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STABLE ISOTOPE INVESTIGATION OF GOLD QUARTZ VEINS AT THE ORIENTAL MINE, ALLEGHANY DISTRICT, CALIFORNIA

by

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INTRODUCTION

The Oriental mine, Alleghany district, California, contains one of the geologically best documented gold quartz vein systems in the Sierra Nevada foothills metamorphic belt (Ferguson and Gannett, 1932; Coveney, 1981; Böhlke, 1986). The veins exposed in the Oriental mine are similar, chemically and mineralogically, to many others throughout the Sierran foothills, and in other metamorphic belts worldwide (e.g. Knopf, 1929; Kerrich, 1981; Phillips and Groves, 1983). The veins at Alleghany apparently were formed in Cretaceous time from CO_2-bearing fluids that flowed through concordant and discordant fractures in older igneous and metamorphic rocks (Böhlke and Kistler, 1986). On the basis of oxygen isotopic data, phase equilibria, and fluid inclusion data, hydrothermal alteration of the wall rocks and much of the vein filling are estimated to have occurred at 320 ±60°C (Marshall and Taylor, 1981; Böhlke and Kistler, 1986; Böhlke, 1986) and at pressures greater than 670 bars (Coveney, 1981) or roughly 2000 ±1000 bars (Böhlke, 1986), in the presence of a fluid with roughly 10 mole per cent CO_2 (Coveney, 1981; Böhlke, 1986). The ultimate source(s) of the hydrothermal fluids is not known. The CO_2-bearing fluids clearly were not in chemical equilibrium with the vein wall rocks near the sites of ore deposition, as the veins are almost everywhere enclosed within metasomatic (carbonatized) alteration halos (Böhlke, 1986). Much of the disseminated gold mineralization appears to have been associated genetically with specific wall rock alteration reactions (Böhlke, 1986), but occurrences of high-grade gold in the veins may imply that other precipitation mechanisms also were locally important. Fluid inclusions containing varying amounts of CO_2 and exhibiting a wide range of homogenization temperatures occur in vein quartz, but their relationships to each other and to gold deposition have not been resolved.

The data reported here constitute a preliminary study, begun in the early 1970's by Coveney and Rye, of the isotopic compositions of fluids and hydrothermal minerals in the Oriental mine vein system (Figure 1). In addition to providing support for inclusion of the Alleghany veins in the general class of gold quartz (low-sulfide) veins in Archean to Phanerozoic metamorphic terranes (Böhlke, 1982), the data also yield information on possible sources of the hydrothermal fluids and dissolved species, and they constrain the interpretation of chemical changes in the hydrothermal fluids during alteration and mineralization.

The major vein structures sampled for this study are the Oriental vein and the Carrolson vein (Figures 1 and 2). Both veins are composed mainly of massive to vuggy quartz, with minor amounts of carbonates, sulfides, chlorite, mica, feldspar, carbon, and gold (Ferguson and Gannett, 1932; Coveney, 1981). Arsenopyrite is a distinctive accessory phase commonly associated with gold mineralization. Wall rock alteration is conspicuous within centimeters to meters of the veins. The dominant alteration reactions involved varying amounts of carbonatization, de-silicification, chloritization, sericitization, albitization, and sulfidization, depending on distance from the veins and on the initial wall rock lithology (Coveney, 1981; Böhlke, 1985; 1986).

Ultramafic rocks are pervasively serpentinized throughout the district. Locally (near the veins) serpentinite was progressively altered to talc + magnesite + chlorite, to magnesite + quartz + chlorite, and finally to dolomite + magnesite + quartz + mica (mariposite). High grade gold ore commonly occurs within the quartz veins near their intersections with altered ultramafic rocks in the district (Ferguson and Gannett, 1932). The altered serpentinite resembles the carbonate- and fuchsite-bearing "listwänites" that accompany gold quartz mineralization in many parts of the world (e.g. Kashkai and Allakhverdiev, 1965; Boyle, 1979).

A distinctive body of porous albitized, mineralized granite is exposed in the footwall of the Oriental vein on the 1300 level. Altered Oriental granite is weakly carbonatized, and portions of it contain a few per cent of arsenopyrite and pyrite. Disseminated gold is most abundant in samples with the highest arsenic contents, but the gold occurs largely as inclusions in pyrite (Coveney, 1981; Böhlke, 1986). Porosity was created by leaching of igneous quartz near subhorizontal fractures. Desilicified granitic rocks ("episyenites") are commonly associated with hydrothermal gold and uranium deposits elsewhere (e.g. Leroy, 1978; Harris, 1980).

The Oriental mine affords an exceptional opportunity to study a variety of important geochemical processes related to gold transport and precipitation in relatively deep CO_2-rich hydrothermal systems. Samples of quartz, sulfides, carbonates, albite, rutile, and graphite for this study were taken from veins and from hydrothermally altered wall rocks (see Figure 2). Vuggy and massive vein quartz samples were collected near and far from high-grade gold ore pockets, in order to determine whether δD (of fluid inclusions) or $\delta^{18}O$ (of quartz) could be used as guides to ore. Isotopic evidence was sought in part to test also (1) Coveney's (1971,1972) suggestion that high-grade gold ores precipitated from fluids undergoing reduction by H_2 from serpentinite (cf. Barton's (1970) discussion of Hg ores), and (2) the suggestion that episyenites lose silica to high pH fluids that are created by extensive CO_2 exsolution (e.g. Maisonneuve et al., 1984). Serpentinite and altered (talc-bearing) serpentinite were also sampled, to see if the same fluid might have caused both serpentinization and carbonatization, as implied recently by Nesbitt et al. (1986) for some gold districts in the Canadian cordillera.

The bulk of the samples were collected in 1970-1973 by R.M. Coveney, Jr. and analysed in the laboratory of R.O. Rye in 1974. The others were collected in 1980-1982 by J.K. Böhlke and analysed in the laboratory of I. Barnes. Some of the data were discussed in summary fashion by Böhlke and Kistler (1986). All isotope ratios presented here are given in conventional delta (δ) notation as per mil deviations from the ratios of standards (SMOW for oxygen and hydrogen; PDB for carbon; CDT for sulfur) (see O'Neil, 1979).

OXYGEN ISOTOPES

The $\delta^{18}O$ values of quartz from veins and veinlets range from +17.3 to +19.8 per mil (Tables 1 and 2). The highest values are from clear euhedral crystals that grew on albite after de-silicification of the altered granite. Within the Carrolson vein, there is no significant difference in $\delta^{18}O$ between quartz collected from gold ore pockets and that collected far from ore pockets (Table 3), though it is not yet known whether any of the gold-bearing quartz actually was precipitated at the same time as the gold. Clear drusy crystals from vugs have a slightly higher average $\delta^{18}O$ (+18.3 per mil) than massive milky quartz (+17.7 per mil) in the Carrolson vein. This may reflect a slightly lower temperature of deposition, or deposition from slightly higher $\delta^{18}O$ fluids; however, more detailed work is needed to establish whether or not the difference is significant.

The bulk of the vein quartz probably was deposited at 320 ±60°C from a fluid with $\delta^{18}O_{H_2O}$ between +8 and +14 per mil (Marshall and Taylor, 1981; Böhlke and Kistler, 1986). These large positive $\delta^{18}O_{H_2O}$ values are similar to those expected for "metamorphic" fluids (Taylor, 1974); however, they are interpreted in part to reflect extensive interaction between the vein-forming fluids and metamorphic wall rocks along their flow paths, such that the fluid "source" characteristics may be masked. The range of vein quartz $\delta^{18}O$ values from the Oriental and Carrolson veins is restricted to the middle of the range of values obtained from other veins near Alleghany (+16 to +22 per mil, e.g. Böhlke and Kistler, 1986). The hydrothermal fluids that formed the Oriental and Carrolson veins interacted locally with mixed metamorphosed igneous and volcano-sedimentary rocks. Fluids with higher $\delta^{18}O_{H_2O}$ values elsewhere in the district probably interacted with more siliceous (chert-rich) metasedimentary rocks before depositing vein quartz with $\delta^{18}O$ values as high as +22 per mil; whereas fluids with lower $\delta^{18}O_{H_2O}$ values apparently interacted with more ultramafic to intermediate meta-igneous rocks before depositing vein quartz with $\delta^{18}O$ values as low as +16 per mil (see Marshall and Taylor, 1981; Böhlke et al., 1984; Rosenbaum and Taylor, 1984; Böhlke and Kistler, 1986).

Crystals of rutile (crystal structure confirmed by X-ray diffraction) occur in albitized granite and on the walls of some veinlets with ankerite ± albite. Needles of rutile were commonly overgrown by euhedral quartz crystals in open vugs and veinlets. The average $\delta^{18}O$ values for rutile (+3.4) and albite (+16.0) (Table 2) yield $\Delta_{albite-rutile} \approx +12.6$ per mil. The albite and rutile could have been in equilibrium at approximately 260-280°C with a fluid with $\delta^{18}O_{H_2O} \approx +9$ to +10 (rutile-water fractionation from Matthews et al., 1979; albite-water fractionation from O'Neil and Taylor, 1967, or Bottinga and Javoy, 1973), or at approximately 360-370°C with a fluid with $\delta^{18}O_{H_2O} \approx +12$ (rutile-water fractionation extrapolated from Addy and Garlick, 1974). Similar ranges of temperatures may be obtained from the fractionation between rutile and hydrothermal quartz from the granite (quartz-water fractionation from Bottinga and Javoy, 1973), indicating that quartz, albite, and rutile all could have been approximately in isotopic equilibrium. The calculated equilibrium temperatures for rutile, albite, and quartz in veinlets and vugs in altered porous granite span the range calculated by Böhlke and Kistler (1986) from oxygen isotope analyses of mica and quartz in veinlets in altered serpentinite from other mines in the district (320 ±60°C). In both instances, the observed fractionations are fairly constant, and most of the uncertainties in temperatures are due to uncertainties in the experimental or theoretical fractionation factors.

HYDROGEN ISOTOPES

Four samples of serpentinite collected at various distances from the Carrolson vein on the 2300 level have not been obviously altered mineralogically by the vein-forming fluid. These serpentinites have fairly uniform δD values (-107 to -124 per mil) (Table 4). The low δD values suggest that these "unaltered" serpentinites last equilibrated with meteoric water (cf. Wenner and Taylor, 1974).

Samples of serpentinite showing progressive degrees of hydrothermal alteration (increasing talc contents) near the vein have increasingly higher δD values. The difference between the δD values of "unaltered" and "altered" samples (>50 per mil) is probably larger than the hydrogen isotope fractionation between serpentine and talc (e.g. Bowers and Taylor, 1985); therefore, the fluid that caused the hydrothermal alteration probably had a significantly higher δD_{H_2O} value than the fluid that equilibrated with the serpentinites.

The δD_{H_2O} values of fluid inclusions released from quartz by crushing vary between -39 and -103 per mil (Table 1). There is no significant difference in δD_{H_2O} between quartz fluids collected from gold ore pockets and quartz fluids collected far from ore pockets (Table 3). However, there apparently is a substantial difference between the δD_{H_2O} values of fluids in clear crystalline quartz and fluids in massive milky quartz; inclusion fliuds from milky quartz have consistently more negative δD_{H_2O} values than fluids from vug quartz.

The fluid inclusion δD_{H_2O} data are tentatively interpreted to reflect mixtures of two fluid types: (1) an isotopically heavier hydrothermal fluid that occupies dominantly primary fluid inclusions in unstrained (clear) quartz, and (2) an isotopically lighter fluid that occupies secondary fluid inclusions in strained, microfractured, and partially annealed (milky) quartz. This interpretation is similar to that given by Foley et al. (1982) for low δD_{H_2O} values obtained from fluid inclusions in hydrothermal quartz from Creede, Colorado (see also Böhlke and Kistler, 1986; Pickthorn et al., 1987).

Because all of the quartz samples probably contain mixtures of the two populations, the δD_{H_2O} values of the two hypothetical fluid end members cannot be determined with certainty. However, it is possible that secondary inclusions dominate some of the milky quartz samples, and that they contain a meteoric fluid component with a δD_{H_2O} of approximately -100 ± 5 per mil. One analysis of late (post-quartz) calcite yielded a similar δD value (-105 per mil)(Table 1).

The inclusion fluids with the highest δD_{H_2O} values are most likely to be those responsible for altering serpentinite to talc-bearing rock, and ultimately to mica-bearing rock (Böhlke and Kistler, 1986). They also are likely to have had relatively high CO_2 contents, as both talc and mica were formed during carbonatization reactions in the ultramafic rocks (Böhlke, 1986). The relationship between δD_{H_2O} and CO_2/H_2O for the inclusion fluids sampled in the present study is shown in Figure 3. Although there is considerable scatter, it appears that there may be a positive correlation between δD_{H_2O} and CO_2/H_2O. This is qualitatively consistent with Marshall and Taylor's (1981) statement that higher δD_{H_2O} values are obtained from Alleghany fluid inclusions with higher

CO_2 contents. Five hydrothermal micas in altered ultramafic rocks from the Alleghany district yield δD values of -49. ± 9. per mil (Böhlke and Kistler, 1986). The heaviest inclusion fluid ($\delta D_{H_2O} = -39$ per mil) could be isotopically similar to the ones that formed some of the micas (T = 320 ± 60°C; Böhlke and Kistler, 1986; Böhlke, 1986; Bowers and Taylor, 1985); however, it should be noted that the measured CO_2/H_2O ratio of this sample is almost two orders of magnitude lower than that inferred for the metasomatic fluid, and observed in many individual fluid inclusions ($X_{CO_2} \approx 0.1$) (Coveney, 1981; Böhlke, 1986).

Paleomagnetic data from plutonic rocks of Late Jurassic to Early Cretaceous age in the Sierra Nevada indicate that the Early Cretaceous gold quartz veins in the foothills metamorphic belt probably formed at approximately 40 to 50° N latitude (Irving, 1979). Stratigraphic studies of onlapping sedimentary sequences in the Great Valley indicate that the metamorphic belt was exposed subaerially, and that the shoreline was approximately 110-120 kilometers west of Alleghany, between 100 and 125 m.y. ago (Ingersoll, 1982). The shoreline moved east to approximately 60 kilometers from Alleghany by 75 m.y. ago (op cit.). Present-day groundwaters in the vicinity of 38 to 40° N along the California coast decrease in δD away from the shoreline (White et al., 1973; Ingraham and Taylor, 1984) (see Figure 4). According to analyses of springs reported in White et al. (1973), modern δD values range from approximately -35 to -65 per mil at 30 to 100 kilometers from the Pacific Ocean. According to Ingraham and Taylor (1984), modern δD values of "ground waters" range from -50 to -75 per mil at 0 to 55 kilometers from the Pacific. Modern meteoric waters in the Sierra Nevada are between approximately -80 and -130 per mil, generally decreasing with elevation (Friedman, et al., 1964; Barnes et al., 1981; Ingraham and Taylor, 1984).

From these data (which appear partly to be in conflict) it is not possible to state with certainty if the inclusion fluids with the lowest δD_{H_2O} values could be Early Cretaceous meteoric water that entered the veins during or soon after the high δD_{H_2O} fluids, or if they must have entered much later when the coastal configuration and the elevation of the Sierra Nevada were more like they are today. Note that the inferred δD_{H_2O} values for secondary fluid inclusions in the Oriental mine (-100 ±5 per mil) are close to those expected for modern meteoric water at Alleghany (Figure 4). It is possible that the highest δD_{H_2O} values observed in quartz inclusions (and implied by the micas) are higher than those of Cretaceous meteoric waters. They could well have been in equilibrium with low- to medium-grade hydrous metamorphic rocks (i.e. "metamorphic fluids") (e.g. Taylor, 1974). Fluids which entered the system in a variety of ways (e.g. degassed from magmas or prograding metamorphic minerals, or as deeply circulating meteoric water) may be isotopically indistinguishable if they equilibrated with metamorphic rocks at low water:rock ratios. Alternatively, it is possible that the higher δD_{H_2O} values are representative of Early Cretaceous meteoric water (Nesbitt et al., 1986). This would require that the inferred secondary inclusions were trapped considerably later, possibly within the last few tens of millions of years, when the Great Valley was exposed above sea level (Ingersoll, 1982), or within the last ten million years, when the Sierra Nevada underwent rapid uplift (e.g. Bateman and Wahrhaftig, 1969; Huber, 1981). Fluid inclusion studies have not been detailed enough yet to determine if both fluids were present in the veins at the same time.

CARBON ISOTOPES

With one exception, the $\delta^{13}C$ values of calcite, dolomite, and ankerite are between -1.5 and -3.1 per mil (Table 2). The values for three ankerite samples from altered granite are identical ($\delta^{13}C \approx -2.9$ per mil). The relative constancy of these values suggests that major fluid reduction did not accompany the deposition of these carbonates, and it is consistent with fluid inclusion evidence that $CO_2 \gg CH_4$ in the vein fluids (e.g. Coveney, 1981; Böhlke and Kesler, unpublished data). The data are consistent with the lack of fluid inclusion evidence for fluid unmixing in these veins (Coveney, 1981; Böhlke, unpublished data). Deposition of dolomite and ankerite apparently was not accompanied by large degrees of $CO_2 - H_2O$ unmixing, as exsolution of a CO_2-rich vapor phase at ≥200°C would have left the remaining dissolved carbonate species depleted in ^{13}C (Friedman and O'Neil, 1977).

The $\delta^{13}C$ values of all but one of the Oriental mine carbonates are within the relatively narrow range of values (-7 to -1 per mil) reported previously for vein carbonate minerals from Alleghany, the Sierra Nevada foothills, and similar deposits worldwide (e.g. Marshall and Taylor, 1981; Böhlke and Kistler, 1986; Colvine et al., 1984; Burrows et al., 1986; Weir and Kerrick, 1987). These $\delta^{13}C$ values are ambiguous with regard to the source of carbon (e.g. Colvine et al., 1984). The narrow range of $\delta^{13}C$ values, their rough similarity from place to place, and their similarity to those of "magmatic carbon", have been used to argue in favor of a magmatic source for CO_2 in Archean gold quartz vein fluids (e.g. Burrows et al., 1986). However, many of the values (including those in Table 2) are also close to those expected from sedimentary carbonate sources, and all reported $\delta^{13}C$ data from similar veins could be explained by mixtures of oxidized and reduced sedimentary reservoirs in proportions not much different from the crustal average (e.g. Ohmoto and Rye, 1979). A juvenile or magmatic source for much of the carbonate may be consistent with the geochronologic and isotopic data from Alleghany, but so could a carbonate-bearing sedimentary source that was tapped during metamorphism below (at higher temperature than) the mineralized veins.

The one exceptional $\delta^{13}C$ value (-10.5 per mil) is from a sample of late calcite filling vugs in a quartz veinlet in porous altered granite. Earlier ankerite on the walls of the same veinlet has $\delta^{13}C \approx -2.9$ per mil. The late calcite could have precipitated from a main-stage fluid that had either lost an immiscible CO_2 vapor phase or incorporated organic carbon from the vein wall rocks. It is also possible that the late fluids were unrelated to main stage fluids, and either mixed with or followed the main stage fluids in the vein fractures.

One sample of graphite from an altered wall rock (crystal structure confirmed by X-ray diffraction) was analysed ($\delta^{13}C = -26.2$ per mil). This carbon is probably too light to have been in isotopic equilibrium with any of the carbonates in the veins at 320 ±60°C (Friedman and O'Neil, 1977), but it is isotopically similar to sedimentary organic carbon (Schwarz, 1969). The $\delta^{13}C$ value of this graphite sample is slightly lower than those reported by Taylor (1981) for carbon in veins at Alleghany ($\delta^{13}C \approx -20$ to -25).

SULFUR ISOTOPES

With one exception, the $\delta^{34}S$ values of pyrite are between 1.6 and 3.1 per mil (Table 5). The values themselves, and the fairly small variation, are consistent with the interpretation that the pyrites were precipitated from a relatively homogeneous, reduced fluid, and that no major oxidation events accompanied their deposition (e.g. Ohmoto, 1972; Rye and Ohmoto, 1974). The $\delta^{34}S$ values of the Oriental mine pyrite are indistinguishable from the majority of values reported for pyrite in Archean gold quartz veins from Western Australia (+1 to +4 per mil, Lambert et al., 1984) and from around the world (+1 to +6 per mil, Colvine et al., 1984). These values are generally slightly more positive than those of most magmatic sulfides, but could indicate various sources within the metamorphic belts (e.g. Ohmoto and Rye, 1979; Lavigne and Crocket, 1984; Lambert et al., 1984).

The one exceptional $\delta^{34}S$ value (-25.8 per mil) is from drusy pyrite crystals which grew on the surfaces of coarse-grained (Type I) arsenopyrite crystals relatively near the present-day ground surface. Relatively late pyrite with a similar habit is commonly associated with calcite ± quartz or chalcedony throughout the Alleghany district, and could have formed from relatively oxidized ground waters some time after the main flux of deep-source hydrothermal fluids through the veins. Isotopically anomalous iron disulfides are a common feature of many hydrothermal deposits, and are believed to be related to downward flooding of systems by overlying high f_{O_2}, low pH waters (e.g. Plumlee and Rye, 1987).

Arsenopyrite $\delta^{34}S$ values are all between 2.1 and 3.2 per mil, except for one at 5.5 per mil. Two arsenopyrite-pyrite assemblages from altered granite yield apparent values of $\Delta_{arsenopyrite-pyrite}$ equal to 0.76 and 0.92. Another pair of samples from the 200 level of the vein yields $\Delta_{arsenopyrite-pyrite}$ of 1.02. The consistency of the arsenopyrite $\delta^{34}S$ values, and of the apparent arsenopyrite-pyrite fractionations, and the general absence of cross-cutting or replacement textures, suggest that both minerals formed under roughly the same physical-chemical conditions and possibly precipitated contemporaneously.

The differences in $\delta^{34}S$ between the galena sample and the three vein pyrite samples (not from the same mine level) yield apparent fractionation temperatures of approximately 245, 305, and 375°C (Ohmoto and Rye, 1979), which are roughly consistent with the range of possible temperatures suggested by $\delta^{18}O$ fractionations between hydrothermal quartz and mica in the Alleghany district (320 ±60°C, Böhlke and Kistler, 1986), and with the compositions of coexisting calcite and siderite from the Oriental mine ($\leq$350°C, Böhlke, 1986). The sphalerite could have been in isotopic equilibrium with coexisting pyrite (near the heavy end of the observed pyrite range) at similar temperatures.

No consistent variations in fluid $\delta^{34}S$ with elevation are indicated, nor are there systematic differences between vein and wall rock sulfides (all of which are interpreted to be hydrothermal, not indigenous). The heaviest sulfur was found in a single sample of acicular arsenopyrite (Type III), which is interpreted by Coveney (1981) to have formed somewhat later than Types I and II.

CONCLUSIONS

Though limited, the isotope data from the Oriental mine suggest the following general conclusions:

(1) The carbon, sulfur, and oxygen isotope data are similar to those reported from many other gold quartz vein systems in metamorphic terranes worldwide, and are consistent with the majority of such systems having formed in similar physical and chemical environments.

(2) The sources of carbon and sulfur are not uniquely determined, but could include metamorphic rocks and (or) juvenile components.

(3) The $\delta^{18}O_{H_2O}$ of the hydrothermal fluids must have been buffered in large part by exchange with the metamorphic rocks. Oxygen isotope ratios in vein quartz from the Oriental mine have a narrower range than the ratios in all vein quartz from various parts of the Alleghany district (cf. Marshall and Taylor, 1981; Böhlke and Kistler, 1986). This is consistent with the interpretation that vein-forming fluids in different parts of the district exchanged oxygen with different wall rocks along their flow paths.

(4) Hydrogen isotope data indicate that vein quartz samples contain mixed populations of fluid inclusions. There is some evidence that inclusions with higher CO_2/H_2O ratios have higher δD_{H_2O} values. The highest δD_{H_2O} values are consistent with a metamorphic origin for the CO_2-bearing fluids, but a meteoric origin cannot be ruled out on this basis alone; however, if the higher δD_{H_2O} values do correspond to unexchanged Cretaceous meteoric water, then the fluid inclusions with low δD_{H_2O} values must have been trapped a long time after the main stages of alteration and vein formation.

(5) The fluids responsible for hydrothermal alteration (carbonatization and steatization) had higher δD_{H_2O} values than the fluids responsible for much of the serpentinitization This is consistent with, but does not confirm, the interpretation that serpentinization and gold mineralization occurred independently (cf. Böhlke, 1986).

(6) Most of the isotopic data are consistent with the interpretation that the "main stage" hydrothermal fluid had concentrations of sulfide > sulfate, and $CO_2 > CH_4$, and that the bulk of the sulfide and carbonate deposition was not accompanied by large excursions in the oxidation state of the fluid (see Böhlke, 1986).

(7) No direct isotopic evidence can be cited yet to support the contention that H_2 gas from serpentinite had a role in precipitating high-grade gold ores in the veins (see Coveney, 1981).

(8) There is no isotopic evidence for extensive fluid unmixing during or immediately following de-silicification of the Oriental mine granite to form episyenite, or during precipitation of vein dolomite. It is possible (but certainly not proven) that CO_2 exsolution accompanied or preceded relatively late calcite deposition in open cavities in vein quartz.

(9) Sulfur, carbon, and hydrogen isotope data all are consistent with the possibility that fluids from shallower reservoirs entered the veins after much of the hydrothermal alteration and vein filling by CO_2-bearing fluids from deeper sources took place. It remains to be demonstrated whether or not the different fluids mixed and caused precipitation of some of the gold in the veins. It is possible that fluid mixing and wall

rock alteration caused high-grade and disseminated mineralization, respectively (cf. Böhlke, 1986; Böhlke et al., 1987; Kirschbaum et al., 1987).

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Table 1. Oxygen isotope analyses of vein quartz and calcite, and hydrogen isotope and CO_2/H_2O analyses of fluid inclusions, Carrolson vein

Sample #	**Mineral**	$\delta^{18}O$	δD_{H_2O}	CO_2/H_2O	**Remarks**
Q-62	quartz	17.5	-56	.007	1900 level, east end; vug crystal
"	"	17.6	--	--	"
19-7E	quartz	17.5	-89	.0014	1900 level, east end; massive milky quartz
19-2W	quartz	18.2	-80	.0012	1900 level, near gold pocket; massive milky quartz
Q-63	quartz	18.8	-66	.0012	2300 level, near gold pocket; vug crystal
Q-17	quartz	19.4	-39	.0018	2300 level, east end; vug crystal
9-23-216	quartz	17.3	-79	.0019	2300 level, high grade gold pocket; massive milky quartz
"	"	17.7	-92	.0007	"
"	"	18.0	--	--	"
23-1W	quartz	17.3	-103	<.0006	2300 level; massive milky quartz
23-4W	quartz	--	-91	--	"
23-6W	quartz	--	-100	--	"
23-9W	quartz	17.8	-84	--	"
9-205-94	calcite	22.5	-105	--	2050 level; vug crystals, associated with pyrite; $\delta^{13}C$ is -3.1 per mil

Note: analysed by R.O. Rye, using techniques described by Rye (1966) and Rye and Rye (1974); $\delta^{18}O$ and δD in per mil relative to SMOW; $\delta^{13}C$ in per mil relative to PDB; CO_2/H_2O is the molar ratio.

Table 2. Oxygen and carbon isotope analyses of hydrothermal minerals from the 1300 level, Oriental vein and granite

Sample #	Mineral	$\delta^{18}O$	$\delta^{13}C$	Remarks
OG-1[a]	quartz	19.8	--	euhedral clear crystals in veinlets
OG-1[a]	albite	16.2	--	euhedral clear crystals in veinlets
OG-1[a]	ankerite	17.0	-2.9	brownish (partially oxidized), mixed with silicates
OG-1[a]	rutile	3.5 (2)	--	2-3 mm long acicular crystals from vugs
527[a]	ankerite	17.1 (2)	-2.9 (2)	mixed with silicates
559[a]	albite	15.8	--	euhedral clear crystals in veinlets
356[a]	quartz	19.7	--	euhedral clear crystals in veinlets
356[a]	quartz	18.2	--	massive milky vein filling
356[a]	ankerite	15.7 (2)	-2.9 (2)	vein lining (on rutile?)
356[a]	rutile	3.2	--	vein lining (under ankerite?)
356[a]	calcite	21.0	-10.5	late vein filling
552[b]	quartz	18.2	--	veinlets in altered serpentinite
552[b]	dolomite	16.2 (2)	-1.5 (2)	veinlets in altered serpentinite (trace magnesite ± quartz?)

Note: [a] = 1300 level, veinlets and porous alteration zones in granite; [b] = 1300 level, Oriental vein, near west end; data followed by (2) are averages for two splits of the sample analysed separately; "ankerite" has the approximate composition $Ca_{.53}Mg_{.30}Fe_{.17}CO_3$; analysed by I. Barnes, M. Burns, and S. Hauser, using techniques summarized by Böhlke and Kistler (1986) for silicates and carbonates, and those of Borthwick and Harmon (1982) for rutile; yields of CO_2 from "ankerite", mixed with silicates, are not known, but the isotopic data may be assumed to be accurate to within ±1 per mil (see Becker and Clayton, 1972; Rosenbaum and Sheppard, 1986); $\delta^{18}O$ relative to SMOW; $\delta^{13}C$ relative to PDB.

Table 3. Averages and standard deviations of $\delta^{18}O$ and δD_{H_2O} for several categories of vein quartz (11 samples)

Quartz (group)	**$\delta^{18}O$ (SMOW)**	**δD_{H_2O} (SMOW)**
from gold ore pockets	18.0 ±0.6 (5)	-79 ± 10 (4)
from barren zones	18.0 ±0.8 (6)	-80 ±24 (7)
vug crystals	18.3 ±0.9 (4)	-54 ±14 (3)
massive milky quartz	17.7 ±0.3 (7)	-90 ±9 (8)

Note: compiled from data in Table 1.

Table 4. Oxygen and hydrogen isotope analyses of serpentinites from the 2300 level, Carrolson vein

Sample #	$\delta^{18}O$	δD	**Remarks**
RC-1	8.9	-107	from DDH at 35 feet; relatively unaltered
RC-2	8.7	-117	from DDH at 110 feet; relatively unaltered
O-23-511	--	-118	face of drift; relatively unaltered
O-23-509	8.7	-124	100 feet west of 507; relatively unaltered
O-23-507	11.4	-96	near ore pocket; minor talc veining
9-238-60	12.5	-63	near ore pocket, 50 feet below 2300 level; abundant hydrothermal talc (+magnesite?)

Note: analysed by R.O. Rye and J.F. Whelan, using techniques described by Rye and Rye (1974); $\delta^{18}O$ and δD relative to SMOW.

Table 5. Sulfur isotope analyses of sulfide minerals in veins and in altered wall rocks

Sample #	**Mineral**	**Yield (%)**	$\delta^{34}S$	**Remarks**
VEINS				
73-2-800	pyrite	92	-25.8	200 level; late crystal on arsenopyrite I[a]
73-2-14	pyrite	89	3.1	200 level
9-13-138	pyrite	91	2.3	1300 level
O-16-444	pyrite	98	1.7	1600 level
73-2-800	arsenopyrite I[a]	109	2.1	200 level
9-19-133	arsenopyrite I[a]	108	2.8	1900 level
9-197-132A	arsenopyrite I[a]	88	2.8	1970 level
9-197-132B	arsenopyrite II[a]	106	2.8	1970 level
9-237-189	arsenopyrite II[a]	107	3.2	2370 level
9-19-141	galena (with gold)	80	-0.8	1900 level
73-2-14	sphalerite	96	2.3	200 level; includes ~10% chalcopyrite
WALL ROCKS				
O-13-514	pyrite	91	1.6	1300 level; in altered granite
O-13-420	pyrite	93	2.3	1300 level; in altered granite
O-23-396	pyrite	80	2.0	2300 level
O-13-514	arsenopyrite II[a]	115	2.6	1300 level; in altered granite
O-13-420	arsenopyrite II[a]	109	3.1	1300 level; in altered granite
80W	arsenopyrite II[a]	107	3.2	1300 level; in altered granite
O-23-110	arsenopyrite III[a]	104	5.5	2300 level; near ore shoot

Note: [a] arsenopyrite types as defined by Coveney (1972): I= coarse (1-30cm) grains, usually with native gold inclusions (all three analysed specimens contained gold), II= fine (1-10mm) equant euhedral twinned crystals, III= fine (1-10mm) acicular crystals (higher As/S ratios, and possibly younger, than I and II); yields are given in per cent of sulfur expected for pure stoichiometric minerals; microprobe and X-ray diffraction analyses of arsenopyrites indicate that they are As-deficient (Coveney, 1981; Böhlke, unpublished data); therefore, a "yield" as high as ~110 per cent may be reasonable; analysed by R.O. Rye and J.F. Whelan, using techniques described by Rye and Rye (1974); $\delta^{34}S$ relative to CDT.

FIGURE CAPTIONS

Figure 1. Geologic sketch maps of (A) a portion of the Alleghany district, western Sierra Nevada metamorphic belt, California, and (B) the 1300 level of the Oriental mine, showing the locations of the Oriental and Carrolson veins. The elevations of the surface outcrop (heavy solid line) and the 1300 level (heavy dashed line) of the Oriental vein are indicated on the small scale map; dotted lines indicate portions of the underground workings; figure modified from Böhlke (1986).

Figure 2. Locations of samples collected underground in and near the Oriental and Carrolson veins, Oriental mine. The Carrolson vein (A) is approximately 50 meters north of the Oriental vein (B) at the 1300 level, and both veins dip toward the north (see Figure 1); the distribution of vein quartz on mined levels is shown in black, with thicknesses exaggerated for clarity; high grade gold ore shoots are indicated by asterisks whose relative sizes reflect relative production; the eastern limit of footwall serpentinite (sp) is shown on most levels; vein samples are designated by open circles; wall rock samples are designated by filled circles; see Ferguson and Gannett (1932), Coveney (1972a; 1981), and Böhlke (1986) for descriptions of the mine geology; the mine is currently flooded below the 1300 level.

Figure 3. δD_{H_2O} versus the mole ratio CO_2/H_2O of fluid inclusions released by crushing vein quartz. The data are listed in Table 1.

Figure 4. δD values for modern meteoric waters versus distance (due east) from the Pacific coast, at 38 to 40° north latitude. Data sources are: circles = Friedman et al. (1964), various surface waters (broken circles designate samples for which the coordinates given in the original reference do not appear to be consistent with the geographic locations of the samples); triangles = White et al. (1973), creeks and springs; crosses = Barnes et al. (1981), "soda springs"; lines = Ingraham and Taylor (1984), representation of trends for "ground waters" across northern California; parentheses enclose data for Clear Lake (Coast Ranges) and Lake Tahoe (Sierra Nevada), which appear to be shifted systematically to higher δD values than surrounding springs or streams.

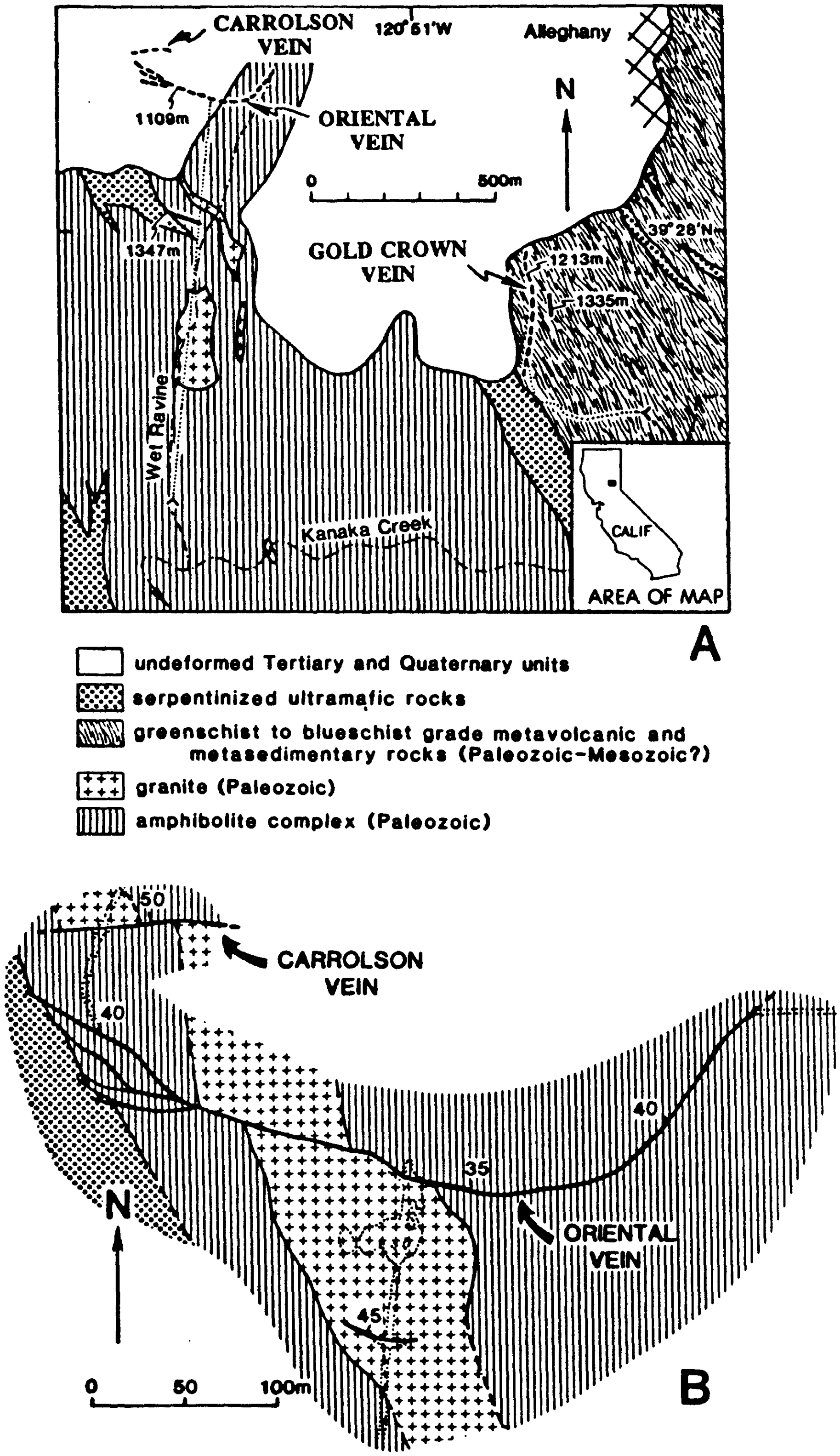
CARROLSON VEIN
120°51'W
Alleghany
1109m
ORIENTAL VEIN
N
0
500m
1347m
GOLD CROWN VEIN
1213m
1335m
39°28'N
Wet Ravine
Kanaka Creek
CALIF
AREA OF MAP
A
undeformed Tertiary and Quaternary units
serpentinized ultramafic rocks
greenschist to blueschist grade metavolcanic and metasedimentary rocks (Paleozoic-Mesozoic?)
granite (Paleozoic)
amphibolite complex (Paleozoic)
50
CARROLSON VEIN
40
40
35
N
ORIENTAL VEIN
45
0
50
100m
B

Figure 1.

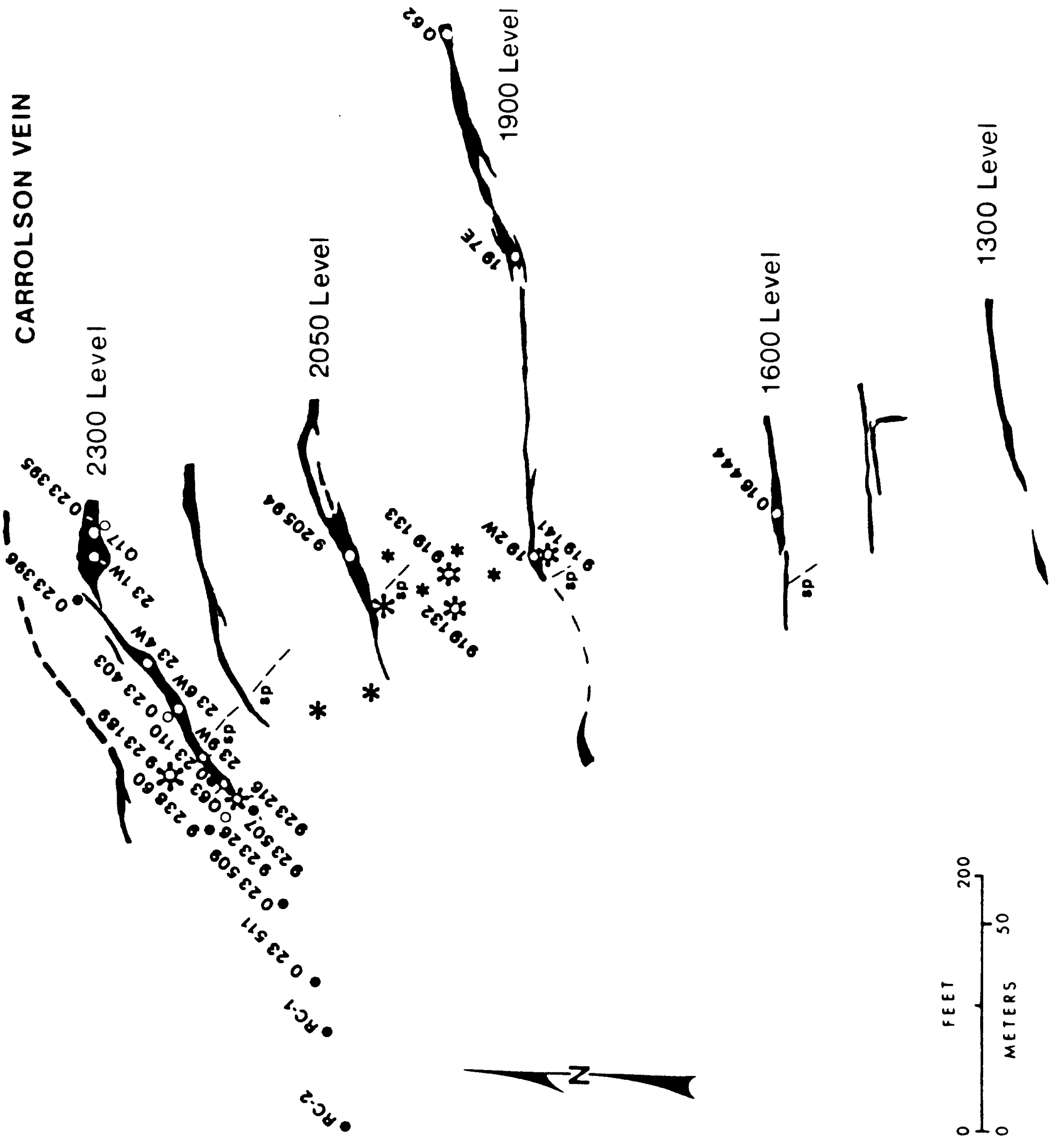
CARROLSON VEIN
2300 Level
2050 Level
1900 Level
1600 Level
1300 Level
023395
017
23 1W
023396
23 4W
23 6W
023 403
23 9W
023 110
923 189
083
923 860
923 216
923 507
923 26
923 509
023 511
RC-1
RC-2
920594
919 133
919 132
19 2W
919 141
19 7E
062
016444
ds
N
FEET
METERS
0
50
200

Figure 2A.

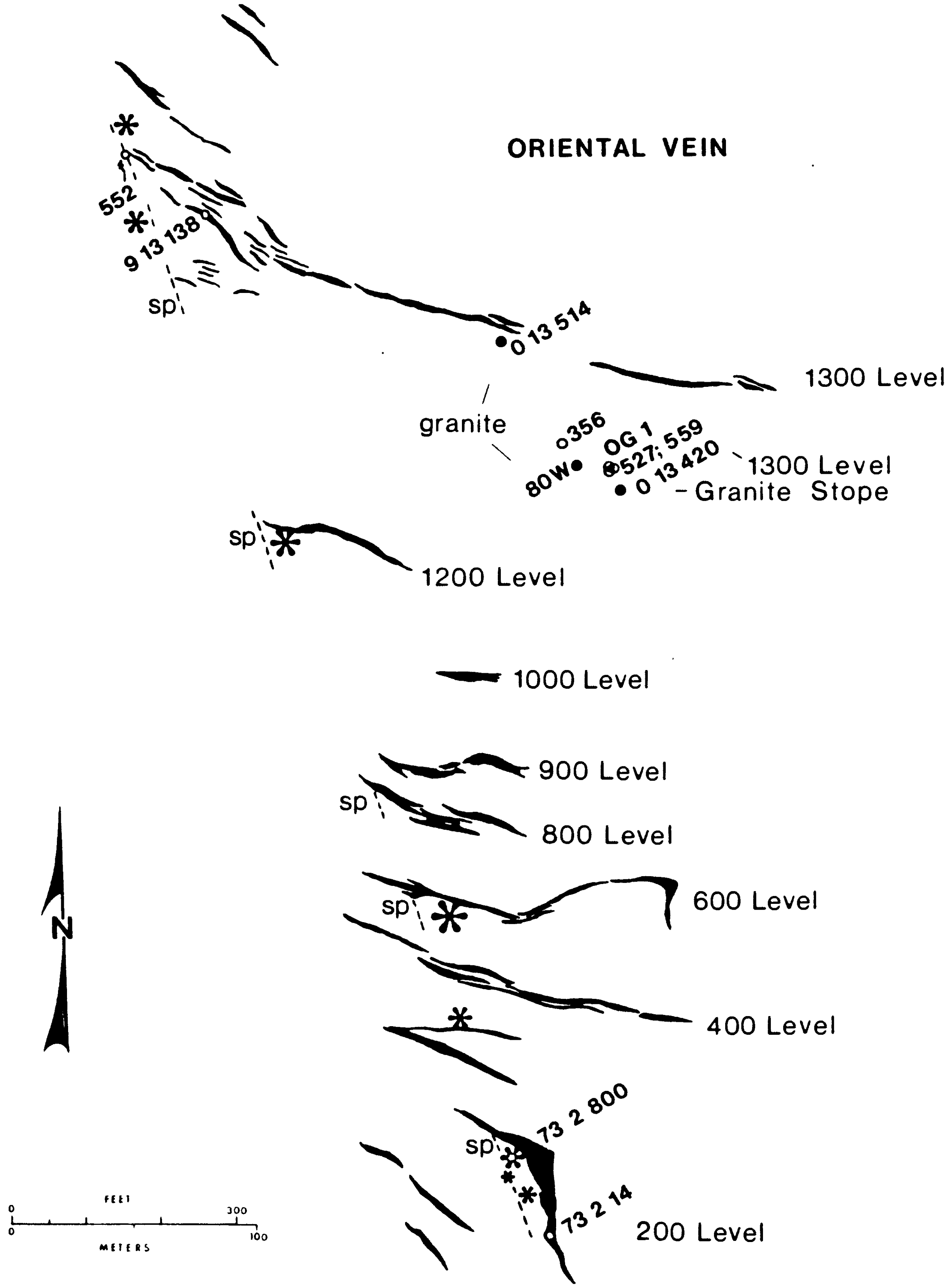
ORIENTAL VEIN
552
9 13 138
sp
0 13 514
1300 Level
granite
o356
OG 1
527; 559
0 13 420
80W
1300 Level
- Granite Stope
sp
1200 Level
1000 Level
900 Level
sp
800 Level
sp
600 Level
400 Level
73 2 800
sp
73 2 14
200 Level
N
FEET
0
300
0
100
METERS

Figure 2B.

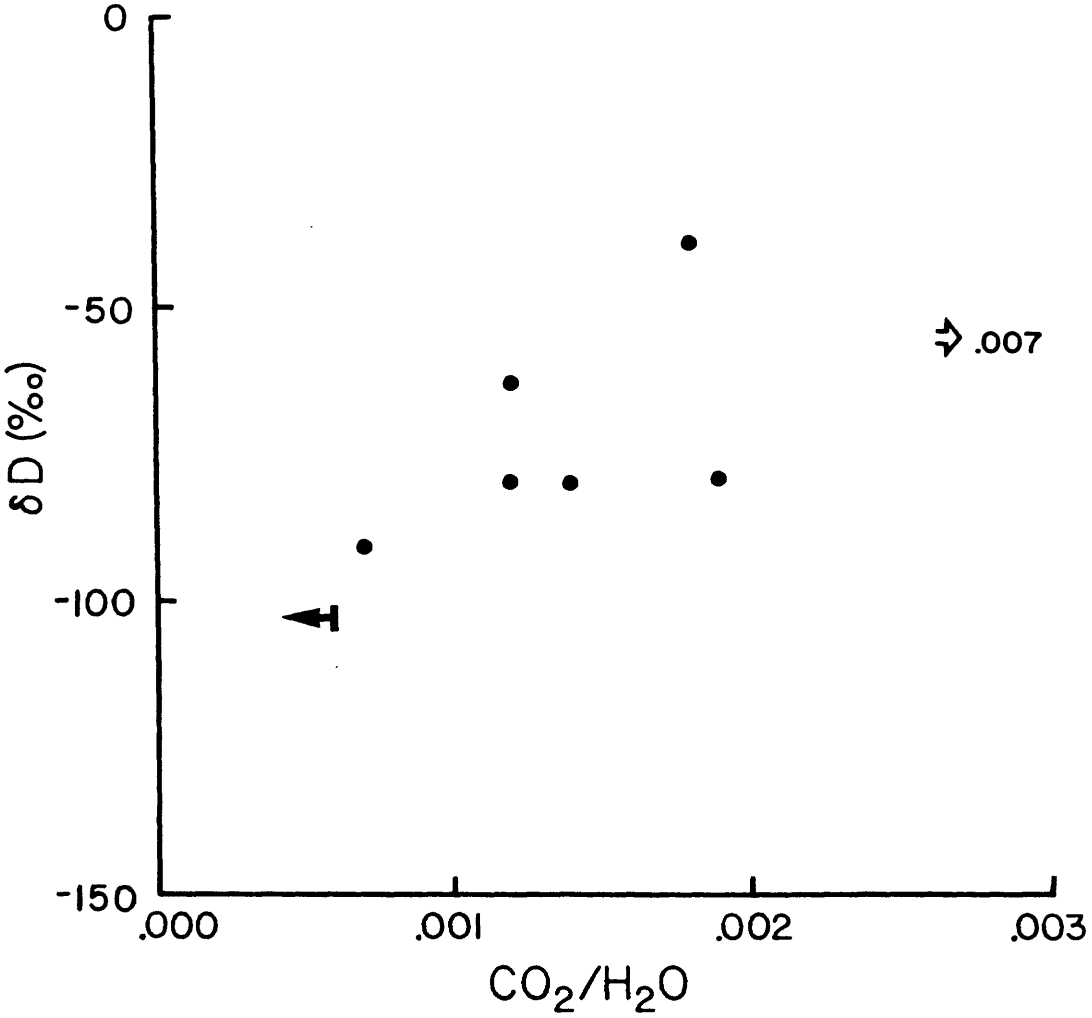

Figure 3

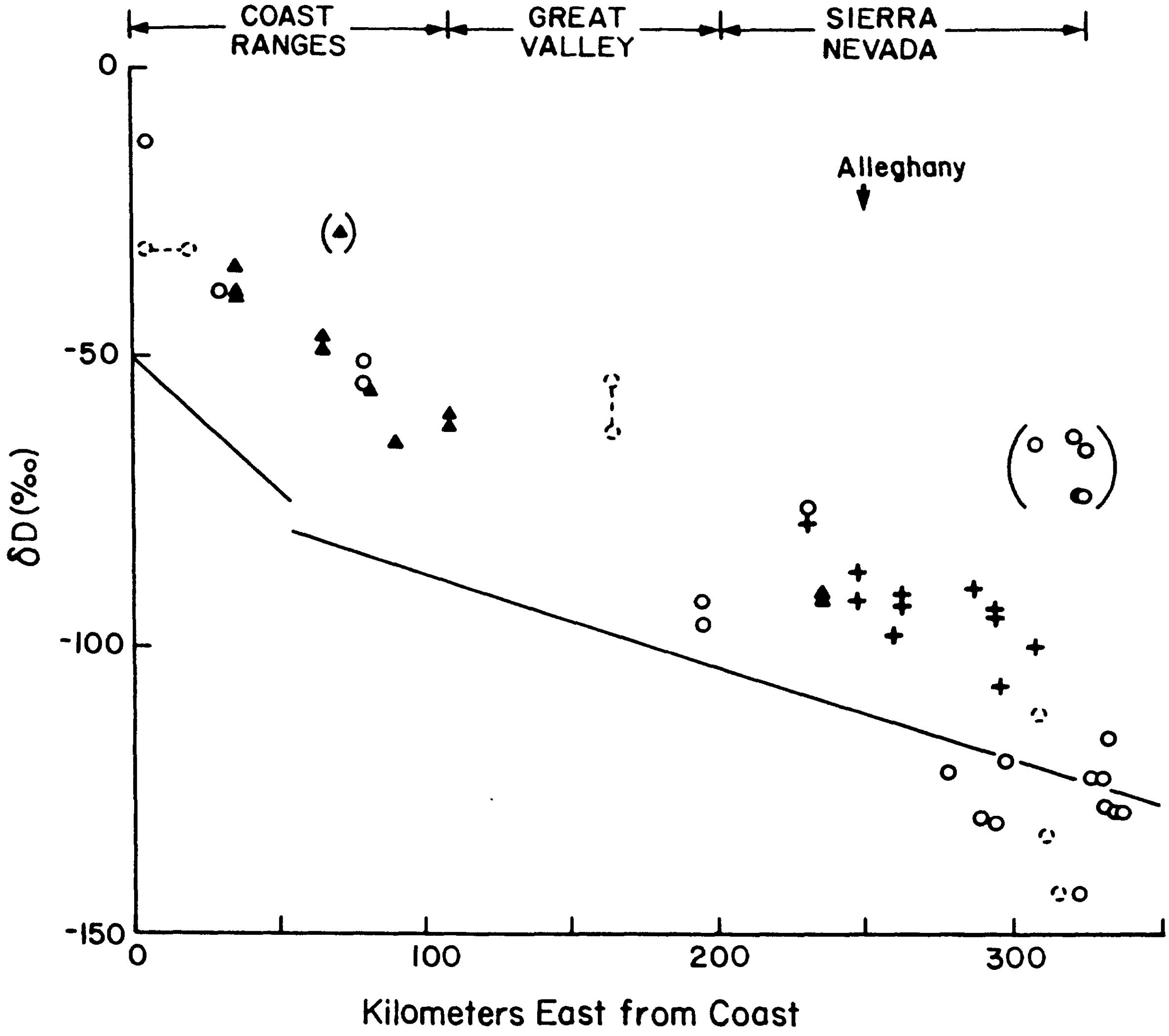
COAST RANGES
GREAT VALLEY
SIERRA NEVADA
Alleghany
δD (‰)
0
-50
-100
-150
0
100
200
300
Kilometers East from Coast

Figure 4