Overview of the Lithophile Element-Bearing Magmatic-Hydrothermal System at Birch Creek, White Mountains, California

MARK D. BARTON

Center for Mineral Resources, Department of Geosciences, University of Arizona, Tucson, Arizona 85721

Abstract

A large lithophile element-bearing hydrothermal system is associated with the well-exposed Birch Creek biotite-muscovite granite and its metamorphic aureole in the White Mountains of eastern California. Elements enriched include F, Be, W, Zn, Pb, Ag, Cu, Au, Bi, and Sn although historic production (of Pb, Ag, Au, W) has been minor and likely resources are small. This system is one of several dozen Late Cretaceous centered hydrothermal systems that are associated with two-mica granites along the Cordilleran miogeocline in the Great Basin. As a group, these occurrences resemble other Mesozoic W-Sn-F-Be bearing systems in circum-Pacific; however, these seemingly lack the economic deposits—an observation that begs the question: Why not? These granites are also of considerable interest because of their structural characteristics, notably intensely deformed and attenuated margins. They have figured prominently in literature on emplacement mechanisms for granites. This paper provides an overview of collaborative studies by several groups looking at the magmatic, structural, metamorphic, and hydrothermal development of the Birch Creek system.

Field, petrological, and geochemical studies demonstrate that the Birch Creek pluton (82 Ma, U-Pb, Ar-Ar) was episodically emplaced with alternating major pulses from at least two, probably three distinct magma sources (two crustal, one subcrustal). Compositions are peraluminous and range from biotite granodiorite to muscovite-biotite granite with episodic aplite formation. Hydrothermal features developed concurrently with each of these events. Field relations reveal a detailed history of fluid release from the evolving magma chamber. These fluids created high-temperature K feldspar-bearing quartz veins that are early and proximal to the magma chamber at any given time. With time, assemblages become K feldspar-destructive (albitization: albite-muscovite-fluorite ± quartz), and finally, only muscovite-stable (greisenization: muscovite-pyrite-fluorite ± quartz). This pattern is consistent with simple models of fluid evolution from the magma. Early veins, like the aplite dike swarms, have concentric and radial orientations consistent with formation in a localized, magma-chamber focused stress regime. Fractures hosting the later associations are consistently northeast oriented and controlled by far-field stress.

Features developed in Upper Proterozoic-Lower Cambrian carbonate and clastic host rocks can be linked to the intrusive history via map patterns and crosscutting relationships. Early grossular-rich garnet- and diopsidic pyroxene-bearing skarnoids (in mixed siltstone-limestone units) and marbles (in massive limestone and dolomite) form and are then deformed (flattened) during early stages of pluton emplacement. Stable isotope data demonstrate at least some of this local metasomatic exchange was accompanied by magmatic fluid influx, whereas the marbles were largely impermeable and escaped metasomatism. “Anhydrous” calcic skarns consist of more iron-rich garnet plus salitic pyroxenes, idocrase, sodic plagioclase, quartz, and fluorite. In dolomite, equivalent vein skarns consist of humite-group minerals plus calcite and variable quantities of diopside, chlorite, spinel and grossular. Hydrous skarn assemblages formed next: an older group is characterized by combinations of clinozoisite-epidote, albitic plagioclase, fluorite, chlorite, and Mg-rich biotite; a younger group is characterized by muscovite, fluorite, pyrite, and fluorophlogopite. Scheelite, beryl, sphalerite, and other sulfides accompany these hydrous skarns. Structurally controlled, distal quartz-carbonate-sulfide veins and replacement bodies extend over 5 km from the intrusion. The skarns, replacement bodies and veins all formed from magmatic fluids as inferred from isotopic and fluid inclusion data.

An integrated time-space view of the hydrothermal, structural, and magmatic development is obtained from the field relationships and other physical and chemical constraints. Deformation in the form of locally intense subsolidus foliations, folded dikes and veins, and many local shears can be linked unequivocally to particular magmatic events during emplacement. These relationships demonstrate that much, perhaps all, of the intense deformation found around the western margin of the intrusion is syn-intrusive and driven by magma emplacement. Furthermore, these patterns help establish clear links between hydrothermal events in the intrusion and in the host. Skarns, both anhydrous and hydrous, formed in response to fluid release from each of the principal intrusive pulses. Overprinting events are found locally, yet their scarcity is consistent with a slowly evolving temperature field around the intrusion and predominant fluid flow upwards rather than outward.

Similarities and differences with other magmatic-hydrothermal systems, particularly those related to granitic composition rocks, are interesting to consider for this system. Level of exposure, early and continuous fluid production, and the lack of large-scale internal communication in the evolving magma chamber(s) all may have contributed to the lack of economically significant mineralization.

† E-mail, barton@geo.arizona.edu
Introduction

The Birch Creek intrusion is located in the southern White Mountains of eastern California and is one of many such systems in the Great Basin (Fig. 1; Barton, 1987). It is a composite Late Cretaceous biotite + muscovite granitoid pluton with a multiphase hydrothermal system that can be tied through mapping, petrology, and geochemistry to the igneous, metamorphic, and structural development of the pluton and its aureole. Mapping at various scales allows a rigorous assessment of the time-space development of this system.

This paper provides context for the Society of Economic Geologists 2000 field trip stop at Birch Creek. More extensive treatments of the area are in preparation. The purpose of the day at Birch Creek is to examine a well-exposed purely magmatic-sourced hydrothermal system and to consider issues of how these features can be linked in time and space, what they reveal about the nature of lithophile element systems in the Great Basin, and what one might learn relevant to the development of economic lithophile element deposits elsewhere.

Previous work and this study

The geology of the southern White Mountains and surrounding areas has been extensively treated, most recently in a volume edited by Gary Ernst (Ernst and Nelson, 1998). Early work was motivated by mining activity in the White and Iyo Mountains and resulted in a reconnaissance of the region and descriptions of a number of mineral occurrences and small mines including several that are part of the Birch Creek system (Knopf, 1918). A wilderness study by the U.S. Geological Survey in the early 1980s constituted the only other published work on mineral resources (Diggles, 1983; Diggles et al., 1983). Most of the geologic work in the area has been done by Clem Nelson and his colleagues and students at the University of California, Los Angeles (UCLA; Nelson, 1966; Nelson and Sylvester, 1971; Nelson, unpub. map; Ernst et al., 1993, Ernst, 1996).

More recent work at Birch Creek was inspired by the need to understand the magmatic history of the Cretaceous granitoids associated with the lithophile element mineralization (Barton and Trim, 1991; Fig. 1) and to elucidate aspects of granitoid emplacement (Saint Blanquat et al., 1998). Birch Creek was chosen for detailed work because of the good exposures both in the intrusion and in the aureole. Others examined are exposed at higher structural levels or have superimposed events. A total of 10 months field work was done between 1986 and 1992 by the author and Heather Trim (Trim, 1990), assisted most summers by field camp and other students from UCLA or the University of Arizona. Collaborators on aspects of the geochemistry, geochronology, petrology, and structure include Peter Holden, Jeffrey Grossman, Gregory Ghidotti, Laurel Goodwin, Matthew Heizeler, Richard Law, Michel Saint Blanquat, and Sven Morgan. Their many contributions are acknowledged as they are highlighted in this paper.

Geologic Framework

Birch Creek is in the southern White Mountains, which are within the Late Proterozoic-Paleozoic Cordilleran miogeocline, immediately to the east and south of the Cretaceous batholiths of the Sierra Nevada and the northern White Mountains (Fig. 1). Greenschist facies clastic and carbonate rocks of Late Proterozoic and Early Cambrian age are folded in a large south-plunging antiform (Nelson, 1962; Ernst, 1996; Stevens et al., 1997). Multiple granitoid plutons of Jurassic and Cretaceous age (McKee and Conrad, 1996) intrude the antiform. Folds are tightened and strata and structures are displaced near the plutons, most notably near the Birch Creek pluton which intrudes the steeply dipping southeastern limb of the White-Inyo antiform (Fig. 2; Nelson and Sylvester, 1971).

Sedimentary rocks

The Upper Proterozoic to Lower Cambrian metasedimentary host rocks of the Birch Creek system belong to the Wyman Formation, Reed Dolomite, Deep Spring Formation and Campito Formation (Nelson, 1962). Oldest is the Upper Proterozoic Wyman Formation which has minimum thickness of 2,750 m. Much of this unit consists of thin-bedded (5–10 cm) calcareous argillites and siltstones with sparse thin limestone lenses. Lenses of light bluish-gray limestone, locally as much as 10 m thick, become abundant in the upper part of the Wyman. The uppermost Wyman Formation in the Birch Creek aureole has some dolomitic zones that may be analogous to those described in the northern Inyo Mountains (Zenger, 1976). These complicate picking the boundary between the Wyman and overlying Reed Dolomite especially where deformation and metamorphism are intense. Away from pluton contacts, the Wyman argillites are metamorphosed to variably phyllitic lower greenschist facies assemblages.

Fig. 1. Location of Birch Creek and other Late Cretaceous two-mica granites in the Great Basin. Modified from Barton and Trim (1991).
Fig. 2. Maps of the Birch Creek area showing lithologies, principal structures, and limits of hydrothermal alteration, metamorphism and deformation. Based on Nelson and Sylvester (1971), Trim (1990), Barton and Trim (1991), and unpublished mapping by Mark Barton, Heather Trim and Clem Nelson. Compiled by Mark Barton, Heather Trim, and Rick Law.
containing combinations of chlorite, biotite, clinozoisite, muscovite, quartz, and albite (Ernst, 1996). Near plutons, Ernst reports rare cordierite and K feldspar in addition to common diopside and grossularitic garnet.

The Reed Dolomite, which spans the Proterozoic-Cambrian boundary, mainly consists of ~650 m of massively bedded white- to tan-colored, fine- to medium-grained, variably oolitic, and remarkably clean dolomite (Ernst and Paylor, 1996). A medial impure sandy unit is present in the southern part of the area (Nelson, 1966). Within approximately 1 km of the Birch Creek contact, Wyman and Red carbonates are recrystallized to medium- to coarse-grained variably foliated marbles. These retain their characteristic bluish-gray and white to tan colors and, for the most part, their calcite-rich and dolomite-rich bulk compositions.

The Lower Cambrian Deep Spring Formation overlies the Reed. It consists of ~500 m of interbedded quartzite, dolomite, and limestone (Nelson, 1962). Low iron calc-silicate minerals are widespread in the Deep Spring, becoming more abundant near the intrusions.

The overlying Lower Cambrian Campito Formation consists of two units. The lower unit comprises ~850 m of medium- to thick-bedded dark-colored magnetite-rich quartzite, whereas the upper unit comprises ~350 m of shale and siltstone. The Campito does not abut the Birch Creek intrusion, but it does host distal base-metal veins. Throughout the region it commonly contains metamorphic biotite and chlorite and is upgraded to sillimanite-bearing assemblages where in contact with Jurassic plutons (Ernst, 1996).

Igneous rocks

Jurassic and Cretaceous granitoids constitute about half the exposure in the White Mountains (e.g., Krauskopf, 1968; McKee and Conrad, 1996). Four plutons plus numerous small felsic and mafic dikes are present in the southern White Mountains. The Birch Creek two-mica granite (82 Ma: U-Pb, Ar-Ar, this study; and K-Ar, McKee and Nash, 1967) is accompanied by the Jurassic Beer Creek biotite-hornblende quartz monzonite / granodiorite (ca. 175 Ma, Ar-Ar, K-Ar), the Redding Canyon biotite quartz monzonite (≥92 Ma), and the Sage Hen Flat biotite-hornblende quartz monzonite (ca. 141 Ma, Ar-Ar, K-Ar) plutons (age data summarized from McKee and Conrad, 1996). Of these, only the Beer Creek pluton is close enough to Birch Creek to have an overlapping of metamorphic aureoles (Fig. 2; see below). Late Cretaceous granites are widespread in the Whites and Inyos; however, only the muscovite-bearing Papoose Flat pluton (83 Ma, U-Pb, Miller, 1996) in the northern Inyo Mountains closely resembles Birch Creek.

Fine-grained granitic dikes are common, particular with 1 km of the contact with Beer Creek pluton (Nelson, 1966). Two generations (116–118 Ma, 147–172 Ma) of thin metamorphic dikes cut the sedimentary rocks in the vicinity of the Birch Creek pluton as they do elsewhere in the region (Iman, 1996; Ernst, 1997). Originally hornblende diabases and quartz diorites, they now have greenstained facies assemblages (chlorite-epidote-actinolite-albite). In the inner aureole of the Birch Creek intrusion, they contain amphibolite facies assemblages. These metamafic dikes do not resemble the rare dismembered biotite-rich mafic dikes and enclaves within the Birch Creek pluton that are described below with the granites. They are significant, however, because they commonly intrude normal faults in the Birch Creek area. Basaltic dikes that would correlate with Miocene flows to the north have not been described around Birch Creek.

Regional structure

The main structural feature of the range, the White-Inyo Anticlinorium, predates the large Jurassic intrusions along the eastern side of the White Mountains. Folding is inferred to be of both Antler (Late Devonian-Early Mississippian) and Last Chance thrust (Permo-Triassic) ages with principal deformation associated with the latter event (Morgan and Law, 1998). Mesozoic intrusives cut these folds but also tighten them and displace their trends, especially near the Late Cretaceous two-mica granites at Birch Creek and Papoose Flat.

As reviewed by Nelson and Sylvester (1971) the regional south-plunging White Mountain antiform is dismembered by faulting and has its trace disrupted as it tightens and wraps around the Birch Creek intrusion (Fig. 2). Open, generally upright folds in the Wyman to the north of the intrusion become isoclinal and are overturned as they track along the north and northwest sides of the intrusion (Figs. 2, 3). Likewise, the trace of faults, such as the Mollie Gibson fault, swing sympathetically around the northwestern side. foliation is well developed in the rocks of the aureole, most notably in the Reed Dolomite, where an obvious planar fabric defined by the carbonate grains extends as much as 2 km to the northwest. In contrast to relationships along the western flank of the system, units along the eastern flank are little deflected from their regional trends and are not as strongly deformed.

High angle faults and small thrust faults are widespread in the region (Nelson, 1966). Small thrust faults are only present well south of Birch Creek. Although some of the faults in Figure 2 may have reverse sense of displacement, most have normal displacement. The most significant of these is the Mollie Gibson fault system, which runs from Payson Canyon in the south northward around the west and northern sides of the Birch Creek intrusion before being lost in folded Wyman. This fault system localizes much of the distal quartz-carbonate-base metal mineralization. West of the intrusion where Campito is placed against Wyman, down-to-the-west displacement is at least 1,150 m. Many small, undeformed metamorphic dikes have been found in the Mollie Gibson fault and parallel structures requiring that movement be Early Cretaceous or older. A north-northeast–trending zone through the eastern part of the pluton is prominent on aerial photographs and on topographic maps (e.g., see fig. 1 in Barton, 2000), but no offsets have been identified where this zone crosses contacts either in the intrusion or in the aureole (cf. Fig. 2).

There is little evidence for major tilting of this system. Miocene basalts 10 km to the north dip gently (ca. 10°) to the east. Undoing this tilt would bring the antiform to a more upright position, but rotation of more than 30° to the west would create asymmetry to the west which would be inconsistent with folding linked to the eastward vergent Last Chance thrust system. Finally, in the intrusion itself, the orientations of radial aplite dike swarms and late greisen veins are predominantly vertical, rotation of more than about 10° would
systematically cant them away from vertical. Neither map patterns nor cross sections require much tilt and the Birch Creek systems itself seems sensibly upright (at least within a few tens of degrees).

Metamorphism

Regionally extensive lower to upper greenschist facies metamorphism is developed across the area. It likely reflects Paleozoic to Mesozoic burial and the superimposed thermal effects of the many Mesozoic plutons (Barton and Hanson, 1989; Ernst et al., 1993; Ernst, 1996, 1997). Well-defined contact metamorphic aureoles extent on the order of 1 km from major intrusive contacts. The contact aureole of the large Jurassic Beer Creek pluton (Fig. 2) overlaps with the northeastern edge of the Birch Creek aureole (Nash, 1962).

Ernst (1996, 1997) estimates peak metamorphic conditions away from the plutons to be about 350°C at 3 ± 1 kbars pressure and attaining temperatures of 500°–550°C adjacent to intrusive contacts (cf. Nyman et al., 1995). The former values give upper limits on background conditions possible during emplacement of the Birch Creek pluton. Their direct relevance is unclear given the complex thermal and deformational history of the region. Estimated ambient conditions during Late Cretaceous emplacement were 250°–300°C and 1.5–2.5 kbar based on Ar-Ar thermochronology, petrology, and stratigraphic considerations.

Birch Creek Pluton

The Birch Creek pluton is composite. It was emplaced as multiple pulses each from several distinct bodies of magma. Repeated episodes of aplite diking, aqueous fluid release, and intense penetrative deformation accompanied emplacement. The intrusion is broadly divided into two large units: the Border Suite and the Central Suite. These are intruded by texturally and chemically distinctive porphyritic granites, the Early and Central Porphyries (Figs. 2, 3). Other noteworthy features include regular variations in K feldspar megacryst contents, sparse biotite-rich cumulates, rare mafic dikes and enclaves, and aplite-pegmatite dike swarms that are coeval with hydrothermal pulses, and intense penetrative deformation. Internal contacts are both sharp and gradational. Sharp contacts define the boundaries between major units, but they also exist locally within the major units.

Most of the pluton was mapped at a scale of 1:6,000. Internal and external contacts were followed where practical, but most observations came from systematic examination of representative exposures and recording textures, mineralogy, and alteration and attitudes of various features. Locally, scales as large as 1:300 and 1:60 were used for key areas and critical outcrops.

Modal mineralogy is simple. Nearly all the rocks are mafic-muscovite-bearing biotite granodiorites and granites. Magnetite, ilmenite, allanite, monazite, zircon, and apatite are accessory minerals in most units. Ubiquitous igneous magnetite in this intrusion contrasts with other Great Basin two-mica granites most of which lack magnetite and are ilmenite-poor (Miller and Barton, 1990).

In each of the major suites, igneous muscovite and K feldspar increase in abundance and are ultimately joined by Fe-Mn garnet (substituting for biotite) in the progression from early phases to late aplites. Overall, the Border Suite is less evolved than the Central Suite and the Porphyries in that the early parts of the Border Suite are atypically mafic—biotite-rich granodiorite—and that Border Suite aplites commonly have biotite and rarely have garnet. In the other groups, rocks are mainly monzogranites and biotite-bearing aplites are subordinate to garnet-bearing aplites.

Most Border Suite and Central Suite units have equigranular or seriate textures. Grain sizes range from medium to coarse, and the abundances of large K feldspar and quartz crystals vary systematically (Fig. 4A-D). Interior to aplite dike swarms, K feldspar megacrysts become abundant, locally constituting up to half of the rock (Fig. 4D; see also Barton, 2000, figs. 3, 4). The Early and Central Porphyries have subhedral, but distinct bimodal grain-size populations with a medium-grained groundmass (1 to 4 mm grains) with sparse K feldspar, quartz and rare muscovite phenocrysts (Fig. 4E). Where foliation is weak, textures are typically hypidiomorphic granular with subhedral quartz and euhedral K feldspar in a
matrix of quartz, feldspars and micas (e.g., Fig. 4B, C). With increasing deformation, these textures are destroyed, although K feldspar augen and elongate aggregates of recrystallized quartz commonly testify to their original presence.

Magmatic and subsolidus foliation and lineations are defined by alignment of micas, feldspars, and aggregates of recrystallized quartz. Macroscopic and microscopic textures are consistent with early, predominantly magmatic foliations and lineations, preserved mainly in the central and eastern parts of the pluton, and strong subsolidus fabrics around the periphery but best developed in the western parts (Fig. 4F, G; Laurel Goodwin, pers. commun. 1996; Saint Blanquat et al., 1998). In addition to quartz, feldspars are recrystallized in the intensely deformed rocks. This evidence of high-temperature deformation is consistent with field relationships that show these fabrics in many areas are cut almost immediately after

Fig. 4. Photographs of some representative granitoids and alteration assemblages. White bars are 1 cm long. A. Strongly foliated Border Suite biotite (± muscovite) granodiorite. Here and in photos B-E, K feldspar is stained yellow, plagioclase is light colored to pink, quartz is gray, and biotite is black. B. Moderately foliated Border Suite muscovite-biotite granite. Note the large ovoid quartz eyes and the K feldspar megacryst. C. Weakly foliated K feldspar megacrystic Central Suite biotite-muscovite granite. D. K feldspar megacryst-crowded biotite-muscovite granite from the base of the aplite dike swarms. E. Megacrystic biotite-muscovite granite from the Central Porphyry. This is typical of Early and Central Porphyries. F. Strongly foliated (subsolidus) Border Suite biotite granodiorite (cf. A). Crossed polars. Field of view ca. 2.5 mm. G. Unfoliated Central Suite granite (cf. E). Crossed polars. Field of view ca. 2.5 mm. Photos F and G courtesy of Laurel Goodwin. H. Microphotograph (crossed polars, 2.5 mm wide) of typical albitization: albite + muscovite + fluorite ± quartz. I. Clinoozoisite-fluorite ± chlorite-albite vein with biotite-fluorite-chlorite envelope. These represent reaction between aplite-generated fluids and Reed Dolomite. J. Muscovite-fluorite-pyrite greisen vein with muscovite-quartz-fluorite inner envelope and muscovite-albite-chlorite-quartz outer envelope in biotite-muscovite granite. K. Retrograded diopside-calcite-chlorite vein with humite-group mineral-calcite envelope in dolomitic marble. Tremolite replaces diopside and serpentine replaces humite-group mineral. L. Section through greisen muscovite-fluorite-beryl vein with banded fluorite-phlogopite envelope in dolomitic marble. M. Clinohumite (clh)-calcite envelopes on calcite-chlorite-clinohumite veins in Reed dolomite. Cut by later, thin tremolite-calcite veins. N. Wyman Formation skarnoid cut by dikes of biotite granodiorite (no reaction envelope) and aplite (strong reaction envelope) to Fe-bearing garnet in the marble and salitic pyroxene in the calc-hornfels. Field of view approximately 2 m across.
crystallization by less deformed dikes emanating from adjacent intrusive units (see below; also see Barton, 2000).

Minor igneous units include extensive leucocratic dikes (mainly aplites and pegmatites), zones of K feldspar-biotite-rich unidirectional solidification textures, minor biotite-rich accumulations, and very rare mafic dikes and enclaves (Fig. 5). All four types are spatially associated.

Aplite dikes contain minor Fe-Mn garnet or biotite in addition to muscovite. They grade into simple pegmatites and quartz-feldspar veins (Fig. 5G). Inward, toward their sources, they transition into fine- and medium-grained muscovite–biotite-bearing leucogranites. Unidirectional solidification textures (USTs, or comb layers; see Moore and Lockwood, 1973; Shannon et al., 1982, Kirkham and Sinclair, 1988) consist of single or multiple layers of large (up to 20 cm long) elongate K feldspar crystals growing in parallel and branching toward the magma chamber (Fig. 5A, B). They are interpreted here, as they are elsewhere, to represent zone of fluid accumulation or flow along the margins of the active magma chamber. Spherulitic layering locally accompanies the USTs (Fig. 5E). Aplite dike swarms emerge from contacts decorated with USTs. For specific examples, see the descriptions and maps in Barton (2000).

Also in these same zones are biotite-rich banded rocks (Fig. 5C) and K feldspar-megacryst-crowded granites (Fig. 5D). The latter commonly choke the base of leucocratic dikes. The former have abundant biotite, magnetite, zircon, and monazite plus feldspars and quartz. Both types are interpreted to be accumulations of crystals based on field relationships and whole-rock compositions (see petrologic discussion below).

No more than a dozen occurrences of fine-grained equigranular biotite quartz monzodiorite dikes and enclaves have been found (Fig. 5D). These mostly occur with the biotite-muscovite granites in the Border and Central Suites just interior to UST/aplite zones. Some are cut by contemporaneous leucocratic dikes. Several in the Border Suite are strongly foliated. Their mineralogy and isotopic compositions indicate equilibration with their local host granitoids, but the quartz-poor compositions and their dike-like geometries point to injection of a distinct mafic magma contemporaneous with em-

![Fig. 5. Photographs of igneous features. A. UST zone showing left-lateral skewing of large K feldspar crystals and biotite-rich accumulations above quartz vein. This is at the contact of megacrystic muscovite-biotite granite with biotite-muscovite granite in Border Suite. Lens cap 6 cm for scale. B. Multiple layers of dendritic K feldspar from principal UST zone on Central Suite contact. C. Biotite-rich accumulation from near Central Suite-Border Suite contact. White bar is 1 cm long. D. Mafic dike cutting Central Suite granite along the eastern contact. These are cut by Central Suite pegmatite dikes. E. Orbicular layering in Border Suite muscovite-biotite granite near UST zones. F. Sheeted aplite dikes typical of more intense part of swarms. G. Quartz K feldspar vein center to aplite dike in aplite dike complex. H. Foliated biotite-muscovite granite cut by aplite and pegmatitic dikes perpendicular to foliation. These are offset by quartz-K feldspar vein (diagonal up to right) and then by thin muscovite-pyrite-filled greisen zones.](image-url)
placement of the pluton. One can speculate that the coincidence with the aplite events reflects deeper injection of mafic magma that triggers large-scale fluid release, similar to interpretations in some volcanic and porphyry systems. Alternatively, it could be that magma chamber dynamics associated with aplite dike and fluid-release allow mafic melts already present to reach higher levels where they freeze.

Spatial patterns in the intrusion

The Border Suite constitutes roughly two-thirds of the exposed area with the Central Suite making up most of the remainder (Fig. 2). Early and Central Porphyry dikes make up about 5 percent of the outcrop, somewhat more than the sum of aplite dikes and rarer features. Numerous internal contacts within these units have been defined locally. Contacts between the Porphyries and the other units are generally simple and sharp (e.g., see fig. 9 in Barton 2000), whereas the Central Suite-Border Suite contact is defined by a well-developed zone of USTs that forms the base of the prominent radial aplite dike swarms. The latter contact is commonly gradational over 50–100 m with the formal boundary being put at the innermost major set of USTs (cf. fig. 15 in Barton, 2000). Map patterns in the south-central section of the pluton indicate that there may be multiple pulses of emplacement and diking within the inner Border Suite, the later becoming gradually more like the Central Suite. The largest of these lies 0.5 to 1.5 km outside the Central Suite contact. There, a prominent aplite zone radiates from a discontinuous UST zone. These internal contacts schematically indicated by the distribution of aplites in Figure 2.

In general, the earliest Border Suite rocks are biotite granodiorites that contain minor muscovite that is difficult to see in hand specimen. They are followed by biotite > muscovite and biotite > muscovite granitoids that become progressively more leucocratic about halfway in and then revert to rocks with biotite substantially greater than muscovite. Unlike the Central Suite, which has a fairly regular contact with the surrounding rocks, the Border Suite was clearly emplaced as dikes (see the outer contact of the pluton, Fig. 2; cf. fig. 4A in Barton, 2000). Internal variation in the Border Suite can be inferred to progress generally from relatively biotite-rich to muscovite-rich and back to biotite > muscovite before reaching the Central Suite contact. K feldspar megacrysts are widespread in the muscovite-bearing parts of the Border Suite, but become abundant (>25 per m²) only locally, commonly immediately internal to discontinuous UST zones and associated aplite swarms (e.g., Camp Hill area, Fig. 4, Barton, 2000).

The Central Suite constitutes the northeastern third of the pluton and is emplaced against Border Suite rocks around its western two-thirds and against Wyman, Reed and Deep Spring around its eastern third (Fig. 2). It is mainly coarse-grained, biotite-muscovite granite with variable contents of K feldspar megacrysts. Fine-grained leucogranites form discordant bodies up to about a hundred meters in length within the Central Suite and it is intruded by the Central Porphyry (below). Megacrysts are most conspicuous along the margins of the Central Suite, just interior to the major aplite swarms. Numbers exceed 100 per m² over large areas (fig. 3 in Barton, 2000).

The Early and Central Porphyries are smaller bodies, several kilometers in length but rarely more than a few hundred meters wide. They are texturally fairly uniform. The main variability within these units is the abundance of K feldspar and quartz phenocrysts. The older porphyry units are peripheral, foliated, and cut by aplite dikes from the Central Suite. Contacts dip outward and these early bodies have a broadly concentric distribution (Fig. 2). The Central Porphyry shows no subsolidus deformation and dips steeply.

Aplite dikes swarms emanate from muscovite-rich phases in all major units. In the Border Suite, the aplites cut early biotite-rich phases because the muscovite-rich rocks are well inside the outer contact. Aplites are largely external to the uniformly muscovite-rich Central Suite and Central Porphyry. They can extend outward over 1 km from their sources. Aplites generated from the Border Suite both parallel and radiate from the elongate masses of muscovite-rich granite from which they are sourced. They typically are foliated but are rarely folded. Central Suite aplites are distinctly radial, emanating from the contact is a large starburst pattern (fig. 3 in Barton, 2000). Among the porphyries, aplites are abundant only with the Central Porphyry, where they parallel the main mass. Orientations of these dikes, plus those of foliations and major vein types, are shown in Figure 6A.

Foliation is prominent on the margins of the intrusion, particularly on the northwest side (Nelson and Sylvester, 1971; Saint Blanquat et al., 1998). Figure 6B shows compilation of the qualitative field classification of foliation intensities. This is broadly similar to patterns obtained from quantitative petrographic and anisotropy of magnetic susceptibility studies (Saint Blanquat et al., 1998, Laurel Goodwin, pers. commun.). Deformation is predominantly accomplished by flattening. Small ductile shear zones and stretching lineations are also common. Consistent patterns in these features commonly make sense, for example, a consistent sense of shear is commonly found on the margins of some of the large aplite swarms (e.g., see fig. 13 in Barton, 2000).

In a number of cases, the field relationships demonstrate that much or all of the penetrative fabric in the rocks accumulated in the interval that it took to crystallize a few tens to a few hundreds of meters of the adjoining granites. Both on the Camp Hill and on the western tip of the Central Suite (Fig. 2), early quartz veins are folded with axial planes subparallel to the foliation in the host granitoid. In each case, these folds are cut by unfolded, little foliated aplite dikes, and quartz veins that emanate from identified units within 100 m inward (see Barton, 2000).

Metamorphism and Metasomatism

Hydrothermal mineral assemblages are ubiquitous in the Birch Creek pluton and extend at least 5 km north and south of its margins (Fig. 2). These comprise quartz-feldspar, albite, and greisen veins in the intrusion, and calcic and magnesian anhydrous and hydrous skarns and quartz-carbonate veins and replacement bodies in the metasedimentary rocks (Table 1). The skarns and veins overprint early thermal metamorphism in the aureole, but in many cases the skarns are themselves deformed with intrusion-linked fabrics. Relative timing can be established by local crosscutting relationships. Furthermore, by use of the aplite dike swarms as markers, a
system-wide relative chronology can be established. A time-space synthesis for the system based on these relationships is presented at the end of this paper.

**Intrusion**

Quartz-feldspar veins with variable amounts of accessory biotite, muscovite, magnetite, and ilmenite form the earliest metasomatic features at any given location in the intrusion (Fig. 5H). They may predate, be contemporaneous with, or rarely, postdate aplite and pegmatite dikes. The earliest veins in most areas dip outward toward the pluton contacts and are cut by dikes from broadly contemporaneous aplite swarms (e.g., see figs. 6, 11, 12 in Barton, 2000). Most of these veins are fairly wide (1–50 cm).

Aplite dikes commonly grade into radially oriented pegmatitic quartz veins that show “vein dike” features indicative of the magmatic-hydrothermal transition (Fig. 5G). More often than not, these veins cut slightly older aplites and pegmatites (Fig. 5H). Although envelopes are absent on most of these veins, a distinct style of narrow (0.1 to 1 cm wide) quartz veins with obvious K feldspar envelopes are present near the major UST zones (fig. 14C in Barton, 2000). Sulfides are absent except in quartz-K feldspar-biotite veins that are abundant in the northeastern corner of the Central Suite, near the eastern end of the Central Porphyry. Chalcopyrite is a minor constituent (<1 vol percent) in the quartz veins and in the nearby skarn (Trim, 1990).

Quartz-feldspar veins locally constitute upwards of 10 vol percent of the rock. The areas of highest abundance are within zones no more than a few hundred meters wide immediately outside of the aplite-generating contacts. Repeated, mutually crosscutting relationships and map patterns allow a detailed picture of fluid release and dike generation to be formed in several areas (e.g., field trip stops IN-2 to IN-7 in Barton, 2000). In a few places, early quartz-feldspar veins are distinctly more iron-rich (as hydrothermal magnetite and biotite) than later veins from the same aplite/vein generation event (see fig. 15 in Barton, 2000).

Superimposed on the quartz-feldspar veins are fracture-controlled zones of albitization and greisenization. Unlike the older veins, these zones have a pronounced northeasterly strike and are more evenly distributed through the intrusion (lower inset in Fig. 2, Fig. 6A). They form the prominent northeasterly grain in the intrusion as seen from the air (fig. 1 in Barton, 2000). Quartz-muscovite-K feldspar veins transition to quartz-muscovite veins with envelopes of albite (first after plagioclase, then after K feldspar) + muscovite + quartz ± chlorite (after biotite). Subsequent veins and replacement zones of albite + muscovite + fluorite ± pyrite ± chlorite show evidence of quartz removal as well as K feldspar destruction (Fig. 4H). This association is succeeded by late greisen veins. These consist of narrow open-space muscovite + pyrite ± quartz with envelopes of muscovite + quartz ± pyrite ± fluorite ± albite (Fig. 5J). In places, these envelopes coalesce to produce pervasive greisenization over widths of a few meters and lengths of a few tens of meters.

The later veins commonly have slicken lines and other evidence of brittle movement, but they are almost never penetratively deformed, even where they apparently formed relatively early in the history of the system as on Camp Hill. Open space is common. In thin section, quartz from these veins can be strained but it rarely shows subgrain development. In the

![Figure 6: Orientation, distribution, and intensity of selected structural elements within the Birch Creek intrusion. Based on unpublished mapping by M.D. Barton. A. Lower hemisphere equal area plots for selected areas in the intrusion. Each stereonet plot is based on at least 20 observations. Note the consistent intrusion-centered orientations for early features (foliation, aplites) and the regional control expressed in the later greisen veins. B. Qualitative intensity of foliation determined during mapping. See text for discussion.](image)
older intrusive units, these are the only materials that have usable, primary (?) fluid inclusions (Van Averbeke, 1996).

Endoskarn is widespread around the margins of the intrusion and in dikes and sills that cut the carbonate rocks. It occurs in small zones which are rarely more than a few meters wide but can extend for many tens of meters. Endoskarn can contain garnet, pyroxene, actinolite, epidote, and chlorite. It is typified by formation of abundant albitic plagioclase or muscovite replacing igneous feldspars. Fluorite is common in some. The calc-silicate zones are typically narrow (<1 m) and form as veins or selvages. The plagioclase-rich zones are commonly depleted in quartz and have actinolite, epidote, or chlorite in the biotite sites. In several cases, endoskarn is overprinted by magmatic-sourced alteration (see fig. 8D, E in Barton, 2000). Shreddy texture biotite replaces actinolite in endoskarn where it is cut by quartz (-K feldspar veins). Greisen veins (muscovite-fluorite with muscovite-chlorite-fluorite envelopes) also cut endoskarn. The lack of foliation in these younger K mica assemblages shows that endoskarn formed almost immediately after emplacement and ductile deformation of the host biotite granodiorite.

Metasedimentary rocks

Contact metamorphism at Birch Creek was studied extensively by Heather Trim (1990) as part of her Ph.D. dissertation. Earlier work by Nelson and Sylvester (1971) outlined the general extent of metamorphism. Nash (1962) examined the area where the aureole of the Beer Creek pluton and that of the Birch Creek pluton overlap. Map patterns, crosscutting relationships with dikes, and timing relative to deformation

Table 1. Alteration Features in the Birch Creek System (listed in local sequence) 1,2,3

<table>
<thead>
<tr>
<th>Igneous rocks</th>
<th>Dolomite</th>
<th>Limestone</th>
<th>Clastic rocks</th>
</tr>
</thead>
<tbody>
<tr>
<td>Early</td>
<td>Endoskarn (can be several events): Plagioclase [feldspars] ± actinolite ± epidote ± chlorite ± garnet ± clinopyroxene</td>
<td>Recrystallization to foliated marbles; local tremolite; talc; diopside; phlogopite</td>
<td>Recrystallization to foliated marbles; grossular-rich garnet ± idocrase wollastonite adjacent to clastic units</td>
</tr>
<tr>
<td>Intermediate</td>
<td>Pegmatitic quartz veins</td>
<td>Andraditic garnet ± magnetite ± salite (epidote + actinolite ± chalcopyrite overprint)</td>
<td>Granitie garnet ± idocrase ± wollastonite ± salite; commonly boudinaged</td>
</tr>
<tr>
<td>Late</td>
<td>Quartz + muscovite (± K feldspar) // ± albite + muscovite</td>
<td>Mn-rich clinzoisoite (or epidote) + fiorite + chlorite ± scheelite // chlorite + fiorite + albitie ± actinolite</td>
<td>Epidote ± idocrase ± fiorite ± chlorite ± scheelite</td>
</tr>
</tbody>
</table>

1 Based on Trim (1990) and Barton (unpub. data)
2 The mineral associations are listed in general paragenetic sequence (i.e., the typical order of formation in any given location). The left-to-right correspondence only roughly reflects local equivalence between different lithologies. This is not the correlation across the system which has multiple events in which many of these assemblages are repeated, and in which some form distally to others. See the section on Time-Space evolution for discussion
3 Symbolism is: minerals in parentheses are generally minor in abundance (<10 volume percent), ± indicates presence or absence; the double slash (“//”) indicates the division between the vein and vein envelope; a single slash (“/”) indicate divisions within veins or envelopes; minerals in brackets ([plagioclase]) indicate mineral that was replaced
demonstrate that thermal metamorphism and a wide variety of metasomatic features (Table 1, Figs. 4, 7, 8) formed during and immediately following magma emplacement.

“Thermal” metamorphism: The Wyman, Reed, and Deep Spring are thoroughly metamorphosed within about 1 km of the pluton contact (Fig. 2, upper inset). Mixed lithologies are converted to calc-hornfels and garnet-bearing skarnoid, whereas the more massive limestones and dolomites are converted to medium to coarsely crystalline marbles (1–10 mm grain size). Pluton-centered foliations and sparse Ca-Mg silicates (tremolite-talc-phlogopite) occur sporadically outward to about 4 km away from the contact (Fig. 2, lower inset). The latter are localized mainly along faults and fractures. They probably reflect metamorphic decarbonation reactions driven by fluid influx.

Skarnoid is well developed in the thin-bedded mixed argillite-limestone beds in the Wyman Formation. It is especially well developed where early dikes are abundant as, for example, north of Camp Hill. It typically consists of reddish brown weathering calc-hornfels (diopsidic pyroxene, clinozoisite, biotite, plagioclase, quartz, minor grossularitic garnet). Limy interbeds are recrystallized and contain variable amounts of grossular-rich garnet. Wollastonite is present in some areas in this association and requires, like the clinozoisite, infiltration of water-rich fluids to keep XCO2 at low values (e.g., see Einaudi et al., 1981).

Skarnoid shows little evidence of major chemical changes other than loss of volatile constituents. Much of it, however, has been isotopically exchanged with lighter oxygen. The distribution of the skarnoid around the intrusion margins, its presence in reactive mixed lithologies, and the isotopic compositions that are shifted toward magmatic values all are consistent with the passage of significant amounts of magmatic waters (Fig. 9). Simple mass balance conditions indicate that more than adequate amounts of fluid would have been released by the magma (Trim, 1990).

Massive carbonates in the Wyman, Reed, and Deep Spring are converted to folded and foliated marbles which generally lack calc-silicate minerals except along veins and adjacent to dikes (Fig. 7A). Penetrative deformation in the Wyman and other units is early: It postdates skarnoid and some of the anhydrous skarns, but predates many aplite dikes and nearly all of the hydrous skarns. Reed marble located in screens and pendants within the intrusion is fine-grained (<0.5 mm) with a sugary texture that shows no evidence of the foliation that is well developed in adjoining dikes. These zones must have recrystallized statically at relatively high temperature following the termination (locally) of penetrative deformation.

Fig. 7. Photographs of some features from the aureole. A. Deformed calc-silicate bands in Wyman marble. This view looks south of outcrop west of pluton. B. Similarly oriented quartz vein with pronounced top-to-the-right skew of crystals. From Camp Hill, west of the pluton. C. Pink clinozoisite vein with fluorite-chlorite-biotite-tremolite envelope cutting calcite-humite zone.

Fig. 8. Late veins in intrusion and aureole. The majority of late veins have northeast orientations regardless of their position in the system. A. Northeast-trending greisen veins in Border Suite granite, west side of intrusion. Outcrop is about 10 m high. B. Close-spaced tremolite-calcite veins in foliated Reed Dolomite, west side.

Fig. 9. Carbon-oxygen isotope plot for carbonates from various assemblages in the aureole (from Trim, 1990). Magmatic (quartz) oxygen is quite uniform near 10.5 per mil. See text for discussion.
Early, “anhydrous” skarn: Magnesian vein skarns form in the Reed Dolomite within about 1,000 m of the contact (upper inset, Fig. 2). Early skarns are characterized by huminite-group minerals and locally more massive garnet-bearing skarns on the immediate contact. The early set consists of calcite plus a huminite-group mineral (clinohumite, Mg₆(SiO₄)₄(F,OH)₂; chondrodite, Mg₃(SiO₄)(F,OH)₂; or norbergite Mg₂(SiO₄)(F,OH)₂), typically in a graphic intergrowth (Figs. 4M, 7C). The graphic texture represents the complete consumption of dolomite to make huminite-group mineral + calcite, for example, by the reaction:

\[
3\text{CaMg}((\text{CO}_3)_2 + \text{SiO}_2\text{aq}) + 2\text{HF} = 
\text{Mg}_6(\text{SiO}_4)(\text{F,OH})_2 + 3\text{CaCO}_3 + \text{H}_2\text{O} + 3\text{CO}_2
\]

Veins centers commonly chloritize, spinel, talc and tremolite. As the pluton contact is approached, first pyroxene and then garnet join the vein centers (Trim, 1990). Many of these veins parallel foliation and they can be folded (Fig. 5 in Barton, 2000). These zones can be quite abundant, in places their envelopes coalesce completing replacing the dolomite. The mineralogical changes require significant addition of silicate, fluorine, aluminum and, in the spinel-bearing variants, titanium. From their spatial distribution and early timing relative to fabric development, Trim inferred that the huminite veins are linked to small bodies of massive magnetite-andradite skarn near the contact and to biotite-magnetite-bearing quartz veins in the granites.

Granodiorite garnet skarns in calcite marble are overprinted by late subcalcic garnet-quartz-idocrase epidote-plagioclase-bearing zones. The latter commonly fill pull-apart structures in thin zones of massive garnet. Their analogue in the Mg skarns is found in widespread calcite + chlorite (+ other minerals) open-space fillings in sheeted and chocolate tablet-type extensional structures (see Figs. 5, 8, in Barton, 2000; rare aplites fill these openings too). In a number of places, aplites clearly correlate with garnet-pyroxene skarns (Fig. 4N).

Hydrous skarns: Hydrous skarn assemblages are the most common type and many have strongly aluminum and fluorine-rich mineral assemblages (Table 1; and distinctive the Great Basin systems as a whole; Barton, 1987). In dolomitic marbles, clinzoisite-plagioclase-chlorite-fluorite veins occur throughout the inner aureole (Fig. 2; Fig. 4I, Fig. 7C). These are almost never strongly deformed. In a few places, single veins can be followed back into aplite dikes that show desilication, K feldspar replacement and chlorite-fluorite-biotite envelopes where they enter the dolomite. Serpentinite, talc, and tremolite retrograde earlier huminite-group minerals and diopside (Fig. 5K). These hydrous minerals commonly form independent, typically straight veins that cut foliation and have attitudes similar to late veins in the intrusion (Fig. 8; see Trim, 1990). In calcite marbles and in garnet skarn, older epidote-idocrase-plagioclase-fluorite association is succeeded by chlorite-biotite-muscovite-fluorite-bearing assemblages. The latter commonly contains scheelite, beryl and sulfides. All types of host rocks and skarns are cut by distinctive and widespread, but volumetrically minor muscovite-fluorite-beryl veins (Fig. 4L). These muscovite-rich veins occur as far as 3 km from the main contact on the northwest side, although that area is probably underlain by a shoulder of the granite (Fig. 2, lower inset).

Quartz-carbonate-sulfide veins: This group extends from the inner aureole, where it overprints all other types, to more than 5 km from the pluton where the rocks bear no other evidence of the Birch Creek system. These bodies are generally texturally simple and small—vein and stockworks are 0.1 to 2 m wide, replacement bodies are 1 to 10 m across. In proximal locations, the veins can have sparse contents of scheelite, chalcopyrite, feldspar, fluorite, and muscovite. With distance, only quartz, carbonates (calcite or dolomite, rarely Fe-Mn carbonates), chlorite, and sericite are present with minor pyrite and base-metal sulfides. Some early veins of this association are deformed (Fig. 7B) and must therefore link to early events in pluton emplacement. Most lack evidence for overall timing except that Ar-Ar results (M. Heizler, unpublished data) demonstrate that they are synchronous with the rest of the intrusion (+1 m.y.). Stable isotopes and fluid inclusion compositions resemble magmatic values and do not show compelling evidence for external fluids although some host-rock contribution is required (Trim, 1990; Van Averbeke, 1996).

Geochemistry and Petrology

Petrologic and geochemical studies show that the intrusion formed from at least two different magmas and evolved only modestly at the level of emplacement. Aqueous fluids were almost entirely magma-sourced. Over 600 thin sections and 200 rocks were examined by a variety of petrographic, fluid inclusion, electron microprobe, isotopic, and whole-rock analytical techniques (e.g., Trim, 1990; Barton et al., 1994; Van Averbeke, 1996). This work will be presented in detail in several papers in preparation. Key observations and interpretations are summarized here.

Magmatic system

Ar-Ar analyses of muscovite and a U-Pb analysis of monazite show that the magmatic system formed close at 82 ± 1 Ma (M. Heizler and P. Holden, unpub. data) and cooled below muscovite closure temperatures quickly. Whole-rock Rb-Sr data from the geochemically coherent Border and Central Suites are consistent with this age (Fig. 10).

Most notable about the intrusion is the isotopic evidence that shows that the Border and Central Suites represent a different magma than that of the Early and Central Porphyries (Fig. 10A; P. Holden, unpub. data). The Porphyries are distinctly less radiogenic in their strontium and more radiogenic in their neodymium isotopic compositions. Lead and oxygen isotope compositions are consistent with this variation, but their differences are less pronounced. These results could indicate separate sources, or more plausibly, different degrees of mixing of several sources somewhere deeper in the magmatic plumbing. That these compositional differences persisted to be emplaced as two alternating sequences seems remarkable (isotopically heterogeneous igneous systems being otherwise common). This also testifies to the lack of mixing at higher levels in the magmatic systems, which has consequences for fluid evolution and mineralization. Overall, a combination of deeper crustal materials with a small contribution of mafic magma best accounts for the geochemistry of this system as it does for other Late Cretaceous granitoids in the Great Basin (for data and discussion, see Barton, 1990,
subequal amounts of $K_2O$ and $Na_2O$, and with SiO$_2$ contents of about 72 wt percent. Particularly Na- or K-rich variants have not been found. Aplites are significantly more evolved in their trace element contents. For example, rare earth elements (Fig. 10C) show strong LREE and Eu depletions compared to their progenitors (J. Grossman, unpub. data).

These and other trace element changes are compatible with local removal of monazite, allanite, and plagioclase, with some loss too of biotite and K feldspar. The complexity of local compositional evolution is illustrated in Fig. 11, which shows normative quartz and feldspar compositions for rocks collected near the Border Suite-Central Suite contact near the detailed map area (figs. 3, 15 in Barton, 2000). Most granites lie to the feldspar-rich side of the 2 kbar water-saturated granite minimum, whereas the aplites and leucogranites scatter around it. A 2 kbar lithostatic pressure is consistent with independent depth estimates. The biotite-dominated late Border Suite is distinctly plagioclase- and mafic-rich compared with older phases. In the immediate vicinity of the contact, K feldspar-rich granites (Fig. 4D) and the biotite-rich layered rocks (Fig. 5C, but including USTs) plot markedly toward the K feldspar apex. The latter also contain unusually high concentrations of accessory phases and their corresponding trace elements (e.g., LREE, Zr). Well inside the Central Suite contact (>200 m?), granite compositions revert back to more typical values (Fig. 11). The interpretation is that the aplites represent the separated liquid fraction for an evolved margin to the outer portions of the Central Suite (to be expected from the field relationships). However, the evolution combined modest crystal fractionation relative to the sidewall (biotite-rich granite and banded rocks) with selective filter pressing of K feldspar megacrysts and other phenocrysts during the aplite-forming events. The megacrystic granite by itself can not mass balance the aplites (the key rocks—main units, aplites, megacrystic granite—are not colinear).

What do the megacrysts indicate? Clearly they correlate with the aplite swarms, being immediately internal to them and best developed where the swarms are strongest. The texture may be largely driven by crystallization kinetics; however, they seem likely to be promoted by the fluid-release/diking events. Perhaps they result from shifts in the feldspar coticic with changes in melt water content (Fig. 11, lower arrow). If so, the megacrysts may indicate the scale of fluid concentration and release within the magma chamber. This local variation contrasts with far more extensive megacryst distribution seen in some other kinds of magmatic-hydrothermal systems.

**Hydrothermal system**

As noted above, stable isotope evidence points to the predominance of magmatic waters in the intrusion and in the aureole (Trim, 1990; Barton, unpub. data). Oxygen isotope values are consistent with simple cooling of magmatic fluid from near-magmatic temperatures (500°–600°C) for quartz-feldspar veins down to 300°–400°C for greisen assemblages. Likewise, skarns show strong shifts in their oxygen and carbon isotopes to values compatible with a magmatic fluid (Fig. 9). Skarnoid and other thermal calc-silicate assemblages show a similar shift which is best interpreted in silicate-rich rocks as a combination of infiltration (to shift oxygen) and decarbonation reactions (to shift carbonates). Sulfur and hydrogen both
show evidence of mixing (Trim, 1990), yet even the most extreme hydrogen values are consistent with mainly magmatic fluids.

These results are compatible with fluid inclusion observations (Van Averbeke, 1996). Where there are usable fluid inclusions, she found moderate salinities (mostly 5–10 wt percent NaCl equiv) throughout the system—from the highest-temperature igneous-hosted alteration to the distal quartz-carbonate-base metal veins. Moderate CO₂ contents were inferred that the hydrothermal system operated under near-lithostatic conditions for much, perhaps all of its history. The key facts include the following: common outward-dipping veins, lack of evidence of external fluids, abundant evidence for near-continuous magmatic fluid production, semi-continuous ductile flow in the intrusion and inner aureole, and several independent rock- and fluid-pressure estimates, all near 2 kbar. This fits nicely with hydrodynamic models for deeper, water-bearing magmatic systems (Hanson, 1995). These results are similar to those obtained on Late Cretaceous systems in the Great Basin. Only in the uppermost parts of the other systems is much meteoric water incorporated (e.g., at McCullough Butte, NV, see Barton, 1987; Barton and Trim, 1991).

**Time-Space Evolution**

Figure 12 shows an interpretation of the temporal and spatial evolution of the Birch Creek system. The field trip illustrates many of these features, with a focus on exposures containing critical relationships that help establish this framework. The need for a time-space framework motivated my original study at Birch Creek. None of the other two-mica-granite-centered districts in the Great Basin have the exposures necessary to choose between fundamentally different models for system development, particularly the central links to igneous history (for example, compare McCullough Butte, Nevada, Fig. 1, Barton, 1982). The key issues are how the system changes with time and how spatially distinct features correlate. Understanding these relationships is essential to any full description of system formation and is central to using those results in practical applications.

This Birch Creek time-space diagram (Fig. 12) shows the evolution of events during magma emplacement and cooling as a function of the horizontal distance from the locus of magma emplacement at any given time. The key to establishing the relationships in this diagram are the aplite dike swarms, indicated by horizontal gray bars. These are tied by their distributions, fabrics, and compositions to particular events in the overall development of the intrusion and the inner aureole. The diagram illustrates how contacts shifted during emplacement of the magmas (analogous to construction of the volcanic piles at Yerington and Humboldt) with direct implications for the origin and timing of the intense intrusion-centered deformation. The figure is drawn for a trajectory from the Central Porphyry contact west toward the Mexican mine (see Figs. 2, 3). In this respect, it contrasts with the diagrams for the tilted systems at Yerington and Humboldt, which show analogous relationships as a function of paleodepth instead of in the horizontal. In all three cases fluids flow outward in the illustrated sections.

**Border Suite:** The system commenced with emplacement of multiple 20- to 100-m-wide dikes of biotite (± muscovite) granodiorite of the Border Suite. These dikes generated quartz-feldspar ± Fe-Ti oxide) veins. These veins were folded and then immediately cut by weakly foliated aplites and by recrystallized but unfolded quartz(-feldspar) veins that emanate from adjacent, more muscovite-rich phases of the Border Suite (see figs. 4, 6, in Barton, 2000). Correlative metamorphic events in the inner aureole included formation of skarnoid and high-T Ca and Mg skarns. These are also deformed (folded, boudinaged) concurrently with the development of strong foliations in the proximal marbles and the marginal granitoids. This deformation accompanied emplacement of the more muscovite-rich units as evidenced by the crosscutting quartz veins and aplite dikes in the intrusion and by widespread, correlative clinzoisite-plagioclase veins in the aureole (Trim, 1990). This deformation must have resulted from inflation of the Border Suite-time magma chamber: the flattening foliations parallel the more muscovite-rich younger units and record only high-temperature events—i.e., their recrystallized igneous feldspars and affected quartzfeldspar) veins but do not affect later aplites or the ubiquitous...
Endoskarn (not indicated in Fig. 12; formed by local influx of aureole fluids into the pluton) likely formed in this interval, as it overprints foliations but is cut and potassically altered by the slightly later quartz-feldspar veins and by greisenization.

Time scales for conductive cooling require that the early Border Suite events could have taken no more than a total of about 10,000 yr. It could have been much less. This early part of the overall history was a time when much of the aureole was heating, both by thermal conduction and by infiltration of fluids expelled from the magma. Inferred fluid-to-rock ratios of 1 or more for much of the Wyman skarnoid are large enough to affect the thermal budget. Overall, the isotherms would have moved outward early on (cf. Fig. 12). Reversal in their movement (where they have vertical slopes on the time-space diagram) marks the transition between prograde and retrograde thermal histories. (One might expect a more complex thermal pattern than that indicated on Fig. 12 given the episodic release of aqueous fluids and influxes of new magma; however, such events are largely unconstrained by petrologic evidence and thus are not illustrated.)

Early Porphyries: The next set of significant events are not shown in Figure 12. Early Porphyry dikes were emplaced following consolidation of their local Border Suite host, but prior to emplacement of the Central Suite. These bodies are well foliated and are cut by unfoliated Central Suite aplite dikes and quartz veins. Also, as described above, aplite dike swarms and internal contacts provide evidence for additional fluid-release pulses within the Border Suite (following more magma emplacement?). Two scenarios for Early Porphyry emplacement are plausible: first, the dikes could have been injected from their chemically distinct magma chamber into the crystallized carapace of the evolving Border Suite-Central Suite chamber (as implied by the continuous presence of magma on the left side of Fig. 12). Alternatively, they could have been injected during a hiatus in emplacement without a residual magma presence elsewhere (near the modern surface). In either case, the system had to remain hot—the Early Porphyries only cut high-temperature features such as quartz-feldspar veins, ductile fabrics and older dikes. They never cut early quartz-muscovite veins, albitization zones, or greisen veins.

Central Suite: Central Suite emplacement mimicked the Border Suite pattern. Relatively biotite-rich, muscovite-bearing granites outside the main contact (marked by USTs) were accompanied by early outward dipping, variably deformed quartz-feldspar veins. These units had several generations of veins and aplices superimposed (e.g., see figs. 12, 15, 16A in Barton, 2000). In Figure 12, the Central Suite is shown to include some fraction of these marginal biotite-rich granites, compatible with the petrologic discussion above. Mapping, however, has yet to reveal any abrupt break between this biotitic material and other parts of the Border Suite.
Central Suite fluids evolved in a way that first created iron-enriched quartz feldspar veins but ultimately led to late, iron-poor, fluorine-rich veins. This transition occurred concurrently with the local shift (near this contact) from pluton-centered to regionally-oriented stress. The issue of timing of albitionization and greisenization is vexing. They are always the latest features at any given location and could be late as a whole. Yet, when taken together, (1) their selective abundance in zones near other veins and dikes, and (2) the systematic rotation of vein orientations with time seem most consistent with local derivation of fluids. In other words, they probably formed in multiple episodes associated with the last gasps of fluid evolution from each stage of the intrusion.

Likewise, K feldspar megacrysts have no definitive timing features other than that they must predate consolidation of their host magma. Nonetheless, their overall marked concentrations at the base of dike swarms indicates that they may derive from the same overall process that releases fluids. Thus they may evidence the scale of fluid release and late-stage magma motion.

Summarizing Central Suite development: Inflation of the Central Suite chamber was accompanied by early fluid release from the water-saturated carapace (making the outward dipping quartz veins) and by sidewall crystallization that drove to magma to modestly more evolved compositions. Increasing overall water content accelerated vein formation and initiated radial diking. The rapid magmatic and hydrothermal evolution is recorded in the complex features comprising the 50- to 100 m-zone that lies immediately outside the main UST zone. The abundant megacrysts that are immediately inward of the UST zone might also record the same overall fluid accumulation and release. Catastrophic fluid release during diking terminated most local magma movement and thus local foliation development. Fluids continued to be released thereafter (from interstitial residual melt?—as seen in the sparse inward-dipping pegmatite dikes) and percolated outward to form the widespread fluorine-rich alteration in regional stress-controlled fracture systems.

Central Porphyry: Unlike the Early Porphyries, the Central Porphyry, which terminated the known magmatic history, generated abundant aplites and a prominent set of hydrothermal features. These are indicated by the third and highest gray bar in Figure 12. The sequence duplicates that seen in the Border and Central Suites except that ductile deformation is not an obvious part of Central Porphyry emplacement.

Aureole: As noted above for the Border Suite, in the inner aureole skarns and other metamorphic features can be tied directly to igneous history. The time-space diagram shows linkages for similar relationships with the Central Suite and Central Porphyry (see Trim, 1990). Timing of the distal assemblages is problematic. Their magmatic (+ metasedimentary rock) geochemical signature requires a close tie overall, but these assemblages may reflect the accumulated effect of fluids evolved through much of the magmatic history. Consequently, the external fluids are illustrated as having a protracted and largely undistinguished history. In some areas, proximity to the contact and field relationships indicate at least some of these formed during particular events (e.g., in the Camp Hill area on the west side; see descriptions above).

Concluding Remarks

Broader context

Late Cretaceous lithophile element hydrothermal systems are common across the Great Basin in the hinterland of the Cretaceous arc (Fig.1). These systems share many features, including an association with strongly peraluminous, generally two-mica granites and a distinctive style of fluorine and aluminum-rich hydrothermal alteration. The apparent lack of economically viable mineralization in the western United States examples has precluded serious interest by most economic geologists, although they inspired some exploration interest at least into the 1980s. In the broader context, these systems have many similarities to Late Mesozoic granitoids and their W-Sn-Mo etc. mineralization around the northern Pacific (Newberry, 1998). In this respect, their study may help others to better understand the origins of mineralized and “barren” systems (see below).

The structural and petrotectonic aspects have generated more recent activity. In the 1970s field and microstructural studies by Clem Nelson, Art Sylvester and others concluded that Birch Creek and the similar, but larger, Papoose Flat intrusion forcefully distended (“ballooned”) their country rocks during emplacement (Nelson and Sylvester, 1971; Nelson et al., 1978; Sylvester et al., 1978). More recently, the process of forceful emplacement has been challenged and reevaluated in many papers, a number of which have focused on structural geology at Papoose Flat (Paterson et al., 1991; Nyman et al., 1995; Paterson and Vernon, 1995; Morgan et al., 1998). What has not been done there (or elsewhere), in part because of the exposures, is to document in detail the magmatic history. In contrast, the field relationships seen in the smaller Birch Creek pluton unequivocally demonstrate that most or perhaps all of the deformation took place during emplacement. This conclusion was a serendipitous, but important result of detailed field studies driven by other questions.

Why isn’t Birch Creek economically mineralized?

In most studies of mineral deposits, the focus is on documenting economically interesting deposits and developing models that are consistent with their formation. In the case of Birch Creek, the system is not of economic interest. Nonetheless, it shares many features with well-mineralized systems worldwide. Better understanding of such non-ore-forming (“barren”) systems should help better understand the key characteristics and processes of better mineralized systems.

Birch Creek, like similar Great Basin systems, has a large hydrothermal system that contains large amounts of many commodities. Why does it lack ore? Is it a matter of exposure level? Birch Creek is likely eroded at least 1 km below the top of the pluton. Perhaps the volume having the most-focused fluid flow and thus the best mineralization has been removed by erosion (see Fig. 3). This could be a factor, although many major lithophile element-rich deposits are apparently eroded to comparable levels (e.g., Shizhuyuan, Hunan; Mao and Li, 1995; Mao et al., 1996). And, elsewhere in the Great Basin, shallow levels are preserved in similar systems (McCullough Butte and Mount Wheeler areas, see Fig. 1). The latter can be large (>200 Mt of F-Zn-Be-W-Mo mineralization at McCullough Butte) but they also are low grade except in
environments with focused traps (replacement zones containing ca. 1% BeO, 0.5% WO₃, 20% CaF₂ in thin carbonate horizon in clastic rocks at Mt. Wheeler). The metal inventories in these other systems are large; thus, it is not simply a lack of material.

More likely, the cause of differences is in the nature of the magmatic systems themselves. Although the albitites have some of the characteristics of W-Sn granites from other parts of the world (cf. Newberry, 1998), the main phases are not evolved. Is this the key? Maybe, but the large metal inventories in some Great Basin systems belie the need for highly specialized compositions. Alternatively, it may be that these plutons form from relatively water-rich melts (evidenced by continuous water release) and thus bleed volatiles in an unfoison manner while causing the systems to freeze at unfavorably great depths. This is a smaller factor controls the metallogenetic potential. Rather, it is the overall physical and chemical magmatic evolution—the consequence of many phenomena—that is ultimately responsible.

Acknowledgments

I thank John Dilles, Greg Ghidotti, Eric Jensen, and David Johnson for their helpful comments on this overview paper and on the field guide. I am particularly grateful to Clem Thompson for last minute heroics in helping this be included. Many people have contributed to the results presented here, most are acknowledged above. Special thanks are due Clem Nelson who originally introduced me to the area and generously provided unpublished regional geologic mapping. Funding for field and lab work was provided by the National Science Foundation (86-07542, EAR 89-03773) and the White Mountain Research Station (University of California).

REFERENCES


