The Granite Problem as Exposed in the Southern Snake Range, Nevada

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Abstract. A geochemically and mineralogically diverse group of granitoids is present within an area of 900 km$^2$ in the southern Snake Range of eastern Nevada. The granitoids exposed range in age from Jurassic through Cretaceous to Oligocene and include two calcic intrusions, two different types of two-mica granites, and aplites. The younger intrusions appear to have been emplaced at progressively more shallow depths. All of these granitoid types are represented elsewhere in the eastern Great Basin, but the southern Snake Range is distinguished by the grouping of all these types within a relatively small area.

The Jurassic calcic pluton of the Snake Creek-Williams Canyon area displays large and systematic chemical and mineralogical zonation over a horizontal distance of five km. Although major element variations in the pluton compare closely with Daly's average andesite-dacite-rhyolite over an SiO$_2$ range of 63 to 76 percent, trace element (Rb, Sr, Ba) variations show that the zonation is the result of in situ fractional crystallization, with the formation of relatively mafic cumulates on at least one wall of the magma chamber. Models of trace element and isotopic data indicate that relatively little assimilation took place at the level of crystallization. Nonetheless, an initial $^{87}$Sr/$^{86}$Sr value of 0.7071 and $^{18}$O values of 10.2 to 12.2 permil suggest a lower crustal magma that was contaminated by upper crustal clastic sedimentary rocks before crystallization. The involvement of mantle-derived magmas in its genesis is difficult to rule out. Two other Jurassic plutons show isotopic and chemical similarities to the Snake Creek-Williams Canyon pluton.

Cretaceous granites from eastern Nevada that contain phenocrystic muscovite are strongly peraluminous, and have high initial Sr-isotope ratios and other features characteristic of S-type granitoids. They were probably derived from Proterozoic metasediments and granite gneisses that comprise the middle crust of this region.

Another group of granitoids (including the Tertiary aplites) show chemical, mineralogic, and isotopic characteristics intermediate between the first two groups and may have been derived by contamination of magmas from the lower crust by the midcrustal metasediments.

Introduction

Faure and Powell (1972, p 43) stated: "the crux of the granite problem is that it is possible that granitic rocks of igneous aspect can form as products of different processes..." Within an area of 900 km$^2$ in the southern Snake Range, White Pine County, Nevada, six igneous masses are well exposed in discrete outcrops. These granitoids range in age from Jurassic to Oligocene, in SiO$_2$ content from 63 to 76 wt.% in $^{18}$O from $-2.6$ to $+13.2$ permil, and in initial $^{87}$Sr/$^{86}$Sr from 0.7071 to 0.7157. The petrologic types exposed include two calcic (Peacock 1931) intrusions, two different kinds of two-mica granites, and aplites. The purpose of this report is to summarize and integrate the information in some 30 papers on the petrology, age relations, and systematic mineralogy of these rocks, and to speculate on the processes that may have been involved in their formation.

We first describe a well-studied pluton from the southern Snake Range using models of its trace element and isotopic composition to demonstrate the role of fractional crystallization in developing its strong compositional zonation. We then compare it with other plutons from the area and attempt to explain their origins from Sr and O isotope relationships.

In this report we refer to a classification of granites according to source materials (Chappell and White 1974). In the scheme, granites inferred to be derived from predominantly sedimentary protoliths are S-types, whereas those derived from igneous source materials are I-types. The chemical, mineralogical, and isotopic criteria used to distinguish between S- and I-type granites have been summarized by Chappell and White (1974), O'Neil and Chappell (1977), Beckinsale (1979), and Furgeson et al. (1980). As is apparent in the discussion to follow, application of these criteria to a particular pluton sometimes leads to ambiguous results.

The central part of the area studied is about 60 km southeast of Ely, Nevada, in the southern part of the Snake Range of eastern Nevada (Fig. 1). The granitoid rocks of this area intruded part of a lower Paleozoic miogeosynclinal sequence of predominantly quartzite and carbonate rocks; they are exposed east of the Mississippian Antler orogenic belt (Gilluly 1963, pp 139, 140) and just west of the Cretaceous Sevier orogenic belt (Armstrong 1968, p 436). None of these plutons has been deeply unroofed. With the possible exception of the Osceola intrusion, probably none of these granitoid masses has been eroded to a depth of more than 300 m. All of the plutons to be described (Fig. 2) are exposed beneath the Snake Range décollement. Parts or all of the area studied have been mapped by Drewes (1958), Misch and Hazzard (1962), Whitebread (1969), and Hose and Blake (1976).
Intrusive Types Exposed

Snake Creek-Williams Canyon Intrusion

Geology. The Middle Jurassic (155 ± 4 m.y.; Lee et al. 1983) pluton of the Snake Creek-Williams Canyon area is the oldest igneous mass present in the southern Snake Range. Within a horizontal distance of 5 km this well-exposed pluton grades from a tonalite (63% SiO₂) where the host rock is limestone to granite (76% SiO₂) where the host rock is quartzite. (The IUGS classification is used throughout this paper; Streckeisen 1973). When the large and systematic variation in the petrology of the pluton (Fig. 3) was recognized, it became apparent that this intrusive mass is ideal for a study of igneous processes. The intrusive rock is well exposed in contact with chemically distinct host rocks, and the chemical and mineralogical trends are developed over large enough distances to be defined with assurance, yet the distances are still small enough to allow detailed field study of the concentration gradients.

The eroded roof of this granitoid body probably was not much higher than the present outcrop. Thus the intrusion probably has not been unroofed to a depth of much more than 300 m. This intrusion is essentially domelike (Drewes 1958, p 231) and roughly concordant with bedding in the overlying sedimentary rocks. Both the general concordance and the most notable discordance are well shown where the pluton crops out at the crest of the range between Snake Creek and Williams Canyon. At the northern edge of this outcrop, beneath Pyramid Peak, the intrusive contact nearly parallels the flat bedding of the overlying quartzite, but at the southern edge of this outcrop, the intrusive contact cuts across about 300 m of quartzite bedding at a very high angle.

Locally, along steep contacts, the host rock has been dragged upward and appears to have been shouldered aside (Drewes 1958, p 231). This is apparent in the extreme east part of the intrusive outcrop, where Pioche Shale and Pole Canyon Limestone have been folded into an overturned syncline. However, just north of this shale outcrop, an apophysis of the intrusion has engulfed part of the limestone. This, as well as the discordant intrusive rock-quartzite contact at the crest of the range, indicates that some stoping took place.

The eastern half of the intrusive outcrop in Snake Creek drainage contains many xenoliths-fragments of Pioche Shale that are partly assimilated. The smallest of these is the size and shape of a walnut; the largest crops out over an area of about 9,000 m². In several places where the larger xenoliths are about a foot long, they comprise as much as 5% of the intrusion. These xenoliths tend to be ellipsoidal and to stand out on weathered surfaces. Strikes tend to parallel the nearest intrusive contact, and dips usually are within 20° of vertical. From this we infer that at least this part of the intrusive must have undergone magmatic flow...
before solidification. West of the 2.0% CaO isopleth, the intrusion is almost devoid of xenoliths.

There is little evidence of contact metamorphism around the intrusion exposed in the Snake Creek-Williams Canyon area. It is not uncommon for the limestone to be gnarled and deformed within a few meters of the intrusive contact, but there is virtually no skarn. Contact action on the Pioche Shale is no more impressive. Within a few feet of the igneous rock, the shale may be baked and recrystallized, but there is no indication of metamorphism. The intrusion’s effect on the Prospect Mountain Quartzite is difficult to study because the quartzite-intrusive contact is exposed only in a few places owing to the tendency of the quartzite to form talus over the intrusion. However, micropegmatitic apophyses were observed along bedding and jointing in the quartzite. North of the mouth of Williams Canyon on the west side of the range contact effects on the Osceola Argillite are more pronounced. In places the rock is changed to a spotted hornfels (Whitebread et al. 1962), but this metamorphism apparently was isochemical in terms of hand specimen-sized samples.

Chemical petrology. The sedimentary rocks exposed in intrusive contact with the granodiorite rocks (Fig. 3) are the Precambrian Osceola Argillite (of Misch and Hazzard 1962), the Precambrian and Cambrian Prospect Mountain Quartzite, and the Cambrian Pioche Shale and Pole Canyon Limestone. Isopleths based on analyses of more than 80 rocks show CaO values increasing from less than 1.0% where the intrusion is in contact with quartzite on the crest of the range to more than 4.0% where the eastern part of the intrusion is in contact with limestone and shale (Fig. 3). In the extreme western part of the same continuous igneous outcrop, where the granodiorite rock (samples 27, 28, 42 and 59) is in contact with Osceola Argillite, CaO again increases and is locally 2.0% or more, whereas SiO₂ contents decline to about 71%. However, because most of the intrusion in this area is altered, no attempt was made to contour CaO values there. Almost all samples contain normative corundum, but only four of the 86 analyzed specimens contain more than two percent normative corundum. By comparing other chemical variables with CaO, it can be shown (Lee and Van Loenen 1971, Fig. 4) that major element variations in this igneous mass compare closely with Daly’s average andesite-dacite-rhyolite (Barth 1962, p 164) over an SiO₂ range of 63–76%. Lee and Van Loenen (1971, p 29–31) give petrographic descriptions of five representative samples with SiO₂ contents ranging 63.5–74.9%.

Mineralogy. Pure fractions of most or all of the minerals (except quartz) present in each of 20 rocks were recovered for study. The results of analytical work on these mineral fractions are summarized in the following discussion; detailed information is given in the cited papers on the individ-
Fig. 4. General relations between CaO content and mineralogy for granitoid rocks of the Snake Creek-Williams Canyon area. (From Lee and Van Loenen 1971, Fig. 11)

The biotite crystallizing from the magma followed an iron enrichment trend, because oxygen fugacity decreased with temperature. These biotites indicate that the most mafic parts of the pluton began to crystallize at a temperature of 780°C and 10^{-13} atm fO_2, and that the most felsic parts finally crystallized at a temperature of about 735°C and about 10^{-13} atm fO_2 (Lee and Van Loenen 1970). The magnetite and/or ilmenite coexisting with these biotites indicate lower temperatures and lower fugacities of oxygen, probably due to differences in degree of subsolidus equilibration of different mineral phases (Lee and Van Loenen 1979).

Amounts and compositions of the feldspars present in these rocks vary with rock chemistry much as one would expect in a series of differentiates. Plagioclase feldspars range from andesine to oligoclase as bulk rock compositions become more silicic. Alkali feldspar is either absent or interstitial in trace amounts along the eastern margin of the intrusion. The alkali feldspars are maximum microcline microperthites, indicating temperatures perhaps below 400°C (Lee et al. 1980). As with the opaque oxides, lack of agreement between this temperature and those deduced from study of the coexisting biotites is probably to be explained by differences in degree of subsolidus equilibration.

Accessory minerals present respond to rock chemistry, indicating that their development was controlled by liquid crystal equilibria. Thus amount, crystal habit, and composition of both zircon (Lee et al. 1968) and apatite (Lee et al. 1973) change gradually from the mafic to the felsic parts of the exposure. The amount and composition of both allanite and monazite (Lee and Bastron 1967) and the amounts of sphene (Lee et al. 1969), epidote (Lee et al. 1971), and the opaque oxides also vary with rock composition.

The gradual and systematic change of accessory mineral types precipitated during crystallization is one of the most striking features of this pluton. The most mafic parts of the mass (63% SiO_2) contain 2 wt.% epidote, 1% sphene, 0.5% magnetite, and well-developed allanite, apatite, and zircon. This is an I-type assemblage. As SiO_2 increases, the amounts and types of accessory minerals change gradually, until the most felsic parts of the intrusion (76% SiO_2) contain only ilmenite, garnet, monazite, and trace amounts of apatite and zircon, a S-type assemblage. The entire suite of accessory minerals comprises only about 0.1 wt.% of the most felsic parts of the intrusive. About 90% of the rare earths present in the pluton are contained in the accessory minerals.

**Trace Elements and Isotopic Variations**

**General.** Since the large and systematic variation in the chemical petrology of this intrusion was first recognized, the essential question has turned on the relative importance of assimilation and fractional crystallization in the development of the pluton. Inasmuch as the mafic parts of the intrusion are in contact with shale and limestone, the most felsic parts with quartzite, and because the Ca-rich accessory minerals (epidote, sphene, apatite, and allanite) are strongly concentrated in the mafic parts of the intrusion, Lee and Van Loenen (1971) were led to the conclusion that assimilation played a dominant role. However, a reconsideration of this conclusion is presented here on the basis of trace element and isotopic data obtained since 1971.

**Fractional crystallization.** Elemental variations of Rb, Sr, and Ba (Lee and Doering 1974, 1980) are shown in Figs. 5–7, on which we will base our discussion of the various models for the crystallization and possible contamination of the pluton. First we will outline the major role of fractional crystallization, and then we will show that neither assimilation nor resitic control had much influence on the distribution of these elements.

Recent petrogenic studies of granitic rocks have shown the importance of cumulate and “mixed” cumulate-melt compositions in plutonic settings (e.g. McCarthy and Hasty 1976; McCarthy and Groves 1979; Bateman and Chappell 1979). Thus the trace element models described here show the variation in the compositions of the evolving liquid and of the instantaneous solids crystallized from the melt.

Assuming that Rb, Sr, and Ba all obey Henry’s law, their concentrations in a melt (C_l) that is evolving by fractional crystallization can be described by the Rayleigh equation:

\[ C_l = C_0 F^{(D - 1)} \]  \hspace{1cm} (a)

where \( C_0 \) is the original concentration of an element in the liquid, \( F \) is the fraction of liquid remaining, and \( D \) is the bulk partition coefficient. Since

\[ D = C_i / C_l \]  \hspace{1cm} (b)

where \( C_i \) is the concentration of an element in the instantaneous solid precipitating from the melt

\[ C_i = D C_0 F^{(D - 1)} \]  \hspace{1cm} (c)

As noted by McCarthy and Groves (1979), where significant changes in the liquidus mineral assemblage occur (and hence changes in \( D \)), the compositional trend of the solid is discontinuous; however, the evolving liquid composition shows only an inflection in its trend. In order to model the trace element variation of the pluton, a variety of initial
conditions \((C_o, D)\) were chosen and iteratively varied until computer-generated trends most closely approximated those of the observed trends. Even though these models are fairly sensitive to changes in \(C_o\) or \(D\), they remain approximate. Any composition that falls within the solid-liquid envelope can be explained by the model — keeping in mind that the rock compositions may be the product of solid-melt mixtures. Compositions that result from the entrapment of residual liquid would theoretically lie on tie lines joining points of equal \(F\) on the liquid and solid evolution curves.

Three discrete groups of samples from the Snake Creek-Williams Canyon pluton appear on the Rb-Ba diagram (Fig. 5). Group 1 has the lowest Rb concentration, moderate Ba, and CaO between 2.6 and 4.5 wt. %. Group 2 has high Ba (>1,400 ppm), moderate Rb, and CaO between 2.0 and 2.5%. Group 3 trends toward lower Ba at higher Rb concentrations and has CaO less than 2.0%. Samples from these groups, defined simply on the basis of CaO concentration, are shown with different symbols on all of the diagrams. Their spatial disposition in the pluton is shown in Fig. 3. If the chemical variation is the result of fractional crystallization, the sharp discontinuity in Ba concentration at about 100 ppm Rb must reflect the introduction of a new cumulate phase with a high \(D_{\text{Rb}}\), and also implies that samples from Group 1 and 2 are essentially cumulate in their nature. Modal data (Table 5, Lee and Van Loenen 1971) suggest that the precipitation of K-feldspar \((D_{\text{Kf}} \sim 6\), Crecraft et al. 1981) coincides with this discontinuity. Many samples from the eastern margin of the pluton lack K-feldspar and others contain only small amounts of interstitial K-feldspar that may have crystallized from trapped liquid. The trace element variation can be approximated by increasing \(D_{\text{Rb}}\) to 2 at \(F=0.8\) (Table 1). The relatively sudden appearance of K-feldspar in the cumulate mineral assemblage is indicated by the sharp demar-
Table 1. Summary of trace element models

<table>
<thead>
<tr>
<th>C&lt;sub&gt;o&lt;/sub&gt; (ppm)</th>
<th>D at F=1.0-0.8</th>
<th>D at F=0.8-0.1</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ba 1.150</td>
<td>0.9</td>
<td>2.0</td>
</tr>
<tr>
<td>Sr 400</td>
<td>2.0</td>
<td>2.0</td>
</tr>
<tr>
<td>Rb 145</td>
<td>0.5</td>
<td>0.5</td>
</tr>
<tr>
<td>Li 27-30</td>
<td>2.1</td>
<td>1.1</td>
</tr>
<tr>
<td>Cs 2-3</td>
<td>0.3-0.4</td>
<td>0.3-0.4</td>
</tr>
<tr>
<td>Zn 32-38</td>
<td>1.9-2.2</td>
<td>1.2</td>
</tr>
</tbody>
</table>

C<sub>o</sub> = Initial concentration in melt
D = Bulk distribution coefficient
F = Fraction of liquid remaining

Cation of the high-Ba group. Although reaching this important phase boundary greatly affects the modeled trend of the liquid (Fig. 5), it is not readily apparent in the composition of the samples in Group 3 which must be most like the liquid. The inflection may be obscured by mixed cumulate-melt compositions, and because most variation in liquid compositions occurs when F < 0.8. If the initial concentrations used are approximately correct, samples 23 and 28 (Fig. 3) may be like the original liquid that intruded into the upper crust.

The Blue Tier batholith, Tasmania, and the Loch Doon intrusion, Scotland, both show analogous Rb—Ba relations that can be interpreted as the result of the introduction of K-feldspar as a cumulus phase (McCarthy and Groves 1979; Tindle and Pearce 1981).

The Ba—Sr variation diagram (Fig. 6) shows groupings similar to those on the Rb—Ba projection. The variation can be approximated by a model that holds D<sub>sc</sub> constant at 2 and varies D<sub>lu</sub> from 0.9 to 2.0 at F = 0.80. Again the appearance of K-feldspar and the consequent increase in D<sub>lu</sub> seems to have occurred over a short interval of crystallization. The almost linear correlation of Ba and Sr in group 3 samples shows that D<sub>lu</sub> ≈ D<sub>sc</sub>; on this plot, samples with a pure liquid composition are indistinguishable from pure cumulates if formed at F < 0.5. Apparently the observed decrease in the proportions of plagioclase and epidote in the cumulates is approximately balanced by the increase in alkali feldspar as no marked changes in D<sub>sc</sub> appear to be necessary to explain the observed trace element variation.

The Rb—Sr variation diagram (Fig. 7) does not show the pronounced groupings of the other two diagrams, presumably because D<sub>rb</sub> and D<sub>sc</sub> remained fairly constant.

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Fig. 6A–D. Barium-strontium variation diagrams (in ppm) for the Snake Creek-Williams Canyon intrusion (Lee and Doering 1974, 1980). Explanations as for Fig. 5A–D. For clarity the liquid evolution curve is displaced upwards.
Neither U nor Th values lend themselves to simple modeling due to marked scatter. The concentrations of both of these elements generally increase from Groups 1 to 3. Total variation, including the cumulates, is about 1–3 ppm for U and 10–20 ppm for Th. The pronounced scatter is probably due to post-magmatic processes and to variations in the concentrations of U- and/or Th-rich accessory minerals (allanite, monazite, and zircon). Relatively high (4 to 7) Th/U ratios are found in samples from Groups 1 and 2, and, if magmatic, indicate that a decrease in the Th/U ratio of a melt is likely if it evolves by fractional crystallization.

If these trace element models are correct, they imply that samples with melt, cumulate, and melt/cumulate chemistries were sampled over a range of F from 1.0 to 0.25. Group 1 samples appear to be solids (with some trapped liquid) formed between F = 1.0 to 0.8, when alkali feldspar was not a significant crystallizing phase. Group 2 appears to have formed when F was between 0.8 and 0.75, and there was cumulus alkali feldspar. Although dominated by cumulate compositions, the scatter along a trend toward liquids formed at the same F suggests that some samples of this group contain as little as 50 percent cumulus material. Group 3 is a diverse assemblage with both melt and cumulate chemistries. It contains samples similar to everything from the original liquid (F = 1.0) to the most evolved liquid (F ≈ 0.25). However, most samples of this group appear to contain a significant cumulate component, judging from their displacement toward the cumulate trend in both the Rb–Sr and Rb–Ba diagrams.

The evolving nature of the cumulus minerals is indicated by the modal data (Fig. 4). On the other hand the analysis to follow assumes that the relative amounts of cumulus minerals remained constant over major periods of crystallization, and thus only gross averages of the mineral proportions can be calculated from the trace element data. Using the D's determined from the modeling of the latter portion of the differentiation and mineral D's from Hanson (1978), equations of the form

$$D_i = D^T_i(X_\alpha) + D^F_i(X_\beta) + \ldots$$

where \( \alpha \) and \( \beta \) are mineral phases, and \( X_\alpha \) and \( X_\beta \) are their weight fractions, can be defined. For Sr, Ba, and Rb the only important mineral phases are biotite and alkali and plagioclase feldspar. Epidote is important to the Sr budget of the mafic minerals, but its contribution to the bulk partition coefficient is small and is ignored here, based on \( D_{Sr}^T \approx 5 \) (Lee et al. 1971), and the fact that it comprises...
a maximum of 2 wt.% of some samples. Thus three equations of this type can be generated:

\[
\begin{align*}
D_{Sr} &= 2.39(X_{AF}) + 4.4(X_{PL}) + 0.12(X_{BIOT}) \\
D_{Ba} &= 2.61(X_{AF}) + 0.3(X_{PL}) + 6.4(X_{BIOT}) \\
D_{K} &= 0.5(X_{AF}) + 0.4(X_{PL}) + 3.3(X_{BIOT})
\end{align*}
\]

where \(X_{AF}, X_{PL},\) and \(X_{BIOT}\) are the weight fractions of alkali feldspar, plagioclase, and biotite. The individual mineral D's have large uncertainties associated with them, further complicating the exact solution of the equations. This problem can be visualized graphically. If the last term in equation (d) is allowed to go to zero, the equation can be represented as shown in Fig. 8, which plots the weight fraction of plagioclase versus that of alkali feldspar. Thus, if no alkali feldspar were crystallizing from the melt, \(X_{PL} = 0.45.\) However, even moderate (10%) changes in \(D_{Sr}\) allow \(X_{PL}\) to range from 0.4 to 0.5. Such variations in \(D_{Sr}\) could occur over the course of crystallization and are certainly within the range of measured \(D_{Sr}\) in various silicic rocks. Simultaneous solutions of equations (e) and (f) with the elimination of the \(X_{BIOT}\) term produces another equation that can be plotted in this projection. Intersection of these curves yields an area of approximate solution to the equations for the latter stages of crystallization -- \(X_{AF} = 0.22 - 0.26 - 0.34,\) \(X_{AF} = 0.16 - 0.20 - 0.24\) (preferred solutions in parentheses). From equations (d) and (f) \(X_{BIOT} \approx 0.11,\) and by difference \(X_{QUARTZ} + X_{ACCESSORIES} \approx 0.40.\) This result is in agreement with the model data (Fig. 4) for samples with \(2.0\) to \(2.5\% CaO (Group 2),\) which appear to be cumulates formed while alkali feldspar was crystallizing.

Estimates of the composition of the cumulate assemblage from the early sequence can be obtained by setting \(X_{AF} = 0.\) In this case \(X_{PL} = 0.4 - (0.45) - 0.5, X_{BIOT} = 0.11 \text{ to } 0.20\) (depending critically on \(D_{Ba}^{\text{Bi}}\)), and \(X_{QUARTZ} + X_{ACCESSORIES} \approx 0.40.\) Again this solution is in reasonable agreement with the modal mineralogy of Group I samples, which appear to be early cumulates.

**Table 2. Compositional variety in the Snake Creek-Williams Canyon pluton**

<table>
<thead>
<tr>
<th>Sample no.</th>
<th>1</th>
<th>23</th>
<th>48</th>
<th>84</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>76.9</td>
<td>74.7</td>
<td>71.1</td>
<td>63.0</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.08</td>
<td>0.21</td>
<td>0.39</td>
<td>0.78</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>13.6</td>
<td>13.8</td>
<td>15.1</td>
<td>18.2</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>0.59</td>
<td>1.8</td>
<td>2.5</td>
<td>5.4</td>
</tr>
<tr>
<td>MnO</td>
<td>0.02</td>
<td>0.06</td>
<td>0.06</td>
<td>0.05</td>
</tr>
<tr>
<td>MgO</td>
<td>0.17</td>
<td>0.39</td>
<td>1.2</td>
<td>2.1</td>
</tr>
<tr>
<td>CaO</td>
<td>0.48</td>
<td>1.8</td>
<td>2.4</td>
<td>4.4</td>
</tr>
<tr>
<td>Na₂O</td>
<td>3.7</td>
<td>3.4</td>
<td>3.7</td>
<td>3.8</td>
</tr>
<tr>
<td>K₂O</td>
<td>4.5</td>
<td>3.7</td>
<td>3.3</td>
<td>2.3</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.01</td>
<td>0.06</td>
<td>0.12</td>
<td>0.22</td>
</tr>
<tr>
<td>Total</td>
<td>100</td>
<td>100</td>
<td>100</td>
<td>100</td>
</tr>
</tbody>
</table>

1 = Evolved sample from Group 3
23 = Similar to parent liquid composition
48 = Sample from Group 2
84 = Sample from Group 1

* From Lee and Van Loenen (1971, Table 5), recalculated volatile -- free to 100 percent, with total Fe as Fe₂O₃

\[X_{ACCESSORIES} = 0.3 \text{ to } 0.4.\] Again this solution is in reasonable agreement with the modal mineralogy of Group I samples, which appear to be early cumulates.

**Composition of Original Liquid**

The initial composition of the liquid used in the trace element models is shown in Table 1. If this composition is used as a criterion the initial liquid appears to have had a CaO content of 1.8 to 2.0 wt.%. Inflections or discontinuities in several chemical (K₂O, Na₂O, P₂O₅, Ba) and mineralogic (modal quartz and mica) parameters occur at this point (Lee and Van Loenen 1971). These observations lend credence to the suggestion (Cain 1974) that the initial magma was fairly silicic. Thus sample 23 of Lee and Van Loenen (1971) appears to be most like the parent liquid of the intrusion. Its composition is shown in Table 2 and is compared with the compositions of samples from Groups 1 and 2 and with one of the most evolved rocks in the intrusion. This comparison clearly shows the dangers of assuming that individual samples from granitic plutons are compositionally similar to the liquids from which they crystallized. These considerations also show that it is totally inappropriate to model the evolution of plutonic rocks as liquids, especially if the sample suite is small. The major element variations in this cumulate-dominated intrusion are similar to those in the classic andesite-dacite-rhyolite series, but appear to have been produced by crystallization from a liquid with greater than about 73% SiO₂.

**Resistite Control**

White and Chappell (1977) suggested that the geochemical variation observed in some plutons results from the unmixing of refractory phases residual from a partial melting event (resistite) and a liquid. Simple mixing curves representing this process connect refractory marginal compositions of the pluton with a high-SiO₂ phase (representative of what could be a minimum melt) in Figs. 5-7. Strong deviations of the data away from these curves show that "resistite" separation was not the dominant factor in the geochemical
evolution of the Snake Creek-Williams Canyon pluton. However, solid-liquid mixing was an important process as the incomplete separation of liquids from crystallizing solids appears to have been important.

Assimilation

The composition (Ba, Rb, Sr) of the parent liquid was used in an attempt to place limits on the amount and type of assimilation of country rock that may have occurred. DePaolo (1981) derived equations that describe the trace element and isotopic evolution of liquids affected by simultaneous assimilation and fractional crystallization in terms of the parameter r, which is the ratio of the mass assimilation rate to the mass crystallization rate. Figs. 5-7 show the results of these calculations for both simple mixing and combined assimilation-fractional crystallization. The liquid and three compositions representative of the sedimentary section at the level of the intrusion were employed in the calculations. Major and trace element analyses of these sediments were presented by Lee and Van Loenen (1971) and Lee and Doering (1974, 1980).

In the Rb-Ba diagram it is apparent that simple assimilation of any of the country rocks cannot duplicate the spectrum of compositions observed. It is especially difficult to explain Group 2 by simple assimilation. However, if fractional crystallization occurred simultaneously with assimilation the trace element trends are not markedly different from those produced solely by fractional crystallization. The scatter of the data is such that it would be difficult to rule out assimilation rates that yield values of r ≤ 0.2.

The effects of assimilation on the Ba-Sr evolution of the pluton are analogous to those in Rb-Ba space. Considered in isolation from the other trends, simple assimilation of ~10% of the Pole Canyon Limestone could explain the composition of Group 1 samples, but this mechanism does not explain the Rb-Ba variation. The amount of assimilation during fractional crystallization is not easily limited. However, in the case of limestone assimilation, r ≤ 0.05 to eliminate “buffering” the Sr content of the melt to noticeably higher levels than are observed.

The trend of the limestone mixing curves in the Sr-Rb diagram is inconsistent with simple mixing to produce the Group 1 samples. Assimilation of even 8 percent of the Pole Canyon Limestone would raise the CaO content of the contaminated melt to the amount observed in the marginal portions of the pluton but would not provide sufficient strontium to explain the strontium abundance in the Group 1 samples. However, assimilation during fractional crystallization can only be limited to r's of less than 0.2.

Thus it appears that simple contamination did not play an important role in producing the observed trace (Rb, Sr, Ba) element variation in the Snake Creek-Williams Canyon intrusion. These observations do not rule out bulk contamination and rehomogenization of the magma before crystallization commenced at the level of exposure. Moreover, it is difficult to rule out assimilation if it occurred during fractional crystallization. Possible values for the parameter r range from 0.05 for the Pole Canyon Limestone to 0.2 for the Pioche Shale or the Prospect Mountain Quartzite.

A consideration of the isotopic variation across the pluton provides a more rigorous test of the relative importance of assimilation and fractional crystallization. A large variation in δ18O (10.2-12.2 permil) accompanies the chemical and mineralogical trends described for this pluton. The Carich eastern margin of the intrusion is depleted in 18O, and the more evolved western portion is enriched in 18O (Fig. 9). As noted by Lee et al. (1982), the isotopic trends are inconsistent with simple wall rock contamination of the magma at the level of exposure, for high-δ18O carbonates are the wall rocks along the eastern margin where the intrusion has the lowest values of δ18O.

The rubidium and strontium isotope systematics also tend to rule out substantial contamination of the magma during fractional crystallization (e.g. Taylor 1980). Equations for quantifying these effects were presented by DePaolo (1981), and results of such calculations for the Snake Creek-Williams Canyon intrusion are shown in Fig. 10. The assimillaint was assumed to be similar to the Osceola Argil- lite and the Prospect Mountain Quartzite which are intruded by the western part of the pluton. The calculated curves show that large variations of 87Sr/86Sr accompany modest changes in δ18O. If the O isotope variation is considered to be caused solely by contamination, 30 to 50% of the crystallized mass must have been assimilated. This would dramatically alter the Rb-Sr systematics of the intrusion, producing “ages” of 30 to 40 m.y. at the time of crystallization, along with slightly curved pseudoisochrons. The isochronous nature of the Rb-Sr system of the intrusion (Lee et al. 1983, Fig. 3), and its agreement with other isotopic ages (zircon U-Th-Pb, Lee et al. 1968, and Lee et al. 1981; hornblende K-Ar, Lee et al. 1970) apparently limit r to values of less than 0.05. Nevertheless, three samples from the extreme western margin of the exposure fall above the Rb-Sr isochron and appear to have been contaminated by the Osceola Argillite. The δ18O variation is limited to less than 1.0 permil by this value (0.05) of r (Fig. 10) for the values of F cited earlier. These results imply that fractional crystallization of the intrusion may have produced much of the observed 18O variation. The fractionation of oxygen isotopes follows a Rayleigh distillation law of the form

\[ R_i = R_0 F^{(i-1)} \] (a)
Fig. 10. $\delta^{18}O$ and $^{87}Sr/^{86}Sr$ shifts calculated for the case of combined assimilation – fractional crystallization. The initial magma was assumed to have $\delta^{18}O = 10$, $^{87}Sr/^{86}Sr = 0.707$, 400 ppm of Sr, and 150 ppm of Rb. The contaminant was assumed to be similar to a composite of the Pioche Shale and the Prospect Mountain Quartzite, with $\delta^{18}O = 13$, $^{87}Sr/^{86}Sr = 0.750$, 40 ppm of Sr, and 225 ppm of Rb. $D_{O}$ was assumed to be 2, $D_{Sr}$ was assumed to be 1, and O concentrations were taken to be equal in the contaminant and in the initial magma. The rapid change of $^{87}Sr/^{86}Sr$ with small changes in $\delta^{18}O$ indicate that only small amounts of assimilation occurred during fractional crystallization. The tick marks indicate values of $F$. On the curve representing $r = 0.5$, these tick marks are labeled, but most of the labels have been omitted from the other mixing curves for the sake of simplicity.

Fig. 11. Plot of $\delta^{18}O$ versus $F$ (the fraction of liquid remaining in a crystallizing magma), showing the progressive change of $\delta^{18}O$ for $\alpha = 0.999$ and $\alpha = 0.9999$. Curves for the liquid and instantaneous cumulates are shown. The initial magma ($F = 1$) was assumed to have a $\delta^{18}O$ of 11.

where $R_{O}$ is the $^{18}O/^{16}O$ ratio ($\delta^{18}O + 1.000$) of the original liquid, $R_{L}$ is this ratio in the evolving liquid, and $\alpha$ is the fractionation factor ($R_{\text{crystal}}/R_{\text{liquid}}$). It follows that $R$ of the cumulates ($R_{C}$) can be calculated from

$$R_{C} = \alpha R_{O} e^{(\alpha - 1)}$$  \hspace{1cm} (b)

At magmatic temperatures $\alpha$ probably varies from $\geq 0.999$ to $\leq 0.9999$. Both values for $\alpha$ were used and the results are shown in Fig. 11. A value of 11.0 permil was chosen for the initial $\delta^{18}O$ of the liquid. This choice implies $\alpha \approx 0.999$ to produce the $^{18}O$-depleted cumulates but seems reasonable based on the nature of the biotite-, magnetite-, epidote-, and sphe-ric-rich eastern margin and the $\delta^{18}O$ values measured for these minerals (Lee et al. 1982).

In the case where $\alpha = 0.999$, $\delta^{18}O$ of the cumulate varies from $\approx 10$ to 12.5 as the liquid varies from 11 to 13.5 permil. The curves clearly show that for $\alpha = 0.999$ the entire $\delta^{18}O$ variation could occur in the cumulates with $F$ ranging from 1 to 0.1. But if, as shown earlier, the intrusion contains samples of "liquids" and "cumulates", the entire $\delta^{18}O$ range could be produced over the apparent range of $F$ (1 to 0.25) derived from the trace element models.

In the case where $\alpha = 0.999$ only small changes in $\delta^{18}O$ occur and cumulates and liquids are only slightly different in composition. Apparently $\alpha$ was closer to 0.999.

**Rare Earth Elements**

The rate earth element (REE) patterns for six samples (Cain 1974) from the pluton are shown in Fig. 12. Although some systematic variation of the patterns correlate with rock chemistry, it is difficult to model the REE evolution of the crystallizing pluton. This is especially true because of the high proportion of REE-rich accessory minerals in the cumulate portions of the pluton: apatite, allanite, and sphene, and epidote constitute more than 3 wt. % of some rocks. Phene contains a significant proportion (> 50%; Cain 1974) of the total REE content of these samples. The rapid change in the modal amounts of these accessories (Fig. 4;
Lee and Van Loenen 1971, Fig. 10) and their high but generally unknown partition coefficients for REE’s make quantitative interpretations difficult with this small set of samples. All of the samples analyzed except number 15 appear to be cumulates, based on their Rb—Sr—Ba relationships.

Nonetheless, some inferences can be drawn from the variations in the REE patterns. From its position on the Rb—Sr—Ba diagram, sample 15 appears to be the closest (of those analyzed for REE) to the original composition of the parent liquid. The presence of a negative Eu anomaly suggests but does not demand equilibration with a feldspar-bearing or Eu-depleted source, indicative of an origin within the crust. Even though sample 6 also has a negative Eu anomaly, its position on the Rb—Ba—Sr diagrams suggests that it is a cumulate, albeit from a relatively evolved liquid. Sample 22, also a Group 3 cumulate, has a slight positive Eu anomaly, consistent with its cumulate nature. Samples 66, 78, and 86 all are Group 1 samples, early cumulates. Comparison of their patterns with that of sample 15 shows that the early cumulates were probably LREE-enriched and HREE-depleted compared to the magma from which they crystallized. Thus the evolving liquid must have become LREE-depleted and HREE-enriched. Sample 6, though itself a cumulate, shows signs of having crystallized from such a melt. This type of REE variation is considered by some to be diagnostic of liquid-state differentiation (Hildreth 1979, and others), but it is seen here to be a natural consequence of crystal fractionation and the formation of cumulates along the margins of a pluton. The mineralogy of Group 1 samples is clearly consistent with this interpretation, for these samples contain large amounts of the strongly LREE-enriched phases allanite (Lee and Bastron 1967) and epidote (Stuckless and Miesch 1981).

The REE pattern of sample 22 shows that by the time it crystallized the melt had been substantially depleted in LREE, or that the concentrations of LREE-rich phases in the cumulate had decreased. Both factors were probably important as a result of the liquid evolution described above and the disappearance of modal sphene and allanite and greatly reduced amounts of epidote (Fig. 4; Lee and Van Loenen 1971, Fig. 10). Sample 6 crystallized from a relatively evolved liquid after monazite had begun to crystallize from the magma and hence it may have elevated REE concentrations because of the extremely high concentrations of REE in this phase (Lee and Bastron 1967).

In summary the trace element and isotopic data suggest that while there may have been bulk contamination and rehomogenization of the magma at depth, there was little contamination during crystallization. This is apparent from the amounts of Rb, Sr, and Ba present in various parts of the pluton, and especially from the oxygen and strontium isotopic variations across the pluton. From a modeling of these parameters on variation diagrams it is clear that fractional crystallization played a major role, and that neither assimilation nor restite control had much effect on the trace element (Rb, Sr, Ba) distribution or on the oxygen and strontium isotope variations observed. However, it is difficult to rule out assimilation completely if it occurred during fractional crystallization. The value of the parameter r may have been as large as 0.05 for the assimilation of sedimentary rocks at the level of exposure. In spite of relatively high $\delta^{18}O$-values (> 10 permil), the mineralogy and chemistry of the pluton indicate that it was not derived solely from a metasedimentary protolith.

Osceola Intrusion

The area of intrusive rock referred to as the Osceola stock by Armstrong (1966) is shown in Fig. 2. Preliminary study indicates that this granitoid rock is similar to the Snake Creek-Williams Canyon intrusion described above. Both plutons become more felsic from east to west. The Osceola stock has $\delta^{18}O$ values (9.6–9.9 permil) somewhat lower than those (10.2–12.2 permil) determined for the Snake Creek-Williams Canyon intrusion (Lee et al. 1982), but rubidium-strontium isotopic results for the two plutons are very similar (Lee, Kistler and Robinson 1983). Data for the Osceola stock give an age of 145 m.y. with an initial $^{87}Sr/^{86}Sr$ ratio of about 0.7075. The results of U—Pb age work on zircon recovered from sample 177-MW-60 (Fig. 2) are complex but show signs of an inherited component and only an approximate age of 160–170 m.y. (Lee, Stacey and Fischer 1983).

Willard Creek Intrusion

The Willard Creek intrusion is one of the equigranular two-mica granites of northeastern Nevada discussed by Lee et al. (1981), who regarded it as a felsic phase of the Osceola intrusion (Fig. 2) just described. However, recent rubidium-strontium results for sample 295-DL-64 (Lee, Kistler and Robinson 1983) suggest that the Willard Creek mass may not be comagmatic with either the Osceola intrusion or the Snake Creek-Williams Canyon pluton, even though all three are Middle Jurassic in age. Muscovite recovered from sample 295-DL-64 gave a K—Ar age of 151 ± 4 m.y. (Lee et al. 1970, Table 1), very similar to the Rb—Sr ages determined for the Osceola intrusion (~ 145 m.y.) and the Snake Creek—Williams Canyon intrusion (155 ± 4 m.y.). Sample 295-DL-64 gave an initial $^{87}Sr/^{86}Sr$ ratio of 0.7086 (± 0.0002) for the Willard Creek mass at an age of 151 m.y. While this value is somewhat higher than the initial ratios determined for the plutons of the Osceola and Snake Creek-Williams Canyon areas, it is appreciably lower than the values (0.7111 to 0.7250) determined for four other two-mica granites in northeastern Nevada. This would indicate that the Willard Creek pluton was generated from less radiogenic source materials than the other two-mica granites in the area. On the other hand the $\delta^{18}O$ value of sample 295-DL-64 was found to be 13.2 permil (Lee et al. 1981), and such a high value suggests that sedimentary materials were an important component in the generation of this two-mica granite.

In addition to its relatively low initial $^{87}Sr/^{86}Sr$ value of 0.7086, the Willard Creek mass differs from most other two-mica granites of northeastern Nevada in that it contains primary magnetite, usually regarded as characteristic of an I-type granite. Nonetheless, biotite from a single sample has more than 3 Å“ ions per 22 anions (Lee et al. 1981), similar to those from strongly peraluminous granites from northeastern Nevada and elsewhere.

Pole Canyon-Can Young Canyon Intrusion

The igneous mass exposed in the Pole Canyon-Can Young Canyon area is roughly elliptical in outline and domal in form. This intrusion consists of two phases that are subequal in outcrop area — a “host” intrusion and late aplite and pegmatitic dikes as much as six feet wide. Where these
Fig. 13. Chondrite-normalized REE patterns for a samples from the Pole Canyon-Can Young Canyon intrusion (Cain 1974) and from a two-mica granite of Cretaceous age from the Ruby Mountains, Nevada (Kistler et al. 1981).

dikes occur in swarms they all but obliterate any vestige of the host intrusion. The mineralogy of these apalites and pegmatites is simple. Except for quartz, muscovite, and alkali feldspars, garnet is virtually the only mineral seen in the field. Fluorite is a common but minor accessory mineral. In addition, careful mineral separation work on a few apalites has revealed the presence of exceedingly minute amounts of garnite, columbite-tantalite, and two unidentified minerals. A few beryl crystals as much as one cm across appear to be confined to a single aplite.

The host intrusion of the Pole Canyon-Can Young Canyon area is an unusual muscovite-phenocrystic (Lee et al. 1981) two-mica granite to granodiorite that was described in detail by Lee and Van Loenen (1971, pp 5, 38, 39). This rock is distinguished in part by the presence of muscovite phenocrysts as large as two cm across, which contain small inclusions of euhedral biotite. It is not uncommon to find as many as 20 biotite crystals included within a single large grain of muscovite. Petrographically, muscovite and biotite, both of which are very fresh looking, are seen to be asso-
ciated and often intergrown. They appear to have formed in equilibrium. The muscovite contains no opaque inclusions, nor does it present any other evidence that it is secondary after biotite. Indeed the rocks are devoid of magnetite and ilmenite, these minerals being undetected not only in thin section but also during systematic mineral separation work.

Chemical and modal analyses and norms of nine samples of this muscovite-phenocrystic two-mica granite are presented by Lee and Van Loenen (1971), who show that the samples contain 2.8–3.4 percent normative corundum, and each is strongly peraluminous (molecular \( \text{Al}_2\text{O}_3 \geq [\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O}] \)). The REE pattern (Fig. 13) of one sample from the Pole Canyon-Can Young Canyon intrusion shows a relatively high (La/Yb ratio ~40) and only a small Eu anomaly (Cain 1974). A two-mica granite from the Ruby Mountains (Kistler et al. 1981) has about the same amounts of HREE, but much lower LREE concentrations (20X chondrite).

The biotites present with muscovite in these rocks contain only about 6.0 percent MgO, leading Lee and Van Loenen (1970) to suggest that they may have initially equilibrated with respect to a relatively low oxygen fugacity. This inference is further substantiated by the complete absence of opaque oxides already noted. The only accessory minera-
als present in these rocks are monazite, an allanite-like mineral, and epiapatite containing many tiny inclusions of acicular zircon.

The \( ^{187}\text{O} \) values for nine samples of this two-mica granite range from 10.6 to 12.1 permil and the minerals recovered from one of these samples appear to be in isotopic equilibrium (Lee et al. 1982). Rubidium-strontium data for four whole-rock samples showed a scatter not unusual for muscovite granites. The \( 87\text{Rb}/86\text{Sr} \) and \( 87\text{Sr}/86\text{Sr} \) values for the whole rocks were therefore averaged to 1.44 and 0.71736, respectively. A regression (York 1969) of this value for the whole-rock specimens, and data for a muscovite, a microcline and a aplite-pegmatite gave an initial \( 87\text{Sr}/86\text{Sr} \) ratio of 0.71574 ± 0.0003 at an age of 79.1 ± 0.5 m.y. (Lee, Kistler and Robinson 1983).

The data summarized suggest that the muscovite-phenocrystic two-mica granite of the Pole Canyon-Can Young Canyon area is a S-type granite, derived from a metasedimentary protolith. Although this granite is unique among the igneous types present in the southern Snake Range, its counterpart, the Tungstonia Granite of Best et al. (1974) is exposed in the Kern Mountains, about 83 km to the north, and a similar though not identical muscovite-phenocrystic granite is present in the Toana Range, about 225 km to the north. Lee et al. (1981) concluded that all three of these muscovite-phenocrystic two-mica plutons resulted from anatectic of Precambrian metasediments under conditions of relatively low oxygen fugacity, and Lee, Stacey and Fischer (1983) found that this anatexis probably took place during the Late Cretaceous, with an indicated source rock age of 1,970 ± 330 m.y. Such an origin would be consistent with the locations and strongly S-type nature of these very unusual rocks.

**Lexington Creek Intrusion**

The map of Whitebread (1969) shows the Lexington Creek intrusion truncated by a thrust fault; Lee et al. (1980) sug-
gest that the foliation noted by Whitebread and by Drewes (1958) may have resulted from movement on the thrust before the pluton was completely crystallized. The original mineralogy of this pluton has been somewhat obscured by deuteric alteration that may have accompanied development of the foliation. However, it is apparent that the unaltered intrusion was a two-mica granite in which the biotite and muscovite were present as discrete crystals. Thus its texture contrasts with that of the Pole Canyon-Can Young Canyon pluton, which has biotite euhedra present within muscovite phenocrysts.

A single analysis of zircon recovered from sample 378 suggests that the Lexington Creek pluton is Cretaceous in age and was derived at least in part from a preexisting crustal source (Lee, Stacey and Fischer 1983). The relatively high \( ^{187}\text{O} \) values of 10.4 and 10.6 permil (Lee et al. 1982) determined for samples GR47 and GR48 (Fig. 2) are consistent with this idea. Three muscovites recovered from this pluton gave K–Ar ages of 77.9, 82.2, and 86.3 m.y. (Lee et al. 1980). The 86.3 m.y. age is assumed to be most nearly representative of the time of emplacement of this intrusion, because there may have been partial degassing of the analyzed micas as a result of heat stress related to late movement along the overlying thrust fault that truncates the intrusive mass. Using 86 m.y. as the emplacement age of this pluton results in a calculated initial \( 87\text{Sr}/86\text{Sr} \) of 0.7114.
Young Canyon-Kious Basin Intrusion

Most of the granitoid rock exposed in the Young Canyon-Kious Basin area (Fig. 2) has been catastically deformed as a result of stresses related to late movement on the Snake Range decollement. The cataclastic part of this pluton is separated from the undeformed part by a fault that probably represents the location of the Cambrian Pioche Slate before the pluton was emplaced (Lee and Van Loenen 1971, Pl. 1). Biottites recovered from two samples from the undeformed part of the pluton, and muscovite from spatially related aplite gave K – Ar ages of 29.9, 30.9, and 34.0 m.y. respectively. Concentrates of secondary muscovite recovered from the cataclastic portion of the intrusion gave K – Ar ages of 17–18 m.y., presumably the time of deformation and thus of the most recent movement on the Snake Range decollement (Lee et al. 1970).

The δ¹⁸O values determined for three samples of the undeformed portion of the intrusion are 8.7, 9.8, and 10.0 permil, among the lowest of unaltered plutonic rocks from this area, but still relatively high. However, most samples of the cataclastic portion have very low (to −2.6 permil) δ¹⁸O values. In general the most intensely deformed samples have the lowest δ¹⁸O values, leading Lee, Friedman and Gleason (1983) to the conclusion that this pluton underwent oxygen isotope exchange with meteoric water at the time of cataclasis.

The cataclasite of the pluton complicates interpretation of the Rb–Sr systematics. Lee, Kistler and Robinson (1983) report an age of 37.4 ± 1.5 m.y. and an initial ⁸⁷Sr/⁸⁶Sr ratio of 0.7120 (± 0.0001), using two cataclasized samples and one undeformed sample. Including the analysis of a spatially associated aplite, the age is 28.3 ± 1.5 m.y., with an initial ⁸⁷Sr/⁸⁶Sr ratio of 0.7126 ± 0.0004), in approximate agreement with the K – Ar ages of the aplite and the undeformed parts of the intrusion. In either case the high initial ⁸⁷Sr/⁸⁶Sr ratio of this Oligocene intrusion and the relatively high δ¹⁸O values indicate that some metamorphic materials were involved in its generation.

The undeformed portion of the Young Canyon-Kious Basin intrusion is an equigranular two-mica granite. Accessory minerals include allanite, monazite, magnetite, zircon, ilmenite, and very sparse sphe, apatite, and spessartine-garnet. The biotites are similar in composition to those in the felsic part of the Middle Jurassic pluton of the Snake Creek-Williams Canyon area, with 2.9 to 3.0 A¹⁰³ per 22 anions and Fe²⁺/Fe³⁺ + Mg of about 0.53 (Lee and Van Loenen 1970). The biotite compositions and the presence of spinel and magnetite suggest that the pluton crystallized under fairly oxidizing conditions. All of the samples analyzed are concordant normative. Undeformed samples have 0.9 to 1.5% normative corundum, much less than the strongly peraluminous Pole Canyon-Can Young Canyon intrusion.

The mineralogy of the Oligocene aplitic is similar to that of the older aplites of the Pole Canyon-Can Young Canyon area, with muscovite, garnet, and fluorite in addition to quartz and feldspars. The aplites are also corundum normative (1.5 to 3.0%), and two samples have δ¹⁸O values (7.4 and 10.3 permil) similar to those found for samples of the undeformed mass in Young Canyon-Kious Basin.

Summary

Some features of the granitic plutons in the southern Snake Range are summarized in Table 3. Magmatic activity began with the emplacement of three plutons of Middle Jurassic age in the western part of the study area (Fig. 2). These are the equigranular two-mica granite of the Willard Creek area and the calcic (Peacock 1931) intrusions of the Osceola and Snake Creek-Williams Canyon areas. The calcic plutons have initial ⁸⁷Sr/⁸⁶Sr ratios of 0.7075 and 0.7071, respectively. These are the lowest initial ratios determined for any of the intrusive rocks of the southern Snake Range. The initial ratio of the Willard Creek pluton is somewhat higher at 0.7085, but this value is low compared with those of younger two-mica granites of the area.

After a period of quiescence lasting about 75 m.y., intrusive activity resumed with the emplacement of the Late Cretaceous two-mica granites of the Pole Canyon-Can Young Canyon and Lexington Creek areas, somewhat to the east of the three Middle Jurassic plutons. Although these two-mica granites appear in outcrop only ten km apart and each crystallized from a relatively evolved (S-type) magma, they are very different rock types (Lee et al. 1981). For example, the Lexington Creek pluton contains discrete grains of both muscovite and biotite, whereas the biotite in the Pole Canyon-Can Young Canyon pluton only occurs as euhedra within large phenocrysts of muscovite.

The next period of quiescence lasted almost 50 m.y. and ended with the emplacement of the 37 m.y. (?) old equigranular two-mica granite of the Young Canyon-Kious Basin area, just to the east of the older plutons, and the penecontemporaneous development of an aplitic phase in the same general area. The granite has a relatively high initial ⁸⁷Sr/⁸⁶Sr of 0.7120.

Field studies indicate that the tops of the southern Snake Range plutons are present at about the same level, even though they range in age from Jurassic to Oligocene. Combined with the stratigraphic evidence that the region has been emergent since Late Jurassic, this indicates that the younger masses were emplaced at progressively more shallow depths. All of the plutons of the southern Snake Range appear at or below the level of the Snake Range decollement. Indeed, the Young Canyon-Kious Basin and the Lexington Creek intrusions were affected by late movement on this structure (Lee et al. 1970, 1980), and there is evidence to indicate that such spatial relationships between intrusive rocks and thrust faults are not uncommon in the eastern Great Basin (Lee, Friedman and Gleason 1983).

There appears to be a confluence of regional igneous types in the southern Snake Range. We have noted the similarities between the distinctive muscovite-phenocrystic two-mica granite of the Pole Canyon-Can Young Canyon area and those of the Kern Mountains and the Toano Range, Nevada. The N-S trend defined by these distinctive granites is 225 km long (Lee et al. 1981). The Middle Jur-
Table 3. Summary of data for intrusive rocks of the southern Snake Range, Nevada

<table>
<thead>
<tr>
<th>Intrusion</th>
<th>Rock type</th>
<th>Age</th>
<th>Initial $^{87}$Sr/$^{86}$Sr</th>
<th>$\delta^{18}$O permil</th>
<th>Accessory minerals</th>
</tr>
</thead>
<tbody>
<tr>
<td>Snake Creek-Williams</td>
<td>Tonalite – granodiorite – granite</td>
<td>155 m.y.</td>
<td>0.7071</td>
<td>10.2–12.2</td>
<td>Felsic phase garnet, monazite</td>
</tr>
<tr>
<td>Canyon</td>
<td>(Rb – Sr)</td>
<td></td>
<td></td>
<td></td>
<td>ilmenite, zircon</td>
</tr>
<tr>
<td></td>
<td></td>
<td>160 m.y.</td>
<td></td>
<td></td>
<td>Mafic phase epidote, magnetite sphene,</td>
</tr>
<tr>
<td></td>
<td>(U – Pb)</td>
<td></td>
<td></td>
<td></td>
<td>allanite apatite, zircon</td>
</tr>
<tr>
<td>Osceola</td>
<td>Granodiorite</td>
<td>~145 m.y.</td>
<td>0.7075</td>
<td>9.6–9.9</td>
<td>Epidote, allanite, magnetite, sphene,</td>
</tr>
<tr>
<td></td>
<td>(Rb – Sr)</td>
<td></td>
<td></td>
<td></td>
<td>apatite zircon</td>
</tr>
<tr>
<td></td>
<td>151 m.y. (Musc K – Ar)</td>
<td>0.7086</td>
<td></td>
<td></td>
<td>Monazite, ilmenite magnetite, apatite,</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>zircon, garnet</td>
</tr>
<tr>
<td>Willard Creek</td>
<td>Equigranular</td>
<td>79 m.y.</td>
<td>0.7157</td>
<td>10.6–12.1</td>
<td>Allanite, monazite, acicular zircon in</td>
</tr>
<tr>
<td></td>
<td>two-mica granite</td>
<td></td>
<td></td>
<td></td>
<td>equant apatite</td>
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<tr>
<td>Pole Canyon – Can Young</td>
<td>Muscovite – phenoocrystic granodiorite – granite</td>
<td>86 m.y. (?)</td>
<td>0.7114</td>
<td>10.4–10.6</td>
<td>Apatite, zircon, fluorite, apatite,</td>
</tr>
<tr>
<td>Canyon</td>
<td>(Rb – Sr)</td>
<td></td>
<td></td>
<td></td>
<td>epidote $^b$, ziosite $^b$</td>
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<td>Lexington Creek</td>
<td>Equigranular</td>
<td>37 m.y.</td>
<td>0.7120</td>
<td>8.7–10.0</td>
<td>Allanite, monazite, magnetite, zircon,</td>
</tr>
<tr>
<td></td>
<td>two-mica granite</td>
<td></td>
<td></td>
<td></td>
<td>ilmenite: very sparse sphene, apatite,</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>and garnet</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Young Canyon – Kious Basin</td>
<td>Equigranular two-mica granite</td>
<td>29 m.y. (?)</td>
<td>0.7120</td>
<td>7.4–10.3</td>
<td>Garnet, fluorite, trace amounts of</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>garnite, columbite-tantalite.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Beryl found in one aplite</td>
</tr>
</tbody>
</table>

$^a$ According to the IUGS classification (Streckeisen 1973) $^b$ May be deuteric

assic calcic plutons in the Osceola and the Snake Creek-Williams Canyon areas are similar in age and petrology to the Notch Peak intrusion (169 ± 3 m.y.) of the House Range in western Utah (Lee, Stacey and Fischer 1983) and to the southern intrusion (152 m.y.) in the Gold Hill area in Utah (Stacey and Zartman 1978). The middle Tertiary igneous event apparent in the Young Canyon-Kious Basin area is represented by an Oligocene (38 m.y.) pluton that makes up the northern part of the intrusive outcrop of the Gold Hill, Utah, area (Stacey and Zartman 1978).

Finally, we note that the sequence of igneous events described here is remarkable similar to that found in the Ruby Mountains of Elko County, about 200 km northwest of the area of Fig. 1. Jurassic (160 ± 2 m.y.), Cretaceous (83 ± 1.3 m.y.), and Tertiary (31.9 ± 1.2 m.y.) intrusive rocks all are present in the Ruby Mountains (Kistler et al. 1981).

Discussion

The Sr and O isotope compositions of the granitoid rocks from this region may yield some clues to the origin of the contrasting magma suites. Lee et al. (1982) showed that minerals from the Snake Creek-Williams Canyon and Pole Canyon-Can Young Canyon plutons appear to be in isotopic equilibrium. Although not all whole rock samples employed in the following analysis were studied in this manner, it seems probable that most are representative of the magmas from which they crystallized. Samples of the cataelastic part of the Young Canyon-Kious Basin pluton show marked depletion of $^{18}$O (Lee et al. 1982), and their $\delta^{18}$O values are not considered further here. It is difficult to assess the role of postmagmatic modification of single whole rock samples that represent entire plutons (e.g. those from Lee et al. 1981). These samples are distinguished from others on Fig. 14. From the correlation of O and Sr isotopes two groups of plutons can be distinguished. Those with $^{87}$Sr/$^{86}$Sr less than 0.709 (all Jurassic in age) trend toward high $\delta^{18}$O (up to 13.2 permil). The other group diverges from this trend and has much higher initial $^{87}$Sr/$^{86}$Sr at comparable values of $\delta^{18}$O. This latter trend is similar to the correlation observed in the Maladeta igneous complex, Spain (Michard-Vitrac et al. 1980), and within granites of the inner Piedmont of the southwestern U.S.A. (Wenner 1981) – rocks that show S-type affinities in their chemistry and mineralogy. The trend of the low $^{87}$Sr/$^{86}$Sr Jurassic plutons is similar to that observed for the eastern portion of the Peninsular Range batholith, California (Taylor and Silver, 1978), to the British Caledonides (Halliday et al. 1980), and to the Sierra Nevada batholith (Masi et al. 1981) – rocks which appear to be chemically similar to the Snake Creek-Williams Canyon pluton.

The covariation of Sr and O isotopic compositions is usually attributed to the mixing of isotopically distinct reservoirs in the mantle and crust. The mixing may be accomplished through bulk contamination of a rising magma, isotope exchange between magma and wall rock, or by physical mixing of the reservoirs within the source region. It is very difficult to distinguish among these mechanisms based simply on isotopic variations.

A schematic cross section of the crust and upper mantle of this portion of the northern Great Basin is shown in Fig. 15. Approximate values of $\delta^{18}$O and $^{87}$Sr/$^{86}$Sr are given for each reservoir, based on the work of Kistler et al.
Fig. 14. δ18O versus 87Sr/86Sr diagram showing the compositions of plutonic rocks from northeastern Nevada. O Osceola pluton; SC Snake Creek-Williams Canyon; W Willard Creek; L Lexington Creek; Y Young Canyon-Kious Basin; RK Cretaceous granite, Ruby Mountains; PJ Jurassic granite, Ruby Mountains; P Pole Canyon-Can Young Canyon; K Kern Mountains; T Toana Range. Plutons represented by a single sample are shown (+) separate from the fields for other intrusions. Superimposed are curves showing simple mixing between a melt derived from the lower crust (400 ppm Sr, 87Sr/86Sr = 0.706, δ18O = 7), and two hypothetical metamorphic components of the middle crust. One of these is a granite gneiss (200 ppm Sr, 87Sr/86Sr = 0.715, δ18O = 12), the other a schist (200 ppm Sr, 87Sr/86Sr = 0.732, δ18O = 10), both similar to rocks in the metamorphic terrane exposed in the Ruby Mountains (Kistler et al. 1981). Also shown is a curve denoting mixing between a lower crustal magma and a hypothetical quartzite (8 ppm Sr, 87Sr/86Sr = 0.760, δ18O = 15) representing eustatic sedimentary rocks from the region (Lee and Doering 1980; Lee et al. 1982; Lee, Kistler and Robinson, 1983; and unpublished). The dashed lines show the effects of contaminating a mantle-derived magma (assumed to have 300 ppm Sr, 87Sr/86Sr = 0.704, and δ18O = 6) from two of the crustal reservoirs.

(1981), Lee et al. (1981, 1982), and Lee, Kistler and Robinson (1983). The stratigraphic reconstruction is from Stewart (1980) and ignores post-Cambrian sedimentation and tectonism. (The stratigraphy summarized by Stewart implies that the Jurassic and Cretaceous plutons in the southern Snake Range were emplaced at a depth of about 10 km from the surface.) The compositions of these reservoirs are shown in schematic fashion on Fig. 14.

From the positions of the reservoirs relative to the composition of the granite rocks it is clear that the Pole Canyon-Can Young Canyon magma (and a few other strongly peraluminous plutons like those in the Toana Range and in the Ruby and Kern Mountains) could have been formed solely by anatectic metasediments and granite gneisses, thought to have been at depth greater than 16 km during the Cretaceous. Indeed, Kistler et al. (1981) have suggested that the Cretaceous and Jurassic granites in the Ruby Mountains have not moved significant distances from their sources. These granites intrude Precambrian metasediments and granite gneisses and are chemically similar to the Pole Canyon-Can Young Canyon pluton. The derivation of these S-type granites from a mid-crustal reservoir of metasedimentary rocks is consistent with the presence of an inherited component in the U–Pb systematics of their zircons (Lee, Stacey and Fischer 1983; Kistler et al. 1981, p. 1448), Nd model ages (Farmer and DePaolo 1980), their mineralogy which implies crystallization at low fO2 (possibly as a result of equilibration with graphite), and their strongly peraluminous chemistry which resulted in the crystallization of muscovite and aluminous biotite. From the apparent composition of the crustal reservoirs it is unlikely that the S-type magmas were generated by partial melting of the late Precambrian sedimentary rocks. During the Cretaceous these sedimentary rocks (mostly quartzites with minor shale) had 87Sr/86Sr ratios of 0.760 to 0.780, much higher than those observed in the Cretaceous two-mica granites (0.711–0.732), and they were at depths of less than about 16 km. An extremely high geothermal gradient would have been necessary to produce partial melting at this depth.

The lack of S-type plutons with high 87Sr/86Sr (> 0.71) in western Utah suggests that the mid-crustal reservoir, interpreted to consist of Proterozoic metasediments (possibly derived from Archean parents; Farmer and DePaolo 1980) and plutonic rocks like those described by Kistler et al. (1981), must not contribute substantially to magmatic processes there (Farmer and DePaolo 1981; Moore et al. 1979; Zartman 1974). This may be because the mid-crustal reservoir does not occur at appropriate depths for contamination or magma generation. This possibility is consistent with the regional stratigraphy and the tectonic reconstructions of Stewart (1980), which demonstrate that eastern Nevada was the site of subsidence and the accumulation of considerable volumes of sediments (> 10 km thick) from the late Precambrian until at least the Triassic, depressing the basement there. In Utah, the mid-crustal reservoir occurs beneath a relatively thin sequence of Paleozoic sedimentary rocks. Alternatively, Proterozoic sediments may have attained great thicknesses only off the western margin of the Archean craton in the present site of eastern Nevada (Fig. 15).
Other plutons (Lexington Creek and Young Canyon-Kious Basin) on the high $^{87}$Sr/$^{86}$Sr trend could conceivably be explained as a result of the mixing of a component from the mid-crustal reservoir with a component of another reservoir having both lower $\delta^{18}O$ and $^{87}$Sr/$^{86}$Sr. Based on the extrapolation of the trend to low $^{87}$Sr/$^{86}$Sr, the $\delta^{18}O$ of this primitive reservoir appears to have been in the range of 7–9 permil, values more consistent with the lower crust than with the mantle. The chemistry and mineralogy of the plutons are also consistent with a lower crustal source. Mixing parabolas join the composition of this hypothetical reservoir ($\delta^{18}O = 7$ permil and $^{87}$Sr/$^{86}$Sr = 0.706) to portions of the mid-crustal reservoir as approximated from Kistler et al. (1981). The isotopic compositions of the plutons from this group fall within this rather broad envelope. Those which we have termed S-types fall at the high $^{87}$Sr/$^{86}$Sr portion of the envelope. Contamination near the site of emplacement would involve the carbonate or clastic reservoirs shown in Fig. 14, and may have further altered the position of the plutons in the $\delta^{18}O$–$^{87}$Sr/$^{86}$Sr diagrams.

The low $^{87}$Sr/$^{86}$Sr trend is more difficult to rationalize. As noted earlier, this trend is similar to that encountered in many plutonic and volcanic environments (generally thought to be related to the subduction process). The correlation is generally explained as the result of increasing amounts of contamination of a primitive mantle-derived magma by sedimentary materials, either at the source or as the magma passes through the crust. If mixing with a noncarbonate sediment is important, then in order to produce the marked $\delta^{18}O$ variation with only small changes in $^{87}$Sr/$^{86}$Sr, a reservoir with very low Sr concentration and high $\delta^{18}O$ must be involved. A possible candidate for this reservoir could be the clastic sedimentary rocks that make up the upper Precambrian-Lower Cambrian section in this region (Fig. 14). The data of Lee et al. (1981, 1982, and unpublished), and Lee, Kistler and Robinson (1983) shows that these sedimentary rocks have $\delta^{18}O$ values of 11 to 15 permil and $^{87}$Sr/$^{86}$Sr = 0.760 (at 150 m.y.). A mixing parabola joining the high $\delta^{18}O$ end of this range with a magma derived from the lower crustal reservoir is shown (Fig. 14). Large amounts of contamination (up to about 70%) or isotopic exchange are required to produce the spectrum of compositions observed. Contamination by carbonates is deemed unlikely because of the covariation of $\delta^{18}O$ and the increasing activity of Al$_2$O$_3$ in the sequence of melts represented by the Osceola, Snake Creek-Williams Canyon, and Willard Creek intrusions. Nonetheless, it seems unlikely that this was accomplished by bulk contamination of a magma.

In summary, the isotopic compositions of the southern Snake Range plutonic rocks define two trends—one toward high $\delta^{18}O$ at moderate $^{87}$Sr/$^{86}$Sr, composed solely of Jurassic plutons, the other toward high $^{87}$Sr/$^{86}$Sr at moderate $\delta^{18}O$. These variations are consistent with the contamination of magmas derived from the lower crust by a component from upper crustal sedimentary rocks producing the first trend. The second trend may involve mixing a lower crustal magma with a component from the middle crust composed of metasedimentary and igneous rocks. The S-type granites appear to have a large proportion of the mid-crustal component.

Other models could also explain the isotopic compositions observed. For example, in terms of Sr and O isotope it is extremely difficult to distinguish between the mixing of a mantle-derived basalt with a component from the lower crust and the simple derivation of a melt from the lower crust. However, we see little compelling evidence for the involvement of mantle-derived melts in producing the O and Sr isotopic relations observed; an end member with the appropriate isotopic composition (low $\delta^{18}O$ and $^{87}$Sr/$^{86}$Sr) probably exists in the lower crust, for it can be discerned in western Utah (E.H. Christiansen, unpublished data). Certainly the aluminaeous nature of all of the granitoid rocks (high Al$_2$O$_3$, with epidote or garnet and/or muscovite), their relatively silicic compositions, and the nature of the liquidus phases (plagioclase, biotite, quartz) in even the more primitive plutons suggests their ultimate derivation from the crust. The apparent absence of mafic igneous rocks of the appropriate age is permissive evidence against involvement of mantle-derived magmas in the generation of these granitic rocks.

Farmer and DePaolo (1981) have interpreted the Nd–Sr isotopic systematics of plutonic rocks from the northern Great Basin to indicate the absence of a lower crustal reservoir in eastern Nevada. This inference is based on a apparent lack of low to moderate $^{87}$Sr/$^{86}$Sr (0.706–0.708) ratios in the region at appropriate Nd isotope ratios. We suggest that with the occurrence of plutons with $^{87}$Sr/$^{86}$Sr lower than 0.708, a significant reservoir with isotopic characteristics of the lower crust ($^{87}$Sr/$^{86}$Sr = 0.706–0.709; $\delta^{18}O$ = 6 to 8 permil) may exist beneath the region.

Conclusions

The chemical, mineralogical and isotopic diversity of the granitic plutons from the southern Snake Range, Nevada,
and surrounding regions is largely a result of differences in the nature and composition of their sources. Strongly peraluminous Cretaceous granites with phenocrystic muscovite were probably derived from a mid-crustal source composed of Proterozoic metasediments. They have high initial strontium-isotope ratios (over 0.715), moderately high delta18O (9 to 13 permil) and other characteristics of S-type granites. Metaluminous to peraluminous granites (Young Canyon-Kios Basin and Lexington Creek plutons) may have derived part of their character from a similar source but they have lower initial strontium-isotope ratios and in some cases bear magnetite and sphene, indicating crystallization under relatively oxidizing conditions. Thus we suggest that a component from the lower crust (or perhaps the mantle) was important in their source region. Three Jurassic plutons with accessory magnetite form a distinct group that has initial Sr-isotope ratios less than 0.709 but with relatively high delta18O (up to 13 permil). We interpret these as originating from magmas that were contaminated as they passed through a thick Precambrian-Cambrian sequence of quartzite-dominated clastic rocks. We suggest that their ultimate source lies in the lower crust but it is difficult to rule out the involvement of mantle-derived mafic magmas in their genesis. One of these plutons has been studied in detail and it appears that little assimilation took place during its crystallization. In situ fractional crystallization nonetheless drove residual liquids to higher delta18O (up to 12 permil) and led to the replacement of sphene and magnetite by ilmenite and garnet as crystallization phases.

Acknowledgements. F.C.W. Dodge and R.W. Kistler of the U.S. Geological Survey provided helpful advice in the field and constructive reviews of this manuscript. We have also benefited from the constructive reviews of Lang Farmer of U.C.L.A. and Craig Simmons of the Colorado School of Mines. Eric H. Christiansen was supported by a National Research Council – U.S. Geological Survey Postdoctoral Research Associateship while part of this work was performed.

References
DePaolo DJ (1981) Trace element and isotopic effects of combined wallrock assimilation and fractional crystallization. Earth and Planet Sci Lett 53: 189-202
Faure G, Powell JL (1972) Strontium isotope geology. Springer-Verlag, p 188
Hanson GN (1978) The application of trace elements to the petrogenesis of igneous rocks of granitic composition. Earth Planet Sci Lett 38: 26-43
Lee DE, Friedman I, Gleason JD (1983) Negative delta18O values found for eastern Nevada plutonic rocks deformed by stresses resulting from postcrystallization movement along spatially related thrust faults. In: Shorter Contributions to Isotope Research, US Geol Surv Prof Pap, Chap. N
Lee DE, Marvin RF, Mehrdt HH (1980) A radiometric age study of Mesozoic-Cenozoic metamorphism in eastern White Pine County, Nevada, and nearby Utah. US Geol Surv Prof Pap 1158-C: 17-28


Lee DE, Van Loenen RE (1971) Hybrid granodiorite rocks of the southern Snake Range, Nevada. US Geol Surv Prof Pap 668 p 48


Moore WJ, Hedge CE, Sorensen ML (1979) Variations in 87Sr/86Sr ratios of igneous rocks along the Uinta trend, northwestern Utah. Geol Soc Am Abst Prog 11, 297


Stacey JS, Zartman RE (1978) A lead and strontium isotopic study of igneous rocks and ores from the Gold Hill mining district, Utah. Utah Geol 5: 1–15


Streckeisen AL (1973) Plutonic rocks; Classification and nomenclature. Geotimes 18, 26–30


Taylor HP, Silver LT (1978) Oxygen isotope relationships in plutonic rocks of the Peninsular Ranges Batholith, southern and Baja California. US Geol Surv Open-File Rep 78–701, 423–426


Whitebread DH, Griggs AB, Rogers WB, Myttyn JW (1962) Preliminary geologic map and sections of the Wheeler Peak quadrangle, White Pine County, Nevada. US Geol Surv Mineral Investigation Field Studies Map MF-244


Received July 7, 1982; Accepted March 20, 1983