OXYGEN AND HYDROGEN ISOTOPE CHARACTERISTICS OF HYDROTHERMAL ALTERATION AT THE ANN-MASON PORPHYRY COPPER DEPOSIT, YERINGON, NEVADA #

By

John H. Dilles*
Department of Geology
Stanford University
Stanford, CA 94305

G. Cleve Solomon**
Division of Geological and Planetary Sciences
California Institute of Technology
Pasadena, CA 91125

Hugh P. Taylor, Jr.
Division of Geological and Planetary Sciences
California Institute of Technology
Pasadena, CA 91125

Marco T. Einaudi
Department of Applied Earth Sciences
Stanford University
Stanford, CA 94305

* Current address: Department of Geosciences, Oregon State University, Corvallis, OR 97331-5506
**Current address: Department of Geology, California Lutheran University, Thousand Oaks, CA 91360
# Contribution No. 4858, Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena, CA 91125
ABSTRACT

Late Cenozoic normal faulting has tilted the Middle Jurassic Ann-Mason porphyry copper deposit ≈90°W and exposed it in cross-section from the near-surface to deep environments (1 to 6 km paleodepth). The orebody and the top, bottom, and sides of the hydrothermal system are exposed in a single fault block that measures 3 by 5 km. Oxygen and hydrogen isotopes have been used to deduce sources and flow paths of hydrothermal fluids. Fresh samples of the host Yerington batholith pluton have typical calc-alkaline magmatic isotopic compositions ($\delta^{18}O_{\text{whole-rock}} = +6.7$ to $+7.0$; $\delta D_{\text{biotite}} = -85$ to $-88$). The various types of hydrothermally altered rocks have similar $\delta D$ and $\delta^{18}O$, only slightly isotopically shifted from original magmatic values, except along an axial zone within and above the orebody. In this zone, altered rocks are enriched in $^{18}O$ by $+1$ to $+3.5$ per mil, and $\delta^{18}O$ increases with decreasing paleodepth. The $\delta^{18}O$ of the whole-rock and/or feldspar and the $\delta D$ of biotite, chlorite, and/or actinolite in the various alteration zones are as follows: propylitic alteration is widespread and is isotopically similar to fresh rock: $\delta^{18}O = +6.2$ to $+7.2$; $\delta D = -82$ to $-92$. Potassic alteration, characterized by 100 percent biotitization of hornblende and Cu mineralization, occurs at ≈3.5 km paleodepth along the axis of the mineralizing granite porphyry dike swarm: $\delta^{18}O = +6.5$ to $+8.4$; $\delta D = -68$ to $-96$. Sodic-calcic alteration, characterized by hydrothermal oligoclase-actinolite-sphene, occurs laterally, alongside and below the ore zone at 3-6 km paleodepth: $\delta^{18}O = +5.7$ to $+8.4$; $\delta D = -67$ to $-90$. Late-stage sodic (albite-chlorite) alteration occurs at 1 to 4 km paleodepth in the vicinity of the axial zone: $\delta^{18}O = +6.7$ to $+9.6$; $\delta D = -79$ to $-85$. The youngest, late-stage sericitic alteration occurs at the shallowest levels, primarily at <1 to 2 km paleodepth, and displays a distinct $^{18}O$- and D-enrichment: $\delta^{18}O = +9.8$ to $+10.4$; $\delta D = -61$. Integration of stable isotopic data, geologic field relations, and phase
equilibria constraints indicates that the high-salinity fluids responsible for potassic alteration and Cu mineralization were of internal (magmatic) origin, but that the synchronous, deep, convecting, saline, sodic-calcic fluids were largely derived from external (non-magmatic) sources. A third type of fluid, also external but lower in temperature and much more dilute was prevalent at shallow levels, and was responsible for late-stage sericitic alteration. This was either an isotopically heavy, coastal meteoric water or possibly a sea water.

INTRODUCTION

Although porphyry copper deposits are well-studied, the sources of metals and mineralizing fluids remain debatable. Oxygen and hydrogen isotopes have proven extremely useful in deducing paths and sources of hydrothermal fluids (e.g., Taylor, 1974; 1979). Previous isotopic studies of porphyry copper deposits suggest that early, high-temperature potassic alteration is caused by magmatic fluids, whereas late, low temperature sericitic (phylllic) alteration is commonly caused by meteoric fluids (Sheppard et al., 1969, 1971; Sheppard and Taylor, 1974; Sheppard and Gustafson, 1976). According to most interpretations, copper is introduced during potassic alteration by high-temperature, high-salinity magmatic fluids (e.g., Gustafson and Hunt, 1975; Burnham, 1979), but an alternative model is that non-magmatic fluids leach copper from adjacent country rock, convect into the porphyry pluton, and deposit it there (e.g., Norton, 1982). Both processes may have occurred at Yerington (Dilles and Einaudi, 1988).

The Yerington district is uniquely suited for addressing the source of fluids and metals because the deep and lateral portions of the hydrothermal systems are well-exposed due to \(\approx 90^\circ\)W tectonic tilting that exposes the geology in cross-section. The district contains about \(6 \times 10^6\) tons of copper in three
porphyry deposits and several skarn deposits of Jurassic age. Excellent surface, mine, and drill core exposures allowed Anaconda Company geologists to map in detail rock types, structures, features of altered and mineralized rocks, and relative age-relations of the deposits (Proffett, 1970, 1977, 1979; Einaudi, 1970a, 1977; Gustafson, 1969). Since these early studies, continued research based at Stanford University has added new geologic data and focussed on petrological, geochemical, and absolute radiometric age-dating studies (Harris, 1979; Harris and Einaudi, 1982; Einaudi, 1982; Carten, 1981, 1986; Dilles, 1984, 1987; Proffett and Dilles, 1984; Dilles and Einaudi, 1988; Dilles and Wright, 1988).

The Yerington district represents an unsurpassed natural laboratory in which to characterize the isotopic signatures of porphyry-related alteration types, map their isotopic patterns in three dimensions, and deduce the sources of water in hydrothermal fluids within a detailed space-time context. Of particular interest were sodic-calcic alteration at deep structural levels and sodic alteration at shallow levels (Proffett, 1970; Carten, 1981, 1986; Dilles, 1984); these alteration types have not been described previously at most other porphyry deposits nor have they been isotopically characterized. The integrated isotopic and geologic data presented below, therefore, add important new constraints to genetic theories of porphyry copper deposits.

Below, we present and interpret oxygen and hydrogen isotopic analyses of whole rocks and minerals from the Ann-Mason porphyry deposit, Yerington district. Ann-Mason was chosen because the present surface exposes a 5 km (vertical) by 3 km (horizontal) section through the center and side of a large (>25 km³) hydrothermal alteration system (Dilles, 1984, 1987). The results are also generally applicable to the analogous Yerington Mine porphyry deposit (Einaudi, 1970b; Proffett, 1970, 1979; Carten, 1986). A companion paper (Dilles
and Einaudi, in prep.) will present the details of wall-rock alteration, mineralization, and sources of copper.

DISTRICT GEOLOGY

The Yerington district lies at the western margin of the Basin and Range Province. Miocene andesites and older rocks have been tilted 60-90°W during Miocene to Recent normal faulting that accommodated >100% east-west extension (Fig. 1, Proffett, 1977). In the Ann-Mason area, structural attitudes (Proffett and Dilles, 1984) and paleomagnetic data (Geissman et al., 1982) indicate that Mesozoic rocks were tilted ≈90°W during late Cenozoic normal faulting with a small additional component of pre-Oligocene tilting. The lower Tertiary erosion surface resting on Mesozoic rocks at Ann-Mason is thought to be at about 1 km depth relative to the Jurassic paleosurface, and therefore the Ann-Mason deposit contains exposures ranging from 1 to 6 km paleodepth (Fig. 2, Dilles, 1984, 1987). Thus, the Jurassic Yerington batholith and associated ore deposits are presently exposed in cross-section with tops to west (Figs. 1 & 2).

The Mesozoic rocks that host ore in the district range from Middle or Late Triassic to Middle Jurassic (Proffett and Dilles, 1984). The oldest rocks are andesite and rhyolite of the McConnell Canyon volcanics. They are overlain by carbonate, volcaniclastic, tuffaceous, and argillaceous sedimentary rocks capped by gypsum, sandstone, and the Artesia Lake volcanics. The Triassic-Jurassic volcanic and sedimentary rocks crop out south of Ann-Mason near Ludwig in a folded and metamorphosed pendant between two Jurassic plutons (Fig. 1).

Copper ores in the Yerington district are associated with the Middle Jurassic Yerington batholith, which intrudes cogenetic (?) andesite-dacite volcanic rocks of Artesia Lake (Proffett, 1979; Einaudi, 1982; Proffett and Dilles, 1984; Dilles, 1987). This composite pluton ranges from quartz
monzodiorite to granite (IUGS classification system of Streckeisen, 1976). Hornblende quartz monzodiorite of McLeod Hill (QMD) is the earliest intrusion; it is cut by the Bear pluton, which is compositionally zoned from hornblende quartz monzonite (QM) to a border granite (BG). These two early intrusions are grouped together as "main phases" on Figure 1. The final phase of batholith emplacement is associated with porphyry copper mineralization and alteration, and consists of the porphyritic granite (PG) of Luhr Hill and cogenetic granite porphyry (GP) dikes. At Ann-Mason the porphyritic granite forms a cupola whose apex is $\approx 3.6$ km beneath the lower Tertiary erosion surface (Fig. 1, 2); a granite porphyry dike swarm emanates from the cupola and extends to the Tertiary surface (Fig. 1; and see Fig. 2 of Dilles, 1987). Hydrothermal alteration and mineralization at Ann-Mason are centered on the dike swarm and are broadly synchronous with its emplacement. U-Pb zircon dates of the early McLeod Hill quartz monzodiorite and a granite porphyry dike yield Middle Jurassic ages of 169.4 and 168.5 Ma, respectively (Dilles and Wright, 1988). The essential mineralogy of all intrusive phases of the batholith is plagioclase (oligoclase $\pm$ andesine), K-feldspar, quartz, hornblende, biotite, magnetite, sphene, apatite, ilmenite, and zircon; augite occurs in some quartz monzodiorite (Dilles, 1987).

Two other Middle Jurassic units post-date the Yerington batholith and associated hydrothermal alteration (Fig. 1): the quartz latite volcanics of Fulstone Spring (Jaf) and the hornblende-biotite granite of the Shamrock batholith (Jsg). The two units have U-Pb zircon dates of 165-167 Ma (Dilles and Wright, 1988).

SAMPLING AND ANALYTICAL PROCEDURES

Mapping of geology and features of altered and mineralized rocks in the Ann-Mason area at 1:4,800 scale was done concurrently with sampling for isotopic
analysis. Alteration assemblages were delineated by field mapping, X-ray
diffraction analyses, petrographic studies, and confirmed by selected microprobe
analyses (Dilles, 1984). Sample descriptions are given in Tables 1 and 2 and
Appendices A and C.

Forty-six samples of representative fresh and hydrothermally altered
Jurassic igneous rocks were analyzed for $^{18}\text{O}/^{16}\text{O}$ and/or D/H. Yerington batholith
samples are all from the Ann-Mason area; all but two are of the early McLeod Hill
quartz monzodiorite or the mineralizing Luhr Hill granite. Whole rock oxygen
isotope analyses were made on 21 variably altered samples of Yerington batholith,
andesite of Artesia Lake, and the Shamrock batholith (Table 1). From an
additional 25 samples in the Ann-Mason area, mineral separates of quartz,
feldspar, and/or whole rock were prepared for $^{18}\text{O}/^{16}\text{O}$ analyses and amphibole or
mica separates were made for analyses of water content and D/H (Table 2).
Mineral separates were made by standard techniques of heavy liquids,
hand-picking, and short cold HF bath (for quartz), and are of >95 percent purity
(100% for quartz), except as noted in Table 2.

Mineral and whole-rock samples were prepared for isotopic analysis by
standard techniques. $O_2$ was extracted from 20 mg heated silicates by reaction
with $F_2$ gas and then converted to $CO_2$ on a hot carbon rod (Taylor and Epstein,
1962); for six samples, a new technique was employed in which 1 mg silicate was
heated by laser and reacted with $BrF_5$ (Sharp, in press). Water was extracted
from 50-125 mg hydrous minerals by melting in vacuo and was converted over heated
uranium to $H_2$ gas for isotopic analysis (Bigeleisen et al., 1952). Water
contents of hydrous minerals were calculated from $H_2$ gas yields measured by
manometer. Isotopic compositions of the prepared gas were measured on gas ratio
(Nier type) mass spectrometers and are reported using the standard $\delta$ notation
relative to SMOW (standard mean ocean water) in units of per mil (parts per
thousand or %). Oxygen analyses were done in the laboratories at Caltech and University of Lausanne, and hydrogen analyses were done in the U. S. Geological Survey laboratories at Menlo Park. The precision of determinations by laboratory replicate analysis (one sigma) is 0.2 per mil for $\delta^{18}O$ and 2-5 per mil for $\delta D$. The raw $\delta^{18}O$ values were corrected to the SMOW scale using a value of +8.45 per mil for the Caltech Rose Quartz standard. The $\delta^{18}O$ value of NBS-28 (fused silica) is +9.60 per mil on this scale. The raw $\delta D$ values were corrected to the SMOW scale by comparison with reference water standards V-SMOW, SLAP, and GISP. The $\delta D$ value of NBS-30 is -64.5 per mil in the Menlo Park laboratory.

HYDROTHERMAL ALTERATION AT ANN-MASON:

The Ann-Mason deposit contains 495 M tons of 0.4% Cu (Einaudi, 1982) within a large and complex volume of altered rock (Fig. 2). The center of the ore deposit is located 2.5 km east of the lower Tertiary erosion surface at $\approx$3.5 km paleodepth, but exposures of altered rocks extend from 1 to 6 km paleodepth and up to 3 km laterally southward forming a structurally intact 3 by 5 km block. Alteration and mineralization patterns on the south flank of the deposit were defined by mapping the surface exposures (Dilles, 1984). The ore zone, which is covered by up to 200 m of Tertiary volcanic rocks, is known only from drill core in the area north of the Singatse fault (Gustafson, 1969). A summary of alteration and mineralization features within this block is presented in Figure 2; in the text that follows, the spatial distribution of hydrothermal features are discussed relative to their original, pre-tilt (cross-sectional) position.

Three alteration groups occur at Ann-Mason: (1) endoskarn, (2) main-stage potassic and sodic-calcic, (3) late-stage sodic and sericitic. The characteristic assemblages are listed in Appendix A. Endoskarn alteration shows no clear-cut relation to porphyry copper alteration. It appears to pre-date,
or to be an early high-temperature and lateral form of, sodic-calcic alteration. Endoskarn alteration is developed at deep levels in quartz monzodiorite and locally in granite porphyry adjacent to skarnoid- and skarn-altered Triassic-Jurassic calcareous sedimentary rocks characterized by andradite garnet and diopsidic salite (Einaudi, 1977, 1982; Harris and Einaudi, 1982). Endoskarn alteration is characterized by salitic pyroxene and plagioclase (An\textsubscript{21-36} ± grandite garnet; based on phase equilibria, the assemblage grandite-labradorite-quartz formed at approximately 550-620°C (Dilles, 1984). This alteration required addition of calcium, presumably from adjacent metasedimentary rocks.

Main-stage alteration consists of two contrasting alteration types: potassic and sodic-calcic (assemblages in Appendix A). The two alteration types developed broadly synchronously (Einaudi, 1970a, 1970b) and are both associated with high-salinity fluid inclusions (Dilles, 1984). Both assemblages formed at relatively high temperatures; based on phase equilibria, sodic-calcic assemblages formed at 360-480°C (Carten, 1981, 1986; Dilles, 1984). Potassic alteration is restricted to an axial zone along the granite porphyry dike swarm above the porphyritic granite cupola, where it entirely encloses the >2000 ppm Cu contour (Fig. 2). Sodic-calcic alteration is developed along side and below the ore zone, where it is restricted to >3.5 km paleodepth and extends 2 km south from the potassic alteration zone to the endoskarn and skarnoid alteration zones. It extends upward (westward) 1 km from the porphyritic granite contact both along granite porphyry dikes and in structurally controlled zones (Fig. 2).

Potassic alteration is characterized by introduction of quartz veins, Cu-Fe sulfides, hydrothermal biotite, and local hydrothermal K-feldspar, and by destruction of hornblende ± oligoclase ± sphene (Appendix A). Conversely, sodic-calcic alteration is characterized by bleaching, addition of hydrothermal
oligoclase, actinolite, and sphene and destruction of biotite, magnetite, and K-feldspar. The spatial and temporal distribution of alteration assemblages and granite porphyry dikes, in conjunction with interpretation of phase equilibria suggests that potassic alteration was caused by fluids migrating upward and cooling along the granite porphyry dike swarm, whereas sodic-calcic alteration was caused by fluids migrating inward and heating toward the porphyritic granite contact, as in the case of hydrothermal convection (Carten, 1981, 1986; Dilles, 1984).

Propylitic assemblage PA (Appendix A) may have been broadly synchronous with the main-stage sodic-calcic and potassic alteration types. The propylitic PA alteration type is broadly distributed peripheral to the potassic alteration zone and orebody in the upper two-thirds of the hydrothermal system (Fig. 2). Though widespread, this alteration type is much weaker in intensity than the main stage alteration referred to above.

Late-stage alteration consists principally of sodic (albitic) and sericitic assemblages (Appendix A), which overprint and destroy main-stage potassic assemblages and form a funnel-shaped zone that expands upward from the center of the orebody (Fig. 2). Late-stage assemblages are characterized by high total sulfide (3-5%), a high ratio of pyrite to chalcopyrite, and fluid inclusions with lower salinities (Dilles, 1984). Sodic (albitic) alteration is characterized by albite replacing primary feldspars, and by chlorite, vermiculite, sericite, and/or tourmaline replacing primary mafic minerals. Sodic assemblage A-1 may be transitional to higher temperature (ca. 400°C) sodic-calcic assemblages because it is characterized by albite-sodic oligoclase in feldspar sites and intergrown biotite, vermiculite, and chlorite in mafic sites (Table 2). Sodic assemblages are destroyed in zones of sericitic alteration, which is the classic porphyry assemblage of quartz, sericite, and pyrite. Tourmaline-quartz breccias
are associated with, but cross-cut, all other late-stage assemblages. Late-stage alteration is thought to be due to shallow convection of dilute, low-temperature (175-250°C) fluids into the cooling hydrothermal system (Tables 1 and 2; Dilles, 1984). Chloritic assemblages PC, CB, and C are characterized by chloritization of biotite and amphibole and may also be late-stage because they post-date the main-stage alteration.

OXYGEN ISOTOPE ANALYSES

All whole-rock and feldspar from the Yerington batholith show a relatively restricted range of δ¹⁸O from +5.7 to +10.4 (Fig. 2, Tables 1 and 2). Nine relatively fresh samples of the Yerington batholith have whole rock δ¹⁸O of +6.1 to +7.5 (Tables 1 and 2). Analyses mineral separates from the freshest rocks gave values of +6.9 for oligoclase from quartz monzodiorite and +7.0 for K-feldspar from granite porphyry, and +9.0 for quartz from each (Table 2). Using the five freshest whole rock samples (excluding Y-329, which may have assimilated adjacent wallrock) together with a whole rock value of +6.7 for Y-767 (calculated from mineral modes and δ¹⁸O's), we obtain a uniform average δ¹⁸O = +6.7 ±0.2 per mil for the original quartz monzodiorite pluton. The one analysis of fresh granite porphyry (Y-781) yields a calculated whole rock δ¹⁸O of +7.0, which is consistent with the common observation of a slight increase in δ¹⁸O during differentiation of magmas (e.g., Taylor and Sheppard, 1986; Grunder, 1987). The primary igneous δ¹⁸O values of the Artesia Lake volcanics and the Shamrock batholith are not known, but are inferred also to be about +7.0 ±1.0, assuming they are typical of intermediate to silicic island-arc volcanic rocks and calc-alkaline batholiths elsewhere (Taylor, 1968; 1974; 1986).

Feldspar and whole rock δ¹⁸O/δ¹⁰ data for altered rocks are considered together below. This is a valid approximation because feldspars are the dominant minerals (>60%) in the Yerington batholith and mafic minerals and quartz
isotopically offset one another, so that whole rock (WR) and feldspar (F) values are approximately equal ($\Delta$WR-F < 0.5 per mil) except in the most intensely hydrothermally altered rocks.

Moreover, fine-grained volcanic rocks and coarse-grained igneous feldspars are both very susceptible to oxygen isotopic exchange with water in alkali-chloride solutions, even down to relatively low temperatures (ca. 200°C) (O'Neil and Taylor, 1967; Taylor, 1974). Thus, the variable oxygen isotopic compositions of feldspars at Ann-Mason are attributable to exchange with hydrothermal solutions (Table 2). On the other hand, medium- to coarse-grained quartz exchanges very sluggishly with hydrothermal solutions at submagmatic temperatures (Taylor, 1974). The measured $\delta^{18}O$ of igneous quartz in altered rocks ranges from +8.1 to +9.6, with many analyses similar to primary magmatic value of +9.0, which suggests that quartz has only partially equilibrated with hydrothermal fluids.

Samples of altered rock have relatively systematic $\delta^{18}O$ values corresponding to alteration type and paleodepth: endoskarn and main stage alteration are within ±1.5 per mil of original, magmatic values and display normal quartz(Q)-feldspar fractionations ($\Delta$Q-F = +1.0 to +2.5); late-stage assemblages range widely from magmatic values to values enriched in $^{18}O$ by up to +3.4 per mil. Two feldspars from endoskarn ($\delta^{18}O = +7.5$) are slightly $^{18}O$-enriched relative to fresh quartz monzodiorite (Table 2).

Main stage alteration around the Ann-Mason porphyry copper deposit is characterized by slight positive and negative shifts in $\delta^{18}O$ relative to fresh rock. All sodic-calcic samples except one are either unshifted or slightly depleted in $^{18}O$, with values ranging from +5.7 to +7.0 per mil (Table 2). The single $^{18}O$-enriched sample (+8.4) is close to the ore zone, where potassic samples are also $^{18}O$-enriched (Fig. 2). Although the potassic alteration samples
tend to be somewhat more $^{18}$O-enriched than the sodic-calcic samples, they are also unshifted or only slightly enriched in $^{18}$O (+6.5 to +8.4). The highest $^{18}$O-enriched samples of potassic alteration are from the shallowest (westernmost) part of the Ann-Mason orebody (Figs. 2 and 3a). However, in this area late stage sodic and chloritic (PC) alteration, characterized by $^{18}$O-enrichment, is commonly superimposed on earlier potassic alteration as evidenced by chlorite or vermiculite replacement of hydrothermal biotite. Thus, the $^{18}$O-enriched values of +7.7 to +8.4 for potassic alteration might be due in part to late-stage alteration; however, note that none of the analyzed samples contained more than 5 vol percent of chlorite and epidote (Tables 1 and 2, Appendix C).

Unlike main-stage alteration, the lower temperature late-stage alteration at shallow structural levels is characterized by erratic but consistently enriched $\delta^{18}$O values, ranging from +6.5 to +10.4. This is clearly shown by the distribution of $\delta^{18}$O contours on Figure 2. Also, note that one sample of sodic (albitic) assemblage A-2 has a reversed quartz-feldspar isotopic fractionation ($\Delta Q-F = -1.0$). This indicates that low temperature alteration did not achieve isotopic equilibrium, as the quartz has $\delta^{18}$O of +8.6 (nearly unshifted from the magmatic value of +9.0), whereas hydrothermal albite ($\delta^{18}$O = +9.6) is strongly $^{18}$O-enriched relative to the fresh rock feldspar (=+7.0) it replaced. As shown in Table 2, sodic assemblages have plagioclase $\delta^{18}$O values of +6.8 to +9.6; sericitic assemblage SQ has whole-rock values of +9.8 to +10.4; and deep-level chloritic assemblage C (chlorite-quartz) has a whole-rock value of +7.4 (Tables 1 & 2). One late-stage alteration sample, tourmaline breccia sample Y-309, has $\delta^{18}$O = +6.5; as discussed later below, it is estimated to have formed from fluids similar to, or slightly lower in $\delta^{18}$O than sericitic fluids. Thus, all late-stage samples but one show clear-cut $^{18}$O-enrichment relative the fresh quartz monzodiorite $\delta^{18}$O value of +6.7.
Seven samples of "propylitic alteration" have $\delta^{18}O$ ranging from +6.2 to +7.1 (Tables 1 and 2). The samples are either from propylitic actinolite-bearing assemblage PA from Ann-Mason or from chlorite-bearing assemblage PC of quartz monzodiorite or Artesia Lake andesite that occur distal to Ann-Mason or other porphyry copper alteration (Figs. 1 & 2). The chlorite-bearing samples are petrographically similar to "chloritic PC" alteration in the Ann-Mason area, but are isotopically similar to actinolite-bearing assemblages. All seven samples are slightly shifted by ±0.5 per mil relative to a fresh quartz monzodiorite $\delta^{18}O$ value of +6.7.

Three chloritic alteration samples of assemblages PC and CB have $\delta^{18}O$-enriched whole-rock $\delta^{18}O$ of +7.6 to +8.2 and contain chloritized biotite and/or amphibole; the samples are within or immediately adjacent to the upper (western) part of the axis of the Ann-Mason alteration pattern where late-stage alteration dominates (Fig. 2). Thus, chloritic alteration shows a spatial distribution and $\delta^{18}O$-enrichment similar to those of late-stage alteration, suggesting that both were caused by the same or similar hydrothermal fluids.

The whole-rock and feldspar oxygen isotope values for the Ann-Mason area are plotted and contoured on Figure 2 to illustrate the pattern of large-scale isotopic shifts outlined above. Most values from Ann-Mason, including unaltered and strongly altered rocks, fall between +6.5 and +7.0, almost identical to the original magmatic values. The area of $\delta^{18}O$-enrichment in the southeast part (deep-lateral) of the map corresponds to the zone of endoskarn alteration and possible wallrock-contaminated magma near carbonate-bearing metasediments. The small areas of slight $\delta^{18}O$-depletion ($\delta^{18}O < +6.5$) correspond to the fringe of sodic-calcic alteration and distal propylitic alteration. The axial zone of $\delta^{18}O$-enrichment (> +7.0), extending from the ore zone upward (westward) to the lower Tertiary erosion surface and increasing in magnitude upward, corresponds
to the area of late-stage sodic and sericitic alteration, chloritic assemblages PC and CB, and the top of the potassic alteration. Note that deepest (easternmost) samples within the potassic zone show little $\delta^{18}O$-shift, whereas those near areas of late-stage alteration at shallow levels show $\delta^{18}O$-enrichment, which suggests that $\delta^{18}O$-enrichment was caused by late-stage alteration. Although $\delta^{18}O$-enrichment in altered rocks generally increases upward (westward) toward the Jurassic paleosurface (Fig. 3a), it varies with alteration type and lateral distance from the central axis of the alteration system. With decreasing depth there is an increase in both the magnitude and range of $\delta^{18}O$ in strongly altered rocks (Fig. 3a). Thus, at shallow (1-3 km) paleodepth, there is a 4 per mil range of $\delta^{18}O$, which increases from magmatic values in propylitic PA assemblage to a 1-2 per mil enrichment in chloritic PC and CB, potassic B, and sodic A-1 assemblages, and finally to strong $\delta^{18}O$-enrichment of 3-4 per mil in sodic A-2 and sericitic SQ assemblages.

There are insufficient data to hypothesize about district-scale oxygen isotope variations. Samples from the distal, propylitic (?), chloritic altered andesite of Artesia Lake (assemblage PC) show little $\delta^{18}O$-shift (Fig. 1), similar to Ann-Mason propylitic samples. One sodic-calcic altered sample of the Shamrock batholith has a $\delta^{18}O$ of +7.9. Because the primary $\delta^{18}O$ of the Shamrock batholith is unknown, the altered rock value is difficult to interpret. However, this sample is significantly $\delta^{18}O$-enriched relative to $\delta^{18}O$ values of sodic-calcic alteration from the Yerington batholith (Tables 1 & 2).

Fractionation of $\delta^{18}O$ between two or more minerals can be used to estimate temperatures of last equilibration. Although isotopic equilibrium for Ann-Mason quartz-feldspar pairs cannot be demonstrated, because this would require $\delta^{18}O$ of three or more coexisting minerals, isotopic equilibrium can be tested by comparison of calculated fractionation temperatures with those estimated by phase
equilibria and fluid inclusion studies. We limit these comparisons to high-temperature assemblages, because isotopic equilibrium is unlikely during low-temperatures hydrothermal alteration where quartz and feldspar have different susceptibilities to isotopic exchange. Furthermore, isotopic fractionation factors are not experimentally well-calibrated below 500°C (see O'Neil, 1986). Nevertheless, application of the equations of Bottinga and Javoy (1973, 1975), calibrated for 500-800°C but extrapolated by us to <400°C (Table 2), yields some calculated temperatures that are in rough agreement with petrologic and fluid inclusion data (Carten, 1981, 1986; Dilles, 1984, 1987): 615°C for endoskarn alteration; 490 to 440°C for mineral reequilibration during cooling of unaltered Yerington batholith; and 425°C for sodic-calcic alteration (Table 2). The calculated \(^{18} \text{O}/^{16} \text{O}\) temperatures that are not in agreement with independent estimates include <400°C for potassic alteration (probably too low) and 570°C for weak sodic-calcic alteration sample Y-645 (too high).

HYDROGEN ISOTOPE DATA

The twenty isotopic analyses of hydrous minerals from fresh and altered Yerington batholith have a range of \(\delta D\) from -61 to -109. Two biotites from freshest quartz monzodiorite and granite porphyry have \(\delta D\) values of -88 and -85, respectively, and these are considered to be primary magmatic values. Most altered rocks are slightly D-enriched relative to these magmatic values.

Two samples are anomalously light, and probably do not give us any information about porphyry-related fluids. One endoskarn sample (Y-40) of diopside partly replaced by actinolite plus epidote yielded \(\delta D = -109\) and 0.40 wt. per cent \(\text{H}_2\text{O}\) (Table 2). Because the utility of D/H measurements of epidote is controversial, and because these hydrous minerals are a mixture and post-date the formation of skarn diopside, they do not reflect skarn \(\delta D\). The second sample
is a "biotite" (Y-505B) from deep potassic alteration, which has a $\delta D = -109$ and an anomalously high water content (4.95 wt.%) as a result of partial alteration to vermiculite. About 69% of the $H_2O$ in this sample can be attributed to vermiculite (14.6 wt.% $H_2O$) and 31% to F-bearing, Mg-rich biotite (2 wt.% $H_2O$ from analysis Y-540B; Dilles, 1984). If we assume that the original igneous or potassic alteration biotite had a $\delta D = -85$, and assume a simple two-component mixed-mica model, the measured $\delta D = -109$ would imply a vermiculite $\delta D$ of about $-120$. This is a plausible $\delta D$ for vermiculite formed by late Cenozoic weathering, based on various estimates of modern and late Cenozoic meteoric waters in the Great Basin ($\delta D = -110$ to -70; Sheppard et al., 1969, 1971; Taylor, 1974). An alternative is that vermiculite replaced biotite during a deep, late-stage alteration similar to sodic assemblage A-2, and therefore that the $\delta D = -109$ represents moderate temperature late-stage fluid. Regardless, little is known at present about the isotopic behavior of vermiculite, so this D/H analysis is not very informative.

All fifteen other minerals analyzed for D/H from hydrothermally altered rocks are relatively uniform in $\delta D$ between -96 and -61. This range is within $\pm 25$ to $-10$ per mil of the magmatic $\delta D$ range of -88 to -85. Actinolites from propylitic samples have $\delta D$ of -92 to -82 similar to the magmatic values. Actinolites from sodic-calcic alteration and biotites from potassic alteration (excluding $\delta D = -109$ of Y505B) extend this range to somewhat higher $\delta D$ values (-95 to -67). Chlorite and biotite-vermiculite mixed-layer micas from late-stage, sodic alteration also have near magmatic values ($\delta D$ of -85 to -79). Sericites from sericitic SQ alteration are moderately D-enriched at -61 per mil (Table 2), and this is the only alteration type whose mean $\delta D$ is distinct from the original magmatic values.

There are only slight variations of $\delta D$ with paleodepth, and these are
mainly due to the strong D-enrichment in the shallow, late-stage sericitic alteration. If we exclude the propylitic alteration samples, there is a weak correlation of D-depletion with increased paleodepth (Fig. 3b). This trend is also seen within one individual alteration type as a slight decrease of biotite δD from potassic alteration with increasing paleodepth between 3.0 and 5.5 km. However, an opposite trend is observed in the slight increase of δD of actinolites in sodic-calcic alteration with increased paleodepth in the same interval.

A plot of δ¹⁸O of feldspar or whole rock versus δD of hydrous mineral (Fig. 3c) reveals that the only truly distinctive field of hydrothermal alteration is the ¹⁸O-rich and D-rich late-stage, sericitic alteration. Fresh Yerington batholith and propylitic alteration are indistinguishable. Both main stage alteration types (potassic and sodic-calcic) overlap with fresh rock δ¹⁸O and δD values, but are distinguishable from them by their greater isotopic variability. Both late-stage sericitic and sodic A-2 alteration assemblages are distinctly enriched in ¹⁸O and D relative to fresh rocks.

DISCUSSION: ISOTOPIC COMPOSITION OF HYDROTHERMAL FLUIDS

The most striking feature of the isotope data presented above for the Ann-Mason porphyry copper orebody is the lack of strong variation. This contrasts sharply with hydrothermal ore deposits of appropriate paleogeography (inland, high elevation, or high latitude) where isotopic data indicate much low-temperature hydrothermal alteration was caused by isotopically light meteoric waters such as Butte, Montana, and many other deposits in Nevada, Utah, and Idaho (Sheppard et al., 1971; Taylor, 1974; Sheppard and Taylor, 1974). The small isotopic variations at Ann-Mason appear to be the result of its coastal paleogeography, similar to the porphyry copper deposit at El Salvador, Chile
(Sheppard and Gustafson, 1976).

Because of the small isotopic variations at Yerington, the isotopic compositions and origins of hydrothermal fluids can be deduced only by applying geologic constraints from alteration assemblages, including formation temperature, relative age, and inferred fluid flow path. When isotopic data are combined with temperature estimates based on phase equilibria and fluid inclusion studies, they support the conclusion of early studies of wall-rock alteration patterns at Yerington that suggested at least two origins for the waters associated with potassic and sodic-calcic alteration (e.g., Carten, 1981, p. 152). The isotopic data are also compatible with geologic and phase equilibria evidence that suggest a third type of water was responsible for late-stage sericitic alteration.

We have utilized measured oxygen and hydrogen isotopic fractionation factors between minerals and water to calculate of the isotopic compositions of waters in equilibrium with these minerals at the specified temperature ranges estimated for each alteration type by Dilles (1984) (Table 2). Equilibrium oxygen isotope fractionation factors for many mineral-H$_2$O systems are relatively well known from experiment, but hydrogen isotope fractionation factors are poorly known in general. We have used oxygen fractionation factors for feldspar-water and muscovite-water from experimental data of O'Neil and Taylor (1967, 1969) and for quartz-water from the empirical curve of Taylor (1974), because these are the calibrations that have been used in most previous studies of porphyry copper deposits (e.g., Sheppard et al., 1969, 1971). Use of the feldspar- and quartz-water curves of Matsuhisa et al. (1979) yields fractionation factors that differ from O'Neil-Taylor curves by as much as 1 per mil (or greater at <400°C), but there are new data by Clayton et al. (1989) that the O'Neil-Taylor curves are more nearly correct than the Matsuhisa calibrations. Tourmaline-water oxygen
isotope fractionations are poorly known, but for granitic rocks they are intermediate between albite-water and muscovite-water (France-Lanord et al., 1988); we have assumed this behavior extends to 200°C.

We have used the hydrogen isotope fractionation factors for amphibole- and mica-water given by Suzuoki and Epstein (1976):

$$1000 \ln(\alpha) = -22.4 \times 10^8 T^{-2} + 28.2 + (2X_{Al} - 4X_{Mg} - 68X_{Fe}),$$

where $T$ is temperature (°K), $\alpha$ is the fractionation factor $(1000 + \deltaD_{\text{mineral}})/(1000 + \deltaD_{\text{water}})$; and $X_{Al}$, $X_{Mg}$, and $X_{Fe}$ are the mole fractions of Al, Mg, and Fe cations in octahedral coordination. This equation is calibrated only for the interval 450-850°C, but we have extrapolated it to 300°C. We believe the equation is the best available approximation. It has been used previously by the largest number of investigators (e.g., Taylor and Friedrichsen, 1983; Brigham and O'Neil, 1985), and therefore it allows comparison of our results with previous work; however, it may not be wholly valid because it ignores some octahedral cations such as Ti, and experimental work by Graham et al. (1984, 1986) indicate that amphibole- and chlorite-water fractionation may not be completely dependent on octahedral cation compositions. At 350-550°C, use of the Graham et al. (1984) curves for amphibole would decrease the D/H fractionation factors by as much as 18 per mil. At 200-250°C we have employed the empirical D-fractionation curves of Taylor (1974) for chlorite- and muscovite-water, but have also listed water values calculated from the chlorite- and muscovite-water curves of Graham et al. (1986) and Lambert and Epstein (1980), respectively (Table 2). The hydrogen isotope fractionation factor for epidote-water of Graham and Sheppard (1980) was also used. However, we do not place much credence in the utility of D/H studies of epidote, and these data are not used in the discussion below. Other fractionation factors employed are listed in Table 2. Because of the uncertainties in fractionation factors, temperatures of formation,
and isotopic analyses, compositions of calculated waters have errors of approximately ± 1 to 2 per mil for $\delta^18O$ and ± 15 to 30 per mil for $\delta D$ at 700°C and 225°C, respectively.

The isotopic compositions of the calculated waters vary moderately on a $\delta^18O$ versus $\delta D$ plot (Fig. 4). Freshest quartz monzodiorite and granite porphyry are in equilibrium at 700°C with waters of magmatic composition ($\delta^18O = +7.4; \delta D = -65$). At the low temperature extreme, two samples of sericitic alteration are in equilibrium at 225°C with an isotopically heavy water ($\delta^18O = 0; \delta D = -20$), a composition similar to that of low-latitude, coastal meteoric water or sea water. Most of the other alteration types yield calculated water compositions that lie between these two end members (Fig. 4).

Field relations and petrologic phase equilibria indicate that main stage potassic and sodic-calcic alteration occurred synchronously (Einaudi, 1970b) at relatively high temperature as a result of hydrothermal fluids with different flow paths and probably of different origins (Carten, 1981, 1986; Dilles, 1984). Using alteration temperatures estimated from mineral phase relations and fluid inclusions, the isotopic data allow calculation of isotopic compositions of waters (Table 2) that reinforce the above conclusion, as elaborated upon below.

Waters for potassic alteration, calculated at 500-600°C, have $\delta^18O$ similar to magmatic waters, but range from D-depleted to D-enriched compositions ($\delta D = -34$ to -79, Fig. 4). The Ann-Mason potassic alteration waters are similar to, but more variable in $\delta D$ and with lower average $\delta D$ than, the waters calculated by Sheppard et al. (1971) to be in equilibrium with potassic alteration biotites at 650°C. These authors concluded that such fluids were primarily magmatic in origin. Of the five samples from Ann-Mason, one sample may be anomalously D-depleted (Y-505B), possibly due to Cenozoic weathering or late-stage alteration, as discussed above. Weathering has not affected the other samples,
two of which are from drill core. Two potassic samples yield δD values of -62 and -71 similar to "primary magmatic waters" from Yerington; however, the two last potassic samples yield calculated water compositions that are markedly D-enriched relative to "primary magmatic waters".

At equilibrium at 500°C the two D-enriched biotites from potassic alteration would have coexisted with waters having δD = -34 and -42. Such fluids cannot have been in equilibrium with "normal" Yerington magmas. The relatively high δD values conceivably could be derived from a hypothetical D-enriched magmatic hydrothermal fluid produced by Rayleigh fractionation during degassing of the magma, as proposed by several studies of granites and rhyolites (Nabelek et al., 1983; Taylor et al., 1983; Brigham and O'Neil, 1985). A more likely scenario is that the samples were affected by late-stage alteration fluids enriched in D. The two D-enriched potassic samples occur near the top (western part, Fig. 2) of potassic alteration in an area of much late-stage sodic alteration, and they contain 10 and 35% chlorite interlayered with biotite (Fig. 3b; Table 2). Influx of small amounts of 18O- and D-enriched late-stage fluids could have retrograded minor amounts of biotite to water-rich and D-rich chlorite, but would have caused only a small 18O-shift, because of the much lower effective water/rock ratio for oxygen relative to hydrogen (Table 2, Fig. 4).

Sodic-calcic hydrothermal fluids in equilibrium with alteration feldspar and actinolite at 400-450°C yield calculated waters with δ18O = +3.1 to +6.3 and δD = -39 to -61 (Table 2, Fig. 4). Such fluids are distinctly 18O-depleted relative to magmatic and potassic alteration fluids. However, sodic-calcic alteration fluids are indistinguishable in δD (within analytical error) from magmatic and potassic fluids, although their average δD value is slightly higher than either. The 18O-depletion of 0 to 2 per mil of sodic-calcic fluids relative to magmatic fluids is probably real, because for three samples of S-3 assemblage,
\( \delta^18O \) of fluid is calculated from measured \( \delta^18O \) values of oligoclase finely intergrown with 1-20% quartz; this intergrowth has completely replaced igneous K-feldspar, and the alteration temperature is well-constrained (360-480°C; Carten, 1986; >400°C, Dilles, 1984). This replacement reaction is caused by exchange of Na\(^+\) in fluid with K\(^+\) in feldspar, and alkali exchange phase equilibria (Orville, 1963) require that if the fluid were originally in equilibrium with both Na- and K-feldspar, as must be the case at Ann-Mason, then the fluid must have followed a heating path to cause the reaction (Carten, 1986). Due to this heating constraint, sodic-calcic fluids must have flowed toward the porphyritic granite cupola and into and along granite porphyry dikes (Figs. 1 and 2), and thus they are likely of non-magmatic origin (Carten, 1981; Dilles, 1984). Detailed field mapping has established that sodic-calcic alteration is locally superimposed upward and inward on broadly contemporaneous potassic alteration, which also requires inward fluid flow along a heating path (Carten, 1981, 1986). Fluid inclusions in sodic-calcic alteration are halite-bearing (>34 wt.% equivalent NaCl; Dilles, 1984). Thus, sodic-calcic hydrothermal fluids were probably high-salinity, external, non-magmatic waters with \( \delta D \approx -50 \) and \( \delta^18O \approx +3 \) to +6 that convected into the hydrothermal system along a increasing thermal gradient. Because the \( \delta^18O \) of hydrothermal feldspar and calculated fluids in sodic-calcic alteration increases toward the central zone of potassic alteration (Fig. 2) along the presumed flow path, sodic-calcic fluids probably evolved to more \(^{18}O\)-enriched compositions during wallrock exchange (see Taylor, 1974). Therefore, the most likely parent would have originally been less \(^{18}O\)-enriched than these calculated waters. The most suitable parent is a saline formation water, possibly evolved from meteoric waters or a mixture of meteoric waters and sea waters, with \( \delta D \approx -50 \) and \( \delta^18O < +3 \).
Because the Jurassic-Triassic section intruded by the Yerington batholith contains an evaporite gypsum in its upper portion, formation waters could have been quite saline (Carpenter and Grethen, 1979). Overlying the gypsum are windblown sandstone of the Preacher's Formation (Noble, 1962) and subaerial andesitic to dacitic flows, breccias, conglomerate, and tuffs of the Artesia Lake (Proffett and Dilles, 1984, 1990). These units immediately pre-date emplacement of the Yerington batholith and indicate a transition from an early shallow marine setting to a coastal, subaerial volcanic arc regime. Thus, when the batholith was emplaced, surficial waters were likely meteoric. Moreover, the Jurassic paleo-latitude of Yerington was closer to the equator on the basis of the North American paleomagnetic pole position than at present (Van Alstine and deBoer, 1978; Geissman et al., 1982). Therefore, Jurassic meteoric waters and formation waters derived from them were likely only slightly lighter than SMOW, consistent with our calculations. Thus, such formation waters could have caused sodic-calcic alteration without any additional component of magmatic, meteoric, or sea water. A similar hypothesis has been proposed at the porphyry copper deposit at Bingham, Utah, for the origin of the lateral, 350°C propylitic alteration from formation waters with δ²H = +4.2 and δD = -40 (Bowman et al., 1987).

A second, less likely, possibility is that sodic-calcic fluids were mixtures of small amounts of meteoric waters similar to late-stage alteration fluids with a large amount of magmatic waters. Such a mixed fluid source is compatible with the isotopic data (Fig. 4), but is considered unlikely because of the inferred inward flow of the sodic-calcic fluids toward the source of magmatic fluids.

We are uncertain that minerals in propylitically altered samples of assemblage PA completely equilibrated with hydrothermal fluids, because of
evidence that alteration was relatively weak; the PA assemblage contains actinolite pseudomorphs of igneous hornblende and relict igneous plagioclase compositions (Appendix C). It also has δ¹⁸O and δD values indistinguishable from unaltered Yerington batholith. The temperature of formation of this widespread alteration type is also poorly constrained and probably varied considerably, but on the basis of the stability of actinolite-chlorite-albite-epidote, a range of 200-400°C can be estimated. Assuming isotopic equilibrium was achieved, waters calculated at 300°C have compositions lying between sodic-calcic and sericitic fluids (Fig. 5).

Late-stage and low-temperature sodic alteration was caused by fluids ranging in temperature from <250° to 400+°C and low to high(?) salinity (Dilles, 1984). Sodic assemblage A-1 contains high-salinity fluid inclusions and albite ± oligoclase feldspar, and yields calculated waters at 350-400°C similar to sodic-calcic fluids (Table 2, Fig. 4). Sodic assemblage A-2, which is associated with sericitic alteration at shallow depths and contains albite-chlorite-sericite, yields a calculated water at 250°C intermediate between the sodic-calcic and sericitic fluids (Fig. 4). Thus, sodic (albitic) hydrothermal fluids may track increased contribution of dilute, "meteoric" type waters into the hydrothermal system. Alternatively, they could record the decrease of temperature and increase of water/rock ratios of formation water convecting into the system. Because of the large error attached to the δD values of the calculated waters at low temperatures, the data cannot distinguish between these alternatives. Regardless, sodic hydrothermal fluids are distinct from magmatic waters, and require a non-magmatic component.

Based on fluid inclusion studies, Dilles (1984) concluded that late-stage sericitic alteration was caused by structurally controlled, low temperature (175-250°C), low salinity (1-12 wt.% and mostly 2-5 wt.% NaCl equiv.)
hydrothermal fluids. The low salinity, low formation temperature, strong structural control, and strong hydrogen ion metasomatism of sericitic alteration in contrast to the older, main-stage alteration types suggest a different origin for sericitic fluids. The isotopic data for sericitic alteration yield waters calculated for equilibrium at 225°C with values of δ¹⁸O ≈ 0 and δD = -20 (Table 2, Fig. 4; δD = -55, using fractionation factors of Lambert and Epstein, 1980). Similarly, the late-stage tourmaline-breccia, composed of ≈50% quartz, ≈34% tourmaline, ≈12% sericite, and <5% rutile and goethite, yields an estimated water composition of δ¹⁸O ≈ -3 at 225°C. The isotopic compositions of late-stage fluids can be best explained as "isotopically heavy" meteoric waters that are ¹⁸O-enriched by an isotopic shift of +2 to +4 per mil (+6 to +8 using δD = -55) from the meteoric water line. The moderate ¹⁸O-shift from meteoric water compositions and the ¹⁸O-depleted whole-rock values require that effective water/rock ratios were higher during sericitic alteration than in the slightly earlier sodic alteration. The large uncertainty (± 30 per mil) for δD of sericitic waters (-20 ± 30; or -55 ± 30 using Lambert and Epstein fractionation factors) precludes use of δD to distinguish absolutely between meteoric, sea water, and magmatic waters.

The Ann-Mason data do not show the divergence of meteoric waters from the earlier, higher temperature trends of magmatic and formation waters, which is seen in most inland and high latitude porphyry copper deposits (Sheppard et al., 1971; Taylor, 1974; Bowman et al., 1987). Yerington clearly lacked the isotopically light meteoric waters typical of inland areas and high latitudes that cause strongly ¹⁸O- and D-depleted values for late stage sericitic and argillic alteration at many porphyry copper deposits (e.g., Santa Rita; Sheppard et al., 1971). A very similar set of isotopic data from El Salvador, Chile, also shows moderate ¹⁸O-depletion and moderate D-enrichment in sericitic alteration.
relative to potassic alteration, and was interpreted to have been caused by $\text{^18}O$-shifted meteoric waters with $\delta D = -35$ (Sheppard and Gustafson, 1976).

Sea water as a source of sericitic fluids is not precluded by the isotopic data (see our preliminary conclusions in Solomon et al., 1983), because sea water would be difficult to distinguish from "isotopically heavy" meteoric water originating at a low latitude, coastal, island arc/back arc setting such as Yerington in the Middle Jurassic, as outlined above. Sea water salinity (3.5 wt.% salts) also falls in the range of salinities of sericitic fluid inclusions (1-12 wt.%, mostly 2-5 wt.% NaCl equiv.). However, two lines of evidence suggest meteoric water is more likely than sea water. First, the calculated $\delta D$ and $\delta^18O$ of sericitic fluids are more similar to $\text{^18}O$-shifted meteoric water than to sea water (Fig. 4); in particular, the calculated sericitic $\delta^18O = 0$ and estimated tourmaline $\delta^18O = -3$ of hydrothermal fluids indicate that if these fluids had undergone any $\text{^18}O$-enrichment or if equilibrium had occurred at $<225^\circ C$, they must have had original values of $\delta^18O < 0$, and therefore could not have been sea water. Secondly, geologic evidence outlined above indicates that Yerington was a subaerial, but coastal, environment when the Yerington batholith was emplaced, and therefore that surficial waters were meteoric waters. In comparison, similar alteration types of similar age in a more continental setting in the Rodman-Ord Mountains, Mojave desert, California, have much lower $\delta^18O$ values and therefore are more clearly of meteoric origin (Solomon and Taylor, in prep.).

Hydrothermal alteration in the Yerington porphyry copper deposits differs from many other porphyry deposits because of the large amounts of sodic-calcic and sodic alteration. The presence of sodic-calcic alteration could be due largely to the exposure of the deep parts of the hydrothermal alteration system at Yerington (Proffett, 1970; Einaudi, 1970b; Carten, 1986); deep exposures
are lacking at most other porphyry deposits, but have been reported with associated sodic-calcic alteration at several, including Ajo and Silver Bell, Arizona (Gilluly, 1946; Cox and Eijun, 1984; Proffett, 1970). The dominance of sodic (albitic) alteration at late stages in the central to upper part of the hydrothermal system at Ann-Mason contrasts with dominance of late sericitic alteration due to evolution of low pH fluids at most other porphyry deposits in the cratonal United States (Meyer and Hemley, 1967). The uppermost part of the Yerington hydrothermal system appears to be represented by sericitic and advanced argillic assemblages exposed in the Buckskin Range, which were not sampled in this study (Hudson, 1983; Proffett and Dilles, 1990; Fig. 1); these exposures may be similar to the low pH alteration at the tops of other porphyry deposits (e.g., El Salvador, Chile; Gustafson and Hunt, 1975). Previous stable isotope data indicate that such low pH fluids at other deposits are largely meteoric (Sheppard et al., 1971). In many porphyry copper deposits in island-arc tectonic settings (Southwest Pacific; British Columbia), albite- and chlorite-bearing alteration assemblages are widespread (Carson and Jambor, 1976; Chivas, 1978; Gustafson, 1978) and could have been caused by seawater. The paleogeographic setting of Yerington in the Middle Jurassic indicates a shallow, marine setting (closed basin?) was succeeded by a near-coastal, subaerial, volcanic-arc environment (Geissman et al., 1982; Proffett and Dilles, 1984). Thus, formation waters that evolved in the Jurassic-Triassic volcanic and sedimentary rocks could be mixtures of trapped meteoric and marine waters. The hydrogen isotopic data at Ann-Mason suggest that a meteoric component was necessary for sodic-calcic and sodic fluids. Such formation waters probably evolved by interaction with evaporite and marine volcaniclastic and sedimentary rocks to attain their final chemical composition. The evolved chemical composition of these fluids, particularly the increased salinity, may have been necessary to cause the large
volumes of sodic-calcic and late stage sodic alteration observed at Yerington. The numerous occurrences of sodic-calcic alteration within plutons intruding volcanic arc terranes in western United States suggest this process may be widespread (e.g., Sparta granite, Oregon, Gilluly, 1933; Ajo porphyry copper district, Arizona, Gilluly, 1946; Mojave desert, Fox, 1988, and Solomon and Taylor, in prep.; Shamrock batholith, Yerington district, Battles, 1990).

CONCLUSIONS

This oxygen and hydrogen isotopic study of the Jurassic Ann-Mason deposit, Yerington district, Nevada, leads us to the following conclusions:

1) Fresh rocks from Yerington porphyry copper batholith have whole rock $\delta^{18}O$ of +6.7 to +7.0 and biotite $\delta D$ of -85 to -88. These are both typical "magmatic values" and are calculated to be in equilibrium with water of about $\delta^{18}O = +7.4$ and $\delta D = -65$ at 700°C.

2) Hydrothermally altered rock is generally little isotopically shifted from fresh rock, but along the axis of hydrothermal alteration from the center of the orebody upward for 3 km is $^{18}O$-enriched by +1 to +3.5 per mil. This is largely due to late-stage alteration.

3) Several potassically altered rock samples at 3.1-5.2 km paleodepth have whole rock or feldspar $\delta^{18}O = +6.5$ to +8.2 and biotite $\delta D = -68$ to -109 and are calculated to be in equilibrium at 500-600°C with a "magmatic water" ($\delta^{18}O = +6.5$ to +7.5; $\delta D = -60$ to -80). However, two biotite samples from 3.2-3.5 km paleodepth yield slightly D-enriched waters ($\delta D = -34$ to -42), probably due to partial chloritization during late-stage alteration by younger fluids.

4) Sodic-calcic (oligoclase-actinolite) alteration at 3 to 6 km paleodepth (oligoclase $\delta^{18}O = +5.7$ to 8.4; actinolite $\delta D = -67$ to -90) occurs
laterally and below potassic alteration and Cu mineralization, and was dominantly caused by non-magmatic fluids ($\delta^18O = <+3 \text{ to } +6$; $\delta D = -50$) that likely originated as isotopically "evolved", $^{18}O$-enriched formation waters, which circulated convectively into the hydrothermal system at ca. 400°C.

5) Late-stage sodic (albitic) and sericitic alterations are characterized by slight $^{18}O$- and D-enrichment (albite or whole rock $\delta^18O = +6.5 \text{ to } +10.4$; chlorite or muscovite $\delta D = -61 \text{ to } -85$). They are superimposed on main stage alteration at <1 to 4+ km paleodepth, and were produced by 200-400°C, convectively circulating fluids. Fluids with calculated (225°C) $\delta^18O \approx 0$ and $\delta D \approx -20 \text{ to } -55$ caused sericitic alteration at relatively high water/rock ratios and were most likely $^{18}O$-shifted meteoric water or, less likely, sea water.

ACKNOWLEDGEMENTS

Field studies by Dilles were supported by the Anaconda Company and its managers J. C. Wilson, K. L. Howard, Jr., and J. DeLong. We thank J. M. Proffett, R. B. Carten, K. L. Howard, Jr., and others for discussions of their insights into the Yerington district geology. Financial support of the oxygen isotope laboratory at Caltech and field support for Solomon were provided by National Science Foundation grants EAR-8313106 and EAR88-16413 to Taylor. We thank Richard Robles for doing several of the oxygen isotope analyses, Z. Sharp and J. Hunziker for use of the Lausanne oxygen isotope laboratory, and R. E. Criss doing ten of the D/H isotopic analyses. J. R. O'Neil, D. White, and I. Barnes provided access to D/H laboratories of the U.S.G.S., and we thank them and L. Adami and M. Huebner for their assistance.
REFERENCES CITED


Carten, R. B., 1986, Sodium-calcium metasomatism; chemical, temporal, and spatial relationships at the Yerington, Nevada, porphyry copper deposit: Econ. Geol., v. 81, p. 1495-1519.

Cathles, L. M., 1977, An analysis of the cooling of intrusives by groundwater convection that includes boiling: Econ. Geol., v. 72, p. 804-826.


Dilles, J. H., and Einaudi, M. T., in prep., Hydrothermal alteration at the Ann-Mason porphyry copper deposit, Yerington district, Nevada


Einaudi, M. T., 1977, Petrogenesis of the copper-bearing skarn at the Mason Valley Mine, Yerington district, Nevada: Econ. Geol., v. 72, p. 769-795.


Harris, N. B., and Einaudi, M. T., 1982, Skarn deposits in the Yerington district, Nevada: metasomatic skarn evolution near Ludwig: Econ. Geol., v. 70, p. 877-898.


Norton, D., 1978, Source lines, source regions, and path lines for fluids in hydrothermal systems related to cooling plutons: Econ. Geol., v. 73, p. 21-28.


Solomon, G. C., and Taylor, H.P., Jr., in prep., Oxygen isotope studies of a Jurassic hydrothermal system associated with rift-related magmatism, Rodman-Ord Mountains, California.


Taylor, H. P., Jr., 1974, The application of oxygen and hydrogen isotope studies to problems of hydrothermal alteration and ore deposition: Econ. Geol., v. 69, p. 843-883.


### TABLE 1
Whole-Rock Oxygen Isotopic Analyses

<table>
<thead>
<tr>
<th>Sample #</th>
<th>$^{18}O$</th>
<th>Rock Type</th>
<th>Alteration Assemblage*</th>
</tr>
</thead>
<tbody>
<tr>
<td>Y300</td>
<td>+7.9</td>
<td>Jsg</td>
<td>Sodic-calcic: Olg + Ksp + Act + Qtz + Sph</td>
</tr>
<tr>
<td>Y30</td>
<td>+6.8</td>
<td>QMD</td>
<td>&quot;Fresh&quot; with minor Chl replacing Hbl</td>
</tr>
<tr>
<td>Y23</td>
<td>+7.6</td>
<td>QMD</td>
<td>Chloritic PC: Ep + Chl + Calc + Plag + Ksp + Qtz + Sph</td>
</tr>
<tr>
<td>Y309*</td>
<td>+6.5</td>
<td>QMD</td>
<td>Sericitic TBx: tourmaline-quartz breccia with SQ clasts</td>
</tr>
<tr>
<td>Y310</td>
<td>+7.9</td>
<td>QMD</td>
<td>Chloritic CB: chloritized 2nd Bi + Ep and Qtz + Tm veinlets</td>
</tr>
<tr>
<td>Y311</td>
<td>+8.2</td>
<td>QMD</td>
<td>Chloritic PC: fresh Hbl and Mgt, weak Ep + Chl</td>
</tr>
<tr>
<td>Y315</td>
<td>+6.5</td>
<td>QMD</td>
<td>&quot;Fresh***</td>
</tr>
<tr>
<td>Y317</td>
<td>+7.0</td>
<td>QMD</td>
<td>Weak sodic-calcic SW: Ep-Ab veinlets; Act + Sph</td>
</tr>
<tr>
<td>Y320</td>
<td>+6.8</td>
<td>QMD</td>
<td>&quot;Fresh&quot; with Ab (+ Ep) on fractures</td>
</tr>
<tr>
<td>Y321</td>
<td>+6.5</td>
<td>QMD</td>
<td>&quot;Fresh&quot; with Ep (+ Ab) on fractures</td>
</tr>
<tr>
<td>Y323</td>
<td>+7.1</td>
<td>BG</td>
<td>Chloritic PC: chloritized mafics (+ Mgt)</td>
</tr>
<tr>
<td>Y325</td>
<td>+6.1</td>
<td>PG</td>
<td>&quot;Fresh&quot; with trace Ep ± 2nd Bi</td>
</tr>
<tr>
<td>Y326</td>
<td>+6.9</td>
<td>PG</td>
<td>Weak sodic-calcic SW(?): Ep-Ab veinlets</td>
</tr>
<tr>
<td>Y327</td>
<td>+6.9</td>
<td>QMD</td>
<td>Endoskarn ES-2: Qtz + Diop + Olg + Sph cut by Ep veins</td>
</tr>
<tr>
<td>Y329</td>
<td>+7.5</td>
<td>QMD</td>
<td>&quot;Fresh&quot;: 0.5 m from contact with Triassic metavolcanics</td>
</tr>
<tr>
<td>Y331</td>
<td>+7.4</td>
<td>QMD</td>
<td>Massive chlorite C: Chl + Qtz + Rut</td>
</tr>
<tr>
<td>Y333</td>
<td>+6.8</td>
<td>QMD</td>
<td>&quot;Fresh&quot; with weak Ep + 2nd Bi</td>
</tr>
</tbody>
</table>

#### Ann-Mason Area

#### Blue Hill Area

#### Buckskin Range

* Alteration assemblages are shown in Appendix A and abbreviations in Appendix B
**"Fresh" refers to fresh or very weakly altered rocks with no alteration minerals (except those noted) observable by hand lens.

$ Fluid inclusions from sample homogenized at 130-220°C with mode at 175-200°C and yielded 1-13 wt.% equivalent NaCl by freezing point depression (Dilles, 1984).

Rock type abbreviations are: Shamrock granite (Jsg); Artesia Lake andesite (Jaf); Yerington batholith units including McLeod Hill quartz monzodiorite (QMD), border granite of Bear pluton (BG), and Luhr Hill porphyritic granite (PG).
FIGURE CAPTIONS

Figure 1: Simplified map of the Yerington district, modified from Proffett and Dilles (1984), showing district sample locations and oxygen-isotope values.

Figure 2: Contoured whole-rock and feldspar oxygen isotopic map of the Ann-Mason porphyry copper deposit, with values projected to the surface north of the Singatse Fault. Data are from Tables 1 and 2; contours are in per mil and outline areas of slight $^{18}O$-depletion ($<+6.5$ per mil), corresponding to some sodic-calcic and distal propylitic alteration, and of moderate $^{18}O$-enrichment, corresponding to areas of late stage sodic, sericitic, and chloritic (PC and CB) alteration. Map area is identical to geologic map in figure 2 of Dilles (1987); additional alteration boundaries are shown on Figure 5.

Figure 3: Plots of mineral and whole rock $\delta D$ and $\delta^{18}O$ values from Ann-Mason. A. $\delta^{18}O$ versus paleodepth, where Lower Tertiary erosion surface is assigned a depth of 1 km. The vertical band indicates the range of $\delta^{18}O$ of unaltered Yerington batholith. Note $^{18}O$-enriched samples at shallow depths, particularly in sericitic, albitic, and chlorite-altered rocks affected by late stage fluids. B. $\delta D$ versus paleodepth. Note slight changes in $\delta D$ as a function of depth within and between alteration types, as discussed in text. C. Plot of $\delta D$ of hydrous mineral versus $\delta^{18}O$ of feldspar or whole rock. Note that propylitic samples are similar to unaltered Yerington batholith, but that all strongly altered rocks have a moderate range of isotopic composition and corresponding minor, but distinct shifts from magmatic values.
Figure 4: Calculated isotopic compositions of water in equilibrium with altered rock minerals (see text and Table 2). Compositions of Gulf Coast formation waters, Geysers, California geothermal waters, and magmatic waters are from Taylor (1974). "Porphyry copper biotites" is the field of calculated waters in equilibrium with hydrothermal biotite at 650°C from the Santa Rita, Bingham, and Ely porphyry copper deposits (Sheppard et al., 1971). Note the similarity of Ann-Mason potassic alteration fluids to "magmatic water", and of sericitic fluids to "O-shifted coastal, meteoric water or sea water; other fluids have intermediate compositions.

Figure 5: Model of fluid origin and circulation at the Ann-Mason porphyry copper deposit, Yerington, Nevada, as shown in schematic cross-sectional view (alteration boundaries from Dilles, 1984). Endoskarn alteration, main-stage potassic alteration and copper mineralization, synchronous main-stage sodic-calcic alteration, and late-stage sodic and sericitic alteration and pyritic mineralization are due to hydrothermal fluids of different age, temperature, temperature gradient, flow path, and isotopic compositions, as indicated. Abbreviations are shown in Appendix B, except Triassic-Jurassic sedimentary rocks (T_Js) and Triassic volcanic rocks (T_V).
OXYGEN ISOTOPE MAP of
ANN - MASON AREA
Yerington District, Nevada

EXPLANATION

GEOLOGY

Contact of PG
Geologic contact

OPPER GRADE
Area of >2000ppm Cu

ALTERATION BOUNDARIES

Limit of strong late stage alteration
Limit of sodic - calcic alteration S-1, S-3 and S-4
Limit of 100% biotized hornblende (strong potassic alteration)
Limit of endoskarn

ISOTOPIC DATA

$\delta^{18}O$ oxygen isotope contours in per mil

SAMPLE DESCRIPTION

$\delta^{18}O$ value, in per mil

SAMPLE TYPES

- Whole rock
- K-feldspar or albite
- Calcic oligoclase
- Oligoclase and albite
- Epidote

Fig. 2 (Dilles et al.)
ALTERATION KEY

- SODIC-CALCIC
- POTASSIC
- ENDOSKARN
- UNALTERED
- PROPYLITIC-ACTINOLITE
- CHLORITIC CB & PC
- SERICITIC
- SODIC (ALBITIC)

A. ALTERED ROCKS

- ENVELOPE OF ALTERED ROCKS

- ENVELOPE OF UNALTERED & PROPYLITIC-ACTINOLITE

- PLAGIOCLASE
- ALBITE OR K-FELDSP.
- WHOLE ROCK

B. Unaltered

- MUSCOVITE
- CHLORITE
- ACTINOLITE
- BIOTITE
Fig. 3c
Dilles et al.
Figure 4

Dilles et al.
Alteration Type

- Late-stage:
  - Sericitic SQ,TBx
  - Sodic (albitic) A1,A2
  - Propylitic PA
  - Sodic-calcic S-2,S-5
  - Potassic B
  - Endoskarn ES-2,ES-1
  - Skarn & skarnoid

Hydrothermal Fluid

- Meteoric Water or Seawater, 200-250°C
  - 250-400°C
- Evolved, non-magmatic 400°C
- Magmatic, 700-400°C
- Skarn, 700-400°C

Fig. 5
Dilles et al.
### Hydothermal Alteration Assemblages at Ann-Mason

| Alteration Type | Assemblage | New (Added) and Recrystallized Minerals | Relic Minerals
<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Endoskarn</td>
<td>ES-1</td>
<td>Grandite + Labradorite (An21-25) + Salite + Qtz + Hbl + Sph + Qtz (± Ep)</td>
<td>Olg/And + Sph + Hbl ± Ksp</td>
</tr>
<tr>
<td></td>
<td>ES-2</td>
<td>0lg/And(An21-25) + Salite + Hbl + Sph + Qtz (± Ep)</td>
<td></td>
</tr>
<tr>
<td>Sodic-Calcic</td>
<td>S-1</td>
<td>0lg/And + Sph + Qtz ± Rt</td>
<td>Olg/And + Sph</td>
</tr>
<tr>
<td></td>
<td>S-2</td>
<td>0lg/(Ab) + Act + Qtz + Sph + Ep (± Py)</td>
<td>Olg + Sph</td>
</tr>
<tr>
<td></td>
<td>S-3</td>
<td>0lg/(Ab) + Qtz + Sph + Ep ± Act (± Chl)</td>
<td>Olg/And + Hbl + Sph ± Ksp</td>
</tr>
<tr>
<td></td>
<td>S-4</td>
<td>0lg/Ab + Qtz + Sph + Ep + Act</td>
<td>Olg + Sph</td>
</tr>
<tr>
<td></td>
<td>S-5</td>
<td>0lg/Ab + Qtz + Sph + Ep + Act</td>
<td>Olg + Sph</td>
</tr>
<tr>
<td></td>
<td>S-6</td>
<td>Olg/Ab + Qtz + Sph + Ep + Act</td>
<td>Olg + Ksp + Sph</td>
</tr>
<tr>
<td></td>
<td>S-7</td>
<td>Olg/Ab + Qtz + Sph + Ep + Act</td>
<td>Olg + Ksp + Sph</td>
</tr>
<tr>
<td>Weak Sodic Calcic SW</td>
<td>Ep + Sph (± Olg)/Ab ± Qtz ± Act ± Bi ± Ser ± Py (± Chl) ± Rt</td>
<td>Olg + Ksp + Hbl + Sph (± Mgt)</td>
<td></td>
</tr>
<tr>
<td>Potassic</td>
<td>K</td>
<td>Ksp + Ab² + Bi + Qtz + Cp ± Bn ± MoS₂ + Rt + Chl¹</td>
<td>Olg + Ksp + Bi</td>
</tr>
<tr>
<td></td>
<td>B</td>
<td>Bi + Ep + Qtz + Ab² + Cp ± Bn or Py + Rt + Ser² + Calc + Chl¹</td>
<td>Bi ± Olg ± Ksp + Mgt ± Sph (± Ilm)</td>
</tr>
<tr>
<td></td>
<td>MB</td>
<td>B1 + Ep ± Cp ± Qtz</td>
<td>Olg + Ksp + Hbl + Mgt + Sph</td>
</tr>
<tr>
<td>Propylitic</td>
<td>PA</td>
<td>Act ± Bi + Chl + Ab² + Ep + Hm + Calc ± Ser ± Rt ± Py ± Tm (± Cp)</td>
<td>Olg/And + Ksp + Hbl ± Bi + Mgt + Sph</td>
</tr>
<tr>
<td>Chloritic</td>
<td>C</td>
<td>Chl + Rt ± Qtz</td>
<td>Ab ± Ksp ± 2nd Bi ± B1 + Mgt ± Ep ± Rt (± Cp)</td>
</tr>
<tr>
<td></td>
<td>CS</td>
<td>Chl + Hm + Py ± Ser ± Calc ± Ab² ± Ep (± Cp ± VerM)</td>
<td>Ksp + Mgt ± Olg</td>
</tr>
<tr>
<td></td>
<td>PC</td>
<td>Chl + Ab² + Hm + Rt ± Ep + Py ± Ser ± Calc (± Cp ± VerM)</td>
<td></td>
</tr>
<tr>
<td>Sodic (Albitic)</td>
<td>A-1</td>
<td>0lg/Ab + Chl + Rt ± Sph ± Py ± Qtz ± VerM</td>
<td>Olg ± 2nd Bi ± Bi ± Rt ± Cp (± MoS₂) (± Cp)</td>
</tr>
<tr>
<td></td>
<td>A-2</td>
<td>Ab + Chl + Ser + Rt + Py (± Tm ± VerM)</td>
<td>± Ksp (± Cp)</td>
</tr>
<tr>
<td></td>
<td>ATm</td>
<td>Ab + Tm + Rt + Py ± Chl ± Ser</td>
<td></td>
</tr>
<tr>
<td></td>
<td>SQ</td>
<td>Ser ± Qtz + Py ± Rt (± Cp ± Tm)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>TBx</td>
<td>Tm + Qtz + Py + Rt ± Ser (± Cp)</td>
<td>(± Cp)</td>
</tr>
</tbody>
</table>

( ) Mineral occurs in trace amounts.
$ $ Quartz, apatite, and zircon are present as relic minerals in all assemblages.
* * Albite forms pseudomorphs after primary igneous plagioclase.
# # Mineral is commonly present, but textures suggest that it partially or entirely post-dates formation of assemblage and is therefore largely retrograde.
### APPENDIX B: ABBREVIATIONS

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jaf</td>
<td>andesite from volcanic rocks of Artesia Lake</td>
</tr>
<tr>
<td>Jsg</td>
<td>Middle Jurassic (166 Ma) Shamrock batholith</td>
</tr>
<tr>
<td>YB</td>
<td>Middle Jurassic (169 Ma) Yerington batholith</td>
</tr>
<tr>
<td>QMD</td>
<td>quartz monzodiorite of YB</td>
</tr>
<tr>
<td>BG</td>
<td>border granite of Bear intrusion of YB</td>
</tr>
<tr>
<td>PG</td>
<td>porphyritic granite of Luhr Hill of YB</td>
</tr>
<tr>
<td>GP</td>
<td>granite porphyry dike of YB</td>
</tr>
<tr>
<td>Ab</td>
<td>albite</td>
</tr>
<tr>
<td>Act</td>
<td>actinolite</td>
</tr>
<tr>
<td>An</td>
<td>anorthite</td>
</tr>
<tr>
<td>And</td>
<td>andesine plagioclase</td>
</tr>
<tr>
<td>Bi</td>
<td>biotite</td>
</tr>
<tr>
<td>Calc</td>
<td>calcite</td>
</tr>
<tr>
<td>Chl</td>
<td>chlorite</td>
</tr>
<tr>
<td>Cp</td>
<td>chalcopyrite</td>
</tr>
<tr>
<td>Diop</td>
<td>diopside</td>
</tr>
<tr>
<td>Ep</td>
<td>epidote</td>
</tr>
<tr>
<td>Fsp</td>
<td>feldspar</td>
</tr>
<tr>
<td>Hm</td>
<td>hematite</td>
</tr>
<tr>
<td>Hbl</td>
<td>hornblende</td>
</tr>
<tr>
<td>Ilm</td>
<td>ilmenite</td>
</tr>
<tr>
<td>Ksp</td>
<td>K-feldspar</td>
</tr>
<tr>
<td>Mgt</td>
<td>magnetite</td>
</tr>
<tr>
<td>Olg</td>
<td>oligoclase plagioclase</td>
</tr>
<tr>
<td>Plag</td>
<td>plagioclase</td>
</tr>
<tr>
<td>Py</td>
<td>pyrite</td>
</tr>
<tr>
<td>Po</td>
<td>pyrrhotite</td>
</tr>
<tr>
<td>Qtz</td>
<td>quartz</td>
</tr>
<tr>
<td>Rt</td>
<td>rutile</td>
</tr>
<tr>
<td>Ser</td>
<td>sericite (muscovite)</td>
</tr>
<tr>
<td>Sph</td>
<td>sphene (titanite)</td>
</tr>
<tr>
<td>Tm</td>
<td>tourmaline</td>
</tr>
<tr>
<td>Verm</td>
<td>vermiculite</td>
</tr>
<tr>
<td>WR</td>
<td>whole rock</td>
</tr>
</tbody>
</table>
APPENDIX C: PETROGRAPHIC DESCRIPTIONS

Y-781: Biotite GP dike: 45.8% aplitic quartz-K-feldspar groundmass (0.05 mm)
with 64.2% phenocrysts. Mode is 36.5% plagioclase (An_{38-18}), 31.2% K-feldspar (in
g.m. and 1 cm microperthitic megacrysts), 21.8% quartz, 5.0% biotite (X_{A1}=.044,
X_{Fe}=.365, X_{Mg}=.553), 3.2% hornblende, <1% epidote, magnetite, sphene, zircon, and
chlorite.

Y-767: Medium-grained hornblende QMD: 39% plagioclase (An_{30-20}), 20% microcline,
15% quartz, 20% hornblende, 2% sphene, 2% magnetite, 1% Biotite (X_{A1}=.051,
X_{Fe}=.361, X_{Mg}=.548), trace ilmenite, inclusions in sphene.

Y-765A: Medium-grained QMD altered to endoskarn assemblage ES-2: 65% plagioclase
(An_{29}) partially replacing K-feldspar, 15% quartz, 17% hornblende, 2.5% sphene,
<0.5% actinolite, zircon, apatite, magnetite, and augite.

Y-29: GP with relict igneous 7% K-feldspar, 22% plagioclase (An_{23-19}), and 3%
quartz phenocrysts; mafics altered to 15% diopside (pseudomorphs hornblende),
1% epidote plus actinolite, and 1% sphene; 50% 0.02 mm groundmass of albite
(?) plus quartz.

101JD87: Quartz vein with abundant goethite replacing Ludwig limestone, adjacent
to plagioclase- diopside±garnet±wollastonite hornfels and skarn, Ludwig Cu Mine.

Y-508: PG with 30% microcline, 38% plagioclase (An_{25-20}), 20% quartz; mafics
consist of 1% relict hornblende, 3% igneous and 2% hydrothermal biotite, 2%
chlorite, 2% sphene, 1% epidote, and <1% each goethite, sericite, and clay.

Y-505B: Medium-grained PG in footwall of Singatse fault, potassically (B
assemblage) altered and containing 1% goethite and 0.5% malachite after primary
sulfides. Rock contains 30% microcline, 35% plagioclase 10% altered to sericite
and clay, 25% quartz, 8% igneous book (1-2 mm) and hydrothermal shredded (0.1-0.3
mm) red-brown to very pale golden brown pleochroic biotite; <1% sphene, zircon,
apatite, magnetite, and rutile. Veins of quartz ± epidote ± goethite ± malachite ±
rutile contain local muscovite and have K-feldspar selvages against
plagioclase.

Y-540-B: GP intruding PG with 14% microperthite 1 cm megacrysts, 36% plagioclase
(An_{25-19}), and 5% quartz.phenocrysts in 35% 0.05 mm K-feldspar-quartz groundmass;
3% igneous and 4% hydrothermal biotite (X_{A1}=.038, X_{Fe}=.392, X_{Mg}=.513); <0.5% each
sphene, magnetite, epidote, chlorite, goethite, ilmenite, pyrite, chalcopyrite,
rutile.

D-322-1661': (drill core D-322 at 1661 ft): GP from orebody containing 2%
aluminum and pyrite and altered to potassic assemblage B. Hornblende sites
are replaced by biotite, epidote, chalcopyrite, and pyrite; plagioclase (An_{28-18})
is locally altered along quartz veinlets and fractures to K-feldspar + epidote;
K-feldspar and igneous biotite are stable. Veins of quartz ± epidote ± muscovite
+ sulfides are associated with weak alteration of plagioclase to muscovite ±
epidote ± calcite and 10% alteration of biotite to chlorite. Biotite from nearby
D-222-917' has X_{A1}=.060, X_{Fe}=.333, X_{Mg}=.571.
V-313: QMD with 10% quartz, 15% microcline, 55% plagioclase (An_{38-18}) 20% altered to clay, epidote, and biotite; mafics are 1% relic hornblende, 2% igneous and 8% olive green hydrothermal biotite, 5% chlorite, 2% epidote, 1% sphene, 1% magnetite, and <0.5% rutile and apatite.

D-222-917': QMD with 10% quartz, 2% K-feldspar, 40% plagioclase (An_{1?}); and 18% sericite, 4% clay, epidote, and chlorite replace feldspar sites; mafics include 1% igneous and 12% olive green hydrothermal biotite (X_{A1}=.059, X_{Fe}=.345, X_{Mog}=.555), 5% epidote (partly in veins), 4% chlorite, 1% chalcopyrite, and <0.5% each rutile, pyrite, magnetite, apatite.

D-109-2682': GP dike with potassic K assemblage from ore zone with 2% chalcopyrite + bornite and quartz + K-feldspar ± epidote veinlets; groundmass is flooded with pink K-feldspar; plagioclase (28%, now albite) is 75% altered to microcline, sericite, and clay; mafic minerals are altered to hydrothermal 5% biotite, 2% chloritized biotite, 0.5% rutile, 1% calcite, and 0.5% sphene.

V-548-5": PG with 8% K-feldspar megacrysts replaced by oligoclase (An_{22?}) + 1% quartz, 2.5% 1mm quartz, 60% plagioclase (An_{27-17}), 15% aplitic .05 mm quartz plus plagioclase, 9% actinolite, 4% epidote, and 2% sphene.

V-756: PG, pervasively altered to sodic-calcic assemblage S-2. Rock contains 65% plagioclase (An_{21-24}) replacing igneous plagioclase and K-feldspar, 18% quartz, 15% actinolite (X_{A1}=.0055, X_{Fe}=.152, X_{Mog}=.838) replacing hornblende and biotite, 1.5% sphene, and <0.25% each of epidote, apatite, zircon, and rutile.

V-774D: GP dike altered to sodic-calcic assemblage S-2/S-6. Rock contains 55% .05 mm plagioclase-quartz groundmass; 2% quartz phenocrysts; 29% plagioclase (An_{6}) replacing primary plagioclase and K-feldspar phenocrysts; 2% epidote, 1% clay. Hornblende is pseudomorphed by 5% actinolite (X_{A1}=.042, X_{Fe}=.229, X_{Mog}=.767); igneous biotite is altered to 6% actinolite, calcite, sphene, and rutile; there are traces of apatite and zircon.

V-316: GP dike with 36% plagioclase (An_{25-19}), 1% 0.5 mm quartz, and 5% K-feldspar sites altered to oligoclase and quartz in a ratio of 4:1 within 50% groundmass of 0.03 mm quartz-plagioclase; mafics are 4% actinolite, 3% epidote, 1% sphene, and trace calcite and apatite.

Y-530: GP, pervasively altered to sodic-calcic assemblage S-6. Rock contains 4% 0.2-1 mm quartz phenocrysts. 26% plagioclase phenocrysts are unzoned (=An_{60}) with trace sericite; groundmass (50%) is .01 mm and altered to albite and quartz, 5% epidote, and trace calcite, apatite, and sphene; 5% K-feldspar megacrysts are altered to epidote and 3% quartz; mafics are altered to 10% actinolite, 5% epidote, and 0.5% sphene.

Y-645: GP dike with two superimposed alteration types: (1) early weak B assemblage potassic alteration characterized by 80% replacement of igneous hornblende by shredded biotite, epidote, and quartz; (2) later weak sodic-calcic alteration (assemblage SW), characterized by 10-25% alteration of K-feldspar to rims of albite + 10% quartz or massive epidote; plagioclase alteration to albite (An_{60}), epidote, calcite, and clay; biotite 60% altered to chlorite; and stable actinolite, sphene, and 0.25% pyrite.
Y-793: Propylitic PA-altered fine-grained QMD. Rock contains 50% plagioclase (An_{30-38}) containing minor albite and 10% sericite, clay, and epidote; 11% K-feldspar, 7% quartz, 18% pale green actinolite, including euhedral pseudomorphs of hornblende and fibrous aggregates ($X_A=.047, X_F=.264, X_M=.681$), 5% clay, 3% chlorite, 2% epidote, 2% magnetite, 1% sericite, 1% sphene, and <1% ilmenite and zircon.

Y-307: Propylitic PA-altered fine-grained QMD near Y-793. Rock contains 47% plagioclase (An_{28-18}) 5% altered to epidote + clay, 15% K-feldspar, 15% quartz, 12% pale green actinolite as pseudomorphs of hornblende and fibrous aggregates; 3% magnetite + hematite, 1.5% leucoxene (spheine?). Rock contains 2% chlorite associated with epidote + leucoxene ± actinolite that replaced biotite and contains rare 0.1 mm calcite and epidote veinlets.

Y-40: PG with 25% quartz, 60% plagioclase (An_{50-18}), 5% microperthitic K-feldspar (85% of K-feldspar was altered to chessboard-twinned plagioclase plus quartz), 8% pale brown biotite, 0.75% chlorite, <0.5% epidote, rutile, apatite, calcite, muscovite, and apatite.

D-114A-1497': QMD altered to albitic assemblage A-1 from orebody. Rock contains 14% quartz; plagioclase and K-feldspar altered to 65% milky plagioclase (An_{28-18}), 4% clay, 0.5% calcite, and 1% epidote; mafic minerals altered to 12% shrdy pale brown biotite-vermiculite intergrowths, 0.5% chlorite, 0.5% rutile, 0.5% sphene, 1.5% chalcopyrite; and traces of apatite, zircon, pyrite, and sericite. Biotite-vermiculite composition from nearby D-222-2863 is $X_A=.041, X_F=.068, X_M=.884$.

Y-799A: GP altered to albitic assemblage A-2. Rock contains 60% <0.05 mm quartz-albite altered groundmass; 5% quartz phenocrysts; plagioclase and K-feldspar phenocrysts altered to 23% albite (An_{22}) and 2% sericite; mafic minerals altered to 4% chlorite ($X_A=.225, X_F=.304, X_M=.452$), 2% sericite; 1-2% each of rutile and goethite (after pyrite); traces of apatite and zircon.

Y-694C: GP dike altered to sericitic assemblage SQ. Rock contains 55% quartz, 40% sericite, 2% jarosite and trace goethite after pyrite, 1% clay, and 0.5% leucoxene (rutile?).

100JD87: BG? pervasively altered to 80% quartz, 19% sericite, 0.5% jarosite, 0.5% rutile, trace hematite; mafic sites contain more sericite and aggregates of rutile and jarosite; contains 5% quartz veins. Sericite from nearby Y-309 has $X_A=.920, X_F=.031, X_M=.044$. 