[AMERICAN JOURNAL OF SCIENCE, VOL. 287, SEPTEMBER, 1987, P. 720–755]

HYDROTHERMAL ALTERATION AND FLUID GEOCHEMISTRY OF THE MEAGER MOUNTAIN GEOTHERMAL SYSTEM, BRITISH COLUMBIA

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ABSTRACT. Drilling for geothermal fluids at Meager Mountain, in southwestern Canada, has provided an opportunity to study hydrothermal processes and fluid flow beneath an active stratovolcano. Drill holes have encountered temperatures as high as 264°C in altered crystalline basement rocks which act as the geothermal reservoir. Petrographic, mineralogic, and trace element studies have been used to establish the paragenetic relationships among the several thermal events that have affected these rocks. These relationships indicate that fault and fracture zones, steeply dipping dikes, and hydrothermal breccias related to recent volcanic activity have focused the upward movement of the geothermal fluids.

Four chemically distinct groups of thermal fluids occur at Meager Mountain. Three are NaCl in character and are associated with a well-defined thermal anomaly on the southern flank of the volcano. The fourth group consists of $NaHCO_3(-SO_4)$ fluids which represent steam-heated groundwaters. The fluids range from low-temperature and very saline with moderate isotope shifts to high-temperature and moderately saline with large isotope shifts.

The chemical and isotopic compositions of the NaCl waters show that little mixing of the different fluid types has occurred. In contrast, extensive fluid mixing is a common feature of highly productive geothermal systems in other volcanic terrains. In these systems, mixing is generally believed to result from convectively driven fluid flow in rocks with high permeabilities. We suggest that the lack of mixing at Meager Mountain reflects fluid flow through a few discrete fracture zones in low permeability rocks and that fluid movement is driven by topographically-controlled head differences. Calculated water/rock weight ratios, based on oxygen and deuterium isotopic shifts of the reservoir fluid, range from 0.005 at 123°C to 0.022 at 126°C.

INTRODUCTION

During the past several decades many productive geothermal systems associated with active volcanic centers have been studied in detail (see, for example, Browne, 1978; White and Guffanti, 1979; Henley and Ellis, 1983). Features common to these systems include widespread surficial activity and extensive subsurface mixing of the geothermal fluids. These features are generally believed to result from

convectively driven fluid movement through rocks characterized by high permeabilities (Studt, 1958; Mahon, Klyen, and Rhode, 1980).

Most volcanic centers in the mountainous terrain of the Cascade Range and its northern extension, the Garibaldi Volcanic Belt of Canada, display little surficial evidence of large, high-temperature geothermal systems. Despite the lack of surficial activity, temperatures near 265°C have been encountered at both Newberry Volcano and Meager Mountain (Sammel, 1981; Souther, 1985). Meager Mountain, the subject of this study, is located 150 km north of Vancouver at the northern end of the Garibaldi Volcanic Belt (fig. 1). Since 1973 twenty-five diamond drill holes, ranging in depth from 45 to 1297 m, and three rotary holes, ranging in depth from 3000 to 3500 m, have been drilled at Meager Mountain to test its geothermal potential (Reader and Fairbank, 1983; Moore, Adams, and Stauder, 1985).

Various aspects of the geothermal system at Meager Mountain have been described by Hammerstrom and Brown (1978), Shore (1978), Fairbanks and others (1981), Clark and others (1982), Moore, Adams, and Stauder (1983, 1985), Reader and Fairbank (1983), Adams, Moore, and Forster (1985), and Ghomshei, Croft, and Stauder (1986). These studies show that, in contrast to most geothermal systems in volcanic terrains, there is no correlation between the salinity, isotopic composition, and temperature of the thermal fluids at Meager Mountain. However, no detailed model of this thermal system has yet been presented. In this paper we discuss the hydrothermal alteration and fluid geochemistry at Meager Mountain and present a conceptual model of the thermal system that explains its hydrothermal and chemical features.

ANALYTICAL METHODS

Mineral assemblages in approx 250 samples from six wells drilled on the southern flank of Meager Mountain (MC-1, -2, -3, M-7, -10, -12; refer to fig. 2) were studied using petrographic and X-ray diffraction techniques. Whole-rock X-ray diffraction patterns were prepared for many of the samples examined in thin section. The clay minerals (<5 micron fraction) were routinely separated from the bulk samples by sonic disaggregation of lightly crushed rock and X-rayed after air drying, vapor glycolation, and heating for 1 hr at 250° and 550°C. Mineral identifications were confirmed by electron microprobe analysis.

Isotopic analyses of carbonate minerals, whole-rock samples, and waters from these wells were performed at the University of Indiana by E. Ripley. The carbon and oxygen isotope compositions of the carbonate minerals were determined from analysis of CO_2 gas produced by reacting the samples with anhydrous phosphoric acid (McCrea, 1950). Where possible, carbonates were removed from the veins with a dental drill.

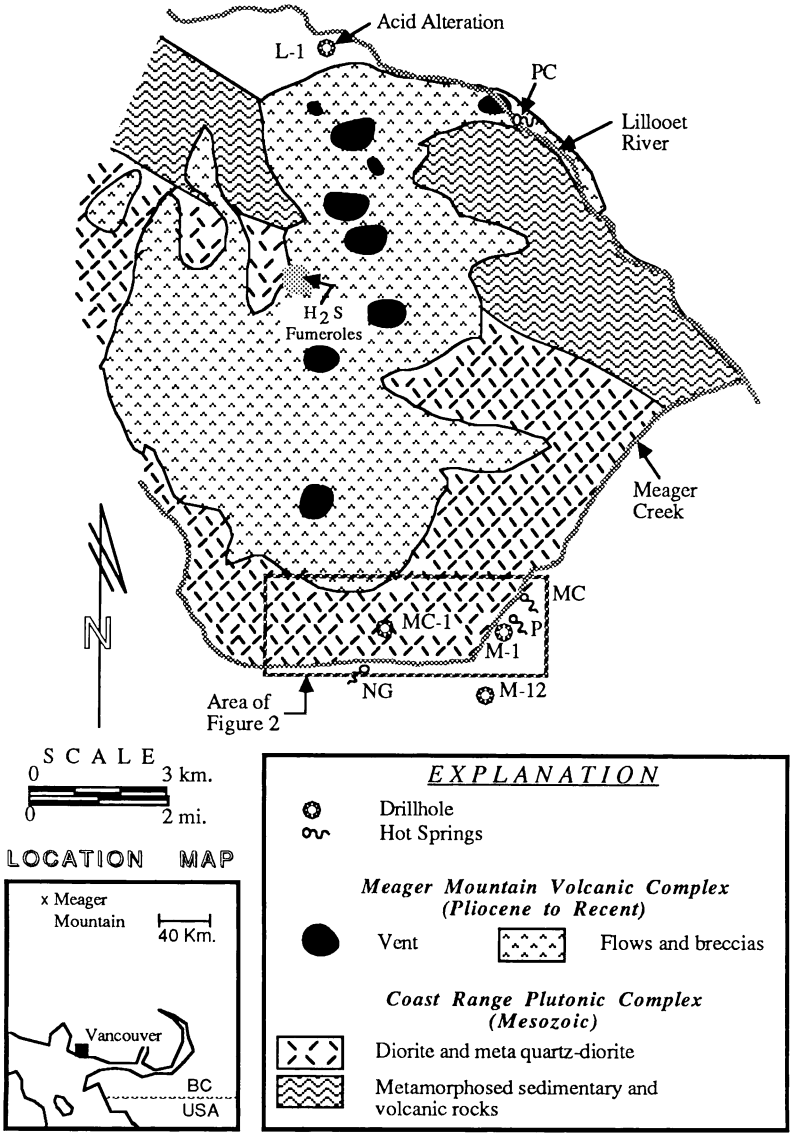


Fig. 1. Generalized geologic map of Meager Mountain showing the distribution and ages of the volcanic vents, thermal features, and fluid types associated with the geothermal system (modified from Read, 1979). Abbreviations: MC = Meager Creek Hot Springs, P = Placid Hot Springs, PC = Pebble Creek Hot Springs, and NG = No Good Hot Springs. The locations of wells L-1, MC-1, M-12, and M-1 are also shown. MC, P, and NG discharge low-TDS NaCl fluids; PC discharges low-TDS $NaHCO_3(-SO_4)$ fluids. High-TDS NaCl fluids flow from wells M-12 and M-1, and moderate-TDS NaCl fluid flows from the deep production well MC-1.

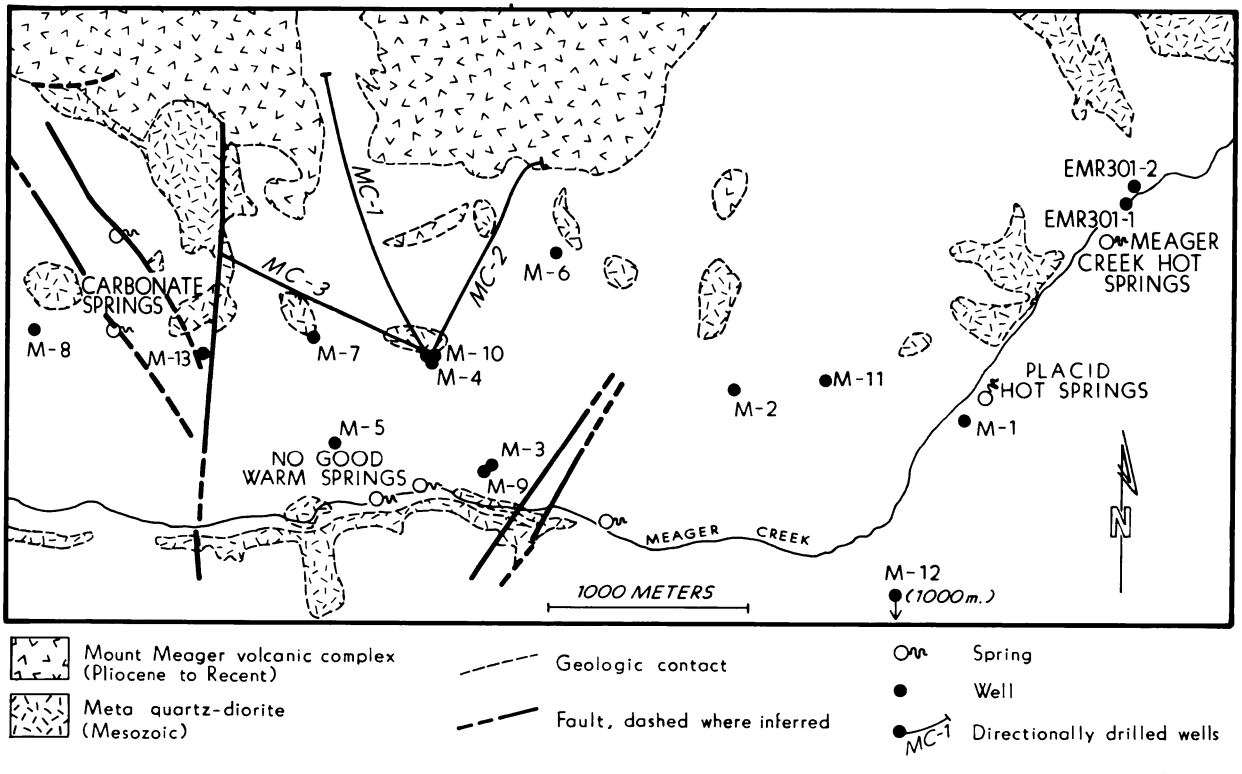


Fig. 2. Detailed map showing the geology and the location of the drill holes and hot springs on the southern flank of Meager Mountain (from Nevin, Sadlier-Brown, and Goodbrande, unpub. map). Wells MC-1, -2, and -3 were directionally drilled and are shown in plan view. Depths given in the text are vertical depths. All other wells were drilled vertically. Surficial deposits are shown in white. Refer to figure 1 for map location.

Hydrogen in the whole-rock samples was extracted by melting the samples under vacuum at 1200°C. The H_2O was collected, purified, and reduced with zinc at 450°C in sealed pyrex tubes to liberate hydrogen. Oxygen was liberated with BrF_5 and converted to CO_2 by reaction with graphite.

Oxygen in the water samples was analyzed by allowing the samples to equilibrate with CO_2 at 50°C. Hydrogen was liberated from these samples by reduction with zinc. Analysis of CO_2 was performed on a 6″ 60° Nuclide mass spectrometer. A 3″ 60° Nuclide mass spectrometer was used for the hydrogen analyses. Duplicate analyses were performed on all samples. The isotopic ratios are reported in the standard δ notation relative to SMOW for oxygen and deuterium and to PDB for carbon.

In order to determine the compositional changes that occurred during hydrothermal alteration of the reservoir rocks, samples from core holes M-7, -8, -9, -10, -12, and -13 (refer to fig. 2) were analyzed for 36 major, minor, and trace elements. With the exception of As and Hg, these analyses were made by inductively coupled argon plasma spectrometry using the techniques and instrumentation described by Christensen and others (1980). Arsenic was determined colorimetrically, and Hg by gold-film detector. In addition, the major and minor element contents of several representative samples of the cyrstalline basement rocks were determined. SiO_2 was analyzed by colorimetric techniques.

GEOLOGIC SETTING

The geology and age relationships of the rocks at Meager Mountain have been described by Read (1979). A generalized geologic map of the area is shown in figure 1. Figure 2 illustrates the distribution of rock types and the location of the geothermal wells on the southern flank of Meager Mountain where most of the exploration efforts have been concentrated.

Meager Mountain rests on a heterogeneous sequence of metamorphic and intrusive rocks of the Mesozoic Coast Range Plutonic Complex. Metasedimentary and metavolcanic rocks of the Triassic Cadwallader and Cretaceous Gambier Groups are widely exposed on the northern flank of the volcano. On the southern side, amphibolite and biotite-muscovite schist occur as lenses in variably foliated meta quartz-diorite of Cretaceous age. During the Cretaceous, the metamorphic rocks were intruded by stocks of intermediate composition. Biotite from a quartz-diorite exposed on the eastern flank of the volcano has a K-Ar age of 82 ± 3 my (Read, 1979).

The crystalline rocks encountered in the geothermal wells on the southern flank of Meager Mountain consist primarily of meta quartz-diorite, minor amphibolite, and granodiorite. The meta quartz-diorite is the host for the high-temperature geothermal fluids produced by MC-1. The average composition of three representative samples of meta

quartz-diorite from M-12 is presented in table 1. The rocks consist of andesine, quartz, hornblende, biotite, epidote, sphene, and opaques.

Granodiorite is the dominant lithology below 1600 m in MC-1 and in the lower 500 m of MC-3. The intrusive consists of a porphyritic phase adjacent to the contact and an equigranular phase at greater depths. Chemical analyses of these intrusive rocks are given in table 1. The porphyritic phase consists of phenocrysts of plagioclase, biotite, and quartz in a matrix that also contains potassium feldspar. The equigranular phase, in addition, contains hornblende, which is approximately equal to biotite in abundance. Although the age of the grandiorite has not yet been established, fission-track and alteration data discussed below suggest that this intrusive may be correlative with the Cretaceous quartz-diorite exposed on the east side of Meager Mountain.

During the late Cenozoic, the basement complex was again intruded by stocks of intermediate composition. Intrusive rocks as young as 10.1 ± 0.6 my are exposed on the northern side of Meager Mountain (Read, 1979).

Eruption of the Meager Mountain Volcanic Complex began approx 1.9 my ago from vents located on the southern side of the volcano (Read, 1979). Meager Mountain is composed of at least nine overlapping volcanic sequences with a total volume of about 70 km^3 (Souther, 1985). The oldest sequences consist predominately of andesite flows and breccias, whereas the younger volcanic units are mainly dacite and rhyodacite flows and domes. K-Ar age dating indicates that during Pliocene to Recent time, the eruptive centers migrated northward across Meager Mountain. The most recent eruptions occurred on the

TABLE 1

Chemical analyses of representative reservoir rocks

	(1)	(2)	(3)
SiO_2	60.13	67.40	71.10
TiO_2	0.53	0.45	0.28
Al_2O_3	18.40	15.71	14.85
Fe_2O_3	5.55	3.53	2.21
MnO	0.13	0.07	0.03
MgO	2.05	1.67	0.95
CaO	6.58	3.71	2.07
Na_2O	4.47	4.57	4.13
K_2O	0.94	2.37	3.71
BaO	0.07	0.09	0.15
P_2O_5	0.23	0.15	0.11
LOI	1.27	0.35	0.68
TOTAL	100.35	100.07	100.27

Data in weight percent; total iron calculated as Fe_2O_3; LOI = loss on ignition. Sample 1: Meta quartz-diorite from M-12. Average of three samples. Sample 2: Coarse-grained granodiorite from MC-1; sample from a depth of 2253 m. Sample 3: Porphyritic granodiorite from MC-1; sample from a depth of 1710 m.

northern flank of the volcano and produced the Bridge River Ash 2440 ± 140 yrs ago (Nasmith, Mathews, and Ronse, 1967).

Hydrothermally altered dikes, similar in composition to the extrusive rocks at Meager Mountain (Fairbank and others, 1981; Read, unpub. data) were intersected in all the drill holes examined by us. The dikes are typically porphyritic and contain variably altered phenocrysts of plagioclase, quartz, biotite, and hornblende(?). In addition, dikes of fine- to medium-grained andesite and diabase were encountered in the lower portions of the deep wells.

Hydrothermal breccias containing dike fragments of the Meager Mountain Volcanic Complex were noted in M-10 (B. Fairbank, written commun.) and MC-2 (B. C. Hydro and Power Authority, unpub. report) during the initial logging of these wells. The breccias in M-10 occur between depths of 318 and 917 m. None of the breccias has an apparent thickness greater than 5.5 m. The fragments entrained in the breccias are variably rounded and range up to 60 cm across. Two core samples of the hydrothermal breccias from M-10 were examined as part of this study. One is from a relatively narrow breccia between 478.7 and 479.2 m. The second, taken from 620 m, is from a thin (0.7 m) hydrothermal breccia that separates the upper contact of a flow-banded dike from the adjacent meta quartz-diorite. A similar breccia occurs along the lower contact of the dike (B. Fairbank, written commun.). The core samples of these breccias contain fragments of the crystalline basement rocks, some of which are intensely sericitized, and fragments of highly altered porphyritic dikes in a matrix of comminuted country rock, quartz, carbonate, chlorite, and pyrite.

The breccia encountered in MC-2, at a depth of 1606 m, consists dominantly of altered andesite and meta-quartz diorite fragments in a matrix of epidote, actinolite, calcite, chlorite, and quartz (B. C. Hydro and Power Authority, unpub. rept). Similar mineral assemblages were observed during our examination of cuttings from an interval just above the breccia and are described below.

Hydrothermal breccias have been recognized in many active and fossil geothermal systems, and their formation is now generally ascribed to explosive boiling triggered by hydraulic or tectonic fracturing of sealed fractures (see for example, Grindley and Browne, 1976; Fournier, 1983; Hedenquist and Henley, 1985). At Meager Mountain, the close association between the dikes and hydrothermal breccias suggests these dikes may have provided both heat and volatiles to fluids trapped beneath sealed or low permeability zones, allowing pore pressures to become large enough to fracture the enclosing rocks. The resulting rubblized zones may have been important, but transient, channels that connected shallow parts of the thermal system to high-temperature zones at depth.

GENERAL ASPECTS OF THE GEOTHERMAL SYSTEM

Geology.—Geothermal exploration on the southern flank of the volcano has resulted in the discovery of a thermal anomaly centered

approximately on M-7 (Fairbank and others, 1981; refer to fig. 2). The temperature and lithologies encountered in six of the core holes drilled in this area provide an illustrative cross section of the shallow portions of the thermal anomaly and are presented in figure 3. The maximum temperatures recorded in the deep wells range from 233°C in MC-3 to 264°C in MC-2 (not shown in fig. 3). Despite the high rock temperatures encountered in these wells, only MC-1 produced fluids above 100°C. This well is capable of producing approx 20,000 kg of steam and water

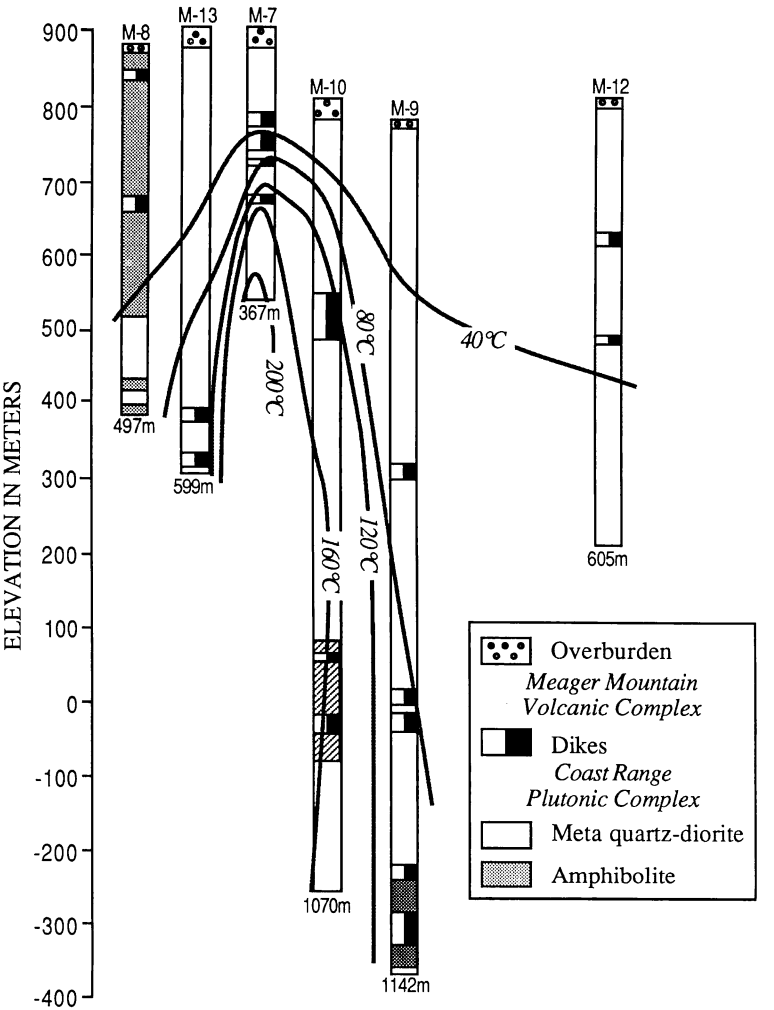


Fig. 3. Distribution of temperatures and lithologies encountered in the wells shown in figure 2. The horizontal distances are not to scale. Only the major dikes and dike swarms are shown.

per hour (Souther, 1985). The fluid enters the well at a depth of 1300 m with a temperature of 194°C.

The thermal anomaly appears to be located at the intersection of major north-and east-trending structural zones (Fairbank and others, 1981). The north-trending zone is defined by the alignment of volcanic vents across the volcano (fig. 1) and may be represented within the thermal anomaly by faults that offset the volcanic and basement rocks. The surface trace of the east-trending zone is approximately coincident with Meager Creek. Low electrical resistivities (Shore, 1978) near Meager Creek and fractured rocks exposed along the creek and in M-9, -10, MC-1, and -2 suggest that the fault zone dips approx 40° to the north (see also Reader and Fairbank, 1983). Thus, the fluids produced by MC-1 could have migrated up-dip along the Meager Creek fault zone from a deeper source area located north of the shallow thermal anomaly (Fairbank and others, 1981).

Fluid geochemistry.—Hot springs and fumaroles presently discharge thermal fluids from several locations around Meager Mountain (figs. 1 and 2). Meager Creek Hot Springs is the most significant of these springs, discharging approx 40 l/sec of 45° to 56°C water from 21 individual vents (Lewis and Souther, 1978; Clark and others, 1982). Smaller quantities of chemically similar warm water also discharge from two other groups of hot springs located 1 km (Placid Hot Springs) and 6 km (No Good Hot Springs) upstream from Meager Creek Hot Springs. These hot spring fluids have temperatures of 45°C and 20° to 40°C, respectively. Pebble Creek Hot Springs, located on the northern side of Meager Mountain, discharges 55° to 60°C water at a rate of approx 10 l/sec. Fumarolic activity at Meager Mountain is restricted to an area on the upper flank of the volcano above Pebble Creek Hot Springs.

The thermal fluids are diverse in their chemical and isotopic compositions. Chemical and isotopic analyses of waters from the Meager Mountain area are presented in table 2. For comparison, the ion ratios of these waters are shown on a Piper plot in figure 4. The waters can be divided into four groups based on their compositions. Three of the groups are NaCl in character and include: low-TDS fluids with slight oxygen isotope shifts; a moderately high-TDS fluid with a strong isotope shift (MC-1 discharge); and high-TDS fluids with moderate isotope shifts. The fourth group consists of low-TDS, $NaHCO_3(-SO_4)$ fluids with slight isotope shifts. Acid-sulfate water typical of many high-temperature geothermal systems has not been reported at Meager Mountain. However, the presence of acid alteration and fumaroles that discharge H_2S makes the occurrence of such fluid likely.

The low-TDS NaCl fluids discharge from the hot springs along Meager Creek and from EMR-1, drilled at the site of the Meager Creek Hot Springs. These fluids have a TDS averaging about 2000 ppm and a maximum temperature of 56°C. The moderately high TDS, NaCl fluid, discharged by the deep well MC-1, has a post-flash TDS content of 4160 ppm. The high-TDS fluids are characterized by high Na/K ratios and

TABLE 2

Representative chemical and isotopic compositions of thermal fluids

	1	2		3	4
		Moderate TDS NaCl fluid			
	Low TDS NaCl fluid	Flashed	Reservoir	High TDS NaCl fluid	Low TDS $NaHCO_3$ ($-SO_4$) fluid
Na	435	1385	1132	3600	423
K	56	103	84	136	17
Ca	85	39	58	490	33
Mg	26	0.8	0.65	240	5
SiO_2	189	326	266	164	85
B	3.3	9.0	7.35	38	2.4
Li	1.2	3.19	2.61	7.6	0.7
HCO_3*	330	85	2660	3535	724
SO_4	152	117	96	1820	329
Cl	661	2130	1740	4230	82
F	0.3	2.1	1.72	0.32	8.8
TDS	1938	4160	4745	12463	1695
pH (20°C)	6.6	8.4	5.9**	7.1	6.9
Temp. (°C)	57	97	194	22	60
$\delta^{18}O$ (SMOW)	−16.4	−7.7	−8.7	−14.2	−19.6
δD (SMOW)	−126	−86	−91.7	−120	−141

Chemical concentrations in ppm, isotopic concentrations in permil. Sample 1: Meager Creek Hot Springs, chemical data from Michel and Fritz (1979), isotopic data from Clark (ms). Sample 2: fluid from MC-1, chemical and isotope data of this study. Reservoir composition calculated from gas and liquid analyses assuming calcite equilibria (see text). Sample 3: fluid from M-12, chemical data from B. C. Hydro (unpub. data), isotope data from Clark (ms). Sample 4: Pebble Creek Hot Springs, chemical data from Michel and Fritz (1979), isotope data from Clark (ms).

*Total alkalinity including all titratable species, except for calculated reservoir composition where HCO_3 represents total CO_2.

**Calculated pH at 194°C.

TDS contents that range from 4600 to over 12000 ppm. These fluids discharge from two wells, M-1 and -12, and several springs. M-12 and the high-TDS springs are located along South Fork Creek, south of Meager Creek. M-1 is located along Meager Creek near Placid Hot Springs (see fig. 2). The TDS content and temperature of the fluid from M-12 are 12463 ppm and 22°C while those of M-1 are 9009 ppm and 56°C, respectively. The springs have TDS contents ranging from 4600 to 7000 ppm. The $NaHCO_3(-SO_4)$ fluids are confined to Pebble Creek Hot Springs. Their TDS content average 1680 ppm.

The composition of the fluid discharged from the deep well, MC-1, reflects the effects of boiling from the downhole temperature of 194°C to an ambient atmospheric boiling temperature. The fluid composition was also modified by deposition of calcite in the wellbore. The pre-flash reservoir fluid composition can be reconstructed by combining analyses of steam and gas samples taken at the wellhead with an analysis of the flashed fluid. The fraction of fluid lost to the steam phase at the wellhead, 0.18, was calculated using the steam tables of Keenan and others (1969). The H_2S and CO_2 concentrations in the steam phase were

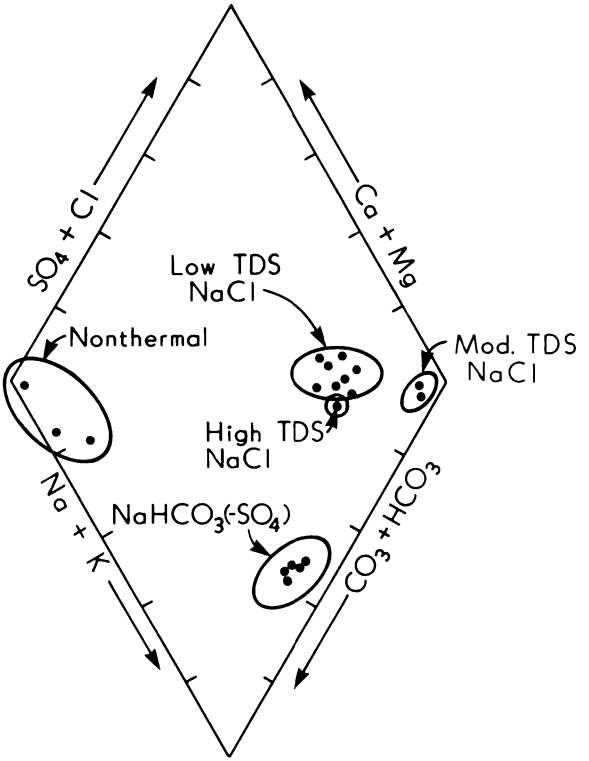


Fig. 4. Piper plot of waters from the Meager Mountain area.

1.9 and 262 mmol/kg of steam, respectively (C. Haway, personal commun.).

The reconstructed composition of the deep fluid, listed in table 2, is corrected for concentration, gas depletion, and calcite deposition due to boiling. Gas depletion of a fluid during boiling causes an increase in pH and a subsequent precipitation of calcite. The pH of the reservoir fluid, 5.9, was calculated using the programs PH (Henley and others, 1984) and PHREEQE (Parkhurst and others, 1980) combined with a WATEQ2 data base (Ball, Jenne, and Nordstrom, 1979).

The subsurface temperatures of the thermal fluids can be estimated using the Na–K–Ca(–Mg) and silica geothermometers (Fournier and Truesdell, 1973; Fournier and Potter, 1979; Fournier, 1981). These geothermometers are the most dependable and widely used, especially at low temperatures. The calculated subsurface temperatures of the low-TDS NaCl fluids, based on the Na–K–Ca(–Mg) geothermometer, range from 50° to 70°C. In most cases, similar temperatures are obtained using the amorphous silica geothermometer. However, a

few of the springs are characterized by both low silica and low TDS contents. These springs have amorphous silica geothermometer temperatures lower than those calculated from the cation geothermometer. These relationships indicate that all the fluids in the low-TDS NaCl group are in equilibrium with amorphous silica at depth.

The high-TDS NaCl fluids have calculated Na–K–Ca(–Mg) temperatures ranging from 29° to 71°C and calculated amorphous silica geothermometer temperatures ranging from 21° to 44°C. The discrepancy between the ranges of predicted temperatures may be due to the high ratio of Mg to K and Ca in the fluids. This high ratio is above the upper end of the data used to calibrate the Na–K–Ca(–Mg) geothermometer (Fournier and Potter, 1979). The moderately high TDS, NaCl fluid from the production well has a Na–K–Ca(–Mg) temperature of 190°C, which is in accord with both the measured temperature and the calculated quartz geothermometer temperature of 200°C.

The calculated Na–K–Ca(–Mg) temperatures of the Na-HCO_3(–SO_4) fluids from Pebble Hot Springs indicate subsurface temperatures of 96° to 101°C. The chalcedony geothermometer also yields temperatures of 96° to 101°C. Chalcedony is a common equilibrium polymorph for thermal fluids at these temperatures in volcanic rocks (Fournier, 1981). However, as discussed in a later section, Na-HCO_3(–SO_4) fluids may be reactive and thus out of equilibrium. Thus the concordance of geothermometer temperatures may be coincidental.

The deuterium and oxygen isotope compositions of the thermal waters are shown in figure 5. For comparison, the local meteoric water line (Clark, ms) is also shown.

HYDROTHERMAL ALTERATION

The secondary mineral assemblages at Meager Mountain reflect the effects of multiple thermal events. Complex mineral relationships are particularly apparent in the Mesozoic rocks that have been affected by regional metamorphism and intrusion during Mesozoic and Cenozoic time, as well as by geothermal activity associated with volcanism at Meager Mountain. In contrast, the hydrothermal mineral assemblages in the dikes and breccias are no older than 2 my and as such are the most clearly related to the modern geothermal system. The mineral assemblages in the dikes and crystalline rocks are discussed separately in the following paragraphs in order more clearly to distinguish pre-volcanic from post-volcanic alteration.

Secondary mineral assemblages in the dikes.—Sheet silicates are the most common secondary phases occurring in the dikes. The distribution of the sheet silicates in two representative high temperature wells, M-10 and MC-1, and in M-12, a low temperature well drilled on the margin of the thermal anomaly (refer to fig. 2), is illustrated in figure 6.

With the exception of the diabase dikes, the clay fraction of the dikes within the thermal anomaly consists of either smectite, chlorite,

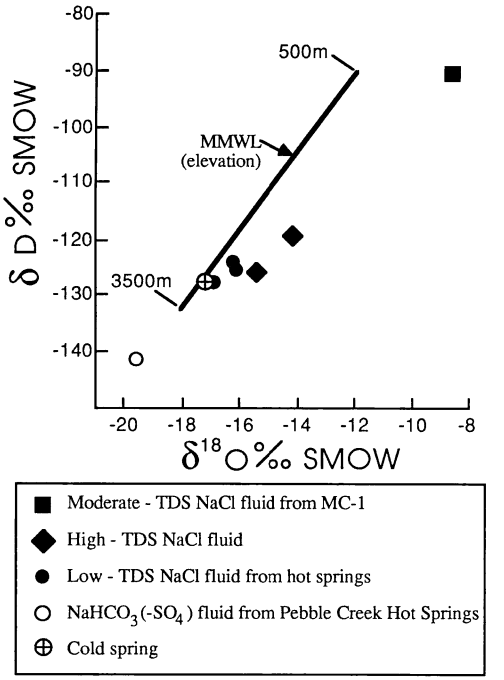


Fig. 5. δD-$\delta^{18}O$ relationships of thermal waters from Meager Mountain. The Meager Mountain meteoric water line (MMWL; Clark, ms) is shown with respect to elevation.

illite, and interlayered illite-smectite or chlorite and illite. Kaolinite was identified in one sample of a dike from the upper part of M-10. Illite occurs in the dikes as an alteration phase in the groundmass and as fine- to medium-grained aggregates in plagioclase and biotite phenocrysts. Chlorite replaces biotite, amphibole, and plagioclase phenocrysts.

Saponite and minor talc were identified optically in the diabase cuttings from MC-1, and their presence was confirmed by X-ray and partial electron microprobe analyses. These minerals typically occur as zoned, fine-grained birefringent aggregates consisting of a core of talc surrounded by faintly pleochroic brown saponite.

In contrast, smectite is the only secondary sheet silicate present in the dikes from M-12, where it occurs as a fine-grained alteration phase in the groundmass.

Calcite, quartz, and hematite are common alteration phases associated with the sheet silicates. Calcite is ubiquitous as a replacement of plagioclase and mafic minerals and as vesicle fillings. Quartz occurs in the matrix of the breccias, and hematite is found mainly as a replacement of iron-titanium oxides and mafic minerals.

Epidote, potassium feldspar, and actinolite are associated with chlorite in a hydrothermally altered andesite dike from a depth of 1625

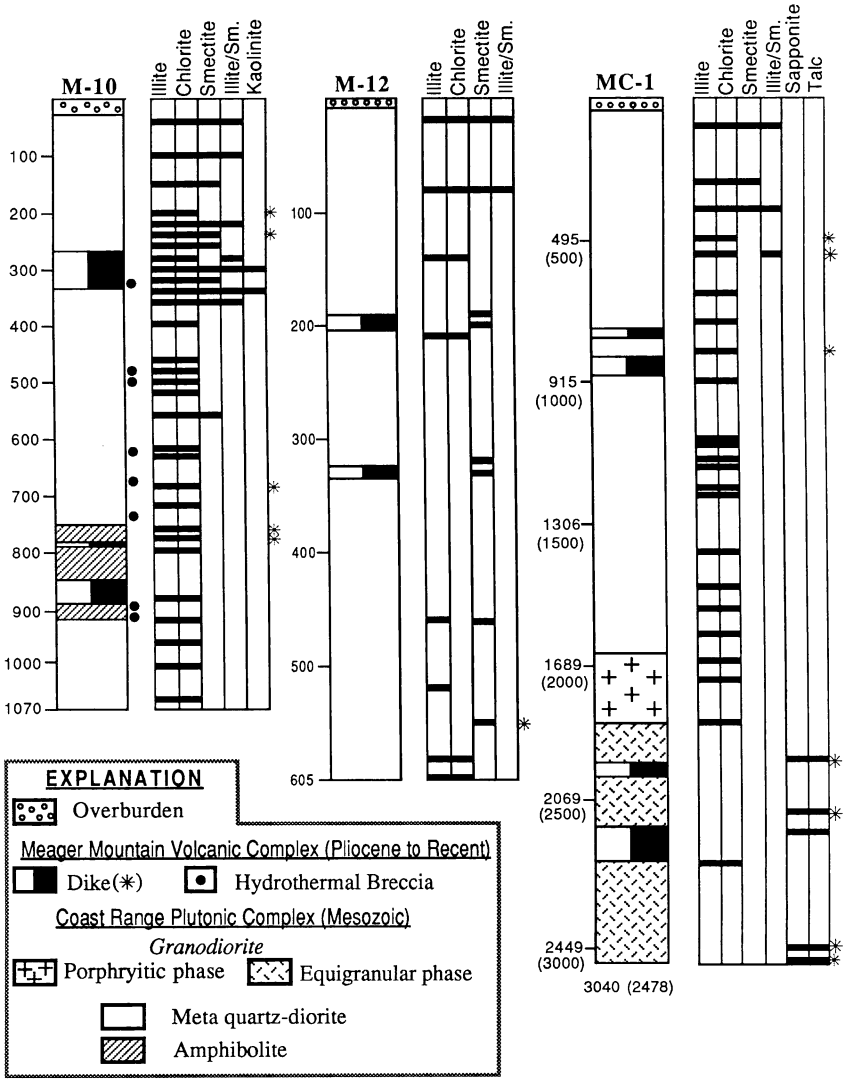


Fig. 6. Distribution of sheet silicates in M-12, -10, and MC-1. Illite/sm = interstratified illite-smectite. MC-1 was directionally drilled. Both the true vertical depth and the length along MC-1 (in parentheses) are shown. Dikes shown below a vertical depth of 1900 m in MC-1 are diabase. Samples from dikes that are too narrow to show on the figure are indicated by asterisks to the right of the alteration assemblage log.

m in MC-2. Similar secondary minerals were reported in the hydrothermal breccia penetrated immediately below this interval (B. C. Hydro and Power Authority, unpub. rept.). Epidote and potassium feldspar replace plagioclase, whereas pale green actinolite occurs as fine acicular crystals after hornblende. Chlorite is found as an interstitial phase between plagioclase laths, where it appears to represent replacement of glass.

Sulfides are sporadically distributed in the volcanic rocks. Pyrite occurs in the breccias as cubes up to several millimeters across and has been reported to occur in the dikes, particularly in M-10 (B. Fairbank, written commun.). P. Read (written commun.) has also reported the occurrence of sphalerite in one of the dikes from M-7.

Relatively few veins were observed in the dikes and hydrothermal breccias. The majority of the veins consist of carbonate ± quartz ± hematite; chlorite or barite are rarely present. Veins dominated by quartz or silicate minerals were encountered in three samples. Quartz stringers and veins of quartz + pyrite + carbonate + hematite cut one of the dikes in M-7 (180 m), and these minerals occur in the matrix of the hydrothermal breccia in M-10. Aggregates containing various proportions of epidote, potassium feldspar, actinolite, quartz, magnetite, and carbonate are associated with the altered andesite chips in MC-2. These aggregates are interpreted as vein fillings.

Secondary mineral assemblages in the crystalline basement rocks.—The Mesozoic metamorphic and intrusive rocks are characterized by pervasive propylitic alteration to chlorite, illite, Fe-Ti oxides, epidote, and carbonate. The intensity of alteration ranges from strong in the meta quartz-diorite within the thermal anomaly to weak to moderate in the granodiorite and in the meta quartz-diorite outside the thermal anomaly (M-12). X-ray analysis of the clay fraction of the crystalline rocks indicates that smectite and interlayered illite-smectite and kaolinite are also present in the metamorphic rocks. The distribution of the sheet silicates in the three wells, M-10, -12, and MC-1 is illustrated in figure 6.

Millimeter thick veins and fracture coatings consisting dominantly of carbonate are common, particularly in the highly altered metamorphic rocks. These veins can be separated into two groups on the basis of their mineralogies. One group is characterized by the simple assemblage carbonate ± hematite. Carbonate + chlorite ± magnetite/hematite ± quartz ± pyrite characterize the second group. The carbonate + chlorite veins occur at depths as shallow as 40 m in M-7. Many display evidence of brecciation during their formation, a feature not associated with the carbonate veins occurring in the dikes. Crosscutting relationships among the carbonate veins are, however, often complex and indicate that the carbonate ± hematite veins both predate and postdate carbonate + chlorite veins. Less commonly, quartz veins or veins consisting dominantly of silicate minerals (epidote, chlorite, potassium feldspar, actinolite, biotite), and containing minor carbonate, pyrite, magnetite, anhydrite, or rarely base-metal sulfides (sphalerite, galena,

chalcopyrite, bornite, molybdenite) occur in the metamorphic rocks. Crosscutting relationships between the carbonate, quartz, and silicate veins are not commonly observed. However, when these relationships are found, the carbonate veins typically can be shown to be younger than the quartz veins.

Both carbonate and silicate-rich veins occur in the metamorphic rocks in MC-2. Carbonate veining and alteration is found primarily in the upper 1800 m of the well. At greater depths, carbonate is relatively uncommon. Actinolite, quartz, potassium feldspar, epidote, and biotite occur as aggregates in the chips and in veins below 1700 m. Veins containing potassium feldspar and quartz, but lacking actinolite and biotite, are more broadly distributed and occur at depths below 850 m in MC-2.

Additional insight into the paragenesis of the veins in MC-2 is provided by core taken from a depth of 2005 m. This sample is a propylitically altered meta quartz-diorite. Textural relationships suggest that the youngest phase of hydrothermal activity in the sample is represented by a vein consisting dominantly of fine-grained quartz and minor secondary biotite, epidote, and actinolite. Crosscutting relationships show that the wall rock along the vein was intensely altered to biotite, chlorite, and epidote and in places brecciated prior to emplacement of the quartz vein. The matrix of this microbreccia consists predominately of biotite.

The alteration assemblages in the crystalline rocks appear to define at least two different hydrothermal events. The first event occurred prior to volcanic activity and is most clearly represented by the intense propylitic alteration in some of the basement rocks. The second is post-volcanic (<2 my) in age and is represented in most of the crystalline rocks by the sheet silicates that characterize the altered dikes. In MC-2, the similarity between the secondary assemblages in the altered andesite and those in the crystalline rocks indicates that post-volcanic alteration has produced actinolite and biotite-bearing assemblages in this well. However, as discussed below, the propylitic assemblages in the altered rocks from the lower portions of the deep wells are consistent with temperatures during both pre- and post-volcanic events and thus could have been produced by either hydrothermal system.

A minimum age of the propylitic alteration and the associated carbonate + chlorite veins can be deduced from the fission-track data of Kelley and Blackwell (1984) combined with fluid inclusion data from M-7. Rocks in this well are characterized by intense propylitic alteration and widespread carbonate veining. Zircon fission-track ages from the upper 200 m of this well and from M-12 indicate these rocks have not been heated above about 200° ± 25°C for any significant length of time during the last 60 to 80 my. Primary fluid inclusions in calcite from a carbonate + chlorite vein sampled at 190 m and from secondary inclusions in quartz adjacent to the vein homogenize at 245° to 250°C. Similarly, the widespread occurrence of secondary epidote, chlorite, and

illite in these rocks indicates that temperatures must have been at least 250°C during alteration (see discussion below). Thus, the propylitic alteration in the metamorphic rocks is almost certainly Cretaceous in age. The most likely cause of this alteration was intrusion of the granodiorite encountered in MC-1 and -3.

Trace element zoning.—Rocks in active geothermal systems frequently display chemical changes that can be correlated with temperature and the distribution of fluid flow paths (see, for example, Christensen, Capuano, and Moore, 1983). Because some of these chemical changes can be detected even before appreciable hydrothermal alteration has occurred, chemical data provide information that cannot be obtained from routine petrographic investigations. During this study, the composition of 270 rock samples taken at 10 or 20 m intervals from the six wells shown in figure 3 and the calcium carbonate discharge precipitates from M-1 and MC-1 were determined.

Chemical analyses of the calcium carbonate precipitates from the two wells provide a view of the trace element mobilities at two different levels within the thermal system. Elements available for deposition at intermediate depths are contained within the discharge deposits from M-1. A chemical analysis of these deposits is shown in table 3. The salient features of this analysis are the high concentrations of Sr, As, and Bi and the traces of Zn.

Discharge precipitates from MC-1 provide information on trace metal mobilities at greater depths. The discharge precipitates from this well are characterized by higher contents of Ag, Cu, Ni, and Zn

TABLE 3

Chemical composition of calcium carbonate discharge precipitates from M-1 and MC-1

	M-1		MC-1
	(1)	(2)	(3)
Fe_2O_3 (%)	6.9	0.8	3.9
Sr (%)	13.8	14.5	1.2
BaO	90	100	260
MnO	340	1140	950
Pb	24	<10	<10
Zn	58	15	132
As	8750	375	<25
Bi	<100	263	<100
Li	6	6	8
Be	4	<0.5	2
Hg (ppb)	7	17	n.a.

Analytic data in ppm unless otherwise noted. n.a. = not analyzed. Sample 1: red discharge precipitate from M-1. Sample 2: white discharge precipitate from M-1. Sample 3: precipitate from the Weir box of MC-1. Sample also contains 4 ppm Cr, 11 ppm Ni, 10 ppm Cu, and 115 ppm Ag. Another sample from MC-1 contains 208 ppm As and 145 ppm Sb.

compared to those from M-1 (table 3). Silver contents as high as 115 ppm were also measured in the precipitate from MC-1.

Only the dikes in M-7, -9, and -10 display any significant enrichments in these elements. Dikes from M-7 and -10 contain anomalous concentrations of Zn in addition to As and Hg. In contrast dikes in M-9 are enriched only in Hg and Zn. The distribution of these enrichments is shown in figure 7.

In contrast to the trace element distributions in the dikes, the meta quartz-diorite and associated amphibolite are characterized by enrichments in: (1) Hg + Cu + Zn + Pb + Ba + K; (2) Hg + Zn ± As; (3) Hg; and (4) Sr. With the exception of two samples enriched in Hg + Zn ± As, the distributions of the trace element signatures in the metamorphic rocks bear no apparent relationship to the present thermal anomaly or to those of the dikes. Instead, these enrichments appear to correlate most closely with the degree of propylitic alteration and the extent of carbonate veining (Moore, Adams, and Stauder, 1983; Ward and others 1985). Similar trace element signatures have been described from copper porphyry deposits of the Guichon Creek Batholith of British Columbia, among others (Olade and Fletcher, 1976). Thus, we interpret these signatures as being the result of the high temperature alteration that affected the metamorphic rocks during the Cretaceous.

Concentrations of Hg ranging from 5 to 1000 ppb occur primarily in the metamorphic rocks in M-8, -13, the lowermost 150 m of M-9, and the interval between 750 and 1000 m in M-10 (Moore, Adams and Stauder, 1983). The absence of more extensive Hg enrichments within the thermal anomaly, particularly in the metamorphic rocks in M-7, -9, and -10 was not expected since Hg is extremely mobile in active geothermal systems (Matlick and Busek, 1976; Christensen, Capuano, and Moore, 1983). The limited distribution of Hg in these rocks is consistent with the lack of fluid production from these wells and suggests there is little connection between the deep fluid conduits and the metamorphic rocks in the upper 1000 m of the present thermal system.

Implications of hydrothermal alteration.—The maximum temperatures of the reservoir rocks during the last 2 my can be estimated by comparing the occurrence of hydrothermal minerals at Meager Mountain with those from other well-documented geothermal systems. Despite their generally spotty occurrence, the secondary minerals and trace elements are strongly zoned with respect to depth and temperature. Four distinct alteration zones can be distinguished. These zones are characterized by: (1) smectite, (2) interlayered illite-smectite, (3) illite and chlorite, and (4) epidote, actinolite, potassium feldspar, and biotite or saponite and talc. As in many other systems, the boundary between the sheet silicate zones is gradational (Cole and Ravinsky, 1984). The overlap between these zones is clearly illustrated in MC-1 and M-10 (refer to fig. 6). Smectite, interlayered illite-smectite, illite, and chlorite are all present in the altered dikes occurring at shallow levels in these wells. In contrast, only the lowest-temperature mineral

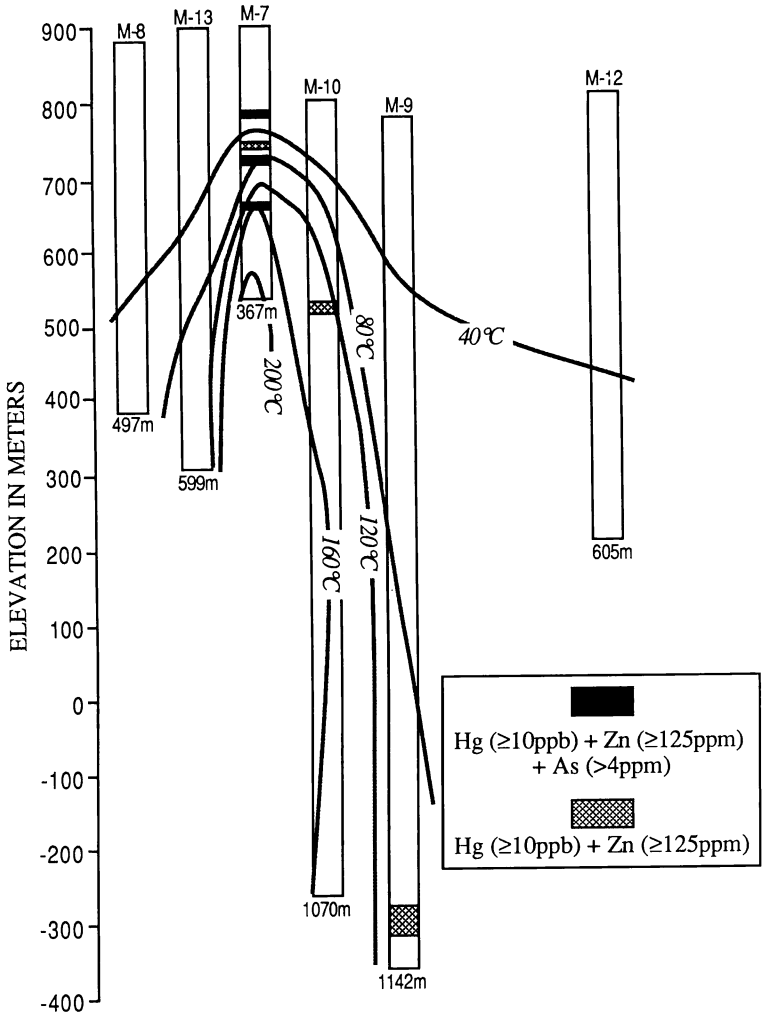


Fig. 7. Distribution of Hg, Zn, and As in altered rocks from the wells shown in figure 3. With only two exceptions, enrichments in these elements are restricted to the dikes. For reference the present-day temperatures are also shown.

zone is present in dikes from M-12. The pattern of sheet silicates at Meager Mountain is similar to that mapped in other high-temperature geothermal fields (Kristmannsdottir, 1976; Browne, 1978; McDowell and Elders, 1980; Cathelineau and others, 1983; Cole and Ravinsky, 1984; Hulen and Nielson, 1986). Despite wide variations in rock and fluid chemistry among high-temperature systems, most common geothermal minerals appear to be stable only within relatively narrow temperature ranges (for example, Henley and Ellis, 1983). In general,

dioctahedral smectite is stable from near-surface conditions (<50°C) to temperatures of approx 175°C. At Meager Mountain, dioctahedral smectite is common only where present temperatures are less than 160°C, although traces of this mineral do persist to higher temperatures. Interlayered illite-smectite occurs in most geothermal systems at temperatures ranging from slightly less than 150° to 225°C. Above 200°C, chlorite and illite are the common sheet silicates, although both chlorite and illite may appear as stable phases at temperatures as low as 150°C (Muffler and White, 1969; Elders, Hoagland, and McDowell, 1979).

The minerals occurring in the highest temperature zone (assemblage 4) at Meager Mountain are found only in modern geothermal systems that have been heated above 200°C. Epidote is found where temperatures have exceeded 200° to 250°C, whereas actinolite, biotite, and talc are characteristic of modern geothermal systems with temperatures above 300°C (McDowell and Elders, 1980; Bird and others, 1984).

Although the present-day temperatures are slightly lower than those deduced from mineral relationships, the trace element signatures of the dikes suggest that the center of the shallow anomaly (refer to fig. 3) has not changed position radically since alteration of the dikes occurred. Figure 7 shows that the trace element distributions in the dikes are crudely zoned with respect to the present temperatures. Dikes characterized by enrichments in Hg + Zn + As occur only in the highest temperature portions of the anomaly, whereas dikes enriched in Hg + Zn occur in slightly lower temperature regions. Furthermore, the trace element and alteration patterns of the rocks within the anomaly suggest that the dikes have been the primary conduits for upwelling fluids in the upper 1000 m of the thermal system.

Temperature differences between the wells are also reflected in the isotopic compositions of calcite in the dikes. Calcite occurs in veins and as a replacement of the phenocrysts and groundmass in the dikes in M-7 and -10, whereas it is found as vesicle fillings in M-12. Figure 8 shows that samples from M-12 are depleted in carbon-13 but enriched in oxygen-18 relative to the calcite from the other two wells. Since no data are currently available on the isotopic compositions of the fluids responsible for the alteration in these wells, the samples were evaluated with respect to the present-day fluids. The light carbon values from M-12 suggest that the calcite could have been deposited by fluids isotopically similar to some of the local cold spring waters or groundwaters. Calculations based on the isotopic contents of these waters yield temperatures of about 25°C. In contrast, the heavier carbon contents of calcite from the dikes in M-7 and -10 suggest these rocks were altered by fluids that could have been isotopically similar to the hot-spring waters. Calculations based on the isotopic composition of these fluids yield minimum temperatures of 50° to 100°C. The carbon and oxygen isotopic fractionation factors used in these calculations are those given

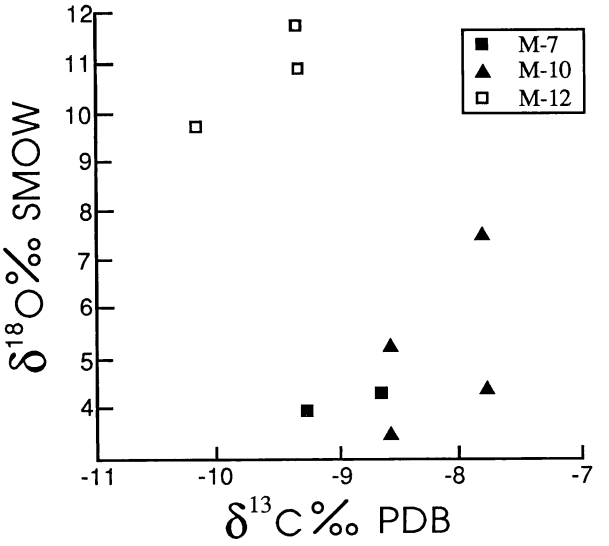


Fig. 8. Isotopic variations of carbon and oxygen in calcite from dikes in M-7, -10, and -12.

by Ohmoto and Rye (1979) and O'Neil, Clayton, and Mayeda (1969). The fluid equilibrium program WATEQ (Truesdell and Jones, 1974) was used to determine the carbon speciation in the waters, and the isotopic compositions of the Meager Mountain fluids were taken from Clark (1980).

ORIGIN OF FLUID VARIATIONS

Isotopic ratios of oxygen and hydrogen in geothermal systems show consistent patterns. As shown in figure 9, deep geothermal waters are generally shifted to higher $\delta^{18}O$ values with respect to regional meteoric waters, whereas the δD values of thermal and meteoric waters are similar. Variations in the isotopic compositions of the Meager Mountain fluids are also shown in figure 9. Dilute NaCl fluids from hot springs and shallow wells exhibit a slight $\delta^{18}O$ shift and little or no δD shift with respect to the average cold spring (fig. 5). $NaHCO_3(-SO_4)$ fluids from Pebble Creek Hot Springs, in contrast, are isotopically depleted in both oxygen-18 and deuterium with respect to the average cold spring composition. Moreover, the deep fluid from MC-1 is isotopically enriched in both oxygen-18 and deuterium with respect to all other thermal fluids sampled at Meager Mountain. We have examined each of the several possible causes for the isotopic variations at Meager Mountain. These are discussed below.

Isotopic variations in water due to varying elevation or latitude of the recharge source are known to occur in geothermal areas (Craig,

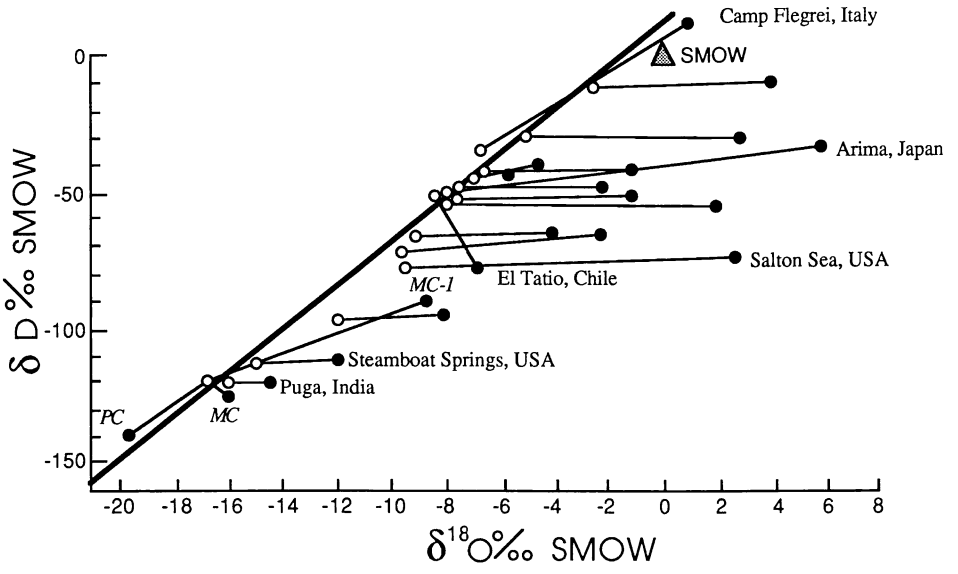


Fig. 9. δD and δ^{18}O relationships in geothermal waters. Open symbols are cold springs or local precipitation, closed symbols are geothermal fluids. Selected fields are labeled. Abbreviations: MC = Meager Creek Hot Springs water, PC = Pebble Creek Hot Springs water, and MC-1 = deep reservoir fluid. Figure is adapted from D'Amore and Panichi (1985). The fluid compositions are shown with respect to the global meteoric water line of Craig (1963).

1963). Variations due to elevation at Meager Mountain have been documented by Clark (ms). These data show that the Pebble Creek Hot Springs fluid and the deep fluid from MC-1 would have to recharge from the highest and lowest elevations, respectively, in the region of Meager Mountain. However, in mountainous terrains the depth of penetration of groundwater is proportional to the height of the fluid column, that is, the elevation of recharge zone. Thus the combination of a low elevation recharge zone for the deep fluid and a high elevation recharge zone for the shallow fluid is unlikely. It is more likely that recharge zones for both shallow and deep fluids are at high elevations, and that subsequent interactions affect their isotopic compositions. The processes known to affect the isotopic composition of geothermal waters are evaporation, mixing, boiling, and water-rock exchange.

Surface evaporation should result in an enrichment in oxygen-18 and deuterium. The slope of a line connecting the δ^{18}O and δD of evaporated water with recharge water will vary from +2 to +3.6 if no water-rock exchange occurs (Giggenbach and Stewart, 1982). This explanation is inadequate for the Meager Mountain fluids for several reasons. First, the slope for the deep fluid, 8.5, is much larger than would be expected for this process. Secondly, there is no evidence that

the deep fluid discharges at the surface, precluding evaporation. Finally, Pebble Creek Hot Springs fluid is isotopically lighter than the average cold spring fluids. The direction of this shift is opposite the effect expected from evaporation.

Mixing of different fluids may also modify their isotopic compositions and is common in geothermal systems (Henley and others, 1984). Enthalpy-chloride diagrams are useful in evaluating and defining such mixing (Fournier, 1979). Figure 10 shows the enthalpy-chloride relationships for the Meager Mountain fluids and the trends discussed below. Enthalpy and chloride can act as conservative quantities during mixing, and consequently mixtures of the endmembers should define a linear trend on an enthalpy-chloride plot (Fournier, 1979). If enthalpy does not act conservatively, as is the case for conductive cooling, the relationship between the variously cooled fluids will define a vertical line on an enthalpy-chloride plot. Cooling by boiling yields an approximately

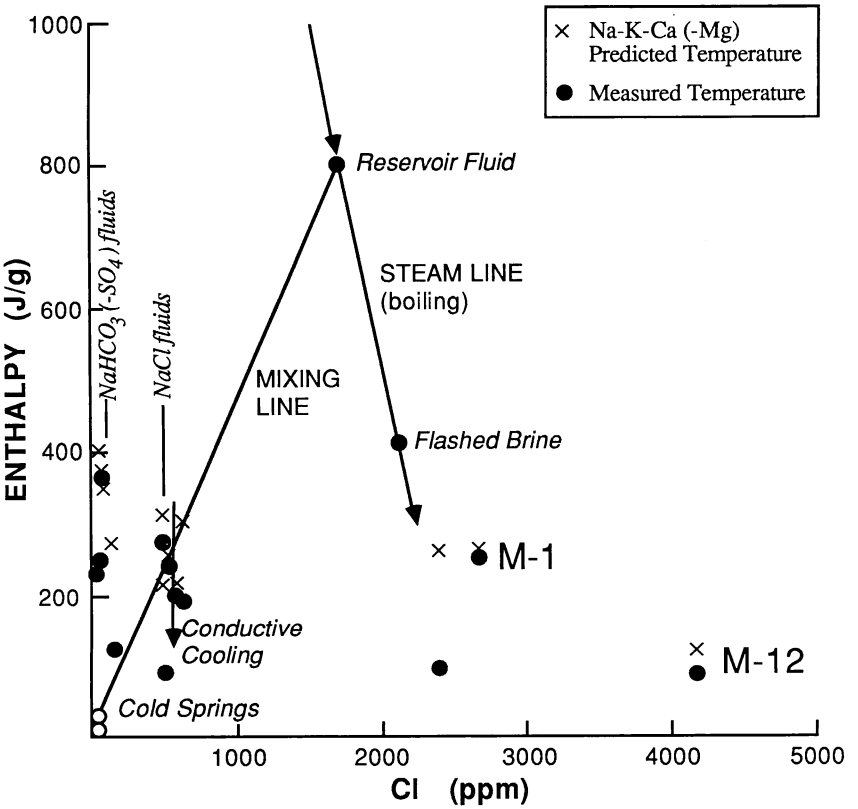


Fig. 10. Enthalpy-chloride plot of fluids from Meager Mountain. Mixing lines shown in figures 10 and 11 are contradictory, as discussed in text.

linear relationship, since the original fluid can be considered a mixture of steam and boiled fluid.

Deuterium-chloride relationships can be used to determine the validity of conclusions based on enthalpy-chloride relationships (Truesdell, Mathenson, and Rye, 1977). Fluids related by mixing will define linear trends on both plots. Fluids related by boiling will display a curvilinear relationship instead of the essentially linear relationship on an enthalpy-chloride plot. Thus, boiling can be distinguished from mixing coincident with the boiling line. As a further test to determine if fluids are related by mixing, the proportions of endmembers in the mixed fluids may be determined by application of the lever rule both to the enthalpy-chloride and the deuterium-chloride plots. If the proportions from both plots do not agree, then the linear relationship may be coincidental.

The enthalpy-chloride and deuterium-chloride relationships of the fluids at Meager Mountain are shown in figures 10 and 11, respectively. The enthalpy-chloride relationships appear to relate the dilute NaCl fluids of the hot springs and shallow wells to the deep fluid and meteoric water by mixing and also suggest a possible boiling relationship between the deep fluid and fluid from M-1. However, these relationships do not

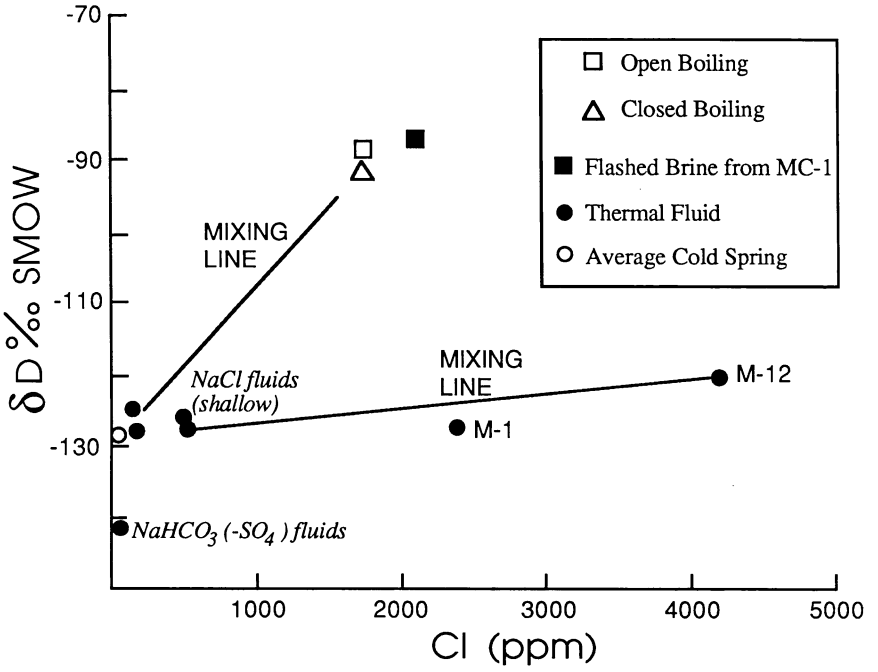


Fig. 11. δD-chloride plot of fluids from Meager Mountain.

hold true upon examination of the deuterium-chloride plot. It is apparent that some of the linear trends are entirely different, and that the proportions of the endmembers are not the same when derived from the two different methods.

In order to evaluate the effects boiling could have on the isotopic composition of the geothermal fluids, four different possible cases were considered. These are adiabatic boiling in an open or closed system and isothermal boiling in an open or closed system. In open boiling, steam is separated from the remaining liquid as soon as it is formed, whereas in closed boiling, the steam is allowed to remain with the fluid and is in equilibrium with the fluid at the final boiling temperature. In natural systems boiling is generally intermediate between open and closed (Truesdell, Mathenson, and Rye, 1977; Drummond and Ohmoto, 1985) and is primarily adiabatic. Isothermal boiling was considered as an extreme case, because it can produce a large steam fraction and thus a large isotopic separation. The effect of closed boiling under adiabatic conditions is calculated from the relationships (Giggenbach and Stewart, 1982):

$$\delta_w = \delta_o + \epsilon y_s \tag{1}$$

and

$$\delta_s = \delta_o - \epsilon(1 - y_s) \tag{2}$$

where δ_w and δ_s are the isotopic compositions of the liquid and steam phases after boiling, δ_o is the istopic composition of the fluid before boiling, y_s is the steam fraction, and ϵ is defined by the relationship:

$$\epsilon = 1000 \ln \alpha \approx 1000(\alpha - 1) \approx \delta w - \delta s \tag{3}$$

where α is the fractionation factor for liquid water and steam. This approximation for ϵ produces little error if 1000 ln α is less than 10 as is the case for oxygen-18 and deuterium fractionation at temperatures above 170°C. The steam fraction, y_s, is determined from:

$$y_s = (H_o - H_w)/(H_s - H_w) \tag{4}$$

where H_w and H_s are the enthalpies of the liquid and steam phases, respectively, at the temperature of boiling, and H_o is the enthalpy of the fluid prior to boiling.

The effects of closed boiling under isothermal conditions are defined by the same relationships with the exception that the steam fraction must be arbitrarily chosen.

Calculations for adiabatic boiling in an open system are more complex. These calculations depend on a numerical integration of the liquid and steam enthalpies throughout the temperature range of boiling. The calculations and integral values are given by Truesdell, Mathenson, and Rye (1977). The effects of isothermal boiling in an open system were calculated from the Rayleigh distillation equation which,

when combined with the δ notation, is given by the relationship:

$$\delta^w = (\delta_o^w + 1000)\ (1 - y)^{\{(1/\alpha)-1\}} - 1000 \qquad (5)$$

where δ_o^w and δ^w are the isotopic compositions of the fluid before and after a fraction y of steam is lost by continuous boiling, and α is the liquid-vapor distribution coefficient. Liquid-vapor fractionation factors used for oxygen-18 were calculated by Bottinga (quoted in Truesdell, Mathenson, and Rye, 1977) from the equation:

$$1000 \ln \alpha = -3.493 + 1.2051\ (10^3/T) + 0.7664\ (10^6/T^2)$$

where T is in kelvins. Fractionation factors for deuterium were taken from the compilation of Truesdell, Mathenson, and Rye (1977).

The calculations indicate that the effects of boiling are not capable of relating the deep fluid to any of the fluids sampled at the surface. For example, the effects of open and closed adiabatic boiling of the deep fluid sampled from MC-1 are shown in figure 11. In these calculations, the fluid was boiled from 194°C to ambient temperature and pressure. The final composition does not approach the isotopic values of the shallow NaCl fluids. The effects of isothermal boiling, not shown in figure 11, were calculated for a temperature of 280°C, since this is the temperature at which deuterium is most depleted in the liquid during boiling. The results of the calculations indicate that for open boiling, 0.99995 of the liquid would have to be boiled off. This is clearly an unlikely steady-state situation. For closed boiling the fractionation factor for deuterium is never large enough to account for the observed relationships. Thus, the deep and shallow NaCl fluids do not appear to be related by concentration of the deep fluid due to boiling.

Steam condensation from a deep boiling fluid, however, can be used to derive a relationship between the Pebble Creek Hot Springs fluid and a deeper flow system. Boiling of a fluid at less than 221°C produces an isotopically light steam phase. Condensation of this steam phase in meteoric water can produce an acidic (from CO_2 and H_2S), reactive, thermal fluid (Mahon, Klyen, and Rhode, 1980). This phenomenon has been observed in many geothermal systems (Giggenbach and Stewart, 1982; Henley and others, 1984). Subsequent reaction of this fluid with rock increases the concentration of reactive ions such as Na, K, and Ca faster than the conservative ions such as Cl. Thus, fluids resulting from steam-heating of groundwater will be $NaHCO_3(-SO_4)$ in character. Field relationships strongly support such an origin for the Pebble Creek Hot Springs. Key observations include: (1) the Na/Cl weight ratio for Pebble Creek Hot Springs, which is 5.16, in contrast to that of the Meager Creek Hot Springs, which is 0.66; (2) H_2S emissions occur on the northern flank of Meager Mountain, hydraulically upstream of Pebble Creek Hot Springs suggesting boiling of a thermal fluid, and (3) acid alteration and thermal fluid occur in L-1, hydraulically downstream of the H_2S emissions (refer to fig. 1 for locations).

Calculations were made to determine if the carbon and sulfur content, the isotopic composition, and the enthalpy of the $NaHCO_3(-SO_4)$ fluid could be reproduced by boiling of the deep fluid. We found that the enthalpy and the carbon and sulfur contents could be duplicated. However, the isotopic composition of the $NaHCO_3(-SO_4)$ fluid could not be duplicated, because steam from the deep fluid would always be too heavy. Thus another, less isotopically evolved, boiling fluid must provide the steam that produces the $NaHCO_3(-SO_4)$ fluid.

The effects of water-rock interaction are less well known in geothermal systems than those of boiling or mixing, but they may also be significant. Water-rock isotopic exchange is a difficult process to model due to the number of variables involved. These variables are temperature, mineral species participating in the isotopic exchange, isotopic fractionation factor choices, initial fluid and rock composition, kinetics and water to rock ratio. In our calculations, we considered temperatures of 100° to 250°C in order to bracket the measured temperature of the deep fluid. The mineral species we considered were epidote, chlorite, muscovite, smectite, and feldspar (An_{35}). These are the predominant minerals in the crystalline rocks at Meager Mountain. The initial isotopic composition of the rock was obtained from four analyses of meta quartz-diorite. The oxygen isotopic composition of the rocks analyzed from Meager Mountain ranged from 3.2 to 7.1 permil, and the hydrogen isotopic composition ranged from −116 to −86 permil. This compositional range is in accord with that found in other intrusive-metamorphic terrains in the Cascade range (Taylor, 1979). Because of the broad range of isotopic values that typify such intrusive-metamorphic terrains an average isotopic composition was used. An average of the same four samples was also used for the oxygen and hydrogen concentrations in these rocks. These concentrations ranged from 48 to 49 percent for oxygen and from 0.25 to 0.49 percent for hydrogen.

Several estimates of isotopic fractionation factors were used to evaluate the extent of water-rock interaction. Regression equations for these temperature-dependent factors have been published by a number of researchers based on several methods, including experimental data, theoretical models, and in-situ studies of basins. The mineral fractionation factors used for oxygen were taken from Bottinga and Javoy (1973; plagioclase), Yeh and Savin (1977; smectite), Cole (ms; smectite), O'Neil and Taylor (1967; muscovite), and Taylor (1974; chlorite).

The choice of which mineral fractionation factors to use for deuterium was more difficult. Fractionation factors can be found in Graham, Sheppard, and Heaton (1980; epidote), Yeh (1980; smectite), and Taylor (1974; muscovite and chlorite). However, the low-temperature values for chlorite and muscovite, the most abundant hydrous minerals at Meager Mountain, must be extrapolated from Taylor (1974) based on the linear portion of the temperature versus $10^3 \ln \alpha$ curve below the dogleg at 400°C. This linear extrapolation is inaccurate because other hydrous minerals for which experimental data exist show

that a large number of minima, maxima, inflections, and crossovers occur in the temperature range of interest (O'Neil, 1986). Thus, rather than use fractionation factors for a single mineral, values for $10^3 \ln \alpha$ ranging from -60 to -15 were used at all temperatures and w/r ratios to obtain a best fit of the measured data. This procedure is discussed in more detail later in this section.

The kinetics of isotope exchange between water and rock have been studied by Cole (ms) and Cole, Ohmoto, and Lasaga (1983). These authors have shown that isotope exchange takes place by two processes, diffusion and dissolution-precipitation. Of the two, dissolution-precipitation is significantly faster than diffusion-controlled isotope exchange. According to Cole (ms), the fraction of oxygen isotope exchange with respect to equilibrium ranges from 0.14 to 0.90 for producing geothermal systems. Thus an equilibrium model of Meager Mountain may be an approximation. However, as discussed below, the reasonable fit of the measured isotopic composition of the deep fluid to an equilibrium model may indicate a near-equilibrium state between water and rock.

The equilibrium composition of the fluid is calculated by combining the relationships:

$$w(\delta_w^f - \delta_w^i) + r(\delta_r^f - \delta_r^i) = 0 \text{ (Taylor, 1974)} \tag{7}$$

and

$$\alpha_{r-w} = (\delta_r^f + 1000)/(\delta_w^f + 1000) \tag{8}$$

where

- w = atom percent of water in the system,
- r = atom percent of rock in the system,
- δ_r^f = final isotopic composition of the rock,
- δ_w^f = final isotopic composition of the water,
- δ_r^i = initial composition of the rock,
- δ_w^i = initial composition of the water, and
- δ_{r-w} = fractionation factor between water and rock.

Combining these relationships yields

$$\delta_w^f = \{(w/r)\,\delta_w^i - 1000\alpha_{r-w} + 1000 + \delta_r^i\}/(\alpha_{r-w} + w/r) \tag{9}$$

where

$$(w/r) = (w/r)_{weight}\,(Ci_w/Ci_r), \tag{10}$$

$Ci_{w,r}$ = concentration in moles of hydrogen or oxygen per weight of water (w) or rock (r), (w/r) = water/rock atomic ratio, and $(w/r)_{weight}$ = water/rock weight ratio. The form of eq 9 differs from that of Taylor (1974) because the approximation

$$1000 \ln (\alpha_{r-w}) \approx \delta_r^f - \delta_w^f$$

was not used. This approximation is not valid for the large water-mineral fractionation factors of deuterium in the temperature range considered here.

An exact fit of the measured composition was obtained by using the following algorithm. A minimum temperature was first obtained by matching the oxygen value at extremely low (w/r) weight ratios, where the final water oxygen composition approaches a constant value. Water/rock ratios were then varied to fit exactly the $\delta^{18}O$ of the deep fluid at 1 °C intervals above the minimum temperature. This was done first, because the temperature dependence of the oxygen fractionation factor is better known than that of deuterium. The final step was to fit the deuterium value at each temperature increment by varying the fractionation factor. This last step also provides a maximum temperature for exchange, because the fractionation factor needed to fit the measured composition rapidly decreases with each temperature increment. A fractionation factor of $10^3 \ln \alpha = -60$ was used to produce the maximum temperature, because we considered it to be the lowest fractionation factor consistent with the hydrous minerals likely to be found in the Meager Mountain reservoir (see O'Neil, 1986, for compilation of fractionation factors).

The combinations of values that gave an exact fit for the isotopic composition of the deep fluid are shown in figure 12A. The isotopic composition of the initial rock used to produce this figure was a $\delta^{18}O$ of 5.3 permil and a δD of -103 permil. Plagioclase (An_{35}) was assumed to be the mineral controlling oxygen fractionation. The deep reservoir fluid falls on the curves defined by temperatures of 123° to 126°C, (w/r) weight ratios of 0.005 to 0.022, and deuterium $10^3 \ln \alpha$ values of -22.6 to -57.7. Figure 12B shows the positions of the initial and final fluids and the average initial rock composition relative to the (w/r) weight-temperature curves. These curves were produced using a constant deuterium fractionation factor of -22.6.

The estimated (w/r) weight ratios of 0.005 to 0.022 fit within the lower range of (w/r) weight values for geothermal systems (Cole, ms; Henley and Ellis, 1983) and for hydrothermal systems within and adjacent to batholiths (Taylor, 1979). These low (w/r) weight ratios correlate well with the low productivity of the Meager geothermal system. For comparison, (w/r) weight ratios range from 0.2 to 0.7 for the highly productive New Zealand geothermal systems.

CONCEPTUAL MODEL OF THE GEOTHERMAL SYSTEM

A schematic model illustrating the major structural and hydrologic features of the geothermal system at Meager Mountain is presented in figure 13. The system at Meager Mountain is characterized by several isotopically and chemically distinct fluids. Four of these fluids are known to be present, while two are hypothetical precursors to the Na-$HCO_3(-SO_4)$ hot spring fluids, as discussed below. In this model, all the fluids originate as meteoric water near the top of the volcano.

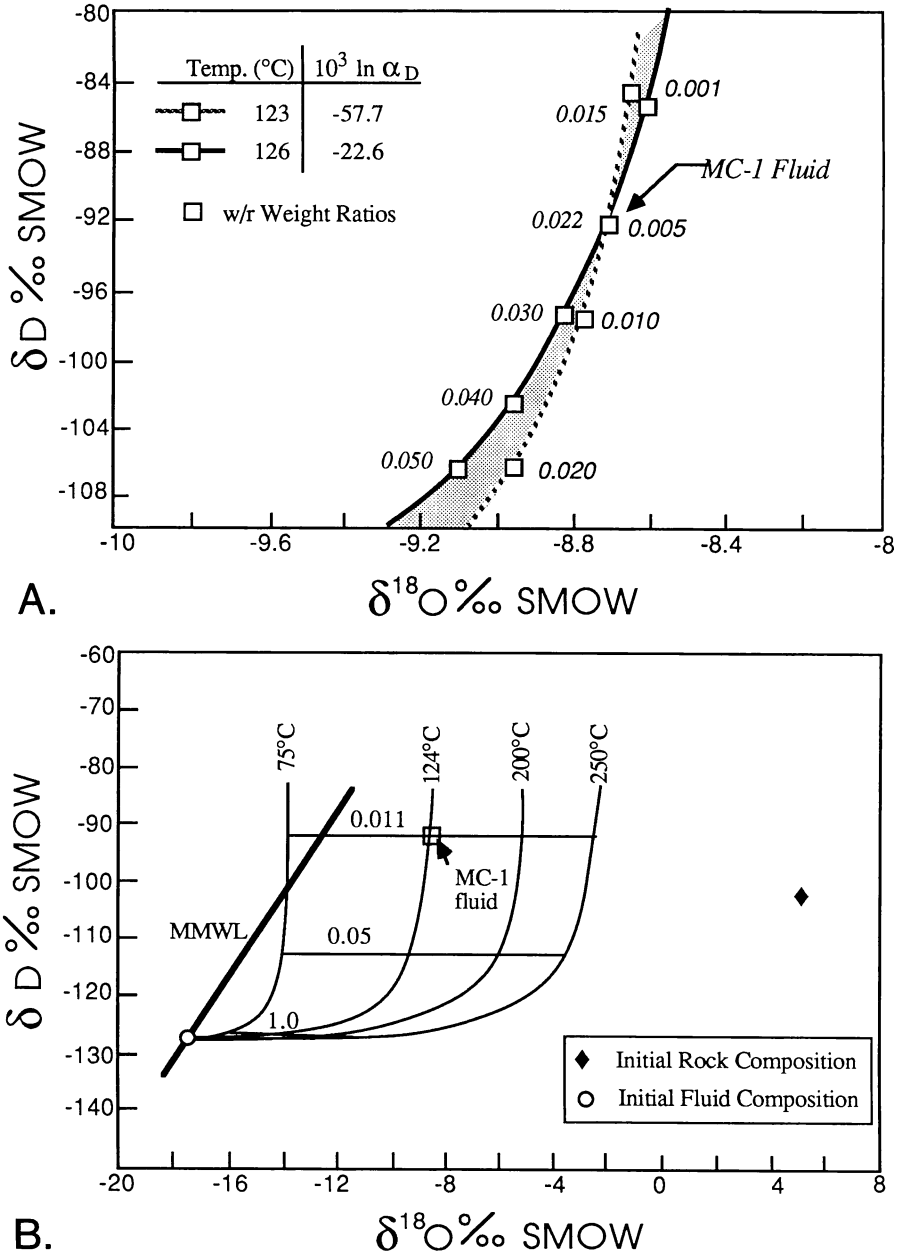
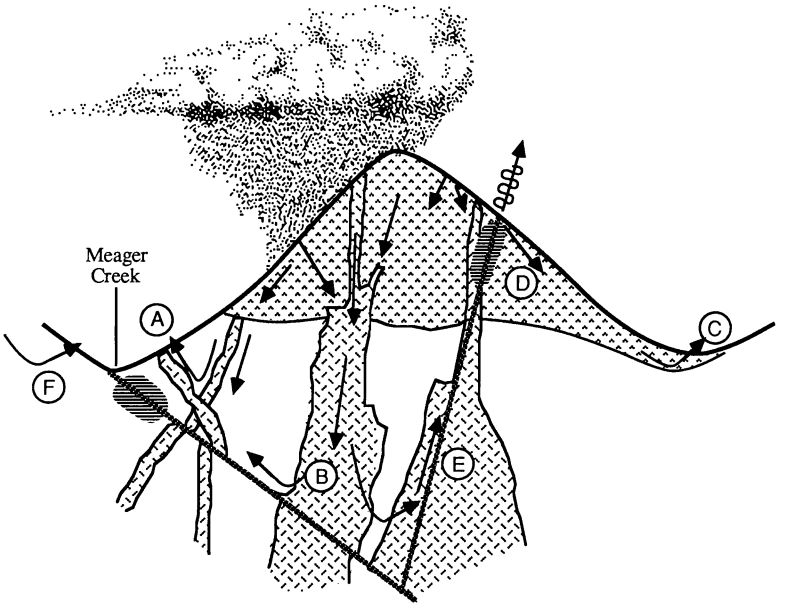


Fig. 12(A) Range of rock-water ratios, temperatures, and deuterium fractionation factors that can be used to produce the deep fluid by isotopic exchange of meteoric water with meta quartz-diorite. An average of four samples was used for the initial isotopic composition of the rock.

(B) Final isotopic composition of fluid as a function of temperature and water-rock ratio at Meager Mountain. Calculated with an average Meager Mountain meta quartz-diorite composition. Deuterium fractionation factor used was −33.8. Water-rock ratios are shown as numbered subhorizontal lines.



EXPLANATION

Fault or Fracture Zone

Zone of Mineral Deposition

Volcanics

Dikes

Crystalline Basement

Flow Path

Gas / Steam Flux

Fluid	TDS	Temp.°C	δD (‰)	δ^{18}O(‰)
(A) NaCl	Low	60	-126	-16.4
(B) NaCl (deep)	Mod	190	-91	-8.7
(C) Neutral HCO_3(-SO_4)	Low	60	-141	-19.6
(D) Acid Sulfate (hypothetical)	Low	~100	?	?
(E) NaCl (hypothetical)	Mod	<220	<-110	<-10
(F) NaCl	High	22	-120	-14.2

Fig. 13. Conceptual model of the Meager Mountain geothermal system showing the distribution of fluid types and major fluid pathways. Orientation of schematic is Northeast–Southwest. See text for discussion.

Shallow flow on the southern side of the volcano produces fluid A, which is a NaCl fluid characterized by a low TDS content, small isotope shifts, and geothermometer temperatures below 70°C. Fluid B is produced by deep flow below the volcano. This NaCl fluid is characterized by moderate TDS contents, large isotope shifts, and geothermometer

temperatures approaching 200°C. Fluid B occurs only in well MC-1, a deep drillhole that intersects the Meager Creek fault zone. This fault zone may also control the shallow flow of fluid A. However, no evidence of mixing between the two fluids has been found. Thus communication between fluids A and B may be cut off by mineral precipitation and sealing of the fractures.

The large isotopic shift and the low salinity of fluid B suggest a long flow-time and a slow velocity for this deep fluid. In general, the degree of isotopic exchange in geothermal fluids is related to the log of chlorinity (Cole, ms). Where fluid velocities are fast, the more rapid process of dissolution is favored over isotopic exchange resulting in relatively high chlorinities compared to the extent of isotopic equilibrium. At Meager Mountain, however, the deep fluid has been strongly isotope-shifted despite its low salinity, suggesting that the fluid velocity is low.

Fluid C is a $NaHCO_3(-SO_4)$ fluid found on the northern slope of the volcano and is characterized by a low TDS content, light oxygen and deuterium isotope signatures, and geothermometer temperatures near 100°C. This fluid was produced by steam heating, oxidation and acidification, and subsequent neutralization of meteoric water. The acid precursor of the $NaHCO_3(-SO_4)$ fluid, fluid D, is hypothetical, but its presence is indicated by acid alteration and H_2S fumaroles on the northern slope of Meager Mountain.

Fluid E, the source of the steam that produced fluids C and D, must have originated under conditions similar to fluid B, the deep fluid sampled on the south side of the volcano. However, the isotopic composition of fluid E, inferred from that of fluid C, is less isotope shifted than that of fluid B. Thus, fluid E is shown on figure 13 with a shallower circulation path than fluid B. An alternate explanation is that fluid E is produced from the same depths but has a higher fluid velocity than fluid B.

Fluid F is a NaCl fluid found in warm springs and wells M-1 and -12 immediately south of Meager Mountain. The fluids discharge within 2 km of Placid Hot Springs (fig. 2). Fluid F has a high TDS content and moderate isotope shifts but has geothermometer temperatures of less than 71°C. The element ratios and isotope shift of this fluid indicate that it must have originated as a geothermal fluid but has subsequently cooled and reequilibrated. These characteristics indicate a very long residence time. Fluid F is shown on figure 13 as flowing from the south because it flows to the surface on the south side of the fluid divide defined by Meager Creek, making any direct connection to the hydrothermal system of Meager Mountain unlikely (C. Forster, personal commun.).

None of the fluids described above can be related by mixing or conductive cooling and subsequent reequilibration, although they all appear to have been derived from isotopically similar meteoric water. This suggests that many of the fluid channels do not interconnect and that each of the fluids evolved independently. Furthermore, the erratic

and minor geothermal mineralization, low fluid productivities, and low water/rock ratios imply that overall permeabilities are low. Thus the fluid flow must be restricted to a few discrete fault and fracture zones. In view of these conditions and the extreme relief, it is likely that fluid flow is driven exclusively by topographically-controlled head differences, rather than by density differences as has been theorized for the more productive geothermal systems.

Shallow dipping faults, such as the Meager Creek fault zone, may play a significant role in allowing the deep, high temperature brines formed beneath the central parts of the volcano to move upward and laterally. Within the upper 1000 m of the surface at Meager Mountain, only the dikes and hydrothermal breccias display extensive geothermal alteration, suggesting that they were the primary conduits for upward movement of the thermal fluids. Subsequent sealing of the fractures in the dikes and breccias by mineral precipitation has substantially reduced the permeabilities of these rocks.

CONCLUSIONS

The Meager Mountain geothermal system is located in crystalline rock beneath an andesitic stratovolcano in the Garibaldi Volcanic Belt, the northern extension of the Cascade Mountain Range. Unlike other producing geothermal systems, wells and springs at Meager Mountain discharge chemically and isotopically distinct fluids that cannot be related by mixing or boiling. High-temperature fluid produced from one of the deep wells at Meager Mountain is characterized by isotopic shifts in both deuterium and oxygen-18. Water-rock weight ratios calculated from the isotopic composition of this fluid range from 0.005 to 0.022 at about 125°C. Erratically distributed and generally weak alteration and trace element enrichments related to modern geothermal activity are also indicative of low water-rock ratios within the explored portions of the system. These chemical and thermal features appear to be the result of fluid flow within discrete channels, rather than through pervasively fractured rock.

ACKNOWLEDGMENTS

This investigation was funded by the U.S. Department of Energy, San Francisco Operations Office under Contract No. DE-AC03-84SF12196. We are particularly grateful to J. Stauder of B. C. Hydro and Power Authority for permission to sample the wells and publish interim results of this work. The staff of Nevin, Sadlier-Browne, and Goodbrande, especially B. Fairbank and R. Openshaw, generously shared their ideas, maps, and logs and provided the samples used in the study. C. Forster introduced us to some of the complexities of fluid flow in mountainous terrains and helped shape our ideas about the Meager Mountain system. R. Capuano made many helpful suggestions during the early phases of the project. Special thanks are also due to R. Kroneman, K. Yorgason, and B. Miller for performing the chemical

analyses, L. McPherson for the X-ray analyses, and E. Ripley for the isotopic analyses. J. Hulen provided valuable assistance in interpreting the X-ray patterns. J. Ballantyne, D. Foley, D. Nielson, H. Ross, and P. Wright reviewed an early version of the manuscript and made many helpful suggestions. P. Daubner drafted the illustrations, and J. Pingree and L. Orvin typed the manuscript. We are grateful for their help.

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