Mesozoic magmatism and deformation in the northern Owyhee Mountains, Idaho: Implications for along-zone variations for the western Idaho shear zone

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ABSTRACT

The northern Owyhee Mountains of southwestern Idaho contain granitoid rocks that are the same age as the Cretaceous western border zone of the Idaho batholith to the north of the Snake River Plain. They contain a well-developed and consistently oriented 020° foliation, zircon yielding U-Pb dates of ca. 160–48 Ma, and initial 87Sr/86Sr isotopic compositions that show a steep west-to-east transition in values from 0.704 to 0.708 over a distance of ~30 km. The rocks of the northern Owyhee Mountains are interpreted to be the southward continuation (Owyhee segment) of the western Idaho shear zone. Similar to a well-studied section of the western Idaho shear zone by McCall (McCall segment), the Owyhee segment displays steep foliation and lineation orientations, deformation of 98–90 Ma plutons, steep Sr isotopic gradients, and syntectonic tonalite intrusions. However, the Owyhee segment has three major differences from the McCall segment: (1) significantly less well-developed solid-state strain fabric foliations; (2) trend of 020° rather than 000°; and (3) a wider transition zone in initial Sr ratios from 0.704 to 0.708. We present a simple tectonic model to explain these differences, assuming a 20° along-zone difference in the initial orientation of the western margin of the Laurentia, a rigid-body collision, homogeneous material behavior, and transpressional kinematics. For the Owyhee segment, the model predicts a lower oblique-convergence angle, less convergent displacement, more dextral transcurrent displacement, and an overall lower finite strain relative to the McCall segment.

INTRODUCTION

Along-zone variations in the strike orientation of major deformation zones are typically accompanied by differences in structural characteristics (e.g., McCaffrey, 1991; Avé Lallemant and Guth, 1990). For example, in the San Andreas fault system, California, both the Transverse Ranges and the Ridge Basin formed as a result of along-zone variations in fault-zone strike orientation (Crowell, 1975; Kellogg and Minor, 2005). In the roots of exhumed ancient orogens, along-zone variations in orientation result in distinctive foliation and lineation patterns in different shear-zone segments (Lin and Jiang, 2001). These along-zone variations in structural characteristics develop because a distinctive deformation field (e.g., relative percentages of pure and simple shear components) characterizes each segment of the deformation zone. The orientation of the deformation zone boundaries relative to the regional displacement field, in turn, controls the strain path. As a result, along-zone variations in both orientations and structural characteristics are particularly relevant for the study of vertical deformation zones having large horizontal displacements (e.g., wrench and transpressional zones).

In neotectonic settings, displacement can be quantified using geodesy or other techniques, as was done along the San Andreas fault system (e.g., Segall, 2002; Titus et al., 2006). However, characterization of along-zone variations in orientation and structural style is difficult for the deep crustal levels commonly exposed in ancient orogens. Displacement in these orogens commonly must be calculated using palinspastic reconstructions or integration of regional-scale strain data. Palinspastic reconstructions are particularly difficult along terrane boundaries where few to no piercing points exist to calculate displacement.

This paper addresses the possible southward continuation of the lithospheric-scale western Idaho shear zone and attempts to quantify the finite and infinitesimal strains of the western Idaho shear zone by focusing upon the along-zone differences in orientation and in structural characteristics. The western Idaho shear zone overprints the Salmon River suture zone, which is the north-south–trending boundary between the western continental margin of Laurentia and accreted terranes (e.g., Fleck and Criss, 1985, 2007; Lund, 1988; Lund and Snee, 1988; Strayer et al., 1989; Manduca et al., 1992). Steep gradients in strontium (Armstrong et al., 1977; Manduca et al., 1992) and oxygen (Criss and Fleck, 1985; Fleck and Criss, 1985; Manduca et al., 1992; King et al., 2007) isotopic composition characterize the western Idaho shear zone, a 2–10-km-wide zone of amphibolite-facies deformation in midcrustal plutons (Zen and Hammarnstrom, 1984; Lund and Snee, 1988; Manduca et al., 1993). North-south–striking fabric associated with a steep east dip and a downdip mineral lineation characterizes the western Idaho shear zone in west-central Idaho. Taubeneck (1971) noted that similar structural fabrics are present in the Owyhee Mountains of southwest Idaho. This paper provides detailed documentation of lithology and fabric in the northern Owyhee Mountains, corroborating the reconnaissance mapping of Taubeneck (1971), and U-Pb zircon dates from seven plutons. Details for the Owyhee segment are compared to similar data from the McCall segment, 120 km north (Fig. 1A). Relative to the McCall segment of the western Idaho shear zone, the shear zone in the northern Owyhee Mountains, known as the Owyhee segment, differs in strike orientation, amount of deformation, and width. The greater width of the Owyhee segment also corresponds to a less steep gradient in strontium isotope
values. These differences can be explained by an along-zone variation in the orientation of the western Idaho shear zone. A model for the orientation of the Owyhee segment predicts a smaller oblique-convergence angle relative to the McCall segment, which correlates with structural characteristics involving a smaller pure shear component, less finite strain, and wider shear zone boundaries.

GEOLOGIC HISTORY OF THE WESTERN IDAHO SHEAR ZONE

Salmon River Suture Zone

The Jurassic-Cretaceous, north-south–striking Salmon River suture zone of west-central Idaho (Fig. 1A) is a critical geological structure for understanding the development of the U.S. Cordillera. The Salmon River suture zone records juxtaposition, prior to 128 ± 3 Ma (Getty et al., 1993), of island-arc terranes of the Blue Mountain Province with continental rocks of the ancestral western Laurentia (Hamilton, 1963; Lund, 1988; Lund and Snee, 1988; Selverstone et al., 1992).

Rocks in the western part of the suture zone have island-arc affinities and consist mainly of...
volcanic, metavolcanic, and metavolcaniclastic rocks of the Blue Mountains Province (Hamilton, 1963; Snee et al., 1995; McClelland et al., 2000). Rocks in the eastern part are primarily metasedimentary, continental rocks that correlate with Belt Supergroup and Late Proterozoic or Paleozoic sedimentary units (Lund, 1988). Intrusion of younger plutons and deformation and metamorphism associated with the Late Cretaceous western Idaho shear zone obscure the precise location of the contact between the volcanic-arc rocks and the continental rocks, but it is taken as the location of the initial strontium 0.704–0.706 isopleth in plutonic rocks across the region (Fleck and Criss, 1985, 2007; Lund and Snee, 1988; Manduca et al., 1992; Snee et al., 1995).

Western Idaho Shear Zone

A zone of high strain, the western Idaho shear zone defines the present boundary between the accreted volcanic arc terranes and the western margin of Laurentia. The shear zone formed syn- to postmagmatically with, and is oriented parallel to, the highly tabular intrusions and intrusive suites that obscure the arc-continent boundary. Taubeneck (1971) mapped these intrusive bodies as part of the “western gneissic border zone” of the Idaho batholith. The mapped western Idaho shear zone is a midcrustal exposure of the Late Cretaceous intra-arc shear zone within the U.S. Cordillera of North America (Manduca et al., 1993; McClelland et al., 2000). The <6-km-wide shear zone overprints the eastern portion of the Jurassic–Early Cretaceous Salmon River belt within the Salmon River suture zone (Manduca et al., 1993; Tikoff et al., 2001; Gray and Oldow, 2005). The western Idaho shear zone overprinting the Salmon River belt resulted in a dramatic shortening of the continental boundary and a modification of the Salmon River suture zone between accreted terranes and continental lithosphere (Giorgis et al., 2005a). Based on fabric and U-Pb zircon dating, the western Idaho shear zone deformation began after ca. 105 Ma and had ceased by ca. 90 Ma (Giorgis et al., 2008). The Salmon River suture zone and the western Idaho shear zone span from Clearwater, Idaho, south through McCall, Idaho, before disappearing beneath the western Snake River Plain (Fig. 1A).

Previous work that focused on the <6-km-wide McCall segment of the western Idaho shear zone indicates the presence of three tabular map units that become more felsic and younger from west to east (Fig. 1B). The westernmost intrusion, the Hazard Creek complex, consists of two main compositions: a coarse-grained, tonalitic hornblende orthogneiss and a biotite granite. The Hazard Creek complex has a U-Pb zircon date of 118 ± 5 Ma and 114.4 ± 2 Ma (Manduca et al., 1993; Unruh et al., 2008). The main porphyritic granodioritic orthogneis of the Little Goose Creek complex, to the east, was emplaced at 105.2 ± 1.5 Ma (Manduca et al., 1993; Giorgis et al., 2008; Unruh et al., 2008). Coarse-grained hornblende and the presence of titanite characterize the easternmost and youngest unit, the Payette River tonalite. This unit is dated at 91.5 ± 1.1–89.7 ± 1.2 Ma (Giorgis et al., 2008; Unruh et al., 2008).

Across the entire Salmon River suture, there are steep gradients in initial 87Sr/86Sr ratios—(87Sr/86Sr)i—from plutons in west-central Idaho that correspond spatially with the McCall segment of the western Idaho shear zone. Rocks in the western part of the high strain zone have (87Sr/86Sr)i ≤ 0.704, whereas rocks in its eastern part have (87Sr/86Sr)i ≥ 0.706 (Armstrong et al., 1977). The low (87Sr/86Sr)i in the west part of the western Idaho shear zone indicates little to no continental crustal contamination, whereas the higher (87Sr/86Sr)i value in its eastern part suggests significant continental crust interaction (Fleck and Criss, 1985; Manduca et al., 1992). The transition in (87Sr/86Sr)i values delineates the arc-continent boundary between the Blue Mountain Province accreted terranes and the western continental margin of ancestral Laurentia.

In the McCall segment, the western Idaho shear zone records transpressional kinematics, displaying dextral shear-sense indicators on lineation-normal foliation outcrop surfaces (Giorgis et al., 2004, 2006b). The structural fabric of the western Idaho shear zone in this region is generally well developed. Foliation strikes north-south and dips steeply to the east and contains a prominent, downdip stretching lineation (Manduca et al., 1993; Giorgis et al., 2004). In the McCall segment, younger normal faults have been used to restore the pre-Miocene surface and determine an original subvertical foliation and lineation (Tikoff et al., 2001; Giorgis et al., 2006a). These solid-state fabric elements are best developed within the porphyritic Little Goose Creek complex, but they also occur in the eastern Hazard Creek complex and western Payette River tonalite. Based on the orientation of the foliation, stretching lineation, and feldspar porphyroclast populations (Giorgis and Tikoff, 2004), the western Idaho shear zone is interpreted to have experienced a minimum of 45 km of dextral transcurrent displacement (McClelland et al., 2000; Tikoff et al., 2001; Giorgis and Tikoff, 2004). The McCall segment was also interpreted to have experienced an angle of oblique arc-continent convergence (α) between 45° and 75° (Dair and Giorgis, 2006; Giorgis et al., 2006b).

We compare deformation in the Owyhee Mountains with deformation in the McCall segment of the western Idaho shear zone because the strain and kinematics are well studied in this locale (e.g., Manduca et al., 1993; Giorgis and Tikoff, 2004). The western Idaho shear zone continues both northward and southward from these areas. To the north, the Klopton Creek–Hammer Creek–Mount Idaho deformation zone (near Grangeville, Idaho; Fig. 1A) exhibits two phases of post–western Idaho shear zone deformation, which locally affect the orientation of the shear zone (Schmidt and Lewis, 2007). Farther north, the younger Orofino shear zone cuts across the western Idaho shear zone (McClelland and Oldow, 2007). South of the Idaho-Nevada border, there is wider separation between the 0.704 and 0.708 (87Sr/86Sr)i isopleths recorded by Mesozoic granitic rocks (e.g., Kistler and Peterman, 1978; Lund et al., 2008). Wyld and Wright (2001) document deformation in the western Nevada shear zone in NW Nevada, which they suggest may link to the western Idaho shear zone. The western Nevada shear zone consists of a series of three distinct shear zones, all of which record dip-slip shear sense. Two of the shear zones were active during the Early Cretaceous, and the youngest (Pueblo shear zone) has a cooling age (Ar/Ar on biotite) of 95 Ma, making them all older than the youngest deformation on the western Idaho shear zone (e.g., Giorgis et al., 2008). Given that the western Nevada shear zone has different kinematics, is older, and is located significantly west of the 0.704 (87Sr/86Sr)i isopleth, it is not a useful comparison for our study.

West-central Idaho is the only documented location in the U.S. Cordillera where such a large change in (87Sr/86Sr)i values occurs over such a small distance (Armstrong et al., 1977; Fleck and Criss, 1985; Manduca et al., 1993). This large variation suggests that the gradient is: (1) a relict of the steepness of the arc-continent boundary within the Salmon River suture zone prior to western Idaho shear zone deformation, or (2) a result of Late Cretaceous western Idaho shear zone deformation superposed on the Salmon River suture zone (Giorgis et al., 2005a) support the latter hypothesis based on the exact spatial correlation between the steep isotopic gradient and the solid-state fabrics in the shear zone. Between 30 and 110 km of shortening is estimated to have occurred during western Idaho shear zone transpression (Giorgis et al., 2005b).
direction and 40 km wide east to west. The mountain range consists of Jurassic and Cretaceous intrusive rocks and Tertiary sedimentary and extrusive rocks (Ekren et al., 1981). Cretaceous granitoids crop out in a series of elongate north-south exposures in the northern part of the Owyhee Mountains. In a geochemical study of southwestern Idaho, Norman and Leeman (1989) documented granodiorite as the dominant lithology in the northern Owyhee Mountains with lesser amounts of tonalite, quartz monzonite, and quartz diorite. Taube-neck (1971) referred to the granitoids of the northern Owyhee Mountains as outliers of the Idaho batholith and part of the “western gneissic border” because of the general orientation of their foliation and the presence of magmatic epidote (Zen and Hammarstrom, 1984). Tertiary basalt flows, sedimentary rocks, and normal faults separate the isolated granitoid exposures (Fig. 2).

Geochronological and geochemical analysis by Armstrong and colleagues (Armstrong, 1975; Armstrong et al., 1977) strengthened the link between the exposed plutons of the northern Owyhee Mountains and the main part of the Idaho batholith. Whole-rock geochronological analyses of granitoid rocks of the northern Owyhee Mountains indicates K-Ar cooling dates of 87–60 Ma that display an eastward decrease in ages (Armstrong, 1975). These dates correlate well with 40Ar/39Ar dates for the McCall segment of the western Idaho shear zone (e.g., Snee et al., 1995; Giorgis et al., 2008).

INTRUSIONS OF THE NORTHERN OWYHEE MOUNTAINS

The present study in the northern Owyhee Mountains focuses on identifying major lithologic units, understanding timing of emplacement and deformation of granitoids, and constraining its Late Cretaceous geologic history. Prior to this study, descriptions of the Late Cretaceous plutonic rocks in the northern Owyhee Mountains were limited, and lithologies were not clearly separated.

The northern Owyhee Mountains consist of at least two compositionally homogeneous intrusions and two heterogeneous intrusive suites. In the northern Owyhee Mountains, from west to east, the intrusions and intrusive suites are the Chipmunk Meadow Intrusive Suite, the Whiskey Ridge tonalite, the Dropoff Intrusive Suite, and the Wilson Peak granodiorite (Fig. 2). Brief descriptions of lithology and foliation and lineation fabric elements follow; more information on composition is provided in Benford (2007).
Chipmunk Meadow Intrusive Suite

The westernmost Chipmunk Meadow Intrusive Suite is exposed over ~18 km² and includes granite, granodiorite, and tonalite (Table 1; Fig. 3). Trace minerals include apatite, allanite, zircon, and Fe-Ti oxides, and secondary chlorite and sericite. The suite is an orthogneiss, with ribbon quartz and K-feldspar porphyroclasts defining the planar fabric. The average field-measured foliation is oriented 006°, 79°E, and the stretching lineation pitches 77°N (Fig. 4A). Lineation consistently pitches steeply north throughout the suite, and foliation strikes between 310° and 020°. In the western part of the suite, the strike of foliation has greater variability (Fig. 2).

Whiskey Ridge Tonalite

The Whiskey Ridge tonalite is exposed over ~25 km² area east of the Chipmunk Meadow Intrusive Suite (Fig. 2), and is compositionally homogeneous in comparison to the other deformed units in the northern Owyhee Mountains (Fig. 3). The tonalite contains quartz, plagioclase, K-feldspar, hornblende, and biotite. It also contains minor amounts of epidote, titanite, apatite, zircon, and Fe-Ti oxides (primary), as well as chlorite (secondary) (Table 1). The epidote gives no indication of replacement of plagioclase and thus is interpreted as magmatic. The Whiskey Ridge tonalite contains dominantly magmatic fabric, although solid-state fabric occurs locally, such as by the contact with the Chipmunk Meadow Intrusive Suite. The average foliation is oriented 009°, 75°E, and lineation pitches 82°N (Fig. 4B). Foliation orientation is consistent throughout the suite, and lineation varies slightly.

The contact between the Whiskey Ridge tonalite and the Chipmunk Meadow Intrusive Suite is not exposed, but outcrops of each occur 5 m apart in one locality. There is no evidence for faulting, such as cataclasite, fault breccia, or fault gouge, or an increase in fractures relative to the surrounding outcrops, and thus the contact is interpreted to be intrusive.

Dropoff Intrusive Suite

The Dropoff Intrusive Suite is exposed over ~15 km² area east of the Whiskey Ridge tonalite and is compositionally heterogeneous and similar to the Chipmunk Meadow Intrusive Suite (Table 1; Fig. 3). Titanite, apatite, zircon, and Fe-Ti oxides are primary minerals; sericite and chlorite are secondary minerals. Epidote was observed in one location. The suite is an orthogneiss, with ribbon quartz and K-feldspar porphyroclasts defining the fabric. The average foliation is oriented 007°, 81°E, and the lineation pitches 75°N, as defined by biotite clots (Fig. 4C). Foliation consistently dips steeply east, whereas the strike varies as much as ±30°. The best developed foliations in the suite strike 020°.

The nature of the contact between the Dropoff Intrusive Suite and the Whiskey Ridge tonalite is unknown due to the lack of continuous outcrop. However, in the center of the map area (43°08′N), there is no structural or geomorphic evidence for a fault where the contact is limited to a <30-m-wide zone. The two units are considered separate based on differences in mineralogy and modal composition: the Whiskey Ridge tonalite contains hornblende and lacks large K-feldspar porphyroclasts and sericite.

Normal faults are inferred to separate the deformed units in the northern Owyhee Mountains. The composition of the Dropoff Intrusive Suite is locally similar to the Wilson Peak granodiorite, but it contains more K-feldspar and is compositionally heterogeneous, suggesting the two units are separate intrusions.

Wilson Peak Granodiorite

The 55 km² Wilson Peak granodiorite is primarily located in the northeastern Owyhee Mountains (Fig. 2) and consists of a single lithologic unit (Table 1). Titanite, allanite, epidote, apatite, zircon, and Fe-Ti oxides are primary minerals, and sericite and chlorite occur rarely as secondary minerals (Table 1). Crosscutting aplitic and pegmatitic dikes are common, especially in the northeastern part of the granodiorite. The Wilson Peak granodiorite has subparallel magmatic and solid-state fabrics (Fig. 5), with the magmatic fabric being dominant. Foliation is oriented 022°, 87°E, with lineation pitching 88°S (Fig. 4D). Lineation, defined by the alignment of biotite and Fe-Ti oxide minerals, occurs in both fabric types. A weak magmatic fabric is characteristic. Fabric at the thin-section scale is defined primarily by the alignment of biotite. The magmatic fabric is characterized by undeformed K-feldspar phenocrysts and

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**TABLE 1. MODAL COMPOSITION OF DIFFERENT UNITS DETERMINED FROM STAINED ROCK SLABS**

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Chipmunk Meadow Intrusive Suite</th>
<th>Whiskey Ridge tonalite</th>
<th>Dropoff Intrusive Suite</th>
<th>Wilson Peak granodiorite</th>
<th>Hardtrigger orthogneiss</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartz</td>
<td>34</td>
<td>26</td>
<td>32</td>
<td>34</td>
<td>31</td>
</tr>
<tr>
<td>K-feldspar</td>
<td>18</td>
<td>2</td>
<td>15</td>
<td>13</td>
<td>22</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>46</td>
<td>59</td>
<td>51</td>
<td>50</td>
<td>43</td>
</tr>
<tr>
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<td>2</td>
<td>7</td>
<td>2</td>
<td>3</td>
<td>2</td>
</tr>
<tr>
<td>Hornblende</td>
<td>–</td>
<td>6</td>
<td>–</td>
<td>–</td>
<td>&lt;1</td>
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<tr>
<td>Zircon</td>
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<tr>
<td>Fe-Ti oxides</td>
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<tr>
<td>Sericite</td>
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</tbody>
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**Figure 3. Intrusive rocks of the northern Owyhee Mountains plotted on International Union of Geological Sciences alkali feldspar (A)–quartz (Q)–plagioclase (P)–ternary diagram (modified from Le Maitre et al., 1989). Key denotes intrusive suites and intrusions of the northern Owyhee Mountains.**
no alignment of quartz crystals (Fig. 5). Some quartz crystals contain chessboard patterns, and some myrmekite occurs at the boundaries of K-feldspar phenocrysts.

The Wilson Peak granodiorite forms a non-conformable contact with Miocene basalt flows on Soldier Cap Mountain, south of Wilson Peak. Based on the 16°W dip of the overlying basalt units, the granodiorite can be partially restored to its pre-Miocene orientation. The average foliation for the granodiorite is 022°, 79°W; an aligned biotite lineation pitches 88°S (Fig. 6).

**Hardtrigger Orthogneiss**

An ~1 km² area of granodiorite orthogneiss occurs near Hardtrigger Creek (Fig. 7). It may be part of the Wilson Peak granodiorite due to the similarity in lithology, composition, and location, but the relationship between the units is unknown. The Hardtrigger orthogneiss typically contains more K-feldspar and lacks the homogeneous composition of the Wilson Peak granodiorite. Titanite, epidote, allanite, apatite, zircon, and opaques are primary trace minerals, and chlorite and white mica are secondary minerals. Solid-state fabrics are oriented 017, 88°E, with a downdip mineral lineation (Fig. 4E). Elongate, polycrystalline mineralogic domains define fabric orientation. Hornblende laths and biotite grains define lineation, where present. Locally, the gneissic fabric is isoclinally folded.

**Microstructures**

The microstructures in the Chipmunk Meadow Intrusive Suite are similar to those of the Whiskey Ridge tonalite, the Dropoff Intrusive Suite, and the Hardtrigger orthogneiss, and they are described next for all intrusions and suites together. High-temperature and low-temperature solid-state fabrics occur in all intrusive units. Weak alignment of mafic phases, chessboard patterns, and elongation parallel to foliation of quartz grains characterize the high-temperature solid-state fabric. Quartz crystals are coarse (~1.5–2.0 mm along the long axis) and have lobate boundaries, suggesting the presence of melt (e.g., Sawyer, 2001). In some areas of the Chipmunk Meadow Intrusive Suite, and more commonly in the Hardtrigger orthogneiss, submagmatic fractures occur within plagioclase grains, indicating deformation while melt was still present (Fig. 5) (Bouchez et al., 1992). In the Hardtrigger orthogneiss and Dropoff Intrusive Suite, myrmekite is common around the edges of K-feldspar porphyroclasts (Fig. 5). Well-aligned biotite, local hornblende, and quartz crystals characterize the low-temperature solid-state fabric. The low-temperature fabrics consist of a bimodal distribution of medium- (~0.3 mm along the long axis) and coarse-size quartz crystals (~1.5 mm along the long axis).

Microstructurally, the Hardtrigger area is a blastomylonite, with high-temperature solid-state microstructures (Fig. 5). Biotite and Fe-Ti oxide phases show alignment, quartz crystals have chessboard patterns and are elongate parallel to foliation, and plagioclase feldspar grains commonly contain subcracking but are otherwise undeformed. Additionally, “globular” quartz occurs near the plagioclase, indicating the
presence of the residual melt during formation of the submagmatic fractures (Bouchez et al., 1992).

FABRIC ANALYSES

The degree of fabric development in the northern Owyhee Mountains was measured using shape-preferred orientation (SPO) analyses that can be used to evaluate shear-zone kinematics and quantitatively define fabric strength (e.g., Launeau et al., 1990; Ikeda et al., 2000; Ohtani et al., 2001; Titus et al., 2005). Using this type of analysis, the fabric ellipsoid allows the minimum amount of finite strain to be constrained.

Shape-preferred orientation analyses were performed on the intrusive units to compare quantitatively the SPO fabric with the field fabric—the foliation and lineation measured in the field. Samples were cut on three mutually perpendicular oriented faces, stained, and analyzed using image analysis software. Face size varied from 17 to 150 cm². Samples were stained with sodium cobaltinitrate and amaranth to differentiate between K-feldspar and plagioclase, respectively, for image analysis. Several parameters were calculated, including: (1) a three-dimensional fabric ellipsoid (SPO), (2) P, the degree of anisotropy or the measure of deviation from a sphere (Jelinek, 1981), and (3) T, the shape parameter or the ellipsoid shape (i.e., prolate, T = −1; oblate, T = 1) (Jelinek, 1981).

Three minerals (K-feldspar, plagioclase, and quartz) and mafic phases (biotite, hornblende, and Fe-Ti oxides) for each of the 10 samples...
were measured. Table 2 summarizes the results of the analyses. All samples show a weak fabric with \( P_j \) not exceeding 1.320. Similar \( P_j \) values of the analyses. All samples show a weak fabric were measured. Table 2 summarizes the results with solid-state deformation areas and for those containing only magmatic fabrics. For all phases, the value of \( T \) varies between \(-0.787 \) and 0.930 (Table 2; Fig. 8). There is no consistent relationship for oblate and prolate fabric shape within the intrusive units. No consistent agreement exists between field fabric and the SPO fabric of the individual phases (Fig. 9). Phases show agreement in orientation with field fabric at some sites, but no single mineral orientation consistently agrees or disagrees with field fabric. The single sample from the Hardtrigger orthogneiss is unlike any of the other samples.

All phases have \( P_j \) values \( \geq 1.114 \), plot in the oblate field, have consistent orientations of solid-state fabric, and agree with field fabric (Fig. 8). The values of \( P_j \) range from 1.114 for K-feldspar to 1.197 for quartz. The value of \( T \) for K-feldspar (the inferred rigid marker) is 0.194, whereas this parameter is nearly double that in the other phases (Table 2).

**Interpretation**

The inconsistent agreement between SPO fabric orientations for each phase and field-fabric orientations suggests that none of the individual phases accurately characterizes fabric in the Owyhee segment. The SPO of the rigid (K-feldspar) and passive (quartz) markers behave similarly, suggesting little ductile deformation.

Most SPO analyses coincide at any site with mesoscopic fabric orientation (Fig. 9), suggesting that the fabric analysis technique is recording the fabric of the Owyhee segment. This relationship also implies that, in different areas, different phases record the fabric. The mean orientation for the SPO fabric ellipsoid of all of the phases yields a foliation of 022°, 75°E, and a lineation (the long axis of the ellipsoid) pitching 65°S. The intermediate axis forms a girdle within the foliation plane, whereas the minimum axis has a subhorizontal plunge and a wide range of trends (Fig. 10). This result is consistent with SPO fabric analyses of K-feldspar porphyroclasts performed along the McCall segment in the Little Goose Creek complex porphyritic orthogneiss (Giorgis and Tikoff, 2004).

**GEOCHRONOLOGY**

**U-Pb Geochronology Methods**

U-Pb geochronology was conducted to determine the timing of the igneous intrusions in the Owyhee Mountains. Mapping and fabric analyses focused on the deformed intrusions. In order to put the northern Owyhee Mountains in a regional context, samples were collected at other locations of intrusive igneous rocks in the Owyhee Mountains. These samples include two samples (SM06-1 and SM06-109) east and two samples (SM06-97 and SM06-98) southwest of the deformed northern Owyhee Mountains (Fig. 11). Zircon crystals were separated from granitoid samples, mounted in epoxy, polished until the centers were exposed, and imaged with a scanning electron microscope cathodoluminescence (CL) detector at the University of Wisconsin–Madison. The CL images guided the selection of crystals for dating; crystals having obvious inherited cores or that were entirely inherited were avoided. All imaged crystals have sector and oscillatory zoning that is typical of zircon crystallized in granitoid magmas. Analyses were conducted upon single crystals extracted from the mounts, except for a crystal from one sample (SM06-109) from the eastern Owyhee Mountains, which was broken into three fragments with a steel tool and the fragments analyzed separately.

Zircon was subjected to a modified version of the isotope dilution–thermal ionization mass spectrometry (ID-TIMS) chemical abrasion method of Mattinson (2005), reflecting a preference to prepare and analyze carefully selected single crystals. Crystals were placed in a muffle furnace at 900 °C for 60 h in quartz beakers, transferred to individual 3 mL Teflon beakers, extracted from the mounts, and isolated by hand-pick. The isolated crystals were mounted in epoxy, polished until the centers were exposed, and irradiated for 7 h at a fluence of 4 × 10¹⁷ neutrons cm⁻² in the reactor at the Los Alamos National Laboratory. The irradiated zircon crystals were mounted in epoxy, polished until the centers were exposed, and imaged with a scanning electron microscope cathodoluminescence (CL) detector at the University of Wisconsin–Madison. The CL images guided the selection of crystals for dating; crystals having obvious inherited cores or that were entirely inherited were avoided. All imaged crystals have sector and oscillatory zoning that is typical of zircon crystallized in granitoid magmas. Analyses were conducted upon single crystals extracted from the mounts, except for a crystal from one sample (SM06-109) from the eastern Owyhee Mountains, which was broken into three fragments with a steel tool and the fragments analyzed separately.

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TABLE 2. Pj AND T VALUES FOR EACH OF THE SITES SAMPLED FOR SHAPE-PREFERRED ORIENTATION (SPO) ANALYSES

<table>
<thead>
<tr>
<th>Site</th>
<th>Unit*</th>
<th>K-feldspar</th>
<th>Quartz</th>
<th>Plagioclase</th>
<th>Other phases†</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Pj</td>
<td>T</td>
<td>Pj</td>
<td>T</td>
</tr>
<tr>
<td>SM06-01</td>
<td>Unnamed intrusion (s)</td>
<td>1.098</td>
<td>–0.289</td>
<td>1.139</td>
<td>0.930</td>
</tr>
<tr>
<td>SM06-08</td>
<td>Dropoff Intrusive Suite (s)</td>
<td>1.147</td>
<td>–0.276</td>
<td>1.320</td>
<td>0.596</td>
</tr>
<tr>
<td>SM06-70</td>
<td>Chipmunk Meadow Intrusive Suite (s)</td>
<td>1.101</td>
<td>0.103</td>
<td>1.116</td>
<td>–0.048</td>
</tr>
<tr>
<td>SM06-86</td>
<td>Hardtrigger orthogneiss (s)</td>
<td>1.114</td>
<td>0.194</td>
<td>1.197</td>
<td>0.374</td>
</tr>
<tr>
<td>SM06-90</td>
<td>Chipmunk Meadow Intrusive Suite (s)</td>
<td>1.041</td>
<td>–0.702</td>
<td>1.093</td>
<td>0.425</td>
</tr>
<tr>
<td>SM06-94</td>
<td>Wilson Peak granodiorite (m)</td>
<td>1.088</td>
<td>–0.498</td>
<td>1.102</td>
<td>0.491</td>
</tr>
<tr>
<td>SM06-105</td>
<td>Whiskey Ridge tonalite (s)</td>
<td>1.079</td>
<td>0.685</td>
<td>1.128</td>
<td>0.184</td>
</tr>
<tr>
<td>SM06-106</td>
<td>Dropoff Intrusive Suite (s)</td>
<td>1.133</td>
<td>–0.466</td>
<td>1.073</td>
<td>0.307</td>
</tr>
<tr>
<td>SM06-116S</td>
<td>Chipmunk Meadow Intrusive Suite (s)</td>
<td>1.111</td>
<td>0.199</td>
<td>1.075</td>
<td>0.339</td>
</tr>
<tr>
<td>SM06-116W</td>
<td>Whiskey Ridge tonalite (s)</td>
<td>1.062</td>
<td>0.542</td>
<td>1.073</td>
<td>–0.785</td>
</tr>
<tr>
<td>Solid-state fabric mean</td>
<td>1.088</td>
<td>–0.498</td>
<td>1.102</td>
<td>0.491</td>
<td>1.084</td>
</tr>
<tr>
<td>Magmatic fabric mean</td>
<td>1.098</td>
<td>–0.001</td>
<td>1.112</td>
<td>0.174</td>
<td>1.120</td>
</tr>
</tbody>
</table>

*s—solid-state fabric; m—magmatic fabric.
†Biotite, hornblende, and/or Fe-Ti oxides, depending on unit.

EXPLANATION

- ▲ Chipmunk Meadow Intrusive Suite
- ■ Whiskey Ridge tonalite
- ● Dropoff Intrusive Suite
- □ Wilson Peak granodiorite
- △ Hardtrigger orthogneiss
- ○ Unnamed intrusion

Figure 8. Hsu plots showing degree of anisotropy versus shape factor based on shape-preferred orientation (SPO) analyses for (A) K-feldspar, (B) plagioclase, (C) quartz, and (D) biotite, hornblende, and Fe-Ti oxide minerals. Degree of anisotropy is generally low in all phases. Samples having a positive shape factor are an oblate shape, and those with a negative value are a prolate shape. An outlier in SPO of biotite, hornblende, and Fe-Ti oxide minerals is from site SM06–70 in the Chipmunk Meadow Intrusive Suite (D).
Figure 9. Lower-hemisphere, equal-area net plots of orientations of shape-preferred orientation (SPO) fabrics compared to field fabrics. Oblate fabric is plotted as solid lines, and prolate fabric is plotted as dashed lines. Thickness of great circles is based on degree of anisotropy. Site for each plot is in the lower right corner. (A–C) Chipmunk Meadow Intrusive Suite, (D–E) Whiskey Ridge tonalite, (F–G) Dropoff Intrusive Suite, (H) Wilson Peak granodiorite, (I) Hardtrigger orthogneiss, and (J) unnamed intrusion (SM06-1) located southeast of the Wilson Peak granodiorite.
Perfluoralkoxy (PFA) beakers containing ultrapure \( \text{H}_2\text{O} \), and then loaded into 300 \( \mu \)L Teflon PFA dissolution microcapsules. Fifteen microcapsules were placed in a large-capacity Parr vessel, and the crystals were partially dissolved in 120 \( \mu \)L of 29 \( \% \) HF for 10–12 h at 180 °C. The contents of each microcapsule were then returned to 3 mL Teflon PFA beakers, and the residual crystals were rinsed in ultrapure \( \text{H}_2\text{O} \), immersed in 3.5 \( M \) \( \text{HNO}_3 \), ultrasonically cleaned for an hour, and flushed on a hotplate at 80 °C for 1 h. The \( \text{HNO}_3 \) was removed, and the crystals were rinsed several times using ultrapure \( \text{H}_2\text{O} \) before being reloaded into the same 300 \( \mu \)L Teflon PFA dissolution microcapsules (rinsed and fluxed in 6 \( M \) HCl during crystal sonication and washing) and spiked with the Boise State University mixed \( ^{233}\text{U}^{238}\text{U}^{206}\text{Pb} \) tracer solution. The crystals were dissolved in Parr vessels in 120 \( \mu \)L of 29 \( \% \) HF having a trace of 3.5 \( M \) HNO\(_3\), at 220 °C for 48 h, dried to fluorides, and then redissolved in 6 \( M \) HCl at 180 °C for ~12 h. U and Pb were separated from the zircon matrix using an HCl-based anion-exchange chromatographic procedure (Krogh, 1973), eluted together, and dried with 2 \( \mu \)L of 0.05 \( N \) \( \text{H}_3\text{PO}_4 \).

Pb and U were loaded on a single outgassed Re filament in 2 \( \mu \)L of a silica-gel–phosphoric acid mixture (Gerstenberger and Haase, 1997), and U and Pb isotopic measurements were conducted on an IsotopX Isotope-Probe-T multicollector thermal ionization–mass spectrometer equipped with an ion-counting Daly detector. Pb isotopes were measured by peak-jumping all isotopes on the Daly detector for 100–150 cycles, and corrected for 0.22 ± 0.04%/a.m.u. (atomic mass unit) mass fractionation. Transitory isobaric interferences due to high-molecular-weight organics, particularly on \(^{206}\text{Pb}\) and \(^{207}\text{Pb}\), disappeared within ~30 cycles, while ionization efficiency averaged 10^4 cycles per second/picogram (cps/pg) of each Pb isotope. Linearity (to ~1.4 × 10^6 cps) and the associated dead time correction of the Daly detector were monitored by repeated analyses of NBS982 and have been constant since installation of the instrument. Uranium was analyzed as \( \text{UO}_2^+ \) ions in static Faraday mode with 10^4 ohm resistors for 150–200 cycles, and was corrected for isobaric interference of \(^{210}\text{U}^{18}\text{O}^{16}\text{O} \) on \(^{208}\text{U}^{16}\text{O}^{16}\text{O} \) with an \( ^{18}\text{O}^{16}\text{O} / ^{16}\text{O} \) of 0.00205. Ionization efficiency averaged 20 mV/ng of each U isotope. U mass fractionation was corrected using the known \(^{233}\text{U}/^{238}\text{U} \) of the tracer solution.

U-Pb dates and uncertainties were calculated using the algorithms of Schmitz and Schoene (2007) and the U decay constants recommended by Steiger and Jager (1977). \(^{206}\text{Pb}/^{238}\text{U} \) ratios and dates were corrected for initial \(^{208}\text{Pb}/^{232}\text{U} \) disequilibrium using a Th/U[magma] of 3 ± 1, resulting in a systematic increase in the \(^{206}\text{Pb}/^{238}\text{U} \) dates of ~90 k.y. All common Pb in analyses was attributed to laboratory blank and subtracted based on the measured laboratory Pb isotopic composition and associated uncertainty. U blanks are difficult to precisely measure but were ~0.1 pg.

Dates presented herein are \(^{206}\text{Pb}/^{238}\text{U} \) dates, and the internal errors on analyses of single crystals and fragments are given at the 2σ level (Table 3; Fig. 12). Crystalization ages of the samples were interpreted from the weighted means of the \(^{206}\text{Pb}/^{238}\text{U} \) dates, based on three to four crystals per sample that are equivalent in age, calculated using Isoplot 3.0 (Ludwig, 2003). Errors on the weighted mean dates are the internal errors based on analytical uncertainties only, including counting statistics, spike subtraction, and blank and initial common Pb subtraction. It is given at the 95% confidence interval, which is the internal 2σ error for samples having mean square of weighted deviates (MSWD) <1.2 and the internal 2σ error expanded by the square root of the MSWD and the Student’s t multiplier of \( n – 1 \) degrees of freedom for samples with MSWD >1.2. This error should be considered when comparing our dates with those from other laboratories that used the same Boise State tracer solution or a tracer solution that...
Figure 11. Left: Map of the Owyhee Mountains showing sample locations for geochronologic studies. Locations are labeled with the site number, the U-Pb date (where possible), and the initial strontium ratio. Right: Normal faults (solid) and inferred normal faults (dashed) in the Owyhee Mountains (modified from Ekren et al., 1981).
## Table 3. U-Th-Pb Isotopic Data

<table>
<thead>
<tr>
<th>Sample</th>
<th>Th/U</th>
<th>206Pb/238U</th>
<th>207Pb/206Pb</th>
<th>208Pb/206Pb</th>
<th>206Pb*</th>
<th>207Pb*</th>
<th>208Pb*</th>
<th>% err§§</th>
<th>Corr. coeff.</th>
<th>Isotopic dates</th>
</tr>
</thead>
<tbody>
<tr>
<td>SM06-1 Unnamed intrusion granodiorite</td>
<td>0.245</td>
<td>2.7922</td>
<td>99.79</td>
<td>134</td>
<td>0.48</td>
<td>8830</td>
<td>0.079</td>
<td>0.047804</td>
<td>0.091</td>
<td>0.089152</td>
</tr>
<tr>
<td>SM06-2 Chipmunk Meadow Intrusive Suite</td>
<td>0.514</td>
<td>1.9393</td>
<td>99.72</td>
<td>107</td>
<td>0.45</td>
<td>6567</td>
<td>0.164</td>
<td>0.047922</td>
<td>0.102</td>
<td>0.01286</td>
</tr>
<tr>
<td>SM06-4 Wilson Peak granodiorite</td>
<td>0.244</td>
<td>0.5042</td>
<td>98.66</td>
<td>202</td>
<td>0.45</td>
<td>1304</td>
<td>0.010</td>
<td>0.048026</td>
<td>0.067</td>
<td>0.015162</td>
</tr>
<tr>
<td>SM06-5 South Mountain granodiorite</td>
<td>0.543</td>
<td>2.1407</td>
<td>99.98</td>
<td>171</td>
<td>0.51</td>
<td>2309</td>
<td>0.116</td>
<td>0.047817</td>
<td>0.137</td>
<td>0.002311</td>
</tr>
<tr>
<td>SM06-6 South Mountain hornblende</td>
<td>0.445</td>
<td>0.5177</td>
<td>97.54</td>
<td>12</td>
<td>0.66</td>
<td>756</td>
<td>0.144</td>
<td>0.047679</td>
<td>0.597</td>
<td>0.047996</td>
</tr>
<tr>
<td>SM06-7 Idaho batholith granodiorite</td>
<td>0.036</td>
<td>0.3836</td>
<td>98.87</td>
<td>26</td>
<td>0.50</td>
<td>1645</td>
<td>0.140</td>
<td>0.048628</td>
<td>0.430</td>
<td>0.048348</td>
</tr>
</tbody>
</table>

*2z1, z2, etc., are labels for analyses composed of single zircon grains or fragments; z7a, z7b, z7c are fragments from the same grain; analyses used in weighted mean calculations are shown in bold; grains were annealed and chemically abraded after Mattinson (2005).

*2Model Th/U ratio calculated from radiogenic 206Pb/238U ratio and 238U/206PbU age.

*2“Pb* and Pb represent radiogenic and common Pb, respectively; mol % 206Pb* is given with respect to radiogenic, blank, and initial common Pb.

*2**Corrected for fractionation, spike, and common Pb; all common Pb was assumed to be procedural blank: 206Pb/204Pb = 18.60 ± 0.80%; 207Pb/204Pb = 15.69 ± 0.32%; 208Pb/204Pb = 105.9 ± 0.14.

*2**Errors are 2σ.

*2**Correlation coefficients are based on analyses of NBS-981 and NBS-982.

*2**Calculations are based on the decay constants of Jaffey et al. (1971). 206Pb/238U and 207Pb/206Pb dates are corrected for initial disequilibrium in 230Th/238U using Th/U(magma) = 3.
Figure 12. Left: U-Pb age of zircon grains and fragments. Note: Two breaks in time occur between 49 and 76 Ma and 99 Ma and 150 Ma. Samples SM06-97 and SM06-98 are from South Mountain, SM06-90 is from Chipmunk Meadow Intrusive Suite, SM06-11 is from Whiskey Ridge tonalite, SM06-94 is from Wilson Peak granodiorite, SM06-1 is from an eastern unnamed intrusion, and SM06-109 is from the eastern Owyhee Mountains, inferred to be Idaho batholith proper. Error bars on the dates are 2σ. Right: Cathodoluminescence (CL) images of zircon grains, with numbers corresponding with those to the left.
was calibrated using EARTHTIME gravimetric standards. When comparing Owyhee Mountains dates with those derived from other tracer solutions and decay schemes (e.g., \(^{40}\)Ar/\(^{39}\)Ar, \(^{187}\)Re-\(^{187}\)Os), the uncertainties in the tracer calibration (0.08%) and \(^{238}\)U decay constant (0.11%; Jaffey et al., 1971) should be added to the internal error in quadrature. Over the course of the experiment, isotope analyses of the TEMORA zircon standard yielded a weighted mean \(^{206}\)Pb/\(^{238}\)U age of 417.43 ± 0.06 (\(\mu = 11\), MSWD = 0.8).

GEOCHRONOLOGIC RESULTS

Concordant U-Pb dates were obtained from 40 of the 41 analyses from seven samples. Dates from individual grains and fragments from a given sample are often not equivalent within their internal errors (generally ±0.05 m.y.), with the largest spread being 100 m.y. and the smallest being 0.3 m.y. Interpretations of igneous crystallization ages thus require assumptions as to the cause(s) of the spread. We interpret differences at the hundred-thousand to million-year scale as resulting from an extended period of zircon crystallization in magmas and/or crystal mush networks, as described for large composite granitoid plutons (Matzel et al., 2006; Miller et al., 2007). Such sub-million-year scatter is not resolvable with other lower precision, in situ (e.g., ion probe, laser ablation) U-Pb geochronological methods. Inherited cores in the crystals, which are also common in evolved granitoids, produce differences at the scale of several million to tens of millions of years. With the exception of one sample, SM06-109, such cores are necessarily less than a few million years older than magmatic overgrowths because large age differences would have resulted in the U-Pb dates being discordant and more scattered. The method of grain mount CL imaging was clearly efficient at rejecting crystals having significantly older inherited cores and highlights the effective in situ capability of low-blank ID-TIMS analysis.

Based on these inferences, igneous crystallization of the sampled granitoid rocks is conservatively interpreted to have occurred during or after the youngest dated zircon. For several samples, this crystallization date includes three to four dates that are equivalent at the young end of the date spread, and thus the end of igneous crystallization can be confidently assigned to this cluster of youngest dates. An exception to this interpretation is the single Jurassic sample, SM06-98, which has a pattern of zoning and dates that suggest the younger dates are younger than the igneous crystallization age, a result of minor zircon recrystallization during subsequent rock deformation and Cretaceous magmatism.

South Mountain

In order to complete a transect across the entire Owyhee segment of the western Idaho shear zone, two samples were collected west of the western Idaho shear zone at South Mountain, ~40 km southwest of the northern Owyhee Mountains (Table 4). Samples of granodiorite (SM06-97) and hornblendite (SM06-98) were collected 1 km apart.

CL imaging of zircon from sample SM06-98 showed no apparent inherited cores. Some of the grains have small, CL-bright domains at the margins of the crystals that appear to have formed during recrystallization. Four of the six analyzed crystals yielded equivalent dates having a weighted mean age of 159.59 ± 0.13 Ma (MSWD = 3.1). Dates from two other crystals are ~1 and 8 m.y. younger. The igneous crystallization age is interpreted to be the weighted mean date. Minor recrystallization at ≤151 Ma is thought to have resulted in younger dates in the two other grains.

Chipmunk Meadow Intrusive Suite

Half of the CL images of zircon from sample SM06-90 appear to show inherited cores. Dates from four crystals are spread between 98.13 ± 0.05 and 97.91 ± 0.05 Ma, and another is slightly younger at 97.47 ± 0.06 Ma. A precise igneous crystallization age cannot be interpreted; however, it is likely that crystallization occurred at ≤97.5 Ma.

Whiskey Ridge Tonalite

CL imaging of zircon from sample SM06-11 showed a small proportion of crystals that appear to have inherited cores. Three of the five analyzed crystals yielded equivalent dates having a weighted mean age of 90.43 ± 0.03 Ma (MSWD = 1.2). Dates from two other crystals are ~0.2 and 1.0 m.y. older. The igneous crystallization age is interpreted to be the weighted mean age, and the two older crystals are thought to have longer magma residence histories or inherited cores.

Wilson Peak Granodiorite

CL imaging of zircon from sample SM06-94 showed a small proportion of crystals that appear to have inherited cores. Ages from three grains are spread between 89.55 ± 0.05 and

<table>
<thead>
<tr>
<th>Table 4: Strontium isotopic data</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sample Unit Lithology T (Ma) Rb (ppm) Sr (ppm) (^{87})Rb/(^{86})Sr (^{87})Sr/(^{86})Sr ± (cm) (^{87})Sr/(^{86})Sr(ij) Latitude (°N) Longitude (°W)</td>
</tr>
<tr>
<td>SM06-1 Unnamed intrusion Granodiorite 85.8 61.5 789.0 0.2254 0.706702 ± 4 0.706391 43°11′16″ 116°51′41.4″</td>
</tr>
<tr>
<td>SM06-6 Dropoff Intrusive Suite Granodiorite 95° 52.4 484.7 0.3126 0.706419 ± 4 0.706004 43°07′30.4″ 116°46′19.0″</td>
</tr>
<tr>
<td>SM06-11 Whiskey Ridge tonalite Tonalite 90.4 37.2 819.8 0.1313 0.706557 ± 3 0.706391 43°11′01.3″ 116°47′45.1″</td>
</tr>
<tr>
<td>SM06-70 Chipmunk Meadow Intrusive Suite Granodiorite 97.5° 83.6 485.5 0.4977 0.706391 ± 3 0.705712 43°18′36.5″ 116°46′55.9″</td>
</tr>
<tr>
<td>SM06-66 Hardtrigger Intrusive Suite Granodiorite 90° 78.4 503.1 0.4508 0.707110 ± 3 0.706542 43°22′33.7″ 116°45′34.4″</td>
</tr>
<tr>
<td>SM06-90 Chipmunk Meadow Intrusive Suite Granodiorite 97.5° 50.1 751.0 0.1929 0.706224 ± 4 0.705961 43°10′57.8″ 116°51′41.4″</td>
</tr>
<tr>
<td>SM06-94 Wilson Peak granodiorite Granodiorite 89.5° 54.9 610.5 0.2599 0.707073 ± 4 0.706748 43°15′19.2″ 116°46′16.9″</td>
</tr>
<tr>
<td>SM06-97 South Mountain Granodiorite 47.9° 49.5 582.6 0.2727 0.706674 ± 3 0.706492 42°44′15.3″ 116°55′02.9″</td>
</tr>
<tr>
<td>SM06-98 South Mountain Hornblendite 159.6° 4.0 152.1 0.0769 0.704746 ± 3 0.704574 42°43′35.3″ 116°54′30.4″</td>
</tr>
<tr>
<td>SM06-109 Idaho batholith Granodiorite 76° 38.1 581.8 0.1894 0.708098 ± 3 0.707896 43°04′06.2″ 116°37′11.0″</td>
</tr>
</tbody>
</table>

*Denotes estimated age.
89.44 ± 0.05 Ma, having a weighted mean age of 89.50 ± 0.13 Ma (MSWD = 4.9). Dates from two other crystals are 0.1 and 0.2 m.y. older. The igneous crystallization age is interpreted to be the weighted mean date, and the two older crystals are thought to have longer magma residence histories or inherited cores.

Sample SM06-1 is from a small granitic exposure southeast of the Wilson Peak granodiorite (Fig. 11), and it is referred to as the unnamed intrusion because not enough is known about the areal extent of the unit. CL imaging showed that most crystals appear to have inherited cores. Dates from six crystals are spread between 87.71 ± 0.05 and 85.75 ± 0.05 Ma. A precise igneous crystallization age cannot be interpreted; however, it is likely that the two youngest crystals constrain crystallization at ca. 85.8 Ma.

Eastern Owyhee Mountains–Idaho Batholith

Sample SM06-109 was collected from farther east in the northern Owyhee Mountains (Table 4). It is a granite containing biotite and muscovite and is interpreted as part of the Idaho batholith proper. CL imaging showed that most zircon crystals appear to have inherited cores, and crystals from this sample are more heterogeneous than crystals from the other samples. Dates from six crystals are spread between 84.74 ± 0.05 and 75.93 ± 0.04 Ma. Dates from two other crystals are ~50 and 100 m.y. older. Dates from three fragments from a single crystal are spread between 84.74 and 79.52 Ma, confirming the suite crystallization history of these crystals. A precise igneous crystallization age cannot be interpreted; however, it is likely that crystallization occurred after ca. 76 Ma.

STRONTIUM ISOTOPIC SIGNATURE

Initial 87Sr/86Sr values have been critical to understanding the geometry and deformatinal history of the western Idaho shear zone, especially along the McCall segment (Manduca et al., 1992; Giorgis et al., 2005b). Steep gradients in the (87Sr/86Sr) of plutons in west-central Idaho correspond spatially with the western Idaho shear zone. Plutons in the western part of the shear zone have (87Sr/86Sr) of 0.704 or less, whereas the eastern part has (87Sr/86Sr) values of 0.706 or greater (Armstrong et al., 1977; Fleck and Criss, 1985, 2007; Manduca et al., 1992).

Results from 10 samples from across the Owyhee segment were added to those from Armstrong et al. (1977) to gain regional coverage of the Owyhee Mountains. The data were placed on a transect oriented at 110°, perpendicular to the trace of the dominant 020° foliation of the Owyhee segment of the western Idaho shear zone. Because the Cretaceous units are fault bounded and largely covered by younger volcanic rocks, it was not possible to plot the data on a single, spatially coherent transect. Rather, a series of segments was constructed through each of the pre-Miocene granitoid exposures, starting at South Mountain in the west and ending at the unnamed intrusion on the east side of the Owyhee Mountains (Fig. 11). Using this approach, all major intrusive units are represented in a 31-km-long geochronological transect.

One hundred to 300 g of rock powder for each sample were dissolved in a 3:1 mixture of HF:HNO3 on a hot plate for 48 h, followed by conversion to a 6 M HCl solution. A 10% aliquot of the resulting solution was spiked with 87Rb and 84Sr tracers. Rb and Sr were quantitatively separated by standard cation exchange chemistry for both the isotope composition and the isotope dilution aliquots. Using isotope dilution techniques, samples were analyzed for Rb and Sr concentrations and 87Sr/86Sr (Table 4). Isotope ratios were measured on the Isoprobe-T at Boise State University. The uncertainty reported for each analysis is the internal standard error. The external reproducibility of the NBS-987 standard is 0.710251 ± 3 (1σ); uncertainty in [Rb], [Sr], and 87Rb/86Sr is estimated at ≤0.3% (1σ). Initial 87Sr/86Sr was calculated assuming an 87Rb decay constant of 1.398 × 10⁻¹¹ yr⁻¹ (Nebel-Jacobsen et al., 2005):

$$\frac{87\text{Sr}}{86\text{Sr}} \frac{87\text{Rb}}{86\text{Sr}} = e^{\frac{1.398 \times 10^{-11}}{t}} - 1 \quad (1)$$

where t is time elapsed since intrusion. Ages were estimated using a combination of U-Pb zircon dates on a subset of samples from across the Owyhee Mountains (see previous) and associated field relationships.

Results

Including the data of Armstrong et al. (1977), the (87Sr/86Sr), values range between 0.704574 and 0.707892 (Table 4). Data were projected onto a line trending orthogonal to the 020° contacts of the units (110°–290°), and the effects of Basin and Range extension were removed to juxtapose the fault-bounded granitoid intrusive blocks. Initial 87Sr/86Sr increases from west-northwest to east-southeast over ~31 km (Fig. 13). This is a minimum distance because the blocks were retrodeformed in order to juxtapose them. One site from South Mountain was removed from the analysis because its U-Pb zircon age (47.80 ± 0.05 Ma) indicates that it was not involved in Late Cretaceous deformation.

![Figure 13. Plot of (87Sr/86Sr) versus distance. Distance is measured perpendicular to the shear zone boundaries (along a traverse trending 110°). Samples from this study are denoted by solid black diamonds, and samples from Armstrong et al. (1977) study are plotted as gray squares. Larger error bars are a result of the unknown age of some of the intrusive units, which are estimated to have an error of ±20 m.y. Black line is the best-fit line of the data, and the equation for this line and the $R^2$ value are in the lower right of the plot.](https://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/2/2/93/3038140/93.pdf)
One of the westernmost samples from South Mountain (SM06-98) is Jurassic and has the lowest (\(^{87}\)Sr/\(^{86}\)Sr) (Fig. 13). The easternmost sample (SM06-109) has the highest (\(^{87}\)Sr/\(^{86}\)Sr), one that is more typical of magmatic arcs that intrude through continental-arc crust (Manduca et al., 1992; Langenheim et al., 2004). This high Sr value and a two-mica granite composition are consistent with the Idaho batholith proper. The main map area (Chipmunk Meadow Intrusive Suite, Whiskey Ridge tonalite, Dropoff Intrusive Suite, and Wilson Peak granodiorite) has rocks that have a mean (\(^{87}\)Sr/\(^{86}\)Sr) \textasciitilde 0.706, typical of magmas generated from mixed oceanic-continental crust (Manduca et al., 1992; Vijan et al., 2003). The Owyhee segment has a steep gradient in (\(^{87}\)Sr/\(^{86}\)Sr), from west to east (Fig. 13). The (\(^{87}\)Sr/\(^{86}\)Sr), gradient across the Owyhee segment indicates a transition from magmas with an accreted arc signature to ones with a continental signature that is similar to, but not as steep as, the McCall segment.

**EVOLUTION OF MAGMATISM AND DEFORMATION**

The northern Owyhee Mountains consist of a series of tabular plutons and intrusive suites that range in compositions and in crystallization ages. The composition of the intrusive suites ranges from granitic to tonalitic, and the ages range from 97.5 Ma in the west to 85 Ma in the east. In the northern Owyhee Mountains, there is a consistent planar fabric orientation of 020° that has an associated steep eastward dip and a downdip lineation of the most deformed rocks. Overall, the degree of fabric development is weak, but deformed granitoid exposures occur over >15 km perpendicular to strike of the zone, indicating a wider deformed zone than the McCall segment. Solid-state fabric is best developed in the Chipmunk Meadow (97.5 Ma) and the Dropoff (undated) Intrusive Suites and the Hardtrigger orthogneiss. The Whiskey Ridge tonalite (90.5 Ma) has a weaker solid-state fabric, and the Wilson Peak granodiorite (89 Ma) contains a magmatic fabric, indicating that these units intruded during the waning stages of western Idaho shear zone deformation.

There is a gradient in initial Sr ratios across the Owyhee segment. The lowest values occur in the westernmost intrusions, suggesting little continental crust contamination, whereas the easternmost intrusions have higher values, indicating significant continental crust contamination. This transition indicates that the Owyhee segment records the juxtaposition of oceanic lithosphere and continental lithosphere. The transition occurs over ~31 km, which we interpret as the actual width of the deformed zone in the Owyhee segment due to the lack of exposure.

Solid-state fabric in the Chipmunk Meadow Intrusive Suite indicates that deformation was ongoing at 97.5 Ma. Magmatic fabric in the Whiskey Ridge tonalite and Wilson Peak granodiorite indicates that deformation was waning by ca. 90 Ma. The Whiskey Ridge tonalite is interpreted to have intruded during the waning stages of deformation based on the parallelism of magmatic and solid-state fabrics. The Wilson Peak granodiorite is interpreted as a late-syn-tectonic intrusion based on the magmatic fabric being parallel to the regional solid-state foliation in other units.

**ALONG-ZONE VARIATIONS**

Intrusions in the McCall and Owyhee segments are similar in geometry (tabular), composition (granitic, granodioritic, and tonalitic), age, and age progression (young to the east). The segments differ in degree of fabric development and orientation and width. In the McCall segment, solid-state fabric occurs in the eastern part of the Hazard Creek complex, throughout the Little Goose Creek complex, and in the western part of the Payette River tonalite. In the Owyhee segment, a solid-state fabric occurs in the Chipmunk Meadow Intrusive Suite, the Dropoff Intrusive Suite, Hardtrigger orthogneiss, and locally in the Whiskey Ridge tonalite. The Whiskey Ridge tonalite is nearly identical in composition and age to the Payette River tonalite of the McCall segment (Manduca et al., 1993). The Dropoff Intrusive Suite of the Owyhee segment is most similar to the highly deformed Little Goose Creek complex of the McCall segment.

Fine-grained quartz, biotite aligned with foliation, and K-feldspar porphyroclasts with tails characterize both suites, although the planar and linear fabric elements are much better developed in the Little Goose Creek complex.

Fabric in the Owyhee and McCall segments differs in degree of development and presence of shear sense indicators. Differences in the degree of fabric development are largely related to the difference in width of the deformed zone of the two segments; deformation in the Owyhee segment occurs over a distance about five times greater than the McCall segment. The mean degree of anisotropy of the K-feldspar SPO ellipsoid is 1.097 in the Owyhee segment compared to 1.64 for the McCall segment (Fig. 14) (Giorgis et al., 2004). This lower degree of fabric development indicates that strain is much lower in the Owyhee segment. Consistent with the low strain, in the northern Owyhee Mountains recrystallized tails on feldspar are rare and do not indicate a shear direction.

**Figure 14. Hsu plot showing degree of anisotropy versus shape factor based on shape-preferred orientation (SPO) analyses of K-feldspar grains for the Owyhee and McCall segments.**

The most developed foliation in the Owyhee segment strikes ~020°, whereas in the McCall segment, it strikes north-south. Foliation for both segments dips steeply and is associated with a downdip biotite-aggregate lineation. Solid-state foliation changes from a northerly strike to a north-northeasterly strike at the southernmost extent of the western Idaho shear zone, north of the western Snake River Plain, and south of Snowbank Mountain, ~50 km south of McCall, Idaho. Along-zone variation in orientation for the two segments may account for the variations noted in fabric development and shear zone width between the Owyhee and McCall segments of the western Idaho shear zone.

The isotopic evidence shows that the deformed zone of the Owyhee segment is much wider than that of the McCall segment. Qualitatively, solid-state and magmatic fabrics observed in the Owyhee segment occur over a greater distance. Quantitatively, the gradient in (\(^{87}\)Sr/\(^{86}\)Sr) measured in the Owyhee segment can be compared to gradients in the McCall segment and in the Sierra Nevada batholith (Fig. 15). Only ratios of the Sierra Nevada batholith within the range of the Owyhee segment (\(^{87}\)Sr/\(^{86}\)Sr) are included in this comparison. The Sierra Nevada batholith is chosen as a comparable analogue because it represents the best example of an unmodified
continental magmatic arc built on the edge of western Laurentia during the Late Cretaceous, and fabric analyses by Giorgis et al. (2005b) restore the western Idaho shear zone to a width similar to the Sierra Nevada batholith. The transition in Sr ratios from oceanic to continental crust in the Owyhee segments occurs across ~31 km and coincides with the zone of solid-state deformation. This transition occurs over ~6 km in the McCall segments and ~71 km in the Sierra Nevada batholith (Fig. 15). More precisely, the ($^{87}$Sr/$^{86}$Sr) data of the Owyhee segment define a gradient of ~0.001/12.5 km, compared to ~0.0012 km for the McCall segment, and ~0.0012 km for the Sierra Nevada batholith.

The width of the western Idaho shear zone does not always correlate to its orientation. In particular, the western Idaho shear zone in the Syringa embayment has an orientation of ~015° (e.g., Lund et al., 2008) and closely spaced ($^{87}$Sr/$^{86}$Sr) isopleths. This section of the western Idaho shear zone is affected by younger deformation, however, specifically, the Klotpon Creek–Hammer Creek–Mount Idaho deformation zone and Orofino shear zone (McClelland and Oldow, 2007; Schmidt and Lewis, 2007). Following Schmidt and Lewis (2007), we attribute the ~015° orientation of the western Idaho shear zone in the Syringa embayment to reflect reorientation of an originally N–S–oriented western Idaho shear zone.

Along-zone variations in Sr isotopic patterns in western Idaho clearly occur (Fleck and Criss, 1985, 2007). Giorgis et al. (2005b) emphasized that solid-state deformation steepened the gradient of the ($^{87}$Sr/$^{86}$Sr) transition in west-central Idaho. The gradient is a product of both the deformation and the original steepness of the transition from oceanic to continental lithosphere. These factors create three possibilities for along-zone variations of the western Idaho shear zone: they are a result of differences in (1) initial width of the oceanic and continental lithospheric contact, (2) dip of the boundary, or (3) magnitude of finite strain. In the following sections, the last possibility is assumed, and a tectonic model is developed to determine the tectonic consequences.

**TECTONIC MODEL**

In a transpressional high-strain zone, differences in orientation relative to the regional oblique-convergence angle strongly affect the relative amounts of convergent and transcurrent movements. Along-zone variations in pluton ages, ($^{87}$Sr/$^{86}$Sr) gradients, and solid-state and magmatic fabric development due to along-zone variations in orientations and widths lead to along-zone variations in finite and infinitesimal strains. The orientations of the McCall and Owyhee segments of the western Idaho shear zone differ by ~20°. Because previous work (e.g., Giorgis and Tikoff, 2004) constrained the initial maximum width of the western Idaho shear zone, these limits and variations in the width of the high-strain zone between the two segments enable the oblique-convergence angle to be calculated.

In transpressional settings, the strain path can be quantified using the angle of oblique convergence ($\alpha$). Pure transcurrent displacement has an $\alpha$ value of 0°, and pure convergent displacement has an $\alpha$ value of 90° (Sanderson and Marchini, 1984; Fossen et al., 1994). The McCall segment of the western Idaho shear zone was interpreted at 45° ≤ $\alpha$ ≤ 75° (Giorgis and Tikoff, 2004; Dair and Giorgis, 2006; Giorgis et al., 2006b). If this is a regional convergence angle, then locally in the Owyhee segment, values for $\alpha$ would be between 25° and 55° with respect to the western Idaho shear zone solid-state deformation fabric boundaries, because the Owyhee segment is oriented 20° clockwise from the McCall segment. This angle of $\alpha$ potentially results in larger transcurrent displacements and smaller convergent displacements for the Owyhee segment.

**Tectonic Model for the Western Idaho Shear Zone**

A tectonic model was developed based on the assumption that the initial width of the western Idaho shear zone was the same for both segments and that the final widths of the McCall and Owyhee segments are 6 km (based on Giorgis et al., 2005b) and 31 km (this study), respectively. Using the gradients of the initial Sr ratios for both segments to define an initial width, $\alpha$ can be constrained for both segments (Fig. 16). Assumptions for the model are: (1) the initial width of the ($^{87}$Sr/$^{86}$Sr) gradient prior to deformation was the same for both segments; (2) colliding terranes on either side of the western Idaho shear zone were completely rigid during deformation; and (3) post-Cretaceous extension was comparable in both locations. The second assumption is supported by a lack of major Cretaceous deformation in the adjacent terranes of the Blue Mountain Province (e.g., Avé Lallemant, 1995; Schwartz et al., 2010).

For both areas, the amount of convergence is the input initial width ($W_i$) minus the final width ($W_f$) (Fig. 16). For the McCall segment, the amount of convergent displacement is:

$$s = W_o - W_{6C}.$$  

Similarly, for the Owyhee segment, the amount of convergent displacement is:

$$p = W_o - W_{6O}.$$
The transcurrent displacement for the two segments is dependent on the McCall oblique-convergence angle (α). For the McCall segment, the transcurrent displacement, d, is:

\[ d = \frac{s}{\tan \alpha} \]  

(4)

For the Owyhee segment, the transcurrent displacement, e, is also dependent on the difference in orientation of the western Idaho shear zone boundaries (θ) between the two segments:

\[ e = \frac{p}{\tan(\alpha - \theta)} \]  

(5)

For the case of the western Idaho shear zone, θ is 20°. In this model, the oblique-convergence angle for the Owyhee segment is defined as α – θ. The value for α, however, must be solved for:

\[ \sin \alpha = \frac{s}{h} \]  

(6)

where h is the displacement vector (Fig. 16), and

\[ \sin(\alpha - \theta) = \frac{p}{h} \]  

(7)

Equations 6 and 7 can be set equal to each other by solving for h:

\[ \frac{s}{\sin \alpha} = \frac{p}{\sin(\alpha - \theta)} \]  

(8)

By using the trigonometry identity

\[ \sin(\alpha - \theta) = \sin \alpha \cos \theta - \cos \alpha \sin \theta. \]  

(9)

the oblique-convergence angle (α) can be isolated:

\[ \alpha = \tan^{-1} \left[ \sin \theta \frac{\sin \alpha}{\cos \theta - \frac{p}{s}} \right] \]  

(10)

By substituting Equations 2 and 3, Equation 10 becomes

\[ \alpha = \tan^{-1} \left[ \sin \theta \frac{\sin \alpha}{\cos \theta - \frac{W_o - W_{fo}}{W_o - W_{sm}}} \right] \]  

(11)

Using calculated values for α, the amount of transcurrent displacement for the two segments, d and e, can be determined using Equations 4 and 5. Limits to the initial width for the western Idaho shear zone during the Late Cretaceous are based on an ~85–100 km width estimated by Giorgis et al. (2005b) for the McCall segment and based on the distance of the same transition in values of Sr isotope ratios for the Sierra Nevada batholith (~71 km). We will utilize an original width of 71 km, based on the ($^{187}$Rb/$^{87}$Sr) gradient in Sierra Nevada batholith, which is the most conservative of the estimates.

**Results of Tectonic Model**

For Equation 11, certain variables remain constant for any initial width. In the Owyhee segment, $W_{fo} = 31$ km and θ = 20°, and for the McCall segment $W_{fM} = 6$ km. With these variables defined, the oblique-convergence angle, α, can be determined for any original width. For an original width ($W_o$) of 71 km, Equation 11 becomes:

\[ \alpha = \tan^{-1} \left[ \frac{\sin 20\degree}{\cos 20\degree - \frac{71 - 31}{71 - 6}} \right] \]  

(12)

Thus, α = 46.5° for an initial width of 71 km, which is the oblique-convergence angle for the McCall segment. In the northern Owyhee Mountains, the oblique-convergence angle would be 26.5°, because of the 20° along-zone variation in orientation of the segments. Based on the proposed range of the initial width for the western Idaho shear zone between 71 and 100 km, the tectonic model produces an oblique-convergence angle between 46.5° and 59° (Fig. 17A). These results are consistent with and further constrain the predictions along the McCall segment using 71 and 100 km initial widths (Giorgis and Tikoff, 2004; Dair and Giorgis, 2006; Giorgis et al., 2006b), where the oblique-convergence angle was predicted to be between 45° and 75°. The difference in orientation of the segments translates to a greater amount of transcurrent displacement (80 km) and a smaller amount of convergent displacement (40 km) in the Owyhee segment relative to the McCall segment, which would be 67 km and 65 km, respectively (Fig. 17B).

The variable θ, the difference in orientation of the two segments, is the most sensitive variable for constraining the oblique-convergence angle and the amount of transcurrent displacement for the two segments. The convergent component is not affected because it is constrained by the initial and final widths of the two segments and is not dependent on θ. However, each degree change in θ results in a 2° change in the oblique-convergence angle for both segments for oblique-convergence angle ~45°. As θ decreases, the transcurrent displacement increases in both areas, with a greater increase in the McCall segment. If θ were 15° instead of 20°, then there would be 22 km more transcurrent displacement for the Owyhee segment and 26 km more transcurrent displacement for the McCall segment. More simply, a smaller varia-

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**Figure 16. Map view of geometry of the oblique convergence in the western Idaho shear zone.** (A) For the McCall segment, α correlates directly to the amount of shortening and wrenching of the western Idaho shear zone. (B) In the Owyhee segment, the convergent component is $W_o - W_{fo}$ and the transcurrent component (e) is $e = (W_f - W_{fo})/\tan(\alpha - \theta)$. The variables are α = McCall oblique-convergence angle; $W_o$ = final width of the Owyhee segment = 31 km; $W_{fo}$ = original width of the shear zone; $\theta$ = angle between the orientation of the McCall segment and the orientation of the Owyhee segment, 20°, and $\alpha - \theta$ = the Owyhee oblique-convergence angle.
tion in along-zone orientation results in similar amounts of transcurrent displacements for the two segments.

From Displacements to Finite Strain

The convergent and transcurrent displacements differ significantly between the McCall and Owyhee segments. Using 71 km as the original width (the current width of the initial strontium ratio transition zone in the Sierra Nevada batholith), the McCall segment experienced 63% more convergent displacement and 23% less transcurrent displacement than the Owyhee segment. The difference in geometry and components of convergent and transcurrent displacement can be directly related to the amount of finite strain of both segments. Assuming homogeneous material behavior, a single deformation matrix that incorporates simultaneous pure shear and simple shear components can express the deformation for both segments (Tikoff and Fossen, 1993):

\[
D = \begin{bmatrix}
1 & \gamma_x(1-k_2) & 0 \\
0 & \ln(1/k_2) & 0 \\
0 & 0 & 1/k_2 
\end{bmatrix} \tag{13}
\]

In Equation 13, \(\gamma_x = \) (amount of transcurrent movement)/(original width of the shear zone) and \(k_2 = \) (final width of the shear zone)/(original width of the shear zone). For an original width of 71 km for the McCall segment, \(\gamma_x = \) (62 km)/(71 km) or 0.87, and \(k_2 = \) (6 km)/(71 km) or 0.08. The simple shear component, \(d_{13} \), equals \((0.87(1-0.08))/(\ln(1/0.08)) \) or 0.32. The deformation matrix for the McCall segment for an original width of 71 km is then

\[
D = \begin{bmatrix}
1 & 0.32 & 0 \\
0 & 0.08 & 0 \\
0 & 0 & 11.83 
\end{bmatrix} \tag{14}
\]

and the deformation matrix for the Owyhee Mountains is

\[
D = \begin{bmatrix}
1 & 0.77 & 0 \\
0 & 0.44 & 0 \\
0 & 0 & 2.29 
\end{bmatrix} \tag{15}
\]

The larger transcurrent component, and thus larger simple shear component \(d_{13} \), is greater in the Owyhee segment than in the McCall segment. In the McCall segment, \(d_{13} \) is much larger as a product of \(\alpha \) because the pure shear component, resulting from the larger convergent displacement, is much greater than in the Owyhee segment.

The larger transcurrent displacement in the Owyhee segment and the larger convergent displacement in the McCall segment indicate that the segments experienced different magnitudes of finite strain. See Appendix for calculations of finite strain and infinitesimal strain magnitudes and orientations for the limits for the initial widths for each segment. In summary, wider initial widths result in larger \(S_1 \) and smaller \(S_3 \) and \(S_{\perp} \) for both the Owyhee and McCall segments. The orientations of \(S_1 \) and \(S_3 \) are vertical in all cases, and the orientations of \(S_3 \) and \(S_{\perp} \) become more orthogonal to the boundary of the deforming zone.
A direct comparison of finite strain between the Owyhee and McCall segments is possible for any width. If the example of an original width of 71 km for both segments is employed again, then the lengths of the principal finite strain axes in the Owyhee segment are: \( S_1 = 2.29, S_2 = 1.080, \) and \( S_3 = 0.341, \) in contrast to the McCall segment, where they are: \( S_1 = 11.830, S_2 = 1.056, \) and \( S_3 = 0.075. \) In both areas, \( S_1 \) is vertical in orientation, and \( S_2 \) and \( S_3 \) are horizontal in orientation. In the Owyhee segment, \( S_1 \) is oriented 077°, and \( S_3 \) is oriented 57° clockwise from the shear-zone boundary. However, in the McCall segment, \( S_1 \) is oriented 089°, which is subperpendicular to the shear zone boundary of 000°.

A comparison of the infinitesimal strain for both segments can also be evaluated. Magnitudes and orientations of infinitesimal strain axes were calculated for 100 increments of deformation by using the matrix of Tikoff and Fossen (1993):

\[
D_{\text{incr}} = \begin{bmatrix} (k_1)^{1/n} & A & 0 \\ 0 & (k_2)^{1/n} & 0 \\ 0 & 0 & (k_3)^{1/n} \end{bmatrix}
\]

where \( A = \frac{\gamma_n}{n} \left( \frac{(k_1)^{1/n} - (k_3)^{1/n}}{\ln[(k_1)^{1/n} / (k_3)^{1/n}]} \right). \) See Appendix for a complete description. For the case of an original width of 71 km in the McCall segment, \( \gamma_n = 0.87, k_1 = 1, k_2 = 0.085, \) and \( n = 100, \) where \( n \) is the number of increments of deformation. Using these values, the equation becomes

\[
A = \frac{0.87}{100} \left( \frac{1^{1/100}}{0.085^{1/100}} \right) \ln[1^{1/100} / 0.085^{1/100}].
\]

In the 71 km example, the lengths of the infinitesimal principal strain axes in the McCall segment are: \( S_1 = 1.025, S_2 = 1.001, \) and \( S_3 = 0.975, \) compared to the Owyhee segment, where the axis lengths are: \( S_1 = 1.010, S_2 = 1.002, \) and \( S_3 = 0.987. \) Similar to the major axis of finite strain ellipsoid, \( S_1 \) is oriented vertically in both segments. \( S_3 \) is also similar in orientation to \( S_3 \). In the Owyhee segment, \( S_1 \) is oriented vertically in both segments, \( S_3 \) is also similar in orientation to \( S_3 \), and \( S_3 \) is oriented 066°, or 46° clockwise from the western Idaho shear zone boundary in the Owyhee segment. In the McCall segment, \( S_1 \) is oriented 080°, subperpendicular to the western Idaho shear zone boundary. For both areas, \( S_1 \) is oriented more orthogonal than parallel to the western Idaho shear zone boundaries, indicating that the western Idaho shear zone is characterized by pure-shear-dominated transpression, rather than simple-shear-dominated transpression, especially in the McCall segment (see Appendix).

The orientations of \( S_3 \) for the segments agree with the along-zone variation in degree of fabric development documented for both segments.

**DISCUSSION**

**Implications of the Tectonic Model**

The results of the model indicate the varying structural styles of deformation associated with the Owyhee and McCall segments of the western Idaho shear zone. The transcurrent component for both segments is highly constrained. For the Owyhee segment, only a 3 km range (82–85 km) is possible, and for the McCall segment, a 10 km range is possible (56–66 km). Previous work suggests that a minimum of 45 km of transcurrent displacement is possible along the McCall segment (McClelland et al., 2000; Tikoff et al., 2001; Giorgis and Tikoff, 2004). The convergent displacement is less constrained because this component is highly reliant on the initial width of the western Idaho shear zone in the model: the range is 40–69 km of convergent displacement for the Owyhee segment and 65–94 for the McCall segment. However, these limits fall within the range of Giorgis et al. (2005a) of 30–110 km based on strain data, and further constrain the convergent component.

The difference in contributions of the pure and simple shear components of finite strain due to along-zone variation in width and orientation of the western Idaho shear zone significantly affects the structural character of deformation for the different segments. In the Owyhee segment, the lower amount of pure shear resulted in a less steep Sr isotopic gradient and a lower amount of finite strain than in the McCall segment. The magnitude of the maximum principal finite strain axis (\( S_1 \)) for the Owyhee segment is between 2.29 and 3.23, whereas it is between 11.83 and 16.67 for the McCall segment. This larger \( S_1 \) for McCall is associated with significantly different intermediate and minimum finite strain axes for the two segments. Both areas record flattening fabric, although the solid-state fabric along the McCall segment is much better developed (Fig. 18).

The large value of \( S_1 \) for the McCall segment relative to the Owyhee segment would produce much greater amounts of vertical uplift (Fig. 19). This stretching would predict that the rocks of the McCall segment most likely deformed at greater depths than the rocks in the northern Owyhee Mountains. Further work to determine the pressure and temperature conditions of the Owyhee intrusions could test this hypothesis.

There are two additional caveats to our analysis. First, the magnitude of Cenozoic volcanic cover is sufficiently high in the northern Owyhee Mountains and may cover a large portion of the shear zone. This cover is important, particularly if it covers up high-strain areas. This concern is somewhat alleviated by the similarity of fabrics throughout the shear zone, an aspect shared by the McCall segment. Second, the lower strain values recorded in the Owyhee segment may result from the intrusions being
younger than those in the McCall segment. Alternatively stated, deformation of \( \geq 105 \) Ma rocks in the McCall segment may not be comparable to deformation of \( \leq 97 \) Ma rocks in the Owyhee segment, as the former may have accumulated more deformation during the period 105–97 Ma. This discrepancy is not a problem if: (1) Deformation in the western Idaho shear zone occurred after 97 Ma (we cannot constrain the initiation of western Idaho shear zone deformation); or (2) if there are younger intrusions in the McCall segment that contain well-developed fabrics. At present, there are few U-Pb dates from the central portion of the western Idaho shear zone.

The along-zone variation in the degree of flattening and wrenching, such as those documented for the Owyhee segment, is a common result of curved boundaries (e.g., Avé Lallémant and Guth, 1990; McCaffrey, 1991). Lin and Jiang (2001) demonstrated how differently oriented segments of the Southern Knee Lake shear zone in Manitoba, Canada, record different structural characteristics. An east-southeast–oriented segment is dominated by the pure shear component, whereas a south–southeast oriented segment is dominated by the simple shear component. These two segments are comparable to the McCall and Owyhee segments, respectively.

**Figure 19.** (A) Three-dimensional view of western Idaho shear zone prior to deformation. Dark-gray area corresponds to original width of the shear zone. (B) Map view of variation in simple shear component and horizontal pure shear component of deformation for the two segments. Arrows show consistent displacement vector along the shear zone. Note differences in final widths for the two segments. (C) Illustration of differences in vertical and horizontal components of only the pure shear deformation for the two segments.

**Cause of Different Orientations Along Trend of the Western Idaho Shear Zone**

We interpret the resultant deformations for the two segments of the western Idaho shear zone as a product of along-zone differences in strike orientation. The different orientation of the Owyhee segment relative to the McCall segment can be explained by two models: (1) the Owyhee segment is part of a bend in a larger shear zone that continues to the south, or (2) the Owyhee segment is the termination of the western Idaho shear zone.

The first possibility is that the western Idaho shear zone is part of a much larger dextral shear-zone system, which is continuous from the Sierra Nevada, California, to west-central Idaho. Dextral shearing is documented along the trend of the Sierra Nevada continental magmatic arc in California during this time. For example, the Sierra Crest shear zone system initiates at ca. 92 Ma (e.g., Tikoff and de Saint Blanquat, 1997). Dextral displacements occurred from the east-central Sierra Nevada (Toulumne Intrusive Suite; Tikoff et al., 2005) to where the southern Sierra Nevada continental magmatic arc terminates against the Garlock fault (Wood and Saleeby, 1998). If the Sierra Crest shear zone and the western Idaho shear zone are connected, then deformation occurred for over 1500 km,
preferentially located along the trace of the Sierra Nevada continental magmatic arc. Alternatively, the western Idaho shear zone may be continuous with Cretaceous dextral shear zones that occurred in the northwestern Great Basin in central Nevada (Oldow et al., 1993; Wyld and Wright, 2001).

The other possibility is that the Owyhee segment is the southern termination of the western Idaho shear zone. Previous work has shown that terminations in dextral shear zones bend clockwise to accommodate displacement (Freund, 1974; Ramsay, 1980; Simpson, 1983). For example, the dextral Hope fault in New Zealand exhibits ~15° of clockwise rotation at its southern terminus (Freund, 1971). The Owyhee segment is oriented clockwise from the McCall segment and displays a similar geometry to the Hope fault. If this scenario is correct, the question becomes why did the shear zone terminate at this location? One possible explanation involves the Late Cretaceous collision of the Insular superterrane (Baja–British Columbia), which has been speculated to be the cause of western Idaho shear zone transpressional deformation (Giorgis et al., 2005a). Palinspastic reconstructions based on minimal-estimate paleomagnetic reconstructions (Dickinson, 2004) and fault-offset reconstructions (Wyll et al., 2006) suggest that the Insular superterrane extended only as far south as the Idaho-Nevada border. If these interpretations are correct, it is reasonable to assume that a shear zone termination would occur in the vicinity of the northern Owyhee Mountains. At present, we cannot distinguish between these two possibilities. We do note, however, that other paleomagnetic studies require significantly more offset of the Insular superterrane (e.g., Wynne et al., 1995) and adjacent portions of Oregon (Housen and Dorsey, 2005), potentially favoring the first hypothesis.

CONCLUSIONS

The Late Cretaceous western Idaho shear zone deforms granitoid intrusions and intrusive suites in the northern Owyhee Mountains, south of the western Snake River Plain. The deformed intrusions of the Owyhee segment from west to east are the Chipmunk Meadow Intrusive Suite, the Whiskey Ridge tonalite, the Dropoff Intrusive Suite, and the intrusions in the Hardtrigger orthogneiss. The solid-state fabric in the Owyhee segment of the western Idaho shear zone has a consistent orientation. Foliation strikes 020° and dips steeply with a downdip lineation. The same orientation of fabric is observed at the farthest southern extent of the shear zone that occurs north of the western Snake River Plain and south of the McCall segment. Fabric analysis techniques (SPO) document a weak solid-state fabric throughout the western portion of the Owyhee segment, suggesting low finite strain. The strongest solid-state fabrics occur in the Dropoff Intrusive Suite. The weaker magmatic fabric and, in some areas, solid-state fabric of the Whiskey Ridge tonalite parallel the solid-state fabric of the intrusive suites, suggesting that this body is a syntectonic intrusion. The Whiskey Ridge tonalite is identical in composition, age, and geometry to the Payette River tonalite located near McCall. The Wilson Peak granodiorite contains a magmatic fabric parallel to the regional solid-state fabric and also is considered to be a syntectonic intrusion.

Plutonic rocks in the Owyhee segment yield U-Pb zircon dates of 160–50 Ma. Only Late Cretaceous plutons older than 90 Ma were affected by the deformation; Jurassic plutons occur farther southwest at South Mountain. In particular, deformation in the Owyhee segment ceased by ca. 90 Ma, consistent with relations in the McCall segment (Giorgis et al., 2008).

A steep gradient in (87Sr/86Sr) occurs over ~30 km from west to east and coincides with the zone of solid-state deformation in the Owyhee segment. Initial Sr ratios increase from values indicative of little continental crust contamination (e.g., 0.704574) to values signifying significant continental crust involvement (e.g., 0.707892). This gradient is atypically steep in comparison with most of the U.S. Cordillera (occurring over ~31 km), but it is less steep relative to the McCall segment (occurring over ~6 km) of the western Idaho shear zone (Kistler, 1974; Armstrong et al., 1977; Farmer and DePaolo, 1983; Giorgis et al., 2005b; Unruh et al., 2008).

A proposed tectonic model explaining the low fabric development and steep gradient in (87Sr/86Sr) assumes a rigid-body collision, transpressional kinematics, and an along-zone difference in the shear-zone strike orientation. Because of the variation in along-zone orientation, the displacement field, the fabric development, the (87Sr/86Sr) gradient, and the finite and infinitesimal strains differ for the Owyhee and McCall segments. For the Owyhee segment, the model predicts an oblique-convergence angle of 26°–39°, 65–94 km of convergent displacement, and 40–69 km of transcurrent displacement. For the McCall segment, the model predicts an oblique-convergence angle of 46°–59°, 65–94 km of convergent displacement, and 56–66 km of transcurrent displacement. The model predicts that both segments experienced pure-shear–dominated transpressional deformation, consistent with observed flattening fabrics and vertical lineations. The variation in orientation of the western Idaho shear zone may reflect the original orientation of the deforming zone boundary or the southern termination of the transpression zone.

APPENDIX

Calculating Magnitudes and Orientations of Finite Strain and Infinitesimal Strain Axes

Example: Owyhee segment, original width = 100 km.

Finite Strain

For the Owyhee segment with an original width of 100 km, the deformation matrix is:

\[
D = \begin{bmatrix}
1 & 0.50 & 0 \\
0 & 0.31 & 0 \\
0 & 0 & 3.23
\end{bmatrix}
\] (A1)

An initial width of 100 km results in the most strain compared to smaller initial widths (i.e., 71 km or 85 km) for the two segments. The tensor (DD^T) to make the matrix symmetric is:

\[
DD^T = \begin{bmatrix}
1 & 0.50 & 0 \\
0 & 0.31 & 0.50 \\
0 & 0 & 3.23
\end{bmatrix}
\] (A2)

which can be simplified to:

\[
DD^T = \begin{bmatrix}
1.25 & 0.16 & 0 \\
0.16 & 0.96 & 0 \\
0 & 0 & 10.43
\end{bmatrix}
\] (A3)

The eigenvalues for this matrix are: 1.27, 0.08, and 10.43. The square root of these values must be calculated to determine the actual eigenvalues because it is a tensor. The eigenvalues then become: 1.13, 0.27, and 3.23. The eigenvectors of A3 are:

\[
eigVC = \begin{bmatrix}
0.99 & -0.13 \\
0.13 & 0.99 \\
0 & 0
\end{bmatrix}
\] (A4)

From Equation A4, it can be concluded that finite strain for the Owyhee segment with an original width of 100 km is:

\[
S_1 = 3.23 \text{ and oriented } (0, 0, 1)
\]

\[
S_2 = 1.127 \text{ and oriented } (0.99, 0.13, 0)
\]

\[
S_3 = 0.276 \text{ and oriented } (-0.13, 0.99, 0)
\]

Infinitesimal Strain

For the Owyhee segment with an original width of 100 km and 100 increments, the deformation matrix is:

\[
D = \begin{bmatrix}
1 & 0.50 & 0 \\
0 & 0.31 & 0 \\
0 & 0 & 3.23
\end{bmatrix}
\] (A5)
For an increment of deformation, the matrix becomes (Tikoff and Fossen, 1993):

\[
D_{\text{acc}} = \begin{bmatrix}
(k^+)^{1/6} & A & 0 \\
0 & (k^-)^{1/6} & 0 \\
0 & 0 & (k^+)^{1/6}
\end{bmatrix}.
\]  \hspace{1cm} (A6)

where \( A \) is:

\[
A = \frac{\mu}{2\gamma} \left( 1 - \frac{\gamma^2}{6} \right)
\]  \hspace{1cm} (A7)

For the case of an original width of 100 km in the northern Owyhee Mountains, the straining increment is 100 km and 100 increments is: \( n = 100 \), where \( n \) is the number of increments. With these values substituted, the equation becomes:

\[
A = \frac{2.85 \times 10^{-6}}{100} (1 - (0.31)^2) \]  \hspace{1cm} (A8)

and simplifies to \( A = 0.0085 \). The variable \( A \), as well as the other variables, can now be substituted into Equation \( A6 \).

\[
\begin{bmatrix}
1.000 & 0.008 & 0 \\
0.008 & 0.976 & 0 \\
0.012 & 0.001 & 1.002
\end{bmatrix}.
\]  \hspace{1cm} (A12)

The eigenvalues for this matrix are: 1.003, 0.973, and 1.024. The square root of these values must be calculated to determine the actual eigenvalues because it is a tensor. The eigenvalues then become: 1.001, 0.986, and 1.012. The eigenvectors of \( A12 \) are:

\[
\begin{bmatrix}
0.954 & -0.301 & 0 \\
0.301 & 0.954 & 0 \\
0 & 0 & 1
\end{bmatrix}.
\]  \hspace{1cm} (A13)

From Equation \( A13 \), it can be concluded that infinitesimal strain for the Owyhee segment with an original width of 100 km and 100 increments is:

\[
\begin{bmatrix}
1.012 \\
0.986 \\
0.954
\end{bmatrix}
\]

From Equation \( A13 \), it can be concluded that infinitesimal strain for the Owyhee segment with an original width of 100 km and 100 increments is:

\[
\begin{bmatrix}
1.012 \\
0.986 \\
0.954
\end{bmatrix}
\]

\[
\sum_{1}^{2} = 1.001 \text{ and is oriented } (0.954, 0.301, 0), \\
\sum_{1}^{3} = 0.986 \text{ and is oriented } (-0.301, 0.954, 0).
\]

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