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ScienceDirect
Lithos 97 (2007) 219–246
LITHOS
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Are Cenozoic topaz rhyolites the erupted equivalents of Proterozoic rapakivi granites? Examples from the western United States and Finland

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Received 6 December 2005; accepted 25 January 2007
Available online 2 February 2007

Abstract

Eruptions of topaz rhyolites are a distinctive part of the late Cenozoic magmatic history of western North America. As many as 30 different eruptive centers have been identified in the western United States that range in age from 50 to 0.06 Ma. These rhyolite lavas are characteristically enriched in fluorine (0.2 to 2 wt.% in glass) and lithophile trace elements, such as Be, Li, Rb, Cs, Ga, Y, Nb, and Ta. REE patterns are typically flat with large negative Eu anomalies; negative Nb–Ta anomalies are small or nonexistent; and F/Cl ratios in glasses are high (>3). These features, together with high Fe/Mg ratios and usually low fO_2, set them apart from subduction-related (I-type) silicic rocks. The rhyolites are metaluminous to only slightly peraluminous, lack indicator minerals of strongly peraluminous magmas, and have low P and B contents; these features set them apart from S-type silicic magmas. Instead, topaz rhyolites have the major and trace element, mineralogic, and isotopic characteristics of aluminous A-type or within-plate granites. Topaz rhyolites were formed during regional extension, lithospheric thinning, and high heat flow.

Topaz rhyolites of the western United States crystallized under subsolvus conditions, and have quartz, sanidine, and Na-plagioclase as the principal phenocrysts. Fluorite is a common magmatic accessory, but magmatic topaz occurs only in a few complexes; both are mineralogical indicators of F-enrichment. Many also crystallized at relatively low fO_2 (near QFM) and contain mafic silicate minerals with high Fe/(Fe+Mg) ratios. Some crystallized at higher oxygen fugacities and are dominated by magnetite and have titanite as an accessory mineral. Post-eruption vapor-phase minerals include topaz, garnet, red Fe–Mn-rich beryl, bixbyite, pseudobrookite, and hematite. They are genetically related to deposits of Be, Mo, F, U, and Sn. Topaz rhyolites erupted contemporaneously with a variety of other igneous rocks, but most typically they form bimodal associations with basalt or basaltic andesite and are unrelated to large collapse calderas.

In their composition and mineralogy, topaz rhyolites are similar to the evolved members of rapakivi granite complexes, especially those of Proterozoic age in southern Finland. This suggests similarity in origin and lessons learned from these rocks may help us better understand the origins of their more ancient counterparts. For example, all topaz rhyolites in western North America seem to be intrinsically related to extension following a regional period of subduction-related volcanism. Cratonized Precambrian crust is found beneath almost all of them as well. Trace element models, Sr–Nd isotopic data, and geologic associations indicate that topaz rhyolites probably form by fractional crystallization of silicic magma which originated by small degrees of melting of hybridized continental crust containing a significant juvenile mantle component not derived from a subduction zone (i.e., intrusions of within-plate mafic magma). The Sr and Nd isotopic compositions of the topaz rhyolites lie between the fields of contemporaneous mafic magmas and older calc-alkaline dacites and rhyolites. Intraplate mafic magmas and their derivatives appear

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doi:10.1016/j.lithos.2007.01.010

to have lodged in the crust and were then re-melted by subsequent injections of mafic magma. In turn, the mafic mantle-derived magma probably formed as a result of decompression related to lithospheric extension *or* to convective-flow driven by the foundering of a subducting lithospheric plate. Although significant uncertainty remains, we suggest that topaz rhyolites (and by extension rapakivi granites) are probably not simply melts of mid-crustal granodiorites, nor are they derived solely from felsic crust that was previously dehydrated or from which melt had been extracted as proposed in earlier papers.
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Keywords: Topaz rhyolite; Cenozoic; Rapakivi granite; Proterozoic; Fluorine; A-type; Anorogenic

1. Introduction

Topaz in rhyolitic lavas was first discovered in 1859 in western Utah and reported in the scientific literature by Simpson (1876). Since then, topaz-bearing rhyolitic lavas have been identified in much of western United States (Christiansen et al., 1983), Mexico (Huspeni et al., 1984), the Yukon Territory of Canada (Sinclair, 1986), eastern Russia and Mongolia (Kovalenko and Kovalenko, 1984). Topaz-bearing rhyolite dikes of Proterozoic age have also been found in southern Finland (Haapala, 1977) and central Arizona (Kortemeier and Burt, 1988). Christiansen et al. (1986) concluded that topaz rhyolite lavas from the western United States are generally similar to some A-type granites. This report summarizes the characteristics of Cenozoic topaz rhyolites–emphasizing new data from the Wah Wah Mountains of southwestern Utah–and compares them to Proterozoic rapakivi granites of the Fennoscandian shield (Fig. 1). Southern Finland is the type locality of these unique anorogenic granitic rocks, which Haapala and Rämö (1992) have redefined as A-type granites with rapakivi texture. Once the similarities between Cenozoic topaz rhyolites and Proterozoic rapakivi granites are clear, we consider a new model for the origin of topaz rhyolites and its implications for the petrogenesis of rapakivi granites in particular and A-type granites in general.

2. Methods of study

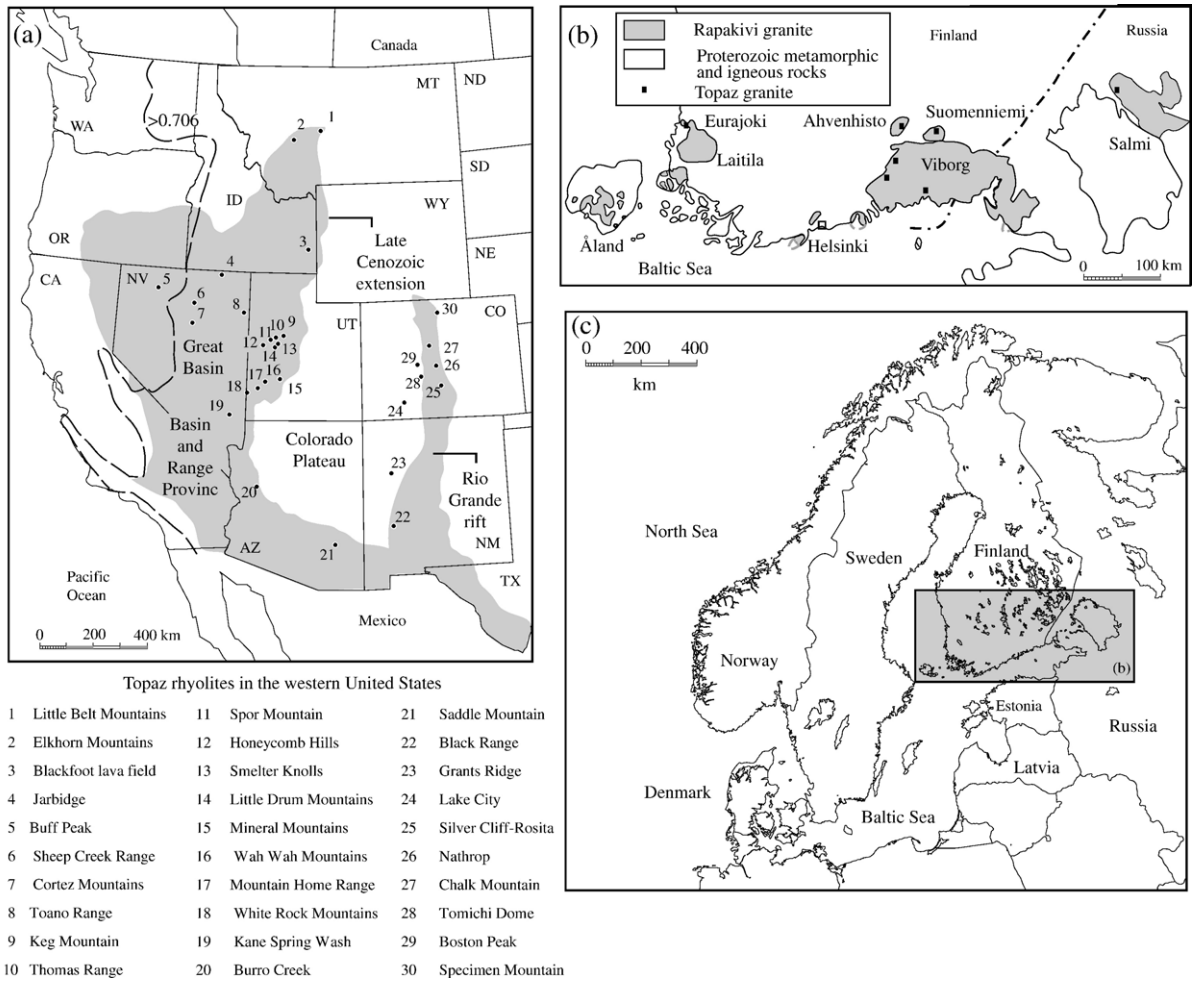
New geochemical data in this paper are presented for rhyolite lavas from in and near the Wah Wah Mountains of southwestern Utah (Fig. 2). Major and trace element compositions were collected by X-ray fluorescence spectrometry at Brigham Young University. Analyses of international materials for 31 elements can be found at http://www.geology.byu.edu/faculty/ehc/. Elemental and isotopic compositions of other topaz rhyolites and for Finnish rapakivi granites are taken from the references cited below.

New Sr, Nd, and Pb isotope compositions were measured on a GV Instruments Sector 54 thermal ionization mass spectrometer (TIMS) at the University of Wisconsin–Madison Radiogenic Isotope Laboratory following standard procedures (e.g., Johnson and Thompson, 1991). Samples were crushed in a steel jaw crusher and powdered in an agate ball mill. For Sr and Nd, 50 mg aliquots of whole-rock powders were spiked with ^{84}Sr- and ^{150}Nd-enriched tracers and dissolved in a mixture of HF and HNO_3, the elements were then separated using standard ion-exchange chromatography. Total procedural blanks were <0.1 ng for both Sr and Nd, which are negligible. For Pb isotope ratios, 100 mg aliquots of whole-rock powder were dissolved in a mixture of HF and HNO_3, and Pb was separated using HBr and HCl on an ion-exchange column. Total procedural blanks for Pb were also negligible at <2 ng. Both Sr and Nd isotope compositions were exponentially corrected for mass fractionation using $^{86}Sr/^{88}Sr = 0.1194$ and $^{146}Nd/^{144}Nd = 0.7219$, respectively. Within-run errors in measured $^{87}Sr/^{86}Sr$ ratios for dynamic analyses are determined as ±2 standard error (2SE) using $n = 120$ (number of measured ratios). The $^{87}Sr/^{86}Sr$ ratio measured for NBS-987 during this study was 0.710265±8 (2SE, $n = 13$). Neodymium was measured as NdO^+ and is presented as εNd values relative to present day CHUR, taken to be equal to BCR-1, measured during the analytical session as 0.512636±5 (2SE). Within-run errors in measured $^{143}Nd/^{144}Nd$ ratios for dynamic analyses are reported as 2SE where $n = 150$ (number of measured ratios). Twelve analyses of an internal Ames Nd standard yielded a $^{143}Nd/^{144}Nd$ ratio and precision of 0.511977±3 (2SE). Lead isotope ratios were corrected for mass fractionation by +0.14% per atomic mass unit based on fourteen analyses of NBS-981 (±0.005%; 2SE) and NBS-982 (±0.008%; 2SE) standards.

3. Distribution and ages

3.1. Distribution of topaz rhyolites in western United States

Topaz rhyolites are widespread in western North America (Fig. 1) and have been found as far north as



Topaz rhyolites in the western United States

1	Little Belt Mountains	11	Spor Mountain	21	Saddle Mountain
2	Elkhorn Mountains	12	Honeycomb Hills	22	Black Range
3	Blackfoot lava field	13	Smelter Knolls	23	Grants Ridge
4	Jarbidge	14	Little Drum Mountains	24	Lake City
5	Buff Peak	15	Mineral Mountains	25	Silver Cliff-Rosita
6	Sheep Creek Range	16	Wah Wah Mountains	26	Nathrop
7	Cortez Mountains	17	Mountain Home Range	27	Chalk Mountain
8	Toano Range	18	White Rock Mountains	28	Tomichi Dome
9	Keg Mountain	19	Kane Spring Wash	29	Boston Peak
10	Thomas Range	20	Burro Creek	30	Specimen Mountain

Fig. 1. (a) Distribution of topaz rhyolites (filled circles) from western United States. All are found within the extensional terranes of the Basin and Range Province and the Rio Grande Rift. Modified from John et al. (2000) and Dickinson (2002). The $^{87}Sr/^{86}Sr$ line marks the western edge of Precambrian basement (modified from Kistler and Peterman, 1978; Wooden et al., 1998; Tosdal et al., 2000). (b) Precambrian rapakivi granites of southern Finland (modified from Lukkari, 2002; Haapala and Lukkari, 2005). (c) Index map showing location of (b) in Fennoscandia.

Montana and extend southward into central Mexico. Most known topaz rhyolites in the western United States lie within the eastern and southern Basin and Range province and along the Rio Grande rift and thus appear to surround the Colorado Plateau. Nearly all topaz rhyolites lie east of the initial $^{87}Sr/^{86}Sr=0.706$ line as determined for Mesozoic plutonic rocks (Fig. 1, Kistler and Peterman, 1978; Wooden et al., 1998; Tosdal et al., 2000). This line is taken by these investigators to mark the westernmost extent of the Precambrian craton in the western United States. To the west is a series of allochthonous or accreted terranes composed of ocean-floor or island arc crust (e.g., Speed, 1979). These terranes may have formed as oceanic crust at the margin of North America during the Paleozoic and early Mesozoic Eras and were later accreted (Oldow, 1984). The topaz rhyolite lava at Buff Peak, Nevada (Castor and Henry, 2000) is the only one that has been found in the northwestern Basin and Range province, in spite of common bimodal (basalt–rhyolite) volcanism and extensional faulting throughout the province. The young lithosphere in this region, with its mafic crust, does not generally appear to have a composition appropriate for the generation of topaz rhyolites. The distribution of topaz rhyolites is entirely included in the region of Cenozoic extensional faulting (Fig. 1). Their emplacement appears to have spanned most of the Cenozoic Era with isotopic ages ranging from 50 Ma (Little Belt Mountains of Montana) to 0.06 Ma (Blackfoot lava field of southern Idaho), although all but 3 are younger than 30 Ma.

In the Wah Wah Mountains and vicinity of southwestern Utah (Fig. 2), which are considered in more detail in this paper, there were two episodes of topaz

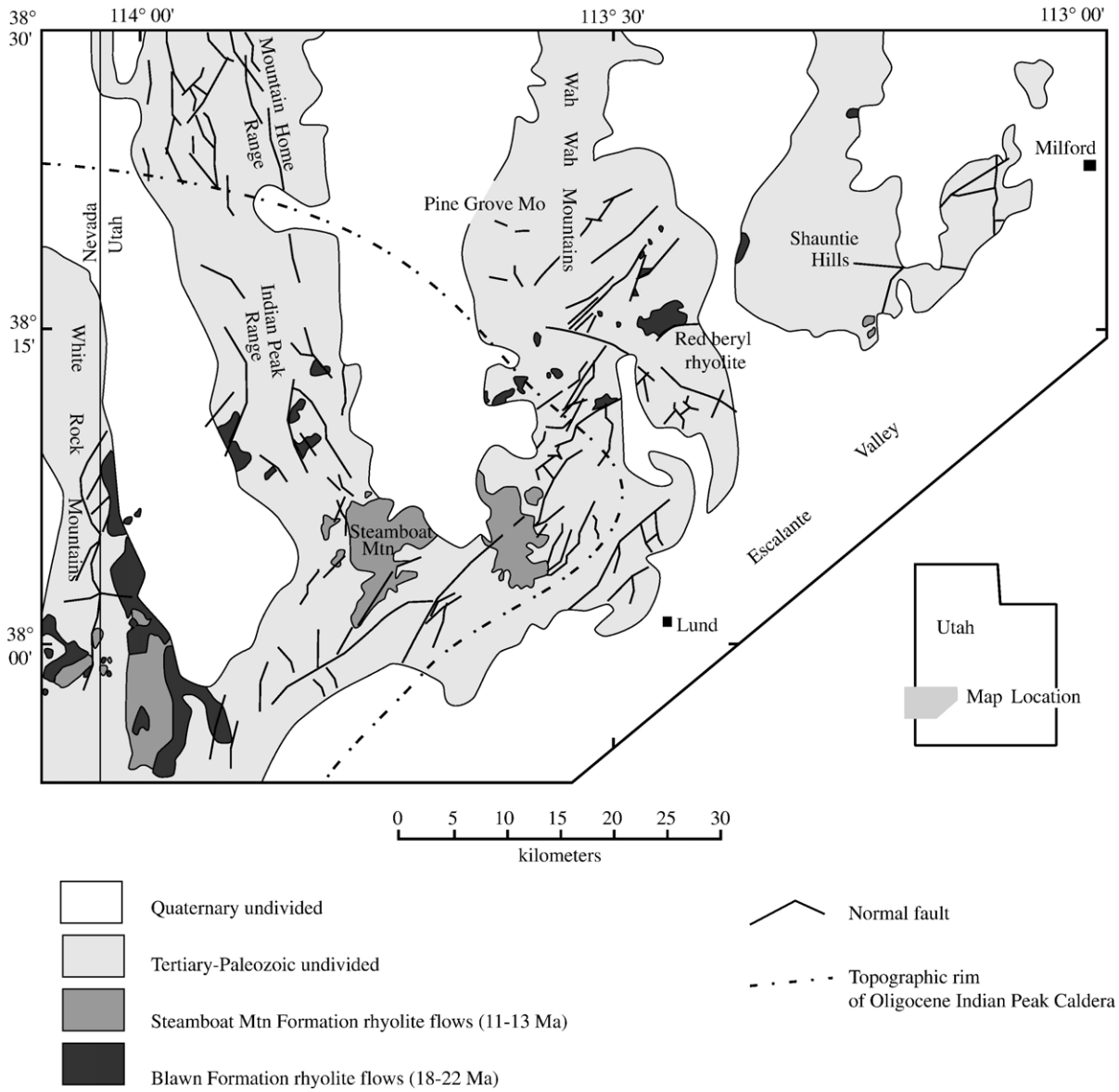


Fig. 2. Simplified geologic map of southwestern Utah showing the distribution of topaz-bearing rhyolites of two different ages included in the Steamboat Mountain Formation and the Blawn Formation. These rhyolites erupted across the Oligocene calc-alkaline andesite to rhyolite suite centered on the Indian Peak caldera. North-trending ranges are bounded by buried Miocene and younger normal faults. Modified from Best et al. (1987).

rhyolite volcanism—one at 22 to 18 Ma and a second at about 13 to 11 Ma (Thompson, 2002). Fission track, structural, and stratigraphic studies suggest that extension in the eastern Great Basin began about 22 to 17 Ma (Rowley et al., 1978; Stockli et al., 2001; Dickinson, 2002) and eventually formed a series of north-trending horsts and grabens (Fig. 2). Thus the onset of extension is closely tied to the eruption of the oldest topaz rhyolites in this area.

3.2. Distribution of Fennoscandian rapakivi granites

The rapakivi granites of Fennoscandia form large composite batholiths and smaller satellitic stocks across central Sweden, southern Finland, into Russia, beneath the floor of the Baltic Sea and in the Baltic countries (Fig. 1b). The rapakivi granites and associated mafic rocks can be divided from east to west into four area-constrained age groups: the Salmi batholith in Russian

Karelia (1.55 to 1.53 Ga), the Wiborg batholith and satellites in southwestern Finland and Estonia (1.67 to 1.62 Ga), the rapakivi batholiths and satellites in southwestern Finland and Latvia (1.59 to 1.54 Ga), and the rapakivi–gabbro complexes of central Sweden (1.53 to 1.47 Ga) (e.g., Rämö et al., 2000; Haapala et al., 2005). A variety of tectonic environments has been proposed for the generation of the Fennoscandian rapakivi granites, but an incipient extensional setting is indicated by oriented swarms of mafic dikes, shallow grabens imaged geophysically, and thinning of the crust across the region (Haapala et al., 2005). The rapakivi granites of Fennoscandia were emplaced in Proterozoic crust that is a few hundreds of millions of years older than the granites themselves (Rämö and Haapala, 1995).

4. Magma–tectonic associations

The nature of igneous rock associations and contemporaneous tectonic activity give some clues about the generation of magmas. This is especially important for Cenozoic topaz rhyolites of the western United States, where we can compare young igneous rocks of "known" tectonic setting with the much older, but geochemically similar, rapakivi granites.

4.1. Magmatic associations for topaz rhyolites

In the Wah Wah Mountains (Fig. 2), for example, following a Cretaceous episode of folding and thrust faulting, subduction-related calc-alkaline volcanism began in the early Oligocene (about 32 Ma) and continued until the early Miocene (Best et al., 1989). This volcanism produced widely scattered, partly clustered, composite volcanoes with andesitic to dacitic lava flows as well as small volumes of isolated andesitic lavas. Widespread dacitic to rhyolitic ash flows erupted from large collapse calderas in the Wah Wah Mountains and Indian Peak Range and other nearby ranges (Fig. 2). Dacite gradually gave way to high K_2O trachydacite ignimbrites by about 26 Ma. Following a local cessation of volcanism, the older topaz rhyolite domes (Blawn Formation, 22 to 18 Ma) erupted along with trachyandesite lava flows (62 to 54% SiO_2) to form a bimodal suite (see Fig. 5). A Miocene lull in volcanic activity in the eastern Great Basin was followed by renewed bimodal magmatism that began about 13 Ma in and near the Wah Wah Mountains (Steamboat Mountain Formation; Best et al., 1987; Fig. 2). Topaz-bearing rhyolites were again accompanied by the eruption of trachybasalts to trachyandesites in the Wah Wah Mountains.

Keith et al. (1994) describe a graben into which a topaz–beryl-bearing rhyolite erupted 22 Ma, suggesting that extension may have begun just prior to the eruption of the rhyolites. Pronounced extension began in the region sometime between 22 Ma (Rowley et al., 1978) and 17 Ma (Stockli et al., 2001) and eventually formed the present system of horsts and grabens (Fig. 2).

Indeed, extensional tectonism appears to be the common factor in almost all areas where topaz rhyolites erupted in the western United States. Episodes of topaz rhyolite magmatism coincide with periods of lithospheric extension: 1) in the eastern Great Basin where normal faulting may have begun as early as 22 Ma ago as noted above and then was renewed under a different stress orientation about 14 Ma which has persisted to the present (Zoback et al., 1981); 2) along the northern Nevada rift that opened 16 Ma (Stewart et al., 1975; Zoback and Thompson, 1978; John et al., 2000); 3) in Montana where normal faulting began about 40 Ma ago (Chadwick, 1978) and intra- or back-arc graben formation may have begun as early as 50 Ma (Armstrong, 1978); 4) along the Rio Grande rift and its northern extension into Colorado which initially developed about 30 Ma ago (Eaton, 1979); and 5) in western Arizona where detachment faulting and crustal extension were underway by before 15 Ma (Suneson and Lucchitta, 1983). The intimate association of extensional tectonics and topaz rhyolite magmatism in the western United States implies a strong genetic connection.

The magmatic associations of topaz rhyolites may also place important constraints on their origins. As in the Wah Wah Mountains, topaz rhyolite magmatism consistently follows a slightly older episode of subduction-related calc-alkaline magmatism (*magnesian* in the sense of Frost et al., 2001a). Lipman et al. (1972) and Christiansen and Lipman (1972) first concluded that the Cenozoic magma–tectonic evolution of the western United States may be divided into two fundamentally different stages. An early suite of calc-alkaline magmas was associated with subduction of the Farallon plate. Eruptions of andesitic lavas, dacites, and rhyolites were common, many of which are potassium rich. The silicic magmas were erupted mostly as ash flows associated with caldera collapse and many of the andesites formed large fields of coalesced stratovolcanoes. This magmatism swept southward across much of the western United States. It started in Montana about 50 million years ago and moved southward and then stagnated in southern Nevada and Utah between about 30 and 25 Ma. At the same time, similar magmas were also erupted across southern Arizona and New Mexico as well as

throughout northwestern Mexico. In all of these areas, subduction zone magmatism was replaced by a bimodal suite of mafic magmas and rhyolite. The timing of the switch is not the same across the entire region. Ultimately, the younger magmatism became associated with lithospheric extension and normal faulting. In many places, the transition was marked by a magmatic gap of several million years. In other cases, the transition was gradual. For example, the early Miocene rhyolites of the Wah Wah and Needle Ranges (Fig. 2) form a bimodal (trachyandesite–rhyolite) association locally, but contemporaneous volcanism elsewhere in southwestern Utah was still broadly calc-alkaline and continuous from andesite to rhyolite. However, by the time the younger topaz rhyolites erupted in the Wah Wah Mountains, calc-alkaline andesite to rhyolite magmatism had ceased throughout the region. The mafic members of the younger bimodal suites varied widely and included potassic trachybasalt and trachyandesite as well as alkaline and tholeiitic basalt. Topaz rhyolites at Kane Springs Wash volcanic center in southern Nevada are associated with contemporaneous peralkaline trachytes and rhyolites.

4.2. Magmatic associations for rapakivi granites

Magmas associated with Finnish rapakivi complexes were likewise diverse, but generally were not calc-alkaline. Rapakivi granites in Finland are part of a bimodal sequence that includes tholeiitic gabbros (as dikes, sills, and small screens between intrusions), norites, anorthosites, and ferrodiorites (Rämö, 1991; Salonsaari, 1995). The silicic members include various petrographic types of granite, rhyolite (quartz porphyry) dikes and local lavas, as well as rare syenite dikes.

5. Mode of emplacement

5.1. Emplacement of topaz rhyolites

The eruption and emplacement of the topaz rhyolites of the Wah Wah Mountains were typical of most others in the western United States. The rhyolites of both eruptive episodes occur as isolated intrusive plugs without significant pyroclastic deposits, as isolated endogenous lava domes or flows with underlying pyroclastic breccias and tuffs and as groups of coalesced domes and flows with interlayered tephra deposits (Fig. 3). Vent-clearing explosions locally created breccias that underlie tuffs. These near vent explosion breccias contain abundant lithic inclusions of the local country rocks. The explosion breccia is commonly overlain by remnants of a tuff ring

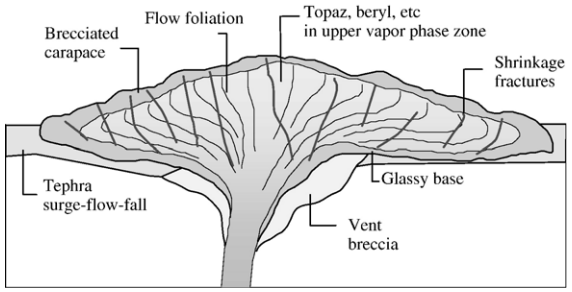


Fig. 3. Synthetic cross section of a topaz-bearing rhyolite dome and flow. The details are modeled after rhyolite lavas in the Wah Wah Mountains of southern Utah (Christiansen, 1980; Christiansen et al., 1996, Thompson, 2002). A typical dome is 0.5 to 3 km across and a few 100 m high. Flows may extend for a few kilometers away from the vent.

consisting of stratified pyroclastic-surge units produced during pulsing unsteady eruptions. Some short and thin (less than 1 m) lithic-rich ash flows probably resulted from minor collapse of low eruption columns. Mantling ash-fall units punctuate the record of explosive volcanism at several localities. Vitrophyres a few meters thick are present at the bases of some lava flows that overlie the pyroclastic deposits. Others have basal flow breccias (about 1 m thick) produced as the flow front crumbled, slumped, and was overridden by the advancing flow. Rapidly quenched vitrophyric blocks from the flow front are common in this part of the flow (Fig. 3). In the upper portions of the flows, felsitic, flow-layered lavas with abundant vapor-phase cavities are typical.

These features suggest that the pyroclastic eruptions were initiated as rising magmas explosively mixed with groundwater (hydromagmatic eruptions). Once the vent was cleared, relatively quiet eruption of rhyolite lava proceeded. The transition from pyroclastic to lava eruptions may also correlate with the eruptive degassing of the magma or with the evisceration of a volatile-rich cap to a small magma chamber (Byrd and Nash, 1993).

The volume of magma in individual domes or flows ranges from less than 0.2 km^3 to a probable maximum of about 10 km^3. However, fairly large volumes (10 to 50 km^3) of coalesced domes and flows accumulated over short (about 1 m.y.) time intervals in the Wah Wah Mountains and Thomas Range of western Utah, and in New Mexico's Black Range (Duffield and Dalrymple, 1990). It is important to contrast both the small volumes and mode of emplacement of these F-rich magmas with other Cenozoic rhyolites from the same region which generally erupted from large collapse calderas (e.g., Indian Peak, Central Nevada, and San Juan caldera complexes, the Snake River Plain, or the southwest Nevada volcanic field). Dacitic and rhyolitic ash flows

from these calderas have volumes 1 to 3 orders of magnitude larger (e.g., Smith, 1979; Best et al., 1989). On the other hand, the eruption of large volumes of F-rich magma over geologically short time intervals in the Thomas Range, Utah, and in the Black Range, New Mexico, suggests that some topaz rhyolites may emanate from magma chambers with volumes approaching those of caldera-related plutons. Only one Cenozoic topaz-bearing granite has been found in the western United States—the 21 Ma Sheeprock granite of western Utah (Christiansen et al., 1988; Richardson, 2004). The pluton has all of the geochemical characteristics of topaz rhyolites from the eastern Great Basin. It was emplaced at a shallow level and covers an area of about 20 km^2. Calzia and Rämö (2005) have identified two A-type granite plutons from the Death Valley region. Both are Miocene in age (12.4 and 10 Ma) and display rapakivi textures. However, compared to contemporaneous topaz rhyolites, they are more oxidized, less enriched in incompatible trace elements (e.g., F, Rb, HREE), and lack topaz.

5.2. Emplacement of rapakivi granites

Rapakivi granites in Fennoscandia typically occur as sharply discordant, composite high-level batholiths and stocks that intrude metamorphic bedrock. In southern Finland, the largest is the Wiborg batholith which is almost 200 km across. The batholiths may be sheetlike bodies 5 to 10 km thick (Laurén, 1970; Haapala and Rämö, 1992; Korja et al., 1993). Individual plutons are as small as a few kilometers across, such as the topaz-bearing parts of the Eurajoki (Haapala, 1977), Artjärvi and Sääskärvi (Lukkari, 2002), Suomenniemi (Rämö, 1991), and Kymi (Haapala and Lukkari, 2005) intrusions (Fig. 2). The zoned, composite Ahvenhisto pluton is larger and covers about 240 km^2 (Edén, 1991), but only a small portion of the pluton is topaz bearing. The depth of emplacement of the Finnish rapakivi granites is, thus far, poorly constrained. Elliott (2001), using amphibole compositions, calculated crystallization pressures of 3.6 to as much as 5 kb, but these were not corrected for the significant effect of low fO_2. Consequently, these estimates are probably over-estimates (Anderson and Smith, 1995). Miarolitic cavities are present in some of the topaz-bearing phases; rapakivi-age volcanic and subvolcanic rocks probably formed much of the roof of the Wiborg batholith. Subvolcanic quartz-feldspar porphyry dikes are often associated with the plutons as are swarms of tholeiitic diabase dikes (Rämö and Haapala, 1995). Pegmatites are also rare. These associations point to a shallow level of emplacement for the granites.

6. Petrography and mineralogy

6.1. Mineralogy of topaz rhyolites

Topaz rhyolites are generally flow-layered and nearly aphyric to sparsely crystalline, but a few contain as much as 40% phenocrysts. The major phenocrysts in topaz rhyolites in the Wah Wah Mountains are typical of others and include sanidine, smoky quartz, sodic plagioclase, and sparse Fe-rich biotite, in order of abundance. Magmatic topaz has not yet been found in the Wah Wah rhyolites. In the western United States it has only been identified in the Honeycomb Hills complex (Congdon and Nash, 1991). Sanidine in topaz rhyolites is generally Or_{40} to Or_{60} and plagioclase is typically sodic oligoclase. Biotite generally has high Fe/(Fe+Mg) (Fig. 4) reflecting the high Fe/(Fe+Mg) of the magma (in many cases, molar Mn and Ti exceed Mg), and the prevalence of relatively low fO_2 during crystallization. At comparable Fe/(Fe+Mg) ratios, the Al_{tot} in biotites from topaz rhyolites is less than from strongly peraluminous two-mica granites (Fig. 4). Biotites from topaz rhyolites also have high F-contents (up to 5 wt.%). Fluorine concentrations this high for Fe-rich biotites suggest crystallization at high fHF and fHF/fH_2O. F/Cl ratios in the biotites also suggest crystallization at high $fHF/fHCl$ (e.g., Byrd and Nash, 1993). Rare phenocryst phases include clinopyroxene (in high T, F-poor magmas from the Thomas Range, Utah, and Jarbidge, Nevada, rhyolites), fayalite (Kane Springs Wash, Nevada), Fe-rich hornblende, or Fe–Mn garnet. Magmatic accessory

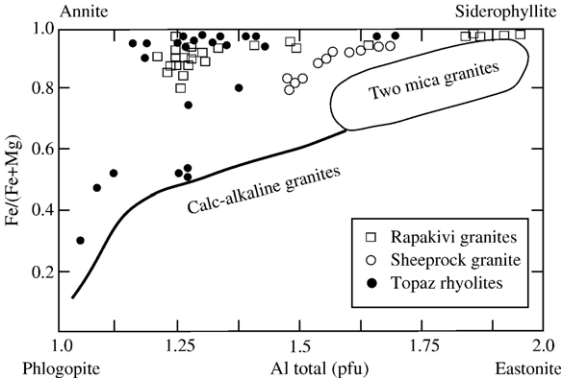


Fig. 4. Biotite compositions of topaz rhyolites and granites from the western United States compared with those of rapakivi granites of southern Finland. Data from Haapala (1977), Christiansen et al. (1986), Rogers (1990 for the Sheeprock granite), Salonsaari (1995), Rieder et al. (1996), and Elliott (2001). Fields for two-mica (broadly S-type) and calc-alkaline (broadly I-type) granites are included for comparison (Christiansen et al., 1986).

Table 1
Representative chemical compositions of topaz rhyolites from the western United States and topaz granites from southern Finland

	Reference	SiO_2	TiO_2	Al_2O_3	Fe_2O_3*	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	Total	LOI	Anal total
Topaz Rhyolites, western United States														
Blawn Formation (22 to 18 Ma)														
LAM-1-38-2		77.06	0.07	12.50	1.14	0.13	0.10	0.61	3.66	4.73	0.01	100.00	1.02	99.96
TET-9-43-2		77.01	0.09	12.44	0.94	0.11	0.10	0.71	3.91	4.67	0.03	100.00	0.76	99.86
RED-04	Thompson (2002)	76.88	0.05	12.91	0.98	0.13	0.07	0.34	3.15	5.49	0.01	100.00	3.58	99.42
RED-06	Thompson (2002)	77.89	0.10	11.87	1.06	0.09	0.26	0.63	3.12	4.98	0.00	100.00	0.74	99.16
RED-12	Thompson (2002)	77.44	0.10	11.96	1.07	0.05	0.32	0.58	3.02	5.45	0.01	100.00	0.84	99.87
RED-33	Thompson (2002)	76.85	0.05	12.72	1.08	0.10	0.02	0.40	4.15	4.62	0.01	100.00	0.51	99.22
RED-34	Thompson (2002)	76.73	0.04	12.88	1.09	0.11	0.03	0.32	4.19	4.61	0.00	100.00	0.37	99.04
RED-40	Thompson (2002)	76.43	0.04	13.02	1.08	0.09	0.04	0.44	4.13	4.73	0.00	100.00	0.42	98.07
RED-41	Thompson (2002)	77.58	0.06	13.06	0.34	0.01	0.15	0.43	3.39	4.97	0.00	100.00	1.12	98.96
Steamboat Formation (13 to 11 Ma)														
BBS-8-229-2		77.49	0.08	12.36	1.07	0.08	0.13	0.25	3.75	4.76	0.01	100.00	0.54	100.12
LAM-9-103-2		75.77	0.07	13.18	1.07	0.11	0.11	0.69	4.06	4.90	0.05	100.00	1.05	99.60
WW-8014		76.41	0.08	12.20	1.23	0.08	0.11	0.99	4.16	4.69	0.05	100.00		99.40
WW-8016		76.30	0.04	12.70	1.14	0.12	0.00	0.42	4.71	4.57	0.00	100.00		99.09
WW-8018		76.29	0.04	12.76	1.16	0.07	0.03	0.47	4.67	4.51	0.00	100.00		99.04
WW-8020		76.45	0.09	12.60	1.26	0.08	0.08	0.46	3.97	5.01	0.00	100.00		99.21
WW-8029		76.52	0.04	12.75	1.16	0.13	0.00	0.31	3.74	5.35	0.00	100.00		99.00
WW-8011A		76.87	0.05	12.77	0.67	0.05	0.00	0.37	4.57	4.65	0.00	100.00		98.83
WW-8011B		76.35	0.08	12.74	1.11	0.10	0.00	0.43	4.49	4.70	0.00	100.00		99.24
Spor Mountain Rhyolite (21 Ma)														
	Christiansen et al. (1986)	75.25	0.05	13.58	1.45	0.06	0.16	0.65	3.75	5.04	0.00	100.00		99.11
Topaz granites associated with rapakivi granites, Finland														
Suomenniemi-4	Rämö and Haapala (1995)	75.40	0.10	13.52	1.36	0.03	0.09	0.83	3.61	5.03	0.03	100.00	0.53	99.66
Eurajoki-2	Rämö and Haapala (1995)	75.99	0.06	13.86	1.45	0.04	0.02	0.87	3.32	4.33	0.06	100.00	0.50	100.25
Eurajoki 5/IH/2001	Haapala et al. (2005)	76.48	0.02	13.75	0.92	0.05	0.00	0.65	3.73	4.39	0.02	100.00	0.24	99.93
Saaskjarvi-17E	Lukkari (2002)	74.38	0.20	13.45	2.14	0.01	0.21	0.92	3.01	5.58	0.10	100.00	0.50	100.30
Ahvenisto	Edén (1991)	75.14	0.02	14.25	2.00	0.04	0.01	0.71	2.86	4.96	0.01	100.00		99.06
Kymi	Haapala et al. (2005)	74.29	0.02	15.24	0.87	0.02	0.10	0.77	4.09	4.59	0.01	100.00	0.40	100.15

	F	Cl	Sc	V	Cr	Ni	Zn	Ga	Rb	Sr	Y	Zr	Nb	Ba	La	Ce	Nd	Sm	Pb	Th	U
Topaz Rhyolites, western United States																					
Blawn Formation (22 to 18 Ma)																					
LAM-1-38-2	2900	100	1	2	3	6	113	24	626	6	85	156	117	25	47	91	43	12	49	62	16
TET-9-43-2	5100	100	3	3	2	7	78	25	636	7	52	155	149	30	43	78	31	10	48	72	22
RED-04	2883	778	1	2	4	6	90	26	813	2	82	159	150	17	34	84	38	9	51	49	30
RED-06	4038	79	3	6	3	8	53	20	446	8	73	158	83	10	63	116	53	12	34	63	11
RED-12	3660	79	3	5	3	11	40	20	477	12	64	157	82	10	52	96	44	11	37	61	10
RED-33	5240	117	3	5	2	9	52	26	709	3	90	148	136	7	50	89	41	10	42	54	15
RED-34	4708	77	1	4	2	7	83	27	789	2	74	136	152	3	57	87	47	10	50	49	12
RED-40	3938	77	2	2	–	9	74	27	796	5	94	144	153	23	38	71	31	8	51	51	19
RED-41	1845	143	2	1	2	14	108	26	641	14	59	162	153	9	34	30	25	7	44	27	13
Steamboat Formation (13 to 11 Ma)																					
BBS-8-229-2	2000	60	2	2	1	6	43	23	553	4	79	147	90	18	40	95	39	9	48	57	17
LAM-9-103-2	900	100	2	4	1	7	20	29	780	12	31	111	139	52	37	92	37	9	56	60	20
WW-8014	5804	59	1	7	6	3	46	21	522	8	112	170	86	12	46	110	45	13	40	62	16
WW-8016	5720	97	2	3	6	1	79	28	727	2	131	164	122	18	32	80	39	11	58	70	32
WW-8018	5461	113	0	3	5	1	44	27	679	3	136	177	126	20	31	81	37	9	44	64	18
WW-8020	4621	90	1	8	6	1	65	22	511	3	104	168	81	17	72	122	52	15	51	61	15
WW-8029	4172	1721	0	5	6	0	105	27	715	0	128	173	124	10	31	81	37	9	57	70	28
WW-8011A	4900	78	1	8	6	1	46	25	671	3	90	172	143	16	32	81	38	9	40	70	21
WW-8011B	3845	1238	1	5	5	1	77	23	521	24	93	160	91	155	41	91	37	10	49	55	21
Spor Mountain Rhyolite (21 Ma)																					
	10205	609	3	7	6	2	80	36	1048	8	118	121	131	21	59	142	63	15	43	76	31
Topaz granites associated with rapakivi granites, Finland																					
Suomenniemi-4	7100		4				47	24	567	21	116	118	51	163		161					
Eurajoki-2	10400							74	965	10		70	60	150	40	91	34	9		30	9
Eurajoki 5/IH/2001	11900	100	12				197	60	1050	8	56	51	70	28	47	97	37	9	88	28	6
Saaskjarvi-17E	2500	195	3	2	18		48	26	412	65	40	171	22	282					49	32	12
Ahvenisto	14270								740	20		70	80	70	20	100				40	10
Kymi	14500	110	8				67	61	978	22	7	12	58	151	65	154	35	8	130	31	

Fe_2O_3* = Total Fe as Fe_2O_3.

Major element concentrations normalized to 100% on a volatile-free basis. Trace element concentrations in ppm.

minerals include zircon, thorite, uraninite, allanite, apatite, fluorite, and Fe–Ti–Mn–Nb oxides.

Mineral geothermometry indicates that most topaz rhyolites crystallized at low temperatures around 650 to 700 °C. Oxides reveal that fO_2 was commonly low, near the QFM oxygen buffer, although some topaz rhyolites crystallized under fairly oxidizing conditions as indicated by oxide and biotite compositions and the presence of titanite in Lake City and Chalk Mountain, Colorado, Mineral Range, Utah, Sheep Creek Mountains and Jarbidge, Nevada, rhyolites. It thus appears that there are oxidized and less oxidized topaz rhyolites, analogous to Proterozoic anorogenic granites from the western United States which consist of an ilmenite- and a magnetite-series (Anderson, 1983; Anderson and Morrison, 2005). All topaz rhyolites in the western United States are two-feldspar rhyolites, in contrast to many other bimodal rhyolites—e.g., many of the rhyolites of the Snake River Plain, Idaho (Leeman, 1982) and the peralkaline rhyolites of the western Great Basin. In general, one-feldspar rhyolites crystallize at higher temperatures than those inferred for topaz rhyolites.

Topaz, fluorite, alkali feldspar, spessartine garnet, hematite, bixbyite, pseudobrookite, and silica minerals line gas cavities and occur in the devitrified matrix of some lavas. Gem-quality red beryl (up to 1 cm long) occurs in one lava flow from the older episode of rhyolitic volcanism in the Wah Wah Mountains (Fig. 2), but is also found in the Thomas Range, Utah, and in the Black Range of New Mexico. Peralkaline minerals (aegerine, riebeckite, etc.) are absent in the topaz-bearing flows, but Nevada's Kane Springs caldera erupted peralkaline trachyte and comendite ash-flow tuffs shortly before intracaldera topaz-bearing lavas (Novak, 1984).

6.2. Mineralogy of rapakivi granites

Finnish rapakivi granite plutons range from subsolvus hornblende granite to biotite granite and further to leucocratic topaz-bearing subsolvus granite. Some of the more mafic varieties have fayalite or an Fe-rich clinopyroxene. Typically, potassium feldspar is the most abundant mineral accompanied by quartz, plagioclase (oligoclase to andesine), and Fe-rich biotite (Fig. 4). In the biotites, Fe/(Fe+Mg) generally exceeds 0.8 and Al_{tot} is also similar to that in biotite from topaz rhyolites (Rieder et al., 1996). On the other hand, F in biotite is comparatively low (ranging from 0.2 to 0.7 wt.%) in rocks that lack topaz (Elliott, 2001). Accessory minerals are fluorite, zircon, apatite, ilmenite, magnetite, and allanite or monazite. Rapakivi texture is common only in the larger, more mafic plutons and rare or absent in the smaller, more evolved plutons (Rämö and Haapala, 1995). Peralkaline rocks are rare, but Rämö (1991) described peralkaline hypersolvus syenite dikes in the Suomenniemi complex.

The youngest phases of several Finnish rapakivi granites are felsic, porphyritic or equigranular microcline-albite granites that often contain topaz and are associated with greisen-type mineralization (Fig. 2). In these topaz granites, the dark mica is F-rich siderophyllite (Fig. 4). Fluorine is as high as 5 wt.% in these micas (Haapala, 1977; Haapala and Lukkari, 2005). Characteristic accessory minerals are fluorite, monazite, bastnaesite, ilmenite, cassiterite, columbite, and thorite, along with rare zircon, apatite, and magnetite. Miarolitic cavities are common and subsolidus reactions are evident in their textures and mineral compositions (Haapala, 1977; Rämö and Haapala, 1995; Lukkari, 2002; Haapala and Lukkari, 2005). The presence of miarolitic cavities and small pockets of pegmatite suggests that the topaz rapakivi granites became water-saturated during crystallization. Less evolved rocks probably crystallized in undersaturated conditions. Temperatures of crystallization have been estimated for the Eurajoki stock of 750 °C at 2 kb for fayalite–biotite–hornblende granite (Haapala, 1977). Using amphibole-plagioclase geothermometry, Elliott (2001) deduced temperatures averaging about 725 °C for several granites from the Wiborg batholith. Two-feldspar temperatures in the same rocks were typically 650 °C. The Fe–Ti oxides in most rapakivi granites are dominated by ilmenite, but Kosunen (1999) has shown that the Obbnas pluton of southern Finland has magnetite and titanite. Magnetite and titanite are also typical accessory minerals in the 1.6 Ga anorogenic granites of Estonia (Kirs et al., 2004).

7. Geochemistry and differentiation trends

7.1. Geochemistry of topaz rhyolites from the Wah Wah Mountains

Representative analyses of topaz rhyolites from the Wah Wah Mountains of southwestern Utah are given in Table 1, along with the composition of an average topaz rhyolite from Spor Mountain, central Utah. The major element composition of topaz rhyolites from the Wah Wah Mountains is fairly restricted. All are high-silica rhyolites with high Na, K, F, Fe/Mg and low Ti, Mg, Ca, and P (Table 1; Fig. 5). Alkali oxides range between 8% and 10%. In general K_2O/Na_2O ratios are greater than one (typically about 1.2 to 1.4 by weight) but this ratio declines with differentiation. Most topaz rhyolites contain 12% to 14% Al_2O_3 (Table 1). In spite of the

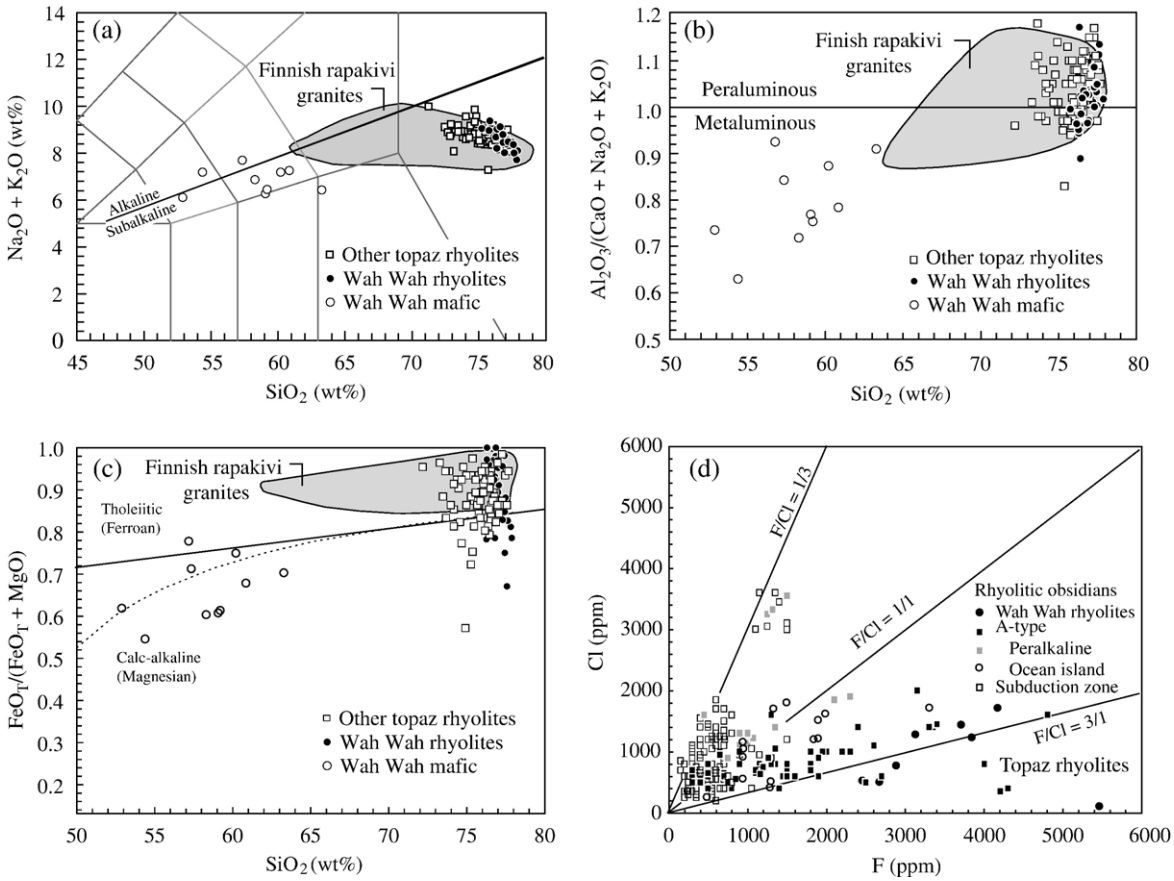


Fig. 5. The compositions of topaz rhyolites from the western United States and Finnish rapakivi granites are similar as shown on these variation diagrams. Field for Finnish rapakivi granites from Rämö and Haapala (1995). (a) Total alkalies versus silica. Gridlines are from the IUGS chemical classification of volcanic rocks. The bimodal character of the Wah Wah volcanic rocks is evident by a gap between 63 and 75% SiO_2. (b) Al-saturation index versus silica. Like rapakivi granites, most topaz rhyolite suites straddle the dividing line between metaluminous and peraluminous compositions. Alkali loss after eruption may be responsible for some of the peraluminous rocks. (c) Most topaz rhyolites are ferroan on a FeO/(FeO+MgO) versus silica diagram. Mafic rocks from the Wah Wah Mountains are both ferroan and magnesian. Solid line is from Miyashiro (1974) and dashed line is from Frost et al. (2001a). (d) F and Cl concentrations in glassy rhyolites including topaz rhyolites. Data from Macdonald et al. (1992) on rhyolite obsidians included.

continued misperception, topaz rhyolites are neither peralkaline nor strongly peraluminous (Fig. 5). The presence of garnet and topaz (absent as vapor-phase minerals in peralkaline volcanic rocks) reveals their aluminous character. Vitrophyres are either slightly peraluminous or metaluminous. However, topaz rhyolites are not the eruptive equivalent of S-type granites (e.g., White and Chappell, 1983) and are decidedly different from the P-rich strongly peraluminous topaz-bearing granites that are associated with some of them (London et al., 1999; Chappell and Hine, 2006). The high Fe/Mg ratios of topaz rhyolites mark them as mostly tholeiitic (or ferroan in the sense of Frost et al., 2001a) in contrast to the older calc-alkaline (or magnesian) magmatism that preceded them in most areas of the western United States (Fig. 5).

Of course, the most discriminating feature of topaz rhyolites is their high fluorine content. For Wah Wah rhyolites, fluorine concentrations in vitrophyres range from 0.2 to 0.5 wt.% (Fig. 5). Topaz appears as an identifiable vapor-phase mineral in lavas whose vitrophyres contain over 0.2 wt.% F. Fluorine concentrations over 1 wt.% are only known from vitrophyres from Spor Mountain and the Honeycomb Hills complex (which has magmatic topaz), both in western Utah. Comparisons of vitrophyre–felsite pairs almost universally show that F is lost during devitrification. Chlorine is even more strongly depleted during devitrification of volcanic glass; no mineral phase concentrates Cl in contrast to F with its mineralogical hosts, topaz and fluorite. Thus, meaningful halogen concentrations can only be obtained by analysis of vitrophyres or obsidians.

In such fresh rocks, Cl concentrations are typically less than 0.2 wt.% and generally much lower than this. The high F/Cl ratios (greater than about 3) of topaz rhyolites set them apart from both calc-alkaline and peralkaline rhyolites which have much lower F/Cl ratios (Christiansen and Keith, 1996). Studies of melt inclusions in the Spor Mountain rhyolite show that only a small amount of the Cl and little F was lost during eruption (Zhang and Christiansen, unpublished data). Rapakivi granites, in contrast, have much lower concentrations of Cl, probably as a result of post-magmatic fluid loss. However, topaz-bearing varieties have F concentrations that commonly exceed 1 wt.% (Haapala, 1977; Rämö, 1991; Lukkari, 2002; Haapala and Lukkari, 2005).

The trace element compositions of Wah Wah rhyolites show the strong enrichments of Rb, U, Th, Pb, Ga, Nb, Ta, Y, Cs, Zn, Sn, Be, and Li typical of topaz rhyolites. In contrast, Sc, Ni, Co, Cr, V, Zr, Hf, Ba, Sr, Eu, and P are all strongly depleted. P_2O_5 concentrations are below 0.02% in almost all topaz rhyolites, including those from the Wah Wah Mountains (Table 1). In contrast, topaz granites and pegmatites associated with strongly peraluminous magmas are phosphorous-rich (Chappell and White, 1992; Christiansen and Keith, 1996; London et al., 1999). High Ga/Al ratios of the Wah Wah rhyolites are typical with average 10,000 * Ga/Al of about 6. This index is used by Whalen et al. (1986) as a key indicator of A-type granites when over 2.5.

REE patterns for topaz rhyolites show some variation (Christiansen et al., 1986), but they generally have low La/Ce_N, La/Yb_N and Eu/Eu* (0.45 to 0.01 for analyzed specimens). REE patterns in samples from the Wah Wah Mountains are similar to most topaz rhyolites and to Finnish topaz granites (Fig. 6). Light REE concentrations generally do not exceed 200 times chondrite values and typically they are less than about 100 times chondrite.

As expected, trace element patterns (Fig. 6B) show strong depletions in Ba, Sr, P, and Ti, but they are most notable for their strong enrichments of Rb, Th, and U and small negative Nb anomalies. Deep negative Nb anomalies are common in subduction zone rhyolites such as those that erupted in the Oligocene of the Great Basin (e.g., in the Wah Wah Mountains the Oligocene dacites and rhyolites of the Indian Peak volcanic field). However, negative Nb anomalies are also apparent in the mafic magmas that accompanied the eruption of the Wah Wah topaz rhyolites.

Silica contents vary little with differentiation as seen in individual dome complexes. For example, silica ranges from 76.2 to 77.8 wt.% in the fresh rhyolites of the Wah Wah Mountains, whereas incompatible elements

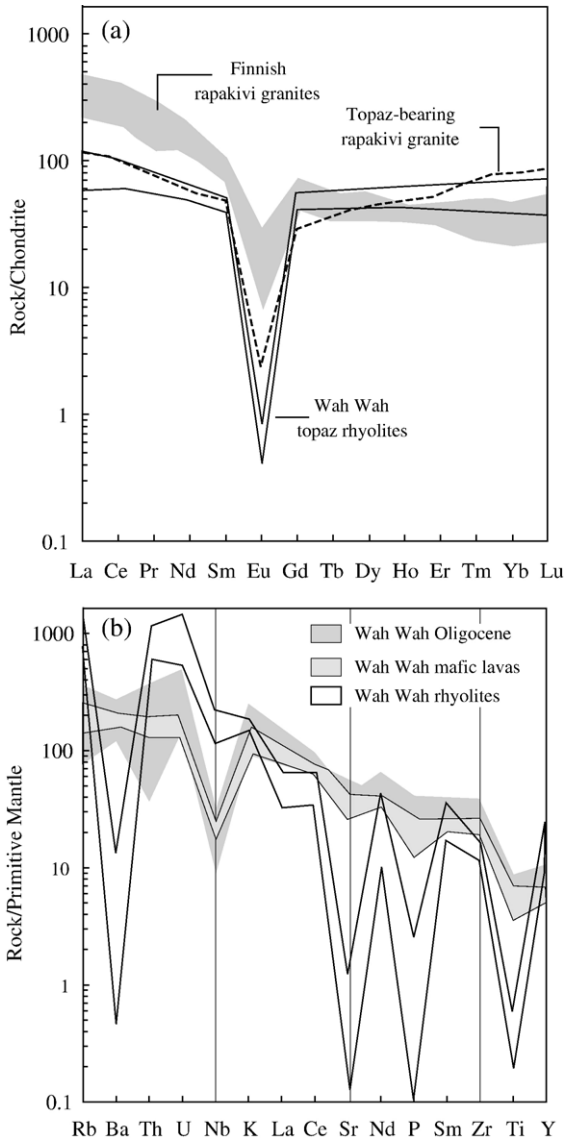


Fig. 6. Trace element compositions of Miocene volcanic rocks from the Wah Wah Mountains compared with Proterozoic rapakivi granites from southern Finland. (a) Chondrite-normalized REE patterns of topaz rhyolites from the Wah Wah Mountains compared with Finnish rapakivi granites (Haapala and Lukkari, 2005). (b) Primitive mantle normalized extended trace element patterns for contemporaneous mafic and silicic lavas from the Wah Wah Mountains. Note the prominent negative Nb anomaly in these rift-related mafic lavas. Normalizing values from McDonough and Sun (1995).

such as Rb, Nb, and U double in concentration. Silica contents are actually lower in rhyolites which are extremely enriched in fluorine and incompatible elements such as the lavas at Spor Mountain and Honeycomb Hills, Utah. K generally declines with increasing concentrations of F and other decidedly incompatible elements, whereas Na increases. An enrichment in fluorine, Na/K,

and incompatible elements is also seen in the Eurajoki and Kymi topaz granites of the Finnish rapakivi complexes (Haapala, 1977; Haapala and Lukkari, 2005). Haapala (1977) and Christiansen et al. (1984) interpreted these trends in granites and rhyolites as resulting from crystallization near the minimum in the granite system with elevated fluorine concentrations (Manning, 1981).

Incompatible elements increase in concentration, locally dramatically, and include Rb, U, Th, Pb, Ga, Nb, Ta, Y, Sn, Li, Cs, and Be. Elements that decline during differentiation include Ti, Fe, Mg, Sc, Ni, Co, Cr, and V (removed by the fractionation of mafic silicates and oxides), Ca, Ba, Sr, and Eu (depleted by feldspar crystallization), as well as P, Zr, and Hf (apatite and zircon are relatively insoluble in metaluminous melts, so removal of these phases keeps element concentrations low). Fe/Mg ratios increase with differentiation because at low oxygen fugacities biotite Fe/Mg ratios are substantially lower than Fe/Mg of melt and the fractionation of magnetite is not pronounced. Differentiation trends for the REE show decreases in LREE and Eu and increases in HREE concentrations. The decline in LREE probably results from the fractionation of small amounts of allanite, chevkinite, monazite, or other REE-rich accessory phases. Likewise, in extremely evolved magmas, U, Th, Y, and perhaps Nb may also decline as uraninite, thorite, xenotime, titanite, Nb-rich ilmenite, and Nb oxides reach saturation and fractionate from the magma (Funkhouser-Marolf, 1985).

In short, the elemental composition of cogenetic lavas reveals the importance of crystal fractionation near the granite minimum. Fractionation involved the removal of sanidine > quartz > plagioclase ≫ biotite > Fe–Ti oxides ≫ apatite > zircon > allanite/monazite/chevkinite. The extreme depletions and enrichments can be explained by fractional crystallization of 70 to 85% of a parental rhyolite (Christiansen et al., 1984; Moyer and Esperança, 1988). Variable degrees of partial melting will not produce the extreme depletions of compatible elements seen in differentiated topaz rhyolites.

7.2. Geochemistry of rapakivi granites from Fennoscandia

Granites in rapakivi suites have high Si, K, F, Rb, Ga, Zr, Hf, Th, U, Zn, and REE (except Eu), and low Ca, Mg, P, and Sr abundances compared to granitic rocks in general (Rämö and Haapala, 1995). Rapakivi granites also have high Fe/Mg, K/Na, and total alkalies. Similar to topaz rhyolites of western United States, the Finnish rapakivi granites straddle the peraluminous–metaluminous boundary (Fig. 5) and have high contents of alkalies (average $Na_2O+K_2O=8.4$ wt.%). Typically the K_2O/Na_2O ratios are above 1. The Fe/Mg ratios are also high, with Fe/(Fe+Mg) averaging about 0.9 (Fig. 5). The extreme enrichments of incompatible trace elements are exemplified by F (ranging from 0.04% to 1.53% with an average of 0.35%) and Rb (to over 1000 ppm). High Ga/Al ratios are typical with 1000 * Ga/Al varying from 1.8 to 15 and an average of 4.2.

As in topaz rhyolites, differentiation trends within individual rapakivi complexes show increases in Si, F, Ga, Rb, Sn, Nb, and decreases in Ti, Al, Fe, Mg, Mn, Ca, Ba, LREE, Eu, Sr, Sc, and Zr (Rämö and Haapala, 1995). Topaz-bearing biotite granites are usually the youngest and the most highly evolved units. Compared to less differentiated (parental?) rocks, they also have lower K/Na and higher Fe/Mg ratios. The highest Rb/Sr, Rb/Ba, and Ga/Al ratios are all found in the topaz-bearing granites. La/Yb_N ratios in rapakivi granites average about 9, but are much lower (about 1) in the topaz-bearing phases. These flat REE patterns also have low Eu/Eu* (Fig. 6).

High Zr and Hf concentrations are the only major or trace element features of common rapakivi granites that are not characteristic of topaz rhyolites. For example, hornblende, biotite–hornblende, and biotite granites from the Suomenniemi batholith have Zr concentrations that range from 600 to 400 ppm (Rämö and Haapala, 1995). Zirconium concentrations are only 150 to 180 ppm in the Wah Wah rhyolites (Table 1); even lower Zr concentrations are found in other topaz rhyolites. However, if the topaz-bearing phases of rapakivi granite complexes are considered, this difference disappears. Zirconium concentrations range from 160 to 180 ppm in the Sääskijärvi topaz granite (Lukkari, 2002), to as low as 70 ppm in the Eurajoki granite (Rämö and Haapala, 1995), and range from 12 to 30 ppm in some phases of the highly evolved Kymi granite (Haapala and Lukkari, 2005). Hafnium and Zr are compatible elements in aluminous granitic magmas such as these; their concentrations decline with differentiation because zircon solubility is limited at low temperatures.

Fig. 7 shows the composition of topaz rhyolites and Finnish rapakivi granites on several tectonic or magmatic discrimination diagrams. In each diagram, the topaz rhyolites overlap with the Finnish rapakivi granites. In general, these distinctive rhyolites and granites plot in the within-plate or A-type fields as contrasted with the I-, S-, and M-type granites distinguished on the diagrams of Whalen et al. (1986) and volcanic arc or syn-collisional granites of Pearce et al. (1984). In the Rb–Y+Nb diagram, topaz rhyolites from the western United States and many rapakivi granites plot near the boundary

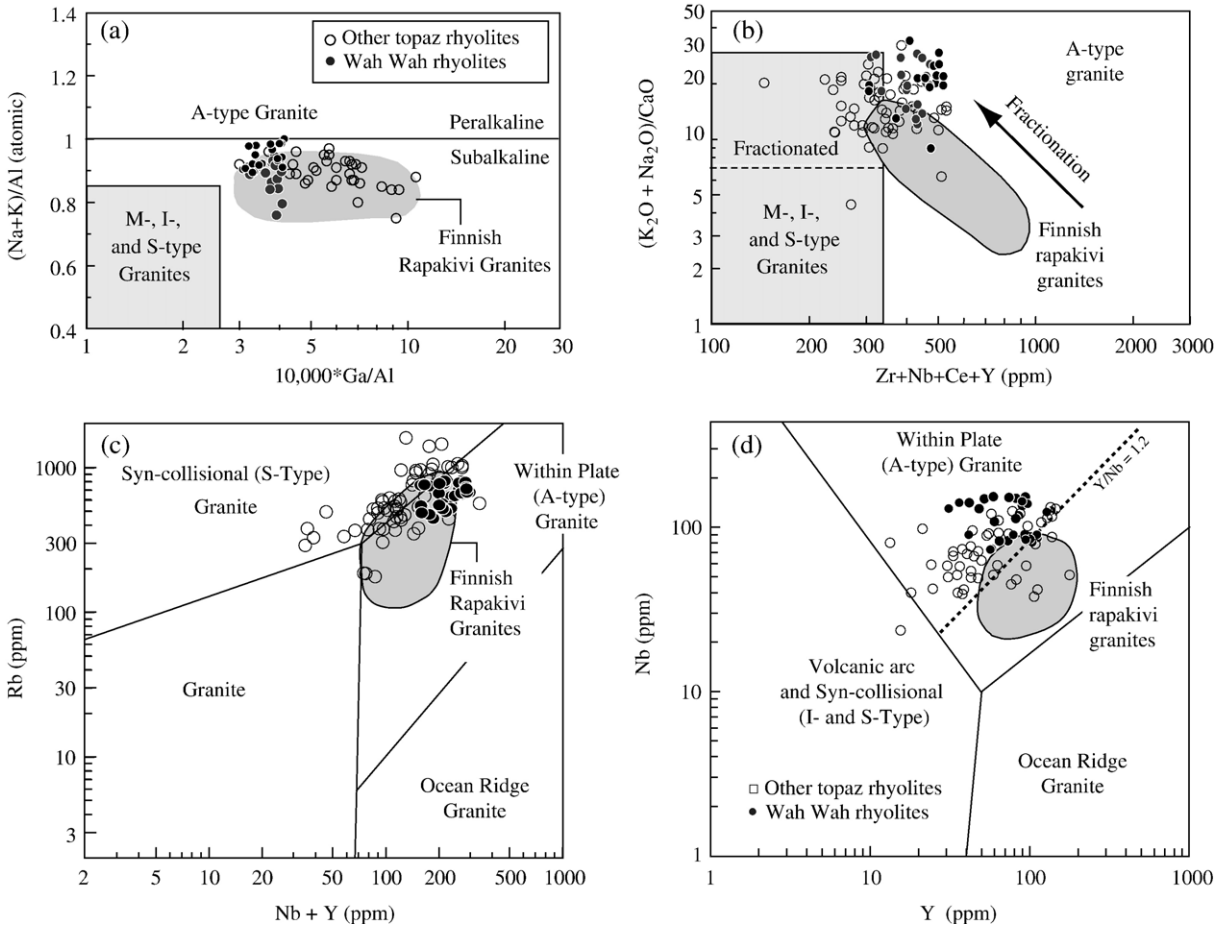


Fig. 7. Topaz rhyolites plot with other metaluminous A-type silicic rocks on the discrimination diagrams of Whalen et al. (1986) and Pearce et al. (1984). Rapakivi granites, especially those with topaz, plot in similar positions except on the Nb–Y diagram where most topaz rhyolites are richer in Nb and/or poorer in Y. Subdivision of within-plate granite field in (d) from Eby (1992).

between WPG (within-plate granites) and syn-COLG (syn-collisional granites) (Fig. 7). On the Nb–Y diagram many topaz rhyolites are richer in Nb and/or poorer in Y than most rapakivi granites. However, here again, most of the topaz granites have lower Y/Nb ratios than parental rapakivi granites. Fractionation of xenotime or titanite may raise the partition coefficients for Y and thereby decrease Y/Nb to lower values during differentiation.

8. Metallogeny

8.1. Metallogeny of topaz rhyolites

Mineral deposits associated with topaz rhyolites in the Wah Wah Mountains include small deposits of fluorite, uranium, alunite, and native S, as well as gem red beryl in one of the older flows (Christiansen et al., 1996; Thompson, 2002). A large, but unexploited, Climax-type porphyry Mo deposit is located at Pine Grove (Fig. 2). The hydrothermal alteration at Pine Grove includes topaz. Younger topaz rhyolite dikes cut the buried intrusion, but the relationship to the erupted topaz rhyolites is unclear. The Mo-mineralized magma system did erupt a garnet-bearing rhyolite tuff (Keith et al., 1986). Elsewhere in the western United States, large deposits of Be (as bertrandite) and Climax-type Mo(W) deposits, small deposits of U and F, and sub-economic occurrences of red beryl, Li, Cs, and Sn are directly related to topaz rhyolites. The rhyolites are contemporaneous with the deposits, co-magmatic with mineralized intrusions, and, in some cases, hosts of the ore. The marked magmatic enrichment of these same elements strongly suggests that the ore elements were derived from the rhyolites (or their intrusive forerunners in the case of Climax-type Mo deposits; Burt et al., 1982). Other types of mineralization (alunite, S, Hg, Au–Ag) are spatially and temporally associated with some topaz rhyolites (e.g., John, 2001). The association

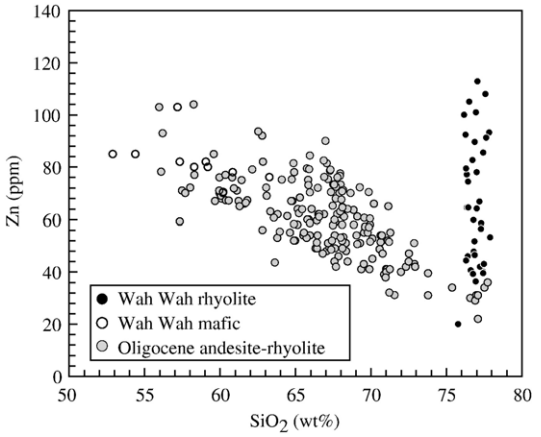


Fig. 8. SiO_2 versus Zn for volcanic rocks from the Wah Wah Mountains of southern Utah. Topaz rhyolites (and other young bimodal rhyolites) are much richer in Zn than calc-alkaline subduction-related silicic rocks of Oligocene age that erupted about 10 m.y. earlier.

of these latter deposits with the rhyolites may rely more on magmatic heat content and volcanologic style for their generation than on any particular compositional feature of topaz rhyolites.

8.2. Metallogeny of rapakivi granites

Greisen-type Sn–Be–W–Zn mineralization (with beryl, genthelvite, and bertrandite as the Be minerals) is associated with the topaz-bearing granites of southern Finland (Haapala, 1977; Edén, 1991; Haapala, 1995, 1997). This element association is similar to that of topaz rhyolites, with the possible exception of Zn. But it should be noted that many bimodal rhyolites (as well as rapakivi granite suites) are exceptionally rich in Zn for their high SiO_2 contents (Fig. 8).

9. Isotopic compositions

9.1. Isotopic compositions of topaz rhyolites from the Wah Wah Mountains

Initial Sr, Nd, and Pb isotope ratios for topaz rhyolites from the Wah Wah Mountains are reported in Table 2. The Pb isotope ratios of the silicic rocks are indistinguishable from those of the contemporaneous mafic lavas, except for the $^{208}Pb/^{204}Pb$ ratios which suggest that the silicic magmas were derived from sources with slightly higher Th/Pb ratios than the sources of the mafic magmas. Considerable uncertainties exist in the initial $^{87}Sr/^{86}Sr$ ratios because of the extremely high Rb/Sr ratios of the rocks. The error bars show the effects of recalculation by assuming a 1 ppm difference in the Sr concentration of these low Sr rocks. Our best estimate is that initial $^{87}Sr/^{86}Sr$ ratios for topaz rhyolites from the Wah Wah Mountains are between 0.706 and 0.710. A further constraint is that the initial $^{87}Sr/^{86}Sr$ ratio of the topaz-bearing Sheeprock granite from western Utah is 0.7064 as taken from a Rb–Sr isochron (Christiansen et al., 1988). Care must be taken because a small amount of upper crustal contamination would significantly raise the Sr isotope ratios of these low Sr magmas. For example, Reece et al. (1990) modeled an increase of the initial $^{87}Sr/^{86}Sr$ ratio from 0.705 to 0.712 as resulting from the assimilation of only 1% of radiogenic upper crustal wall rocks into Taylor Creek Rhyolite of New Mexico's Black Range.

These low initial $^{87}Sr/^{86}Sr$ ratios and the implied low Rb/Sr ratios of the sources were originally attributed to derivation of topaz rhyolites from felsic granulitic rocks in the lower continental crust with little or no involvement of mantle or "juvenile" crust (Christiansen et al., 1983, 1988). However, new Nd isotopic data from the Wah Wah

Table 2
Strontium, neodymium, and lead isotopic compositions of Miocene lavas from the Wah Wah Mountains, Utah

Sample #	Rock Type	$^{87}Sr/^{86}Sr$ (T)	ε_{Nd} (T)	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
Steamboat Formation (13 to 11 Ma)						
BBS-8-229-2	Topaz rhyolite	0.710*	−11.4	18.1	15.6	39.4
LAM-9-103-2	Topaz rhyolite	0.710*	−9.7	18.2	15.6	39.3
LAM-9-103-3	Bas trachyandesite	0.7054	−7.2	17.5	15.5	37.6
LAM-956-1	Trachyandesite	0.7052	−6.8	18.1	15.6	38.1
Blawn Formation (22 to 18 Ma)						
LAM-1-38-2	Topaz rhyolite	0.710*	−9.3	18.2	15.6	39.4
TET-9-43-2	Topaz rhyolite	0.710*	−9.5	18.1	15.6	39.3
BAN-8-127-2	Bas trachyandesite	0.7065	−4.4	18.7	15.6	38.9
FRSC-2-62-3	Trachyandesite	0.7060	−8.8	18.3	15.6	38.6

Procedures and errors are described in the text.
* $^{87}Sr/^{86}Sr$ (T) for topaz rhyolites have uncertainties on the order of +/−0.005 because of extremely high Rb/Sr ratios.

Mountains are inconsistent with that interpretation (Table 2 and Fig. 9). The Nd isotope ratios are higher (εNd −9 to −11) in the topaz rhyolites from the Wah Wah Mountains than in the slightly older calc-alkaline suite (εNd −12 to −19) and other granites and rhyolites from the eastern Great Basin, but the Nd isotope ratios are slightly lower than in the contemporaneous mafic magmas (εNd −4 to −9). Also plotted on Fig. 9a, are the compositions of average mafic and felsic xenoliths from the Colorado Plateau (Condie et al., 1999), which may be the best current estimates of the lower crustal composition in this part of the western United States.

Although it is clear that the topaz rhyolites of the Wah Wah Mountains cannot be derived simply by partial melting of a felsic component in the crust, they could be derived from partial melts of a mixture of felsic and mafic crustal components of Proterozoic age. Alternatively, the Sr and Nd isotopic data could be explained by the inclusion of an important mantle component or "juvenile" crustal component composed of young mantle-derived mafic magmas (represented by the contemporaneous mafic volcanic rocks) trapped in the crust and then re-melted.

9.2. Isotopic composition of rapakivi granites

Rämö and Haapala (1995) and Haapala et al. (2005) have reviewed the isotopic compositions of many rapakivi granites. As noted earlier, most rapakivi granites are only a few hundred million years younger than the Proterozoic crust into which they intruded. Thus, it is difficult to clearly ascertain the nature of their sources using slowly changing Nd isotope ratios. For example, rapakivi granites from southern Finland (εNd 0 to −3) lie within the broad isotopic evolution field for 1.9 Ga Svecofennian crust, but because the granites are only 300 million years younger than this crust, their derivation entirely from continental crust is not assured. The εNd values (+1.6 to −1.7) of rapakivi-age diabase dikes and other mafic rocks largely overlap those of the rapakivi granites (Rämö, 1991; Haapala et al., 2005). Rapakivi granites that intrude older Archean terranes are rare, but the isotopic evidence for the nature of their sources is clearer. For example, rapakivi granites and gabbro-anorthosites of the Salmi batholith in Russian Karelia (εNd at 1540 Ma −6.2 to −9; Rämö, 1991; Neymark et al., 1994) and the rapakivi granites of the Shachang complex in northeastern China (εNd at 1685 Ma about −5.7; Rämö et al., 1995; Haapala et al., 2005) have much higher εNd values (5 to 7 epsilon units) than the Archean crust they intrude or than crustal evolution curves with typical Sm/Nd ratios for

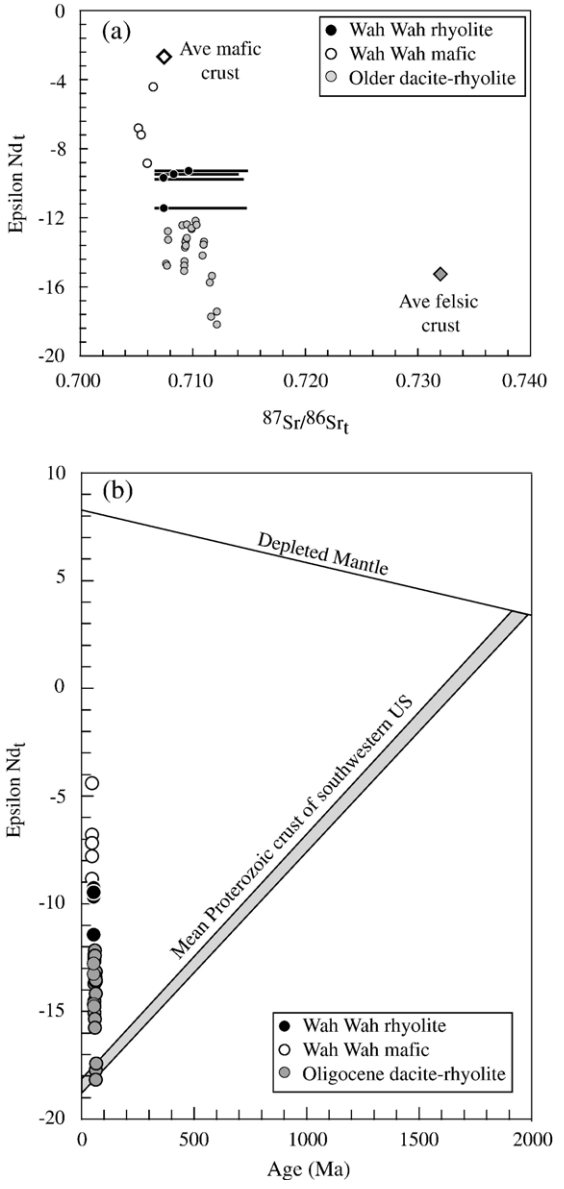


Fig. 9. Isotopic composition of volcanic rocks from the Indian Peak volcanic field of the Wah Wah Mountains, southern Utah. (a) Topaz rhyolites (black circles) are very distinct from the felsic xenoliths from the lower and middle crust (Ave felsic crust) of the Colorado Plateau analyzed by Condie et al. (1999). If topaz rhyolites are derived from ancient crust, then the crust must be more similar to mafic xenoliths (Ave mafic crust) of the Colorado Plateau examined by Wendlandt et al. (1993) which have much higher Sm/Nd ratios than typical Proterozoic crust. Alternatively, young mantle-derived magmas may have been a significant component in the sources of topaz rhyolites from the western United States. (b) Miocene topaz rhyolites from the Wah Wah Mountains have distinctly higher epsilon Nd values than Oligocene dacite and rhyolite (gray circles) from the region and plot far above the evolution curve for Proterozoic crust (Miller and Wooden, 1994).

felsic crust. This could be the result of mixed crustal sources consisting of Archean and Proterozoic components *or* it could mean that juvenile mantle-derived components were involved in their generation. In addition, isotopic evidence shows that parts of the Sherman batholith of Wyoming cannot be derived solely from the felsic Archean crust it intrudes (Frost et al., 1999, 2001b). The Cretaceous bimodal A-type granite complexes of western Namibia (including the Spitzkoppe topaz-bearing granites described by Frindt et al., 2004b) that are related to continental rifting and mantle-plume activity show geochemical and isotopic (Nd, Sr, O) compositions that suggest varying mixing relations between plume-related mantle magmas and lower crustal rocks (Trumbull et al., 2004; Frindt et al., 2004a). The initial $^{87}Sr/^{86}Sr$ ratios and εNd values of the topaz-bearing granite are distinct from the metasedimentary gneisses it intrudes (McDermott and Hawkesworth, 1990; Seth et al., 1998) but overlap substantially with one type of the contemporaneous Etendeka flood basalts. εNd ranges from −4 to −7 for the LTZ.L flows, which are interpreted to have interacted extensively with mafic lower crust (Ewart et al., 1998a,b).

10. Discussion

10.1. Topaz rhyolites are the eruptive equivalents of rapakivi granites

Based on similarities in the mineralogy, calculated volatile fugacities, elemental and isotopic compositions, and metallogeny, we conclude that topaz rhyolites of the western United States are the Cenozoic equivalents of highly evolved rapakivi granite magmas. As such, they may help us better understand the origin and evolution of rapakivi granites and other A-type felsic magmas. In the following paragraphs, we try to synthesize the most important interpretations that derive from this conclusion.

10.2. Textures

Rapakivi textures have not been found in any topaz rhyolite from the western United States, in spite of other evidence for magma mixing, rapid decompression that accompanied eruption, and other changes in the physical and chemical conditions that have been suggested as causing the distinctive rapakivi texture (Rämö and Haapala, 1995; Eklund and Shebanov, 1999). The rapakivi texture is also absent in the topaz-bearing phases of rapakivi complexes of southern Finland. Instead, the texture is most common in the more mafic hornblende-bearing phases of the composite intrusions. Thus, topaz rhyolites may not provide a test for the origin of the rapakivi texture.

10.3. Bimodality

Topaz rhyolites are mostly, but not exclusively, members of bimodal volcanic series. This is apparent for the Wah Wah Mountains in Fig. 5 and for the northern Basin and Range Province as a whole in Fig. 10. Before about 25 Ma, volcanism was dominated by a nearly continuous series of andesite, dacite, and rhyolite compositions. After this date, intermediate magmatism declined in importance and basaltic magmas erupted. Topaz rhyolites formed during this younger, bimodal volcanism.

Likewise, rapakivi granites are often associated with mafic rocks ranging from gabbros to anorthosites, ferrodiorites, and basaltic dikes (Rämö and Haapala, 1995). The common bimodality of these distinctive A-type magmas must be accounted for in any successful petrogenetic model.

10.4. Tectonic setting: orogenic, anorogenic, or taphrogenic

None of the topaz rhyolites of the western United States were erupted during periods of contractional

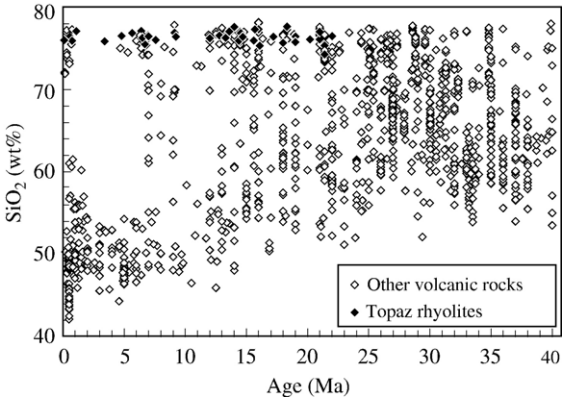


Fig. 10. SiO_2 (normalized on anhydrous basis) content versus time for igneous rocks of the eastern Great Basin (east of $^{87}Sr/^{86}Sr$ line in Fig. 1). Middle Cenozoic magmatism was dominated by intermediate to silicic magmatism. Beginning between 25 and 20 Ma mafic lavas were erupted together with high-silica rhyolites and intermediate compositions were rare. Topaz rhyolites erupted as an important part of this late Cenozoic bimodal volcanism. Data from middle Cenozoic compilation of Barr (1993), augmented by data from Vitaliano and Vitaliano (1972), Clark (1977), Best et al. (1980), Ekren et al. (1980), Novak (1984), Christiansen et al. (1986), Fitton et al. (1988), Coleman and Walker (1992), Moore (1993), Miller and Wrucke (1995), Brooks et al. (1995), Rogers et al. (1995), Fleck et al. (1996), Beard and Johnson (1997), Nelson and Tingey (1997), Cunningham et al. (1998), Smith et al. (1999), DePaolo and Daley (2000), and Clark (2003).

orogenesis; almost all are clearly related to extension (taphrogeny). A few of the older rhyolites may have erupted during a nearly neutral tectonic environment with the least principal stress oriented vertically—a truly anorogenic setting. For example, laccoliths in southern Utah and Montana are about the same age as some of the older topaz rhyolites in these regions. Fig. 11 outlines our interpretation of the tectonic evolution of the western United States (e.g., Best and Christiansen, 1991). Before about 45 Ma, shallow subduction of oceanic lithosphere beneath the western United States shut off magmatism over a broad area and caused widespread folding, thrust faulting, and crustal thickening during the Sevier and Laramide orogenies. Between 45 Ma and about 22 Ma (Fig. 11b), the shallow oceanic slab dropped away from the lithosphere (slab rollback). The immersion of the slab into the hotter asthenosphere caused it to dehydrate over a broad area and induced the formation of magmas with subduction zone signatures far from the coastal trench. The rollback of the slab appears to have started beneath Montana and progressed southward as an east-trending bend or as a series of tears in the plate propagated in that direction. This is evidenced by the southward moving front of calc-alkaline magmatism that swept across the western United States during this time period. These magmas are calc-alkaline andesites to rhyolites with strong subduction-related geochemical signatures. Composite volcanoes and caldera complexes were the principal volcanic features. After about 22 Ma (Fig. 11c), counterflow of hot asthenosphere to replace the cold lithosphere appears to have dominated the tectonism in the region. It was accompanied by extension and lithospheric thinning below what is now the Basin and Range province. Extension and decompression were accompanied by enhanced heat flow from the counterflow of the asthenosphere. As a result, partial melting created basaltic magmas that erupted or were lodged in the base of the crust. Rhyolites, many of which bear topaz, formed and erupted concurrently. Slab rollback followed almost immediately by extension may help explain the conflicting evidence for nearly simultaneous formation of extension- and subduction-related igneous rocks.

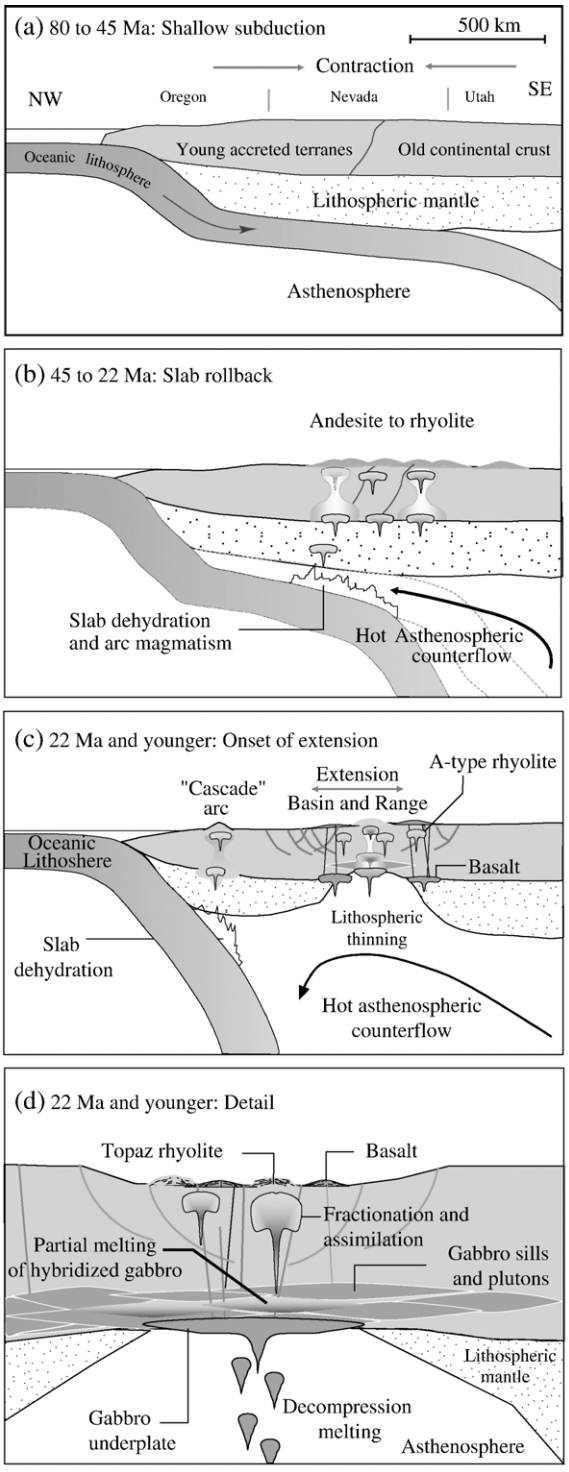


Fig. 11. Cenozoic tectonic evolution of the western United States is shown in these schematic cross sections. (a) Before 45 Ma, shallow subduction of oceanic lithosphere beneath the western United States shut off magmatism over a broad area and caused contractional deformation. (b) Between 45 and about 22 Ma, the shallow slab dropped away from the lithosphere creating a southward moving front of calc-alkaline magmatism. (c) After about 22 Ma, counterflow of hot asthenosphere and lithospheric extension caused the lithosphere to thin. Decompression melting and enhanced heat flow created within-plate basaltic magmas that erupted or intruded the crust becoming hybridized in the process. (d) Partial melting of the hybridized lower crust created rhyolitic magma which differentiated and assimilated crust to become topaz rhyolites.

Recent studies of the tectonic setting of the Fennoscandian rapakivi granites show several similarities to that of the Cenozoic rhyolites of the western United States. For example, there is usually no evidence of concurrent compressional orogenic movement in the fabric of Finnish rapakivi granites (Rämö and Haapala, 1995). Instead, they are contemporaneous with silicic and mafic dikes suggesting concurrent extension. Geophysical studies show that the crust, and especially the lower crust, is markedly thinner beneath the rapakivi granites. Proterozoic graben structures and faults, suggesting intracontinental rifting, have been identified by seismic reflection data (Korja et al., 1993), and there is new evidence that after emplacement of the rapakivi granites and related rocks, thick fluvial sediments started to fill developing rift zones (Kohonen and Rämö, 2005). Haapala et al. (2005) concluded that the mafic and silicic magmas were produced during incipient rifting of the Proterozoic continental crust. Apparently, extension was important in the setting in which rapakivi granites formed and local upwelling of the mantle may also have been important. There is no evidence for contemporaneous subduction-related orogenesis in Poland or the Baltic states to the south of the rapakivi batholiths. However, in southwestern Sweden and southern Norway, 500 to 1500 km to the southwest of the rapakivi batholiths, there are 1.69 to 1.50 Ga calc-alkaline plutonic and volcanic rocks that represent a subduction-related magmatic arc at the margin of the Fennoscandian shield (Åhäll et al., 2000; Andersen et al., 2004). Subduction-related mantle flow at the margin of the continent may have contributed to the mantle upwelling in the inner parts of the thickened continent (e.g., Hoffman, 1989; Åhäll et al., 2000; Haapala et al., 2005). Thus, a taphrogenic origin far behind a contemporaneous magmatic arc seems likely for the Svecofennian rapakivi granites.

10.5. Volumes of magma and the role of fractional crystallization

Topaz-bearing phases of rapakivi batholiths are usually late and small (at most a few kilometers across), but most if not all, are parts of larger complexes from which the topaz granites probably evolved by fractional crystallization (e.g., Haapala, 1997). For topaz rhyolites, estimates of eruption volumes range from a few tenths of a cubic kilometer for small domes less than 1 km across to as much as 50 km^3 for the composite lava fields in the Thomas Range of Utah (Christiansen et al., 1986) and the Black Range of New Mexico (Duffield and Dalrymple, 1990). The large composite flow fields are about 20 km across. Distinctive fractional crystallization trends have been identified in many of these composite flow fields. Minor contamination by wall rocks has also been identified in the Black Range (Reece et al., 1990).

Clearly there are great difficulties in comparing the volumes of intrusive and extrusive rocks. However, some important relationships are apparent. Finnish rapakivi granites are exposed in four large and eleven smaller composite magmatic complexes. Outcrop areas of the composite batholiths range from almost 20,000 km^2 (the Wiborg massif) to less than 10 km^2 for the smallest complexes. The topaz-bearing intrusions (Eurajoki, Kymi, Ahvenisto, Suomenniemi, Artjärvi and Sääskärvi) are at most a few kilometers across. The small sizes of these intrusions compared with the much larger hornblende–biotite– and biotite–granite intrusions are consistent with the geochemical evidence for extensive fractional crystallization of silicic magma to form highly F-rich residual magmas. On the other hand, the topaz granites have more variable εNd values than surrounding "normal" rapakivi granites (Rämö, 1991), which points to possible differences in their sources. In any case, the close association of topaz granites and topaz rhyolites in the apical parts of the epizonal rapakivi complexes in Finland may imply that the topaz rhyolites of western United States are underlain by much larger composite granitic batholiths that have not been exposed by uplift and erosion.

10.6. Halogen concentrations

Both fluorine and chlorine are incompatible elements; Cl is probably more incompatible than fluorine. Thus, because A-type silicic magmas are enriched in all incompatible elements, it is expected that halogen concentrations also will be high. Generally, this is the case with F; concentrations in glassy topaz rhyolites are 3 to 15 times higher than in average glassy calc-alkaline rhyolites (e.g., Macdonald et al., 1992). However, F is much more enriched than Cl imparting the high F/Cl ratios found in these rock suites (Fig. 5). The origin of the high F/Cl ratios is still problematic. Hypotheses include: 1) high F/Cl ratios are related to a granulitic crustal source that had high F/Cl ratios in residual biotite that broke down during fluid-absent melting (Collins et al., 1982; Christiansen et al., 1983), 2) high F/Cl ratios are not primary characteristics of the magmas, but instead result from separation of a Cl-rich magmatic fluid, perhaps even during eruption (Christiansen et al., 1986; Webster, 1992), 3) high halogen concentrations and high F/Cl ratios are the result of fractional crystallization of

mantle-derived magmas that were unrelated to subduction and had high F/Cl ratios, or 4) high F/Cl ratios are caused by partial melting of underplated mafic rocks recently derived from the mantle.

As noted by many others, granulites have small amounts of hydrous phases–either amphibole or biotite–that are enriched in F compared to Cl and water because of the higher thermal stability of F-rich end members. Decomposition of the small amount of hydrous minerals could produce a small amount of F-rich, high F/Cl ratio silicic melt. However, the isotopic and trace element data do not support the generation of topaz rhyolites solely from typical crustal materials, including felsic granulites derived by metamorphism of calc-alkaline igneous rocks with or without a sedimentary component. Thus, although attractive at first, we have rejected this hypothesis as untenable.

Studies of melt inclusions in topaz rhyolites are in their infancy, but thus far show that melt inclusions in quartz have high F/Cl ratios similar to the obsidians and vitrophyres (unpublished studies by X. Zhang and E.H. Christiansen). This suggests that the high F/Cl ratios are not simply the result of devolatilization during eruption and preferential loss of Cl into the fluid or vapor. High F/Cl ratios have been found in studies of "melt" inclusions in the topaz granites of Finland as well (Haapala and Thomas, 2000).

Few potentially parental mafic magmas have the requisite high F/Cl ratios to be direct parents of aluminous A-type granites (Fig. 5). For example, arc magmas have characteristically low F/Cl ratios (0.2 to 1; e.g., Bacon et al., 1992; Christiansen and Keith, 1996) as a result of the enrichment of Cl introduced into the mantle wedge by dehydration of the subducting slab of oceanic lithosphere. The much higher solubility of Cl than F in hydrothermal fluids (Webster, 1992) enriches the fluid and then the resulting partial melt in Cl. Closed system differentiation of these magmas produces silicic magmas and eventually ore deposits with low F/Cl ratios unlike topaz rhyolites.

Glassy rinds and melt inclusions of midocean ridge basalts do have high F/Cl ratios (Byers et al., 1986), but it is very unlikely that any of the other elemental or isotopic characteristics of A-type magmas could be the result of differentiation of MOR basalt. Glassy, plume-related rocks have moderate F/Cl ratios (≤ 2) and their differentiation produces magmas with similar ratios, but high halogen concentrations. These are the characteristics of peralkaline A-type magmas which are commonly thought to be derived by fractionation of mildly alkalic mafic magmas (Mahood and Baker, 1986; Scaillet and Macdonald, 2001). On the other hand, crystallization of mafic magma trapped in the lower crust would produce a solidified rock with relatively high F/Cl ratios because F can be incorporated in some silicates and phosphates in preference to Cl which is probably lost in a hydrothermal fluid. Subsequent partial melting of this mafic plutonic rock would yield a magma with a high F/Cl ratio but low overall halogen concentrations.

We prefer the last explanation for the high F/Cl ratios by a process of elimination of other hypotheses and because it agrees with the isotopic evidence for a significant "mantle" component in the topaz rhyolites. This proposal does not place much of a constraint on the ultimate origin of the source rock, however, because even calc-alkaline subduction zone *magmas* (which have low F/Cl ratios) crystallize to form *rocks* with high F/Cl ratios. For example Christiansen and Lee (1986) concluded that the average F/Cl ratio of granitic rocks from the Basin and Range province is 3.3, but F and Cl concentrations are low (e.g., 0.043 and 0.013 wt.% respectively).

10.7. Sources and partial melting history

Several sources for the generation of A-type granites have been proposed, including: 1. anatexis of felsic calc-alkaline (I-type) crust, 2. partial melting of felsic granulites depleted in incompatible trace elements by earlier high-grade metamorphism or melt extraction, 3. partial melting of distinctive ancient mafic lower crust, 4. fractional crystallization of mantle-derived magma (with or without assimilation of felsic crust), and 5. partial melting of juvenile mafic or hybridized intermediate crust created by intrusion of basaltic magma. We consider each below.

10.7.1. Anatexis of felsic calc-alkaline crust

Patino-Douce (1996) concluded, on the basis of partial melting experiments with a granodiorite and a tonalite from the Sierra Nevada batholith, that A-type granites could form by low pressure ($\sim$4 kb) partial melting of normal (= felsic calc-alkaline) continental crust. He based this conclusion on a comparison of the major element characteristics of A-type granites with the partial melting experiments. Important discriminating factors were modest K+Na/Al, high Fe/Mg, and high Ti/Mg ratios in the low pressure experiments. However, the high Fe/Mg and Ti/Mg ratios in the experimental liquids were the result of partial melting at a very low fO_2 imparted by using a graphite-based cell assembly. Estimated fO_2 of the experiments was 1 log unit below the QFM oxygen buffer. Apparently, the low oxygen fugacity and low pressure restricted the crystallization of

titaniferous magnetite (ilmenite and rutile are the only Fe–Ti oxides noted) and elevated Fe and Ti in the melt. Abundant Fe^{2+} made the residual pyroxenes Fe-rich. In contrast, normal calc-alkaline granodiorites and tonalites are strongly oxidized having initially crystallized at oxygen fugacities as much as 10^4 to 10^5 times higher than the experimentally imposed buffer. Moreover, as argued by Carmichael (1991), the oxygen fugacity of a magma or its source is particularly resistant to change. Thus, it is unlikely that partial melting of normal crustal granodiorites would occur at such a low fO_2. Skjerlie and Johnston (1992) also were able to produce F-rich granitic melts with high Fe/Mg ratios from partial melting of tonalite. However, their experiments were also done at low oxygen fugacities near the QFM buffer.

A related problem with melting calc-alkaline crust to get A-type magmas, including topaz rhyolites and rapakivi granites, involves the generation of the low fO_2 found in the reduced varieties (e.g., Frost and Frost, 1997). Unless it includes large volumes of reduced sedimentary carbon, continental crust composed of normal calc-alkaline igneous rocks is oxidized (2 or more log units above the QFM oxygen buffer, e.g., Carmichael, 1991) because of its generation in subduction zones. Partial melting of such oxidized crust also gives oxidized magmas, not the reduced tholeiitic (or ferroan types) so common among A-type magmatic suites.

Moreover, partial melting of a typical calc-alkaline granodiorites or diorites will not produce the trace element characteristics of A-type granites, such as those discussed here. For example, Rb/Nb ratios in calc-alkaline (I-type) granitic rocks (apparent in volcanic arc, subduction zone, and in average continental crust) are rather high and not easily changed by partial melting since both Rb and Nb are incompatible elements. Negative Nb anomalies such as those found in typical continental crust are hard to erase by partial melting processes (Christiansen and Keith, 1996). In fact, Nb is probably slightly more compatible than K, Rb, U, and Th during dehydration melting involving biotite and Rb/Nb ratios may increase during crustal melting (e.g., Frindt et al., 2004a,b).

Nd and Sr isotopic compositions of topaz rhyolites and A-type granites in general are difficult to explain by partially melting calc-alkaline igneous rocks in the crust. Typical Rb/Sr and Sm/Nd ratios and the Precambrian age of many of the country rocks and underlying lithosphere create higher $^{87}Sr/^{86}Sr$ and lower $^{143}Nd/^{144}Nd$ ratios than are found in these rocks. As noted above, even young middle Cenozoic calc-alkaline dacites, rhyolites, and granites have higher $^{87}Sr/^{86}Sr$ and lower $^{143}Nd/^{144}Nd$ than found in the topaz rhyolites.

Another important problem with any model involving melting of felsic crust is the common association of A-type granites with peralkaline magmas. Peralkaline silicic rocks are not only the same age as some topaz rhyolites and rapakivi granites, but they also occur within the same magmatic complexes. Small volumes of topaz-bearing lava erupted in the middle of the Kane Springs Wash caldera, Nevada (Novak, 1984), and peralkaline syenite dikes are included in the Suomenniemi complex of southern Finland (Rämö, 1991). Generating peralkaline magmas by melting metaluminous source rocks typical of the continental crust would require unusual circumstances.

10.7.2. Partial melting of felsic granulites

A lower crustal felsic granulite is a more likely magma source because depletion of incompatible and/or soluble lithophile elements (like Rb) could erase or decrease the Rb/Nb ratio, produce low Rb/Sr ratios, and water-poor, but F-rich rocks. Inherently small degrees of melting could then give rise to incompatible element-rich rocks. Thus, the derived magma would have a small Nb anomaly and a low Sr isotope ratio—important characteristics of topaz rhyolites and rapakivi granites (e.g.,Christiansen et al., 1988). However, felsic granulites have Sm/Nd ratios that are similar to other crustal rocks, and thus develop low εNd values over time (Rudnick and Fountain, 1995; Condie et al., 1999). As noted above, the relatively high Nd isotope ratios of Cenozoic topaz rhyolites are inconsistent with derivation solely from ancient (i.e., Proterozoic) felsic crust.

Contrasting opinions prevail for some Precambrian A-type granites. For example, DePaolo (1981) and Bennett and DePaolo (1987) used Nd isotopic evidence to conclude that the Proterozoic anorogenic granites of the western United States were derived by melting of pre-existing crust. However, as pointed out by Johnson (1993), basaltic rocks derived from the lithospheric mantle can have εNd values as low as −11‰. If such mafic rocks are part of the source of the Proterozoic anorogenic granites of the southwestern United States, then the "mantle" component could be as high as 60%.

10.7.3. Partial melting of mafic lower crust

Topaz rhyolites of the western United States could be the result of partial melting of a distinctive lower crustal reservoir of Proterozoic age not typically "sampled" by rising magmas in orogenic settings. To explain the relatively high εNd values of topaz rhyolites, this lower crustal source would need to have a high Sm/Nd ratio similar to some of the mafic xenoliths transported to the surface of the Colorado Plateau and in the transition

zone with the Basin and Range province (Esperança et al., 1988; Wendlandt et al., 1993; Chen and Arculus, 1995). For example, Condie and Selverstone (1999) speculate that the lower crust of the Colorado Plateau is composed mostly of amphibolite and mafic granulite with an additional 25% tonalite or diorite. These mafic xenoliths have an average εNd of −3.5 and range from 7 to −13 (Esperança et al., 1988; Wendlandt et al., 1993), in spite of their Proterozoic ages (Fig. 8). Important questions that need to be answered about this potential source include: Would silicic partial melts have the requisite trace element characteristics—especially high incompatible element concentrations, low Rb/Nb ratios, and high F? All of the mafic xenoliths from the Colorado Plateau analyzed by Mattie et al. (1997) have negative Nb anomalies and high Rb/Nb ratios. Could mafic lower crust melt to produce low fO_2 magmas with high Fe/Mg and F/Cl ratios? Most of the xenoliths examined so far have low Fe/Mg ratios like other magnesian (or calc-alkaline in the classification of Miyashiro, 1974) rocks and would probably melt to produce magmas with relatively high fO_2. These questions can only be answered by further mineralogical investigations of mafic lower crustal xenoliths from the region.

Predominantly mafic lower crustal xenoliths with ages close to the Archean–Proterozoic boundary have been found in kimberlites in eastern part of the Fennoscandian shield. Mafic xenoliths from southern Kola peninsula show episodes of magmatic growth (2.5–2.4 Ga) and reworking (1.7 Ga) and have widely varying Nd isotopic compositions (εNd values at 1.54 Ga commonly between −5 and −9), overlapping the compositions of both the rapakivi granites and the associated gabbroic rocks of the Salmi batholith (Neymark et al., 1994; Kempton et al., 2001). Mafic lower crustal xenoliths from the Kuopio area in eastern Finland imply episodes of major growth (2.7 Ga) and reworking (including K-metasomatism at 1.8 Ga) and have εNd (at 1.64 Ga) values (−3.5, −2.5, and +2.8) grossly matching the Wiborg rapakivi granites (see Hölttä et al., 2000; Peltonen and Mänttäri, 2001). However, partial melting models of hornblende-rich xenoliths and plagioclase–clinopyroxene-rich garnet-bearing xenoliths from the Kuopio area (Elliott, 2003) do not support the interpretation that rapakivi granites are derived by simple partial melting of these mafic rocks.

10.7.4. Fractional crystallization of mantle-derived magma

Topaz rhyolites and rapakivi granites could be the result of fractional crystallization of basaltic magma formed during continental rifting. (Elsewhere passage over a mantle plume would produce the same type of magmatic evolution.) This could explain their high εNd values and strong enrichment in incompatible elements. For example, McCurry et al. (in press) concluded that small volumes of extremely evolved A-type rhyolites on the Snake River Plain were produced by fractional crystallization of basaltic parents through ferrolatitic compositions. The principal problem with this hypothesis is the lack of intermediate composition magmas in topaz rhyolite associations and in many, but not all, rapakivi granite complexes. Perhaps the dense, Fe-rich intermediate composition magmas are too dense to be shallowly emplaced and are trapped in the middle crust (e.g., Christiansen and McCurry, in press).

10.7.5. Partial melting of juvenile or hybridized mafic crust

Finally, topaz rhyolites (and by extension rapakivi granites) could be partial melts of mafic intrusive systems within the crust (Fig. 11d). Coeval mafic magma may have lodged in the crust as dikes and sills and become variably hybridized by interaction with older, more felsic crustal rocks. Re-melting could have been caused by subsequent intrusions of hot mafic magma and heat from the rising asthenosphere (e.g., Frost and Frost, 1997; Streck, 2002; Christiansen and McCurry, in press). Strong fractional crystallization of this partial melt–accompanied by more assimilation of older continental crust–could then produce the highly evolved topaz rhyolites and granites. Many late Cenozoic basalts from the western United States lack negative Nb anomalies unless contaminated by continental crust (Lum et al., 1988; Moyer and Esperança, 1988; Barr, 1993; Smith et al., 1999). Such a young gabbroic source would explain the high εNd, the overlap of the Pb isotopic composition of the mafic and silicic rocks, the typically high Fe/Mg and F/Cl ratios, the low fO_2, low Rb/Nb ratios and lack of large negative Nb anomalies, the association with mafic magma in bimodal volcanic fields, and the association with peralkaline magmas, which may be derived by fractionation of the mafic end member or partial melting of alkali basalt lodged in the lower crust.

The most significant problem with this hypothesis may be the large volumes of some of the composite rapakivi batholiths. Partial melting experiments (Helz, 1976; Spulber and Rutherford, 1983) and MELTS models (Ghiorso and Sack, 1995) show that 10 to 30% melting of gabbro or ferrodiorite can yield felsic magma with about 65% to 73% silica, which could then differentiate to high-silica rhyolite. However, if the rapakivi batholiths are 5 to 10 km thick, as noted above, and if they were produced by 10% partial melting, then

melt must have been extracted from a melting interval in the crust that is 50 to 100 km thick. Greater degrees of partial melting would, of course, lower the thickness of the melting interval, as would focusing of melt formed over a larger area into a small region. Additionally, greater volumes of silicic magma could have been produced if the intrusive complex was intermediate in composition as the result of hybridization with felsic crust.

11. Conclusions

The Cenozoic topaz rhyolites from the western United States and rapakivi granites from southern Finland are very similar. The rhyolites are especially similar to the late, highly differentiated, topaz-bearing phases of the intrusive complexes. Similarities include elemental and isotopic compositions, mineral assemblages, mineral compositions, volatile fugacities, metallogenic associations, and inferred tectonic settings. We conclude that they must have formed and differentiated by similar processes, even though separated by thousands of kilometers and billions of years. Consequently, the rhyolites may shed light on the origin of rapakivi granites and A-type granites in general. Likewise, the granitic rocks should tell us about the nature of batholiths related to topaz rhyolites.

We find that modern topaz rhyolites and A-type rhyolites are common in extensional (taphrogenic) settings. Young A-type rhyolites and granites are also found above mantle plumes. Together, we consider these global tectonic environments to be anorogenic in that they contrast with the orogenic settings where most granitic magmas are formed. The link between tectonic setting and A-type magma characteristics seems to lie with the mantle-derived magmas found at rifts and plumes, principally tholeiitic to mildly alkaline basalt with high melting temperatures, high Fe/Mg, low fO_2, low fH_2O, high F/Cl, radiogenic Nd isotopic compositions, unradiogenic Sr isotopic compositions, and small or absent Nb–Ta–Ti–Pb anomalies. Such mafic magmas are not common in orogenic settings where magnesian, calc-alkalic magmas with low Fe/Mg, high fO_2, high fH_2O, low F/Cl, low Sm/Nd, low concentrations of incompatible elements, and pronounced Nb–Ta–Ti–Pb anomalies are the rule (essentially calc-alkaline I-type granitoids). Most felsic continental crust is formed in such orogenic settings and inherits many of these characteristics and over time develops very low εNd values and high $^{87}Sr/^{86}Sr$ ratios. All of these characteristics are different from those of topaz rhyolites and rapakivi granites. Consequently, the most controversial conclusion of this paper is that felsic continental crust with its orogenic geochemistry is in general not the dominant source of topaz rhyolites or rapakivi granites.

Instead, we maintain that the sources of these distinctive A-type silicic magmas must include a significant mantle-derived component of within-plate character. Some A-type rhyolites could be formed by extreme fractional crystallization of mantle-derived basaltic magma. If so, a crustal density filter may serve to suppress the eruption or shallow emplacement of dense, Fe-rich intermediate composition magmas creating the characteristic bimodal associations. In most cases, however, this mantle component probably comes from partial melting of coeval more or less hybridized, gabbro–diorite intrusive complexes to produce rhyolitic (rapakivi granite) magma that fractionates to highly evolved topaz rhyolite (topaz granite). The isotopic, chemical, and bimodal character of many examples of this type of magmatism would thus be explained. In either case, subsequent assimilation of middle or upper crustal rocks may further mask the mantle signature creating magmas with intermediate isotopic compositions (in Sr, Nd, and O) and the small negative Nb anomalies.

On the other hand, many investigators of aluminous A-type magmas conclude that they are solely derived by partial melting of older continental crust with the heat derived from underplating of mafic mantle-derived magma (e.g., Creaser et al., 1991; Haapala and Rämö, 1992; Patino-Douce, 1996; Frindt et al., 2004a; Anderson and Morrison, 2005).

Important tests of these contrasting conclusions will be centered on comparisons of A-type magmas through geologic time. Future studies should focus on isotopic, trace element, and mineralogic studies of the composition of potential crust and mantle sources, including studies of basement outcrops and deep crustal and mantle xenoliths. More isotopic systems must be brought to bear on the problem, including S, O, and Hf. For example, Goodge and Vervoort (2006) have measured Hf isotopic compositions of zircons from many Proterozoic A-type granites from Laurentia and found that they are indistinguishable from the crust they intrude. Such Hf isotopic studies of young rhyolites and granites are sorely needed, especially where there is a large difference between the age of the basement and the age of the anorogenic magmatism. Another fruitful avenue of research lies in better estimates of the volatile fugacities and oxidation states of A-type magmas: do they commonly crystallize at low fO_2 consistent with involvement of a reduced mantle component (e.g., Frost and Frost, 1997) or is there a wide variation in fO_2 implied by the presence of magnetite-series A-type

granites and titanite (e.g., Anderson and Morrison, 2005; Dall'Agnol et al., 2005; Bogaerts et al., 2006)? Are fHF/fHCl ratios characteristically high in the parent magmas, and hence a clue to magma sources, or are the high ratios a result of late degassing?

In short, many questions still remain about the ultimate sources of A-type granites. These questions can most fruitfully be addressed by comparative studies of rhyolites and granites from multiple complexes throughout Earth's long history.

Acknowledgments

This paper is an outgrowth of a presentation made in honor of the retirement of Ilmari Haapala in Helsinki, January 2003. We appreciate the technical assistance of David Tingey at Brigham Young University and extensive collaboration with Jeffrey D. Keith, Timothy Thompson, and Myron G. Best on all aspects of magmatism in the western United States and O. Tapani Rämö on the rapakivi granites. The comments of David John and O. Tapani Rämö on earlier drafts of the paper were greatly appreciated.

References

Åhäll, K.I., Connelly, J.N., Brewer, T.S., 2000. Episodic rapakivi magmatism due to distal orogenesis? Correlation of 1.69–1.50 Ga orogenic and inboard, "anorogenic" events in the Baltic Shield. Geology 28, 823–826.

Andersen, T., Griffin, W.L., Jackson, S.E., Knudsaen, T.-L., Pearson, N.J., 2004. Mid-Proterozoic magmatic arc evolution at the southwest margin of the Baltic Shield. Lithos 73, 289–318.

Anderson, J.L., 1983. Proterozoic anorogenic granite plutonism of North America. Geological Society of America, Memoir 161, 133–154.

Anderson, J.L., Morrison, J., 2005. Ilmenite, magnetite, and peraluminous Mesoproterozoic anorogenic granites of Laurentia and Baltica. Lithos 80, 45–60.

Anderson, J.L., Smith, D.R., 1995. The effects of temperature and fO_2 on the Al-in-hornblende barometer. American Mineralogist 80, 549–559.

Armstrong, R.L., 1978. Cenozoic igneous history of the U.S. Cordillera from latitude 42° N to 49° N. Geological Society of America, Memoir 152, 265–282.

Bacon, C.R., Newman, S., Stolper, E., 1992. Water, CO_2, Cl, and F in melt inclusions in phenocrysts from three Holocene explosive eruptions, Crater Lake, Oregon. American Mineralogist 77, 1021–1030.

Barr, D.L., 1993. Time, space, and composition patterns of middle Cenozoic mafic to intermediate composition lava flows of the Great Basin, western USA. MS thesis, Brigham Young University, Provo, Utah, 68 pp.

Beard, B.L., Johnson, C.M., 1997. Hafnium isotope evidence for the origin of Cenozoic basaltic lavas from the southwestern United States. Journal of Geophysical Research 102, 20,149–20,178.

Bennett, V.C., DePaolo, D.J., 1987. Proterozoic crustal history of the western United States as determined by neodymium isotopic mapping. Geological Society of America Bulletin 99, 674–685.

Best, M.G., Christiansen, E.H., 1991. Limited extension during peak Tertiary volcanism, Great Basin of Nevada and Utah. Journal of Geophysical Research 96, 13509–13528.

Best, M.G., McKee, E.H., Damon, P.E., 1980. Space–time-composition patterns of late Cenozoic mafic volcanism, southwestern Utah and adjoining areas. American Journal of Science 280, 1035–1050.

Best, M.G., Keith, J.D., Mehnert, H.H., 1987. Early Miocene tectonism, magmatism, and mineralization in and near the southern Wah Wah Mountains, southwestern Utah. Oligocene and Miocene Volcanic Rocks in the Central Pioche–Marysvale Igneous Belt, Western Utah and Eastern Nevada. U.S. G.S. Professional Paper, pp. 29–47.

Best, M.G., Christiansen, E.H., Blank, R.H., 1989. Oligocene caldera complex and calc-alkaline tuffs and lavas of the Indian Peak volcanic field, Nevada and Utah. Geological Society of America Bulletin 101, 1076–1090.

Bogaerts, M., Scaillet, B., Vander Auwera, J., 2006. Phase equilibria of the Lyngdal granodiorite (Norway): implications for the origin of metaluminous ferroan granitoids. Journal of Petrology 47, 2405–2431.

Brooks, W.E., Thorman, C.H., Snee, L.W., 1995. The $^{40}Ar/^{39}Ar$ ages and tectonic setting of the middle Eocene northeast Nevada volcanic field. Journal of Geophysical Research 100, 10,403–10,416.

Burt, D.M., Sheridan, M.F., Bikun, J.V., Christiansen, E.H., 1982. Topaz rhyolites—distribution origin and significance for exploration. Economic Geology 77, 1818–1836.

Byers, C.E., Garcia, M.D., Muenow, D.W., 1986. Volatiles in basaltic glasses from the East Pacific Rise at 21 degrees N; implications for MORB sources and submarine lava flow morphology. Earth and Planetary Science Letters 79, 9–20.

Byrd, B.B., Nash, W.P., 1993. Eruption of rhyolite at the Honeycomb Hills, Utah; cyclical tapping of a zoned silicic magma reservoir. Journal of Geophysical Research 98, 14,075–14,090.

Calzia, J.P., Rämö, O.T., 2005. Miocene rapakivi granites in the southern Death Valley region, California, USA. Earth Science Reviews 73, 221–243.

Carmichael, I.S.E., 1991. The redox states of basic and silicic magmas; a reflection of their source regions? Contributions to Mineralogy and Petrology 106, 129–141.

Castor, S.B., Henry, C.D., 2000. Geology, geochemistry, and origin of volcanic rock-hosted uranium deposits in northwestern Nevada and southeastern Oregon, USA. Ore Geology Reviews 16 (1–2), 1–40.

Chen, W., Arculus, R.J., 1995. Geochemical and isotopic characteristics of lower crustal xenoliths, San Francisco volcanic field, Arizona, U.S.A. Lithos 36, 203–225.

Chadwick, R.A., 1978. Geochronology of post-Eocene rhyolitic and basaltic volcanism in southwestern Montana. Isochron/West 22, 25–28.

Chappell, B.W., Hine, R., 2006. The Cornubian Batholith: an example of magmatic fractionation on a crustal scale. Resource Geology 56, 203–244.

Chappell, B.W., White, A.J.R., 1992. I- and S-type granites in the Lachlan fold belt. Royal Society of Edinburgh Transactions 83, 1–26.

Christiansen, E.H., 1980. Uranium mineralization associated with fluorine-enriched rhyolites in southwestern Utah: U.S. Department of Energy Open-File Report GJBX-225, 80, pp. 415–458.

Christiansen, E.H., Keith, J.D., 1996. Trace element systematics in silicic magmas. A metallogenic perspective. Geological Association of Canada Short Course Notes Volume 12, Trace Element Geochemistry of Volcanic Rocks. 24–26 May 1996, Winnipeg, Manitoba Canada, pp. 115–151.

Christiansen, E.H., Lee, D.E., 1986. Fluorine and chlorine in granitoids from the Basin and Range province, western United States. Economic Geology 81, 1484–1494.

Christiansen, R.L., Lipman, P.W., 1972. Cenozoic volcanism and plate tectonic evolution of the western United States. II. Late Cenozoic. Philosophical Transactions of the Royal Society of London, Series A 271, 249–284.

Christiansen, E.H., McCurry, M., in press. Contrasting origins of Cenozoic silicic volcanic rocks from the western Cordillera of the United States. Bulletin of Volcanology.

Christiansen, E.H., Burt, D.M., Sheridan, M.F., Wilson, R.T., 1983. Petrogenesis of topaz rhyolites from the western United States. Contributions to Mineralogy and Petrology 83, 16–30.

Christiansen, E.H., Bikun, J.V., Sheridan, M.F., Burt, D.M., 1984. Geochemical evolution of topaz rhyolites from the Thomas Range and Spor Mountain, Utah. American Mineralogist 69, 223–236.

Christiansen, E.H., Burt, D.M., Sheridan, M.F., 1986. The geology of topaz rhyolites from the western United States. Geological Society of America Special Paper 205 (82 pp.).

Christiansen, E.H., Stuckless, J.S., Funkhouser, M.J., Howell, K.A., 1988. Petrogenesis of rare-metal granites from depleted crustal sources—an example from the Cenozoic of western Utah, USA. In: Taylor, R.P., Strong, D.F. (Eds.), Recent Advances in the Geology of Granite-Related Mineral Deposits. Canadian Institute of Mining and Metallurgy, vol. 39, pp. 307–321.

Christiansen, E.H., Keith, J.D., Thompson, T.J., 1996. Origin of gem red beryl in the Wah Wah Mountains, Utah. Mining Engineering 49, 37–41.

Clark, E.E., 1977. Late Cenozoic volcanic and tectonic activity along the eastern margin of the Great Basin, in the proximity of Cove Fort, Utah. Brigham Young University Geology Studies 24 (1), 87–114.

Clark, D., 2003, Geologic map of the Sage Valley quadrangle, Juab County, Utah Geological Survey, MP-03-02.

Coleman, D.S., Walker, J.D., 1992. Evidence for the generation of juvenile granitic crust during continental extension, Mineral Mountains Batholith, Utah. Journal of Geophysical Research 97, 11,011–11,024.

Collins, W.J., Beams, S.D., White, A.J.R., Chappell, B.W., 1982. Nature and origin of A-type granites with particular reference to southeastern Australia. Contributions to Mineralogy and Petrology 80, 189–200.

Condie, K.C., Selverstone, J., 1999. The crust of the Colorado Plateau: new views of an old arc. Journal of Geology 107, 387–397.

Condie, K.C., Latysh, N., Van Schmus, W.R., Koxuch, M., Selverstone, J., 1999. Geochemistry, Nd and Sr isotopes, and U/Pb zircon ages of granitoid and metasedimentary xenoliths form the Navajo Volcanic Field, Four Corners area, southwestern United States. Chemical Geology 156, 95–133.

Congdon, R.D., Nash, W.P., 1991. Eruptive pegmatite magma; rhyolite of the Honeycomb Hills, Utah. American Mineralogist 76, 1261–1278.

Creaser, R.A., Price, R.C., Wormald, R.J., 1991. A-type granites revisited: assessment of a residual-source model. Geology 19, 163–166.

Cunningham, C.G., Unruh, D.M., Steven, T.A., Rowley, P.D., Naeser, C.W., Mehnert, H.H., Hedge, C.E., Ludwig, K.R., 1998. Geochemistry of volcanic rocks in the Marysvale volcanic field, west-central Utah. U.S. Geological Survey Bulletin B-2158, 223–231.

Dall'Agnol, R., Teixeira, N.P, Rämö, O.T., Moura, C.A.V., Macambira, M.J.B., de Oliveira, D.C., 2005. Petrogenesis of the Paleoproterozoic rapakivi A-type granites of the Archean Carajás Metallogenic Province, Brazil. Lithos 80, 101–129.

DePaolo, D.J., 1981. Neodymium isotopes in the Colorado Front Range and crust–mantle evolution in the Proterozoic. Nature 291, 193–196.

DePaolo, D.J., Daley, E., 2000. Neodymium isotopes in basalts of the southwest Basin and Range and lithospheric thinning during continental extension. Chemical Geology 169, 157–185.

Dickinson, W.R., 2002. The Basin and Range province as a composite extensional domain. International Geology Review 44, 1–38.

Duffield, W.A., Dalrymple, G.B., 1990. The Taylor Creek Rhyolite of New Mexico; a rapidly emplaced field of lava domes and flows. Bulletin of Volcanology 52, 475–487.

Eaton, G.P., 1979. A plate tectonic model for late Cenozoic crustal spreading in the western United States. In: Riecker, R.E. (Ed.), Rio Grande Rift. Tectonics and Magmatism. American Geophysical Union, Washington, D.C., pp. 7–32.

Eby, G.N., 1992. Chemical subdivision of the A-type granitoids; petrogenetic and tectonic implications. Geology 20, 641–644.

Edén, P., 1991. A specialized topaz-bearing rapakivi granite and associated mineralized greisen in the Ahvenisto complex, SE Finland. Bulletin Geological Society of Finland 63, 25–40.

Ekren, E.B., Byers, F.M., Hardyman, R.F., Marvin, R.F., Silberman, M.L., 1980. Stratigraphy, preliminary petrology, and some structural features of Tertiary volcanic rocks in the Gabbs Valley and Gillis ranges, Mineral County, Nevada. U.S. Geological Survey Bulletin B 1464 (54 pp.).

Eklund, O., Shebanov, A.D., 1999. The origin of rapakivi texture by sub-isothermal decompression. Precambrian Research 95, 129–146.

Elliott, B.A., 2001. Crystallization conditions of the Wiborg rapakivi batholith, SE Finland: an evaluation of amphibole and biotite mineral chemistry. Mineralogy and Petrology 72, 305–324.

Elliott, B.A., 2003. Petrogenesis of the postkinematic magmatism of the central Finland Granitoid Complex II; sources and magmatic evolution. Journal of Petrology 44, 1681–1701.

Esperança, S., Carlson, R.W., Shirey, S.B., 1988. Lower crustal evolution under central Arizona. Sr, Nd, and Pb isotopic and geochemical evidence form the mafic xenoliths of Camp Creek. Earth and Planetary Science Letters 90, 26–40.

Ewart, A., Milner, S.C., Armstrong, A.R., 1998a. Etendeka volcanism of the Goboboseb Mountains and Messum Igneous Complex, Namibia: part I. Geochemical evidence of early Cretaceous Tristian plume melts and the role of crustal contamination. Journal of Petrology 39, 191–225.

Ewart, A., Milner, S.C., Armstrong, R.A., Duncan, A.R., 1998b. Etendeka volcanism and of the Goboboseb Mountains and Messum Igneous Complex, Namibia: part II. Voluminous quartz latite volcanism of the Awahab magma system. Journal of Petrology 39, 227–253.

Fitton, J.G., James, D., Kempton, P.D., Ormerod, D.S., Leeman, W.P., 1988. The role of lithospheric mantle in the generation of late Cenozoic basic magmas in the western United States. Journal of Petrology 1988, 331–349.

Fleck, R.J., Turrin, B.D., Sawyer, D.A., Warren, R.G., Champion, D.E., Hudson, M.R., Minor, S.A., 1996. Age and character of basaltic rocks of the Yucca Mountain region, southern Nevada. Journal of Geophysical Research 101, 8,205–8,227.

Frindt, S., Trumbull, R.B., Romer, R.L., 2004a. Petrogenesis of the Gross Spitzkoppe topaz granite, central western Namibia: a geochemical and Nd–Sr–Pb isotope study. Chemical Geology 206, 43–71.

Frindt, S., Haapala, I., Pakkanen, L., 2004b. Anorogenic Gross Spitzkoppe granite stock in central western Namibia: part I. Petrology and geochemistry. American Mineralogist 89, 841–856.

Frost, C.D., Frost, B.R., 1997. Reduced rapakivi-type granites: the tholeiite connection. Geology 25, 647–650.

Frost, C.D., Frost, B.R., Chamberlain, K.R., Edwards, B.R., 1999. Petrogenesis of the 1.43 Ga Sherman batholith, SE Wyoming, USA: a reduced, rapakivi-type anorogenic granite. Journal of Petrology 40, 1771–1802.

Frost, B.R., Barnes, C.G., Collins, W.J., Arculus, R.J., Ellis, D.J., Frost, C.D., 2001a. A geochemical classification for granitic rocks. Journal of Petrology 42, 2033–2048.

Frost, C.D., Bell, J.M., Frost, B.R., Chamberlain, K.R., 2001b. Crustal growth by magmatic underplating: isotopic evidence from the northern Sherman batholith. Geology 29, 515–518.

Funkhouser-Marolf, M.J., 1985. The mineralogy and distribution of uranium and thorium in the Sheeprock granite, Utah. M.S. thesis, University of Iowa, Iowa City, 60 p.

Ghiorso, M.S., Sack, R.O., 1995. Chemical mass transfer in magmatic processes. IV. A revised and internally consistent thermodynamic model for the interpolation and extrapolation of liquid–solid equilibria in magmatic systems at elevated temperatures and pressures. Contributions to Mineralogy and Petrology 119, 197–212.

Goodge, J., Vervoort, J., 2006. Origin of Mesoproterozoic A-type granites in Laurentia: Hf isotope evidence. Earth Planetary Science Letters 243, 711–731.

Haapala, I., 1977. Petrography and geochemistry of the Eurajoki stock; a rapakivi–granite complex with greisen-type mineralization in southwestern Finland. Bulletin of the Geological Society of Finland 286 (128 pp.).

Haapala, I., 1995. Metallogeny of the rapakivi granites. Mineralogy and Petrology 54, 149–160.

Haapala, I., 1997. Magmatic and postmagmatic processes in tin-mineralized granites: topaz-bearing leucogranites in the Eurajoki rapakivi granite stock, Finland. Journal of Petrology 38, 1645–1659.

Haapala, I., Lukkari, S., 2005. Petrological and geochemical evolution of the Kymi stock, a topaz granite cupola within the Wiborg rapakivi batholith, Finland. Lithos 80, 347–362.

Haapala, I., Rämö, O.T., 1992. Tectonic setting and origin of the Proterozoic rapakivi granites of southeastern Fennoscandia. Transactions of the Royal Society of Edinburgh. Earth Science 83, 165–171.

Haapala, I., Thomas, R., 2000. Melt inclusions in quartz and topaz of the topaz granite from Eurajoki, Finland. Journal Czech Geological Society 45, 149–154.

Haapala, I., Rämö, O.T., Frindt, S., 2005. Comparison of Proterozoic and Phanerozoic rift-related basaltic–granitic magmatism. Lithos 80, 1–32.

Helz, R.T., 1976. Phase relations of basalts in their melting ranges at $P(H_2O) = 5$ kb; part II, melt compositions. Journal of Petrology 17, 139–193.

Hoffman, P.J., 1989. Speculations on Laurentia's first giga-year (2.0–1.0 Ga). Geology 17, 135–138.

Hölttä, P., Huhma, H., Mänttäri, I., Peltonen, P., Juhanoja, J., 2000. Petrology and geochemistry of mafic granulite xenoliths from the Lahtojoki kimberlite pipe, eastern Finland. Lithos 51, 109–131.

Huspeni, J.R., Kesler, S.E., Ruiz, J., Tuta, Z., Sutter, J.F., Jones, L.M., 1984. Petrology and geochemistry of rhyolites associated with tin mineralization in northern Mexico. Economic Geology 79, 87–105.

John, D.A., Wallace, A.R., Ponce, D.A., Fleck, R.B., Conrad, J.E., 2000. New perspectives on the geology and origin of the northern Nevada Rift. In: Cluer, J.K., Price, J.G., Struhsacker, E.M., Hardyman, R.F., Morris, C.L. (Eds.), Geology and Ore Deposits 2000: the Great Basin and Beyond: Geological Society of Nevada Symposium Proceedings, May 15–18, pp. 127–154.

John, D.A., 2001. Miocene and early Pliocene epithermal gold–silver deposits in the northern Great Basin, western United States; characteristics, distribution, and relationship to magmatism. Economic Geology 96, 1827–1853.

Johnson, C.M., 1993. Mesozoic and Cenozoic contributions to crustal growth in the southwestern United States. Earth and Planetary Science Letters 118, 75–89.

Johnson, C.M., Thompson, R.A., 1991. Isotopic composition of Oligocene mafic volcanic rocks in the northern Rio Grande rift: evidence for contributions of ancient intraplate and subduction magmatism to evolution of the lithosphere. Journal of Geophysical Research 96, 13,593–13,608.

Keith, J.D., Christiansen, E.H., Tingey, D.G., 1994. Geological and chemical conditions of formation of red beryl, Wah Wah Mountains, Utah. Utah Geological Association Publication 23, 155–169.

Keith, J.D., Shanks, W.C., Archibald, D.A., Farrar, E., 1986. Volcanic and intrusive history of the Pine Grove porphyry molybdenum system, southwestern Utah. Economic Geology 81, 553–577.

Kempton, P.D., Downes, H., Neymark, L.A., Wartho, J.A., Zartman, R.E., Sharkov, E.V., 2001. Garnet granulite xenoliths from the northern Baltic Shield; the underplated lower crust of a Palaeoproterozoic large igneous province? Journal of Petrology 42, 731–763.

Kirs, J., Haapala, I., Rämö, O.T., 2004. Anorogenic magmatic rocks in the Estonian crystalline basement. Proceedings of the Estonian Academy of Sciences. Geology 53 (3), 210–225.

Kistler, R.W., Peterman, Z.E., 1978. Reconstruction of crustal blocks of California on the basis of initial strontium isotopic compositions of Mesozoic granitic rocks. U.S. Geological Survey Professional Paper 1071 (17 pp.).

Kohonen, J., Rämö, O.T., 2005. Sedimentary rocks, diabases, and late cratonic evolution. In: Lehtinen, M., Nurmi, P.A., Rämö, O.T. (Eds.), Precambrian Geology of Finland — Key to the Evolution of the Fennoscandian Shield. Elsevier B.V., Amsterdam, pp. 563–604.

Korja, A., Korja, T., Luosto, U., Heikkinen, P., 1993. Seismic and geoelectric evidence for collisional and extensional events in the Fennoscandian Shield—implications from Precambrian crustal evolution. Tectonophysics 219, 129–152.

Kortemeier, W.T., Burt, D.M., 1988. Ongonite and topazite dikes in the Flying W Ranch area, Tonto Basin, Arizona. American Mineralogist 73, 507–523.

Kosunen, P., 1999. The rapakivi granite plutons of Bodom and Obbnas, southern Finland; petrography and geochemistry. Bulletin of the Geological Society of Finland 71, 275–304.

Kovalenko, I., Kovalenko, N.I., 1984. Problems of the origin, ore-bearing and evolution of rare-metal granitoids. Physics of the Earth and Planetary Interiors 35, 51–62.

Laurén, L., 1970. An interpretation of the negative gravity anomalies associated with the rapakivi granites and the Jotnian sandstone in southern Finland. Geologiska Föreningens i Stockholm Förhandlingar 92, 21–34.

Leeman, W.P., 1982. Rhyolites of the Snake River Plain–Yellowstone Plateau province, Idaho and Wyoming: a summary of petrogenetic models. In: Bonnichsen, B., Breckenridge, R.M. (Eds.), Cenozoic Geology of Idaho. Idaho Bureau of Mines and Geology Bulletin, vol. 26, pp. 203–212.

Lipman, P.W., Prostka, H.J., Christiansen, R.L., 1972. Cenozoic volcanism and plate-tectonic evolution of the western United States. Part 1, early and middle Cenozoic. Philosophical Transactions of the Royal Society of London 271, 217–248.

London, D., Wolf, M.B., Morgan VI, G.B., Garrido, M.G., 1999. Experimental silicate-phosphate equilibria in peraluminous

granitic magmas, with a case study of Albuquerque Batholith at Tres Arroyos, Badajoz, Spain. Journal of Petrology 40, 215–240.

Lukkari, S., 2002. Petrography and geochemistry of the topaz-bearing granite stocks in Artjärvi and Sääskijärvi, western margin of the Wiborg rapakivi granite batholith. Bulletin of the Geological Society of Finland 74 (1–2), 115–132.

Lum, C.L., Leeman, W.P., Foland, K.A., Kargel, J.A., Fitton, J.G., 1988. Isotopic variations in continental basaltic lavas as indicators of mantle heterogeneity: examples from the western U.S. Cordillera. Journal of Geophysical Research 94, 7871–7884.

Macdonald, R., Smith, R.L., Thomas, J.E., 1992. Chemistry of the subalkalic silicic obsidians. U.S. Geological Survey Professional Paper 1523 (214 pp.).

Mahood, G.A., Baker, D.R., 1986. Experimental constraints on depths of fractionation of mildly alkalic basalts and associated felsic rocks; Pantelleria, Strait of Sicily. Contributions to Mineralogy and Petrology 93, 251–264.

Manning, D.A.C., 1981. The effects of fluorine on liquidus phase relationships in the system Qz–Ab–Or with excess water at 1 kb. Contributions to Mineralogy and Petrology 76, 206–215.

Mattie, P.D., Condie, K.C., Selverstone, J., Kyle, P.R., 1997. Origin of the continental crust in the Colorado Plateau: geochemical evidence from mafic xenoliths from the Navajo Volcanic field, southeastern USA. Geochimica et Cosmochimica Acta 61, 2007–2021.

McCurry, M., Hayden, K., Morse, L., in press. Genesis of post-hotspot A-type rhyolite of the eastern Snake River Plain volcanic province by extreme fractional crystallization of olivine tholeiite basalt. Bulletin of Volcanology.

McDermott, F., Hawkesworth, C.J., 1990. Intracrustal recycling and upper-crustal evolution: a case study from the Pan-African Damara mobile belt, central Namibia. Chemical Geology 83, 263–280.

McDonough, W.F., Sun, S.S., 1995. The composition of the Earth. Chemical Geology 120, 223–253.

Miller, C.F., Wooden, J.L., 1994. Anatexis, hybridization and the modification of ancient crust; Mesozoic plutonism in the Old Woman Mountains area, California. Lithos 32, 111–133.

Miller, R.J., Wrucke, C.T., 1995. Age, chemistry, and geologic implications of Tertiary volcanic rocks in the Last Chance Range and part of the Saline Range, northern Death Valley region, California. Isochron/West 62, 30–36.

Miyashiro, A., 1974. Volcanic rock series in island arcs and active continental margins. American Journal of Science 274, 321–355.

Moore, D.K., 1993. Oligocene East Tintic volcanic field, Utah: geology and petrogenesis. M.S. thesis, Brigham Young University, Provo, Utah, 65 p.

Moyer, T.C., Esperança, S., 1988. Geochemical and isotopic variations in a bimodal magma system. The Kaiser Spring volcanic field, Arizona. Journal of Geophysical Research 94, 7841–7859.

Novak, S.W., 1984. Eruptive history of the rhyolitic Kane Springs Wash volcanic center, Nevada. Journal of Geophysical Research 89, 8603–8615.

Nelson, S.T., Tingey, D.G., 1997. Time-transgressive and extension-related volcanism in southwest Utah and vicinity. Geological Society of America Bulletin 109, 1249–1265.

Neymark, L.A., Yu, Larin, A.M., 1994. Pb–Nd–Sr isotopic constraints on the origin of the 1.54–1.56 Ga Salmi rapakivi granite–anorthosite batholith (Karelia, Russia). Mineralogy and Petrology 50, 173–194.

Oldow, J.S., 1984. Evolution of a late Mesozoic back-arc fold and thrust belt, northwestern Great Basin, U.S.A. Tectonophysics 102, 245–274. doi:10.1016/0040-1951(84)90016-7.

Patino-Douce, A.E., 1996. Generation of metaluminous A-type granites by low-pressure melting of calc-alkaline granitoids. Geology 25, 743–746.

Pearce, J.A., Harris, N.B.W., Tindle, A.G., 1984. Trace element discrimination diagrams for the tectonic interpretation of granitic rocks. Journal of Petrology 25, 956–983.

Peltonen, P., Mänttäri, I., 2001. An ion microprobe U–Th–Pb study of zircon xenocrysts from the Lahtojoki kimberlite pipe, eastern Finland. Bulletin of the Geological Society of Finland 73, 47–58.

Rämö, O.T., 1991. Petrogenesis of the Proterozoic rapakivi granites and related basic rocks of southeastern Fennoscandia: Nd and Pb isotopic and general geochemical constraints. Geological Survey of Finland Bulletin 355 (161 pp.).

Rämö, O.T., Haapala, I., 1995. One hundred years of rapakivi granite. Mineralogy and Petrology 52, 129–185.

Rämö, O.T., Haapala, I., Vaasjoki, M., Yu, J., Fu, H., 1995. 1700 Ma Shachang complex, northern China: Precambrian rapakivi granite not associated with Paleoproterozoic orogenic crust. Geology 23, 815–818.

Rämö, O.T., Korja, A., Haapala, I., Eklund, O., Fröjdö, S., Vaasjoki, M., 2000. Evolution of the Fennoscandian lithosphere in the mid-Proterozoic: the rapakivi magmatism. Institute of Seismology. University of Helsinki, Report S-41, 129–136.

Reece, C., Ruiz, J., Duffield, W.A., Patchett, P.J., 1990. Origin of Taylor Creek rhyolite magma, Black Range, New Mexico, based on Nd–Sr isotope studies. In: Stein, H.J., Hannah, J.L. (Eds.), Ore-Bearing Granite Systems; Petrogenesis and Mineralizing Processes. Geological Society of America Special Paper, vol. 246, pp. 263–273.

Richardson, P.D., 2004. Pluton zonation unveiled by gamma-ray spectrometry and magnetic susceptibility: case study of the Sheeprock Granite, western Utah. M.S. thesis, Brigham Young University, Provo, Utah, 72 p.

Rieder, M., Haapala, I., Povondra, P., 1996. Mineralogy of dark mica from the Wiborg rapakivi batholith, southeastern Finland. European Journal of Mineralogy 8, 593–605.

Rogers, J.R. 1990. Origin of beryl in the Miocene Sheeprock granite, west-central Utah. M.S. Thesis, Brigham Young University, 69 p.

Rogers, N.W., Hawkesworth, C.J., Ormerod, D.S., 1995. Late Cenozoic basaltic magmatism in the western Great Basin, California and Nevada. Journal of Geophysical Research 100, 10,287–10,301.

Rowley, P.D., Anderson, J.J., Williams, P.L., Fleck, R.J., 1978. Age of structural differentiation between the Colorado Plateaus and Basin and Range provinces in southwestern Utah. Geology 6, 51–55.

Rudnick, R.L., Fountain, D.M., 1995. Nature and composition of the continental crust: a lower crustal perspective. Reviews of Geophysics 33, 267–309.

Salonsaari, P., 1995. Hybridization in the subvolcanic Jaala–Iitti complex and its petrogenetic relation to rapakivi granites and associated mafic rocks of southeastern Finland. Bulletin of the Geological Society of Finland 67 (part 1b), 1–104.

Scaillet, B., Macdonald, R., 2001. Phase relations of peralkaline silicic magmas and petrogenetic implications. Journal of Petrology 42, 825–845.

Seth, B., Kröner, A., Mezger, K., Nemchin, A.A., Pidgeon, R.T., Okrusch, M., 1998. Archean to Neoproterozoic magmatic events in the Kaoko Belt of NW Namibia and their geodynamic significance. Precambrian Research 92, 341–363.

Simpson, J.H., 1876. Report of Exploration Across the Great Basin of the Territory of Utah for a Direct Wagon-route from Camp Floyd to Genoa in Carson Valley in 1859. Engineer Department, U.S. Army, Washington, D.C. 518 pp.

Sinclair, W.D., 1986. Early Tertiary topaz rhyolites and associated mineral deposits in the northern Canadian Cordillera; products of anorogenic magmatism. Program with abstracts — Geological Association of Canada. Mineralogical Association of Canada Annual Meeting, vol. 11, pp. 127–128.

Skjerlie, K., Johnston, A.D., 1992. Vapor-absent melting at 10 kbar of a biotite- and amphibole-bearing tonalitic gneiss: implications for the generation of A-type granites. Geology 20, 263–266.

Smith, E.I., Sanchez, A., Walker, J.D., Wang, K., 1999. Geochemistry of mafic magmas in the Hurricane volcanic field, Utah; implications for small- and large-scale chemical variability of the lithospheric mantle. Journal of Geology 107 (4), 433–448.

Smith, R.L., 1979. Ash-flow magmatism. Geological Society of America Special Paper 180, 5–27.

Speed, R.C., 1979. Collided Paleozoic microplate in the western United States. Journal of Geology 87, 279–292.

Spulber, S.D., Rutherford, M.J., 1983. The origin of rhyolite and plagiogranite in oceanic crust; an experimental study. Journal of Petrology 24, 1–25.

Stewart, J.H., Walker, G.W., Kleinhampl, F.J., 1975. Oregon–Nevada lineament. Geology 3, 265–268.

Stockli, D.F., Linn, J.K., Walker, J.D., 2001. Miocene unroofing of the Canyon Range during extension along the Sevier Desert Detachment, west central Utah. Tectonics 20, 289–307.

Streck, M.J., 2002. Partial melting to produce high-silica rhyolites of a young bimodal suite. Compositional constraints among rhyolites, basalts, and metamorphic xenoliths from the Harney Basin, Oregon. International Journal of Earth Sciences 91, 583–593.

Suneson, N.H., Lucchitta, I., 1983. Origin of bimodal volcanism, southern Basin and Range province, west-central Arizona. Geological Society of America Bulletin 94, 1005–1019.

Thompson, T.J., 2002. A model for the origin of red beryl in topaz rhyolite, Wah Wah Mountains, Utah, USA. M.S. thesis, Brigham Young University, Provo, Utah, 99 p.

Trumbull, R.B., Harris, C., Frindt, S., Wigand, M., 2004. Oxygen and neodymium isotope evidence for source diversity in Cretaceous anorogenic granites from Namibia and implications for A-type granite genesis. Lithos 73, 21–40.

Tosdal, R.M., Wooden, J.L., Kistler, R.W., 2000. Geometry of the Neoproterozoic continental breakup, and implications for location of Nevadan mineral belts. In: Cluer, J.K., Price, J.G., Struhsacker, E.M., Hardyman, R.F., Morris, C.L. (Eds.), Geology and Ore Deposits 2000, The Great Basin and Beyond: Geological Society of Nevada Symposium, Reno, 2000, Proceedings, pp. 451–466.

Vitaliano, C.J., Vitaliano, D.B., 1972. Cenozoic volcanic rocks in the southern Shoshone Mountains and Paradise Range, Nevada. Geological Society of America Bulletin 83, 3269–3280.

Webster, J.D., 1992. Water solubility and chlorine partitioning in Cl-rich granitic systems; effects of melt composition at 2 kbar and 800 degrees C. Geochimica et Cosmochimica Acta 56, 679–687.

Wendlandt, E., DePaolo, D.J., Baldridge, W.S., 1993. Nd and Sr isotope chronostratigraphy of Colorado Plateau lithosphere: implications for magmatic and tectonic underplating of the continental crust. Earth and Planetary Science Letters 116, 23–43.

Whalen, J.B., Currie, K.L., Chappell, B.W., 1986. A-type granites; geochemical characteristics, discrimination and petrogenesis. Contributions to Mineralogy and Petrology 95, 407–419.

White, A.J.R., Chappell, B.W., 1983. Granitoid types and their distribution in the Lachlan fold belt, southeastern Australia. Geological Society of America, Memoir 159, 21–33.

Wooden, J.L., Kistler, R.W., Tosdal, R.M., 1998. Pb isotopic mapping of crustal structure in the northern Great Basin and relationships to Au deposit trends. In: Tosdal, R.M. (Ed.), Contributions to the Gold Metallogeny of Northern Nevada. U.S. Geological Survey Open-File Report, pp. 20–33.

Zoback, M.L., Thompson, G.A., 1978. Basin and Range rifting in northern Nevada—clues from a mid-Miocene rift and its subsequent offsets. Geology 6, 111–116.

Zoback, M.L., Anderson, R.E., Thompson, G.A., 1981. Cainozoic evolution of the state of stress and style of tectonism of the Basin and Range province of the western United States. Philosophical Transactions of the Royal Society of London, Series A: Mathematical and Physical Sciences 300, 407–434.