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Isotope Geochemistry of Thermal and Nonthermal Waters in the Valles Caldera, Jemez Mountains, Northern New Mexico

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Over 100 stable isotope and 45 tritium analyses from thermal and nonthermal waters of the Jemez Mountains region, New Mexico, have been used to define the hydrodynamics of the Valles caldera (Baca) geothermal system and related geothermal fluids of the region. Evaluation of 36 cold meteoric waters yields an equation for the Jemez Mountains meteoric water line of $\delta D = 8\delta^{18}O + 12$, while further evaluation of nine cold meteoric waters yields an equation relating recharge elevation to deuterium content of $E(\text{meters}) = -44.9\ (\delta D) - 1154$. Based on the deuterium content of five Baca well waters (223°–294°C), the average recharge elevation of the Valles geothermal system ranges from 2530 to 2890 m. This range of elevations falls between the elevations of the lowest point of the caldera floor (2400 m) and the summit of the resurgent dome inside the caldera (3430 m). Thus stable isotopes indicate that the caldera depression probably serves as a recharge basin for the deep geothermal system. Although cold spring waters of the Jemez Mountains region consist of meteoric water, tritium analyses show that most of them contain water between 20 and 75 years old. The two major streams draining the Valles moat zone, San Antonio Creek and East Fork Jemez River, may contain more than 50% of this relatively old groundwater depending on the season. In contrast, streams draining the central resurgent dome of the caldera contain present-day meteoric water. Using piston flow and homogeneously mixed reservoir models as end-member cases, the tritium contents of the Baca fluids (0.18–1.10 tritium units (TU)) indicate that the mean residence time of water in the reservoir is between 60 and 10,000 years old. Deep geothermal fluids display a positive oxygen 18 shift of not less than 2‰ because of rock-water isotopic exchange at 220°–300°C. The Valles geothermal system is capped by a vapor zone that is roughly 600 m thick and best developed at Sulphur Springs. Fumarolic steam at Sulphur Springs is unusually depleted in oxygen 18, suggesting that it is probably derived by boiling of near-surface groundwater at 200°C. Surface acid-sulfate waters are mixtures of condensed steam and surficial groundwaters that display the isotopic processes of evaporation and exchange between H_2O and CO_2. A lateral outflow plume discharges from the Valles geothermal system down the Jemez fault zone and feeds two sets of thermal springs in San Diego Canyon. Isotopic evidence shows that these springs consist of three components: (1) deep geothermal fluid, (2) surficial and/or near-surface groundwater, and (3) relatively old, but cold, mineralized water. This latter component presumably circulates in Paleozoic strata underlying the western caldera flank. Hot, Precambrian, pore fluid brine occurs beneath the main Valles caldera hydrothermal system and may be generated by metamorphic processes in the relatively impermeable conductive regime above the magmatic heat source of the caldera.

INTRODUCTION

The Valles caldera (Figures 1 and 2) occurs in the center of the Jemez Mountains volcanic field located in north central New Mexico at the boundary between the southeastern Colorado Plateau and the western Rio Grande rift [*Smith et al.*, 1970]. Volcanic activity culminated during the last 1.45 Ma with the formation of the Toledo and Valles calderas and eruption of intracaldera rhyolites as young as 0.13 Ma [*Doell et al.*, 1968; *Gardner at al.*, this issue; *Heiken et al.*, this issue]. This magmatic activity has provided a potent heat source to generate a high-temperature hydrothermal system inside the caldera [*Goff and Grigsby*, 1982; *White et al.*, 1984]. The U.S. Geological Survey created a known geothermal resource area (KGRA) during the late 1960's that covers most of the two calderas, and Union Oil Company discovered the Valles caldera (Baca) geothermal field (220°–300°C) by drilling Baca 4 in October 1970 [*Dondanville*, 1971, 1978]. On the western margin of the Valles caldera, Los Alamos National Laboratory has developed the prototype hot dry rock (HDR) geothermal system that reaches temperatures of 325°C at depths of 4.5 km in impermeable basement rocks [*Dash et al.*, 1983]. Reviews of the Valles caldera geothermal projects have been published by *Laughlin* [1981] and *Goff and Grigsby* [1982].

In the past, most thermal springs in the Jemez Mountains region were used for bathing, but presently, only the spa of Jemez Springs is commercially active [*Summers*, 1976]. Although the numerous thermal manifestations have been indexed and studied [*Trainer*, 1974, 1975, 1978; *Goff et al.*, 1981, 1982; *Goff and Grigsby*, 1982], comprehensive isotopic data have been accumulated on the thermal and nonthermal waters only during 1978 to 1983. The isotopes ^{18}O, ^{2}H, and ^{3}H represent the best tracers of the water molecule and thus provide excellent information on the origin, history, and flow paths of groundwater [*Fontes*, 1980]. The objectives of this stable isotope and tritium investigation are (1) to compute the average recharge elevation and to determine the probable recharge area of the Valles geothermal system, (2) to discuss the relative ages of the various thermal and nonthermal fluids, and (3) to refine earlier hypotheses on the hydrodynamics of the geothermal system as defined by chemical data. Field and laboratory procedures for the collection and isotopic analysis of water samples (Table 1 and Figures 1 and 3) have been described by *Goff et al.*, [1982] and *Vuataz et al.* [1984].

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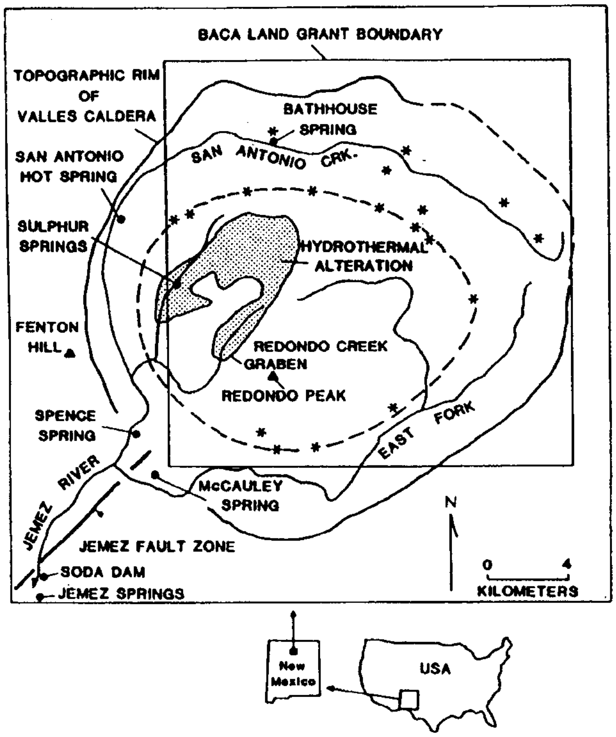

Fig. 1. Map of Valles caldera showing major tributaries, major hot spring areas, and Jemez fault zone; dashed line indicates ring-fracture zone; stars denote intracaldera rhyolite vents, and dotted pattern shows area of most intense surface alteration.

GEOLOGIC AND HYDROLOGIC SETTING

The Jemez Mountains volcanic field (> 13 to 0.13 Ma) consists of domes, flows, tuffs, and volcaniclastic sediments that overlie Tertiary basin fill sedimentary rocks of the Rio Grande rift on the east and Paleozoic through Precambrian rocks on the west [*Doell et al.*, 1968; *Bailey et al.*, 1969; *Smith et al.*, 1970; *Riecker*, 1979; *Gardner and Goff*, 1984; *Gardner et al.*, this issue]. Volumetrically, the field is dominated by two-pyroxene andesites and rhyolite ash flow tuffs. Many local aquifers occur due to great variations in the permeability of the volcanic rocks. Tertiary rocks filling the Rio Grande rift consist primarily of nonindurated sandstones of the Santa Fe Group and display comparatively high permeability and porosity. Extensive aquifers occur in these sandstones at depths ≥600 m beneath the Pajarito Plateau east of Valles caldera [*Purtymun and Johansen*, 1974]. Permian red beds of the Abo Formation crop out beneath the Tertiary rocks throughout the western Jemez Mountains region. These red beds act as aquitards because of their shale contents and diagenetic cements. Pennsylvanian Madera Limestone underlying the Abo contains both thermal and nonthermal aquifers that are controlled by stratigraphic horizons and faults [*Trainer*, 1974]. Precambrian granitic and metamorphic rocks are relatively impermeable; thus groundwater is commonly encountered in porous sedimentary strata just above the Precambrian interface.

The tectonic evolution of the Jemez Mountains has been recently discussed by *Gardner and Goff* [1984] and *Aldrich* [this issue]. Because the volcanic rocks straddle the western boundary of the Rio Grande rift, volcanic rocks and Tertiary sediments thicken to the east (Figure 2). Several north and northeast trending fault zones occur along the western Rio Grande rift, but two of them, the Jemez and Pajarito fault zones (Figure 3), have profound influence on the hydrology of the region [*Goff and Sayer*, 1980; *Goff et al.*, 1981]. The Jemez fault zone acts as a conduit for a lateral outflow plume of deep geothermal water that leaks out of the southwestern side of Valles caldera. In contrast, the Pajarito fault zone, the ring faults along the eastern side of Valles caldera, and the Precambrian basement between these faults apparently act as a barrier to eastward flow of geothermal fluid because no known hot springs or other thermal manifestations occur east of the caldera.

Structures created during the development of the Jemez Mountains and Rio Grande rift are overprinted by the structural collapse of Toledo and Valles calderas. These collapse features presumably serve as traps for the collection and deep circulation of meteoric waters. Uplift of the central resurgent dome of Valles caldera [*Smith and Bailey*, 1968] resulted in extension and formation of the northeast trending Redondo Creek Graben, which lies on strike with the precaldera Jemez fault zone [*Goff and Gardner*, 1980]. Although thermal gradient analysis has defined as many as three major zones of convective upflow to the Valles geothermal system [*Swanberg*, 1983], only one upflow zone circulates in the Redondo Creek graben, and it is this upflow zone that comprises the Baca geothermal field.

Surface drainage of the Jemez Mountains is controlled by the Rio Grande and several tributaries. The most important tributary is the Jemez River which collects the interior streams of Valles caldera: San Antonio Creek, Sulphur Creek, Redondo Creek, and Jaramillo Creek [*Trainer*, 1974]. Within Valles caldera, surface water is shed toward the caldera moat zone from the rim and from the central resurgent dome [*Laughlin*, 1981]. A large amount of water apparently infiltrates through moat sediments and volcanic rocks, but the rest drains away by the Jemez River which breaches the southwest wall of Valles caldera and flows down San Diego Canyon.

The rugged topography, variety of rock types, and rift- and caldera-related structures of the Jemez Mountains region result in formation of numerous local groundwater systems (i.e., perched aquifers, contact springs, karstic springs, basin aquifers) as well as a geothermal system linked to the volcanic heat source. In their comparative studies characterizing the thermal and nonthermal waters of the Jemez Mountains, *Goff et al.* [1982] and *Grigsby et al.* [1984] have defined six groups of waters according to their geochemistry and geohydrology. These water types are described as follows.

Type A: Cold Groundwaters

Waters included within this group issue from wells and springs on the Pajarito Plateau and from springs, wells, and creeks within Valles caldera and the Jemez Mountains. Many of the springs discharge along the walls of San Diego Canyon that cuts through the volcanic sequence into Paleozoic and Precambrian rocks. These cold, relatively dilute waters occur within three primary geologic settings: (1) Miocene to Quaternary volcanic rocks, (2) late Tertiary basin fill sediments of the Rio Grande rift, and (3) Paleozoic and subordinate Mesozoic sedimentary rocks of the Colorado Plateau. Chemically, HCO_3 is the dominant anion, while Cl and trace elements such as B are low. Oxygen and hydrogen isotopic compositions indicate the waters have a meteoric origin (Table 1). Cold groundwaters are easily distinguished from thermal and mineral waters by their low Cl and trace element contents (Figure 4).

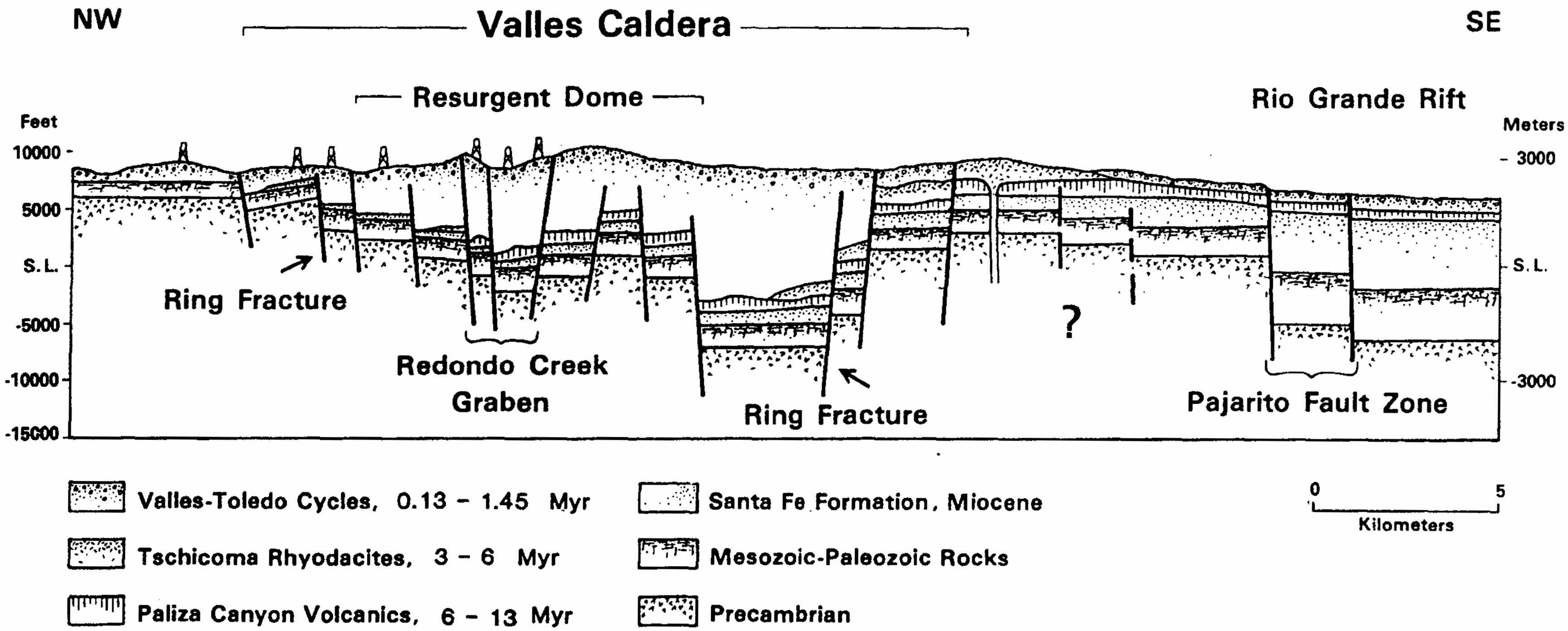

Fig. 2. Schematic NW-SE cross section across Valles caldera [*Goff*, 1983] based on stratigraphy of many geothermal wells projected into the plane of the section and the gravity interpretation of *Segar* [1974]; intracaldera volcanic rocks are omitted for clarity. Structure between Valles caldera and Pajarito fault zone is poorly constrained due to paucity of geophysical and drill hole information. Structure east of Pajarito fault zone is discussed by *Goff and Grigsby* [1982] and *Gardner and Goff* [1984].

TABLE 1. Isotopes and Selected Geochemical Data of the Thermal and Nonthermal Waters of the Valles Caldera–Southern Jemez Mountains Region, New Mexico

Name	Map[a]	Date	Temperature,[b] °C	Emergence Elevation, m	$\delta^{18}O^c$, ‰	dD^c, ‰	^3H, TU	Cl, mg/L	SO_4, mg/L	HCO_3, mg/L	B, mg/L	SiO_2, mg/L
				Los Alamos and Pajarito Plateau Area[d]								
Gallery Spring[e]	1	Aug. 1978	11	2440	−12.20	−84.3	···	<1	<5	52	<0.05	43
T-3 well	2	Aug. 1978	13	1790	−10.65	−73.8	···	4	5	102	<0.05	15
T-2 well	3	Aug. 1978	11	1790	−10.60	−73.5	···	2	5	78	<0.05	5
Sacred Spring[e]	4	Aug. 1978	14	1720	−11.80	−81.8	···	2	7	114	<0.05	34
Basalt Spring[e]	5	Aug. 1978	15	1830	−10.85	−76.5	···	12	18	98	<0.05	44
L-6 well	6	Sept. 1978	27	1740	−13.45	−94.7	···	4	6	170	<0.05	33
L-1B well	7	Sept. 1978	30	1710	−14.30	−103.0	···	15	32	326	0.45	36
PM-2 well	10	Sept. 1978	24	1780	−11.40	−77.5	···	3	<5	65	0.25	83
PM-1 well	11	Sept. 1978	28	1760	−10.95	−74.1	···	6	6	146	0.25	82
G-6 well	12	Sept. 1978	31	1780	−11.25	−76.0	···	2	5	94	<0.05	55
G-4 well	14	Sept. 1978	26	1780	−11.10	−76.3	···	2	5	92	0.12	53
G-2 well	16	Sept. 1978	30	1720	−11.95	−83.1	···	4	5	122	0.12	77
G-1A well	17	Sept. 1978	28	1740	−11.80	−82.5	···	1	5	100	0.55	78
G-1 well	18	Sept. 1978	29	1740	−11.65	−81.0	···	1	5	97	<0.05	84
White Rock Canyon Spring[e]	19	Sept. 1978	19	1640	−11.00	−76.8	···	1	<5	74	<0.05	71
Pajarito Spring[e]	32	July 1979	20	1680	−10.90	−74.5	···	6.4	7.5	100	0.05	67
Unnamed cold spring[e]	52	July 1980	15	2020	−11.55	−80.5	···	7.1	3.4	75	<0.01	65
Unnamed cold spring[e]	53	July 1980	17	2050	−11.60	−80.1	···	7.5	2.4	73	<0.01	69
				Sulphur Springs Area								
Men's Bathhouse Mudpot	21	Jan. 1979	78	2600	−3.25	−50.2	···	2.5	786	0	<0.1	221
Men's Bathhouse Mudpot	21	Aug. 1981	82	2600	−3.65	−47.1	···	8.5	1890	0	0.03	240
Men's Bathhouse Mudpot	21	Jan. 1982	72	2600	···	···	2.1 ± 0.5	···	···	···	···	···
Women's Bathhouse Spring[f]	21	Aug. 1981	86	2600	−8.45	−60.8	···	<1	6400	0	0.2	168
Women's Bathhouse Spring	21	Jan. 1982	89	2600	···	···	19 ± 2	···	···	···	···	···
Lemonade Spring	21	Sept. 1980	58	2600	−11.05	−69.7	···	17	2740	0	0.03	238
Lemonade Spring	21	Aug. 1981	57	2600	−11.10	−68.7	···	3.5	2370	0	<0.1	229
Electric Spring	21	Aug. 1981	32	2600	−11.90	−75.8	···	8.6	2850	0	<0.1	212
Footbath Spring[f]	21	Aug. 1981	39	2600	−20.40	−82.1	···	<1	7900	0	0.2	214
Footbath Spring	21	Jan. 1982	18	2600	−23.15	−107.0	13 ± 2	···	···	0	···	···
Unnamed hot spring	21	Jan. 1979	63	2600	−8.80	−60.7	···	3.7	2110	0	<0.1	230
Unnamed spring	21	Aug. 1981	19	2600	−10.05	−75.4	···	7.7	287	320	<0.1	42
Main Fumarole, steam[g]	21	March 1979	93	2600	−14.05	−86.0	···	<0.5	4.5	0	<0.01	<1
Main Fumarole, steam	21	Sept. 1980	93	2600	−13.80	−86.4	···	···	···	0	···	···
Sulphur Creek	21	Aug. 1981	16	2600	−9.75	−72.9	···	9.6	238	48	<0.1	44
Sulphur Creek	21	Jan. 1982	1	2600	···	···	39 ± 4	···	···	···	···	···
Alamo Canyon Pool[e]	23	March 1979	1	2650	−13.45	−97.3	···	4.9	109	0	<0.1	44
Short Canyon Spring[e]	24	Sept. 1980	8	2650	−12.20	−86.4	···	5.9	199	0	<0.1	55
Short Canyon Spring[e]	24	Aug. 1981	15	2650	−12.75	−89.2	···	7.8	517	0	<0.1	70
				Redondo Creek Area								
Redondo Creek[e]	61	May 1983	4	2440	−12.50	−88.7	34.1 ± 0.9	10.7	12.5	27	0.06	25
Baca 4 well[h]	62	June 1982	294	2840	−9.96	−88.3	0.49 ± 0.08	1898	35	155	14.0	518
Baca 13 well[h]	62	June 1982	278	2832	−9.96	−86.3	1.0 ± 0.4	1907	37	168	14.9	488
Baca 15 well[h]	62	July 1982	268	2779	−8.61	−83.9	0.25 ± 0.05	2580	36	70	20.1	539
Baca 19 well[h]	62	Oct. 1982	223	2779	−8.41	−83.1	0.47 ± 0.08	2665	38	111	21.4	451
Baca 24 well[h]	62	June 1982	259	2664	−8.43	−81.8	1.10 ± 0.09	2555	39	72	21.3	503
				Hot Dry Rock System, Fenton Hill[i]								
Make-up water well[e]	64	···	20	2650	−12.75	−90.0	···	19.0	13.3	118	0.02	68
GT-2 well (initial)	64	···	155	2650	−8.55	−77.8	···	1502	520	373	41.6	245
GT-2 well[j]	64	March 1980	180	2650	−7.40	−74.7	···	1750	341	538	42.3	267
GT-2 well[j]	64	March 1980		2650	−9.45	−81.4	···	921	248	493	25.0	232
GT-2 well[j]	64	March 1980		2650	−10.90	−84.6	···	583	227	400	14.5	214
GT-2 well[j]	64	March 1980		2650	−10.80	−85.0	···	528	260	471	13.5	214
GT-2 well[j]	64	April 1980	170	2650	−11.10	−85.3	···	387	222	483	10.9	211
GT-2 well[j]	64	April 1980		2650	−11.00	−85.0	···	412	258	454	11.2	227
GT-2 well[j]	64	May 1980		2650	−10.70	−84.6	···	485	286	450	13.4	233
GT-2 well[j]	64	June 1980		2650	−10.75	−81.5	···	380	229	490	11.8	228
GT-2 well[j]	64	July 1980		2650	−10.55	−82.7	···	314	190	471	9.8	211
GT-2 well[j]	64	July 1980		2650	−10.75	−81.5	···	351	205	467	10.9	225
GT-2 well[j]	64	Nov. 1980	160	2650	−11.05	−82.6	···	302	208	434	10.2	199
				Ring Fracture Zone								
Spence Hot Spring	25	July 1978	45	2240	−12.35	−86.4	···	8	16	144	0.15	66
Spence Hot Spring	25	March 1982	42	2240	−12.25	−86.5	1.9 ± 0.5	7.2	18	135	···	···
Spence Hot Spring	25	Jan. 1983	42	2240	···	···	0.20 ± 0.07	8.2	17	140	0.12	66
McCauley Spring	26	July 1978	31	2240	−12.60	−88.4	···	6	7	86	0.24	56
McCauley Spring	26	March 1982	32	2240	−12.45	−89.1	2.3 ± 0.5	3.9	6.6	81	···	···
McCauley Spring	26	Jan. 1983	32	2240	···	···	0.27 ± 0.07	4.4	5.8	87	<0.05	54
San Antonio Hot Spring	27	July 1978	42	2550	−12.65	−92.0	···	2	7	56	<0.05	79
San Antonio Hot Spring	27	March 1982	41	2550	−12.70	−91.6	5.1 ± 1.0	2.2	7.6	50	<0.01	80

TABLE 1. (continued)

Name	Map[a]	Date	Temperature,[b] °C	Emergence Elevation, m	$\delta^{18}O$,[c] ‰	dD,[c] ‰	^3H, TU	Cl, mg/L	SO_4, mg/L	HCO_3, mg/L	B, mg/L	SiO_2, mg/L
				Ring Fracture Zone (continued)								
San Antonio Hot Spring	27	March 1983	41	2550	...	...	0.85 ± 0.11	7.0	9.5	57	<0.1	74
Bathhouse Hot Spring	28	Feb. 1979	38	2570	−12.40	−86.4	...	2.4	15	71	<0.1	96
Bathhouse Hot Spring	28	March 1982	37	2570	−11.80	−85.0	8.3 ± 1.1	8.6	12.7	61	<0.01	103
Bathhouse Hot Spring	28	March 1983	38	2570	...	...	0.44 ± 0.07	6.9	15.2	76	<0.1	105
San Antonio Creek	28	Feb. 1979	2	2570	−12.85	−92.9	...	...	...	...	...	...
San Antonio Creek	28	March 1982	3	2570	−12.85	−90.5	6.3 ± 1.2	1.6	9.8	46	<0.01	56
San Antonio Creek	27	Sept. 1982	22	2540	...	...	6.4 ± 0.9	1.2	8.7	51	<0.01	...
San Antonio Creek	28	March 1983	1	2570	...	...	15.5 ± 0.4	6.1	3.5	39	<0.1	38
Battleship Seep	34	Aug. 1979	19	2070	−12.50	−92.9	...	284	373	1745	4.15	18
Battleship Seep	34	Arpil 1980	11	2070	−12.80	−92.7	...	323	290	1980	4.47	17
Battleship Seep	34	May 1983	12	2070	...	...	3.40 ± 0.17	299	295	1966	4.45	17
East Fork Jemez River[e]	34	March 1982	4	2070	−12.40	−88.0	13 ± 2	1.9	4.6	50	...	...
Valle Grande Spring[e]	39	Aug. 1979	15	2630	−12.40	−85.0	...	9.1	2.6	34	<0.01	52
Valle Grande Spring[e]	39	March 1982	14	2630	−12.50	−85.1	2.3 ± 0.8	1.0	2.3	34	...	...
Valle Grande Spring[e]	39	Jan. 1983	14	2630	...	...	1.93 ± 0.13	1.6	1.5	46	<0.05	...
Horseshoe Spring[e]	42	Aug. 1979	14	2420	−12.65	−90.2	...	6.1	6.2	145	0.01	46
Horseshoe Spring[e]	42	Jan. 1983	11	2420	−12.20	−88.9	2.32 ± 0.11	4.0	5.1	151	<0.05	56
				Soda Dam and Jemez Springs Area								
Soda Dam Spring	29	July 1978	47	1930	−10.60	−84.9	...	1480	37	886	11.5	43
Soda Dam Spring	29	April 1980	47	1930	−10.70	−85.4	...	1520	39	1000	15.0	46
Soda Dam Spring	29	Dec. 1980	47	1930	−10.60	−85.2	...	1560	36	1250	13.9	44
Soda Dam Spring	29	March 1982	47	1930	...	...	2.9 ± 0.6	1480	35.5	1490	...	...
Soda Dam Spring	29	Aug. 1982	48	1930	...	...	1.39 ± 0.10	1614	37.5	1455	14.1	...
Soda Dam Spring	29	Jan. 1983	47	1930	−10.35	−84.0	1.35 ± 0.10	1536	34	1458	15.7	47
Soda Dam Spring	29	May 1983	47	1930	...	...	1.62 ± 0.12	1477	35	1488	13.9	42
Grotto Spring	29	July 1978	38	1930	−10.65	−84.6	...	1480	41	834	11.6	38
Hidden Warm Spring	29	May 1979	29	1930	−10.95	−84.9	...	1195	69.1	1400	10.6	44
Hidden Warm Spring	29	March 1982	32	1930	−10.65	−85.1	5.7 ± 1.1	1240	48.3	1370	...	...
Hidden Warm Spring	29	Jan. 1983	32	1930	...	...	3.63 ± 0.18	1294	53	1324	13.4	43
Jemez River, Soda Dam[e]	29	April 1980	5	1930	−13.20	−94.4	...	4.0	9.0	52	0.09	17
Jemez River, Soda Dam[e]	29	March 1982	2	1930	−12.50	−89.5	13 ± 2	19.5	19.8	107	...	...
Jemez River, Soda Dam	29	Jan. 1983	1	1930	...	...	11.8 ± 0.5	10.3	14.7	101	...	54
Jemez River, Soda Dam[e]	29	May 1983	9	1930	−12.00	−85.4	32.6 ± 0.8	11.3	14.2	57	0.11	32
Main Jemez Spring	30	Jan. 1979	55	1890	−10.6	−82.3	...	904	40.9	711	7.9	93
Main Jemez Spring	30	Jan. 1979	36	1890	−10.4	−81.4	...	968	45.4	699	8.0	85
Main Jemez Spring	30	March 1982	46	1890	−10.40	−82.1	1.2 ± 0.4	926	45.5	720	...	...
Travertine Mound Spring	30	Jan. 1979	70	1890	−11.30	−83.6	...	829	36.1	723	7.8	93
Travertine Mound Spring	30	Dec. 1980	72	1890	−11.35	−83.1	...	910	42.4	436	7.0	92
Travertine Mound Spring	30	March 1982	72	1890	...	...	6.7 ± 0.9	869	39.8	715	...	...
Travertine Mound Spring	30	Jan. 1983	73	1890	...	...	2.92 ± 0.13	906	38	697	7.02	90
Buddhist Spring	30	March 1982	43	1890	−11.05	−83.6	10 ± 1	617	35.1	660	...	...
Jemez well at 24 m	30	Jan. 1979	68	1890	−11.3	−84.0	...	705	45.0	642	6.1	70
Jemez well at 152 m	30	Jan. 1979	61	1890	−11.8	−85.9	...	243	38.0	479	2.2	24
Jemez well at wellhead	30	Jan. 1983	73	1890	...	...	3.58 ± 0.17	862	39	686	6.79	89
				Jemez Mountains, Other Areas								
Panorama Spring	31	May 1979	13	2070	−11.80	−86.9	...	21.7	56.7	519	<0.1	62
Panorama Spring	31	March 1982	7	2070	−11.90	−87.4	4.3 ± 1.0	13.1	21.6	480	...	...
Panorama Spring	31	May 1983	9	2070	...	...	1.56 ± 0.09	14.1	26	477	0.08	51
Sino Spring[e,h]	35	Aug. 1979	21	2300	−12.30	−88.0	...	6.0	5.0	82	0.01	78
Sino Spring[e]	35	March 1982	17	2300	−11.90	−87.2	1.1 ± 0.6	3.0	4.8	76	...	...
Sino Spring[e]	35	Sept. 1982	21	2300	...	...	0.29 ± 0.09	3.2	4.0	71	<0.01	...
Indian Valley well[e]	38	Aug. 1979	18	2460	−12.30	−87.1	...	...	...	...	...	...
Indian Valley well[e]	38	Oct. 1979	18	2460	−12.75	−91.1	...	7.3	6.2	137	<0.01	77
Las Conchas Spring[e]	40	Aug. 1979	15	2770	−11.40	−78.2	...	6.9	13.1	11	<0.01	45
Las Conchas Spring[e]	40	May 1983	3	2770	−11.65	−79.3	44.3 ± 1.2	1.5	11.7	29	<0.02	34
Unnamed cold spring[e]	41	Aug. 1979	10	2490	−13.75	−98.2	...	7.4	17.4	90	<0.01	48
Unnamed cold spring[e]	43	Aug. 1979	13	2330	−13.60	−96.5	...	3.5	4.4	57	<0.01	74
Seven Springs[e]	44	Aug. 1979	12	2480	−14.25	−99.1	...	3.6	8.7	49	<0.01	41
Seven Springs[e]	44	May 1983	12	2480	−13.50	−96.2	20.6 ± 0.6	1.9	5.9	54	<0.02	30
Henson's Well[e]	46	Dec. 1980	19	2390	−11.30	−81.2	...	7.7	20.5	305	0.01	62
Unnamed cold spring[e]	48	June 1980	11	2500	−12.60	−88.6	...	5.4	9.9	40	<0.01	30
Unnamed cold spring[e]	49	June 1980	9	2650	−12.75	−90.5	...	5.5	6.0	24	<0.01	34
Unnamed cold spring[e]	50	June 1980	7	2560	−12.55	−87.6	...	9.8	6.7	39	<0.01	38
Apache Spring[e]	51	July 1980	9	2530	−12.25	−85.1	...	8.0	8.3	57	<0.01	58
Turkey Spring[e]	54	July 1980	18	2130	−11.00	−76.0	...	8.0	4.2	103	0.03	61
Turkey Spring[e]	54	May 1983	18	2130	−10.90	−75.7	1.65 ± 0.12	3.6	3.7	106	<0.02	57
				San Ysidro–Jemez Pueblo Area								
Ponderosa Spring	36	Aug. 1979	17	1740	−12.35	−84.6	...	267	193	848	3.16	16
Ponderosa Spring	36	May 1983	15	1740	...	...	2.07 ± 0.08	263	202	862	3.01	15
Canon Spring	37	Aug. 1979	20	1830	−12.35	−86.6	...	71	127	367	0.41	47

TABLE 1. (continued)

Name	Map[a]	Date	Temperature,[b] °C	Emergence Elevation, m	$\delta^{18}O$[c], ‰	dD[c], ‰	^3H, TU	Cl, mg/L	SO_4, mg/L	HCO_3, mg/L	B, mg/L	SiO_2, mg/L
San Ysidro–Jemez Pueblo Area (continued)												
San Ysidro mineral spring	55	May 1983	27	1680	−10.40	−86.6	···	1820	2330	1860	8.32	14
San Ysidro mineral spring	55	May 1983	22	1680	···	···	0.40 ± 0.10	1671	1183	1740	9.77	16
Zia Hot Well	56	Aug. 1979	56	1840	−11.25	−89.8	···	3000	3740	1440	6.52	30
Zia Hot Well	56	April 1980	54	1840	−12.55	−89.0	···	3210	3430	1068	7.41	33
Zia Hot Well	56	Feb. 1983	53	1840	···	···	0.05 ± 0.06	2984	3338	1398	7.8	33
Unnamed well	57	Aug. 1979	21	1920	−13.65	−101.4	···	···	···	···	···	···
Salt Spring	58	Aug. 1979	29	1690	−10.20	−84.9	···	2380	702	1870	9.2	···
Log Spring	59	Aug. 1979	29	2190	−12.25	−87.7	···	···	···	···	···	···
Owl Spring	60	Aug. 1979	18	1760	−12.15	−86.2	···	27.8	33.8	311	0.12	22

[a]Map numbers correspond to those on Figure 3.
[b]Temperature was measured to the tenth of a degree and rounded off to the nearest unit.
[c]Precision of stable isotope analyses: $\delta^{18}O = \pm 0.15‰$; $dD = \pm 0.5‰$.
[d]Elevation for Los Alamos wells correspond to the static water level averaged for the sampling year.
[e]Waters used to calculate the local meteoric water line.
[f]Chemical constituents are from water sampled in September 1980.
[g]Chemical constituents are from condensed steam sampled in March 1982.
[h]Recalculated values before flashing (except ^3H); Baca 13 data from *Goff et al.* [1985]; other stable isotope data from *Truesdell and Janik* [this issue]; other tritium data from *White* [this issue]; other chemical data from P. Trujillo and D. Counce, LANL.
[i]Data from *Goff et al.* [1982] and *Grigsby et al.* [1984].
[j]GT-2 well production fluid from the 280-day experiment.
[k]Chemical constituents are from water sampled in December 1980.

Type B: Thermal Meteoric Waters

Waters of this group consist of the Bathhouse, San Antonio, Spence, and McCauley warm springs, which issue from the western ring fracture zone of Valles caldera. Chemically and isotopically, they resemble dilute cold groundwaters of type A (Table 1 and Figure 4), although thermal meteoric waters generally contain higher SiO_2 concentrations. All thermal meteoric waters discharge from fractured rhyolite of the Valles moat zone or, near the Jemez fault zone, from the contacts of the moat rhyolites with underlying Paleozoic red beds.

Type C: Acid-Sulfate Waters

Thermal and nonthermal acid-sulfate waters issue from springs and mud pots in the Sulphur Springs area on the

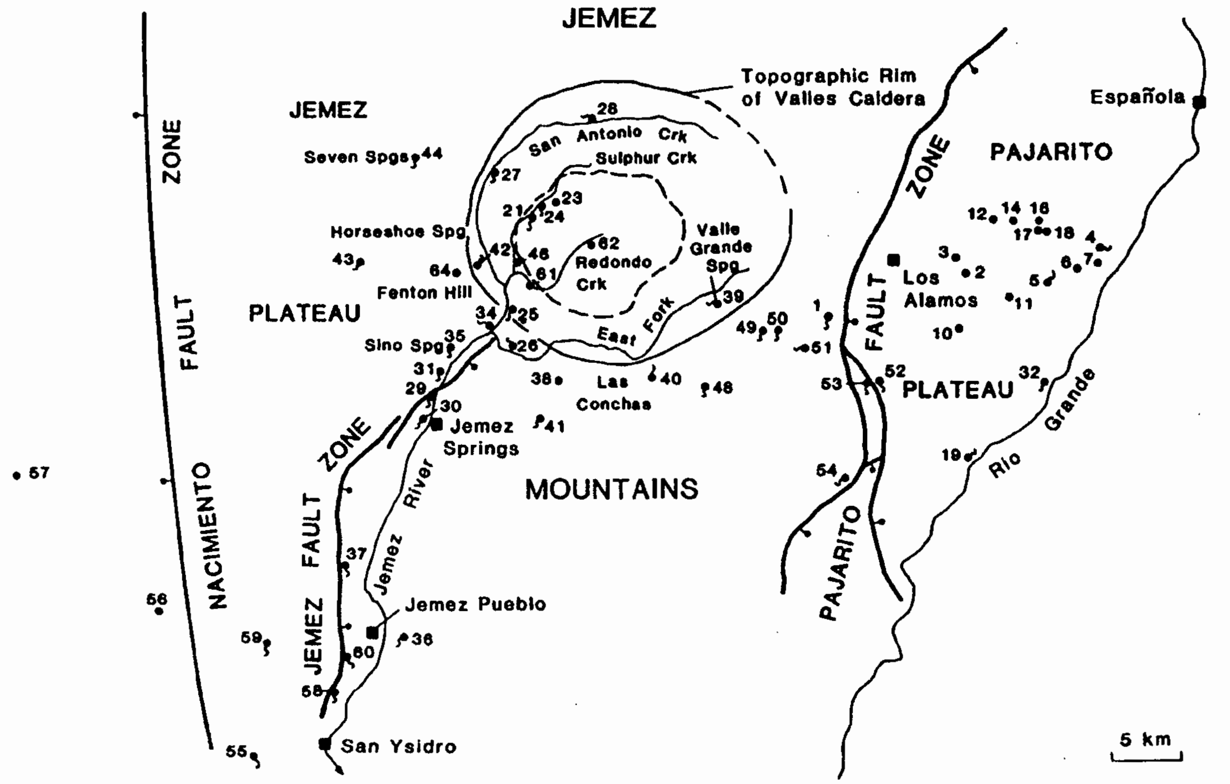

Fig. 3. Location map of Jemez Mountains region showing the locations of all springs, wells, and streams listed in Table 1; dashed line inside Valles caldera represents the base of the central resurgent dome. Numbers correspond to the spring and well locations shown on Figure 2 of *Goff et al.* [1982].

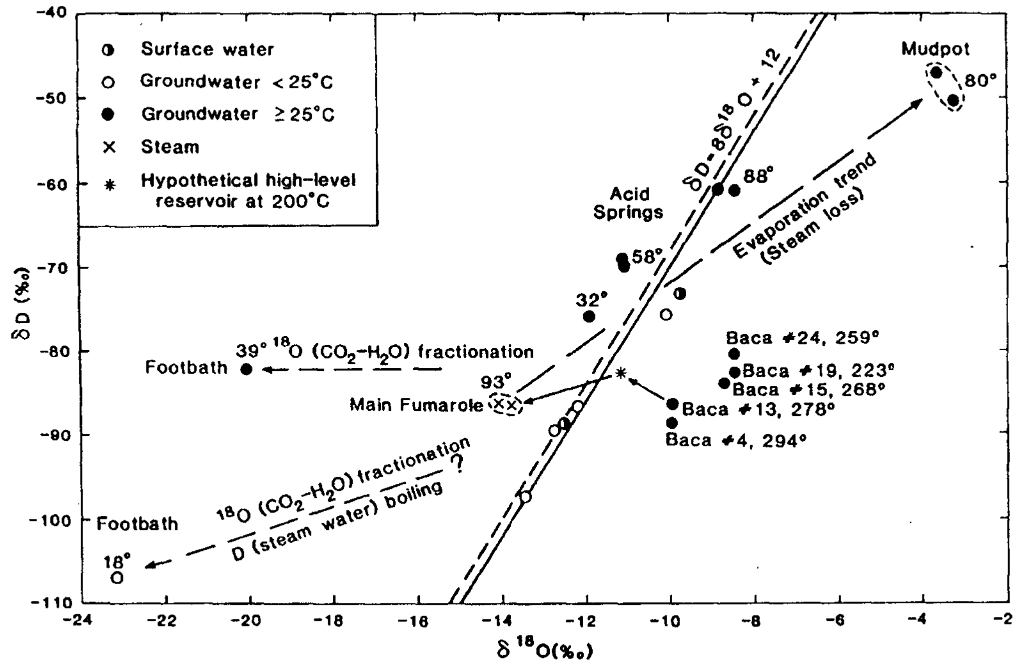

Fig. 8. Plot of δD versus $\delta^{18}O$ for the waters of the Sulphur Springs area and Baca reservoir fluids, Valles caldera, New Mexico (modified from *Goff et al.* [1985]); legend same as Figure 4.

thermal system because of their areal extent and accumulations of sediment and because most streams draining the surrounding heights flow into these valleys. Valle Grande Spring, for example, has a deuterium value of −85‰ or about midway between the range in deuterium values of the Baca well samples. It is clear, however, that the entire recharge zone consists of the 22-km-diameter caldera depression and the highlands immediately surrounding the depression.

Origin of Surface Thermal Waters

Perhaps the most common modification of isotopic composition observed in high-temperature geothermal systems is the positive oxygen 18 enrichment caused by high-temperature isotopic exchange between water and country rock [*White*, 1968]. These rock-water interactions result in shifts of 1–15‰ in $\delta^{18}O$ depending on temperature, relative volumes of water and rock, original isotopic compositions of water and rock, and residence time of fluid in the reservoir rocks. Other important modifications in isotopic composition can occur in geothermal systems as described by *Truesdell et al.* [1977], *Truesdell and Hulston* [1980], and *Fritz and Frape* [1982]. Briefly summarized, these additional processes include (1) mixing between near-surface waters and various hydrothermal and mineral fluids, (2) boiling in the upper part of the hydrothermal reservoir, (3) nonequilibrium surface evaporation in springs, hot pools, and mud pots, and (4) isotopic exchange between water and another fluid, such as CO_2, rather than rock. Hot spring fluids in the Valles caldera system display all of these isotopic modifications.

Although the deep geothermal fluids in the Baca reservoir exhibit the typical oxygen 18 shifts observed in geothermal systems (about 2.5‰ in Baca 13 and 3.5‰ in Baca 15), isotopic variations between different wells and their significance to reservoir dynamics are discussed by *Truesdell and Janik* [this issue] and *White* [this issue]. Our object in this section is to discuss the isotopic variations of the surface thermal fluids while making only general reference to their relationship to the deep hydrothermal system.

Sulphur Springs Area

Stable isotope variations in the acid-sulfate waters of Sulphur Springs (Figure 8) have been recently discussed by *Goff et al.* [1985]. Most variations can be attributed to processes of subsurface boiling and surface evaporation. Based on geochemical constraints and drilling data, the Sulphur Springs system consists of a vapor zone 600 m thick that overlies a boiling hydrothermal fluid of about 200°C. The oxygen isotope composition of steam in the vapor zone is best represented by steam from the Main Fumarole (93°C, about the boiling point at 2600 m above sea level), which is isotopically depleted in oxygen 18 relative to local meteoric water by roughly 2‰. *Goff et al.* [1985] pointed out that if the deep hydrothermal fluid beneath Sulphur Springs is isotopically identical to the composition of Baca 13 fluid (278°C), it is impossible to generate this steam by single-stage boiling at known reservoir temperatures. Instead, the steam could result from boiling of rapidly percolating, near-surface groundwater at 200°C or by at least two stages of boiling of deep hydrothermal fluid. The isotopic composition of a hypothetical, intermediate-level reservoir at 200°C is shown in Figure 8. Because Baca 15 is even more enriched in oxygen 18 relative to Sulphur Springs steam, it is impossible to generate that steam from Baca 15 by any reasonable boiling process at known reservoir temperatures.

Once steam reaches the surface at Sulphur Springs, it generally condenses into hot springs and mud pots that bubble freely due to residual noncondensible gases such as CO_2 and H_2S. Data for noncondensed steam at the Main Fumarole and for highly evaporated water at Men's Bathhouse Mudpot lie along the ends of an "evaporation line" (Figure 8) that results from kinetic isotopic fractionation during surface evaporation.

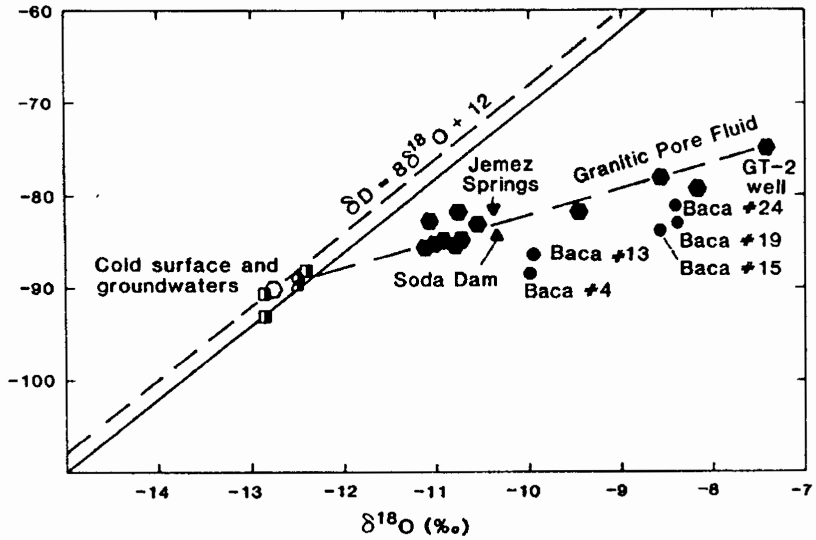

Fig. 9. Plot of δD versus $\delta^{18}O$ for the fluids produced from the 280-day circulation experiment, Fenton Hill, compared with Baca reservoir fluids and springs in San Diego Canyon, Valles caldera, New Mexico; heavy isotope enrichment of the Precambrian pore fluid is indicated (legend same as Figure 4).

One relatively cool spring (Footbath Spring, 18°–39°C) displays an oxygen 18 depletion of about 10‰ (Figure 8). Such depletion has been seldom described in the geothermal literature. *Fritz and Frape* [1982] suggested that oxygen isotope exchange between CO_2 and H_2O could explain the extreme reverse shifts observed in cold, carbonated brines of the Canadian Shield. C. Fouillac (Bureau de Recherches Geologiques et Minieres, personal communication, 1984) has offered a similar explanation for the isotopically depleted carbonated mineral springs of Ardeche, France. Because CO_2 concentrates oxygen 18 relative to H_2O by 30–40‰ at temperatures between 90° and 25°C [*Friedman and O'Neil*, 1977], the amount of oxygen 18 depletion in CO_2-charged water should increase at lower temperatures in pools having little discharge of water but vigorous discharge of CO_2, as in Footbath Spring. The $\delta^{18}O$ of CO_2 at Footbath Spring is 12.60‰ [*Goff et al.*, 1985], whereas the $\delta^{18}O$ of H_2O at Footbath Spring is −20.4 to −23.15‰ (Table 1).

Hot Dry Rock Fluid in Precambrian Basement

Fluids produced from HDR well GT-2 during the 280-day circulation experiment [*Grigsby et al.*, 1984] were mixtures of dilute surface water ("make-up" water) and 200°C pore fluid brine (or "formation" water) trapped in Precambrian granodiorite at 3 km depth. A linear mixing pattern between the two components is clearly displayed in the plot of δD versus $\delta^{18}O$ (Figure 9). As discussed by *Grigsby et al.* [1984], an uncontaminated sample of pore fluid was never obtained; thus

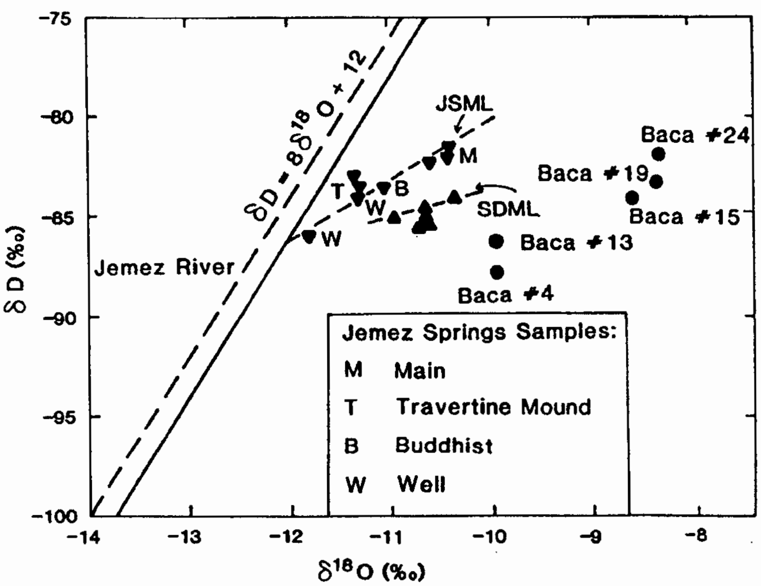

Fig. 10. Plot of δD versus $\delta^{18}O$ for the thermal and nonthermal waters at Soda Dam and Jemez Springs, San Diego Canyon, New Mexico, showing local mixing trends; data from Baca reservoir fluids are shown for comparison (legend same as Figure 4).

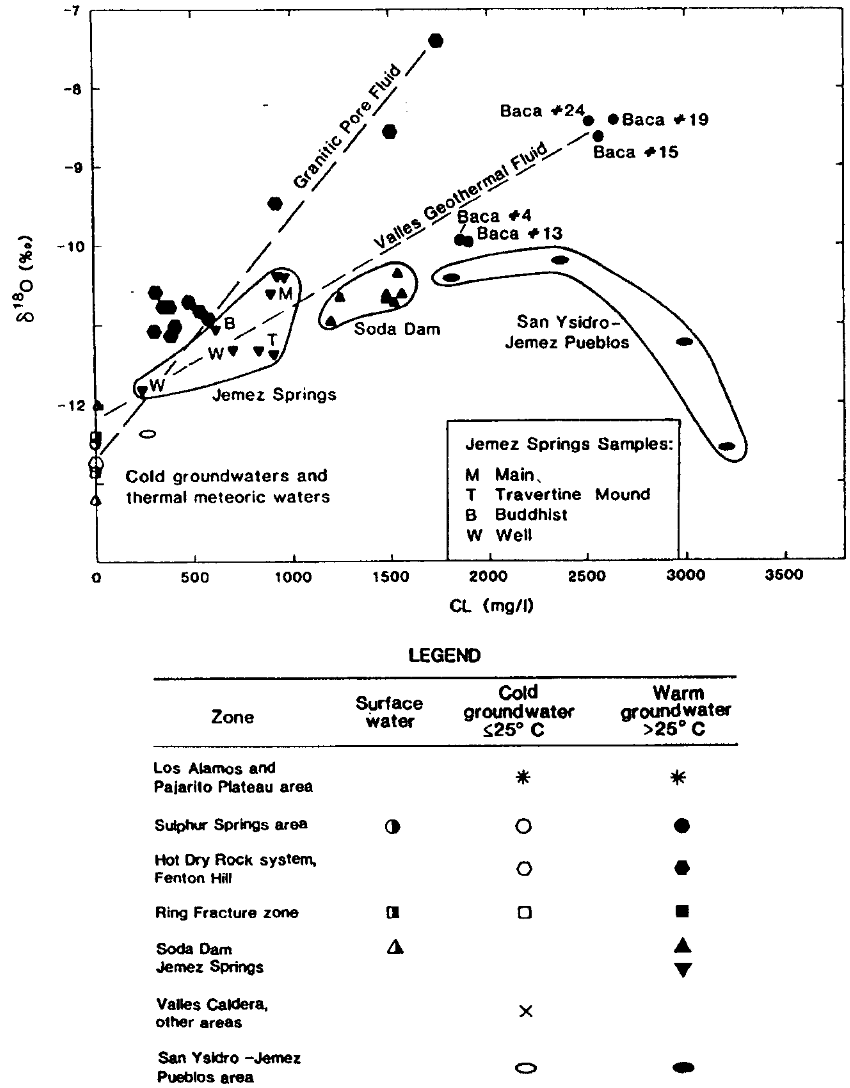

Fig. 4. Plot of $\delta^{18}O$ versus chloride showing compositions of various thermal and nonthermal waters of the Valles caldera–Jemez Mountains region, New Mexico; acid-sulfate waters are not shown for clarity but would plot along the $\delta^{18}O$ axis; mixing relations between surface water, granitic pore fluid brine, and Baca geothermal fluid are shown by dashed lines.

western flank of the resurgent dome inside Valles caldera [*Goff et al.*, 1985]. These waters have formed by surface mixing of cold groundwater with SO_4 from oxidation of H_2S and with condensed steam from a boiling geothermal system at depth. Chemically, SO_4 is the major anion, trace elements such as B are low, and *p*H ranges from 0.6 to 5, while stable isotope ratios show extreme variability. Gases and thermal fluids rise along postcaldera faults and fault intersections [*Goff and Gardner*, 1980] and have altered the intracaldera tuffs and rhyolites to argillic and advanced argillic mineral assemblages [*Charles et al.*, this issue].

Type D: Deep Geothermal and Derivative Waters

Geothermal fluid from the deep, liquid-dominated reservoir was first discovered by Union Oil Company during exploration drilling in the Redondo Creek graben. Reservoir rocks consist primarily of caldera fill tuffs and precaldera andesites. Fluid flow is controlled by fracture and subordinate stratigraphic permeability [*Dondanville*, 1978; *Hulen and Nielson*, 1982]. The neutral chloride fluid (7000 mg/L total dissolved solids (TDS)) ranges from about 220° to 300°C, contains significant trace element concentrations, and is isotopically enriched in oxygen 18 [*White et al.*, 1984; *White*, this issue; *Truesdell and Janik*, this issue]. The reservoir has been tapped by several production wells, but the volume of the reservoir was not considered large enough to justify construction of a commercial 50-MW_e (megawatts electric) power plant [*Kerr*, 1982].

Deep reservoir fluid is distinct from other thermal and mineral waters of the Valles region in a plot of Cl versus $\delta^{18}O$ (Figure 4). Because of the many samples that were obtained from the Baca wells in 1982, several authors have recently noted that two types of hydrothermal fluid circulate in the Redondo Creek fault zone [*White et al.*, 1984; *Smith and Kennedy*, 1985; *Truesdell and Janik*, this issue; *White*, this issue]. The first fluid represented by Baca 4 and 13 has lower Cl, higher $^3He/^4He$, higher temperature, and deuterium values

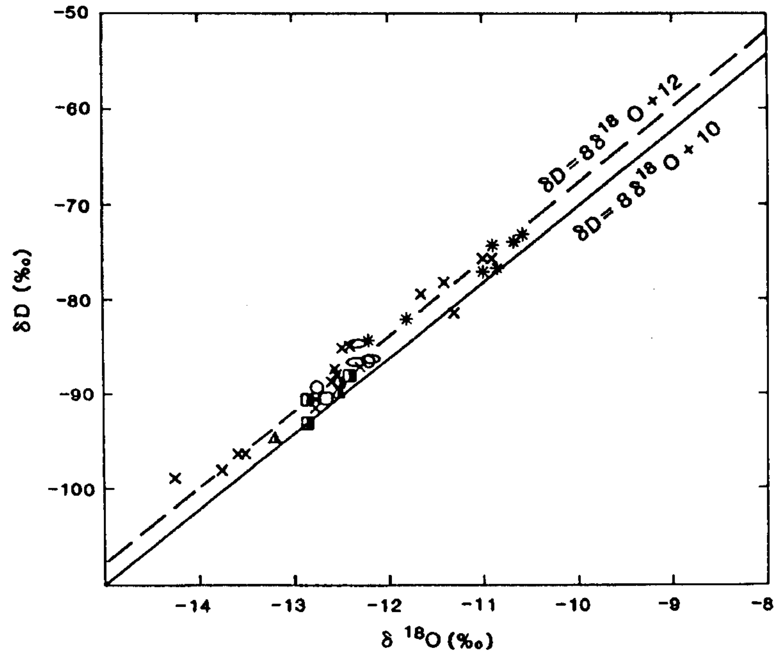

Fig. 5. Plot of δD versus $\delta^{18}O$ for selected cold groundwaters and surface waters, Jemez Mountains region, New Mexico; the Jemez Mountains meteoric line (dashed line) is compared to the world meteoric line (solid line) of *Craig* [1961] (legend same as Figure 4).

averaging about −87‰. The second fluid represented by Baca 24 and 15 has higher Cl, lower $^3He/^4He$, lower temperature, and deuterium values averaging about −83‰. Baca 19, which has had 4 times as much water injected into it (mostly flashed water from Baca 4) as it has produced, contains a fluid that superficially resembles Baca 15 and 24 but may be very contaminated and mixed. Regardless of Baca 19, the various data indicate that the two Baca fluids have different flow paths and histories in the reservoir, but many controversies remain regarding the development of the fluids (compare discussions of *Smith and Kennedy* [1985], *Truesdell and Janik* [this issue], and *White* [this issue]).

Outside the caldera, two groups of thermal springs (Soda Dam and Jemez Springs) issue from the Jemez fault zone in San Diego Canyon. Based on geochemical and isotopic similarities to the deep fluids inside the caldera and their geologic and structural setting, these hot springs apparently originate by lateral, subsurface flow of hydrothermal fluid along strands of the Jemez fault zone and by local mixing with dilute waters [*Trainer*, 1974; *Goff et al.*, 1981]. Slight geochemical variations between the two groups of springs, particularly their inverse relationship of Cl content and temperature, suggest that different flow paths and fluid histories are required to produce these two groups of waters [*Goff et al.*, 1981; *Trainer*, 1984; *White et al.*, 1984].

Type E: Precambrian Pore Fluid

Another type of deep geothermal fluid has been recognized in relatively impermeable Precambrian basement rocks at depths of 3000 m beneath the Fenton Hill hot dry rock (HDR) site on the west flank of the caldera. This fluid, which is the most saline ever observed in the Jemez Mountains region (as high as 20,000 mg/L TDS), is apparently a pore fluid whose origin is not completely resolved [*Grigsby et al.*, 1984]. Due to injection of cold, dilute water into the HDR system during circulation tests, uncontaminated samples of the pore fluid have never been obtained (Table 1 and Figure 4). The pore fluid brine is recognized as a concentrated pulse of thermal water that is ejected from the HDR system when circulation starts. Isotopically, the pore fluid resembles the deep geothermal waters inside Valles caldera, but differences in the trace element chemistry of conservative components strongly suggest a unique origin. Possibly, the pore fluid brine represents the formation waters that reside in the impermeable, conductive regime between the magmatic heat source and the convecting hydrothermal system [*Grigsby et al.*, 1984].

TABLE 2. Selection of Waters From Table 1 for Determination of Recharge Elevation

Map	Name	Elevation, m: Emergence	Elevation, m: Recharge	$\delta^{18}O$ ‰	δD, ‰	Number of Analyses
1	Gallery Spring	2440	2590	−12.20	−84.3	1
24	Short Canon Spring	2610	2770	−12.50	−87.8	2
34	East Fork Jemez River	2070	2740	−12.40	−88.0	1
39	Valle Grande Spring	2630	2740	−12.45	−85.1	2
46	Henson's Well	2390	2440	−11.30	−81.2	1
48	Unnamed cold spring	2500	2770	−12.60	−88.6	1
50	Unnamed cold spring	2560	2870	−12.55	−87.6	1
51	Apache Spring	2530	2710	−12.25	−85.1	1
54	Turkey Spring	2130	2260	−10.95	−75.9	2

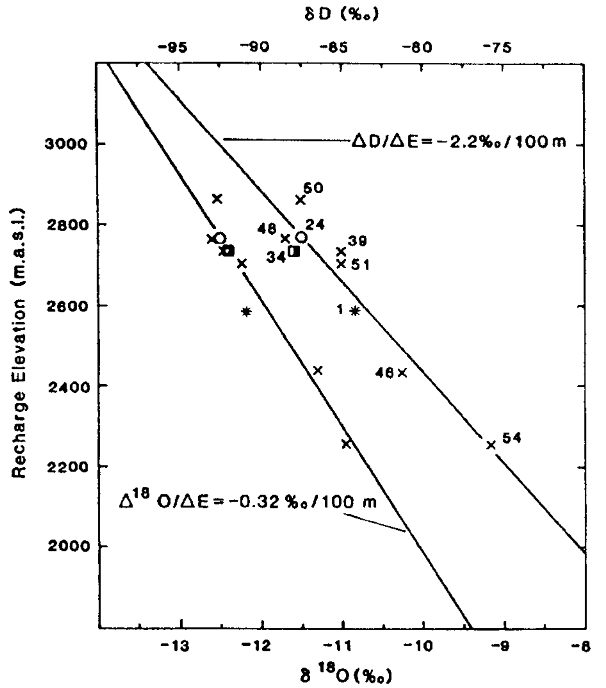

Fig. 6. Plot of δD and $\delta^{18}O$ as functions of recharge elevation for a selection of meteoric waters, Jemez Mountains, New Mexico; numbers identify the map numbers in Table 1 and Figure 3 (legend same as Figure 4).

Type F: Mineral Waters of San Ysidro Region

Mineral waters of variable temperature discharge along the southwestern side of the Nacimiento fault zone, 35 km west of Valles caldera (Figure 3). Most of the springs are <30°C, but the 250-m-deep Zia Hot Well drilled in 1926 (Kaseman 2 oil test) encountered copious amounts of mineralized water at 54°C. This well was never successfully plugged and flows to this day. Although superficially resembling deep geothermal waters of Valles caldera, distinct geochemical and isotopic characteristics suggest that the San Ysidro waters have a different origin (Table 1 and Figure 4) [*Goff et al.*, 1981]. Because these mineral waters circulate in Mesozoic to Paleozoic rocks of the San Juan Basin west of the Nacimiento Fault, it is most likely their chemistry results from deep circulation and solution of evaporite minerals in the basin.

STABLE ISOTOPES (OXYGEN 18 AND DEUTERIUM)

Derivation of Local Meteoric Water Line

To determine the local meteoric water line of the Valles caldera region from our stable isotope data (Table 1), a set of empirical conditions were defined in order to select representative cold, dilute meteoric waters. These conditions require that the waters have (1) temperature <25°C, (2) deuterium excess, $d \geq 9‰$ (where $d = \delta D - 8\delta^{18}O$), (3) low mineralization (TDS < 500 mg/L), and (4) significant flow rate (>1 L/min). When these criteria are applied to the cold waters marked by footnote e in Table 1, 36 samples from 28 different locations yield an equation for the Jemez Mountains meteoric water line (Figure 5) of

$$\delta D = 8\delta^{18}O + 12 \quad (1)$$

Extreme d values of 9.2‰ and 14.9‰ of some waters used in the calculation cannot be readily explained but are within ±2 standard deviations of equation (1). For oceanic precipitation the deuterium excess is typically around 10‰, whereas higher values such as 12‰ are expected for inland precipitation [*Fontes*, 1980]. Strong evaporative conditions such as closed-basin lakes in desert climates may cause d values to exceed 20‰ [*Gat*, 1980]. Thus the d value of 12‰ obtained for the Jemez Mountains region is entirely consistent with its inland location 1200 km away from the nearest ocean.

Relation Between Recharge Elevation and Isotopic Composition

The determination of the elevation of groundwater recharge in mountainous areas is usually not difficult because of the pronounced gradient of isotopic composition of precipitation with altitude [*Fontes*, 1980]. Once determined, this relation can be used to evaluate the source regions and circulation paths of groundwaters, such as those that recharge geothermal reservoirs. Of the cold, dilute waters used to calculate the Jemez Mountains meteoric water line above, seven springs, one well, and one river were selected for further evaluation because their recharge elevations could be estimated with great reliability based on topographic and geologic considerations (Table 2). From these nine samples, linear equations and correlation coefficients, r, have been calculated for the relations between δD and $\delta^{18}O$, and their respective recharge elevations E (meters) (Figure 6) in which

$$E(\mathrm{m}) = -44.9(\delta D) - 1154 \; (r = -0.96) \quad (2)$$

and

$$E(\mathrm{m}) = -314(\delta^{18}O) - 1161 \; (r = -0.97) \quad (3)$$

The corresponding isotopic gradients to these equations are

$$\Delta D/\Delta E = -2.2‰ \text{ per } 100 \text{ m} \quad (4)$$

and

$$\Delta^{18}O/\Delta E = -0.32‰ \text{ per } 100 \text{ m} \quad (5)$$

The Valles oxygen 18 gradient is very similar to those calculated from studies of other mountainous environments. For example, *Moser and Stichler* [1970] obtained a value of about −0.3‰ per 100 m on Mount Kilimanjaro, Africa, and the European Alps; *Vuataz* [1982] calculated a value of −0.27‰ per 100 m from the north and west sections of Switzerland; *Zuppi et al.* [1974] obtained a range of −0.28 to −0.34‰ per 100 m for central Italy; and *Payne and Yurtsever* [1974] calculated −0.26‰ per 100 m in Nicaragua. These isotopic gradients are relatively constant in spite of climatic and latitude differences.

Although the correlation coefficients of equations (2) and (3) are identical, the deuterium gradient is preferred in most cases for the determination of recharge elevation. This is because oxygen 18 will almost always undergo stronger isotopic modification after infiltration than deuterium (exchangable hydrogen is in low abundance in most rocks). If evaporation of a water sample is suspected, it is necessary to recalculate the isotopic composition before recharge elevation is estimated. In Figure 7, all analyzed deuterium values for waters listed in Table 1 (except Baca well waters) have been plotted against their emergence elevations and compared with the corresponding isotopic gradient.

Waters falling to the right side of the deuterium gradient line (Figure 7) reflect isotopic shifts due to boiling, mixing, and rock-water interactions. Hot acid-sulfate springs from Sulphur Springs have suffered the effects of subsurface boiling and sur-

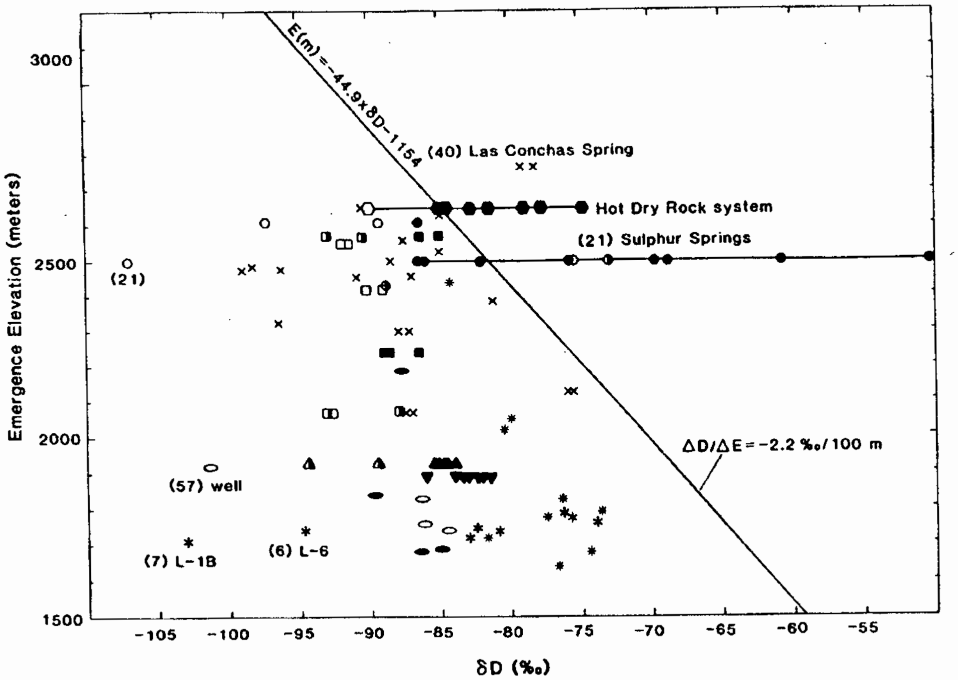

Fig. 7. Plot of elevation versus deuterium content for various types of thermal and nonthermal waters of the Jemez Mountains region, New Mexico; numbers refer to map numbers in Table 1 and Figure 3 (legend same as Figure 4).

face evaporation [*Goff et al.*, 1985]. The fluids produced by circulation of the hot dry rock system at Fenton Hill show the effects of mixing cold groundwater (make-up water) with geothermal pore fluid. The precise isotopic composition of the pore fluid is not known but the least contaminated sample displays the strongest oxygen 18 enrichment and mineralization. Presumably, the long-term residence in Precambrian rocks and rock-water interactions that have affected this deep pore fluid would preclude accurate estimation of its recharge elevation (if it is indeed meteoric in origin). One cold dilute spring, Las Conchas Spring, has been collected twice, and both data points fall to the right of the line for reasons that are unknown.

A few waters fall to the extreme left of the line (Figure 7), suggesting that their recharge elevations average from 3000 to 3500 m. Such high average recharge elevations do not seem plausible because Redondo Peak, the highest point inside Valles caldera culminates at 3430 m. Two wells (L-1B and L-6) from the Los Alamos well field on the Pajarito Plateau that discharge warm water from a large confined aquifer, exemplify this problem. Their deuterium contents are −94.7 and −103.0‰, respectively; thus their estimated mean recharge elevations are 3100 and 3470 m, which seems too high for the Jemez Mountains. Because *Purtymun and Johassen* [1974] have demonstrated that these wells tap an artesian aquifer at depths below 600 m in westward dipping Santa Fe Group rocks, *Goff and Sayer* [1980] suggested that this aquifer may be recharged from the Sangre de Cristo Mountains bordering the eastern side of the Rio Grande rift. These mountains reach elevations of 4000 m; thus they seem much more realistic as a source for the two isotopically depleted well waters.

Another group of three cold dilute springs (41, 43, and 44, Table 1) have apparent recharge elevations of about 3200 m even though the mountains immediately above them barely reach these elevations. These springs were sampled during the spring and summer months; their temperatures are ≤ 10°C, and their mineralization is very low (< 200 mg/L TDS). One of the springs has a relatively high tritium content (21 TU), implying a short underground transit time, perhaps a few months. Apparently, the samples we collected represent snowmelt infiltration.

Recharge Elevation of Valles Geothermal Reservoir

Stable isotope analyses of five Baca geothermal wells tapping the hydrothermal reservoir beneath the Redondo Creek area of Valles caldera are listed in Table 1. The analyses are corrected for steam flash. Temperatures of produced fluids range from about 225°C to 300°C. Additional stable isotope data from the Baca wells can be found in the work by *Truesdell and Janik* [this issue] and *White* [this issue]. The deuterium contents of the five wells vary from about −82 to −90‰ indicating from equation (2) a range in average recharge elevation of 2530–2890 m for the Baca wells.

Several authors have claimed that the depression of Valles caldera and the mountains around its rim and within serve as recharge zones for the deep geothermal system [*Bodvarsson et al.*, 1980; *Faust et al.*, 1984; *Goff et al.*, 1985]. The stable isotope data and calculated values of average recharge elevation are entirely consistent with these views. The floor of Valles caldera ranges in elevation from about 2400 m in the southwest where the Jemez River exits to about 2680 m near the headwaters of San Antonio Creek. The northern and eastern rims of Valles caldera exceed 3100 m, whereas the summit of Redondo Peak inside the caldera is 3430 m. Examination of the deuterium contents of cold springs, wells, and creeks inside the caldera (Table 1 and Figure 3) shows they range from −73‰ (Sulphur Creek) to −97‰ (Alamo Canyon Pool) with most values between −82 to −90‰.

We speculate that the broad moat valleys in the north and east sectors of the caldera, Valle San Antonio, Valle Toledo, and Valle Grande act as principal recharge zones to the geo-

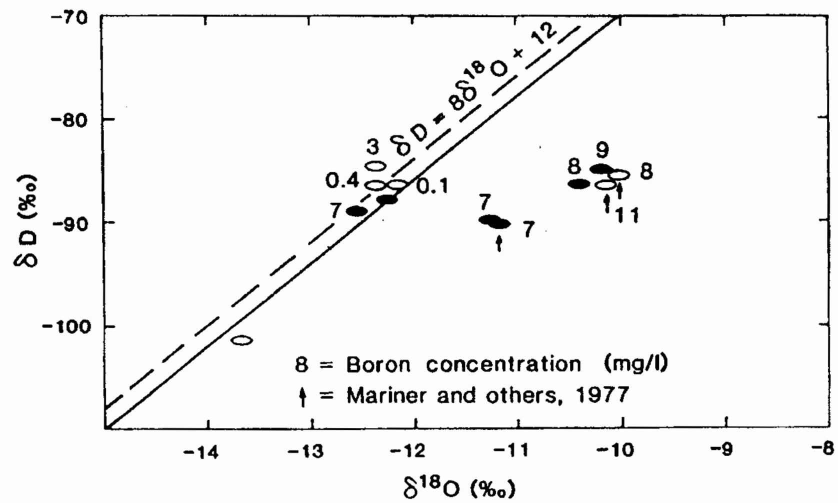

Fig. 11. Plot of δD versus $\delta^{18}O$ for the cold and mineral waters of the San Ysidro–Jemez Pueblo areas, New Mexico (legend same as Figure 4).

the most oxygen 18 enriched fluid from GT-2 can only approximate the isotopic composition of the end-member brine. The magnitude of the oxygen 18 shift (3‰) indicates long-term residence time in the granitic rocks and/or high-temperature, rock-water interaction presumably driven by the magmatic heat source beneath Valles caldera. The deuterium value of the most isotopically enriched GT-2 fluid is similar to deuterium values of Jemez Mountains meteoric waters emerging at lower elevations.

Although superficially similar, the HDR pore fluid brine is distinct from the Baca hydrothermal fluids in being more enriched in both oxygen 18 and deuterium and in having different ratios of conservative ions such as B/Cl [*Grigsby et al.*, 1984]. This suggests to us that the HDR and Baca fluids have unique origins. Perhaps the pore fluid brine is representative of fluids generated by metamorphic processes in the conductive regime between the Valles magma chamber and the overlying hydrothermal system.

Soda Dam and Jemez Springs

Numerous investigators have suggested that the mineralized hot springs in San Diego Canyon are derived from the hydrothermal system in Valles caldera [*Dondanville*, 1971; *Trainer*, 1974, 1975; *Trainer and Lyford*, 1979]. In support of this hypothesis, *Goff et al.* [1981] and *Pearson and Goff* [1981] presented geologic, geochemical, and geophysical arguments backed up by drill hole data from the Jemez Springs geothermal well showing that a hydrothermal plume was leaking out of Valles caldera down the Jemez fault zone. In addition, *Goff and Grigsby* [1982] demonstrated that thermal fluids were also flowing through permeable Paleozoic rocks, above the Precambrian basement, adjacent to the fault zone. *Faust et al.* [1984] used a three-dimensional finite element code based on this model and upon balances between heat content and mineral load in an attempt to show that the thermal plume contributed to the base flow of the Jemez River. In a recent synthesis, *Trainer* [1984] calculated dilution factors, estimated the volume of thermal water entering the Jemez River and described the underground conduits of thermal flow. Several authors have noted the very linear mixing trends obtained by plotting Na, Li, B, and Br versus Cl for hot spring waters in San Diego Canyon and Baca well waters [*Goff et al.*, 1981; *White et al.*, 1984; *White*, this issue]. Because of a negative correlation between chloride content and discharge temperature, *Goff et al.* [1981] and *Trainer* [1984] have argued that the two groups of springs at Soda Dam and Jemez Springs, have unique flow paths and mixing conditions.

Isotopic compositions of the spring groups at Soda Dam and Jemez Springs, and of the five Baca wells, are plotted on Figure 10. Isotope compositions of the many waters at Jemez Springs fall on a well-defined mixing line (labeled JSML, Figure 10) in which Main Jemez Spring is at one end and meteoric water is at the other. Isotope values from the two springs at Soda Dam lie on a poorly defined mixing line (SDML, Figure 10). Because the isotopic compositions and mixing trends of the two groups of springs are distinct, their waters must have evolved from the Valles geothermal fluids by mixing with meteoric fluids of different isotopic composition. It is impossible, however, based on stable isotope data to choose which Baca well fluid is most representative of the hot end-member fluid in the geothermal system because the well fluids and spring waters do not define any mixing line. To generate the isotopic trend defined by the Jemez Springs mixing line from any Baca well fluid would require three-component mixing with two meteoric fluids at the least. The spread in all the data allows other mixing scenarios as well. Although geologic, geophysical, and some geochemical data show a genetic link between Soda Dam and Jemez Springs waters and hydrothermal fluids in the caldera, stable isotope data show that waters from each spring group have evolved from possibly different hot end-member fluids by mixing with different meteoric waters.

San Ysidro–Jemez Pueblo Area

Although roughly 25 mineral springs discharge in this area, they have not been sampled extensively due to lack of access. From available data, mineralized groundwaters in this area do not possess the same geochemical characteristics as the thermal fluids in San Diego Canyon derived from the hydrothermal system in Valles caldera. In particular, ratios of conservative species (i.e., B/Cl, Na/Cl) are significantly different and the concentration of SO_4 is much higher in the San Ysidro waters. A significant isotopic shift is observed for these mineralized fluids (Figure 11), but no correlation exists between isotopic composition and temperature or total mineralization. Only B and HCO_3 concentrations seem to increase as

TABLE 3. Annual Tritium Concentration of the Colorado River and Precipitation in Southwestern United States.

	Colorado River							Precipitation																	
	Cisco (Utah)			Imperial Dam (Ariz.-Cal.)			Colorado River	Salt Lake City (Utah)			Denver (Colorado)			Flagstaff (Arizona)			Albuquerque (New Mexico)			Socorro (New Mexico)			Mount Withington (New Mexico)			Precipitation
Year	m	sd	n	m	sd	n	Average	m	sd	n	m	sd	n	m	sd	n	m	sd	n	m	sd	n	m	sd	n	Average
1956		...			...		...		...			...			...			...		55	—	1		...		55
1957		...			...		...		...			...			...			...		108	70	8		...		108
1958		...			...		...	534	304	4		...			...		365	184	4	324	154	5	125	0	2	337
1959		...			...		...		...			...			...		2190	933	2	534	430	3	202	99	3	975
1960		...			...		...		...			...			...			...		228	49	5	257	43	5	243
1961	142	16	2		...		142	187	161	4	258	144	4		...		129	81	2	246	121	6	261	278	6	216
1962	457	—	1		...		457	2113	575	4	428	237	2	945	521	11	483	213	7	671	731	8	610	346	6	875
1963	1084	319	9		...		1084	3267	2661	11	2874	2685	10	1509	586	10	2000	838	10	1554	976	6	5797	7782	2	2834
1964	1137	590	12	330	—	1	734	2228	1554	11	2972	2039	11	1369	1077	11	2029	1309	10	1262	1612	8	623	479	4	1747
1965	816	249	12	442	37	11	629	914	704	12	1258	779	10	649	532	12	587	321	11	702	325	2		...		822
1966	691	73	12	697	116	12	694	584	294	10	879	371	8	615	298	4	327	191	9		...			...		601
1967	496	36	9	717	92	12	607	446	143	7	431	188	4		...		220	104	4		...			...		366
1968	398	44	10	615	54	12	507	301	81	4	294	90	4	115	—	1	206	113	4		...			...		229
1969	362	45	12	530	66	12	446	297	197	11	338	151	2	178	106	8	199	106	4		...			...		253
1970	303	23	12	428	47	12	366	230	132	11		...		108	54	11	234	130	4		...			...		191
1971	264	21	12	347	23	12	306	216	149	12		...		128	86	9	238	174	4		...			...		194
1972	212	23	11	291	24	11	252	96	41	11		...		60	38	7	74	45	3		...			...		77
1973	178	17	10	250	25	12	214	86	34	12		...		136	—	1	69	27	4		...			...		97
1974	152	21	11	211	11	11	182	107	88	10		...		69	14	5	50	18	2		...			...		75
1975	137	11	10	181	13	12	159	72	42	12		...			...		74	48	3		...			...		73
1976	119	13	12	151	16	12	135	53	22	9		...			...		60	47	3		...			...		57
1977	108	6	12	132	6	12	120	61	26	11		...			...		50	23	4		...			...		56
1978	102	5	10	114	15	12	108	100	60	7		...			...		50	24	4		...			...		75
1979	84	8	12	97	7	9	91	40	—	1		...			...		23	5	3		...			...		32
1980	71	7	12	83	4	11	77	31	16	10		...			...		30	17	4		...			...		31
1981	72	4	12	79	5	11	76	47	27	9		...			...		35	16	4		...			...		41
1982	57	6	11	69	1	4	63	24	10	12		...			...		34	31	4		...			...		29
1983	54	—	1		...		54	15	1	2		...			...			..			...			...		15

Annual tritium concentrations (in TU) are computed with data obtained from a data bank of the U.S. Geological Survey [*WATSTORE*, 1983]; m, mean; sd, standard deviation; n, number of analyses.

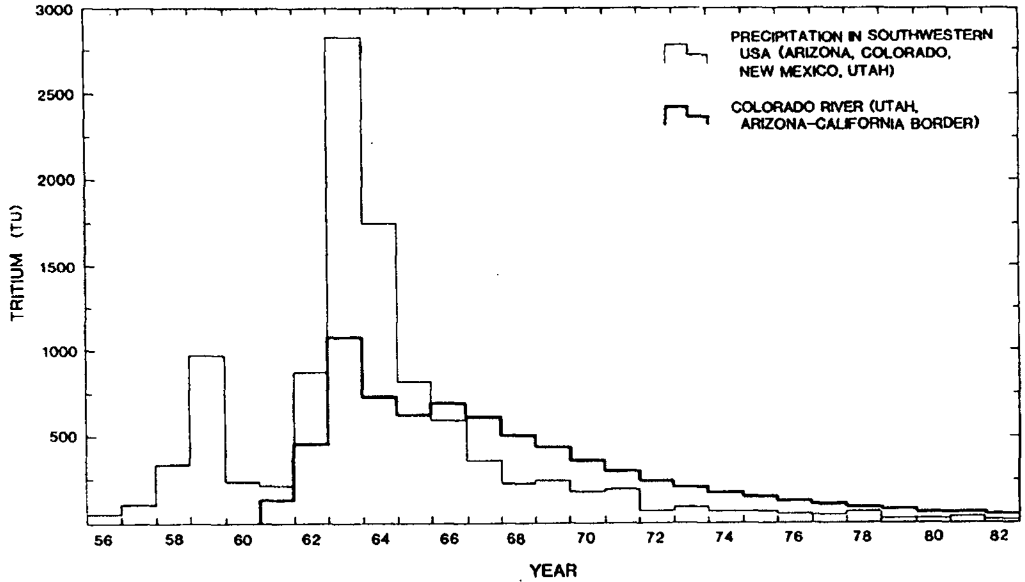

Fig. 12. Histograms showing tritium-time variations in meteoric waters of the southwestern United States (see Table 3 for data).

oxygen 18 enrichment increases, suggesting a complex low-temperature interaction between the waters and their sedimentary host rocks of the eastern San Juan Basin. Evaporite minerals such as gypsum are very common in the stratified Paleozoic-Mesozoic rocks, particularly the Todilito Formation from which the springs issue.

TRITIUM-BASED AGES OF THERMAL WATERS

Before the first atmospheric nuclear tests produced artificial tritium (1953), the natural concentration of tritium in surface water reached 5–20 TU [*Fontes*, 1980]. In 1963, artificial tritium concentrations peaked at about 1000 TU in surface waters and several thousand TU in precipitation. Since the moratorium on atmospheric testing, tritium levels in precipitation have been steadily decreasing. Tritium data from six stations collecting precipitation (Utah, Colorado, Arizona, and New Mexico) and from two stations monitoring the Colorado River (Utah and Arizona-California border) have been averaged (Table 3) and the results plotted on a histogram (Figure 12) showing the evolution of tritium content in the surface waters of the southwestern United States as a function of time. During the years 1978–1982, average tritium content in precipitation fell from 75 to 40 TU and in streams from 110 to 60 TU [*WATSTORE*, 1983].

Water collected by a stream having a large watershed, like the Colorado River, represents a complex combination of groundwaters from various origins plus precipitation. Tritium variations from streams of this size are buffered compared to the variations observed in direct precipitation. Systematic recharge and discharge of these groundwaters cause a phase lag in the tritium response of large rivers when compared to direct precipitation. This phase lag is observable during periods of both increasing and decreasing atmospheric tritium. This explains why the Colorado River is systematically lower in tritium than rain water before 1965 and higher after 1965 (Figure 12). Using the 12.3-year half-life of tritium and the records of tritium content in surface waters from previous years, it is possible to date or at least to give an apparent age to groundwaters younger than 100 years.

Tritium in Cold Surface Water and Groundwater

Five streams were sampled in the Jemez Mountains during our investigation: San Antonio Creek, Sulphur Creek, Redondo Creek, East Fork Jemez River at its mouth near Battleship Rock, and the Jemez River just above Soda Dam. The first four streams drain the inside of Valles caldera and unite

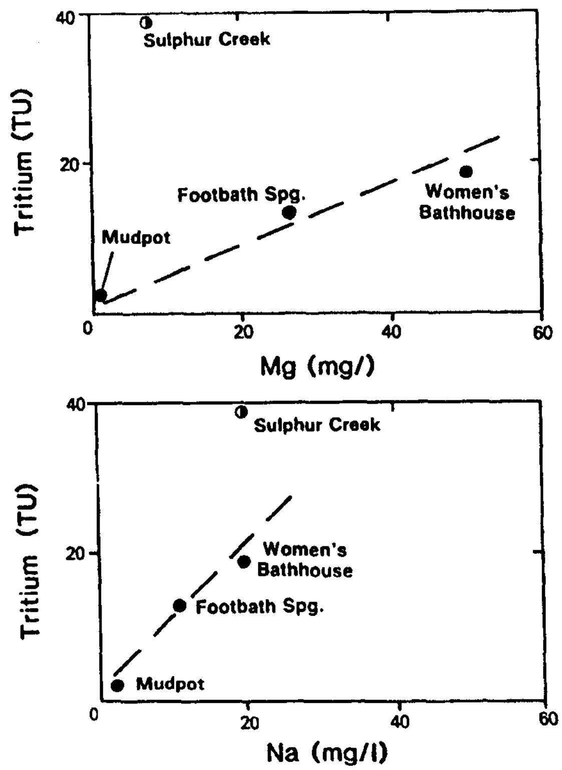

Fig. 13. Plots of tritium versus sodium and magnesium from acid-sulfate waters of the Sulphur Springs area and Sulphur Creek, Valles caldera, New Mexico (legend same as Figure 4).

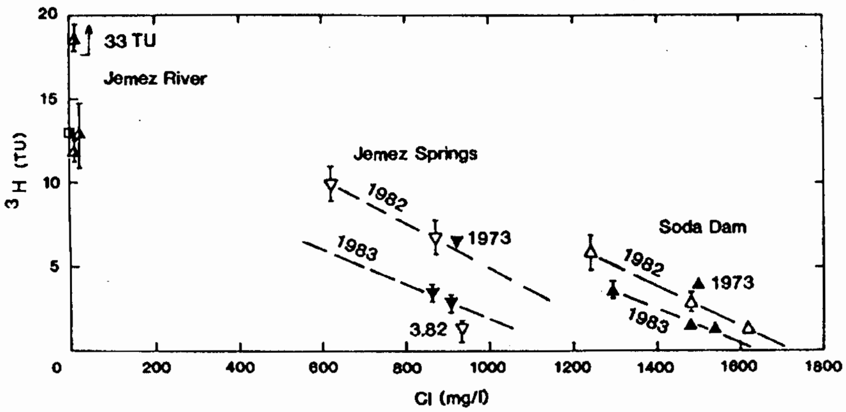

Fig. 14. Plot of tritium versus chloride for the thermal and nonthermal waters of Soda Dam and Jemez Springs, San Diego Canyon, New Mexico. Legend is same as Figure 4 except open triangles are 1982 samples, solid triangles are 1983 samples, and points labeled 1973 are from *Trainer* [1984].

just outside the southwestern topographic rim of the caldera at Battleship Rock to form the main Jemez River. These comparatively small streams do not display tritium values that are buffered like the Colorado River because they are highly sensitive to local and seasonal variations in the tritium content of precipitation. During 1982 and 1983, the five streams varied in their tritium content from 39 to 6 TU (Table 1), suggesting that the average age of their water varies from 5 to 0 years. San Antonio Creek and Jemez River at Soda Dam displayed nearly threefold variations in tritium content between seasons. These variations partially mimic the seasonal variations in tritium observed in precipitation at Albuquerque, New Mexico (about 100 km south of Valles caldera), where tritium content reaches a maximum by the end of spring and a minimum at the beginning of winter [*WATSTORE*, 1983]. Unfortunately, we do not have seasonal samples of each stream during both years to accurately document seasonal tritium variations.

Nine cold springs collected from the Valles caldera region display very different tritium contents, ranging from 0.3 to 44 TU. Other parameters are also highly variable, like discharge rate (<2 to >2000 L/min) and total mineralization (90 to 3700 mg/L TDS). No general correlation exists among the parameters of these springs, but the two springs having the highest tritium contents (Seven Springs, 20.6 TU, and Las Conchas Spring, 44.3 TU) have low mineralization and high discharge from volcanic rocks. These data indicate that the two latter springs contain water having rapid, shallow underground circulation representative of nearly contemporaneous precipitation. In contrast, Sino Spring, which discharges dilute water from the base of the Tertiary volcanic section, has only 0.29 TU. Thus an apparent age of mean underground transit time of 50–75 years is estimated for the water of Sino Spring assuming piston flow.

Other cold springs like Panorama, Turkey, Horseshoe, and Valle Grande Springs contained just 1–2 TU during 1982–1983 and have an estimated age of their waters at 30–40 years, assuming piston flow. The latter two springs discharge from the moat zone of Valles caldera. Two cold mineral springs (Battleship Seep and Ponderosa Spring) contained 3.4 and 2.1 TU, respectively, suggesting that they may contain a mixture of young and old waters. Clearly, any groundwater presently having 5 TU or more tritium must contain a young (post-1953) component, but without repeat samples during complete hydrologic cycles it may be difficult to accurately assess the flow and mixing history of the individual springs.

Tritium in Thermal Water

Thermal meteoric waters. As mentioned above, this group of dilute thermal waters is represented by four main springs that issue from rhyolitic rocks in the western ring fracture zone of Valles caldera [*Goff and Grigsby*, 1982]. Bathhouse, San Antonio, Spence, and McCauley springs all consist of $NaCO_3$ waters with low Cl and very little, if any, enrichment in trace elements such as B (Table 1). Only SiO_2 concentrations are comparatively high reflecting the elevated temperatures of the springs and, probably, the high solubility of volcanic glass.

During the winter of 1983, the tritium content of these four springs was below 1 TU, yielding apparent ages for their water of 50 ± 10 years. In March 1982, however, the tritium levels of the same springs ranged from 2 to 8 TU, while their mineral load was slightly decreased. In a parallel study, *Phillips et al.* [1984] obtained a $^{36}Cl/Cl$ ratio of 879×10^{-15} from San Antonio Hot Spring or about 25% higher than the background ratio for New Mexico (the half-life of ^{36}Cl is 3.01×10^5 years). Both the tritium and ^{36}Cl data show that some postbomb water is contributing to the spring flow. Possibly snowmelt was more advanced during March 1982 than in winter 1983, and the 1982 samples reflect some subsurface mixing of snowmelt and normal hot spring water. In any case, the spring waters appear to be composed primarily of groundwaters that circulate only in the upper 500 m of the moat zone of the caldera and are heated by the high heat flux of the volcanic heat source [*Goff and Grigsby*, 1982]. The springs do not appear to be composed of deeply circulating hydrothermal fluids or derivatives resulting from mixture with such fluids.

Acid-sulfate waters. Tritium content of acid-sulfate springs is highly variable because they are generally mixtures of condensed steam and near-surface groundwaters. This variability is well displayed by the acid-sulfate thermal waters of the Sulphur Springs area (Figure 13). For example, Men's Bathhouse Mudpot, which displays no apparent discharge but bubbles violently with CO_2 and H_2S, contains only 2.1 TU (Table 1). This mud pot is composed primarily of condensed steam with small additions of surface groundwater or precipitation. In contrast, the flowing acid-sulfate spring at Women's Bathhouse Spring contains a relatively high proportion of surface groundwater and 19 TU. Because the groundwaters flow through acid-altered rocks containing soluble sulfate salts, the total mineralization of the flowing springs is greater than in waters that are composed only of steam condensates.

Deep geothermal waters and derivatives. Isotopic dating of deep geothermal reservoirs has usually revealed relatively tritium-free fluids indicating that the waters are mostly more than 80–100 years old [*Truesdell and Hulston*, 1980]. The range of tritium contents from five Baca wells sampled twice during summer 1982 is 0.18–1.1 TU (Table 1; see also *White* [this issue]). It is surprising that some well fluids contain as much tritium as they do, but because the wells have often been used for injection, some young meteoric water has been introduced to the reservoir. Using the 12.3-year half-life of tritium and value of 5 to 20 TU as background, the age of the Valles geothermal reservoir is not less than 60–90 years, assuming that 0.18 TU represents the uncontaminated tritium value of the reservoir.

Another way of assessing the age of the fluids in the Baca reservoir is to assume it is a homogeneously mixed reservoir and not a reservoir responding only to piston flow. *Pearson and Truesdell* [1978] state that a homogeneously mixed reservoir model takes into account hydrodynamic-dispersive mixing and the likely possibility that there are many meteoric sources having different ages of water recharging the geothermal system. Using the homogeneously mixed reservoir model, the average residence time of water in a reservoir having the tritium concentrations observed in the Baca wells is roughly 10,000 years [*Pearson and Truesdell*, 1978, Figure 1]. Piston flow and homogeneously mixed reservoir models can be considered as end-member cases. Thus the mean age of water in the Baca reservoir based on tritium data is between 60 and 10,000 years.

Another notion of the relative age of the Baca reservoir fluids was obtained by *Phillips et al.* [1984], who report that the $^{36}Cl/Cl$ ratio of Baca 13 is 35.3×10^{-15}, a value consistent with secular equilibrium between hydrothermal fluid and the neutron flux expected in tuffaceous reservoir rocks. Several half-lives (3.01×10^5 years) are required to approach seqular equilibrium, thus, the reservoir Cl must have a mean residence time of more than 0.5 Ma. Although the background $^{36}Cl/Cl$ ratio of New Mexico shallow groundwaters is 700×10^{-15}, the amount of Cl in meteroric waters that would be expected to recharge the geothermal system (≤ 10 mg/l) is small compared with the amount of Cl in the reservoir (1900 to 2700 mg/L). The relation between chloride residence time and water residence time has not been determined, yet the low $^{36}Cl/Cl$ ratio of Baca 13 fluid indicates that the flux of meteoric ^{36}Cl through the geothermal system is low and that the mean residence time of water in the reservoir is probably several hundred years at least.

As mentioned above, the groups of springs at Soda Dam and Jemez Springs are apparently mixed waters issuing from a hydrothermal outflow plume originating inside Valles caldera. Based on single analyses in 1973 of samples from Soda Dam and Jemez Springs containing approximately 6 and 4 TU, respectively, *Trainer* [1984] concluded that the dilute groundwater mixing with geothermal water in the hydrothermal plume had an average underground residence time of 20 years. If Trainer's two-component model were correct, the two groups of thermal springs would have reached their highest tritium load in the early 1980's because the diluting component would have percolated into the ground in the early 1960's, during the peak of atmospheric nuclear testing. Our recent data, which consist of repeated analyses from several springs during 1982 and 1983, indicate a more complicated mixing relation between surface groundwaters and thermal fluids because tritium concentrations have not substantially changed between 1973 and 1983. A negative correlation between tritium and Cl contents of the thermal springs is shown in Figure 14. Because the number of data points is still comparatively small and differences exist in the absolute tritium values from 1982 to 1983, only general trends can be extrapolated to end-members. Nonetheless, extrapolation to the cold, dilute component yields a tritium value between 12 and 20 TU, the same value as the adjacent Jemez River during the same time period.

Extrapolation of the trends to the tritium-free end-member indicates substantially different chloride values for the thermal component: 1100–1300 mg/L for Jemez Springs and 1600–1700 mg/L for Soda Dam. These values are lower than the chloride content of about 1900–2700 mg/L (corrected for steam loss) in the deep hydrothermal fluid of Valles caldera (Table 1) and strongly suggest that a component of old, tritium-free groundwater is mixing with the thermal plume. This deduction supports the ^{36}Cl study of the Soda Dam fluid by *Phillips et al.* [1984], who found a $^{36}Cl/Cl$ ratio of only 13.3×10^{-15}, or less than half the value of Baca 13. *Phillips et al.* [1984] interpreted this low value to mean that the fluid of the hydrothermal plume is mixing with cold mineralized groundwater depleted in ^{36}Cl. Presumably, this "dead" Cl is leached from the thick Paleozoic sedimentary section surrounding the western side of Valles caldera. Although we have not found a tritium-free groundwater issuing from Paleozoic rocks in the immediate Valles area, tritium-free mineral waters emerge from similar rocks in the San Ysidro area 40 km away (see below).

Because the two sets of hot springs have a negative correlation between temperature and mineralization, previous investigators concluded that different flow paths, different amounts of conductive heat loss, slightly different parent fluid compositions, or different proportions of cold near-surface groundwater caused this correlation. Our tritium and stable isotope results add yet another variable to the picture because they show that the hydrothermal plume consists of old geothermal water from the caldera, young near-surface groundwater, and old mineralized groundwater from the Paleozoic section. Clearly, some mixing with shallow tritiated water occurs late in the mixing cycle as in the case of Buddhist Spring (10 TU in 1982), which emerges only 10 m from the edge of the Jemez River.

Mineral waters of San Ysidro. As mentioned previously, mineral waters from the San Ysidro area are geochemically distinct from the hydrothermal fluids in Valles caldera (Figure 4). Tritium concentration is relatively low in these waters (Table 1). In fact, the concentration of the Zia Hot Well (0.05 TU) is the lowest value we found in the Jemez Mountains region, suggesting an age of at least 90 years for that fluid assuming piston flow. Since 1926, about 10×10^6 m^3 of 54°C geothermal water has emerged from this well by artesian flow [*Summers*, 1976]. The tritium content of surface mineral springs is somewhat higher than Zia Hot Well, suggesting contamination with surface precipitation. Although the tritium data are not sufficient to discriminate between the systems at Valles caldera and San Ysidro, low tritium contents and high total dissolved solids indicate that the San Ysidro waters are relatively old and deeply circulating.

CONCLUSIONS

Stable isotope and tritium data indicate that four groundwater regimes occur in the Valles caldera-Jemez Mountains region. These are (1) a surficial regime of young water with rapid runoff (age ≤ 5 years), (2) a somewhat deeper regime of slightly older water circulating primarily in Tertiary volcanic rocks (age 5–50 years), (3) a deep regime of groundwaters in Tertiary basin fill sediments filling the western Rio Grande rift (age unknown), and (4) a deep regime of groundwaters circulating in Precambrian to Mesozoic rocks surrounding the western margins of Valles caldera, and in Tertiary rocks filling the caldera collapse structure (age > 50 years).

Valles caldera possesses a "type" example of a volcanically driven hydrothermal system (compare models of *Goff and Grigsby* [1982] and *Faust et al.* [1984] to idealized model of *Henley and Ellis* [1983]). The essential elements of this system are (1) meteoric recharge from local sources, (2) upward convection of relatively old geothermal fluid above a volcanic heat source, (3) subsurface boiling of hydrothermal fluid and production of an overlying vapor-rich zone with steam-heated groundwaters, and (4) lateral outflow of hydrothermal fluid. Stable and radiogenic isotopes in combination with geochemical data on conservative ions support every aspect of this hydrodynamic model. In addition to thermal waters comprising the essential features of the deep hydrothermal system, dilute thermal meteoric waters circulate in caldera moat deposits above the hydrothermal system, and pore fluid brines exist in hot Precambrian rocks beneath the hydrothermal system.

Our isotopic studies, together with the ^{36}Cl investigation of *Phillips et al.* [1984], have yielded some results that are specific to the Valles caldera hydrothermal system and not necessarily applicable to all volcanically driven systems. For example, although the Valles hydrothermal system is recharged by local meteoric water, the recharge rate must be slow because radiogenic isotopes indicate that the hydrothermal fluids are old, probably several hundred years. In addition, thermal springs derived from the lateral outflow plume of the hydrothermal system consist of at least three components: an old geothermal component, a young meteoric component, and an old mineralized component. This latter fluid can only originate in Paleozoic rocks surrounding the caldera.

Stable isotope and supporting geochemical data verify that the mineral fluids of the San Ysidro area are derived from a unique, low-temperature geothermal system in which waters circulate along the Nacimiento fault zone bordering the east side of the San Juan Basin [*Goff et al.*, 1981].

Finally, we note that most of the cold springs in the Jemez Mountains region, commonly issuing from the base of the Tertiary volcanic section, consist of surprisingly old water (5–50 years). Two cold springs inside the Valles caldera moat zone, Valle Grande and Horseshoe springs, also consist of older water. Streams that drain the moat zones of Valles caldera (San Antonio Creek and East Fork Jemez River) contain unusually low, but seasonally variable tritium concentrations, indicating that the streams are fed from both surficial waters and slightly older groundwaters. In contrast, streams draining the central resurgent dome (Redondo Creek and Sulphur Creek) contain water with present-day meteoric levels of tritium.

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REFERENCES

Aldrich, M. J., Jr., Tectonics of the Jemez lineament in the Jemez Mountains and Rio Grande rift, *J. Geophys. Res.*, this issue.

Bailey, R. A., R. L. Smith, and C. S. Ross, Stratigraphic nomenclature of volcanic rocks in the Jemez Mountains, New Mexico, *U.S. Geol. Surv. Bull.*, *1274-P*, 19 pp., 1969.

Bodvarsson, G. S., S. Vonder Haar, M. J. Wilt, and C. F. Tsang, Preliminary estimation of the reservoir capacity and longevity of the Baca geothermal field, New Mexico, *Rep. 9273*, 11 pp., Soc. of Pet. Eng., AIME, Richardson, Tex., 1980.

Charles, R. W., R. J. Vidale Bulen, and F. Goff, An interpretation of the alteration assemblages at Sulphur Springs, Valles caldera, New Mexico, *Jour. Geophys. Res.*, this issue.

Craig, H., Isotopic variations in meteoric waters, *Science*, *133*, 1702–1703, 1961.

Dash, Z. M., et al., Hot Dry Rock geothermal reservoir testing: 1978 to 1980, *J. Volcanol. Geotherm. Res.*, *15*, 59–99, 1983.

Doell, R. R., G. B. Dalrymple, R. L. Smith, and R. A. Bailey, Paleomagnetism, potassium-argon ages, and geology of rhyolites and associated rocks of the Valles caldera, *Mem. Geol. Soc. Am. 116*, 211–248, 1968.

Dondanville, R. F., The hydrothermal geology of the Valles caldera, Jemez Mountains, New Mexico, open file consultant report, 36 pp., Union Oil Co., Santa Rosa, Calif., 1971.

Dondanville, R. F., Geologic characteristics of the Valles caldera geothermal system, New Mexico, *Trans. Geotherm. Res. Counc.*, *2*, 157–160, 1978.

Ellis, A. J., and W. A. J. Mahon, *Chemistry and Geothermal Systems*, 397 pp., Academic, Orlando, Fla., 1977.

Faust, C. R., J. W. Mercer, S. D. Thomas, and W. P. Ballean, Quantitative analysis of existing conditions and production strategies for the Baca geothermal system, New Mexico, *Water Resour. Res.*, *20*, 601–618, 1984.

Fontes, J. Ch., Environmental isotopes in groundwater hydrology, in *Handbook of Environmental Isotope Geochemistry*, vol. 1A, *The Terrestrial Environment*, edited by P. Fritz and J. Ch. Fontes, pp. 75–140, Elsevier, Amsterdam, 1980.

Friedman, I., and J. R. O'Neil, Compilation of stable isotope fractionation factors of geochemical interest, Data of Geochemistry, 6th ed., edited by M. Fleischer, Chapter KK, *U.S. Geol. Surv. Prof. Pap.*, *440-KK*, 10 pp., 1977.

Fritz, P., and S. K. Frape, Comments on the ^{18}O, ^{2}H, and chemical composition of saline groundwaters on the Canadian Shield, in *Isotope Studies of Hydrologic Processes*, edited by E. C. Perry, pp. 57–63, Northern Illinois University Press, De Kalb, 1982.

Gardner, J. N., and F. Goff, Potassium-argon dates from the Jemez volcanic field: Implications for tectonic activity in the north-central Rio Grande rift, *Field Conf. Guideb. N.M. Geol. Soc.*, *35*, 75–81, 1984.

Gardner, J. N., F. Goff, S. Garcia, and R. C. Hagan, Stratigraphic relations and lithologic variations in the Jemez volcanic field, New Mexico, *J. Geophys. Res.*, this issue.

Gat, J., The isotopes of hydrogen and oxygen in precipitation, in *Handbook of Environmental Isotope Geochemistry*, vol. 1A: *The Terrestrial Environment*, edited by P. Fritz and J. Ch. Fontes, pp. 21–47, Elsevier, Amsterdam, 1980.

Goff, F., Subsurface structure of Valles caldera: A resurgent cauldron in northern New Mexico, *Geol. Soc. Am. Abstr. Programs*, *15*, 381, 1983.

Goff, F. E., and J. N. Gardner, Geologic map of the Sulphur Springs area, Valles caldera geothermal system, New Mexico, *Los Alamos Natl. Lab. Rep. LA*, *8634-MAP*, 1980.

Goff, F., and C. O. Grigsby, Valles caldera geothermal systems, New Mexico, USA, *J. Hydrol.*, *56*, 119–136, 1982.

Goff, F., and S. Sayer, A geothermal investigation of spring and well waters of the Los Alamos region, New Mexico, *Los Alamos Sci. Lab. Rep. La*, *8326-MS*, 21 pp., 1980.

Goff, F. E., C. O. Grigsby, P. E. Trujillo, D. Counce, and A. Kron, Geology, water geochemistry, and geothermal potential of the

Jemez Springs area, Cañon de San Diego, New Mexico, *J. Volcanol. Geotherm. Res.*, *10*, 227–244, 1981.

Goff, F., T. McCormick, P. Trujillo, D. Counce, and C. Grigsby, Geochemical data for 95 thermal and nonthermal waters of the Valles caldera–southern Jemez Mountains region, New Mexico, *Los Alamos Natl. Lab. Rep. LA, 9367-OBES*, 51 pp., 1982.

Goff, F., J. Gardner, R. Vidale, and R. Charles, Geochemistry and isotopes of fluids from Sulphur Springs, Valles caldera, New Mexico, *J. Volcanol. Geotherm. Res.*, *23*, 273–297, 1985.

Grigsby, C. O., F. Goff, P. E. Trujillo, and D. A. Counce, Geochemical behavior of a hot dry rock geothermal reservoir, *Field Conf. Guideb. N.M. Geol. Soc.*, *35*, 265–270, 1984.

Heiken, G., F. Goff, J. Styx, S. Tamanyu, M. Shafiqullah, S. Garcia, and R. Hagan, Intracaldera volcanic activity, Toledo caldera and embayment, Jemez Mountains, New Mexico, *J. Geophys. Res.*, this issue.

Henley, R. W., and A. J. Ellis, Geothermal systems ancient and modern, *Earth Sci. Rev.*, *19*, 1–50, 1983.

Hulen, J. B., and D. L. Nielson, Stratigraphic permeability in the Baca geothermal system, Redondo Creek area, Valles caldera, New Mexico, *Trans. Geotherm. Resour. Counc.*, *6*, 27–31, 1982.

Kerr, R. A., Extracting geothermal energy can be hard, *Science*, *218*, 668–669, 1982.

Laughlin, A. W., The geothermal system of the Jemez Mountains, New Mexico, and its exploration, in *Geothermal Systems: Principles and Case Histories*, edited by L. Rybach and L. J. P. Muffler, pp. 295–320, John Wiley, New York, 1981.

Mariner, R. H., T. S. Presser, and W. C. Evans, Chemical, isotopic, and gas compositions of selected thermal springs in Arizona, New Mexico, and Utah, *U.S. Geol. Surv. Open File Rep.*, *77-654*, 42 pp., 1977.

Moser, H., and W. Stichler, Deuterium measurements on snow samples from the Alps, in *Isotope Hydrology*, pp. 43–57, International Atomic Energy Agency, Vienna, 1970.

Payne, B. R., and Y. Yurtsever, Environmental isotopes as a hydrogeological tool in Nicaragua, in *Isotope Techniques in Groundwater Hydrology*, pp. 193–202, International Atomic Energy Agency, Vienna, 1974.

Pearson, C., and F. Goff, A Schlumberger resistivity study of the Jemez Springs region of northwestern New Mexico, *Trans. Geotherm. Resour. Counc.*, *5*, 119–122, 1981.

Pearson, F. J., Jr., and A. H. Truesdell, Tritium in the waters of Yellowstone National Park, *U.S. Geol. Surv. Open File Rep.*, *78-701*, 3 pp., 1978.

Phillips, F. M., F. Goff, F. Vuataz, H. W. Bentley, D. Elmore, and H. E. Gove, ^{36}Cl as a tracer in geothermal systems: Example from Valles caldera, New Mexico, *Geophys. Res. Lett.*, *11*, 1227–1230, 1984.

Purtymun, W. D., and S. Johansen, General geohydrology of the Pajarito Plateau, *Field Conf. Guideb. N.M. Geol. Soc.*, *25*, 347–349, 1974.

Riecker, R. (Ed.), *Rio Grande Rift: Tectonics and Magmatism*, 438 pp., AGU, Washington, D. C., 1979.

Segar, R. L., Quantitative gravity interpretation, Valles caldera area, Sandoval and Rio Arriba counties, New Mexico, open file report, 12 pp., Union Geotherm. Co., Santa Rosa, Calif., 1974.

Smith, R. L., and R. A. Bailey, Resurgent cauldrons, *Mem. Geol. Soc. Am.*, *116*, 613–662, 1968.

Smith, S., and B. Kennedy, Nobel gas evidence for two fluids in the Baca (Valles Caldera) geothermal reservoir, *Geochim. Cosmochim. Acta*, *49*, 893–902, 1985.

Smith, R. L., R. A. Bailey, and C. S. Ross, Geologic map of the Jemez Mountains, New Mexico, scale 1:125,000, *U.S. Geol. Surv. Misc. Invest. Map*, *I-571*, 1970.

Summers, W. K., Catalog of thermal waters in New Mexico, *N. M. Bur. Mines Miner. Resour. Hydrol. Rep.*, *4*, 80 pp., 1976.

Swanberg, C. A., Geothermal resources of rifts: A comparison of the Rio Grande rift and the Salton Trough, *Tectonophysics*, *94*, 659–678, 1983.

Trainer, F. W., Groundwater in the southwestern part of the Jemez Mountains volcanic region, New Mexico, *Field Conf. Guideb. N.M. Geol. Soc.*, *25*, 337–345, 1974.

Trainer, F. W., Mixing of thermal and nonthermal waters in the margin of the Rio Grande rift, Jemez Mountains, New Mexico, *Field Conf. Guideb. N.M. Geol. Soc.*, *26*, 213–218, 1975.

Trainer, F. W., Geohydrologic data from the Jemez Mountains and vicinity, north-central New Mexico, *U.S. Geol. Surv. Water Resour. Invest.*, *77-131*, 146 pp., 1978.

Trainer, F. W., Thermal springs in Cañon de San Diego as a window into Valles caldera, New Mexico, *Field Conf. Guideb. N.M. Geol. Soc.*, *35*, 249–255, 1984.

Trainer, F. W., and F. P. Lyford, Geothermal hydrology in the Rio Grande rift, north-central New Mexico, *Field Conf. Guideb. N. M. Geol. Soc.*, *30*, 299–306, 1979.

Truesdell, A. H., and J. R. Hulston, Isotopic evidence on environments of geothermal systems, in *Handbook of Environmental Isotope Geochemistry*, vol. 1A, *The Terrestrial Environment*, edited by P. Fritz and J. Ch. Fontes, pp. 179–226, Elsevier, Amsterdam, 1980.

Truesdell, A. H., and C. J. Janik, Reservoir processes and fluid origins in the Baca geothermal system, Valles caldera, New Mexico, *J. Geophys. Res.*, this issue.

Truesdell, A. H., M. Nathenson, and R. Rye, The effects of subsurface boiling and dilution on the isotopic compositions of Yellowstone thermal waters, *J. Geophys. Res.*, *82*, 3694–3704, 1977.

Vuataz, F. D., Hydrogeologie, geochimie et geothermie des eaux thermales de Suisse et des regions alpines limitrophes, Mater. Geol. Suisse, *Ser. Hydrol.*, vol. 20, 174 pp., Kümmerly and Frey, Berne, 1982.

Vuataz, F. D., J. Stix, F. Goff, and C. Pearson, Low temperature geothermal potential of the Ojo Caliente warm springs area, northern New Mexico, *Los Alamos Natl. Lab. Rep. LA, 10105-OBES*, 56 pp., 1984.

WATSTORE, Water data storage and retrieval system, U.S. Geol. Surv., Albuquerque, NM, 1983.

White, A. F., J. A. Delaney, A. F. Truesdell, C. Janik, F. Goff, and H. Crecraft, Fluid chemistry of the Baca geothermal field, Valles caldera, New Mexico, *Field Conf. Guideb. N.M. Geol. Soc.*, *35*, 257–263, 1984.

White, D. E., Environments of generation of base-metal ore deposits, *Econ. Geol.*, *63*, 301–335, 1968.

White, A. F., Chemical and isotopic characteristics of fluids within the Baca geothermal reservoir, Valles caldera, New Mexico, *J. Geophys. Res.*, this issue.

Zuppi, G. M., J. Ch. Fontes, and R. Letolle, Isotopes du milieu et circulations d'eaux sulfurees dand le Latium, in *Isotope Techniques in Groundwater Hydrology*, pp. 341–361, International Atomic Energy Agency, Vienna, 1974.

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