Geochronology of the Long Valley Caldera as related to the Long Valley magma chamber

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The region of Long Valley, California, nestled east of the White Mountains and west of the Sierra Nevada crest, has a long history of volcanism. The Long Valley volcanic field occupies an area of ~4,000km2 and is the largest of three volcanic fields in the Owens Valley Rift. Its first phases of volcanism occurred as early as ~4Ma during Pliocene time and subsequent volcanic activity has continued intermittently through the Holocene (Bailey, 2004). The development and rise of the Long Valley magma chamber occurred from 1.9 to 0.6 Ma, spanning a period of at least 1.3 million years (Bailey, 1976). The three major phases of volcanism that can be attributed to the Long Valley magma chamber, the rhyolites of Glass Mountain, the Bishop Tuff and the post-caldera rhyolites (Bailey, 1976) are all centered on the present Long Valley caldera. This paper will trace the geochronology of volcanic events within the Long Valley caldera, in particular, those associated with the Long Valley magma chamber.

Prior to the inception of Pliocene volcanism, Long Valley region was a well-eroded upland with streams that drained to the west (Bailey, 1989b). Jurassic and Cretaceous granodiorite and granites of the Sierra Nevada batholith and Paleozoic and Mesozoic metamorphic rocks of the Mount Morrison and Ritter range roof pendants make up the underlying, pre-Tertiary basement rocks (Bailey, 1976). The present topography centers around the Long Valley Caldera, an elongate elliptical depression stretching 32 km from east to west and 16 km from north to south located 50 km northwest of Bishop, California and 30 km south of Mono Lake. (Figure 1) The floor of the caldera is characterized by low, sage and grass-covered relief in the eastern half, where the elevation starts near 2070 m, and by higher, more densely forested hills in the western half, where the average elevation reaches 2440 m (Bailey, 1976). A group of faulted and dissected hills distinguishes itself from the west central caldera floor with elevations of 2590 m, creating an annular moat around the edge of the caldera. Steep walls

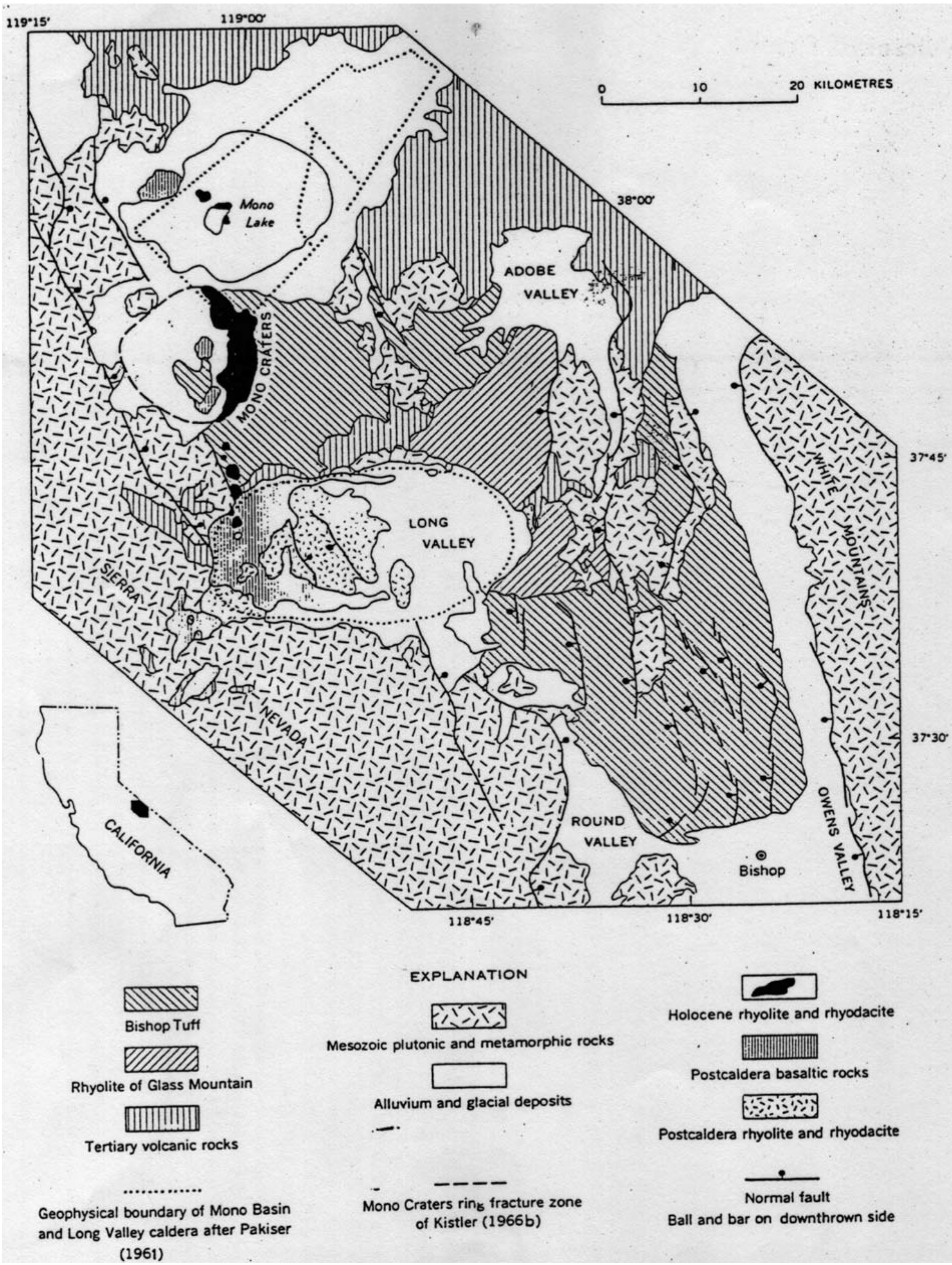

Figure 1. Index and generalized geologic map of the Long Valley- Mono basin area
Taken from Bailey (1976)

enclose the caldera floor on all sides, except to the east and southeast. Here, the southeastern rim gently arches into the Volcanic Tableland, with a mere relief of 250 m from the caldera floor. In contrast, the tops of McGee and Laurel Mountains dominant the southern rim at an elevation of 3050 m, placing remnants of a late Tertiary erosion surface1000 m above the caldera floor (Bailey, 1976). The west wall is 500m of relief formed by the eastern edge of the San Joaquin Mountains and the Two Teats. To the north and northeast, Bald Mountain and Glass Mountain create the highest wall of 1200 m, coming to an elevation of 3350 m.

The first phase of volcanism in the Long Valley region started 3.6 to 3.0 Ma with eruptions of trachybasaltic-trachyandesitic lavas to the northwest of the present caldera (Bailey, 1976). Because remnants of these initial eruptions are now distributed over an area of 4000 km2 in the caldera vicinity (Bailey, 1989b), Hildreth (2004), after Moore and Dodge (1980), attributes these eruptions to partial melting of the mantle caused by extensional forces associated with the Basin and Range province. Mafic magma was extruded across a wide belt- stretching from the present Long Valley caldera, to 40 km southwest into the Sierra Nevada and 30 km northeast into the Adobe Hills (Hildreth, 2004). Bailey (1989b) concurs that the irregular scattering of these volcanic rocks implies a broad mantle source region. This is a unique period in the volcanic history of the Long Valley region, however, as the subsequent volcanic activities were all based around local concentrations of magma (Hildreth, 2004). The first magmatic center erupted precaldera rhyodacites from a 20 km wide zone on the northwestern rim of the caldera, creating the main mass of San Joaquin Mountain, the Two Teats, and Bald Mountain (Hildreth, 2004). These younger flows, dated from 3.2 to 2.6 Ma, are not directly related to the Long Valley magma chamber but most likely represent the beginning of accumulation and differentiation of magma deep within the crust from which the Long Valley magma chamber eventually formed (Bailey, 1989b).

The second phase of volcanism in Long Valley emerged when this center of mantle-driven crustal-melt moved 20 km eastward, to the northeast rim of the present caldera, underneath Glass Mountain (Hildreth, 2004). (Figure 2) Glass Mountain (2.2 to 0.8 Ma) rises 1190 m (Metz and Manhood, 1985) above the caldera floor and constitutes domes, flows and tuffs of aphyric to sparsely porphyritic rhyolite. (Bailey, 1976). Well preserved ashflow tuff, pumice falls and alluvial debris associated with lavas of the Glass Mountain center can be seen on the north and southeast side of the mountain. It is assumed that similar deposits existed on the southwestern side before the mountain was down faulted with the collapse of the caldera (Bailey, 1976). These ashes are identified as far away as Ventura County, California and Beaver County, Utah (Bailey, 1976). The total volume of magma extruded from the Glass Mountain center is approximately 80-120 km3 (Hildreth, 2004). The location of Glass Mountain's vents and intrusive centers, which run in an arc nearly parallel to the northeast wall of the caldera, suggests that these eruptions represent early leakage from the Long Valley magma chamber caused by an insipient caldera ring fracture (Bailey, 1976).

The emplacement of Glass Mountain involved at least 60 eruptive units of high silica rhyolite, all containing between 76.6 -77.7 % silica, with as many as 17 of these events being Plinian eruptions of ashfall and pumice (Hildreth, 2004). K-Ar ages based on sanidine from Glass Mountain obsidian and biotite rhyolite range from 0.90 Ma to 1.92 Ma and suggest that the eruptions of this volcanic complex took place over a period of at least 1 million years (Bailey, 1976). Two discernable eruptive sequences can be determined based on the composition of the lavas (Hildreth, 2004). An initial sequence of eruptions (2.113-1.20 Ma) (Metz and Manhood,

1985) included at least twenty-four eruptive units and produced high silica, but chemically varied rhyolites (Hildreth, 2004). The chemical variation between these rhyolites suggests that they were extruded from several different magma bodies that existed at varying stages of evolution (Hildreth, 2004). The second sequence of eruptions (1.10-.79 Ma) (Metz and Manhood, 1985) consisted of at least 35 eruptive units of lavas with relatively similar chemical composition (Hildreth, 2004). The second episode was also marked by eruptions of higher volumes of material (Metz and Manhood, 1985). The relative chemical homogeneity of the younger lavas suggests that they were erupted from an integrated magma chamber (Metz and Manhood, 1985). Thus, during the eruptive lull between the two episodes, the magma chamber evolved from its period of coalescence to the mature, high-level Long Valley magma chamber (Metz and Manhood, 1985).

The third phase of volcanism in the Long Valley region – the cataclysmic eruption of the Bishop Tuff and the resulting caldera subsidence- is associated with this mature magma chamber. Hildreth (2004) suggests this concentration of crustal-melt completed its third and final westward migration sometime between the final Glass Mountain eruption (0.79 Ma) and before the cataclysmic eruption of the Bishop Tuff (0.76 Ma). This final location – 20 km west of Glass Mountain – would be the magmatic center of subsequent volcanic activity within the Long Valley Caldera (Hildreth, 2004). (Figure 2) 0.76 Ma (Lipshie, 2001), the Long Valley magma chamber underwent a catastrophic eruption of an enormous volume of rhyolite magma (~600km3) (Bailey, 1976), now known as the Bishop Tuff (Izett, 1988, from others, Gilbert, 1938, Bailey et al., 1976). The sudden partial emptying of the magma chamber caused the roof to collapse into itself, leaving the elongate depression of the Long Valley Caldera (Bailey, 1989b).

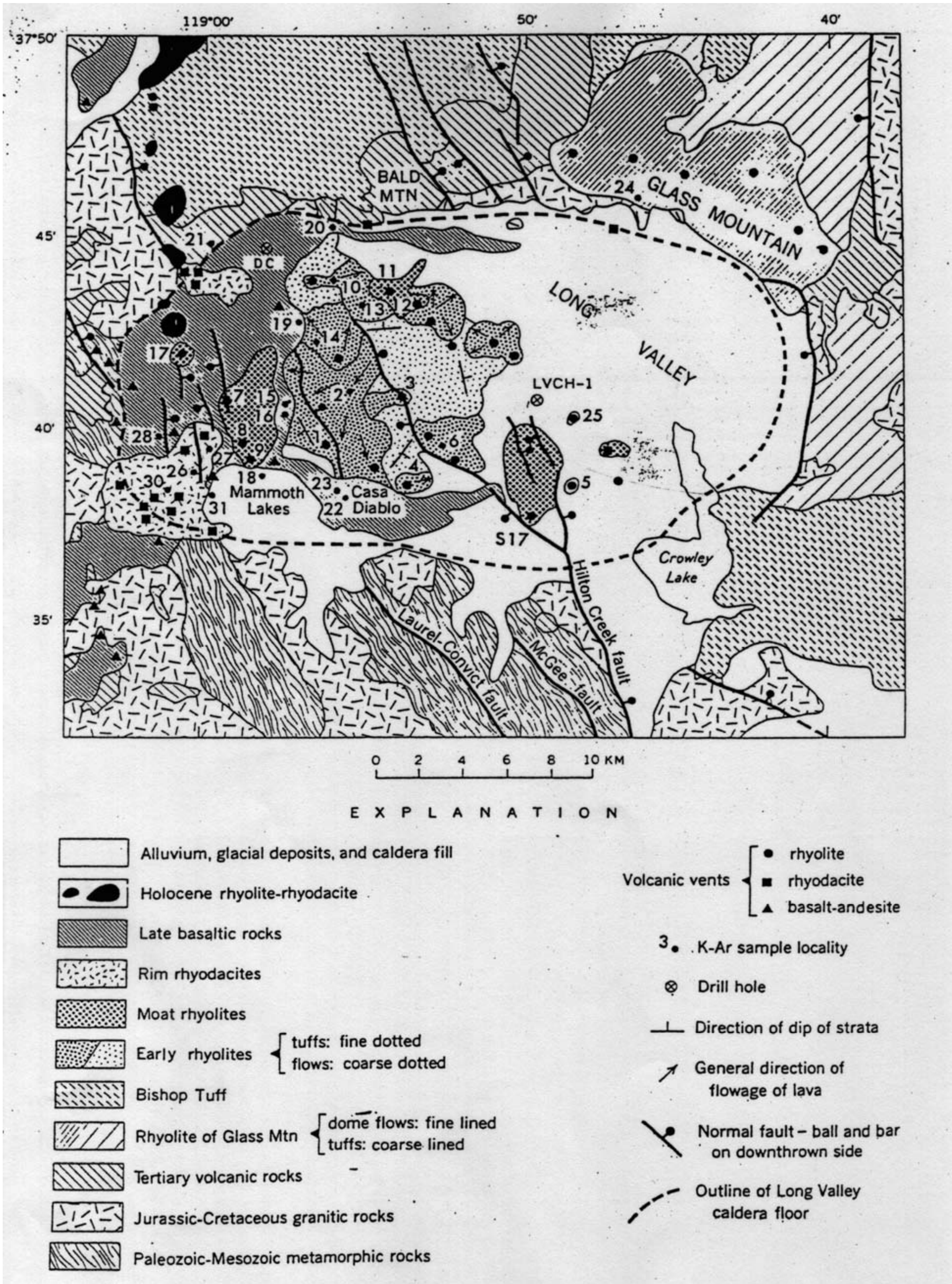

Figure 2. Generalized geologic map of Long Valley caldera
Taken from Bailey, (1976)

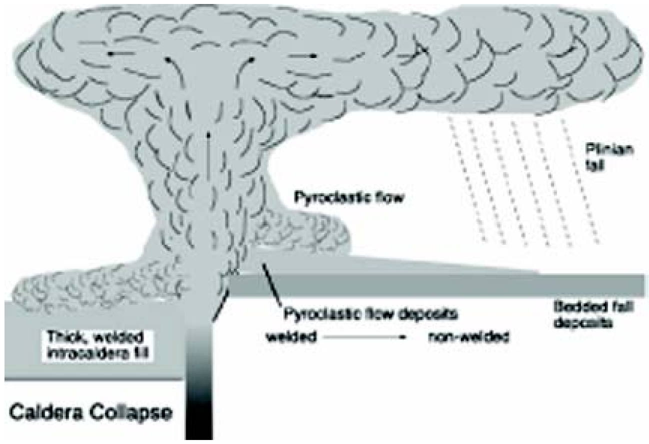

Figure 3. Schematic illustration of depositional process and facies in explosive volcanic eruptions.

Taken from Wallace et al, (2002)

Lack of erosion or reworking of the material between emplaced eruptive units suggests that the eruption occurred extremely rapidly, possibly in a matter of days or weeks (Hildreth and Manhood, 1986). Izett et al (1988) have described two distinct eruptive stages based on the stratiography of the Bishop Tuff. (Figure 3) First, initial Plinian eruptions sent massive amounts of pyroclastic material into the stratosphere, depositing a layer of air-fall pumice at the base of Bishop (Bailey, 1976). Fine-grained pyroclastic material, known as Bishop Ash, was also produced by these eruptions and then blanketed by atmospheric winds across more than a million square kilometers of the Western United States (Izett, 1988). Deposits of the Bishop Ash have been identified as far north as southern Idaho, as far east as central Nebraska, and as far south as southern New Mexico (Izett, 1988). (Figure 4) This Plinian eruptive cycle was followed by effusive eruptions of thick ash flows that radiated from the caldera, ponding to the south in the lower relief of the Owens River, to the north in the Adobe Valley, to the northwest into Mono basin, and to the west over what is now occupied by Mammoth Mountain (Bailey, 1976). (Figure 2) The thickness of these deposits varies in each location but the Volcanic Tableland, formed by welded Bishop south of the caldera, measures up to 200m (Bailey, 1989b). These outflow sheets of Bishop Tuff extend across an area of 1040-1150 km2, with an estimated volume of 140km3 (Bailey, 1976) but they represent less than half of the total magma extruded as Bishop Tuff. The remaining two-thirds of

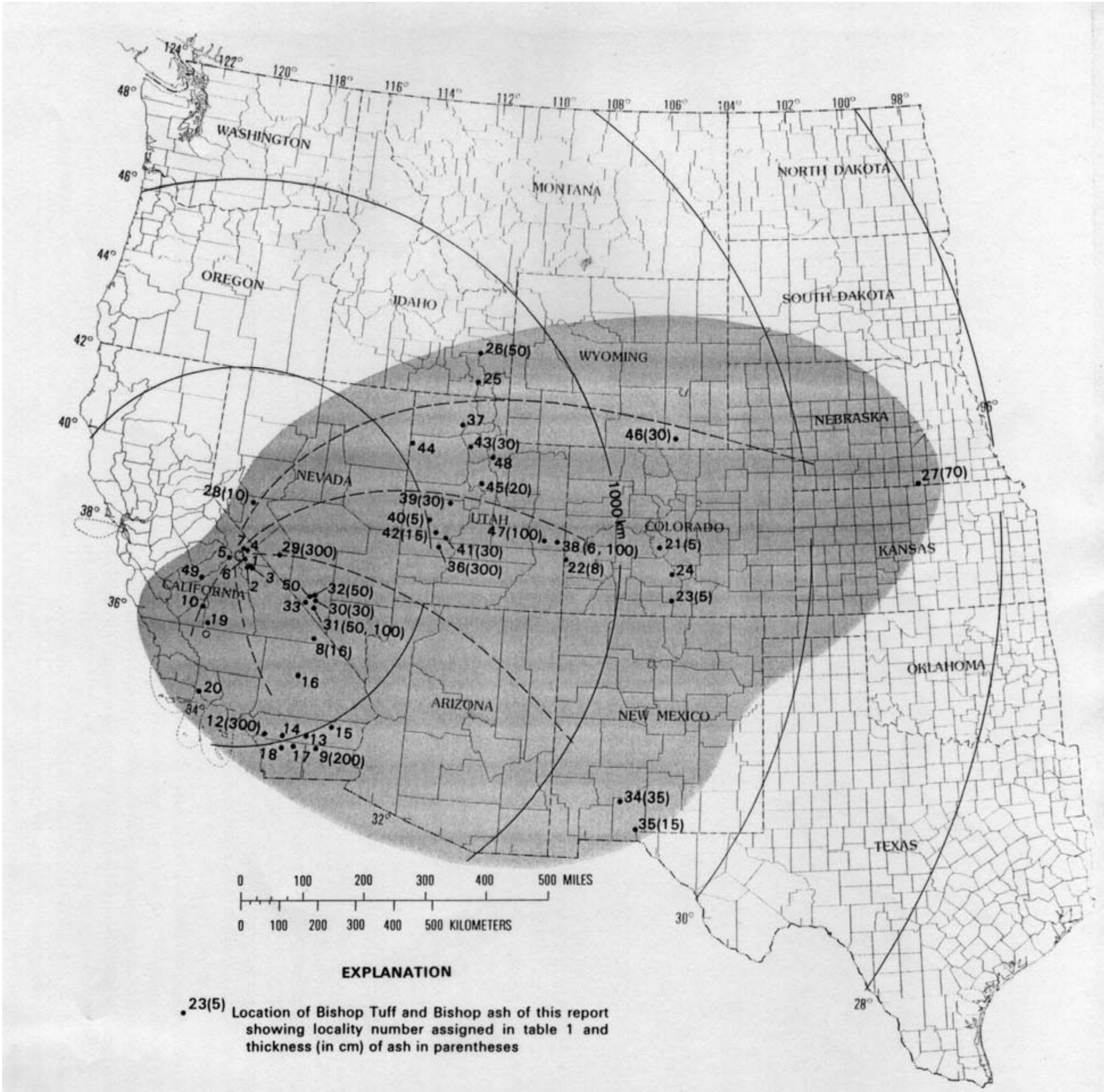

Figure 4. Map showing thicknesses of Bishop Ash beds in the Western United States. Shaded area indicated inferred extent of original Bishop ash fall.
Taken from Izett, (1988)

the Bishop Tuff accumulated within the depression of the caldera (Bailey, 1976). No intracaldera Bishop Tuff is exposed but drilling core samples have identified the presence of as much as 1500 m of Bishop Tuff buried within the caldera beneath younger volcanic and sedimentary rocks (Bailey,1989b). (Figure 5)

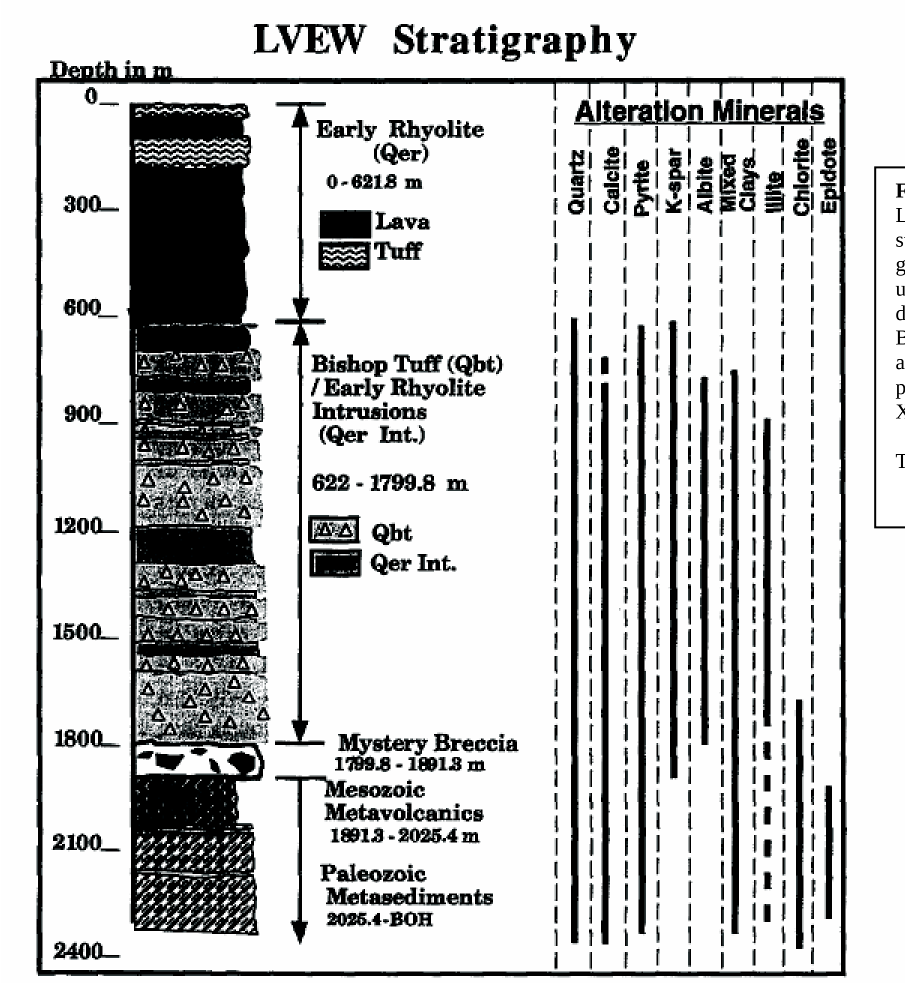

Figure 5. Major stratigraphic units in LVEW are identified. The scale of the stratigraphic diagram precludes displaying greater detail such as thin Qer intrusive units. 'Mystery Breccia' is an informal designation for the breccia unit underlying Bishop Tuff. The location and type of alteration mineralogy is based on petrographic thin section observations and X-ray diffractometry

Taken from McConnell et al., (1996)

For the most part, the composition of the Bishop Tuff suggests an orderly, progressive tapping of a "classic zoned magma chamber" (Hildreth and Manhood, 1986). (Figure 6) The Bishop Tuff is composed of rhyolite with 5-25% (Bindeman and Valley, 2002) phenocrysts of quartz, sanidine, plagioclase, biotite, and Fe-Ti-oxides (Bailey, 1976). Bindeman and Valley (2002) describe the magma chamber has having a relatively homogenous composition of major elements and a temperature gradient of 70 degrees Celsius. There is an observed upward progression towards less-evolved pumice compositions and higher Fe-Ti oxide temperatures throughout the Bishop Tuff deposits (Hildreth and Manhood, 1986). The contents of the top of the magma chamber are represented by deposits of aphyric, allanite-bearing, pyroxene free, high-

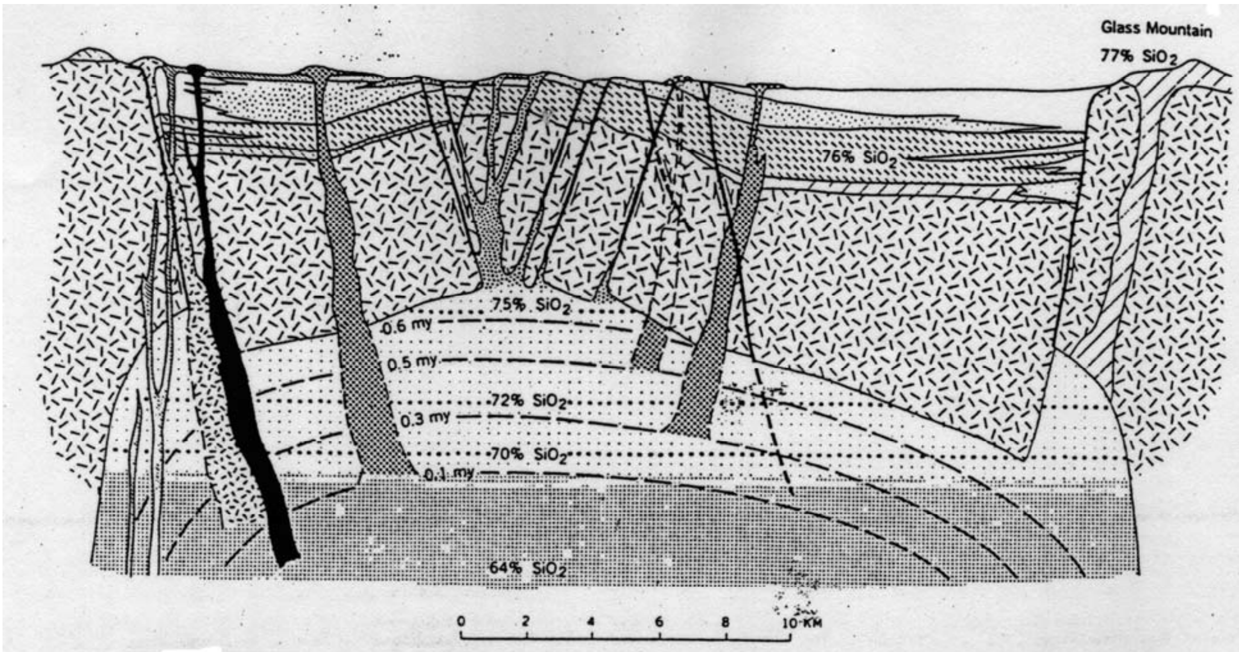

Figure 6. Schematic east-west cross section through Long Valley cadlera and its subjacent showing hypothetitical changes in chemical composition and depth of crystallization with time. The heavily dotted part of the chamber is rhyodacite magma; lightly dotted part it rhyolitic magma. Horizontal dotted lines show silica gradient in the vertically zoned chamber. Curved dashed lines show depth of crystal-liquid interface (depth to residual magma) at specified times (millions of years ago). Formation patterns are the same as in **Figure 2** except basement plutonic and metamorphic rocks (hachured) are undivided. The vertical scale is unspecified because of uncertainties in thickness of intracaldera units and depths in chamber.

Taken from Bailey, (1976)

silica rhyolite pumice, with a low Fe-Ti-oxide temperature of ~720 degrees Celsius (Bindeman, 2002). The deeper sections of the magma chamber are represented by the ignimbrite sheets that have higher concentrations of crystals, are pyroxene-bearing, and exhibit gradual increases in Fe-Ti-oxide temperatures up to ~790 degrees Celsius (Bindeman and Valley, 2002). A correlation can thus be made between the rise in temperature of the magma chamber and the composition and the variety of phenocrysts in the Bishop Tuff deposits (Hildreth and Manhood, 1986). Hildreth and Manhood (1986) propose that these changes in composition, for example the first appearance of pyroxene and the last disappearance of allanite in the emplaced units of Bishop Tuff, can be used as loose indicators of the eruptive sequence.

The studies of Hildreth and Manhood (1986) indicate that the eruption of the Bishop Tuff followed a pattern common to caldera-forming eruptions, transitioning from a single vent, Plinian eruption phase (that may or may not include associated ashflows fed by the caldera collapse) to a ring-vent phase that later results in the caldera subsidence. The composition of basal airfall and the earliest ashflow deposits of Bishop Tuff, which lack prevolcanic lithic fragments but are rich in metasiltstone, metapelite, quartzite and Wheeler Crest Quartz Monozonite, suggests that the initial Plinian vent opened in the south central caldera, near the Hilton Creek Fault (Hildreth and Manhood, 1986). Multiple vents began opening in a counter-clockwise direction around the circumference of the caldera after ~50km3 of magma was erupted as fallout and ~100km3 magma was erupted as outflow sheets (Hildreth and Manhood, 1986). The ring fracture phase thus accounts for ~ 100km3 of the Bishop Tuff that was deposited outside the caldera and another ~400km3 of Bishop Tuff that remained inside the subsiding caldera (Hildreth and Manhood, 1986). Assuming that the total volume of magma expelled as the Bishop Tuff was ~500km3, Hildreth and Manhood (1986) calculated that the caldera started to form after 20% of the total volume of Bishop Tuff had been erupted.

The circumference of the caldera collapse most likely occurred around a string of arcuate ring faults, now buried within the caldera (Bailey, 1976). (Figure 7) The only outer ring fault that is now exposed is visible along the east wall for a distance of 12km (Bailey, 1976). Dimensions of the original ring fault zone (an oval of 12x22km), as inferred by gravity, drill holes and vent distribution, are significantly smaller than the topographic area now recognized as the Long Valley Caldera. The area (~220km2) of this original ring fault zone only represents 55% of the area (~400km2) of the present caldera. Bailey (1976) suggests this enlargement is due to slumping of the caldera walls during the initial, unstable period of formation as well as by subsequent erosion.

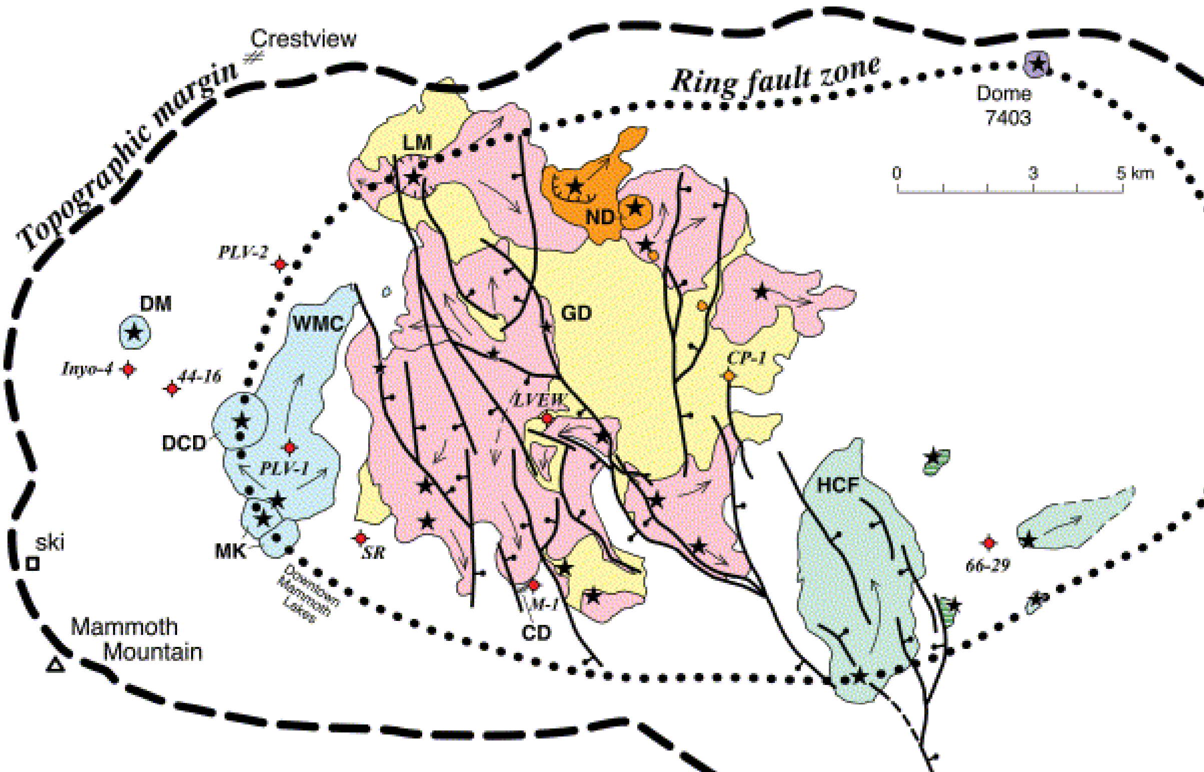

Figure 7. Postcaldera Long Valley Rhyolites. Used to show the Ring Fault Zone. Early Rhyolite (760–650 ka) lavas in red and tuffs in yellow. Black stars indicate 13 Early Rhyolite vents exposed (as well as vents for younger rhyolites). Three clusters of Moat Rhyolite lavas crop out in north (527–481 ka, orange), southeast (362–288 ka, green), and west (161–101 ka, blue); all are crystal-rich except three (of the five) units in the southeastern cluster, which are phenocryst-poor rhyolite lavas (distinguished in unpatterned pale green). Arrows generalize lava flow directions. Place–name abbreviations: CD=Casa Diablo geothermal plant; DCD=Dry Creek dome; DM=Deer Mtn; GD=Gilbert Dome; HCF=Hot Creek flow; LM=Lookout Mtn; MK=Mammoth Knolls (two domes); ND=North dome; Ski=Mammoth Mtn ski complex; WMC=West Moat Coulee. Drillholes mentioned in text are: LVEW=Long Valley Exploratory Well; SR=Shady Rest; others are named as designated on map—CP-1, M-
Taken from Hildreth, (2004)

The collapse of the caldera was followed by an almost immediate renewal of volcanic activity that constitutes the fourth phase (750-100 Ka) of volcanism in the Long Valley region. Hildreth (2004) recognizes these post-caldera eruptions along the caldera's original ring-fault zone, including the Early Rhyolites and the Moat Rhyolites, as being derived from a reorganized, convectively mixed, and thermally reconstructed Long Valley magma chamber. As no mafic magma was extruded with the rhyolite of the Bishop Tuff (Bindeman and Valley, 2002), quenched globular basaltic inclusions in these post-caldera rhyolites suggest that mafic magma had been re-introduced into the Long Valley magma system after an absence of 1.8 million years (Bailey, 2004).

The Early Rhyolites are an accumulation of ~100km3 crystal-poor rhyolite tuffs, domes and flows in the central part of the caldera. K-Ar measurements date the formation of the Early Rhyolites between 0.73-0.63 Ma, meaning the eruptions started only 0.03 million years after the eruption of the Bishop Tuff (Bailey, 1976). The Early Rhyolites are aphyric to sparsely porphyric (Bailey, 1976) and are compositionally similar to the latest eruptions of Bishop Tuff, although they exhibit a much lower percentage of phenocrysts (Hildreth, 2004). Sanidine and quartz, which have a significant presence (15-25%) in the Bishop Tuff, are largely absent in the Early Rhyolite tuffs which only contain 0-3% of plagioclase, orthopyroxene, Fe–Ti oxides, and biotite, as well as traces of apatite, zircon, and pyrrhotite (Hildreth, 2004). Bailey (1976) suggests the low crystal content to be evidence of near-liquid temperatures of the lavas at the time of eruption. At least 13 source vents (Hildreth, 2004) have been identified for the Early Rhyolites but Bailey (1976) suggests the probability of additional vents buried within the southeastern caldera moat. Hildreth (2004) proposes that these eruptions were most likely subplinian and volumetrically small as no distal ash deposits associated with the Early Rhyolites have been identified outside Long Valley.

Concurrent to the eruption of the early rhyolites, a resurgent dome formed between 730 and 650 ka (Bailey, et al., 1989b). The subcircular structural uplift has a diameter of 10km and rises 500m above the surrounding moat, with the highest point marked by Gilbert Dome at an elevation of 2626m (Hildreth, 2004). The northwest orientation of the keystone graben, which can be seen in a 5km-wide fault through the resurgent dome, is consistent with other structural trends along the east of the Sierra front (Bailey, 1976). Bailey (1976) suggests lines of structural weakness within the Mount Morrison roof pendant determined the resurgent dome's west-central location on the caldera floor. During most of its history, the resurgent dome existed as an island in the Pleistocene Long Valley Lake as suggested by the remnants of lake terraces and glacial erratics (Bailey, 1976). First, the elevations of outward tilting lake terraces on the resurgent dome exceed elevations of the highest lake terraces on the caldera walls by as much as 35m (Bailey, 1976). Secondly, beach pebbles on the dome reach elevations at least 80 m higher than those found on the caldera walls (Bailey, 1976). Lastly, granitic and metamorphic glacial erratics were found 200 m higher on the dome than the highest terrace on the caldera walls. These several pieces of evidence also verify the positive uplift of the dome rather than relative down dropping of the surrounding caldera moat (Bailey, 1976). Because the radial tilting decreases progressively with younger flow units on the dome, Bailey (1976) suggests that the uplift was slowing down towards the final eruptions of the Early Rhyolites and K-Ar ages indicate the resurgence occurred within 100,000 years of the caldera collapse (Bailey et al., 1989b) but no definitive ending date can be determined (Bailey, 1976). The cause of the resurgent dome is unclear, however (Hildreth, 2004). Hildreth (2004) suggests this structural uplift was not necessary associated with post-caldera eruptive activity but perhaps renewed buoyancy or intrusion of the remaining viscous magma in the Long Valley magma chamber.

After the uplift of the resurgent dome, the caldera experienced about 100,000 years of quiescence. Then, crystal-rich rhyolite began to erupt from vents in the caldera moat (Bailey, 1976), probably from ring fractures associated with the resurgent dome (Bailey, et al.,1989b). Bailey (1976) informally identified these eruptions as the Moat Rhyolites but Hildreth (2004) specifies three eruptive centers among these rhyolites: (1) the north-central rhyolite chain, (2) the southeastern cluster and (3) the west moat rhyolites. Starting in the north and progressing in clockwise succession around the resurgent dome, these centers erupted in 200,000-year intervals at 500, 300, and 100ka, respectively. These rhyolites are coarsely porphyrite hornblende-biotite with as much as 20% phenocrysts of hornblende, biotite, quartz, sanidine, and plagioclase (Bailey, 1976). The common forms of the Moat Rhyolites are steep-sided domes and thick flows (Bailey, 1989b). The silica content (72-74%) of the Moat Rhyolites resembles that of the Early Rhyolites but the Moat Rhyolites contain much less K2O (4.7%) and Ba (680–715 ppm) (Hildreth, 2004). Both the structure and crystal content of the Moat Rhyolites suggest they erupted with higher viscosity and lower temperatures than the early rhyolites and could indicate the start of cooling and crystallization in the magma chamber (Bailey, 1989b).

The total volume of magma extruded as the Moat rhyolites is insignificant when compared with the previous eruptions of the Long Valley magma chamber (Hildreth, 2004). The north-central rhyolite chain erupted ~1km3 of magma from five extrusive vents over a period of 46,000 ±22,000 Ka based on K-Ar dating (Hildreth, 2004). Bailey (2004) identified mafic enclaves (silica content 53.5%) in lava from the northern most vent of this chain but no others have been found in the younger rhyolites from the Long Valley magma chamber. The second group of eruptions, the southeastern rhyolite cluster, produced a volume of ~1.5km3 over an interval as long as 75,000 years (Hildreth, 2004). The five, randomly distributed vents are located where the Hilton Creek fault intersects the ring fracture zone (Bailey, 1976). Three lava units identified with this group of Moat Rhyolites have significantly low crystal content (1–3% feldspars, biotite, cpx, Fe–Ti oxides), which represents a period of reversion ~330Ka to phenocryst-poor magma in the Long Valley reservoir (Hildreth, 2004). This reversion is unique in the evolution of the Long Valley magma chamber because all other Long Valley Rhyolites (527-100Ka) exhibit high crystal content (Hildreth, 2004). The final cluster of Moat Rhyolites, the West Moat Rhyolites, include four small lava domes and the west moat coulee (Hildreth, 2004). The west moat coulee, which measures 574 m thickness, constitutes ~4km3 of lava but the other four domes combined equal merely ~1km3. Bailey (1976) suggests that the three discrete locations of these eruptions are controlled by the intersection of major northwest-trending precaldera faults and the ring fracture zone the bound the resurgent dome.

In summary, during the period between 3.6 Ma to 100ka, the Long Valley magma chamber produced a mafic-to-silicic sequence of volcanism that centered around the Long Valley caldera (Bailey, 1989b). (Figure 8) After the emplacement of Glass Mountain, the Bishop tuff and the early rhyolites, the Long Valley magma chamber appears to have reached an isostatic equilibrium in the upper crust, most likely due to this loss of mass (Bailey, 1976). The magma chamber has been experiencing gradual cooling and downward congealing over the past 0.6 million years, periodically accumulating enough volatile pressure by crystallization to extrude the moat rhyolites (Bailey, 1976). Due to the progressive decrease in the temporal and volumetric rates of the last eruptions, the once dynamic Long Valley magma chamber is thought to be moribund (Hildreth, 2004). Further evidence suggests the near complete crystallization of the Long Valley magma chamber, such as the crystal rich, low temperature lava composition for all the moat rhyolites younger than 300 ka. Finally, the attribution of the recent (1979-2003) 80cm uplift of the resurgent dome to an intrusion of a mafic dike further suggests that the rhyolitic reservoir below the caldera is penetrable and, thus, moribund (Hildreth, 2004). Hildreth (2004) suggests that the most recent volcanism of the greater Mammoth Mountain and Mono-Inyo chains does not represent renewal of the Long Valley Magma chamber. Instead, Hildreth (2004) proposes that both chains are new and mutually independent fields of crustal-melt.

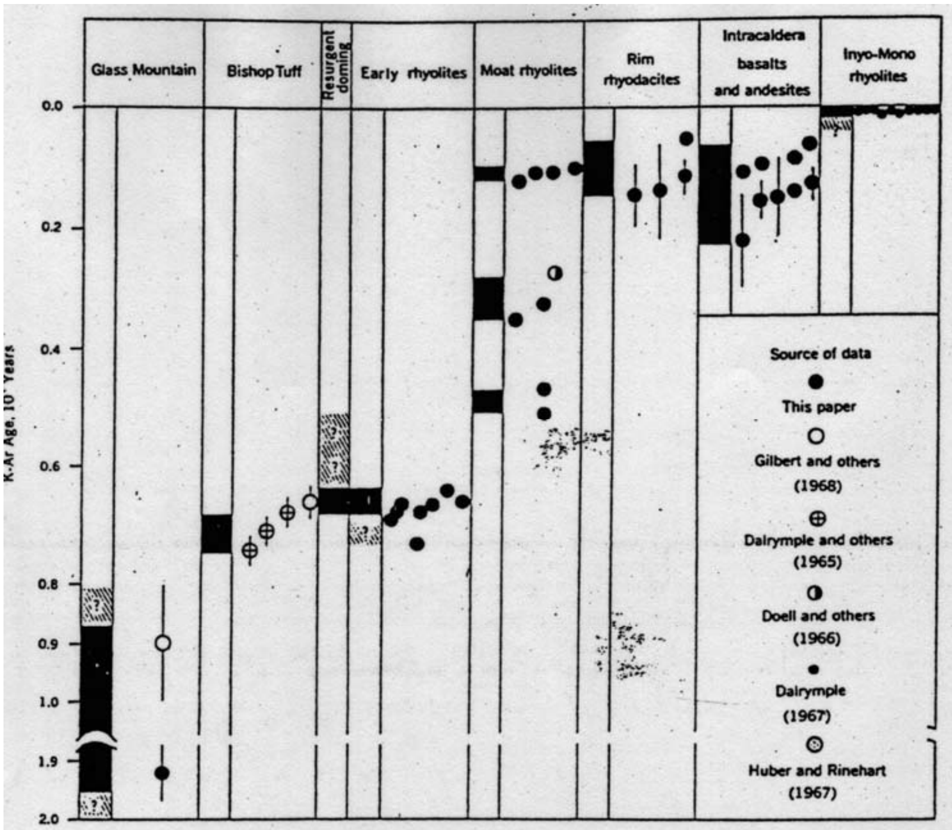

Figure 8. K-Ar ages and inferred geochronology of volcanic events associated with Long Valley caldera. The dark bars indicate the approximate time and duration of frequent activity; lined bars with queries indicate uncertainty.
Taken from Bailey, (1976)

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