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Oxygen and Hydrogen Isotope Studies of Contact Metamorphism in the Santa Rosa Range, Nevada and Other Areas*

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Abstract. The O^{18}/O^{16} and D/H ratios have been determined for rocks and coexisting minerals from several granitic plutons and their contact metamorphic aureoles in the Santa Rosa Range, Nevada, and the Eldora area, Colorado, with emphasis on pelitic rocks. A consistent order of O^{18}/O^{16} and D/H enrichment in coexisting minerals, and a correlation between isotopic fractionations among coexisting mineral pairs are commonly observed, suggesting that mineral assemblages tend to approach isotopic equilibrium during contact metamorphism. In certain cases, a systematic decrease is observed in the oxygen isotopic fractionations of mineral pairs as one approaches the intrusive contacts. Isotopic temperatures generally show good agreement with heat flow considerations. Based on the experimentally determined quartz-muscovite O^{18}/O^{16} fractionation calibration curve, temperatures are estimated to be 525 to 625° C at the contacts of the granitic stocks studied.

Small-scale oxygen isotope exchange effects between intrusive and country rock are observed over distances of 0.5 to 3 feet on both sides of the contacts; the isotopic gradients are typically 2 to 3 per mil per foot. The degree of oxygen isotopic exchange is essentially identical for different coexisting minerals. This presumably occurred through a diffusion-controlled recrystallization process. The size of the oxygen isotope equilibrium system in the small-scale exchanged zones varies from about 1.5 to 30 cm. A xenolith and a re-entrant of country rock projecting into an intrusive have both undergone much more extensive isotopic exchange (to hundreds of feet); they also show higher isotopic temperatures than the rocks in the aureole. The marginal portions of most plutons have unusually high O^{18}/O^{16} ratios compared to "normal" igneous rocks, presumably due to large-scale isotopic exchange with metasedimentary country rocks when the igneous rocks were essentially in a molten state. The isotopic data suggest that outward horizontal movement of H_2O into the contact metamorphic aureoles is very minor, but upward movement of H_2O is important. Also, direct influx and absorption of H_2O from the country rock appears to have occurred in certain intrusive stocks. The D/H ratios of biotites in the contact metamorphic rocks and their associated intrusions show a geographic correlation that is similar to that shown by the D/H ratios of meteoric surface waters, perhaps indicating that meteoric waters were present in the rocks during crystallization of the biotites.

Except in the exchanged zones, the O^{18}/O^{16} ratios of pelitic rocks do not change appreciably during contact metamorphism, even in the cordierite and sillimanite grades; this is in contrast to regional metamorphic rocks which commonly decrease in O^{18} with increasing grade. Thus, contact metamorphic rocks generally do not exchange with large quantities of "igneous" H_2O, but regional metamorphic rocks appear to have done so.

Introduction

The purpose of the present study is to investigate how the isotopic compositions of rocks and minerals change in non-carbonate rocks during contact metamorphic processes. In contact metamorphic aureoles, one commonly can trace a single lithologic unit from an area where the rocks are essentially unaffected to a locality where the rocks are in contact with intrusive igneous bodies. The physical parameters, particularly the temperature, can be at least qualitatively inferred from

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the spatial configurations of the country rocks and the intrusions. More specifically, the present studies aim to determine: (1) The O^{18}/O^{16} and *D/H* ratios of contact metamorphic minerals and rocks. (2) The extent of isotopic exchange between the country rock and the intrusive. (3) The effect of metamorphic reactions upon the isotopic compositions of metamorphic rocks. (4) The isotopic compositions of H_2O liberated by contact metamorphic dehydration reactions. (5) Whether oxygen isotopic geothermometers are applicable in a contact metamorphic environment.

Two areas were studied in detail, the Santa Rosa Range, Nevada, and the Eldora area, Colorado. A companion study, dealing principally with the O^{18}/O^{16} and C^{13}/C^{12} variations that result from contact metamorphic decarbonation reactions, also describes oxygen and hydrogen istotopic variations of pelitic rocks in the contact zone of a pluton near Birch Creek, California (SHIEH and TAYLOR, 1969). Extensive reference to those data are made in the present study.

Previous oxygen isotope studies of metamorphic rocks are by: BAERTSCHI (1957); ENGEL *et al.* (1958); JAMES and CLAYTON (1962); TAYLOR and EPSTEIN (1962); TAYLOR *et al.* (1963); SHARMA *et al.* (1965); SCHWARCZ (1966); GARLICK and EPSTEIN (1967); ANDERSON (1967); TAYLOR (1968a); TAYLOR and COLEMAN (1968); and SHEPPARD and SCHWARCZ (1968). The only previous oxygen isotope study of a contact metamorphic aureole, however, is that of PERRY and BONNICHSEN (1966) who found that the quartz-magnetite isotopic fractionations in the metamorphosed Biwabik iron formation varied fairly systematically from 7.4 to 11.5 per mil outward for 15 to 210 m from the Duluth gabbro contact.

Hydrogen isotope variations in igneous and metamorphic minerals have been studied by GODFREY (1962), FRIEDMAN *et al.* (1964a), and TAYLOR and EPSTEIN (1966).

The experimental oxygen isotopic calibrations of several mineral-H_2O systems are pertinent to the present study. These are: quartz-water and magnetite-water by O'NEIL and CLAYTON (1964), alkali feldspar-water and anorthite-water by O'NEIL and TAYLOR (1967), and muscovite-water by O'NEIL and TAYLOR (1966), for various temperatures from 250 to 800° C.

Experimental Techniques

Sample Collections

The exact locations and field relations of the samples are very important in order to interpret the isotopic data, particularly for samples near the intrusive contacts. For samples less than 1 foot from the contact, the distances were marked before they were broken from the outcrop. The rock samples adjoining each contact were sliced at $^1/_2$ to 1 inch intervals parallel to the contact surface before they were crushed for mineral separation.

Sample-distances within 50 feet of the contact were measured by a steel tape. Sample-distances greater than 50 feet from the contact were measured by pacing. At distances greater than 500 feet, sample localities were located on a geologic or topographic map. Whenever possible, samples of a traverse were collected along the strike of a single bed or lithologic unit.

Sample Preparations

For oxygen isotopic analysis, mineral separations were made on as small a volume of rock as possible. This is particularly important for samples very close to the intrusive contacts because of the steep isotopic compositional gradient. If samples are very coarse-grained, hand-picking was used. Otherwise, routine magnetic and heavy liquid procedures were performed. Quartz was separated from feldspar by dissolving away the feldspars in cold hydrofluoric acid. By combining the above methods, a purity of 95 to 100 percent was obtained.

For hydrogen isotope analysis, purity is not critical with respect to non-hydrous minerals. Muscovite was easily separated magnetically from biotite, and since coexisting biotite and hornblende invariably have similar D/H ratios (GODFREY, 1962; TAYLOR and EPSTEIN, 1966) it is not necessary to make completely pure separates of these two minerals.

Oxygen Extraction from Silicates and Oxides

The extraction of oxygen from silicates and oxides was performed by reacting with BrF_5 at 500 to 650° C in nickel reaction vessels, as described by CLAYTON and MAYEDA (1963) and GARLICK (1964). Mass spectrometric analyses were performed on CO_2 gas obtained from this oxygen by combustion with a resistance-heated graphite rod as described by TAYLOR and EPSTEIN (1962).

For most silicates, a grain size as coarse as 50 to 100 mesh is easily reacted at 500° C and a time of 8 h. However, andalusite and magnetite are exceptions. These minerals have to be ground exceedingly fine in order to obtain high yields. If coupled with higher reaction temperatures ($\sim$630° C) and longer reaction time (24 to 48 h), near 100% yields can usually be obtained. Andalusite was not analyzed previous to the present study; a study of yields vs. isotopic compositions was therefore made and the oxygen isotopic compositions are, within experimental error, independent of per cent yields.

Hydrogen Extraction

The experimental technique for extraction of hydrogen from OH-bearing minerals is similar to that described by FRIEDMAN (1953) and GODFREY (1962). Samples, after degassing at room temperature under high vacuum for about 1 h, were heated by an induction furnace to a maximum temperature of 1,300 to 1,500° C to liberate hydrogen and water. Water was converted to hydrogen by passing it over hot uranium at about 750° C. The volume of total hydrogen evolved was measured and the gas was then ready for mass spectrometric analysis.

Mass Spectrometry and Standards

Mass spectrometric analyses of oxygen isotopes are performed on CO_2 gas, and of hydrogen isotopes on hydrogen gas. The mass spectrometers used are 60 degree, single-focusing, double-collecting, dual gas-feed, Nier-McKinney type instruments. The analytical error for oxygen samples is about 0.1 to 0.2 per mil and for hydrogen samples about 2 to 3 per mil. The oxygen and hydrogen isotope data reported in this work are relative to Standard Mean Ocean Water (SMOW) as defined by CRAIG (1961) and CLAYTON and MAYEDA (1963). During the course of this research, a working standard (St. Peter sandstone, $\delta = 10.9$) was usually included in each set of six samples analyzed.

Notation

The fractionation factor α_{A-B} between two chemical compounds or phases A and B is defined by:

$$\alpha_{A-B} = R_A / R_B$$

where $R = (O^{18}/O^{16})$ or (D/H).

The experimental data are reported in terms of the quantity δ defined by

$$\delta_A = \frac{R_A - R_{\text{std}}}{R_{\text{std}}} \times 1{,}000$$

where R_{std} is the reference standard (SMOW).

If α_{A-B} is very close to unity, then $1{,}000 \ln \alpha_{A-B} \approx \delta_A - \delta_B$. For abbreviation, $1{,}000 \ln \alpha_{A-B}$ will be denoted by Δ_{A-B} in the present work.

Analytical Results (Presented in Table 1)

Santa Rosa Range, Nevada

Geological Relationships. The contact metamorphism in the Santa Rosa Range of Nevada has been studied by COMPTON (1960). The metamorphic rocks of the area

Table 1. *Oxygen and hydrogen isotope analyses of minerals and rocks*

Sample	Distance from contact (feet)	Mineral	δ O^{18} per mil	δ D per mil
Sawtooth traverse I				
SRO-18	−50[a]	Quartz	11.7 ± 0.1[c] (3)[d]	
Trondhjemite		Plagioclase (An_{15})	9.2 (1)	
		Biotite ($H_2O = 4.6\%$)[b]	5.5 ± 0.1 (6)	−95 ± 0 (2)
		Magnetite	−0.3 ± 0.1 (2)	
SRO-17	−10	Quartz	11.8 ± 0.1 (2)	
Trondhjemite		Plagioclase (An_{15})	9.7 (1)	
SRO-16	−8	Quartz	11.5 ± 0.2 (2)	
Quartz vein in Trondhjemite				
SRO-19AI	−0.5	Quartz	14.5 ± 0.2 (3)	
Trondhjemite		Plagioclase (An_{15})	12.1 (1)	
		Muscovite	11.2 ± 0.1 (3)	−64 (1)
		Biotite ($H_2O = 4.1\%$)	7.9 ± 0.1 (4)	−105 ± 1 (2)
		Magnetite	0.2 ± 0.0 (2)	
SRO-15C	−0.17	Quartz	16.0 ± 0.1 (4)	
Trondhjemite		Plagioclase (An_{15})	13.4 (1)	
		Muscovite	12.7 ± 0.2 (3)	−62 (1)
		Biotite ($H_2O = 3.6\%$)	9.7 ± 0.1 (2)	−104 ± 2 (2)
SRO-15A	−0.04	Quartz	15.9 ± 0.1 (4)	
Trondhjemite		Plagioclase (An_{15})	14.2 (1)	
		Muscovite	13.1 ± 0.1 (4)	−66 (1)
		Biotite ($H_2O = 3.7\%$)	10.0 ± 0.1 (2)	−103 ± 4 (2)
SRO-15B	+0.08	Quartz	16.1 ± 0.2 (4)	
Porphyroblastic		Andalusite	15.2 ± 0.1 (2)	
andalusite schist		Muscovite	13.2 ± 0.1 (5)	−65 (1)
		Biotite ($H_2O = 4.3\%$)	9.9 ± 0.1 (5)	−97 ± 1 (2)
		Whole rock ($H_2O = 2.9\%$)	13.5 (calc.)	−75 (1)
SRO-19AII	+0.5	Quartz	16.7 ± 0.0 (2)	
Porphyroblastic		Andalusite	14.4 ± 0.1 (2)	
sillimanite-		Muscovite	14.1 ± 0.2 (4)	−59 ± 0 (2)
staurolite-andalusite		Biotite ($H_2O = 3.6\%$)	10.9 ± 0.1 (4)	−86 ± 3 (8)
schist		Magnetite	0.3 (1)	
		Whole rock	14.5 (1)	
SRO-19B	+1.0	Quartz	18.7 ± 0.1 (2)	
Porphyroblastic		Andalusite	15.8 ± 0.1 (4)	
andalusite schist		Muscovite ($H_2O = 4.6\%$)	15.3 ± 0.2 (3)	−57 ± 2 (2)
		Biotite ($H_2O = 4.4\%$)	11.8 ± 0.2 (4)	−110 ± 2 (2)
		Whole rock ($H_2O = 2.6\%$)	15.9 (1)	−83 (1)
SRO-19b	+1.0	Quartz	17.4 ± 0.2 (2)	
Aplite dike		Muscovite	16.1 ± 0.2 (2)	
		Whole rock	17.1 (1)	

[a] Negative distances are measured inward into the intrusive; positive distances are measured outward into the country rock.

[b] H_2O contents are given in weight per cent.

[c] Analytical error shown is average deviation from the mean.

[d] Numbers in parentheses represent number of separate analyses.

Table 1 (continued)

Sample	Distance from contact (feet)	Mineral	δO^{18} per mil	δD per mil
SRO-19C Andalusite schist	+50	Quartz	19.2 ± 0.1 (2)	
		Muscovite	15.5 ± 0.1 (2)	−69 (1)
		Biotite ($H_2O = 4.3\%$)	12.0 ± 0.0 (3)	−98 ± 0 (2)
		Whole rock ($H_2O = 1.5\%$)	16.6 (1)	−81 (1)
		Quartz pod	18.0 ± 0.1 (3)	
SRO-19D Porphyroblastic phyllite	+600	Quartz	19.1 ± 0.2 (2)	
		Andalusite	16.2 ± 0.1 (2)	
		Muscovite	16.4 ± 0.1 (2)	−73 ± 2 (2)
		Chlorite	12.2 ± 0.1 (2)	
		Whole rock ($H_2O = 3.6\%$)	16.8 (1)	−79 (1)
		Quartz in quartzite	18.6 ± 0.1 (3)	
SRO-19E Phyllite	+2,000	Whole rock ($H_2O = 0.8\%$)	19.8 (1)	−63 (1)
SRO-19F Phyllite	+2,500	Whole rock ($H_2O = 3.0\%$)	16.2 (1)	−80 (1)
		Quartz	18.9 ± 0.0 (2)	
		Grey mica*	14.0 (1)	
		Muscovite (calc.)**	15.0	
		Chlorite (calc.)**	11.0	
		* X-ray diffraction shows: muscovite ~75% chlorite ~25% **Assuming δ muscovite — δ chlorite = 4.		
SRO-20A Phyllite	+7,000	Whole rock ($H_2O = 2.4\%$)	18.0 (1)	−79 (1)
Sawtooth traverse II				
SRO-22F Trondhjemite	−50	Quartz	11.4 ± 0.1 (2)	
		Plagioclase (An_{15})	9.4 (1)	
		Biotite	5.3 ± 0.1 (2)	
SRO-22E Trondhjemite	−1.5	Quartz	11.3 ± 0.0 (2)	
		Plagioclase	9.8 (1)	
		Biotite	5.6 ± 0.0 (2)	
SRO-22D Gneiss	+0.17	Quartz	12.1 ± 0.1 (2)	
		Muscovite	9.2 ± 0.0 (2)	
		Biotite	6.5 ± 0.0 (2)	
		Whole rock	10.2 (1)	
SRO-22C Porphyroblastic andalusite schist	+5	Quartz	15.1 ± 0.0 (3)	
		Andalusite	14.3 ± 0.0 (2)	
		Muscovite	12.6 ± 0.2 (2)	
		Biotite	9.7 ± 0.1 (2)	
		Whole rock	14.3 (1)	
SRO-22A Porphyroblastic staurolite-andalusite schist	+85	Quartz	15.3 ± 0.0 (2)	
		Andalusite	14.0 ± 0.0 (2)	
		Muscovite	13.3 ± 0.2 (2)	
		Biotite	10.3 ± 0.1 (2)	
		Whole rock	13.9 (1)	

Table 1 (continued)

Sample	Distance from contact (feet)	Mineral	δ O^{18} per mil		δ D per mil	
SRO-25b Chloritoid phyllite	+2,000	Quartz	18.0	(1)		
		Grey Mica*	14.3	(1)		
		Whole rock	15.7	(1)		
		Muscovite (calc.)**	15.3			
		Chlorite (calc.)**	11.3			
		* X-ray diffraction shows a mixture of muscovite and chlorite roughly in a proportion of 3:1. ** Assuming δ muscovite — δ chlorite = 4.				
SRO-25 Phyllite	+2,000	Whole rock	14.9	(1)		
Sawtooth stock, other samples						
SRO-24 Staurolite phyllite	+2,000	Quartz	17.7±0.1	(2)		
		Grey mica*	14.6	(1)		
		Quartz pod	17.6	(1)		
		Muscovite (calc.)**	15.6			
		Chlorite (calc.)**	11.6			
		* X-ray diffraction shows a mixture of muscovite and chlorite in roughly 3:1 ratio. **Assuming δ muscovite — δ chlorite = 4.				
SRO-A Trondhjemite	−500	Quartz	11.4±0.1	(2)		
		Plagioclase (An$_{15}$)	9.5	(1)		
		Biotite	5.3±0.1	(2)		
SRO-B Porphyroblastic andalusite schist	+1.0	Quartz	15.8±0.0	(2)		
		Andalusite	14.8±0.0	(2)		
		Muscovite	12.2±0.1	(2)		
		Biotite	9.4±0.0	(2)		
SRO-C Garnet hornfels	+20	Quartz	19.3	(1)		
SRO-D Pegmatite dike	+30	Pure quartz core	18.3	(1)		
		Quartz	17.8	(1)		
		K-feldspar	12.4	(1)		
		Muscovite	15.7	(1)		
Flynn north traverse						
SRO-4 Granodiorite	−300	Quartz	12.6±0.1	(3)		
		Plagioclase (An$_{30}$)	10.0	(1)		
		Biotite	6.6±0.2	(3)		
		K-feldspar (leucocratic band)	11.8±0.1	(3)		
SRO-3C Granodiorite	−15	Quartz	14.3±0.0	(2)		
		Plagioclase (An$_{20}$)	11.6±0.2	(2)		
		Hornblende	9.4±0.1	(3)		
		Biotite	8.9±0.1	(2)		
SRO-3A-1 Granodiorite	−0.5	Quartz	14.5±0.2	(3)		
		Plagioclase (An$_{20}$)	12.3±0.1	(3)		
		Hornblende	9.1±0.2	(3)	−91	(1)
		Biotite	9.2±0.1	(3)		

Table I (continued)

Sample	Distance from contact (feet)	Mineral	δ O^{18} per mil	δ D per mil
SRO-3 A-2	+0.5	Quartz	14.8 (1)	
Amphibolite		Plagioclase	12.9 ± 0.1 (3)	
		Hornblende	9.6 ± 0.1 (3)	−90 (1)
		Chlorite	8.8 ± 0.2 (3)	
SRO-3 B	+10	Plagioclase	12.3 ± 0.2 (4)	
Amphibolite		Hornblende	8.6 ± 0.1 (4)	
		Biotite	8.5 ± 0.2 (3)	
SRO-9	+250	Plagioclase	12.3 ± 0.1 (4)	
Amphibolite		Hornblende	8.0 ± 0.2 (4)	
		Biotite	8.5 (1)	
SRO-5	+5	Semi-hornfels	16.7 ± 0.1 (2)	
SRO-7	+20	Semi-hornfels	16.5 ± 0.0 (2)	
SRO-8	+150	Semi-hornfels	17.3 ± 0.0 (2)	
SRO-10	+400	Semi-hornfels	17.2 ± 0.1 (2)	
SRO-12	+800	Porphyro-phyllite	16.7 ± 0.1 (2)	
SRO-11	+3,000	Phyllite	16.3 ± 0.2 (2)	
SRO-13	+5,000	Phyllite	16.6 ± 0.2 (2)	
Flynn south traverse				
SRO-39	−75	Quartz	10.7 ± 0.1 (2)	
Granodiorite		Hornblende	6.4 (1)	
		Biotite	5.0 ± 0.1 (2)	−90 (1)
SRO-38	−35	Quartz	10.5 (1)	
Granodiorite		Plagioclase (An$_{30}$)	8.2 (1)	
		Hornblende	5.9 (1)	
		Biotite	4.6 (1)	
SRO-37	+45	Quartz	17.3 ± 0.1 (2)	
		Muscovite	15.2 ± 0.2 (2)	
		Biotite	12.0 ± 0.1 (2)	
SRO-36	+150	Quartz	17.3 ± 0.1 (2)	
Andalusite cordierite		Andalusite	15.0 ± 0.1 (2)	
hornfels		Biotite	12.2 ± 0.0 (2)	
SRO-42	+200	Quartz	17.1 ± 0.2 (2)	
Andalusite-		Andalusite	15.0 ± 0.1 (2)	
cordierite		Muscovite	15.5 (1)	
hornfels		Biotite	12.2 (1)	
SRO-46	+3,200	Quartz	16.9 (1)	
Porphyroblastic phyllite		Andalusite	14.8 ± 0.1 (2)	
SRO-48	+4,500	Andalusite	15.1 ± 0.1 (2)	
Porphyroblastic phyllite		Whole rock	16.3 (1)	
SRO-49 Phyllite	+5,000	Whole rock	15.6 (1)	
SRO-50 Phyllite	+5,300	Whole rock	16.3 (1)	

Table 1 (continued)

Sample	Distance from contact (feet)	Mineral	δO^{18} per mil	δD per mil
Santa Rosa stock				
SRO-29F	−150	Quartz	13.1 ± 0.2 (2)	
Granodiorite		Plagioclase (An_{20})	10.4 (1)	
		Biotite (groundmass)	6.8 ± 0.1 (2)	
		Biotite (phenocryst)	7.7 (1)	
SRO-29A	+0.08	Quartz	14.6 ± 0.1 (2)	
Migmatized hornfels		Muscovite	11.5 ± 0.1 (2)	
		Biotite	8.6 ± 0.1 (2)	
		Whole rock	11.8 (1)	
SRO-29B Migmatized hornfels	+0.4	Whole rock	11.8 (1)	
SRO-29C Hornfels	+0.9	Whole rock	13.6 (1)	
SRO-29D	+2	Quartz	16.9 ± 0.0 (2)	
Hornfels		Muscovite	13.7 ± 0.0 (2)	
		Biotite	10.7 ± 0.1 (2)	
		Whole rock	14.0 (1)	
SRO-29E Hornfels	+30	Whole rock	15.7 (1)	
SRO-28-1	−0.17	Dike	12.2 (1)	
SRO-28-2	−0.01	Dike	13.4 (1)	
SRO-28-3	+0.02	Hornfels	13.6 (1)	
SRO-28-4	+0.17	Hornfels	14.9 (1)	
SRO-28-5	+0.4	Hornfels	15.7 (1)	
SRO-28-6	+1.7	Hornfels	15.0 ± 0.2 (2)	
SRO-31 Semi-hornfels	+2,500	Whole rock	16.1 (1)	
SRO-32 Semi-hornfels	+2,600	Whole rock	16.1 (1)	
SRO-34	+2,500	Quartz	14.0 ± 0.1 (2)	
Granodiorite, from plug, 300 × 600 ft.		Biotite	8.1 ± 0.1 (2)	
13-85-7a	−9,000	Quartz	9.5 ± 0.0 (2)	
Adamellite		Plagioclase (An_{20})	8.0 ± 0.1 (2)	
		Biotite	4.4 ± 0.1 (2)	−92 (1)
13-85-8	−5,000	Quartz	9.6 ± 0.1 (2)	
Adamellite		Plagioclase (An_{20})	7.9 (1)	
		Biotite	4.6 (1)	
Eldora traverse				
Biotite samples (HART, 1964)				
C0A	−5	Biotite	5.4 ± 0.1 (3)	−109 (1)
C20A	+20	Biotite	6.6 ± 0.0 (2)	
C58A	+58	Biotite	6.5 ± 0.1 (2)	

Table 1 (continued)

Sample	Distance from contact (feet)	Mineral	δO^{18} per mil	δD per mil
Biotite samples (Hart, 1964)				
C248A	+248	Biotite	7.0 ± 0.2 (4)	
C6B	+1,130	Biotite	6.4 ± 0.2 (2)	
C7B	+2,400	Biotite	5.4 ± 0.0 (2)	
C4C	+3,600	Biotite	7.3 ± 0.2 (2)	−119 (1)
BC-5C	+5,200	Biotite	6.7 ± 0.2 (2)	
C14C	+14,100	Biotite	6.7 ± 0.1 (2)	
C12C	+22,500	Biotite	8.0 ± 0.1 (2)	
N-3	−450	Quartz	9.9 ± 0.0 (2)	
Quartz monzonite		Biotite	4.9 ± 0.1 (2)	
EL-1	−50	Quartz	9.9 ± 0.1 (2)	
Quartz monzonite		Biotite	4.8 ± 0.1 (2)	
El-2	+2	Quartz	10.7 ± 0.1 (2)	
Gneiss		K-feldspar	9.1 ± 0.0 (2)	
		Hornblende	6.7 ± 0.0 (2)	
		Biotite	6.7 ± 0.0 (3)	
		Magnetite	2.7 ± 0.1 (2)	
N-2	+2	Quartz	10.4 ± 0.1 (2)	
Amphibolite		Plagioclase	9.3 ± 0.1 (2)	
		Hornblende	7.0 ± 0.1 (2)	
		Biotite	5.8 ± 0.0 (3)	
		Magnetite	2.3 ± 0.1 (2)	
EL-12	+12	K-feldspar	9.0 ± 0.1 (2)	
Pegmatite		Biotite	6.6 ± 0.0 (2)	
N-1	+18	Plagioclase (An_{45})	9.2 ± 0.1 (2)	
Amphibolite		Hornblende	7.0 ± 0.1 (2)	
C-20A	+20	Quartz	10.4 ± 0.1 (2)	
Pegmatite		K-feldspar	9.2 (1)	
		Biotite	6.6 ± 0.0 (2)	
C4B	+950	Quartz	12.2 ± 0.0 (2)	
Gneiss		Biotite	6.9 ± 0.1 (2)	
		Magnetite	3.7 ± 0.1 (2)	
		Quartz lens	12.8 (1)	
Feldspar samples (Steiger and Hart, 1967)				
Chittenden Mountain traverse				
CH-1	+75	K-feldspar	9.1 (1)	
CH-4	+800	K-feldspar	10.6 (1)	
CH-8	+2,200	K-feldspar	9.3 (1)	
CH-9a	+2,890	K-feldspar	10.0 (1)	
CH-10	+2,900	K-feldspar	9.6 (1)	
CH-11	+3,100	K-feldspar	10.4 (1)	
CH-12	+4,300	K-feldspar	10.9 (1)	

Table 1 (continued)

Sample	Distance from contact (feet)	Mineral	δO^{18} per mil	δD per mil
Ute Mountain traverse				
UM-4	+1,200	K-feldspar	9.6 (1)	
UM-5	+1,380	K-feldspar	9.9 (1)	
UM-6	+1,650	K-feldspar	10.6 (1)	
UM-7	+1,800	K-feldspar	9.8 (1)	
UM-10	+2,900	K-feldspar	8.6 (1)	
UM-11	+3,300	K-feldspar	10.0 (1)	
UM-12	+3,900	K-feldspar	9.2 (1)	
Mineral Mountain traverse				
MM-1a	+10	K-feldspar	9.9 (1)	
MM-1b	+30	K-feldspar	9.5 (1)	
MM-8	+1,510	K-feldspar	10.4 (1)	
Bryan Mountain traverse				
BM-1	+20	K-feldspar	11.4 (1)	
BM-2	+200	K-feldspar	10.6 (1)	
BM-3	+600	K-feldspar	10.6 (1)	
BM-4	+800	K-feldspar	10.4 (1)	
BM-5	+900	K-feldspar	11.6 (1)	
BM-6	+1,090	K-feldspar	9.7 (1)	
Caribou traverse				
HES-8	−530	Plagioclase (An_{25})	9.0 (1)	
Monzonite		Hornblende	6.5 ± 0.1 (2)	
		Biotite	4.5 ± 0.0 (2)	
		Magnetite	1.7 ± 0.1 (2)	
HES-4a	−0.04	Plagioclase (An_{25})	8.5 ± 0.0 (2)	
Monzonite		Biotite	5.4 ± 0.1 (2)	
		Magnetite	1.8 ± 0.0 (2)	
HES-4b	+0.04	Quartz	10.1 (1)	
Gneiss		Biotite	5.5 ± 0.0 (2)	
		Magnetite	1.9 ± 0.0 (2)	
HES-4c	+0.6	Quartz	10.1 ± 0.0 (2)	
Gneiss		K-feldspar	9.6 (1)	
		Biotite	6.1 ± 0.2 (2)	
		Magnetite	3.4 ± 0.0 (2)	
HES-1	+1,300	Quartz	12.8 ± 0.0 (2)	
Sillimanite schist		Perthite	10.9 (1)	
		Biotite	8.0 ± 0.1 (2)	
		Magnetite	4.5 ± 0.0 (2)	
		Whole rock	10.4 (1)	

include a 20,000-foot section of Upper Triassic (and Jurassic?) sediments that was folded, metamorphosed, locally faulted, and intruded by granitic stocks in the late Cretaceous and early Tertiary. The country rocks are dominantly metapelitic; they were converted to metamorphic assemblages of quartz, albite, muscovite, chlorite, and carbonates during the low-grade regional metamorphism.

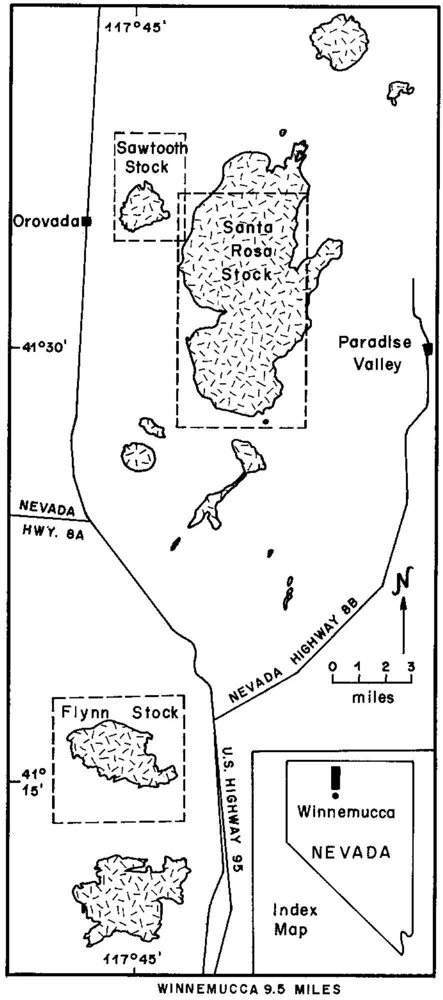

Fig. 1. Map of Santa Rosa Range, Nevada, showing distribution of the granitic stocks (after Compton, 1960)

There are nine different stocks which vary in composition from sodic granodiorite to adamellite, trondhjemite, and tonalite. Structural relationships show that the stocks are predominantly crosscutting bodies. The distribution of the stocks is shown in Fig. 1.

The metamorphic aureoles are divided into two zones by Compton (1960): an outer zone of porphyroblastic phyllite and an inner zone of schist or hornfels where the biotite in the groundmass can be identified megascopically.

Four mineral assemblages were developed in the outer and inner parts of the aureoles: (1) Cordierite-biotite; (2) Cordierite-andalusite-biotite; (3) Staurolite-andalusite-biotite; (4) Andalusite-biotite; all with quartz, muscovite, and plagioclase.

Sawtooth Stock and Its Aureole. The Sawtooth stock is exposed as a circular body with a diameter of about 1.5 miles. A contact metamorphic aureole was developed

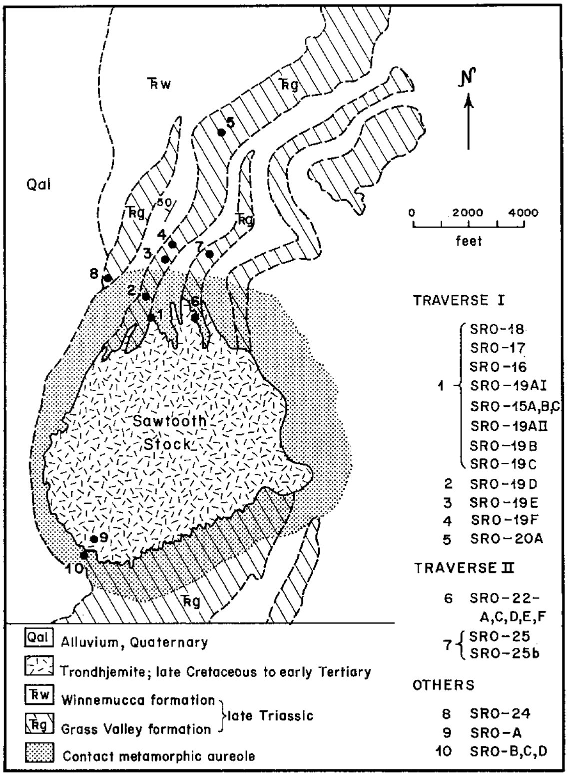

Fig. 2. Generalized geologic map of the Sawtooth stock and its contact aureole, Santa Rosa Range, Nevada, showing sample locations (geology after Compton, 1960)

around the stock with variable thickness from 1,500 to 2,500 feet. The stock and its aureole are unique among the stocks in this area in many respects. Among these are: (1) the trondhjemite intrusive produced highly schistose contact metamorphic rocks in the aureole; (2) a very sharp, zigzag contact exists between the intrusive and the country rocks; (3) quartz veins and quartzo-feldspathic dikes are abundant; (4) chloritoid and staurolite occur in the aureole; and (5) andalusite, muscovite, biotite, and staurolite in the schists are very coarse grained.

Two traverses from the north side of the stock were carefully sampled for isotopic analysis. The geology and sample locations are shown in Fig. 2. Traverse I, corresponding to Compton's No. 1 suite, was sampled from the tip of an outward-projecting tongue of the intrusive. Traverse II, corresponding to Compton's No. 2 suite, was taken from the center of a re-entrant of the country rock which projects into the intrusive. SRO-24 is equivalent to Compton's sample 3B that contains staurolite.

Figs. 3 and 4 are plots of O^{18}/O^{16} ratios vs. distance from the contact for traverse I and traverse II, respectively.

Hydrogen isotopic compositions and water contents of minerals and rocks from traverse I are plotted in Fig. 5.

Flynn Stock and Its Aureole. The Flynn stock is an oval-shaped body in plan, with an exposed area of about 7 square miles. Its contact metamorphic aureole

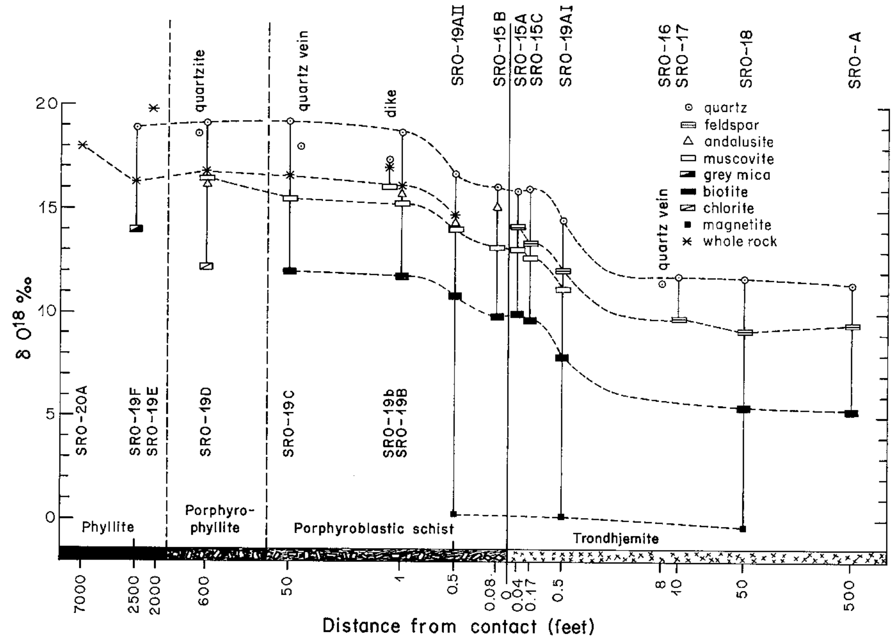

Fig. 3. Plot of O^{18}/O^{16} ratios vs. distance for samples from traverse I of the Sawtooth stock contact zone, Santa Rosa Range (equivalent to Compton's suite 1)

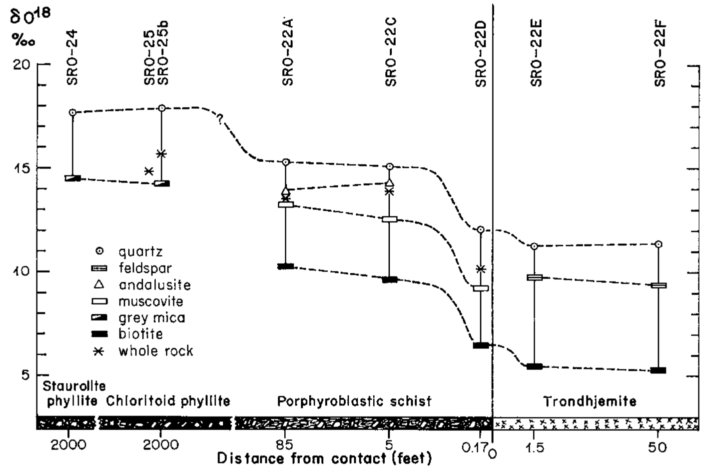

Fig. 4. Plot of O^{18}/O^{16} ratios vs. distance for samples from traverse II of the Sawtooth stock contact zone, Santa Rosa Range (equivalent to Compton's suite 2)

is 2,500 to 7,500 feet in width. The principal rock types in the aureole are hornfels, semi-hornfels, and porphyroblastic phyllite. The simplified geologic map and sample locations are shown in Fig. 6.

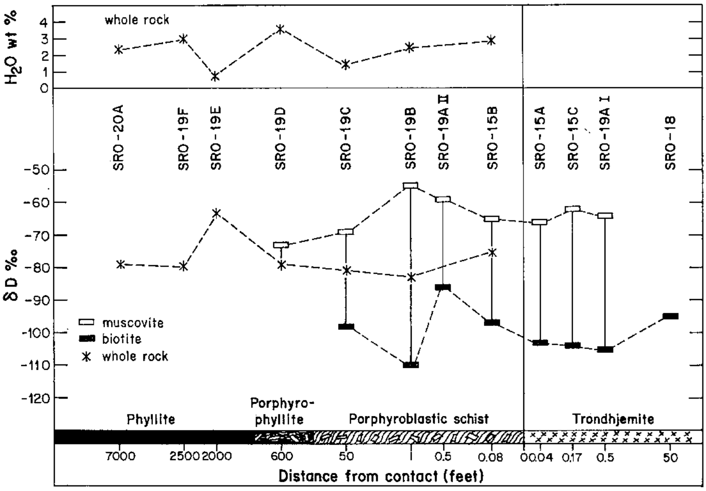

Fig. 5. Plot of D/H ratios and water contents vs. distance for samples from Sawtooth traverse I Santa Rosa Range

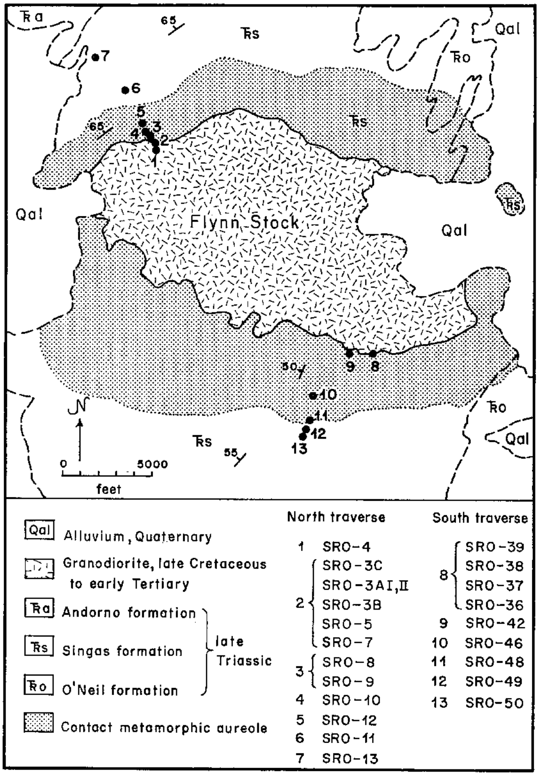

Fig. 6. Generalized geologic map of the Flynn stock and its contact aureole, Santa Rosa Range, Nevada, showing sample locations (geology after Compton, 1960)

The Flynn stock and its aureole differ from the Sawtooth stock and aureole in many ways, among which are:

Flynn	*Sawtooth*
1. Hornblende-biotite granodiorite	1. Hornblende-free, biotite trondhjemite
2. Fine-grained, compact, massive hornfels	2. Coarse-grained, muscovite-rich schist
3. Cordierite present, no staurolite or chloritoid	3. Staurolite and chloritoid present, no cordierite
4. K-feldspar formed in the hornfels near the intrusive contact	4. No K-feldspar formed in the schist

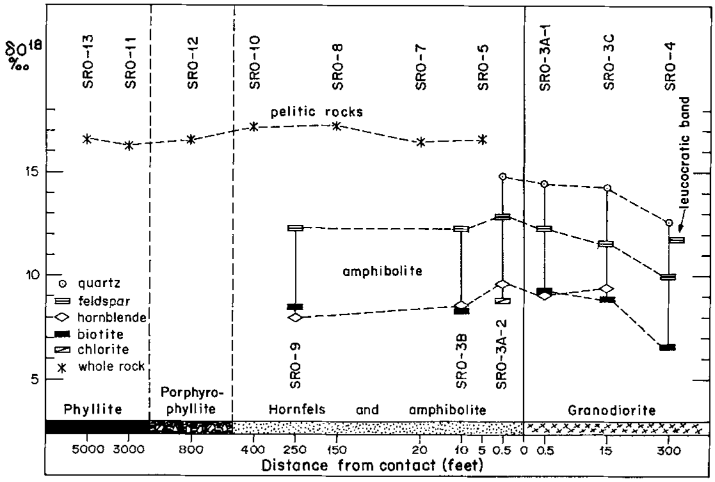

Fig. 7. Plot of O^{18}/O^{16} ratios vs. distance for samples from north traverse of Flynn stock contact zone, Santa Rosa Range

Two traverses were sampled for isotopic analysis (Fig. 6). The north traverse was taken from an amphibolite-granodiorite contact. The associated hornfels and semi-hornfels, which are the major rock types, were also collected. The south traverse, which is approximately equivalent to Compton's No. 15 suite, was sampled from the unaffected phyllite to the completely recrystallized, extremely compact hornfels near the intrusive contact. The contact is not well exposed in the south traverse, so no samples were collected within 25 feet of the contact. The isotopic data are illustrated in Figs. 7 and 8.

Santa Rosa Stock and Its Aureole. The Santa Rosa stock is the largest stock in the Santa Rosa Range (about 36 square miles in outcrop area). The stock is mainly granodiorite and has produced a contact metamorphic aureole 2,500 to 10,000 feet in width. The principal contact metamorphic rocks are hornfels and semi-hornfels. This is the only aureole that contains all the four mineral assemblages occurring in the area.

A traverse was taken across the southeast side of the stock (see Fig. 9). The samples near the intrusive contact are equivalent to Compton's sample No. 9B.

Fig. 10 is a plot of δO^{18} vs. distance-to-contact for several samples analyzed. The change of O^{18}/O^{16} ratios in the vicinity of a small granodiorite dike (apophysis

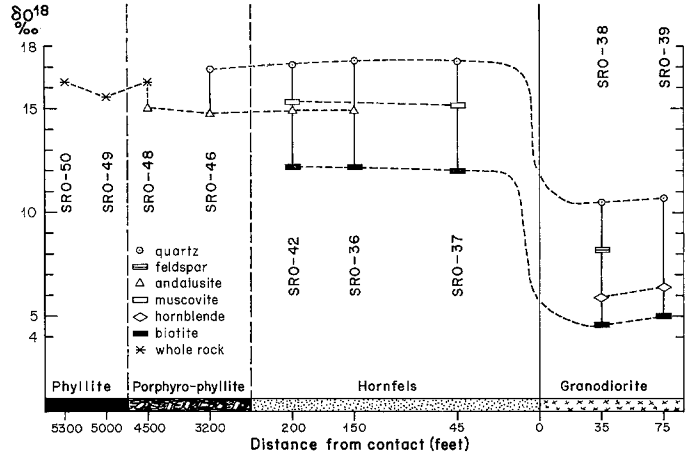

Fig. 8. Plot of O^{18}/O^{16} ratios vs. distance for samples from south traverse of Flynn stock contact zone, Santa Rosa Range (equivalent to COMPTON'S suite 15)

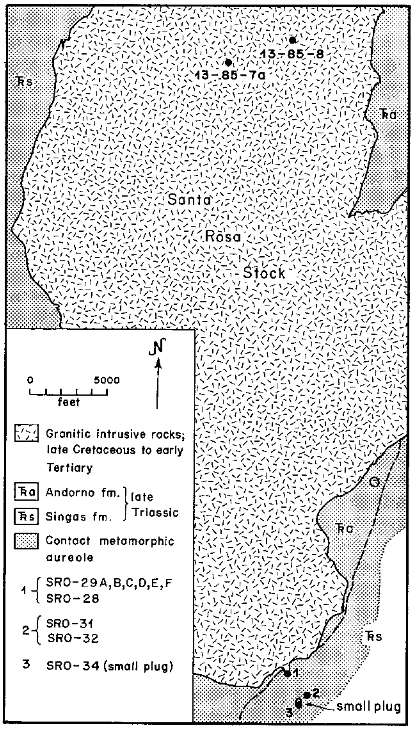

Fig. 9. Generalized geologic map of the Santa Rosa stock and its contact aureole, Santa Rosa Range, Nevada, showing sample locations (geology after COMPTON, 1960)

of the stock) has also been studied. The result is shown in Fig. 11. The dike is 5 inches thick and its contact is extremely sharp. There is no obvious mineralogical or textural change in the hornfels, either in hand specimen or under the microscope, as a result of the intrusion of the dike.

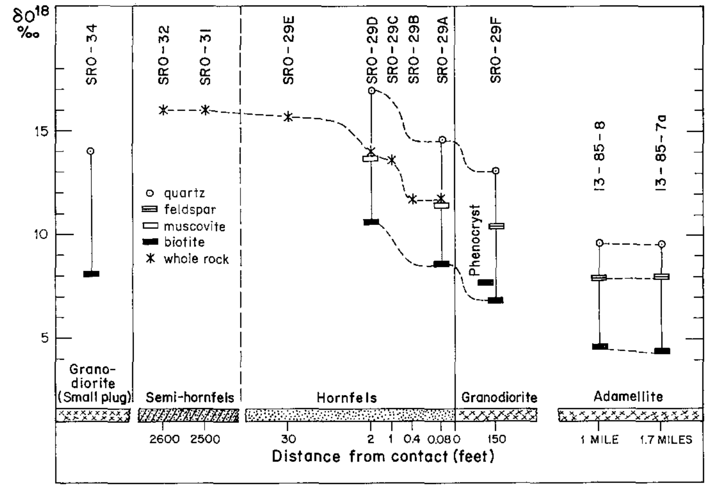

Fig. 10. Plot of O^{18}/O^{16} ratios vs. distance for samples from the Santa Rosa stock contact zone, Santa Rosa Range (equivalent to Compton's 9B and 10B)

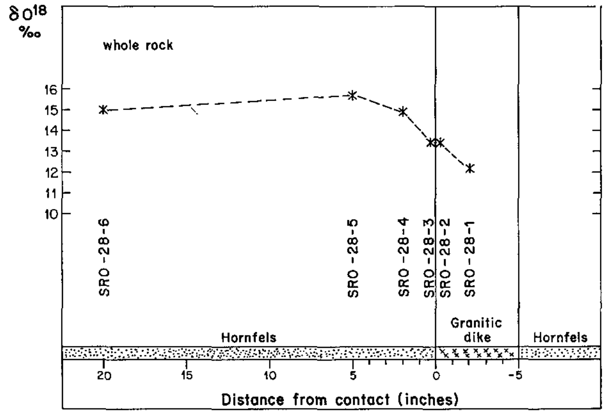

Fig. 11. Plot of O^{18}/O^{16} ratios vs. distance for whole-rock samples from a granitic dike-hornfels contact in the Santa Rosa aureole, about 30 feet from the main intrusive contact

Eldora Area, Front Range, Colorado

The oldest rocks in the Eldora area are Precambrian schists, gneisses, and amphibolites of the Idaho Springs formation. The rocks were highly and uniformly metamorphosed 1,200 to 1,500 million years ago (Hart, 1964). These Precambrian crystalline rocks were intruded during Laramide time by a series of stocks extending southwestward across the Front Range. The contact metamorphic zone of the Eldora stock was chosen by Doe and Hart (1963) and Hart (1964) for studies on the effects of contact metamorphism on isotopic mineral ages of different minerals. Steiger and Hart (1967), and Wright (1967) carried out studies on the microcline-orthoclase transition within the Eldora contact zone.

In the present study, oxygen isotopic analyses were made on the samples originally used by Hart (1964), and Steiger and Hart (1967). Samples were also collected from an additional traverse in the Caribou stock contact zone about 1 mile north of the Eldora stock. The general geology and all sample locations are shown in Fig. 12.

The Eldora stock is a quartz monzonite with an exposed diameter of about 2 miles. The only major mineralogical change observed by Hart (1964) and Steiger and Hart (1967) is the transition of the alkali feldspar from a perthitic microcline of high obliquity in the unaffected country rock to non-perthitic orthoclase near the contact. The transition can be sharp or gradational, and the width of the orthoclase zone is variable, depending strongly upon the configuration of the contact. The mineralogic ages of different minerals also responded differently due to the contact heating (Doe and Hart, 1963; Hart, 1964). The results of Hart (1964) are shown in the upper half of Fig. 13.

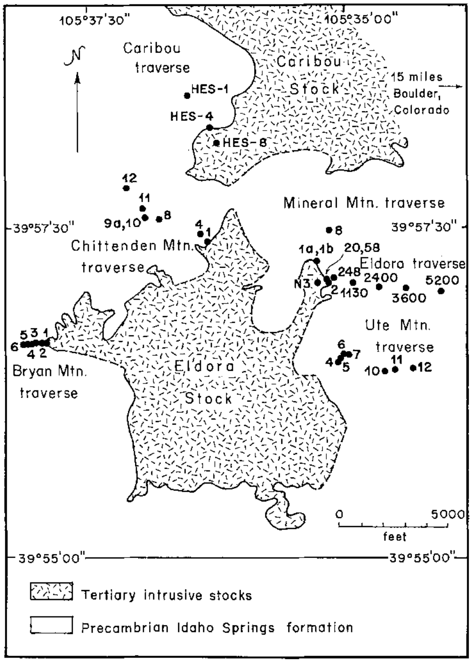

Fig. 12. Generalized geologic map showing sample locations in the Eldora and Caribou stock contact zones (after Steiger and Hart, 1967)

The oxygen isotopic results from the Eldora traverse are shown in the lower half of Fig. 13. The O^{18}/O^{16} ratios measured on the identical K-feldspar samples used by Steiger and Hart (1967) for their microline-orthoclase transition studies are shown in Fig. 14. A spread of about 2 per mil is observed in each of the four traverses studied, and the δ-values show no apparent change across the microcline-orthoclase transition zone.

A traverse was taken along the west side of the Caribou stock where the intrusive is a monzonite and the country rocks are sillimanite schist, gneiss, pegmatite, and amphibolite. The sample locations are shown in Fig. 12. The isotopic data are illustrated in Fig. 15.

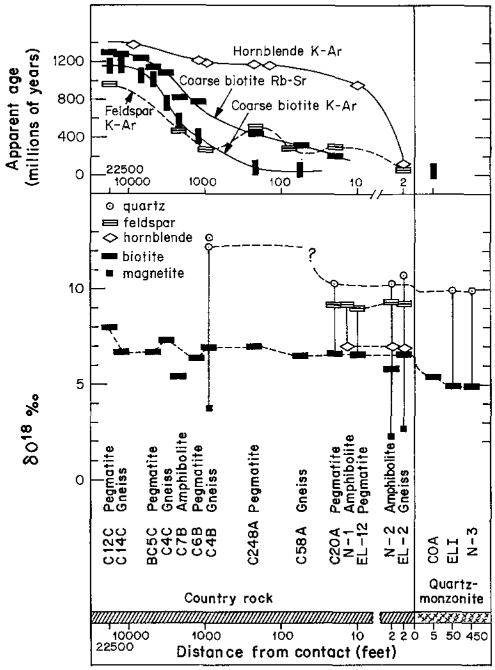

Fig. 13. Variation of mineral ages (upper figure after Hart, 1964) and O^{18}/O^{16} ratios (lower figure) as a function of distance from the intrusive contact for the Eldora traverse

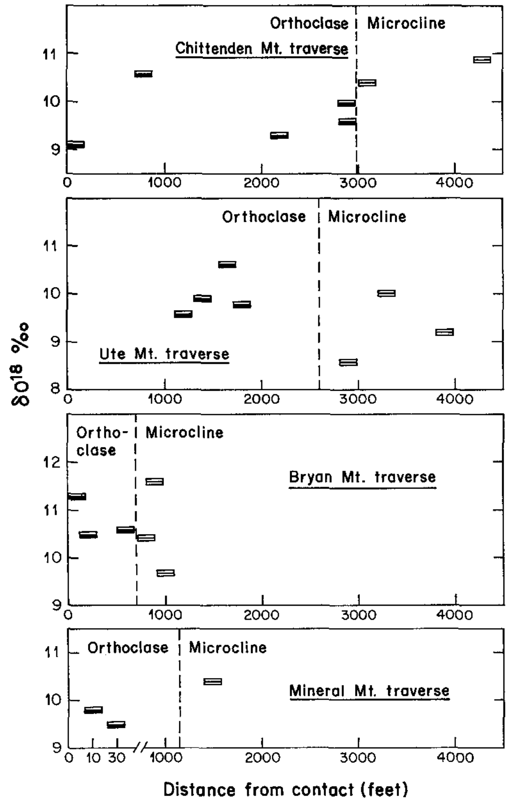

Fig. 14. Plot of O^{18}/O^{16} ratios vs. distance, for the identical K-feldspar samples used by Steiger and Hart (1967) for their microcline-orthoclase transition studies (Eldora stock contact zone)

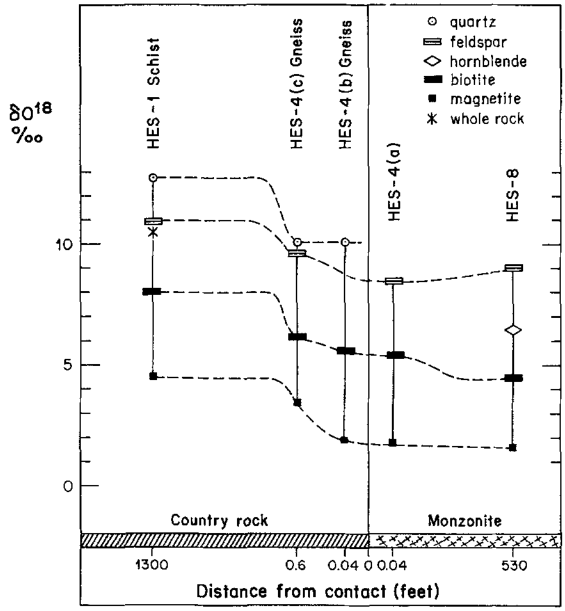

Fig. 15. Plot of O^{18}/O^{16} ratios vs. distance for samples from the Caribou stock contact zone

Discussion and Interpretation

Relative O^{18}/O^{16} and D/H Ratios in Coexisting Minerals of Contact Metamorphic Rocks

One of the most striking systematics observed in the present research is the consistent order of O^{18}-enrichment in the coexisting minerals in a given rock. Listed in order of decreasing O^{18}/O^{16}, the minerals are: Quartz; Feldspar; Andalusite; Muscovite; Hornblende; Biotite; Magnetite.

The above sequence is identical to that reported by Taylor *et al.* (1963), Garlick and Epstein (1967), and Taylor and Coleman (1968) in regional metamorphic rocks, except that andalusite is added in the present study. If one assumes that the Al_2SiO_5 polymorphs andalusite and kyanite do not fractionate oxygen isotopes, the order of muscovite and kyanite suggested by Garlick and Epstein (1967), and Taylor *et al.* (1963) is reversed from the present sequence. However, note that only three kyanite analyses were made in the above references; two of them do not coexist with muscovite, and the third is based on only a 50—60 per cent oxygen yield.

Only five rocks in the present study exhibit an "abnormal" oxygen isotopic sequence. Two rocks (an amphibolite and a granodiorite) in the north traverse of the Flynn stock show a reversal in the positions of hornblende and biotite. The biotite in the amphibolite (SRO-9) is an alteration product of the hornblende. The granodiorite was collected 6 inches inward from its contact with another amphibolite sample which also shows biotite altering from hornblende. The isotopic positions of andalusite and muscovite are reversed in two rocks: one occurs in the fine-grained hornfels of the south traverse of Flynn stock and the other in the porphyroblastic phyllite of Sawtooth traverse I. The former rock also shows an unreasonably small quartz-muscovite fractionations ($\Delta = 1.6$), and in the latter rock most of the muscovite is clearly a retrograde product of andalusite porphyroblasts. Most of the above examples of isotopic disequilibrium can therefore be correlated with textural or mineralogical evidence of disequilibrium.

The fifth "abnormal" rock is a pegmatite dike (SRO-D), crosscutting country rock on the south side of the Sawtooth stock, in which K-feldspar is lower in δO^{18} than coexisting muscovite by 3.3 per mil. It appears that the K-feldspar has become depleted in O^{18} relative to the coexisting quartz, because this quartz is isotopically similar to adjacent metasedimentary quartz. The extremely large quartz-K-feldspar fractionation ($\Delta = 5.4$) implies an absurdly low temperature of less than 100° C, so this is unquestionably a non-equilibrium pair. On the other hand, the quartz-muscovite fractionation ($\Delta = 2.1$), corresponding to a temperature of 830° C, does not appear to be far away from equilibrium. The δ-values of quartz and muscovite are similar in both the pegmatite and the country rock, suggesting that the isotopic composition of the pegmatite is controlled by isotopic exchange with the adjacent metasedimentary rock. The δ-value of the feldspar then must be a result of exchange down to what appear to be unreasonably low temperatures, or it must be a result of interaction with low-O^{18} fluids (see TAYLOR, 1968b, p. 66). The plausible sources of such O^{18}-depleted fluids are meteoric water that may have been present in the metasediments, or relatively low-temperature fluids derived from a late stage of crystallization of the trondhjemite. Such fluids may also be responsible for the very low O^{18}/O^{16} ratios of magnetite samples from the Sawtooth aureole. These magnetites are exceedingly fine-grained and form only about 0.1 per cent of the rock; hence they are very amenable to isotopic exchange.

Muscovite, biotite and hornblende are the three hydroxyl-bearing minerals analyzed in the present study. It is found, without exception, that muscovite concentrates deuterium relative to coexisting biotite; and hornblende and biotite have almost identical D/H ratios, in agreement with the results of TAYLOR and EPSTEIN (1966) and GODFREY (1962).

Oxygen Isotopic Fractionations among Coexisting Minerals

In Figs. 16 and 17 are plotted the oxygen isotopic fractionations quartz-muscovite vs. quartz-biotite and feldspar-biotite vs. quartz-biotite. The data of TAYLOR and EPSTEIN (1962), TAYLOR *et al.* (1963), GARLICK and EPSTEIN (1967), and TAYLOR (1968b) are also included. The plots show good corrrelations; therefore most of these assemblages must represent approximate isotopic equilibrium. Note that the ratios of Δ quartz-muscovite to Δ quartz-biotite from the present research are consistently smaller than those obtained by GARLICK and EPSTEIN (1967) in regional metamorphic rocks, but almost identical to those reported by TAYLOR (1968b), and TAYLOR and EPSTEIN (1962) in granitic igneous rocks (Fig. 16 and Table 2).

The oxygen isotopic fractionations among coexisting minerals from the Sawtooth traverses are shown in Fig. 18. The isotopic fractionations in traverse I are consistently larger than in traverse II, indicating a lower temperature environment in traverse I. The quartz-biotite and quartz-muscovite fractionations of traverse I

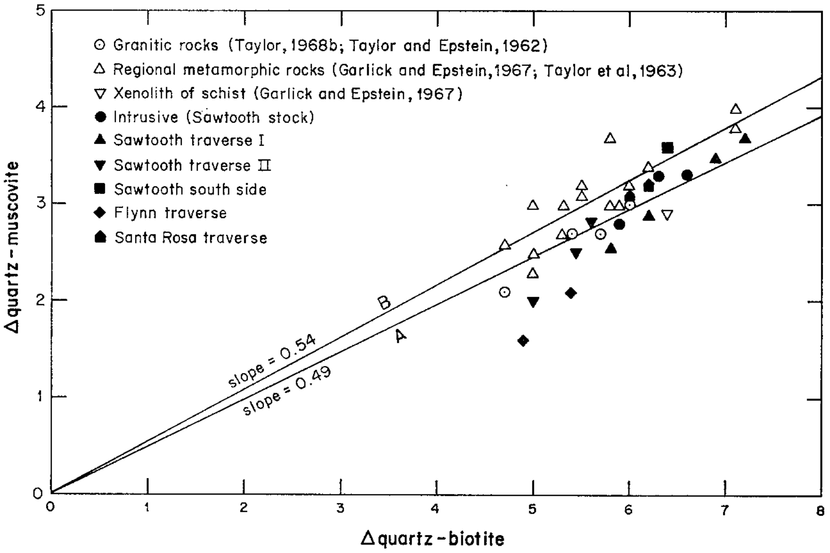

Fig. 16. Plot of oxygen isotopic fractionations Δquartz-muscovite vs. Δquartz-biotite for quartz-muscovite-biotite assemblages from contact metamorphic, regional metamorphic and granitic igneous rocks. Slope *A* is derived from average values in contact metamorphic and granitic igneous rocks. Slope *B* is derived from average values in regional metamorphic rocks

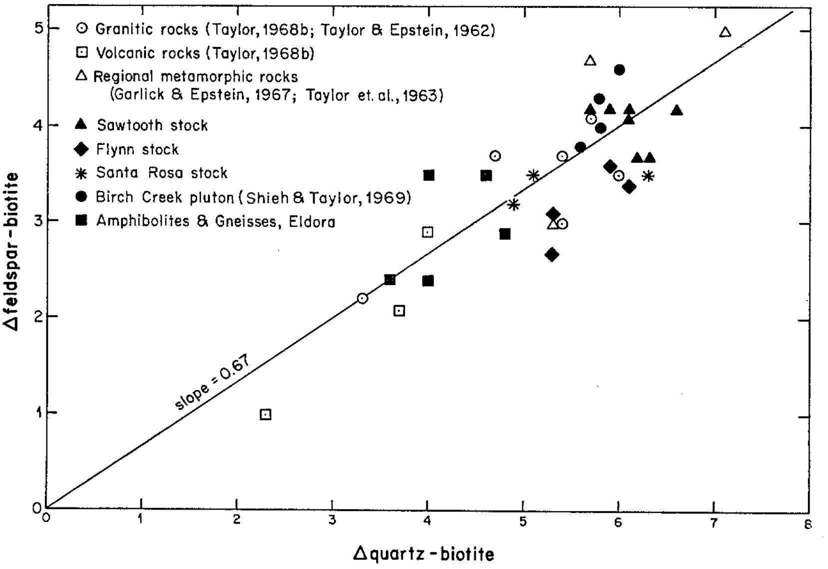

Fig. 17. Plot of oxygen isotopic fractionations Δfeldspar-biotite vs. Δquartz-biotite for quartz-feldspar-biotite assemblages from contact metamorphic, regional metamorphic, granitic plutonic and volcanic rocks. The slope is derived from the average values of all samples

increase quite regularly as a function of sample-distance outward from the contact. Traverse II, within the re-entrant, shows the opposite trend, but the muscovite-biotite fractionations in both traverses decrease regularly approaching the

Table 2. *Comparison of the ratios of quartz-muscovite and quartz-biotite O^{18}/O^{16} fractionations in different rock types*

Rock type	Δquartz-muscovite/ Δquartz-biotite	Ave. dev.	No. samples
Contact metamorphic[a]	0.49	0.03	12
Regional metamorphic[b]	0.54	0.03	15
Granitic igneous[c]	0.48	0.02	4

[a] This work, excluding samples SRO-37 and SRO-42 which show evidence of isotopic disequilibrium.

[b] Garlick and Epstein (1967), excluding the schist xenolith which has a ratio of 0.45.

[c] Taylor (1968b), and Taylor and Epstein (1962).

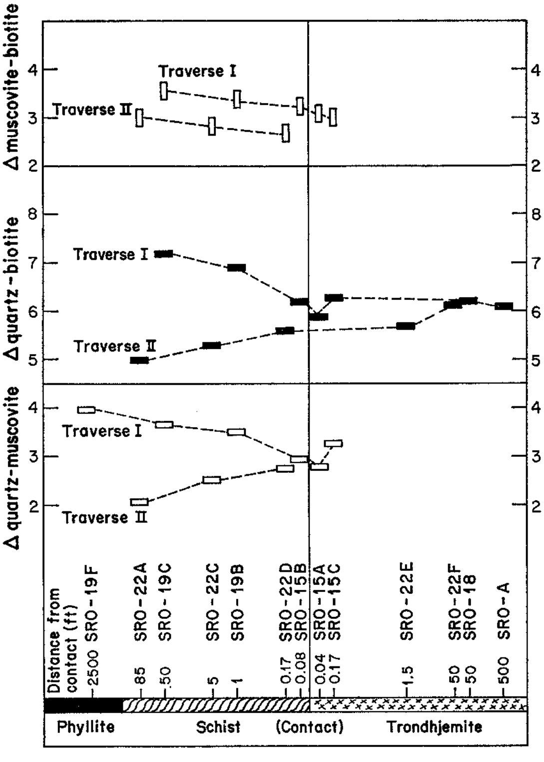

Fig. 18. Plot of oxygen isotopic fractionations for the mineral pairs quartz-biotite, quartz-muscovite, and muscovite-biotite for samples from traverses I and II of the Sawtooth stock contact zone, arranged in order of sample distances from the contact. Three assemblages in traverse I are not plotted; two occur 0.5 ft. from the contact in the partially exchanged zone, and the one 600 ft. away contains muscovite that is clearly a retrograde alteration of andalusite. The "reversed" trend shown by traverse II is probably due to lack of equilibrium, as the 5 ft. and 85 ft. samples are only 30 to 50 per cent exchanged (see Table 3)

contact. A regular increase of the quartz-biotite isotopic fractionation outward from the intrusive contact is also observed in a contact metamorphic aureole east of the Sierra Nevada near Birch Creek, California (Shieh and Taylor, 1969); these data are shown in Fig. 19. The Birch Creek pluton and contact metamorphic aureole are described by NASH (1962) and shown in Fig. 3 of MCKEE and NASH (1967). In the Flynn north traverse, plagioclase-hornblende fractionations increase outward from 3.2 and 3.3 at the contact to 3.7 at 10 ft. and 4.3 at 250 ft.

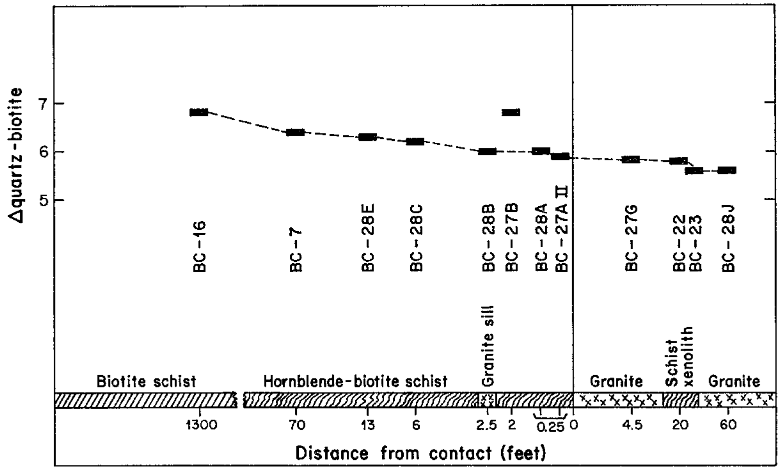

Fig. 19. Oxygen isotopic fractionations between coexisting quartz and biotite plotted vs. distance from intrusive contact, for samples from the Birch Creek contact zone (SHIEH and TAYLOR, 1969)

Geothermometry

Oxygen Isotopic Temperatures in Contact Metamorphic Rocks and Intrusions. All the country rock samples near intrusive contacts that show incomplete isotopic exchange with the intrusive, or vice versa, very likely indicate a non-equilibrium situation. The calculated isotopic temperatures in these rocks show a large scatter, which is understandable considering the very steep isotopic gradients preserved in the vicinity of the contact. The optimum situation for testing equilibrium is represented by samples collected right at the contact. In these samples, the coexisting minerals usually show 100% isotopic exchange between the country rock and the intrusive, and have attained the same isotopic composition from opposite directions, in the same way that true equilibrium is demonstrated in the laboratory (e.g. see O'NEIL and TAYLOR, 1967, p. 1417).

In Table 3 are listed the oxygen isotopic "temperatures" of various assemblages, the per cent O^{18}/O^{16} exchange between the country rock and the adjoining intrusive (or vice versa), and the isotopic compositions of waters in equilibrium with these assemblages, based on the experimentally determined quartz-muscovite curve (see TAYLOR, 1967, p. 120—123). Since quartz-biotite and quartz-magnetite fractionation curves have not yet been determined experimentally, or are in need of revision, tentative curves were derived from natural assemblages. The quartz-biotite curve was derived by assuming Δ quartz-muscovite $= 0.49 \times \Delta$ quartz-biotite (see Fig. 16); the quartz-magnetite curve utilized is the one estimated by TAYLOR and COLEMAN (1968) from mineral assemblages in regional

Table 3. *Comparison of isotopic temperatures and heat flow temperatures in various traverses, together with calculated O^{18}/O^{16} ratios of coexisting H_2O and percent exchange between country rock and intrusive*

Sample[a]	Mineral	δO^{18} per mil	% Exchange[b]	Quartz-mineral temperature (°C)	Heat flow[c] temperature (°C)	δH_2O[d]
Sawtooth Traverse I						
SRO-18 (−50)	Quartz	11.7	0			8.8
Trondhjemite	Biotite	5.5	0	515	>550	
	Feldspar	9.2	0			
	Magnetite	−0.3	?	390		
SRO-19AI (−0.5)	Quartz	14.5	65			10.7
Trondhjemite	Muscovite	11.2	58	470	550	
	Biotite	7.9	53	475		
	Feldspar	12.1	55			
	Magnetite	0.2	?	315		
SRO-15C (−0.17)	Quartz	16.0	100			12.2
Trondhjemite	Muscovite	12.7	91	470	550	
	Biotite	9.7	93	505		
	Feldspar	13.4	83			
SRO-15A (−0.04)	Quartz	15.9	100			13.6
Trondhjemite	Muscovite	13.1	100	570	550	
	Biotite	10.0	100	545		
	Feldspar	14.2	100			
SRO-15B (+0.08)	Quartz	16.1	100			13.5
Schist	Muscovite	13.2	100	550	550	
	Biotite	9.9	100	520		
SRO-19AII (+0.5)	Quartz	16.7	78			15.0
Schist	Muscovite	14.1	61	630	550	
	Biotite	10.9	52	560		
	Magnetite	0.3	?	265		
SRO-19B (+1.0)	Quartz	18.7	16			14.4
Schist	Muscovite	15.3	9	440	460	
	Biotite	11.8	10	450		
SRO-19C (+50)	Quartz	19.2	0			14.5
Schist	Muscovite	15.5	0	420	455	
	Biotite	12.0	0	430		

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[a] Numbers in parentheses indicate sample distance from contact. Negative distances measured inward into intrusive; positive distances measured outward into country rock.

[b] In samples SRO-22C, 22A and BC-22, 23 the % exchange includes large-scale exchange; all others include only small-scale exchange.

[c] Sawtooth Traverse I = Model E; Flynn South Traverse = Model B; Santa Rosa Aureole = large intrusion of Model A; Birch Creek Traverses = Model D; Caribou and Eldora Traverses = Model B.

[d] Calculated from quartz-muscovite isotopic temperature, or lacking that pair, calculated from quartz-biotite temperature. (For HES-8 and HES-4a, the plagioclase-biotite temperature is utilized).

Table 3 (continued)

Sample[a]	Mineral	δO^{18} per mil	% Exchange[b]	Quartz-mineral temperature (°C)	Heat flow[c] temperature (°C)	δH_2O[d]
Sawtooth Traverse II						
SRO-22F (−50)	Quartz	11.4	0			8.6
Trondhjemite	Biotite	5.3	0	530	>550	
SRO-22E (−1.5)	Quartz	11.3	0			9.0
Trondhjemite	Biotite	5.6	25	570	550	
SRO-22D (+0.17)	Quartz	12.1	100			9.7
Gneiss	Muscovite	9.2	100	570	550	
	Biotite	6.5	100	580		
SRO-22C (+5)	Quartz	15.1	49			13.7
Schist	Muscovite	12.6	44	665	550	
	Biotite	9.7	40	610		
SRO-22A (+85)	Quartz	15.3	45			15.7
Schist	Muscovite	13.3	33	900	~550	
	Biotite	10.3	28	680		
Flynn South Traverse						
SRO-38 (−35)	Quartz	10.5	0			7.9
Granodiorite	Biotite	4.6	0	550	>650	
SRO-37 (+45)	Quartz	17.3	0			15.6
Hornfels	Muscovite	15.2	0	830	640	
	Biotite	12.0	0	630		
SRO-36 (+150)	Quartz	17.3	0			15.9
Hornfels	Biotite	12.2	0	670	620	
Santa Rosa Aureole						
13-85-8 (−5000)	Quartz	9.6	0			8.4
Adamellite	Biotite	4.6	0	695	>650	
SRO-29F (−150)	Quartz	13.1	0			11.2
Granodiorite	Biotite	7.7	0	600	>650	
SRO-29A (+0.08)	Quartz	14.6	100			11.5
Hornfels	Muscovite	11.5	100	510	650	
	Biotite	8.6	100	535		
SRO-29D (+2)						
Hornfels	Quartz	16.9	~50			13.4
	Muscovite	13.7	~60	490	650	
	Biotite	10.7	~50	505		
Birch Creek Traverses (Shieh and Taylor, 1969; also see Fig. 19)						
BC-281 (−10)	Quartz	12.9	0			
Granite						
BC-22 (+5)	Quartz	13.0	~100			10.6
Xenolith	Biotite	7.2	~100	560	>550	
BC-23 (+1)	Quartz	12.7	~100			10.6
Xenolith	Biotite	7.2	~100	590	>550	

Table 3 (continued)

Sample[a]	Mineral	δO^{18} per mil	% Exchange[b]	Quartz-mineral temperature (°C)	Heat flow[c] temperature (°C)	δH_2O[d]
BC-28H (−2.5)	Quartz	12.9	0			
Granite						
BC-27G (−4.5)	Quartz	12.2	0			9.8
Granite	Biotite	6.4	0	560	>550	
BC-27AI (−0.4)	Quartz	12.7	42			10.2
Granite	Biotite	7.0	54	570	550	
BC-27AII (+0.25)	Quartz	14.0	90			11.4
Schist	Biotite	8.1	92	545	550	
BC-27B (+2)	Quartz	19.4	0			15.6
Schist	Biotite	12.6	0	460	550	
BC-28C (+6)	Quartz	19.2	0			16.3
Schist	Biotite	13.0	0	515	548	
Caribou Traverse						
HES-8 (−530)	Plagioclase	9.0	0			7.7
Monzonite	Biotite	4.5	0	500[e]	>650	
	Magnetite	1.7	0	550[e]		
HES-4a (−0.04)	Plagioclase	8.5	~100			9.4
Monzonite	Biotite	5.4	100	800[e]	650	
	Magnetite	1.8	100	610[e]		
HES-4b (+0.04)	Quartz	10.1	100			9.5
Gneiss	Biotite	5.5	100	750	650	
	Magnetite	1.9	100	625		
HES-4c (+0.6)	Quartz	10.1	100			10.4
Gneiss	Biotite	6.1	42	940	650	
	Magnetite	3.4	42	810		
HES-1 (+1300)	Quartz	12.8	0			11.9
Schist	Biotite	8.0	0	710	460	
	Magnetite	4.5	0	615		
Eldora Traverse						
EL-2 (+2)	Quartz	10.7	?			11.0
Gneiss	Biotite	6.7	?	940	650	
	Magnetite	2.7	?	645		
N-2 (+2)	Quartz	10.4	?			9.8
Amphibolite	biotite	5.8	?	750	650	
	Magnetite	2.3	?	635		
C-4B (+950)	Quartz	12.2	0			10.5
Gneiss	Biotite	6.9	0	630	490	
	Magnetite	3.7	0	595		
EL-1 (−50)	Quartz	9.9	0			8.6
Adamellite	Biotite	4.8	0	660	>650	
N-3 (−450)	Quartz	9.9	0			8.7
Adamellite	Biotite	4.9	0	680	>650	

[e] Plagioclase-mineral temperatures.

metamorphic rocks. The latter curve gives temperatures about 50—70° C lower than the quartz-magnetite temperatures of GARLICK and EPSTEIN (1967) in the range Δ quartz-magnetite = 8 to 11.

As might be expected, samples showing *complete* oxygen isotopic exchange between the intrusive and the country rocks give the most reasonable and concordant temperatures. The quartz-mica isotopic contact temperature at Sawtooth stock is about 555° C and at Santa Rosa stock 525° C. The quartz-magnetite isotopic contact temperature at Caribou stock is 625° C, but the quartz-biotite temperature of 750°C is probably too high. The contact temperature at the Birch Creek pluton is 540° C, and two samples collected from a xenolith embedded in the marginal zone of the Birch Creek pluton give quartz-biotite temperatures of 560 and 590° C, 20 to 50° C higher than the main contact temperature (SHIEH and TAYLOR, 1969). This is reasonable because a xenolith enclosed in the magma should in general have a distinctly higher temperature than the intrusive contact.

Those samples showing *no evidence* of exchange might also be expected to represent assemblages formed in isotopic equilibrium, because similar samples usually give similar quartz-mica isotopic temperatures and these temperatures are geologically reasonable. Examples are: the trondhjemites in the Sawtooth stock average at 525° C, the granodiorites in the Flynn stock 580° C, the granodiorite in the margin of the Santa Rosa stock 600° C, the adamellites in the central portion of the Santa Rosa stock 680° C, the quartz-monzonites in the Eldora stock 670° C, and the granite in the pluton of Birch Creek 570°C. The country rock temperatures are rather uniform outward for 1 to 70 feet from the contact: at Birch Creek 515 to 500° C and at Sawtooth 445 to 425° C. The temperatures in the Sawtooth aureole appear to be abnormally low in comparison with the contact temperatures (but see discussion below). The schists and gneisses in the Caribou and Eldora traverses give rather high quartz-biotite temperatures (630 to 710° C) for samples 1,000 feet away from the contact, but the quartz-magnetite temperatures (595 to 615° C) seem more reasonable, and these conceivably could represent the original regional metamorphic temperatures.

Temperatures Calculated from Heat Flow Models. The temperature distribution in the neighborhood of an intrusive-country rock contact is governed by (1) the temperature of the intrusion, (2) the temperature range of solidification of magma, (3) the thermal properties of intrusion and country rock, (4) the latent heat of solidification, (5) the extent of convection in the magma, (6) the heat of metamorphic reactions in the country rocks (including heat of vaporization of porewater), and (7) amount of heat carried away by material transport or by radiation. Factor (2) may be simplified to a definite melting point at the liquidus temperature, because JAEGER (1957) found that the width of the range of solidification is not important to the contact temperature or to the temperatures outside the intrusion.

For factor (5) one may take the extreme case by keeping the whole of the magma well stirred so that its temperature is uniform throughout its volume until solidification is completed. This will give an upper limit for the contact temperature and an over-estimate of temperatures in the country rock near the contact.

Factor (7) may be neglected because the isotopic data do not allow any appreciable massive movement of aqueous fluids from the intrusive into the country rock because of the very steep oxygen isotopic gradient (3 per mil per foot) preserved near the contact. Also, radiation is probably not important in a deep-buried body at the temperature of interest.

In factor (6), if the metamorphic reaction is exothermic, the country rock temperatures would be higher; if it is endothermic, the temperature would be lower. The vaporization of pore-water would also lower the temperature, but JAEGER (1959) has shown that the lowering of contact temperatures is not important for moderate to low values of rock porosity. Most of the rocks studied in this research probably have low porosities because of the antecedent regional metamorphism and depth of burial. For simplicity, factor (6) will not be considered.

Since we are particularly interested in the contact temperatures, because of our greater confidence in the oxygen isotopic geothermometers in these regions, the various contact "temperatures" based on different heat flow models are listed in Table 4 (based on calculations analogous to those of JAEGER, 1959).

Table 4. *Initial contact temperatures of an infinite igneous dike based on different heat flow models*

Heat flow model	Contact temperature °C	
	Magma = 700	Magma = 900
1. Equal thermal properties of country rock and intrusive, no latent heat of crystallization ($\sigma = 1$, $L = 0$).	400	500
2. Equal thermal properties of country rock and intrusive, with latent heat of crystallization ($\sigma = 1$, $L = 100$)	504	614
3. Equal thermal properties of country rock and intrusive, latent heat accounted for by exaggerating the magma temperature ($\Delta T = L/C = 300°$ C, $\sigma = 1$)	550	650
4. Same as model 1, except different σ ($\sigma = 2$, $L = 0$)	500	633
5. Some as model 2, except different σ ($\sigma = 2$, $L = 100$)	610	760
6. Convection (extreme case)	700	900

Initial country rock temperature = 100° C; $C_1 = C_2 = 0.25$ cal/gm° C;
L = latent heat (cal/gm); $\sigma = \left(\frac{K_2\,\varrho_2\,C_2}{K_1\,\varrho_1\,C_1}\right)^{1/2}$;
K = thermal conductivity; ϱ = density; C = specific heat; subscript 1 refers to country rock; 2 refers to solidified magma.

In the Sawtooth stock, the trondhjemite magma probably intruded below 750° C, and in the Santa Rosa and Flynn stocks the initial magma temperatures were probably close to 900° C (COMPTON, 1960). The thermal properties of granite and shale give a value for σ around 2 (LOVERING, 1936), but for a phyllite-granite contact, the σ-value is probably less than 2. If we assume the magma intruded at 700° C where the country rock was 100° C, the contact temperatures in Table 4 range from 700 to 400° C, according to different heat flow models. Model 1 assumes no latent heat; therefore it would give too low a temperature. Model 6 assumes the extreme case of convection; this would give too high a temperature. If the above two cases are disregarded, then the most probable initial contact temperature would lie somewhere between 500 and 610° C. If one assumes the magma

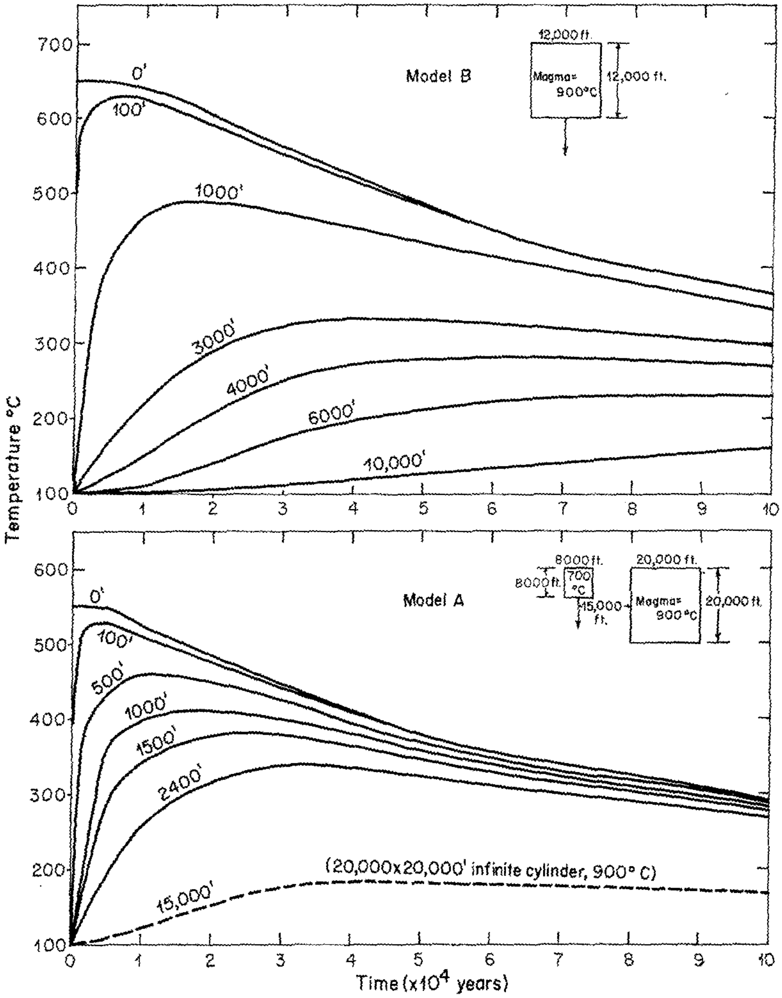

Fig. 20. Temperature vs. time curves for various infinite cylinder heat flow models, assuming initial country rock temperature = 100° C and diffusivity = 0.009 cm²/sec. Latent heat of solidification accounted for by exaggeration of intrusive temperature by 300° C. Model A is constructed analogous to the conditions of intrusion of the Sawtooth stock, and model B of the Flynn stock

temperature to be 900° C, the initial contact temperature probably would lie between 610 and 750° C, depending on the different kinds of heat flow models used. Since model 3, in which the latent heat is accounted for by exaggeration of the intrusion temperature, is mathematically very simple and gives a contact temperature almost in the middle of the most probable temperature range, it is the model utilized to construct the thermal history of the country rock as shown in Figs. 20, 21 and 22.

Fig. 20 shows two graphs containing various temperature vs. time curves. Model A was constructed by assuming an infinite cylinder of square cross section 8,000 × 8,000 feet, with initial intrusive temperature 700° C, country rock tem-

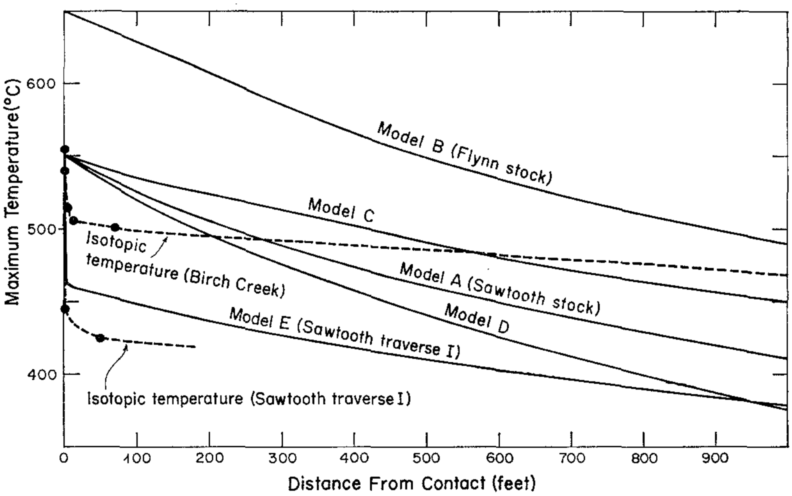

Fig. 21. Maximum temperature vs. distance curves for various heat flow models. Models A and B are described in Fig. 20. Models C and D both represent intrusions with initial temperatures of 700° C; Model C represents an infinite dike 8,000 ft. thick, and Model D is an infinite cylinder of square cross-section 8,000 × 8,000 ft. The latter is identical to Model A, but without the contribution of the nearby stock. Model E is described in Fig. 22. All parameters used in the calculations are listed in Table 4 (3). The dashed lines are isotopic temperature gradients measured for Sawtooth traverse I and for the Birch Creek traverse of Shieh and Taylor (1969). The Sawtooth traverse I isotopic "temperatures" should be analogous to Model E, and although we are dealing with a composite pluton, the Birch Creek isotopic "temperatures" should be roughly analogous to Model D. Except for Model E, the isotopic "temperatures" near the contact have a steeper gradient than do the maximum heat flow temperatures. The Birch Creek isotopic temperature gradient extrapolates to a value of 460° C at 1,300 ft. (sample BC-16). The Sawtooth I isotopic temperature gradient is not shown at distances greater than 150 ft., but it will be very gentle because the O^{18}/O^{16} fractionations at 600 ft. and 2,500 ft. are similar to or only slightly larger than those at 50 ft

perature 100° C, thermal diffusivity 0.009 cm^2/sec, and an exaggeration of the intrusive temperature by 300° C to account for the latent heat of crystallization. The model should be closely analogous to the conditions of intrusion of the Sawtooth stock. The heat contribution from the nearby Santa Rosa stock was also taken into account by assuming another parallel infinite cylinder of dimensions 20,000 feet × 20,000 feet and magma temperature 900° C located 15,000 feet away. Model B was constructed analogous to the conditions of intrusion of the Flynn stock (dimensions 12,000 × 12,000 feet and intrusive temperature 900° C). Fig. 21 is a plot of maximum temperature vs. distance for different models, together with the isotopic temperature gradients measured for Sawtooth traverse I and the Birch Creek traverse. The outer limit of the Sawtooth metamorphic aureole (~1,500 feet) will attain a maximum temperature of 380° C, and that of the Flynn aureole (~3,000 feet) will be 340° C. These seem to be "reasonable" temperatures for the beginning of metamorphic reactions in the lowest-grade rocks in the Santa Rosa Range.

Comparison of Oxygen Isotopic Temperatures with Heat Flow Calculations and with Mineral Parageneses. The oxygen isotopic temperatures and the temperatures calculated from heat flow models for a variety of samples are given in Table 3. The oxygen isotopic temperatures at the contacts of the Sawtooth and Birch Creek plutons are very similar: 555 and 540° C, respectively. These temperatures are quite close to those calculated from heat flow models 2, 3, 4, and 5 in Table 4, assuming an initial magma temperature of 700° C and country rock temperature of 100° C. The quartz-magnetite temperature of 625 °C at the contact of the Caribou stock is also very close to the heat flow temperatures calculated from models 2, 3, and 4, assuming an initial intrusive temperature of 900° C and country rock temperature of 100° C. The quartz-biotite temperature of 630° C 45 ft. from the contact of the Flynn stock is also compatible with heat flow Model B.

The central part of the Santa Rosa stock exhibits quartz-biotite temperatures of 660 to 695° C. These data, in conjunction with the heat flow models, suggest that the isotopic contact temperature of the Santa Rosa stock (525° C), is abnormally low, but this probably reflects the peculiar geometry of the intrusive at this locality, similar to the situation at Sawtooth traverse I as discussed below.

The oxygen isotopic temperatures of SRO-19B and SRO-19C, two schist samples collected 1 and 50 feet, respectively, away from the contact of the Sawtooth stock, are far too low compared with that of SRO-15B, a schist sample collected 0.08 feet away from the contact. SRO-19B gives a temperature of 445° C and SRO-19C of 425° C, about 90 to 110° C lower than the temperature of SRO-15B. Such a steep temperature gradient is not compatible with heat flow model A. However, because the intrusive at this locality forms a narrow wedge projecting into the country rock (see Fig. 2), the temperatures in the country rock should be considerably lower than the temperatures in the case of a plane contact. A more realistic model (E) representing such an intrusive wedge may be constructed by placing two small infinite cylinders of igneous rock, of square cross sections 500 × 500 feet and 50 × 50 feet, side by side at the surface of the 8,000 × 8,000 foot cylinder of the original model A, as illustrated in Fig. 22. The temperature vs. time curves in the 1-foot and 50-foot samples can then be calculated by superimposing temperatures obtained from Model A and the two additional cylinders. As may be seen in Fig. 22, the 1-foot sample attains a maximum temperature of 510° C, but it stays above 500° C for only 2 months; it cools below 400° C and then gradually rises to a maximum of 460° C and remains above 450° C for about 12,000 years. The 50-foot sample reaches a maximum temperature of 455° C, and remains above 450° C for about 9,000 years. These temperatures are in much better agreement with the quartz-mica isotopic temperatures.

The geometry of the intrusive in the Santa Rosa traverse also closely resembles that of Model E (see Fig. 9); the abnormally low oxygen isotopic temperatures in the country rock (0.08-foot sample 525° C and 2-foot sample 500° C) may largely be explained by Model E, but retrograde exchange and/or non-equilibrium are also plausible interpretations.

At the Birch Creek pluton, three schist samples collected 6 to 70 feet away from the intrusive contact give quartz-biotite isotopic temperatures from 500 to 515° C, 20 to 35° C lower than the temperature obtained from a sample 0.25 feet from the contact. The isotopic temperature gradient in this case is 35° C/70 feet; this is in fair agreement with a heat flow temperature gradient of 20° C/70 feet, based on

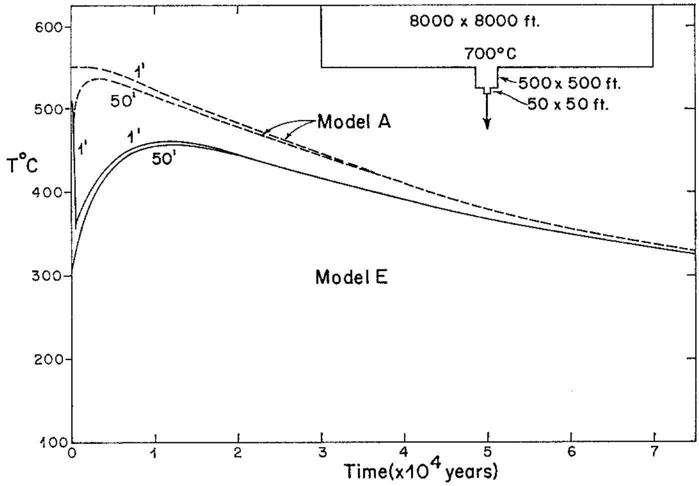

Fig. 22. Temperature vs. time curves for heat flow model E. The model is constructed by placing two more infinite cylinders of square cross sections 500 × 500 ft. and 50 × 50 ft. side by side at the surface of the 8,000 × 8,000 foot cylinder of Model A. Solid lines: cooling curves for 1-foot and 50-foot samples in Model E. Dashed lines: cooling curves for 1-foot and 50-foot samples in Model A. The maximum temperatures for the 1-foot and 50-foot samples are about 80 to 90° C lower in Model E than in Model A. Note that there are two maximum temperatures for the 1-foot sample in Model E (510 and 460° C), but the first maximum lasts for only a short time (the sample remains above 500° C for only 2 months). Model E closely resembles the intrusive geometry at the localities represented by Sawtooth traverse I and the Santa Rosa traverse (see Figs. 2 and 9), and may explain the abnormally low country-rock isotopic temperatures near both contacts

the cooling of a magma body (initial temperature = 700° C) in the shape of an infinite cylinder of square cross section 8,000 × 8,000 feet with plane contact (Model D in Fig. 21). The geometry of the latter cylinder is similar to that of the southern tip of the composite pluton in Birch Creek (see McKee and Nash, 1967, Fig. 2); however, note that the Birch Creek contact zone is adjacent to a younger granite intrusion only 250 ft. wide, and a steeper temperature gradient is to be expected at the border of such a small intrusion.

The isotopic temperatures in the marginal zones of the intrusions are often lower than the isotopic contact temperatures. For example, the quartz-biotite isotopic temperature difference between the marginal intrusive and the contact is about 80° C at the Eldora and Caribou stocks, about 30° C at the Sawtooth, and about 60° C at the Flynn stock. It is thus possible that the oxygen isotopic fractionations might be "quenched in" not far below the temperature maximum at the various contacts, but might continue to re-equilibrate in the marginal portions of the intrusions until a somewhat lower temperature is reached. Fracturing and shearing are very likely in the marginal zones of intrusive stocks during late stages of solidification, and this might allow access to aqueous fluids and promote deuteric recrystallization. Kennedy (1955) has suggested that the marginal portions of the magma could have crystallized at a lower temperature than the central portion because of the higher P_{H_2O} in the margins. Also, as will be discussed below, migration of water from the country rocks into the intrusive might be significant in certain stocks studied in this research.

It was observed by COMPTON (1960) and by the present writers that in the innermost part of the Flynn and Santa Rosa aureoles K-feldspar has formed from muscovite and biotite, and sillimanite has apparently formed by breakdown of both andalusite and micas. In the Sawtooth aureole no K-feldspar has been found, but sillimanite has formed at the contact from some biotite and andalusite grains. Using the experimentally determined muscovite plus quartz breakdown curve (EVANS, 1965), and the Al_2SiO_5 polymorphic transition curves (WEILL, 1966; FYFE and TURNER, 1966; RICHARDSON *et al.*, 1968), and assuming that P_{H_2O} is approximately equal to P_{Total}, it can be determined that the temperature near the contacts of the Flynn and Santa Rosa stocks was probably between 550 and 600° C, and the water pressure was between 1,000 and 1,500 bars (see TURNER, 1968, p. 248—252). These mineralogical "temperatures" are in good agreement with temperatures derived from oxygen isotopic geothermometers and from heat flow models. The P_{Total} also agrees with the independent estimate of 1 to 2 Kb by COMPTON (1960) based on geological evidence. Inasmuch as no K-feldspar has been found in the Sawtooth aureole, and sillimanite has formed from andalusite, either the temperature was lower and/or P_{H_2O} somewhat higher in the Sawtooth aureole than in the Flynn and Santa Rosa aureoles. The isotopic data are definitely compatible with higher P_{H_2O} (see below), and the isotopic contact temperatures of the Sawtooth stock also appear to be somewhat low.

Extent of Isotopic Exchange and Dimensions of the Isotopic Equilibrium System

General Statement. Studies of the variations of oxygen isotopes in the vicinity of an intrusive contact offer an excellent opportunity to investigate the extent and possibly the mechanism of isotopic exchange between the intrusive and its country rock, because the oxygen isotopic composition of igneous rocks and minerals are quite different from those of sedimentary and low-grade regional metamorphic rocks and minerals. There is always a tendency for different isotopes to undergo exchange in the presence of an isotopic compositional gradient. Isotopic equilibrium is established in a system only if the isotopic compositions of a particular mineral are equal everywhere in that system, provided that system is everywhere at a constant temperature.

Three types of oxygen isotopic exchange will be distinguished in the following discussion. The small-scale oxygen isotopic exchange observed close to the contact involves a steep isotopic gradient that generally extends only 2 to 3 feet on both sides of the contact. This exchange probably takes place essentially in the solid state through a diffusion-controlled process. The large-scale oxygen isotopic exchange is observed in the marginal zones of the stocks or plutons where "abnormally" high O^{18}/O^{16} ratios exist; this probably takes place when the intrusions were largely molten. A third type of isotopic exchange (also large-scale) is observed in the Birch Creek xenolith (SHIEH and TAYLOR, 1969) and Sawtooth reentrant. Widespread movement of aqueous fluids is probably an important factor in both types of large-scale isotopic exchange.

Small-Scale Oxygen Isotope Exchange. The small-scale oxygen isotopic exchange occurs at the immediate contact between the intrusion and the country rocks where a steep (1 to 3 per mil per foot) oxygen isotopic compositional gradient is observed. Beyond a few feet away from the contact, both on the intrusive side and on the country rock side, the O^{18}/O^{16} ratios of the individual minerals or whole rock samples remain quite constant (see the various δO^{18}-distance plots in Figs. 3, 4, 7, 8, and 10). Note that even in the presence of such steep isotopic gradients a "normal" sequence of O^{18}-enrichment of the coexisting minerals is preserved in all the traverses. The shapes of the curves in the δO^{18}-distance plots indicate that O^{18}/O^{16} exchange between the intrusive and the country rock has occurred to varying degrees depending upon distance from the contact. If 100% exchange occurs, the O^{18}/O^{16} ratios of a particular mineral in the intrusive and in the country rock will be equal. If no exchange occurs at all, a particular mineral will retain its original O^{18}/O^{16} ratio. Inasmuch as the O^{18}/O^{16} ratios of the individual minerals remain quite constant beyond a few feet away from the contact, such samples are assumed to represent 0 percent exchange. The percent isotopic exchange for different minerals from various traverses has been calculated. The results are presented in Table 3 and Fig. 23.

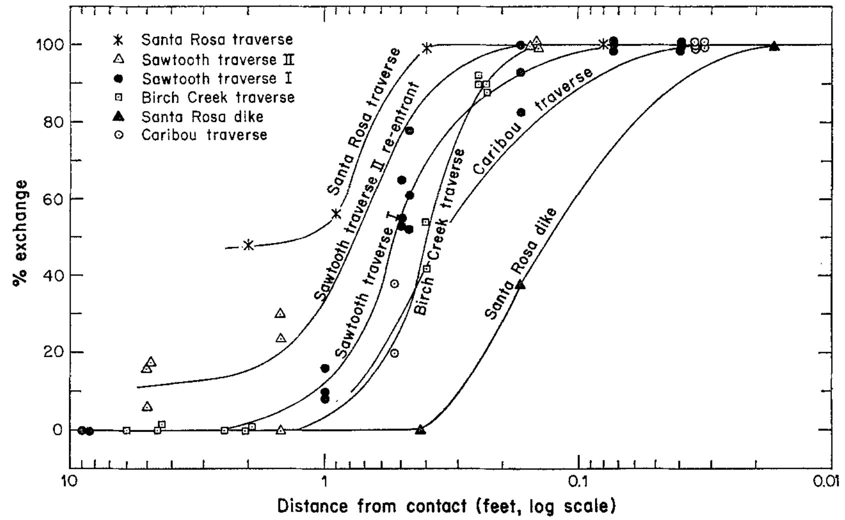

Fig. 23. Percent oxygen isotope exchange (small-scale) between intrusive and country rock for several traverses studied in the present work. Note that the Caribou curve is calculated from assumed "country-rock" δ-values for feldspar (9.8) and biotite (6.6) obtained from average values of these minerals in gneiss of the Eldora area (see Tab. 1)

The extent of oxygen isotopic exchange in most cases does not exceed 25% for samples beyond 2 feet from the contact, possibly with the exception of the Santa Rosa stock where a 2-foot country-rock sample appears to have undergone 50% exchange. There is no marked difference in the degree of exchange among the coexisting minerals analyzed in a rock, except for the 7-inch sample from the Caribou traverse in which quartz has undergone 100% exchange whereas magnetite and biotite have only experienced 40% exchange. It should be pointed out that this magnetite is much coarsergrained than the Sawtooth magnetites discussed previously.

Garlick and Epstein (1966, 1967) analyzed the O^{18}/O^{16} ratios of coexisting minerals in a schist xenolith enclosed in trondhjemite and the hydrothermal wall-rock alteration envelopes at Butte, Montana. They found that garnet undergoes oxygen isotopic exchange in the xenolith only with difficulty, whereas ilmenite, muscovite, and biotite are the easiest minerals to exchange. In the hydrothermally altered zone at Butte, K-feldspar has experienced much isotopic exchange whereas quartz is not appreciably affected by the alteration process. The same effects have also been observed by Taylor (1968b) in red-rock granophyres. No garnet was analyzed in the present study, but the feldspar samples analyzed from two traverses in the Sawtooth stock show essentially the same degree of exchange as coexisting quartz and biotite. The apparent discrepancy between the present result and that of Taylor (1968b) and Garlick and Epstein (1967) is not completely understood. However, it is known that variations in temperature and in the alkali concentrations of the hydrothermal fluids can have important effects on exchange processes involving feldspars (see O'Neil and Taylor, 1967).

The dimensions of the oxygen isotopic equilibrium system during such small-scale oxygen isotopic exchange can be determined by measuring the maximum distance between samples that have undergone essentially 100% isotopic exchange, such that a particular mineral has attained essentially the same O^{18}/O^{16} ratio in the samples. It is found that such an equilibrium system is quite small, being about 1 foot in the Santa Rosa traverse, 0.5 feet in Sawtooth traverse II, 0.2 feet in Sawtooth traverse I, 0.3 feet in Birch Creek, 0.1 feet in Caribou traverse, and 0.05 feet in the Santa Rosa dike (compare with Anderson, 1967).

The size of such an exchange system may depend upon (1) temperature, (2) time of heating, and (3) availability of an exchange medium (oxygen-bearing fluids). The dimensions of oxygen isotopic equilibrium systems estimated above correlate quite well with combinations of such parameters. For example, the Santa Rosa stock is the biggest stock; as a consequence, the country rock should be held above a specific temperature for a much longer time than in the case of the smaller stocks. The Sawtooth stock is probably the "wettest" of the intrusives. Traverse II of Sawtooth is in the reentrant where the temperature should have been abnormally high. All these factors (higher temperatures, longer heating, and higher water pressure) should facilitate the isotopic exchange.

Note that even though the isotopic contact temperatures of the Santa Rosa stock are unusually low (possibly as a result of a local convexity in the configuration of the contact), the extent of exchange shown in Fig. 23 is larger than for any other stock. This suggests that the length of time the contact remains above a specified temperature is more important than slightly higher contact temperatures.

The Caribou stock contact zone was probably the "driest' of those studied, because the country rock had already been regionally metamorphosed to sillimanite grade prior to the contact metamorphism; note that the Caribou traverse indeed does show the smallest extent of exchange. The Santa Rosa dike is only about 5 inches thick, and crosscuts hornfels about 30 feet away from the main intrusive contact. The heating effect of the country rock by this dike must have been very small, but it still shows significant exchange (Fig. 23); this is probably largely a result of the long "heating time" in the Santa Rosa aureole.

It is interesting to inquire into the mechanism of such small-scale isotopic exchange. Isotopic exchange may occur through the mechanism of (1) recrystallization, perhaps by diffusion along cleavage cracks or crystal imperfections, accompanied by oxygen isotopic equilibration in a fluid film at the interface between exchanged and unexchanged portions of mineral grains, as proposed by O'NEIL and TAYLOR (1967) in their experiments on oxygen and cation exchange in feldspars; or by (2) solid-state diffusion through the crystal lattice of each mineral. Mechanism (2) is a very slow process. The diffusion coefficient of oxygen in quartz relative to the *c*-axis was determined by CHOUDHURY *et al.* (1965) at 667° C to be 8×10^{-14} ($\perp$) and 4×10^{-12} ($\parallel$) cm²/sec, using the nuclear reaction O^{18} (p, α) N^{15}; using $D = 4 \times 10^{-12}$ cm²/sec and $t = 10^4$ years, the average distance travelled by the oxygen along the *c*-axis of quartz would be only 1.6 cm. The distance would be much smaller at temperatures of 500 to 600° C. On the other hand, mechanism (1) is a very effective process, because intergranular or surface diffusion coefficients are several orders of magnitude larger than lattice diffusion coefficients, particularly in the presence of a fluid film; essentially complete isotopic exchange in mineral grains is possible within reasonable experimental times (see O'NEIL and TAYLOR, 1967).

Another argument against the effectiveness of simple solid-state diffusion is the fact that the isotopic fractionations among coexisting minerals apparently were "frozen in" at approximately equilibrium values at temperatures just below the maximum contact temperatures. All the minerals are involved in the exchange, suggesting that the mineral assemblages have undergone complete recrystallization during the small-scale exchange, and that only minor adjustments of O^{18}/O^{16} ratios occurred by solid diffusion during the further temperature decline. This is particularly clear in Sawtooth traverse I, where a very steep isotopic temperature gradient is preserved and the isotopic contact temperature of about 555° C should be very close to the maximum temperature attained at this contact.

The oxygen isotopic data in the present study do not seem to allow any transfer mechanisms which require abundant or massive movement of solution or fluid phase across the solidified intrusive contact, because of the two-direction exchange, the existence of steep oxygen isotope gradients (as steep as 3 per mil/ft.), and the very narrow zone (less than a few feet) that has taken part in the isotopic exchange. It is more likely that the oxygen isotopes were transferred from the intrusive to the country rock, and vice versa, by diffusion along grain-boundaries in the presence of an essentially static interstitial fluid film.

Oxygen isotopic zoning of mineral grains in the exchanged zones does not exist, at least in so far as the present experimental accuracy is concerned. The steepest isotopic gradient observed in the present study is about 0.015 per mil/mm, and the largest average grain-size of minerals other than andalusite is about 2 mm. Therefore, in the exchange zone the maximum δ-value difference between two adjacent grains of a mineral is not more than 0.03 per mil.

In all of the traverses studied except Sawtooth traverse I, the maximum δO^{18}-lowering in the country rock is greater than the maximum δO^{18}-enrichment in the intrusive in the small-scale isotopic exchange zones (Table 5). This presumably is due to the fact that such small-scale isotopic exchange would begin to take place as soon as the country rock is intruded, but the isotopic gradient on the intrusive side would not be well preserved until after the magma has become solidified. However, in traverse I of the Sawtooth stock, where the intrusion forms a narrow wedge surrounded by the country rock, the δO^{18}-lowering in the country rock is less than the δO^{18}-increase of the intrusive, in keeping with the undoubtedly more rapid crystallization of the trondhjemite near this contact. It is also noted that the ratio δO^{18}-lowering/δO^{18}-increase is slightly greater in quartz than in biotite for all the traverses studied in this research.

Large-Scale Oxygen Isotopic Exchange in the Igneous Rocks. Large-scale oxygen isotopic exchange effects are here defined as those observed over a scale of hundreds of feet or more. Unlike the small-scale exchange discussed above, except for two examples discussed below, these effects are confined to the igneous rocks.

Table 5. *Comparison of maximum δO^{18}-lowering in the country rock vs. δO^{18}-increase in the intrusive for various small-scale isotopic exchange zones*

Traverse	Mineral	δO^{18}-lowering (country rock) (per mil)	δO^{18}-increase (intrusive) (per mil)	Ratio of δO^{18}-lowering to δO^{18}-increase
Sawtooth I	Quartz	3.2	4.3	0.7
	Biotite	2.0	4.5	0.4
Sawtooth II	Quartz	3.2	0.8	4.0
	Biotite	3.8	1.2	3.2
Santa Rosa	Quartz	3.4	1.5	2.3
	Biotite	3.2	1.8	1.9
Birch Creek[a]	Quartz	6.0	1.2	5.0
	Biotite	4.9	1.3	3.8
Caribou and Eldora	Quartz	2.7	0.2	13.0
	Biotite	2.5	0.9	2.8
	Magnetite	2.6	0.2	13.0

[a] Shieh and Taylor (1969).

They have been observed in the marginal portions of the Santa Rosa, Flynn, Sawtooth, and Birch Creek plutons as much as 300 to 500 feet inward from the contact. A small igneous plug (300 × 600 feet) intruded into the outer aureole of the Santa Rosa stock shows even more marked effects than the examples mentioned above.

The evidence for such large-scale isotopic exchange is the "abnormally" high O^{18}/O^{16} ratios of minerals in the marginal portions of the intrusives compared to those obtained from the central portion of the Santa Rosa stock and from the "normal" plutonic granitic rocks reported by Taylor and Epstein (1962), as shown in Table 6. It is not known as yet exactly how far these isotopic effects extend inward into the intrusions; further sampling is required.

It seems certain that the "abnormally" high O^{18}/O^{16} ratios of the marginal parts of these intrusions are the result of contamination of an originally "normal" igneous granitic rock (mostly in a molten state) with O^{18}-rich material. The only logical source of such O^{18}-rich material is the metasedimentary country rock that has been intruded. A "normal" igneous rock (or magma) can be isotopically contaminated by the country rock in the following ways: (1) simple melting or assimilation of stoped blocks, (2) exchange with xenoliths enclosed in the magma, (3) exchange with country rock by means of aqueous fluids, and (4) absorption of water from the country rock. Mechanism (4) is actually a variant of (3), except that the aqueous fluids are permanently incorporated into the intrusive instead of merely moving through the magma or the hot, solidified stock, and exchanging oxygen with it.

Simple melting or assimilation of country rock can be important if the magma is intruded at high temperatures. There is geologic evidence that many of the stocks in the Santa Rosa Range are dominantly crosscutting (and stoped) bodies (COMPTON, 1960, p. 1397). Some assimilation of the country rock by the magma almost certainly occurred at the time of intrusion. However, the compositional pattern of the Santa Rosa stock does not support the idea of simple assimilation of appreciable quantities of country rock at the present level of exposure (COMPTON, 1960, p. 1401). A material-balance calculation from the oxygen isotopic data shows that sample SRO-29F, collected 150 feet inward from the Santa Rosa stock contact, requires that 30 atom per cent of its oxygen be derived from the country rock if simple assimilation is the only mechanism. Such large amounts of assimilated metasediments would certainly be reflected in the chemical composition of the marginal zone of the intrusion.

Table 6. *Comparison of O^{18}/O^{16} ratios of quartz and biotite from marginal portions of various intrusions with those from the central portion of Santa Rosa stock and from "normal" plutonic granitic rocks*

Pluton		δ quartz (per mil)		δ biotite (per mil)	
Central zones of plutons					
"Normal" plutonic granitic rocks (TAYLOR and EPSTEIN, 1962)		9.9 ± 0.3[a]	(6)[b]	5.0 ± 0.4	(5)
Central portion of Santa Rosa stock		9.5 ± 0.0	(2)	4.5 ± 0.1	(2)
Marginal zones of plutons					
Sawtooth stock		11.5 ± 0.1	(4)	5.4 ± 0.1	(4)
Flynn stock	North	13.5 ± 0.8	(2)	7.8 ± 1.1	(2)
	South	10.6 ± 0.1	(2)	4.8 ± 0.2	(2)
Santa Rosa stock		13.1	(1)	6.8	(1)
Small intrusive plug		14.0	(1)	8.1	(1)
Birch Creek pluton[c]		12.3 ± 0.6	(4)	6.0 ± 0.4	(2)
Eldora and Caribou stocks		9.9 ± 0.0	(2)	4.8 ± 0.2	(3)

[a] The uncertainty shown is average deviation from the mean.
[b] Numbers in parentheses represent number of samples.
[c] SHIEH and TAYLOR (1969).

Mechanism (2) should be very effective for large-scale oxygen isotopic exchange. A xenolith immersed in the magma may undergo isotopic exchange to such an extent that the O^{18}/O^{16} ratios of minerals in the xenolith become essentially identical to the ratios in the intrusive without melting the xenolith. An example of such an exchange mechanism has been observed in a schist xenolith in the Birch Creek pluton.

O^{18}-enrichment of a magma can also be the result of isotopic exchange with country rock by means of aqueous fluids (i.e., mechanism 3). This can occur by diffusion of H_2O through the silicate melt, or if the magma is saturated with H_2O, by bubbles of H_2O rising through the melt. Diffusion is much more rapid in a liquid than in a solid and transfer will also be enhanced by convective mixing. Two examples may possibly serve to illustrate this effect. Sample SRO-D is a pegmatite dike 3 inches thick crosscutting schist of the Sawtooth aureole. The pegmatite consists of a pure quartz core about 0.5 inch thick bordered on both sides by very coarse-grained K-feldspar, muscovite and minor quartz. The δO^{18}-values are: quartz (core) 18.3, quartz (border) 17.8, K-feldspar 12.4 and muscovite 15.7 per mil. As discussed above, the O^{18}/O^{16} ratios of the quartz and muscovite in this pegmatite are almost identical to those of the corresponding minerals in the schist. This cannot be the result of simple small-scale isotopic exchange with the schist after solidification of the pegmatite, because the quartz in the core is even more O^{18}-rich than in the border zone and the quartz and muscovite are both extremely coarse-grained. In addition, the same effects observed in the pegmatite can also be seen in a 0.5 inch thick aplite dike crosscutting SRO-19B (schist). The aplite is fine-grained (0.3 mm) and is composed of quartz ($\delta = 17.4$), feldspar, and muscovite ($\delta = 16.1$). The quartz in the aplite is 1.3 per mil lower in δO^{18} than the quartz in the adjoining schist. However, inasmuch as the muscovite in the aplite is 0.8 per mil higher in δO^{18} than the muscovite in the schist, this cannot have been the result of simple isotopic exchange in the crystalline state. It is very likely that the aplite and pegmatite were derived from the Sawtooth stock, indicating that they have both been enriched in O^{18} by at least 6 per mil. The fact that similar effects are observed in two granitic dikes with such vastly different grain sizes and textures suggests that isotopic exchange in both occurred at least partially with liquid magma.

For purposes of calculation, let us assume that mechanism (3) is entirely responsible for the exchange effects observed in the Santa Rosa stock and small plug. If we assume that $\delta_{H_2O} = 14.0$ in the country rock, then a simple material-balance calculation indicates that, respectively, about 35 and 50 weight percent H_2O must move through the marginal zone of the stock and small plug. This is based on an assumed temperature of 800° C in the intrusive and utilization of the data of O'Neil and Taylor (1967) in conjunction with that in Table 6.

Mechanism (4), the absorption of water from the metasedimentary rocks into the magma, can be significant if the magma is relatively "dry" (i.e., if it is strongly undersaturated in H_2O). There is geologic evidence that the Santa Rosa and Flynn stocks were "drier" than the Sawtooth stock when emplaced (Compton, 1960, p. 1408); a much larger quantity of water might migrate from the country rocks into the "dry" stocks than into the "wet" stock. This is apparently borne out by the oxygen isotopic data, because the marginal portions of the Santa Rosa stock, the north side of the Flynn stock, and the whole portion of the small plug in the Santa Rosa aureole show the greatest O^{18}-enrichment. As the Sawtooth stock was presumably much more nearly saturated in H_2O, there would have been less tendency for influx of H_2O from the country rocks, and the Sawtooth stock does indeed show significantly less marginal isotopic contamination. In the Eldora and Caribou stocks the country rocks were regionally metamorphosed to a very high grade prior to the intrusion of the stocks; one would not expect any appreciable influx of water from such "dry" country rocks, and the marginal intrusive samples in fact show no signs of oxygen isotopic contamination (Table 6).

In summary, each of the four mechanisms discussed above is probably involved in the O^{18}-enrichment of the minor intrusions and the marginal portions of the larger igneous bodies. The relative importance of each mechanism will depend largely upon the physical conditions of emplacement of the plutons. It should be pointed out that very significant oxygen isotopic effects have occurred with little or no evidence of bulk chemical changes in the rocks. In most of the stocks examined, xenoliths are very rare, and mechanisms (3) and (4) may be predominant. In the Birch Creek pluton, however, xenoliths are abundant near the contact and they have undergone extensive isotopic exchange, suggesting that mechanism (2) perhaps prevails at this locality (Shieh and Taylor, 1969).

Large-Scale Isotopic Exchange in Xenolith and Re-entrant. The only places where large-scale oxygen isotopic exchange effects were found in metasedimentary rocks are in the Birch Creek xenolith and in the Sawtooth re-entrant. Shieh and Taylor (1969) analyzed two samples collected from the center and the margin of a schist xenolith having an exposed area 10 × 50 feet in Birch Creek. The O^{18}/O^{16} ratios of minerals in these two samples are essentially identical to those of the same minerals in the adjacent granite (Table 3). If we assume that the xenolith originally had O^{18}/O^{16} ratios similar to the schist outside the pluton, then a lowering of δ-values by about 6 per mil is indicated; the size of the oxygen isotopic equilibrium system in the xenolith is at least 10 feet (the width of the xenolith). This is more than 30 times larger than at the main contact.

In the re-entrant of Sawtooth traverse II where an embayment of country rock projects into the intrusive, the O^{18}/O^{16} ratios of quartz in two schist samples collected 5 and 85 feet away from the contact show a uniform lowering of about 3 per mil in comparison with O^{18}/O^{16} ratios for samples outside the re-entrant. This cannot be due to the same mechanism responsible for small-scale isotopic exchange discussed earlier because a further abrupt lowering of O^{18}/O^{16} ratios, which is characteristic of the small-scale exchange, occurs within 1 foot of the contact.

It is likely that the samples collected from the xenolith and the re-entrant were maintained at high temperatures for very long periods of time. Both heat flow considerations (Jaeger, 1961) and oxygen isotopic data (smaller fractionations among coexisting minerals) support this view. Although diffusion would be faster in such an environment than at the main contact because of the higher temperatures involved, the widths of the isotopically exchanged zones would not be expected to be so vastly different. Therefore, a much more effective method of material transport seems to be required to explain the large-scale exchange in the xenolith and re-entrant. In this connection, it may be significant that the Sawtooth and Birch Creek plutons appear to have been more H_2O-rich than the other intrusions studied in this work. It is logical to infer that a hot, low-density, aqueous fluid would tend to move upward in a gravitational field even if it were present only as a fluid film in the pore spaces of rocks. Therefore, one can expect much larger-scale effects in country rocks that are underlain by or surrounded by an intrusive igneous rock that is a source of volatiles. The only two examples of such geometry in the present study or that of SHIEH and TAYLOR (1969) are the xenolith and re-entrant. For all other cases, the intrusive-country rock contact is essentially vertical, and upward-moving volatiles would produce little isotopic exchange between the intrusive and country rock. Thus, whereas horizontal outward movement of H_2O from the magma is negligible, upward transport may be quite important.

Isotopic Fractionations in Dehydration Reactions

The effect of dehydration reactions upon the oxygen isotopic compositions of the pelitic rocks can be studied by examining the variations of the oxygen isotopes across the metamorphic zones. The best sets of samples on which to study such isotopic effects are those from the Santa Rosa Range. The O^{18}/O^{16} ratios, either of whole rock or of individual minerals, do not show any noticeable change across the metamorphic zones in any of the traverses studied. Three explanations are possible: (1) No significant fractionations of oxygen isotopes between water and pelitic rock accompany dehydration reactions. (2) Oxygen isotopic fractionations occur during dehydration reactions but the amount of water escaping from the system is negligible compared with the total oxygen in the rock. (3) Isotopic fractionations occurred, but later isotopic homogenization has swept out all the isotopic inequalities. This last possibility is rejected because of the demonstrably small size of the country-rock systems that have undergone homogenization of oxygen isotopes.

Possibility (1) can be evaluated in terms of experimental data. The oxygen isotopic compositions of pelitic rocks and the coexisting waters in equilibrium with various mineral assemblages are given in Tables 1 and 3. It is noted that δ rock and δH_2O are similar, particularly for samples close to the contact. The few samples collected from the main parts of the contact aureoles show δ rock higher than δH_2O by about 2 to 4 per mil. Therefore, if the dehydrated water is equilibrated with the rock before escaping, the dehydration process will either produce no effect, or at most will increase the δ rock by only 0.1 to 0.2 per mil (see below). There is a faint suggestion in the isotopic data that rocks in the aureole are perhaps slightly richer in O^{18} than their unmetamorphosed equivalents (see Fig. 7).

Possibility (2) can be evaluated by calculating the proportion of oxygen liberated during the dehydration reactions. COMPTON (1960, p. 1408) has derived the following balanced dehydration reactions in which the actual stoichiometric amounts of reactants and products are taken into account:

$$\begin{aligned}1.1\ \text{chlorite} + 1\ \text{muscovite} = 1\ \text{biotite} + 1.5\ \text{andalusite} + 0.25\ \text{staurolite}\\ + 1.2\ \text{quartz} + 4.1\ H_2O\,.\end{aligned} \tag{1}$$

$$\begin{aligned}1.4\ \text{chlorite} + 1\ \text{muscovite} + 1.3\ \text{quartz} = 1\ \text{biotite} + 1\ \text{cordierite}\\ + 1\ \text{andalusite} + 5.6\ H_2O\,.\end{aligned} \tag{2}$$

$$1\ \text{chlorite} + 1\ \text{muscovite} = 1\ \text{biotite} + 2.5\ \text{andalusite} + 1\ \text{quartz} + 4\ H_2O \tag{3}$$

From the above dehydration reactions, it is calculated that the atom percent of oxygen liberated as dehydrated water relative to the total oxygen in the reacting minerals is 12.9%, 14.1%, and 13.3%, respectively, for these three reactions. Inasmuch as the rocks are composed of other non-hydrous constituents as well, any kinetic fractionation during dehydration would have to significantly exceed 10 per mil in order to produce as much as a 1 per mil change in the country rock; in actual fact, the aureoles appear to have lost 2 to 3 weight per cent H_2O during dehydration (COMPTON, 1960, p. 1403). Thus, both the actual oxygen isotopic data and material-balance considerations indicate that dehydration reactions would not produce any appreciable oxygen isotopic changes during contact metamorphism.

The D/H ratios of the pelitic rocks also do not show any obvious change across the contact metamorphic zones. Therefore, no significant fractionations of hydrogen isotopes accompany contact metamorphic dehydration reactions, either.

Source and Movement of Water During Contact Metamorphism

Water in contact metamorphism can be derived either from the country rock or from the intrusive. Water in the country rock probably comes mainly from metamorphic dehydration reactions, but meteoric water may also be present in the pore spaces and fissures of the country rock. The amount of water in the magma depends in part upon the solubility of water in the magma. Inasmuch as the solubility of water in a silicate melt is strongly pressure-dependent, an ascending magma may separate an aqueous phase during its course of intrusion. The total pressure on the stocks in the Santa Rosa Range during their intrusion is estimated to be about 1 to 2 Kb; a granitic melt under such conditions could contain at most about 5.5 weight percent H_2O (BURNHAM, 1967).

In textbooks of petrology, it is generally conceded without much justification that "near granite contacts a permeable rock undergoing metamorphism may be continuously flushed with outward-flowing water" (TURNER, 1968, p. 244). Similar statements emphasizing the idea of outward movement of water through contact metamorphic aureoles from granitic intrusions can be found in HARKER (1939, p. 23), BARTH (1962, p. 262), and TURNER and VERHOOGEN (1960, p. 658). The oxygen isotopic data are pertinent to the questions as to the source and movement of water during contact metamorphism, because the water equilibrated with the magma can definitely be distinguished from that equilibrated with the sediments or low-grade metamorphic rocks.

From Table 3, it is seen that the calculated O^{18}/O^{16} ratios of waters are about 5 to 7 per mil higher in the metamorphic rocks than in the intrusives. Also note that the O^{18}/O^{16} ratios of waters are rather uniform within the intrusive or within the country rocks in an individual contact zone, and that intermediate O^{18}/O^{16} ratios are obtained only right at the contacts. Inasmuch as the exchanged zone in the country rock at these contacts is less than a few feet wide, horizontal outward movement of large quantities of H_2O across the contact during the crystallization of the magma is ruled out. The idea that a large amount of H_2O is expelled outward through the entire contact aureole from the granitic intrusions, or that H_2O acts as a significant outward heat carrier, is not compatible with the isotopic data presented here.

The oxygen isotopic data, however, are compatible with the concept that upward movement of H_2O is probably far more important and operative at a much larger scale than the horizontal outward movement of H_2O from the magma. The large-scale isotopic exchange observed in the xenolith and in the re-entrant was cited earlier as possible evidence of appreciable upward movement of H_2O. Also, massive upward movement of volatiles could occur within the contact aureole, and no isotopic trace of such a phenomenon would remain, because such H_2O would be isotopically similar to the country rock. From chemical analyses of the metasedimentary rocks in the Santa Rosa Range, COMPTON (1960) concludes that 2 to 3 weight percent H_2O was driven out of the contact aureoles during prograde metamorphism. The volume of rock in each contact aureole is 1.5 to 2 times that of the adjoining intrusive. Therefore, even if the magmas were nearly saturated with H_2O (and the Sawtooth stock is the only body for which this is even a possibility), at least as much H_2O is liberated from the country rocks as from the intrusion. In such a situation there is probably little or no tendency for H_2O to move either outward or inward across the intrusive contacts, because H_2O is being simultaneously released by igneous crystallization as well as contact metamorphism.

If a "dry" magma is emplaced into relatively "wet" sediments or metasedimentary rocks, however, considerable migration of H_2O from the country rocks into the stock certainly should occur. It was pointed out above that the abnormally high O^{18}/O^{16} ratios of the marginal portions of the Santa Rosa and Flynn stocks may in large part be due to such an effect. Note that COMPTON (1960, p. 1408) recognized this possibility, even though emphasizing the importance of igneous H_2O in the contact aureoles. It is therefore likely that "wet" intrusions such as the Sawtooth stock display evidence of higher P_{H_2O} in their contact aureoles, not necessarily because of massive outward movement of H_2O, but because there is much less *inward* movement of dehydration H_2O and hence less "dessication" of the country rocks.

The rough correlation between the D/H ratios of biotite and those of local surface waters (see below) may possibly be interpreted to mean that, in addition to the dehydration H_2O, some meteoric ground water was incorporated into the shallow intrusions studied in the present research. TAYLOR (1968b) has shown on the basis of O^{18}/O^{16} data that much larger quantities of ground water have been incorporated into certain other shallow plutons such as the Mull, Skye, Ardnamurchan, and Skaergaard intrusions. Except perhaps for the SRO-D pegmatite feldspar and the magnetites of the Sawtooth contact zone, no oxygen isotopic evidence for meteoric water has been found in the present study. However, much smaller quantities of meteoric water are sufficient to produce a change in the D/H ratio of a rock than to change its O^{18}/O^{16} ratio.

Comparison of Isotopic Relationships in Contact and Regional Metamorphic Rocks

One of the major problems in metamorphic petrology is whether truly magmatic or "juvenile" H_2O plays an important role during metamorphism. Oxygen isotopic analysis of metamorphic rocks may help to unravel the problem. If the metasedimentary rocks have undergone isotopic exchange with significant quantities of pore-fluids in communication with or derived from plutonic igneous rocks during metamorphism, one should expect a decrease in O^{18}/O^{16} ratios with increasing metamorphic grade. If metamorphism only involves simple loss of water through dehydration reactions, then the O^{18}/O^{16} ratios of metasedimentary rocks should remain essentially constant during progressive metamorphism.

Fig. 24 shows a plot of the O^{18}/O^{16} ratios of pelitic rocks from different metamorphic grades and zones of regional and contact metamorphism. The O^{18}/O^{16} ratios from regional metamorphic rocks were taken or calculated from data of

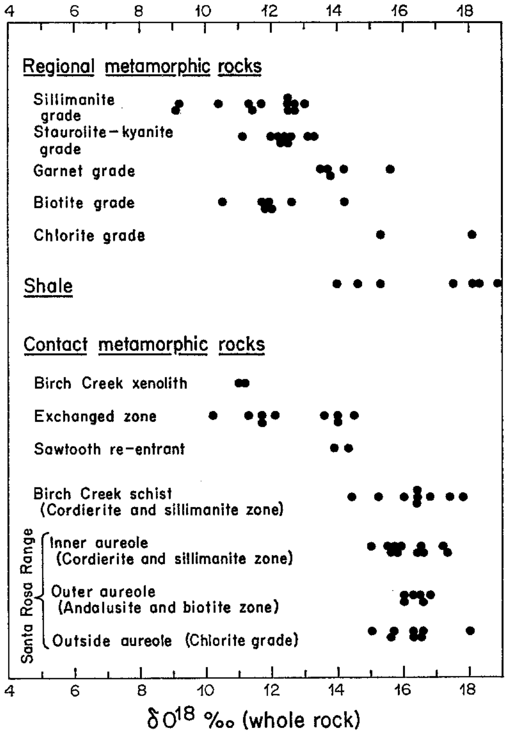

Fig. 24. O^{18}/O^{16} ratios of pelitic rocks plotted against metamorphic grade for both regional and contact metamorphism. Data are from the present study, Garlick and Epstein (1967), Clayton and Mayeda (1963), Taylor (1968a), Taylor and Epstein (1964), Shieh and Taylor (1969), and Savin (1967)

Garlick and Epstein (1967), Clayton and Mayeda (1963), and Taylor (1968a). A few garnet-grade pelitic schists from southeastern Vermont were excluded because of the abnormally low O^{18}/O^{16} ratios observed in that region. The O^{18}/O^{16} ratios of shale were taken from Taylor and Epstein (1964) and Savin (1967). For contact metamorphic rocks, only the pelitic rocks from the Santa Rosa Range and from Birch Creek (Shieh and Taylor, 1969) were plotted in the figure because these are the only samples studied for which distinct metamorphic zones are recognizable.

It is rather striking from Fig. 24 that in regional metamorphic rocks, the O^{18}/O^{16} ratios of pelitic rocks tend to decrease with increasing metamorphic grade. In contact metamorphic rocks, the O^{18}/O^{16} ratios remain exceedingly constant throughout the entire contact aureole except for samples from the very narrow exchanged zones, the re-entrant, and the xenolith, where the O^{18}/O^{16} ratios are considerably lower. Except for the latter samples, it is also noted that the O^{18}/O^{16}

ratios of pelitic rocks in the contact metamorphic aureoles are within the range shown by shales and chlorite-grade regional metamorphic rocks. The O^{18}/O^{16} ratios of samples from the xenolith, the re-entrant, and the exchanged zones lie in about the same range as those in the higher-grade regional metamorphic rocks; all these samples must have exchanged with a large external reservoir of low-O^{18} oxygen during metamorphism. In the case of the contact metamorphic samples this reservoir is obviously the immediately adjacent igneous intrusions (and the aqueous fluids derived therefrom). For the regional metamorphic rocks, there are commonly no directly associated igneous rocks that one can appeal to as a plausible low-O^{18} reservoir. Therefore, we must postulate exchange with relatively large quantities of metamorphic pore fluids (H_2O-rich ?) that are ultimately derived from deep-seated plutonic igneous bodies of some type. This is somewhat of a paradox, in that the high-grade metamorphic rocks that apparently exhibit the most extensive oxygen isotope exchange with igneous rocks or fluids are those that are least directly associated with intrusive igneous bodies. The explanation of this phenomenon probably lies in (1) the enormously greater time interval involved in regional metamorphism, during which much larger quantities of H_2O could move through the rocks; (2) the intense structural deformation during regional metamorphism compared with the essentially static geometrical relationships between intrusive and country rock during contact metamorphism; and (3) the observation that most belts of regional metamorphism are in fact broadly associated in space and time with plutonic igneous activity.

The range of D/H ratios for samples from the contact metamorphic areas are compared with data from regional metamorphic and plutonic igneous rocks in Fig. 25. The minerals in contact metamorphic rocks and their associated plutons in general tend to have lower D/H ratios than regional metamorphic rocks. The D/H ratios in the country rocks surrounding the Sawtooth, Flynn, Birch Creek, and Eldora plutons are almost indistinguishable from the D/H ratios in the intrusives. This implies that there was some type of communication of H_2O (or H_2, which is even more mobile) between the aureole and the intrusive.

In the contact metamorphic rocks and associated plutons, there is a rough correlation between D/H ratios and geographic locations of samples, as the Rocky Mountain samples (e.g., the Eldora stock in Colorado) tend to be low in deuterium. FRIEDMAN *et al.* (1964b) showed that the D/H ratios of continental surface waters tend to decrease with an increase in the latitude and altitude of the ground surface; the Rocky Mountain surface waters are most depleted in deuterium, the Basin and Range waters next, and the California waters least depleted, of the three general areas studied in the present work. Geographically these areas correspond, respectively, to the Eldora, Santa Rosa, and Birch Creek localitites; it is therefore conceivable that the rough correlation between the D/H ratios in these contact metamorphic minerals and the nearby surface waters may be a result of exchange with local meteoric ground waters during the early Tertiary contact metamorphism.

We may note that if δD biotite is plotted against ΔD muscovite-biotite for the Sawtooth traverse (see data in Fig. 5) a very systematic correlation is observed. Inasmuch as δD muscovite is fairly uniform, the variation in ΔD muscovite-biotite (one of which, $\Delta = 53$, is the largest fractionation yet observed for this mineral pair) is almost wholly due to variation in δD biotite. It is possible that the *D/H* ratios of some of the biotites were lowered by exchange with light meteoric pore fluids during the retrograde temperature decline following contact metamorphism; biotite would become unstable before muscovite during such a temperature decline, and thus might be more susceptible to isotopic exchange.

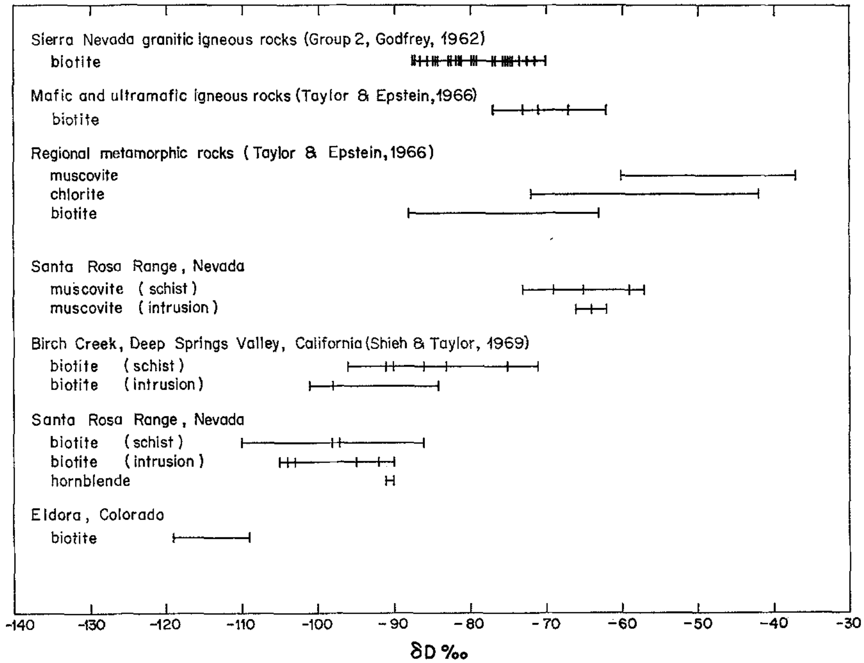

Fig. 25. Comparison of *D/H* ratios of minerals from contact metamorphic aureoles and their associated intrusions with *D/H* ratios of minerals from regional metamorphic and plutonic igneous rocks

The feasibility of the above explanations is enhanced by the observation of Taylor (1968b) that several shallow intrusions in the Scottish Hebrides have interacted with meteoric ground waters on a far larger scale than those studied here. Also, from isotopic analyses of hydrated volcanic glass, Taylor (1968b) and Friedman and Smith (1958) have shown that even as far back as the early Tertiary, meteoric water isotopic patterns in the western U.S.A. were similar to those prevailing today.

Summary and Conclusions

1. The consistent order of O^{18}/O^{16} and *D/H* enrichment among the coexisting minerals of contact metamorphic rocks is similar to relationships previously observed in regional metamorphic and igneous rocks; this strongly suggests that mineral assemblages in general tend to approach isotopic equilibrium during contact metamorphism. Samples at the intrusive-country rock contacts are clearly

demonstrated to contain mineral assemblages formed in oxygen isotope equilibrium, because identical O^{18}/O^{16} ratios of a particular mineral were attained by approaching isotopic equilibrium from opposite directions.

2. A correlation between oxygen isotopic fractionations and sample distance from contact is observed in certain instances; the best traverses in this respect are: Birch Creek, where a regular decrease of quartz-biotite fractionations (from 6.8 to 5.9) is observed in 7 samples collected from 1300 to 0.25 feet of the contact; and Sawtooth traverse I, where a regular decrease in quartz-muscovite fractionations (from 3.7 to 2.9) and quartz-biotite fractionations (from 7.2 to 6.2) is observed in 3 apparent equilibrium assemblages collected from 50 to 0.08 feet of the contact.

3. Small-scale oxygen isotope exchange effects between igneous intrusions and adjacent country rocks are observed within 0.5 to 3 feet of the intrusive contacts. The existence of a steep isotopic gradient (about 3 per mil per foot) in the exchanged zones, on both the intrusive side and the country rock side of the contact, suggests that the small-scale isotopic exchange occured essentially in the solid state by a diffusion-controlled recrystallization process. This presumably took place by grain-boundary diffusion in a static interstitial fluid film. At the contacts we typically observe a lowering of the country rock O^{18}/O^{16} ratios by about 3 to 4 per mil and an increase in the intrusive O^{18}/O^{16} ratios by about 1 to 2 per mil.

4. The size of the oxygen isotope equilibrium system in the small-scale exchanged zones is very small; it ranges from about 1.5 to 30 cm. The size of such an equilibrium system shows a good correlation with the presumed physical conditions at each contact during magmatic intrusion. The larger-size systems are associated with higher temperatures, larger intrusions, longer times of heating, and greater availability of aqueous fluids.

5. Based on oxygen isotope fractionation calibration curves, the contact temperatures are estimated to be 555° C at the Sawtooth stock, 525° C at the Santa Rosa stock, 540° C at the Birch Creek pluton, and 625° C at the Caribou stock. The isotopic contact temperatures agree reasonably well with temperatures estimated from heat flow calculations and from mineral parageneses, except that the Santa Rosa value is too low, suggesting either lack of equilibrium or that retrograde isotopic exchange has been important at this locality. Deuteric or retrograde exchange has also been important in the marginal zones of all the intrusives, where isotopic temperatures of 525 to 600° C are typically observed.

6. The apparently low isotopic country rock temperatures in the Santa Rosa and Sawtooth I traverses are compatible with a heat flow model (E) based on their occurrence at the tips of tongue-like wedges of the intrusives that project outward into the country rock. The isotopic temperatures obtained from samples in a xenolith and in a re-entrant of country rock that projects inward into an intrusion are higher than those obtained on samples collected in the vicinity of planar contacts; this also agrees with heat flow considerations.

7. Samples from tiny intrusive bodies and dikes and from the marginal portions of most of the larger plutons have unusually high O^{18}/O^{16} ratios relative to "normal" igneous rocks from the central portions of plutons. This is interpreted to be the result of large-scale oxygen isotopic exchange between essentially molten

igneous rock and metasedimentary country rock, either through a medium of aqueous fluids or by contamination with xenolithic blocks of country rock. Direct influx and absorption of water from the country rock may also be significant in certain intrusive stocks that are initially undersaturated with respect to H_2O. The D/H ratios of biotites from a given intrusion and its associated contact metamorphic aureole are essentially identical, implying communication of H_2O (or H_2) between the aureole and its intrusion.

8. The fact that only a very narrow zone of country rock has undergone isotopic exchange with the intrusive implies that horizontal outward movement of water into the contact metamorphic aureole is almost negligible during crystallization of the granitic magmas studied in this research. This indicates that most of the H_2O involved in contact metamorphism is derived from the country rocks themselves rather than from the intrusive.

9. The oxygen isotope exchange effects in a xenolith and in a re-entrant were found to be much more extensive than the exchange effects observed at main intrusive-country rock contacts, indicating that large-scale isotopic homogenization can occur in such instances, perhaps through upward transfer of H_2O-rich volatile fluids.

10. The O^{18}/O^{16} ratios of pelitic rocks throughout all the contact metamorphic aureoles, except for the re-entrant, xenolith, and small-scale exchanged zones, are identical to the O^{18}/O^{16} ratios of their unmetamorphosed parent rocks and similar to the range of O^{18}/O^{16} observed in shales and chlorite-grade regional metamorphic rocks. This is in contrast to the behavior of pelitic rocks in regional metamorphism, in which a decrease of O^{18}/O^{16} ratios commonly accompanies increasing metamorphic grade. Deep-seated regional metamorphism therefore usually involves exchange with an external oxygen reservoir (see TAYLOR and EPSTEIN, 1962; TAYLOR *et al.*, 1963; and EPSTEIN and TAYLOR, 1967), whereas contact metamorphism generally does not.

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