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THE YITT-B PEGMATITE SWARM AT BERNIC LAKE, SOUTHEASTERN MANITOBA: A GEOCHEMICAL AND PARAGENETIC ANOMALY

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ABSTRACT

The YITT-B granitic pegmatite swarm, located in the Bird River Greenstone Belt of the Bird River Subprovince, western Superior Province of the Canadian Shield, consists of steeply dipping to subhorizontal dykes hosted by cordierite-bearing metaturbidites. Internal structure of the pegmatites is highly variable and, most significantly, is marked by mutual textural transitions and virtual paragenetic identity of zones that would normally be distinguished as layered albite-rich aplite and saccharoidal albite. The pegmatites are best classified as peraluminous, LCT-family, beryl – columbite – phosphate subtype of the rare-element class, but they are unique in many respects. Beryl and tourmaline are extremely rare, whereas Nb,Ta-oxide minerals and cassiterite are abundant. Primary phosphates and the Nb,Ta,Sn-oxide minerals are very fine-grained and dispersed in the albite-rich zones. With the exception of the phosphates, no other minerals show secondary alteration. Geochemically, the pegmatites are enriched in (Ta>Nb) and Sn, but are very poor in Li, Be, B, and F. The level of Fe–Mn fractionation is moderate, but that of Nb–Ta is very advanced. The YITT-B pegmatites have only a limited number of approximate analogs on a global scale, and they are unique within the Cat Lake – Winnipeg River pegmatite field. The diversity of pegmatite categories in this field, accentuated by the exotic features of the examined pegmatite swarm, reflects localized generation of granitic magmas from a broad spectrum of protoliths. Separate paths of ascent and internal differentiation of the parent granites compounded the effect of the heterogeneous source lithologies on the end-product pegmatites.

Keywords: granitic pegmatite, derivation, fractionation, tantalite, ferrotapiolite, ixiolite, phosphates, YITT-B pegmatite, Manitoba.

SOMMAIRE

L'essaim de venues de pegmatites granitiques de YITT-B, située dans la ceinture de roches vertes de Bird River, dans la sous-province de Bird River de la Province du Supérieur du Bouclier Canadien, comprend des filons à fort pendage ou bien subhorizontaux mis en place dans des métaturbidites à cordierite. La structure interne des pegmatites est très variable et, faits encore plus marquants, les transitions texturales sont mutuelles et la paragenèse, identique entre les aplites litées riches en albite et les zones à albite saccharoïdale. Les pegmatites sont peralumineuses, de la famille dite LCT, du sous-type à béryl – columbite – phosphate de la catégorie à éléments rares, mais elles s'en détachent à plusieurs niveaux. Le béryl et la tourmaline sont excessivement rares, tandis que les oxydes à Nb,Ta et la cassitérite sont abondants. Les phosphates primaires et les oxydes de Nb, Ta et Sn ont cristallisé en grains très fins dispersés dans les zones riches en albite. A l'exception des phosphates, les minéraux ne montrent aucun signe d'altération secondaire. Parmi les aspects géochimiques, les pegmatites sont enrichies en Ta, Nb (Ta > Nb) et Sn, mais sont très appauvries en Li, Be, B, et F. Le niveau de fractionnement impliquant le Fe et le Mn est moyen, mais le degré d'évolution en termes de Nb et de Ta est très avancé. Les pegmatites du cortège YITT-B n'ont que très peu d'analogues à l'échelle mondiale, et semblent uniques dans l'essaim de pegmatites du lac Cat et de la rivière Winnipeg. La grande diversité des catégories de pegmatites dans cette région, accentuée par les caractéristiques des pegmatites examinées ici, résulte de la génération localisée de volumes de magma granitique à partir d'un spectre assez large de compositions du protolithe. La grande variabilité des pegmatites dépendrait donc des effets composites des tracés individuels de l'ascension de ces volumes de magma et de leur différenciation interne, ajoutés à la variabilité de la composition de la source.

(Traduit par la Rédaction)

Mots-clés: pegmatite granitique, dérivation, fractionnement, tantalite, ferrotapiolite, ixiolite, phosphates, pegmatite YITT-B, Manitoba.

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INTRODUCTION

The Cat Lake – Winnipeg River pegmatite field in southeastern Manitoba is famous for its abundance and variety of granitic pegmatites, which include one of the largest pegmatite deposits of the world, the Tanco pegmatite at Bernic Lake (Černý *et al.* 1981, 1996). Although fairly well examined in the field and laboratory, this field still harbors some local pegmatite populations that have not been studied in detail. One of them is located north of the Tanco mine, on the YITT claims discovered in 1988 during examination of a regional lithogeochemical anomaly. Routine sampling and assaying of the pegmatites revealed somewhat elevated contents of Ta (Hood 1988). Anderson (1992) examined a specific segment of this pegmatite population, designated the YITT-B swarm, during a follow-up to the general exploration performed by the Tantalum Mining Corporation of Canada Ltd., W. C. Hood and Rothesay Resources Ltd. in 1986 – 1991. This swarm is considered representative for the local area.

Anderson's (1992) study indicated that the pegmatites examined represent a conspicuous anomaly, rich in Ta and Sn, but strongly impoverished in Li, Be, Mn, B, and F, with implications for genetic relationships among the pegmatite groups of the host field. Over and above its regional significance, the YITT-B swarm also shows paragenetic and geochemical features that are rarely encountered on a global scale, and their petrological implications are scarcely mentioned in the literature. We present here the results of the thesis research, complemented by additional examination of several mineral groups, and by general considerations of the YITT-B swarm in terms of pegmatite systematics.

LOCATION AND GENERAL GEOLOGY

The YITT-B swarm is located approximately 190 km ENE of Winnipeg, close to the Manitoba – Ontario boundary. The pegmatites outcrop less than 3 km northeast from the Tanco mine site (Fig. 1).

Geologically, the area is underlain by the Bird River Greenstone Belt of the volcano-plutonic Bird River Subprovince, in the Archean Superior Province of the Canadian Shield. The belt, flanked by domes of metatonalitic basement and granitic batholiths, forms an E–W-trending synclinorium comprised of metavolcanic and proximally derived metasedimentary rocks (Trueman 1980, Černý *et al.* 1981). These lithologies comprise six formations intercalated with, or intruded by, synvolcanic to late-tectonic intrusive rocks. The greenstone belt is dissected by numerous E–W-striking subvertical faults, and locally offset by NNW–SSE-striking cross-faults (Fig. 1).

The YITT-B swarm is hosted by the Booster Lake Formation, the youngest sedimentary suite of the greenstone belt, located in the core of the synclinorium. In the broader vicinity of the pegmatites examined, the Booster Lake Formation outcrops in an E–W-trending section about 1 km wide (Fig. 1). It consists of a well-bedded, metamorphosed sequence of greywacke – mudstone turbidites, locally with the greywacke component distinctly dominant over fine discontinuous layers of the mudstone – siltstone. The bedding has a very consistent orientation, striking 90–100° with a dip of 70–80° to the south.

The greywacke consists of felsic metavolcanic detritus, with subrounded fragments of quartz as the dominant component, ~5% subrounded grains of feldspar, and ~10% fine-grained biotite. The biotite has a preferred orientation defining a subtle schistosity parallel to the bedding. In contrast, the main constituents of the mudstone – siltstone are very fine-grained quartz with ~20% foliated biotite. Cordierite, usually strongly to completely transformed to chlorite, is common and locally attains concentrations as high as 30%. The biotite + cordierite assemblage is consistent with low-pressure lower amphibolite-facies metamorphism, in accord with the regional grade recognized for this area by Trueman (1980), as described in Černý *et al.* (1981).

THE PEGMATITES

Structural setting and deformation

The pegmatites constituting the YITT-B swarm, shown in Figure 2, can be subdivided into three principal types on the basis of their distribution and attitudes. Most of the dikes of the North-west, Raft, and Big outcrops are conformable to the bedding of the enclosing metaturbidites, but some are parallel to it only in strike-section and cross-cut it at low angles in dip-section. The South-west dike is vertical and strikes ~25–30° west of a vertical fault (with ~1 m dextral offset in outcrop) and ~65° east of this fault. The easternmost cluster of pegmatites consists of small, subparallel, vertically dipping dikes striking parallel with metaturbidite bedding, and one large subhorizontal dike.

The South-west dike attains a length of 215 m in outcrop, and several of the dikes are as much as 10 m thick. However, the overall dimensional relationships and shapes of the bodies are obscured because of lack of three-dimensional observations. The South-west dike seems to be a fairly simple body, but the pegmatites in the other outcrops are probably interconnected below the erosion surface, and may represent parts of a single anastomosing system.

The host metaturbidites underwent significant ductile deformation prior to intrusion of the pegmatites (Trueman 1980). In addition, the host rocks underwent substantial deformation subsequent to emplacement and solidification of the pegmatites, as indicated by the locally intense transposition and drag-folding of bedding immediately adjacent to the pegmatite contacts.

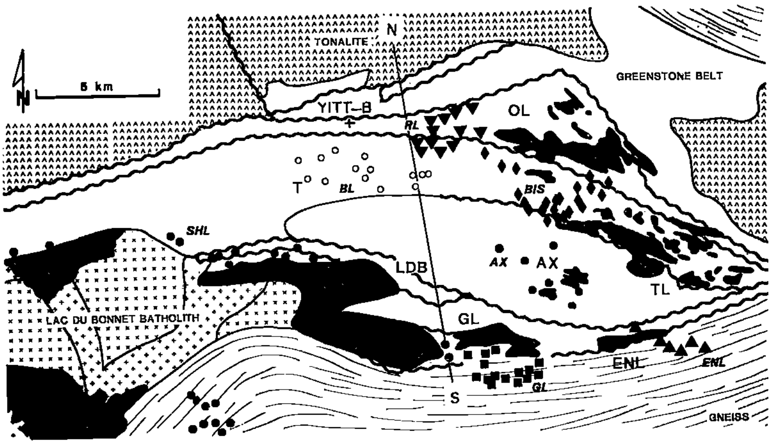

FIG. 1. Location of the YITT-B pegmatite swarm, northeast of the Tanco deposit (T), in the Winnipeg River segment of the Cat Lake – Winnipeg River pegmatite field in southeastern Manitoba. Distribution of fertile leucogranites (black; LDB, GL, ENL,TL, AX, OL) and pegmatite groups (*LDB, SHL, GL, ENL, AX, BIS, RL, BL*) in the Bird River Greenstone Belt (unpatterned); wavy lines denote major faults and shears; the N–S line marks the course of the section shown in Figure 12.

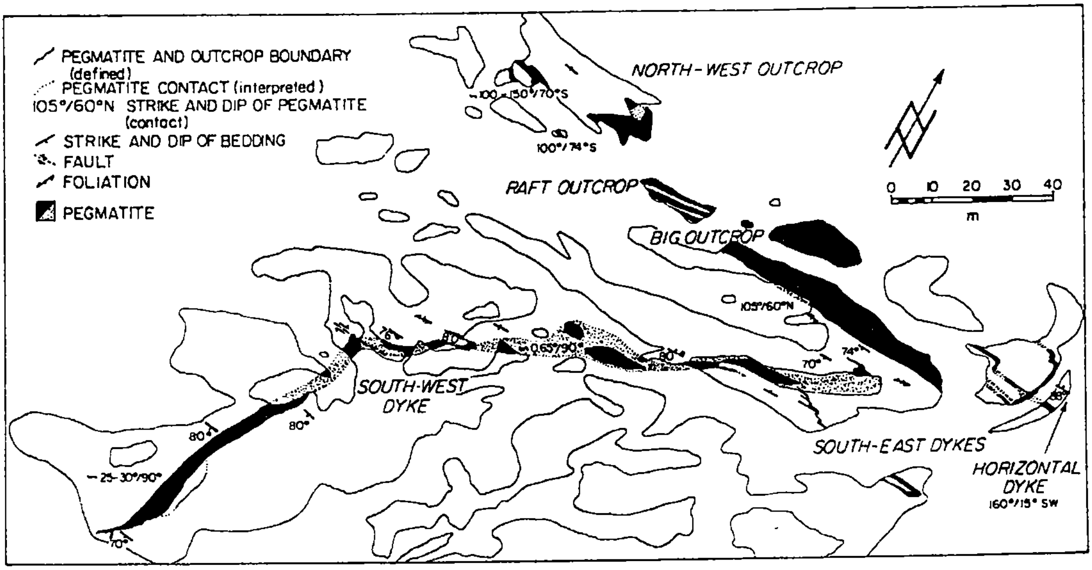

FIG. 2. Map of the YITT-B pegmatite swarm showing location of the six main bodies commented on in the text, and the principal structural features of the host rocks and pegmatites.

However, deformation structures within the pegmatites are sparsely developed, and include fractured and kinked crystals of feldspar and muscovite, and a local muscovite-defined foliation, developed only in the outer margins. On a larger scale, the South-west dyke has been boudinaged in the plane of the metaturbidite bedding in one location, and one of the dykes in the Raft outcrop is cross-cut by a dilational quartz-filled fracture, which is truncated at the pegmatite – host rock contact. These features emphasize the significant contrast in competence between the pegmatites and the metaturbidites during the deformation. Late faulting in

TABLE 1. COMPOSITE SCHEME OF INTERNAL ZONING OF THE YITT-B PEGMATITES, SOUTHEASTERN MANITOBA

hanging-wall exocontact	metasomatic	Ms + Qtz (± Grt,Tur)
wall zone	f.-gr.aplitic	Ab + Qtz + Ms (± Grt) {Ap, ± Tur}
intermediate zone 1	m.- to c.-gr. pegmatitic	Kfs + Qtz + Ms ± Ab {Grt, ±Nb,Ta,Sn minerals, sulfides}
intermediate zone 2a*	layered albitic aplite	Ab + Qtz > Ms > Grt (±Kfs) {Nb,Ta,Sn minerals, sulfides, phosphates}
core**	blocky	Kfs + Qtz (± book Ms, Ab) {beryl}
intermediate zone 2b*	saccharoidal albite	Ab > Qtz (± Ms, Grt) {Nb,Ta,Sn minerals, sulfides, phosphates}
intermediate zone 1	m.- to c.-gr. pegmatitic	Kfs + Qtz + Ms ± Ab {Grt, ±Nb,Ta,Sn minerals, sulfides}
wall zone	f.-gr. aplitic	Ab + Qtz + Ms (±Grt) {Ap}
footwall exocontact	metasomatic	Ms + Qtz (±Grt)

( ): subordinate to minor, { }: accessory.
* 2a and 2b are interchangeable, locally mutually transitional along strike; they may occur at either side of the adjacent intermediate zone 1, and may be missing in the footwall part of the sequence; ** blocky core may not be developed, and intermediate zone 1 may occupy the central parts; the zoning is commonly asymmetrical.

this area is apparently restricted to the above-mentioned dextral fault separating the two segments of the South-west Dyke.

Internal structure

The YITT-B pegmatites are heterogeneous bodies in terms of mineralogy, grain size and texture. Table 1 lists the rock-forming and accessory minerals of the different zones composing the pegmatites, from contacts with the host rock toward the central parts of the dikes. However, close approximation to such a sequence is developed only in the Raft and Horizontal outcrops, and in an asymmetrical manner. The other pegmatites consist of the same zonal assemblages, but the sequence of zones is much less regular than in the two dikes above. The juxtaposition of the zones is particularly haphazard in the dikes that are the most irregular and contorted in their overall shape. Nevertheless, the garnetiferous albite-rich aplite is in most cases close to the contacts, and the coarse-grained assemblages carrying blocky K-feldspar strongly tend to occupy the interior of the dikes. Despite the irregularities in disposition of the zones, cross-cutting of finer-grained assemblages by coarser K-feldspar-bearing or muscovite-rich aggregates is only rarely observed, and replacement phenomena along the contacts of the saccharoidal albite with coarse-grained assemblages are scarce.

The distribution and relative volumes of the zones also are diversified among the bodies of pegmatite. The most prominent in this respect is the saccharoidal albite zone, which is very dominant in the South-west dike, but distinctly minor in the Raft outcrop and in some dikes of the North-west outcrop. Also, the platy albite seems to be restricted to the main dike of the North-west outcrop; albite is very fine-grained in all other pegmatites.

The limited distribution of some of the accessory minerals must be emphasized. Beryl was found only in the K-feldspar-bearing core of the northeastern part of the South-west pegmatite, whereas tourmaline was identified only in the border zone and adjacent altered raft of metasedimentary rocks in the Big outcrop.

The relationship of the albite-rich aplite and sacharoidal albite zones deserves a special note. In their typical conformations, these two zones seem to be quite different in the field, the layered and commonly garnetiferous aplite in the outer parts of the dikes as opposed to the more internally located, non-layered, and largely garnet-free saccharoidal albite. However, transition of one into the other along strike can locally be observed, as for example in the North-west outcrop, and despite some significant quantitative differences, the assemblage of accessory minerals is qualitatively the same in both zones. Thus the field-based distinction of the two zones becomes rather blurred on closer inspection.

EXPERIMENTAL METHODS

Atomic absorption spectroscopy was used for partial analysis of K-feldspar, muscovite and beryl. The electron microprobe (EMP) was employed for chemical analysis of other minerals. Instrumental conditions of the Cameca Camebax SX–50 for the analysis of feldspars and garnet are given in Teertstra *et al.* (1997), for Nb,Ta-oxide minerals, in Černý *et al.* (1992b), for tourmaline, in Selway *et al.* (1998), and for phosphate minerals, in Smeds *et al.* (1998).

TABLE 2. REPRESENTATIVE PARTIAL COMPOSITIONS OF K-FELDSPAR, MUSCOVITE AND BERYL, YITT-B PEGMATITES, SOUTHEASTERN MANITOBA

	K-feldspar			Muscovite			Beryl
	1	2	3	4	5	6	7
K_2O, wt.%	13.84	13.73	14.36	9.60	9.56	9.50	0.24*
Na_2O	1.34	1.02	1.20	0.73	0.74	0.77	0.62
Li_2O	<0.01	<0.01	<0.01	0.026	0.032	0.027	0.14
Rb_2O	0.36	0.44	0.50	0.46	0.52	0.66	0.01
Cs_2O	0.006	0.023	0.021	0.017	0.025	0.047	0.22
K/Rb	35	29	26	19	17	13	14

*The relatively high K content suggests slight contamination by muscovite.

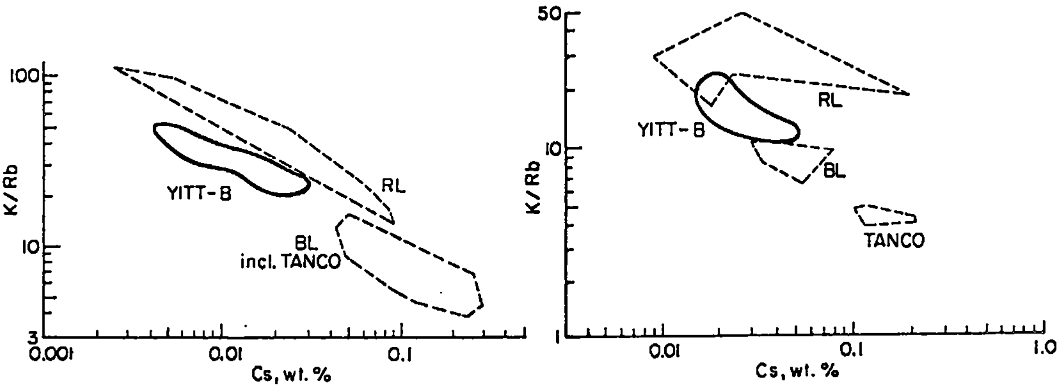

FIG. 3. Compositional fields of the YITT-B K-feldspar (left) and muscovite (right) in terms of K/Rb *versus* Cs, compared to those characterizing the Rush Lake and Bernic Lake pegmatite groups of the Cat Lake – Winnipeg River field (Černý *et al.* 1981); the latter group includes the Tanco deposit.

X-ray powder diffraction (XRD) was used for routine identification of individual minerals, with the Philips automated diffraction system PW 1710.

MINERALOGY AND GEOCHEMISTRY

Because of their very fine-grained and sparsely disseminated character, the Nb,Ta-oxide minerals and phosphates could be meaningfully investigated only in heavy-media and magnetic separates mounted in polished sections. This unavoidably restricted the choice of laboratory methods required to characterize multiple samples of sizeable populations, and it also severely limited textural and paragenetic information on these mineral groups.

K-feldspar

Microcline perthite occurs dominantly in the blocky core zones but is also found in lesser quantities in the intermediate and albite-rich aplite zones. In the intermediate zone, microcline forms small lath-shaped crystals ≤2 cm in size. In the albite-rich aplite, it is found as 2 to 10 cm crystals randomly oriented in the aplitic matrix. Blocks attaining 70 cm in length are typical of the core zones.

Enrichment in rare alkalis is moderate (Table 2), as is the overall range of the K/Rb ratio and Cs content (Fig. 3). The level of fractionation closely corresponds to that characterizing K-feldspar from moderately evolved pegmatites that do not contain minerals of lithium elsewhere in the Cat Lake – Winnipeg River field (Fig. 3; Černý *et al.* 1981). Reconnaissance XRD study of the structural state showed high values of triclinicity (0.80–0.96; Goldsmith & Laves 1954), indicating a highly ordered distribution of Si and Al.

Albite

Three principal types of albite can be distinguished: the most widespread fine-grained lath-shaped albite of the aplitic and saccharoidal zones, medium- to coarse-grained albite of the coarse-grained intermediate zones, and relatively rare platy albite prominent in the North-west outcrop but virtually absent in most of the other pegmatites.

Except the plagioclase of the wall zone immediately adjacent to the wallrock, which correspond to oligoclase An_{9-13}, EMP analysis shows albite from all other zones to be virtually pure (Ca from below detection limit to 0.07 wt.% CaO). Random samples of different paragenetic types of albite consistently yield XRD powder patterns of low albite.

Quartz

Quartz is a substantial component of all zones, mainly fine to medium grained, but rather coarse in the blocky cores. In this latter association, it tends to be bluish grey in color. Graphic intergrowths with feldspars or muscovite were not found in this pegmatite swarm.

Muscovite

Muscovite is the only mica encountered in the YITT-B pegmatites, but it is found in several varieties: fine-grained slightly brownish in the exocontacts and greenish yellow in the albite-rich zones, medium-grained in the coarse-grained intermediate zone, and coarse to book-like in the blocky cores. All these varieties consist of the $2M_1$ polytype.

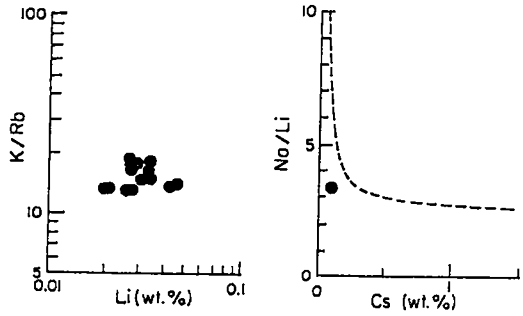

FIG. 4. The K/Rb *versus* Li plot of muscovite (left), emphasizing its low Li content, and Na/Li *versus* Cs of beryl (right), showing a rather advanced alkali fractionation; the dashed line indicates the average course of alkali fractionation in beryl from granitic pegmatites (from Černý 1975).

TABLE 3. REPRESENTATIVE COMPOSITIONS OF TOURMALINE, YITT-B PEGMATITES, SOUTHEASTERN MANITOBA

	5C	5I	5R		5C	5I	5R
SiO_2 wt.%	35.90	35.80	35.70	*T* Si *apfu*	5.915	5.937	5.928
TiO_2	0.83	0.89	0.99	Al	0.085	0.063	0.072
Al_2O_3	33.10	32.20	32.30	Σ	6.000	6.000	6.000
FeO*	7.39	9.69	9.83	*Z* Al	6.000	6.000	6.000
MnO	0.06	0.17	0.20	*Y* Al	0.342	0.230	0.250
CaO	0.59	0.34	0.34	Mg	1.400	1.187	1.114
MgO	5.70	4.80	4.50	Fe^{2+}	1.018	1.344	1.365
ZnO	0.05	0.04	0.14	Mn	0.008	0.024	0.028
Na_2O	1.78	2.13	2.02	Zn	0.006	0.005	0.017
K_2O	0.03	0.02	0.03	Ti	0.103	0.111	0.124
B_2O_3**	10.55	10.48	10.47	Σ	2.877	2.901	2.898
H_2O**	3.41	3.29	3.17	*X* Na	0.569	0.685	0.650
F_2	0.49	0.68	0.92	Ca	0.104	0.060	0.060
$-O{=}F_2$	−0.21	−0.29	−0.39	*R*	0.006	0.004	0.006
				Σ	0.679	0.749	0.716
Σ	99.67	100.25	100.22	F	0.255	0.357	0.483
				OH	3.745	3.643	3.517
				Σ	4.000	4.000	4.000

* ΣFe as FeO; ** calculated by stoichiometry; C: core, I: intermediate zone, R: rim. Atomic contents normalized to 31 anions, B = 3, OH + F = 4.

Partial analyses of the coarse-grained muscovite gave alkali contents and ratios shown in Figure 3 and Table 2. As in the case of the K-feldspar, the degree of fractionation attained by muscovite is moderate, and the variations are restricted. The Li content is very low (Fig. 4), particularly if compared to that of muscovite from the primitive pegmatites of the Birse Lake group (0.03–0.80 wt.% Li_2O), or the Greer Lake pegmatites lacking lithium minerals (0.16–0.70 wt.% Li_2O; Černý *et al.* 1981). Reconnaissance EMP analysis of the different types of YITT-B muscovite shows about the same degree of Rb and Cs enrichment in all varieties, and slightly elevated Fe and Mg contents in the phengitic exocontact and wall-zone muscovite; it also reveals very low content of F (0.06–0.13 wt.%).

Beryl

Rare finds of beryl are restricted to the northeastern termination of the South-west dike. White euhedral crystals a few cm in size are located in the blocky core of the pegmatite. Alkali contents of one analyzed crystal show an appreciable degree of fractionation (Table 2); we classify this beryl as transitional between the typical sodic–lithian and lithian–cesian types (Fig. 4; Černý 1975).

Tourmaline

This mineral is as scarce as beryl, found only in the endocontact and exocontact of the North-west pegmatite with a raft of metasediments. Black-brown crystals less than 2 mm in length are concentrated mainly in the fine-grained albite-rich aplite of the wall zone. The crystals are compositionally zoned, but the variations are very modest; Ti and Fe are slightly enriched in the outer zones, but exceptions also are encountered (Table 3). The chemistry generally corresponds to intermediate members of the schorl – dravite series, with low Ca content. Calculation of potential Li content by stoichiometry (Burns *et al.* 1994) gives from 0.19 to 0.27 wt.% LiO_2 (0.13–0.18 atoms per formula unit, *apfu*, Li). However, the presence of lithium in this tourmaline is doubtful because the overall low content of Li in the whole pegmatite. Consequently a Li-free recalculation of the atomic contents is preferred in Table 3, although the rare tourmaline may serve as a local sink for Li in a similar manner as beryl does in its own limited occurrence.

Garnet

Garnet is found dominantly in the albite-rich aplite and, to a lesser degree, in the coarse-grained intermediate zone. Subhedral to euhedral crystals average <1 mm in the former, but attain up to 10 mm in the latter zone. Garnet shows an exceedingly narrow compositional range, from $Sps_{51}Alm_{49}$ to $Sps_{60}Alm_{40}$ in both associations. Much of this spread is covered within the zoning of individual crystals, which are in most cases enriched in Mn, Y and P in the cores, and in Fe and Ca in the outermost parts (Table 4). Figure 5 shows the conspicuously low contents of Ca and Mg, and the intermediate level of fractionation in terms of Fe–Mn, in the YITT-B garnet relative to the compositional fields of its counterparts from all other pegmatite groups of the Cat Lake – Winnipeg River field (Černý *et al.* 1981).

The compositions of garnet consistently show deficient Si, excess of Al, and excess of M^{2+} cations in variable proportions. In the absence of significant Na

TABLE 4. REPRESENTATIVE COMPOSITIONS OF GARNET, YITT-B PEGMATITES, SOUTHEASTERN MANITOBA

	#3911		#421		#3891	
	core	rim	core	rim	core	rim
SiO_2 wt.%	34.80	35.60	35.30	35.40	34.90	35.90
TiO_2	0.03	0.00	0.00	0.00	0.02	0.00
P_2O_5	0.42	0.06	0.26	0.20	0.38	0.14
Al_2O_3	20.60	20.50	20.90	20.80	20.70	20.80
Y_2O_3	0.03	0.00	0.03	0.00	0.04	0.01
FeO	20.60	20.90	19.20	20.40	18.30	19.80
MnO	22.70	22.50	23.70	22.60	24.70	23.80
CaO	0.10	0.14	0.09	0.13	0.09	0.07
MgO	0.07	0.06	0.07	0.05	0.02	0.00
$\sum$	99.35	99.76	99.55	99.58	99.15	100.52
Atomic contents normalized to 12 oxygen atoms per formula unit						
Si	2.907	2.960	2.932	2.941	2.916	2.957
Ti	0.002	0.000	0.000	0.000	0.001	0.000
P	0.030	0.004	0.019	0.019	0.027	0.010
Al	2.028	2.009	2.046	2.037	2.039	2.019
Y	0.001	0.000	0.001	0.000	0.002	0.001
Fe^{2+}	1.439	1.453	1.334	1.418	1.279	1.364
Mn	1.606	1.584	1.668	1.591	1.748	1.661
Ca	0.009	0.012	0.008	0.012	0.008	0.006
Mg	0.009	0.007	0.009	0.006	0.002	0.000
$\sum$cat.	8.032	8.030	8.016	8.019	8.022	8.018

and Y, these relationships suggest locally fluctuating substitutions of two types: $(AlP)Si_{-2}$ and $MnP(AlSi)_{-1}$ (Taylor & Wise 1995, Burt 1996).

Oxide minerals of niobium and tantalum

Three phases of this category are present in the pegmatites investigated: columbite – tantalite, ixiolite and ferrotapiolite. All three are found in about equal proportions in the YITT-B pegmatite system, with only minor variations among the individual dikes. The only exception is the northwestern part of the Big outcrop, where tapiolite is dominant over columbite – tantalite, and ixiolite is absent. Minor differences also are found among parts of a single dike; for example, central parts of the South-West dike contain ferrotapiolite > ixiolite > columbite – tantalite, whereas the southwestern termination has columbite – tantalite > ixiolite ferrotapiolite.

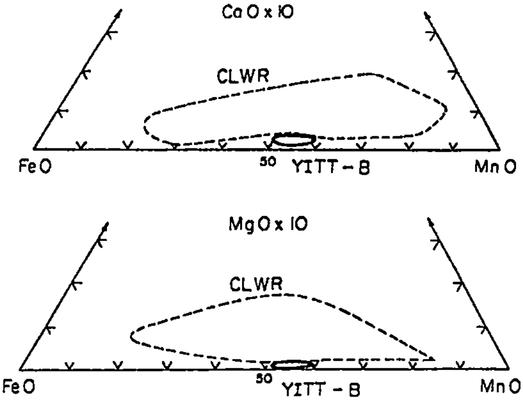

FIG. 5. The composition of garnet in the YITT-B suite, in terms of FeO – MnO – CaO × 10 and FeO – MnO – CaO × 10 (wt.% proportions). The garnet (heavy line) is compared to the compositional fields of garnet from other pegmatite groups of the Cat Lake – Winnipeg River pegmatite field (dashed outline; from Černý *et al.* 1981).

TABLE 5. REPRESENTATIVE COMPOSITIONS OF THE Nb,Ta,Sn-OXIDE MINERALS, YITT-B PEGMATITES, SOUTHEASTERN MANITOBA

	Columbite-tantalite			Ixiolite		Ferrotapiolite		Cassiterite
	391-11	393-1	385-14	391-6	385-6	389-13	389-11	388-F6
WO_3	0.19	0.08	0.40	0.16	1.05	0.00	0.00	0.00
Ta_2O_5	38.47	62.30	68.93	60.16	66.16	78.01	78.53	2.43
Nb_2O_5	41.89	20.64	14.75	9.95	5.34	5.20	5.49	0.17
ZrO_2	0.00	0.00	0.00	0.51	0.42	0.07	0.00	0.05
SnO_2	0.06	0.15	0.26	13.16	11.90	0.77	0.51	97.02
TiO_2	0.23	0.26	0.51	1.89	2.06	0.61	0.40	0.09
Bi_2O_3	0.00	0.08	0.00	0.05	0.00	0.00	0.00	0.02
Sb_2O_3	0.01	0.00	0.02	0.00	0.08	0.00	0.02	0.07
Fe_2O_3	1.15	0.05	0.21	0.74	0.78	0.00	0.05	0.00
FeO	9.64	8.85	5.65	7.69	4.51	13.29	12.66	0.34
MnO	7.34	6.72	9.36	4.41	7.37	0.70	1.51	0.01
CaO	0.01	0.03	0.02	0.01	0.11	0.00	0.01	0.00
MgO	0.01	0.00	0.03	0.00	0.02	0.01	0.01	0.01
Σ	99.01	99.15	100.14	98.73	99.80	98.66	99.18	100.21
Atomic contents normalized to 24 oxygen atoms and 12 cations*								
W^{6+}	0.013	0.006	0.032	0.013	0.086	0.000	0.000	0.000
Ta^{5+}	2.801	5.116	5.782	5.090	5.688	7.046	7.063	0.199
Nb^{5+}	5.069	2.818	2.058	1.401	0.763	0.783	0.812	0.023
Zr^{4+}	0.000	0.000	0.000	0.077	0.065	0.011	0.000	0.007
Sn^{4+}	0.006	0.018	0.032	1.632	1.500	0.102	0.067	11.640
Ti^{4+}	0.046	0.059	0.118	0.442	0.491	0.152	0.099	0.020
Bi^{3+}	0.000	0.000	0.000	0.004	0.000	0.000	0.000	0.002
Sb^{3+}	0.001	0.000	0.003	0.000	0.010	0.000	0.003	0.009
Fe^{3+}	0.232	0.011	0.048	0.174	0.184	0.000	0.011	0.000
Fe^{2+}	2.158	2.236	1.459	2.000	1.193	3.692	3.502	0.086
Mn^{2+}	1.664	1.720	2.447	1.162	1.973	0.199	0.423	0.003
Ca^{2+}	0.003	0.010	0.007	0.003	0.037	0.000	0.004	0.000
Mg^{2+}	0.004	0.000	0.014	0.000	0.009	0.005	0.005	0.004
Σ	12.000	12.000	12.000	12.000	12.000	11.991	12.000	11.992

All compositions recalculated to the unit-cell content of ordered columbite-tantalite, to facilitate comparison; Fe_2O_3 calculated to eliminate excess cations over 12.0. The compositions are quoted in weight % oxides.

Representative compositions of all three minerals are given in Table 5. Plots of the bulk compositions in the (Nb + Ta) – (Mn + Fe) – (Sn + Ti) diagram show that most of the columbite – tantalite and ferrotapiolite are fairly pure, with negligible contents of the tetravalent cations (Fig. 6). In contrast, ixiolite is considerably enriched in Sn, but falls slighly short of the ideal ratio $(Mn,Fe)_{25}(Sn,Ti)_{25}(Ta,Nb)_{50}$ of its ordered analog wodginite. The three phases are distinctly separated in the columbite quadrilateral (Fig. 7), with the ixiolite compositions typically in the central parts of the tantalite – tapiolite gap (*cf.* Černý & Ercit 1989). Ferrotapiolite populates the central part of its compositional field. Columbite – tantalite shows a broad spread of Ta/(Ta + Nb), but a narrow range

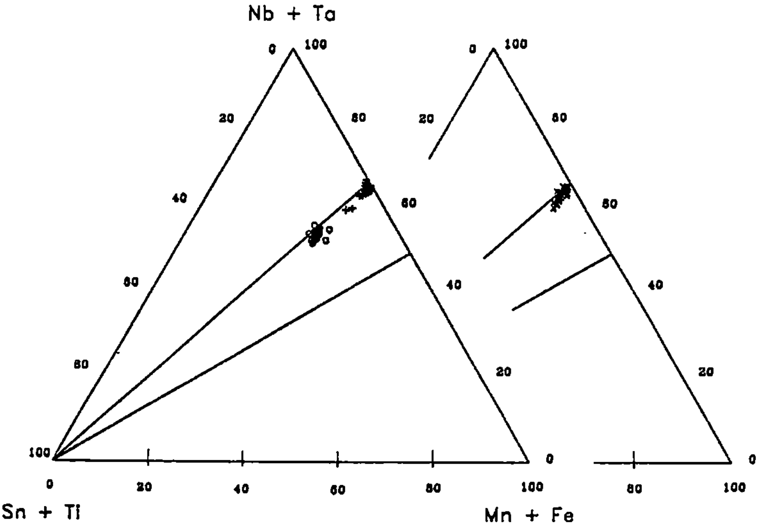

FIG. 6. Compositional fields of columbite – tantalite (crosses), ixiolite (circles) and ferrotapiolite (×) in terms of (Nb,Ta) – (Mn,Fe) – (Sn,Ti) (atomic proportions); the compositions of ixiolite approach but do not attain the ideal proportions in wodginite, $(Mn,Fe)_{25}(Sn,Ti)_{25}(Nb,Ta)_{50}$.

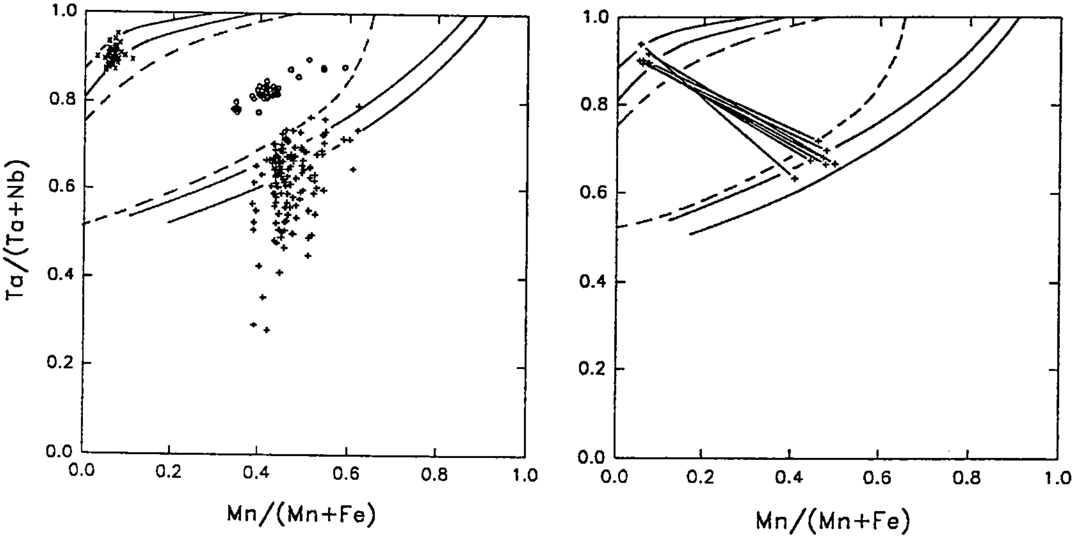

FIG. 7. Compositions of the Nb,Ta-oxide minerals in the columbite quadrilateral (atomic ratios). Left: all data compared to the empirical two-phase field of Černý *et al.* (1992a), as outlined by pairs of coexisting tantalite and ferrotapiolite (belts outlined by solid lines) and by compositions of single phases (dashed); crosses: columbite – tantalite, circles: ixiolite, ×: ferrotapiolite. Right: compositions of coexisting tantalite and ferrotapiolite compared to the same two-phase field boundaries.

of Mn/(Mn + Fe) values centered at ~0.45. Quite a few compositions of tantalite, and some of the ferrotapiolite, populate the space between the boundaries limiting the compositions of coexisting phases and those of single minerals (*cf.* Černý *et al.* 1992a). However, a single composition of tantalite marginally violates this space if pairs of demonstrably coexisting tantalite + ferrotapiolite are considered alone (Fig. 7).

Ferrotapiolite and ixiolite are generally very homogeneous, in contrast to the columbite – tantalite, which shows bimodal distribution of grain size and strong oscillatory zoning. Most of the columbite – tantalite fall into the coarser category, >0.5 mm (but rarely over 1.5 mm), which is composed of brick-shaped euhedral crystals. A lesser but still substantial portion of columbite – tantalite forms distinctly finer-grained, thin bladed crystals ~0.10–0.15 mm in size. In both types of crystals, the oscillatory zoning reflects variations in Ta/(Ta + Nb), whereas the Mn/(Mn + Fe) value of individual crystals is virtually constant. Most of the Ta/(Ta + Nb) range shown in Figure 7 is generated by the oscillatory zoning; the bulk compositions of individual crystals would show much less variation.

Figures 8 and 9 show an example of oscillatory zoning typical of the coarser-grained crystals, closely repeated in all crystal fragments examined. The overall composition and range of oscillation are virtually the same in the finer-grained crystals, although the pattern of oscillation is simpler. No systematic correlation is observed between the oscillatory pattern of the fine-bladed crystals and any segment of that characterizing the coarser-grained columbite – tantalite.

In terms of minor cations typical of the broader columbite family, the three minerals show slight to substantial differences (Fig. 10). The content of Fe^{3+} is the lowest in ferrotapiolite, and only marginally higher in ixiolite relative to columbite – tantalite. Ixiolite displays the highest Sn content, tapiolite being slightly more stanniferous of the other two phases. Titanium is slightly but distinctly enriched in ixiolite as compared to ferrotapiolite, and considerably lower in columbite – tantalite relative to the other minerals. The content of Zr is mainly below the detection level in columbite – tantalite and ferrotapiolite, but it shows a prominent enrichment in ixiolite (in agreement with the results of Černý *et al.*, in prep., on Zr-rich ixiolite and wodginite).

A powder XRD survey performed on mineral mixtures, with only occasional and moderate enrichment in individual species recognizable under the binocular microscope (*e.g.*, the finer-grained bladed columbite – tantalite), generated unit-cell dimensions of a rather poor quality. However, the structural state of both types of columbite – tantalite was identified as moderately disordered [*cf.* Ercit *et al.* (1995) for reference data], whereas ferrotapiolite was found to be highly ordered [*cf.* Wise & Černý (1996) for reference grid]; ixiolite is strongly to totally disordered by definition (*e.g.*, Wise & Černý 1986).

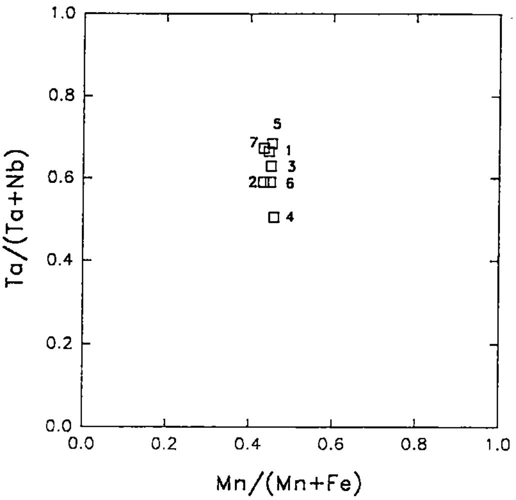

FIG. 8. Typical sequence of compositions in oscillatorily zoned columbite – tantalite in the columbite quadrilateral (atomic ratios), from 1 in the center to 7 in the outer rim of the crystal shown in Figure 9.

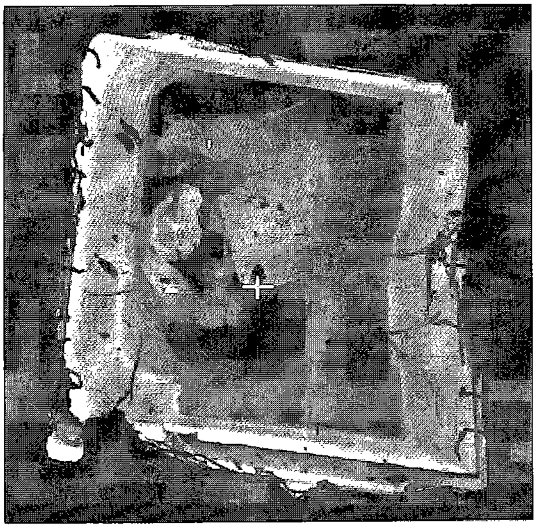

FIG. 9. Typical compositional zoning of a crystal of columbite – tantalite, about 250 μm across. The variation of brightness in the back-scattered electron image reflects solely the changes in Nb/Ta, quantified in Figure 8.

Overall, the compositional and structural features of the YITT-B oxide minerals are distinctly different from those of any others in the Cat Lake – Winnipeg River pegmatite field that were examined in sufficient detail [Tanco: Ercit (1986), Greer Lake group: Černý *et al.* (1986), Lower Tanco: Ferreira (1984), Huron Claim: Paul (1984), Shatford Lake group: H.M. Buck, in progress].

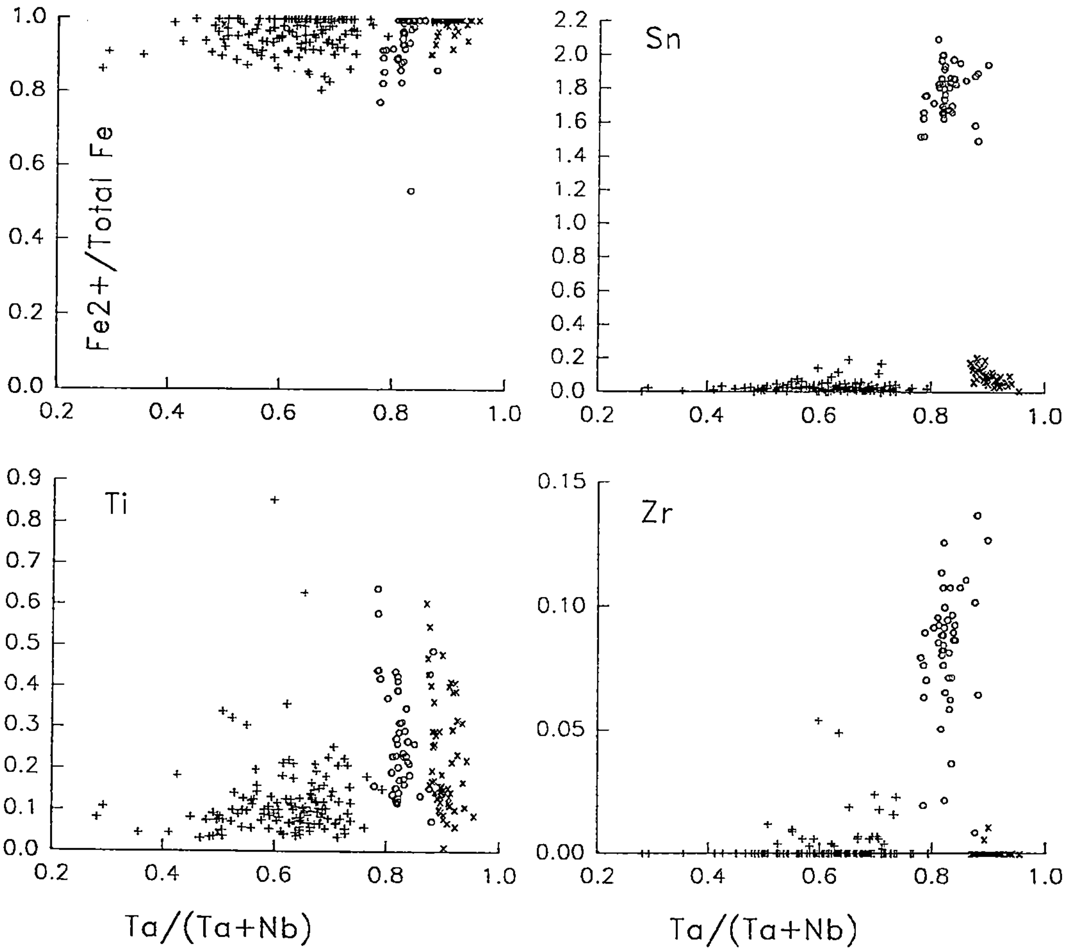

FIG. 10. Selected compositional features of columbite – tantalite (crosses), ixiolite (circles) and ferrotapiolite (x), shown as the atomic ratio Fe^{2+}/total Fe and Sn, Ti, Zr (*apfu*) *versus* the atomic ratio Ta/(Ta + Nb). Note the low Fe^{3+} in ferrotapiolite, Sn higher in ferrotapiolite than in columbite – tantalite, Ti concentrated preferentially in ixiolite and ferrotapiolite, and Zr strongly enriched in ixiolite.

Cassiterite

Fine-grained subhedral crystals of cassiterite are very subordinate relative to the above Nb,Ta-oxide minerals. They were found in a substantial concentration only in a single concentrate from the North-west outcrop, which is also rich in ixiolite. Cassiterite seems to be absent in several other pegmatites, notably in the northwestern part of the Big outcrop, which also lacks ixiolite. Cassiterite is only moderately enriched in the ferrotapiolite component (Table 5).

Beusite

Homogeneous grains of this mineral are fairly abundant in the phosphate-enriched samples of the fine-grained albite-rich assemblages from the south-western termination of the South-west dyke and from the South-east dykes. This beusite belongs to the Ca-rich varieties of the graftonite – beusite series, with Ca variable from 12.1 to to 14.0 wt.% CaO (0.75 – 0.85 *apfu* Ca); the atomic ratio Mn/(Mn + Fe) is fairly constant at ~0.68 (Table 6).

Triplite

Triplite seems to be very rare, as it was identified only in three grains. Its chemistry shows considerable (OH), subordinate Ca and a high Mn/(Mn + Fe) value, 0.88 (Table 6).

Alluaudite

Alluaudite also is abundant in the beusite-bearing concentrates mentioned above. It forms isolated grains up to 1 mm in size dispersed in fine-grained albite, and

TABLE 6. REPRESENTATIVE COMPOSITIONS OF BEUSITE, TRIPLITE AND ALLUAUDITE, YITT-B PEGMATITES, SOUTHEASTERN MANITOBA

	Beusite		Triplite			Alluaudite	
	2-18	3-11	3-12			2-20	2-22
P_2O_5	40.87	41.50	31.56			44.08	43.68
SiO_2	0.00	0.00	0.00			0.01	0.03
Fe_2O_3	-	-	-			26.45	23.85
FeO	14.34	15.20	7.32			0.00	0.08
MnO	30.94	28.92	54.54			20.38	21.78
MgO	0.00	0.00	0.17			0.06	0.05
CaO	12.09	14.00	1.25			0.08	0.20
SrO	0.00	0.01	0.00			0.04	0.00
ZnO	0.36	0.45	0.04			0.02	0.00
Na_2O	0.00	0.00	0.00			8.54	9.95
K_2O	0.00	0.00	0.01			0.03	0.01
H_2O*	-	-	1.48			-	-
F_2	0.00	0.00	5.37			0.00	0.00
Cl_2	0.00	0.00	0.02			0.00	0.00
$-O=F_2,Cl_2$	-	-	-2.26			-	-
Σ	98.60	100.08	99.50			99.69	99.61
			Atomic contents *pfu***				
P	2.006	2.001	0.996	*T*	P	3.006	3.000
Fe^{2+}	0.695	0.724	0.228		Si	0.001	0.002
Mn	1.520	1.397	1.722		Σ	3.007	3.002
Mg	0.000	0.000	0.009				
Ca	0.751	0.855	0.050	*A*	Na	1.000	1.000
Zn	0.015	0.019	0.001				
Σ	2.981	2.995	2.010	*B*	Na	0.334	0.567
					K	0.003	0.001
					Ca	0.007	0.017
					Sr	0.002	0.000
					□*	0.654	0.415
F	-	-	0.632		Σ	1.000	1.000
Cl	-	-	0.001				
OH	-	-	0.367	*C*	Mn	1.000	1.000
				D	Fe^{3+}	1.604	1.458
					Mn	0.391	0.498
					Fe^{2+}	0.000	0.005
					Mg	0.007	0.006
					Zn	0.001	0.000
					Σ	2.003	1.967

* calculated by stoichiometry; ** beusite normalized to 8 oxygen atoms, triplite to 5 anions, alluandite to 12 oxygen atoms. The compositions are quoted in weight % oxides and volatiles.

inclusions in late apatite interstitial to albite (discussed below). Many compositions of alluaudite established by EMP analysis indicate alteration and leaching (low wt.% oxide totals, and P in excess of 3 *apfu*), but others lead to a rather simple formula corresponding to an extremely Ca-poor, Na□-type of alluaudite with Mn^{2+} as the only *M*(1) cation and Fe^{3+} dominant at the *M*(2) site, which is extremely poor in Fe^{2+} (Table 6; *cf.* Moore & Ito 1979). If all Fe is considered, then the atomic ratio Mn/(Mn + ΣFe), 0.47–0.56, would indicate a range of Mn- to Fe-dominant compositions. Textural relationships seem to suggest that alluaudite is a primary mineral, predating the second generation of primary apatite (see below). However, the Fe^{3+}-rich composition contradicts such a paragenetic role; the apatite-hosted alluaudite probably represents pseudomorphs after an altered primary phosphate, possibly beusite.

Triphylite

A few grains of triphylite were identified by Selway *et al.* (1998) in heavy fractions of crushed samples. In contrast to the other ferromanganoan phosphates analyzed here (Table 6), triphylite is Fe-dominant, with an average composition of $Li_{1.00(calc.)}(Fe_{0.55}Mn_{0.42}Ca_{0.01})P_{1.01}O_4$ (Selway *et al.* 1998, Table 2).

Apatite

Three generations of apatite are distinguished by their paragenetic association, morphology and chemistry (Table 7). (i) Early primary apatite forms euhedral crystals, occasionally visible in hand specimens; it consists of manganoan fluorapatite (0.94 F *apfu*) with up to 3.6 wt.% MnO. (ii) Late primary apatite typically forms anhedral fillings of interstices among crystals of saccharoidal albite (Fig. 11); it contains up to 7.3 wt.% MnO but is largely F-poor (0.50–0.63 F *apfu*), and can be characterized as a manganoan hydroxylapatite – fluorapatite. (iii) Microscopic secondary apatite, closely associated with alluaudite and beusite, which it in part replaces, also is Mn-bearing (up to 5.8 wt.%

TABLE 7. REPRESENTATIVE COMPOSITIONS OF APATITE, (PARA)SCHOLZITE AND PHOSPHOPHYLLITE, YITT-B PEGMATITES, SOUTHEASTERN MANITOBA

	Apatite						Phospho-phyllite	(Para) scholzite
	euhedral		interstitial		secondary			
	3-3p	3-9p	18AP	18AD	1-1s	2-153	1-8p	1-9s
P_2O_5	41.84	41.76	42.40	41.30	38.31	38.00	32.80	35.19
SiO_2	0.00	0.00	-	-	0.00	0.33	0.06	0.16
Al_2O_3	0.00	0.06	-	-	0.02	0.25	0.00	0.00
FeO	0.57	0.14	0.14	0.24	0.79	2.33	8.92	0.86
MnO	3.58	2.80	2.08	7.33	1.25	5.82	5.07	2.69
MgO	0.00	0.00	0.00	0.00	0.00	0.04	0.03	0.00
CaO	51.97	53.06	54.50	48.70	51.41	44.82	0.53	11.96
SrO	0.06	0.26	0.06	0.46	0.28	0.30	0.00	0.07
ZnO	0.13	0.00	-	-	0.00	0.01	37.08	40.10
Na_2O	0.00	0.03	0.00	0.00	0.81	0.41	0.00	0.00
H_2O*	0.07	0.04	0.65	0.79	1.40	0.79	16.40	9.03
F_2	3.55	3.58	2.37	1.83	0.13	1.28	0.00	0.00
Cl_2	0.09	0.16	0.01	0.16	0.80	0.98	0.00	0.00
$-O=F_2,Cl_2$	-1.51	-1.55	-1.00	-0.81	-0.23	-0.76	-	-
Σ	100.38	100.44	101.27	100.00	95.05	94.73	100.88	100.10
				Atomic contents *pfu***				
P	2.994	2.985	2.976	2.986	2.919	2.933	2.032	1.981
Si	0.000	0.000	-	-	0.000	0.030	0.004	0.011
Al	0.000	0.006	-	-	0.002	0.027	0.000	0.000
Σ	2.994	2.991	2.976	2.986	2.921	2.990	2.036	1.992
Fe^{2+}	0.040	0.010	0.009	0.016	0.059	0.178	0.547	0.048
Mn	0.256	0.200	0.146	0.531	0.095	0.449	0.315	0.151
Mg	0.000	0.000	0.000	0.000	0.000	0.005	0.003	0.000
Ca	4.706	4.799	4.891	4.458	4.957	4.376	0.042	0.852
Sr	0.003	0.013	0.003	0.023	0.015	0.016	0.000	0.003
Zn	0.008	0.000	-	-	0.000	0.001	2.004	1.968
Na	0.000	0.005	0.000	0.000	0.141	0.072	0.000	0.000
Σ	5.013	5.027	5.049	5.028	5.267	5.097	2.911	3.022
F	0.949	0.956	0.622	0.492	0.037	0.369	-	-
Cl	0.013	0.023	0.001	0.023	0.122	0.151	-	-
OH	0.038	0.021	0.377	0.485	0.841	0.480	-	-
Σ	1.000	1.000	1.000	1.000	1.000	1.000		
H_2O	-	-	-	-	-	-	4.000	2.000

*calculated by stoichiometry; ** apatite normalized to 13 anions, phosphophyllite to 12 oxygen atoms, (para)scholzite to 10 oxygen atoms; low P_2O_5, Σ oxides, Σ tetrahedral cations and excess of M^{2+} in secondary apatite suggests the presence of carbonate anions.

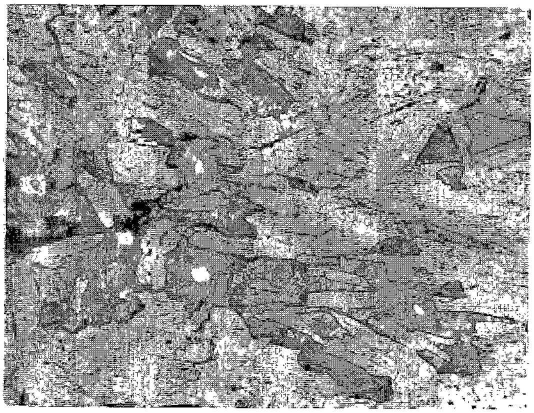

FIG. 11. Late, primary, anhedral apatite (dark grey) interstitial to lath-shaped saccharoidal albite (off-white); the shorter edge of the photo equals 2 mm.

MnO) but enriched in Fe (0.8–2.7 wt.% FeO), and is F-poor and Cl-bearing (averaging at 0.35 F and 0.15 Cl *apfu*). Low contents of P and low totals of wt.% oxides in this secondary apatite (Table 7) indicate a substantial presence of CO_2, rather common in this generation of apatite in pegmatites of the adjacent Bernic Lake group (Černý & Lenton 1995, Černý *et al.* 1996). It is best classified as manganoan carbonate-hydroxylapatite.

Phosphophyllite and (para)scholzite

The results of EMP analysis unambiguously identify phosphophyllite, yielding an almost ideal formula with a Mn/(Mn + Fe) value of 0.37 (Table 7). The chemistry of (para)scholzite also is well defined, if we assume that Fe^{2+} enters the Zn site, rather than that of Ca, which is evidently shared by Mn. However, the material available for study did not permit an X-ray-diffraction study, which alone would distinguish between the two polymorphs. Both phosphophyllite and (para)scholzite are found only as microscopic veinlets and granular aggregates replacing alluaudite.

Unknown phosphate

Along with alluaudite, the late interstitial apatite contains microscopic inclusions of a dark-colored phase that defies unambiguous identification. The EMP-determined chemical composition, which is the only characteristic available, suggests a Mn- and particularly Fe-rich apatite, but the stoichiometry is not satisfactory. The composition is close to that of secondary phases encountered in current studies of other populations of phosphates in pegmatites (Smeds *et al.* 1998), consistently associated with alteration products of primary phosphates. In our case, this mineral probably also replaces a precursor that did form true inclusions in the host apatite, in a manner similar to that exhibited by the alluaudite described above.

Sulfides

Sulfide minerals are dispersed in the albite-rich aplite and saccharoidal albite aggregates, as well as in the blocky core zones. Arsenopyrite is the most abundant phase, commonly with inclusions of löllingite, bismuthinite and several unidentified minerals consisting of diverse combinations of As, Bi, Fe and S; the inclusions are generally <30 μm in size.

DISCUSSION

Paragenetic relationships

The pegmatites of the YITT-B swarm are comparable to the beryl type of rare-element pegmatites of the LCT family, and more specifically to the beryl – columbite – phosphate subtype. They are peraluminous and carry the typical assemblage of minor phases in qualitative terms. However, the array of subordinate to accessory minerals differs quantitatively and texturally from the vast majority of pegmatites that define this category, as classified by Černý (1991a).

Beryl, one of the principal rare-element accessory minerals of this subtype, is extremely rare. Observations in the field are restricted to the subhorizontal erosional surface, and beryl might occur in a relatively greater abundance at undisclosed or eroded levels of the pegmatite dikes. However, this seems improbable, as the exposed dikes display highly variable attitudes, are extensively variable in internal structure, and show a restricted but distinct range in the degree of fractionation. Thus it is unlikely that the present exposure would completely miss vertical levels potentially enriched in beryl, which is also very rare in other related pegmatites. Also, beryl shows a prominent tendency to occur together with Nb,Ta-oxide minerals (*e.g.*, Beus 1960, Černý 1989a), which are abundant and widespread in the albite-rich zones of the YITT-B swarm. Thus the lack of beryl can be considered real, rather than an effect of restricted exposure.

Tourmaline is yet another phase that is extremely rare. Although not participating in the definition of the above-mentioned type and subtype of pegmatite, it is very common in beryl-type pegmatites; its virtual absence is yet another unusual feature of the YITT-B swarm.

The absence of lithium minerals is characteristic of the beryl category of pegmatites, and it extends in this case to the assemblage of phosphates. Phosphate minerals commonly include the first Li-bearing phases (such as graftonite – triphylite exsolution lamellae and triphylite *per se*) in regionally zoned populations of pegmatites, before the appearance of lithium silicates (Smeds *et al.* 1998). In cases of Li-poor systems, these Li-bearing phosphates may express the maximum accumulation of Li in the most exterior pegmatites of a given pegmatite group (Abella *et al.* 1995, Malló *et al.* 1995).

The essential identity of the albite-rich aplite and saccharoidal albite zones in the YITT-B pegmatites must be emphasized. In most pegmatitic granites and pegmatites, these two units are spatially distinct, and differ in internal textures and mineral assemblages (*e.g.*, Beus 1950, 1960, Černý *et al.* 1981). Albite-rich aplites are usually confined to the margins of pegmatite bodies, are layered, predate the crystallization of coarser-grained pegmatite zones, and display simple rock-forming and accessory mineralogy; they are interpreted as primary magmatic in origin. In contrast, typical units of saccharoidal albite occupy internal parts of pegmatite dikes, lack any regular textural patterns, display metasomatized margins against neighboring pegmatite zones, and carry minerals of rare elements such as Nb, Ta, Ti, Sn, Be, Zr, Hf and B. Traditionally considered products of replacement, the saccharoidal albite units were experimentally documented as largely primary (London *et al.* 1989). In the YITT-B pegmatites, layered texture and relative abundance of garnet are the only features locally distinguishing the albite-rich aplite from saccharoidal albite, which is poor in garnet and lacks any distinctly organized internal texture. Otherwise, the mineral assemblage and composition of individual minerals are the same, and mutual textural transitions are observed in the outcrops. Consequently, garnetiferous layered assemblages cannot be taken for early magmatic members of pegmatite bodies at their face value, but must be examined in detail and compared with all aspects of other albite-rich units that may be present. Typical as the above-mentioned characteristics of the layered aplites are in most of their localities, occurrences of layered garnetiferous albite-rich assemblages inside pegmatite bodies are well known from numerous pegmatite populations (*e.g.*, Cameron 1955, Hutchinson & Claus 1956). Attention must be paid to all types of their occurrences, as "the textural development of layered aplites remains one of the major enigmas of pegmatitic rocks" (London 1992, p. 526); the distinction from, or identity with, sacharoidal albite assemblages is part and parcel of the problem.

A prominent, and to the best of our knowledge unique feature of the YITT-B albitic assemblages, is the dispersed, fine-grained character of the primary phosphates of Ca, Fe and Mn. These minerals commonly form sizeable to giant blocky crystals in intermediate to core-margin zones (*cf.* Moore 1982), and are relatively coarse-grained even if associated with saccharoidal albite (Abella *et al.* 1995, Malló *et al.* 1995). In our case, the average grain-size of the phosphates is distinctly less that 0.5 mm, rarely attaining 1 mm. It does not differ from the grain size of the interstitial apatite and the Nb,Ta-oxide minerals, which are routinely fine-grained in saccharoidal albite zones (*e.g.*, Ercit 1986, Abella *et al.* 1995, Novák *et al.*, in prep.).

Last but not least, a general lack of subsolidus alteration phenomena in the YITT-B pegmatites must be pointed out. The only minerals that yielded low-temperature reaction products are the Ca,Fe, Mn-phosphates, which are the most reactive of all primary components of granitic pegmatites (Moore 1982). Beryl, tourmaline, garnet and most notably the Nb,Ta-oxide minerals are intact. The last case is particularly surprising; secondary minerals such as microlite or fersmite are routinely encountered in even the most primitive members of the beryl type of pegmatites (*cf.* Černý & Ercit 1985, 1989).

Geochemical signature: general significance

Some geochemical features of the YITT-B pegmatites are quite obvious from the paragenetic characteristics outlined in "Mineralogy and Geochemistry" above, and in the preceding section. The pegmatites are considerably enriched in Sn and in Ta, which distinctly predominates over Nb. In contrast, the abundances of Li, Be, B and F (and, to a degree, Mn) are very low.

Tin is of no particular concern because its abundance or paucity depends first of all on its concentration in the magmatic source, and it can be distributed over most stages of melt fractionation, migration and consolidation *via* a multitude of transport mechanisms (*e.g.*, Taylor 1979). The fate of Sn is not solely connected with one or another of the other elements under consideration, to the exclusion of other factors.

The same can be said about beryllium, which can be transported by a variety of complex anions and released in the cationic form available for crystallization of minerals at many stages of pegmatite consolidation (Beus 1960). Although rarely documented in rare-element pegmatite populations worldwide, Be was in our case probably greatly depleted in the source, or extracted from the differentiating parent granitic melt at pre-pegmatite stages. The very few examples of the "beryl" category of pegmatites, in which this mineral is virtually absent, include the Plex pegmatite on Baffin Island (Tomascak *et al.* 1994), the Nyanga pegmatite in Uganda, discussed by Černý *et al.* (1989b), the Prašivá pegmatites in Slovakia (Uher *et al.* 1998a, b), and most of the phosphate-rich pegmatites in the Bavarian Oberpfalz (*e.g.*, Strunz *et al.* 1975).

If tourmaline is to be taken as representative of the concentration of B in the YITT-B pegmatite-forming melt, then the overall abundance of boron must have been extremely low. However, London *et al.* (1994) demonstrated experimentally that a concentration of at least 2 wt.% B_2O_3 is required to stabilize tourmaline in

TABLE 8. GEOCHEMICAL FEATURES OF PEGMATITE GROUPS IN THE SOUTHERN PART OF THE CAT LAKE - WINNIPEG RIVER FIELD

Pegmatite group	Characteristic rare elements																				
Shatford Lake		(Rb)		**Be**			LREE	**HREE**	**Y**	**Zr**	(Hf)	U	Th	(Sn)	Ti	**Nb**	Ta			**F**	
Lac du Bonnet	**Li**	(Rb)	(Cs)																(P)		
Greer Lake	(Li)	Rb	(Cs)	**Be**	(Zn)		(LREE)			(Zr)				(Sn)	(Ti)	Nb	Ta				
Eaglenest Lake		(Rb)		Be																	
Birse Lake		(Rb)																**B**			
Rush Lake	**Li**	Rb	Cs	Be										**Sn**	(Ti)	Nb	**Ta**	B	**P**	(F)	(S)
Bernic Lake	**Li**	**Rb**	**Cs**	**Be**		**Ga***				(Zr)	(Hf)			**Sn**	(Ti)	Nb	**Ta**	B	**P**	F	S
YITT-B		(Rb)												**Sn**	(Ti)	Nb	**Ta**		P		S

boldface - abundant; ( ) - minor; * except for the Bernic Lake group, data on Ga are very limited.

a peraluminous, hydrous, F-poor granitic melt. Thus some concentration of B must have been present, but generally below the tourmaline-stabilizing threshold, and it must have been dispersed, unless it was incorporated into micas (Černý *et al.* 1995). Analysis of the rock-forming and exocontact muscovite for B was not feasible in this study; however, concentrations of boron in ostensibly B-free minerals and their implications for the B content of the parent magma should be examined in the future, as boron is suspected to serve as a complexing agent for transport of rare elements (Wolf & London 1993, Wolf *et al.* 1994). Examples of beryl-type pegmatites lacking tourmaline are sparse: the best known are those of the Preissac – Lacorne field in Quebec (Mulja *et al.* 1995a, b), Greer Lake in the Cat Lake – Winnipeg River field (Černý *et al.* 1981, 1986), and the PEG group in the Yellowknife field (Meintzer 1987, Wise 1987).

As mentioned in the preceding section, the absence of lithium minerals is characteristic of pegmatites of the beryl type, except Li-bearing phosphates in the most evolved subtype of this category. Trace amounts of Li are, however, routinely encountered in other minerals of these pegmatites such as K-feldspar, muscovite, tourmaline or beryl. In this respect, the YITT-B pegmatites are among the most Li-poor ever encountered in the beryl subtype: triphylite is extremely rare, Li is below detection limit of the atomic-absorption spectroscopy analysis (1–2 ppm) of K-feldspar, extremely low in muscovite (Table 2, Fig. 4), very low in rare beryl and probably so in rare tourmaline. A Li-poor source of the examined pegmatite swarm must be invoked, as beryl-type pegmatites commonly exhibit concentrations of Li in rock-forming minerals that are much higher than those listed in mineral descriptions above (*cf.* Černý *et al.* 1981, Meintzer 1987, Anderson 1984, Mulja *et al.* 1995a, b, Roda *et al.* 1995).

Fluorine is yet another element that is present in very low concentrations in the pegmatites examined: compositions of muscovite and tourmaline show negligible F contents, and no F-bearing species were identified among the Nb,Ta-oxide minerals. Early euhedral apatite is about 95% fluorapatite; however, the late, primary but interstitial apatite in the albite-rich zones has ~$(F_{0.60-0.50}OH_{0.40-0.50})$ (Table 7). The only other phosphate with a substantial F content, triplite, is apparently very rare.

The fluorine deficiency is in marked contrast with the high degree of fractionation of Ta from Nb, which is generally assigned to the difference in thermal stabilities of F-complexes of these metals (*e.g.*, Černý & Ercit 1989, Keppler 1993). However, the concentration of F in the melt could have been distinctly higher than indicated by the mineralogy of the solidified pegmatites, and could have been sufficient to promote fractionation in terms of Nb–Ta, as the average assayed content of the Nb,Ta-oxide minerals does not exceed ~0.05 wt.%. Recent experimental results of Linnen & Keppler (1996), however, indicate that differential solubilities in a pegmatite-forming melt may be responsible for relegating Ta-rich minerals to the late stages of solidification, and the role of F-complexing may be not so essential. In contrast, fractionation of Mn from Fe seems to be hindered by low F-content of the parent melt, and particularly by low μ(K,Na)F. Manganese is routinely extremely fractionated in the Nb,Ta-oxide minerals of the lepidolite subtype of complex pegmatites, but its enrichment in the YITT-B minerals is only moderate. The beryl-type pegmatites of the Yellowknife field (Meintzer 1987, Wise 1987) and the Nyanga (Uganda) and Spittal (Austria) pegmatites (Černý *et al.* 1989a, b) can be quoted as close analogs with very low F contents, poor fractionation of Mn from Fe, but considerable enrichment in Ta relative to Nb. In conspicuous contrast, the Scheibengraben pegmatite in northern Moravia, Czech Republic, a member of the beryl – columbite (–phosphate) subtype, shows moderate content of F increasing toward the late stages of consolidation, concurrent with extreme extension of the range of fractionation of Nb from Ta and considerable enrichment in Mn in late stages (Novák *et al.* 1994, in prep.).

We can conclude that the pegmatites under study are, within their basic beryl – columbite – phosphate subtype, exceptionally poor in Li, Be, B, and F, with modest fractionation of Mn from Fe, but very advanced separation of Ta from Nb.

Geochemical signature: regional context

In view of the fact that the YITT-B pegmatites represent a geochemical anomaly with very few (and only approximate) analogs on global scale, it is not surprising that they are quite exotic in the spectrum of pegmatites in the southern part of the host Cat Lake – Winnipeg River field. Table 8 shows condensed geochemical features of the pegmatite groups populating this district in relation to the YITT-B swarm. Some of the groups show individual mutually comparable features, but the overall signatures are invariably different. For example, the Greer Lake group is the only other one to be B-deficient, but it also is extremely poor in P and S, which are well represented in the YITT-B swarm.

The cause of the broad diversity in the geochemistry of pegmatite groups must be sought in the different lithologies that yielded the anatectic granitic melts that were intruded and consolidated as fertile leucogranites parental to the individual pegmatite groups. Although no isotopic dating is available for the YITT-B pegmatites, there is no reason for divorcing them from the LCT family of peraluminous pegmatites of the Winnipeg River area, dated at 2640 ± 7 Ma (etched columbite – tantalite U–Pb: Baadsgaard & Černý 1993), statistically coeval with the NYF family of pegmatites of the Shatford Lake group and the Lac du Bonnet batholith [gadolinite U–Pb 2652 ± 5 Ma: Baadsgaard & Černý (1993); zircon U–Pb 2665 ± 20 Ma: Krogh *et al.* (1977), respectively]. Whereas the Shatford Lake group and its parent granite bear a strong juvenile, mantle-derived isotopic and trace-element imprint, the peraluminous pegmatites and their parent leucogranites display a variety of signatures indicative of derivation from tonalitic basement and mainly metasedimentary greenstone-belt lithologies (Longstaffe *et al.* 1981, Černý 1989b, 1990, 1991b).

The paragenetic, trace-element and isotopic diversity of the Winnipeg River granite + pegmatite systems is readily understandable when the structural setting, intrusive pathways and probable protoliths are considered. Figure 12 shows a section across the Bird River Greenstone Belt with projected loci of the individual leucogranites and their ascent trails. Thermal doming, associated in space and time with the intrusion of the Lac du Bonnet batholith and combined with frictional heat along major faults (Goad & Černý 1981;

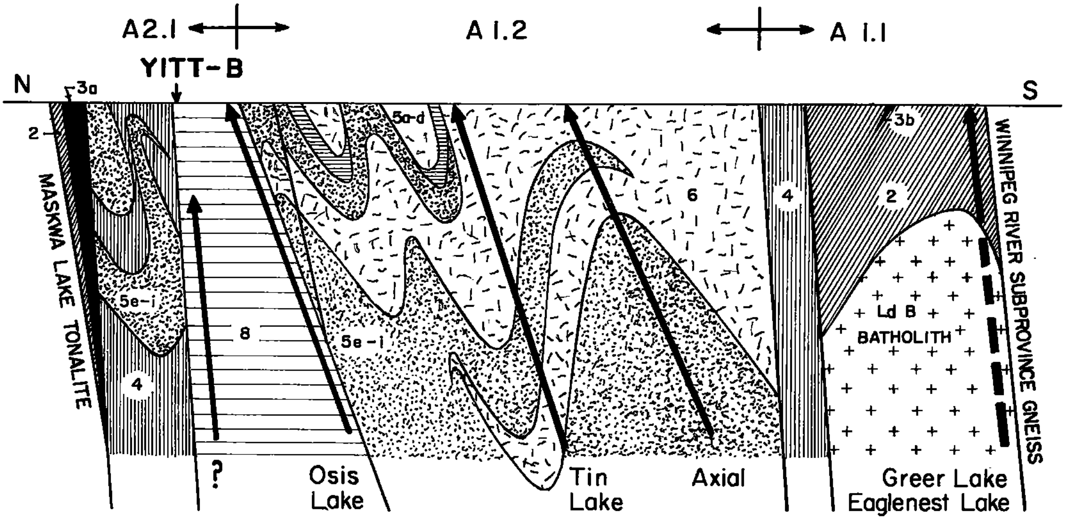

FIG. 12. A 12.7-km-long N–S section across the fault-bound Bird River Greenstone Belt, modified from Trueman (1980) and Černý *et al.* (1981). Units: 2: metabasalts of the Lamprey Falls Formation, 3a: serpentinite, peridotite, picrite and metagabbros of the Bird River sill, 3b: metagabbro and related synvolcanic intrusions, 4: metarhyolite and related volcaniclastic lithologies of the Peterson Creek Formation, 5a–d: metabasalt (to metarhyolite) and 5e–i: metaconglomerates, cordierite – anthophyllite and garnet – biotite schists of the Bernic Lake Formation, 6: composite synvolcanic intrusions of metagabbro, metadiorite, metagranodiorite, 8: metaconglomerates, metapelites and metagreywackes of the Booster Lake Formation. Heavy arrows mark projected pathways of emplacement of the fertile leucogranites. A1.1 and A1.2: upper greenschist subfacies (transitional to lower amphibolite subfacies); A2.1: lower amphibolite subfacies of the low-pressure sequence of Winkler (1967).

cf. Strong & Hanmer 1981) triggered partial melting in deep parts of the greenstone belt and in the adjacent metatonalitic basement, generating local pools of leucogranitic magmas strongly dependent on the petrochemistry of the locally available lithologies. The specific geochemical signature of the YITT-B swarm, different from any other pegmatite population of the region, is evidently due to the isolated location of this swarm north of the Bernic Lake group and west of the Osis Lake group. Lithologies from which the parent magma was extracted could have been distinctly different from those which served as protoliths for the other granite + pegmatite groups. Metamorphosed equivalents of rhyolite lavas and pyroclastic rocks probably were significantly involved, in contrast to source regions of other leucogranites (Fig. 12).

CONCLUSIONS

(i) The YITT-B pegmatite swarm belongs to the beryl – columbite – phosphate subtype of peraluminous, LCT-family granitic pegmatites of the rare-element class, but it is rather unique in several respects.

(ii) Beryl and tourmaline are extremely rare, and lithium minerals (including phosphates) are absent. In contrast, Nb,Ta-oxide minerals and cassiterite are abundant.

(iii) Except for their textural features and abundances of garnet, zones that would be routinely recognized as layered garnetiferous albite-rich aplite and saccharoidal albite display identical mineral content, mineral chemistries and mutual transitions.

(iv) Primary phosphates (apatite, beusite, triplite) and the Nb,Ta-oxide minerals (columbite – tantalite ≈ tapiolite > ixiolite in abundances) are very fine-grained and dispersed in the albite-rich zones.

(v) Late secondary alteration is restricted to the phosphates, generating alluaudite, carbonate-hydroxylapatite, phosphophyllite and (para)scholzite. The primary Nb,Ta-oxide minerals and other phases are intact.

(vi) The geochemical signature shows abundant Ta, Nb and Sn, but very low contents of Li, Be, B and F. Moderate fractionation of Mn from Fe is in conspicuous contrast with very advanced separation of Ta from Nb.

(vii) The paragenetic and geochemical features are quite exceptional on global scale, with only a limited number of approximate analogs in other pegmatite fields. The swarm also is distinctly different from other pegmatite populations in its parent Cat Lake – Winnipeg River field. Localized anatexis in greatly variable lithologies of the basement, greenstone belt and adjacent metamorphic suites, separate paths of magma ascent, and internal differentiation of individual fertile leucogranitic magmas generated the diversity of pegmatite types composing the field.

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