with SiO_2 (Fig. 12) but, nevertheless, they diminish with increasing degree of evolution, as measured by the K/Rb ratio, *i.e.*, total concentrations of the *REE* rapidly decrease from 273 to 6.2 ppm as K/Rb drops from 376 to 44. It is striking that the fractionation processes operating in this peraluminous granite – pegmatite system can reduce the total *REE* contents of its more fractionated units to levels comparable to those in many so-called primitive rocks.

Representative *REE* data are presented in Table 2 and in Figures 18, 19 and 20, normalized to the chondritic data of Wakita *et al.* (1971). Considerable diversity in chondrite patterns is evident for the Ghost Lake batholith units and adjacent rare-element pegmatites of the Mavis Lake Group, representing one of the most complete *REE* fractionation sequences yet documented for an Archean peraluminous granite – pegmatite system (see also Černý & Meintzer 1988). Five types of *REE* patterns, described below, are apparent for this system.

Ghost Lake batholith

The highest total *REE* and $(Ce/Yb)_N$ values typify the chemically most primitive units (Fig. 18): GLB–1 with a negative Eu anomaly [174–273 ppm, mean $(Ce/Yb)_N = 95$] and GLB–2 [128–79 ppm, mean $(Ce/Yb)_N = 46$]. Such *Type-1* patterns possibly reflect the retention of *HREE* by garnet in the restite. Although garnet has not been observed in restite entrained in any internal unit, it is a widespread phase in the mesosome of metasedimentary migmatites west of the batholith. The Type-1 pattern is very similar to that of orthopyroxene – cordierite – biotite peraluminous granites distributed within the granulite zones of the English River Subprovince, as exemplified by the Churchill Lake Batholith (Breaks 1991), and plotted in Figure 18.

In GLB–1, three samples with *Type-2* patterns are marked by flatter curves [$18 < (Ce/Yb)_N < 61$], considerably lower total *REE* (30–54 ppm) and a small positive Eu anomaly (Eu/Eu* between 1.6 and 2.0). Such patterns possibly represent feldspar cumulate rocks, as suggested by the higher than normal contents of CaO, Na_2O or K_2O. There is, however, a marked difference between these rocks and low-melt-fraction leucosomes from metasedimentary migmatites west of the batholith, as the latter exhibit lower total *REE*, lower $(Ce/Yb)_N$, between 3.2 and 6.5, and more appreciable positive Eu anomaly (Figs. 18, 20).

It is uncertain what role cordierite plays in the geometry of Type-2 patterns. Few data on *REE* concentrations in cordierite exist in the literature; Bea (1991, p. 1866), for example, showed that cordierite from the low-melt-fraction Peña Negra anatexites of central Spain is typified by a $(Ce/Yb)_N$ ratio similar to that of the bulk rock, but it differs in having a substantial negative Eu anomaly.

Type-3 patterns (Fig. 18), as exemplified by GLB–3, mark a striking change in shape to a much flatter slope [$(Ce/Yb)_N$ between 9 and 10) and fractionation toward lower total *REE* (31–44 ppm). *Type-4* patterns (Fig. 19) characterize pegmatitic units within the eastern GLB lobe; they are marked by pronounced depletion in *LREE* and much larger negative Eu anomalies (Eu/Eu* <0.06) relative to the above patterns. Total concentrations of the *REE* have been drastically reduced (<1.5–35 ppm), particularly Eu and the *HREE*, whose abundances are commonly below the 50 ppb detection limit. Potassic pegmatite units (GLB–6, –7 and –8) and fine-grained leucogranite (GLB–5) exhibit the most severe depletion in total *REE* (0.4–4.8× and 1.6–6.3× chondrite, respectively). Although GLB–7 patterns display some overlap with those of the fine-grained leucogranite field (Fig. 19), these represent chemically the more primitive potassic pegmatite units (237 < K/Rb < 310), which occur outside the eastern lobe of the batholith and intrude GLB–1 and –3. *Type-5* patterns, of rare positive slope [$(Ce/Yb)_N = 0.35$], occur only in aplite interlayered with potassic pegmatite at the Contact Beryl Occurrence.

The highest concentration of the *REE* in the Ghost Lake batholith is found in "fibrolite"-rich veins hosted in GLB–1 (Fig. 19). Its chondrite-normalized pattern is very similar to that of biotite–cordierite restite entrained in GLB– and –3. About one-half of its 845 ppm total *REE* is accounted for by Ce, suggesting retention by monazite or apatite (or both).

Many of the fertile granite units reveal curious inflections at Nd, Tb and Ho in their chondrite-normal-

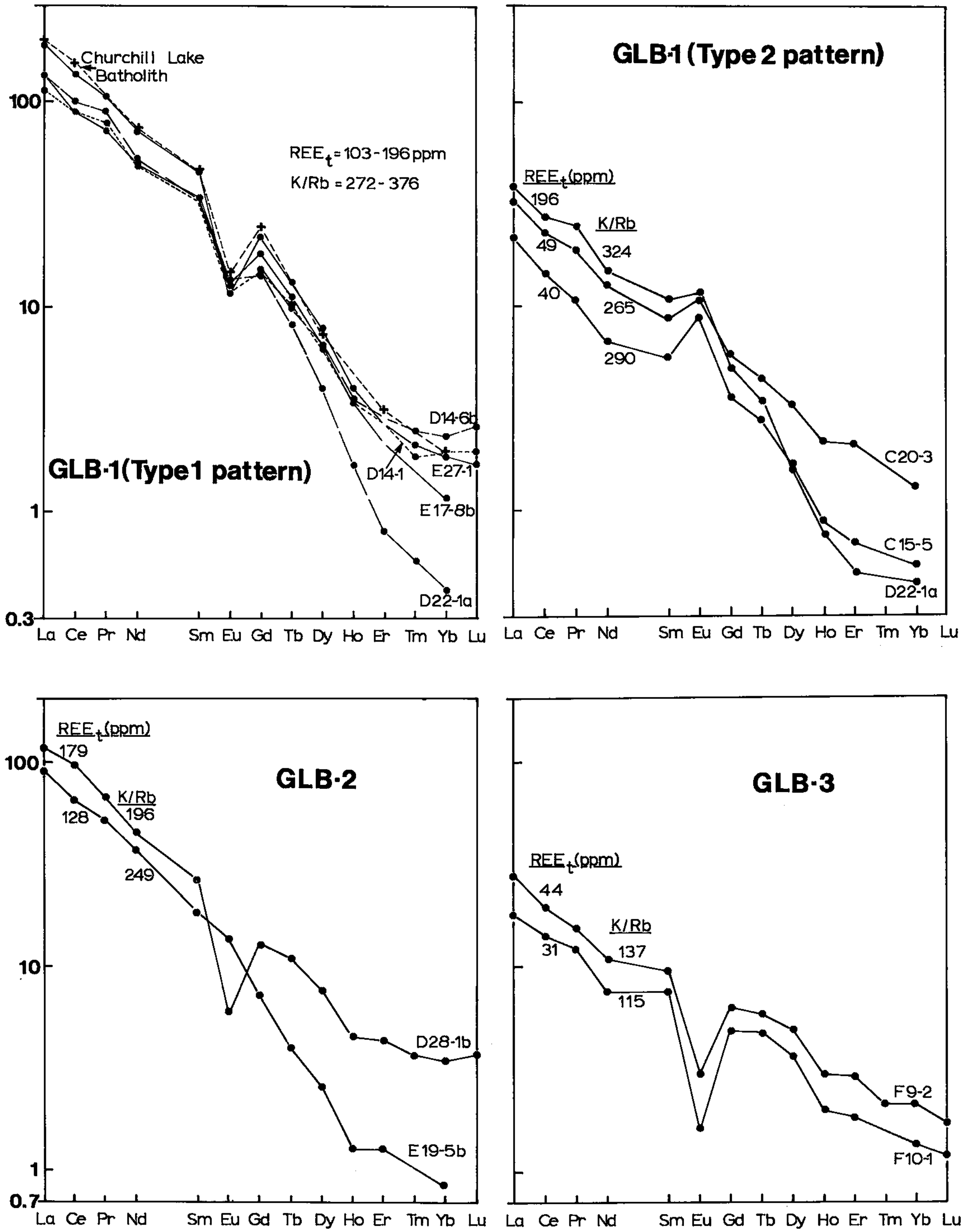

FIG. 18. Chondrite-normalized *REE* diagrams for units GLB–1, –2 and –3.

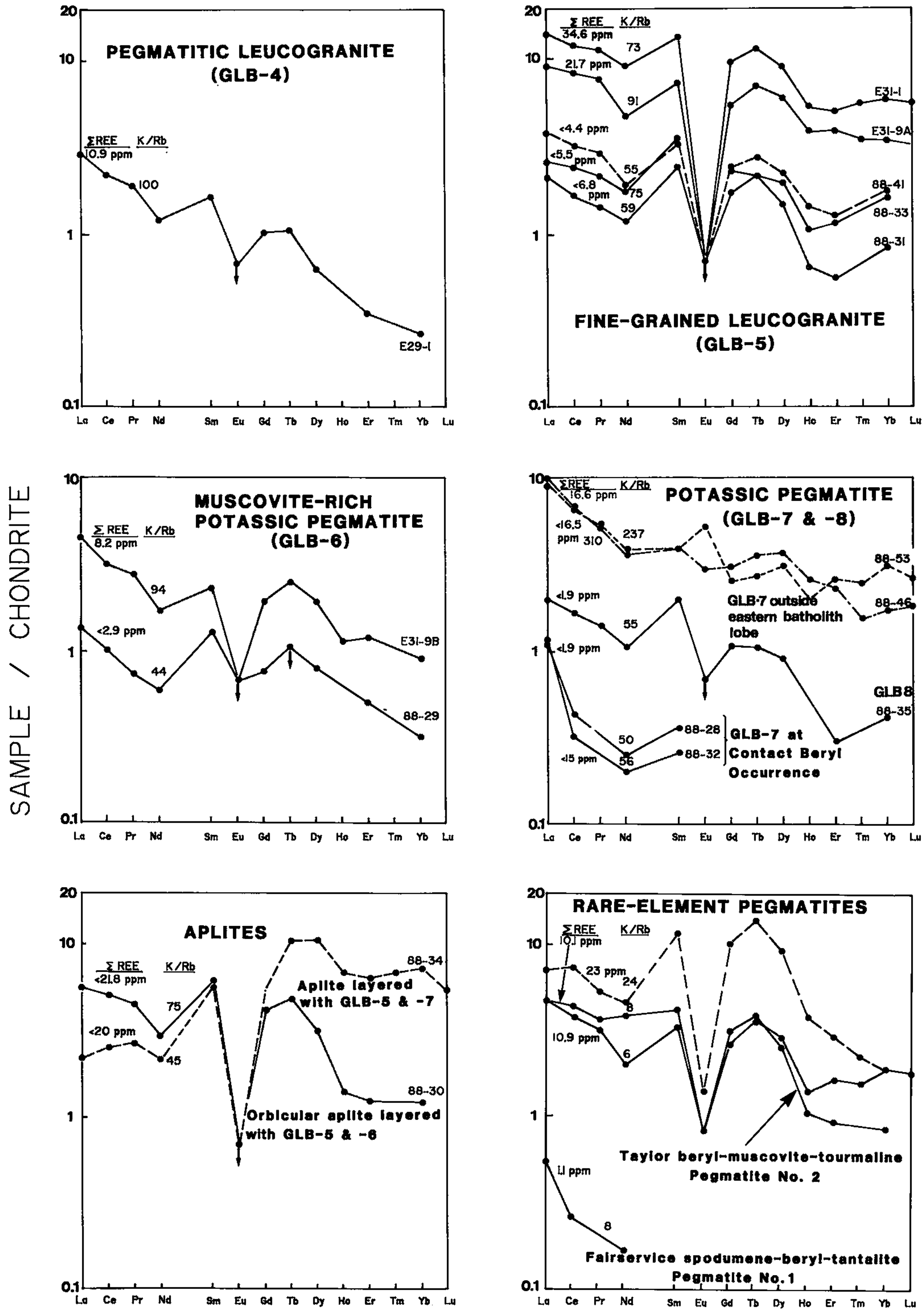

FIG. 19. Chondrite-normalized *REE* diagrams for various pegmatitic granites from the eastern lobe of the Ghost Lake batholith and adjacent rare-element pegmatites of the Mavis Lake Group.

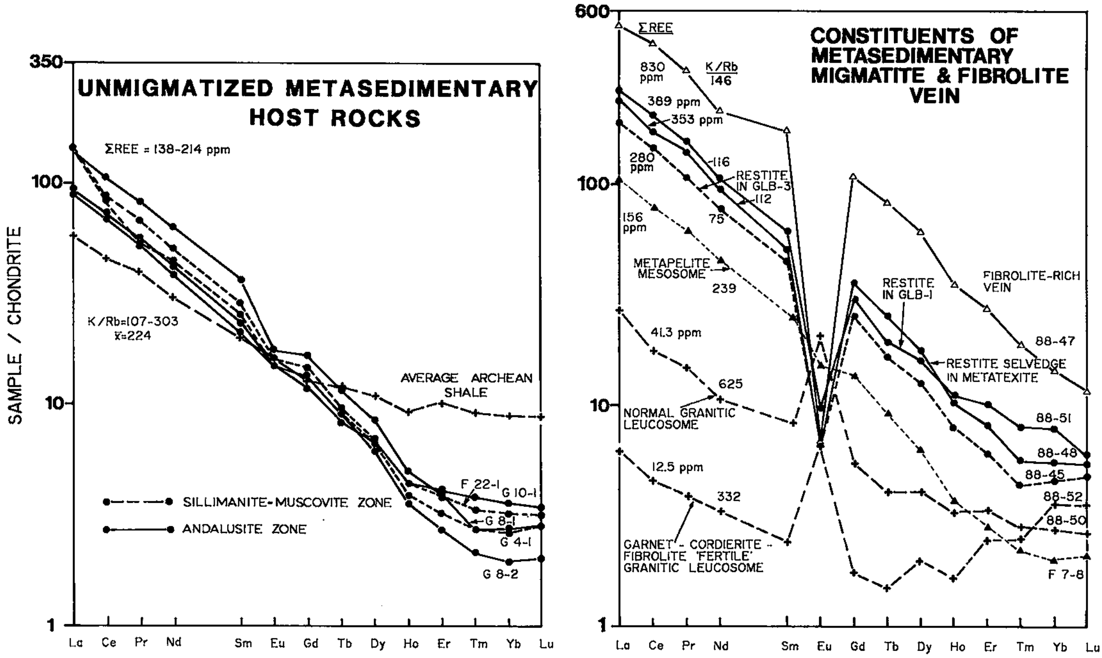

FIG. 20. Chondrite-normalized *REE* diagrams for unmigmatized Zealand metasedimentary rocks and derived migmatite components.

ized profiles (Figs. 18, 19). In the case of units GLB–5, –6, –7, aplite and the Taylor beryl pegmatites, the latter two inflections occur within a sinuous pattern for the *HREE*, and have been documented elsewhere by Kontak (1990, p. 809). It is uncertain what process or mineral(s) produced the above inflections; however, similar inflections of generally lower values also occur in GLB–1 (type-2 pattern), GLB–2 and –3 (Fig. 19) and may be another expression of consanguinity of the batholith units.

Rare-element granitic pegmatites

Partial to complete chondrite-normalized patterns for four rare-element pegmatites are given in Figure 19. Patterns of the Taylor beryl pegmatite No. 2 display a striking similarity to those of the fine-grained leucogranite and potassic pegmatite units within the eastern lobe of the Ghost Lake batholith, as revealed by total *REE* between 1 and 12× chondrite, $(Ce/Yb)_N$ between 2.8 and 3.1, and Eu/Eu* less than 0.11. The *REE* are most severely depleted in the bulk sample of the Fairservice No. 1 pegmatite, which contains less than 1.08 ppm.

Černý *et al.* (1981, p. 146) were among the first to point out that shallow to flat chondrite-normalized patterns featuring drastic depletion in total *REE*, coupled with large negative Eu anomalies are characteristic of Archean fertile granites and related rare-element pegmatites. The greater degree of *LREE* depletion relative to *HREE*, coupled with the significant drop in total abundance, have been ascribed to presence of biotite and apatite in the residue (Puchelt & Emmermann 1976, Emmermann *et al.* 1975, Hanson 1978), fractional crystallization of monazite and allanite (Miller & Mittlefehldt 1982), and fluoride complexing into an aqueous phase (Flynn & Burnham 1978, Alderton *et al.* 1980, Muecke & Clarke 1981). The petrographic data favor the role of fractional crystallization of monazite and apatite as the main control in *LREE* depletion in the transition from GLB–1 and –2 to the first stage of fertile granite (GLB–3). Some units (GLB–5 and aplite) of the second-stage fertile granite, however, possess a much larger negative Eu anomaly and generally lower total *REE*. As these units are devoid of monazite and apatite, we suggest that fluoride-complexing into an aqueous phase may have locally influenced the *REE* patterns developed in the eastern lobe of the batholith, particularly around the Contact Beryl Occurrence. This is implied by modest to high levels of F found in the syn- and late magmatic contact-metasomatic units and the tendency for *REE*-enrichment in these altered units (Breaks 1989). Such metasomatic units also contain sporadic but generally higher levels of Ba (172–668 ppm) and Rb (128–3370 ppm) than most adjacent GLB–5 to –8 units; this may also reflect mobilization *via* a fluorine-rich complex.

Volatiles

Boron is an important volatile element in GLB–5 to –8, in syn- and late magmatic metasomatic assemblages related to the Ghost Lake batholith, and in rare-element pegmatites of the Mavis Lake Group and its associated exomorphic aureoles. Boron undergoes enrichment in the progression from GLB–1 (mean 14 ppm, range 5–19 ppm) to GLB–3 (mean 27 ppm, range 10–130 ppm) and peaks in GLB–5 (mean 1850 ppm, range 1200–2500 ppm). Extreme concentrations up to 8700 ppm occur in the dravite – phlogopite – actinolite contact-metasomatic assemblage.

Fluorine exhibits a less definite trend; it varies between 60 and 770 ppm in most batholith units. Anomalies occur in the restite entrained in GLB–1 and –3 (1040–2520 ppm), "fibrolite"-rich veins (840–2520 ppm), metasomatic selvedges associated with GLB–1 (2250 ppm) and GLB–7 (1500–1860 ppm), and synmagmatic dravite – phlogopite – actinolite contact-metasomatic unit (0.86–2.04%). The last plausibly represents a depository for a significant quantity of fluorine, which infiltrated from pegmatite-forming melts that lacked an appropriate mineralogical host (*e.g.*, fluorite or apatite) for the excess fluorine. The legacy of fluorine enrichment of these melts is, however, reflected in modest contents of F in primary muscovite (1500–3750 ppm) in the derivative units of fertile pegmatitic granite.

Chlorine in the Ghost Lake batholith has a range similar to that of F (<30 to 532 ppm); however, its trend toward enrichment is converse to fluorine, *i.e.*, the highest values are reached in GLB–1, and concentrations are comparatively lower in the pegmatitic granites (<30–60 ppm).

Petrochemistry of Metasedimentary Rocks and Derived Migmatite Constituents

Very few studies of the genesis of rare-element pegmatites have provided meaningful data regarding distribution of rare elements in potentially important protoliths (*e.g.*, Černý 1989b: Wekusko Lake and Cat Lake – Winnipeg River pegmatite fields). Several investigators have considered clastic metasedimentary rocks as the initial concentrators of such elements, along with important volatiles B, F and Cl (Norton 1973, Stewart 1978, Breaks *et al.* 1985, Kretz *et al.* 1989). In the Dryden area, there are unmigmatized and migmatized clastic metasedimentary rocks adjacent to the Ghost Lake batholith, that demonstrate the generation of anatectic melts, and thus are a plausible source of the batholith and associated rare-element pegmatites. In this section, we consider the disposition of the rare elements and boron during successive stages of crustal anatexis promoted by muscovite- and biotite-dehydration reactions.

Previous investigations

The fate of lithium in the metamorphic segment of the geochemical cycle is poorly understood relative to its behavior in the surficial, sedimentary and igneous environments (Heier & Adams 1964, Ronov *et al.* 1970, Heier & Billings 1972). Only a few investigators have examined the variation of Li in the transition from medium-grade to high-grade migmatitic rocks. Holland & Lambert (1972) concluded that anatexis of granitic gneisses and metasedimentary rocks would produce a leucosome component depleted in Li and a corresponding biotite-rich restite enriched in this metal. Hence, they inferred that biotite dehydration under granulite-facies conditions would be necessary in order to effectively release Li to anatectic melts. Sighinolfi & Gorgoni (1978) encountered a pattern of depletion of Li in metasedimentary rocks of the western Italian Alps, marked by an 80% decrease in abundance of Li from the amphibolite to the granulite facies.

An important question concerns the fate of rare elements and B during successive stages of crustal anatexis abetted by dehydration reactions involving muscovite and biotite. Breaks (1982, 1991) found that metasedimentary rocks (wackes and pelites) in the English River Subprovince underwent a pronounced depletion (80%) in Li content in the transition from the muscovite–sillimanite zone to the sillimanite – K-feldspar migmatite zone (Table 4). A further 56% decrease in Li is observed in the metapelites in low-pressure granulite zones, whereas wackes show little additional loss of Li (Table 3). These data demonstrate that loss of Li in regionally metamorphosed clastic sedimentary terranes is step-wise and that significant release of Li is associated with muscovite-dehydration reactions. Therefore, the anatectic fractionation of Li is not strictly contingent upon biotite-dehydration reactions in the restite component, as has been commonly proposed in the literature (Holland & Lambert 1972, Černý 1982, p. 441). Furthermore, exchange reactions between the phlogopite component of biotite (common in restite) and a chloride-bearing fluid may potentially release Li (Bos 1990) to anatectic melts that have achieved water saturation without requiring the decomposition of biotite.

Dryden area

In order to determine the distribution of rare elements under a spectrum of metamorphic conditions represented by the low-, medium- and high-grade migmatitic facies of the Zealand metawackes and metapelites (1.58 < A/CNK_{mol} < 3.05). Concentrations of B, Be, Cs, F, and Li were determined in 100 samples, and *REE*, Rb and

TABLE 4. SUMMARY OF PETROGENETIC MODEL FOR THE ASSOCIATION GHOST LAKE BATHOLITH - RARE-ELEMENT PEGMATITE OF THE DRYDEN FIELD

Petrogenetic stage	Age (if known)	Process(es)	Resulting Rock Unit(s)	Mechanism(s) of rare element concentration
1	2714 Ma[1]	Clastic sedimentation into large basins.	Wacke, mudstone, arkose, banded iron formation	Surface adsorption and substitution in structure of clay minerals.
		High grade metamorphism and intitial migmatization of Zealand metasedimentary rocks due to muscovite dehydration	Metatexite	Anatectic fractionation of some rare elements (Li, Be) and B into "fertile" leucosomes and coexisting melanosomes.
2	2710+5/-2 Ma to 2701.9+3.9/-3.0 Ma[2]	Crustal thickening effected by volcanic arc-continent collision. Granulite-grade metamorphism causes extensive anatexis of Zealand metasedimentary rocks via biotite dehydration.	Out-of-sequence supracrustal rocks in Dryden-Sioux Lookout region	Redistribution of rare elements via base cation leaching. Fertile leucosomes incorporated into more extensive granitic melt systems via increasing partial melting.
	2685[3]	Emplacement of Ghost Lake Batholith	Primitive S-type peraluminous granite (GLB-1 and -2: biotite and cordierite-biotite granite)	Crystal-melt fractionation progressively enriches residual melt in rare elements. Also rare elements held in entrained restite, and cordierite in phenocrysts and clotty aggregates with quartz. Further rare element redistribution by base cation leaching.
3		Extraction of residual melt from GLB-1	First fertile granite mass GLB-3: biotite-muscovite granite and related potassic pegmatite	Continued concentration of rare elements, B and F via crystal-melt fractionation. Final event of base cation leaching overprints GLB-3.
4		Extraction of residual melt from GLB-3	Second fertile granite mass (GLB-4 to -8) forming a pegmatitic granite facies with local endogeneous rare element mineralization.	Maximum accumulation of rare elements, B, F in batholith apical zone. Widespread exsolution of volatile saturated phase. Egress of fluid phase enriched in these elements to host rocks prior to crystallization of pegmatite granite units.
5		Dispersion of rare-element-enriched melt/fluid masses along host rock anisotropies.	Regionally zoned rare-element pegmatite field (Mavis Lake Pegmatite Group)	Concentration mechanisms uncertain. Possibly crystal-melt fractionation in combination with liquid state thermogravitational diffusion.
		Metasomatic interaction of pegmatite-derived fluids with mafic host rocks.	Amphibolites enriched in tourmaline, holmquistite and rare-element-enriched biotite.	Diffusion and possibly volatile transfer.

1. Davis and Edwards (1982)
2. Corfu (1988)
3. D. Davis (preliminary monazite age)

Ta, in part of the suite [Breaks (1989), summarized in Table 3]. Lithium and boron will be used here as the principal "tracers", because better-quality and more complete data are available for them than for the other elements.

The contents of lithium in these metasedimentary rocks displays a strong relationship with metamorphic grade. Metapelites show varying degrees of enrichment in Li over the average Archean shale of Cameron & Garrels (1980: 42 ppm), attaining substantial concentrations reaching 375–590 ppm, amongst the highest recorded in the literature [only Shaw's (1954) value of 630 ppm Li is higher]. Average concentrations of Li in the chlorite–biotite and andalusite zones are broadly similar; however, enrichment of 2.3–2.8 times the value for the average Archean shale is evident for the sillimanite–muscovite zone. A sudden drop in average levels of Li to 33% of that in the sillimanite–muscovite zone occurs in the migmatitic K-feldspar – sillimanite zone, which is virtually identical to metapelitic inclusions in GLB–1. A similar pattern of distribution and extent of Li loss (35%) is obvious for metawacke samples in the transition into the migmatitic zones (Table 3).

The distribution of boron in the metasedimentary rocks also varies significantly with metamorphic grade. Metapelites exhibit a similar pattern of average enrichment and depletion of B to that of Li, increasing initially by a factor of 2× in transition from the andalusite (65 ppm) to the sillimanite–muscovite (142 ppm) zones before more drastically diminishing to less than 8 ppm in the K-feldspar – sillimanite migmatitic zone. It should also be noted that the mean concentrations of boron of the unmigmatized Zealand metasedimentary rocks are 2–4 times higher than that of the average Archean shale (33 ppm: Cameron & Garrels 1980). In the migmatitic zone, the level of concentration of B is considerably depleted (range <8–16 ppm) and compares with that of the lower continental crust (9.3 ppm B: Truscott *et al.* 1986), though generally higher than metapelites or metawackes from the nearby granulite zones in the English River Subprovince (Breaks 1991) as at eastern Lac Seul (1.1–2.6 ppm boron: unpubl. results of eight analyses). For metawackes, a similar pattern of enrichment or depletion is once more apparent (Table 3).

Variation in cesium concentrations as a function of metamorphic grade is easily assessed, as many analytical results fall below the 10 ppm detection limit. Nevertheless, many samples of metapelite and metawacke exhibit appreciable enrichment (31–150 ppm), significantly exceeding the ranges given for shales by Canney (1952: <2–18 ppm) and Horstmann (1957:<2–14 ppm). Furthermore, the Rb/Cs ratio, which ranges from 0.8 to 13 (mean = 6.5), suggests significant fractionation of Cs relative to Rb in the Zealand metasedimentary rocks compared to the ratio in the average crust (31: Taylor & McLennan 1985). Elsewhere, however, Nagaytsev & Popova (1988) have shown in regionally metamorphosed metasedimentary rocks of the Baltic Shield that mean Cs contents vary little in the medium-grade zones, but register a significant drop (by a factor of 4.5) as the grade increases from the muscovite–sillimanite to the sillimanite – K-feldspar zones, from 9.9 to 2.2 ppm.

Constituents of migmatites

Limited comparison of the pertinent chemistry of mesosome, leucosome and restite constituents of migmatitic metasedimentary rocks that contain low amounts of leucosome provides some data on the initial fractionation of the rare elements. Two types of low-leucosome-fraction, *in situ* migmatites (*i.e.*, metatexites) can be distinguished in the Dryden area on the basis of chemistry: "normal" leucosomes and "fertile" leucosomes. "Normal" granite leucosomes show a typical impoverishment in rare elements, B and F, which would suggest that these are unlikely sources of fertile granites (analysis 88–50, Table 2). The "fertile" tourmaline – garnet – cordierite – "fibrolite" leucosome (analysis 88–52, Table 2), on the contrary, is very strongly peraluminous (A/CNK_{mol} = 3.45) and displays substantial enrichment in Be (99 ppm) and a modest increase in Li (62 ppm) relative to the normal leucosome. As with fertile granites, these leucosomes have greatly depleted levels of total *REE* (12.5 ppm), similar to several units in the eastern lobe of the batholith. More substantial concentrations of B (2600 ppm), Be (118 ppm), Cs (90 ppm), F (1600 ppm), Ga (50 ppm), Rb (421 ppm), Sn (12 ppm) and Ta (14 ppm) characterize the adjacent biotite – "fibrolite" – tourmaline restite selvedge relative to the normal leucosome.

Chondrite-normalized *REE* patterns for low- to medium-grade metapelites and derived constituents of migmatite are displayed in Figure 20. The unmigmatized metapelites and derived mesosome (F7–8 in Fig. 20) have similar patterns, featuring a much higher $(Ce/Yb)_N$ and small, generally negative Eu anomalies (Eu/Eu^* between 0.58 and 0.91) relative to the average Archean shale of Taylor & McLennan (1981). *REE* patterns for most of GLB–1 and the metapelites are very similar (*cf.* Figs. 18 and 20). Furthermore, there is some similarity between Type-2 *REE* patterns in the Ghost Lake batholith and the normal leucosome in terms of $(Ce/Yb)_N$ (5–18 *versus* 6.5) and total *REE* (25–72 ppm *versus* 41 ppm). There is dissimilarity, however, in the more prominent positive Eu anomaly in the normal leucosome (Eu/Eu^* 3.1 *versus* 1.95). Samples of restite, regardless of whether the material was derived *in situ* or entrained in GLB–1 or –3, have virtually identical, modestly sloped patterns [$(Ce/Yb)_N$ = 27–37] marked by a substantial negative Eu anomaly (Eu/Eu^* between 0.16 and 0.24). This suggests derivation through some common process. The "fibrolite"-rich vein that postdates GLB–1, described previously, also has a similar pattern, but with a significantly higher concentration of the *REE*.

To the writers' knowledge, the formation of fertile leucosomes in metasedimentary protoliths has not been previously described in the literature and offer additional support for the contention that transfer of some rare elements into the leucosome does, in fact, occur during incipient anatexis. The mechanism of concentration of Be and Li may relate to the complexing agency of boron (London 1987).

DISCUSSION

Substantial progress has been made, particularly within the past fifteen years, in the elucidation of the petrographic and chemical attributes, possible petrogenetic mechanisms, tectonomagmatic environments, and experimental investigation of pertinent melt systems responsible for the generation of peraluminous granites and associated mineralization (*e.g.*, White & Chappell 1977, Le Fort 1981, London 1986a, Wickham 1987, Wall *et al.* 1987, Holtz & Johannes 1991). The Ghost Lake batholith exhibits many of the field, mineralogical and chemical characteristics of S-type peraluminous granites (Chappell & White 1974). This interpretation is strongly supported by the geological context of this batholith and similar masses of peraluminous granite found in the English River Subprovince, situated 60 km to the north (Breaks 1991). In the Superior Province, one- and two-mica peraluminous granites are strongly confined to regional belts dominated by high-grade migmatized metasedimentary rocks (Breaks *et al.* 1985, Percival 1989, Williams 1991). In the English River

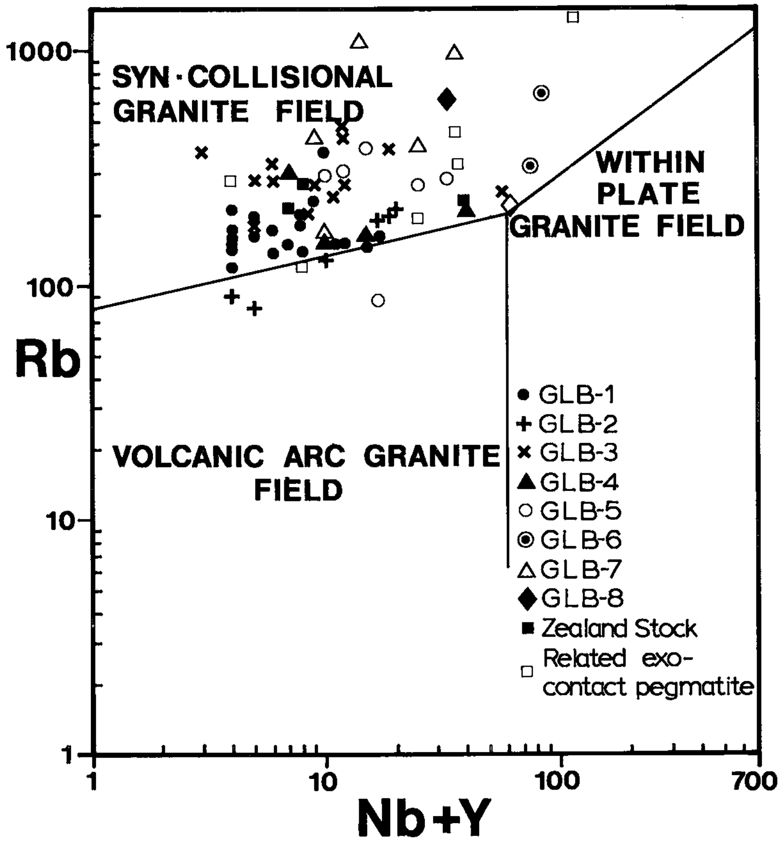

FIG. 21. Rb *versus* Nb+Y plot, after Pearce *et al.* (1984), for internal units of the Ghost Lake batholith, the Zealand stock and related, barren, exocontact pegmatites.

Subprovince, these granites collectively underlie over 5,000 km^2.

The tectonic setting is inferred to be of synorogenic collisional type, marked by crustal thickening within the 15–40 km boundary zone between a 2775–2725 Ma volcanic arc (Wabigoon Subprovince: Blackburn *et al.* 1991) and *circa* 3 Ga sialic microcontinent now only sporadically preserved within the Winnipeg River Subprovince to the north (Beakhouse 1991). This inferred tectonic setting is supported by chemical discrimination diagrams, notably the Rb *versus* Y+Nb plot of Pearce *et al.* 1984, shown as Figure 21 and the A/CNK$_{mol}$ *versus* A/NK$_{mol}$ diagram (Fig. 22) of Maniar & Piccoli (1989), in which the vast number of Ghost Lake batholith samples fall into the syncollisional fields.

The S-type petrographic characteristics of the Ghost Lake batholith are supported in the chemically primitive facies by ubiquity of metasedimentary enclaves and cordierite–biotite restite, in addition to phenocrystic cordierite. White *et al.* (1986, p. 116) regarded such cordierite as "the most reliable mineralogical indicator of strongly peraluminous composition (and therefore S-type affinities)". The Ghost Lake batholith contrasts, however, with the higher-crustal-level variant of S-type granites, as exemplified by those in the Lachlan Fold Belt of Australia (Chappell *et al.* 1987) or some Hercynian plutons of western Europe (*e.g.*, Portugal: Holtz & Barbey 1991) by the lack of contact metamorphic aureoles and by widespread pegmatitic granite segregations.

Furthermore, the Ghost Lake batholith contrasts with the Australian S-type granite complexes (*e.g.*, Phillips *et al.* 1981, Price 1983) and Archean mafic peraluminous granites of the English River Subprovince (Breaks 1991, unpubl. data) in its fractionation path to high Na and generally lower levels of Mg and total Fe. Chappell & White (1974, 1982) considered low Na (<3.2% Na_2O in rocks with 5% K_2O) as characteristic of S-type granites of Australia. However, high-Na examples, such as the Manaslu pluton (Le Fort 1981, France-Lanord & Le Fort

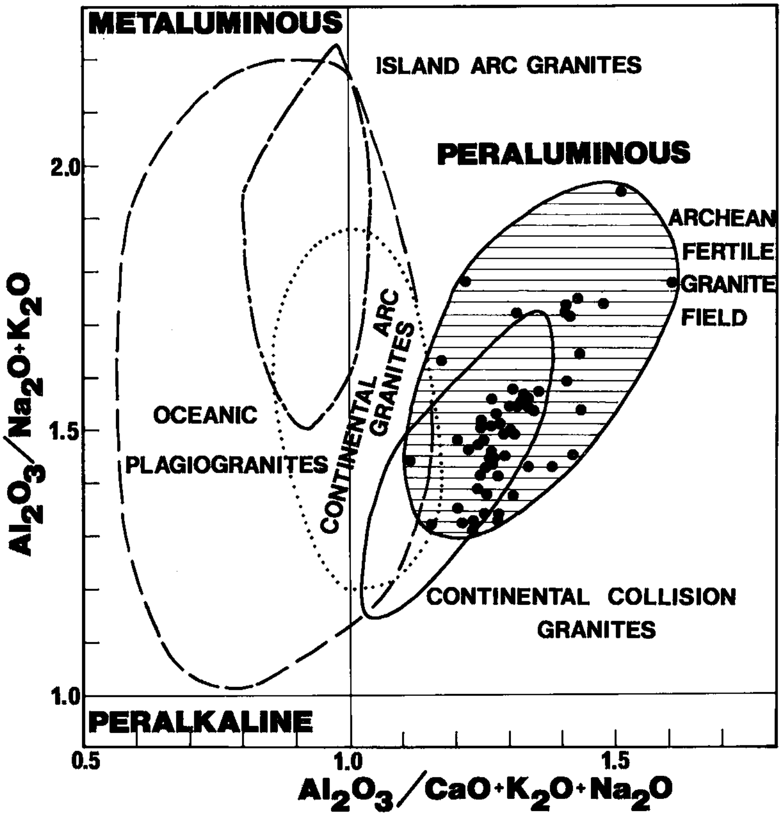

FIG. 22. A/CNK$_{mol}$ *versus* A/NK$_{mol}$ tectonic discrimination plot of Maniar & Piccoli (1989) applied to internal units of the Ghost Lake batholith, the Zealand stock and related, barren, exocontact pegmatites, and field proposed for Archean fertile granites of the Superior Province, based upon the wide range in fractionation provided by this batholith.

1988), may occur in tectonic settings conducive to S-type magma generation, *i.e.*, zones of continent-continent collision.

Sodium-rich granitic melts have been produced by partial melting of natural greywacke protoliths under H_2O-saturated conditions in the presence of a chloride solution (Rensselaer greywacke: Kilinc 1972); however, such melts would have little success in rising to higher levels in the crust. The derivation of the highly fractionated Na-rich leucogranite compositions, similar to GLB–5, directly from anatexis of greywacke protoliths is, furthermore, considered unlikely as corresponding Na-rich leucosomes have not been found in the metatexites of the Zealand metasedimentary unit. Alternatively, we favor a high-Na-generating mechanism due to the increasing influence of H_2O and B. The chemically primitive rocks of GLB–1 and –2 are relatively potassic and enriched in normative quartz in comparison to the more evolved pegmatitic granites (Breaks 1989, p. 254). The experimental investigations of the H_2O-undersaturated system Qz–Or–Ab–Al_2O_3–H_2O–CO_2 by Holtz *et al.* (1992b) indicate a shift of the minimum composition and quartz–feldspar cotectic boundary toward increasing quartz, in comparison to the subaluminous haplogranite system (Tuttle & Bowen 1958, Luth *et al.* 1964). Therefore, the normative quartz- and orthoclase-rich compositions of GLB–1 may indicate an undersaturation in H_2O relative to the more widely pegmatitic stage-1 and -2 fertile granites. The progressive increase in normative albite in GLB–3 of stage-1 and GLB–5, the dominant unit of stage-2, consequently can be interpreted in the light of the experimental data of Holtz *et al.* (1992a,b) as owing to the increasing accummulation of H_2O in the residual granitic melts. In the case of GLB–5, the greater amount of normative albite may have been abetted by higher levels of boron (0.39–0.81%

B_2O_3). Boron is known to move the composition of the minimum melt in the haplogranite system toward the albite apex while significantly expanding the field of quartz (Pichavant & Manning 1984, Pichavant & Ramboz 1985, Benard *et al.* 1985, Pichavant 1987).

Another mechanism capable of generating high levels of Na includes anatexis of a compositionally heterogeneous protolith (Holtz 1989, Ortega & Gil Ibarguchi 1990). This process may have relevance to the study area, as the inferred tectonic setting would be conducive to partial melting of a tectonically interleaved and diverse crustal *mélange* composed of metasedimentary rocks (wacke-mudstone), tonalite, and felsic to mafic metavolcanic rocks. However, an evaluation of this mechanism is beyond the scope of this paper.

The geological setting of the Ghost Lake batholith demonstrates an evolutionary link between low-leucosome-fraction metasedimentary migmatites, extensive masses of anatectic, cordierite-bearing, S-type granite, chemically evolved fertile leucogranites (locally with endogenous mineralization such as beryl and cassiterite), and an exocontact rare-element pegmatite field. In consequence, we propose that a cordierite–biotite granite facies be added to the idealized cross-section of a fertile granite pluton of Černý & Meintzer (1988, p. 182). In other plutons of fertile granite, this cordierite-bearing facies may either lie unexposed or simply have gone unrecognized, as cordierite is commonly altered to a host of secondary minerals.

The most probable source of peraluminous granitic melt, such as was emplaced to form the Ghost Lake batholith, is the extensive anatexis of the Zealand wacke–mudstone succession. Under fluid-absent granulite-facies conditions, biotite dehydration would have yielded copious quantities of melt (Vielzeuf & Holloway 1988) that incorporated rare elements and boron, already enriched in "fertile" leucosomes and biotite-rich selvedges. Although granulites are not exposed in the study area, they may be predicted to occur at deeper levels, as in the geologically similar English River Subprovince (Breaks 1991).

The absence of linear trends in most of the Harker plots of the Ghost Lake batholith, and the overlapping fields of data, particularly for the alkalis and alkali-earth elements, represent a departure from the linear arrays of data of suites of peraluminous granite whose petrogenesis was dominated either by processes of fractional crystallization or restite unmixing (Phillips *et al.* 1981, Chappell *et al.* 1987). Alternatively, it is suggested by the data that other superimposed endogenic processes effected redistribution of many elements. In our opinion, crystal–melt fractionation trends in the Ghost Lake batholith were obscured by the effect of extraction of successive batches of residual melt after the crystallization of the early units GLB–1 and –2, and the initial stage of fertile granite (GLB–3), which may have been compounded by the episodic redistribution of elements *via* base-cation leaching.

The hypothesized extraction of successive batches of fertile granite melt is difficult to establish from field relations. Although the parental melts, now represented by GLB–1 and GLB–3, commonly express local attainment of volatile saturation in the form of ubiquitous pods of pegmatite, no cupola zones were encountered within these units where ponding of pegmatite-forming melt could have occurred. Chemically, the presence of the successive batches of fertile granitic melt may be inferred from the K versus Rb diagram which, in particular, reveals overlapping fields broadly correlative with increasing fractionation. The advancement in chemical evolution in the Ghost Lake batholith also correlates with a decrease in areal extent of the units: GLB–1+2 > GLB–3 > composite mass of GLB–4 to –8.

THE POSSIBLE ROLE OF BASE-CATION LEACHING IN THE GENESIS OF FERTILE GRANITES

Vernon (1979) developed a chemical model to explain the formation of late sillimanite in migmatitic, metapelitic gneisses of the Cooma complex, New South Wales. His model proposes that retrograde replacememt of cordierite released H^+ ions, which subsequently led to muscovite–"fibrolite" replacement of K-feldspar. Elsewhere, several studies summarized by Kerrick (1990, p. 346-352) also have indicated a growing consensus that "fibrolite"-rich veins develop by metasomatism by Al–B–alkali-bearing fluids. These models are applicable to the Ghost Lake batholith and adjacent migmatized metasedimentary rocks, as there are widespread "fibrolite" veins in these rocks. "Fibrolite" and secondary muscovite commonly overprint microcline, cordierite and plagioclase in units GLB–1, –2 and –3 in a manner very similar to that documented by Vernon (1979). In particular, cordierite is variably replaced by a complex symplectite composed of the disharmonious-type sillimanite of Vernon & Flood (1977), secondary biotite, chlorite, muscovite and rare andalusite.

The "fibrolite"-rich veins in the present study can thus be explained as products of base-cation leaching by acidic fluids, which was repeated several times during the early evolution of the Ghost Lake batholith. As a consequence of the leaching process, rare elements previously locked in microcline (Rb, Cs, Ga and Li), biotite (Be, Cs, Ga, Li, Ta, Nb and Rb) and cordierite (Be, Cs, Li and Rb) were released into an infiltrating fluid. In addition, the presence of dumortierite suggests that boron played a role as a complexing agent.

Base-cation leaching may play a role in the subsolidus mobilization of Na, K, Si, rare alkalis and volatiles such as B and F from early units of fertile granite plutons and supracrustal host-units. Fluids enriched in these elements and mobilized along channels in the early, consolidated granitic units also could potentially have interacted with contemporaneous, evolving residual melts. Although a genetic connection of the systems of "fibrolite"-rich veins with the residual melts that were inferred to have been extracted from GLB–1 and –3 cannot be proven from field evidence, it is attractive to consider this, as the process of base-cation leaching spanned the time from early migmatization of the host metasedimentary rocks to after the crystallization of the first-stage fertile granite (GLB–3).

Petrogenesis

We propose that generation of the peraluminous granite – rare-element pegmatite association in the Dryden Field resulted from a complex interplay of petrogenetic processes involving as many as five stages (Table 3).

Stage 1 was marked by the initial anatexis of interbedded wacke–mudstone of the Zealand sequence under high-grade regional metamorphism, which produced low-leucosome-fraction metatexite migmatites, probably *via* muscovite dehydration. The initial concentration of the rare elements and B in the protoliths occurred during the development of regionally extensive clastic sedimentary basins of turbiditic derivation in the Sioux Lookout Terrane and nearby English River Subprovince at about 2714 Ma (Davis & Edwards 1982). Concentration was mainly accomplished through adsorption and substitution into the structures of clay minerals and possibly augmented by detrital phases (*e.g.*, Sn, Nb in oxide minerals) that formed the matrix of the derivative wacke and mudstone units. These concentrations may have been augmented, by unknown processes, during regional metamorphism up to sillimanite–muscovite grade. The initial anatexis records the first depletion of rare elements and B and their redistribution into certain "fertile" leucosomes.

Stage 2 was a period of crustal thickening in the Sioux Lookout Terrane (Beakhouse 1988) by the collision, bracketed between 2710^{+5}_{-2} and <2698 Ma (Corfu 1988, Davis *et al.* 1988, Beakhouse & McNutt 1991), of a 2775–2725 Ma ensimatic volcanic arc that now comprises the Wabigoon Subprovince (Blackburn 1991) with an older, *circa* 3 Ga (Krogh *et al.* 1976) sialic microcontinent forming the Winnipeg River Subprovince.

Ultimately, the process of crustal thickening led to further high-temperature – low-pressure metamorphism, more extensive anatexis of the wacke–mudstone metasedimentary sequence, and the widespread generation of peraluminous, S-type granitic melts of the collisional or Hercinotype variety (Pitcher 1982). Dehydration of biotite under fluid-absent, granulite conditions yielded granitic melt enriched in rare elements and B, that intruded higher crustal levels to form the Ghost Lake batholith and related bodies at *circa* 2685 Ma (preliminary U–Pb monazite age, D.W. Davis, pers. comm., 1992). Crystal–melt fractionation led to gradual accumulation of rare elements, B and H_2O into residual melt and metasomatic selvedges, where they were further concentrated by base cation leaching.

Stage 3 is interpreted as the initial extraction of fertile residual granite melt, which crystallized as muscovite–biotite granite of GLB–3. This unit is marked by further accumulation of most rare elements *via* crystal–melt fractionation and base-cation leaching, but not to levels required to form endogenous mineralization such as beryl in the consanguineous pods of potassic pegmatite.

Stage 4 represents the late magmatic evolution of the Ghost Lake batholith. It is featured by the maximum concentration of rare elements, B and H_2O within the batholith and possibly reflects a second stage of extraction of fertile granitic melts, which was most plausibly derived from GLB–3. This melt ponded within a cupola and crystallized as a complex pegmatitic granite facies within the eastern lobe of the batholith. It is characterized by widespread petrographic features indicative of exsolution of a volatile-saturated phase: marked coarsening of grain size, ubiquitous layering of the constituent units, directionally developed textures and rare miarolitic cavities. Early dispersal of fluid enriched in most rare elements, B, F, Sn and W occurred prior to crystallization of the pegmatitic granite facies. Fluid emigrated along early shear-zones and formed contact-metasomatic deposits characterized by dravite, holmquistite and phlogopite.

Stage 5, the final event, led to dispersion of masses of rare-element-enriched melt + fluid from the roof zone of the batholith upward along planar discontinuities to form an exocontact, regionally zoned rare-element pegmatite aureole known as the Mavis Lake Group. The late emplacement of the rare-element pegmatites is established by intrusion of the synmagmatic metasomatic zone by pods of cassiterite–beryl potassic pegmatite in the beryl zone of the Mavis Lake Group.

Conclusions

Important conclusions derive from this study:

(1) The Ghost Lake Batholith is a *circa* 2685 Ma, peraluminous, S-type granite – rare-element-pegmatite-generating system with a genetic linkage amongst the clearest yet described for any Superior Province pegmatite field.

2) This mineralized system is a ultimate product of a 2709–<2698 Ma arc–continent collision that engendered crustal thickening of clastic metasedimentary and mafic metavolcanic rocks. This resulted in Abukuma-type high-grade metamorphism, probably attaining granulite-facies conditions, which caused extensive intracrustal partial melting under fluid-absent conditions.

(3) The Ghost Lake batholith contains a chemically primitive facies similar to cordierite granites in younger orogens (Bea 1991). These rocks are rarely exposed in fertile granite complexes that have spawned rare-element pegmatites. Petrographic and chemical features demonstrate evolutionary links between a clastic metasedimentary source, primitive and fertile peraluminous granite and rare-element pegmatites.

(4) Rare elements, B and H_2O contents of the parent Ghost Lake batholith were concentrated to levels necessary to spawn rare-element pegmatites by crystal–melt fractionation, two stages of extraction of residual melt, and episodic base-cation leaching.

ACKNOWLEDGEMENTS

This paper is published with the permission of Vic Milne, Director of the Ontario Geological Survey. The manuscript benefitted from reviews by François Holtz, Dan Kontak, Robert F. Martin and an anonymous individual. We also thank François Holtz and Dan Kontak for useful preprints and reprints. The writers acknowledge useful discussions with Jack Satterly, who first recognized boron exomorphism in the Dryden area, Richard P. Taylor and Petr Černý, the latter two of which made several visits to the field area. Hugh de Souza, Srebri Petrov and Jean Richardson of the Geoscience Laboratory staff at the Ontario Geological Survey are thanked for their assistance in the generation of chemical and mineralogical data used in this study.

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Received February 19, 1992, revised manuscript accepted July 18, 1992.