GEOLOGIC MAP OF LONG VALLEY CALDERA, MONO-INYO CRATERS VOLCANIC CHAIN, AND VICINITY, EASTERN CALIFORNIA

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GEOLOGIC SETTING

Long Valley caldera and the Mono-Inyo Craters volcanic chain compose a late Tertiary to Quaternary volcanic complex on the west edge of the Basin and Range Province at the base of the Sierra Nevada frontal fault escarpment. The caldera, an east-west-elongate, oval depression 17 by 32 km, is located just northwest of the northern end of the Owens Valley rift and forms a reentrant or offset in the Sierran escarpment, commonly referred to as the "Mammoth embayment." The Mono-Inyo Craters volcanic chain forms a north-trending zone of volcanic vents extending 45 km from the west moat of the caldera to Mono Lake. The prevolcanic basement in the area is mainly Mesozoic granitic rock of the Sierra Nevada batholith and Paleozoic metasedimentary and Mesozoic metavolcanic rocks of the Mount Morrison, Gull Lake, and Ritter Range roof pendants (map A).

PHYSIOGRAPHY

The oval floor of Long Valley caldera ranges in elevation from about 2,000 m in its eastern half, which is dominated by Lake Crowley and the surrounding grass- and sage-covered surface of Long Valley proper, to about 2,600 m in its western half, which is hillier and mostly forested. The topographic walls of the caldera rise steeply to elevations of 3,000–3,600 m, except on the east and southeast, where the rim is a 100-200-m subdued escarpment that forms the northwest edge of the Volcanic Tableland. The highest hills within the central and western parts of the caldera rise to a maximum elevation of 2,800 m. Between the central hills and the caldera walls is an annular depression, referred to as the caldera moat, which is drained on the north and east by Deadman Creek and the Owens River and on the south by Mammoth Creek and its lower reach known as Hot Creek.

Only the northern part of the Mono-Inyo Craters volcanic chain, the Mono Craters proper, forms a prominent physiographic feature—a 17-km-long arcuate ridge, extending southward from Mono Lake to near June Lake. Composed of 30 or more overlapping volcanic domes and craters, the ridge rises about 610 m above the surrounding terrain to a maximum elevation of 2,800 m. Viewed from the west this ridge forms a striking skyline of barren crags mantled with a pale-gray blanket of tephra. The Inyo Craters chain forms a less prominent, discontinuous line of volcanic domes and craters extending 10 km south of the Mono Craters into the west moat of Long Valley caldera.

VOLCANISM

Long Valley caldera

Volcanism in the Long Valley area (Bailey and others, 1976; Bailey, 1982b) began about 3.6 Ma with widespread eruption of trachybasaltic-trachyandesitic lavas on a moderately well dissected upland surface (Huber, 1981). Erosional remnants of these mafic lavas are scattered over a 4,000-km^2 area extending from the Adobe Hills (5–10 km notheast of the map area), around the periphery of Long Valley caldera, and southwestward into the High Sierra. Although these lavas never formed a continuous cover over this region, their wide distribution suggests an extensive mantle source for these initial mafic eruptions. Between 3.0 and 2.5 Ma quartz-latite domes and flows erupted near the north and northwest rims of the present caldera, at and near Bald Mountain and on San Joaquin Ridge (Chaudet, 1986). These intermediate-composition lavas probably represent the onset of deep-crustal magmatic accumulation and differentiation that eventually culminated in formation of the large shallow Long Valley magma chamber, from which subsequent more silicic eruptions originated. The first of these more silicic eruptions occurred in the vicinity of Glass Mountain, on the northeast rim of the present caldera, where a 1,000-m-thick complex of high-silica rhyolite domes, flows, and tuffs accumulated between 2.1 and 0.8 Ma (Metz and Mahood, 1985). The high, central part of this complex, consisting of overlapping and juxtaposed domes, flows, and intrusions, was surrounded by an extensive apron of tuffs, consisting of tephra, small-volume ash-flows, and block-and-ash flows, now preserved only on the north and southeast flanks of the complex. The older of these Glass Mountain rhyolites are somewhat heterogeneous chemically, but those 1.1 Ma and younger in age are quite homogeneous and more voluminous, suggesting that early isolated source bodies did not coalesce to form a single large chamber until the latter part of the Glass Mountain episode (Metz and Mahood, 1985).

At 0.73 Ma, catastrophic rupturing of the roof of the magma chamber resulted in the expulsion of 600 km^3 of rhyolite magma, mainly as hot incandescent ash flows. This partial emptying of the chamber caused collapse of its roof, forming the 17-by-32-km oval depression of Long Valley caldera. The resulting ash-flow deposits, the Bishop Tuff, inundated 1,500 km^2 of the surrounding countyside and accumulated locally to thicknesses approaching 200 m on the Volcanic Tableland in upper Owens Valley, in Adobe Valley, and in the southwestern part of Mono Basin. Some ash flows spilled westward over the crest of the Sierra

Nevada, filling the canyons of the Middle Fork of the San Joaquin River, although subsequent erosion has removed most of this intercanyon tuff. A large volume of the Bishop Tuff also ponded within the caldera, where drill holes (Smith and Rex, 1977; Gambill, 1981) have penetrated as much as 1,500 m of the Bishop Tuff buried beneath younger caldera fill. Plinian ash clouds associated with this climactic eruption drifted thousands of kilometers downwind, depositing an ash layer (informally designated the Bishop ash bed) recognized as far east as Kansas and Nebraska (Izett and others, 1970; Izett, 1982). Thinner Bishop ash deposits also are recognized in southwest California (Merriam and Bischoff, 1975) and in Pacific Ocean cores (Sarna-Wojcicki and others, 1987). Gilbert (1938) first mapped and described the Bishop Tuff and recognized Long Valley as its probable source. Hildreth and Mahood (1986) have shown, on the basis of the distribution of accidental inclusions in the deposits, that the initial Plinian eruptions vented in the south-central part of the caldera and that, with the transition to ash-flow eruptions, venting migrated around the ring-fracture zone, the final eruptions occurring along the northern caldera margin.

The Bishop Tuff is composed of porphyritic rhyolite ash and pumice lapilli and blocks containing 5 to 25 percent phenocryts of sanidine, quartz, plagioclase, biotite, augite, hypersthene, Fe-Ti oxides, and accessories, and it shows considerable variation in degree of welding and crystallization (Sheridan, 1965, 1970). Exposures in the Owens River Gorge show evidence of multiple cooling units (Sheridan, 1967), suggesting that the tuff was emplaced during at least two major eruptive pulses. Hildreth (1979), on the basis of detailed mineralogical and chemical studies, has shown that the tuff represents the stratigraphically inverted contents of the upper part of the Long Valley magma chamber and that it preserves chemical gradients developed in the chamber prior to eruption.

After collapse of the roof of the magma chamber, eruptions continued within the 2– to 3–km–deep caldera. Early postcaldera pyroclastic eruptions formed a thick sequence of intracaldera bedded tuffs that were followed by extrusion of thin, hot, fluid obsidian flows. Simultaneously, renewal of magma presssure at depth uplifted, arched, and faulted the central part of the caldera floor, forming a resurgent dome (Smith and Bailey, 1968) about 10 km in diameter, 500 m high, and having a northwest-trending medial graben 4 km wide. This uplift occurred within 100,000 years or less after caldera collapse, between about 0.7 and 0.6 Ma, as indicated by K–Ar ages of the contemporaneous obsidian flows (Bailey and others, 1976). The rhyolite tephra and obsidian flows erupted during this episode are informally referred to as *early rhyolite* and are typically aphyric to sparsely porphyritic, containing less than 5 percent crystals of plagioclase, hypersthene, biotite, and Fe–Ti oxides. They contrast strikingly with the preceding crystal-rich Bishop Tuff and appatently erupted at higher temperatures and probably from greater depth.

During rise of the resurgent dome and eruption of the early rhyolite, the caldera was filled by a large lake, Pleistocene Long Valley Lake (Mayo, 1934), terraces and strandlines of which are well preserved along the east wall of the caldera and locally on the flanks of the resurgent dome. Continued rise of the resurgent dome eventually raised the lake surface to the level of the low southeast caldera rim, where overflow subsequently cut the Owens River Gorge. Gradual downcutting of the gorge, accompanied by intermittent tectonic tilting of the southeast rim near the Sierran front, ultimately drained the lake between 100,000 and 50,000 years ago. (Lake Crowley is not a remnant of the Pleistocene lake but a modern reservoir impounded behind a dam contructed in 1940.) During the early life of the Pleistocene lake, glaciers flowing into it from the High Sierra generated debris-laden icebergs that drifted across the lake, depositing large erratics of Sierran granite on the flanks of the resurgent dome, which stood as an island, and on the lake terraces on the east caldera wall.

After an interlude of about 100,000 years of volcanic quiescence, crystal-rich rhyolite began erupting in the caldera moat peripheral to the resurgent dome. This porphyritic rhyolite, informally referred to as *moat rhyolite*, typically contains 15 to 20 percent phenocrysts of plagioclase, quartz, sanidine, biotite, hornblende, and Fe-Ti oxides. It formed thick, steep-sided domes and flows, indicating higher viscosity and lower temperature than the early rhyolite, and probably signaled the onset of cooling and crystallization of the magma chamber. Moat rhyolite erupted at about 200,000-yr intervals—at 0.5, 0.3, and 0.1 Ma in clockwise succession around the resurgent dome, in the north, southeast, and west sectors of the moat, respectively—with four to five domes or flows erupting during each episode. The moat rhyolite consists of two distinct chemical types, low- and high-silica types, which erupted alternately in postcaldera time. The low-silica type (hornblende-biotite rhyolite, units Qmr1, 2, 3) is the most voluminous and most frequently erupted and probably represents the dominant volume (Hildreth, 1981) of the magma chamber. In contrast, the high-silica type (units Qmrh and Qmrm) has chemical characteristics similar to those of the Bishop Tuff, which suggests late development of Bishop-type magma in the upper part of the postcaldera magma chamber (Bailey, 1984). The 0.1-Ma western moat rhyolites appear to be the youngest lavas so far derived from the Long Valley chamber. However, seismological studies (Hill, 1976; Steeples and Iyer, 1976; Ryall and Ryall, 1981; Sanders, 1984; Hill and others, 1985b; Leutgert and Mooney, 1985) suggest that a residual body of magma still underlies the resurgent dome at depths of 5 to 15 km or more, providing a potential source for future eruptions. For such a body of silicic magma to have remained partially liquid and been a source for continuing eruptions throughout the past 700,000 years, additional heat must have been repeatedly added, most likely by periodic injection of basalt from the mantle into the lower part of the silicic chamber (Lachenbruch and others, 1976). Recent tectonic and apparent magmatic unrest within Long Valley caldera (Hill and others, 1985a) suggests that a brief episode of such magmatic injection may have occurred between 1980 and 1986 (Bailey, 1984).

Mono–Inyo Craters volcanic chain

The volcanism associated with Long Valley caldera, beginning about 3.6 Ma with precaldera trachybasalt-trachyandesite and quartz latite, continuing with the rhyolite of Glass Mountain and climactic eruptions of the Bishop Tuff, and concluding with the early- and moat-rhyolite extrusions as young as 0.1 Ma, constitutes a generally mafic-to-silicic sequence. The Mono-Inyo Craters volcanic chain, a 45-km north-trending zone of vents extending from Mammoth Mountain through the west moat of the caldera to Mono Lake, overlaps this Long Valley sequence both spatially and temporally. It also may be interpreted as a mafic-to-silicic sequence derived from younger chambers probably not directly related to Long Valley chamber, although in the area of overlap the interrelation is ambiguous. As here defined, the Mono–Inyo Craters volcanic chain includes postcaldera basaltic and quartz-latitic lavas (including the dacites and rhyodacites in and near Mono Lake) whose vents are localized along the same general north-south trend as the Mono-Inyo Craters rhyolites. Although the actual fissures and faults along which these vents are aligned are evident only sporadically along the chain, they clearly parallel a pervasive north- to northnortheast-trending fracture system in the pre-Tertiary Sierran rocks that is evident in aerial photographs, satellite imagery, and digital topographic maps of the Sierra Nevada.

The Mono-Inyo sequence began between 0.3 and 0.2 Ma with the eruption of postcaldera trachbasalt-trachyandesite from numerous vents in and near the west moat of the caldera, including those at Devils Postpile National Monument, southwest of the caldera. These mafic flows filled the west moat locally to depths of as much as 250 m and poured around the resurgent dome sending long tongues of lava into the northern and southern parts of the caldera moat. Younger mafic vents erupted successively farther north near June Lake (40-20 ka?) and at Black Point on Mono Lake (13.3 ka)—suggesting that the fissure system propagated northward. Intermittently during intervals in this mafic activity, quartz-latite domes and flows vented locally in the western part of the caldera, as well as farther north in and near Mono Lake. The greatest volumes, however, accumulated near the northwest and southwest caldera margins, where the north-south fissure system apparently intersected caldera ring fractures. On the southwest edge of the caldera, between about 200 and 50 ka, extrusion of at least 12 quartz-latite and associated low-silica sodic rhyolite domes and flows built Mammoth Mountain, an imposing cumulovolcano straddling the caldera rim.

Rhyolites began erupting to form the Mono-Inyo Craters about 40–38 ka—first at the Mono Craters and more recently at the Inyo Craters. The arcuate chain of 30 or more overlapping rhyolite domes, flows (coulees), and craters that forms the Mono Craters apparently erupted along the mylonitized border of a subcircular Cretaceous pluton centered on Aeolian Buttes (Kistler, 1966a; see maps A and D). Seismological studies (Achauer and others, 1986) suggest that their probable magmatic source is a subjacent north-south elongate chamber at 10–20 km depth; there appears to be no evidence for a larger deep-crustal chamber. The older Mono domes (40,000–3,000 yr B.P.) are moderately to sparsely porhyritic, whereas the younger ones (3,000–550 yr B.P.) are aphyric, suggesting a recent increase in eruptive temperature or decrease in depth of eruptive source, or both. Radiocarbon dating and dendrochronological studies indicate that the youngest eruptions in the chain occurred at the northern end, most likely between A.D. 1325 and 1365 but no later than A.D. 1368 (Sieh and Bursik, 1986).

The Inyo Craters chain forms a discontinuous 10-km-long line of rhyolite dome-flows and craters that include Wilson Butte, Obsidian Dome, Glass Creek dome, north and south Deadman Creek domes, as well as several other smaller unnamed domes and craters. Wilson Butte and north Deadman Creek dome are chemically and petrographically similar to the rhyolites of the Mono Craters and also are significantly older than the others, about 1,350 and 6,000 yr B.P., respectively (Miller, 1985). Although within the Inyo Craters chain, they are considered southern outliers of the rhyolite of Mono Craters. The younger Inyo domes—Obsidian Dome, Glass Creek dome, south Deadman Creek dome, two smaller unnamed domes, and their associated tephras—all erupted within a span of probably no more than a few years, or possibly months. Radiocarbon dating of the tephra deposits indicates an age of 650–550 yr B.P. (Miller, 1985), but stratigraphic and dendrochronological considerations indicate that they commenced erupting after cessation of the northern Mono Craters eruptions, possibly A.D. 1368 (Sieh and Bursik, 1986), and could be as young as A.D. 1472 (Miller, 1985). The initial pyroclastic eruptions from the three main younger vents blanketed the sourrounding terrain with thick deposits of pumice lapilli and ash; this explosive activity ceased from all three vents before extrusion of any of the lava domes (Miller, 1985). Drilling (Eichelberger and others, 1985) has confirmed that this activity was fed by a shallow rhyolite dike, as suggested by Miller (1985).

The younger Inyo domes consist of two contrasting, fluidly intermixed rhyolites: one light-colored, pumiceous, and very coarsely porphyritic, and the other more silicic, sparsely porphyritic obsidian. The two types superficially resemble the porphyritic moat rhyolite of Long Valley caldera and the aphyric to sparsely porphyritic obsidian of the Mono Craters respectively and one possible explanation for their origin is that they represent the mixing of magmas from the Long Valley and Mono Craters chambers along a connecting fissure (Bailey and others, 1976; Sampson, 1987; Sampson and Cameron, 1987).

During or shortly after the pyroclastic phase of the Inyo vents and probably before extrusion of the domes, phreatic explosions blasted three craters on the summit and south flank of Deer Mountain, a 0.1-Ma moat-rhyolite dome south of Deadman Creek dome (Rinehart and Huber, 1965). Two of these craters now contain small lakes (the Inyo Crater Lakes). Radiocarbon dating and dendrochronological considerations limit the eruptions between A.D. 1340 and 1460 (Wood, 1977). Several additional phreatic explosion craters of similar age erupted about 1 km to the northwest along a northwest-trending fault zone. The three aligned craters on Deer Mountain probably formed by the interaction of ground water with the same rhyolite dike that fed the Inyo domes, but no juvenile rhyolite magma reached the surface during the explosions; only hot mud and blocks of older trachyandesite, rhyolite, glacial debris, and basement rock were ejected. Other phreatic explosion craters and deposits occur 5 km to the south on the north face of Mammoth Mountain, where they are aligned subparallel to the caldera wall, apparently on a subsidiary caldera ring fracture. The associated phreatic debris is overlain by Inyo Craters pumice-fall deposits, but a radiocarbon age of 500 ± 200 yr B.P. obtained from a stump buried in the debris suggests that the eruptions were approximately contemporaneous with those from the Inyo Craters.

Still younger eruptions occurred at the northern end of the Mono-Inyo chain. In Mono Lake, several small dacite cinder cones and flows form Negit Island and the northeast corner of Paoha Island; a partly submerged rhyolite flow forms a small group of islets northeast of Negit; on the east side of Paoha is a small hydrothermally altered rhyodacite dome and a cluster of partly submerged phreatic explosion craters; and on the northern third of Paoha is a craggy rhyolite dome or invasive flow protruding through disrupted lake sediments (Lajoie, 1968). Most of Paoha Island is composed of lake sediments that were uplifted from the lake bottom, probably by intrusion of this latter rhyolite dome. Uplift caused slumping and sliding of these water-saturated sediments away from the center of the island, producing a remarkable landside jumble, particularly west of the island where the lake bathymetry suggests that slides extend under water at least 2 km laterally. Based on detailed study of radiocarbon-dated shorelines around Mono Lake, Stine (1984) concludes that the oldest of the dacites is slightly younger than 2,000 yr B.P. and that the youngest postdates 220 yr B.P.; uplift of Paoha Island and extrusion of the associated rhyolite is younger still and may have occurred between A.D. 1720 and 1850, according to Stine.

The bathymetry of Mono Lake (Scholl and others, 1967; Los Angels Department of Water and Power, 1987) shows that the lake is deepest immediately east and south of Paoha Island, where it attains depths of 40–50 m. Four or five small subcircular depressions on the lake bottom adjacent to Paoha Island suggest a peripheral arc of subaqueous craters. In all, there are at least ten subaerial vents and possibly five subaqueous vents within Mono Lake.

The recency and greater frequency of eruptions along the Mono-Inyo Craters chain during the past 3,000 years, together with evidence for increased magma temperature and (or) shallower eruption depth (Bailey, 1982b) suggests that in spite of the recent (1980–1986) episode of unrest at Long Valley, future eruptions are more likely to occur along the Mono-Inyo chain than at Long Valley (Hill and others, 1985a). The potential hazards associated with possible future eruptions in the area are outlined by Miller and others (1982).

STRUCTURE

As noted above, Long Valley caldera lies at the northern end of the Owens Valley rift, between the east-facing Sierran escarpment and the west-facing White Mountain front. The pre-Tertiary basement in the area is principally Sierran batholithic rocks and associated metamorphic roof pendents (map A). In the eastern half of the area, the basement consists mainly of Triassic granodiorite, which locally encloses masses of Paleozoic metasedimentary rocks and is intruded by small stocks of Jurassic granite. In the western half of the area, the basement consists mainly of Paleozoic metasedimentary rocks of the Mount Morrison roof pendant and metavolcanic rocks of the Ritter Range roof pendant, which dip nearly vertically and trend generally northwest and are intruded by moderately large Cretaceous granitic plutons. The contact between these contrasting eastern and western basement assemblages passes through the middle of Long Valley caldera, approximately coincident with the Hilton Creek fault (map B), although control for this contact is provided by only two deep intracaldera drill holes aided by the aeromagnetic expression of the sedimentary roof pendant rocks (Kane and others, 1976).

Structural trends in the immediate vicinity of the caldera are dominated by the northwest-trending Sierran frontal fault zone (map B), which parallels the northwest elongation of the Sierran plutons and associated roof pendants. Within the caldera, this northwest regional trend is reflected by the northwest trend of the medial graben on the resurgent dome, as well as by alignment of many of the intracaldera vents (map D). East of the caldera, structural trends are more northerly, subparallel to the White Mountain front, which lies 15 km east of the map area. Formation of the Sierran escarpment and most other Basin and Range escarpments in the area mainly postdates eruption of the precaldera lavas (3.6 Ma), and most of the bounding faults continue to be active, as attested by abundant and occasionally strong earthquakes along them (van Wormer and Ryall, 1980; Ryall and Ryall, 1980; Sherburne, 1980).

Major structual elements east of the caldera include the Benton Range, Casa Diablo Mountain, and Black Mountain—east-tilted fault-blocks bounded on the west by the Benton Range-Casa Diablo Mountain fault zone and the Black Mountain fault, respectively (map B).

These faults, and their many subsidiary splinters, displace Tertiary precaldera trachybasalt flows locally as much as 500 m; they also displace the rhyolite tuff fans of Glass Mountain and the Bishop Tuff by lesser amounts.

Major Sierran frontal faults within the map area include the Round Valley, Hilton Creek, Hartley Springs, Lee Vining, Fern Lake, and Silver Lake faults, all with displacements ranging from 500 to greater than 1,000 m down to the east. The Hilton Creek and Round Valley faults merge south of the map area to form the main west boundary of the Owens Valley rift. Northward, the Hilton Creek fault is defined by an anastomosing system of minor intracaldera faults that project into the Alpers Canyon fault and the Bald Mountain graben, both of which die out north of the caldera—a relation that suggests the precaldera Hilton Creek fault decreased in throw northward and terminated within or just north of the present site of the caldera. Crossing and diverging northnortheastward from the intracaldera continuation of the Hilton Creek fault is a narrow, segmented, grabenlike structure, starting on the northwest sector of the resurgent dome and contunuing north of the caldera as the Sage Hen Peak fault and associated graben. This structure merges with the Baxter Spring and Indian Spring faults, which define a northnorthwest-tending graben that narrows northward and continues as a single prominent west-downward normal fault east of Mono Lake. This complex but more or less continuous zone transecting the east-central part of the caldera consists essentially of two major northnorthwest-trending normal faults, the Hilton Creek in the south and the Indian Springs in the north, downthrown east and west respectively, joined by a northnortheast-trending graben, the Sage Hen Peak fault and graben.

Transecting the western part of the caldera is the Hartley Springs fault, which lies on strike with the Lee Vining fault to the north but is separated from it by an apparent structural gap through Pumice Valley. The Hartley Springs fault projects southward through the west moat of the caldera toward the Laurel-Convict fault, but the latter is a Paleozoic of Mesozoic structure within the Mount Morrison roof pendant with no apparent Cenozoic displacement. Continuation of the Hartley Springs fault through the caldera is suggested by the alignment of west moat rhyolite domes (maps B and D) and by a slight oversteepening of west moat trachybasalt flows along a coincident line, suggesting that they overflowed a preexisting east-downward fault scarp. The decline in displacement along the Lee Vining-Hartley Springs zone from greater than 1,000 m near Lee Vining to 600 m near the caldera suggests that the fault dies out southward, just as the Hilton Creek fault does northward to the east. This en echelon relation of the Hartley Springs and Hilton Creek faults suggests the existence of a northerly inclined ramp between the two faults in precaldera time.

The Silver Lake-Fern Lake fault zone intersects the west wall of the caldera, and the alignment of several postcaldera trachybasalt-trachyandesite vents in the west and southwest moat (map D) suggests that its intracaldera continuation crosses the inferred intracaldera segment of the Hartley Springs fault and arcs into the south moat, where it possibly coincides with the caldera ring fault. This inference suggests that in precaldera time, in the present site of the south moat, the eastward extention of the Fern Lake fault formed a north-downward east-west scarp across the north-sloping ramp between the Hartley Springs and Hilton Creek faults.

Transecting these converging structures in the west moat is a young north-trending, right-stepping, en echelon graben related to the Mono-Inyo Craters fissure system. This structure narrows and continues south of the caldera as a zone of minor faults and fissues in the granite of the High Sierra, along which is located Fish Creek (Iva Bell) Hot Springs, near the south edge of the Devils Postpile quadrangle, and possibly also Mono Hot Springs, about 15 km farther south in the Kaiser Peak quadrangle.

Both the Hartley Springs and Hilton Creek faults terminate as major escarpments at the north and south caldera boundaries, respectively, and continue within the caldera only as minor, discontinuous, en echelon splinters. This marked difference in displacement outside and inside the caldera suggests that in postcaldera time the cauldron block was structurally isolated by the ring fault and that the residual subcauldron magma chamber hydraulically dampened or absorbed intracaldera fault movements; possibly only recently has the chamber roof thickened sufficiently, as a result of cooling and crystallization, to behave rigidly and transmit tectonic stresses across the caldera floor. However, incomplete knowledge of the relative amounts of pre-, syn-, and post-caldera throw on these Sierran faults prevents full development of this hypothesis. A similar difference in displacement is evident farther north where Pumice Valley embays the Sierran front west of the Mono Craters; here both the Lee Vining and Hartley Springs faults terminate as major escarpments at the north and south edges of Pumice Valley. However, there appears to be no large shallow magma body beneath Pumice Valley to provide a similar explanation. An alternative explanation, which may apply to Long Valley as well, is offered by Burisk (Sieh and Bursick, 1986), whereby the space generated by extension along the Sierran front west of the Mono Craters is occupied by dike intrusion beneath the craters, or in the case of Long Valley, by periodic refilling of the magma chamber.

Development of the Long Valley magma chamber and location of the caldera were undoubtedly controlled by structural factors. The caldera clearly formed along a segmant of the Sierran frontal fault where left-stepping en echelon offset is unusually well developed and where the northwest-trending Sierran front diverges from the more northerly trend of the White Mountains front. Both of these relations suggest a region of localized extention that would facilitate rise of mafic magma from depth and where opening of fractures would promote magma stoping, fusion, and fractionation and provide space for the more silicic melts thus generated.

The main Long Valley caldera ring fault, presumably a subcircular or elliptical fault or fault zone bounding the subsided cauldron block, is not exposed, but its general location, coincident with the annular caldera moat, is suggested by the distribution of the moat rhyolite vents peripheral to the resurgent dome. The main ring fault appears to lie well within the present topographic boundary, which was enlarged by landslides, avalanches, and rapid erosion of the initially steep, fault-bounded walls during or shortly after caldera collapse. In the moat, on the east side of Long Valley proper, the position of the main ring fault is suggested by an arc of discontinuous minor faults—the Long Valley fault zone (map B). Although not a prominent escarpment, and possibly in part the result of differential compaction of thick intracaldera sediments, this zone coincides with the steep arcuate gravity gradient (Kane and others, 1976) that is inferred to define the main structural boundary in the subsurface. East of this apparent main fault zone are two subsidiary ring faults, the Watterson Canyon fault, which approximately coincides with the present topographic rim of the caldera, and the Clover Patch fault, which is considerably east of the topographic rim. These subsidiary arcuate faults, which are unique to the eastern side of the caldera, are possibly a consequence of the laccolithic form of the Long Valley magma chamber postulated by Lachenbruch and others (1976), wherein evacuation of the shallow, eastern part of the chamber generated lateral rotational sliding of megaslide blocks rather than vertical collapse during caldera subsidence.

Evidence for postsubsidence modification and enlargement of the caldera is best recorded along this eastern margin. Here, the topographic rim is 3-4 km east of the apparent main ring fault. Immediately south of Glass Mountain the rim coincides with the trace of the Watterson Canyon fault; farther south the rim diverges eastward as much as 2 km from the fault trace and forms the edge of the Volcanic Tableland east of Lake Crowley. In this area the fault trace is partly buried by postcaldera sediments, which show that thick rhyolite tuffs of Glass Mountain (unit Qgt), initially forming the fault-bounded rim, were rapidly eroded and redeposited as coarse alluvial fans (unit Qof). These fan deposits in turn were reworked by the caldera lake into tiers of terrace gravels (unit Qtg; see cross section C-C′). The relations indicate that the topographic rim, particularly east of Lake Crowley, has migrated eastward of both the Long Valley and Watterson Canyon faults and is in large part a fault line scarp. In the western, steep-walled part of the caldera, indirect evidence for landslide enlargement of the walls is found in collapse megabreccias (Lipman, 1976) encountered within the Bishop tuff in at least one intracaldera drill hole (No. 5, table 1).

Table 1.—*Principal drill holes in Long Valley-Mono Craters area*

Map No.	Date	Number	Operator/Location	Elevation (m)	Depth (m)	Temperature Max./Bottom (degrees C)		Reference
1	(1971)	PRC 4397.1/1	Geothermal Resources International south shore Mono Lake, T. 1 N., R. 27 E., sec. 20	1954	1253	46	46	Axtell (1972)
2	(1971)	PRC 4372.1/23–1	Getty Oil north shore Mono Lake T. 2 N., R. 26 E., sec. 23	1957	743	58	58	Axtell (1972)
3	(1976)	LV 66–29	Republic Geothermal central Long Valley T. 3 S., R. 29 E., sec. 29	2128	2109	72	72	Smith and Rex (1977)
4	(1979)	CP–1	Union Oil of California east flank, resurgent dome T. 3 S., R. 28 E., sec. 15	2226	1846	142	103	Gambill (1981)
5	(1979)	M–1	Union Oil of California Casa Diablo Hot Springs T. 3 S., R. 28 E., sec. 32	2226	1605	146	146	Gambill (1981)
6	(1982)	82–6	Union Oil of California southeast of Mono Craters T. 1 S., R. 27 E., sec. 33	2341	899	28	28	R.F. Dondanville, written commun., 1987
7	(1982)	PLV–1	Phillips Petroleum west moat of caldera T. 3 S., R. 27 E., sec. 22	2585	715	110	110	Benoit (1984)
8	(1982)	PLV–2	Phillips Petroleum west moat of caldera T. 3 S., R. 27 E., sec. 3	2347	636	68	68	Benoit (1984)
9	(1985)	IDFU 44–16	Unocal Geothermal west moat of caldera T. 3 S., R. 27 E., sec. 16	2451	1796	215	180	Suemnicht, 1987
10	(1986)	SR (RDO–8)	Lawrence Berkeley Laboratory Shady Rest, Mammoth Lakes T. 3 S., R. 27 E., sec. 25	2374	715	202	202	Wollenberg and others (1987)

Postsubsidence intracaldera volcanic vents show a strong preference for alignment along Sierran frontal trends (map D). Most early rhyolite vents are localized along northwest-trending faults bounding the medial graben. Although the moat rhyolites are arranged peripheral to the resurgent dome, individual vents particularly those north and west of the resurgent dome, are aligned along intracaldera projections of the Hilton Creek and Hartley Springs faults, respectively. Similarly, several postcaldera trachybasalt-trachyandesite vents are aligned along the intracaldera projection of the Fern Lake fault. In contast, vents for the Mono-Inyo Craters volcanic chain trend generally north-south, although those at the soutern end near Pumice Butte form a broad zone that trends more northnortheasterly. The rhyolite vents of the Mono–Inyo Craters chain form a generally linear northerly trend, but locally they appear to consist of northnortheast-trending en echelon segments (Fink, 1985). Vents of the central segment of the Mono Craters arc appear to be aligned on such a trend, but the curvature shown by the vents at the northern and southern extremities, together with the location of the granodiorite-metasedimentary rock contact beneath the craters, as exposed in the Mono Craters tunnel (map A), reflects the structural influence of the mylonitized border of the Aeolian Buttes pluton (Kistler, 1966b).

Precaldera volcanic vents do not show any strong preferential alignment (map C). They tend to be concentrated around the northern margin of the caldera, but this probably is more apparent than real, as over much of the area precaldera rocks are buried by the Bishop Tuff and younger tephra, and within the caldera they are downfaulted and buried by caldera fill. It is evident, however, that quartz latite vents are concentrated near the caldera, whereas basaltic vents are more widely distributed. Both north and northeast vent alignments are evident locally, but neither trend is dominant or traceable for far due to younger cover. Few vents, either basaltic or quartz latitic, occur immediately south of the caldera, but to the southwest, regional mapping (Moore and Dodge, 1980; Bailey, 1982a) shows that 20 or more basaltic vents are scattered over a large area in the High Sierra. The precaldera rhyolites of Glass Mountain form an arcuate ridge along the northeast caldera rim, but the actual vents do not have a demonstrably arcuate pattern clearly contolled by incipient caldera ring fractures; as previously suggested (Bailey and others, 1976). Many additional rhyolite vents probably are buried within the caldera.

HOT SPRINGS, FUMAROLES, AND THERMAL ACTIVITY

Long Valley caldera is recognized as a region of high heat flow (Lachenbruch and others, 1976), and hot springs, fumaroles, and active hydrothermal alteration are prevalent in many parts of the caldera, particularly in the southeast moat and adjacent flanks of the resurgent dome. Notable concentrations of fumaroles and (or) hot springs occur at Casa Diablo Hot Springs, Fumarole Valley, Hot Creek Springs, Little Hot Creek Springs, and southeast of Big Alkali and Little Alkali Lakes. Fine-grained black pyrite (FeS_2) and travertine ($CaCO_3$) are being actively deposited in many of these springs, and native sulfur is a common sublimate around fumaroles. Siliceous sinter also occurs in some of these areas, but it is the result of former higher temperature hot-spring activity and is not forming at the present time. (An isolated occurrence of siliceous sinter is reported by Krauskopf and Bateman, (1977) in the faulted Tertiary basalt terrane in the far northeast corner of the map area, but its age and relation to possible thermal sources is not known.) The coarse, silica-cemented, deltaic sediments (unit Qsc) that flank the east and south sides of the resurgent dome are pervasively penetrated by fossil fumarolic pipes, suggesting an unusually intense episode of thermal activity during deposition of this unit about 300,000 yr ago. Quite possibly, eruption of the southeast moat rhyolites at about this time was accompanied by renewed uplift of the resurgent dome and by intensified hydrothermal activity, thereby causing the shedding of coarser detritus off the resurgent dome into the caldera lake and concomitant widespread silica-cementation of the deltaic sediments thus formed. Pervasive hydrothermal alteration of lake beds beneath the deltaic sediments is evident near many faults on the resurgent dome. This alteration is predominantly argillic, and near Little Antelope Valley commercial grade kaolin has been mined intermittently since the 1950's (Cleveland, 1962).

The prevalence of thermal activity in Long Valley has encouraged geothermal exploration, and several commercial and scientific exploratory holes have been drilled to depths ranging from 100 to 2,000 m. (The location of drill holes deeper than 500 m for which data are available are shown on the geologic map and cross sections and are keyed by number to supplementary data and references presented in table 1.) Chemical geothermometry of hot spring and drill-hole waters suggests that deep reservoir temperatures range from 200 to 280 °C, but the maximum temperature attained in drill holes is 220 °C (Suemnicht, 1987). Drilling to date (1987) indicates that thermal waters are confined to relatively shallow aquifers, mainly within the intracaldera Bishop Tuff, at depths above 1,500 m in the west moat and above 700 m in the south moat, but neither the location of the deep thermal source nor where or how the aquifers are fed is known. Presumably residual magma beneath the resurgent dome is the principal heat source, but higher temperatures at greater depths in the western part the of the caldera suggest a possible westerly source (Sorey, 1978, 1985; Blackwell, 1985).

Surface thermal acativity associated with the Mono–Inyo Craters volcanic chain is restricted to the very southern and northern extremities of the chain. In the south, active fumaroles occur on the north and south slopes of Mammoth Mountain, and hot springs occur to the west at Reds Meadow Hot Springs and to the south in Fish Creek at Iva Bell Hot Springs. In the vicinity of Mono Lake, fumaroles and hot springs occur on Paoha

Island, and hot springs occur northeast of Black Point cinder cone. Although surface thermal features are confined to the extremeties of the chain, elevated temperatures do occur in the subsurface along it. During the course of constructing the Mono Craters tunnel (map A), warm water (35 °C) and copious amounts of carbon dioxide were encountered when passing through feeder dikes (Gresswell, 1940), and a bottom-hole temperature of 80 °C was attained in the Obsidian Dome conduit drill hole in the Inyo Craters chain (Eichelberger, and others, 1985).

GLACIATION

The oldest currently recognized glacial deposit in the Long Valley area is the McGee Till (unit Qmg) (Blackwelder, 1931), exposed on the south rim of the caldera on McGee Mountain. It overlies a scoriaceous basaltic vent complex having a radiometric age of 2.6 Ma (Dalrymple, 1963), and although generally considered Pleistocene in age, it could be older. The 3-Ma here-abandoned Deadman Pass Till (Curry, 1966) is now recognized as a product of erosion and weathering of pyroclastic deposits unusually rich in accidental granitic and metamorphic clasts (R.A. Bailey, N.K.Huber, and R.R. Curry, unpub. data, 1987), and it is here mapped as part of the quartz latite of San Joaquin Ridge (unit Tsjt).

During Sherwin glaciation (Blackwelder, 1931), thick, now much eroded, lobate masses of till were deposited at many places along the Sierran front, particularly at the mouth of Rock Creek (Putnam, 1960; Sharp, 1968) and south of Lee Vining Creek (Kistler, 1966b). Scattered remnants of the Sherwin till or related outwash deposits also occur beneath the Bishop Tuff at Aeolian Buttes (Putnam, 1949), between precaldera basalt and the Bishop Tuff near Crestview, and between older Quaternary gravels (unit Qog) and Bishop Tuff near Alpers Canyon on the northwest rim of the caldera. Sporadic granitic blocks, also possible remnants of Sherwin outwash, rest on precaldera basalt east of Bald Mountain in upper McLaughlin Creek drainage. Sherwin outwash, clearly related to the Sherwin Till at the mouth of Rock Creek, also crops out between Tertiary basalt and the Bishop Tuff in the walls of Owens River Gorge (Putnam, 1960; Sharp, 1968). The age of the Sherwin Till is not known precisely, but on the basis of depth of weathering of the till beneath the Bishop Tuff, Sharp (1968) estimates that it may predate the tuff by a few ten-thousand years but not by as much as 100,000 years.

Fragmentary evidence suggest that glaciers were active in the Sierra Nevada at the time of eruption of the Bishop Tuff and for some time after formation of Long Valley caldera. Whether this glaciation was a continuation of the Sherwin glaciation or a distinctly younger stage is uncertain. Evidence for glaciation coeval with emplacement of the Bishop Tuff is displayed at Little Round Valley, just south of Lake Crowley, where a remarkable display of fossil fumarolic pipes and other unusual depositional features in the tuff suggest the proximity of a ice-tuff contact (R.A. Bailey, unpub. data, 1986). Clear evidence for active glaciation during early postcaldera time is provided by the ice-rafted erratics dispersed on early intracaldera lake terraces.

Glacial deposits of Wisconsin age in the Long Valley-Mono Basin area have been studied for many decades, but their subdivision and correlation remain problematic and are to some degree controversial. The glacial subdivisions shown on this map are based on casual observations made over several years, plus two weeks of intensive comparative examination of most of the moraines in the map area, qualitatively applying the semiquantitative critera of Sharp (1969). In most drainages, the subdivisions shown are in general agreement with recent studies; where they differ, or where subdivisions have not been previously made, they are commonly queried (for example, Qmb? in Laurel, Sherwin, and Convict Creeks), and they require verification by more detailed study. In the Long Valley area, from Rock Creek northwest to Mammoth Creek, two distinct morainal lobes can be readily distinguished in most of the major Sierran valleys; in general agreement with most previous work, the older and younger lobes are mapped as deposits of Tahoe and Tioga glaciations, respectively. In Mammoth Creek, where Wisconsin age glaciers spread out as broad piedmont lobes, the youngest moraine overlies a trachybasalt flow with a radiometric age of 64 ka (Bailey and others, 1976) and is mapped as Tioga in age. The Casa Diablo Till (Curry, 1971) lies between trachybasalt and trachanydesite flows equivocally dated at 64 and 126 ka, respectively; it may be Tahoe or Mono Basin in age. In the Mono Basin area, multiple Wisconsin - age glaciations are more readily distinguishable than in Long Valley, three or four distinct pairs of lateral moraines being recognizable in most of the major valleys. These are mapped as deposits of Mono Basin, Tahoe, Tenaya, and Tioga glaciations, in general agreement with mapping by Sharp and Birman (1963) and Gillespie 1982).

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