

# Non-pegmatitic Deposits of Beryllium: Mineralogy, Geology, Phase Equilibria and Origin

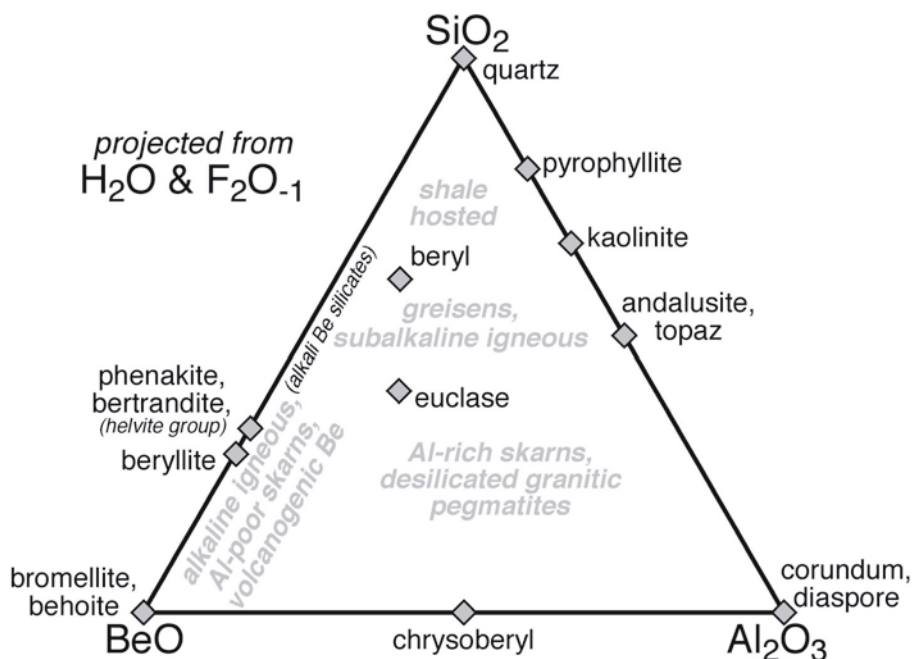
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## INTRODUCTION

Non-pegmatitic occurrences of Be minerals constitute a diverse set of geologic environments of considerable mineralogical and petrological interest; they currently provide the majority of the world's Be ore and emeralds and they contain the greatest resource of these commodities. Of the approximately 100 Be minerals known (see Chapter 1 by Grew; Appendix A), most occur in hydrothermal deposits or non-pegmatitic igneous rocks, where their distribution varies systematically with the setting and origin (Table 1, Fig. 1).



**Figure 1.** Chemography of the principal solid phases in the BeO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub>-H<sub>2</sub>O(-F<sub>2</sub>O<sub>.1</sub>) “BASH” system with the projected positions of helvite group and alkali Be silicates. Also shown are generalized fields for some of the major types natural of occurrences (cf. Table 1, Fig. 4; see text for discussion).

Beryllium minerals are best known from geologic systems associated with felsic magmatism. They also occur in a variety of settings that lack evident igneous associations. Environments range from the surface to the deep crust and host rocks range from feldspathic to carbonate to ultramafic in composition. Genetically related igneous rocks are felsic and share low calcium and high F contents, but are diverse in composition, setting and origin. Compositions range from strongly peraluminous to

**Table 1.** Beryllium minerals in non-pegmatitic deposits: formulas, use and occurrence.<sup>1</sup>

<i>Mineral</i>	<i>Formula</i>	<i>Use</i>	<i>Family</i> <sup>2</sup>	<i>Occurrence</i>
<b>Behoite</b> (Bht) <sup>3</sup>	Be(OH) <sub>2</sub>	ore?	BASH	alkalic pegmatite, skarns, rhyolites
<b>Bertrandite</b> (Brt)	Be <sub>4</sub> Si <sub>2</sub> O <sub>7</sub> (OH) <sub>2</sub>	ore	BASH	granitic and alkalic pegmatites, greisens, skarns, veins, rhyolites
<b>Beryl</b> (Brl)	Be <sub>3</sub> Al <sub>2</sub> Si <sub>6</sub> O <sub>18</sub>	ore, gem	BASH	granitic pegmatites, greisens, skarns, veins, rhyolites
<b>Beryl v. Emerald</b>	Be <sub>3</sub> (Al,Cr,V) <sub>2</sub> Si <sub>6</sub> O <sub>18</sub>	gem	BASH	granitic pegmatites & metamorphosed equivalents, veins, metamorphic rocks
Beryllite (Byl)	Be <sub>3</sub> SiO <sub>4</sub> (OH) <sub>2</sub> ·H <sub>2</sub> O		BASH	alkalic pegmatite
Bromellite (Brm)	BeO		BASH	skarns, desilicated pegmatites
<b>Chrysoberyl</b> (Ch)	BeAl <sub>2</sub> O <sub>4</sub>	ore, gem	BASH	granitic pegmatites, skarns
Clinobehoite (Cbe)	Be(OH) <sub>2</sub>		BASH	desilicated pegmatites
<b>Euclase</b> (Euc)	BeAlSiO <sub>4</sub> (OH)	gem	BASH	granitic pegmatites, greisens, skarns, veins
<b>Phenakite</b> (Ph)	Be <sub>2</sub> SiO <sub>4</sub>	ore, gem	BASH	alkaline & granitic pegmatites, skarns, greisens, veins
<b>Bazzite</b> (Bz)	Be <sub>3</sub> (Sc,Al) <sub>2</sub> Si <sub>6</sub> O <sub>18</sub>		BASH+	alkalic and granitic pegmatites, veins
Magnesiotaaffeite-2N'2S (Taf) ("Taaffeite", "Taprobanite")	BeMg <sub>3</sub> Al <sub>8</sub> O <sub>16</sub>	gem	BASH+	Mg-Al schists (metamorphosed pegmatite?), skarns
Magnesiotaaffeite-6N'3S (Mgr) ("Musgravite")	BeMg <sub>2</sub> Al <sub>6</sub> O <sub>12</sub>	gem	BASH+	metamorphosed pegmatites
Stoppaniite (Spp)	(Na,□)(Fe <sup>3+</sup> ,Al,Mg) <sub>2</sub> -Be <sub>3</sub> Si <sub>6</sub> O <sub>18</sub> ·H <sub>2</sub> O		BASH+	alkaline volcanic
Surinamite (Sur)	Mg <sub>3</sub> Al <sub>4</sub> (BeSi <sub>3</sub> O <sub>16</sub> )		BASH+	metamorphosed pegmatites
Aminoffite (Am)	Ca <sub>3</sub> Be <sub>2</sub> Si <sub>3</sub> O <sub>10</sub> (OH) <sub>2</sub>		alkaline	skarns
<b>Barylite</b> (Bar)	BaBe <sub>2</sub> Si <sub>2</sub> O <sub>7</sub>	ore	alkaline	alkalic pegmatite; skarns; greisens
<b>Bavenite</b> (Bav)	Ca <sub>4</sub> Be <sub>2</sub> Al <sub>2</sub> Si <sub>9</sub> O <sub>26</sub> (OH) <sub>2</sub>		alkaline	alkalic and granitic pegmatites, veins, skarns, greisens
Chkalovite (Chk)	Na <sub>2</sub> BeSi <sub>2</sub> O <sub>6</sub>		alkaline	alkalic pegmatites
<b>Epididymite</b> (Epd)	Na <sub>2</sub> Be <sub>2</sub> Si <sub>6</sub> O <sub>15</sub> ·H <sub>2</sub> O	ore?	alkaline	alkalic pegmatites, skarns
Eudidymite (Eud)	Na <sub>2</sub> Be <sub>2</sub> Si <sub>6</sub> O <sub>15</sub> ·H <sub>2</sub> O		alkaline	alkalic pegmatites
<b>Gadolinite</b> <sup>4</sup> -(Y), -(Ce) (Gad)	Be <sub>2</sub> Fe(Y,REE) <sub>2</sub> Si <sub>2</sub> O <sub>10</sub>	ore	alkaline	alkaline pegmatites and granites, veins, greisens
Gugiaite (Gug)	Ca <sub>2</sub> BeSi <sub>2</sub> O <sub>7</sub>		alkaline	skarns
Hingganite <sup>4</sup> -(Y) (Hin)	Be <sub>2</sub> (□,Fe)(Y,REE) <sub>2</sub> Si <sub>2</sub> O <sub>8</sub> -(OH,O) <sub>2</sub>		alkaline	alkaline pegmatites
Hsianghualite (Hsh)	Ca <sub>3</sub> Li <sub>2</sub> Be <sub>3</sub> (SiO <sub>4</sub> ) <sub>3</sub> F <sub>2</sub>		alkaline	skarns
Hyalotekite (Htk)	(Ba,Pb,K) <sub>4</sub> (Ca,Y) <sub>2</sub> Si <sub>8</sub> -(B,Be) <sub>2</sub> (Si,B) <sub>2</sub> O <sub>28</sub> F		alkaline	Fe-Mn "skarns"; alkaline pegmatites
Joemithite (Jo)	PbCa <sub>2</sub> (Mg,Fe <sup>2+</sup> ,Fe <sup>3+</sup> ) <sub>5</sub> [Si <sub>6</sub> Be <sub>2</sub> O <sub>22</sub> ](OH) <sub>2</sub>		alkaline	Fe-Mn "skarns"

Leifite (Lf)	(Na,□)(H <sub>2</sub> O,□)Na <sub>6</sub> Be <sub>2</sub> [Al, Si,Zn) <sub>3</sub> Si <sub>15</sub> O <sub>39</sub> F <sub>2</sub> ]		alkaline	alkaline pegmatites
<b>Leucophanite</b> (Lph)	CaNaBeSi <sub>2</sub> O <sub>6</sub> F	ore?	alkaline	alkaline pegmatites; skarns
Lovdarite (Lv)	K <sub>2</sub> Na <sub>6</sub> (Be <sub>4</sub> Si <sub>14</sub> O <sub>36</sub> )·9H <sub>2</sub> O		alkaline	alkaline pegmatites
<b>Meliphanite</b> (Mph)	Ca <sub>4</sub> (Na,Ca) <sub>4</sub> Be <sub>4</sub> AlSi <sub>7</sub> O <sub>24</sub> - (F,O) <sub>4</sub>		alkaline	alkaline pegmatites; skarns
<b>Milarite</b> (Mil)	K(□,H <sub>2</sub> O,Na) <sub>2</sub> (Ca,Y, REE) <sub>2</sub> (Be,Al) <sub>3</sub> Si <sub>12</sub> O <sub>30</sub>		alkaline	alkaline & granitic pegmatites; skarns; veins
Odintsovite (Od)	K <sub>2</sub> (Na,Ca,Sr) <sub>4</sub> (Na,Li)Ca <sub>2</sub> - (Ti,Fe <sup>3+</sup> ,Nb) <sub>2</sub> O <sub>2</sub> [Be <sub>4</sub> Si <sub>12</sub> O <sub>36</sub> ]		alkaline	alkaline veins
Roggianite (Rg)	Ca <sub>2</sub> [Be(OH) <sub>2</sub> Al <sub>2</sub> Si <sub>4</sub> O <sub>13</sub> ] •<2.5H <sub>2</sub> O		alkaline	veins, pegmatites
Samfowlerite (Sf)	Ca <sub>14</sub> Mn <sub>3</sub> Zn <sub>2</sub> (Zn,Be) <sub>2</sub> Be <sub>6</sub> - (SiO <sub>4</sub> ) <sub>6</sub> (Si <sub>2</sub> O <sub>7</sub> ) <sub>4</sub> (OH,F) <sub>6</sub>		alkaline	Fe-Mn “skarns”
Semenovite-(Ce) (Sem)	(Ce,La,REE,Y) <sub>2</sub> Na <sub>0.2</sub> (Ca,Na) <sub>8</sub> - (Fe,Mn)(Si,Be) <sub>20</sub> (O,OH,F) <sub>48</sub>		alkaline	alkaline pegmatites
Sorensenite (Ss)	Na <sub>4</sub> SnBe <sub>2</sub> Si <sub>6</sub> O <sub>18</sub> ·2H <sub>2</sub> O		alkaline	alkaline veins
Sverigeite (Sv)	Na(Mn,Mg) <sub>2</sub> Sn[Be <sub>2</sub> Si <sub>3</sub> O <sub>12</sub> (OH)]		alkaline	Fe-Mn “skarns”
Trimerite (Trm)	(Mn <sub>2</sub> Ca)[BeSiO <sub>4</sub> ] <sub>3</sub>		alkaline	Fe-Mn “skarns”
Tugtupite (Ttp)	Na <sub>4</sub> BeAlSi <sub>4</sub> O <sub>12</sub> Cl		alkaline	alkalic pegmatites & veins
Wawayandaite (Ww)	Ca <sub>12</sub> Mn <sub>4</sub> B <sub>2</sub> Be <sub>18</sub> Si <sub>12</sub> O <sub>46</sub> (OH,Cl) <sub>30</sub>		alkaline	Fe-Mn “skarns”
Welshite (Wsh)	Ca <sub>2</sub> Mg <sub>3.8</sub> Mn <sup>2+</sup> <sub>0.6</sub> Fe <sup>2+</sup> <sub>0.1</sub> Sb <sup>5+</sup> <sub>1.5</sub> O <sub>2</sub> - [Si <sub>2.8</sub> Be <sub>1.7</sub> Fe <sup>3+</sup> <sub>0.65</sub> Al <sub>0.7</sub> As <sub>0.17</sub> O <sub>18</sub> ]		alkaline	Fe-Mn “skarns”
<b>Danalite</b> (Dn)	Fe <sub>4</sub> Be <sub>3</sub> Si <sub>3</sub> O <sub>12</sub> S	ore	helvite	skarns, granitic pegmatites
<b>Genthelvite</b> (Gnt)	Zn <sub>4</sub> Be <sub>3</sub> Si <sub>3</sub> O <sub>12</sub> S		helvite	alkaline pegmatites, carbonatite
<b>Helvite</b> (Hlv)	Mn <sub>4</sub> Be <sub>3</sub> Si <sub>3</sub> O <sub>12</sub> S		helvite	veins, skarns, greisens, alkaline and granitic pegmatites
Babephite (Bf)	BaBe(PO <sub>4</sub> )F		non-silicate	placer (alkaline igneous?)
Bearsite (Bs)	Be <sub>2</sub> (AsO <sub>4</sub> )(OH)·4H <sub>2</sub> O		non-silicate	polymetallic porphyry
Bergslagite (Bsg)	CaBeAsO <sub>4</sub> (OH)		non-silicate	Fe-Mn “skarns”
Beryllonite	NaBePO <sub>4</sub>		non-silicate	granite and granitic pegmatites
“Glucine” (Gl)	CaBe <sub>4</sub> (PO <sub>4</sub> ) <sub>2</sub> (OH) <sub>4</sub> ·0.5H <sub>2</sub> O		non-silicate	weathering
<b>Hambergite</b> (Hmb)	Be <sub>2</sub> (OH,F)BO <sub>3</sub>		non-silicate	alkaline and granitic pegmatites
<b>Herderite</b> (Hrd)	CaBePO <sub>4</sub> (F,OH)		non-silicate	greisens, granitic pegmatites
<b>Hurlbutite</b> (Hrb)	CaBe <sub>2</sub> (PO <sub>4</sub> ) <sub>2</sub>		non-silicate	granitic pegmatites, veins
Moraesite (Mr)	Be <sub>2</sub> (PO <sub>4</sub> )(OH)·4H <sub>2</sub> O		non-silicate	granitic pegmatites, weathering
Swedenborgite (Sw)	NaBe <sub>4</sub> SbO <sub>7</sub>		non-silicate	Fe-Mn “skarns”
Uralolite (Ur)	Ca <sub>2</sub> Be <sub>4</sub> (PO <sub>4</sub> ) <sub>3</sub> (OH) <sub>3</sub> ·5H <sub>2</sub> O		non-silicate	greisen, granitic pegmatites
Berberite (Bb)	Be <sub>2</sub> BO <sub>3</sub> (OH,F)·H <sub>2</sub> O		non-silicate	pegmatite, alkaline igneous(?), skarn

<sup>1</sup> More common minerals in **bold**. Compiled from Mandarino (1999), Strunz and Nickel (2001) and Appendix 1 of Chapter 1 of this volume.

<sup>2</sup> Minerals are grouped into the four families by common chemical characteristics: (1) predominantly BeO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub>-H<sub>2</sub>O the “BASH” group, including a subgroup “BASH+” for minerals also containing Mg, Fe, Sc and Na, (2) Na-Ca-K silicates—“alkaline” group, (3) M<sub>4</sub>Be<sub>3</sub>Si<sub>3</sub>O<sub>12</sub>S helvite group, and (4) complex non-silicates (phosphates, borates, arsenates, etc.).

<sup>3</sup> Abbreviations for Be- and other minerals are taken from Kretz (1983) or constructed to be consistent with that paper. They are used in most figures, Table 2 and Appendix A. In alphabetical order (by abbreviation) these are: Act (actinolite), Ad (K-feldspar var. adularia), Aeg (aegerine), Agt (aegerine-augite), Ab (albite),

*Table 1 footnotes, continued.*

<sup>3</sup> Abbreviations, continued: Am (amphibole), Amz (K-feldspar var. amazonite), Anc (analcime), And (andalusite), Ap (apatite), Ath (anthophyllite), Bt (biotite), Cal (calcite), Carb (carbonates), Chl (chlorite), Chr (chromite), Col (columbite), Cpx (Ca-clinopyroxene), Crn (corundum), Cst (cassiterite), Cyl (cryolite), Dsp (diaspore), Drv (dravite), Ep (epidote), Eud (eudialyte), Fa (fayalite), Fl (fluorite), fo (forsterite)Fs (feldspar), Ghn (gahnite), Grt (garnet), Hbl (hornblende), Hdd (spodumene var. hiddenite), Hem (hematite), Kfs (K-feldspar), Kln (kaolinite), Ky (kyanite), Mag (magnetite), Mc (microcline), Mnz (monazite), Mo (molybdenite), Ms (muscovite), Ne (nepheline), Ntr (natrolite), Ofs (oligoclase), Pas (parisite), Phl (phlogopite), Pll (polythionite), Pl (plagioclase), Prl (pyrophyllite), Px (pyroxene), Py (pyrite), Qtz (quartz), Rbk (riebeckite), Sch (scheelite), Sid (siderite), Sdl (sodalite), Sid (siderophyllite), Tlc (talc), Toz (topaz), Tr (tremolite), Ttn (titanite), Tur (tourmaline), Ves (vesuvianite [idocrase]), W (water), Wlf (wlfamite), Znw (zinnwaldite), Zrn (zircon).

<sup>4</sup> Most investigators have not distinguished gadolinite-(Y) and gadolinite-(Ce), so gadolinite-group minerals are simply referred to in the text as “gadolinite”. Similarly, hingganite-group minerals are simply referred to in the text as “hingganite.”

peralkaline and can be silica undersaturated. Beryllium minerals also occur in metamorphic and basinal environments and are redistributed by surface processes. Table 2 summarizes the types and significance of major groups of occurrences by their lithologic associations. Figure 2 shows the global distribution of some important examples and regional belts. For most types, at least one example has been described in some detail and can be used to help evaluate general patterns; however, even in these only rarely has Be been the principal economic interest.

Few papers cover this spectrum of deposits. The classic synthesis studies are from the Soviet literature (e.g., Beus 1966; Vlasov 1968; Zabolotnaya 1977; Ginzburg et al. 1979; Grigor'yev 1986) with few extensive summaries in the western literature (e.g., Warner et al. 1959; Mulligan 1968; Sinkankas 1981). The golden age of investigation was in the 1950s and 1960s, driven by exploration interest in the U.S. and the (then) Soviet Union, with most papers published between about 1960 and 1985. Much quality work was done by Soviet scientists, a moderate amount of which is available in English translation. Unfortunately many of the detailed studies are in limited-distribution monographs and reports that are difficult to access. Many compendia of papers dealing with aspects of rare metal systems have been published that contain related papers (Evans 1982; Hutchison 1988; Taylor et al. 1988; Moeller et al. 1989; Stein et al. 1990; Seltsmann et al. 1994; Pollard 1995b; Kremenetsky et al. 2000b and earlier volumes). Continuing work on Be-bearing magmatic systems, particularly pegmatites, is reviewed by Černý (this volume) and London and Evensen (this volume).

This chapter reviews the principal types of non-pegmatitic Be occurrences—magmatic, hydrothermal, metamorphic and surface-related—covering aspects of their mineralogy, stability, geologic framework, genesis and global distribution. Although there is a continuum between pegmatitic and non-pegmatitic occurrences, granitic pegmatites are only briefly mentioned here. In spite of the considerable study that the non-pegmatitic occurrences have received as possible sources of Be as a commodity or of Be minerals as gems or specimens, there remains a great deal to be learned about the characteristics and origins of these systems.

### **Economic sources of beryllium and beryllium minerals**

***Beryllium ore.*** Prior to about 1970, the main source of Be was hand-picked pegmatitic beryl typically from small, labor-intensive operations. New uses for Be in nuclear and other high-tech applications motivated extensive exploration campaigns for Be and other rare metals from the 1940s through the early 1960s. These efforts resulted in the discovery in the Soviet Union, the United States, and Canada of many significant

occurrences of non-pegmatitic Be mineralization. The Spor Mountain, Utah Be deposits, the world's most important source of Be (Cunningham 2000), were discovered during intensive regional Be exploration in 1959 and began producing in 1969. This exploration was aided by the recognition of the association of Be with chemically evolved felsic igneous rocks, the occurrence with F-rich rocks, and the development of neutron-sourced gamma ray spectrometers ("berylometers", Brownell 1959), which enabled rapid semi-quantitative assay in the field of the Be content of rocks (e.g., Meeves 1966).

Global production of Be in 2000 was 226 tonnes (t) of metal equivalent of which about 75% (180 t) was produced in the U.S. from the Spor Mountain operation of Brush Wellman Corporation (Cunningham 2000). In 1998, Brush Wellman reported reserves for the Spor Mountain district of 7 million tonnes (Mt) at 0.26% Be (0.72% BeO) or about 18,300 t of contained metal. Global production was down from 289 t in 1998 and represents less than half of world capacity. Consumption in 1998 (390 t) was substantially larger and was supported by sales of ore from U.S. government stockpiles. A total value of \$140 million was based on quoted prices for Be-Cu master alloy, the main product.

Presently there is little economic incentive for Be exploration, because the Spor Mountain district alone contains roughly 50 years of resource at current consumption rates and large, sub-economic resources have been identified in a number of other areas (Fig. 3, see Appendix A). Solodov (1977) gave general estimates for types of Be deposit as a function of age, setting, and type. His estimates totaled >100,000 t of contained Be metal of which half is in non-pegmatitic deposits with grades  $\geq 0.05\%$  Be. Many times this amount likely exist in the numerous unevaluated occurrences that resemble the better known deposits (data compiled in Appendix A indicate >200,000 t of contained Be).

**Gems.** Non-pegmatitic deposits are also major sources of gems, notably emerald, aquamarine, red beryl and alexandrite (chrysoberyl). Desilicated granitic pegmatites and veins in ultramafic and mafic rocks provide emerald, chrysoberyl, and some phenakite (Beus 1966; Sinkankas 1981). Shear-zone and vein-type emerald deposits are also important, especially the black shale-hosted deposits of Colombia (Snee and Kazmi 1989; Cheilletz 1998). Most aquamarine occurrences are pegmatitic, however some gem material comes from miarolitic cavities, greisens and veins, and a considerable fraction is reworked by surficial processes into placer deposits. Many of the hard rock occurrences also produce sought-after specimens of other Be minerals such as phenakite and bertrandite (Sinkankas 1981; Jacobson 1993a). In 1999 U.S. production of beryl gemstones totaled approximately \$3 million and U.S. consumption of cut emeralds (~1/3 world total) amounted to about 5 million carats (1,000 kg) worth approximately \$180 million (Olson 2000). Global resource estimates for Be gemstones do not exist.

Although economic deposits of Be and Be gems are limited to Spor Mountain, granitic pegmatites, and a large handful of gem producing districts, the varied occurrence of and popular and scientific interest in Be minerals merit a more general treatment.

## TYPES OF DEPOSITS

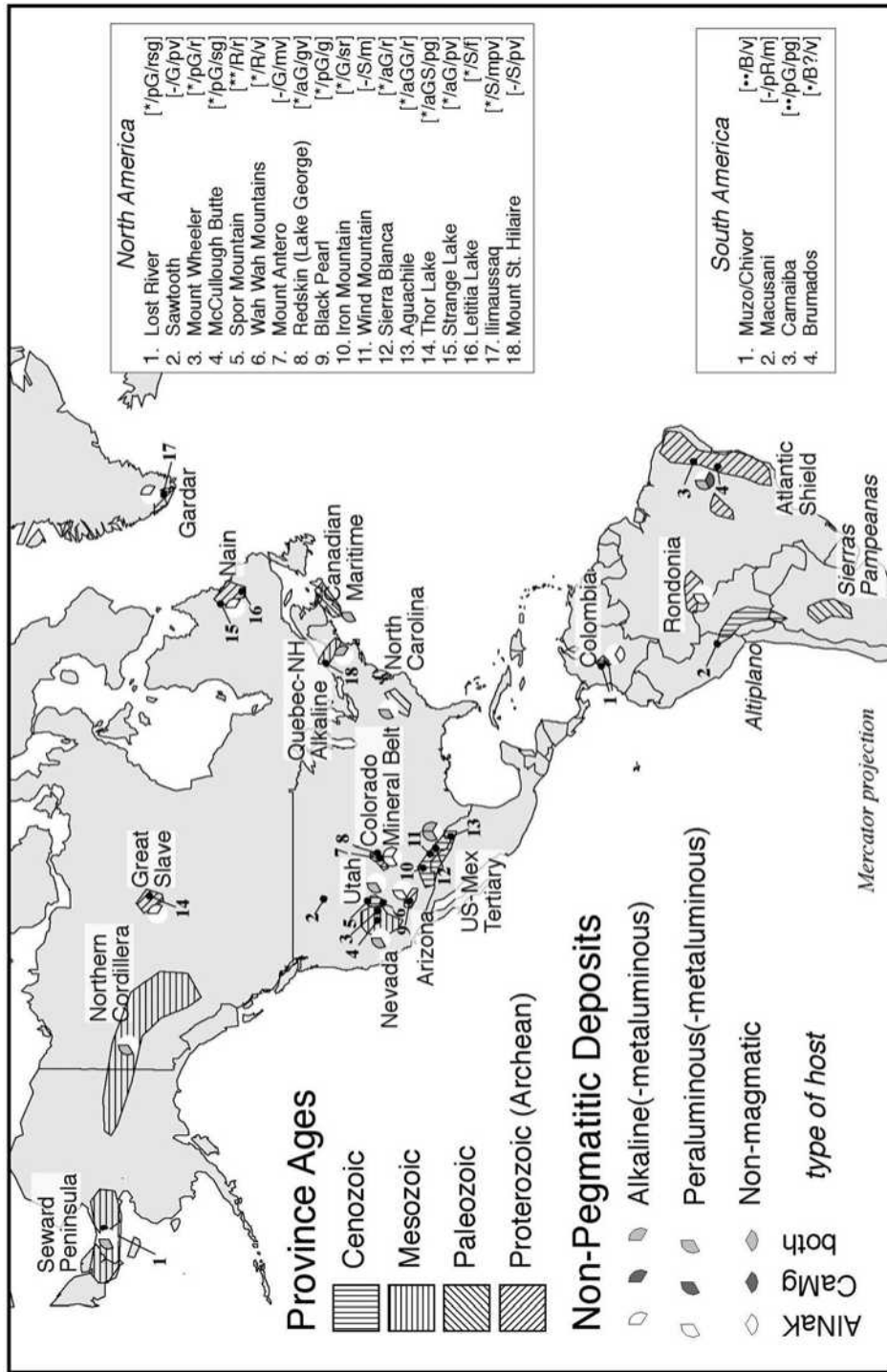
We group Be deposits by geologic setting (Table 2) specifically emphasizing differences in (1) associated sources (magmas or other materials) and (2) depositional environment (magmatic or metasomatic, and the host). Figure 4 illustrates the general geologic environments for the major groups of occurrences. Beryllium deposits naturally divide into igneous-related and non-magmatic types. They divide further by the nature of the associated magma and the host rock. As explained below, host rock and magma compositions exert strong controls on Be mineralogy as a function of their acidity-

**Table 2.** Main types of beryllium occurrences: significance, principal beryllium minerals, and examples.<sup>1</sup>

<i>Association</i>		<i>Igneous</i>			<i>Metasomatic</i>		
<i>type</i>	<i>variety</i>	<i>magmatic</i>	<i>pegmatitic</i>	<i>aluminosilicate</i> (greisen, vein)	<i>carbonate</i> (skarn, replacement)	<i>mafic/ultramafic</i> (blackwall, vein)	
<i>Igneous connection direct</i>							
granite	metaluminous to peraluminous	<ul style="list-style-type: none"> <li>• common? / Be resource?</li> <li>• Li micas, beryl</li> <li>• Beauvoir, France; Sheeprock, USA</li> </ul>	<ul style="list-style-type: none"> <li>• abundant / gems &amp; major source of Be beryl</li> <li>• Minas Gerais, Brazil; Bernic Lake, Canada</li> </ul>	<ul style="list-style-type: none"> <li>• abundant / gems±Be resource(?)</li> <li>• beryl, phenakite</li> <li>• Sherlova Gora, Russia; Mt. Antero, USA; Aqshatau, Kazakhstan</li> </ul>	<ul style="list-style-type: none"> <li>• abundant / Be resource</li> <li>• phenakite, chrysoberyl, bertrandite, helvite gr.</li> <li>• Lost River, USA; Mt. Wheeler, USA; Mt. Bischoff, Australia</li> </ul>	<ul style="list-style-type: none"> <li>• common / emerald</li> <li>• beryl, chrysoberyl, phenakite</li> <li>• Reft River, Russia; Khaltaro, Pakistan; Carnaiba, Brazil</li> </ul>	
rhyolite	metaluminous to peraluminous	<ul style="list-style-type: none"> <li>• common / —</li> <li>• Be in glass or micas</li> <li>• Macusani, Peru; topaz rhyolites, western USA</li> </ul>	<ul style="list-style-type: none"> <li>• rare / gem red beryl</li> <li>• beryl</li> <li>• Wah Wah Mtns, USA; Black Range, USA; (cf. Spor Mtn. USA)</li> </ul>	<ul style="list-style-type: none"> <li>• uncommon / principal source of Be</li> <li>• bertrandite</li> <li>• Spor Mtn, USA; Aguachile, Mexico</li> </ul>	<ul style="list-style-type: none"> <li>• rare / Be resource</li> <li>• phenakite, bertrandite, leucophanite</li> <li>• Ermakovskoe, Russia</li> </ul>	<ul style="list-style-type: none"> <li>• —</li> </ul>	
granite	peralkaline	<ul style="list-style-type: none"> <li>• rare? / —</li> <li>• ?</li> <li>• Khaldzan-Burgtey, Mongolia</li> </ul>	<ul style="list-style-type: none"> <li>• uncommon / Be resource</li> <li>• gadolinite, phenakite</li> <li>• Strange Lake, Canada</li> </ul>	<ul style="list-style-type: none"> <li>• rare / Be resource</li> <li>• phenakite, helvite, gadolinite</li> <li>• Verknee Espee, Kazakhstan</li> </ul>	<ul style="list-style-type: none"> <li>• rare / Be resource</li> <li>• phenakite, bertrandite, leucophanite</li> <li>• Ermakovskoe, Russia</li> </ul>	<ul style="list-style-type: none"> <li>• —</li> </ul>	
rhyolite	peralkaline	<ul style="list-style-type: none"> <li>• rare? / Be resource?</li> <li>• ?</li> <li>• Brockman, Australia</li> </ul>	<ul style="list-style-type: none"> <li>• common / —</li> <li>• epididymite, eudidymite, chkalovite</li> <li>• Lovozero, Russia; Ilimaussaq, Greenland</li> </ul>	<ul style="list-style-type: none"> <li>• rare / Be resource</li> <li>• berylite, eudidymite</li> <li>• Seal Lake, Canada; Thor Lake, Canada</li> </ul>	<ul style="list-style-type: none"> <li>• rare / Be resource</li> <li>• bertrandite</li> <li>• Sierra Blanca USA</li> </ul>	<ul style="list-style-type: none"> <li>• —</li> </ul>	
syenites	peralkaline	<ul style="list-style-type: none"> <li>• uncommon / Be resource?</li> <li>• chkalovite, epididymite</li> <li>• Ilimaussaq, Greenland; Wind Mtn, USA</li> </ul>	<ul style="list-style-type: none"> <li>• common / —</li> <li>• epididymite, eudidymite, chkalovite</li> <li>• Lovozero, Russia; Ilimaussaq, Greenland; Wind Mtn, USA</li> </ul>	<ul style="list-style-type: none"> <li>• rare / Be resource</li> <li>• berylite, eudidymite</li> <li>• Seal Lake, Canada; Thor Lake, Canada</li> </ul>	<ul style="list-style-type: none"> <li>• rare / Be resource</li> <li>• bertrandite</li> <li>• Sierra Blanca USA</li> </ul>	<ul style="list-style-type: none"> <li>• —</li> </ul>	

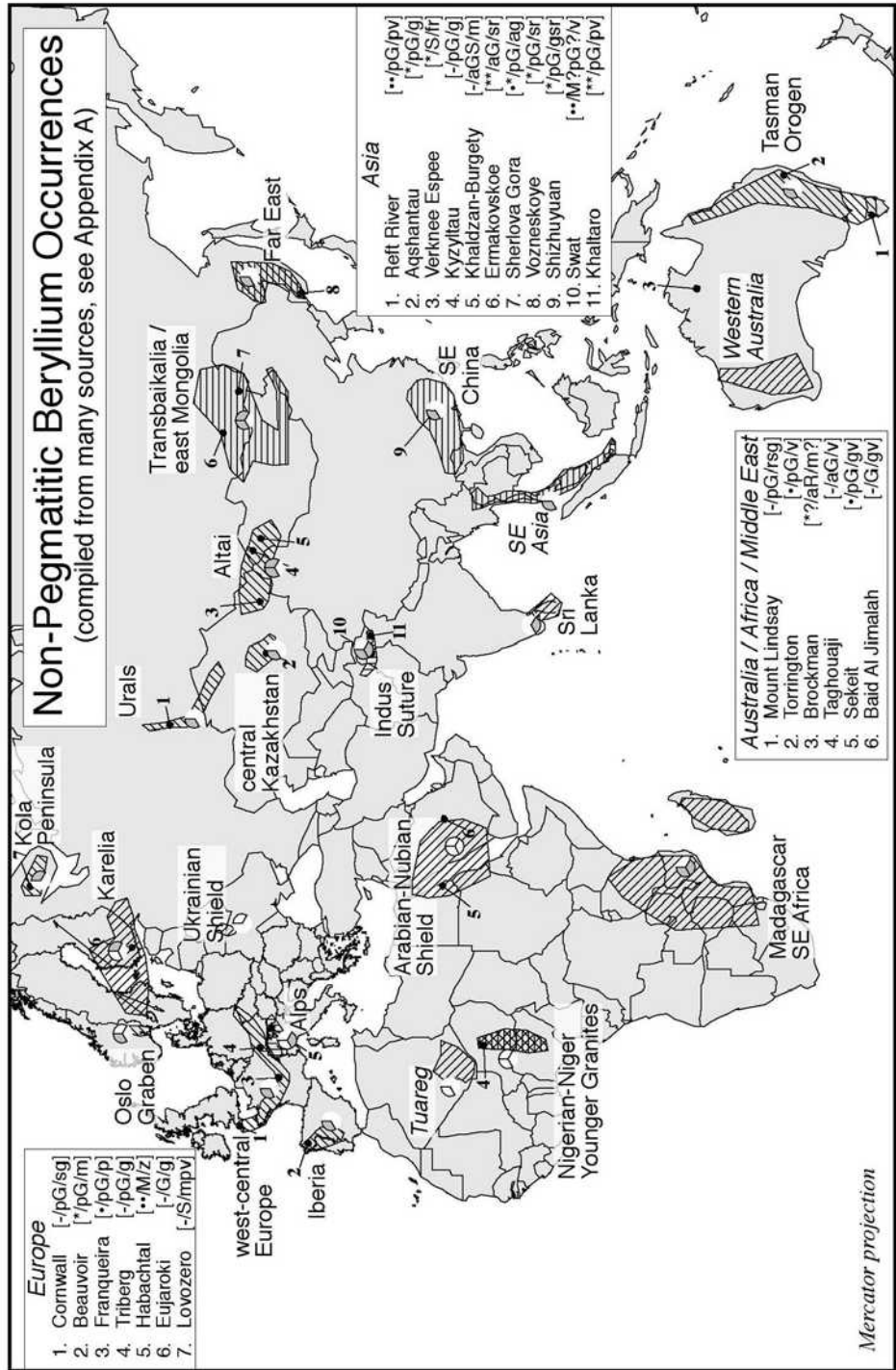
<i>Igneous connection indirect or absent (“non-magmatic”)</i>						
metamorphic	shear / vein	• —	• —	<ul style="list-style-type: none"> <li>• uncommon / specimens</li> <li>• phenakite, milarite, bavenite</li> <li>• Swiss &amp; Italian Alps</li> </ul>	<ul style="list-style-type: none"> <li>• uncommon? / emerald</li> <li>• beryl</li> <li>• Mingora, Pakistan; Brumado, Brazil</li> </ul>	<ul style="list-style-type: none"> <li>• uncommon? / emerald source</li> <li>• beryl</li> <li>• Habachtal, Austria;</li> </ul>
basin	vein	• —	• —	<ul style="list-style-type: none"> <li>• uncommon / premier emerald source</li> <li>• beryl (euclase)</li> <li>• Muzo &amp; Chivor, Colombia</li> </ul>	<ul style="list-style-type: none"> <li>• —</li> </ul>	<ul style="list-style-type: none"> <li>• —</li> </ul>
miscellaneous	hydrothermal	• common / —; Be minerals and enrichments are present in a variety of Mn-rich hydrothermal systems and Fe-Mn oxide-rich deposits including hot-spring systems (Butte, USA, Silverton, USA; Långban, Sweden; Golconda, USA)				
surface	placer	• common / source of gems; most placer Be minerals are believed to be derived from pegmatites or metamorphosed pegmatites, but they can come from many sources of resistant Be minerals (Sri Lanka; Madagascar; Minas Gerais, Brazil)				
surface	supergene	• rare(?) / —; little documented, BASH minerals survive; helvite group & Na-Ca silicates mostly weather; local supergene transport				

<sup>1</sup> For each type of occurrence, the first bullet summarizes the number of occurrences and the economic importance, if any; the second bullet list the main Be-bearing phases, and the third bullet names one or more prominent localities. A dash indicates absence.

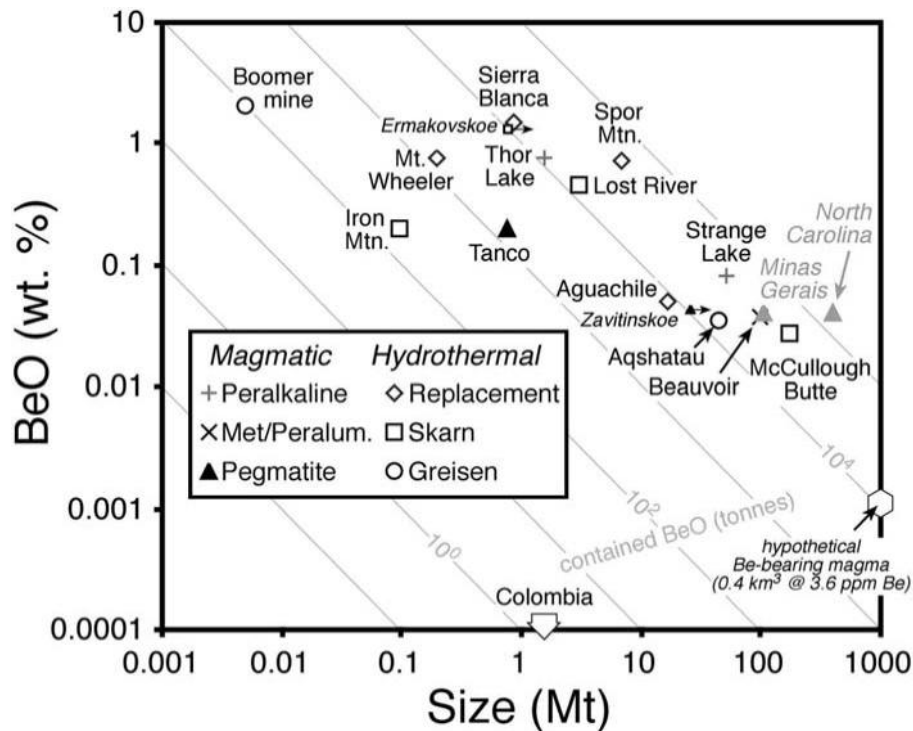


**Figure 2.** Map showing the location of non-pegmatitic Be occurrences and the Be-bearing belts that are mentioned in this paper. Deposit-type assignments can be uncertain or generalized depending on available information and the complexity of the region. The Symbolism for the key deposits is: significance (0-2 stars) / petrologic association (**figure and caption continued next page >>**)





(G = granitoid, S = synite, R = rhyolite, B = basinal; M = metamorphic; for igneous rocks: p = strongly peraluminous m = principally metaluminous; a = peralkaline) / deposit style (f = fenites, g = greisen, m = magmatic, p = pegmatitic, r = replacements s = skarn, v = veins, z = reaction zones). See text for discussion. Compiled from multiple sources summarized in Appendix A.

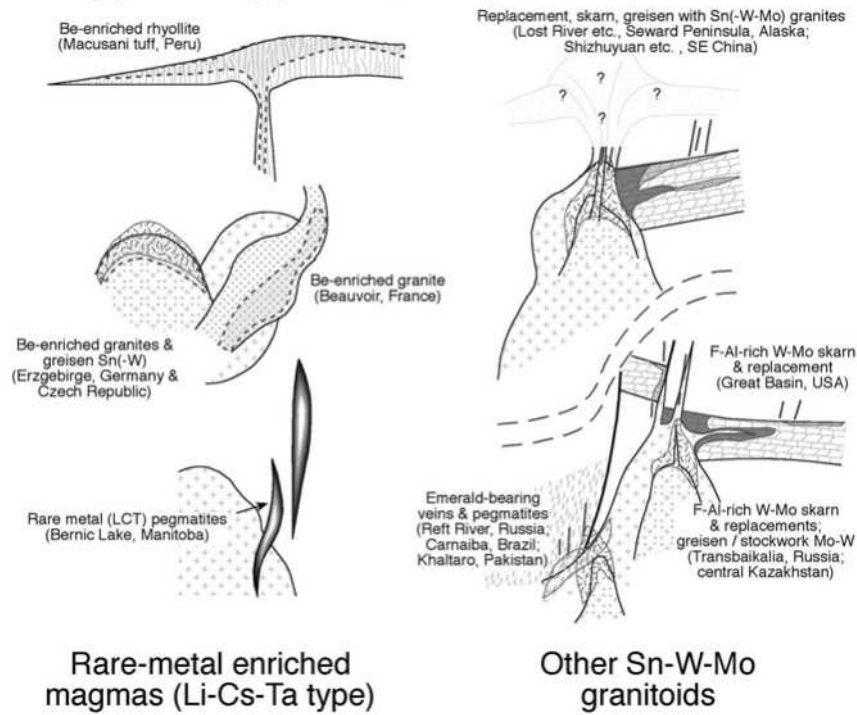


**Figure 3.** BeO concentrations and tonnage for some better documented Be-bearing mineral deposits. These are a mixture of published resource estimates and geologic inventories reflecting the sparse data available for Be occurrences. A considerable fraction in some systems likely resides as isomorphic substitutions in micas or other silicates (e.g., Beauvoir, McCullough Butte). Data and sources are summarized in Appendix A except for the pegmatite deposits (in black; Tanco: Sinclair 1996, Zavintorskoe: Kremenetsky et al. 2000a) or districts (in gray; Minas Gerais: Sinclair 1996; North Carolina tin belt: Griffiths 1954). Most Russian deposits lack tonnages, but grade and minimum sizes are given by Kremenetsky et al. (2000a). The two highest grade systems with the highest rank (size) are plotted at their minimum reported sizes (Zavitinskoe and Ermakovskoe, which are italicized). The point labeled “hypothetical Be-bearing magma” illustrates the small amount of magma required to make a world-class Be deposit compared to 100 km<sup>3</sup> or more for most other metals.

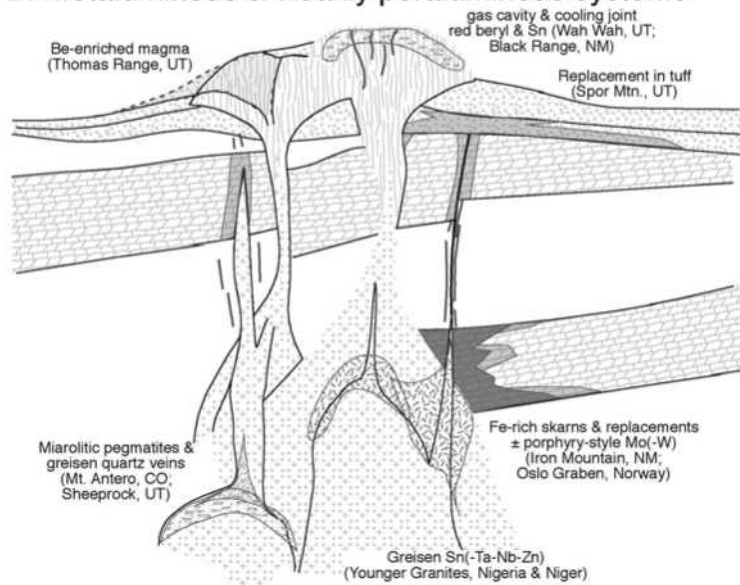
basicity and their degree of silica saturation. Emerald deposits are commonly treated as a group unto themselves (Sinkankas 1981; Snee and Kazmi 1989; Cheilletz 1998); here, we also treat them separately, but group them by origin. The text and Figure 4 are organized around this geological classification in order to emphasize mineralogical and petrological similarities, whereas Appendix A and Figure 2 are organized geographically and can serve as an index to the text via the “types” columns.

Within the igneous-related group, there is a continuum from Be-enriched magmas to complex behavior in pegmatites (London and Evensen, this volume) to the wide variety of hydrothermal deposits considered in this paper. The latter include skarns, replacement bodies, greisens and veins which form in aluminosilicate, carbonate, and ultramafic host rocks (cf. Shcherba 1970). Most non-pegmatitic accumulations form in the upper crust, typically in the upper 5 km. Mineral assemblages and compositions vary systematically with compositional variations of host rocks and related igneous rocks. Magmatic compositions are uniformly felsic but range from strongly peraluminous through metaluminous to peralkaline. Most source rocks are quartz-rich with the important exception of silica-undersaturated syenitic suites (Fig. 5A). Apart from sharing highly felsic compositions, igneous-related systems are chemically diverse (Fig. 5B). Likewise, tectonic settings are quite varied although moderately thick continental crust and late- or post-orogenic timing are common themes. It is the shared low CaO and elevated F

A. Strongly to weakly peraluminous systems

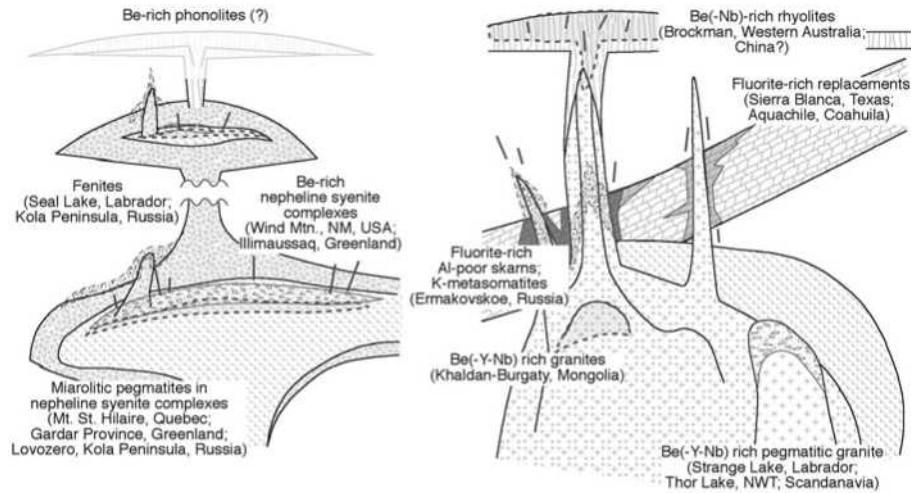


B. Metaluminous & weakly peraluminous systems



**Figure 4.** Sketches illustrating the main types of Be deposits. (A) Deposits associated with strongly peraluminous magmatism. The distinction between the Li-Cs-Ta enriched group and the others is gradational, see text for details. (B) Deposits associated with metaluminous to weakly peraluminous magmas. These rarely have strongly peraluminous and peralkaline phases.

## C. Peralkaline systems



Undersaturated  
(syenitic) systems

Peralkaline  
granites & rhyolites

## D. Non-magmatic systems



## KEY

Lithology		Be-enriched rocks	
	Clastic rocks		Magmatic enrichments (dispersed) — in minerals or volcanic glass
	Carbonate rocks		Miarolitic & pegmatic zones; Alpine-type clefts or pegmatites in metamorphic rocks
	Ultramafic rocks		High-temperature (>300°C) veins — individual, sheeted and stockwork
	Felsic gneisses & schists		Low-temperature (<300°C) veins — mostly individual or sheeted
	Felsic volcanic rocks (lavas & pyroclastic rocks)		Skarns — typically fluorine-rich, with variable Al-Fe-metal contents
	Granitic rocks (fine-grained, porphyritic, coarse-grained)		Carbonate replacement deposits — typically fluorine-rich
	Syenitic rocks (fine-grained, coarse-grained)		Fenites & related alkaline alteration — in alkaline igneous complexes

**Figure 4, continued.** Sketches illustrating the main types of Be deposits. (C) Deposits associated with peralkaline magma types. These are further divided by silica saturation into undersaturated (nepheline syenites) and oversaturated (granites and quartz syenites). (D) Non-magmatic systems of diverse origins. Examples are listed in Tables 2 and Appendix A; locations are shown in Figure 2. See text for further description and discussion.

contents, and not magma sources or other intensive variables such as oxidation state or water content, that probably favor magmatic and post-magmatic Be enrichment (Fig. 5B,C).

Traditionally, the magmas associated with rare metals (e.g., Li, Be, Nb, Ta, REE, W) have been divided into three broad groups by their associated enriched elements (e.g., Tischendorf 1977; Kovalenko 1978; Pollard 1989):

- normal (biotite  $\pm$  muscovite) granites with or without W(-Mo-F-Bi-Sn) mineralization
- Li-F rare-metal enriched granites typically with Sn-Ta(-Nb-Cs) enrichments
- peralkaline granites with associated Nb-Ta-Zr-F concentrations.

This classification does not explicitly distinguish differences in alumina saturation ( $a_{\text{Al}_2\text{O}_3}$ ) or silica saturation ( $a_{\text{SiO}_2}$ ). Given that these variables strongly influence Be and alteration mineral stability, the three traditional groups reflect neither distinct Be mineral assemblages nor distinct types of hydrothermal alteration. In light of this, we divide igneous rocks (volcanic and intrusive, including pegmatitic) and associated Be occurrences into four groups that emphasize differences in magmatic  $a_{\text{Al}_2\text{O}_3}$  and  $a_{\text{SiO}_2}$  (e.g., Shand 1927; Carmichael et al. 1974; cf. Fig. 12, below):

- strongly to weakly peraluminous suites that range from chemically non-specialized with W-Mo mineralization to Li-F-Sn so-called “specialized” granites—these have BeO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub>-H<sub>2</sub>O (“BASH”) family minerals; muscovitic hydrothermal alteration is characteristic,
- metaluminous to weakly peraluminous suites with variable Nb, Ta, F, Sn, Mo and Li enrichments—these have phenakite, bertrandite, and helvite group minerals; Li-Fe micaceous hydrothermal alteration is characteristic,
- peralkaline to metaluminous quartz-saturated suites typically with Nb-Y-F enrichments—these have phenakite, bertrandite, and Ca-Na-Be silicates; feldspathic hydrothermal alteration is characteristic, and
- silica-undersaturated, generally peralkaline suites with high Nb-REE-Y—these have Ca-Na-Be silicates and helvite group minerals; feldspathic hydrothermal alteration is characteristic.

There can be a wide-range of element enrichments (geochemical specialization) within each group. Not surprisingly, this division has parallels with Černý's classification of common and rare-metal pegmatites (Černý 1991a and Chapter 10, this volume). An advantage of using this four-part classification is that it systematizes and makes predictable the principal differences in Be mineral parageneses and alteration mineralogy. Thus it is possible, in principle, to place a deposit into one of these groups based on the mineral parageneses present. These compositional variations also broadly correlate with tectonic setting and with time as is discussed in the concluding section of this paper. In contrast, more traditional approaches that focus on depositional environment (e.g., skarn, vein, replacement, greisen etc.) do not by themselves distinguish fluid sources or broader environments.

Beryllium minerals also occur in a handful of metamorphic, sedimentary and surficial environments (Table 2). At best, these have tenuous connections to felsic magmatism. Some types, such as the Colombian emerald deposits, have distinctive basin-related hydrothermal origins, whereas others, such as some of the “shear-zone” emerald deposits likely form by local redistribution of materials during metamorphism (Grundmann and Morteani 1989). Placer accumulations are best known where coarse, Be minerals are sourced from high-grade metamorphic terrains (Rupasinghe et al. 1984;

**Figure 5 (opposite page).** Plots summarizing whole rock chemical data for selected igneous suites associated with non-pegmatitic Be deposits. Major element data are from sources cited in Appendix A and the text. (A) Total alkalis vs. silica showing fields for rock suites grouped by alumina saturation (same as in B). Compositional ranges for alkaline and subalkaline global volcanic rocks shown for comparison (Wilson 1989). (B)  $\text{Al}_2\text{O}_3$  and CaO contents normalized to  $(\text{Na}_2\text{O}+\text{K}_2\text{O}+\text{CaO})$  for Be-associated igneous suites highlighting the wide range of alkalinities and aluminum saturation index (ASI = molar  $\text{Al}_2\text{O}_3/(\text{Na}_2\text{O}+\text{K}_2\text{O}+\text{CaO})$ ) but low overall CaO. This projection shows the location of the boundaries for strongly peraluminous, weakly peraluminous, metaluminous, and peralkaline compositions while highlighting the relative CaO contents. (C) and (D) Beryllium, F and Li concentrations in glasses (Macusani, Spor Mtn., Topaz Mtn., Khaldzan-Buregtey), other volcanic rocks and intrusive rocks (data from Coats et al. 1962; Tauson et al. 1978; Christiansen et al. 1984, 1988; Černý and Meintzer 1988; Pichavant et al. 1988a; Trueman et al. 1988; Kovalenko et al. 1995b; Rimbault et al. 1995). Also shown on the right-hand side of (C) is beryl solubility at 650°C in granitic melt for ASI values of 1.0 and 1.3 (Evensen et al. 1999). Note the contrasting trends for magmatic evolution—strongly peraluminous systems evolve to Li-Cs-Ta-enriched compositions (“LCT”), whereas most other systems show more subdued rare alkali enrichment (cf. the Nb-Y-F = “NYF” mixed types of erny 1991a).

Dissanayake and Rupasinghe 1995). These commonly provide outstanding gem material (Sinkankas 1981).

Emerald deposits deserve special comment because of their economic importance and popular appeal. They form with granitic pegmatites and magmatic-hydrothermal veins of many types, by local metamorphic redistribution of materials, and in basin-related and metamorphic-derived hydrothermal systems. Like other Be deposits, no single factor controls emerald formation save for the requirement of Cr ( $\pm$  V) from local host rocks to generate their deep green color.

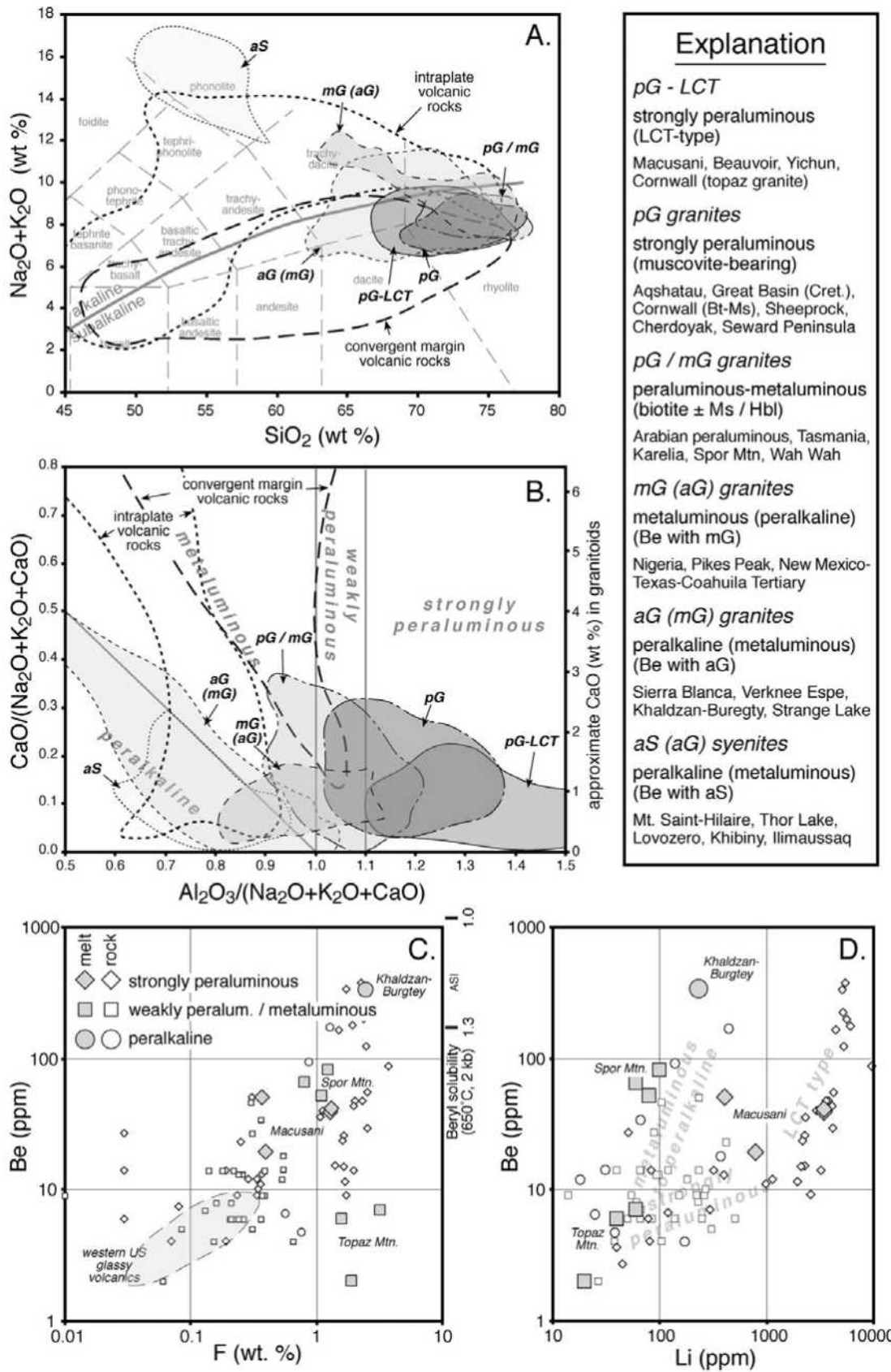
### BERYLLIUM MINERAL COMPOSITIONS

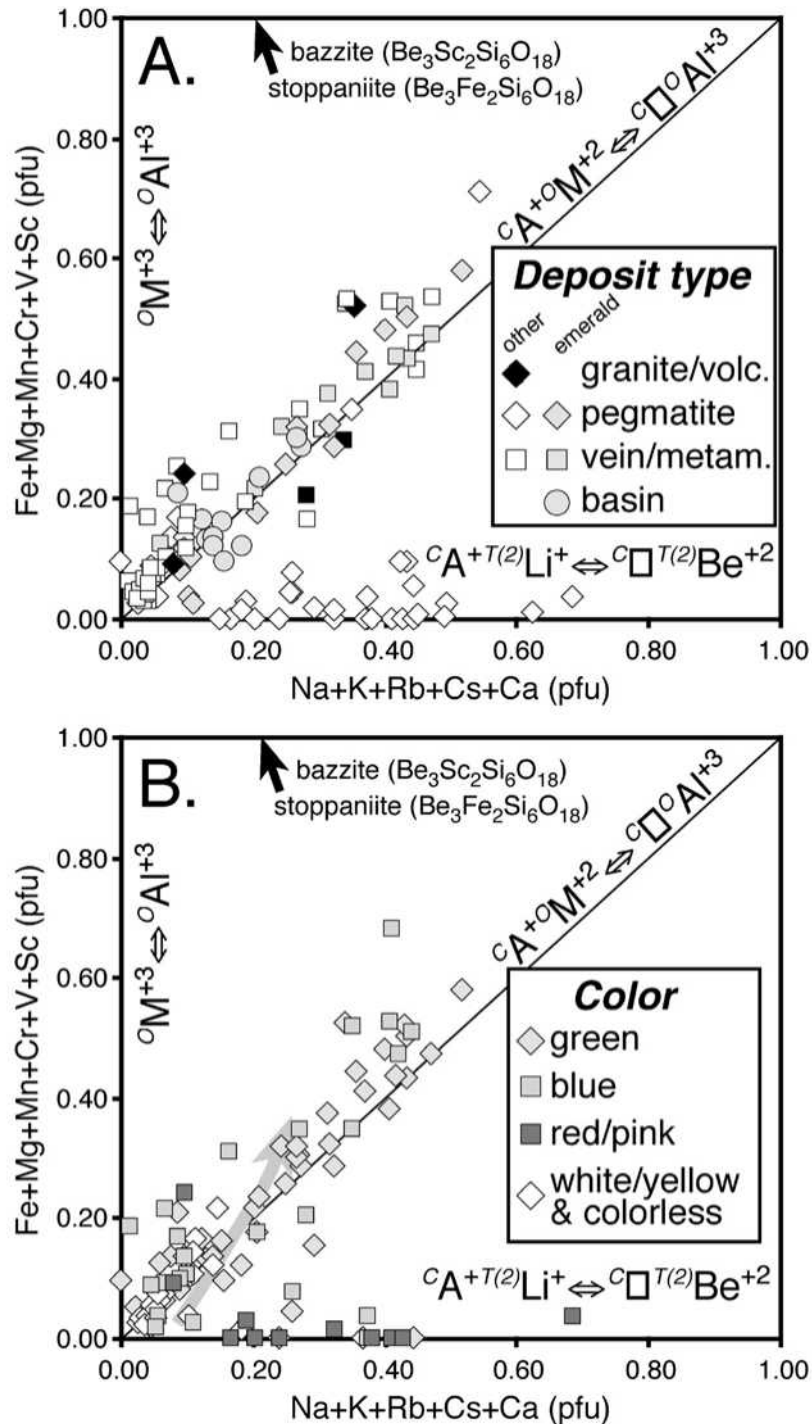
Most of the Be minerals listed in Table 1 exhibit little natural compositional variability (e.g., Chapter 1 by Grew, this volume; Chapter 10 by Černý, this volume). In non-pegmatitic occurrences, the main exceptions are the beryl group (beryl, stoppaniite and bazzite, plus structurally related milarite) and the helvite group (helvite, danalite, and genthelvite), plus minerals including the taaffeite group, the gadolinite group and meliphanite-leucophanite. Given the variably F-rich nature of Be occurrences, substitution of F for OH may be more common than appreciated even though evidence for this substitution mainly restricted to herderite, euclase and bertrandite (Beus 1966; Hsu 1983; Lebedev and Ragozina 1984; see Chapters 10 and 13, this volume by Černý and Franz and Morteani, respectively). A few other minerals such as chrysoberyl have minor, though petrologically and gemologically interesting variations in cation contents. Examination of compositional patterns in the beryl and helvite groups both documents systematic differences with environment and yields insight into differences in the conditions of formation.

**Beryl group**— $(\square, \text{Na}, \text{Cs}, \text{H}_2\text{O})(\text{Be}, \text{Li})_3(\text{Al}, \text{Sc}, \text{Fe}^{+3}, \text{Cr}, \text{Fe}^{+2}, \text{Mg})_2[\text{Si}_6\text{O}_{18}]$

**Composition.** The compositions of beryl and related minerals have long been known to vary with geologic environment (Fig. 6A; Staatz et al. 1965; Beus 1966). The principal chemical substitutions in the beryl structure,  $\square^{T(2)}\text{Be}_3\text{OAl}_2[\text{Si}_6\text{O}_{18}]^{T(1)}$ , can be represented as:

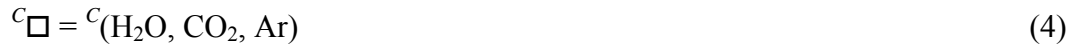






**Figure 6.** Beryl compositions plotted in terms of transition metal and alkali contents (except Li) per formula unit (6 Si). Broadly, this corresponds to octahedral and channel substitutions as noted on the diagram (following Aurisicchio et al. 1988; see Hawthorne and Huminicki, this volume). (A) Data classified by general geologic environment. Compare Figure 7. See text for discussion. (B) Data classified by color (as reported by the authors). The arrow indicates the trend from pale blue to dark green color in the Somondoco, Colombia (Kozłowski et al. 1988) and Khaltaro, Pakistan (Laurs et al. 1996) emerald localities. Many analyses including most alkali beryls have no reported color and are not plotted—most may be colorless or weakly colored. (Data compiled from Deer et al. 1978; Aurisicchio et al. 1988; Kozłowski et al. 1988; Laurs et al. 1996; Calligaro et al. 2000; S. Young and M.D. Barton, unpubl. analyses).





(Auricchio et al. 1988, Černý, this volume, Hawthorne and Huminicki, this volume). The first two coupled substitutions lead, respectively, to “octahedral” (Exchange 1) and “tetrahedral” (Exchange 2) beryls. Both are probably limited to no more than about 0.5 per formula unit (pfu) because they lead to underbonding on one of the oxygens in the beryl structure (Auricchio et al. 1988; cf. Fig. 6A). In contrast, exchange between  $\text{Al}^{+3}$  and other trivalent cations in the octahedral site (Exchange 3) can go to completion, as evidenced by the end-member minerals bazzite ( $\text{Sc}^{+3}$ ) and stoppaniite ( $\text{Fe}^{+3}$ ). Other substitutions are permissible. Li can exchange with Na and Cs in the alkali site as demonstrated by experiment (Manier-Glavinaz et al. 1989b); however, its importance in nature is unclear given that atomic Li rarely exceeds the other alkalis less divalent cations (i.e., the amount required for type 2 exchange).

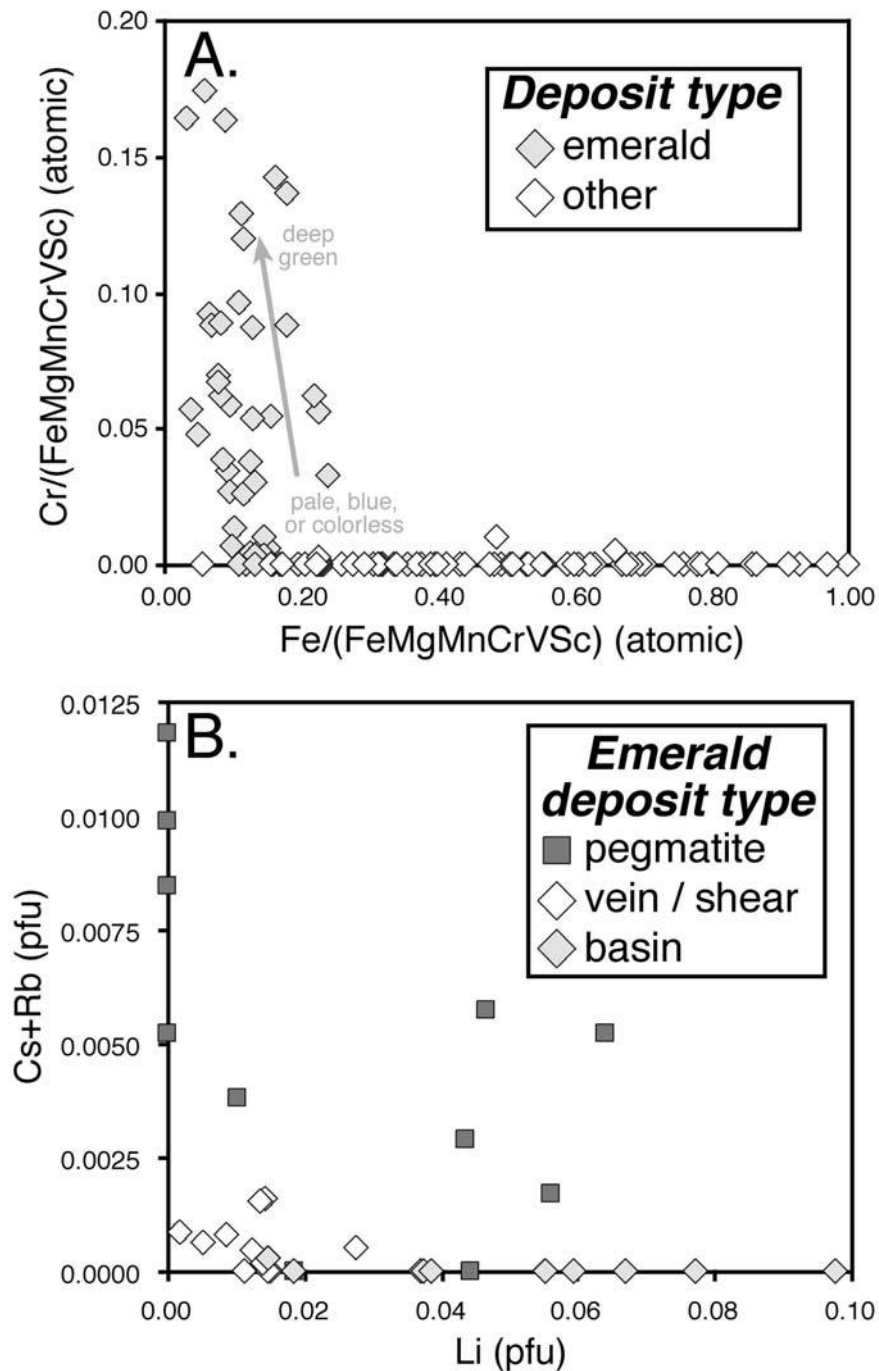
Non-pegmatitic beryls range from end-member beryl to large octahedral substitutions by both Exchanges 1 and 3. In these beryls, tetrahedral substitution is minor (Fig. 6A). In contrast, pegmatitic beryls—except for pegmatite-related emeralds—range from nearly pure compositions with at most limited type 1 exchange (<0.2 pfu) to quite high values of type 2 exchange (~0.5 pfu; Fig. 6). The most extensive type (1) substitution occurs in metamorphic-hosted beryls—both emeralds and non-emeralds (e.g., Franz et al. 1986). The more extensive type (3) substitutions, up to bazzite and stoppaniite, occur in metaluminous granites and syenites as well as in some metamorphic rocks. Channel volatile contents (dominantly  $\text{H}_2\text{O}$ ) can be virtually nil, as in volcanic red beryl (Shigley and Foord 1984), but fall mostly between ~0.6 and 2.8% in both pegmatitic and non-pegmatitic types. Other components typically reflect host rock compositions: V in sedimentary rocks, Mn in chemically evolved pegmatites and volcanic rocks, Cr and Mg in ultramafic, mafic and some sedimentary rocks, and Sc and in  $\text{Fe}^{+3}$  in A-type (mildly alkaline, oxidized) granites. Where chrysoberyl forms in the same settings, for example in desilicated pegmatites as the gem variety alexandrite, it accommodates  $\text{Fe}^{+3}$  and Cr to about the same degree as beryl.

Milarite,  $(\text{K}, \Box, \text{H}_2\text{O}, \text{Na})_2(\text{Ca}, \text{Y}, \text{REE})_2[(\text{Be}, \text{Al})_3\text{Si}_{12}\text{O}_{30}]$ , resembles beryl in having a structure of double six-membered rings interconnected by Be tetrahedra and Ca octahedra (Hawthorne and Huminicki, this volume). As in beryl, alkalis and water can substitute in channels which in milarite are defined by stacking of the double rings. Milarite occurs in skarns, alpine veins and various alkaline-related metasomatic rocks (Appendix A) as well as in various types of pegmatites (Černý, Chapter 10, this volume). Compositional variations of milarite are sparsely documented, but the (Y, REE)-rich varieties appear to be more common in alkaline settings (cf. Černý).

**Color and composition.** Not surprisingly, transition-metal-rich, octahedrally substituted beryls typically have more intense colors, mostly blues or greens, although red is characteristic of volcanic-hosted beryl (Fig. 6B). Pegmatitic beryls can be intensely colored (e.g., aquamarine and emerald), however, most tetrahedrally substituted beryls, if not colorless, tend to be pale in color, typically pink, less commonly yellow, green or blue.

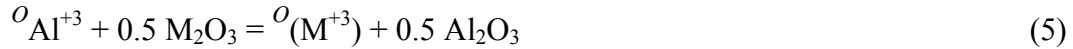
In emerald, the intense green color reflects substitution of  $\text{Cr}^{+3}$  for  $\text{Al}^{+3}$  and a paucity of Fe (Fig. 7A) regardless of setting, whereas the rare alkali content does reflect their environment of origin (Fig. 7B). The latter is true in spite of the fact that all emeralds are dominated by the octahedral substitution (Fig. 6A). Given these patterns and the great interest in emeralds in the gem trade, it is obvious why chemical fingerprinting of emerald provenance has been pursued with some vigor and success (e.g., Dereppe et al.

2000). Emerald is properly restricted to beryl where Cr exceeds other coloring agents by weight (Kazmi and Snee 1989a). The analogous substitution of  $V^{+3}$  in beryl also creates an intense green coloration that is often termed emerald. Even the deep red Mn-rich volcanic-hosted beryl from Utah has been marketed, controversially, as “red emerald” (Spendlove 1992).

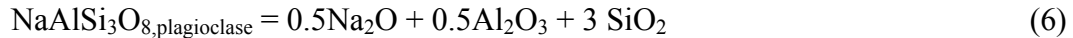


**Figure 7.** Emerald and other beryl compositions from the literature (see Figure 6 for sources). (A) Plot illustrating the elevated Cr contents and low Fe to Mg (etc.) ratios of emeralds compared to other types of beryls. This illustrates the main difference with other environments. Cr is not reported in many of the other analyses; it may have either been below detection or not sought. As in Figure 6, the arrow shows the trend from pale blue to dark green colored beryls at Somondoco, Colombia (basin-related) and Khaltaro, Pakistan (pegmatite). (B) Plot of rare alkalis in emeralds from various settings illustrating variations analogous to those seen in other beryls. See text for discussion.

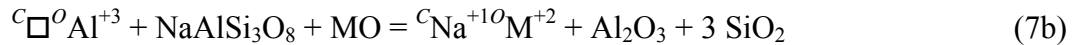
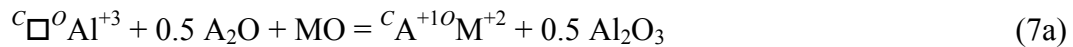
**Petrologic controls on beryl composition.** A simple analysis of the common substitutions in terms of alumina activity ( $a_{\text{Al}_2\text{O}_3}$ ) and the availability of other cations helps rationalize their correlation with geologic environment. In the simplest case, illustrated by equation 5, the type 3 substitution of trivalent Cr, Fe, V and Sc for Al will be promoted by the relative abundance of these elements in certain rocks or by decreasing  $a_{\text{Al}_2\text{O}_3}$ . Alumina



activity will be low in aluminum-deficient assemblages (e.g., many ultramafic and carbonate rocks) and in alkaline igneous rocks. Reaction (6) shows that alkalinity and alumina activity inversely correlate in feldspar-bearing rocks:

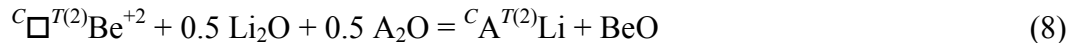


Similarly, any combination of decreasing  $a_{\text{Al}_2\text{O}_3}$ , increasing alkalinity, or increasing availability of (Mg, Fe, Mn)O will promote type 1 (octahedral) substitution:



Thus, as observed, beryl group minerals forming in metaluminous igneous rocks and in ultramafic or carbonate host rocks should generally have higher octahedral substitutions than beryls from peraluminous varieties. For example, emerald and green vanadian beryls are most common in rocks lacking muscovite (e.g., Kazmi and Snee 1989b). Ferric-iron-rich aquamarines, the  $\text{Fe}^{+3}$  end member stoppaniite, and the  $\text{Sc}^{+3}$  end member bazzite are most typical of metaluminous rocks—biotite granites or, in the case of stoppaniite, syenite (Ferraris et al. 1998; Della Ventura et al. 2000). Conversely, in some circumstances Fe contents may be suppressed either by intrinsically low Fe relative to other octahedral cations (as in ultramafic rocks) or by sequestration in other phases (e.g., pyrite in the Colombian emerald deposits, Ottaway et al. 1994).

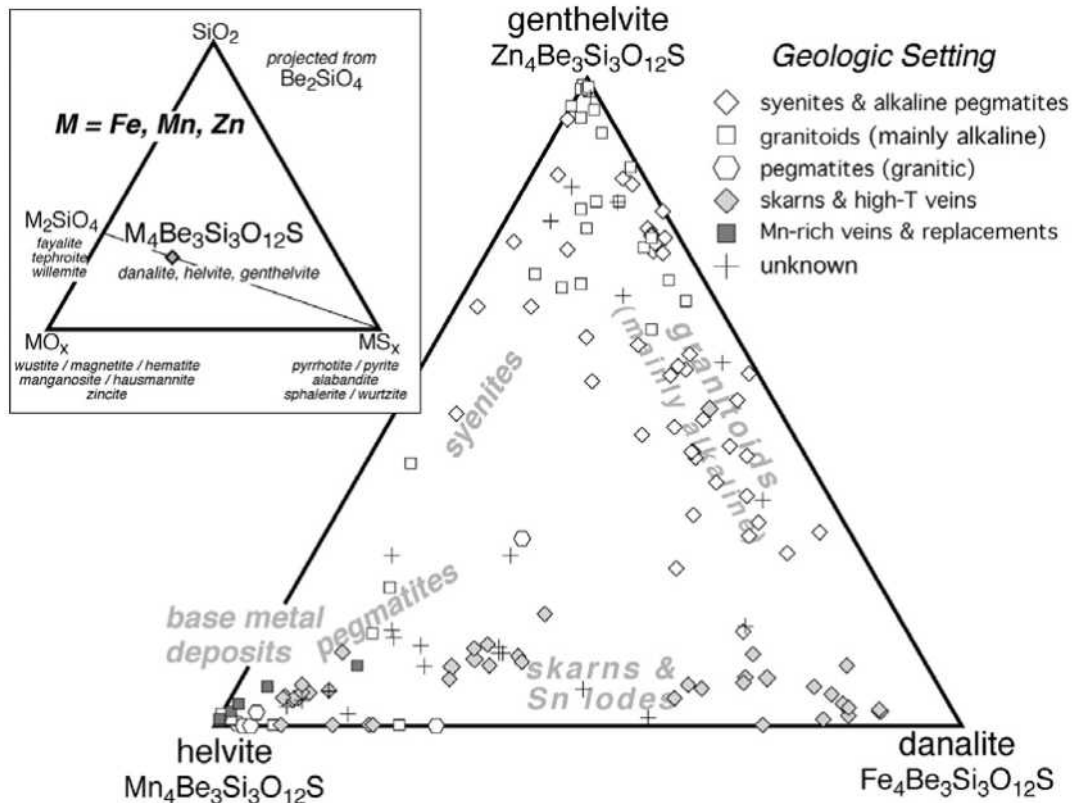
The tetrahedral (type 2) substitution is common in Li-Cs-Ta pegmatites, but apparently is rare elsewhere. It logically follows Reaction (8) where availability of Li or Cs is the key.



Increasing overall alkalinity (reaction 6) is not likely to be a factor given that Li-Cs-Ta pegmatites are strongly peraluminous (Černý 1991a), but it could contribute to tetrahedral substitution in some mildly alkaline greisen-type systems. Unfortunately very few complete beryl analyses are available for the latter. One might expect octahedral substitutions to accompany the tetrahedral except for the fact that highly evolved pegmatites with high Li and Cs have very low contents of Mg and Fe and only modest Mn. This may contribute to the separation of the field for tetrahedrally substituted beryls from the other occurrences in Figure 6A.

### Helvite group—(Mn,Fe,Zn)<sub>4</sub>[BeSiO<sub>4</sub>]<sub>3</sub>S

**Composition.** Helvite-group minerals are present in minor quantities in Be-bearing skarns, alkaline igneous settings, and some hydrothermal veins. Changes in Mn-Fe-Zn ratios spanning all three end-members account for most of compositional variation in the helvite group (Fig. 8). Rarely, Al substitutes for Zn; Finch (1990) proposed that the mechanism is  $2 \text{Al}^{+3} + \Box = 3 \text{Zn}^{+2}$  based on compositional variations in hydrothermal genthelvite from the syenitic Motzfeldt intrusion, Greenland which contains to ~10 wt %  $\text{Al}_2\text{O}_3$ . Other elements might be present, for example Na given the structural similarity with tugtupite ( $\text{Na}_4[\text{BeSiO}_4]_3\text{Cl}$ ), or Cd where genthelvite coexists with greenockite (Nechaev and Buchinskaya 1993).

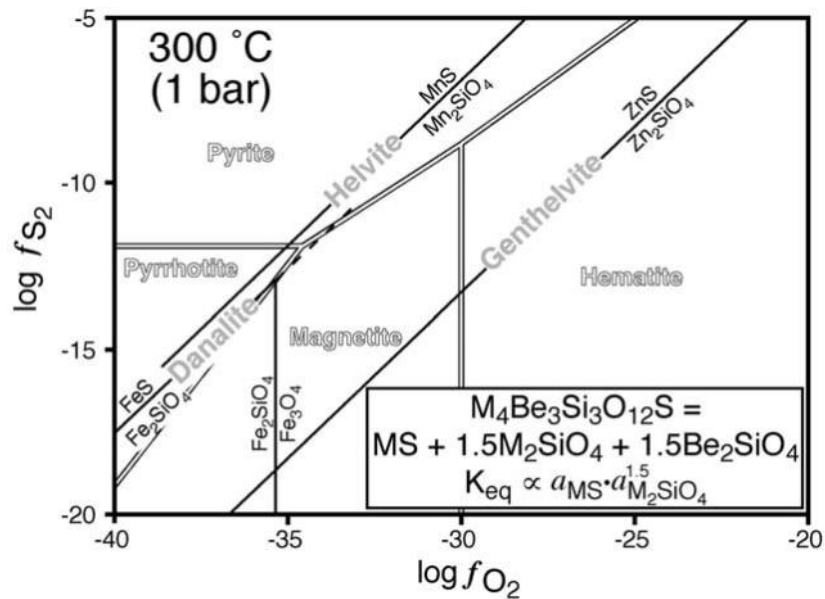


**Figure 8.** Analyzed helvite group minerals plotted in terms of the end member compositions and distinguished by geologic environment (sources of data include: Vlasov 1966b; Dunn 1976; Kwak and Jackson 1986; Larsen 1988; Perez et al. 1990; Ragu 1994a). The inset shows the chemographic relationship of helvite group minerals to silica, phenakite, and Mn-Fe-Zn oxides, sulfides and silicates.

As illustrated in Figure 8, helvite-group compositions differ systematically between genetic environments. Zinc-rich compositions (genthelvitite) with or without Al typically occur in pegmatites, miarolitic cavities or veins associated with metaluminous to peralkaline granites and syenites (Burt 1988; Larsen 1988; Perez et al. 1990). Peraluminous granitic pegmatites and occurrences in base-metal-sulfide veins and replacements are typically Mn-dominated, whereas variable Fe:Mn varies from near end-member danalite to helvite in skarns and Sn lodes (greisens), with danalite being dominant common in the more reduced systems (Burt 1980; Kwak and Jackson 1986).

**Petrologic controls on helvite-group compositions.** The unusual composition of the helvite group—combining  $\text{Be}_2\text{SiO}_4$ , a metal sulfide, and a metal orthosilicate (Fig. 8 inset)—means that these minerals are sensitive to redox and sulfidation states as well as to the activity of phenakite (Burt 1980, 1988). Conditions favorable for formation of the various end-members differ based on the relative stability of the related sulfides and silicates as illustrated in Figure 9. For each of the three, maximum stability occurs along the boundaries where their respective orthosilicates and monosulfides coexist along with phenakite. Departure from the ideal conditions by oxidation, reduction, gain or loss of sulfur, or reducing the activity of phenakite will all be unfavorable. Hence, low  $a_{\text{Al}_2\text{O}_3}$  (“alkaline” conditions) favor helvite group minerals because beryl replaces phenakite and lowers  $a_{\text{Be}_2\text{SiO}_4}$  with increasing  $a_{\text{Al}_2\text{O}_3}$  (see next section). Danalite preferentially occurs in reduced and low sulfidation state environments; helvite dominates in more sulfidized,

Mn-rich settings where pyrite and sphalerite sequester Fe and Zn; and genthelvite is restricted to relatively oxidized but low sulfur settings characteristic of many (per)alkaline rocks where Fe and Mn mainly enter oxides and other silicates (cf. Burt 1980, 1988).



**Figure 9.** Helvite group mineral stability a function of oxidation and sulfidation state relative to some other zinc, iron and manganese minerals. End members should have maximum stabilities on the orthosilicate- monosulfide boundaries (inset; also see Fig. 8 inset). Note that maximum stability for danalite would project along the dashed line were it not for magnetite formation. Calculated using thermodynamic data from Barton and Skinner (1979) and Robie et al. (1978).

### Other minerals

Gadolinite group minerals,  $(Y,REE)_2(Fe,\square)[Be_2Si_2O_8](O,OH)_2$ , leucophanite,  $CaNaBeSi_2O_6F$ , and meliphanite,  $Ca_4(Na,Ca)_4Be_4AlSi_7O_{24}(F,O)_4$ , occur mainly in alkaline or metaluminous pegmatites or miarolitic cavities but are also found in a handful of alkaline-rock related hydrothermal deposits (Table 1, Appendix A). Little is published about gadolinite-group compositions in non-pegmatitic occurrences. Based on the study of Pezzotta et al. (1999) who studied a range of granite-related occurrences in the southern Alps, considerable variation in Y / LREE / HREE would be expected as well as variable B contents. Leucophanite and meliphanite solid solutions are reported from alkaline metasomatites (Ganzeeva et al. 1973; Novikova 1984) presumably reflecting differences in Ca/Na.

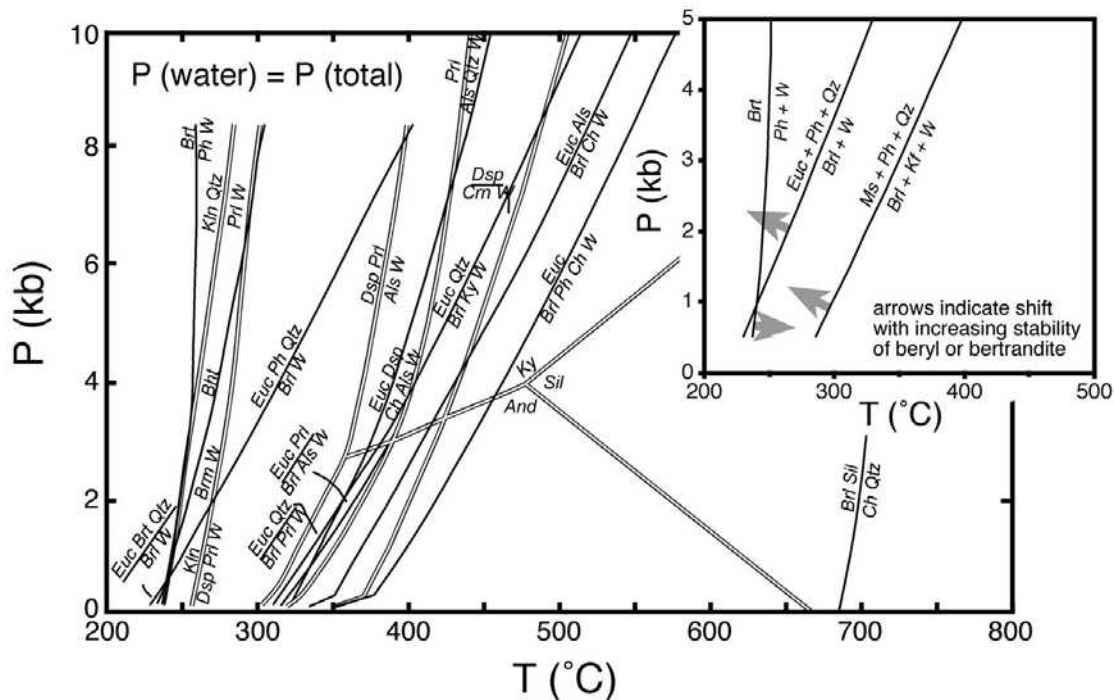
## BERYLLIUM MINERAL STABILITIES

Available data on beryllium mineral stabilities, derived from experiment, theory and natural assemblages, provides a valuable framework for classification and understanding of natural occurrences. Published studies on Be mineral stabilities are summarized in Appendix B and have been reviewed extensively elsewhere (Barton 1986; Burt 1988; Wood 1992; Franz and Morteani, London and Evensen, Chapters 13 and 11, respectively, this volume). Most of this work has focused on the BeO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub>-H<sub>2</sub>O (BASH) system and coexisting melts and aqueous fluids. Here we briefly review mineral equilibria and solubilities of particular relevance to non-pegmatitic deposits and focused on BASH minerals.

The figures presented here were calculated using the internally consistent thermodynamic model for BASH phases and topaz from Barton (1982b, 1986), which were adapted to the SUPCRT database (Johnson et al., 1992) by adjusting for differences in the enthalpy of formation of  $\text{Al}_2\text{O}_3$  between the databases, and refitting the heat capacities to the Meier-Kelly function. Presently, there is a need to reevaluate the thermodynamic data for BASH minerals by including results published since 1985 (Appendix B) in a rigorous fit. In addition, one could also build a thermodynamic model for other phases, such as the helvite group and the Na-Be silicates, by combining available experimental data with constraints from natural assemblages.

### Pressure-temperature-activity relationships

**P-T.** Other than for the BASH system there are essentially no reversed equilibrium data for the pressure-temperature stability fields of Be minerals (Appendix B). In the BASH system, the salient characteristics of pressure-temperature phase relationships (Fig. 10) are (1) that the hydrous minerals (excepting beryl) are stable only at temperatures below  $500^\circ\text{C}$  and (2) that the assemblages are not distinctly pressure sensitive. Bertrandite persists only up to about  $300^\circ\text{C}$ . The lower limit of beryl stability is between  $200$  and  $350^\circ\text{C}$  depending on coexisting minerals (Fig. 10 inset). In quartz-bearing assemblages, chrysoberyl is restricted to near-magmatic and higher temperatures, although the position of the reaction chrysoberyl+quartz = beryl+aluminum silicate is sensitive to beryl composition and its position remains controversial. See Barton (1986) and Franz and Morteani (this volume) for further discussion of these relationships.



**Figure 10.** Pressure-temperature projection of phase relationships in the  $\text{BeO-Al}_2\text{O}_3\text{-SiO}_2\text{-H}_2\text{O}$  (BASH) system. Redrawn from Barton (1986). Limiting reactions for bertrandite and beryl both can depend on solid solution effects, F for OH in bertrandite, and multiple components in beryl (inset).

**T-activity** In contrast to the limited insight available from the P-T relationships, activity diagrams are of considerably greater utility in understanding the occurrence of Be minerals because of the metasomatic origin of most non-pegmatitic Be deposits (Figs. 9, 11-14). The most useful independent variables are: (1) temperature, which varies

markedly in time and space in most Be-bearing geologic systems, and (2) the activities of the major components, notably alumina and silica. Silica and alumina are key because they frame the thermodynamic conditions defined by many rock-forming minerals and, in addition, can be related to alkalinity of melts and fluids through reactions (6) and (9). Reaction (9) relates fluid acidity to alkalinity in the presence of plagioclase when  $a_{\text{Al}_2\text{O}_3}$  and  $a_{\text{SiO}_2}$  are defined.

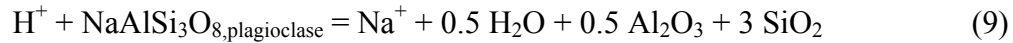


Figure 11 plots BASH mineral assemblages in terms of each  $a_{\text{SiO}_2}$  and  $a_{\text{Al}_2\text{O}_3}$  as functions of temperature. At high T, beryl, phenakite, and chrysoberyl ( $T > 600^\circ\text{C}$ ) are stable at high silica activities (Fig. 11A,C). With decreasing silica activity beryl is replaced by chrysoberyl+phenakite and phenakite is ultimately replaced by bromellite. This is the characteristic sequence found in desilicated pegmatites. A similar progression occurs at lower temperatures except that chrysoberyl is strongly quartz undersaturated and first euclase and then bertrandite become key phases. Skarns and carbonate hosted replacement bodies typically exhibit zoning that reflects these varying degrees of silica saturation and paths from high- to low-temperature across Figure 11A. At  $\leq 1$  kbar solutions can become strongly undersaturated with respect to quartz, whereas at higher pressures they may stay closer to quartz saturation (Fig. 11A inset). These contrasting paths rationalize differences observed in carbonate-hosted hydrothermal systems.

Another useful contrast comes from consideration of  $a_{\text{Al}_2\text{O}_3}$ , a variable which highlights differences between Al-rich and Al-poor assemblages (Fig. 11B). The saturation surface for the Al-only phases, corundum ( $T > 360^\circ\text{C}$ ) and diaspore ( $T < 60^\circ\text{C}$ ), neither of which is stable with quartz, bounds the top of the diagram. Quartz coexists with andalusite at high temperature, but then pyrophyllite followed by kaolinite formed with decreasing temperature. Chrysoberyl and euclase are the characteristic minerals at high  $a_{\text{Al}_2\text{O}_3}$ , whereas beryl occupies an intermediate field (Fig. 11B). In contrast, phenakite and bertrandite are stable only at distinctly lower  $a_{\text{Al}_2\text{O}_3}$  conditions until bertrandite and kaolinite become stable together at about  $225^\circ\text{C}$ . A key boundary is that between K-feldspar and muscovite which separates strongly peraluminous assemblages from others. Considering this reaction, it becomes clear why in most quartz-bearing rocks, beryl is the dominant silicate down to relatively low temperatures barring conditions of unusual acidity (as in some greisens) or basicity (as in peralkaline rocks). On cooling in the presence of muscovite and K-feldspar, only below  $T \approx 300^\circ\text{C}$  does beryl give way to phenakite+quartz (arrow in Fig. 11B). Solid solution will expand the beryl field to still lower temperatures (Fig. 10 inset).

Odintsova (1993) derived an analogous topology as a function of  $a_{\text{BeO}}$  and temperature. She subsequently use it to interpret the paragenesis of ultramafic-hosted emerald deposits in the Ural Mountains (Odintsova 1996). Because BeO is rarely more than a minor component, most assemblages will only have a single saturating Be phase, thus relationships among Be-bearing mineral assemblages are more readily applied when cast in terms of other components.

**Activityactivity** Projecting the variables from Figure 11 into  $a_{\text{Al}_2\text{O}_3}$  -  $a_{\text{SiO}_2}$  space (Fig. 12) provides a particularly revealing look at Be mineral assemblages because reactions among rock-forming minerals separate major rock types on the same diagrams. In Figure 12, quartz-saturated rocks (granitoids, rhyolites, etc.) lie along the top of the diagrams passing downward into undersaturated rocks. The latter are split by key reactions such as  $\text{Mg}_2\text{SiO}_4 + \text{SiO}_2 = \text{Mg}_2\text{Si}_2\text{O}_6$ . Saturation with muscovite and andalusite occurs along the right boundary, defining strongly peraluminous rocks, whereas peralkaline assemblages (and rocks) are located near the acmite-bearing reaction that passes diagonally across the left half of the diagram.

**Figure 11 (opposite page).** (A) Beryllium mineral stability as a function of temperature and concentration of aqueous silica at 1000 bars. Shading outlines the upper  $a_{\text{SiO}_2}$  limit of chrysoberyl, the lower  $a_{\text{SiO}_2}$  limit of beryl, and the upper thermal stability of euclase. The inset shows aqueous silica concentrations at two pressures and 3 alternative fluid paths. Path (a) represents cooling with little decompression and would remain quartz saturated; path (b) represents decompression, whereas path (c) represent isobaric cooling but quartz undersaturation because of the low-P retrograde solubility of quartz. See text for additional discussion. (B) Beryllium mineral stability as a function of temperature and the activity of alumina (corundum) at 1000 bars. Shading outlines stability limits for beryl, the upper  $a_{\text{Al}_2\text{O}_3}$  limit for phenakite/bertrandite, and the upper thermal stability of euclase. Note the arrow and label for the lower limit of stability for beryl in the presence of K-feldspar (cf. inset in Fig. 10). (C) BASH mineral compatibilities at 225, 350 and 500°C projected from H<sub>2</sub>O onto the BeO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub> plane (cf. Fig. 1). Mineral abbreviations from Table 1. (A) and (B) are modified from Barton (1986).

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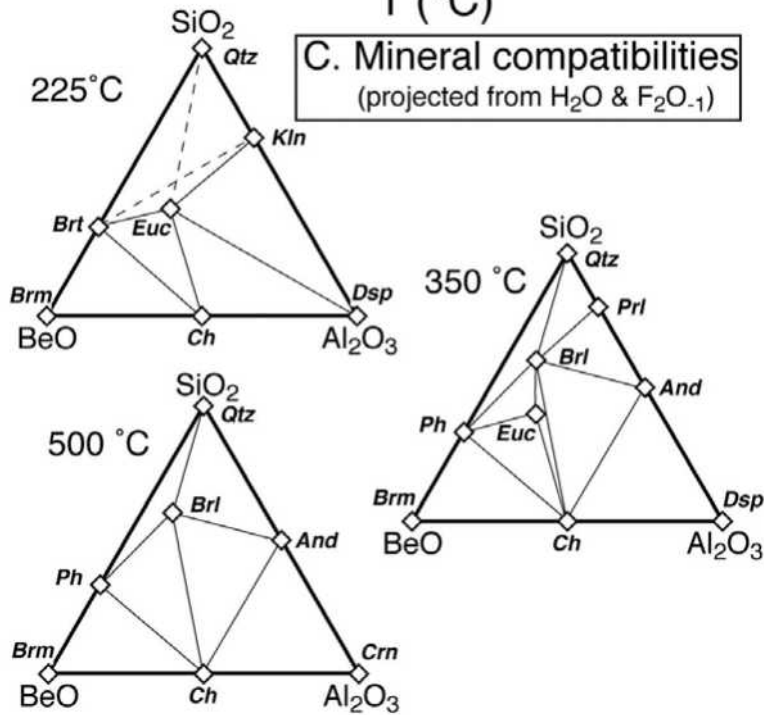
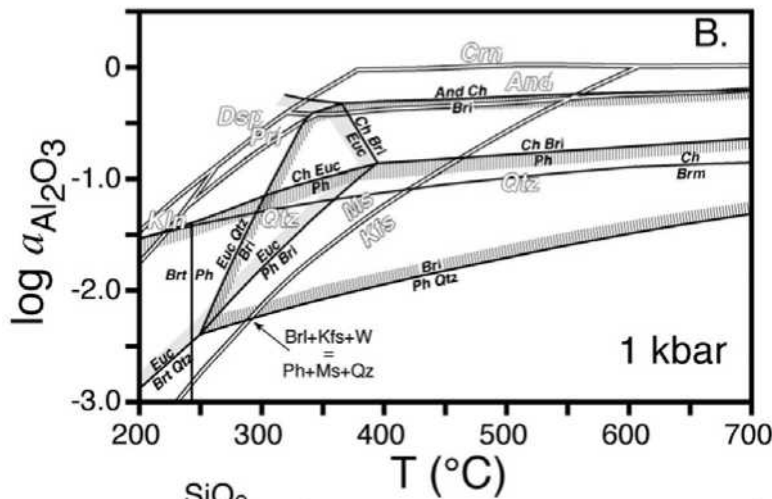
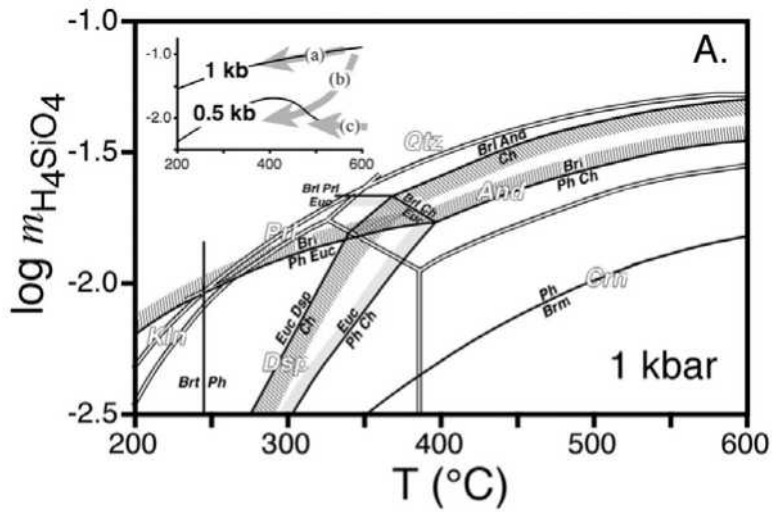
By examination of the superimposed Be mineral stability boundaries at 600°C (Fig 12A), it is clear why beryl is typical of strongly peraluminous granitoids and rocks, why phenakite ( $\pm$  helvite group) is common in metaluminous and peralkaline rocks, and why alkali Be silicates occur in peralkaline silica-undersaturated rocks. Chrysoberyl has a large stability field but only for unusual rocks that must be Al-rich and Si-poor (e.g., desilicated pegmatites). With decreasing temperature, the fields for euclase and, especially, the Al-free Be silicates expand at the expense of the beryl and chrysoberyl fields. Topologically-correct phase boundaries for beryllite, epididymite and chkalovite are shown in the upper left based on occurrences in peralkaline syenites. The breadth of the phenakite/bertrandite fields is consistent with widespread occurrence of these minerals in low-temperature deposits, particularly carbonate replacements. The right hand side matches assemblages found in strongly peraluminous igneous-hosted greisens (top) and in silica-undersaturated greisens developed in carbonate rocks (right; the meaning of greisen is discussed below).

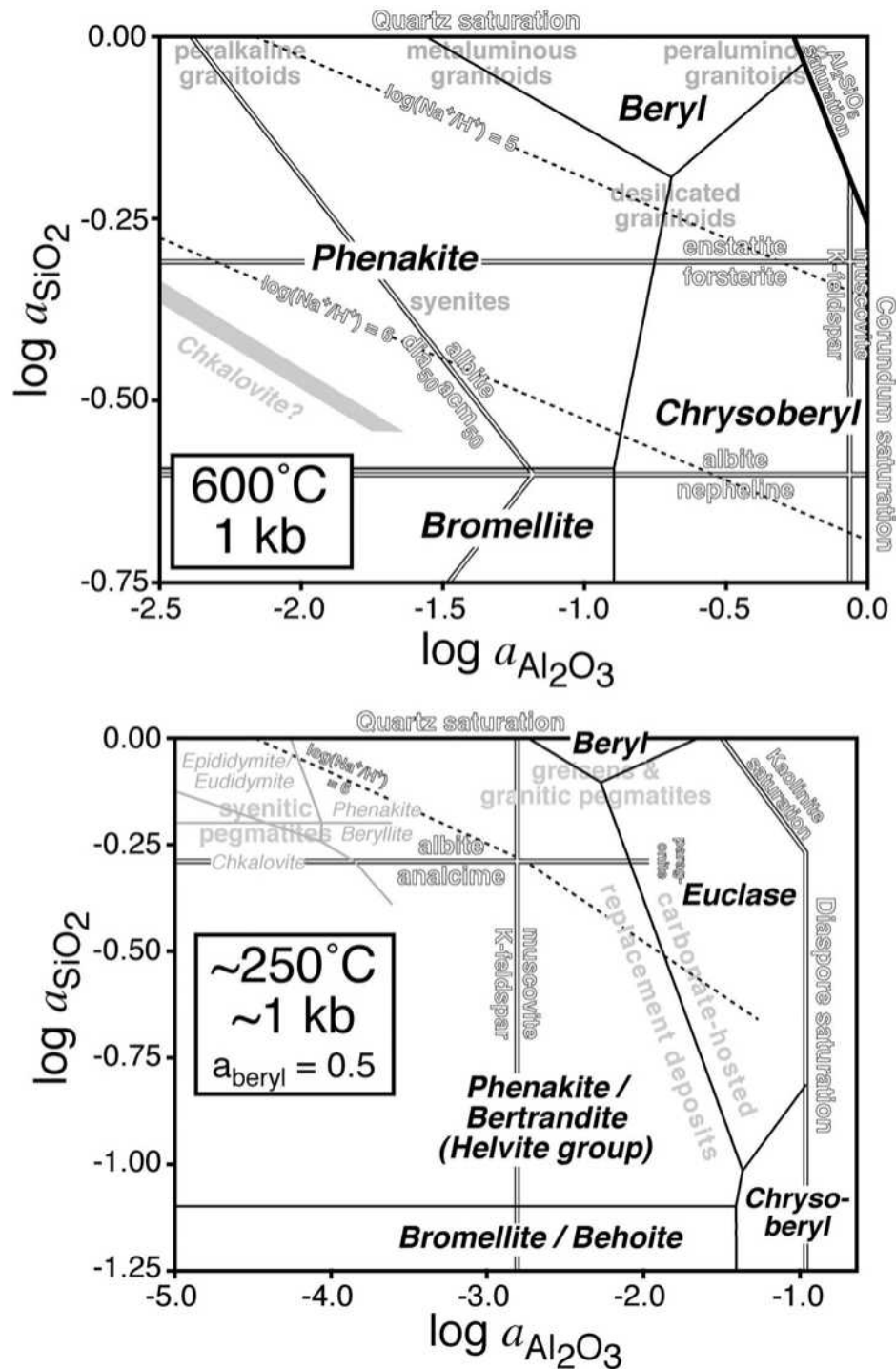
Activity relationships in terms of other components are germane to a number of occurrences, particularly HF, CaO, MgO and P<sub>2</sub>O<sub>5</sub>. Increasing the activity of acid fluoride species leads to topaz replacing other Al-bearing silicates and fluorite replacing other Ca-bearing minerals—these are typical minerals of greisens (Burt 1975, 1981). Phenakite and bertrandite replace beryl and euclase with increasing HF as well as with increasing alkalinity (e.g.,  $\text{K}^+/\text{H}^+$ , see Fig. 13, cf. Fig. 11B) consistent with their widespread occurrence in greisens of various flavors. Fluorine has a similar role in Ca-bearing rocks, where fluorite formation sequesters Ca and leads to more acid (Al-dominated) mineral assemblages. This was considered by Burt (1975) who used natural assemblages to derive topologies for activity diagrams involving P<sub>2</sub>O<sub>5</sub>, CaO and F<sub>2</sub>O<sub>1</sub> and analyze the relationships between beryl, phenakite and various Be phosphates.

Beryllium mineral parageneses in the Ca-Mg silicate assemblages of ultramafic and carbonate hosted deposits can also be usefully visualized by recasting phase relationships in terms of the activities of CaO, MgO and SiO<sub>2</sub>. For example, Figure 14 illustrates possible phase relationships and zoning paths in desilicated pegmatites or quartz-feldspar veins at 500°C and 3 kbar. Starting with a granitic/vein assemblage on the high-silica side, paths can go upward (as in a dolomitic limestone) into the actinolite (or clinopyroxene) field and yield zoning from beryl to chrysoberyl to phenakite or downward into phenakite and ultimately bromellite. Under these particular conditions, beryl is near its stability limit (Fig. 14 inset) and small differences in solid solution can have significant differences in the position of phase boundaries and thus paths.

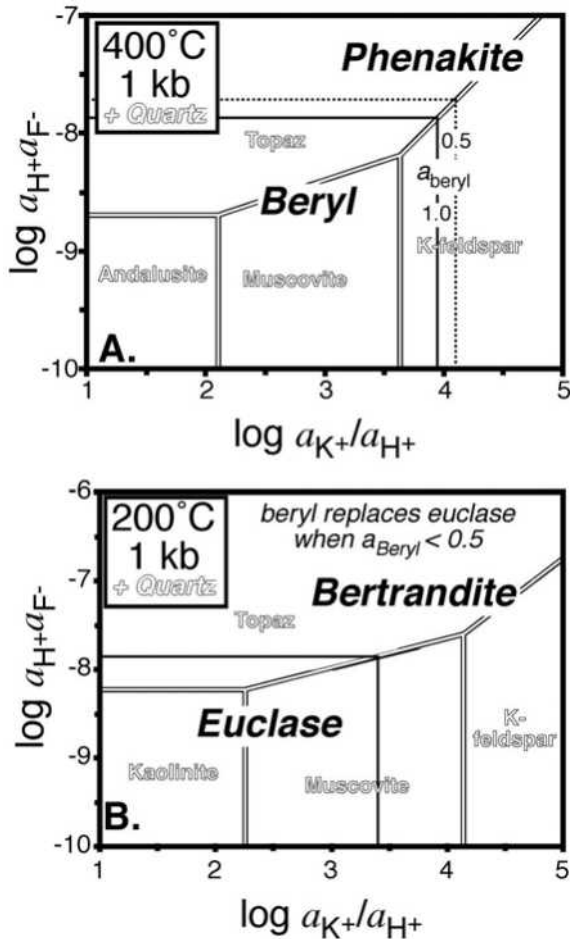
**The meaning of greisen.** Many Be-bearing rocks are referred to as greisen, which refers to a broad spectrum of Al-bearing metasomatic rocks that are typically F-rich and



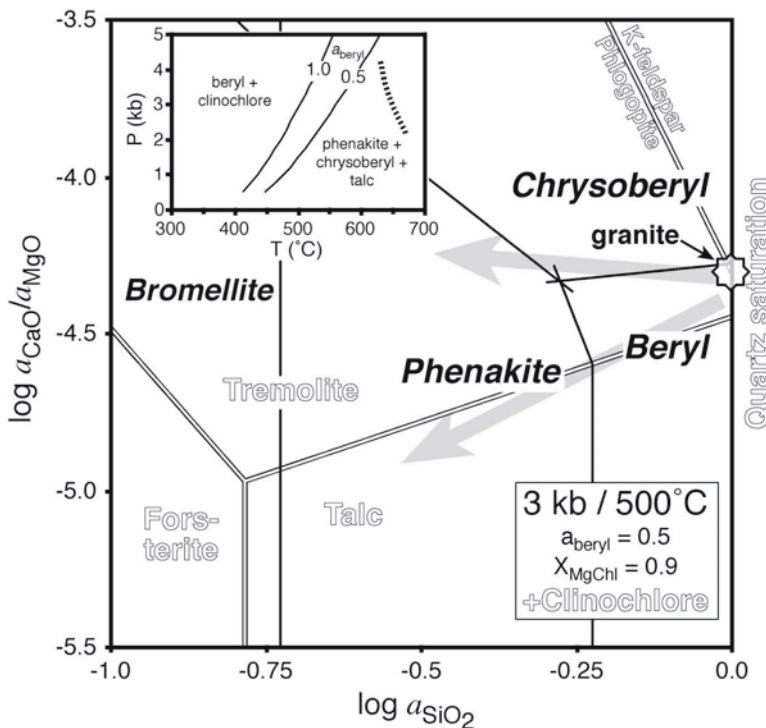




**Figure 12.** Beryllium mineral stabilities as a function of silica and alumina activities. The diagrams illustrate the preeminent control that these rock-defined variables have on mineral assemblages in Be-bearing hydrothermal systems. (A) Phase relationships of Be minerals as a function  $a_{\text{Al}_2\text{O}_3}$  and  $a_{\text{SiO}_2}$  at 600°C and 1 kbar related to mineralogy in felsic igneous rocks. The field for chkalovite stability is speculative although topologically plausible and is consistent with the recent work by Markl (2001). (B) Phase relationships of Be minerals as a function  $a_{\text{Al}_2\text{O}_3}$  and  $a_{\text{SiO}_2}$  at 250°C and 1 kbar related to mineralogy in felsic igneous rocks and some major groups of Be deposits. The activity of beryl = 0.5. The speculative fields for beryllite, chkalovite, epididymite / eudidymite are consistent with their chemography and mineral associations reported from alkaline syenitic pegmatites (stable with albite and analcime).

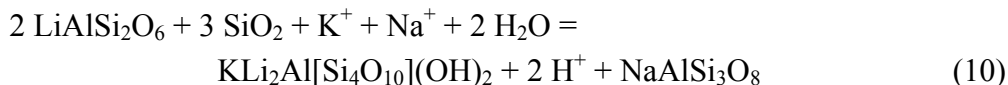


**Figure 13.** Calculated Be mineral stability as a function of H·F and K/H at 1000 bars and temperature of 400°C (A) and 200°C (B). The P-T dependence of limiting reactions is shown in the inset in Figure 10. Similar topologies are discussed by Kupriyanova et al. (1982) and Burt (1981).



**Figure 14.** One topology for phase relationships of Be minerals in ultramafic-hosted deposits at 500°C and 3 kbar. The arrows represent two alternative evolutionary paths that are discussed in the text. Chlorite is present throughout and activities of beryl and clinocllore are reduced. The inset shows a schematic water-saturated granite solidus and the calculated position of the dehydration reaction for  $\text{beryl} + \text{chlorite} = \text{phenakite} + \text{chrysoberyl} + \text{talc}$ , demonstrating that on the activity diagram the field of beryl will expand significantly at lower temperatures, helping to account for the scarcity of phenakite and chrysoberyl in ultramafic-hosted occurrences.

commonly contain one or more newly formed mica group minerals (Shcherba 1970; Burt 1981; Kotlyar et al. 1995). Greisen is most common in feldspathic host rocks, but it is described in many protoliths including carbonate and ultramafic rocks (“apocarbonate” and “apoultramafic” greisens, respectively, see Shcherba 1970). This traditional, broad definition lacks the mineralogical specificity to be petrologically useful. In this paper, rather than restrict the long-ingrained usage, we simply focus on the mineral assemblages and note their implications for intensive variable such as  $a_{\text{Al}_2\text{O}_3}$  or acidity. For example, whereas a beryl-bearing topaz-quartz-muscovite greisen is intrinsically acid (Fig. 13; e.g., from Aqshatau, Kazakhstan), a phenakite-bearing polyolithionite greisen (e.g., from Thor Lake, Canada) is intrinsically alkaline compared to assemblages containing spodumene, as demonstrated by Reaction (10):



### Solubility relationships

Another requirement in understanding Be occurrences is the behavior of Be in fluids—aqueous solutions and silicate melts. Although few experimental data exist (Appendix B; London and Evensen, this volume), the principal results merit comment here because they yield useful insight into the processes and patterns in non-pegmatitic deposits.

**Aqueous fluids.** BeO is only sparingly soluble in pure water, however Be compounds with  $\text{F}^-$ ,  $\text{CO}_3^{2-}$ ,  $\text{Cl}^-$  and  $\text{SO}_4^{2-}$  are all significantly soluble (or decompose) in water at room temperature. These potential ligands plus  $\text{OH}^-$  have received some attention from experimentalists, although not necessarily in experiments designed to yield thermodynamic data (Appendix B). The nearly ubiquitous association of F-bearing minerals with Be deposits has led many investigators to postulate that complexing by  $\text{F}^-$  is important (Beus 1966). A few others have advocated other complexes, particularly for those deposits where F is apparently absent and other potential ligands such as  $\text{CO}_3^{2-}$  or  $\text{SO}_4^{2-}$  are abundant (e.g., Griffiths 1965; Reyf and Ishkov 1999).

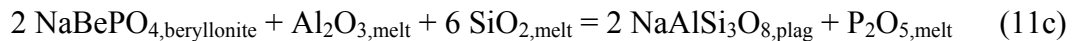
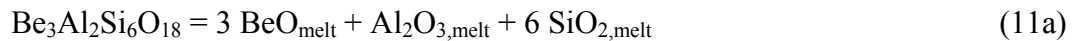
In his review and synthesis of the existing experimental data, Wood (1992) concluded that only  $\text{F}^-$ ,  $\text{F}^- \text{-CO}_3^-$  and  $\text{F}^- \text{-OH}^-$  complexes can generate aqueous Be concentrations  $>1$  ppm in equilibrium with phenakite or bertrandite at temperatures up to  $300^\circ\text{C}$  and at plausible pH conditions. According to Wood’s analysis, fluoride complexes ( $\text{BeF}^+$ ,  $\text{BeF}_2^\circ$ ,  $\text{BeF}_3^-$ ,  $\text{BeF}_4^{2-}$ ) predominate at lower pH (2-5) whereas a mixed  $\text{F}^- \text{-CO}_3^-$  complexes (e.g.,  $\text{BeCO}_3\text{F}^-$ ) may dominate at higher pH (5-7), particularly where  $[\text{F}^-]$  and  $[\text{CO}_3^-]$  both exceed about 0.01 molal. Beryllium concentrations exceeding 1 ppm seem necessary to make many Be deposits, which commonly have  $>1000$  ppm Be. In some settings, lower concentrations may suffice, as for instance in the case of the Colombian emerald deposits where Renders and Anderson (1987) believe that  $\text{OH}^-$  complexes were sufficient to move all the Be necessary to make the emeralds (but cf. Banks et al. 2000).

In spite of their obvious importance to understanding many hydrothermal deposits, aqueous Be concentrations at  $T > 300^\circ\text{C}$  are virtually unexplored except for a very few studies. As is the case at lower temperatures,  $\text{F}^-$  is implicated as though not proven to be the key complexing agent. Beus et al. (1963) found significant Be concentrations in F-bearing solutions that had reacted with beryl, alkali feldspar and quartz at  $490\text{-}540^\circ\text{C}$ . This is consistent with evidence from experiments on fluids equilibrated with Macusani rhyolite at  $650^\circ\text{C}$  and 2 kbar (London et al. 1988). Macusani rhyolite melt (39 ppm Be, 1.3% F) furnishes only 6 ppm Be and 0.35% F to coexisting aqueous fluid (London et al. 1988). Given these results and the fact that beryl solubility in Macusani melts is near 500 ppm Be (Evensen et al. 1999), one can speculate that a plausible maximum Be

concentration in a magmatically-derived aqueous fluid would be on the order of 100 ppm. Such concentrations resemble those calculated by Wood (1992) at lower temperatures for phenakite- and bertrandite-bearing assemblages. They are more than adequate to make a major Be deposit.

**Silicate melts.** Beryllium solubility in felsic melts and its partitioning with coexisting minerals and aqueous fluids has been extensively studied by David London and coworkers (London et al. 1988; Evensen et al. 1999; Evensen and London 2002, London and Evensen, this volume). Others have focused on distribution of Be among silicate minerals in igneous rocks (e.g., Kovalenko et al. 1977; Bea et al. 1994).

Melting of beryllium phases in the end-member systems (Appendix B) has limited geologic relevance, whereas the principal controls on Be solubility in felsic magmas are  $a_{\text{SiO}_2}$ ,  $a_{\text{Al}_2\text{O}_3}$ , and, more rarely other components (Evensen et al. 1999):



The first two reactions were investigated by Evensen and London (1999). They showed that Be mineral solubility is a strong function of temperature, increasing by factors of 2-10 from 650°C to 850°C, and that beryl is the saturating phase ( $\pm$ chrysoberyl) in metaluminous and peraluminous melts (cf. Fig. 12). Their results in compositionally simple haplogranite melts demonstrated that Be solubility decreases with the increasing  $a_{\text{Al}_2\text{O}_3}$  consistent with Reaction (11a). Complexing of Be by other elements is implied by increased beryl solubility in the Li-B-P-F-rich, but nonetheless strongly peraluminous (andalusite- and sillimanite-bearing, Pichavant et al. 1988b) Macusani rhyolite.

Evensen and London's experimental results are roughly consistent with what one would expect from Reaction (11a) and the 1 to 1.5 log unit difference in  $a_{\text{Al}_2\text{O}_3}$  between strongly peraluminous granites (e.g.,  $\text{Al}_2\text{SiO}_5$ -saturated) and metaluminous granites (at the phenakite-beryl boundary) shown in Figure 12. Using Reaction (11a), predicted Be contents of beryl-saturated melt should increase by approximately 0.5 log units (a factor of 3) from the  $\text{Al}_2\text{SiO}_5$  limit to phenakite-saturated conditions. This is compatible with the experimentally observed 3-8 times increase in Be solubility over a similar range of ASI. The differences likely reflect more complex speciation (and thus activity-composition relationships) than this simple analysis allows. Applying the same reasoning to the phenakite-stable field in Figure 12 and using Equation (11b), one predicts that Be contents of phenakite-saturated peralkaline granites would be the same as in metaluminous granites (barring changes in Be melt speciation). Only with decreasing  $a_{\text{SiO}_2}$ , as in undersaturated syenites, would solubilities be substantially higher, perhaps by as much as a factor of two. In melts with exceptionally high  $\text{P}_2\text{O}_5$  activities beryllonite and possibly other Be-bearing phosphates could substitute for beryl (chrysoberyl or phenakite) as the liquidus phase (Reaction 11c, Charoy 1999).

This analysis underscores the conclusion of Evensen and London (1999) that Be mineral saturation in peraluminous melts is plausible for geologically reasonable Be contents and, furthermore that discrete magmatic Be minerals would not be expected in peralkaline and undersaturated systems except, perhaps, in very late pegmatites.

### MAGMATIC BERYLLIUM ENRICHMENTS

Magmatic beryllium enrichments are apparently common, and of interest in their own right, but are they important to make Be deposits? This is uncertain. Enrichment in other elements, notably F for aqueous complexing of Be, may be more much important

for igneous-related hydrothermal systems. Here we review aspects of magmatic enrichments and the compositions of igneous rocks associated with non-pegmatitic deposits.

In felsic magmatic systems, Be concentrations can exceed typical crustal compositions of 2-6 ppm by a factor of 10 or more (Beus 1966; Hörmann 1978). Magmatic Be concentration takes place in intrusive and volcanic rocks which range from strongly peraluminous to peralkaline in composition (as in pegmatites, cf. Černý 1991a, this volume; London and Evensen, this volume). Figures 5C and 5D also illustrate the range of Be contents in magmatic systems and the positive correlation between Be and F contents found in volcanic and hypabyssal rocks (Coats et al. 1962; Shawe and Bernold 1966; Kovalenko et al. 1977; Macdonald et al. 1992). The correlation with F is not seen in many deeper rock suites as in pegmatites, for example, where F may be fugitive (e.g., Černý and Meintzer 1988; London 1997). Post-eruption loss of F also likely accounts for some of the variability in volcanic rock suites.

In peraluminous rocks, magmatic Be contents appear to be limited to a few hundred ppm Be (e.g., Kovalenko and Yarmolyuk 1995; Raimbault et al. 1995) but they may exceed 1,000 ppm in some alkaline rocks (Meeves 1966; Richardson and Birkett 1996). This follows the known pattern of increasing Be solubility with increasing alkalinity of the melt (Evensen et al. 1999, cf. Fig. 5C). In metaluminous and peraluminous systems, Be enrichment commonly accompanies enrichment in Li, Cs, Ta whereas in peralkaline systems, Be enrichment sporadically accompanies enrichments in Zr, Nb, REE and others (Fig. 5D; Tischendorf 1977; Kovalenko and Yarmolyuk 1995; Pollard 1995a). The highest concentrations in most igneous environments are in late-stage pegmatites and post-magmatic hydrothermal alteration. Many systems exhibit a continuum between magmatic and hydrothermal features with Be-bearing igneous rocks having clearly post-magmatic veins and cavities with hydrothermal Be minerals. It is commonly difficult to distinguish magmatic from post-magmatic enrichment. Beryllium- and F-enriched rhyolites (topaz rhyolites, ongonites, etc.) are widespread, typically in the same regions and commonly in the same districts as hydrothermal Be deposits (Shawe 1966; Kovalenko and Yarmolyuk 1995).

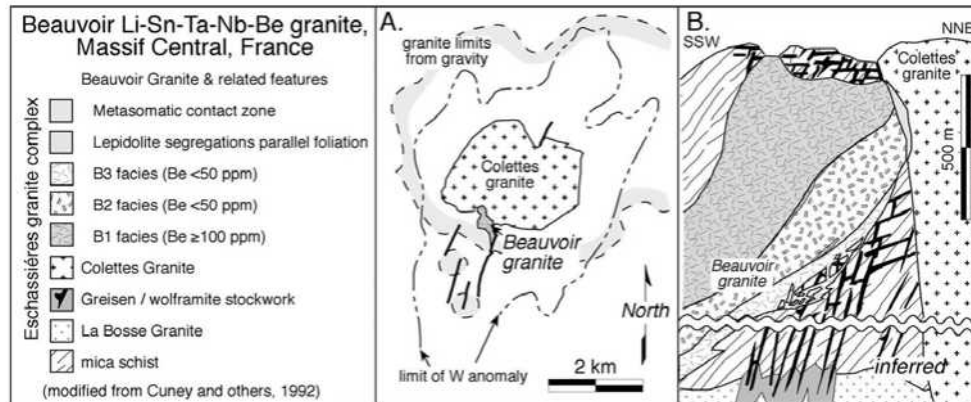
### **Strongly peraluminous to metaluminous systems**

Peraluminous magmas may or may not show strong enrichment in Li with the Be enrichment. Some follow enrichment like that in Li-Cs-Ta pegmatites ("LCT" type of Černý 1991a and this volume). Examples include a number of the highly evolved Hercynian (Variscan) granitoids of Europe (e.g., Raimbault and Burnol 1998; Charoy 1999), the Macusani rhyolite, Peru (Pichavant et al. 1988a), and the Honeycomb Hills, Utah (Congdon and Nash 1991). In contrast, many strongly peraluminous granites do not exhibit this extreme enrichment in rare elements (e.g., Transbaikalia, the western U.S.; Shaw and Guilbert 1990). Nonetheless they have high F contents and late magmatic (miarolitic to pegmatitic) beryl transitional into Be-bearing hydrothermal assemblages. They may evolve along a different path (cf. London 1992). In weakly peraluminous to metaluminous granitoids and volcanic rocks Be enrichments are not accompanied by dramatic (percent level) contents of Li, but they do have elevated values (Fig. 5).

In most peraluminous to metaluminous igneous rocks, Be is dispersed as a trace element in the rock-forming minerals, most commonly the micas and sodic plagioclase (e.g., London and Evensen, this volume; Kovalenko et al. 1977). Accessory magmatic beryl is described in some granites, aplites and miarolitic zones (e.g., Sheeprock Mountains, Utah, Christiansen et al. 1988; Rogers and Christiansen 1989; Argemela, Spain, Charoy 1999; Mt. Antero, Colorado, Jacobson 1993b). Beryllonite is apparently the principal discrete Be mineral in the Beauvoir granite, France, where only modest

amounts of Be (ca. 100 ppm) occur in lepidolite (Charoy 1999).

With the possible exception of the Beauvoir granite (Fig. 15; Cuney et al. 1992; Raimbault et al. 1995), non-pegmatitic peraluminous magmatically enriched rocks lack sufficient Be to be considered Be resources (cf. North Carolina Sn-Ta belt, Griffiths 1954, Fig. 3). At Beauvoir, a composite stock of fine-grained Li-rich leucogranite contains 20-300 ppm Be (>100 ppm in the most evolved unit). It post-dates more voluminous muscovite-biotite granites which have associated greisen-style W and Sn mineralization. Similar patterns are common elsewhere in European Hercynian igneous centers (e.g., Cornwall, England; Manning and Hill 1990).



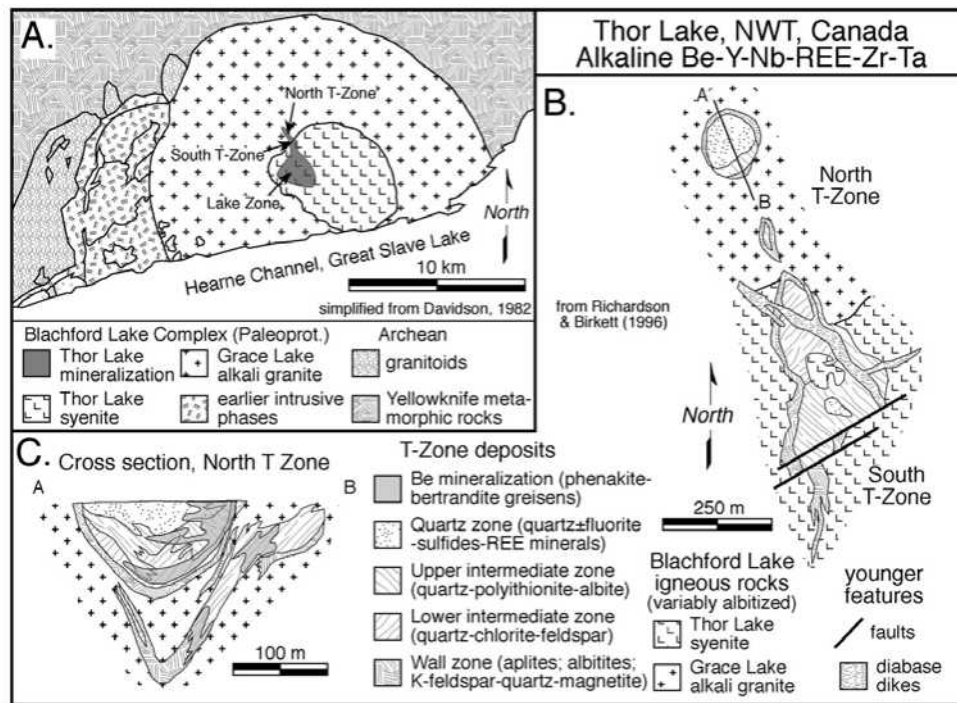
**Figure 15.** Geology of the Beauvoir Li-F-Sn-Ta-Be granite, a strongly peraluminous system with magmatic rare-metal enrichments. (A) The Beauvoir rare metal granite is a late, volumetrically minor phase of the Echassières leucogranite complex; Sn and W mineralization are associated with earlier phases. (B) Cross section through the Beauvoir granite showing three main phases and cross-cutting relationships with earlier granites and W mineralization. Beryllium is concentrated in lepidolite and beryllonite in B1, the final intrusive unit (Charoy 1999). Figures modified from Cuney et al. 1992.

### Peralkaline-metaluminous systems

Peralkaline to metaluminous magmatic systems can have substantial Be enrichments in rocks ranging from riebeckite-aegirine granites to undersaturated syenites and their volcanic equivalents (Richardson and Birkett 1996; Sørensen 1997; Fig. 2, Appendix A). Like the magmatically enriched peraluminous suites, these rocks are typically enriched in F as well as Be but contain a different set of trace elements characterized by Y, Nb, REE with more moderate enrichment in Li (Table 2, Fig. 5; Černý 1991b; Sørensen 1992; Kovalenko et al. 1995a). Associated pegmatitic and hydrothermal deposits are common.

In alkaline granites and quartz syenites Be enrichments can be in the 100s of ppm (Fig. 5; e.g., Khaldzan-Buregtey, Mongolia, Appendix A) and have associated Be-rich alkaline pegmatites. Large deposits with pegmatitic character at Strange Lake and Thor Lake in Canada (Fig. 2, Appendix A) formed during the terminal stages of the development of rare-element-rich alkaline centers. Both have complex internal structures and prominent hydrothermal overprints and the importance of magmatic versus hydrothermal processes concentration is contentious. At Thor Lake (Fig. 16) phenakite, bertrandite, gadolinite and helvite occur in late quartz-fluorite-polyolithionite “greisen” zones in a composite feldspar-dominated “pegmatite” (Trueman et al. 1988). The Be mineralization postdates Ta-Nb-Zr mineralization; both are associated with syenite breccias in syenites and peralkaline granites of the Blachford Lake complex. The deposit post-dates the youngest intrusion, a syenite, and is emplaced in somewhat older alkali granite of the same complex. At Strange Lake, gadolinite, leifite and milarite form in

lenticular zones associated with the latest stages of a Zr-Nb-Y-enriched riebeckite granite complex. A hydrothermal overprint is clear, although it is debated whether the enrichments are fundamentally magmatic (Miller 1996) or hydrothermal (Salvi and Williams-Jones 1996).



**Figure 16.** Geology of the Thor Lake area and rare metal deposits, Northwest Territories, Canada. (A) The Thor Lake deposits are associated with the Thor Lake syenite, the youngest member of the alkaline Early Proterozoic (2.1 Ga) Blachford Lake Complex (Davidson 1982). (B) Be mineralization occurs in the T-zone deposits near the NW margin of the Thor Lake syenite. Phenakite(-bertrandite-gadolinite)-rich hydrothermal quartz-fluorite-polythionite pegmatitic “greisens” are superimposed on a complex set of albite, microcline, and magnetite-rich rocks (Trueman et al. 1988).

The volcanic equivalent of magmatic Be-enriched alkaline granites may be represented by the F-Nb-Zr-Ta-Y-REE-rich trachytic rocks of the Brockman deposit, Western Australia (Ramsden et al. 1993; Taylor et al. 1995a). At Brockman, the hydrothermally altered “Niobium Tuff” averages several hundred ppm Be, which is present (redistributed into?) in quartz-carbonate-bertrandite veins that are restricted to this rare-element enriched stratum. Magmatic concentrations of Be up to 180 ppm occur in the hypabyssal cryolite-bearing, Nb- peralkaline to peraluminous rhyolites of the Sierra Blanca district Texas (Price et al. 1990), which have associated Be-F replacement deposits (see below). Although some rocks from both of these areas are chemically peraluminous, their geological associations, trace-element patterns and associated minerals clearly link them to the peralkaline family.

Beryllium enrichments are also common in the late magmatic phases in undersaturated rocks including examples from the Kola Peninsula, Greenland, and the southwestern United States (Appendix A; Sørensen 1997). Lujavrites (eudialyte-acmite nepheline syenites) from Ilímaussaq, Greenland average 60 ppm Be, while contents up to 1000 ppm have been reported from pegmatitic nepheline syenite at Wind Mountain, New Mexico (Meeves 1966; Steenfelt 1991; Sørensen 1992; Markl 2001). Large Be



inventories have been reported (Fig. 3; Appendix A), but none of the undersaturated alkaline systems appear to host plausible resources due to low grades and dispersion of Be in the rock-forming silicates. Large Be enrichments in phonolites have apparently not been recognized, although the elevated Be seen in shallow intrusive systems like Ilímaussaq and Wind Mountain make eruption of such magmas plausible (Fig. 4). They would be silica-undersaturated, peralkaline analogs of the Macusani rhyolites.

Post-magmatic Be enrichments are widespread in syenitic pegmatites and hydrothermal veins in these locations and others, notably the Oslo province, the Kola Peninsula and Mt. Saint-Hilaire (Appendix A; Beus 1966; Vlasov et al. 1966; Engell et al. 1971; Horváth and Gault 1990; Men'shikov et al. 1999; see below). These undersaturated, typically feldspathoid- or zeolite-bearing rocks contain a distinctive suite of Be minerals, notably the Al-poor, Na-Ca-Be silicates (e.g., epidymite, leifite, leucophanite, chkalovite) plus others such as gadolinite, phenakite, bertrandite, genthelvite and bromellite.

***Magmatic vs. metasomatic albite-rich granitoids.*** Albite-rich granitoids (and some albite-rich syenites) can have either magmatic or metasomatic origins. Both types commonly have Be enrichments but they can be difficult to distinguish from one another. Magmatic varieties have F-enrichments and carry considerable concentrations of rare metals such as Ta, the specific suite corresponding to the overall genetic family (Kovalenko and Yarmolyuk 1995; Pollard 1995a). Such granitoids are extreme differentiates of F-rich magmas (Manning 1982). Fluorine-rich metasomatic albitization is also common in granitic systems and can carry contain broadly similar element enrichments (Charoy and Pollard 1989; Laurs et al. 1996; Haapala 1997). Distinguishing between the two requires textural or geochemical observations (sharp versus gradational geologic contacts; petrographic evidence for replacement, dissolution of earlier minerals such as quartz; high versus low variance assemblages, uniformity of phase proportions). Beryl concentrates in both settings (e.g., Beus 1966; Charoy 1999).

## HYDROTHERMAL OCCURRENCES ASSOCIATED WITH FELSIC MAGMATISM

Hydrothermal Be deposits generated by felsic magmas are numerous and diverse (Table 2, Figs. 2, 4). Depositional environments, particularly the composition of the host rocks, exert the most prominent control on the styles of mineralization regardless of magmatic compositions. Igneous compositions strongly influence mineralogy, element enrichments and zoning. Magmatic Be enrichment can be important in some cases, but overall is apparently subordinate to other factors. We group systems by igneous compositions and foremost, by the degree of alumina saturation because this is predictive of mineral associations (Fig. 12) and correlates broadly with other intensive variables and geologic setting. Boundaries between groups can be arbitrary as there is clearly a continuum among these groups and many igneous centers possess a range of compositions.

Vein, greisen and volcanic (fumarolic) deposits occur in felsic igneous and siliciclastic sedimentary host rocks. These deposits commonly have abundant F. Muscovite-rich alteration, quartz veins and variable amounts of W, Mo, Bi, and Sn typify the beryl-dominated mineralization that forms in strongly peraluminous systems. Li(-Fe) micas and alkali-feldspar alteration become characteristic with decreasing  $a_{\text{Al}_2\text{O}_3}$ , as metal assemblages gain Zr-REE-Nb and lose W. Fenites and quartz-absent hydrothermal veins form in silica-undersaturated systems. At low temperatures (<300°C) bertrandite-bearing quartz veins can form commonly with K-feldspar±carbonate±sericite±fluorite, in some cases with Mn-rich, base-metal sulfide-rich associations.

Skarn, greisen and replacement deposits form in carbonate-bearing host rocks where they generally contain abundant fluorite. They comprise the economically most important deposits including the fluorite-rich replacement deposits in the carbonate-lithic-rich tuff of Spor Mountain. All silica-saturated magma types can produce garnet, pyroxene and vesuvianite-rich Be-bearing skarns where mineral ratios and compositions tend to reflect the redox state of the related granites (cf. Einaudi et al. 1981). Aluminum-rich metasomatism—which produces muscovite, other micas and diaspore (cf. “apocarbonate greisen” of Shcherba 1970)—characterizes the BASH mineral-bearing deposits that form near peraluminous granites. Less aluminous magmas generate skarns and K-feldspar-bearing or Al-poor fluorite-rich replacement deposits. Typical minerals include phenakite, bertrandite, and the helvite group with less common bavenite, leucophanite, gadolinite, milarite and others. A distinctive texture found in many carbonate-hosted systems is rhythmically banded replacement containing alternating light and dark layers with combinations of fluorite, Be minerals (helvite-danalite is typical) and other minerals including silicates and magnetite (“ribbon rock” Jahns 1944b; “wrigglite” Kwak 1987; see photos in Fig. 18C and Fig. 22, below).

Mafic and ultramafic host rocks are relatively uncommon, but they can be important in that they host most emerald deposits, which form where beryl-bearing pegmatites or veins gain Cr and lose silica during original emplacement or subsequent metamorphism. Most such systems are peraluminous. Biotite-producing metasomatism is ubiquitous. This group is treated separately below.

### **Peraluminous magma-related systems**

Hydrothermal Be mineralization occurs with many strongly peraluminous muscovite- or cordierite-bearing granites as well as with weakly peraluminous biotite granites (“pG” in Appendix A). This suite contains some of the more important non-pegmatitic Be deposits, including large sub-economic resources in the Seward Peninsula, Alaska, eastern Nevada and central Kazakhstan. It is also notable for emerald and aquamarine deposits associated with ultramafic and greisen host rocks, respectively (e.g., Reft River, Ural Mtns; Sherlova Gora, Transbaikalia). The salient characteristics of the peraluminous group are aluminum-rich hydrothermal alteration and predominance of BASH minerals.

The peraluminous family can be cast into two groups (cf. Fig. 4A): (1) specialized strongly peraluminous granites, commonly with exceptionally high Li-Cs-Ta (LCT) and other lithophile elements, locally with associated greisen Sn mineralization, and (2a) less specialized but strongly peraluminous granites with or without W-Mo(-Sn) mineralization, or (2b) weakly peraluminous Sn-W(-Mo) systems with elevated rare metal contents. The last group commonly has late muscovite-bearing leucogranites. Although this group can be considered to form a continuum with metaluminous systems, it generally has highly aluminous alteration assemblages in various rock types that are lacking in the latter. Most of hydrothermal systems formed at <5 km depth, but some, particularly those associated with strongly peraluminous muscovite-biotite granites, formed in the 5-10 km range. Geochemical data and geological associations point to a metasedimentary, perhaps dominantly pelitic, source for the magmas of group (1), a mixed crustal source for the magmas of group (2), and a hybrid crust and mantle source for the magmas of group (3) (cf. Černý 1991b; Newberry 1998).

It is unusual to find hydrothermal Be mineralization associated with the more evolved Li-Cs-Ta-type magmas even though many have substantial magmatic Be contents (Fig. 5; e.g., Macusani, Beauvoir, Richemont). For example, most specialized granites of the European Hercynian lack hydrothermal Be occurrences (Stussi 1989; Manning and Hill 1990). In the Cornubian Sn-W district Be minerals occur in greisens,

veins and skarns that are related to the biotite(-muscovite) granites but do not form with the highly specialized topaz granites (Appendix A; Jackson et al. 1989; Manning and Hill 1990). Where Be minerals are mentioned with Li-Cs-Ta-type magmas they form in early assemblages, for example in the Erzgebirge where beryl and hercynite are minor constituents of proximal quartz-topaz greisens at Ehrenfriedersdorf and elsewhere (Baumann 1994). Similar relationships are apparent in the Geiju district, Yunnan, China where Be-W mineralization occurs at the apex of greisenized rare-metal granites and zones outward into Sn and sulfide mineralization (Kwak 1987). The lack of hydrothermal accumulations in many of these systems may reflect their low solidus temperatures, their low water contents and limited ability to exsolve water, and the relatively low partition coefficients for Be into coexisting aqueous fluid (London et al. 1988; Raimbault and Burnol 1998). In contrast to the Li-Cs-Ta-group, the less compositionally extreme peraluminous magmas are associated with many occurrences.

**Feldspathic host rock:** These rocks host three common styles of fracture-controlled Be mineralization: beryl in quartz-K-feldspar(-mica) veins, beryl in albitized rocks, beryl and other Be minerals in muscovite-topaz-fluorite-dominated greisens. The latter are by far the most important. Many areas contain all three styles in a progression from early, proximal and typically deeper K-feldspar-stable assemblages through albitization to late, commonly distal greisen associations. Occurrences are widespread, notable examples found with biotite±muscovite granites occur in China, central Asia, the North American cordillera, and western Europe (Appendix A).

Small, coarse-grained quartz-K-feldspar(-muscovite-biotite) veins containing accessory beryl, molybdenite and wolframite occur with some peraluminous pegmatites and W-Mo(-Sn) affiliated granitoids (e.g., in the Canadian cordillera and maritime provinces, Mulligan 1968). These veins typically lack fluorite and paragenetically later Be-rich veins are rare. Geological context indicates that they formed at considerable depth; they could represent root zones of other deposit types. One variant on this theme is illustrated by the large Verknee Qairaqty and Koktenkol stockwork W(-Mo) deposits in Kazakhstan where minor beryl occurs only in early, 300-400°C quartz-K-feldspar-molybdenite-scheelite veins (Mazurov 1996; Russkikh and Shatov 1996). Beryllium is distributed throughout the paragenesis at Dajishan, Jiangxi, China where quartz-feldspar-beryl veins change with time and distance into helvite-bearing fluorite-muscovite-quartz veins with wolframite, scheelite and molybenite (Raimbault and Bilal 1993). Another variant may be represented by the relatively F-poor Sn-W deposits of SE Asia where Be mainly occurs as beryl in pegmatitic bodies (Suwimonprecha et al. 1995; Linnen 1998). Feldspar-dominated veins are of little economic interest and consequently are thinly documented. Conversely, alkali feldspars are present in many F-rich greisen and albitic assemblages associated with major Be deposits.

Albitized rocks are widespread in Be-rich peraluminous systems where they grade into mica-dominated greisen assemblages. Typically, albite+muscovite±fluorite±chlorite replace igneous feldspars and micas; modal quartz also commonly decreases. These form pipes, veins, vein envelopes (commonly around mica-rich greisen veins), and pervasive zones particularly near the tops of intrusions. Accessory beryl with albitic assemblages is reported in many systems (Beus 1966; Dyachkov and Mairorova 1996). A well described example at Triberg, Germany (Markl and Schumacher 1996, 1997) formed from biotite-muscovite leucogranites that have late beryl-bearing miarolitic pegmatites. Hydrothermal beryl-albite-muscovite-fluorite alteration ultimately grades into beryl(-bertrandite-phenakite)-bearing quartz-muscovite-topaz greisen veins. Like many other beryl-rich two-mica systems, the mineralizing fluids contained <10 wt % NaCl equivalent (NaCl<sub>eq</sub>) and Sn and W were only weakly concentrated. Rogers (1990) describes similar

relationships in muscovite-bearing biotite granite in the Sheeprock Mountains, Utah.

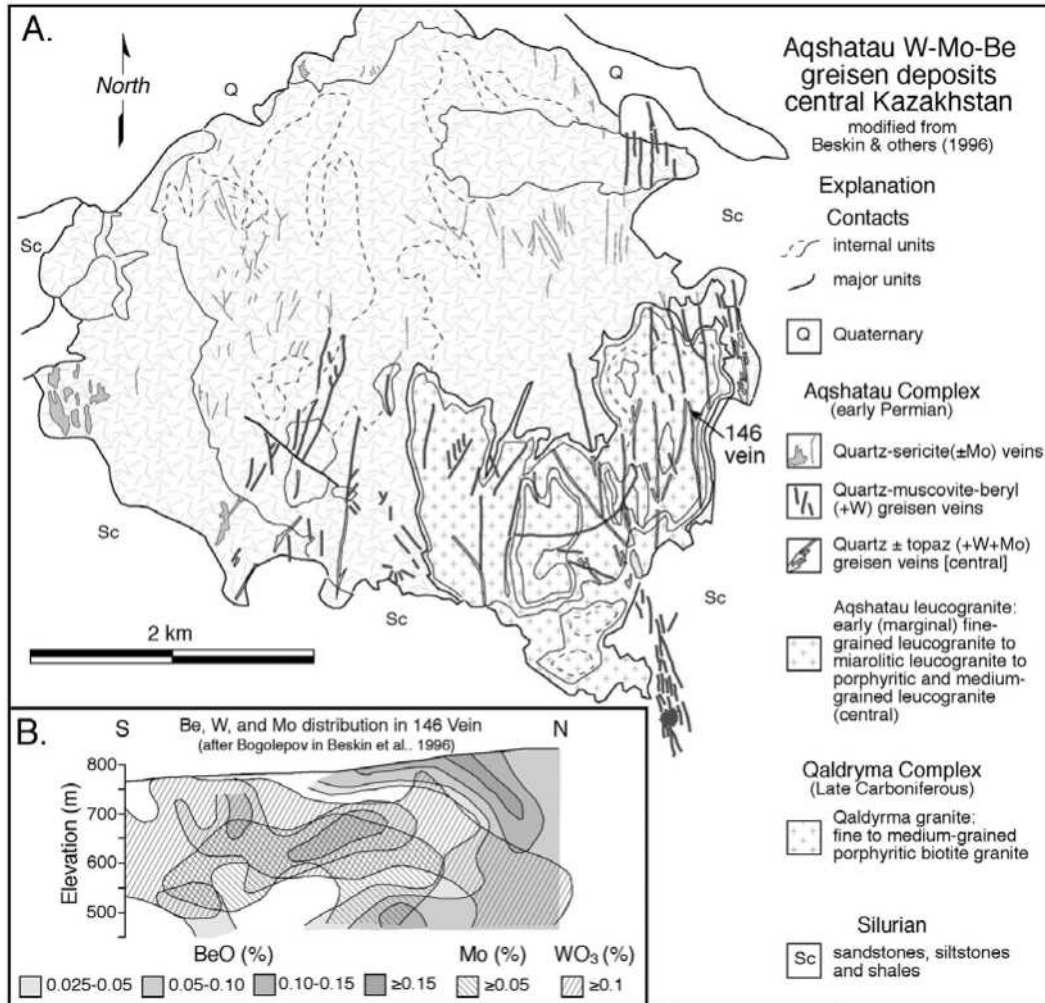
Quartz-rich greisens with abundant accessory muscovite, topaz, fluorite and siderophyllite contain most Be minerals (beryl > phenakite, bertrandite, euclase) found in peraluminous-related deposits. Beryl can be either in the vein fill with quartz and other minerals or it can concentrate at the outer margins of the greisen envelopes against feldspar-stable assemblages (generally albite; Beus 1966). Overall Be distribution varies in greisens; it is typically distal or late within the intrusions and may or may not extend into surrounding veins or skarns. Hematite-bearing alteration and helvite group minerals are rare with the peraluminous group in contrast to greisens associated with fundamentally metaluminous biotite granites.

Many examples from around the world illustrate these patterns (Appendix A). In the Great Basin, reduced (low  $\text{Fe}^{+3}/\text{Fe}^{+2}$ ) Cretaceous muscovite-biotite granites and their clastic host rocks contain minor beryl in muscovite-fluorite-quartz-pyrite±wolframite greisen veins. Paragenetically earlier quartz-K-feldspar veins and albitization typically lack beryl. This is compatible with the paucity of Be in the correlative pre-greisen stages of associated skarns. These systems contain large quantities of F, Be and Zn, with minor Mo (proximal), W and Sn (distal). Associated fluids had moderate  $\text{CO}_2$  contents and salinities (5-10 wt %  $\text{NaCl}_{\text{eq}}$ ) and were of magmatic derivation. In the northern Cordillera, sparse beryl occurs in proximal muscovite- or topaz-bearing alteration in W-Mo (e.g., Logtung, Yukon) and Sn-W (e.g., Lost River, Alaska) systems. Similar relationships hold along the western margin of the Pacific in southeastern China (e.g., Wangfengshan, Guangdong) and eastern Australia (e.g., Mole Granite, New South Wales)

Many Be-bearing W-Mo(-Sn) greisens occur in central and eastern Kazakhstan where they are associated with mainly Late Paleozoic biotite±muscovite leucogranites (Appendix A; Burshtein 1996; Serykh 1996; Ermolov 2000). Beryllium occurs in several modes: as beryl in muscovite-topaz-quartz greisens (e.g., Aqshatau; Fig. 17), as bertrandite±helvite (after beryl) in late fluorite-rhodochrosite-sulfide veins (e.g., East Qonyrat), and as chrysoberyl and other minerals in F-rich skarns (e.g., Qatpar).

Among peraluminous-related deposits greisen deposits are the more common than other types and the only variety from which Be has been produced. The largest reported resource is the Aqshatau district in Kazakhstan (Fig. 17; Appendix A; Beskin et al. 1996) where beryl has been produced from W(-Mo-Bi) greisen veins. Beryl formed in the distal parts of the quartz-topaz-muscovite-wolframite greisen veins where  $\text{BeO}$  contents can exceed 0.1% (Fig. 17B). These veins exhibit zoning centered on a multi-phase muscovite-biotite leucogranite complex that appears to be the fluid source. Extensive study demonstrates that saline (>30 wt %  $\text{NaCl}_{\text{eq}}$ ) magmatic fluids account for most of the mineralization with fluid pressures fluctuating near lithostatic values. These fluids were followed by an influx of dilute, meteoric waters under hydrostatic conditions that formed late quartz-sulfide-carbonate assemblages.

The noted aquamarine locality at Sherlova Gora, Siberia (Sinkankas 1981) has extensive beryl-bearing quartz veins (with  $\text{BeO} \geq 0.02\%$ ) with topaz, siderophyllite, fluorite and muscovite greisens (Beus 1966). These occur in the outer portion of a variably miarolitic and porphyritic biotite±muscovite granite pluton. Within the pluton, beryl is late and tends to be distal in the greisen veins. Beryllium as well as alkalis are removed from intensely greisenized rocks. In the system as whole, beryl and minor Mo-W±Sn mineralization form proximally whereas Sn-polymetallic mineralization extends well away from the intrusions (Troshin and Segalevich 1977). This district, although emplaced at shallower levels than those in the Great Basin, is also characterized by both muscovite-bearing granites and relatively low Sn-W-Mo contents.

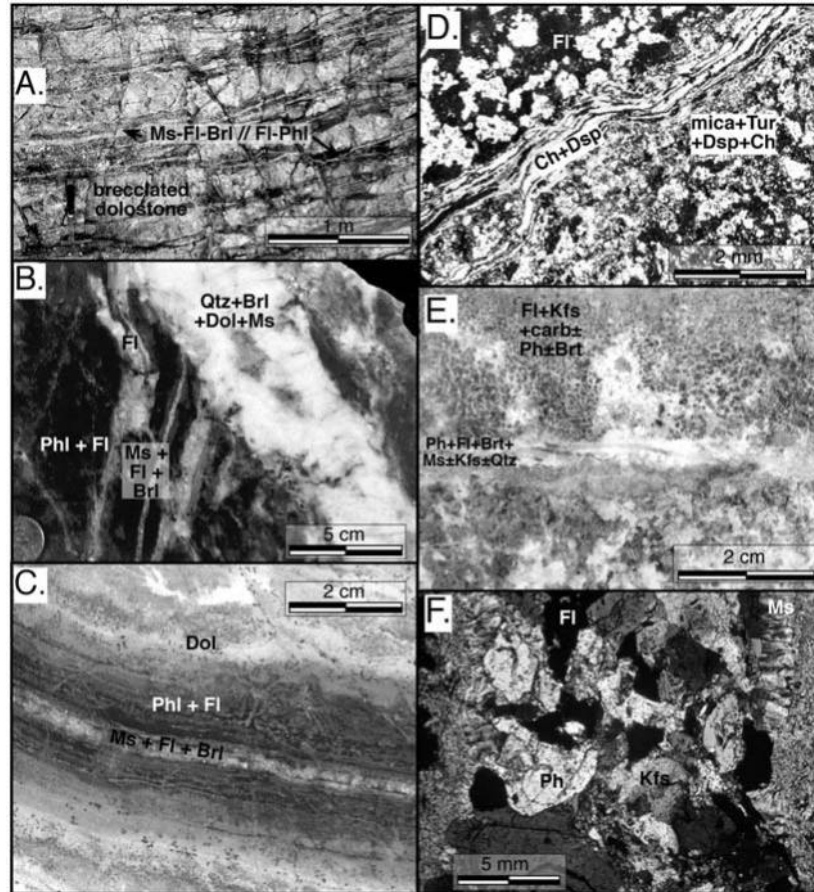


**Figure 17.** Geology of the Aqshatau (Akchatau) greisen W-Mo-Bi-Be district, central Kazakhstan. (A) Geology map showing the distribution of granitoids and hydrothermal features in the central part of the district. Older, Carboniferous granodiorites and granites of the Qaldyrma complex are cut by cupolas of the ore-related Permian biotite-bearing leucogranites of the Aqshatau complex. Note lateral zoning of W-Mo-Be mineralization away from the latter. (B) Zoning of WO<sub>3</sub>, Mo and BeO (values in %) along the 146 vein (see A for location). [A,B redrawn from Beskin et al. 1996.]

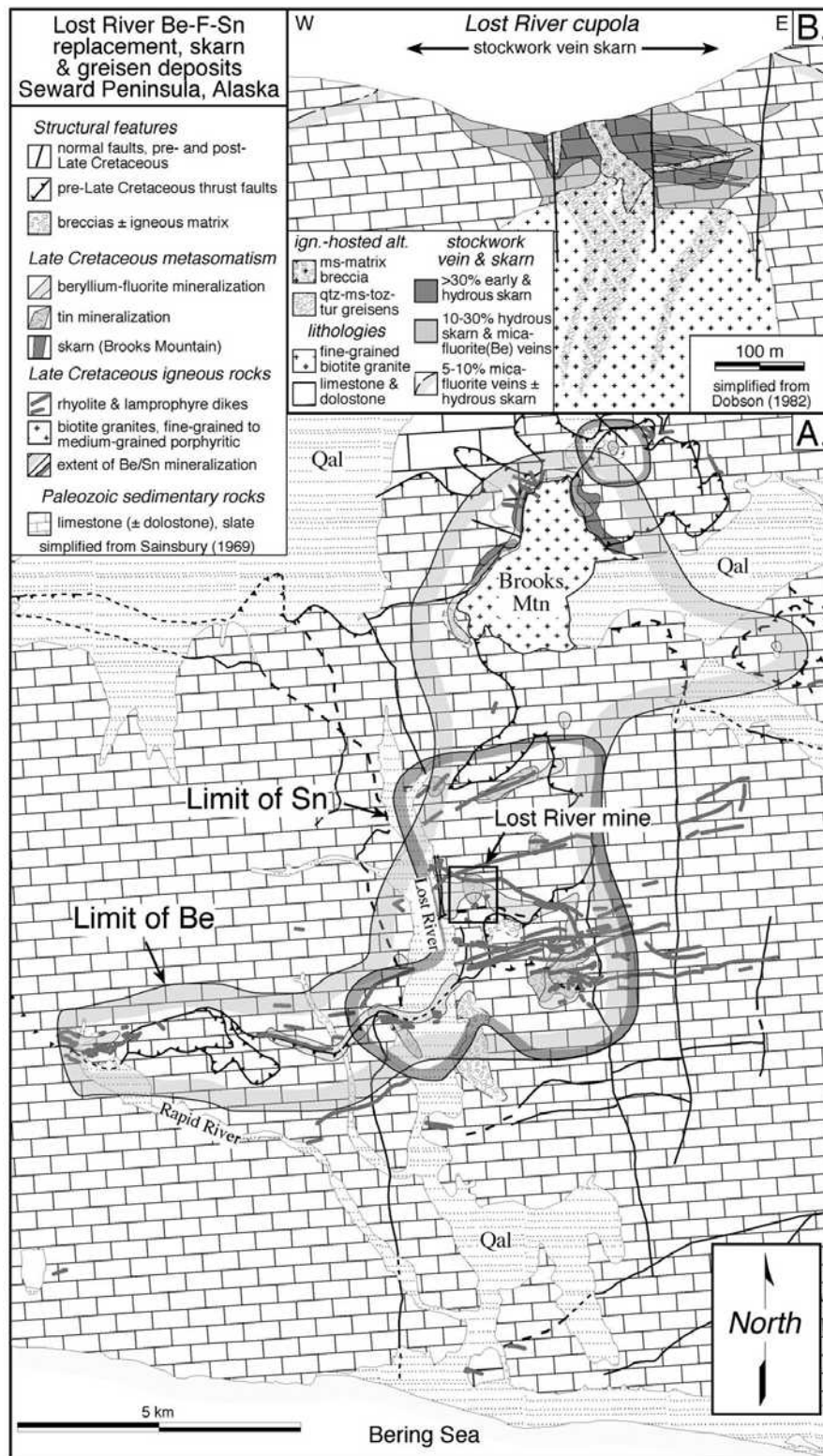
Although greisen-type alteration hosts most Be minerals in peraluminous granitoids, Be minerals are commonly sparse when compared to the amounts present in adjacent carbonate hosted-mineralization. With only a handful of exceptions worldwide (Aqshatau, Sherlova Gora), the peraluminous granite-hosted deposits have not been a significant source of either Be metal or gems. This is well illustrated by Phanerozoic Be-rich magmatic-hydrothermal systems around the circum-Pacific (e.g., Cretaceous Cordillera, Tasman system, SE China; see Appendix A).

**Carbonate host rock** In carbonate rocks, Be deposits related to strongly peraluminous granitoids are characterized by exceptionally high F and Al contents and elevated contents of many other elements including Li, Sn, and W. Beryllium occurs both in skarns and in superimposed or distal apocarbonate greisen or replacement deposits. In skarns, Be is reported to isomorphically substitute in vesuvianite and other silicates (e.g., Beus, 1966), whereas in the greisen or replacement deposits Be clearly forms discrete

phases. Fluorite, F-rich silicates, micas (muscovite, Li-micas, phlogopite), topaz, albite, K-feldspar and quartz can all be abundant. Typical Be minerals include chrysoberyl, phenakite, beryl, bertrandite, euclase and bertrandite. Rarely present are bavenite and Mg-bearing aluminates of the taaffeite group (Table 1). Quartz is typically sparse in carbonate-hosted greisen-style alteration. Iron and base metal contents vary considerably. Pressure (depth of emplacement) appears to be an important factor in quartz and sulfide abundances: higher salinity fluids, more extensive metal transport and silica-undersaturated assemblages occur at <1-1.5 kbar, whereas more siliceous assemblages (due to higher silica mobility) and less concentrated metals (linked to lower fluid salinities) are more common at higher pressures (cf. Fig. 11A).



**Figure 18.** Textures of carbonate-hosted Be mineralization associated with strongly peraluminous granites showing characteristic F- and Al-rich veins and replacements. (A)-(C) Mica-fluorite-beryl (-quartz) veins from the deep W(-Mo) systems at McCullough Butte, Nevada (Appendix A). (A) Trench face with typical muscovite-fluorite(-beryl-pyrite-scheelite-sphalerite-bertrandite-quartz) veins and fluorite-phlogopite envelopes in brecciated dolostone. This exposure averages about 25% CaF<sub>2</sub> and 0.25% BeO. (B) Quartz-aquamarine-dolomite(±muscovite) vein cutting muscovite-fluorite-beryl(white) veins with inner fluorite to outer fluorite+phlogopite envelopes. (C) Rhythmically banded fluorite-phlogopite skarn envelope on muscovite-fluorite-beryl-pyrite vein. (D) Mottled chrysoberyl-rich replacement from the shallow Lost River Sn(-W) district, Alaska. Very fine-grained chrysoberyl+diaspore vein cuts mottled white mica (Li, Be-bearing)+tourmaline+diaspore+chrysoberyl replacement of limestone. (E)-(F) Phenakite-bertrandite-fluorite vein and replacement mineralization Mount Wheeler Mine, Nevada (Appendix A) which typical of distal fluorite-rich mineralization in carbonate rocks associated with many magma types (cf. Zabolotnaya 1977). (E) Phenakite(-bertrandite)-fluorite-muscovite-adularia vein-cutting fluorite-adularia-manganosiderite-phenakite replacement of limestone. Fluorite is dark and comprises more than half of these zones. (F) Micrograph of central vein from (F) showing phenakite-fluorite-adularia (Kfs) vein with muscovite border. All photos by M.D. Barton (except for A—modified from Sainsbury 1969).



**Figure 19.** Geology of Be-F-Sn-W mineralization in the Lost River area, Seward Peninsula, Alaska. (A) Geologic map of the region showing the distribution of Be-F replacement mineralization, Sn and skarn mineralization, and igneous rocks. Note the strong structural control. Qal – Quaternary alluvium. Simplified from Sainsbury (1969). (B) Cross section of the Lost River mine area showing the generalized distribution of early and hydrous skarns and multiple stages (early and late) fluorite-mica-Be veins. Simplified from Dobson (1982).

High-grade Be mineralization in carbonate rocks typically consists of paragenetically complex fine- to coarse-grained fluorite±mica, K-feldspar, diaspore or tourmaline-rich open-space and replacement veins (Fig. 18; Ginzburg et al. 1979). These veins typically comprise stockwork systems that can extend up to several kilometers from known sources and proximal skarns (e.g., McCullough Butte, Nevada; Lost River, Alaska; Fig. 19). Rhythmic layering is common as it is in other Be-bearing systems (Fig. 18C). The ore minerals range from very fine-grained—“curdy” chrysoberyl is common and represents a metallurgical challenge for economic recovery (Fig. 18D; Apollonov 1967)—to quite coarse-grained, which is common with phenakite-bertrandite-fluorite replacements (Fig. 18E; Ginzburg et al. 1979). Chrysoberyl occurs principally in the lower pressure, better metallized districts.

Associated anhydrous calcic skarns range from reduced types with hedenbergitic pyroxene > aluminous garnet+vesuvianite to oxidized types with andraditic garnet±vesuvianite±magnetite > diopsidic pyroxene (cf. Newberry 1998). The latter are more common with shallow, Sn(-W) systems, like those associated with biotite granites in western Alaska or Tasmania, whereas the former are more common with W(-Mo-Sn) skarns associated with biotite±muscovite granites from southeastern China and the western USA (Fig. 4). Magnesian skarns of both types differ in having more abundant humite-group minerals, other Mg-silicates, sellaite, spinel and magnetite (cf. Einaudi et al. 1981; Kwak 1987).

One of the best-documented districts is at Lost River on the Seward Peninsula, western Alaska (Appendix A; Fig. 19). This area contains the second largest U.S. Be resource after the tuff-hosted ores at Spor Mountain, Utah. Biotite granites and late muscovite-bearing leucogranites in the Seward Peninsula are all peraluminous, but they differ from strongly peraluminous granitic suites in that they are commonly more oxidized (higher  $Fe^{+3}/Fe^{+2}$ ) and they mostly lack Al-saturating phases such as muscovite or cordierite (Sainsbury 1969; Swanson et al. 1990). The Be deposits occur as fine-grained, commonly laminated chrysoberyl-diaspore-mica-fluorite veins and replacement bodies in limestones and dolomites (Fig. 18D). Also present are minor phenakite, beryl and euclase. These bodies are developed along faults and adjacent to dikes and extend kilometers from the Sn-rich greisenized granites and skarns. There is a strong vertical control on the distribution of hydrothermal alteration that reflects level of exposure in the systems. The Be-F-Al replacements are mainly distal and structurally high. Proximal iron-rich skarns can contain helvite with fluorite and magnetite, but more commonly are dominated by andraditic garnet, vesuvianite, magnetite, fluorite, scapolite and minor pyroxene overprinted by hydrous skarn with hornblende, biotite, fluorite, cassiterite and sulfides (Dobson 1982; Swanson et al. 1990). The skarns abut topaz-tourmaline-muscovite greisenized fine-grained biotite±muscovite granites (Fig. 19 inset). Carbonate-hosted greisen veins both predate and postdate the skarns and thus indicate multiple fluid release events (Dobson 1982).

Systems similar to Lost River occur elsewhere in Alaska and, notably, down the eastern side of Asia from Siberia into southern China (Fig. 2; Appendix A). Systems in southeastern China are associated with middle to late Mesozoic biotite±muscovite granites, typically with greisen mineralization within the intrusions. Unlike in North America, several of these complex polymetallic districts are economically important for other commodities. The Shizhuyuan, Hunan W(-Sn-Mo) district contains the world's premier W deposit and has Be-rich Mo-W-B-Sn±Cu±Pb skarns, greisens and replacement bodies (Mao et al. 1996b). At Xianghualing, Hunan, late chrysoberyl-fluorite-phlogopite ribbon rock and later fluoborite-chrysoberyl (-taaffeite group mineral) mica veins with spinel envelopes and minor sulfides overprint garnet-vesuvianite skarns



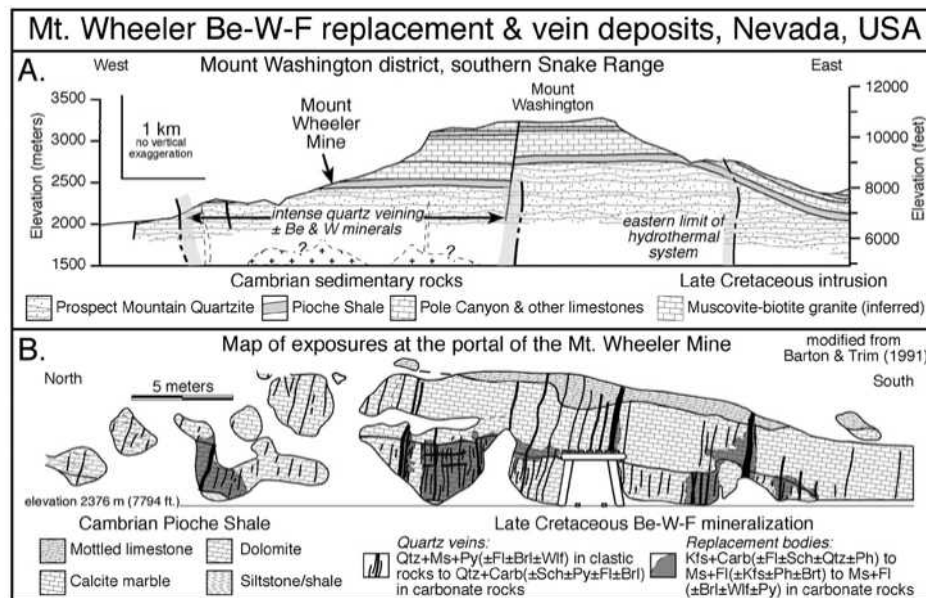
and ribbon magnetite skarns. These magnesian skarns contain chondrodite, vesuvianite, diopside, amphibole, tourmaline and formed in dolomitic host rocks adjacent to beryl-bearing albitized and greisenized biotite granite and (Lin et al. 1995a, p. 238-242). Similar Be-F-rich mica-dominated veins overprint and are distal to complex polymetallic Sn skarns in the Geiju Sn district (Kwak 1987).

In districts with less compositionally evolved magmas, Be tends to be dispersed either in rock-forming silicates or as minor beryl. High Be concentrations occur only in particularly favorable traps. For example, many large hydrothermal systems are associated with Cretaceous two-mica granites in the Great Basin of the western United States (Appendix A). At McCullough Butte, Nevada two-mica granite contains 6-12 ppm Be and 0.03-0.2% F (Barton, unpubl. data), whereas contemporaneous two-mica granitoids emplaced at 20+ km depths in the Ruby Mountains have 0.5-4 ppm Be (Calvin Barnes, written comm., 2000). Around and above the shallower plutons (emplaced at 5-10 km depth), muscovite-phlogopite-fluorite-quartz-bearing veins and replacement zones overprint or are peripheral to reduced W(-Mo-Sn)-bearing pyroxene>garnet skarns (Fig. 18A-C). Some of these hydrothermal systems contain much Be (e.g., McCullough Butte, Fig. 3), but high concentrations (>0.1% BeO) only formed in distal locations where fluids traversed unreactive quartzites prior to encountering carbonate rocks. This is illustrated by McCullough Butte (see Fig. 18A) and, most extensively, around the Mount Wheeler mine, Mt. Washington district, Nevada (Fig. 20). In the Mt. Washington district, a substantial Be(-W-F) resource (Fig. 3) occurs in laterally extensive phenakite-bertrandite-beryl-scheelite-fluorite replacement bodies that formed in the basal carbonate unit above a thick clastic section overlying an unexposed granite (Fig. 20B). In place of skarn minerals, early mineral assemblages in this distal, cool hydrothermal system contain Fe-Mn carbonates and quartz, and were followed by deposition of progressively Al-, F-, and Be-enriched assemblages that culminate in muscovite-fluorite-beryl-quartz veins.

### **Metaluminous magma-related systems**

Notable non-pegmatitic deposits of Be occur throughout the world with metaluminous or weakly peraluminous magmatism (“G” in Appendix A, Fig. 2). The preeminent region comprises Spor Mountain and other deposits associated with mid-Tertiary felsic magmatism in southwestern North America. In contrast to strongly peraluminous-related systems where Be can be economically concentrated in pegmatites and igneous-hosted vein deposits, metaluminous suites lack major intrusion-hosted deposits even though phenakite, beryl, helvite group and gadolinite occurrences are widespread. Instead, carbonate and volcanic rocks host the important deposits which mostly are fluorite-rich, bertrandite-, phenakite- or helvite-bearing veins and replacements. Beryl is normally subordinate to the other Be minerals or is completely absent.

Genetically related igneous rocks are highly felsic—typically biotite-bearing leucogranites and high-silica rhyolites (Fig. 5A,B). Coeval syenites and hornblende-bearing calc-alkaline granites are commonly present, whereas muscovite-bearing variants are scarce. They may occur in bimodal suites with mildly alkaline mafic rocks. Typical associated metals include Mo, Sn(>W), Ta, Nb and Zn. Most, but not all, of these systems are F-rich and many are sulfur-poor as evidenced mineralogically by widespread genthelvite-danalite solid solutions (Fig. 9) and other indicators of relatively low S such as Pb-enriched feldspar (amazonite). Another common feature of metaluminous-related systems, shared to some degree with other types, is an association with distal Mn-rich replacement or vein mineralization. In some areas, for example south-central New Mexico, helvite-bearing base-metal sulfide replacements formed near skarn and volcanic-hosted Be mineralization. A link to proximal Be-enrichments is not evident in other



**Figure 20.** Geological relationships in the Mount Wheeler Mine area, Mount Washington district, Nevada. Beryllium mineralization here is associated with non-specialized strongly peraluminous granites. This district contains the most extensive and highest grade Be mineralization of the >20 occurrences associated with Late Cretaceous two-mica granites in the Great Basin (Barton 1987; Barton and Trim 1991). (A) Cross section through the district showing lateral extent of hydrothermal system. High-grade Be(-W-F) mineralization is laterally extensive but vertically restricted to carbonate rocks in the Pioche Shale, the lowest reactive beds in the sedimentary sequence. Only sparse granite porphyry dikes are exposed at the surface, but large two-mica granite bodies are exposed nearby in the southern Snake Range (Barton, unpubl.). (B) Sketch map showing the localization of high-grade (ca. 1% BeO & WO<sub>3</sub>; 20% CaF<sub>2</sub>) low-temperature (<300°C) replacement bodies that form where sheeted quartz veins intersect the basal carbonate rocks. See Figure 18E,F.

examples, including helvite in the epithermal Mn-rich mineralization at Silverton, Colorado and in the distal polymetallic Mn-mineralization at Butte, Montana district (Warner et al. 1959). These occurrences share features with distal rhodochrosite-bertrandite and Mn-silicate/carbonate occurrences in many districts in central and east Asia (e.g., East Qoyrat, Kazakhstan; Shizhuyuan, China).

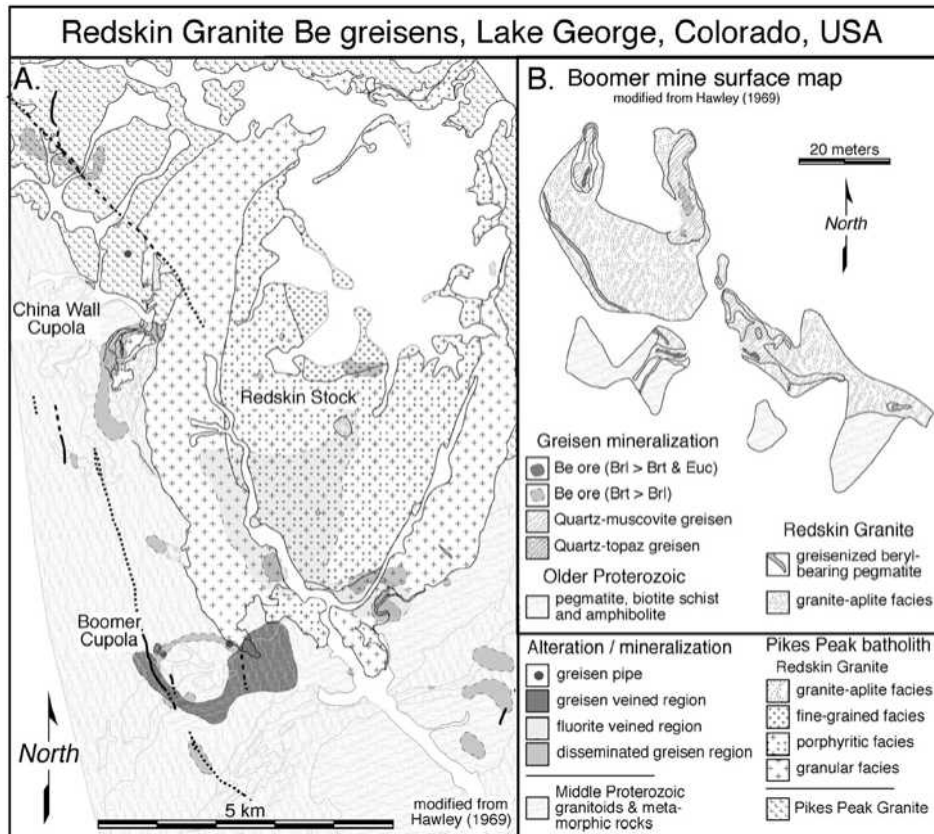
**Felsic host rock in non-volcanic settings:** In intrusive environments pegmatitic, miarolitic, albitized and greisenized bodies contain gadolinite, beryl, bazzite, phenakite, bertrandite, bavenite, and helvite group minerals. Although many features overlap with peraluminous systems, mineral assemblages are typically less acid (higher molar (K+Na)/Al). Cavity filling and alteration mineral associations are dominated by alkali feldspars, quartz and trioctahedral micas (siderophyllite, zinnwaldite, etc.). Muscovite is common in some greisens, but is generally less abundant than in strongly peraluminous examples. Beryl and (rare) euclase are largely restricted to muscovite-bearing greisens. In a number of regions, porphyry Mo style mineralization is associated with the Be-bearing intrusive complexes (e.g., in Norway, southwestern US, central Asia; cf. Geyti and Schönwandt 1979; Burt et al. 1982; Burt and Sheridan 1986). Although few of these areas have detailed petrologic documentation, most appear to have formed at no more than 5 km depth from moderate salinity magmatic fluids. None of the granite-hosted deposits have been major sources of Be.

Biotite granites and monzonites with accessory Be minerals in miarolitic cavities and pegmatitic veins are well known in the southern Alps, Oslo Rift, Colorado (Pikes Peak

batholith; Mt. Antero), and the Younger Granites Province of Nigeria (Appendix A). These occurrences are small and mainly of mineralogical interest. For instance, sparse gadolinite, bazzite, and bavenite are present in pegmatitic quartz-feldspar( $\pm$ beryl) assemblages in biotite granite plutons in northern Italy and in miarolitic cavities in the peralkaline to metaluminous Oslo Rift intrusions. Mid-Tertiary calc-alkaline batholiths in Colorado (Mount Antero area, Mt. Princeton batholith) and Idaho (Sawtooth batholith) have biotite granites with miarolitic cavities and pegmatitic quartz veins that are noted for their aquamarine and other specimen material. At Mt. Antero, beryl, phenakite and bertrandite form sequentially in miarolitic cavities, pegmatites and muscovite-bearing quartz-molybdenite veins all associated with the apex of a small chemically specialized biotite leucogranite (Adams 1953; Jacobson 1993b). Geologic relationships and fluid inclusion data from Mt. Antero indicate that associated fluids were magmatic in origin had fairly low salinities (0.5-8 wt % NaCl<sub>eq</sub>) and spanned a wide temperature range (~600 to 200°C; Kar 1991).

In metaluminous-related Be-rich hydrothermal systems, paragenetically early metasomatic K-feldspar $\pm$ K-Fe mica assemblages are commonly mentioned; however, these assemblages are not well-described and they apparently lack Be minerals. This is a bit surprising given the common co-occurrence with beryl and other minerals in miarolitic cavities. Similarly, albitization with quartz loss and Li-Fe-Al mica growth in metaluminous systems mostly lacks Be minerals (e.g., Beus 1966; Charoy and Pollard 1989). An exception is at Sucuri, Brazil where helvite-group minerals occur with iron and base metal sulfides (Raimbault and Bilal 1993). Albitization of uncertain origin—magmatic or hydrothermal—is widely reported. Haapala (1997) argued that much of the albite is magmatic or due to local post-magmatic redistribution of components without significant sodium metasomatism. Albitization is widely reported in the Nigerian alkaline complexes in the same areas that contain Be-bearing greisens. According to Bowden et al. (1987) the albite has “snowball texture” (albite laths and zinnwaldite enclosed in large microcline and quartz crystals)—a texture that plausibly seems magmatic (e.g., Lin et al. 1995b, but cf. Kempe et al. 1999). Given the focus placed on albite-rich rocks as either an evolved magmatic source or a leached metasomatic source for Be this topic remains a fertile one for additional work.

In felsic host rocks hydrothermal Be mainly occurs in small greisen veins and pipes with or without associated albite-rich assemblages. Well-known regions are commonly Sn-rich and include Karelia, Brazil, Nigeria, Colorado and central Asia (Fig. 2, Appendix A). Hydrothermal quartz, Li-Fe-Al sheet silicates, and fluorite are ubiquitous. Topaz, cassiterite, wolframite and Mo-Zn(-Pb-As-Cu) sulfides are common. Cryolite can be present in fluid inclusions or as a separate phase. Sheet silicates include chlorite, Li-muscovite, siderophyllite and Li-Fe micas with chlorite being most common in the outer part of vein envelopes. Although muscovite is prominent in some deposits, greisens in many areas contain only trioctahedral micas (e.g., in Nigeria, Bowden and Jones 1978). Beryl is the most widely reported Be mineral and bertrandite, genthelvite and phenakite are common. These bodies typically zone from central quartz-rich bodies with topaz through inner mica-quartz envelopes to outer chlorite-mica-K-feldspar envelopes. A characteristic reddening of vein envelopes due to dispersed hematite is widely reported (e.g., Nigeria, Karelia). Beryllium minerals typically occur in the central part of greisen bodies along with other ore minerals. Petrological studies of several systems in Nigeria, Karelia and Mongolia show that mineralizing fluids are of magmatic origin and saline (10 to >40 wt % NaCl<sub>eq</sub>), can have moderate CO<sub>2</sub> contents, commonly show evidence of phase separation, and were trapped at temperatures from 200 to 500°C (Haapala 1977a; Imeokparia 1992; Akande and Kinnaird 1993; Graupner et al. 1999).



**Figure 21.** Geology of the greisen Be deposits associated with the Redskin biotite granite, Tarryall Mountains, Colorado (Hawley 1969). The Redskin granite is late, highly evolved phase of the 1 Ga Pikes Peak batholith (Desborough et al. 1980). Like some other fundamentally metaluminous systems, it has late muscovite-bearing phases which are associated with the Be-rich muscovite-quartz-fluorite-topaz greisens (inset). Mirolitic and other pegmatitic Be mineral occurrences occur widely in nearby parts of the Pikes Peak batholith (Eckel 1997). The distribution of rock types, hydrothermal features, and structures make it seem likely that the complex is tilted at moderate angles to the southwest giving an oblique section.

A well studied example of greisen-type mineralization is that associated with late units of the Redskin biotite granite in Colorado (Fig. 21; Hawley 1969). The Redskin granite is an evolved late phase of the variably alkaline 1.0 Ga Pikes Peak batholith, Colorado (Desborough et al. 1980). As a whole the batholith is known for Nb-Y-F-type miarolitic pegmatites that contain genthelvite, phenakite, bertrandite, gadolinite but lack beryl (Levasseur 1997; see next section). Quartz-lithian muscovite-topaz-fluorite ( $\pm$ wolframite-cassiterite-sulfide) greisens form small (1-20 m) pipes and veins developed within and above variably muscovite-bearing porphyritic and aplitic biotite granites. These occur along the southern and western phases of the intrusion which likely was the upper part of the now tilted intrusive system. Beryllium minerals (beryl+bertrandite $\pm$ rare euclase) were quite localized within the greisen bodies and in places formed unusually high grade deposits ( $>5\%$  BeO; Hawley 1969). The deposit at the Boomer Mine produced the first non-pegmatitic Be ores in the United States, operating between 1956 and 1965 (Meeves 1966; Hawley 1969).

**Carbonate host rock** As in the case of peraluminous-related systems, limestone and dolostone make excellent hosts for F- and Be-rich skarn and replacement deposits. These commonly accompany skarn, replacement and greisen Sn, W and base-metal

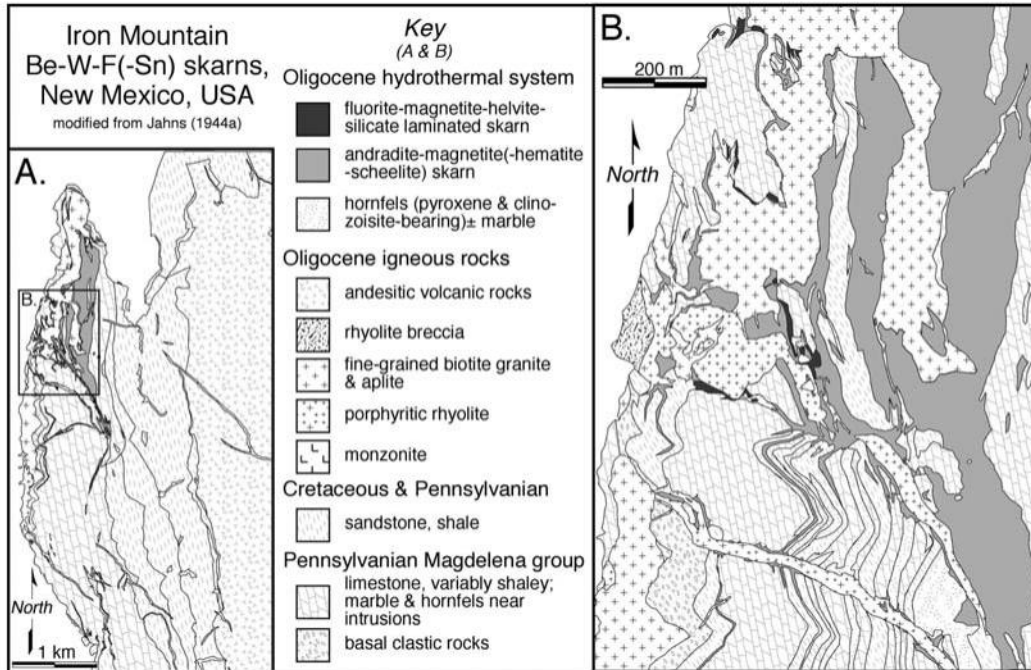
mineralization. In some regions only carbonate-hosted mineralization comprises ore (e.g., Karelia, Haapala 1977b). Styles share some common features with strongly peraluminous systems, notably that anhydrous skarns with little Be are overprinted by hydrous, typically F-rich assemblages with higher Be contents. They differ in that the metaluminous-related skarns tend to be more oxidized (garnet/vesuvianite > pyroxene; cf. Einaudi et al. 1981) and have replacement assemblages that are typified by fluorite plus iron sulfides, oxides and silicates. Topaz- and muscovite-bearing assemblages can occur in greisenized areas, but they are minor compared to the abundant aluminous assemblages (micas+fluorite±plagioclase±diaspore±topaz) found with peraluminous granites. Tourmaline and other boron minerals can be abundant. Typical Be minerals are phenakite, bertrandite, danalite and helvite. Beryl and chrysoberyl are rare, but considerable Be can be bound in vesuvianite and other silicates (up to ~1% BeO, Beus 1966).

Notable examples of carbonate-hosted deposits include Pitkäranta, Russia, Iron Mountain, New Mexico, and a number of Sn skarn-greisen-replacement deposits in Tasmania and elsewhere in eastern Australia (e.g., Mt. Garnet, Queensland; see Appendix A and Fig. 2). This Sn-rich group has been thoroughly reviewed by Kwak (1987) and Newberry (1998). In comparison, a distinct group of Mo-W-bearing andraditic skarns contain only minor helvite (e.g., Oslo Rift, southeastern Arizona). In Tasmania, Be minerals occur in hydrous silicate assemblages that are developed after earlier calcic or magnesian skarns and they occur in iron sulfide/iron oxide-fluorite replacement bodies. Phenakite and bertrandite replace danalite in greisenized areas (Kwak and Jackson 1986). Kwak (1987) summarizes fluid inclusion data for the Tasmanian deposits which formed from high salinity (>30 wt % NaCl<sub>eq</sub>) fluids and spanned a large temperature range. Lower temperature variants on these deposits are best expressed in volcanic-associated Be deposits, however Bulnayevev (1996) links some moderate temperature (140-350°C) carbonate-hosted fluorite-bertrandite-phenakite deposits to subalkaline granitoids.

At Iron Mountain, New Mexico, 29 Ma porphyritic alkali granites intruded Paleozoic sedimentary rocks and formed oxidized Sn-, W-bearing andradite-magnetite skarns which are overprinted by helvite-bearing fluorite-rich assemblages (Fig. 22; Jahns 1944a). Skarns formed during several intrusive events (Robertson 1986) and the intrusive rocks contain abundant small veins consisting of quartz-K-feldspar, quartz-biotite-fluorite, or biotite. The Be ores form small bodies in skarn or marble and consist of rhythmically banded fluorite containing variable amounts of helvite, magnetite, vesuvianite, chlorite, and scheelite (Fig. 23). In his original description Jahns (1944b) coined the term “ribbon rock” for this texture while citing earlier descriptions at Lost River, Alaska and Pitkäranta, Karelia, Russia (formerly Finland). In New Mexico, contemporaneous volcanic rocks in adjacent fault blocks contain higher level bertrandite-rich mineralization (Meeves 1966), a few red beryl occurrences (Kimblar and Haynes 1980), and many volcanogenic cassiterite (wood tin) deposits (Rye et al. 1990). Southern New Mexico also hosts numerous other Be occurrences including helvite-bearing sulfide-carbonate replacement bodies at the Grandview Mine within 30 km of Iron Mountain (Warner et al. 1959).

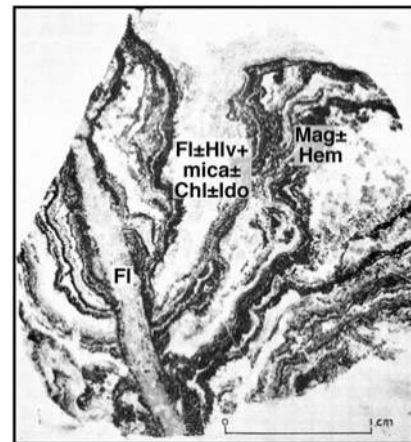
**Volcanic associated deposits:** Shallow low-temperature (150-250°C) replacement and vein deposits of Be are linked to volcanic and hypabyssal high-silica rhyolites and granite porphyries. These “epithermal” deposits are the main source of non-pegmatitic Be ore (Spor Mountain, Utah) and the sole source of gem red beryl (Fig. 2; Appendix A). Although best known in the Basin and Range Province of the western United States, similar deposits are reported from a number of areas in Asia (Zabolotnaya 1977; Kovalenko and Yarmolyuk 1995). Genetically related volcanic and hypabyssal biotite-

bearing topaz rhyolites (ongonites) are weakly peraluminous to metaluminous and are particularly rich in F, Be, and Li (Fig. 5C,D; Kovalenko et al. 1979; Burt et al. 1982; Christiansen et al. 1983). In the western United States they formed in an extensional tectonic setting and they belong to a compositionally expanded magmatic pattern that is bimodal in character and has felsic rocks that range from peraluminous to peralkaline.



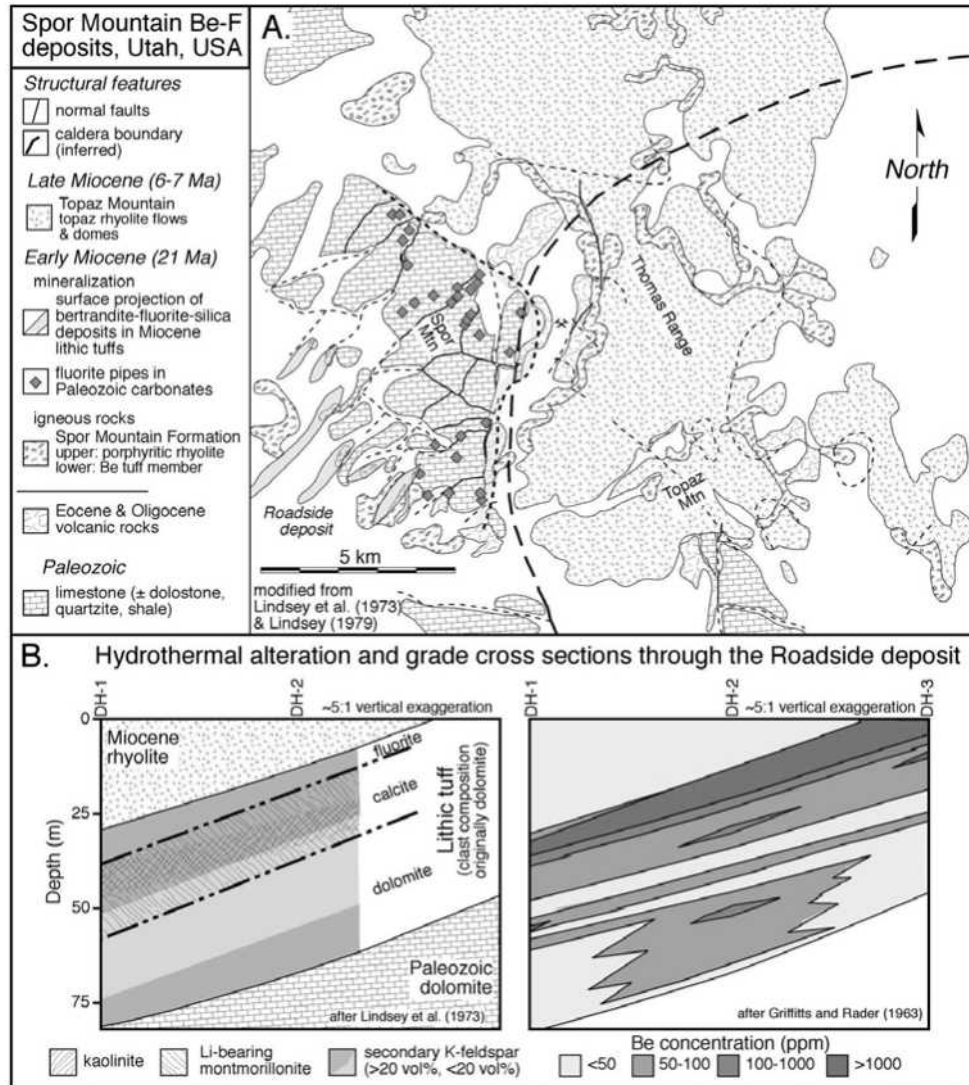
**Figure 22.** Geology of the Iron Mountain area, New Mexico (modified from Jahns 1946). (A) Simplified district geology showing extent of Fe-rich skarn and replacement deposits in Paleozoic carbonate rocks adjacent to Oligocene intrusions. The system is tilted about 30° to the east. (B) Detail of the central part of the district showing distribution of banded helvite-fluorite-iron oxide skarns at the marble contact or superimposed on earlier Sn-bearing andradite and W-bearing magnetite skarns. Be-F mineralization is related to the younger fine-grained granite (Robertson 1986).

**Figure 23.** Rhythmically banded helvite-rich replacement from Iron Mountain, New Mexico. Magnetite±hematite (dark bands) alternating with fluorite-helvite-vesuvianite (light bands; photo modified from Jahns 1944a,b).



Fine-grained bertrandite and, rarely, beryl or behoite are associated with hydrothermal silica (quartz, chalcedony, opal), calcite; fluorite, carbonate, K-feldspar and Li clays in tuffs and breccias. Fluorite, although abundant in some of the better mineralized deposits, is not always present. Although sulfides are absent, Zn, Mo, Li and other metals can be concentrated in Mn oxides and clays (Lindsey 1975). BeO concentrations range from a few hundred parts per million to a few percent.

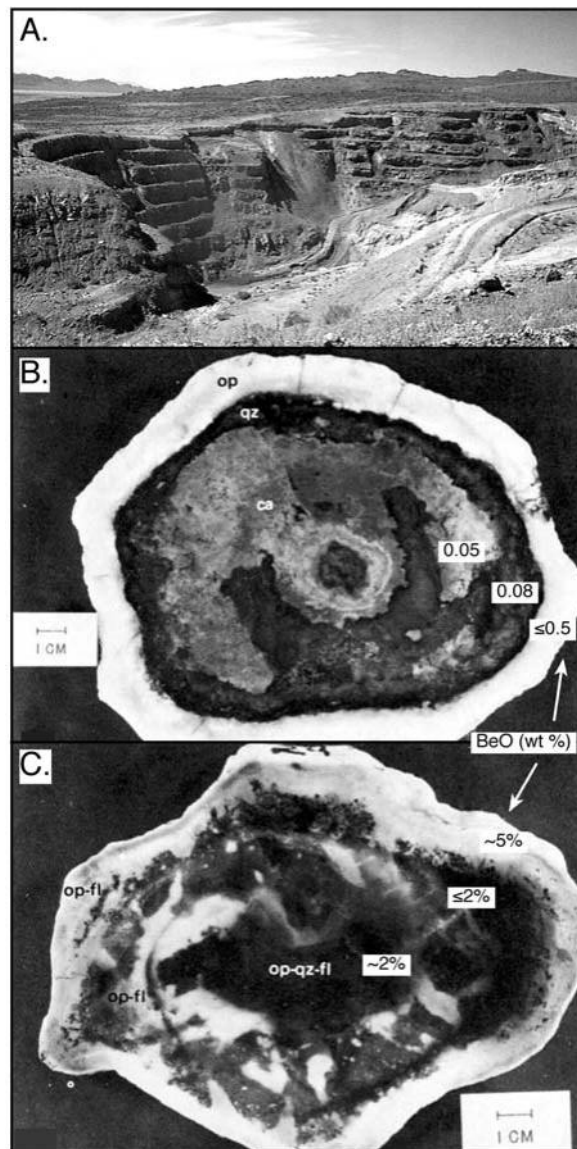
Fluorite-absent quartz/chalcedony veins with bertrandite-adularia-calcite±clays are described in Utah (Rodenhouse Wash, Griffiths 1965) and the former Soviet Union (Rozanov and Ontoeva 1987). Bertrandite-bearing (up to 2.5% BeO) clay-altered rhyolite tuff at Warm Springs, New Mexico also lacks fluorite (Hillard 1969). Most occurrences contain abundant fluorite in addition to silica minerals, bertrandite, calcite, K-feldspar and various clays. Zabolotnaya (1977) describes a quartz-fluorite-bertrandite±adularia stockwork with epithermal textures in subvolcanic Paleogene rhyolites. Similar deposits are present in Mongolia (Kovalenko and Yarmolyuk 1995).



**Figure 24.** Generalized geological relationships and mineralization in the Spor Mountain and Thomas Range area, Utah, locus of the world’s principal Be supply. (A) Geologic map of the district showing distribution of Be deposits in Miocene lithic tuff and regionally associated hydrothermal alteration (K-feldspar, fluorite and argillic types; adapted from Lindsey 1975). (B) Cross sections from the Roadside deposit (see A) showing types and distribution of hydrothermal alteration in early Miocene lithic tuffs and distribution of bertrandite mineralization in carbonate-clast-rich lithic tuff. The tuff is also enriched in Li, Zn and other elements. (Alteration from Lindsey et al. 1973. Be content from Griffiths and Rader 1963.)

The volcanic-hosted bertrandite-fluorite-silica ores of the Spor Mountain district are only one of several dozen Be occurrences in the region (Appendix A; Meeves 1966; Shawe 1966). At Spor Mountain and in the adjacent Thomas Range voluminous Cenozoic volcanic rocks overlie a carbonate-dominated Paleozoic sedimentary section (Fig. 24). Volcanism began with 39-38 Ma latites and andesites, followed by 30-32 Ma rhyolitic ash-flow tuffs. It culminated with early and late Miocene topaz rhyolite flows—the 21 Ma Spor Mountain Rhyolite, and the 6-7 Ma Topaz Mountain Rhyolite (Lindsey 1977). Interestingly, this bimodal distribution of topaz rhyolites occurs throughout western Utah. The topaz rhyolites have been intensively studied (e.g., Christiansen et al. 1984) and rocks of each episode have high concentrations of Be, F, Li and other lithophile elements (Fig. 5D). Uranium-lead dating of uraniferous silica yields an estimated oldest age of 20.8 Ma for Be mineralization but that younger thermal events were also likely (Ludwig et al. 1980).

Beryllium mineralization is localized in stratified tuff breccia immediately beneath rhyolite flows of Spor Mountain Formation (Figs. 24B and 25A). The tuff is extensively altered with a progression from regional diagenetic clays and K-feldspar (Lindsey 1975) to intense K-feldspathization with secondary sericite and smectite in the immediate vicinity of Spor Mountain (Lindsey et al. 1973). Dolostone clasts in the tuff breccia show corresponding alteration from original dolomite to calcite to silica to fluorite (Fig. 25B,C). Figures 24B and 25B,C show how Be grade increases with intensity of alteration in the tuff matrix and in carbonate nodules within the tuff. The source of the Be-bearing fluids is uncertain. Hydrogen and oxygen isotopic data (Johnson and Ripley 1998) are consistent with involvement of surface waters. Fluorite-rich, Be-poor ( $\leq 20$  ppm) breccia pipes cut the Paleozoic carbonate rocks beneath the older rhyolites. These pipes lie along structures that also appear to control the Be orebodies. Lindsey et al. (1973) speculated that a connection to deeper Be mineralization exists; however, the deposits could reflect shallow



**Figure 25.** Photos of spot Mountain mineralization. (A) Open pit with ore body at the bottom beneath unmineralized tuff (photo by Steve Young). (B) and (C) Two mineralized carbonate clasts from lithic tuff in the Roadside deposit illustrating the progression in alteration and BeO contents (cf. Fig. 7 in Lindsey et al. 1973). Mineralogy is labeled (qz = quartz, op = opal, fl = fluorite, ca = calcite), as are approximate BeO contents, (B) Partially replaced carbonate nodule from deeper part of the tuff with calcite core and quartz to opal outer zones with minor fluorite and bertrandite. (C) Carbonate clast fully replaced by fluorite-silica-bertrandite from the upper part of the tuff.



degassing of a magma without deeper mineralization, or they might have formed by leaching of the Be-rich Spor Mountain Rhyolite (cf. Wood 1992).

In the same region and of considerable gemological interest are occurrences of strongly colored, Mn-rich red beryl. These are restricted in occurrence to topaz rhyolites and were first described from the Thomas Range, Utah (Hillebrand 1905). In that area, small (<1 cm) red beryl occurs in gas-phase cavities along with topaz, bixbyite and quartz in 6-7 Ma rhyolites that overlie the Early Miocene rhyolites related to the Spor Mountain deposits. South of the Thomas Range, in the Wah Wah Mountains gem red beryl has been commercially produced from 22-23 Ma topaz rhyolite (Keith et al. 1994; Christiansen et al. 1997). In the Wah Wah Mountains, crystals exceeding 2 cm in length occur in cooling joints near the flow tops. Early bixbyite, topaz and silica polymorphs are followed by red beryl and ultimately kaolin plus mixed layer clays. The beryl is post-magmatic and is interpreted by Keith and coworkers to form by reaction of Be fluoride complexes released from the devitrifying rhyolite and subsequently react with feldspar and bixbyite along the joints. Red beryl also occurs the Sn-bearing rhyolites of the Black Range, New Mexico (Kimble and Haynes 1980) which are close in time and space to the hypabyssal felsic intrusions that are associated with the Iron Mountain skarns and nearby volcanic-hosted bertrandite deposits (Meeves 1966). Given the small amounts of beryl present, it appears that these occurrences require no more than local redistribution of Be from the host topaz rhyolites.

### **Peralkaline magma-related systems**

Sodic amphibole and sodic pyroxene-bearing granites, quartz syenites and nepheline syenites are associated with several large Be deposits and a number of mineralogically interesting occurrences (Appendix A). These are uncommon compared to deposits associated peraluminous and metaluminous igneous systems, likely reflecting the relative rarity of peralkaline magmas. This diverse group shares styles that range from magmatic pegmatitic assemblages to low-temperature (<200°C) hydrothermal systems hosted by a variety of rocks (Table 2; Fig. 4). As such they provide a useful mineralogical counterpoint to those deposits generated by more aluminous magmas. Consistent with the overall alkaline compositions, characteristic Be minerals include Na-Ca silicates (epididymite, chkalovite, leucophanite, milarite, leifite), gadolinite, and the Zn-Mn members of the helvite group (Table 1). Alteration assemblages in felsic rocks are dominated by Na(±K)-rich framework and chain silicates commonly with Li micas with or without quartz. Many of these assemblages can be termed fenites given that the key alteration minerals are K-feldspar or albite plus sodic pyroxenes and amphiboles. The local dissolution of quartz, the peralkaline silicates, and the rarity of acid assemblages provide a striking contrast with analogous post-magmatic metasomatism in peraluminous and most metaluminous systems (recall that Li-micas are not acid minerals; Eqn. 10). Carbonate-hosted systems differ less from the metaluminous environment. Fluorite-rich replacement bodies with or without skarns develop in carbonate rocks with characteristic Be minerals being phenakite, bertrandite, leucophanite and milarite among others.

In many regions, notably extension-linked alkaline felsic provinces like Nigeria, Norway and southwestern North America, metaluminous to weakly peraluminous biotite granites form concurrently with peralkaline aegirine- or riebeckite-bearing granites and syenites (Fig. 2). Where they coexist in the same igneous centers both peralkaline and metaluminous rocks may have rare element enrichments. Within the same intrusive suite, hydrothermal Be deposits more commonly form with the biotite granites, for example as in Nigeria, the Pikes Peak batholith, west Texas and nearby areas, and Norway (see Appendix A for references). Nevertheless, magmatic systems that are largely or entirely peralkaline do host major Be mineralization as described next.

**Peralkaline granites.** Beryllium mineralization generated by peralkaline granites ranges from pegmatitic to low temperature hydrothermal, paralleling the spectrum found in more aluminous systems. The hydrothermal systems have abundant K-feldspar and albite-rich alteration typically with quartz veining. Sodic amphibole, pyroxene and fluorides can be present as can be late hematitization and mica-rich greisens. Beryllium minerals tend to be late and are commonly distal. Zabolotnaya (1977) groups such deposits into helvite-group or leucophanite-bearing types with feldspathic metasomatism. Unfortunately, very few deposits are well described.

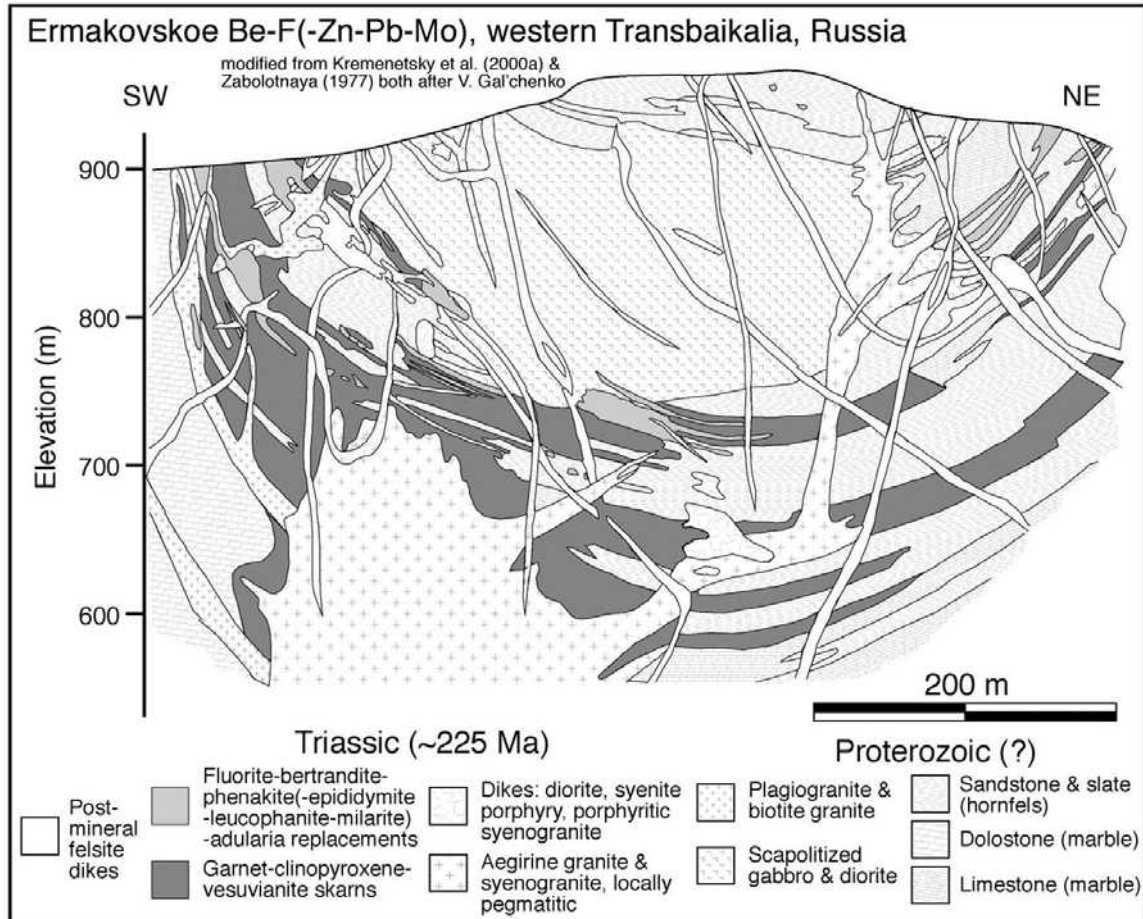
**Feldspathic host rock.** At the high-temperature end of the spectrum, Nb-Y-F-type pegmatites and miarolitic cavities are common. For example, pegmatites in the locally peralkaline granites and quartz syenites from Pikes Peak batholith, Colorado have phenakite, bertrandite, genthelvite, barylite, and gadolinite in pegmatites containing quartz, albite, amazonite, and Li-mica (Levasseur 1997; Kile and Foord 1998; cf. Russian localities: Bazarov et al. 1972; Nedashkovskii 1983). More acid associations containing muscovite and beryl are relegated to greisen zones in the metaluminous Redskin biotite granite (last section). In the pegmatitic Strange Lake deposit, Labrador, the original Zr-Nb-Y-Be concentrations are likely magmatic (Miller 1996). Nonetheless, high-temperature sodic pyroxene hydrothermal assemblages and moderate-temperature calcic hydrothermal alteration culminate in hematite-fluorite-Be mineral (leifite, gadolinite, milarite) assemblages (Salvi and Williams-Jones 1996). This paragenesis parallels the shifts to calcic assemblages that occur in undoubted hydrothermal systems (see Novikova 1983 and carbonate-hosted systems below).

Even though proximal mineralized pegmatites are common, the best developed Be concentrations in most of these systems are hydrothermal, late, and distal. Permian-age fluorite-phenakite-helvite(-gadolinite-milarite-barylite-bertrandite) mineralization occurs on the outer fringes of a hydrothermal system that is associated with variably porphyritic riebeckite granites in the Verknee Espee district in Kazakhstan (Belov and Ermolov 1996). These granites have pegmatitic facies and are feldspathically altered adjacent to proximal Nb-Ta-Zr-REE mineralization. The latter is hosted by K-feldspar-riebeckite-aegirine-fluorite±quartz veins with albite-riebeckite-aegirine-biotite envelopes in clastic rocks and tuffs. Hingganite-rich quartz-fluorite-albite-aegirine stockworks with up to 0.15% BeO occur in apical portions of aegirine-riebeckite granites at Baerzhe, Inner Mongolia (Wu et al. 1996). Analogous albite, fluorite and hematite-bearing styles of hydrothermal alteration formed with the Be-enriched magmas at the Khaldzan-Buregtey Zr-Nb-REE deposit in Mongolia (Kovalenko et al. 1995b).

Variations on this style occur in the northern Ukrainian shield, where genthelvite- and phenakite-bearing Be-Ta-Sn deposits are associated with cryolite-bearing biotite and riebeckite granites of the Mesoproterozoic Perga complex (Appendix A; Esipchuk et al. 1993; Kremenetsky et al. 2000a). Quartz-K-feldspar metasomatic rocks are common but greisen-type alteration with late sulfides, siderite and cryolite is also present (Vynar and Razumeeva 1972). Muscovite-bearing greisens and other acid assemblages in these systems are rare. Zabolotnaya and Novikova (1983) describe a possible example of more acid alteration from an unnamed Mesozoic occurrence in Siberia where dickite+bertrandite occurs with alkaline granites and quartz syenites in a hydrothermal system that is otherwise characterized by alteration to typical K-feldspar, albite, hematite, and fluorite-rich facies.

**Carbonate host rock.** Beryllium-bearing replacement bodies associated with peralkaline magmatic systems are fluorite-rich. Skarn alteration can be abundant or there may only be small amounts of calc-silicate minerals present. The calc-silicate-poor fluorite-bertrandite replacement bodies at Sierra Blanca, Texas and Aguachile, Coahuila

belong to a continuum with this group because they formed in peralkaline intrusive centers, even though the most closely associated intrusions are not themselves peralkaline. Calcic-sodic Be silicates such as leucophanite, meliphanite and milarite are common in some deposits where they either post-date phenakite or bertrandite or are the main Be phase (e.g., Novikova 1984).



**Figure 26.** Cross section through the aegirine-riebeckite granite-related hydrothermal system in the Ermakovskoe district, western Transbaikal, Russia. Fluorite-bertrandite-phenakite-leucophane mineralization is distal to and superimposed on garnet-pyroxene-vesuvianite skarns surrounding Mo-mineralized feldspathically altered alkali granites and granite porphyries (modified from figures based on the work of V. Gal'chenko presented in Zabolotnaya 1977 and Kremenetsky et al. 2000a). This deposit produced Be in the past (Kremenetsky et al. 2000a).

The best-described skarn-related example is the Ermakovskoe deposit in Transbaikalia, Russia (Appendix A; Zabolotnaya 1977; Lykhin et al. 2001). This deposit produced Be ore for the Soviet Union (Kremenetsky et al. 2000a). At Ermakovskoe, proximal skarns and skarn-overprinting to distal Be-F replacement bodies formed around and over a Triassic aegirine granite and related syenitic and granitic dikes (Fig. 26). The intrusion is variably albitized and potassically altered and it contains minor quartz-feldspar-molybdenite mineralization in its upper portions. REE and Zr are also metasomatically enriched. Early metasomatism in the sedimentary and igneous host rocks created K-feldspar-rich assemblages in the aluminous rocks and vesuvianite(beryllian)-garnet-pyroxene skarns in the carbonate rocks. Beryllium occurs mainly as phenakite (deep) and bertrandite (shallow) in late dark fluorite-adularia-calcite-ankerite replacement zones (Novikova et al. 1994). Subsequently, more calcic phases of alteration overprint

the system to form leucophanite, meliphanite, helvite, milarite and bavenite along with carbonates and sodic silicates. Fluid inclusion studies document a complex set of high-salinity, carbonate- and sulfate-rich fluids with Be contents approaching 1000 ppm (Reyf and Ishkov 1999). Kosals et al. (1974) report additional homogenization data from an unnamed deposit that closely resembles Ermakovskoe.

In another undisclosed Siberian location, leucophanite-fluorite-K-feldspar-albite veins and replacements occur in limestones along skarn-bearing contacts with a riebeckite granite (Kosals and Dmitriyeva 1973). Overprinting these is subordinate association containing fluorite, Li-mica, danalite, milarite, phenakite and bertrandite. Alteration in the granites and adjacent clastic rocks is dominated by riebeckite-albite-quartz and riebeckite-microcline-albite assemblages with Ta-Zr mineralization. Gadolinite joins leucophanite where rare fluorite veins cut the granite. The fluid inclusions studied by Kosals and Dmitriyeva (1973) indicate the leucophanite mineralization occurred between 380 and 490°C from variably saline fluids. In yet another unnamed deposit related to a aegirine-riebeckite granite, phenakite mineralization occurs in aegirine-alkali feldspar fenites and later fracture-controlled quartz-hematite-phenakite mineralization (Nedashkovskii 1970).

Fluorite in these systems is commonly dark in color and REE-rich (Hügi and Röwe 1970). Textures can be rhythmically banded or sieve-like with Be minerals and feldspars (Kosals et al. 1974; cf. Fig. 20).

***Silica-saturated peralkaline volcanic settings.*** Deposits with peralkaline volcanic rocks are apparently rare. The best candidate is the Early Proterozoic Brockman Nb-Zr-Ta deposit, Western Australia (Appendix A) where Be is remobilized into bertrandite-quartz-carbonate veins in a weakly metamorphosed rare-metal rich volcanic tuff. Other possible examples include some of the volcanic-hosted bertrandite-fluorite occurrences in Mongolia (Kovalenko and Yarmolyuk 1995) and the helvite-bearing Shixi occurrence in China which is associated with hypabyssal dikes of Na-altered Nb-Ta-Zr-F-rich sodic rhyolite (Lin 1985; Appendix A).

***Nepheline syenites and carbonatites.*** Hydrothermal Be enrichments may be more common with undersaturated igneous rocks than is generally appreciated. This is because the minerals are commonly dispersed in late igneous units or thin veins, they are easily weathered, and they are typically subtle in appearance. Chemically evolved units in nepheline syenite complexes commonly contain elevated Be in associated pegmatites, miarolitic cavities, hydrothermal veins and feldspathized rocks (Ilímaussaq, Mt. Saint-Hilaire, Lovozero, Khibiny, Oslo graben; Appendix A; Sørensen 1997). Where present (e.g., Mt. Saint-Hilaire, Lovozero), pegmatites can have complex internal structures and are dominated by albite, natrolite, sodalite with accessory chkalovite, leucophanite, and epididymite. In addition, nearly all described systems contain multiple types of quartz-free metasomatic assemblages that contain albite, aegerine, analcime and other sodic minerals in both veins and wall rock alteration. Li micas are common whereas muscovite is absent. The absence of quartz and the scarcity of K-feldspar contrasts with the hydrothermal features associated with peralkaline granites. Genthelvite, bertrandite, epididymite, chkalovite, leucophanite are the more common of the large number of hydrothermal Be minerals present. Beryl forms only where quartz-bearing peraluminous rocks are cut by Be-bearing veins (Mt. Saint-Hilaire, Horváth and Gault 1990) and chrysoberyl has been reported in quartz-absent, aluminous(?) xenoliths and veins from the Khibiny massif (Men'shikov et al. 1999); euclase not reported. These mineral associations are consistent with the phase relationships presented in Figures 11 and 12.

Geological relationships are well defined in only a few areas. Engell et al. (1971) describe Be distribution in the Ilímaussaq nepheline syenite complex. Hydrothermal

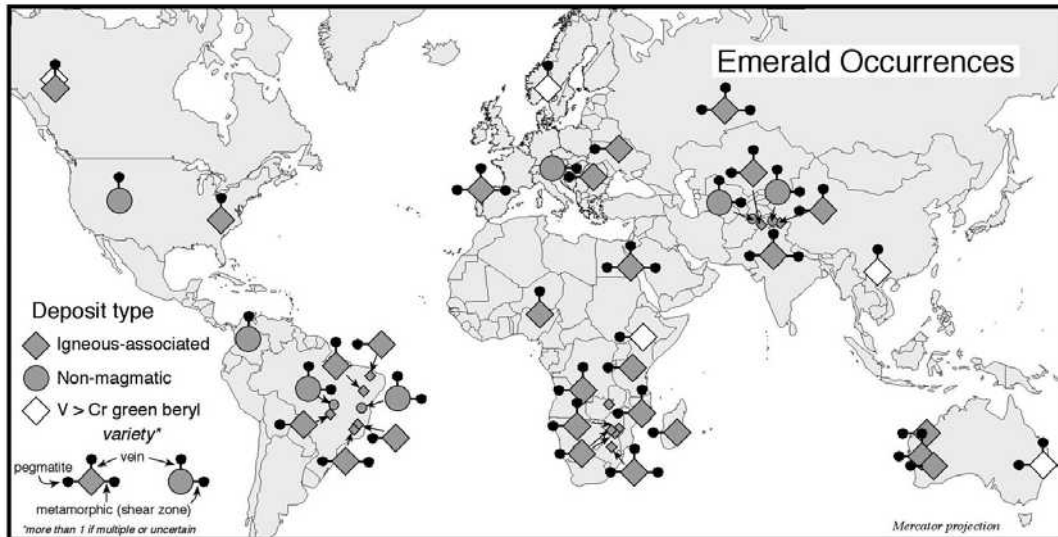
enrichments up to 0.1% BeO occur in zones with abundant veins. At Ilímaussaq more than a half dozen vein types contain combinations of aegirine, arfvedsonite, analcime, albite, natrolite, sodalite, Li mica, ussingite and other phases. These veins and associated albite-aegirine-arfvedsonite-natrolite fenites formed above late differentiates (lujavrites: eudialyte-acmite nepheline syenites) in the intrusion. The latter contain 10-30 ppm Be (30-80 ppm BeO). Hydrothermal Be minerals include chkalovite, tugtupite, bertrandite, beryllite among others (Engell et al. 1971; Sørensen et al. 1981; Markl 2001). Similar occurrences are known in the Lovozero and Khibiny complexes of the Kola Peninsula. Epididymite, leucophanite, chkalovite and many other Be minerals occur in hydrothermal albite-natrolite-polyolithionite-bearing assemblages in pegmatites and veins (Men'shikov et al. 1999). As at Ilímaussaq, these are preferentially associated with late-crystallizing lujavrites. At Letitia Lake, Labrador (Appendix A; Fig. 2), epididymite and berylite constitute up to a few percent of zones with Nb-REE-Zn-bearing alkali-feldspar-rich veins and fenites in shallow syenitic and trachytic rocks. These hydrothermal zones contain about 0.4% BeO. In the Khibiny-Lovozero Complex, Kola, isotopic evidence points to involvement of meteoric waters in the later veins in these otherwise magmatic-fluid dominated systems (Borshchevskii et al. 1987). The atypical abundance of hydrothermal quartz in the syenite-related deposit at Thor Lake, Canada (Fig. 16; described with magmatic deposits) may reflect the local granite host or a hidden quartz-bearing intrusion or simply the cooling of magmatic fluids to low temperatures.

Hydrothermal occurrences in carbonate host rocks are apparently rare. A plausible candidate is at Hicks Dome in southern Illinois where Paleozoic carbonate rocks host bertrandite-bearing fluorite mineralization (Baxter and Bradbury 1980; Kogut et al. 1997). The latter centers on breccias are linked to alkaline magmatism, possibly a carbonatite. Mineralization is interpreted as due to mixing of F-rich magmatic fluids with basinal brines (Plumlee et al. 1995).

Overall, relatively few reported Be enrichments occur with carbonatite complexes, even those with fluorite mineralization. At Muambe, Mozambique massive Y-LREE bearing fluorite with up to 1% BeO replaces the marginal parts of a carbonatite complex adjacent to K<sub>2</sub>O-rich fenitized shallow breccias (Appendix A). In the Magnet Cove, Arkansas, carbonatite complex, Erickson et al. (1963) found Be enrichments in late, thaumasite-bearing hydrothermal veins. At several other localities, Be enrichments occur in carbonatite-bearing alkaline complexes, but are most likely related to associated peralkaline syenites or granites. Berylite occurs in REE-Zr-Th-bearing fenitized rocks adjacent to nepheline syenites and carbonatites at Vishnevogorskii in the southern Urals (Zhabin and Kazakova 1960; Zhabin et al. 1960; Kogarko et al. 1995). Up to 0.6% BeO is reported with Zr-Y-Th-U-HREE breccia bodies associated with carbonatite dikes in the Coldwell alkalic complex, Canada, but are inferred by Smyk et al. (1993) on the basis of REE and rare metal enrichments to be derived from nearby felsic rocks.

### **Igneous-related emerald deposits**

Although amagmatic origins for emeralds are clearly established in some deposits (e.g., see Giuliani et al. 1997, Franz and Morteani, this volume), most investigators link the majority of emerald deposits to metasomatism driven by granitic magmatism (Fig. 27, Appendix A; Kazmi and Snee 1989a; Sinkankas 1989). In the latter group, emerald forms where berylliferous granitic pegmatites or granite-derived quartz-mica-feldspar veins intersect Cr(±V) bearing host rocks. Most such host rocks are ultramafic or mafic in composition. The emerald-bearing veins and dikes typically have complex contact-parallel quartz-poor metasomatic zones which are dominated by biotite(-phlogopite), other Mg(-Ca) sheet silicates, amphiboles, and plagioclase (Figs. 28, 29). Phenakite and chrysoberyl can accompany emerald mineralization in ultramafic rocks (Fig. 14).



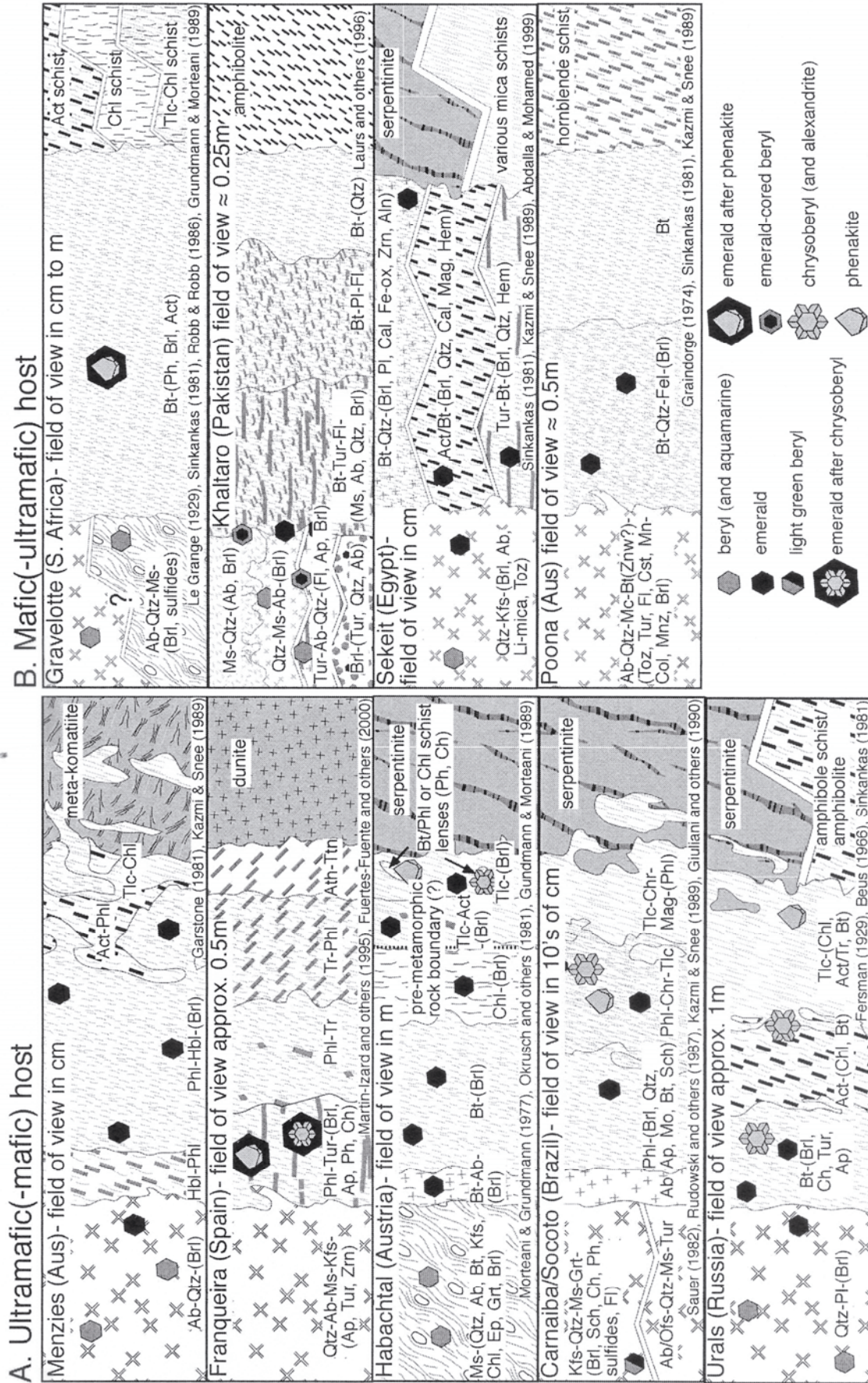
**Figure 27.** Global distribution of emerald and vanadian beryl deposits by geologic type. Compiled from Kazmi and Snee (1989a) and other sources. See text for discussion of and uncertainties in classification.

Mineralogically, these emerald-bearing systems have many parallels with apocarbonate greisens reflecting formation in analogous silica-poor, Mg-Ca rich hosts. In contrast to the ultramafic and mafic-hosted Cr-rich types, vanadian green beryls ( $V > Cr$ ) typically form in pelitic rocks which have higher V/Cr than most mafic and ultramafic rock (Fig. 27). As many workers have pointed out, the rarity of emerald stems from the infrequent pairing of a beryl-forming environment with rocks that contain significant Cr. That said, many areas have this juxtaposition, yet lack reported emeralds and thus should be prospective for them.

Few emeralds form with metaluminous granitoids. Perhaps the only significant examples are the greisen-affiliated deposits related to the Nigerian Younger Granites (Abaa 1991; Schwarz et al. 1996). Syenitic occurrences are virtually non-existent because beryl does not form in these systems except where pelitic host rocks become involved (e.g., V-bearing “emerald” at Eidsvoll, Norway; ordinary beryl in hornfels at Mt. Saint-Hilaire, Quebec). Most deposits arise from peraluminous biotite and biotite-muscovite granites; this connection likely reflects the elevated  $Al_2O_3$  activities that are required to make beryl (Fig. 12). Within this last group, emerald occurs with multiple magma and vein types. Genetically related granites and pegmatites can be chemically specialized as indicated by associated Li, Ta or Sn minerals (e.g., Hiddenite, North Carolina; Poona, Australia; some Brazilian and Egyptian deposits), however many of the better documented examples are associated with mineralogically simple granites and related veins (Reft River, Russia; Khaltaro, Pakistan; Egypt; Carnaiba and Socoto, Brazil; Crabtree, North Carolina; cf. Giuliani et al. 1990).

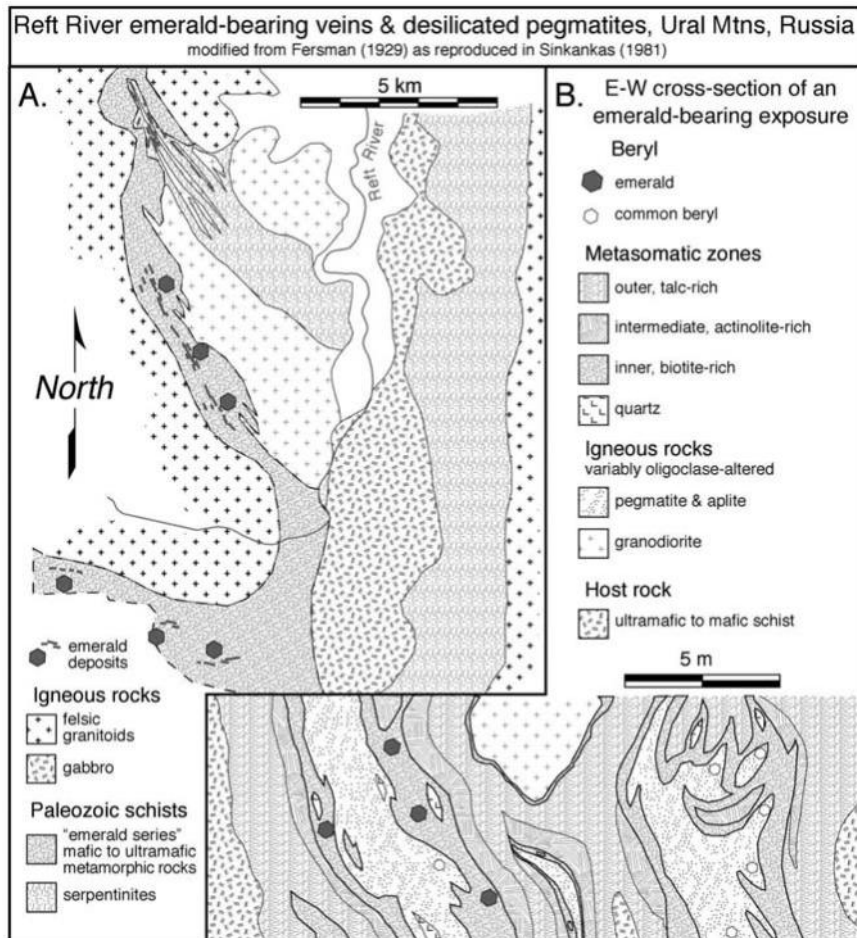
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**Figure 28 (opposite page).** Zoning in igneous-affiliated emerald deposits showing the diversity of source units (quartz veins, aplitic and pegmatitic dikes, felsic schist) and host rocks (silica-rich and silica-poor ultramafic and mafic compositions). For several of these occurrences metamorphic origins that long post-date magmatism have been proposed (Habachtal, Gravelotte, Franqueira, Sekeit; also see discussion under metamorphic occurrences). Compiled from sources listed on the figure.



In both ultramafic and mafic rocks, emerald crystals form both within hybridized veins and dikes and in their biotite-rich envelopes (Fig. 28). Veins typically show evidence of reaction with the country rock and can be dominantly quartz or mica (phlogopite or muscovite) or plagioclase with a wide-variety of additional minerals including K-feldspar, tourmaline, fluorite, molybdenite, and scheelite. Pegmatitic dikes are variably metasomatically modified (“desilicated”), typically expressed by lack of K-feldspar, abundance of plagioclase and diminished amounts of quartz. This process may take place concurrently with emplacement or during later metamorphism, as can generation of the common (though not universal) mineral foliations. Where intense chemical exchange between the host and incoming fluids has affected the entire mass it may be impossible to tell if the causative fluids were magmatic or hydrothermal. Aquamarine and common beryl may be present in the less reacted interiors of pegmatites and veins, whereas emerald is most common in enveloping zones of metasomatic biotite±plagioclase±quartz.

In peridotites or serpentinites, outer metasomatic zones consist of chlorite, talc, actinolite plus other amphiboles and they commonly contain minor chrysoberyl and phenakite. These “blackwall” assemblages are similar to those found worldwide at



**Figure 29.** Geology of the Reft River, Urals, Russia emerald deposits. (A) Regional geology showing distribution of emerald deposits in mafic to ultramafic metamorphic rocks adjacent to two-mica granite plutons. The deposits themselves are localized by dikes and veins from the granites. (B) Cross-section of an exposure with emerald and common beryl occurring adjacent to desilicated pegmatites. These form an apparent continuum with emerald-bearing quartz and/or albite veins (Beus 1966, p. 236-244). Both A and B redrawn after figures from Fersman (1929) as illustrated in Sinkankas (1981).

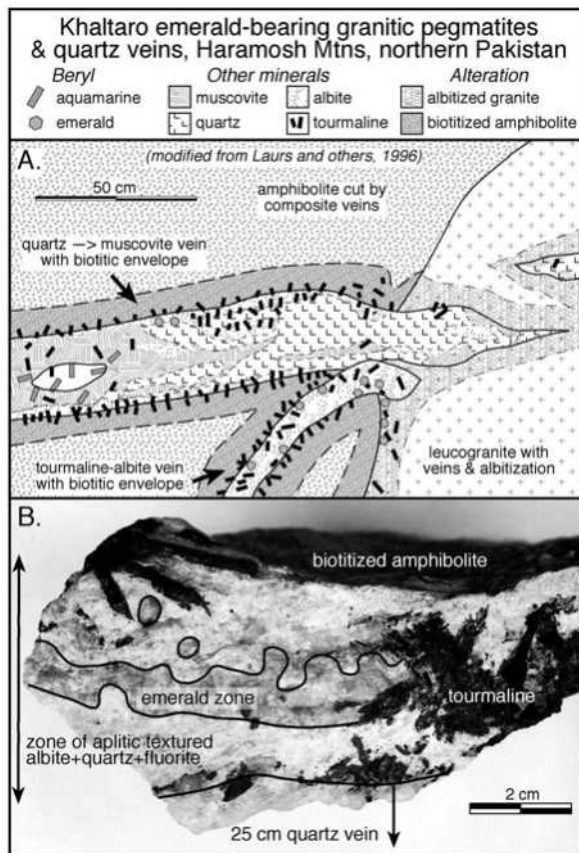


contacts between felsic and ultramafic rocks (e.g., Coleman 1966). Outer zones in ultramafic hosts may be relatively siliceous, for example with talc-bearing assemblages (Gravelotte), however serpentine or olivine bearing silica-undersaturated assemblages are more common. These silica-poor rocks have the best-developed chrysoberyl and phenakite which form in the intermediate to outer metasomatic zones (Beus, 1966) and may in turn be overgrown by emerald (e.g., Martin-Izard et al. 1995; Fig. 28). Desilicated units can also contain rare Be minerals including bromellite (Klement'eva 1970), bavenite (Kutukova 1946), and epididymite and milarite (Černý 1963). Surinamite, which replaces chrysoberyl+talc at high-pressure (Hölscher et al. 1986), may represent the high pressure or metamorphosed equivalent of lower-pressure blackwall assemblages.

One of the better-studied ultramafic-hosted districts is along the Reft River in the Ural Mountains in Russia where pegmatites and veins derived from Devonian two-mica granites cut Paleozoic serpentinites (Fig. 29A). Emeralds are associated with both quartz-feldspar-mica pegmatites and mica (muscovite/phlogopite/margarite), plagioclase-mica, and quartz-albite veins (Beus, 1966; Vlasov, 1968). The mica-rich varieties have been termed “glimmerites.” Metasomatic zoning is typical of the class and is similar around all varieties of veins and dikes—inner biotite to intermediate actinolite-chlorite to outer talc zones (Fig. 29B). Emeralds are concentrated in the biotite-rich rocks whereas phenakite and chrysoberyl are more distal. Some of the world's finest chrysoberyl (variety alexandrite) comes from the intermediate zone in these mines. Egyptian, Brazilian, South African and Australian occurrences (Appendix A; Figs. 27, 28) are similar. The common presence of fluorite and F-bearing micas suggests that the Mg-Ca-rich host rocks triggered F precipitation, which may have led to the precipitation of Be and Al from solution to form emerald, chrysoberyl and other minerals (e.g., Soboleva et al. 1972).

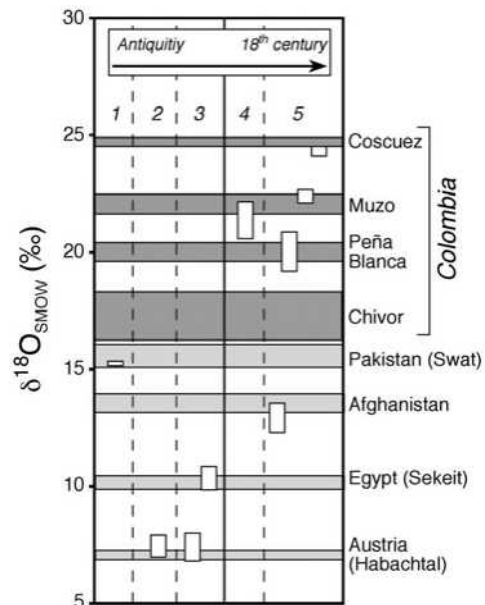
In contrast to the multiple metasomatic zones present in ultramafic host rocks, mafic rocks commonly develop only a biotite-rich envelope between the central veins or dikes and the host (Fig. 28). This likely reflects the higher activities of SiO<sub>2</sub> and CaO and a lower activity of MgO in mafic rocks (Fig. 14). For example, at Khaltaro, Pakistan emeralds formed at the contact between mafic amphibolites and young (~9 Ma) greisenized and albitized leucogranites (Fig. 30; Laurs et al. 1996). These include emeralds formed in the outer plagioclase-rich margins of aquamarine-bearing albite-quartz-muscovite veins and, rarely, in and around tourmaline-bearing simple pegmatites (Fig. 30). From whole rock analyses and modal mineralogy Laurs et al. (1996) deduced that the addition of F, alkalis and other elements to the amphibolite drove reactions making fluorite and biotite which in turn released the Cr and Fe needed for emerald formation.

Although magmatic-hydrothermal origins are well established in many areas, emerald paragenesis remains controversial in some regionally metamorphosed settings where igneous bodies are present. The dilemma stems from the fact that many occurrences are small bodies that could originate either (1) through local bimetasomatic exchange by intergranular diffusion over the relatively long times available during regional metamorphism or (2) by infiltration±diffusion metasomatism generated during shorter lived magmatic / hydrothermal events. In many areas a regional metamorphic overprint is absent and there is unambiguous evidence for magmatic conditions and geochemical signatures. Oxygen isotopic data can discriminate between some igneous-related systems and others of sedimentary or metamorphic origin (Fig. 31; Giuliani et al. 1997). These differences have been used with great success by Giuliani et al. (2000a) to deduce global sources and trading patterns in jewelry. For example, their results show that distinctive <sup>18</sup>O-enriched sedimentary-sourced Colombian emeralds spread throughout Europe within a decade of the discovery of the New World. In many cases, however, it is



**Figure 30.** Khaltaro, Pakistan emerald deposit associated quartz veins and pegmatites from muscovite-biotite leucogranite intruding amphibolites (Laurs et al. 1996). (A) Sketch of relationships between quartz and albite vein. Modified from Laurs et al. (1996). (B) Photograph of emerald-rich sample from the margin of a quartz vein adjacent to the amphibolite. Note that the high abundance of beryl (emerald) far exceeds melt solubility and must be hydrothermal. Photo provided by Brendan Laurs.

**Figure 31.** Oxygen isotopic compositions of emeralds for various deposits (vertical bars) and for different pieces of jewelry (numbered columns) as a function of fabrication age (redrawn from Giuliani et al. 2000a). This demonstrates the variability among in  $\delta^{18}\text{O}$  in emeralds, which is a function of geologic environment (source, indicated by horizontal bands), and it shows that this systematic variation can be applied to the interpretation of trade patterns in gemstones. See Figure 27 for locations.



not clear if one mechanism or the other dominates or if both are involved (e.g., Gravelotte, Franqueira, Seikeit; Appendix A). Other tests might help resolve the mode of origin of some deposits, for example, determining the extent of mobility of Al or accurate dating of intrusions, metamorphism and emerald formation. Unfortunately, for many emerald occurrences, these are not well documented and clearly merit further investigation. In the next section we turn to those deposits where non-magmatic origins are clearly indicated.

## NON-MAGMATIC OCCURRENCES

Non-magmatic processes are involved in many Be mineral occurrences, both in their genesis and in their subsequent modification. Local Be enrichments are noted in a varied group of Fe-Mn-oxide-rich rocks. Hydrothermal deposits related to sedimentary and metamorphic fluids comprise an important group of emerald deposits, notably the basin-related Colombian emerald deposits. Metamorphic and surficial processes redistribute Be through local concentration, mineralogical transformations, and placer processes. Some of these can be important gem sources—especially metamorphic emerald deposits and placer deposits of aquamarine and other materials.

### Fe-Mn(-Zn) oxide-rich occurrences

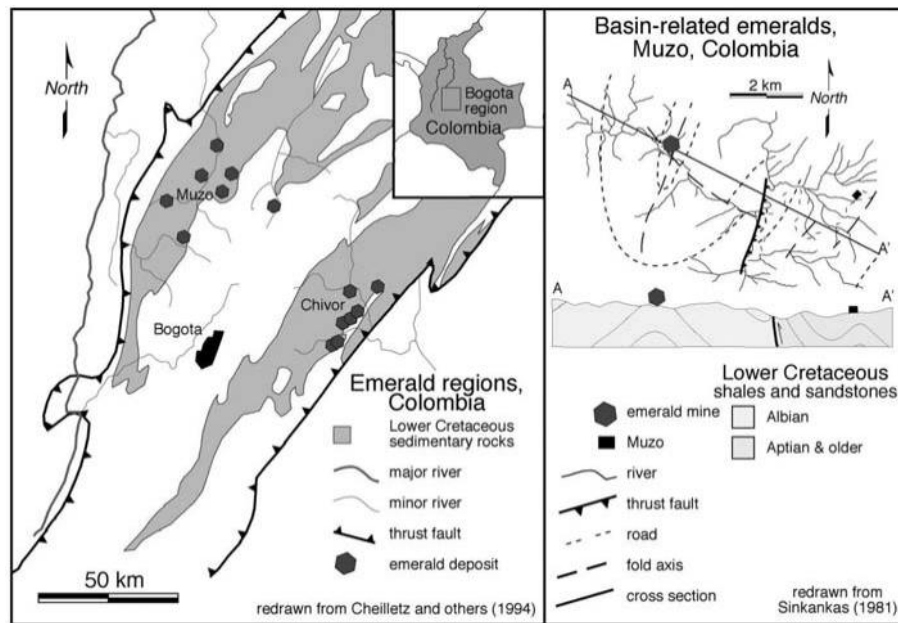
A handful of Fe-Mn( $\pm$ Zn)-oxide-rich rocks contain Be minerals or moderate Be enrichments. These occurrences clearly contrast with the obviously igneous-linked Fe / Mn associations discussed earlier which typically have magnetite-fluorite-helvite or sulfide-bearing helvite-Mn-carbonate/silicate $\pm$ fluorite assemblages. The oxide-rich systems occur in supercrustal rocks of both sedimentary and volcanic origin; the role of igneous activity in their generation is problematic (Grew, this volume). The best-known example is the metamorphosed deposit at Långban, Sweden (Moore 1970) which contains twelve Be minerals (Introduction, this volume). Similar rocks at Franklin, New Jersey (Palache and Bauer 1930) may have a granitic source for their Be. Miscellaneous locations include helvite with Mn-rich jaspers in the Pyrenees Mountains (Ragu 1994b), bergslagite and other Be enrichments with Mn-Fe oxide layers in Switzerland (Graeser 1998; Brugger and Gieré 2000), and milarite with Mn oxide deposits in New South Wales (Kawachi et al. 1994). Mn-Fe-oxide-rich hot spring deposits in Nevada (Golconda, Sodaville) contain up to 60 ppm Be (160 ppm BeO) along with W and other metals (Warner et al. 1959). Zasedatelev (1973) suggested that similar materials occur elsewhere. Beyond a common oxide-rich, sulfide-poor geochemistry, it is not obvious that they share a common genesis.

### Basinal (and metamorphic?) brine-related emerald deposits

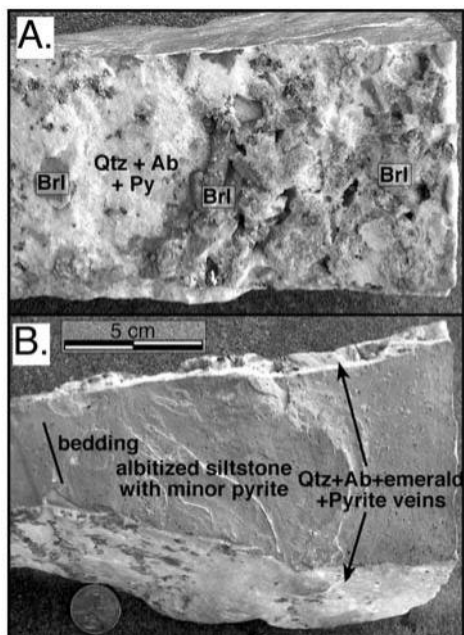
Beryl, mainly emerald with minor aquamarine, is known from several areas around the world that share moderate temperatures of origin, saline fluids, high  $\delta^{18}\text{O}$ , and a lack of associated igneous rocks (Appendix A; Figs. 27, 31). Of these, the Colombian emerald deposits are seemingly unique in their geology as they are in the size, quality and number of emeralds produced. In all of these deposits, the concentration of beryl is low and most authors infer that Be is derived by local concentration from the wall rocks (see also Franz and Morteani, this volume).

Emerald deposits in Colombia are hosted by Cretaceous siltstones, sandstones and shales and occur in two belts, one west and one east of Bogota (Fig. 32). The clastic host rocks belong to a Mesozoic sedimentary basin that contains marine evaporites. Mineralization in the western belt formed during Oligocene compression whereas mineralization in the eastern belt formed in the latest Cretaceous during an extensional episode (Cheilletz et al. 1994; Branquet et al. 1999). Beryl (mainly emerald, rarely aquamarine; euclase occurs at Chivor) occurs in gash veins with carbonate, quartz, albite, pyrite and other minerals in fractured, brecciated, variably albitized rocks (Fig. 33; Beus 1979; Cheilletz and Giuliani 1996). Grades are extraordinarily low (Fig. 3): probably <10 ppb gem emerald, and well under 100 ppb emerald overall even though the rocks themselves contain 500-5000 ppb Be (Feininger 1970; Kozłowski et al. 1988).

Fluid inclusion studies show that the mineralizing fluids contained up to 40 wt %  $\text{NaCl}_{\text{eq}}$  and were between 250 and 400°C (Kozłowski et al. 1988; Cheilletz et al. 1994; Giuliani et al. 1995, 1999; Banks et al. 2000). Stable isotopic studies of beryl and gangue



**Figure 32.** Geologic setting of the Colombian emerald districts. (A) Regional geology showing distribution of emerald districts in thrust belts surrounding inverted Cretaceous basin. Emerald deposits on the eastern side are latest Cretaceous in age and formed during extensional faulting, whereas on the western side they are localized by Eocene-Oligocene thrust and tear faults (Branquet et al. 1999). (B) Sketch of geologic relationships in the Muzo area showing correlation with compressional structures. Both figures redrawn from Sinkankas (1981).



**Figure 33.** Front view (A) and side view (B) of emerald-rich quartz-albite-pyrite veins cutting at high angles across albitized siltstone from Chivor(?), Colombia. Albitization is common in the Colombian emerald districts with concomitant redistribution of Be, Cr and other components (Beus 1979; Cheilietz and Giuliani 1996). Photo and specimen courtesy of Frank Mazdab.

minerals indicate a basinal brine source with high  $\delta^{18}\text{O}$  (16-25‰; Fig. 31) and sulfate reduction from sulfur to form pyrite. Hydrogen isotope ratios and fluid inclusion compositions indicate that two fluids were involved, a basinal brine and another, perhaps surface-derived fluid (Banks et al. 2000; Giuliani et al. 2000b). Basinal fluids are now generally accepted as the key ingredient in order to react with the host sedimentary rocks mobilizing the Be and Cr ( $\pm\text{V}$ ) required to form emerald (Beus 1979; Ottaway et al. 1994; Giuliani et al. 1999) perhaps with deposition due to mixing with a second calcium-bearing fluid (Banks et al. 2000). Local redistribution of Be by breakdown of Fe-Mn

hydroxides (Giuliani et al. 1999) or organic matter (Ottaway et al. 1994) is plausible, but not required.

Several other beryl occurrences have possibly analogous origins with basinal or metamorphic brines. In Uintah region of northeastern Utah, a single emerald crystal is reported from carbonate-rich veins that cut black shales and that have a basinal brine signature (Keith et al. 1996; Olcott et al. 1998). Gem beryl (aquamarine and emerald) occurs in the magnesite deposits of Brumado, Bahia, Brazil where it occurs with uvite, dolomite and topaz in quartz veins (Bodenlos 1954; Cassedanne and Cassedanne 1978). A basinal or metamorphic origin is conceivable given the association with bedded magnesite (which is commonly of basinal brine origin) and the lack of directly associated granitic bodies. A third candidate for basinal origin is the Mingora and nearby emerald deposits of the Swat district, Pakistan where beryl-carbonate-quartz veins cut ultramafic-bearing metamorphic rocks (Kazmi et al. 1989). Oxygen isotope ratios from Mingora are relatively high (Arif et al. 1996) and could indicate fluids either cooled from an S-type leucogranite or from a metamorphic or sedimentary source. Fluid inclusions reveal that temperatures ranged from 250° to 450°C and that the fluids contained  $\leq 0.03$  mole fraction of  $\text{CO}_2 + \text{CH}_4$  with salinities up to 20 wt %  $\text{NaCl}_{\text{eq}}$  (Seal 1989). From these results Seal (1989) inferred that fluid mixing may have contributed to beryl deposition. Some Afghanistani deposits (e.g., Panjsher) that lie westward along the Indus suture share many of the characteristics that are compatible with a sedimentary or metamorphic brine origin (Kazmi et al. 1989; Giuliani et al. 1997; Fig. 31). Nwe and Morteani (1993) interpret similar brines that occur late in the paragenesis of the Gravelotte, South Africa emeralds to be of metamorphic origin and to be linked to the most Cr-rich crystals.

### Metamorphic occurrences

A number of Be mineral localities, notably some emerald deposits and “alpine clefts,” occur in regionally metamorphosed rocks and lack evidence for magmatic involvement (Figs. 2, 27). Their origins are attributed to local redistribution and in some cases by introduction of new material during metamorphism.

**Shear zone or metamorphic emeralds.** Based on studies in the Alps and other regions Morteani, Grundman and coworkers have recognized that a number of emeralds and related minerals show compelling evidence for growth during regional metamorphism (Morteani and Grundmann 1977; Okrusch et al. 1981; Grundmann and Morteani 1989; Appendix A, Fig. 27). These occurrences have been termed metamorphic or shear zone deposits (Grundmann and Morteani 1989) and in some reviews they are combined with other non-magmatic deposits (Giuliani et al. 1998). Shear zone deposits resemble many igneous-related emerald deposits. Both types have emerald  $\pm$  phenakite  $\pm$  chrysoberyl in biotite-rich zones in metaultramafic and metamafic rocks. Although the biotite-rich zones commonly form at the contact with felsic lithologies, this is not always so. Textures show that beryl overgrows other metamorphic minerals, in some cases with curving inclusion trails clearly demonstrating synkinematic growth (Morteani and Grundmann 1977; Grundmann and Morteani 1989). As described above, these emeralds contain more Fe and Mg than most other types (Fig. 6).

At Habachtal, Austria emerald formed at the contact of metaperidotites and felsic gneisses. Pegmatites and quartz veins are absent, but the felsic gneisses contain ample Be to form the emeralds (Okrusch et al. 1981; Grundmann and Morteani 1989). There and elsewhere in the Tauern Window Mg-Fe-Na-rich beryl grew relatively late in the metamorphic history, commonly on preexisting phenakite (Franz et al. 1986). Fluid inclusions have moderate salinities and  $\text{CO}_2$  contents consistent with a metamorphic fluid (Nwe and Grundmann 1990). Similar deposits that lack closely associated igneous rocks occur in Afganistan, Pakistan and Brazil (Kazmi and Snee 1989a; Giuliani et al. 1990).

These systems commonly have CO<sub>2</sub>-bearing inclusions and non-magmatic oxygen isotopic values that are consistent with metamorphic origins (Giuliani et al. 1997, 1998; Fig. 31).

More controversial are deposits associated with pegmatites, veins and granites where Be was likely introduced as part of an igneous event, but emerald formation may significantly post-date magmatism. These include the deposits at Gravelotte, South Africa and Franqueira, Spain (Appendix A; e.g., Franz et al. 1996; Martin-Izard et al. 1996). Although textural and isotopic evidence can be compelling for metamorphic growth, the question of origin arises because available evidence does not necessarily preclude coeval magma emplacement. Synkinematic intrusion accompanied by progressive growth of metasomatic zones could lead to similar beryl overgrowth of phenakite during deformation. Conversely, metamorphism of pegmatitic or vein systems could drive additional local metasomatic exchange and emerald growth.

**Alpine clefts.** Of mineralogical interest only, alpine clefts comprise metamorphic gash veins and a variety of igneous-related cavities and veins, all of which can contain minor accessory Be phases (Appendix A; Stadler et al. 1973; Graeser 1998). Although the term “alpine cleft” has been widely applied to open-space mineral occurrences in other parts of the world (cf. Cook 1998), in order for it to be petrologically useful it should be restricted to types analogous to those seen in the Alps, especially the open-space metamorphic gash veins. The metamorphic occurrences represent a distinct environment for local concentration of Be where phenakite, milarite, bavenite and other minerals accompany quartz, chlorite, adularia, and hematite in open cavities that formed in moderate grade metamorphic rocks of felsic composition (Graeser 1998). Isotopic and fluid inclusion evidences indicate local metamorphic sources for Alpine-type veins (Luckscheiter and Morteani 1980; Mullis et al. 1994; Henry et al. 1996). In some ways this is analogous to the processes inferred for the Habachtal emerald deposits (above; Okrusch et al. 1981). In contrast to the phenakite, milarite assemblages of metamorphic clefts, the numerous Alpine igneous-hosted localities contain beryl, bazzite, bavenite, gadolinite and a variety of other minerals. The latter occur in aplites, miarolitic cavities, pegmatites and granite-related quartz veins of both Hercynian and Alpine age (Hügi and Röwe 1970; Pezzotta et al. 1999). Even among these occurrences, some may be primarily metamorphic in origin. The host felsic biotite granitoids that have modest Be enrichments in their differentiated phases (Hügi and Röwe 1970).

### **Weathering and placers**

With the exception of some placer settings, Be minerals are typically dispersed during weathering and sedimentary processes (Grigor'yev 1986). Local increases in soils can occur where Be dispersed in rock-forming minerals concentrates in montmorillonite or, more rarely, oxides and hydroxides during weathering (Sukhorukov 1989). For the most part, however, Be concentrations are diluted except when particularly stable Be silicates are present.

Beryllium minerals in the BASH group typically resist weathering and thus they occur in placer deposits. Alluvial deposits, particularly in Sri Lanka and related Gondwanan occurrences, are major sources of gem aquamarine, chrysoberyl and taaffeite-group minerals (Menon and Santosh 1995; Dissanayake et al. 2000; Shigley et al. 2000). These materials can be of unusually high quality because alluvial processes preferentially remove damaged (flawed) parts of crystals (Sinkankas 1981). Even in the world class Sri Lanka gem placers Be contents estimated to be 1 to 13 ppm comparable to or only slightly enriched over the local crust (Rupasinghe et al. 1984; Rupasinghe and Dissanayake 1985).

Supergene concentration is unimportant (Grigor'yev 1986). Local concentrations in soils, stream sediments and plants mainly represent residual accumulation and subsequent dispersion during weathering of hypogene mineralization (Sainsbury et al. 1967; Grigor'yev 1997). Beryl, phenakite and chrysoberyl typically do not weather; however, several Russian studies (Ginzburg and Shatskaya 1964; Novikova 1967; Grigor'yev 1997) have shown that beryl and phenakite can break down during weathering of pyrite- and fluorite-rich rocks. In their examples, the products are Be-enriched clay minerals and secondary phosphates such as herderite and moraesite. The authors interpret the process to be one of acid attack by sulfate and fluoride bearing supergene fluids. Given that the association of sulfides, fluorite and Be minerals is common, this process should be fairly widespread. Bacterial enhancement of weathering might also be important. Experiments show that bacteria can increase dissolution rates of helvite, beryl and chrysoberyl by factors of 5 or more without apparent Be toxicity to the bacteria (Mel'nikova et al. 1990). Large supergene accumulations are not reported, but minor upgrading might be expected where F-bearing acid groundwaters flow from Be-bearing rocks into mafic or carbonate rocks.

In contrast to the BASH group, other Be minerals such as those in the helvite, gadolinite and Na-Ca-silicate groups weather readily (Grigor'yev 1986). They react under near-surface oxidizing and weakly acid conditions to produce clays, oxides and some secondary silicates. Supergene changes are more widely reported in alkaline rocks, for example on the Kola Peninsula where hydrothermal chkalovite, epidymite and eudidymite weather to produce beryllite, bertrandite and other phases (Vlasov et al. 1966). These authors report no supergene enrichment.

### SYNOPSIS OF DEPOSIT CHARACTERISTICS AND ORIGINS

In contrast to pegmatitic deposits of Be, which are widely distributed in time but of rather limited geological variability, non-pegmatitic deposits are mainly Phanerozoic in age and are geologically diverse (Figs. 2, 4; Appendix A; cf. Rundkvist 1977; Zabolotnaya 1977). This diversity is all the more remarkable given the ready substitution of Be into many rock-forming minerals at crustal abundances and the overall ineffective concentration of large amounts of Be by crustal processes. The largest districts contain no more Be than that in a few cubic kilometers of granite or rhyolite (cf. Fig. 3) in contrast to deposits of many other elements (e.g., Cu, Au, Mo) which require 10s to 100s of km<sup>3</sup> of crustal source materials. This section summarizes the key patterns in deposit characteristics at global and system scales. These patterns can readily be interpreted in terms of experimentally determined Be mineral stabilities and well-known geologic processes. Ultimately, non-pegmatitic Be mineral occurrences form either (1) by concentration of Be during magmatic or hydrothermal processes, or (2) by local redistribution of Be during metamorphic or sedimentary processes.

#### Global-scale patterns

**With geologic setting.** As reviewed by others (e.g., Strong 1988), chemically evolved felsic magmas and related mineralization form in a wide variety of environments. Examples with prominent Be deposits (Fig. 2) include convergent margin compressional settings (e.g., the circum-Pacific Mesozoic), collisional to post-collisional transpressional to transtensional (e.g., the Indus suture, European Hercynian, Urals and central Asia), to extensional environments in rift or continental extensional environments (e.g., Norway, Niger-Nigeria, US-Mexico Cenozoic, Proterozoic alkaline systems across Laurentia). Amagmatic deposit types—whether metamorphic, basin-related or placer—require orogenesis to drive material transport via fluid flow (Colombian emeralds, Alpine clefts) or local diffusion and recrystallization (shear zone type emeralds, granulite-

sourced placers). Thus, all these settings generally share the requirement of relatively thick continental crust, which is needed for generation and differentiation of the diverse felsic magmas required for igneous-related deposits or for the crustal thickening required to drive regional metamorphism.

Other contributing factors to regional patterns are less well understood. For example, felsic magmatism and lithophile element (Sn-W-Mo-F-Zn) systems are far more widespread than documented Be mineralization. Provincial differences are striking: Southwestern North America has many Be deposits ranging in age from Mesoproterozoic to Cenozoic, yet is devoid of economically important Sn deposits. Conversely, the major Bolivian and Thai-Malaysian Sn provinces lack major known Be occurrences. Could crustal characteristics (thus inheritance) be important as may be the case for elements like Sn? There seems little compelling reason to think so, given that ample Be is present in most felsic igneous rocks; however, other petrogenetic factors could be key to generating a favorable geologic environment (e.g., shallow F-rich hydrothermal systems or a distinctive differentiation path). Of course, worldwide patterns of Be mineralization are poorly known due to the lack of systematic Be exploration.

**Over geologic time.** Examination of Appendix A and Figure 2 shows that non-pegmatitic Be deposits are mainly Phanerozoic in age (cf. Rundkvist 1977; Zabolotnaya 1977) and in this respect resembles the temporal distribution of other types of intrusion-related and epithermal mineral deposits (Meyer 1981). Most major occurrences are younger than 350 Ma with the prominent exception of Proterozoic alkaline-related deposits in Canada and eastern Europe. This age distribution reflects the preservation potential of the generally shallow crust in which Be deposits form and the typically thick and thus elevated crust. Preservation of volcanic and subvolcanic levels is not expected for these environments any more than it is in other epizonal terrestrial mineralization (Meyer 1981). Thus volcanic and hypabyssal systems are most common in the late Mesozoic and Cenozoic, whereas deeper-seated varieties make up most older systems. The increased proportion of alkaline-related systems with age may reflect preservation by the crustal extension that typically accompanies alkaline magmatism (cf. Barton 1996).

Principal episodes of Be mineralization are in the Proterozoic (1.6-1.0 Ga; Laurentia, Brazil), Devonian to Carboniferous (western Europe, Ural Mountains through the Altai into eastern Asia; Tasman belt), the late Mesozoic (180-65 Ma; northern circum-Pacific, east-central Asia; Nigerian rift; Colombia eastern belt), and the later part of the Cenozoic (35-5 Ma; western Colombian emeralds, southwestern North America, Indus suture, Alps). Other times can be regionally significant: for example, the Pan African (Late Proterozoic to early Paleozoic; Arabian-Nubian shield, eastern Brazil). Even within individual episodes the nature of the systems is diverse, as exemplified by the Cenozoic examples in southwestern North America. Thus, apart from the clear influence of preservation on the temporal distribution and the control exerted by magma types reflecting the pulse of orogenesis, the evidence shows that Be deposits have neither a compelling temporal progression nor a discernable global synchronicity.

### **System-scale patterns**

**Geological characteristics.** As emphasized by the organization of this review, the most distinctive features of igneous-related Be systems are their felsic, F-rich character, and the systematic variation of mineralogical characteristics that reflect the variation in igneous compositions from peralkaline to strongly peraluminous and from quartz-rich to quartz-undersaturated (Table 3; Fig. 5). Non-magmatic deposits are much less common. Many of the latter have evidence for local, wall-rock sources of Be. Consequently, very few of them represent large mass accumulations of Be, even though some comprise major gem deposits (e.g., Giuliani et al. 1999).



**Table 3.** Synopsis of the paragenesis of Be minerals in non-pegmatitic occurrences.

| <i>Occurrence type</i>                  | <i>Mineral associations &amp; abundance<sup>1</sup></i>                                                                                                                                                                                                                                                                                           | <i>Time-space distribution</i>                                                                                                                                                                                                                                                                                                            |
|-----------------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Magmatic</b>                         |                                                                                                                                                                                                                                                                                                                                                   |                                                                                                                                                                                                                                                                                                                                           |
| Li-Cs-Ta (LCT) magmas                   | • minor Brl (or Hrd) with Qtz, alkali feldspar, Li micas                                                                                                                                                                                                                                                                                          | • late magmatic phase                                                                                                                                                                                                                                                                                                                     |
| Muscovite(-beryl) granites              | • minor Brl                                                                                                                                                                                                                                                                                                                                       | • late magmatic or post-magmatic in vugs and fractures                                                                                                                                                                                                                                                                                    |
| Nb-Y-F (NYF) miarolitic pegmatites      | • minor Ph, Gad etc. with Li-Fe micas, alkali feldspars                                                                                                                                                                                                                                                                                           | • post-magmatic                                                                                                                                                                                                                                                                                                                           |
| Syenites                                | • dispersed in rock-forming minerals in late differentiates; rare Chk, Epd, etc. in pegmatites                                                                                                                                                                                                                                                    | • late magmatic phase or post-magmatic                                                                                                                                                                                                                                                                                                    |
| <b>Magmatic-Hydrothermal</b>            |                                                                                                                                                                                                                                                                                                                                                   |                                                                                                                                                                                                                                                                                                                                           |
| <b>Strongly peraluminous</b>            |                                                                                                                                                                                                                                                                                                                                                   |                                                                                                                                                                                                                                                                                                                                           |
| Li-Cs-Ta-F-Sn (LCT magmas)              | • rare Brl or Hrd in Qtz-Toz±Ms greisens or albitization                                                                                                                                                                                                                                                                                          | • early, proximal                                                                                                                                                                                                                                                                                                                         |
| Sn (-W) (Bt±Ms granites)                | • feldspathic host: (1) Brl-Ab-Ms-Fl; (2) Brl(-Brt-Euc) in Qtz-Ms-Fl±Toz±Tur veins; (3) uncommon Brl-Qtz-feldspar<br>• carbonate host: (1) Ch-Ph / Brt in Fl-mica±Dsp veins; (2) Hlv / Dn-Fl±Mag-silicate replacement                                                                                                                             | • (1) proximal (outer parts of intrusion), relatively early; (2) intermediate position and time; early and proximal<br>• both groups typically post-date and can be distal to Fe-rich garnet skarns; sulfides later                                                                                                                       |
| W-Mo (Bt+Ms granites)                   | • feldspathic host: (1) Brl common in Qtz-feldspar(-mica-Wlf-Mo) veins; (2) Brl-Ab-Ms-Fl; (3) Brl(-Brt±Euc) in Qtz-Ms-Fl±Toz veins; sulfide-Brt- Fl-carbonate veins<br>• carbonate host: (1) Brl-Ph or Brt in Fl-mica veins; (2) Ph-Brt-Fl-Kfs-Qtz-carbonate veins and replacements<br>• ultramafic/mafic host: Brl(emerald)-Ph-Ch with Bt-Pl±Qtz | • (1) early, proximal veins with central Brl; (2) commonly at intrusion margins, intermediate timing; (3) intermediate in time & position, Brl commonly distal; (4) late or distal<br>• (1) intermediate to late, can overprint in Grt-Px-Ido skarns; (2) distal may or may not be late<br>• Brl is central, other Be minerals are distal |
| <b>Metaluminous-weakly peraluminous</b> |                                                                                                                                                                                                                                                                                                                                                   |                                                                                                                                                                                                                                                                                                                                           |
| Mo(-W-Sn) (Bt granites & rhyolites)     | • feldspathic host: (1) Brl, Ph, Brt or Hlv in Qtz-Toz±Li-Fe mica or Kfs greisen veins; (2) Gnt-Brt-Qtz-Hem veins<br>• carbonate host: (1) Hlv or Ph with Fl-Mag-K-silicates; (2) Brt-Fl±Qtz ±clay replacement<br>• volcanic host: (1) Brt-Fl-silica-Kfs-clay-carbonate; (2) red Brl-Qtz-Mn-Fe oxides                                             | • (1) Be intermediate to late & distal; (2) late, proximal?<br>• (1) Be minerals late & typically distal after andradite-rich skarn; (2) distal, timing uncertain, skarn absent<br>• (1) distal, overall timing uncertain; (2) intermediate timing during cooling of flows                                                                |

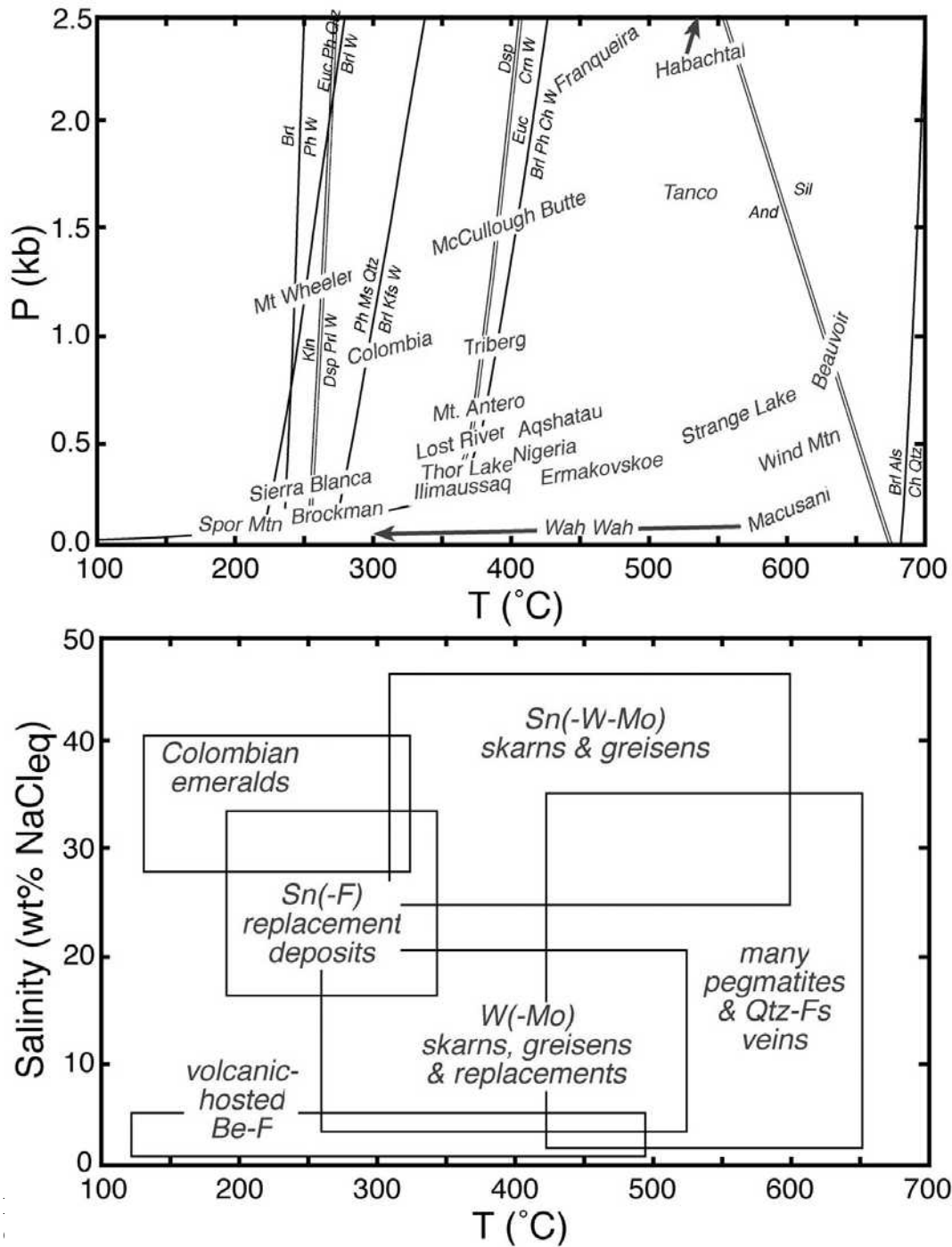
| <b>Peralkaline</b>                             |                                                                                                                                                                                                                                      |                                                                                                                                                                                                                                    |
|------------------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Nb-REE-Y-F (aegirine-Rbk granites & rhyolites) | <ul style="list-style-type: none"> <li>• feldspathic host: Gad-Ph-Hlv-alkali feldspar -Qtz±Li mica Rbk±Mag metasomatism</li> <li>• carbonate host: (1) Ph-Brt-Lph-Fl-Kfs-Qtz replacement; (2) Brt-Fl-clay±Kfs replacement</li> </ul> | <ul style="list-style-type: none"> <li>• proximal (pegmatitic) to distal (replacement) timing variable; proximal Nb-Ta-REE</li> <li>• (1) distal to / overprint Grt-Px Ido skarns; (2) distal in low-T shallow settings</li> </ul> |
| Nb-REE-Y (undersaturated syenites)             | <ul style="list-style-type: none"> <li>• feldspathic host: (1) Eud-Bar-alkali feldspar-Na amphibole; (2) Epd-Lph-Chk etc.-Ab-Anc±Sdl (3) Fl-Brt replacement</li> </ul>                                                               | <ul style="list-style-type: none"> <li>• (1) position &amp; timing unclear; (2) intermediate to late in host intrusion; (3) distal, timing unknown</li> </ul>                                                                      |
| <b>Non-magmatic</b>                            |                                                                                                                                                                                                                                      |                                                                                                                                                                                                                                    |
| Fe-Mn-oxide                                    | <ul style="list-style-type: none"> <li>• rare, mainly non-silicates</li> </ul>                                                                                                                                                       | <ul style="list-style-type: none"> <li>• discrete Be minerals are late</li> </ul>                                                                                                                                                  |
| Brines / basinal & metamorphic                 | <ul style="list-style-type: none"> <li>• sparse Brl ±Euc in Qtz-carbonate-feldspar-Py veins</li> </ul>                                                                                                                               | <ul style="list-style-type: none"> <li>• with main vein-forming event, temporal pattern uncertain</li> </ul>                                                                                                                       |
| Metamorphic deposits                           | <ul style="list-style-type: none"> <li>• shear zone: Brl(-Ph-Ch) with Bt(-Tlc-Act-Chl-Pl)</li> <li>• Alpine cleft: Ph-Mil-Qtz-Chl-Kfs-Hem</li> </ul>                                                                                 | <ul style="list-style-type: none"> <li>• near felsic contacts; syn-or post-peak metamorphism; some could be syn-magmatic</li> <li>• post-peak metamorphism</li> </ul>                                                              |
| Surficial deposits                             | <ul style="list-style-type: none"> <li>• minor Brl, Ch etc. in placers with other resistant minerals</li> </ul>                                                                                                                      | <ul style="list-style-type: none"> <li>• in alluvial systems downstream from pegmatite/granulite sources</li> </ul>                                                                                                                |

<sup>1</sup> Mineral abbreviations after Kretz (1983) or as in Table 1. Feldspathic hosts include granitoids and clastic sedimentary rocks.

As shown in Figure 34, a salient feature of non-pegmatitic Be deposits is the wide range of formation conditions and fluid compositions. Depths range from the surface to >10 km and temperatures from magmatic conditions to surface temperatures, and commonly exhibit a broad range within a single deposit. Fluids vary from hypersaline to dilute; they may or may not have CO<sub>2</sub> and/or CH<sub>4</sub> and have diverse redox states and acidities. These fluid compositional characteristics generally correlate well with igneous compositions and depth. Sn, W and base metal-rich systems tend to be acid, saline and shallow. Oxidation and sulfidation states directly influence alteration and metallic mineralogies. These variables also govern helvite group stability (Fig. 9) and the transition metal contents of beryl. Hydrothermal alteration assemblages directly control the identity of stable Be minerals primarily by imposing values of  $a_{Al_2O_3}$  and  $a_{SiO_2}$  (Figs. 11, 12).

The systematic progression in hydrothermal mineral associations resembles that in many other types of magmatic-hydrothermal ore deposits (Table 3). In feldspathic rocks early, high-temperature pegmatitic and mirolitic cavities form a geological continuum with feldspar-mica±quartz veins and sodic or potassic feldspathization. Later mica-rich and the latest-stage polymetallic assemblages form at somewhat lower temperatures and tend to be more extensive. Fluorine is typically abundant as evidenced by the presence of fluorite, topaz and/or F-bearing mica. Magmatic albite-rich rocks are common, however even more widespread is hydrothermal albitization which is clearly significant in many igneous-related systems and in the Colombian basin-related deposits.

Deposits hosted in carbonate-rich and ultramafic rocks have similar broad similarities, with differences in detail with host rock and magma type. Skarns form early. Vesuvianite, garnet and pyroxene typify calcareous hosts, whereas humite-group minerals and other magnesian silicates occur in magnesian carbonates (e.g., Kwak 1987).



Although there is a tendency for more reduced and pyroxene-rich skarns with increasing aluminum saturation index in associated intrusions, these patterns are complex as has long been recognized in other skarn systems (e.g., Einaudi et al. 1981; Newberry 1998). Skarn mineral assemblages may contain some Be but most ends up in later F-rich veins and replacements. As a class, low-T replacement deposits comprise the bulk of high-grade Be mineralization, whether as fluorite-silica-bertrandite after carbonate clasts at Spor Mountain, fluorite-diaspore-micas-chrysoberyl at Lost River, or fluorite-adularia-phenakite at Ermakovskoe and Mount Wheeler. Replacements of carbonate or skarn by fluorite and iron-rich oxides, sulfides and sheet silicates form another major group, typically with Sn-bearing hydrothermal systems. Reaction between veins and dikes in

ultramafic rocks leads to metasomatic assemblages dominated by micas, plagioclase, other sheet silicates, and amphiboles that are broadly analogous to the skarns and apocarbonate greisens. The lack of garnet and pyroxene in ultramafic-hosted deposits reflects their high Mg/Ca.

Mineral associations are one of the most prominent correlations with magmatic compositions and the phase equilibrium reasons were described at some length under “Beryllium Mineral Stabilities” above. To summarize the key features: Micas change from muscovite-rich in peraluminous systems to Li-bearing trioctahedral micas in more alkaline systems. Mica-rich assemblages are not created equal—trioctahedral-mica dominated greisens are not acid (e.g., low  $K^+/H^+$ ) assemblages (cf. Eqn.10). The latter thus should not be equated chemically with muscovite-bearing greisens or sericitic and related assemblages that occur in most porphyry Cu and Mo systems. BASH phases occur with peraluminous and some metaluminous granites; helvite group, phenakite and bertrandite with lesser beryl occur with metaluminous to peralkaline varieties and helvite group and alkali Be silicates in silica-undersaturated alkaline rocks (see Fig. 12). Amagmatic systems show a corresponding variation with alkalinity of the host materials: BASH phases form in aluminous rocks (e.g., Colombian and shear zone emerald deposits) whereas helvite group, phenakite and other Al-poor Be minerals form in Al-poor rocks (e.g., Fe-Mn oxide group).

***Time-space relationships.*** No single pattern captures the time-space distribution of Be minerals in hydrothermal systems. Beryllium may be precipitated early or late, proximally or distally (cf. pegmatites; see Černý, this volume). Be is proximal in more F-poor systems both in granitic pegmatites and in quartz-feldspar veins, but more commonly it is distal and late. In volcanic-related deposits, Be minerals (bertrandite, red beryl) are generally late and low temperature (e.g., western Utah), whereas in systems with clear magmatic enrichments beryl (in peraluminous rocks), epidymite (in peralkaline rocks) or other Be minerals are prominent in early stages (e.g., Mount Antero; Sherlova Gora, Sheeprock, Mt. Saint-Hilaire, Ilímaussaq, Lovozero), although a large suite of later Be minerals can crystallize through much of the sequence. The classic carbonate-hosted systems, regardless of the aluminosity of associated igneous rocks, tend to have Be concentrated in lower temperature assemblages, typically distal parts of the systems. Length scales can be meters to kilometers. Phenakite, bertrandite and, to a lesser extent chrysoberyl tend to be distal to beryl-bearing assemblages. Although one has the impression in many systems that there is a simple evolution from high-temperature, proximal assemblages to late and overprinting lower-T assemblages, patterns can be more complex with multiple events and reversals in sequence (Lost River, Shizhuyuan, Birch Creek).

### **Origins**

Ultimately, these patterns in the non-pegmatitic occurrence of Be minerals reflect original rock compositions, controls on solubility, and magmatic evolution at the source of fluids. Magmatic concentrations are relatively low due to the limited solubility of Be minerals in silicate melts. Hydrothermal deposits derived from such melts would have proximal Be and might lack significant Be in distal or low-temperature assemblages if aqueous complexing is weak (e.g., in the F-poor quartz-feldspar vein association in Table 3). When Be is strongly complexed (e.g., by F) and is not near saturation at its source, aqueous fluids could travel some distance before Be precipitation occurs in a favorable physical or chemical setting. Precipitation mechanisms include cooling, mixing, neutralization, and removal of F from the fluid. Wall-rock reaction, particularly with carbonate or mafic mineral rich rocks, would be particularly effective for the last two depositional mechanisms; this is abundantly illustrated by fluorite-rich deposits of many

types (Table 3). Differences in timing and zoning may also be due to evolution of the magma itself. Multiple magma batches, commonly of distinct origin, are well documented in many areas (e.g., Birch Creek, Cornwall, Beauvoir, Iron Mountain, Shizhuyuan).

The common features of Be-enriched igneous-related deposits are: (1) a magma that does not strongly remove Be in early formed minerals, and (2) a mechanism for generating and focusing F- and Be-bearing fluids. Melts with abundant Ca and Mg (i.e., most subalkaline types) will be unfavorable for magmatic enrichment because Be is partitioned strongly into sodic plagioclase (cf. London and Evensen, this volume). Likewise, mafic and calcic minerals sequester F in igneous systems either during crystallization (e.g., in biotite or apatite) or by formation of fluorite and other minerals on fluid release (cf. Barton 1987). For these reasons, igneous-related Be deposits are restricted to felsic, alkaline (low  $\text{CaO}/[\text{Na}_2\text{O}+\text{K}_2\text{O}+\text{CaO}]$ , cf. Fig. 5) igneous suites, but are not restricted in terms of alkalinity or silica saturation (in the sense of  $a_{\text{Al}_2\text{O}_3}$  and  $a_{\text{SiO}_2}$ ). Instead, the latter factors govern hydrothermal alteration types and Be mineral assemblages (Fig. 12), and are reflected in the suites of associated elements (Fig. 4; Table 3). Wall rocks as well as magmas may source Be in some of these systems, particularly where sodic plagioclase is destroyed and thus releases Be to the hydrothermal system (Beus et al. 1963). These processes deserve further investigation.

In contrast to igneous-related systems, in most non-magmatic deposits Be is only locally redistributed, Be concentrations are low, and the occurrences are mainly of mineralogical or gemological interest. Original enrichment in Be-bearing felsic rocks plays a role in some emerald deposits. Elsewhere, Be deposits are inferred to originate either by the action of aqueous fluids that release Be by mineralogical changes in the host rocks (e.g., Colombian emeralds, Alpine clefts) or by reworking by surficial processes (e.g., placer deposits).

### ACKNOWLEDGMENTS

Any undertaking of this scope depends heavily on the work of others and we can only acknowledge a fraction of the underlying science through citation. We thank our reviewers, Gerhard Franz, David Lindsey and Steve Ludington for constructive and thorough comments; they bear no responsibility for errors and oversights. Ed Grew's editorial oversight and patience is especially appreciated. National Science Foundation grants to MDB partially supported preparation of this paper (EAR 98-15032) and earlier work on Be-mineral stabilities and granite-related mineralization in southwestern North America (most recently EAR 95-27009). Additional support for compilation and drafting was provided through the Center for Minerals Resources and cooperative work on with the U.S.G.S. Global Mineral Resource Assessment Project. MDB thanks Brendan Laurs and the Gemological Institute of America for help during research, particularly for the use of their library. Brendan Laurs, David Lindsey, and Frank Mazdab provided photographs. Meg Watt, Jared Bond and Steve Hubbs helped with preparation of the manuscript.

### APPENDIX A: SUMMARY OF BERYLLIUM DEPOSITS BY REGION

(next eleven pages)

## Appendix A. Summary of beryllium deposits by region

| Province / Location                                                                                                      | Type*                           | Age          | References                                                                                                    | Comments                                                                                                                                                                                          |
|--------------------------------------------------------------------------------------------------------------------------|---------------------------------|--------------|---------------------------------------------------------------------------------------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <i>NORTH AMERICA</i>                                                                                                     |                                 |              |                                                                                                               |                                                                                                                                                                                                   |
| <b>Cordillera—Cenozoic</b>                                                                                               |                                 |              |                                                                                                               |                                                                                                                                                                                                   |
| <b>Southwest US—Mexico Tertiary</b>                                                                                      | G, aG, S, pG (sk, rpl, vol, ma) | CenM         | Warner et al. 1959; Meeves 1966; Shawe 1966; Burt and Sheridan 1988                                           | Mainly replacement and vein Hlv, Brl and Brt mineralization with mid- and early Cenozoic rocks                                                                                                    |
| Iron Mountain area: Iron Mountain district, Black Mountains, Apache Warm Springs, New Mexico, USA ( <i>see Fig. 22</i> ) | G(sk, volc)                     | CenM (29)    | Jahns 1944a,b; Robertson 1986                                                                                 | Iron Mountain: Hlv-Fl-Ves skarns (0.2Mt @ 0.2-0.7% BeO) superimposed on Mag-Sch-(Sn)-Adr skarn; Black Range volcanics have red Brl and wood Sn; Warm Springs have volcanic-hosted Brt             |
| Wind Mountain (Comudas Mountain), New Mexico, USA                                                                        | aS (m)                          | CenM (35)    | Meeves 1966; McLemore et al. 1996; Nutt et al. 1997                                                           | BeO bearing (0.1%) pegmatitic Ne syenite in district of shallow alkaline laccoliths; Eud identified                                                                                               |
| Sierra Blanca, Texas, USA                                                                                                | aG (rpl)                        | CenM (37)    | McAnulty 1980; Price et al. 1990; Rubin et al. 1990; Henry 1992                                               | Fl-rich replacement bodies in limestone adjacent to Li-Be-Zn-Rb-Y-Nb-REE-Th rich alkaline Cyl-bearing rhyolites; 850 kT @ 1.5% BeO as Brt, Ph (Bht, Bb, Ch); minor grossularitic skarn; clays+Anc |
| Aquachile, Coahuila, Mexico                                                                                              | aS,aG (rpl)                     | CenM         | McAnulty et al. 1963; Griffiths and Cooley 1978                                                               | Brt-Ad-bearing Fl replacement (17 Mt @ 0.1% BeO) adjacent to alkaline rhyolite & Rbk Qtz syenite                                                                                                  |
| Mt. Antero, Colorado, USA                                                                                                | G (m,gr,mp)                     | CenM         | Adams 1953; Sharp 1976; Kar 1991; Jacobson 1993b                                                              | Bt granite with Brl-rich aplite, miarolitic cavities with aquamarine to later Ph and Brt; greisen Qtz veins with Brl-MS-Mo-Toz; gem source                                                        |
| Silverton, Colorado, USA                                                                                                 | volc (vn-Mn)                    | CenM         | Warner et al. 1959; Casadevall and Ohmoto 1977                                                                | Minor Hlv in Mn-rich epithermal Ag-Au veins                                                                                                                                                       |
| <b>northern Basin &amp; Range Province (Utah etc.)</b>                                                                   | rV, G, #pG (rpl, gr, m, vn, sk) | CenM (MiocE) | Warner et al. 1959; Shawe 1966; Lindsey 1977; Burt and Sheridan 1986; Congdon and Nash 1991                   | Mainly western Utah Be belt, most occurrences in volcanic rocks, but also skarn (Hlv-Br) and granite-hosted occurrences—dominant Be producer; magmas up to 80 ppm Be                              |
| Spor Mountain, Utah, USA ( <i>see Fig. 24</i> )                                                                          | rV (rpl), [rV (vn)]             | CenM (21)    | Shawe 1968; Lindsey et al. 1973; Lindsey 1977; Ludwig et al. 1980; Baker et al. 1998; Johnson and Ripley 1998 | 21 Ma Brt in Fl-silica replacement of carbonate clasts in lithic tuff (7Mt @ 0.72% BeO), extensive Li-Zn-bearing Kfs and clay alteration; overlying 6-7 Ma topaz rhyolites have red Brl           |
| Wah-Wah, Utah, USA                                                                                                       | rV (vn)                         | CenM (23)    | Keith et al. 1994; Thompson et al. 1996                                                                       | Topaz rhyolite with late red Brt+Kln in fractures with early Mn-Fe oxides; main gem red Brt source                                                                                                |
| Rodenhous Wash, Gold Hill, Utah, USA                                                                                     | G(?) (vn)                       | CenM?        | Griffitts 1965; Shawe 1966; Lindsey 1977                                                                      | 3 km long Qtz-Ad-Cal-Brt (low-F) fracture zone, (> 1Mt 0.5% BeO)                                                                                                                                  |
| Sheeprock Mtns, Utah, USA                                                                                                | G(m, p, gr)                     | CenM (22)    | Cohenour 1963; Rogers 1990                                                                                    | Late magmatic, miarolitic and Qtz-greisen Brt in late phase of Bt(-Ms) granite (>1Mt, 0.01-0.1% BeO)                                                                                              |

| Sawtooth Batholith, Idaho, USA                    | G (p)               | Cen (Eoc)    | Sinkankas 1981; Kiiisgaard and Bennett 1985                                                                | Brl-bearing miarolitic pegmatites and Qtz veins in shallow Bt granite                                                                                                                                      |
|---------------------------------------------------|---------------------|--------------|------------------------------------------------------------------------------------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Cordillera—Cretaceous</b>                      |                     |              |                                                                                                            |                                                                                                                                                                                                            |
| <b>Seward Peninsula, Alaska</b>                   | pG (rpl, sk, gr)    | MesL (Cret)  | Sainsbury 1968, 1969; Swanson et al. 1990; Newberry 1998                                                   | Several Sn-W Bt granite centers with proximal Be-bearing skams and extensive distal Be-F-Al replacement deposits; similar in NE-most Siberia                                                               |
| Lost River district, Alaska, USA<br>(see Fig. 19) | pG (rpl, sk, gr)    | MesL (CretL) | Sainsbury 1968, 1969, 1988; Dobson 1982                                                                    | Extensive Ch-Dsp-Ms-Fl-replacement of carbonates other BASH minerals in Ms-Fl-Qtz veins (>3Mt @ 0.3-1.75% BeO) early and late relative to Sn-Fe-rich skams (±Hlv) and greisen in Bt granite                |
| <b>Northern Cordillera, Yukon &amp; Alaska</b>    | pG (sk, gr, vn) (E) | MesL (Cret)  | Mulligan 1968, 1971; Dick and Hodgson 1982; Anderson 1988; Newberry et al. 1990; Layne and Spooner 1991    | Early to Late Cretaceous granitoids with a wide range of commonly F-rich Mo-W, W-Mo-Cu and Sn-W systems; Brl and other Be minerals reported in a number of skarn and greisen/vein systems, also pegmatites |
| Crown (Finlayson Lake), Yukon, Canada             | (E) (pG, gr)        | MesL (CretE) | Groat et al. 2000a,b                                                                                       | Qtz-Tur±Sch veins associated with Ms±Bt granite have emerald where they cut Chl-mica schists (green V-rich Brl reported with Lened stock to north)                                                         |
| Logtung, Yukon, Canada                            | pG (gr, sk)         | MesL (CretE) | Noble et al. 1984; Anderson 1988                                                                           | large hornfels and skarn hosted stockwork W-Mo(-Bt) deposit with Brl in early Qtz-Sch-Mo-Ms veins                                                                                                          |
| <b>Great Basin (Nevada)</b>                       | pG (gr, sk)         | MesL (CretL) | Olson and Hinrichs 1960; Shawe 1966; Barton 1987, 1990; Barton and Trim 1991; Van Averbek 1996             | >20 Late Cretaceous Bt-Ms granites exposed at different levels with greisen, skarn and replacement mineralization with F-Zn-W(-Mo)                                                                         |
| Birch Creek, California, USA                      | pG (gr, sk)         | MesL (82)    | Barton 1987, 2000; Barton and Trim 1991                                                                    | Ms-Phl-Fl-Brl veins (cut F-, Al-rich W(-Cu) skarn) and Brl-bearing greisen veins (cut albization, Kfs-Qtz veins) in well-documented multi-event Bt-Ms granitic center                                      |
| McCullough Butte, Nevada, USA                     | pG(sk, gr)          | MesL (84)    | Barton 1982a; Barton et al. 1982; Barton and Trim 1991                                                     | Brl(±Bav-Brt) Ms-Phl-Fl-Pl Zn-W(-Mo) skams superimposed on earlier Grt-Px-Ves skarn (175Mt @ 0.027% BeO, 10% CaF <sub>2</sub> ) above greisenized Bt-Ms granite                                            |
| Mt. Wheeler, Nevada, USA<br>(see Fig. 20)         | pG (rpl, vn)        | MesL (85)    | Stager 1960; Lee and Erd 1963; Barton and Trim 1991, unpubl. data; National Treasure Mines, unpubl. report | Ph-Brt(-Brt) Fl-rich-Ad-Mn Sid-Qtz-Ms-Sch replacements and Qtz veins in basal carbonate unit above hidden intrusion (>0.2Mt @ 0.75% BeO, 20% CaF <sub>2</sub> , 0.2% WO <sub>3</sub> )                     |
| Oreana, Nevada, USA                               | pG (sk, p, vn)      | MesL (CretL) | Kerr 1938; Olson and Hinrichs 1960                                                                         | Brl-Sch in pegmatitic Qtz-Pl-Fl-Ms veins with Bt±Ms granite stock; mined for W (0.1 to 10% Brl)                                                                                                            |
| Uinta Mountains, Utah, USA                        | B (vn) (E)          | PhanL        | Keith et al. 1996; Olcott et al. 1998                                                                      | Emerald in Cal veins in 950 Ma shale; formed at 200-300°C from Mes or Cen oil-field brine (no magmatism)                                                                                                   |

## Appalachians

|                                      |                  |            |                                                                                                               |                                                                                                                                                                                                                                |
|--------------------------------------|------------------|------------|---------------------------------------------------------------------------------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Quebec-New Hampshire Alkaline</b> | S (p,vn), G (sk) | Mes        | Burt 1980; Eby 1985; Foland and Allen 1991                                                                    | Hot-spot-related Monteregian-White Mountain alkaline province has minor occurrences (notably Dn) plus spectacular locality at Mt. Saint-Hilaire                                                                                |
| Mt. Saint-Hilaire, Quebec, Canada    | S (p,vn)         | MesL (123) | Currie et al. 1986; Gilbert and Foland 1986; Currie 1989; Mandarino and Anderson 1989; Horváth and Gault 1990 | 15 (possibly 17) Be minerals (Epd > Lph > Gnt > others; cf. Introduction, this volume) from pegmatites, miarolitic cavities & veins in shallow, composite Ne gabbro to sdI syenite intrusion                                   |
| <b>Canadian Maritime</b>             | pG (gr, p, vn)   | PalM,L     | Mulligan 1968; Currie et al. 1998                                                                             | Scattered occurrences of Brl in granitic pegmatites, Qtz veins and greisens with or without W-Mo±Sn                                                                                                                            |
| <b>North Carolina</b>                | pG (p, vn) (E)   | PalM       | Griffitts 1954; Sinkankas 1981, p. 565-572; Wood 1996                                                         | Simple and Li-Sn pegmatites and Brl-bearing veins, some associated with granodioritic stocks (122kT BeO estimated in Brl @ 0.02-0.1% BeO); emerald associated with desilicated pegmatites (Spruce Pine) and Qtz-Ms veins (Hdd) |

## Proterozoic North America and Greenland

|                                                               |                  |                   |                                                                                                                                           |                                                                                                                                                                                                                                                                                                                                  |
|---------------------------------------------------------------|------------------|-------------------|-------------------------------------------------------------------------------------------------------------------------------------------|----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Pikes Peak Batholith, Colorado, USA</b>                    | (a)G, S (p, gr)  | ProtL (1.0 Ga)    | Gross and Heinrich 1966; Ludington 1981; Simmons et al. 1987; Levesseur 1997; Kile and Foord 1998; Smith et al. 1999                      | Variably rare-metal enriched, composite alkaline batholith; Gad, Ph, Hlv, Gnt, present in miarolitic pegmatites typically with Amz-Znw; local areas of F-rich veins and greisens with Brl-Brt                                                                                                                                    |
| Redskin Granite & Boomer Mine, Colorado, USA<br>(see Fig. 21) | G (gr)           | ProtL (1.0 Ga)    | Hawley et al. 1966; Hawley 1969; Desborough et al. 1980                                                                                   | Qtz-Ms-Toz greisens along margin of Redskin Bt granite (Pikes Peak derivative) were mined for Brl+Brt (>5Kt @ 2% BeO), minor Wlf-Cst-sulfides                                                                                                                                                                                    |
| <b>Arizona-New Mexico</b><br>Black Pearl & Tungstona mines    | pG (gr, p)       | ProtM (1.4 Ga)    | Jahns and Wright 1944; Heinrich 1957; Olson and Hinrichs 1960; Dale 1961; London and Burt 1978; Schmitz and Burt 1990                     | Greisen W(-Mo-Be) veins (Boriana, Black Pearl), apogranites, and complex pegmatites with 1.4 Ga granites; similar in style & age to pegmatites in Colorado & New Mexico                                                                                                                                                          |
| <b>Great Slave</b><br>Thor Lake, NWT, Canada<br>(see Fig. 16) | aG, S (gr, f, p) | ProtE (ArchL)     | Davidson 1978; Trueman et al. 1988; Kretz et al. 1989; Sinclair and Richardson 1994; Richardson and Birkett 1996; Taylor and Pollard 1996 | Thor Lake (T-zone) Ph-Brt(-Gad-Hlv) deposit (1.6 Mt @ 0.76% BeO) in Qtz-F1-PlI "pegmatite" with Na- and Kfs alteration; nearby Nb-Ta-Zr mineralization in syenitic breccia; both with composite alkaline Blachford Lake complex (2.1 Ga); in same area are older, deeper Be(-Li-B) pegmatites & Qtz veins around 2.5 Ga granites |
| <b>Nain</b>                                                   | S, aG (m, p, f)  | ProtM (1150-1340) | Richardson and Birkett 1996; Miller et al. 1997                                                                                           | Peralkaline volcanic and shallow intrusive rocks with widespread Be-Nb-Ta-Zr-Y-REE mineral occurrences                                                                                                                                                                                                                           |



|                                                    |                  |                   |                                                                                                                                                                             |                                                                                                                                                                                                                                                |
|----------------------------------------------------|------------------|-------------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| Strange Lake (Lac Brisson), Labrador, Canada       | aG (m, p, f)     | ProtM (1240)      | Hill and Miller 1990; Birkett et al. 1991; Boily and Williams-Jones 1994; Miller 1996; Richardson and Birkett 1996; Salvi and Williams-Jones 1996, 1997; Miller et al. 1997 | Lf-Gad(-Mtl) and Zr-Nb-Y mineralization (52Mt @ 0.08% BeO) in lenticular zones in rare-metal-rich granite in larger alkali granite center; low-T F-rich hydrothermal overprint may have helped to concentrate rare elements                    |
| Letitia Lake (Mann#1, Seal Lake), Labrador, Canada | S (f), aG (-)    | ProtM (1327-1337) | Heinrich and Deane 1962; Miller 1989; Richardson and Birkett 1996; Miller et al. 1997                                                                                       | Eud-Bar mineralization (1.8Mt @ 0.375% BeO) with Nb(-REE-Zn) in alkali Fs veins & feldspathic replacement in hypabyssal Agt-Rbk syenite and trachyte; alkali granites lack deposits                                                            |
| <b>Gardar Province, Greenland</b>                  | S, aG (m, p, vn) | ProtM (1340-1168) | Upton and Emeleus 1987; Woolley 1987; Finch 1990; Steinfeldt 1991; Macdonald and Upton 1993; Paslick et al. 1993; Goodenough et al. 2000                                    | Alkaline granites (Ivigut), apatitic Ne syenite complexes (Ilimaussaq, Motzfeldt), and alkaline volcanics; magmatic and hydrothermal Be enrichments are common in later parts of apatitic intrusions                                           |
| Ilimaussaq, Greenland                              | S (m, p, vn)     | ProtM (1168)      | Engell et al. 1971; Bailey et al. 1981                                                                                                                                      | Be enrichments in Agt-Ab-Anc etc. hydrothermal veins & fenites (Chk and 18 other Be minerals; cf. Introduction, this volume); plus modest late magmatic enrichment (in Iujavrite; >20Kt BeO in Kjanefield area with Li, Zr, Nb, Zn, REE, Y, F) |

SOUTH AMERICA

Andes—Cenozoic-Mesozoic

|                                                                                            |                  |             |                                                                                                                                                                                                                 |                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                               |
|--------------------------------------------------------------------------------------------|------------------|-------------|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Andean Altiplano</b><br>Macusani, Peru                                                  | rV (LCT, other?) | Cen (Mioc)  | Pichavant et al. 1988a, b; Lehmann et al. 1990; Dietrich et al. 2000                                                                                                                                            | Beryllium minerals apparently rare, but widespread Miocene rare metal enriched felsic intrusive and volcanic centers many with Sn-Ag(-B) ores, Fl is relatively uncommon; Macusani tuff is an end member                                                                                                                                                                                                                                                                                                      |
| <b>Colombia</b><br>(Muza, Chivor, Somondoco, Gachala, Peñas Blancas etc.)<br>(see Fig. 32) | B (vn) (E)       | MesL & CenM | Feininger 1970; Beus 1979; Cheilletz et al. 1994; Ottaway et al. 1994; Giuliani et al. 1995; Giuliani 1996, 2000b; Cheilletz and Giuliani 1996; Zimmermann et al. 1997; Branquet et al. 1999; Banks et al. 2000 | Emerald(@ 0.01-0.05 ppm BeO=rare aquamarine+Euc) in veins in variably albitized Cretaceous black shales and siltstones (2-10 ppm BeO); associated minerals include Qtz, Py, Cal, Na-plagioclase, Ms, Pas; two ages: Muza & other western belt deposits generated during Oligocene compression; Chivor & eastern belt during Late Cretaceous extension; fluid inclusion, isotopic, and other evidence for basinal brine source with sulfate reduction by hydrocarbons; evidence for fluid mixing in both belts |
| <b>Sierras Pampeanas, Argentina</b>                                                        | pG (p, gr/vn?)   | PalL (?)    | Morteani et al. 1995                                                                                                                                                                                            | Hercynian Brl-rich pegmatites and Brl-bearing Qtz veins                                                                                                                                                                                                                                                                                                                                                                                                                                                       |

| <b>Brazil—Proterozoic</b>               |                                |                       |                                                                                                                                           |                                                                                                                                                                                                                                                                         |  |
|-----------------------------------------|--------------------------------|-----------------------|-------------------------------------------------------------------------------------------------------------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|--|
| <b>Atlantic Shield, Brazil</b>          | M? B? pG (p, vn)<br>(E), G (f) | ProtE,<br>ProtL, PalE | Bilal and Fontelles 1988; Kazmi and Snee 1989a; de Medeiros Delgado et al. 1994; Sinclair 1996; Giuliani et al. 1997; Ribeiro et al. 1997 | Many Be deposits (gems, common Brl), mostly pegmatitic or vein related, possible shear zone and basin-related deposits (total Minas Gerais pegmatites: 106Mt @ 0.04% BeO); NYF, LCT and simple pegmatites; Hlv-Ph-Brl mineralization with albization with Sueri granite |  |
| Carmaiba & Socoto, Bahia, Brazil        | pG (p, gr, vn) (E)             | ProtE (1.9 Ga)        | Kazmi and Snee 1989a; Giuliani et al. 1990; Chellett et al. 1993                                                                          | Brl (-Mo-W) greisen, Qtz veins, and pegmatites associated with two-mica granites intrudes serpentinite-bearing metamorphic rocks locally forming emerald(±Ph)                                                                                                           |  |
| Santa Terezinha de Goias, Goias, Brazil | M? G? (vn) (E)                 | PalE (520)            | Kazmi and Snee 1989a; Giuliani et al. 1997; Ribeiro et al. 1997                                                                           | Emeralds developed in intensely folded Bt-rich schists & Tlc-carbonate schists; pegmatite rare & unrelated(?)                                                                                                                                                           |  |
| Brumado, Brazil                         | pG? B/M? (vn) (E)              | Prot?                 | Bodenlos 1954; Cassedanne and Cassedanne 1978; Kazmi and Snee 1989a                                                                       | Emerald/aquamarine in Mgs unit in metamorphic sequence; Qtz, Drv, Toz, Mg-Ca silicates associated; uncertain origin, others suggest hidden granite, basinal / metamorphic origin possible                                                                               |  |
| <b>Rondonia</b>                         | G (gr, f)                      | ProtM                 | Kloosterman 1974; Haapala 1995; Bettencourt et al. 1999                                                                                   | Ph ± other Be minerals with greisen-style Sn mineralization Bt granites; compared with Nigeria and broadly coeval rapakivi systems; cf. with Sueri granite in Goias (above)                                                                                             |  |
| <b>EUROPE</b>                           |                                |                       |                                                                                                                                           |                                                                                                                                                                                                                                                                         |  |
| <b>Alps—Cenozoic</b>                    |                                |                       |                                                                                                                                           |                                                                                                                                                                                                                                                                         |  |
| Alpine clefts - metamorphic             | M (vn)                         | Cen                   | Stadler et al. 1973 Hügi and Röwe 1970; Graeser 1998                                                                                      | Metamorphic Chl-Qtz-Ad fissures contain rare Be minerals (Ph, Brl, Mil, Bav etc., very rare Brl) typically in feldspathic host rocks                                                                                                                                    |  |
| Alpine clefts - granitic                | G (p, vn)                      | Cen                   | Stadler et al. 1973 Hügi and Röwe 1970; Graeser 1998; Pezzotta et al. 1999                                                                | Be-minerals (Brl, Bz, Bav, Gad) in pegmatites, aplites and veins associated with modestly Be-enriched granites                                                                                                                                                          |  |
| Habachtal, Austria                      | M (E)                          | Cen                   | Morteani and Grundmann 1977; Okrusch et al. 1981; Grundmann and Morteani 1989; Kazmi and Snee 1989a; Nwe and Grundmann 1990               | Emerald, Ph & Ch form at contact along metamorphic shear zone where ultramafic rocks after serpentinite are in contact with felsic lithologies; textural evidence for syn-metamorphic growth of emerald after Ph                                                        |  |
| <b>Western Europe—Paleozoic</b>         |                                |                       |                                                                                                                                           |                                                                                                                                                                                                                                                                         |  |
| <b>West-central Europe</b>              | pG (g, m, p, sk, LCT)          | PalL                  | many                                                                                                                                      | Brl, Ph, Brl, Euc, Hlv group, etc. widespread in greisen, vein, pegmatite, skarn occurrences with Bt±Ms granites; LCT (Toz) granites not directly linked to mineralization                                                                                              |  |
| Cornwall, England                       | pG (gr, sk, p, vn)             | PalL                  | Kingsbury 1961; Clark 1970; Embrey and Symes 1987; Manning and Hill 1990                                                                  |                                                                                                                                                                                                                                                                         |  |

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| Massif Central, France<br>e.g., Beauvoir<br>(see Fig. 15) | pG (gr, mLCT)          | PalL             | Burnol 1978; Stussi 1989; Cuney et al. 1992; Rimbault et al. 1995; Rimbault and Burnol 1998                                                | multi-stage LEM granite/veins; Be mainly in Li-mica; rare post-magmatic Brl and Hrd; multi-stage intrusion of progressively more evolved granitic magmas starting with main phase Ms-Bt granites                                                                                                                              |
| Erzgebirge, Germany / Krušné Hory, Czech Republic         | pG (gr)                | PalL             | Baumann 1994                                                                                                                               | Brl & Hrd are rare; found early in Sn-W-Li-F greisens, Be enrichments are present in some of the specialized granites                                                                                                                                                                                                         |
| Triberg, Schwarzwald, Germany                             | pG (gr)                | PalL (315)       | Markl and Schumacher 1996, 1997                                                                                                            | Zoned system in Bt±Ms leucogranite with Brl-bearing mirolitic cavities overprinted by Brl-bearing albitization & Qtz-Ms(-Fl-Toz-Cst) greisens, late Ph & Brl                                                                                                                                                                  |
| Western Iberia                                            | pG (mLCT, gr, p) (E)   | PalL             | Kelly and Rye 1979; Voncken et al. 1986; van Gaans et al. 1995; Fiertes and Martin-Izard 1998; Charoy 1999                                 | Present in Be-enriched magmas (Argemala), pegmatites (Franqueira), and W-Sn greisens (Panasqueira)                                                                                                                                                                                                                            |
| Franqueira, Galicia, Spain                                | pG (p) (E)             | PalL (Carb 315)  | Martin-Izard et al. 1995, 1996; Franz et al. 1996; Fiertes et al. 2000                                                                     | Emerald, Ch, Ph formed around margin of granitic pegmatite in ultramafic rocks; alternative interpretations of syn-intrusion or metamorphic (post-intr.) origin                                                                                                                                                               |
| <b>Oslo Graben</b>                                        | G S (sk, p, vn)        | PalL (Perm)      | Brinck and Hofmann 1964; Ihlen and Vokes 1978; Larsen et al. 1987; Larsen 1988; Jamtveit and Andersen 1993                                 | Be-minerals associated with veins, skarns and pegmatites associated with Oslo Rift-related Bt granites, monzonites & syenites; multiple parageneses, minerals include Hlv group, Brl, Epd, Brl, Lph, Bav, Bar, Brl, Ph, Gad (22 in all)..                                                                                     |
| Eidsvoll (Byrud Minnesund), Oslo region, Norway           | G(p) / S (p, vn) "(E)" | PalL (Perm)      | Kazmi and Snee 1989a                                                                                                                       | Green V(0.9%)>Cr(<0.1%) Brl in Cambrian Alum Shale intruded by Qtz syenite                                                                                                                                                                                                                                                    |
| <b>Kola Peninsula</b><br>Lovozero & Khibiny complexes     | aS (p, vn)             | PalM (360-420)   | Beus 1966, p. 244-255; Vlasov et al. 1966; Gerasimovsky et al. 1968; Borshevsckii et al. 1987; Kogarko et al. 1995; Men'shikov et al. 1999 | Lovozero & Khibiny appaitic syenitic intrusions have elevated Be overall reaching maximum in later units; complex pegmatites of various types follow and overlap with hydrothermal vein (e.g., Ntr-Fs-Pil); many Be minerals with varied paragenesis: Chk, Epd, Eud, Lph, Mil, Brl, Ttp, Lf, Bar, Brl, Byl, Gnt (15 in all).. |
| <b>Northern Europe—Proterozoic</b>                        |                        |                  |                                                                                                                                            |                                                                                                                                                                                                                                                                                                                               |
| <b>Karelia—Finland, Russia</b>                            | G (gr, sk, p)          | ProtM (1.4-1.65) | Haapala 1977b; Haapala 1988; Haapala and Ramo 1990; Sviridenko and Ivashchenko 2000                                                        | Be-bearing Sn(-W-Zn) F-rich greisens (Brl, Hlv, Brl at Kymi, Finland; Eujaroki, below), polymetallic Sn-W skarns (Hlv, Brl, Ph at Pitkänta, Russia)                                                                                                                                                                           |
| Eurajoki, Finland                                         | G (gr, vn)             | ProtM (1.57)     | Haapala 1977a; Haapala 1997                                                                                                                | Multiple phase Brl(-Hbl-Fa) granite to Toz-bearing leucogranite with Be-bearing Sn(-W-Mo) greisens (Qtz-Sid-Toz-Chl etc.) with Brl and Hlv; albitization present; unusual Grt-Gln greisens                                                                                                                                    |

|                                                                   |                                  |             |                                                                                                                     |                                                                                                                                                                                                                                                            |
|-------------------------------------------------------------------|----------------------------------|-------------|---------------------------------------------------------------------------------------------------------------------|------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>Ukrainian Shield</b>                                           | aG S? (f, vn)                    | ProtM       | Vynar and Razumeeva 1972; Zabolotnaya 1977; Sheremet and Panov 1988; Esipchuk et al. 1993; Kremenetsky et al. 2000a | Feldspathic metasomatic Be deposits with Gnt at Perga, Ukraine (>10Kt Be @ 0.54% BeO) and Diabazovoe, Belarus (>10Kt Be @ 0.3% BeO); in region of cyl-bearing rare-metal granites, syenites and Brl-bearing pegmatites                                     |
| <i>ASIA</i>                                                       |                                  |             |                                                                                                                     |                                                                                                                                                                                                                                                            |
| <b>Central Asia—Cenozoic</b>                                      |                                  |             |                                                                                                                     |                                                                                                                                                                                                                                                            |
| <b>Indus Suture—Pakistan, Afghanistan</b>                         | M(B?) (vn) (E); pG (p, vn) (E)   | Cen         | Kazmi and Snee 1989b; Seal 1989                                                                                     | Many emerald deposits of different styles along mafic/ultramafic-rich Indus suture; many other Brl-bearing pegmatites and granite-related veins                                                                                                            |
| Mingora (Charbagh, Makhad, Gujjar Kili), Swat, Pakistan           | M (B?) (vn) (E)                  | Cen         | Kazmi et al. 1986; Kazmi and Snee 1989a; Arif et al. 1996                                                           | Emerald in ultramafic schists with other lithologies nearby; Brl mainly with Qtz- or carbonate-rich zones with accessory Tur & Py; Mingora most important deposit                                                                                          |
| Khallaro, Gilgit, Pakistan<br>(see Fig. 3f)                       | pG (vn,p) (E)                    | Cen (Mioc)  | Kazmi et al. 1989; Laurs 1995                                                                                       | Emerald and aquamarine-bearing Qtz-Ms & Ab-Tur veins in Butized mafic amphibolite plus miarolitic pegmatites all with variably albitized Bt leucogranite                                                                                                   |
| <b>Eastern Asia—mainly Mesozoic</b>                               |                                  |             |                                                                                                                     |                                                                                                                                                                                                                                                            |
| <b>Transbaikalia / eastern Mongolia</b>                           | aG, pG, G, S, rV (rpl, gr, m, f) | Mes         | Kovalenko and Yarmolyuk 1995; Kremenetsky et al. 2000a                                                              | Many deposit types associated with mainly Mesozoic granitoids ranging from peraluminous to peralkaline, volcanic-hosted Be in Mongolia                                                                                                                     |
| Ermakovskoe, Buryatia, Russia<br>(see Fig. 2b)                    | aG (f, rpl, sk)                  | MesE (225)  | Zabolotnaya 1977; Reyf and Ishkov 1999; Kremenetsky et al. 2000a                                                    | Fl-rich Ph (deep) to Brl (shallow) (± Lph, Eud, Mil) replacement deposits (>10Kt contained BeO @ 1.3% BeO) distal to variably porphyritic Agt granite stock in carbonate-clastic-igneous host; Grt-Px-Ves skarn; F-bearing, Na-Kfs alteration in intrusion |
| Sherlova Gora, Transbaikalia, Russian                             | pG (gr,mp)                       | MesM (Jur)  | Beus 1966; Troshin and Segalevich 1977; Kosukhin 1980; Sinkankas 1981; Dukhovskiy 2000                              | Brl-rich Fl±Toz-Qtz greisen (Ms-Sid) veins ± miarolitic pegmatites associated with Bt leucogranites with modest Sn-W mineralization; gem locality                                                                                                          |
| Baerzhe, Inner Mongolia, China                                    | aG (f)                           | MesL (127)  | Wu et al. 1996                                                                                                      | Zr-Nb-HREE-Be deposit (0.02-0.15% BeO) with Hin ±Gnt in Ab-Fl-Qtz-Zrn altered Agt-Rbk granite                                                                                                                                                              |
| <b>Far East / eastern Yakutia, Russia</b><br>(e.g., Voznesenkoye) | pG (sk, gr, rpl)                 | MesL (PalM) | Bredikhina 1990; Kupriyanova and Shipanov 1997; Kremenetsky et al. 2000a; Rodinov 2000; Trumlina et al. 2000        | Mostly Cretaceous (PalM to CenM) Sn(-W) systems (Bt granites) with scattered occurrences of Be mineralization; Vozneskoye (PalM) has well developed Be-rich Fl-mica replacements in skarn (2-10Kt @ 0.06% BeO)                                             |

|                                             |                                   |                 |                                                                                            |                                                                                                                                                                                                                                                                                                                                |
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| <b>Southeast China</b>                      | pG (gr, sk, p, m?), V?            | Mes (M,L)       | Zhang 1992; Lin et al. 1995a; Kamitani and Naito 1998                                      | Many skarn, vein greisen W-Mo-Sn systems with minor to abundant Be mineralization related to Jurassic and Cretaceous Bt±Ms granites                                                                                                                                                                                            |
| Shizhuoyuan, Hunan, China                   | pG (sk,gr)                        | Mes (M?)        | Moh 1988; Mao et al. 1996a,b; Newberry 1998                                                | Polymetallic W(-Mo-Bi-Be-F-Sn) system with multi-stage greisen and F-rich skarn mineralization; Grt-Px-Ves skarn to Fl-Mag-rich hydrous skarns with W-Sn-Mo-Bi and distal Sn-Be-Cu, distal Hlv-bearing Mn base-metal                                                                                                           |
| Xianghualing, Hunan, China                  | pG (sk, gr)                       | Mes             | Lin et al. 1995a, p. 238-242                                                               | Albitized and greisenized Yanshanian polyphase Bt granite with Brl; Grt-Ves skarn, magnesian Mag-rich-chondrodite-Tur skarn in dolostone; Ch / Taf Fl(-Phl-fluorite- etc.) banded skarns                                                                                                                                       |
| Wangfengshan, Guangdong, China              | pG (sk, gr)                       | Mes             | Lin et al. 1995a, p. 242-244                                                               | Yanshanian volcanic and intrusive rocks, variably porphyritic intrusive rocks; intense albitization and fracture-controlled greisenization with Brl and Hlv-Dn early and Brl-Euc late                                                                                                                                          |
| Malipo County, Yunnan, China                | pG (p/vn) (E)                     | MesL (Cret L)   | Kamitani and Naito 1998; Zhang et al. 1999                                                 | V > Cr "emerald" associated with shallow pegmatitic veins with Tur and Seh in area of coeval granite-related W-Sn mineralization                                                                                                                                                                                               |
| Shixi, Zhejiang, China                      | rV (f, gr?)                       | Mes?            | Lin et al. 1995a, p. 254-255                                                               | Hypabyssal dikes of porphyritic sodic rhyolite, albitized and sercitized, high Be (as Hlv) and Nb, Ta, Zr, F                                                                                                                                                                                                                   |
| <b>SE Asia</b>                              | pG / G (p, sk, gr)                | Mes             | Manning et al. 1983; Pollard et al. 1995; Suwimonprecha et al. 1995; Linnen 1998; Zaw 1998 | Major Sn(-W) province related to chemically evolved Bt granites with pegmatites, greisens, skarns etc.; Be mineralization is rare except in pegmatites; generally low F                                                                                                                                                        |
| <b>Central &amp; Western Asia—Paleozoic</b> |                                   |                 |                                                                                            |                                                                                                                                                                                                                                                                                                                                |
| <b>Altai</b>                                | pG (gr, p, LCT), aG / S (m, p, f) | PalML, Mes E    | Kartashov et al. 1993; Kovalenko and Yarmolyuk 1995; Graupner et al. 1996; Kremenskiy 1996 | Be-bearing LCT and NYF pegmatites, greisen and feldspathic metasomatic rocks associated with peraluminous to peralkaline granites (± syenities)                                                                                                                                                                                |
| Kyzyltau, Mongolia                          | pG (vn, gr)                       | PalL (Carb 340) | Graupner et al. 1996                                                                       | Shallow W(-Y-Be-Mo) Qtz (-Fl-Wf-Brl) vein system (+ albitization, greisenization) with multiple phase intrusive complex of Bt granites                                                                                                                                                                                         |
| Khaldzan-Buregtey, Altai, Mongolia          | aS, aG (f, m)                     | PalM (Dev 385)  | Kovalenko et al. 1995b; Kovalenko and Yarmolyuk 1995                                       | Rare metal peralkaline granite phases in larger complex has Zr-Nb(-REE-Y-Be-Sn, Rb±Li etc) mineralization with Gnt & Bav; magmatic enrichment demonstrated by melt inclusions; post-magmatic metasomatism with formation of Pll, silicification, fluoritization, hematization etc. but rare metals only in enriched protoliths |

|                                                                                         |                       |               |                                                                                                                                                                         |                                                                                                                                                                                                                                                                               |
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| Verknee Espee, Kazakhstan                                                               | aG (f)                | PaLL (Perm)   | Belov and Ermolov 1996; Kremenetsky et al. 2000a                                                                                                                        | Proximal Nb-Ta-Zr-REE mineralization with Bt-Ab-Kfs-Rbk veins associated with altered Rbk-Agt granites; Ph-Hlv-Gad-Bar-Mil-Brt occur in distal Be orebody                                                                                                                     |
| <b>Central Kazakhstan</b>                                                               | pG (gr, sk)           | PaLL (Carb L) | Beus 1966, p. 293-298; Zasedatelev 1973; Burshtein 1996; Malchenko and Ermolov 1996; Serykh 1996; Dukhovskiy 2000                                                       | Be-bearing greisen and skarn deposits associated with Hercynian Be-enriched leucogranites; Mo(-W) stockwork East Qonyrat (Kounrad), Mo-W-Be-F skarn and replacement Qatpar (sk), W(-Mo) greisen Aqshantau, Qaraoba (Karaoba); unnamed Mag-mica-Fl-Ch                          |
| Aqshatau, Kazakhstan<br>(see Fig. 17)                                                   | pG (gr, vn)           | PaLL (Carb L) | Beskin et al. 1996; Matveev 1997; Ermolov 2000; Kremenetsky et al. 2000a                                                                                                | Large W(-Mo-Bi-Be) greisen system (16Kt contained BeO @ 0.03-0.07% BeO; Brl distal but early); complex zoned Qtz-Toz-Ms-Fl greisen in Bt granite and clastic rocks; late sulfides                                                                                             |
| <b>Ural Mtns, Russia &amp; Kokshetau, Kazakh.</b><br>Boevskoe, Orlinogorski, Elenovskii | pG, G (gr, sk, p) (E) | PaIM (Dev)    | Zabolotnaya 1977; Sinkankas 1981; Odintsova 1996; Grigor'yev 1997; Kremenetsky et al. 2000a; Letnikov 2000                                                              | Mainly Devonian magmatism: Urals—Ms-Bt granites with Be-bearing pegmatites, greisens & mica-Fl metasomatites (Mo-W type) (Boevskoe: >10Kt Be @ 0.12% BeO); Kokshetau—Bt granites with Brl-Ph-Hlv Sn greisens etc. (up to 0.2% BeO) and alkaline granites with REE-Ta-Nb-Sn-Be |
| Reft River (Marinsky, Troitskiy, etc.), Urals, Russia<br>(see Fig. 29)                  | pG (p, vn, gr) (E)    | PaIM (Dev)    | Fersman 1929; Zabolotnaya 1977; Sinkankas 1981; Kazmi and Shee 1989a                                                                                                    | Two-mica granites in mafic-ultramafic metamorphic host rocks with extensive Brl(-Ch-Ph-Brt-Bav)-bearing Bt-Fl-rich desilicated veins & pegmatites; major emerald occurrence                                                                                                   |
| <b>Southern Asia—Proterozoic &amp; Archean</b>                                          |                       |               |                                                                                                                                                                         |                                                                                                                                                                                                                                                                               |
| Sri Lanka (placers & pegmatites)<br>-- Gondwanan granulites                             | M, (p), P             | Arch / Prot   | Dahanayake 1980; Rupasinghe et al. 1984; Rupasinghe and Dissanayake 1985; Menon et al. 1994; Dissanayake and Rupasinghe 1995; Menon and Santosh 1995; Grew, this volume | Beryllium-bearing pegmatites and granulite-grade metamorphic assemblages (Brl, Taf, Ch); pegmatites may be syn- or post-metamorphic; major gem placers (1-13 ppm Be in placers is similar to bedrock); Pan-African age in adjacent southern India                             |
| <b>AUSTRALIA</b>                                                                        |                       |               |                                                                                                                                                                         |                                                                                                                                                                                                                                                                               |
| <b>Tasman Orogen—Paleozoic</b>                                                          |                       |               |                                                                                                                                                                         |                                                                                                                                                                                                                                                                               |
| <b>Tasman Orogen</b><br>Tasmania, New South Wales, Queensland                           | pG (sk, gr, repl, p)  | PaIM / PaLL   | Taylor 1978; Kwak and Jackson 1986; Blevin and Chappell 1995; Walshe et al. 1995                                                                                        | Five regions of Sn-W mineralization with Bt granites has sporadically Be-bearing greisen, skarn and replacement mineralization; some Brl-bearing pegmatites                                                                                                                   |

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| Tasmania, Australia—Mt Lindsay, Mt Bischoff, Moina, etc.                  | pG (sk, gr, rpl)                     | PalM / L (Dev-Carb)    | Kwak and Askins 1981; Solomon 1981; Kwak 1983; Kwak and Jackson 1986; Bajwah et al. 1995; Halley and Walshe 1995                     | Be minerals (Hlv, Dn, Ph) in complex Fl-rich skarns and sulfide replacement bodies with Sn(-W-Zn) mineralization (up to 0.5% BeO); distal to greisenized Bt granites; boron common; greisen overprint converts Dn to Ph / Brt              |
| Mole Granite (Emmaville, Torrington), NSW, Australia                      | pG (p, vn, gr) ("E")                 | PaIL                   | Taylor 1978; Sun and Eadington 1987; Kazmi and Snee 1989a; Plimer and Kleeman 1991                                                   | Brl widespread with early pegmatites and Sn(-W-Mo) veins & greisens with Mole Bt granite; deep green Brl from deposits in host rocks have both V and Cu > Cr                                                                               |
| <b>Western Australia—Early Proterozoic &amp; Archean</b>                  |                                      |                        |                                                                                                                                      |                                                                                                                                                                                                                                            |
| Brockman, Western Australia                                               | aVr m? rpl?                          | ProtE (1870)           | Ramsden et al. 1993; Taylor et al. 1995a,b                                                                                           | Hydrothermally altered Fl-bearing alkali rhyolite with Nb-Zr-Ta-Be enrichment (4.3Mt @ 0.08% BeO); predates orogenesis                                                                                                                     |
| Western Australia emeralds (Menzies, Poona, Pilgangoora, Wodgina, Curlew) | pG (p, vn?, LCT) (E)                 | ArchL                  | Sinkankas 1981; Kazmi and Snee 1989a; Partington et al. 1995                                                                         | Emeralds form in margin of variably desilicated simple and complex pegmatites in greenstones and ultramafic schists; other major LCT pegmatites in region (Greenbushes)                                                                    |
| <i>AFRICA / MIDDLE EAST</i>                                               |                                      |                        |                                                                                                                                      |                                                                                                                                                                                                                                            |
| <b>Africa - Nigeria-Niger</b>                                             |                                      |                        |                                                                                                                                      |                                                                                                                                                                                                                                            |
| <b>Nigerian-Niger Younger Granites</b>                                    | aG, S, G (gr, vn, p) (E)             | MesM (Jur), ProtL, Pal | Bowden 1985; Bowden et al. 1987; Matheis 1987; Imeokparia 1988; Kuester 1990; Perez et al. 1990; Schwarz et al. 1996                 | Alkaline granites, syenites and volcanic rocks with Bt granites (Jurassic in Nigeria older to north); Brl-, Gnt-, and/or Ph-bearing greisens, veins & albitites with Sn, Zn, Ta-Nb deposits; Pan African Brl-bearing pegmatites in Nigeria |
| Nigerian ring complexes—Saiya-Shokobo, Rirawai (Liruet)                   | G (aG, S) (f, gr, vn, p) (E)         | MesM (Jur)             | Bowden and Jones 1978; Kinnaird 1985; Bowden et al. 1987; Imeokparia 1988; Aboa 1991; Schwarz et al. 1996                            | Brl (including emerald, aquamarine), Gnt, Brt, Ph associated with Sn-Zn-Ta greisens (Li micas), veins and microitic pegmatites mainly with Bt granites but also with peralkaline intrusive rocks                                           |
| Taghouaji, Air Massif, Niger                                              | aS, G (vn, gr)                       | PalE (430)             | Bowden et al. 1987; Perez et al. 1990                                                                                                | Gnt-Brt-Hem with Zn-bearing silicates and Cu-Pb-Zn sulfides in Qtz veins associated with the only alkaline complex with Bt granite in Niger; Brl present in granite                                                                        |
| <b>Northern Africa - Arabia Pan-African</b>                               |                                      |                        |                                                                                                                                      |                                                                                                                                                                                                                                            |
| <b>Arabian-Nubian Shield</b>                                              | pG (gr, p) (E) / aG (m?, p) / M? (E) | ProtL (600-550)        | Jackson and Ramsay 1986; du Bray et al. 1988; Binda et al. 1993; Grundmann and Morteani 1993; Mohamed 1993; Abdalla and Mohamed 1999 | Late Proterozoic - early Paleozoic (Pan African) anorogenic granite province with multiple types of lithophile element suites commonly with Be enrichments; metamorphic emerald                                                            |

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| Sekcit (Sikait) and nearby areas, Egypt | pG (gr, vn) (E), M? (E) | ProtL / PalE    | Soliman 1982, 1984, 1986; Binda et al. 1993; Grundmann and Morteani 1993; Hassanen and Harraz 1996; Abdalla and Mohamed 1999 | Brl ( $\pm$ emerald) occurs in greisens and Qtz veins associated with Pan African peraluminous variably Sn-Ta-W-Mo-enriched granites; emerald in veins etc. in mafic schists with control by regional shear zones; alkaline rare-metal granites & metasomatism also common in region |
| Baid al Jamalah, Saudi Arabia           | pG (gr, vn)             | ProtL (569)     | Jackson and Ramsay 1986; du Bray et al. 1988; Kamilli et al. 1993                                                            | Proximal Brl in greisen W(-Mo-Sn) deposit associated with Be-enriched Bt granite; other systems (e.g., Silsilah) commonly lack Brl                                                                                                                                                   |
| <b>Tuareg massif, Algeria</b>           | pG (p/vn?)              | ProtL (680-610) | Anonymous 1987; Cottin et al. 1990; Goodwin 1996                                                                             | Numerous Be occurrences in area of Pan-African magmatism superimposed on older basement                                                                                                                                                                                              |

#### Southeastern Africa - Archean and Proterozoic

|                                                |                                         |              |                                                                                                                                |                                                                                                                                                                                                 |
|------------------------------------------------|-----------------------------------------|--------------|--------------------------------------------------------------------------------------------------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| <b>SE Africa / Madagascar</b>                  | G (p, vn), M ( $\pm$ E), P, carbonatite | Arch / Prot  | Sinkankas 1981; Kazmi and Snee 1989a; Grew (this volume)                                                                       | Abundant Brl-bearing pegmatites (LCT and other), also some vein systems & placers; some emeralds; granulite-related pegmatites                                                                  |
| Gravelotte (Leysdorp), Transvaal, South Africa | pG / M? (p, vn) (E)                     | Arch         | Sinkankas 1981, p. 492; Robb and Robb 1986; Grundmann and Morteani 1989; Kazmi and Snee 1989a; Nwe and Morteani 1993           | Emeralds in Qtz veins and Bt-Tlc schist; desilicated pegmatites and Qtz veins in Act-Chl-Bt schists; evidence suggests metamorphic redistribution may be key for Ph and later emerald           |
| Zimbabwe—Filabusi, Fort Victoria, Mweza        | pG? (p,vn, gr) (E)                      | Arch         | Sinkankas 1981; Kazmi and Snee 1989a                                                                                           | Emerald ( $\pm$ Ch) in ultramafic host adjacent to pegmatites, Qtz veins, greisens; many Be pegmatites including LCTs (e.g., Bikita)                                                            |
| Kafubu (Kitwe), Zambia                         | G? M? (vn, p?)                          | ProtL / PalE | Graziani et al. 1983; Sliwa and Ngulwe 1984; Kazmi and Snee 1989a; Kamona 1994; Mumba and Bartot 1998, M. Hitzman (pers. com.) | Brl(emerald) $\pm$ Ch in mafic schists in metamorphic complex of Lufilian arc near Copper Belt, with Qtz-Tur veins; spatially associated pegmatites but occurrences hydrothermal or metamorphic |
| Muambe, Mozambique                             | carbonatite (rpl)                       | Prot         | Cilek 1989; Woolley 2001                                                                                                       | Up to 1% BeO+Y-La enrichment in proximal Fl replacement in K <sub>2</sub> O-rich fenite aureole around iron-rich calcic carbonatite                                                             |

\* Abbreviations for deposit types:

- (1) genetic association: G = granite (Bt, Bt-Hbl or Fa), pG = peraluminous granite (Bt $\pm$ Ms, Ctd), pG = peralkaline granite (sodic pyroxene or amphibole present), S = syenitic rocks (commonly silica undersaturated and/or peralkaline), M = metamorphic origin (non-magmatic), B = basin-related hydrothermal, P = placer.
- (2) mineralization styles: m = magmatic, LCT = Li-Cs-Ta-type enrichment, p = pegmatitic (miarolitic or otherwise), f = feldspathic replacement (albitization, K-feldspathization, can be peralkaline), gr = greisen (Toz, Ms, Li-Fe micas all included), vn = veins (lacking prominent envelopes), sk = skarn (metasomatic calc-silicates), rpl = replacement (typically in carbonates, F-rich, lacks calc-silicates)
- (3) emerald occurrences are indicated by "(E)."



**APPENDIX B: SELECTED STUDIES RELEVANT TO BERYLLIUM  
MINERAL STABILITIES**

(see also Franz and Morteani, pp. 561ff, this volume)

| Material        | Synthesis and Reversal Experiments                                                                                                                                                                                                                                                                | Thermodynamic Data                                                                                                            | Equation of State                                                                            | Theoretical & Natural                                                                                                           |
|-----------------|---------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|-------------------------------------------------------------------------------------------------------------------------------|----------------------------------------------------------------------------------------------|---------------------------------------------------------------------------------------------------------------------------------|
| <i>Fluids</i>   |                                                                                                                                                                                                                                                                                                   |                                                                                                                               |                                                                                              |                                                                                                                                 |
| Melt-mineral    | Ganguli et al. 1975; Cemič et al. 1986; London et al. 1989; Icenhower and London 1995; Evensen et al. 1999; Evensen and London 2002; London and Evensen (this volume)                                                                                                                             |                                                                                                                               |                                                                                              | Bea et al. 1994                                                                                                                 |
| Melt-aqueous    | London et al. 1988                                                                                                                                                                                                                                                                                |                                                                                                                               |                                                                                              |                                                                                                                                 |
| Aqueous species | Beus et al. 1963; Soboleva et al. 1977, 1984b; Samchuk and Mitskevich 1980; Barton 1986; Renders and Anderson 1987; Koz'menko et al. 1988; Prasad and Ghosh 1988; Clegg and Brimblecombe 1989                                                                                                     | Samchuk and Mitskevich 1980; Barton 1986; Renders and Anderson 1987; Wood 1992                                                |                                                                                              | Wood 1992                                                                                                                       |
| <i>BASH</i>     |                                                                                                                                                                                                                                                                                                   |                                                                                                                               |                                                                                              |                                                                                                                                 |
| Behoite         |                                                                                                                                                                                                                                                                                                   | Kostryukov et al. 1977; Barton 1986                                                                                           |                                                                                              |                                                                                                                                 |
| Bertrandite     | Bukin 1968; Klyakhin et al. 1981; Hsu 1983; Lebedev and Ragozina 1984; Barton 1986                                                                                                                                                                                                                | Kiseleva et al. 1985, 1986; Barton 1986; Hemingway et al. 1986                                                                | Hazen and Au 1986                                                                            | Kosals et al. 1974; Burt 1978; Kupriyanova 1982; Wood 1992; Odintsova 1993                                                      |
| Beryl           | Syromyatnikov et al. 1972; Franz and Morteani 1981; Klyakhin et al. 1981; Hsu 1983; Aines and Rossman 1984; Franz and Morteani 1984; Polupanova et al. 1985; Barton 1986; Cemič et al. 1986; Renders and Anderson 1987; Manier-Glavinaz et al. 1988, 1989a; Wang et al. 1992; Evensen et al. 1999 | Barton 1986; Hemingway et al. 1986; Kiseleva et al. 1986; Renders and Anderson 1987; Gurevich et al. 1989; Pilati et al. 1997 | Morosin 1972; Deganello 1974; Schlenker et al. 1977; Hazen et al. 1986; Haussuehl 1993       | Burt 1975, 1978; Kupriyanova 1982; Odintsova 1993; Kupriyanova and Shpanov 1997; Markl and Schumacher 1997; Evensen et al. 1999 |
| Bromellite      | Barton 1986                                                                                                                                                                                                                                                                                       | Barton 1986                                                                                                                   | Hazen and Finger 1987; Pilati et al. 1993                                                    |                                                                                                                                 |
| Chrysoberyl     | Franz and Morteani 1981; Hsu 1983; Ospanov 1983; Barton 1986; Cemič et al. 1986; Hölscher et al. 1986; Wang et al. 1992                                                                                                                                                                           | Kiseleva et al. 1985; Barton 1986; Hemingway et al. 1986; Kiseleva et al. 1986; Hofmeister et al. 1987                        | Au and Hazen 1987; Hazen and Finger 1987                                                     | Burt 1978; Kupriyanova 1982; Odintsova 1993; Kupriyanova and Shpanov 1997                                                       |
| Euclase         | Franz and Morteani 1981; Hsu 1983; Ospanov 1983; Barton 1986                                                                                                                                                                                                                                      | Franz and Morteani 1981; Kiseleva et al. 1985; Barton 1986; Hemingway et al. 1986; Kiseleva et al. 1986                       | Hazen et al. 1986                                                                            | Burt 1978                                                                                                                       |
| Phenakite       | Bukin 1968; Syromyatnikov et al. 1972; Franz and Morteani 1981; Ospanov 1983; Lebedev and Ragozina 1984; Soboleva et al. 1984a; Barton 1986; Wang et al. 1992                                                                                                                                     | Matveev and Zhuravlev 1982; Topor and Mel'chakova 1982; Kiseleva and Shuriga 1983; Barton 1986; Hemingway et al. 1986         | Hazen and Au 1986; Hazen and Finger 1987; Yeganeh-Haeri and Weidner 1989; Pilati et al. 1998 | Kosals et al. 1974; Burt 1978; Odintsova 1993; Pilati et al. 1998                                                               |

## Appendix B, continued

|                      |                                     |                                          |                                                    |                              |
|----------------------|-------------------------------------|------------------------------------------|----------------------------------------------------|------------------------------|
| Be-Mg-Al-Si          | Hölscher et al. 1986                |                                          | Schmetzer 1981                                     | Grew 1981; Grew, this volume |
| <i>Helvite group</i> |                                     |                                          |                                                    |                              |
| Danalite             | Fursenko 1982                       | Ospanov 1983                             |                                                    | Burt 1980, 1988              |
| Genthelvite          | Fursenko 1982                       | Ospanov 1983;<br>Mel'chakova et al. 1991 |                                                    | Burt 1980, 1988              |
| Helvite              | Klyakhin et al. 1981; Fursenko 1982 |                                          | Werner and Plech 1995 (also tugtupite)             | Burt 1980, 1988              |
| <i>Others</i>        |                                     |                                          |                                                    |                              |
| Chkalovite           | Ganguli et al. 1975                 | Kiseleva et al. 1984a,b                  | Henderson and Taylor 1989                          | Markl 2001                   |
| Gadolinite           | Ito and Hafner 1974                 |                                          | Demartin et al. 1993                               |                              |
| Phosphates           |                                     |                                          | Klaska and Jarchow 1973; Henderson and Taylor 1984 | Burt 1975                    |

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**14****Non-pegmatitic Deposits of Beryllium:  
Mineralogy, Geology, Phase Equilibria and Origin****Mark D. Barton, Steven Young**

|                                                                                                                                                                   |     |
|-------------------------------------------------------------------------------------------------------------------------------------------------------------------|-----|
| INTRODUCTION.....                                                                                                                                                 | 591 |
| Economic sources of beryllium and beryllium minerals.....                                                                                                         | 594 |
| TYPES OF DEPOSITS.....                                                                                                                                            | 595 |
| BERYLLIUM MINERAL COMPOSITIONS.....                                                                                                                               | 604 |
| Beryl group—(□,Na,Cs,H <sub>2</sub> O)(Be,Li) <sub>3</sub> (Al,Sc,Fe <sup>+3</sup> ,Cr,Fe <sup>+2</sup> ,Mg) <sub>2</sub> [Si <sub>6</sub> O <sub>18</sub> ]..... | 604 |
| Helvite group—(Mn,Fe,Zn) <sub>4</sub> [BeSiO <sub>4</sub> ] <sub>3</sub> S.....                                                                                   | 609 |
| Other minerals.....                                                                                                                                               | 611 |
| BERYLLIUM MINERAL STABILITIES.....                                                                                                                                | 611 |
| Pressure-temperature-activity relationships.....                                                                                                                  | 612 |
| Solubility relationships.....                                                                                                                                     | 618 |
| MAGMATIC BERYLLIUM ENRICHMENTS.....                                                                                                                               | 619 |
| Strongly peraluminous to metaluminous systems.....                                                                                                                | 620 |
| Peralkaline-metaluminous systems.....                                                                                                                             | 621 |
| HYDROTHERMAL OCCURRENCES ASSOCIATED WITH FELSIC MAGMATISM.....                                                                                                    | 623 |
| Peraluminous magma-related systems.....                                                                                                                           | 624 |
| Metaluminous magma-related systems.....                                                                                                                           | 631 |
| Peralkaline magma-related systems.....                                                                                                                            | 639 |
| Igneous-related emerald deposits.....                                                                                                                             | 643 |
| NON-MAGMATIC OCCURRENCES.....                                                                                                                                     | 649 |
| Fe-Mn(-Zn) oxide-rich occurrences.....                                                                                                                            | 649 |
| Basinal (and metamorphic?) brine-related emerald deposits.....                                                                                                    | 649 |
| Metamorphic occurrences.....                                                                                                                                      | 651 |
| Weathering and placers.....                                                                                                                                       | 652 |
| SYNOPSIS OF DEPOSIT CHARACTERISTICS AND ORIGINS.....                                                                                                              | 653 |
| Global-scale patterns.....                                                                                                                                        | 653 |
| System-scale patterns.....                                                                                                                                        | 654 |
| Origins.....                                                                                                                                                      | 658 |
| ACKNOWLEDGMENTS.....                                                                                                                                              | 659 |
| APPENDIX A: Summary of Beryllium Deposits by Region.....                                                                                                          | 659 |
| APPENDIX B: Selected Studies Relevant to Beryllium Mineral Stabilities.....                                                                                       | 671 |
| REFERENCES.....                                                                                                                                                   | 672 |