

DOE/ID/12079--61

DE82 019951

OXYGEN-ISOTOPE GEOCHEMISTRY OF QUATERNARY RHYOLITE  
FROM THE MINERAL MOUNTAINS, UTAH, U. S. A.

by

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March 1982

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## ACKNOWLEDGEMENTS

Financial assistance for this study was provided by the Department of Energy, Division of Geothermal Energy Contract DE-AC07-79ET27002 and continued under Contract DE-AC07-80ID12079. We thank Mr. Ray Lambert and Mr. John Covert for performing the mass spectrometry analysis of these samples.

## ABSTRACT

Oxygen isotope analyses were made of phenocryst and glass separates from four Quaternary rhyolite flows and domes in the Mineral Mountains, southwest Utah. With the exception of biotite in all samples and alkali feldspar in the rhyolite domes, all minerals appear to be in close oxygen isotope exchange equilibrium. The geothermometry equations proposed by Bottinga and Javoy (1973) and Javoy (1977) for quartz, alkali feldspar and magnetite produce the best agreement with temperature results from two-feldspar and iron-titanium oxide geothermometry for these rhyolites. If the rhyolites were generated by partial melting in the crust, their whole-rock (glass)  $\delta^{18}\text{O}$  values (6.3 to 6.9 permil) are consistent with generation from I-type (Chappell and White, 1974, O'Neil and Chappell, 1977; O'Neil et al., 1977) sources.

## INTRODUCTION

In the Mineral Mountains of southwest Utah, just east of the Roosevelt Hot Springs thermal area, a series of Quaternary rhyolite flows and domes occur along the crest of the range. The presence of young silicic volcanics has been considered to be an indicator of the presence of shallow magma chambers and potential heat sources for geothermal resources (Lipman et al., 1978; Evans and Nash, 1975, 1978), including the Roosevelt Hot Springs system. Because of this potential link with the Roosevelt Hot Springs geothermal system, chemical and petrologic studies of these volcanics have been made over the past several years to define petrogenesis, including P-T conditions of crystallization and depth estimates for the magma chamber, and the nature of the magma source area. These studies (Evans and Nash, 1975, 1978; Nash and Evans, 1978) have applied several chemical geothermometers to phenocryst phases in the rhyolites, including the Fe-Ti oxide (Buddington and Lindsley, 1964; Carmichael, 1967; Carmichael et al., 1974) and two-feldspar (Stormer, 1975) geothermometers, to define temperature of eruption.

The objectives of this study are two-fold. The first objective is to apply oxygen-isotope geothermometers to phenocryst phases in the rhyolites to test the compatibility of isotope and chemical geothermometers, and to test whether oxygen-isotope exchange equilibrium has been achieved in relatively viscous rhyolitic volcanics. The second objective is to define the  $\delta^{18}O$  value of the rhyolite magmas in order to place constraints on the source area for these volcanics. This information will help evaluate whether or not the rhyolites were generated by partial melting of a source containing a significant component of pelitic metasedimentary material or one dominated by primitive or mantle-derived material. These two types of source regions were

proposed by Chappell and White (1974) to produce the two geochemically and petrologically distinct groups of Paleozoic granitoids in southeast Australia which they designated 'S-type' (sedimentary) and 'I-type' (igneous). O'Neil and Chappell (1977) subsequently demonstrated that the S-type granitoids had higher  $\delta^{18}\text{O}$  values than the I-type granitoids. In the past several years, several studies (Taylor and Turi, 1976; Taylor and Silver, 1978; Longstaffe, et al., 1980) have utilized oxygen-isotope data to infer the extent of involvement of pelitic metasedimentary material in the formation of intermediate to acidic igneous rocks. To these ends, oxygen-isotope analyses were made on phenocryst and glass separates of four samples of rhyolite flows and domes from the Mineral Mountains area.

#### GENERAL GEOLOGY

The geology of the Mineral Mountains and of the silicic volcanism has been discussed in Evans and Nash (1978) and will not be repeated here. The Quaternary silicic volcanism was divided into three stages and types by Evans and Nash (1978):

1. The earliest stage is recorded by obsidian flows occurring along Bailey Ridge and Wild Horse Canyon. Their macroscopic character and calculations of  $P_{\text{H}_2\text{O}} \geq 3000$  bars (Evans and Nash, 1978; Nash and Evans, 1978) suggest unusual fluidity for this bulk composition. Radiometric dating of the Bailey Ridge flow indicates an age of  $0.77 \pm 0.08$  m.y. (Lipman et al., 1978).
2. Pyroclastic rocks occupy an intermediate position in the stratigraphic sequence and are of ash-fall and ash-flow origin. These rocks are mainly exposed in Ranch Canyon. K-Ar dating on a single obsidian clast from an ash-flow unit in Ranch Canyon yielded

an age of  $0.68 \pm 0.04$  m.y., providing an older limit for the age of the pyroclastics (Lipman et al., 1978). Recent K-Ar dating of anorthoclase phenocrysts from these pyroclastic rocks yields ages of  $0.61 \pm 0.04$  and  $0.66 \pm 0.03$  m.y. (Evans and Brown, 1981).

3. The last stage of rhyolite volcanism is represented by a group of at least 11 domal rhyolites distributed sporadically over 10 km near the crest of the Mineral Mountains. Age determinations on obsidian from Bearskin Mountain and sanidine from Little Bearskin Mountain have yielded K-Ar ages of  $0.58 \pm 0.12$  m.y. and  $0.53 \pm 0.05$  m.y., respectively (Lipman et al., 1978).

Petrography, bulk chemistry and trace element chemistry data on these rhyolites can be found in Evans and Nash (1978) and Nash and Evans (1978).

For this study, mineral and glass separates were made from samples of the rhyolite flows from Bailey Ridge and Wild Horse Canyon and from two of the domal rhyolites--Little Bearskin and North Twin Flat Peak.

#### ANALYTICAL TECHNIQUES

Mineral separates were obtained using standard heavy liquid and magnetic separation techniques. Oxygen from silicate minerals was extracted by reacting 5 to 10 mg of sample with  $\text{BrF}_5$  at  $550^\circ\text{C}$  in nickel reaction vessels for 14 hours (Clayton and Mayeda, 1963). The evolved  $\text{O}_2$  gas was then converted to  $\text{CO}_2$  for mass spectrometric analysis by combustion with graphite (Taylor and Epstein, 1962).

Isotopic measurements for  $\text{CO}_2$  gas were made with a Micromass 602 D mass spectrometer, a double collector,  $90^\circ$  sector magnetic deflection instrument of 6 cm radius. The isotopic data for oxygen are reported relative to SMOW

(Craig, 1961). Analytical error for oxygen isotope ratios is  $\pm 0.2$  permil for magnetite and  $\pm 0.1$  permil for all other phases.

### Notation

All isotopic data are reported in the delta notation, where

$$\delta_A = \frac{R_A - R_{std}}{R_{std}} \times 1000.$$

$\delta_A$  represents the  $\delta^{18}O$  value of sample A, and R is the  $^{18}O/^{16}O$  ratio of the sample or standard. For coexisting phases A and B,

$$10^3 \ln \alpha_{A-B} \cong \delta_A - \delta_B = \Delta_{A-B}$$

where  $\alpha_{A-B}$  is the fractionation factor, defined as

$$\alpha_{A-B} = \frac{R_A}{R_B} = \frac{1000 + \delta_A}{1000 + \delta_B}$$

### ANALYTICAL RESULTS

The oxygen-isotope compositions measured for phenocryst and glass separates from the four rhyolite samples are compiled in Table 1. Also included for thermometry purposes is the isotopic fractionation factor,  $\Delta$ , or difference in  $\delta^{18}O$  value, for the various pairs of phases. The minerals exhibit a consistent order of  $^{18}O$  enrichment, with quartz (Qz) > alkali feldspar (Pg) > biotite (Bi)  $\geq$  magnetite (Mt), indicating an approach to oxygen-isotope exchange equilibrium in all rhyolite samples. The isotopic fractionation between minerals is illustrated in the concordance plots of Figures 1 and 2. Equilibrium isotope fractionation factors inferred from

natural data (Bottinga and Javoy, 1973, 1975; Javoy, 1977) are also plotted in these figures for comparison.

Concordance diagrams have been used in previous studies (Anderson et al., 1971; Bottinga and Javoy, 1973, 1975; Clayton and Epstein, 1961; Javoy et al., 1970) to illustrate systematic correlations of oxygen-isotope data and to evaluate graphically the degree of attainment of isotope exchange equilibrium between mineral phases. In such plots, all mineral pairs that have differences in  $\delta^{18}\text{O}$  values equal to those defined by the fractionation equations are interpreted as having achieved isotope exchange equilibrium. However, there is disagreement between the experimentally and empirically determined fractionation factors for many mineral pairs, including quartz-plagioclase and quartz-magnetite pairs, which exceed analytical error over much of the temperature range 300-1000°C. In these cases, a conclusion regarding attainment of isotope exchange equilibrium depends on which fractionation equations are chosen. Many of the oxygen-isotope fractionations based on experiment, including quartz and plagioclase, yield temperatures different from those based on empirical calibrations. The empirical fractionations appear to have achieved better success at providing geologically reasonable temperatures (Bottinga and Javoy, 1973, 1975; Javoy, 1977) and greater internal consistency among temperatures from individual mineral pairs, and are preferred here on this basis. However, it is not clear whether the observed correlations of oxygen-isotope data from rock samples result solely from frozen-in equilibrium thermal effects. Bottinga and Javoy (1975) noted that the data from rock samples contained scatter which they interpreted as departures from isotope exchange equilibrium. Dienes (1977) has cautioned that natural oxygen-isotope correlations for mineral pairs may be additionally influenced by random errors or systematic retrograde effects,

possibly related to cooling or incipient hydrothermal processes. There are thus uncertainties in both the experimental and empirically derived oxygen-isotope fractionations for many mineral pairs which warrant some caution in their application. Despite the uncertainties in the empirical calibrations, there is good correlation of  $\delta^{18}\text{O}$  values for natural quartz-plagioclase-magnetite triplets (Bottinga and Javoy, 1973, 1975; Deines, 1977; Javoy, 1977). Further, the geothermometry equations based on these correlations have often produced geologically reasonable temperatures (Bottinga and Javoy, 1973, 1975), which is justification for their continued use.

The  $\delta^{18}\text{O}$  values of quartz, plagioclase and magnetite from samples 22 and 25 plot within analytical error of the empirical concordance lines in Figures 1A and 1B, suggesting that these minerals are in oxygen-isotope exchange equilibrium. Sample 24 plots somewhat away from the concordance line in both Figures 1A and 1B. Sample 24 plots below the Qz-Mt/Qz-Pg concordance line and above the Qz-Mt/Pg-Mt line, indicating that the  $\Delta(\text{Qz-Pg})$  fractionation factor is too large and the  $\Delta(\text{Pg-Mt})$  factor is too small relative to the  $\Delta(\text{Qz-Mt})$  factor. These discrepancies indicate that the  $\delta^{18}\text{O}$  value of plagioclase is too low for the phase to be in isotope exchange equilibrium with quartz and magnetite. Plagioclase in sample 24 has suffered at least minor sub-solidus oxygen-isotope exchange, a feature often observed in igneous rocks (Bottinga and Javoy, 1973, 1975; O'Neil and Taylor, 1967; Taylor, 1968, 1971, 1974).

Both samples 22 and 24 plot significantly off the concordance line in Figure 2. Because quartz and plagioclase in sample 22 are in approximate oxygen-isotope exchange equilibrium, then either the biotite in sample 22 has not equilibrated isotopically with the quartz and plagioclase or the fractionation factor used for the concordance line in Figure 2 does not

represent equilibrium fractionation for biotite. None of the other proposed biotite factors (Table 2) produce concordance either, so the biotite may simply not be in isotope exchange equilibrium. Assuming that the  $\Delta(Qz-Bi)$  factor is correct, biotite would also be out of exchange equilibrium with quartz and plagioclase in sample 24, because the point plots above the concordance line, and it has just been demonstrated (Figures 1A and 1B) that  $\Delta(Qz-Pg)$  is too large for sample 24. These figures demonstrate that biotite in both samples 22 and 24 and plagioclase in sample 24 are depleted in  $^{18}O$  relative to  $\delta^{18}O$  values in coexisting quartz and magnetite. Apparently both biotite samples have either not equilibrated with the magma, or not maintained equilibrium with the magma, or have undergone post-solidus interaction with an external,  $^{18}O$ -depleted oxygen reservoir, presumably meteoric water. Hydrogen-isotope analyses would be useful in detecting limited meteoric water-rock interaction, but sufficient quantities of biotite for hydrogen-isotope analysis could not be separated.

#### ISOTOPE THERMOMETRY

The attainment of internal oxygen-isotope exchange equilibrium, as indicated by the concordance diagrams, does not necessarily mean that the phenocryst phases have preserved their original crystallization temperatures. The concordance diagram allows an evaluation of exchange equilibrium and identification of exchange with external reservoirs. It does not allow evaluation of the primary nature of recorded temperatures, as shown by numerous examples of disagreement between isotope and chemical or phase equilibrium thermometers (Bottinga and Javoy, 1975; Javoy, 1977). Experience over the last ten to fifteen years with oxygen-isotope geothermometry has revealed that in slow-cooling systems (metamorphic and plutonic rocks),

coexisting silicate/oxide phases are vulnerable to retrograde isotope exchange. Such exchange can occur under near-equilibrium conditions, producing concordant  $\delta^{18}\text{O}$  values for mineral pairs but isotope temperatures that are lower than true formation temperatures. Fortunately,  $\delta^{18}\text{O}$  values are commonly "frozen" in minerals formed in rapidly cooled basic volcanic rocks (Anderson et al., 1971). Little information is available for more acidic volcanic rocks, however.

A second problem with oxygen-isotope geothermometry is the lack of agreement as to which geothermometer equation to use. Table 2 is a compilation of several geothermometer equations proposed for the minerals found in the Mineral Mountains rhyolites or computed from available mineral-water fractionation expressions. The various equations are based on both experimental measurement [for example Qz-Pg (A)] and empirical calibration of measured  $\delta^{18}\text{O}$  values in igneous and metamorphic rocks [for example Qz-Mt (B)].

The results of all geothermometry equations are presented in Table 3. Unconstrained use of all geothermometry equations produces unacceptably large ranges in isotope temperatures. For example the three alternate Qz - Pg geothermometers give a temperature range of 344 to 765°C for precipitation of quartz and plagioclase in sample 22, with the lower ranges of temperature being far below the solidus for magma of rhyolite composition. However some of the isotope geothermometers produce temperatures in good agreement with those from independent chemical geothermometers. For sample 22, the geothermometry equations proposed by Bottinga and Javoy (1973, 1975) provide temperatures of 765°C, 724°C, and 725°C from quartz-plagioclase, quartz-magnetite, and plagioclase-magnetite pairs respectively. These temperatures agree within analytical uncertainty with the results of Fe-Ti oxide (740°C)

and two-feldspar (770°C) geothermometry (Evans and Nash, 1978) for this sample (Table 4). The consistency of these three isotope temperatures reflects the concordance of the mineral  $\delta^{18}O$  values with the fractionation factors shown in Figures 1A and 1B, which were defined by the same geothermometry equations of Bottinga and Javoy (1973, 1975). The agreement between the isotope and chemical geothermometry results and the fact that these temperatures are geologically quite reasonable suggests these closely represent formation (crystallization) temperatures for these phases in sample 22. These same geothermometer equations produce concordant temperatures for sample 25 as well. However, the temperatures provided by the Bottinga and Javoy 1973) equations for quartz-plagioclase, quartz-magnetite, and plagioclase-magnetite pairs from sample 24 differ by an amount exceeding analytical uncertainty.

The quartz-plagioclase geothermometer (B, Table 2) equation incorporating the experimental data of O'Neil and Taylor (1967) is similar to that proposed by Bottinga and Javoy (1973) and produces quite similar temperature results (Table 3). However, equation (C) for this mineral pair produces temperatures either lower (samples 22 and 25) or higher (sample 24) than the other two versions. The quartz-magnetite geothermometer based on the experimental data of Clayton et al. (1972) for quartz produces temperatures systematically higher than the other isotope geothermometers (with the exception of those incorporating biotite) and the Fe-Ti oxide and two-feldspar geothermometers. All proposed isotope geothermometers involving biotite give unacceptable or even irrational values for temperature, including those based wholly or in part on the calibrations of Bottinga and Javoy (1973) (B and E, Table 3). This disagreement indicates either that biotite is out of isotope exchange equilibrium with the other phenocrysts, or that none of the proposed geothermometer equations adequately accounts for the oxygen-isotope

characteristics of magmatic biotites. One factor that certainly needs to be evaluated in this regard is the effect of Fe for Mg substitution on the oxygen-isotope composition of biotite.

Temperature results from the geothermometers proposed by Bottinga and Javoy (1973) are presented in Table 4 for all four samples, along with results from application of chemical geothermometers (Evans and Nash, 1978) for comparison. There is good agreement between the chemical and Qz - Pg, Qz - Mt and Pg - Mt geothermometers for samples 22 and 25. The Qz - Mt pair from sample 23 provides a geologically reasonable temperature (762°C), but no other geothermometry data exist for comparison. The Qt - Mt pair from sample 24 provides a temperature (682) in satisfactory agreement with the Fe-Ti oxide and two-feldspar geothermometry, but both the Qz - Pg and Pg - Mt results are significantly different. In sample 24, plagioclase has apparently experienced some retrograde exchange and is no longer in isotopic exchange equilibrium with quartz and magnetite. For the other samples, the agreement of the chemical geothermometers with the results from the geothermometer equations proposed by Bottinga and Javoy (1973) and Javoy (1977) supports the contention of the latter author that these geothermometers are internally more consistent and produce temperatures that are often more reasonable geologically than those based on experimental data alone.

#### MAGMA SOURCE

The  $\delta^{18}\text{O}$  values of the glass groundmass closely approximate the  $\delta^{18}\text{O}$  values of the rhyolite magmas, as this phase constitutes greater than 95 volume % of these rocks.  $\delta^{18}\text{O}$  values of the glass range from 6.3 to 6.9 permil, which are in the low range of values for acidic volcanics. These values are consistent with derivation of the Mineral Mountains rhyolites from

an I-type source (Chappell and White, 1974; O'Neil et al., 1977). The oxygen-isotope data indicate that if the rhyolites were generated by partial melting in the crust, the source region cannot contain a significant component of pelitic metasedimentary material.

#### CONCLUSIONS

Application of oxygen-isotope thermometry to phenocryst phases of the Mineral Mountains rhyolites yields some temperature results that are geologically reasonable and consistent with results from application of chemical geothermometers. The geothermometry equations proposed by Bottinga and Javoy (1973, 1975) are most successful and yield temperatures of 716-765°C for the rhyolite flows and 626-686°C for the rhyolite domes. These results are in good agreement with temperatures calculated from Fe-Ti oxide and two-feldspar geothermometers. Some degree of sub-solidus retrograde re-equilibration may be indicated by the isotope thermometry results for the domal rhyolites. However, liquidus temperatures as low as 650°C may be possible if fluorine and water were sufficiently effective in lowering the rhyolite liquidus. The range of  $\delta^{18}\text{O}$  values for the rhyolite-- + 6.3 to + 6.9 permil--is typical of I-type magmas, indicating that no significant pelitic metasedimentary component is involved in the generation of the Mineral Mountains rhyolites.

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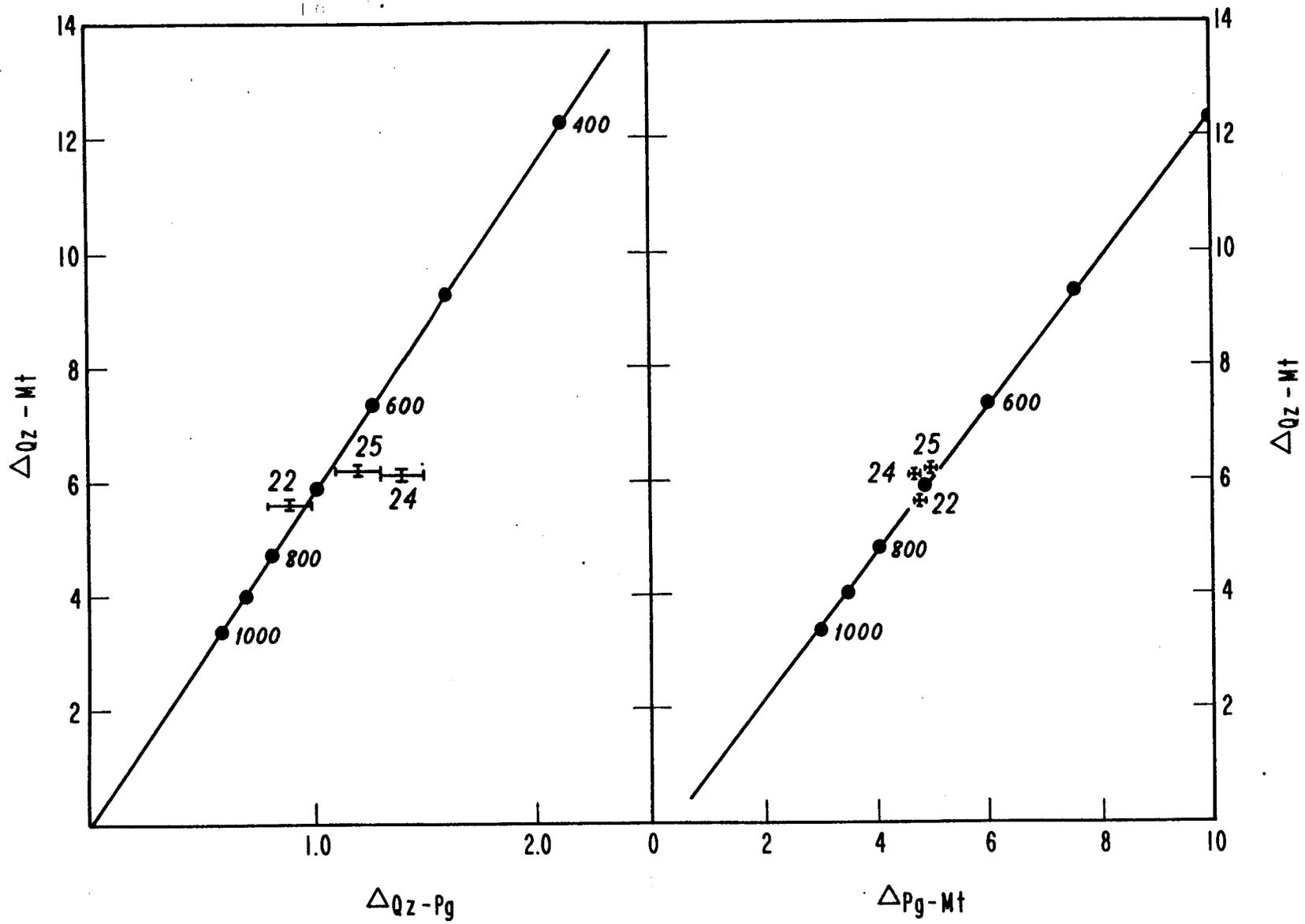
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## FIGURE CAPTIONS

- Figure 1. Plots of the Mineral Range oxygen-isotope data on concordance diagrams of  $\Delta(\text{Qz-mt})$  versus  $\Delta(\text{Qz-Pg})$  in Figure 1A and  $\Delta(\text{Qz-mt})$  versus  $\Delta(\text{Pg-mt})$  in Figure 1B. The concordance lines are based on the data of Becker (1971), Bottinga and Javoy (1973) and Javoy (1977). Error bars for each mineral pair reflect the combined analytical error for individual mineral analysis.
- Figure 2. Plot of Mineral Range oxygen-isotope data on a concordance diagram of  $\Delta(\text{Qz-Bi})$  versus  $\Delta(\text{Qz-Pg})$ . The concordance line is based on the data of Bottinga and Javoy (1973) and O'Neil and Taylor (1967). Error bars for each mineral pair reflect the combined analytical error for individual mineral analysis.



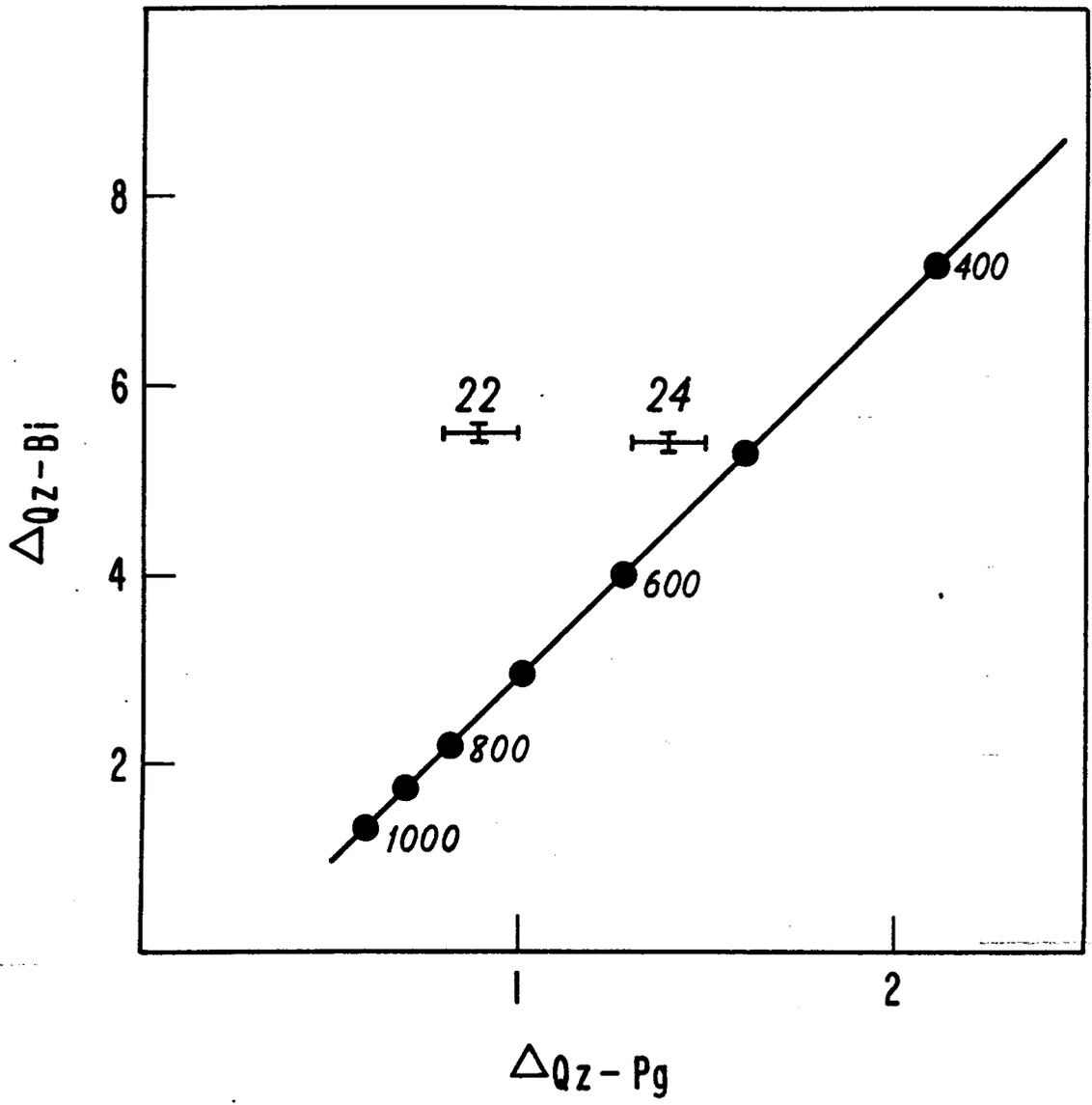


Table 1.—Oxygen-isotope compositions and fractionation factors for mineral pairs from rhyolite flows and domes, Mineral Mountains, Roosevelt Hot Springs Thermal Area

Rock Unit	Sample No.	$\delta^{18}\text{O}$ (SMOW)	$\Delta(\text{Qz-Pg})$	$\Delta(\text{Qz-Mt})$	$\Delta(\text{Qz-Bi})$	$\Delta(\text{Pg-Mt})$	$\Delta(\text{Pg-Bi})$	$\Delta(\text{Bi-Mt})$
Bailey Ridge	MR-76-22							
	quartz (Qz)	7.2	0.9	5.6	5.5	4.7	4.6	0.1
	glass	6.3						
	feldspar (Pg)	6.3						
	biotite (Bi)	1.7						
	magnetite (Mt)	1.6						
Wild Horse Canyon	MR-76-23							
	quartz	7.1		5.2				
	glass	6.3						
	magnetite	1.9						
Little Bearskin	MR-76-24							
	quartz	7.5	1.5	6.1	5.4	4.6	3.9	0.7
	glass	6.9						
	feldspar	6.0						
	biotite	2.1						
	magnetite	1.4						
North Twin Flat Peak	MR-76-25							
	quartz	7.4	1.2	6.2		5.0		
	glass	6.4						
	feldspar	6.2						
	magnetite	1.2						

Table 2. Compilation of oxygen-isotope fractionation factors used in geothermometry calculations.

Mineral Pair	$10^3 \ln \alpha$	Data Sources
Qz-Pg (A)	-0.40 ( $10^6 T^{-2}$ ) + 1.95 (>500°C) 0.47 ( $10^6 T^{-2}$ ) + 0.51 (<500°C)	Clayton et al. (1972); O'Neil and Taylor (1967)
Qz-Pg (B)	1.19 ( $10^6 T^{-2}$ ) - 0.29	Bottinga and Javoy (1973); O'Neil and Taylor (1967)
Qz-Pg (C)	0.97 ( $10^6 T^{-2}$ )	Bottinga and Javoy (1973)
Qz-Mt (A)	3.98 ( $10^6 T^{-2}$ ) + 2.24 (>500°C) 4.85 ( $10^6 T^{-2}$ ) + 0.80 (<500°C)	Becker (1971); Bottinga and Javoy (1973); Clayton et al. (1972)
Qz-Mt (B)	5.57 ( $10^6 T^{-2}$ )	Becker (1971); Javoy (1977)
Pg-Mt (A)	4.38 ( $10^6 T^{-2}$ ) + 0.30	Becker (1971); Bottinga and Javoy (1973); O'Neil and Taylor (1967)
Pg-Mt (B)	3.01 ( $10^6 T^{-2}$ ) + 2.24	Bottinga and Javoy (1973); Clayton et al. (1972)
Pg-Mt (C)	4.60 ( $10^6 T^{-2}$ )	Bottinga and Javoy (1973); Javoy (1977)
Qz-Bi (A)	2.15 ( $10^6 T^{-2}$ ) + 1.27 (>500°C) 3.02 ( $10^6 T^{-2}$ ) - 0.17 (<500°C)	Bottinga and Javoy (1975); Clayton et al. (1972); O'Neil and Taylor (1967)
Qz-Bi (B)	3.74 ( $10^6 T^{-2}$ ) - 0.97	Bottinga and Javoy (1973, 1975); O'Neil and Taylor (1967)
Qz-Bi (C)	3.33 ( $10^6 T^{-2}$ ) - 0.17 (>500°C) 4.20 ( $10^6 T^{-2}$ ) - 1.61 (<500°C)	Bertenrath and Friedrichsen (in prep); Clayton et al. (1972)
Qz-Bi (D)	4.92 ( $10^6 T^{-2}$ ) - 2.41	Bertenrath and Friedrichsen (in prep); Bottinga and Javoy (1973)
Qz-Bi (E)	3.69 ( $10^6 T^{-2}$ ) - 0.60	Bottinga and Javoy (1973)
Pg-Bi (A)	2.55 ( $10^6 T^{-2}$ ) - 0.68	Bottinga and Javoy (1975); O'Neil and Taylor (1967)
Pg-Bi (B)	3.73 ( $10^6 T^{-2}$ ) - 2.12	Bertenrath and Friedrichsen (in prep); O'Neil and Taylor (1967)
Pg-Bi (C)	2.72 ( $10^6 T^{-2}$ ) - 0.60	Bottinga and Javoy (1973)
Bi-Mt (A)	1.83 ( $10^6 T^{-2}$ ) + 0.97	Bottinga and Javoy (1973, 1975); O'Neil and Taylor (1967)
Bi-Mt (B)	0.65 ( $10^6 T^{-2}$ ) + 2.41	Bertenrath and Friedrichsen (in prep); Bottinga and Javoy (1973)
Bi-Mt (C)	1.88 ( $10^6 T^{-2}$ ) + 0.60	Bottinga and Javoy (1973)



Table 4. Comparison of oxygen-isotope (Bottinga and Javoy, 1973; Javoy, 1977) and chemical geothermometry ( $^{\circ}\text{C}$ ) results for the Mineral Mountains rhyolites.

Flow Unit	Qz-Pg (C)	Qz-Mt (B)	Pg-Mt (C)	Fe-Ti Oxide	Two-Feldspar
Bailey Ridge MR-76-22	765 $\pm 115$	724 $\pm 27$	716 $\pm 32$	740	770
Wild Horse Canyon MR-76-23		762 $\pm 29$			
Little Bearskin MR-76-24	531 $\pm 60$	682 $\pm 24$	727 $\pm 33$	640	670
North Twin Flat Peak MR-76-25	626 $\pm 75$	675 $\pm 24$	686 $\pm 29$	650	710