An Investigation of Volcanic Depressions

Part I

Geologic and Geophysical Features of Calderas

by

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C A L D E R A S

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Terminology

Calderas are "large volcanic depressions, more or less circular or cirque-like in form, the diameters of which are many times greater than those of the included vent or vents, no matter what the steepness of the walls or form of the floor" (Williams, 1941). When this statement was written, it was generally assumed that all intra-caldera vents were approximately cylindrical conduits; since then, however, it has been found that many vents within calderas and along their margins are arcuate fissures, presumably surface expressions of cone-sheets and ring-dikes. Nevertheless, the 1941 definition of calderas is still retained.

Deep erosion of calderas commonly reveals 'ring complexes' composed of arcuate and circular intrusive bodies. The first such complex to be described was that of Glencoe, Scotland. This crudely circular structure was referred to as a <u>cauldron subsidence</u> formed by foundering of a cylindrical block of the crust into an underlying body of magma (Clough, Maufe, and Bailey, 1909). While the block sank, magma rose along the encircling fracture, "as liquor in a full bottle rises round a settling stopper", to feed a ring of volcanoes around a surface caldera. As first used, therefore, the term cauldron subsidence referred to a mechanism by which a caldera might originate. Since then, the term has usually been applied to all downdropped blocks enclosed by ring dikes. Examples include the Tertiary cauldron subsidences of Scotland and Ireland, the Jurassic ones of Nigeria, the Triassic ones of New England, and the Permian ones of the Oslo region, Norway. Unfortunately, some recent writers have extended the term 'cauldron subsidence' to all structures "produced by block subsidence, irrespective of size or shape". Experience shows, however, that virtually all large volcanic basins which are not oval or circular in plan, but of irregular shape or bordered by straight walls, owe their origin to a combination of volcanic and tectonic controls; such basins are better spoken of as volcano-tectonic depressions because their forms are largely controlled by the structure of the sub-volcanic basement. The Taupo depression in the North Island of New Zealand is an excellent example.

Classification and origin of calderas

Recent studies of calderas and related forms necessitate a drastic revision of the writer's 1941 classification. It still seems proper to speak of large cirque-like depressions formed by collapse of the flanks of volcanoes as a consequence of steam-blast eruptions as explosion calderas (e.g. Bandai-san, Japan, 1888, and the Chaos Crags, near Lassen Peak, California), but such calderas are extremely rare. It also seems proper to continue usage of the term erosion calderas for unusually large depressions formed within the summits of volcanoes by erosional widening of original craters. Except for these two types, however, all calderas are products of engulfment, and it is the classification of these that concerns us here.

Collapses that produce calderas result either from withdrawal of magmatic support or from ring-fracture stoping. Support may be withdrawn from the roof of a magma reservoir in several ways among which the chief seem to be: 1) rapid and voluminous eruptions of ash, pumice, and scoria, mostly as glowing avalanches; 2) eruption of fluid lavas from rifts on the flanks of shield volcanoes; 3) subterranean injection of magma into rift zones. In each case, it is the drainage of the reservoir that brings about engulfment. But collapse may also result from passive sinking of more or less cylindrical blocks into underlying reservoirs without attendant drainage. Such sinking is usually preceded by formation of annular fractures through tumescence or de-tumescence of the volcanic edifice, i.e. by increase or diminution of magmatic pressure. Subsidence may take place whether the roof-rocks have a lower or a higher density than the underlying magma; if their density is lower they can only subside to an appropriate level in the reservoir; if their density is higher they may subside far enough to be totally submerged.

Five types of collapse-calderas were distinguished in the writer's 1941 classification. These were: 1) Krakatoa type; 2) Kilauea type; 3) Katmai type; 4) Cryptovolcanic type; and 5) Glencoe type. The cryptovolcanic type is no longer considered valid. Prior to 1941, it was generally thought that some calderas were produced by muffled, subterranean explosions, by abortive attempts to form explosion pipes (diatremes). These "cryptovolcanic calderas" are usually characterized by a central area of uplifted and intensely fractured rocks surrounded by a moat of subsided rocks more or less deformed by arcuate folds and faults. The central uplifts were attributed to deep-seated explosions and the surrounding depressions to contraction of underlying magma as it solidified and lost its gas. High-pressure forms of silica (coesite and stishovite) and 'shatter cones' have since been discovered in many of these "cryptovolcanoes", and additional evidence has been obtained to show that the pumice-like suevite associated with the Riesskessel "cryptovolcano" of southern Germany is not volcanic in origin but formed by partial impact-fusion of bedrock granites. There is now general, though not yet complete agreement that the supposed cryptovolcanic depressions were produced by the supersonic impact of meteorites and larger celestial bodies, their central uplifts resulting from elastic rebound and the arcuate folds and faults from shock waves spreading from the explosion foci.

Studies made since 1941 have also made it necessary to redefine the Katmai and Glencoe types of caldera; in addition it seems desirable to intodouce some new types.

Calderas owe their origin to the following causes:

1. <u>Withdrawal of magmatic support</u>. This results chiefly from subterranean migration of magma from its reservoir into rift zones and from partial drainage by copious discharge of magma through surficial vents, generally in the form of ash and pumice. Magmatic support may also be withdrawn, but only to a relatively minor extent, by eruptions of fluid lava from fissures on the flanks of basaltic volcanoes. The explosive eruptions that lead to formation of most calderas may issue from the summit-vents of pre-existing volcanoes or from arcuate fissures on the flanks, but they may also issue from arcuate fissures that bear no genetic or definite geometric relationship to pre-existing volcanoes.

2. <u>Gravitative settling along ring fractures</u>. This may or may not be accompanied by surface eruptions. Settling may take place even though the ring fractures converge downward; if they do, the intra-caldera deposits may settle differentially to produce basin shapes or settling may take place without deformation if the summit of the volcanic edifice tumesces sufficiently to cause appreciable widening of the encircling fractures. If the density of the subsiding block is less than that of the underlying magma, the block can only subside to a certain level determined by the relative densities; if, on the other hand, the density of the subsiding block exceeds that of the magma in the upper part of the reservoir it may sink until arrested by denser layers or by a layer of blocks previously stoped from the reservoir roof, as suggested by Chapman (1966).

A Preferred Classification

Most calderas originate from a combination of causes; hence no rigid classification is possible. It seems advisable, nevertheless, to separate three principal groups on the basis of the relative amounts of lava and pyroclastic ejecta involved in their formation, and to subdivide each of these groups into several types.

A. Calderas associated with voluminous explosive eruptions of siliceous magmas.

1. Krakatoa type. Collapse results from copious eruptions of magma as pumice falls and pumice flows. In part at least, the eruptions issue from the summit-vents of pre-existing composite volcanoes; eruptions may also issue from newly opened fissures on the flanks of the original cones.

2. Katmai type. Collapse results from drainage of the central conduit of a volcano and perhaps also of some of the underlying reservoir by discharge of magma through adjacent conduits. The top of Mount Katmai collapsed in 1912 when its central conduit was drained by eruption of ash- and pumice-flows from vents in the adjacent Valley of Ten Thousand Smokes.

3. Valles type. Collapse follows discharge of colossal volumes of ash and pumice as pyroclastic flows from arcuate fissures unrelated to pre-existing volcanoes. The volumes of the pyroclastic flows generally range from 100 to 1,000 cubic kilometers. Many calderas of this type are characterized by uplift of the floors to produce "resurgent domes".

B. <u>Calderas associated with effusive eruptions of basaltic magma</u>.

1. <u>Masaya type</u>. Form by repeated collapses of vents within the summits of flattish basaltic shield volcanoes as magma migrates at depth. Eruptions from arcuate and radial rifts play no part in caldera formation.

2. <u>Hawaiian type</u>. Form by collapse during late stages of growth of large basaltic shield volcanoes. Engulfment results from subsidence of a summit-block enclosed by steeply inclined ring fractures. Prior tumescence of the shield is accompanied and followed by drainage of magma into rift zones, with or without flank eruptions of lava.

3. <u>Galápagos type</u>. Also form by collapse during late stages of growth of large basaltic shield volcanoes. But engulfment results chiefly from injection of sills, and eruptions of lava through circumferential fractures near the summits and from radial fractures far down the sides of the shields.

C. <u>Calderas associated with mixed eruptions from ring fractures</u>.

1. <u>Glencoe type</u>. Long-continued, intermittent eruptions of lava and pyroclastic ejecta from arcuate fractures surrounding a subsiding cauldron.

2. <u>Suswa type</u>. Form by collapse of the summits of large volcanic cones or cone-clusters along ring-fractures as a result of magma withdrawal or gravitational settling. Eruptions do not precede and initiate collapse, as in the case of calderas of the Krakatoa type, but they result from and follow collapse, and they issue

from the ring-fractures, producing both lavas and fragmental ejecta.

D. <u>Major volcano-tectonic depressions</u>.

These are exceptionally large collapse depressions related to colossal eruptions of pyroclastic flows. Their shapes are controlled primarily by structures in the basement-rocks; hence they tend to be more or less straight-walled linear grabens. The Toba depression of Sumatra was produced by discharge of ash- and pumice-flows with a total volume of approximately 2,000 cubic kilometers; the Taupo Volcanic depression in the North Island of New Zealand was formed by repeated collapses following a succession of voluminous pyroclastic flows.

We may now describe in more detail the foregoing types of calderas and volcano-tectonic depressions and the manner in which they originated.

Calderas of Krakatoa Type (A1)

The two most common types of calderas are the Krakatoa and Valles types, both of which are related to large-scale eruptions of ash and pumice. It must be emphasized, however, that it is difficult and may be impossible to assign a given caldera to one or the other of these two types. The Krakatoan type is formed by foundering of the tops of pre-existing volcanoes, and the explosive eruptions that bring about collapse issue, at least in part, from the summit-vents of these volcanoes, though they may also issue from newly opened fractures; the Valles type, on the other hand, is produced by eruptions through newly opened arcuate fractures independent of pre-existing volcanoes. To make a distinction it is necessary, therefore, to know whether or not the site of the caldera was formerly occupied by a volcanic cone or cone cluster.

Calderas of Krakatoa type, as Smith has suggested (1960), are usually associated with pyroclastic-flow deposits having volumes between 10 and 100 km^3., whereas most calderas of Valles type are associated with still more voluminous pyroclastic flows. The floors of calderas of Krakatoa type are never upheaved to produce "resurgent domes" such as may be seen in many calderas of Valles type; however, domes of Peléan type and cinder cones are present on the floors of almost all Krakatoan calderas.

The present site of the caldera of Krakatoa was once partly occupied by three coalescing cones; the caldera of Crater Lake, Oregon, lies on the site of the former cone of Mount Mazama; that of Towada, Japan, lies on the site of a large andesitic cone; and that of Santorin lies on the site of a former cone-cluster.

The activity that initiates collapse invariably begins by high-pressure Vulcanian eruptions, the ejecta falling in showers to produce well-bedded deposits. And in general, as activity continues, the fragments in the pumice-falls tend to increase in size as the gas-pressure diminishes. Subsequently, magma may be discharged as glowing avalanches (pumice flows) that tend to become increasingly basic and loaded with crystals as deeper levels of the magma chamber are tapped. The volume of the preliminary pumice falls is generally much less than that of the later pumice flows, but there are many exceptions to this rule; indeed, as noted in the sequel, repeated collapses may follow repeated pumice falls that are not accompanied by pumice flows.

A few examples of Krakatoan calderas may now be described.

1. Krakatoa. The great eruptions of 1883 were preceded by a long period of quiet. The initial eruptions, which began on May 20 from the cone of Perboewatan, lasted only a few days, the activity dying down until June 19, when eruptions were resumed from the same vent. Five days later, a new vent opened at the foot of the adjacent cone of Danan (figure 1). When the islands were visited on August 11, for the last time before the final paroxysm, three main vents were said to be in mild eruption. Meager evidence suggests that all of the Vulcanian eruptions of pumice that preceded the climax issued from the cones of Perboewatan and Danan; the cone of Rakata does not seem to have contributed. During the fateful days of August 26-28, the pumice falls were followed by tremendous pumice flows, some of the coarse, unstratified and unsorted deposits of which can still be seen lying on the bedded pumice-fall deposits on the adjacent islands. Most of the pumice flows swept across the floor of the Sunda Straits; hence their volume cannot be calculated. There can be no doubt, however, that their total volume was vastly greater than that of the earlier pumice falls, even though they issued within less than two days. Whether the pyroclastic flows issued from the old vents of Danan and Perboewatan or from new fractures, opened during the eruptions, cannot be told. Several tidal waves accompanied the climactic episode, the largest, more than 35 meters high, inundating the adjacent coasts of Java and Sumatra, drowning more than 36,000 persons.

When explosive activity came to an end, the cones of Danan and Perboewatan had vanished; so had half of Rakata. The sea floor had subsided over an approximately circular area 7 kilometers across, from which two newly formed radial grabens extended. In 1927, after 44 years of quiet, a cone of basaltic cinders, Anak Krakatoa, began to rise from the floor of the caldera, from a point roughly midway between the former vents of Danan and Perboewatan; it has been active intermittently ever since.

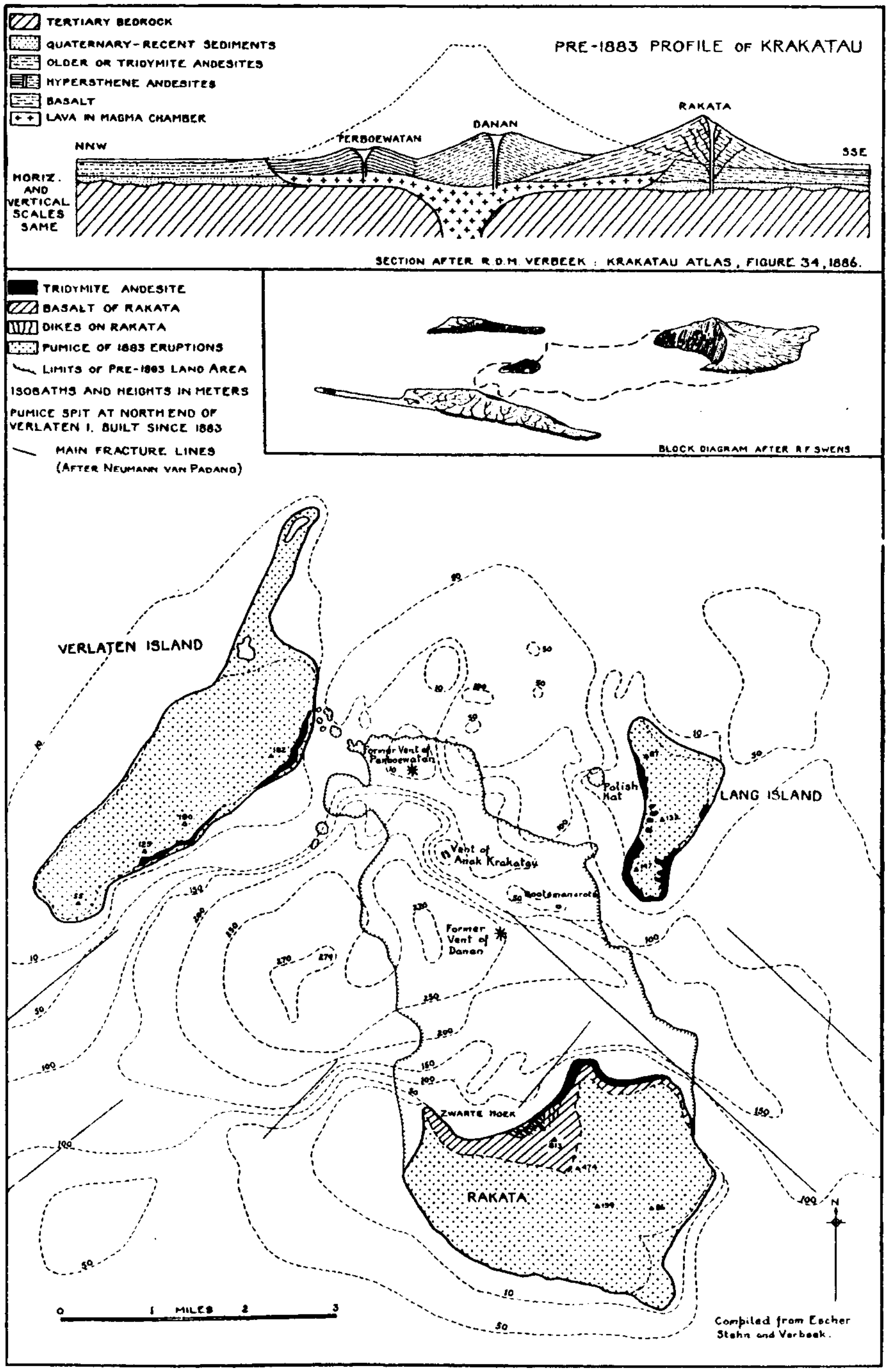
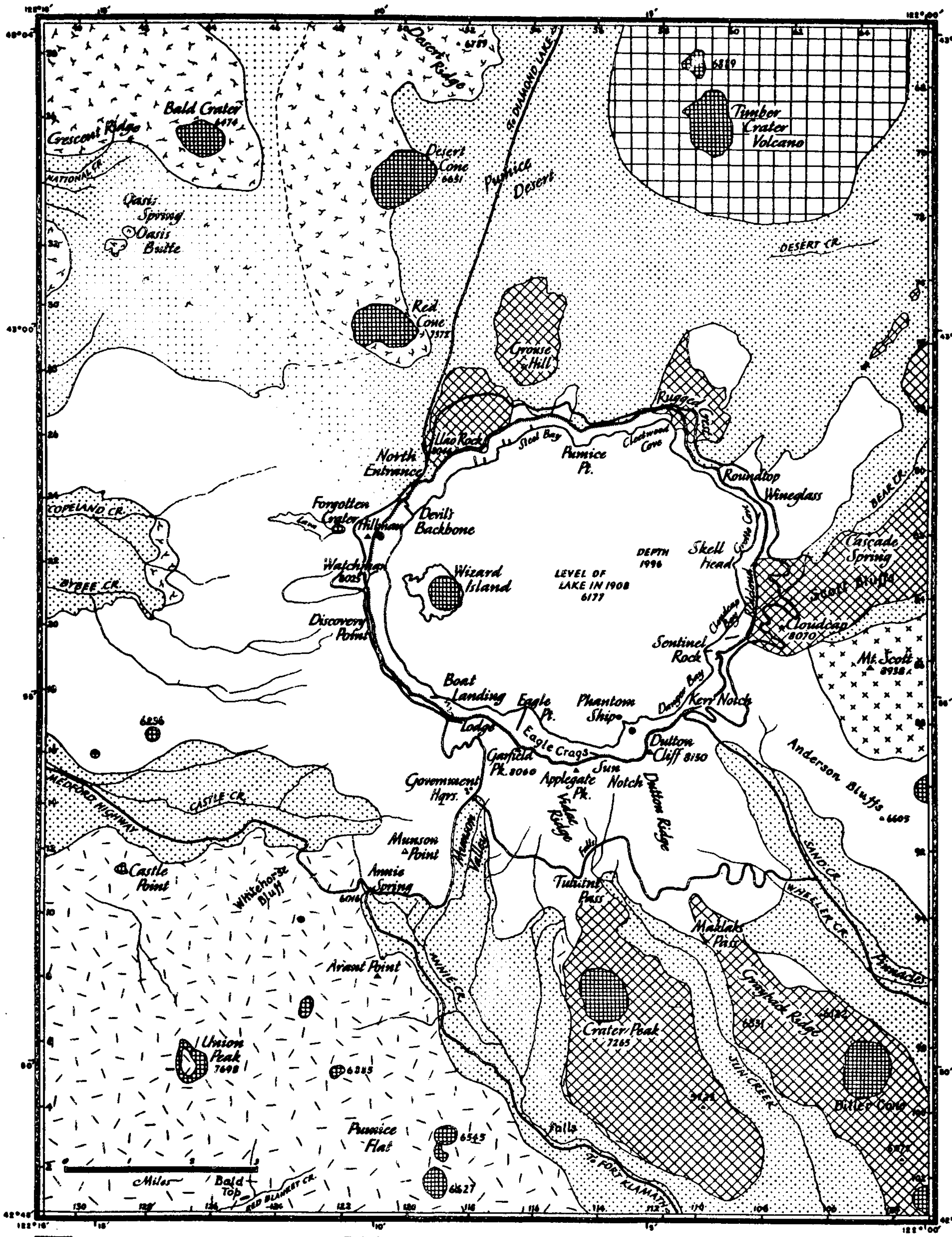


Fig. 2. The Krakatau caldera.

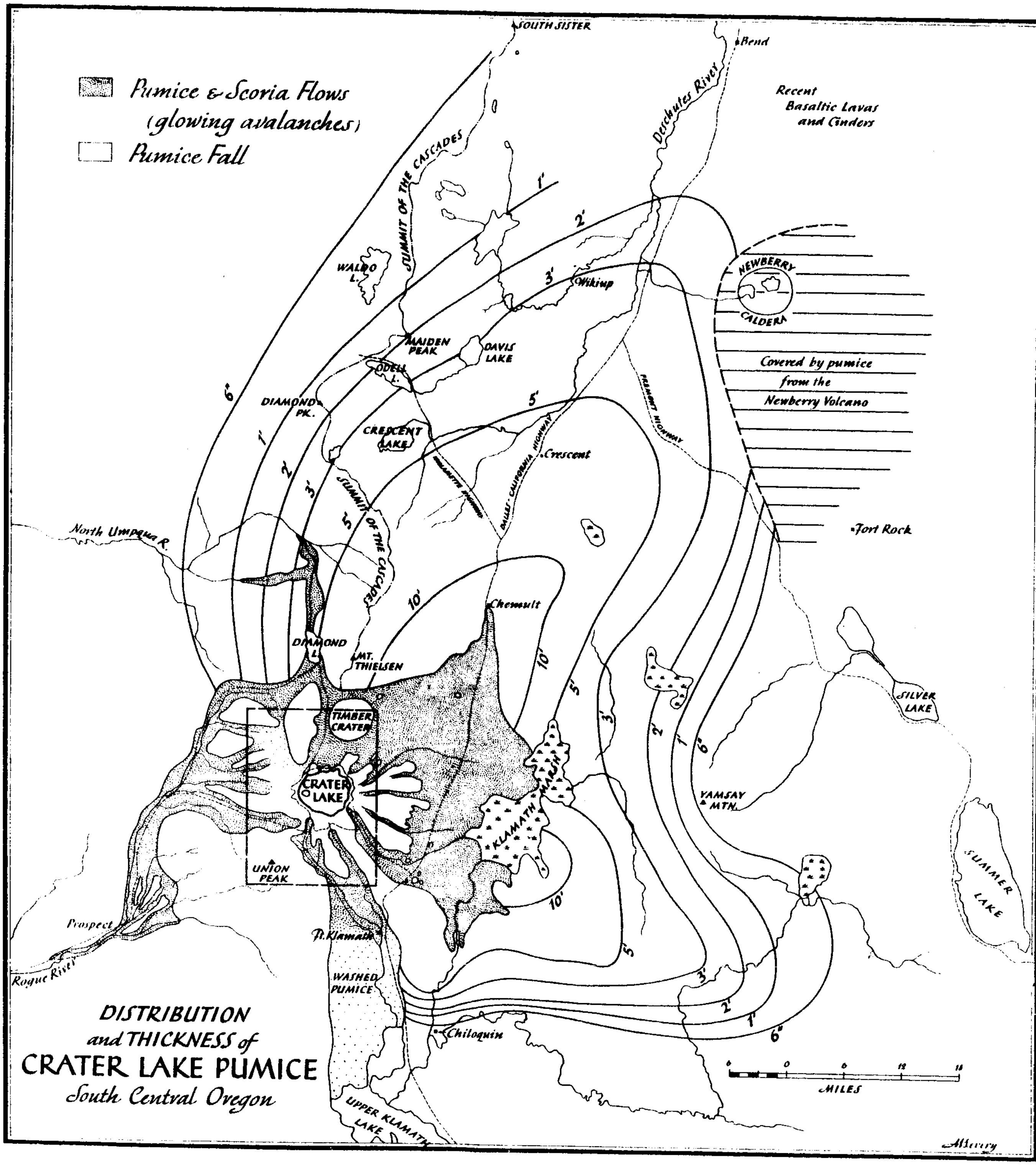
2. Crater Lake, Oregon. This magnificent caldera originated 6,600 years ago by collapse of the top of an ancestral cone, Mount Mazama, and some of its parasites (figure 2). The main volcano had grown to an elevation of approximately 3,500 meters, chiefly by eruption of andesitic lavas, but also, during later stages, by eruption of dacite pumice, when a semicircular fracture opened on its northern flank, between 10,000 and 15,000 years ago. Viscous flows of siliceous andesite and glassy dacite issued from this fracture to produce an arc of parasitic domes and cones; at about the same time, dacite domes and basaltic cinder cones grew on the lower slopes of Mount Mazama. Finally, after a long interval of quiet, the climactic eruptions began. At first, as at Krakatoa, the eruptions were of the high-pressure Vulcanian type. Finely divided dacite pumice was blown high above the vents, to be carried by winds, first toward the east and then, as the pumice became coarser, toward the northeast. These airborne ejecta have been traced over an area of almost 1.5 million square kilometers, extending northward into British Columbia and Alberta and eastward into Montana. Soon after these initial, high-pressure eruptions, the character of the activity changed, as it did at Krakatoa; the pumice fall was followed by glowing avalanches, first of dacite pumice and then of basaltic scoria, that raced down the mountainsides, choking the glacial canyons to great depth and spreading far beyond, on to the adjacent plateau. When the eruptions ended, the top of Mount Mazama had vanished and a caldera about 10 kilometers wide and more than 1,200 meters deep had taken its place.

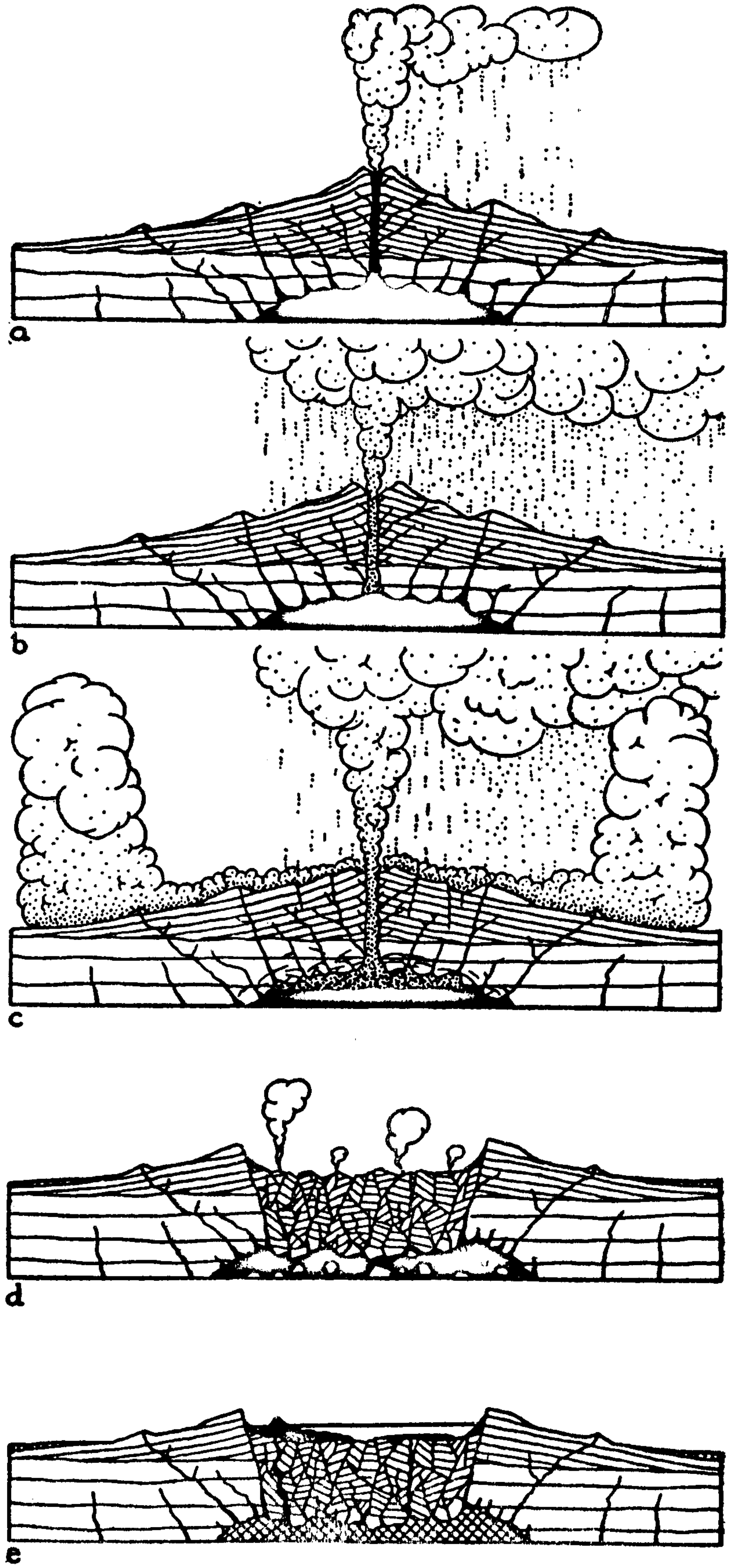
Williams (1942) estimated that the volume of the Mazama pumice fall was approximately 14 cubic kilometers, all but 2 km^3. of this amount lying within the 15 cm. isopach. But discovery of fine Mazama ash as far away as Canada and Montana necessitated drastic revision of this original estimate.



GEOLOGIC MAP OF CRATER LAKE NATIONAL PARK

Numbers along margins refer to boundary markers. Elevations in feet. For details of geologic formations on the caldera walls, see panoramic photographs, plates 23 to 29. For contours, see U. S. Geological Survey sheet of the park. (Drawn by A. W. Severy.)


DISTRIBUTION
and THICKNESS of
CRATER LAKE PUMICE
South Central Oregon
Pumice & Scoria Flows
(glowing avalanches)
Pumice Fall
SOUTH SISTER
Bend
Deschutes River
Recent
Basaltic Lavas
and Cinders
SUMMIT OF THE CASCADES
WALDO L.
NEWBERRY
CALDERA
Wikiup
MAIDEN PEAK
DAVIS LAKE
ODELL L.
Covered by pumice
from the
Newberry Volcano
DIAMOND PK.
CRESCENT LAKE
Crescent
FREMONT HIGHWAY
DALLES - CALIFORNIA HIGHWAY
Fort Rock
North Umpqua R.
Chemult
DIAMOND L.
MT. THIELSEN
SILVER LAKE
TIMBER CRATER
CRATER LAKE
KLAMATH MARSH
YAMSAY MTN.
UNION PEAK
SUMMER LAKE
Prospect
Rogue River
Ft. Klamath
WASHED PUMICE
Chiloquin
MILES
UPPER KLAMATH LAKE
1'
2'
3'
5'
6"
10'


a
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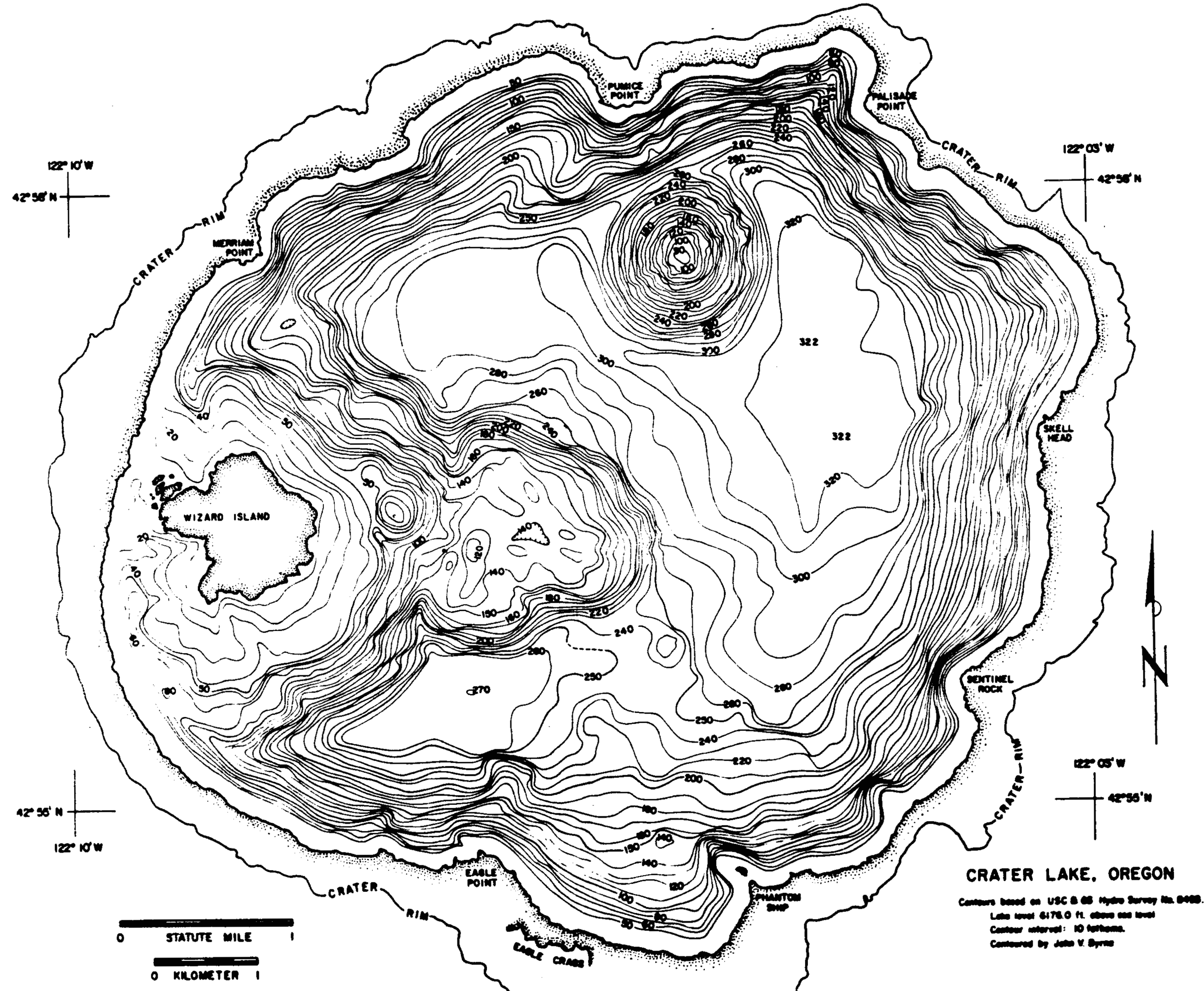
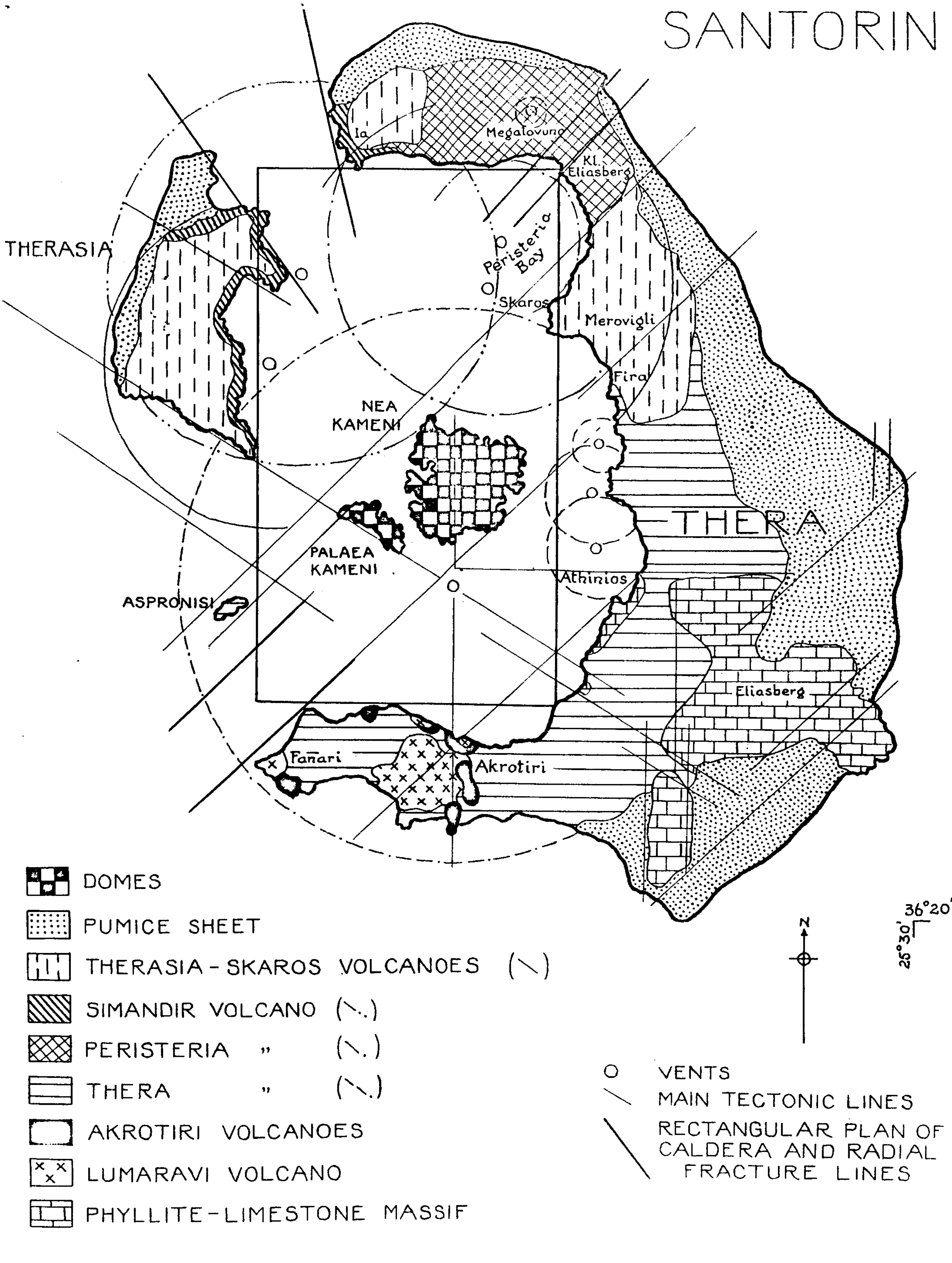
e



CRATER LAKE, OREGON
Contours based on USC & GS Hydro Survey No. 8469.
Lake level 6176.0 ft. above sea level
Contour interval: 10 fathoms.
Contoured by John V. Byrne
WIZARD ISLAND
MERRIAM POINT
PUMICE POINT
PALISADE POINT
SKELL HEAD
SENTINEL ROCK
PHANTOM SHIP
EAGLE POINT
EAGLE CRAGS
CRATER RIM
122° 10' W
42° 58' N
122° 03' W
42° 55' N
STATUTE MILE
KILOMETER
322


SANTORIN
THERASIA
Ia
Megalovuno
Kl. Eliasberg
Peristeria Bay
Skaros
Merovigli
Fira
NEA KAMENI
PALAEA KAMENI
THERA
Athinios
ASPRONISI
Eliasberg
Fanari
Akrotiri
N
36°20'
25°30'
DOMES
PUMICE SHEET
THERASIA - SKAROS VOLCANOES
SIMANDIR VOLCANO
PERISTERIA "
THERA "
AKROTIRI VOLCANOES
LUMARAVI VOLCANO
PHYLLITE-LIMESTONE MASSIF
VENTS
MAIN TECTONIC LINES
RECTANGULAR PLAN OF CALDERA AND RADIAL FRACTURE LINES

Calculations recently made by Dr. Gordon Goles (in press) show that between 16 and 24 km^3. of ash fell beyond the 15 cm. isopach -- i.e. much more than fell within that limit. Accordingly, the total volume of ejecta that fell from the air is now estimated to be between 30 and 38 km^3. To this must be added the volume of the glowing avalanche deposits (24 to 32 km^3.) and that of the final ash-fall (1 km^3.). The aggregate volume is thus between 55 and 71 km^3.

Crystals make up 8 to 14 km^3. of this amount; lithic fragments make up 5 to 8 km^3., and the remaining 33 to 58 km^3. consist of pumiceous fragments of glass. It is impossible, of course, to estimate accurately the vesicularity of the glass-fragments and the pore-spaces between them; it seems likely, however, that they are equivalent to between 15 and 25 km^3. of liquid magma. The actual magma may have contained up to 3 per cent by weight of water in solution, but this would have had only a very slight effect on its volume.

Our conclusion from the wide range of figures just enumerated is that during the culminating eruptions of Mount Mazama the total volume of ejecta -- i.e. of liquid magma, crystals, and lithic fragments -- was between 28 and 47 km^3. For convenience, let us say that the total volume was approximately 40 km^3. The volume of the mountaintop that collapsed to form the caldera was not 68 km^3., as originally estimated, but approximately 60 km^3., for the ancestral peak was only about 3,000 meters high and it must have contained a huge summit-depression within which enough ice accumulated to feed three glaciers that spread southward, down the sunny slopes, to extend beyond the present rim of Crater Lake through Sun Notch, Kerr Notch, and Munson Valley.

Earlier calculations had suggested that the combined volume of lithic fragments, liquid magma, and crystals erupted by Mount Mazama was about 40 km^3. less than the volume of the mountain that collapsed; present calculations reduce the discrepancy to about 20 km^3. If the magma had already begun to vesiculate just before the climactic eruptions, this discrepancy would be further reduced, but probably by only a small amount. Hence, one still seems to be forced to conclude that some of the space necessary to permit the collapse of Mount Mazama was provided by subterranean withdrawal of magma, either through fissures in the walls of the reservoir or by migration at greater depths. It is conceivable, indeed, that such withdrawal may have triggered the explosive eruptions that led to engulfment.

After a brief interval of quiet following the formation of the caldera, a pyroxene andesite cinder cone -- the Merriam Cone -- rose from the floor, close to the northern wall. Subsequently, as the lake-level continued to rise, a similar cone grew near the southwestern wall forming Wizard Island. Long flows of andesite spread from its base, as shown in Figure 5; these either poured round or are capped by a small dome of glassy dacite. The final eruptions appear to have taken place more than a thousand years ago.

3. Shikotsu caldera, Hokkaido. The Japanese island of Hokkaido contains at least ten calderas of Krakatoa-type. Shikotsu is typical (Katsui, 1963). This caldera, which measures 13 by 15 km. across and has a volume of 80 km^3., originated about 20,000 years ago, on the former site of a large composite cone. The ancestral cone, like Mount Mazama, was built chiefly of andesitic lavas and fragmental deposits until, during late stages of growth, these were augmented by falls of dacite pumice. Then followed a long period of quiescence. Physiographic evidence suggests that just before the climactic eruptions, the ancestral cone may have tumesced and been cut by faults. In any event, the outbursts heralding the climax were high-pressure blasts that produced rhyolitic pumice falls having a total volume of approximately 25 km^3., including 4 per cent of lithic fragments. These falls were followed by dacite

pumice flows with a volume of about 100 km^3., including 13 per cent of lithic fragments. All of the lithic material within the pumice fall was derived from the ancestral cone and its Tertiary basement, but the pumice flows also include fragments of granite, quartz diorite, and hornfels derived from greater depths as the level of the effervescing magma sank in the reservoir. When the volume of all the Shikotsu pumice is calculated as liquid magma containing 5 per cent by weight of water and this is added to the volume of lithic debris, the total approximates closely to the volume of the ancestral cone that collapsed to form the caldera. Accordingly, withdrawal of magmatic support by explosive eruptions may well have been the sole cause of engulfment. Subsequently, three andesitic volcanoes grew along a fissure trending N. 30° W. across the caldera, one of which, Tarumai, has been active within historic times.

4. Towada caldera, Honsyu. This caldera, which measures approximately 10 km. across, occupies the former site of a large composite cone of pyroxene andesite, and was formed by collapse following discharge of voluminous flows of dacite pumice, some of which travelled 30 km. from their source.

After the collapse, a new cone was built of andesite essentially similar to that which formed the ancestral cone. The top of the second cone then collapsed following discharge of andesitic pumice to produce a miniature caldera, 3 km. across. Finally, two domes of pyroxene dacite rose in the middle of the main depression.

5. Santorin caldera, Greece. This famous caldera is selected for description as an example of a Krakatoan caldera developed on the former site of a cluster of coalescing volcanic cones (figure 6). The oldest cones are partly preserved at the southern end of the island of Santorin; their activity

was more explosive than effusive, much of it was submarine, and the products varied in composition from basalt to rhyolite. The center of activity then moved northward to what is now the middle of the island of Santorin where the large Thera volcano was built, mainly by eruptions of pumice from a summit-vent and of lava from parasitic vents on the flanks. Before Thera became extinct, other volcanoes began to grow in what is now the northern part of the island and near the island of Therasia. These volcanoes -- the Peristeria-Simandir group -- were less explosive than the earlier ones, erupting many flows of andesite and dacite. Then the Skaros volcano was built, mainly of dacite flows, within a graben separating the Thera and Peristeria volcanoes, while another large lava cone grew to the west, partly on the present site of Therasia. At the close of this long period of volcanism, the area now occupied by the caldera of Santorin was a large, rugged island made up of many overlapping cones, some of which were incised by deep valleys of tectonic origin.

A long interval of quiet ensued prior to the explosions that led to formation of the caldera. The quiet was broken by Vulcanian eruptions that left a layer of airborne, rose-colored pumice, ranging in thickness up to about 3 meters. Variations in the coarseness and thickness of these ejecta indicate that the eruptive vents lay over the northern part of the present caldera, and the scarcity of lithic debris suggests that the vents were already open when activity began. The explosions cannot have been particularly violent, for most of the inhabitants of the island had time to escape, though forced to leave many of their goods and chattels behind. Their buildings remained intact, suggesting that no severe quakes preceded the eruptions.

The "Rose Pumice" is overlain by the "Middle Pumice", a thicker sequence of interbedded, well-stratified airborne and fluviatile materials testifying to a succession of mild explosions and quiet pauses. The culminating explosions then laid down the "Upper Pumice". In thickness, and in the size and abundance of its lithic debris, this far exceeds the earlier deposits. Even so, the final explosions do not seem to have been cataclysmic, like those of Krakatoa; few if any of the deposits were laid down by pumice flows. Lenticular bedding, erosional breaks, and fluviatile sorting seem to imply repeated explosions of moderate intensity from many sources.

An accurate estimate of the total volume of material erupted is impossible, because most of the ejecta fell into the sea, but clearly the volume was sufficient to drain the magma chamber far enough to remove support from the roof, causing it to founder. The collapse took place about 1,470 B.C., and it generated tsunamis (seismic sea waves) between 100 and 165 feet high, that swept within 20 minutes on to the coasts of Crete, 75 miles away, causing such havoc as to bring the Minoan civilization to a sudden end.

Three principal controls determined the shape of the caldera. The scalloped margin is obviously related to concentric fracturing and collapse of separate cones, for the centers of several of the arcuate bays on the island of Santorin coincide closely with the positions of former vents. Each cone in the original Santorin complex was built over a cupola on the upper surface of the regional magma chamber; hence, general lowering of the magma level by the eruptions first drained the cupolas, causing their roofs to be engulfed. The shape of the caldera was also controlled in part by the formation of more or less radial grabens that extend outward between the islands of Santorin and Therasia, perhaps following extensions from the main magma chamber. The third control was volcano-tectonic, i.e. structural trends in the pre-volcanic basement. One such trend is north-south, and another is east-west; between them they account for the crudely rectangular shape of the Santorin caldera.

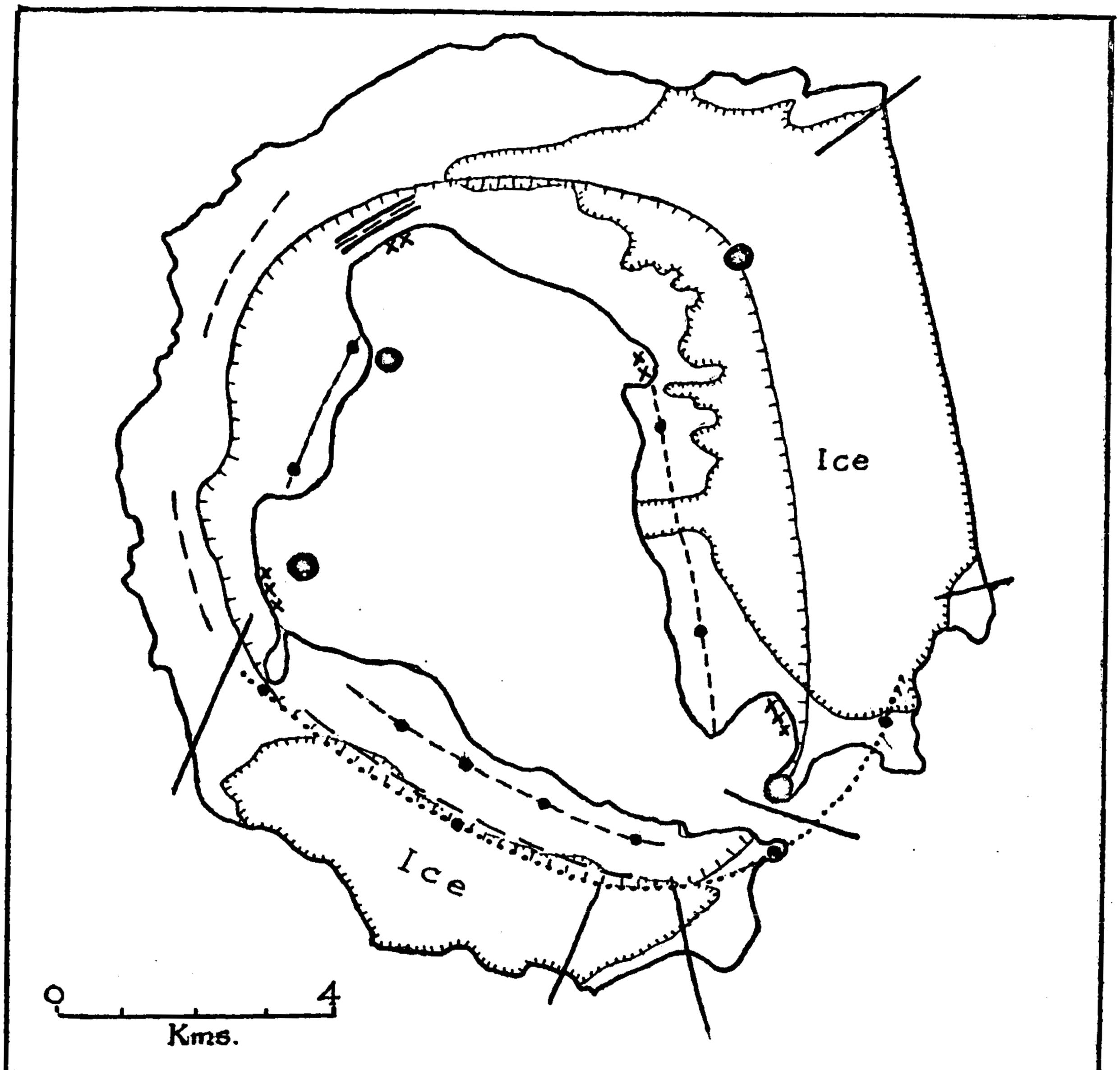
More than a thousand years elapsed after the engulfment before volcanism was resumed. But commencing in 197 B.C. and continuing in 46 B.C., domes of Peléan type, composed of siliceous andesite and dacite, rose from the central horst on the caldera floor to form the island of Palaea Kameni. A second island, Mikra Kameni, rose in 1570-1573 A.D., and a third, Nea Kameni, was born in 1707. This was enlarged by further domical protrusions in 1886 and again in 1925-1928.

6. Deception Island caldera, South Shetland Islands. This caldera is selected for description because it is an unusually good example of a Krakatoan caldera in which collapse was followed by the formation of arcuate lines of small volcanoes along the rim and on the walls (Hawkes, 1961).

The island would be a complete ring were it not for a narrow passage, aptly called Neptunes Bellows, which leads into the sheltered inner bay of Port Foster (figure 7). A roughly circular ridge-crest, that rises to a maximum height of more than 550 meters, marks the caldera rim. The outer slopes, which are largely mantled with ice, descend gently to sea-cut cliffs some 50 meters high; the inner slopes, which mark the caldera wall, are steeper and locally precipitous, and in places they are considerably modified by cones and craters, products of post-caldera eruptions.

Four overlapping volcanoes once occupied the site of the caldera; the vents of two of them lie on the caldera rim, and the vent of another lies beneath the center of a semicircular bay on the western wall. All four volcanoes were built chiefly of andesitic and basaltic lavas.

Paroxysmal eruptions of andesitic pumice then took place, but whether from the four pre-existing vents or from new fractures along and close to the present caldera rim is not known. Deposits of pumice are almost continuously exposed for long distances along the coastal cliffs, and locally they reach a thickness of more than 215 meters. Their original volume cannot be judged, because most of the ejecta fell or poured into the ocean. Some faulting had already occurred prior to the explosions, but the prime cause of the collapse



DECEPTION ISLAND CALDERA

Caldera rim. Limits of main ice fields.
Fault ○ Vents of pre-caldera volcanoes.
First post-caldera eruptive fissures.
Second. Third. ×× Fumaroles.
• Post-caldera vents. After D.D.Hawkes.

which produced the caldera was almost surely the sudden, rapid, and voluminous discharge of pumice. Collapse was controlled almost wholly by concentric fractures, but in minor part it was controlled by radial fractures, one of which accounts for Neptunes Bellows. After engulfment, little remained of the original volcanoes except for three small islands that rose from the caldera rim; the remainder foundered into the ocean.

Three volcanic episodes followed the formation of the caldera, and during each the eruptive vents were localized along concentric, marginal fractures. Four of the first post-caldera vents lie along and close to the southern rim, two of them near Neptunes Bellows. Simultaneous eruptions may have taken place along and close to the eastern rim of the caldera, but this is an area thickly covered by ice. Activity at the four recognizable vents was mainly explosive, but at least once during the history of each vent flows of olivine basalt breached the crater walls.

During the second post-caldera episode, eight small cones grew along arcuate fractures close to the bottom of the caldera walls, near the shore of Port Foster Bay (figure 7). Deep craters still mark the eruptive vents. Activity was again chiefly explosive, but most of the ejecta were andesitic rather than basaltic.

Finally, after a long interval of quiet, during which morainic and fluio-glacial deposits were laid down, the last eruptive episode began, probably continuing into historic times. Eruptions were localized along concentric fractures within and outside the western rim of the caldera, and were characterized by emission of basaltic flows, partly from open fissures and partly from the flanks of scoria cones. Steam rises from several of the beaches bordering the central bay, suggesting that fumarolic activity still continues along and close to the base of the caldera walls.

7. <u>Coatepeque caldera</u>, <u>El Salvador</u>. Almost all calderas of Krakatoa type are surrounded by copious deposits laid down by pumice flows, though the pumice deposits surrounding the Santorin caldera, as noted already, are almost entirely, if not entirely products of pumice falls. On the other hand, none of the deposits around the caldera of Coatepeque can be ascribed to pumice flows, all being airborne deposits, partly reworked by running water.

The original Coatepeque volcano was built chiefly of andesitic and basaltic lavas and fragmental materials. During Late Pleistocene time, a long series of rhyolitic, quartz latitic, and rhyodacitic pumice eruptions ensued. Careful petrographic study has recently made it possible to recognize 18 distinctive pumice falls (Meyer, 1964). One fall has a volume of 32 km^3.; another has a volume of about 13 km^3.; two others measure 8 km^3. each. Their total volume approximates 73 km^3.; the equivalent volume of magma is about 50 km^3., almost exactly the volume of the ancestral cone that collapsed to produce the caldera. Accordingly, it is no longer necessary to postulate that engulfment was caused principally by underground migration of magma to supply the young volcanoes of Santa Ana and Izalco (Williams and Meyer-Abich, 1955). What chiefly distinguishes the mode of formation of the Coatepeque caldera from that of Santorin is the length of time involved in the preceding pumice eruptions. The Santorin eruptions seem to have lasted no more than 50 years, but those from the Coatepeque volcano lasted very much longer, because many pumice-fall deposits are separated by deep zones of weathering. This probably means that the engulfment which produced the Coatepeque caldera was not cataclysmic, as at Santorin, but took place intermittently over thousands of years.

Katmai Type (A 2)

Most interpretations of what took place in 1912 in the Valley of Ten Thousand Smokes, Alaska, and at the adjacent volcano of Katmai have been based on studies made by Fenner. Painstaking work led him to postulate that the ash- and pumice-flows which swept down the Valley of Ten Thousand Smokes issued from an extensive sill of rhyolitic magma contaminated by contact with andesitic debris on the valley floor. He also supposed that almost 20 cubic kilometers of hybrid pumice were blown from the summit crater of Mount Katmai, more than half of this being andesitic material that tumbled from the crater walls to be assimilated in a lake of rhyolitic lava. Accordingly, the Katmai type of caldera was defined as one "produced by a combination of internal solution, pumice explosions, and avalanching of the crater walls". It is now apparent, however, that there was no internal assimilation of Katmai volcano prior to the disappearance of its summit and formation of the caldera; indeed no eruptions took place from the crater of Katmai in 1912, during the eruptions in the Valley of Ten Thousand Smokes.

Observations made by the writer and Dr. Garniss Curtis in 1953, and others made in the following year by Curtis led to a radically different interpretation of the events of 1912. We concluded that no sill of rhyolite was injected under the floor of the Valley of Ten Thousand Smokes; on the contrary, the ash- and pumice-flows that swept down the valley issued from swarms of vertical fissures near the valley head, particularly from fissures near the vent from which the rhyolitic dome of Novarupta was subsequently extruded. The hybrid character of the ash and pumice did not result from contamination of fresh rhyolite magma by assimilation of andesitic rocks, but from mingling of fresh andesitic magma that occupied the fissure-system beneath the Katmai, Mageik, Martin, and Trident volcanoes with fresh rhyolitic magma that occupied the nearby, sub-parallel fissure-system beneath the domes of Falling Mountain and Mount Cerberus. Moreover, Curtis made isopach-maps of the ash- and pumice-falls overlying the deposits of ash- and pumice-flows in the Valley of Ten Thousand Smokes, and found that Mount Katmai could not have been the source of any of the ejecta; it had erupted nothing. It was the rapid discharge of many cubic kilometers of rhyolitic and hybrid magma through vents 1,500 meters below Mount Katmai, within the Valley of Ten Thousand Smokes, that drained the conduit of the volcano, causing its top to founder, producing a caldera approximately 3 kilometers across and 1,300 meters deep. When this was first visited, in 1917, a small cone of andesitic ejecta could be seen within a lake on the floor; andesitic magma had returned to the fissure-system beneath the volcano. By 1953, the andesitic cone was submerged beneath a lake about 600 meters deep.

The foregoing account indicates that the caldera of Katmai was not formed by internal assimilation and subsequent eruptions, but by collapse when magma was withdrawn from the central conduit and perhaps from the underlying reservoir to supply eruptive vents in the adjacent Valley of Ten Thousand Smokes.

No other calderas of this type have yet been recognized, but it seems reasonable to suppose that some will be found. Many closely spaced volcanoes lie on fissures that connect at depth; hence the top of one volcano may collapse because of eruptions from a neighbor or from newly opened fissures not far away.

Reference should also be made under this caption to circular basins that are not genetically related to pre-existing volcanoes but originate through removal of magmatic support by discharge of material from adjacent vents. Such basins were referred to by van Bemmelen as "lateral depressions".

The Pilomasin Basin in south Sumatra is a shallow depression, 4.4 by 3.3 kilometers across, lying immediately east of a line of fissure-vents from which more than 4 cubic kilometers of siliceous pumice were erupted in Vulcanian fashion. While the eruptions were going on, the basin sank spasmodically, so that pumice-fall deposits are interbedded concordantly with inward-dipping, lacustrine clays. The volume of pumice, recalculated as magma, is almost precisely equal to the volume of the basin. It seem likely, therefore, that the subsidence was related genetically to the eruptions (van Bemmelen, 1931).

A comparable, intermittent subsidence may account in part for the basin that holds Mono Lake, California. Repeated eruptions of rhyolitic pumice and obsidian flows from a chain of vents extending southward from the lake may have drained magma from a throughgoing fissure to cause periodic sagging of the basin. Two Quaternary cinder cones, one within the lake and the other on its northern shore, and a central island largely composed of deformed beds of lacustrine pumice and diatomite, denote the presence of an underlying body of magma, fluctuations of which may account not only for the cones and disturbance of the lake sediments but also for the general subsidence.

Valles Type (A 3)

The largest calderas in the world are related to the discharge of colossal volumes of rhyolitic, quartz latitic, and dacitic ash flow tuffs (ignimbrites) from linear and arcuate fissures opened during regional tumescences as bodies of acid magma rise through the crust. Examples include the large calderas in the North Island of New Zealand, and those being discovered in increasing number in the Great Basin of Nevada and Utah. Maucher (1960) has described a caldera approximately 60 kilometers in diameter in the Italian Dolomite Alps, where it is associated with Permian rhyolitic ash flows having a volume of 2,000 cubic kilometers. And Elston (1965) has suggested, on the basis of preliminary studies, that the Mogollon Plateau of southwest New Mexico is a circular caldera or volcano-tectonic depression no less than 150 kilometers across, related to the eruption of Miocene rhyolitic ash flows having an aggregate volume of perhaps 12,000 cubic kilometers. The huge basin is entirely surrounded by a circular graben and has a raised rim apparently formed by eruptions of rhyolitic pumice and protrusion of elongated rhyolitic domes from vents along a ring fracture.

1. The Valles caldera. One of the largest, and certainly one of the most instructive and most carefully studied calderas in the world lies close to the western edge of the Rio Grande graben in northern New Mexico (Smith and Bailey, 1966). It measures 20 by 25 kilometers from rim to rim (figure 8). Its walls surround a circular moat, between 3 and 6 kilometers wide, partly filled by a ring of rhyolite domes. Towering more than 650 meters above the moat is a central mountain, Cerro Redondo, 10 by 13 kilometers across, which is a "resurgent dome" produced by uplift of the caldera floor.

The Valles caldera does not occupy the beheaded top of an ancestral cone; on the contrary, it lies on the western flank of an irregular group of coalescing cones and domes built mainly during Pliocene time by discharge of dacitic and rhyolitic lavas and pyroclastic ejecta, along with subordinate andesites and basalts.

The climactic, caldera-forming eruptions began about 1,400,000 years ago, probably from arcuate fissures close to the northeast edge of the Valles caldera. These eruptions laid down the 'Lower Bandelier Tuff'. The first outbursts were of the high-pressure Vulcanian kind and they produced a mantle-bedded layer of coarse rhyolitic pumice. The preliminary pumice falls were followed immediately by a rapid succession of increasingly hot, mafic, and crystal-rich rhyolitic pumice flows that swept down the mountainsides at great speed, leaving a deposit of approximately 200 cubic kilometers. The source-area then collapsed about 1,000 meters, forming the Toledo caldera.

A million years ago, after a few domes of quartz latite had grown within the Toledo caldera and deep canyons had been cut through the 'Lower

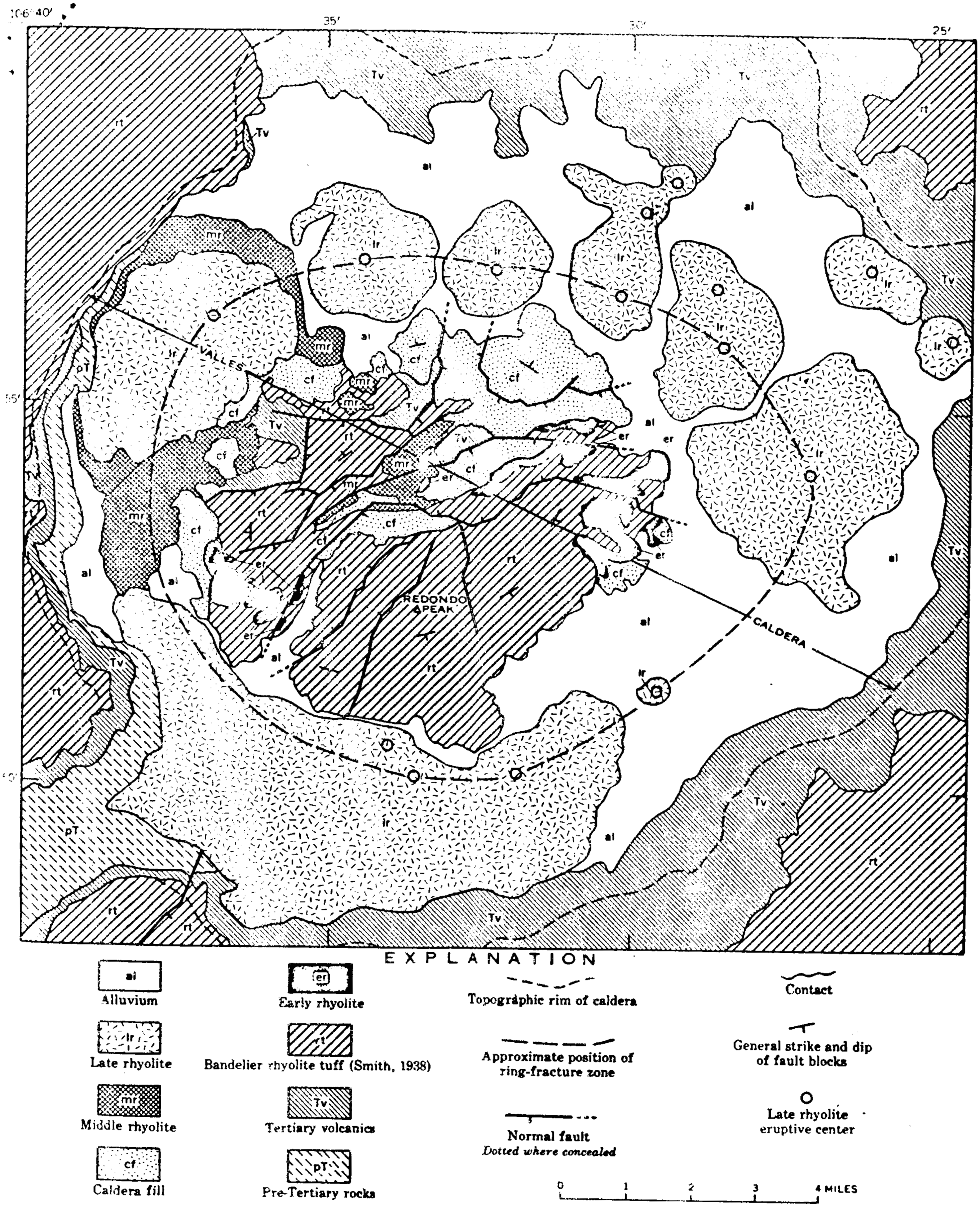


FIGURE 340.1.—Generalized geologic map of the Valles caldera, Jemez Mountains, N. Mex.

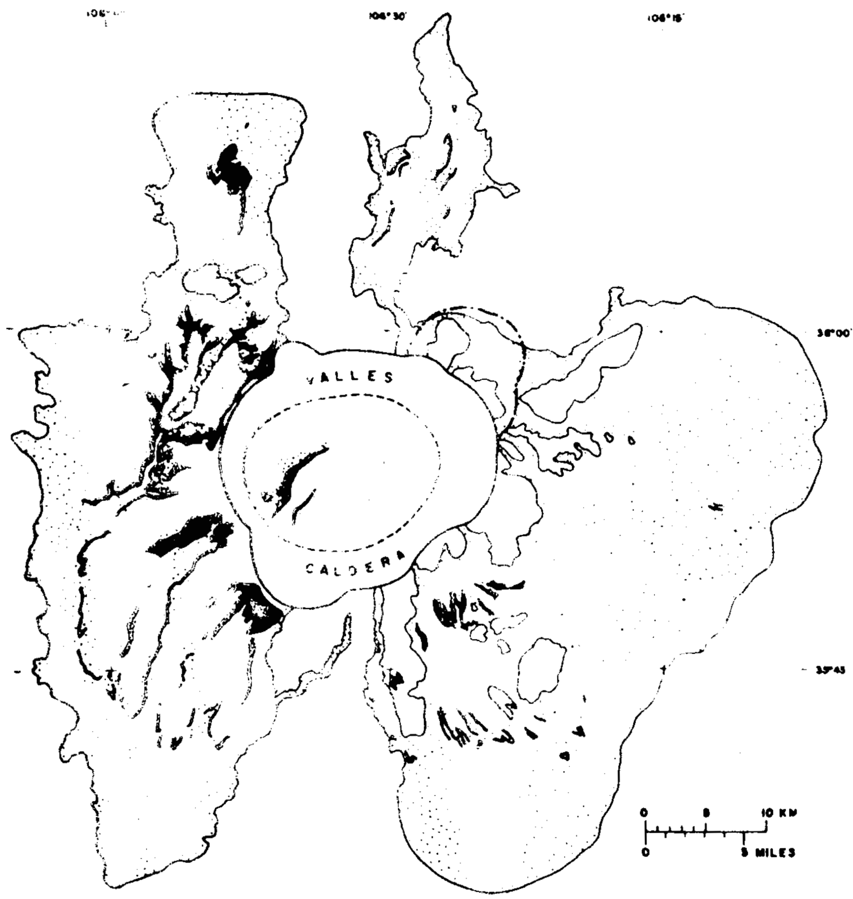


FIG. 4 - Distribution of Bandelier Tuff around the Valles Caldera (topographic rim, solid black line; ring fracture, dashed line) and Toledo Caldera (topographic rim, dash-dot line). Outcrop of lower member of the Bandelier, solid black. Maximum extent of upper member of the Bandelier, dotted.

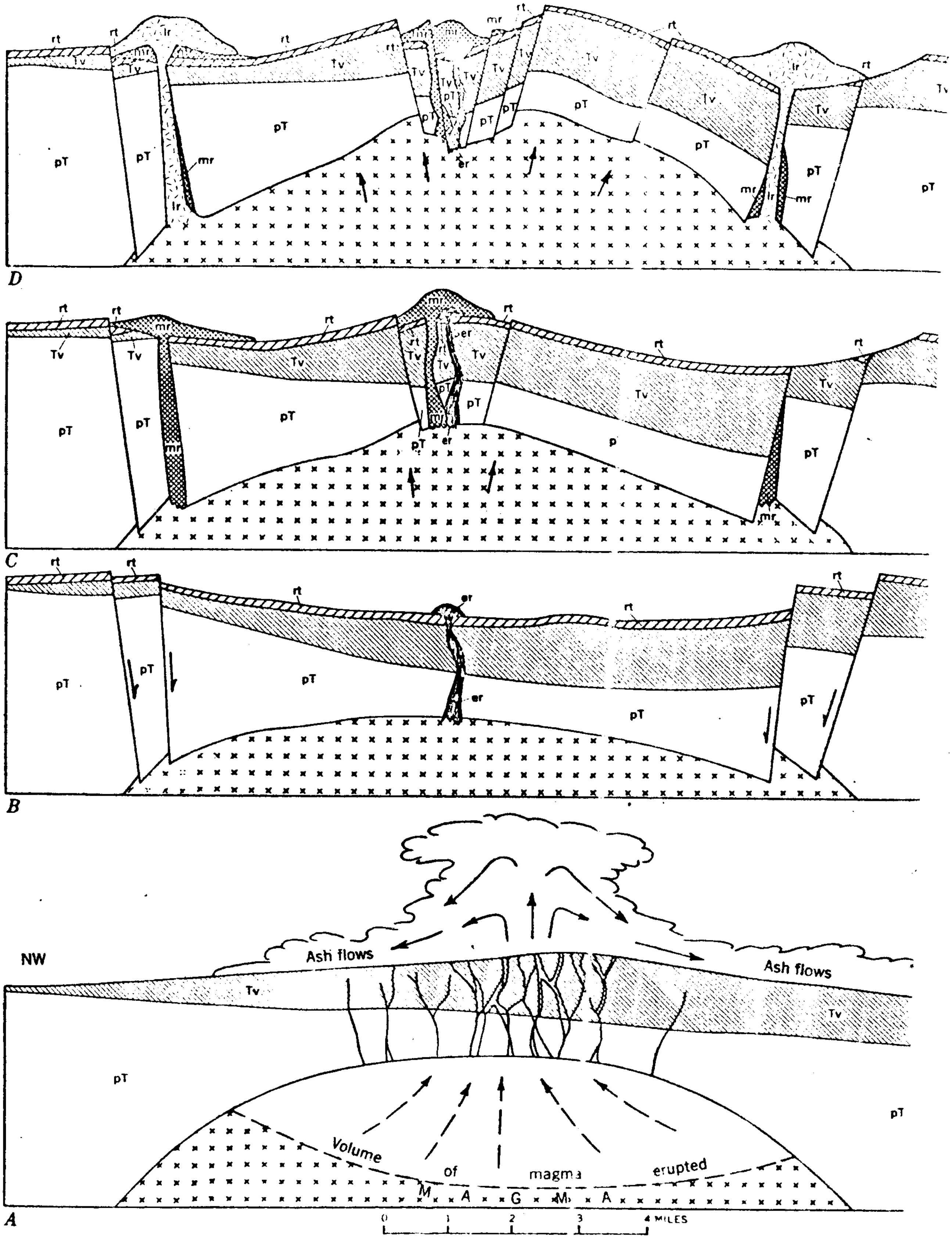


FIGURE 340.2.—Schematic sections showing the evolution of the Valles caldera (no vertical exaggeration). Symbols are the same as for figure 340.1. Caldera fill omitted for simplicity.

Bandelier Tuff', eruptions of rhyolitic ash and pumice were resumed, this time from ring fractures southwest of the Toledo caldera. Once more, the initial showers of pumice were followed by vastly larger pumice flows, the 'Upper Bandelier Tuff', having an aggregate volume close to 200 cubic kilometers. Sagging or partial foundering of the source-area seems to have closed the eruptive conduits temporarily, after the first 80 km^3. of magma were discharged; then followed sporadic eruptions separated by short intervals of quiet, until the final 40 km^3. of magma were rapidly discharged and the final collapse took place, the source-area foundering 1,000 to 1,500 meters to form the Valles caldera.

Shortly after the caldera was formed, 'Early Rhyolites', essentially similar in composition to the 'Upper Bandelier Tuff', issued from northeast-trending fissures near the center of the caldera, partly as viscous flows and partly as pumice showers, and a lake began to grow within the great depression. At or about the same time, the central part of the caldera floor started to rise, probably forced upward by resurgence of new magma into the underlying reservoir. Ultimately, this "resurgent dome" rose more than 1,500 meters, approximately as far as the floor of the caldera had previously subsided. And while the "resurgent dome" rose, sediments continued to accumulate in the surrounding lake, and 'Middle Rhyolite Flows' issued sluggishly from an apical graben on top of the dome and from fissures near the northwest base. Thereafter a group of large domes of rhyolitic lava was built over ring fractures within the moat. The first of these viscous protrusions, which rose about 900,000 years ago, was the easternmost; others rose in turn in an anticlockwise direction, the last of them rising close to the southern wall of the caldera about 400,000 years ago. Weak hydrothermal and hot spring activity still continues near the western base of the "resurgent dome" and within its apical graben.

The ring fractures through which the great volumes of ash and pumice were erupted and along which collapse took place to produce the Toledo and Valles calderas are presumed to be vertical or to dip steeply inward. They were formed and widened by repeated tumescence of the roofs of the magma chambers, thus permitting both the rise of magma to the surface and subsequent foundering.

2. Creede caldera, Colorado. Several calderas originated in the San Juan Mountains during Late Tertiary time as a result of copious eruptions of ash- and pumice-flows (Steven and Ratté, 1965; Ratté and Steven, 1967). The town of Creede lies inside one caldera and near two others. The oldest of these three calderas -- La Garita -- originated several miles northeast of Creede following eruption of a large volume of quartz latite ignimbrite. Rhyolitic and quartz latitic lavas and pyroclastic ejecta were then erupted from vents a short distance to the south, and these were inundated by a composite sheet of rhyolitic ignimbrite between 20 and 40 cubic kilometers in volume. The accompanying collapse formed the Bachelor Mountain caldera, in part coincident with the older, La Garita caldera and with the younger, Creede caldera.

Several collapses, each initiated and accompanied by colossal ash-flow eruptions, produced the final, Creede caldera. The first eruptions issued from vents inside the present depression, laying down rhyolitic lavas and ash-flows. Deposits of succeeding rhyolitic ash flows nearly surround the caldera; their average thickness ranges from 200 to 250 meters and their aggregate volume exceeds 100 cubic kilometers. Their discharge was in part concurrent with the early stages of collapse. By this time, the upper, relatively crystal-poor rhyolitic portion of the magma chamber had been evacuated. Then, as the floor of the caldera continued to sink, a series of crystal-rich quartz latite ash flows was discharged, the last ones, the Snowshoe Mountain ash flows, which are restricted to the center of the caldera, reaching a thickness of more than 2,000 meters and a volume of more than 400 cubic kilometers. The principal collapse took place during and after these eruptions, so that tongues of talus from the walls of the caldera interfinger with some of the final ash flows.

After the caldera had reached its full depth, the central part of the floor was uplifted 1,300 meters by intrusion of a stock or laccolith, producing the "resurgent dome" of Snowshoe Mountain. This dome, like Cerro Redondo within the Valles caldera, is broken by an apical graben and by many other faults, and is separated from the caldera walls by an intervening moat. Viscous flows and bulbous piles of lava were then extruded, not only from vents inside the moat, as in the Valles caldera, but also from vents beyond the margin of the caldera. And here they consisted, not of rhyolite as in the Valles caldera, but of quartz latite. Their eruption closed a long sequence marked by an upward change from crystal-poor rhyolite to crystal-rich quartz latite, and a concurrent increase in lavas at the expense of ash flows, changes reflecting a general diminution in gas content as lower and less siliceous levels of the magma reservoir were tapped.

Lake- and stream-deposits, layers of airborne ash, and travertine accumulated within the caldera during and after the rise of the "resurgent dome" and subsequent eruption of quartz latites; together these intra-caldera beds make up the Creede formation of late Miocene or Pliocene age.

Not all of the faulting associated with the Creede caldera took place along ring fractures; there was also recurrent faulting along north- to northwest-trending lines before, during, and after its development. These linear faults were probably controlled by structures in the sub-volcanic basement; some of them border the apical graben that crosses the Snowshoe Mountain

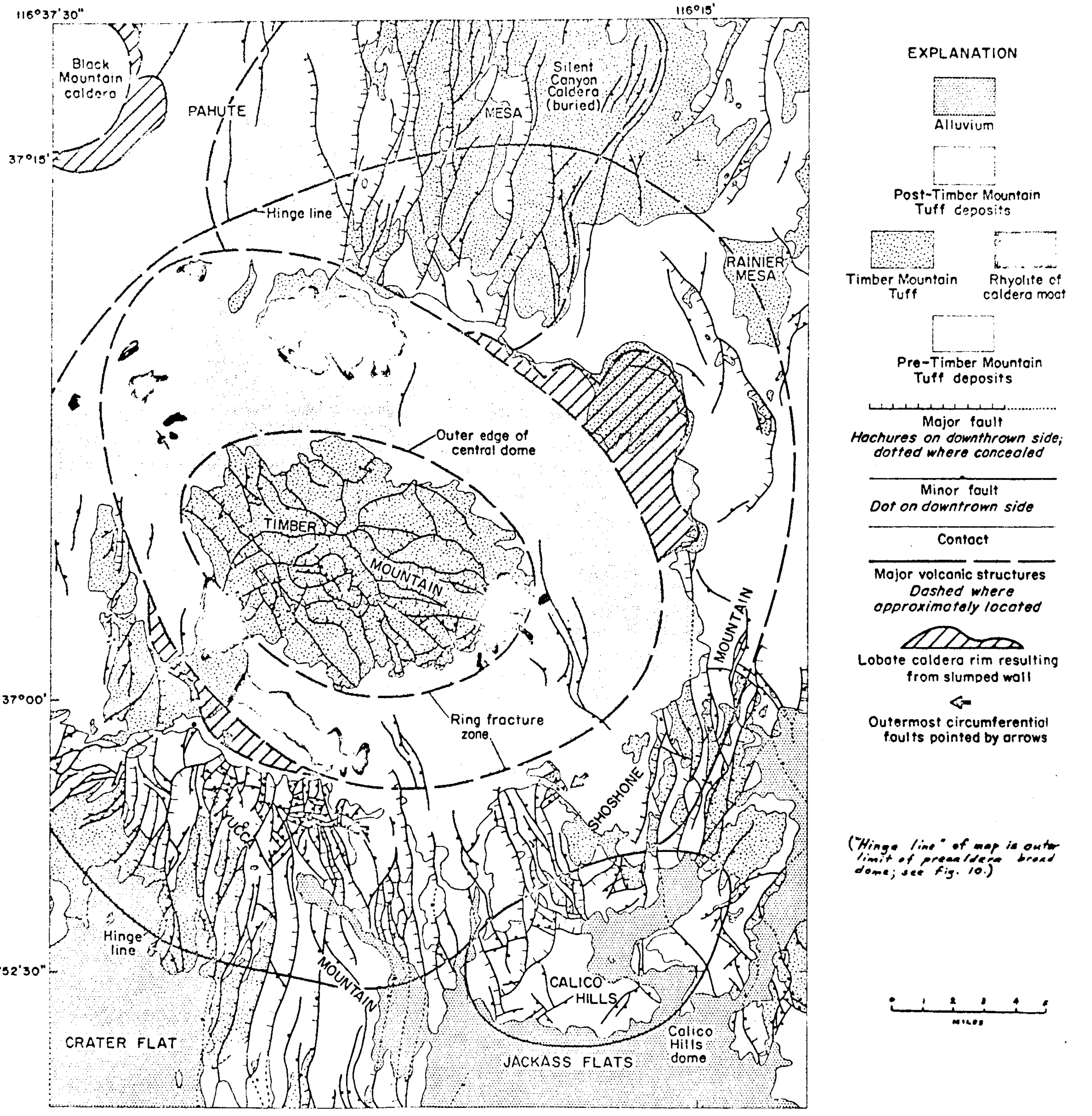


Figure 9. Structure map of Timber Mountain caldera and vicinity.

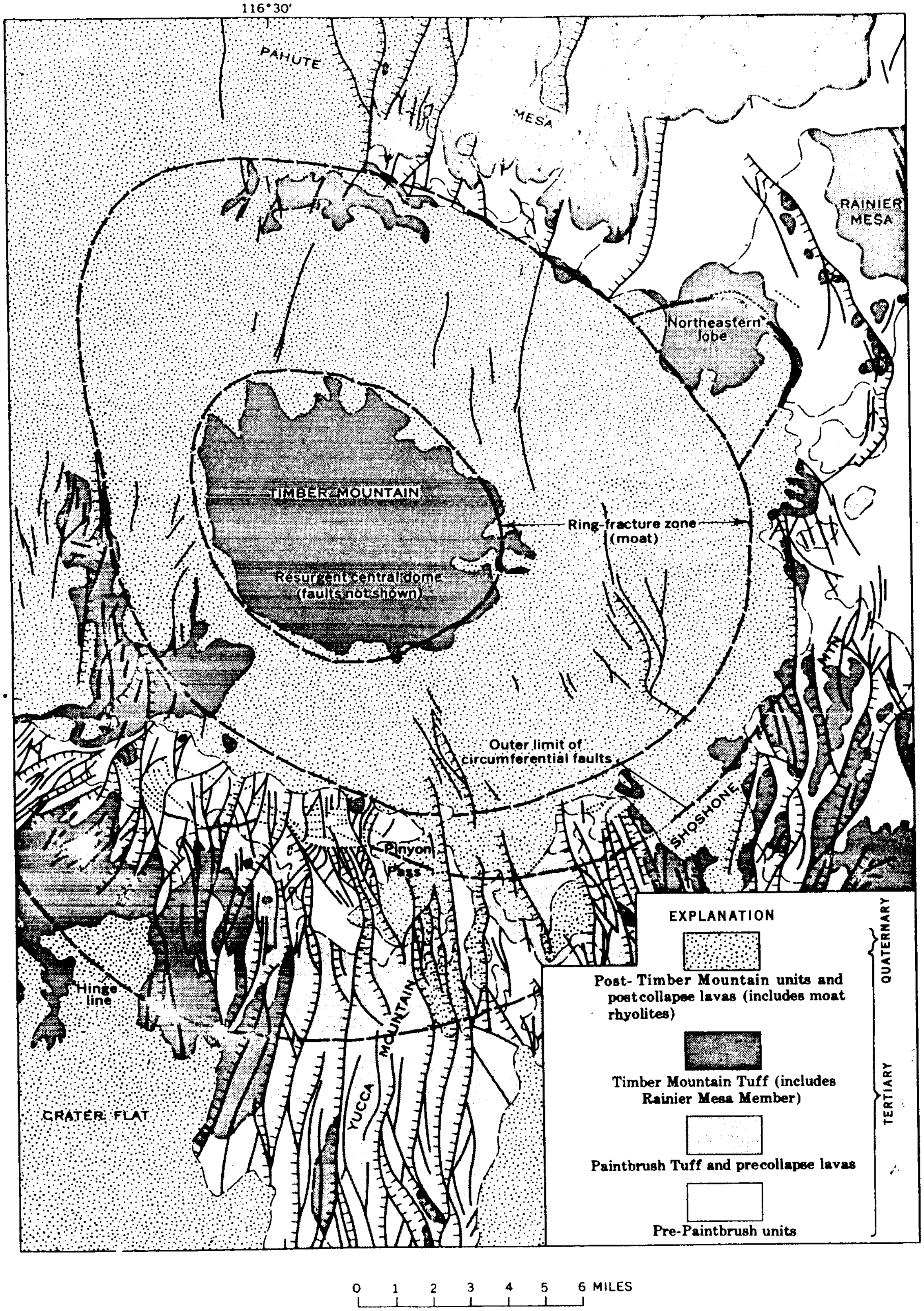


FIGURE 2.—Generalized geologic map of the Timber Mountain area, Nevada. Line symbols: all contacts dashed; hachured lines, major basin-range faults, with hachures on downthrown side; solid lines, minor basin-range faults; dotted lines, circumferential faults.

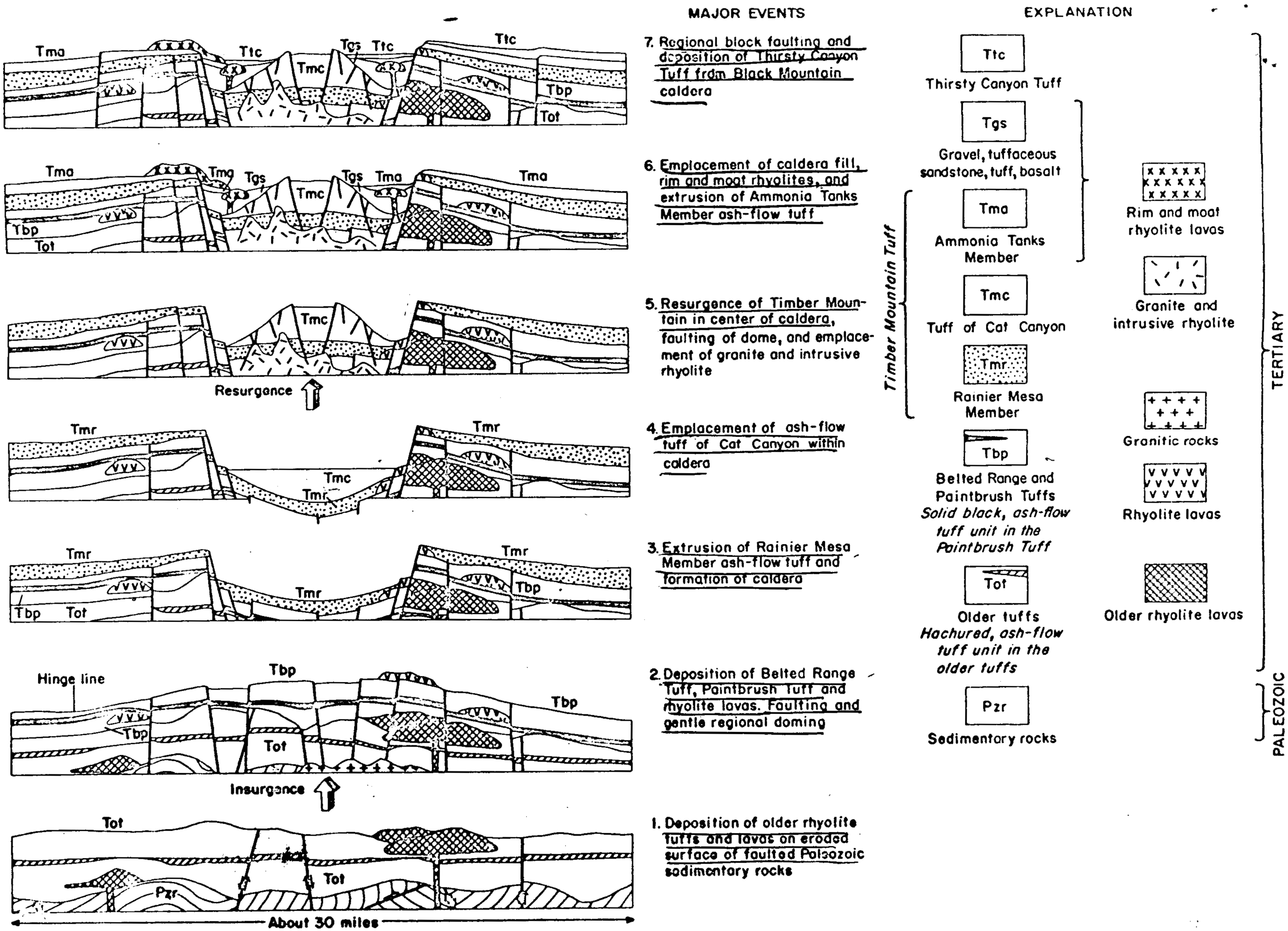


Figure 10.--Diagrams of geologic history of Timber Mountain caldera, Nevada

dome, and some border the Creede graben that extends outward from the north wall of the caldera. Virtually all of the mineralization for which the Creede district is famous took place along these linear faults after the Creede formation had been deposited and after all surface volcanism had come to an end.

3. Timber Mountain caldera, Nevada. The Great Basin of Nevada and western Utah is one of the world's largest ignimbrite provinces. At intervals for approximately 30 million years, from early Oligocene to late Pliocene times, ash-flow tuffs were laid down over an area of about 400,000 square kilometers to an average thickness of at least 300 meters and locally to a thickness of more than 2,000 meters. Detailed studies have thus far been restricted to the eastern margin of the basin and to its southwestern part, within and near the Nevada Test Site. These studies have revealed the presence of many large calderas related to the ash-flow eruptions, and no doubt others will be discovered as field work is extended. We select for description the largest caldera yet described, namely the Timber Mountain caldera within the Nevada Test Site (Carr, 1964, and Christiansen, 1965 and 1966).

Miocene lavas and ash-flow tuffs had already been deposited and had been broadly arched and broken by arcuate faults prior to the eruptions that led to formation of the caldera (figure 12). These eruptions took place about 11 million years ago, and some of them probably issued from the arcuate faults just mentioned. They laid down a thick succession of rhyolitic ash flows which together constitute the Rainier Mesa member of the Timber Mountain Tuff. The deposits cover 8,000 square kilometers and their volume approximates 1,150 cubic kilometers. It was the rapid discharge of this great volume of material that was responsible for most of the collapse which formed the caldera. Eruptions from vents within the caldera then deposited similar rhyolitic ash flows -- the Cat Canyon member of the Timber Mountain Tuff -- to a thickness of more than 900 meters, and the caldera floor probably subsided still farther as a consequence. The central part of the floor was then upheaved at least 1,200 meters to make a "resurgent dome", 13 by 16 kilometers across, elongated in a northwest direction, in line with the adjacent Black Mountain caldera and the Calico Hills dome (figure 11). Many arcuate fractures developed along the margins of the rising dome, the principal ones dipping outward, toward the surrounding moat, and some fractures were intruded by dikes of rhyolite and larger bodies of granite porphyry. Subsequently, as the "resurgent dome" continued to rise, longitudinal and radial grabens developed near its center, and intrusions and extrusions of rhyolite took place both within the dome and along ring fractures within the encircling moat. Shortly thereafter, ash flows were again erupted from vents inside the caldera, some accumulating within the depression to lap against the flanks of the "resurgent dome" and caldera walls, while others swept beyond the walls to inundate most of the area previously buried by the Rainier Mesa ash flows. The volume of the younger flows, which constitute the Ammonia Tanks member of the Timber Mountain Tuff, approximates 720 cubic kilometers, and one can hardly doubt that their eruption was attended by renewed collapse of the caldera floor.

All of the foregoing events occurred within a few hundred thousand years. Subsequently, about 7 million years ago, the Thirsty Canyon ash flows were discharged from the adjacent Black Mountain caldera, some of them pouring into the moat of the Timber Mountain caldera. Upturning of some of the Thirsty Canyon tuffs along the margin of the Timber Mountain "resurgent dome" indicates that the caldera floor continued to rise to a still later time.

The history of the Timber Mountain caldera is thus one of recurrent tumescence and arcuate fracturing. Colossal ash flows and smaller lava flows issued from these fractures during episodes of inflation; collapses took place along them during episodes of deflation.

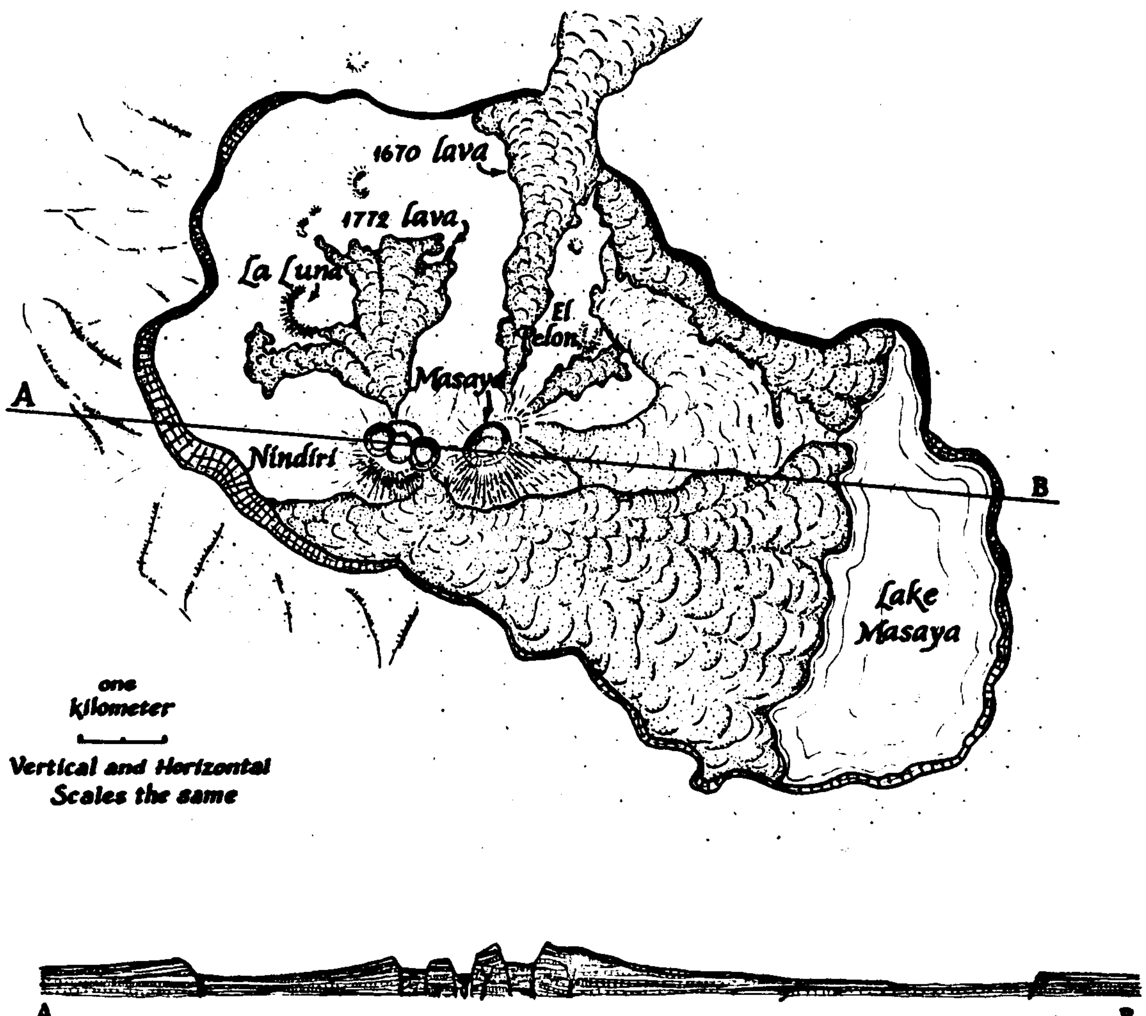
B. Calderas associated with effusive eruptions of basaltic magma.

Masaya type (B 1)

Masaya caldera, Nicaragua. The floor and walls of Masaya caldera consist entirely of iron-rich basaltic lavas and subordinate basaltic scoria. Nonetheless, the caldera lies approximately midway in the chain of Central American composite cones, most of which consist of andesite, and only 5 kilometers to the southwest lies Lake Apoyo, inside a typical Krakatoa-type caldera formed by collapse following copious eruptions of dacite pumice. Perhaps nowhere in the world does a basaltic caldera lie closer to one of Krakatoan type.

The Masaya caldera is elongated in a west-northwest direction, parallel to an adjacent fault scarp and to the general trend of the Quaternary volcanic chain of Nicaragua, and its long axis, if prolonged, passes through the center of the adjacent Apoyo caldera. It measures approximately 6 by 11 kilometers across. On the west, its walls rise about 150 meters above the floor; on the south and east, they generally rise between 80 and 100 meters; northward, however, they dwindle in height and finally disappear. The low southwest wall seems to be the truncated flank of an older composite cone (figure 13).

Calderas of Hawaiian and Galápagos type, to be described in the sequel, lie inside the tops of huge basaltic shield volcanoes; Hawaiian ones form at the intersection of rift zones on the flanks of the shields, from which abundant lavas are discharged; those in the Galápagos Islands are surrounded by circumferential fissures, and the adjacent flanks are cut by radial fissures, from all of which lavas issue in large volumes at short intervals. The Masaya caldera, on the contrary, lies inside the top of a low, broad, asymmetrical shield with slopes so gentle as to be hardly perceptible. No arcuate eruptive vents border the rim and no rift zones cut the outer flanks (McBirney, 1956). All



Masaya Caldera
Nicaragua

eruptive vents lie within the caldera, and their lavas, unable to spread to the west and southwest because of the opposing fault scarp, or to the south and southeast because of the slopes of the older, Apoyo volcano, poured in other directions, chiefly to the north.

About a dozen craters and scoria cones are arranged in a circle, roughly 4 kilometers in diameter, within the western part of the caldera, the five largest craters lying close together on the southern periphery (figure 13). These ring-vents probably outline the boundary of an early collapse-structure. Almost all of the lavas on the caldera floor issued from them; that is why the floor has a gently domical form.

During the 16th and 17th centuries, the crater of Nindiri, one of the five large craters on the ring of vents, was periodically occupied by a lava lake. The level of the lake, though generally close to that of the surrounding caldera floor, fluctuated rapidly; in 1670, the crater was flooded so that lava poured northward to inundate the caldera rim and spread beyond for a distance of 9 kilometers. Since that time, lava has risen into the adjacent craters, but never to form long-lived lava lakes.

No general subsidence of the caldera floor has been observed during historic times. However, the floors of some of the larger craters have subsided as much as 200 meters during cycles of eruption. Two large pit craters originated by collapse in 1858-1859, accompanying or following seismic shocks and mild eruptions of gas. And the floor of Santiago crater sank about 40 meters in 1927 as a result of dynamite explosions. Recorded history, considered along with the scalloped margin of the depression, suggests that Masaya caldera was produced by a succession of more or less cylindrical collapses arranged roughly along tectonic, west-northwest - trending lines, and resulting primarily from periodic migration of magma underground. Explosive eruptions and flank flows played no part in its development.

A small positive gravity anomaly coincides with the vent-area in the western part of the caldera, but no marked anomalies are associated with the boundary faults around the main depression.

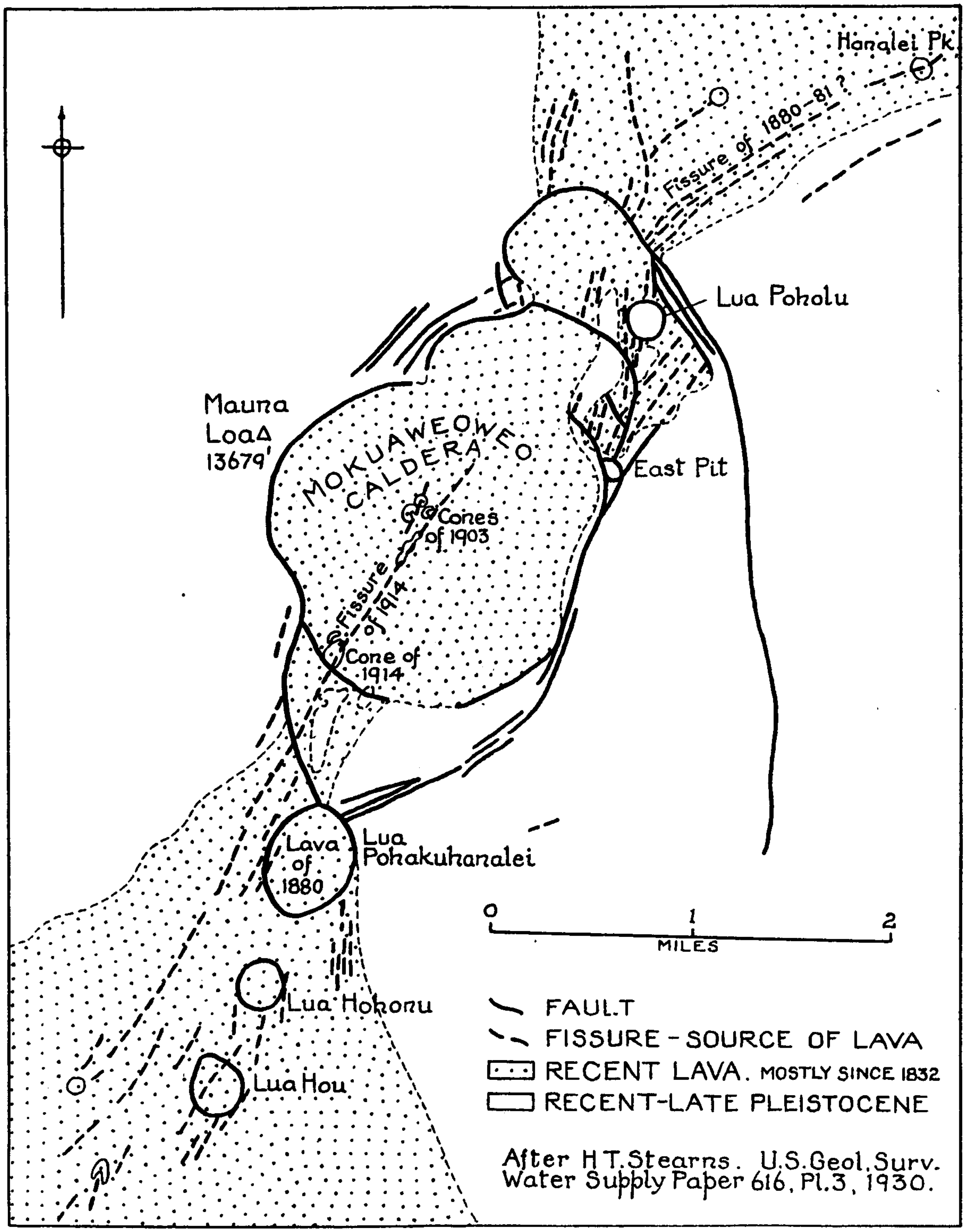
Hawaiian Type (B 2)

Hawaiian shield volcanoes are built mainly by eruption of fluid flows of tholeiitic basalt that follow each other in quick succession, spreading as thin sheets, usually of great extent. Summit calderas originate when the shields have grown almost to their full height, while activity is still vigorous and eruptions continue at short intervals. As the calderas increase in width and depth, lavas are poured from fissures cutting their floors and walls, and from rift zones on the flanks of the shields. In general, three rift zones radiate outward from each caldera at angles of approximately 120°, one of the three being much less distinct than the other two. Ultimately, subsidence of the floor comes to an end, and the caldera-filling stage begins. Thick flows of tholeiitic basalt accumulate inside the caldera; then flows of more alkaline basalt. Eruptions take place at much longer intervals than during the shield-building stage, explosive activity becomes more frequent, and the magma types more varied, ranging from alkaline basalts to trachytes. The calderas are finally buried, and only deep erosion reveals their former presence, as it has on most of the Hawaiian Islands. One exhumed caldera, on the island of Kauai, is the largest of all Hawaiian calderas, measuring 16 by 20 kilometers across (Wentworth and Macdonald, 1953; Macdonald, 1965).

Kilauea and Mauna Loa, the two active volcanoes on the island of Hawaii, both have summit-calderas. That of Kilauea measures 4.4 by 3.3 kilometers across and is approximately 150 meters deep; Mokuaweoweo, the caldera on top of Mauna Loa, measures 2.6 by 4.5 kilometers across and its depth ranges from 130 to 180 meters (figure 14). Both have steep walls, interrupted in places by step faults, and both are partly surrounded by benches, up to 3 kilometers wide, traversed by inward-facing fault scarps. Both are also closely associated with pit craters; indeed the calderas grow in part by coalescence and incorporation of adjoining pits. The outermost caldera faults of Kilauea lie outside the pit craters of Kilauea iki and Keanakakoi, and three pits lie close to Mokuaweoweo, on the southwest rift zone of Mauna Loa. Some pit craters are less than 30 meters across; others are more than a kilometer across. A few small ones originate by collapse of the roofs of lava tubes, but most of the larger ones result from collapse of more or less cylindrical blocks when magma stopes and melts its way almost to the surface and then withdraws into adjacent rifts, leaving the surficial rocks without support.

None can doubt that Hawaiian calderas originate by engulfment. Williams (1941) suggested that collapse results from drainage of the central conduits by subterranean injection of magma into rift zones and by copious eruptions of lava from rift zones far down the flanks of the shields. Subsequent studies have shown that he was unduly impressed by the rapid enlargement of Halemaumau, the inner pit of Kilauea caldera, during the 1924 eruptions when a long-lived lava lake was drained from the pit by injection of magma into the adjacent rift zone and by eruption of lava on to the sea floor from its eastern end. Williams did not attempt to estimate how much the pit of Halemaumau was enlarged by avalanching of the walls, by injection of magma into the rift zone, and by submarine effusions. It is now known, however, that the volume of lava discharged at the surface during any Hawaiian eruption is invariably less, and generally very much less than the volume of the summit-collapse and general subsidence of the surrounding region.

Stearns and Macdonald (1946), realizing this to be the case, and realizing also that the tops of growing shield volcanoes are always unstable


Hanalei Pk.
Fissure of 1880-81 ?
Lua Poholu
Mauna Loa
13679'
MOKUAWEOWEO CALDERA
East Pit
Cones of 1903
Fissure of 1914
Cone of 1914
Lava of 1880
Lua Pohakuhanalei
0
1
2
MILES
Lua Hohonu
Lua Hou
FAULT
FISSURE - SOURCE OF LAVA
RECENT LAVA. MOSTLY SINCE 1832
RECENT-LATE PLEISTOCENE
After H.T. Stearns. U.S. Geol. Surv.
Water Supply Paper 616, Pl.3, 1930.

owing to the presence of abundant liquid magma in underlying fissures, suggested that collapse took place when magma stoped its way upward until the solid cover, becoming too weak and thin to support itself, foundered. In other words, they thought that Hawaiian calderas were cauldron subsidences of the Glencoe type. Subsequently, Macdonald (1965) abandoned this interpretation because of the absence of circumferential dikes around eroded Hawaiian calderas and of circumferential eruptions around the active ones.

Careful geophysical studies during recent years have amply shown that the calderas of Kilauea and Mauna Loa generally, if not always collapse following broad tumescence of the shields, when rift zones are distended and injected by magma, whether or not lavas issue at the surface. The shields seem almost to breathe; they swell and subside as underlying magma rises and falls. At times, the entire upper part of the shield is gently uparched and later depressed; at other times, the caldera floor moves up and down like a gigantic piston.

Gravity studies have lately shown that several Hawaiian volcanoes are underlain at depths of a kilometer or two by ultra-dense rock -- presumably olivine-rich cumulates or protrusions from the sub-crustal mantle -- that extend downward at least to the depth of the ocean floor (Macdonald, 1965). Parts of these growing, high-density columns periodically spread sideways, entering and wedging apart the bordering rift zones. In doing so, they cause the volcanoes to swell, widening the fractures around the summits and the rift zones on the flanks.

Evidence gathers to show that the fracture planes encircling the summit-calderas of Hawaiian shields dip steeply inward, as cone sheets do, because of intermittent upthrusts of magma. The basin-shape of the lavas inside the eroded Koolau caldera on Oahu suggests the sinking of a wedge-shaped block. Collapse of the summit of a shield can therefore take place only after fractures have been widened by general distension of the shield in the manner just suggested. In brief, tumescence is a necessary preliminary to collapse. But why widening of ring fractures around the calderas is never accompanied by eruptions from rim volcanoes remains to be explained.

Galápagos Type (B 3)

The westernmost islands of the Galápagos archipelago, Albemarle and Narborough, constitute one of the most active volcanic regions in the world. Five coalescing basaltic shield volcanoes make up most of Albemarle Island, and a single one, 25 by 30 kilometers across and almost 1,500 meters high, makes up Narborough Island. Each of these six shields, like Kilauea and Mauna Loa, is in the mature stage of vigorous growth, erupting abundant flows of tholeiitic basalt, and each has a large caldera at the summit. Similar calderas were formerly present on most of the other islands of the archipelago, but, like most of those in the Hawaiian archipelago, they are buried beneath alkaline lavas and cinder cones.

The active shield volcanoes of the Galápagos Islands differ chiefly from those on the Island of Hawaii in their profiles, the dimensions of their calderas, and especially in their fissure-systems. Hawaiian shields tend to have the forms of overturned saucers, their flattish tops passing outward to gentle slopes with average inclinations of only 3° to 6°. The Galápagos shields, on the contrary, particularly the three northern ones, Darwin, Wolf, and Narborough, have the shapes of overturned soup plates or even of the tortoises from which the archipelago takes its name! The slopes of Darwin volcano approximate 18°, flattening rapidly near the base and top; the upper part of Wolf volcano is almost flat but the slopes beyond increase to 35°; the slopes of the mighty Narborough shield also increase outward from a wide, flattish bench at the summit to 15° or 25°, diminishing rapidly near the base until they again become almost flat.

The largest Galápagos caldera, that of Sierra Negra, measures 7 by 10 kilometers across, though it is also the shallowest, being less than 110 meters deep. The almost circular Darwin caldera is 5 kilometers wide and between 200 and 230 meters deep. Much more impressive are the calderas of the Wolf and Narborough shields; that of Wolf measures 5 by 7 kilometers from rim to rim, and 615 meters in depth; that of Narborough measures 4.6 by 6.7 kilometers at the rim, and 2.5 by 3.3 kilometers at the floor, and its depth varies from about 770 to 845 meters.

From a genetic standpoint, however, the principal difference between Galápagos and Hawaiian shield volcanoes is in the geometry of their feeding fissure-systems. Most Hawaiian shields, as noted already, have two main rift zones that converge toward the summit-calderas at angles of 120° or more, and a third, much weaker rift zone that bisects the other two; arcuate rift zones around the calderas are conspicuous by their rarity or absence. On the contrary, no volcanoes in the world display more spectacular examples of circumferential fissures around their calderas or radial fissures on their flanks than do the active shield volcanoes of the Galápagos Islands.

The flattish benches surrounding the Galápagos calderas are traversed by arcuate, concentric fissures, some of which are gaping cracks, though most are marked by chains of small spatter- and scoria-cones. Countless basaltic flows have issued from these ring fractures during recent times, but only a few have spread inward to cascade into the adjacent calderas; almost all have poured outward, down the steep flanks of the shields, to reach the sea. This seems to be the principal means by which the shields now grow. The number and length of the arcuate fissures vary considerably, even around the same caldera, and probably no single fissure describes a full circle at the surface, though many may do so at depth. Around parts of the Narborough caldera, there are at least four concentric fissures, the outermost more than a kilometer from the rim. No vertical displacements seem to have taken place on any of these circumferential fissures; rather than being formed by periodic subsidences, they probably owe their origin to periodic distensions of the shields by rising magma. On the other hand, the walls of some Galápagos calderas are cut by concentric, inward-facing fault scarps, at the bottoms of which magma has escaped to produce cinder- and spatter-cones or to form flows that have tumbled to the caldera floors. These arcuate fault scarps, and the arcuate benches within all of the calderas clearly indicate spasmodic engulfments.

The radial fissures, so strikingly developed on the flanks of the Galápagos shields, originated during times of rising magma, when the volcanoes tumesced. The lavas erupted from them have contributed greatly to the growth of the shields, but not nearly as much as the flows from the ring fractures around the summits. All of the radial fissures are marked by lines of spatter- and scoria-conelets, many of which are breached on their lower sides owing to downslope extensions of the rifts during eruptive activity.

Our observations suggest that the steep slopes of the Galápagos shields result mainly from distension as downward-widening ring dikes are injected into concentric fissures encircling the calderas. Injection of sills from the central conduits may be a subsidiary cause. The calderas originate by collapse along ring fractures; this may take place passively, by sinking of the heavy floors into underlying magma, or when support is removed as magma rises up the bordering fractures to issue from the rim vents or breaks through radial fractures to pour down the flanks of the shields. Explosive activity plays a negligible part in the process.

C. Calderas associated with mixed eruptions from ring fractures.

Glencoe Type (C 1)

1. Glencoe. The first cauldron subsidence to be described was that of Glencoe, Scotland, where a thick series of Devonian volcanic rocks

subsided within an oval area, approximately 8 kilometers across, enclosed by a ring intrusion. Clough, Maufe, and Bailey, in their classic paper on this cauldron (1909), supposed that subsidence and marginal intrusion took place almost if not actually at the same time, magma welling up around the subsiding block. They also thought that the surface caldera formed by subsidence probably resembled that of Askja in Iceland, though they could not be certain that rim volcanoes ever surrounded the Glencoe caldera.

The term 'Glencoe type' of caldera was introduced by Williams in 1941 to include "cauldron subsidences resulting from the collapse of the roof of a magma chamber along ring fractures, that is, stoping of cylindrical blocks of the crust". Ring-fracture stoping was thought to be consequent upon either "withdrawal of magmatic pressure at depth or, more likely, foundering of heavy crustal rocks into lighter magma below". Implicit in the definition was the idea that during or after subsidence of the central block, magma might rise along encircling fractures to feed ring volcanoes around a surface caldera.

Recent studies of the Glencoe caldera by Roberts (1963, 1966) and Taubeneck (in press) have shown that the downdropped volcanic rocks within the encircling ring dikes include two groups of rhyolitic ignimbrites interbedded with lavas that range in composition from basalts to andesites, and with volcanic sediments. Subsidence began at an early stage, during or shortly after the first group of ignimbrites was erupted from the peripheral ring fractures. Most of the later lavas, as well as the second group of ignimbrites, were also erupted from the marginal ring-fault zone. Coarse sediments accumulated close to the caldera walls while finer sediments accumulated near the center. Intermittent eruptions from rim volcanoes were accompanied by intermittent subsidence of the caldera floor. It was by a combination of gravitative settling and magma discharge through ring fractures that cauldron subsidence was brought about.

Most, if not all of the calderas associated with the Tertiary ring complexes of Scotland and Ireland, the Triassic ones of New England, and the Permian ones of the Oslo district, Norway, are calderas of the Glencoe type. None of them seems to have originated by collapse of the summit of a large central volcano; on the contrary, they appear to have been formed by long-continued, mixed eruptions from arcuate vents along and close to the margins of subsiding basins. No modern analogue is known in which a ring of eruptive vents surrounds a sinking caldera unless it be one that occupies the site of a former volcanic cone.

2. Lake Atitlan, Guatemala. The basin that holds Lake Atitlan was approximately circular, measuring 19 by 21 kilometers across, before three large, youthful volcanoes, Atitlan, Toliman, and San Pedro grew along its southern side (figures 15 & 16). Elsewhere, the walls of the basin consist of gently folded Pliocene lavas, ignimbrites, tuffaceous sediments, and diatomites that rest in part on plutonic rocks. The faults scarps that enclose most of the basin vary from about 300 to 600 meters in height and are deeply eroded; on the south side, however, particularly near the young volcanoes, the scarps are fresher, steeper, and lower, and they converge to the central vent of the Atitlan volcano.

No volcano ever occupied the present site of the Atitlan basin, and its crudely circular outline transects the structures of the surrounding Pliocene rocks at random. Before the basin developed, its site was crossed by deep canyons separated by narrow divides that dwindled in height southward, toward the sea; but the outline of the basin in no way reflects the former topography. One is compelled to suppose that the site of the basin was once underlain by a body of magma that was partly or entirely drained to feed the huge volcanoes along the southern side, the combined volume of which exceeds

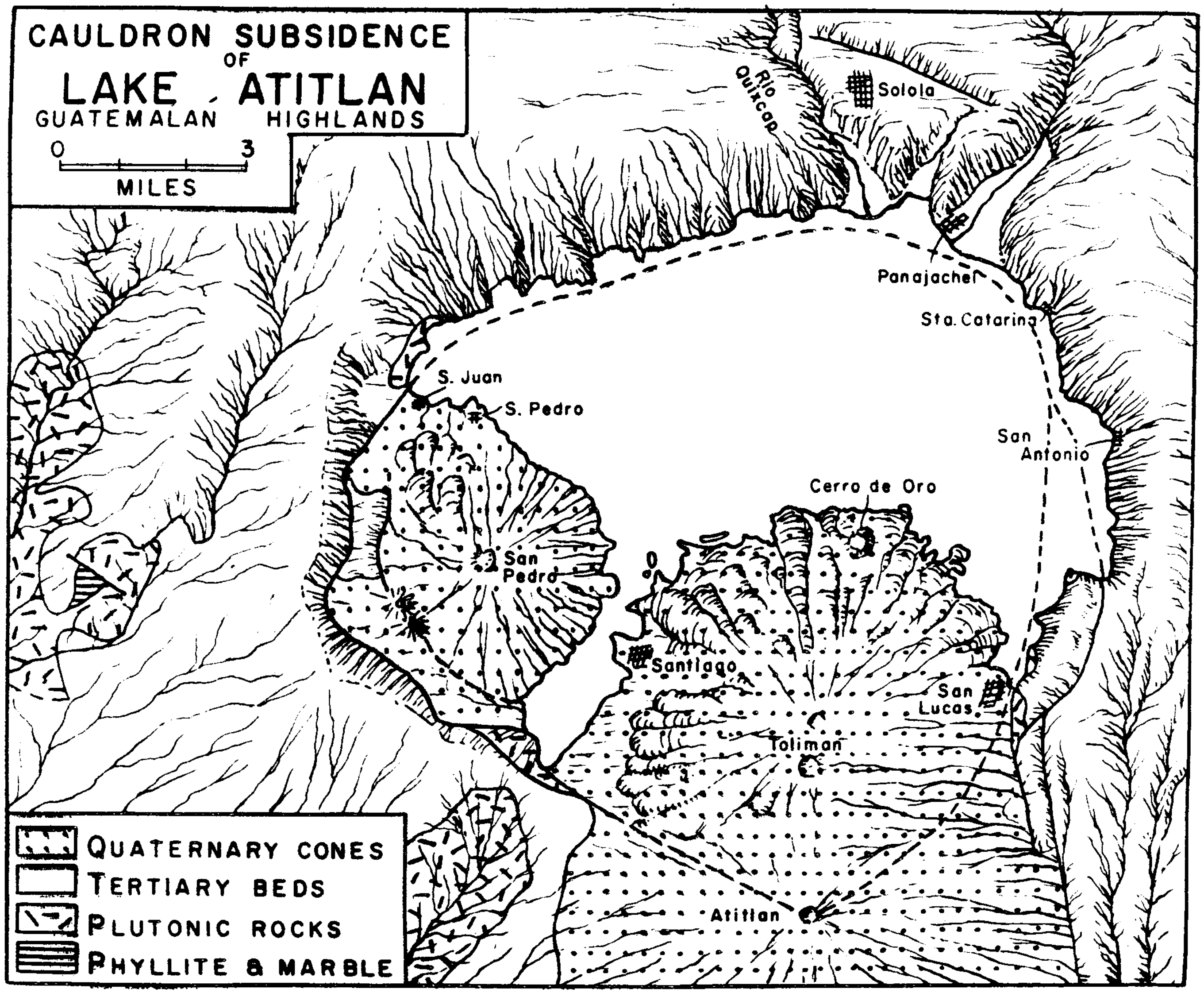
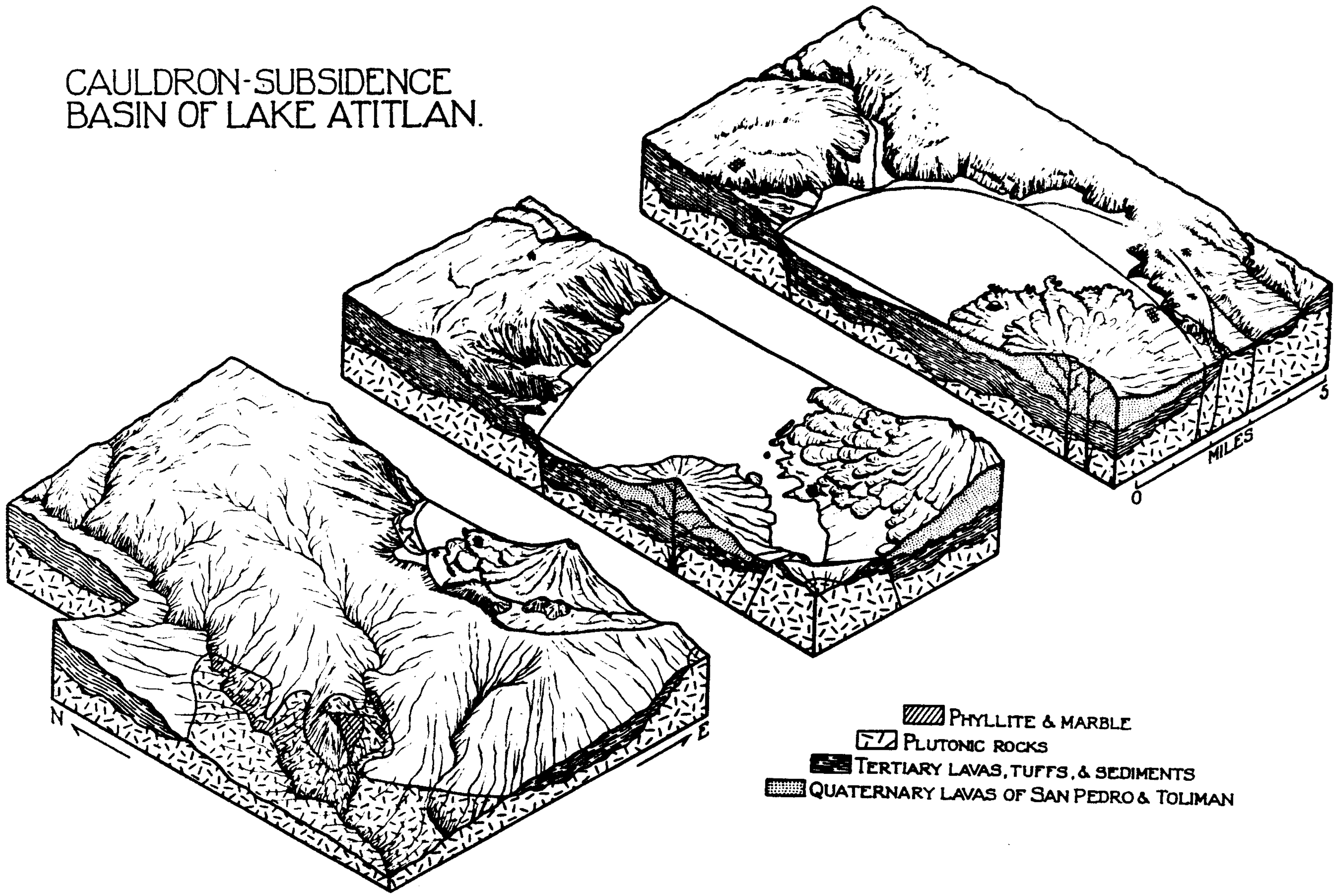


Figure 2. Geologic sketch-map of the area around Lake Atitlan.

CAULDRON-SUBSIDENCE BASIN OF LAKE ATITLAN.



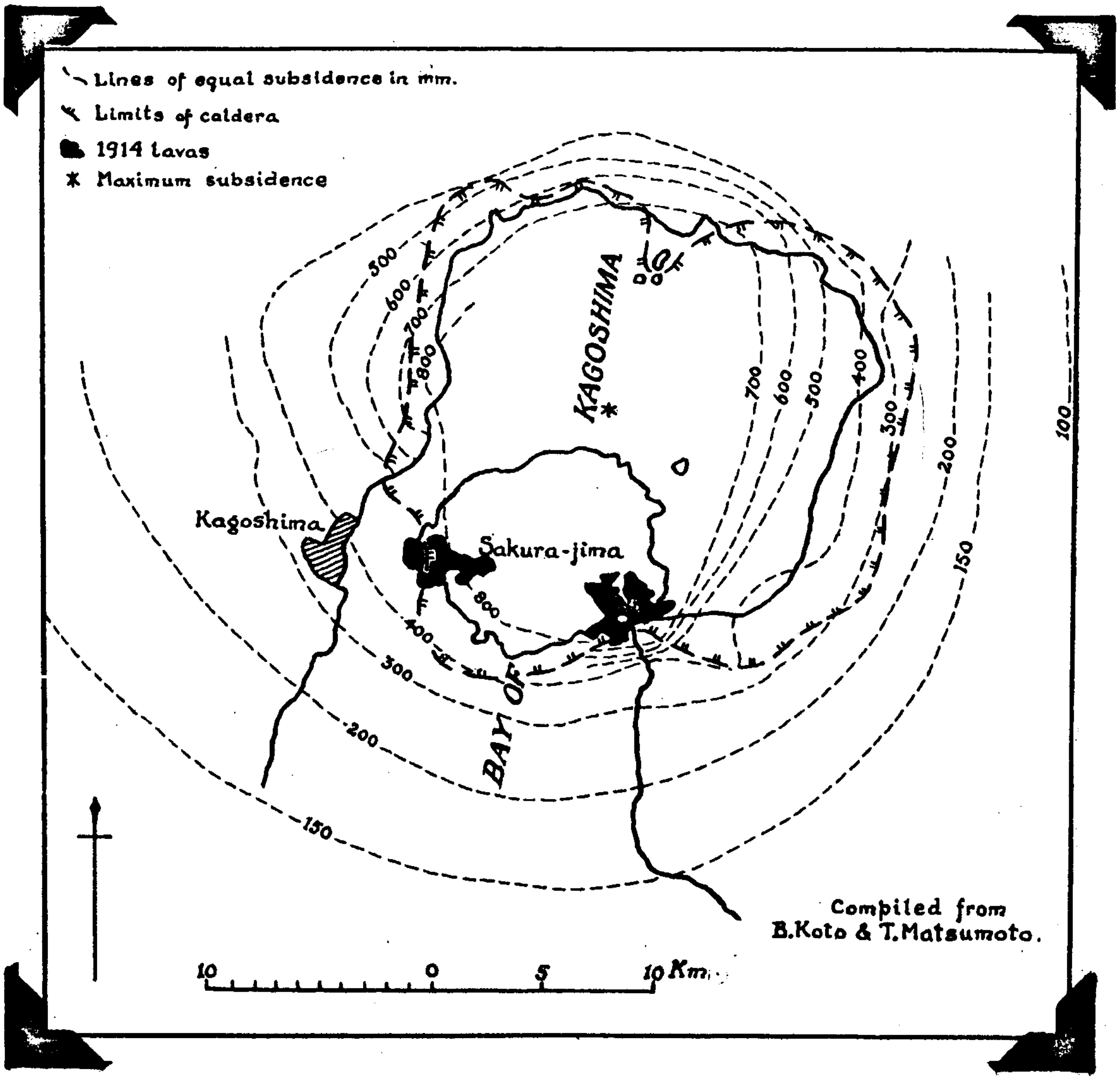
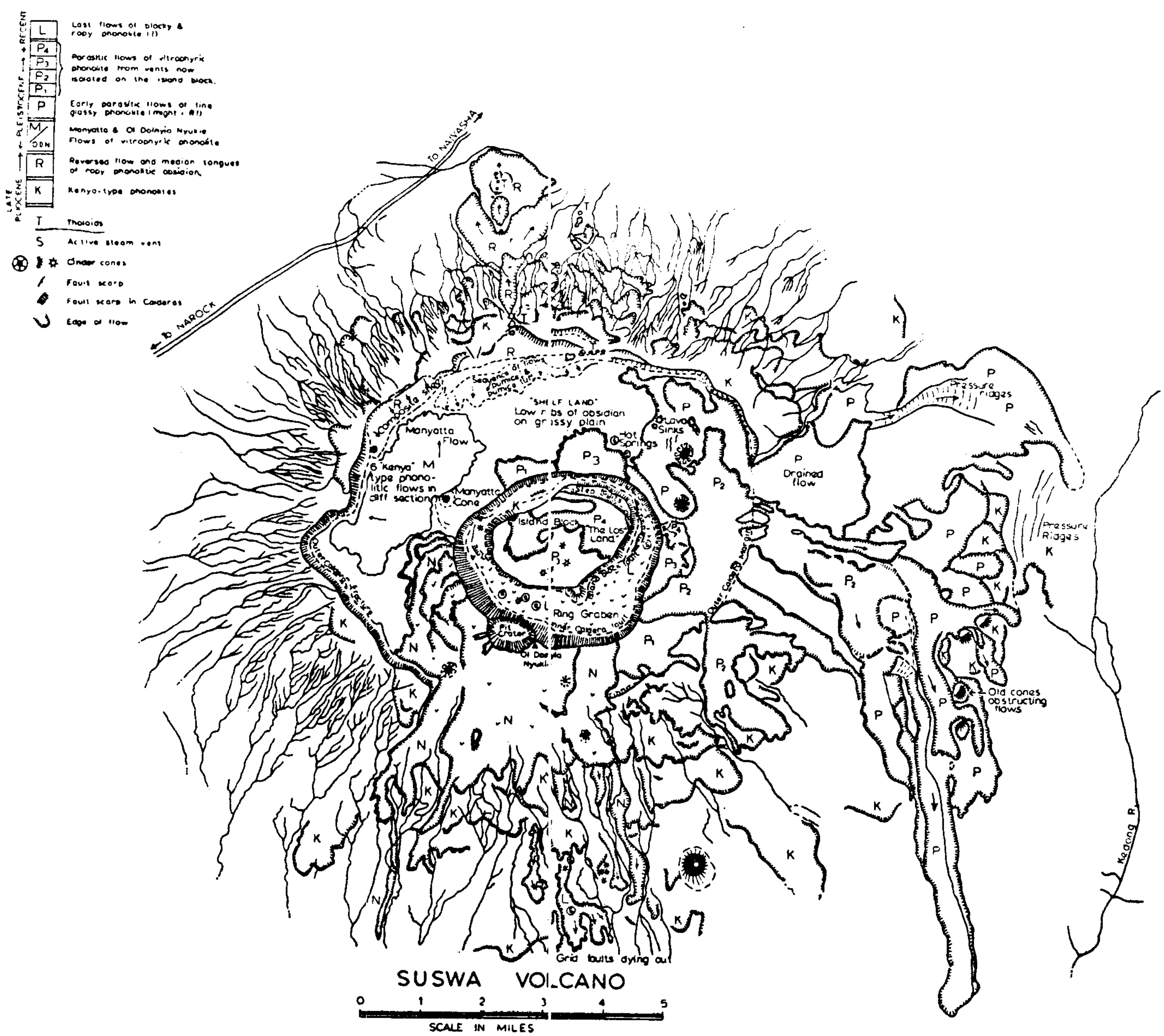


Figure . The Aira caldera and rim volcano of Sakura jima. Shows the lines of equal subsidence produced by the outflow of lava in 1914.



MAP. 1 - General map of the Suswa complex drawn from air-photographs.

Note: The flows marked N are the Ol Doinyio Nyukie Flows (ODN in the Key). The flows marked P are the earliest parasitic flows of the eastern side of the volcano. These may include some early flows of the parasitic eruptions of the second sequence (post-Ol Doinyio Nyukie) together with fine, glassy flows of the terminal phase of the first sequence. More mapping is needed to unravel this complex of tongues.

120 cubic kilometers. It was probably this subterranean migration of magma that caused cauldron subsidence. Atitlan volcano rises from the rim of the cauldron, and it may be supposed that eruptions of lava from its central vent and from the neighboring Toliman and San Pedro volcanoes caused the southern part of the cauldron to subside still farther, thus rejuvenating and steepening the adjacent fault scarps, just as the floor of the Aira caldera at the head of Kagoshima Bay, Kyusyu, subsided when lavas were erupted from the rim volcano of Sakura-jima in 1914 (figure 17).

Suswa Type (C 2)

1. Suswa Volcano, Kenya. The Suswa caldera (figure 18) is taken as an example of a caldera formed by engulfment of the top of a pre-existing volcano during and prior to mixed eruptions from arcuate fissures on the flanks (McCall, 1963; McCall and Bristow, 1965; Johnson, 1966, and in press). The ancestral volcano was an asymmetric, shield-shaped cone, between 700 and 800 meters high, built on the floor of the Eastern Rift Valley of Kenya during Quaternary times by the outpouring of sodalite phonolite lavas. A period of quiet and erosion followed the building of the primitive shield. The summit of the shield then collapsed and eruptions took place through the encircling fractures. Collapse was not the result of the eruptions, as in calderas of Krakatoa type; on the contrary, it was the opening of ring fractures that initiated the eruptions. The first ejecta formed pumiceous lahars, up to 2 meters thick, on the outer flanks of the volcano; then followed much more voluminous deposits of well-stratified and well-sorted pumice, products of airborne showers, locally as much as 60 meters thick. Most of the pumice fell on the outer slopes of the volcano, but much of it accumulated against the caldera wall. Effusions of viscous lava alternated with the explosive eruptions. Many flows spread on

to the upper, outer slopes, but some poured down the steep caldera walls where they were interbedded concordantly between sheets of inward-dipping pumice.

The oval caldera measures roughly 8 by 12 kilometers across and its walls reach a maximum height of 200 meters, but how its volume compares with that of the magma discharged by the mixed eruptions from the encircling fractures is not known.

Many lavas then issued from fissures on the floor of the caldera and a large cone was built in the southwest part. Subsequently, a remarkable annular collapse took place on the floor, producing a ring graben, between 500 and 1,500 meters wide, with precipitous walls, locally more than 200 meters high. Within this steep-walled trough lies an oval, flat-topped "island block", 2.5 by 3.75 kilometers across, most of it lying a little below the level of the top of the surrounding, inward-facing fault scarp. The surface of the "island block" and the rim of the graben consist of intra-caldera lavas. Minor explosive eruptions occurred within the graben, but these contributed nothing to its formation. The prime cause of the graben was probably the rise and subsequent withdrawal of magma from an underlying ring fissure; in other words, the graben is the surface expression of a ring dike that failed to break through its roof. Fumarolic activity still continues within it.

The Menengai caldera, which lies about 150 kilometers north of Suswa, also within the Eastern Rift Valley of Kenya, was formed about 10,000 years ago (McCall, 1957). When first described, it was considered to be a caldera of Krakatoa type, but it is now believed that collapse took place prior to and during explosive eruptions of pumice from marginal ring fractures. Here, however, the explosive eruptions were not accompanied by lava flows as they were at Suswa.

2. Medicine Lake caldera, northern California. A shield volcano, some 33 kilometers wide and 800 meters high, was built in the Medicine Lake Highland during Plio-Pleistocene times by quiet effusions of olivine andesite (Anderson, 1941). Its summit then collapsed along elliptical fractures to produce a caldera having a volume of about 8 cubic kilometers, measuring approximately 6.5 by 11 kilometers across and at least 150 meters deep. Andesitic magma rose up the surrounding fractures as the caldera sank, building rim volcanoes that obliterated the caldera walls and poured lavas down the outer slopes of the shield. Eruptions from the rim volcanoes gradually became more localized, and in addition to andesites they discharged rhyolite and dacite, partly as pumice and partly as lava flows. Finally, during Recent time, a sheet of dacite poured on to the caldera floor from a vent near the center, and copious quantities of rhyolitic pumice were blown from vents a short distance beyond the rim, to be followed by large flows of rhyolitic obsidian. At about the same time, many flows of basalt poured over the lower slopes of the original shield volcano and numerous cinder cones were formed.

D. Major volcano-tectonic depressions

Our only concern here is with those large volcano-tectonic depressions associated with voluminous eruptions of siliceous pumice. It should be noted, however, that many of the world's chief rift valleys, including those of East Africa and the Rhine, are also volcano-tectonic in origin. Such rifts are generally related to the rise of alkaline magmas into the crust along linear belts, particularly of basic, ultrabasic, and carbonatite magmas. The rising magmas uparch their roofs, producing linear welts the apices of which collapse to form rifts. Magmas break through to the surface, both inside and outside the rifts, before, during, and mainly after periods of faulting. But consideration of these complex relationships is beyond the intent of the present study, which is to describe and examine the origin of collapse-depressions formed in direct connection with large-scale explosive eruptions of siliceous magmas.

1. The Toba depression, north Sumatra. Much credit is due to van Bemmelen for his early recognition of how the huge basin that holds Lake Toba was formed. Already, in 1929, he realized that the basin must have been formed by collapse consequent upon the rapid eruption of the enormous volume of pumice which surrounds the lake; and ten years later, though not the first to recognize that part of the lake-floor was upheaved to form the Island of Samosir, he was the first to emphasize the importance of such "resurgent doming", a feature now known to typify many calderas of the Valles type.

The western part of Sumatra is occupied in large part by the lengthy Barisan Range, which was uparched during Pliocene and Early Pleistocene time by the rise of linear plutons, some of which served as feeders to andesitic volcanoes. When the arch reached a critical height, its crest collapsed, forming an apical graben, the Semangka rift zone, analogous in many ways to the Rhenish and African rift valleys already mentioned. Andesitic volcanism continued along and near the Semangka rift zone, in some places down to historic times; locally, however, rhyolitic and dacitic pumice was discharged suddenly and in immense volumes from the tops of domical uplifts or 'tumors' that rose above the general level of the regional arch, as around Lake Ranau in south Sumatra and Lake Toba in north Sumatra.

During Late Pleistocene time, continued rise of the 'Batak tumor', on the present site of Lake Toba, resulted in fracturing of the roof so that the upward-pressing magma was able to issue explosively as rhyolitic pumice (figure 20). During a single paroxysmal phase, approximately 2,000 cubic kilometers of pumice were erupted. The initial eruptions must have been of the high-pressure Vulcanian type, for pumice showered over southern Malaya, 300 to 400 kilometers away, but by far the greater part of the ejecta swept swiftly over the ground as glowing avalanches or pumice flows, inundating 20,000 to 30,000 square kilometers of the Toba region, locally to a thickness of 600 meters. The deposits of these pyroclastic flows are almost wholly unstratified, and they contain conspicuously few fragments of older andesites and 'bedrocks'.

During and immediately following these eruptions, the source-area foundered, producing the Toba depression, the crudely rectangular shape of which was controlled chiefly be pre-existing fractures in the basement. The lake itself is 87 kilometers long; the encircling depression is 100 kilometers long and has a maximum width of 31 kilometers. The amount of subsidence is said to vary from 800 to 2,000 meters, and the original volume of the depression approximates 2,000 cubic kilometers, equalling that of the erupted pumice.

How soon the floor of the great depression began to rise after the collapse cannot be told. It is apparent, however, that the ancestral Lake Toba began to form shortly after the engulfment, so that stratified tuffs and diatomites accumulated on its floor. Rise of magma ultimately lifted part of the floor to produce a "resurgent dome", represented by the Island of Samosir and the adjacent Prapat peninsula (figure 19). Uplifted lacustrine beds on Samosir dip in westerly directions, whereas stratified beds on Prapat peninsula dip in opposite directions. Between them runs a flooded, apical graben, 65 kilometers long and 2 to 8 kilometers wide, comparable to the apical grabens that cut the resurgent domes in the Valles, Creede, and Timber Mountain calderas. Diatomite layers are present on Samosir to a height of 210 meters above the present lake level, but obviously the amount of uplift was considerably greater, reaching a maximum in the region now occupied by the apical graben.

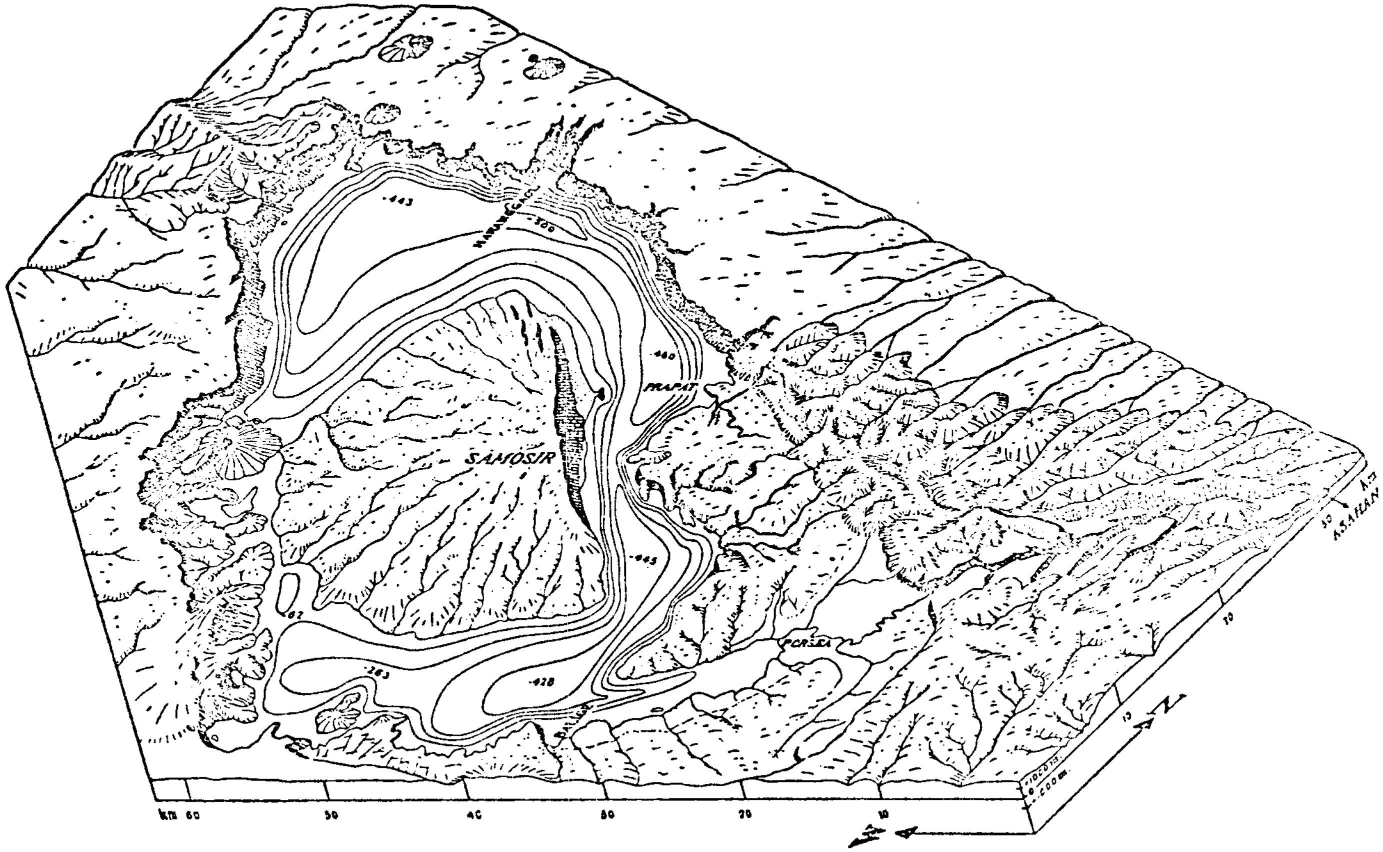


Fig 17.
Isometric block diagram of the Toba cauldron. (Depth figures of the Toba lake in metres).

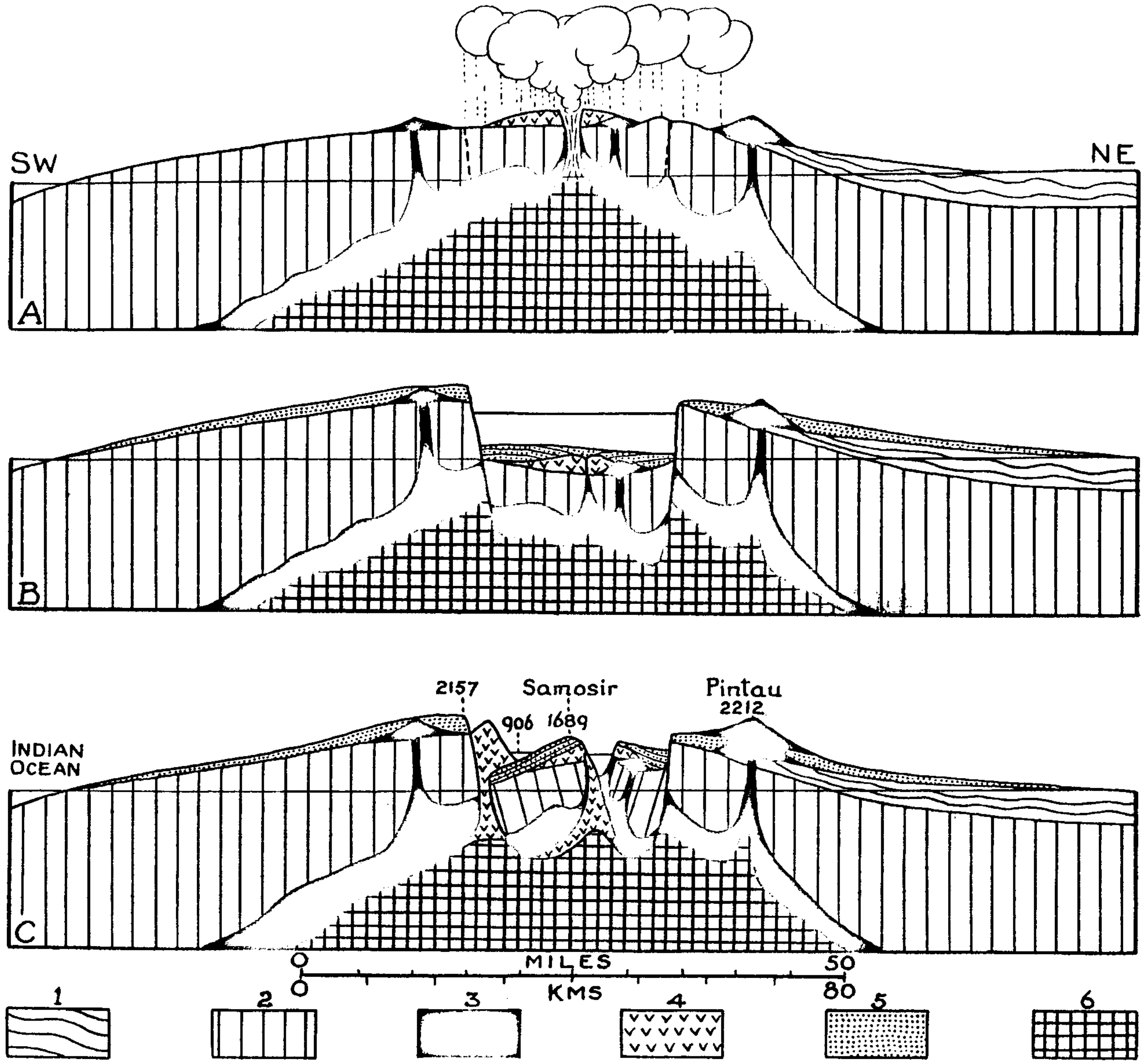


Figure . Origin of Lake Toba, Sumatra – a volcano-tectonic depression.
1, Tertiary sediments; 2, pre-Tertiary crust; 3, basic and intermediate magma and volcanic rocks; 4, acid (siliceous) lavas; 5, acid pumice tuffs (Toba tuff); 6, granitic Toba batholith. For explanation of stages A, B and C, see text. Vertical scale exaggerated 5 times. (Redrawn, with slight modifications, from R.W. van Bemmelen, De Ingenieur in Ned.- Indië, Vol. 4, No. 9, p. 138, 1939). Heights in meters.

During uplift of the floor of Lake Toba -- as in many calderas of the Valles type -- siliceous magma rose into the apical graben and into the moat surrounding the resurgent dome. Several domes of rhyolite were protruded into the graben and dikes of andesite were injected into the walls. A small volcano, formed of siliceous andesite or dacite, also grew between the western shore of the island of Samosir and the adjacent wall of the Toba depression, while other domes of siliceous lava rose near the southern wall. At about the same time, andesitic volcanoes continued to be active outside the depression.

2. The Taupo Volcanic Zone, New Zealand. All of the active volcanoes and almost all of the boiling pools and geysers of New Zealand are concentrated in the North Island, within the north-northwest - trending Taupo volcanic zone, the median strip of the Central Volcanic Region. Within the narrow southern part of the zone lie the large, recently active andesitic volcanoes of the Tongariro National Park, including Ruapehu and Ngauruhoe; within the narrow northern part lie the andesitic cones of Mt. Edgecumbe and White Island. In between, occupying the widest part of the zone and spreading beyond, on to the Kaingaroa and Patatere plateaus, is a vast expanse of rhyolitic ignimbrite and bedded, airfall pumice, associated with many large domical piles of rhyolitic lava. Sporadic occurrences of andesite are present close to the margins of the rhyolitic province; basalts are rare and all are confined to the province, specifically to the three principal rhyolitic eruptive centers (Thompson, 1964).

The Taupo volcanic zone has long been recognized to be a major structural depression. It continues southward into the Wanganui Basin of Late Tertiary and Quaternary marine sedimentation (Gregg, 1961); in the opposite direction it is prolonged through the Kermadec and Tonga Islands, almost to Samoa.

These perfectly aligned links form a volcanic chain crossing the southwest Pacific for a distance of about 2,000 kilometers, bordered for much of the distance by parallel submarine trenches (figure 21).

The Central Volcanic Region of the North Island is a crudely triangular area flanked by southward-converging mountains composed principally of Mesozoic and Paleozoic graywackes (figure 25). Except for the high andesitic cones of the Tongariro National Park, it is essentially a vast, relatively low-lying rhyolitic plateau. To the east, in the rugged Kaimanawa Mountains, the graywackes, the 'basement rocks', rise to elevations of more than 1,500 meters above sea-level; to the west, in lower, discontinuous ranges, some of them rise almost 1,000 meters above sea-level. Yet beneath the Taupo Volcanic Zone, all of these basement rocks lie below sea-level, and generally far below. At the southern end of the zone, under the high andesitic cones, the basement generally lies several hundred meters below sea level, sinking to more than 1,000 meters below sea level under Lake Taupo (figure 24). Still farther north, some of the basement rocks sink to three times that depth. The axis of greatest subsidence coincides approximately with the axis of the Taupo Volcanic Zone, and it therefore seems reasonable to suppose that the subsidence is genetically related to volcanism.

The rhyolitic ignimbrite eruptions of New Zealand, like those of Sumatra, followed a long period of intermittent, shifting uplifts. The major uplift of the western graywacke ranges probably began during Late Miocene times and was completed by the close of Miocene or during Early Pliocene times (Thompson, 1964). Subsequently, during Late Pliocene or Early Pleistocene times, after an interval of stability and erosion, and not long before the first rhyolitic eruptions, there was a minor uplift of a few hundred meters. Uplift of the eastern graywacke ranges generally took place later, beginning in Late Pliocene time and not reaching its maximum until Early or Middle

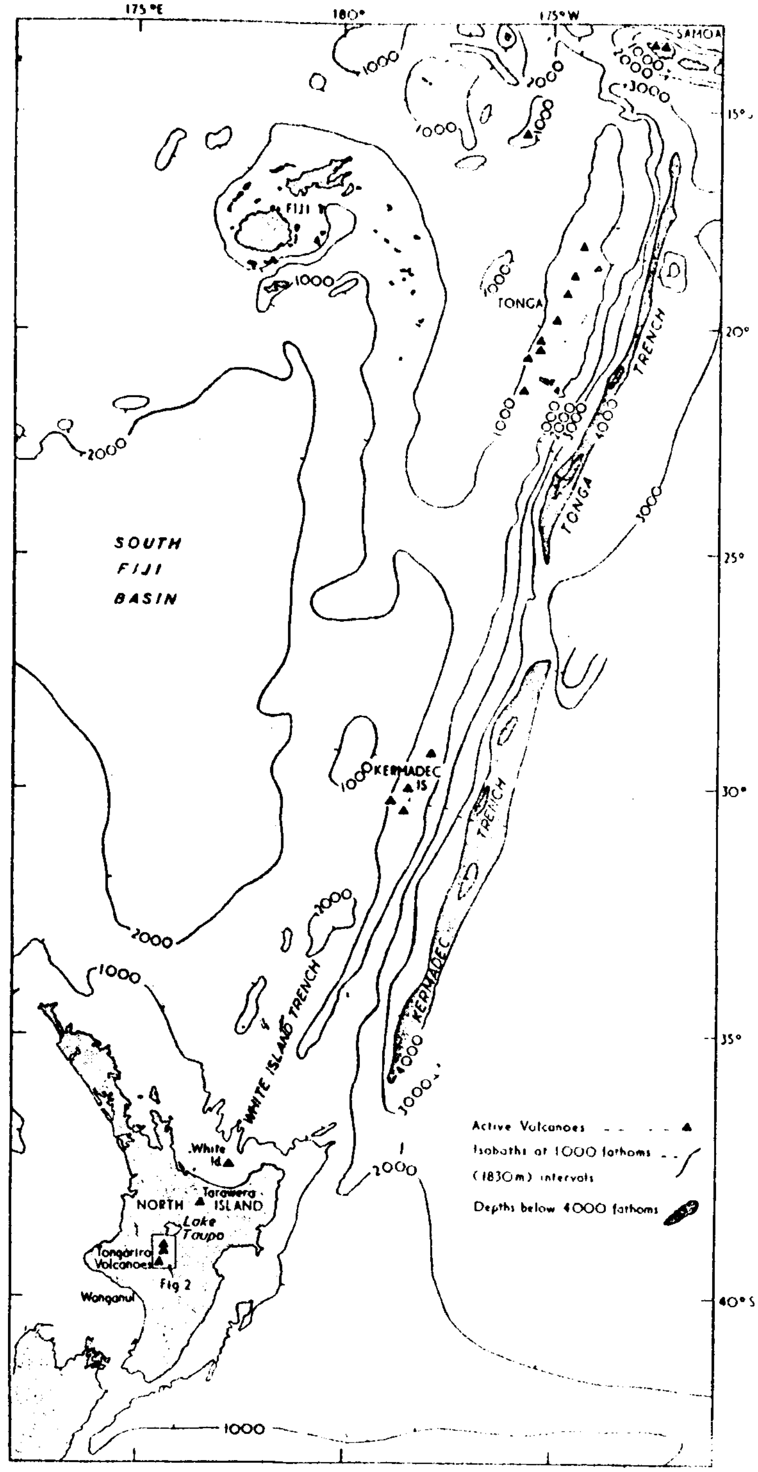

175°E
180°
175°W
SAMOA
FIJI
TONGA
TONGA TRENCH
SOUTH FIJI BASIN
KERMADEC IS
KERMADEC TRENCH
WHITE ISLAND TRENCH
White Id.
Tarawera
NORTH ISLAND
Lake Taupo
Tongariro Volcanoes
Fig 2
Wanganui
15°
20°
25°
30°
35°
40°S
Active Volcanoes
Isobaths at 1000 fathoms (1830m) intervals
Depths below 4000 fathoms

Fig. 1.

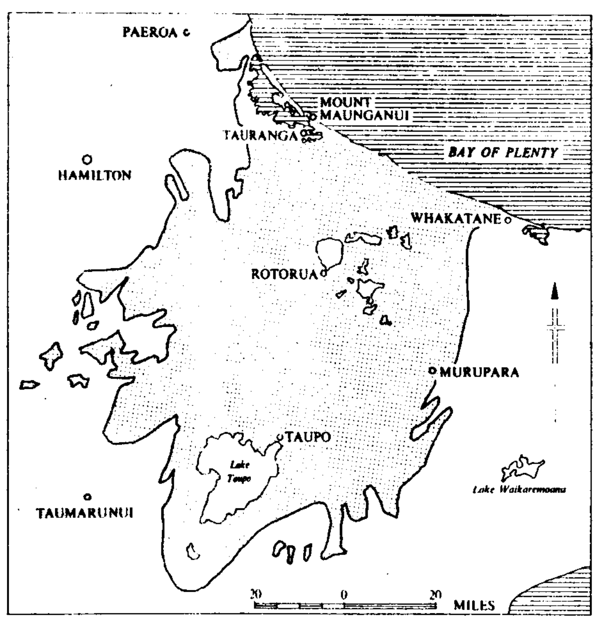


Fig. 2.2. Present extent of ignimbrite in the central North Island (including areas buried by younger deposits).

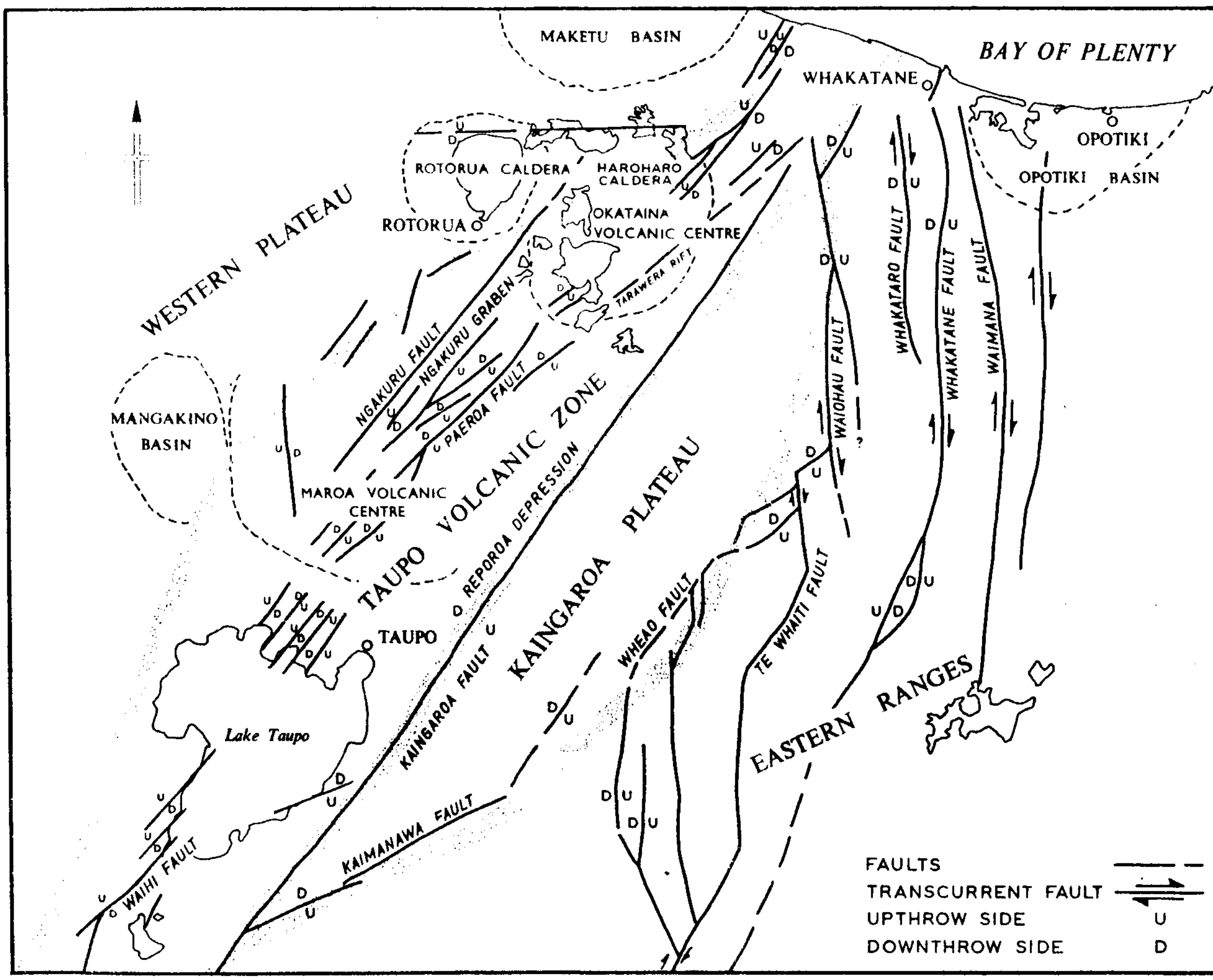


Fig. 2.3. Main structural features of part of the Bay of Plenty region.

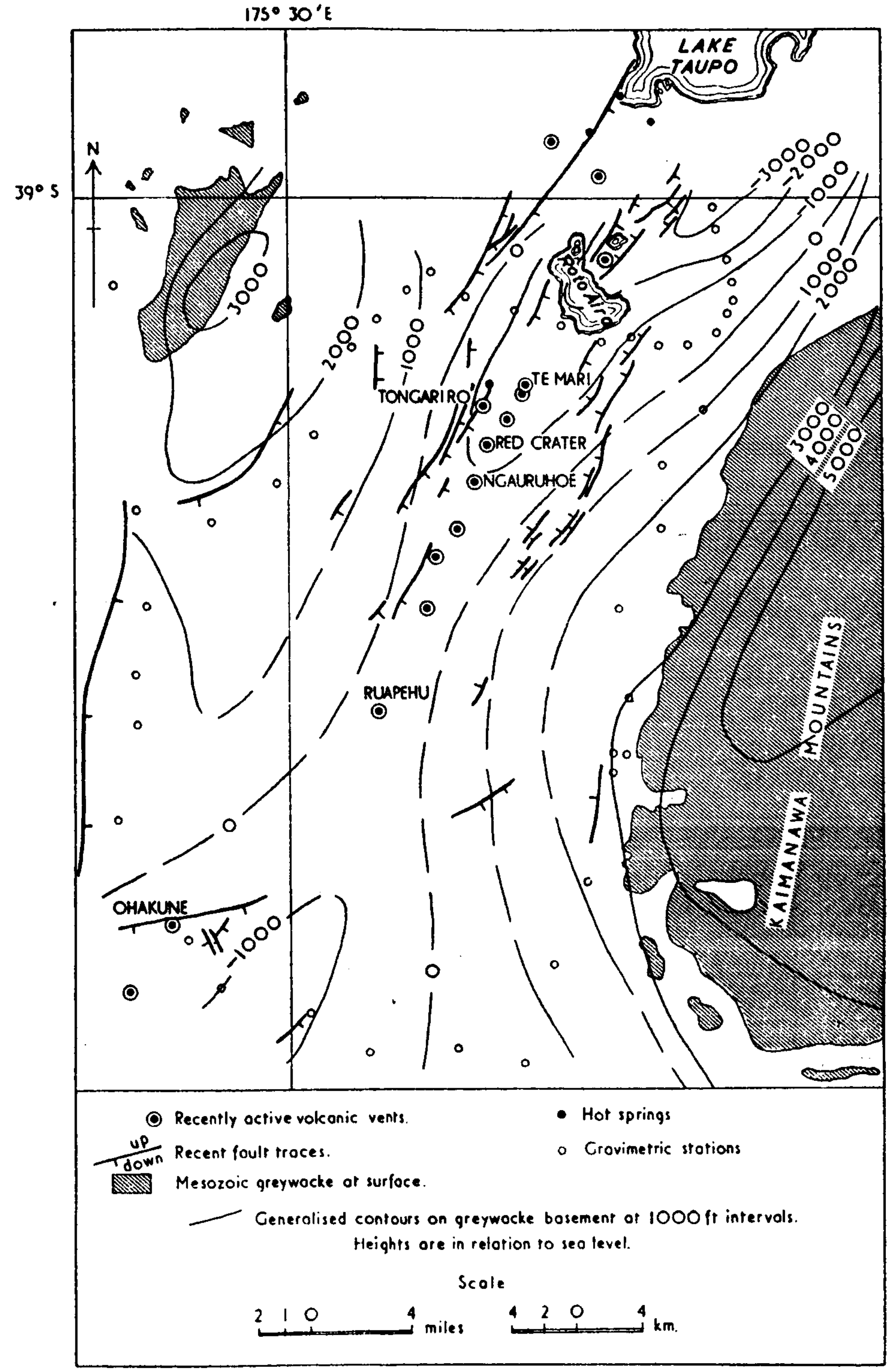

175° 30'E
39° S
N
LAKE TAUPO
TONGARIRO
TE MARI
RED CRATER
NGAURUHOE
RUAPEHU
OHAKUNE
KAIMANAWA MOUNTAINS
3000
2000
1000
0
4000
5000
Recently active volcanic vents.
Hot springs
up
down
Recent fault traces.
Gravimetric stations
Mesozoic greywacke at surface.
Generalised contours on greywacke basement at 1000 ft intervals.
Heights are in relation to sea level.
Scale
2 1 0 4 miles
4 2 0 4 km.

Fig. 2.

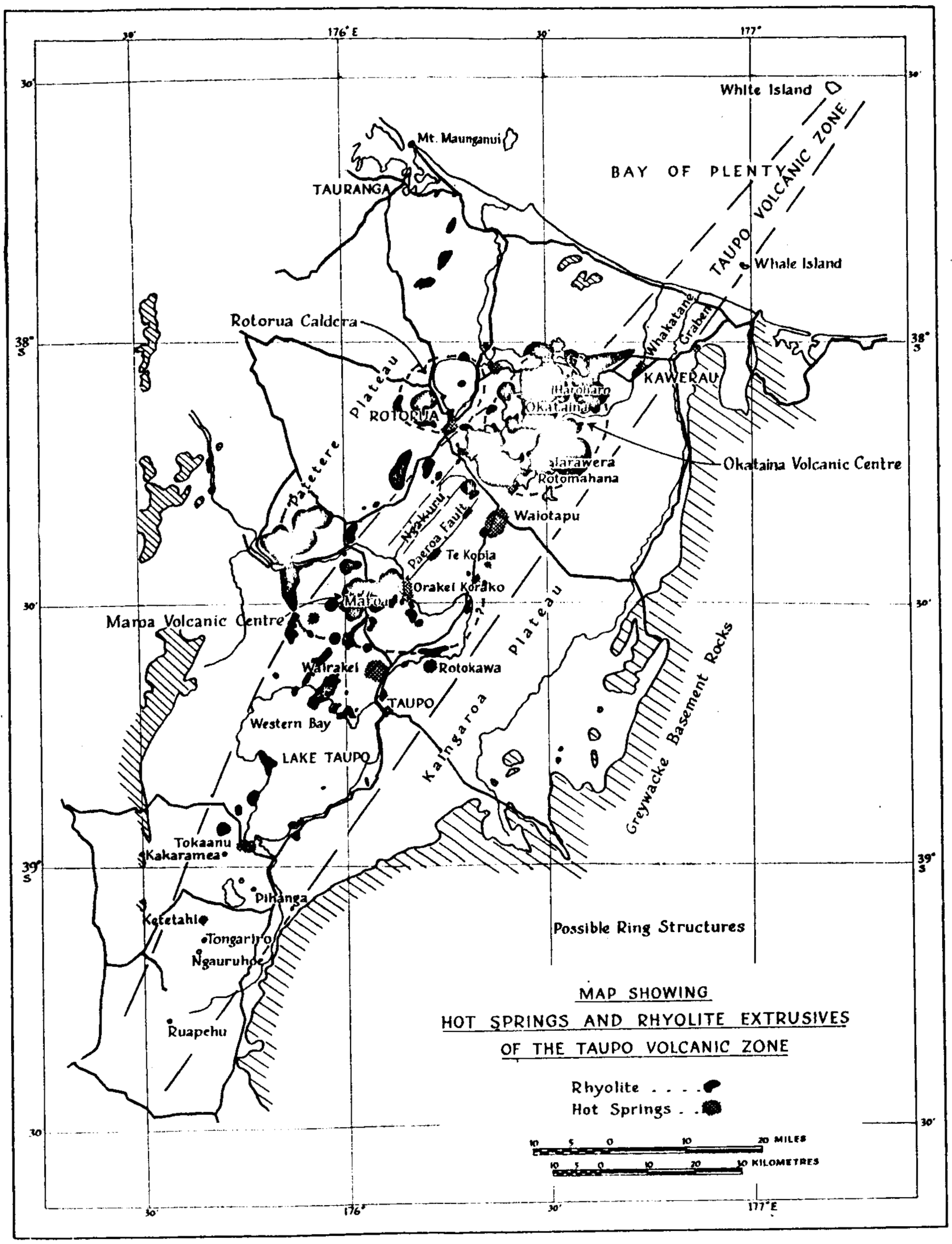


Fig. 1—Map of Taupo Volcanic Zone

Pleistocene time. These uplifts were probably caused by rising bodies of magma, and it was probably the opening of tension fissures on the crests of the uplifts that initiated explosive activity by releasing pressure on the rising magmas. Accordingly, rhyolitic eruptions generally began earlier in the western part of the Central Volcanic Region, and most of the youngest ignimbrite sheets are concentrated in the eastern part.

A shallow tectonic basin seems to have been present before the rhyolitic eruptions began, but most of it must have been above sea level. The present deep trough in the basement was caused by intermittent downwarping and downfaulting that accompanied the discharge of rhyolitic magma. Between Late Pliocene and Late Pleistocene times, pyroclastic flows inundated more than 25,000 square kilometers. Some ignimbrite sheets exceed 200 cubic kilometers in volume; one sheet may have twice that volume. Together they make up approximately 15,000 cubic kilometers (Healy, 1962).

The Kaingaroa Plateau is a huge segment of the Kaimanawa graywacke-range that sank and slid westward, toward the central part of the Taupo Volcanic Zone, where removal of magmatic support by eruptions was greatest. The plateau is, in a sense, a gigantic landslide block that left an arcuate embayment, more than 150 kilometers across, on the western flank of the Kaimanawa Mountains, and it owes its subdued relief to burial by some of the last sheets of ignimbrite related to the Okataina volcanic center.

Countless north-northeast - trending normal faults traverse the Taupo Volcanic Zone, some of them separating it from the adjacent ignimbrite plateaus. Faulting began soon after the first ignimbrites were discharged, and it has continued intermittently ever since. Abundant narrow horsts and grabens characterize the landscape, and most of the hot spring areas are localized on faults (figure 23).

In addition to the linear faults, there are more or less circular depressions bordered by ring fractures. These depressions mark the source-areas of most of the ignimbrites, and each tends to be associated with a distinctive type of ignimbrite (Healy, 1962). All but one of them lies within the Taupo Volcanic Zone; the exception, Rotorua caldera, is tangent to its western margin.

Lake Rotorua lies in the bottom of a circular basin, about 16 kilometers across, formed during Middle and Late Pleistocene times by sinking of part of the western ignimbrite plateau to a depth of more than 400 meters. During and after the subsidence, effusions of viscous rhyolite lavas produced a series of large domes, first along the rim and then on the floor of the caldera, one dome being sliced in half by the collapse.

The largest of the circular depressions is the Okataina volcanic center, which measures approximately 33 kilometers across. It was formed after the Rotorua caldera, during Late Pleistocene times by a subsidence of more than 300 meters. Most of the depression was then filled by a group of coalescing rhyolitic domes, the largest of which forms Mt. Haroharo and Mt. Tarawera. Explosive eruptions of rhyolitic pumice preceded the rise of most, if not all of the domes, and in 1886 explosions of basaltic scoria issued from a rift cutting the Tarawera domes.

A third eruptive center, Maroa, lies between the Okataina center and Lake Taupo. Many ignimbrite eruptions took place at this center, and after it collapsed a cluster of coalescing rhyolite domes developed on its floor, along north-northeast - trending fractures; other domes rose along the west and southwest parts of the caldera rim, forming the Mokai Ring Complex. A fourth circular depression, the Tokaanu volcanic center, adjoins the southern end of Lake Taupo. Though much smaller than the others, it also contains a group of intra-caldera rhyolitic domes, and, like the

Haroharo caldera within the Okataina center, it is associated with Quaternary basalts.

Copious eruptions of ignimbrite are also responsible for the downwarping and downfaulting that produced the basin holding Lake Taupo. Indeed its 'Western Bay' is a collapse caldera on the site of an eruptive center. During the last 10,000 years, since the lake was formed, many eruptions of rhyolitic pumice have taken place from vents beneath its eastern part, leaving extensive sheets of mantle-bedded ejecta over the surrounding region. The last of these eruptions, which occurred only about 1,800 years ago, was as violent perhaps as any witnessed by man in historical times (Healy, 1964).

Ample evidence has lately been found to indicate that the rhyolitic ignimbrites of the Taupo Volcanic Zone were derived by melting of the graywacke basement and of some of the underlying granitic and gneissoid rocks; moreover, melting must have taken place at fairly shallow depths to permit the regional downwarping and circular collapses just enumerated. On the other hand, the andesites that form the high volcanoes of the Tongariro National Park seem to have originated at greater depths, perhaps from fractures that penetrate the crust to the underlying mantle (Healy, 1962).

<u>The Basin of Lake Ilopango</u>, <u>El Salvador</u>. A median trough runs east-southeast across the southern part of El Salvador into the Gulf of Fonseca (Williams and Meyer-Abich, 1955). It was formed during Late Pliocene or Early Pleistocene times by foundering of the crest of a broad geanticline crowned by a belt of volcanic cones. Most of the trough has since been filled by growth of large Quaternary volcanoes, composed principally of andesite and basalt. Midway along the trough lies Lake Ilopango, approximately 8 by 12 kilometers across, enclosed by steep walls, from 150 to 500 meters high, consisting mostly of dacite pumice.

A deep graben was already present on the site of the lake before the pumice deposits began to accumulate. Roughly at the same time as a coalescing cluster of Peléan domes of andesite was growing a few kilometers to the west of the original Ilopango basin, domes of dacite began to grow within it and a short distance to the north, accompanied by explosions of pumice. Activity of this kind continued intermittently for a long time until the basin was largely filled, and glowing avalanches of pumice, along with pumice lahars, were able to pour over the northern rim. Wholesale collapse then took place, controlled by the boundary faults of the graben, so that a new basin was produced, almost commensurate with the original one.

The process of filling was then repeated. During early stages, large volumes of pumice were laid down by streams; subsequently, as new domes rose and explosive activity was resumed, pumice accumulated from airborne showers, glowing avalanches and lahars until the pile again grew high enough to permit some of the latest avalanches and lahars to spread across the northern rim. The presence of deeply weathered zones between many of the pumice sheets denotes that the filling process was often interrupted by long intervals of quiet. Finally, a few thousand years ago, during or immediately after the last pumice eruptions, a second collapse took place, forming the present depression. The volume engulfed was between 40 and 60 cubic kilometers; the volume of magma discharged by the last eruptions was trivial by comparison; it appears, therefore, that the prime cause of engulfment was withdrawal of magma at depth.

Since the second collapse, a northwest-trending line of dacite domes has developed on the floor of the depression, the last of them rising near the middle of Lake Ilopango in 1880, its pinnacled top forming the rocky islets known as the Islas Quemadas.

Distribution of Calderas

The distribution of Quaternary calderas is irregular. There are very few, for example, throughout the length of the Andes, and most of those recorded in the literature measure only 3 or 4 km. across, one exception being the caldera of Antuco in Chile. No calderas are known in Panama, Costa Rica, and the 'Neo-Volcanic Zone' of Mexico, and only one, the Qualibou caldera on the island of St. Lucia, has been described from the West Indies. The Cascade Range, which stretches 1,000 km. northward from Lassen Peak in California to Mount Garibaldi in British Columbia, contains scores of large volcanic cones, yet Crater Lake, Oregon, is its only caldera. On the other hand, at least ten calderas are present in Nicaragua, El Salvador, and Guatemala, and they are abundant in the volcanic chains of the Alaska and Kamchatka peninsulas and in the Aleutian Islands. Several small calderas and many cones with somma-rims a few kilometers across are to be seen in the Kurile Islands. Both Hokkaido and Kyusyu are unusually rich in large calderas, but on the intervening island of Honsyu there are only two, namely Towada and Hakone. Calderas are relatively rare in East Africa, despite the enormous size of many of the volcanoes; for instance, the giant volcanoes of Nyamlagira, Niragongo, and Kilimanjaro have steep-walled summit-pits measuring no more than 2 km. across.

Nowhere in the world are large calderas more abundant and closely spaced than in the Hawaiian and Galápagos Islands. Each of the five coalescing basaltic shield volcanoes on Albemarle Island in the Galápagos archipelago has an exceptionally large caldera, and some are of unusual depth (see page 33). Yet in other Pacific archipelagos, e.g. the Society Islands, calderas are conspicuous by their absence. The only caldera in Iceland is that of Askja, a composite volcano built largely of fragmental ejecta of intermediate composition; the basaltic shield volcanoes, more characteristic features of the Icelandic landscape, contain only small summit-sinks, perhaps because each was built rapidly by almost continuous discharge of fluid lava, as Walker has suggested.

Calderas of the Valles type tend to occur in clusters some of which lie within volcanic chains, as do those in Hokkaido and Kyusyu and those in the North Island of New Zealand, whereas others, such as those formed in Mio-Pliocene times in southwest Nevada, seem to be distributed at random, showing no apparent control by regional structures.

Reynolds (1956) has suggested that calderas in orogenic regions, such as Lake Toba in Sumatra, differ from those in kratogenic regions, such as the British ring complexes, in being associated with much more voluminous ignimbrites. This supposed difference, she thought, might reflect different levels of erosion, the Toba depression being of Quaternary age while the British ring complexes are of Tertiary and Devonian age; alternatively, she added, the difference might "represent an initial difference correlatable with the relatively greater thickness of sialic basement rocks, together with a greater head of gas in orogenic zones". We cannot subscribe to these views. The Devonian ring complex of Glencoe contains at least two thick sheets of ignimbrite, and ignimbrites are plentiful in the Permian ring complexes of the Oslo region, Norway, which lie in a kratogenic region. Calderas are notably scarce in the Andes, as we have noted, although the sialic crust in that region is unusually thick; on the other hand, calderas of the Valles type associated with colossal sheets of ignimbrite are especially numerous in southwest Nevada where geophysical studies indicate the sialic crust to be unusually thin. Clearly, many other factors control the distribution and kinds of calderas, among them the depth of the magma chamber, the nature, particularly the strength of its roof, the volume of magma discharged immediately before collapse takes place, and the rate at which the magma chamber is drained at the surface and replenished from below.

Attitudes of Boundary Faults

E. M. Anderson's often quoted studies (1936, 1937) of the dynamics of formation of cone-sheets and ring-dikes are largely responsible for the idea that calderas and cauldron subsidences are surrounded by outward-dipping fractures, his view being that such fractures develop when magmatic pressures are reduced. The idea gained further support from those who maintained that subsidence necessary for caldera formation cannot take place if the encircling fractures dip inward. All have agreed, however, that none of the boundary fractures depart much from the vertical.

Surprisingly few measurements have been made of the dips of ring dikes enclosing cauldron subsidences. Clough, Maufe, and Bailey, in their account of the first such dikes to be recognized, namely those of Glencoe, said that along the north side of the cauldron the boundary fault dips outward at 50° to 70°, whereas on the opposite side it dips inward. Their conclusion was that the cauldron was initially a vertical cylinder that was later tilted. Bailey, in discussing the famous Loch Bá felsitic ring-dike, the prototype of all ring-dikes, which surrounds the younger of the two calderas of Mull, said that there is "a slight suggestion of a general outward inclination ...... but far more evidence is required upon this point". Richey, writing of the ring-dikes of Ardnamurchan, was not sure whether they are vertical, dip inward or outward. All of the ring dikes of New England are essentially vertical, and it has therefore been assumed that space for their emplacement was produced by stoping of blocks between concentric fractures.

A powerful argument in favor of the idea that the boundary faults around calderas dip inward is the widespread 'basining' of intra-caldera beds.

The cauldron subsidence of the Ossipee Mountains, the first of the New England ring complexes to be described, provides an excellent example. Louise Kingsley (1931), in her account of this sub-circular cauldron, which approximates 15 kilometers in diameter, says that the encircling ring-dike of nordmarkite, which varies from about 30 meters to more than 1.5 kilometers in width, is practically vertical. Yet all of the bedded volcanic rocks within the ring dip inward, at diminishing angles toward the center. This 'basining' seemed to her to have resulted from differential sagging of the volcanic rocks between vertical faults or from compression caused by convergence of the ring-faults at depth.

Pillow basalts inside the early caldera of Mull are also 'basined', suggesting long-continued subsidence of flows discharged on to the sinking floor of a lake. And Macdonald says that beds inside the Koolau caldera on Oahu are also basined, presumably because the boundary faults dip inward. The volcanic and sedimentary rocks inside the Glencoe cauldron are not only 'basined', but locally, close to the margins of the cauldron, they are overturned. These relationships convinced Reynolds and Roberts that the boundary faults must dip inward. Roberts went on to argue that if the boundary faults of ring complexes dipped outward, subsidence would have produced far more complete ring dikes and fewer partial rings than are actually observed; however, Modell and other students of Billings, in their work on the ring complexes of New England, have shown that partial rings can be produced by eccentric and sideways sinking of blocks bounded by outward-dipping fractures.

Gravity surveys of Japanese calderas of the Krakatoa and Valles types led Yokoyama to conclude not merely that the boundary faults dip inward but that they do so at very low angles. He was therefore forced to the view that the calderas were not produced by subsidence but by evisceration of volcanic cones by violent explosions, the resultant depressions being largely filled by infall of light, pumiceous ejecta. These conclusions are inconsistent with the relative scarcity fact of lithic ejecta around calderas; nowhere do they approach the volumes of the calderas themselves.

Moreover, gravity measurements are incapable of detecting faults in rocks of uniform densities; consequently, Yokoyama's interpretation may be based on the presence of loose material covering the floors of the calderas rather than on underlying denser roofs of the downfaulted blocks.

Some ring-dikes enclosing cauldron subsidences are known to dip outward. For example, some crescentic dikes surrounding cauldrons in the Sara-Fier district of Nigeria show outward dips of both their inner and outer contacts. And Hills (1958) points out that in the Cerberean caldera of southeast Australia, which measures 16 by 30 kilometers across, the intra-caldera beds dip inward despite the fact that the enclosing ring-dike dips outward at 75° to 80°.

The visible walls of the summit-caldera of the basaltic volcano O-sima, Japan, are precipitous. A boring made less than 10 meters from the base of the wall penetrated more than 160 meters without intersecting the boundary fault.

A potent argument in favor of the idea of inward-dipping fractures is the fact that such fractures must often develop during the doming that usually precedes the climactic explosions which accompany and follow the formation of calderas of the Valles, Glencoe, and Suswa types. Such doming is known to precede the discharge of fluid lavas that accompanies sinking of the floors of Hawaiian calderas. And perhaps an initial doming produces arcuate, inward-dipping fractures on the flanks of some composite volcanoes that subsequently collapse to form calderas of the Krakatoan type.

Krakatoan calderas originate by sudden catastrophic engulfment. So sudden were the collapses that produced the calderas of Krakatoa and Santorin that they generated devastating tsunamis (see pages 10 and 15). Collapses of this kind can hardly take place if the encircling fractures maintain an inward dip at depth (see pages 60-62). On the other hand, almost all other calderas, except for some of Katmai and Suswa type, originate by intermittent and long-continued subsidence, and their development may well be facilitated by repeated widening of the boundary fractures by recurrent doming. Sudden, large-scale collapses are most likely to occur if the boundary fractures dip outward, or if fractures that dip inward close to the surface pass downward into vertical and outward-dipping fractures; they are also likely to occur where subsidence starts near the center of the interior block and spreads quickly toward the margins.

As large volcanoes continue to grow, swelling and subsiding repeatedly in response to the rise and fall of underlying magma, inward-dipping fractures tend to develop during episodes of resurgence whereas outward-dipping fractures tend to develop during episodes of withdrawal. Many volcanoes therefore tend to be bordered by a plexus of arcuate fractures of varying dip, and perhaps this is one of the reasons why calderas of the Krakatoan type evolve suddenly. Calderas of the Glencoe and Valles type and volcano-tectonic depressions of the Toba type never evolve on the sites of long-lived volcanic cones; on the contrary, most of them evolve over bodies of siliceous, anatectic magma that rise intermittently but rarely recede, so that almost all of the arcuate fractures they produce tend to dip inward.

One of the strongest advocates of the idea that calderas are surrounded by inward-dipping fractures was Escher (1929; for this and other references to his papers, see Williams, 1941). Escher appears to have been unduly influenced in this regard by Perret's classic account of the 1906 eruption of Vesuvius, when the central conduit was greatly enlarged by the coring action of uprushing gas. The sequence of events during that great eruption may be briefly summarized. During the initial "luminous liquid lava phase", from April 4 to 8, the central conduit was rapidly drained by outflows of lava at successively lower levels until, as Perret says, the magma-level sank 200 meters below sea-level. At 3:30 a.m. of April 8, the lava phase gave place suddenly to a "gas phase" that lasted 18 hours. During this brief interval, a column of gas some 500 meters wide rose steadily from the mouth of the conduit at a rate of 500 meters a second to a height of 13,000 meters. The uprushing gas abraded the walls of the conduit, greatly increasing their diameter. Then followed a "dark ash phase" that lasted two weeks. During this phase the oversteepened walls of the conduit repeatedly crumbled and collapsed, so that by the time the eruptions came to an end, the conduit had expanded in diameter from 75 to 500 meters and the slope of its walls had increased from approximately 35° to between 65° and 75°.

Escher argued that much larger cylinders might be cored through volcanic cones in this fashion, their size being a function of the velocity of the uprushing gases. He even suggested that cylinders might be cored from depths of as much as 50 kilometers. Once large cylinders were formed, he thought that the adjacent rocks would founder along inward-dipping "sliding planes" to produce calderas. Elsewhere, Williams (1941) has insisted that the volume and character of the lithic ejecta among the pyroclastic deposits surrounding calderas makes this interpretation untenable. Not only is the volume of lithic debris blown out prior to caldera formation vastly too small, but nearly all such debris comes from shallow depths, whereas much of it would have to originate from deep, sub-volcanic levels if Escher's explanation were correct. His gas-coring mechanism was based on a mistaken view of what took place at Vesuvius in 1906. The short-lived "gas phase" of that eruption was, in our opinion, caused by sudden influx of groundwater into the partly evacuated magma chamber following the initial outpouring of lava. Steam-blast eruptions of this kind cannot originate at very great depths such as those visualized by Escher as the explosion-foci from which cylinders are cored to the surface. The subject might be dismissed at this point were it not for Reynolds' revival of Escher's opinions.

Reynolds (1956) writes as follows: "If the sialic crust were perforated by escaping gas-streams it is practically inevitable that pyroclastic material derived therefrom would be radically changed in composition. All that would be necessary for this to happen is that alkalies, and perhaps other constituents, should be transported by the gas .... If added to this the gas had a temperature high enough to melt the fine-grained acid material it transported then the latter would be sufficiently disguised to appear as what Williams calls "new magma" .... What one requires to know is not how much recognizable crustal material was erupted prior to caldera-formation, but whether sufficient acid lava or pyroclastic material, that could have been derived by transfusion of sialic rocks, was erupted". In reply, it must suffice to say that all of the lithic ejecta erupted with the pumice during formation of the caldera of Krakatoa in 1883 consist of unaltered or at most very weakly altered fragments torn from the pre-existing volcanic cones and from underlying Late Cenozoic sediments; nothing in the composition or mineral-content of the associated pumice suggests derivation by 'transfusion' of rocks derived from the deeper, pre-volcanic basement. The same may be said concerning the pyroclastic ejecta around all calderas of the Krakatoa type. Cylindrical conduits cored from depths of tens of kilometers, as envisaged by Escher, would traverse a wide range of rock types; it stretches the imagination too far to suppose that all of the rocks would be so 'transfused' by gases as to yield pumiceous ejecta of almost uniform composition! Our conclusion is that Escher's theory of caldera formation by insliding of the walls of cylindrical conduits cored from great depths by torrents of gas is not supported by field or petrographic evidence and should be discarded.

Possible Role of Fluidization in the Formation of Calderas

Reynolds advocated the view that the ring-fractures surrounding the caldera of Glencoe were not only enlarged by the coring action of rising gases but that some of the gases were hot enough to produce magma from the rocks through which they rose. Some of the magma formed in this way was expelled at the surface as ignimbrite; the remainder congealed in the ring-fractures to form dikes. These views have since been advocated by Roberts (1963, and 1966 a and b). In his opinion, "build-up of gas pressure in depth initiated the Glencoe downward-converging ring-fractures", and release of pressure consequent upon the discharge of ignimbrites from the fractures permitted the central block to subside and produce a caldera. Agreeing with Reynolds, Roberts goes on to say that the ring dikes of Glencoe "in the main, made way for themselves by gas coring rather than by piecemeal stoping". We concur in their opinion and that of W. H. Taubeneck (oral communication) that the famous "flinty crush rocks" that border part of one of the Glencoe ring dikes were produced by a process of fluidization; but Reynolds and Roberts go much further, maintaining that when temperatures were high enough the fluidized system of gas and solid particles gradually changed to a mixture of gas and liquid droplets. "On reaching the surface the droplets would vesiculate to produce shards" and so give rise to the sheets of rhyolitic ignimbrite present within the Glencoe cauldron. Subsequently, as the velocity of the uprushing gases diminished, "the system passed from the 'entrained' to the 'expanded' phase, the droplets coalesced to form the massive porphyrite" of the ring dikes.

Consider first the suggestion that the inward-dipping ring fractures were formed by accumulation of gas-pressure at depth. Gas-pressures adequate to lift the central block at Glencoe, which at the surface measures 20 kilometers across, could not possibly develop in a magma chamber. The maximum pressures measured during historic eruptions from the muzzle-velocities of large ejecta rarely exceed a few hundred atmospheres and never exceed 1,000 atmospheres. Such pressures would be utterly inadequate to lift a block of the size of the Glencoe cauldron and keep open the surrounding fractures. Moreover, no violent, high-pressure explosions accompanied the opening of the ring fractures; had there been, a ring of large explosion-craters would have been formed and vast deposits of lithic debris would have accumulated inside and outside the cauldron. Instead, what one sees at the present level of erosion are massive dikes of porphyrite occupying the ring fractures. It was not any build-up of gas pressures at depth that produced the ring fractures but the rise of a large body of siliceous magma.

Nor can we accept the view that the ignimbrites originated by vesiculation of droplets of magma produced by fluidization. Reynolds says that "the fractures would provide escape-routes for the gas which, if its velocity were appropriate, would transport the melted (transfused) sialic rocks upward as spray (ignimbrite)". The fact is, however, that probably all ignimbrites, despite their high content of gas, are discharged at relatively low pressures. The pumice falls that normally precede the ash- and pumice-flows (ignimbrites) represent the high-pressure products of the initial outbursts, but the ignimbrites themselves issue as foaming clots and spray that rise no more than a few hundred or a few thousand meters above the vents before sweeping downslope as glowing avalanches. If ignimbrites originated in the manner proposed by Reynolds and Roberts one of their characteristic features would be an abundance of lithic fragments showing all stages of liquefaction; observations in the field and with the microscope show, however, that while lithic debris is common in almost all ignimbrites, fragments that show partial melting are conspicuously rare, if indeed any have been identified with certainty. None of the ignimbrite sheets within the Glencoe cauldron exceeds about 2 cubic kilometers in volume; around many calderas, however, particularly around those of the Valles and Toba types, the volumes of the ignimbrites are measurable in hundreds and occasionally in thousands of cubic kilometers. It is far more reasonable to suppose that these great volumes of ignimbrite-magma were produced, not by the fluidizing action of rising gases, but by slow melting of basement rocks as a result of increasing temperatures, and that the anatectic melts then worked their way upward gradually until they reached levels where they could dome the overlying rocks to open conical fractures, permitting the melts to vesiculate and so escape at the surface as incandescent droplets and spray.

Equally untenable is Roberts' view that as gas-pressures diminished and the velocity of the gases rising up the ring fractures was reduced, the liquid droplets were no longer expelled to form ignimbrites but fell back on each other to accumulate and anneal within the fracture-zones to produce the porphyrite ring dikes. Nothing in the texture nor in the structure of the porphyrites suggests such an origin; on the contrary the fluidal banding parallel to the contacts and the chilled margins suggest that the porphyrites are 'normal' magmatic intrusions heavily charged with phenocrysts. Similar ring dikes around other calderas-- e.g. around the Early Caldera of southeast Mull -- were emplaced so forcefully as to produce concentric folding in the surrounding rocks; surely no such deformation could result from dikes produced by fluidization. The walls of most diatremes admittedly formed by fluidization rarely show any signs of deformation.

Manner of Collapse

Locke (1926; for references, see Williams, 1941) pointed out that rapid subsidences in mines cause the overlying ground to settle en bloc, whereas slow subsidences produce many small adjustments so that the roof-rocks are broken into a jumble of fragments as they founder. The size of the collapsing body must also be taken into account. Small pipes, 30 meters or so across, formed by mineralization stoping, tend to be jumbled throughout, whereas large pipes, of the order of say 300 meters in diameter, may drop as cylinders, showing only minor fracturing and no rotation of the fragments.

Wisser's penetrating analysis (1927) of subsidences in the Bisbee mining district, Arizona, caused as underground ore-bodies shrank through oxidation, indicate that domical fractures first develop above the ore, allowing successive shells to split from the roof. When the heap of spalls piles up until it can support most of the weight of the roof, no more domical cracks develop; instead, irregular cracks develop and these gradually extend upward, shattering the rocks into blocks. As the blocks slip downward, they retain their relative positions, and eventually the whole shattered mass slumps along nearly vertical boundary fissures. In this way, a broken mass may founder between massive, undisturbed walls. The boundary fissures generally steepen at depth, and in their lower parts they dip outward.

Somewhat similar processes may be involved during the emptying of a magma chamber by rapid and voluminous eruptions of ignimbrite. Domical fractures probably develop in the roof of the reservoir as the level of the magma is lowered; indeed, diminution of magmatic pressure may well cause parts of the roof to 'backfire' into the reservoir. Domical cracks would then extend upward, shattering the roof into a mass of blocks. Engulfment would tend to begin near the center, where the roof was thinnest and weakest, especially if pre-existing conduits rose from the top of the reservoir, for insliding of the conduit walls would hasten collapse. Stoping and spalling of blocks from the upper part of the reservoir roof would also be facilitated by the rise of fluid, gas-rich tongues of magma, as the floor of the crater of Vesuvius was undermined and collapsed in 1913 when, after a 7-year period of quiet, tongues of fluid lava rose in the central conduit.

Early infall of the central part of the roof of the reservoir, coupled with insliding of the walls of central conduits may account for the absence of central peaks on the floors of calderas formed on the sites of pre-existing cones, i.e. inside calderas of the Krakatoa, Katmai, and Suswa types? No such peaks are to be expected on the floors of calderas of the Valles and Glencoe types, nor on the floors of volcano-tectonic depressions of the Toba type, since these basins do not originate on the sites of former volcanic cones.

Tendencies to collapse must be accentuated if the magma chamber is extensive relative to the caldera, if it lies close to the surface, if its roof is gently rather than steeply arched, and if it is already punctured by many conduits. At Santorin, as we have noted (page 13) the present site of the caldera was once occupied by a cluster of overlapping cones, each built over a cupola rising from a common magma chamber. When explosive eruptions of pumice took place, the cupolas were presumably the first to be drained of magma so that their roofs collapsed, ultimately to form the semi-circular bays that now indent the caldera walls; only when enough magma had been drained from the main chamber did the principal collapse take place to form the major part of the caldera.

Precisely how collapse takes place must depend to a considerable extent on the nature and attitudes of the roof-rocks above the reservoir. If a pre-existing volcanic cone occupies the site of the incipient caldera, most of the lava flows and interbedded fragmental layers will dip outward, and hence most of the joints, being perpendicular to them, will dip inward at high angles, providing planes of weakness that facilitate engulfment. Wisser (1927) observed in the Bisbee mining district that where thick-bedded limestones overlay shrinking ore-bodies, the marginal fractures of the subsided areas were well-defined, but where the overlying rocks were thin-bedded limestones, the marginal fractures were poorly defined and impersistent, forming indefinite zones rather than clean-cut breaks. Similar differences are to be expected in the walls of calderas. If the initial ring-fractures cut thick, massive lavas, with few or no interbeds of loose fragmental ejecta, the caldera walls are apt to be steep and well-defined, as they are, for example, in the Masaya and Kilauea calderas; if, on the other hand, the ring-fractures cut thin lava flows separated by many fragmental interbeds, the caldera walls are apt to be much less steep and to be somewhat irregular, as they are, for example, around Crater Lake. If the ring-fractures only cut unconsolidated fragmental layers, the caldera walls are apt to be gently inclined and to be modified by irregular benches and minor slump-blocks. Finally, if collapse takes place slowly and intermittently among unconsolidated or poorly consolidated ejecta, the resultant caldera is apt to be a saucer-shaped <u>conca</u> like the Pilomasin Basin in Sumatra, the Conca di Bolsena in Italy, or Mono Lake in California.

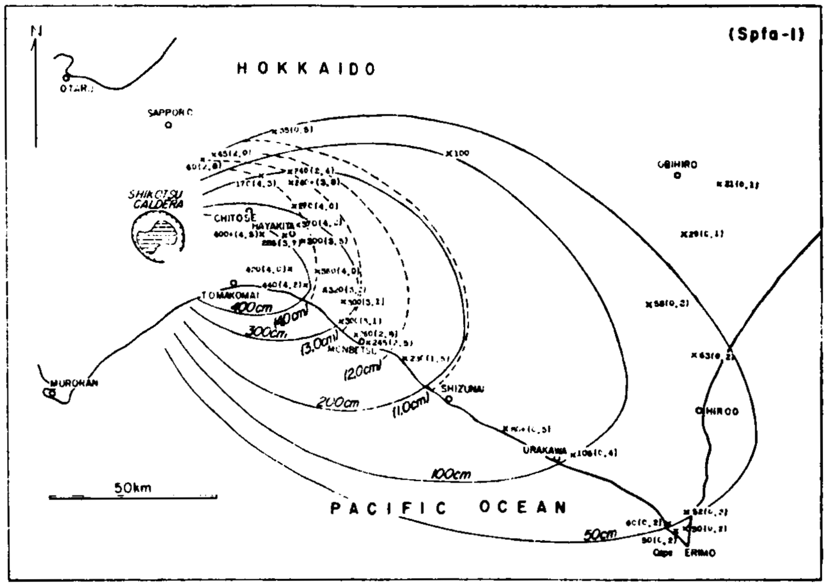
Lithologic variations among pyroclastic ejecta

Our concern here is with the products of explosive eruptions associated with calderas of the Krakatoa and Katmai types on the one hand, and with those accompanying calderas of the Valles type and volcano-tectonic depressions of the Toba type on the other.

In virtually all calderas of the Krakatoan type, activity begins with high-pressure eruptions that hurl lithic and pumiceous ejecta far above the vents to fall in showers on hill and valley alike. Consequently, the products of the initial eruptions are likely to form well-stratified sheets that mantle the topography. And because the ejecta are blown high into the atmosphere and dispersed by winds for long distances, they tend to be fairly well-sorted as to weight and size of constituents. Moreover, an 'eolian differentiation' takes place. Within any given size-range, heavy lithic fragments tend to fall first, then mafic minerals, followed by felsic minerals, and finally by fragments of pumiceous glass. And in general the sizes of the lithic and pumiceous fragments diminish with increasing distance from the source. But obviously many factors tend to modify these relationships. Much depends on the shapes of the pumice fragments, on their content of crystals, and the degree of vesicularity of the glass; much also depends on variations in the velocity and direction of the winds at different levels. More detailed analyses of the characters of deposits left by pumice falls have been presented by Williams (1941) and Katsui (1963). Katsui's diagrams illustrating the distribution and the variations in thickness and grain-size of the pumice-falls related to the calderas of Shikotsu and Mashu, Hokkaido, are reproduced here for convenience (figure 26-28).

In almost all explosive eruptions that lead to formation of calderas of the Krakatoan type, the initial pumice-falls are followed by less violent but more voluminous discharge of foaming magma, the bulk of which pours from the vents as a mass of incandescent clots and spray giving off abundant gas at high temperature. The turbulent mixture then races downslope in the form of glowing avalanches, the deposits of which are spoken of as ash flows, pumice flows, and scoria flows, depending on the coarseness and composition of the materials, or are referred to collectively as ignimbrites.

Because the materials constituting ignimbrites are transported in a turbulent manner and are deposited rapidly, their lithologic characters differ notably from those of pumice-falls. In particular, the individual



full line : thickness contour of the deposit.
broken line: contour of arerage of maximum grain size of pumice.
x : observed point showing thickness and grain size in cm.
Fig. 3. Distribution of the Shikotsu pumice-fall deposit (Spfa-1).

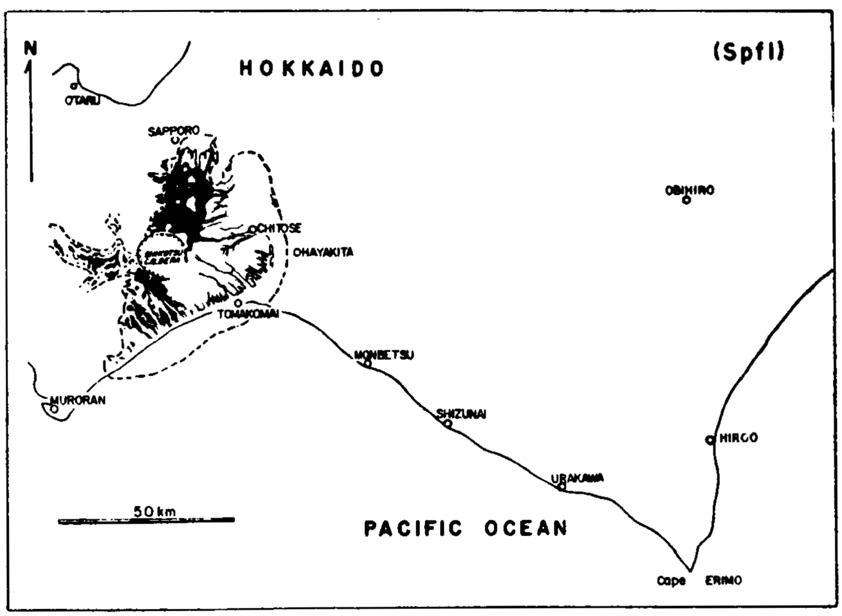
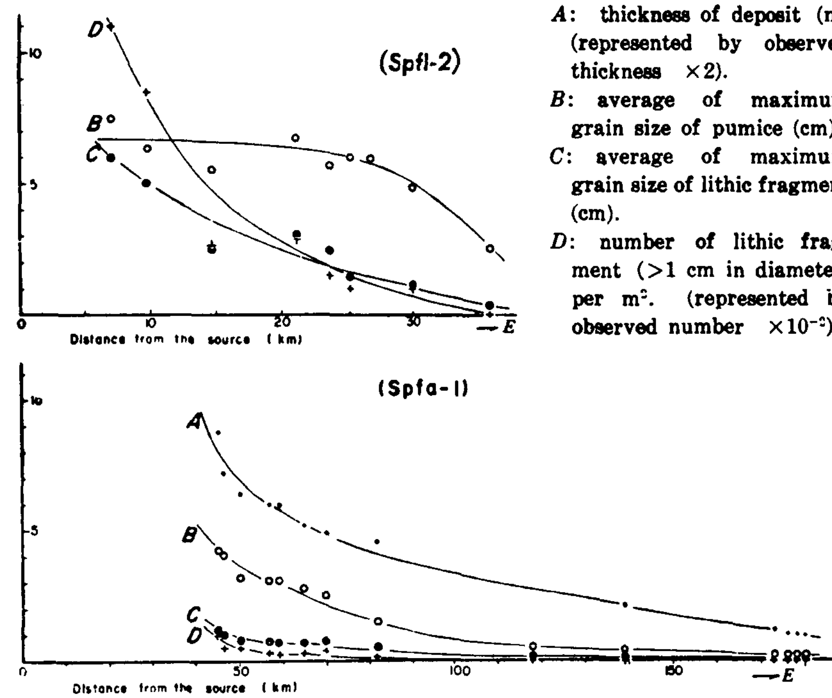


Fig. 4. Distribution of the Shikotsu pumice-flow deposit (Spfl-1 and 2). **after** DOI and OSANAI (1956)

Figure 26. Pumice-fall and pumice-flow deposits around the Shikotsu caldera, Hokkaido. (After Y. Katsui, 1963).



A: thickness of deposit (m) (represented by observed thickness ×2).
B: average of maximum grain size of pumice (cm).
C: average of maximum grain size of lithic fragment (cm).
D: number of lithic fragment (>1 cm in diameter) per m^2. (represented by observed number ×10^{-2}).

Fig. 5. Variation of grain size of the Shikotsu pumice-fall deposit (Spfa-1) and the Shikotsu pumice-flow deposit (Spfl-2).

Figure 27. Characters of the pumice-fall deposits around the Shikotsu caldera, Hokkaido. (After Y. Katsui, 1963).

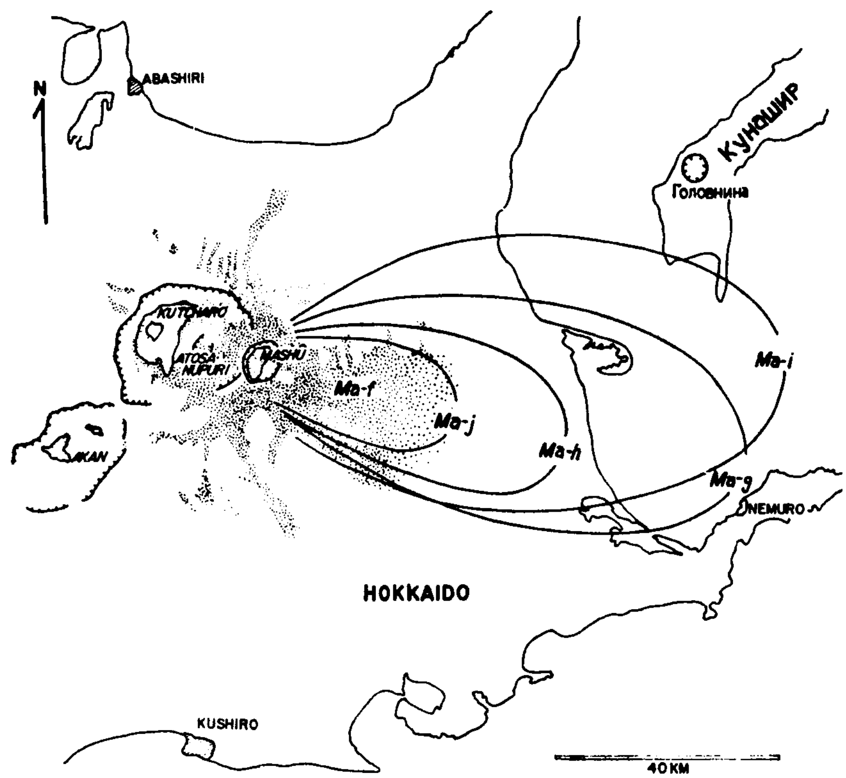


Fig. 7. Map showing the distribution of the Mashû pyroclastic deposits erupted during the culminating phase.
Ash-fall and pumice-fall deposits, Ma-j, i, h and g, are represented by 10cm thickness contour lines; and pumice-flow deposit, Ma-f, is shown by dotted area.

Figure 28. Pyroclastic deposits around the Mashu caldera, Hokkaido. (After Y. Katsui, 1963).

sheets are marked by an almost complete lack of stratification and by being very poorly sorted, lapilli, bombs, and blocks generally being scattered at random in an almost flourlike matrix of pumice dust. A crude sorting may be seen occasionally in the distal parts of ignimbrite sheets, and sometimes, owing to differential rates of advance, straight and sinuous trains and lenses of lithic fragments and pumice bombs become concentrated within the finer materials.

Some deposits of pyroclastic flows related to Krakatoan calderas exhibit crude columnar jointing and various degrees of welding, the pumiceous and scoriaceous fragments being annealed and flattened while still hot; such features are developed on a vastly larger scale among the pyroclastic flows related to calderas of the Valles type. Similarly, while products of vapor-phase crystallization and fumarolic sublimates are not uncommon among the pyroclastic flows related to calderas of the Krakatoan type, they are far more widespread among those related to calderas of the Valles type.

It is only close to the eruptive vents that any difficulty arises in distinguishing between the products of pumice-falls and pumice-flows. Ejecta may fall so rapidly and in such large volume close to the source that the deposits may be completely unstratified and unsorted; indeed they may also retain sufficient heat to produce incipient welding of the glassy constituents and generate secondary fumaroles.

The voluminous ignimbrites related to calderas of the Valles type and to volcano-tectonic depressions such as Lake Toba and the Taupo Volcanic Zone of New Zealand exhibit many lithologic variations, both vertically and laterally with respect to the source. Vertical variations in crystal-content and chemical composition are discussed on a later page; here we note that variations are to be seen in the degree of welding of the deposits, not only in vertical sections but within individual sheets and compound cooling units as they are traced away from the eruptive vents. Particularly thorough studies by Ross and Bailey at the Valles caldera, New Mexico, by Ratté and Steven at the Creede caldera, Colorado, and by other members of the U. S. Geological Survey around the calderas of southwest Nevada have provided a wealth of information on these variations. Close to the calderas, where the deposits accumulate rapidly and almost continuously at high temperatures, it is difficult or impossible to distinguish individual sheets of ignimbrite or to recognize compound cooling units, since all gradations can be seen between them. Farther from the vents, however, "the members change laterally into less welded and less compacted facies, and alternating zones of densely welded and partially welded tuff mark the change from a simple cooling unit to compound cooling units; also near the margins of the deposit, the ash flows become interlayered with volcanic rocks from other sources .... These features indicate episodic accumulation near the margins and possibly reflect variations in the intensity of the eruptions at the source. The decrease in the overall thickness of the ash flows away from the place of their origin and the increase in cooling with distance of transport account for the lateral differences in compaction and welding" (Ratté and Steven, 1967).

Degrees of compaction and welding are influenced also by the strength of the eruptions. Smith and Bailey (1966) say that "the emplacement temperature range seems best explained by the cooling effect of mixing with air in a vertical eruption column before the ash flows formed. The height of the column, and hence the amount of mixing may be related to a changing vapor pressure in the magma. The evidence implies that a falling vapor pressure is conducive to greater heat conservation, hence to successively higher emplacement temperatures during the course of the eruption". As deeper and hotter levels of the magma chambers are tapped by the eruptions, there is a tendency to develop more numerous and thinner zones of dense welding within the ignimbrite sheets. A sudden increase in the content of lithic fragments in a sheet probably reflects increase in the intensity of the eruptions or collapses of the walls of the eruptive vents.

For a detailed account of the lithologic variations exhibited by pyroclastic flows, the reader should consult papers by Ross and Smith (1961) and by Smith (1960) both of which also provide full bibliographies.

Compositions of pyroclastic ejecta

Almost all of the fragmental ejecta related to calderas of the Krakatoa, Katmai, and Valles types and to volcano-tectonic depressions of the Toba type consist of siliceous materials. The most abundant ejecta are rhyolitic; then, in order, follow quartz latites and rhyodacites, dacites, and andesites; pyroclastic flows of basaltic composition are extremely rare. No case has been described in which the erupted magma becomes increasingly siliceous as activity proceeds; the normal sequence is one in which the erupted magma not only becomes increasingly basic but also becomes increasingly rich in crystals as activity continues.

The pyroclastic ejecta around a few calderas of the Krakatoa type, e.g. Crater Lake and Vesuvius, exhibit a pronounced increase in basicity toward the close of a brief eruptive episode, but in general the increase is slight. Among the vastly more voluminous ejecta related to calderas of the Valles type, the change in the composition of the magma is relatively slight, particularly when one considers not only the volumes involved but also the much greater length of the eruptive episodes. The most common change is from rhyolite to quartz latite. The same is true of the still more voluminous pyroclastic flows related to the volcano-tectonic depressions of Sumatra and the Taupo Volcanic Zone of New Zealand.

One of the best examples of these changes is to be seen among the ejecta discharged prior to the formation of Crater Lake. The initial pumice fall and the pumice flows that followed consist of dacite with silica percentages of 66.4 to 68.9. Crystals, mostly of plagioclase and pyroxene, make up approximately 10 to 15 per cent by volume of the pumice fall; in the pumice flows they make up 20 to 25 per cent. After discharge of the dacitic, upper part of the magma chamber, composition of the ejecta changed abruptly, for the final pyroclastic flows are composed of basaltic scoria which were hotter than the dacitic flows, some of them being hot enough to weld and develop columnar structures as they cooled. The silica-percentage of these flows of basic scoria ranges from 53.9 to 56.9, and their bulk composition closely resembles that of the olivine basalts that make up numerous cinder cones in the vicinity. The crystal-content of the scoria flows varies rapidly within wide limits; nowhere is it as low as in the dacite pumice, and usually it ranges from about 40 to 60 per cent; moreover, hornblende replaces pyroxene as the dominant mafic mineral.

A second illustration is provided by Katsui's study of the ejecta of the Shikotsu caldera in Hokkaido. The initial pumice fall consisted of rhyolite with a silica percentage of 72.41; then followed more voluminous pumice flows of dacite with an average silica percentage of 70.87; finally there were smaller pumice flows of felsic andesite with a silica percentage of 63.46.

One more example of variation in the products of Krakatoan eruptions leading to caldera formation must suffice, namely the ejecta discharged by Vesuvius during the great eruption of 79 A.D., when the summit of the volcano collapsed to produce the amphitheatre of Monte Somma. The initial eruptions followed a long period of quiet during which differentiation in the magma chamber led to a concentration of lighter, more gas-rich layers above darker, more basic and crystal-rich layers. Hence the first ejecta consisted of white leucite-phonolite pumice (R.I. - 1.509; SiO_2 - 53.46%; total alkalis -

14.1%) admixed with abundant lithic fragments torn from the walls of the conduit. As activity continued and lower levels of the reservoir were tapped, the pumice changed to a light gray color and became more basic (R.I. - 1.518), and the lithic fragments included bits of Eocene limestone torn from a depth of about a kilometer below sea level. Then followed greenish gray, still more basic pumice (R.I. - 1.525) admixed with lithic fragments of metamorphosed Triassic dolomites and co-magmatic plutonic rocks, torn from the roof of the reservoir at depths of between 5 and 6 kilometers below sea level. All told, these pumice falls covered Pompeii to a depth of 260 cm. An ensuing "gas blast produced sandy comminuted material from the conduit walls, together with a little fragmented lava", the debris falling on Pompeii to a thickness of 5 cm. During the brief pause that followed, parts of the conduit-walls caved inward, and some of the debris was blown out as lithic ash by the next explosions, forming a layer 3 cm. thick. The intensity of the eruptions then increased greatly, and finely divided ash (R.I. - 1.535) was blown out in vast quantity, some of it falling on Pompeii to a thickness of 64 cm. Rains falling during the eruptions mixed with falling ash to produce pisolites, and, ultimately the watersoaked debris formed devastating lahars that swept down the west side of the volcano to bury the town of Herculaneum. Analysis of the tephritic leucite phonolite from Herculaneum shows a silica percentage of 51.64 and a total-alkali content of 12.22 per cent. Thereafter, the intensity of the eruptions gradually diminished, apparently ending with an outflow of leucite tephrite lava (SiO_2 - 49.19%; total alkalis - 8.90%) from the northern foot of the volcano. The roof of the reservoir then collapsed, carrying with it the top of the volcano and leaving in its place the caldera of Monte Somma (Rittmann, 1962).

All of the voluminous pyroclastic flows related to calderas of the Valles type also show a general tendency toward an increase in crystal-content and a corresponding decrease in silica-content. The ash fall that preceded the ash flows constituting the Upper Bandelier Tuff of the Valles caldera contains only 5 per cent phenocrysts, whereas in the first three sub-units of the overlying ash flows the content of phenocrysts increases to 35 per cent. Partial foundering of the roof of the reservoir then brought the eruptions to a temporary halt. Subsequently, while the last third of the total discharge was erupted to produce sub-units 4 and 5, the content of phenocrysts remained at approximately 20 per cent. The silica percentage fell from about 77 in subunit 3 to about 72 in subunit 5; concomitantly there was a progressive decrease in total alkalies and a complementary decrease in FeO relative to MgO. The Lower Bandelier Tuff reveals similar trends (Smith and Bailey, 1966).

At the Creede caldera, the first 100 cubic kilometers of ash flows consist of rhyolite; these were discharged during early stages of collapse; then, as the floor of the caldera continued to sink, crystal-rich ash flows of quartz latite were discharged, totaling 400 cubic kilometers in volume.

Many of the large ash flows associated with the Valles-type calderas of southwest Nevada show a pronounced upward increase in the number of phenocrysts (Lipman, et al., 1966). "In each of these zoned sheets, basal crystal-poor rhyolite grades upward into crystal-rich quartz latite. These compositional changes appear to reflect vertical variations in the magmas from which the ash-flow sheets were erupted". One sheet, the Tonopah Springs Member of the Paint-brush Tuff of Miocene (?) and Pliocene age is typical. This is a "multiple-flow compound cooling unit which originally covered about 700 square miles and had a volume of about 40 cubic miles .... The principal compositional zonation of the tuffs is from basal crystal-poor rhyolite (77 per cent SiO_2; 1 per cent phenocrysts) to capping crystal-rich quartz latite (69 per cent SiO_2; 21 per cent phenocrysts). In addition, concentrations of pumice blocks and lithic inclusions vary systematically, permitting recognition of at least six flow units". The upward increase in the content of crystals is accompanied by systematic variations in their proportions. "In crystal-poor rhyolite, plagioclase and alkali feldspar occur in subequal amounts and clinopyroxene is absent; in crystal-rich quartz latite, alkali feldspar is more abundant than plagioclase, and clinopyroxene is conspicuous .... Crystallization had progressed further in the quartz latite than in the rhyolite at the time of eruption. Three-fourths of the chemical variation results from differences in groundmass composition rather than differences in phenocryst content". All of the variations just mentioned are explicable in two ways, according to Lipman et al., namely: 1) Fractional crystallization and the settling of crystals into a larger continuous magma body underlying the erupted part, and 2) fractional anatectic melting of sialic crust". Both processes probably operated together.

Volumes of Pyroclastic Ejecta

The volume of ejecta discharged explosively during and prior to the formation of most calderas of the Krakatoan, Katmai, and Valles types approximates the volume of material engulfed, when the volume of the pumiceous and scoriaceous glass fragments is recalculated as liquid magma and allowance is made for the content of crystals and of old lithic debris. The caldera of Crater Lake, Oregon, provides a notable exception to this rule, for, as noted earlier (pp. 10-12), the volume of the mountain that collapsed exceeds the volume of magmatic and lithic ejecta by approximately 20 km^3. In such instances, it must be assumed that the discrepancy is accounted for by subterranean withdrawal of magma from the feeding reservoir.

Smith (1960) has pointed out that the volumes of ash-flow deposits associated with calderas of the Krakatoa type, i.e. those formed on the sites of pre-existing volcanoes rarely exceed 50 km^3., though the pyroclastic flows surrounding the Japanese caldera of Shikotsu have twice that volume. And while it is true of almost all calderas of the Krakatoa type that the volume of the associated ash- and pumice-flow (glowing avalanche) deposits greatly exceeds that of the preceding ash- and pumice-fall deposits, there are several exceptions, e.g. the Crater Lake and Lake Coatepeque calderas, already described.

Most calderas lie within or close to ocean basins; hence it is often difficult and sometimes impossible to calculate exactly the volumes of the pyroclastic ejecta involved in their formation. It is commonly said, for example, that 18 km^3. of pumice were erupted prior to the formation of the caldera of Krakatoa in 1883, and some have maintained that this volume equals that of the caldera itself. But this can be no more than speculation, for while a crude estimate can be made of the volume of the preliminary pumice showers, even though most of them fell into the sea, the volume of the succeeding pumice flows remains unknown because most of their deposits were laid down on the ocean floor.

The most voluminous pyroclastic ejecta related to a caldera of the Krakatoa type are those that surround Shikotsu caldera, Hokkaido. Katsui (1963), in his thorough study of this caldera, calculated the volumes of the ejecta as follows:

<table>
<tr><th rowspan="2"></th><th rowspan="2">Vol. km³.</th><th rowspan="2">Mean of apparent density</th><th rowspan="2">Mass 10⁹ tons</th><th colspan="2">Lithic fragments</th><th colspan="2">Juvenile ejecta</th></tr>
<tr><th>Mass 10⁹ tons</th><th>Vol. km³.</th><th>Mass 10⁹ tons</th><th>Vol. as liquid magma: km³.</th></tr>
<tr><td>Pumice flows</td><td>100</td><td>1.3</td><td>130</td><td>17.0</td><td rowspan="2">8.8</td><td>113</td><td rowspan="2">52</td></tr>
<tr><td>Pumice fall</td><td>25</td><td>0.5</td><td>12.5</td><td>0.5</td><td>12</td></tr>
</table>

Volume of collapsed material: approximately 80 km^3.

Volume of erupted material: dry liquid plus lithic fragments - approximately 60 km^3.
wet liquid plus lithic fragments - approximately 80 km^3.

Katsui assumed that just before the Shikotsu eruptions began, water made up 5 per cent by weight of the magma; this, however, would increase the volume of the magma only very slightly if the water were all in solution; only if the magma had begun to vesiculate could its volume have been increased by the amount which Katsui suggests.

The volumes of pyroclastic flows related to calderas of the Valles type and to volcano-tectonic depressions are generally much greater than those just enumerated for calderas of Krakatoan type. For example, the volumes of the pyroclastic flows around the Aira and Aso calderas of Kyusyu each approximate 110 km^3. The Toledo caldera, adjoining the Valles caldera of New Mexico, was formed during and following the discharge of 200 km^3. of ash flows; and the Upper Bandelier Tuffs, related to the Valles caldera, also had a volume of 200 km^3. The final sequence of ash-flow tuffs laid down within the Creede caldera have a thickness of no less than 2,000 meters and a volume of 400 km^3. The Timber Mountain ash flow of southwest Nevada approximates 2,000 km^3. in volume.

The volcano-tectonic depression that holds Lake Toba, Sumatra, resulted from collapse accompanying and following discharge of 2,000 km^3. of rhyolitic pumice, and studies now being made in the Yellowstone region, Wyoming, suggest the presence of similar deposits of comparable volume. Finally, the volume of pyroclastic flows in the Mogollon Plateau of New Mexico, one of the largest ignimbrite fields in the world, is calculated to be approximately 8,000 km^3. (Elston, 1965). Perhaps the Plains of San Augustin, "a grabenlike structure probably larger than the Toba depression" was formed by collapse associated with some of the eruptions, and perhaps a major caldera is also present in the area.

Post-Caldera Eruptions

Within calderas that originate on the sites of former composite volcanoes, i.e. within calderas of the Krakatoa, Katmai, and Suswa types, it is common to see post-caldera domes and cones, but it is only in calderas of the Valles type and in volcano-tectonic depressions of the Toba type that one sees "resurgent domes".

Less than five years after the caldera of Katmai was formed, a small cone of andesite was built on the floor, signifying a return of andesitic magma to the central conduit after it had been drained in 1912 to mingle with rhyolitic magma in the adjacent fissure-system crossing the Valley of Ten Thousand Smokes.

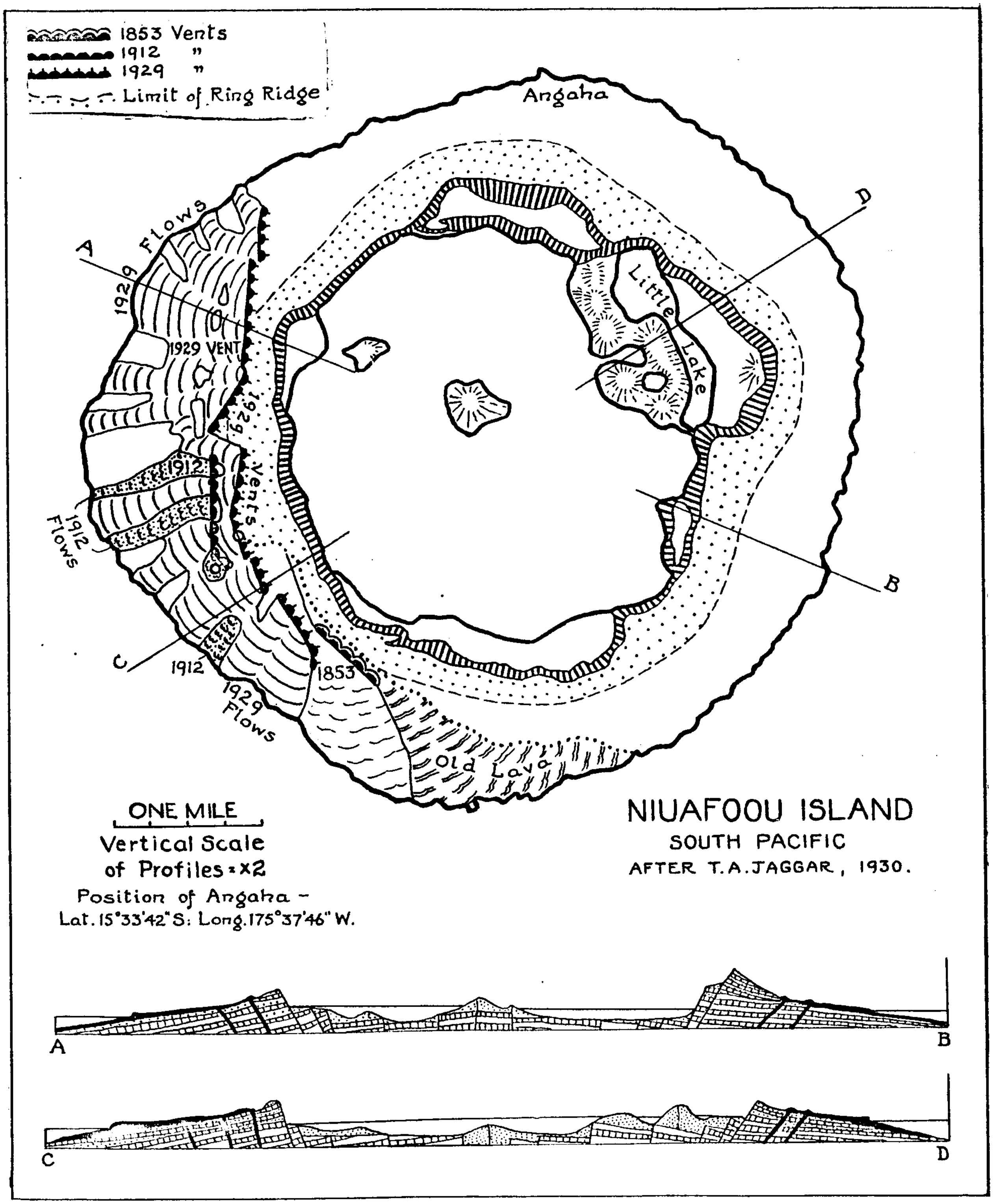
Forty-two years after the eruptions of dacite pumice that led to the formation of the caldera of Krakatoa in 1883, a basaltic cone, Anak Krakatoa, rose from the floor and still continues to erupt. Approximately 1,300 years after the caldera of Santorin was formed, dacite domes were built on a horst crossing the floor, and ever since, at long intervals, similar domes have risen to add to the size of a coalescing group. The caldera of Monte Somma was formed in 79 A.D., but it was not until 1631 that the cone of Vesuvius began to grow in its center. A period of quiet of unknown length followed the formation of the caldera of Crater Lake, Oregon; eruptions of andesitic lava and scoria then produced the submerged Merriam Cone and Wizard Island, and a Pelean dome of glassy dacite was built on the floor of the lake.

In most calderas, including those just mentioned, the interior cones and domes show no apparent control by tectonic structures in the basement-rocks. In some calderas, however, the loci of intra-caldera eruptions seem to be determined by pre-existing structures. The domes and cones within the Towada caldera, Honsyu, for example, are aligned in a north-northwest direction, parallel to a fault that displaces the oldest rocks on the walls of the caldera. Cones within the Suswa caldera, Kenya, the Shikotsu and Aso calderas, Japan, and the Newberry caldera, Oregon, likewise show alignments that reflect structural trends in the basement-rocks.

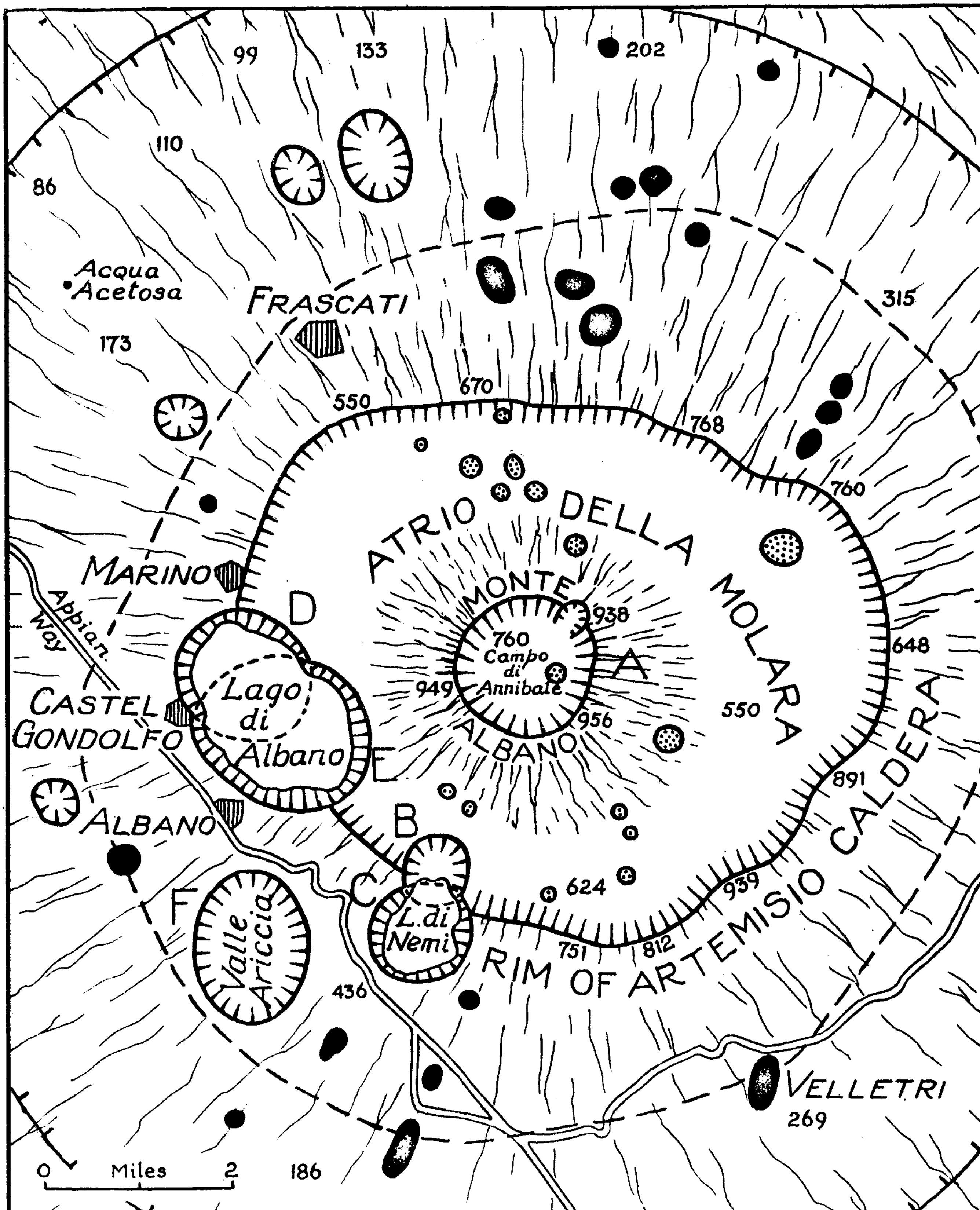
Domes and cones may be built over ring-fractures at or close to the margins of calderas. Rhyolite domes within the caldera that holds Lake Coatepeque in El Salvador provide an example; so do the lava-scoria cones close to the walls of Askja caldera in Iceland. (See figures 7, 8, 29 and 30).

Few calderas of the Valles type contain composite or scoria cones of andesitic or basaltic composition, the Aso caldera in Kyusyu being one of the exceptions. On the other hand, many of them contain domes of rhyolite or quartz latite. These may rise from the rims of the calderas, as some do in the Taupo Volcanic Zone of New Zealand, or within the moats between the caldera walls and the central 'resurgent domes'. Examples of such ring- or moat-domes are illustrated in figures 8 and 11.

'Resurgent domes' were first recognized by van Bemmelen within the Lake Toba depression, Sumatra, and many have since been discovered in calderas of the Valles type in Nevada, Colorado, and New Mexico, though, for reasons not understood, none seem to be present in New Zealand or Japan. These 'resurgent domes' result from uplift of the floors of the calderas by large, rising bodies of siliceous magma, and, as noted already, the structural uplifts may exceed 1,000 meters. Complex fractures develop on the flanks of the domes as they rise, and apical grabens are commonly produced, running along the major axes of the domes, probably following structural trends in the sub-volcanic basement. Viscous, siliceous lavas may be erupted from these fractures during the later stages of domical uplift and subsequently, they may also issue from ring-fractures beneath the encircling moats to form thick flows or bulbous domes.



1853 Vents
1912 "
1929 "
Limit of Ring Ridge
Angaha
A
D
1929 Flows
1929 VENT
1929 Vents
1912
1912 Flows
Little Lake
B
C
1912
1929 Flows
1853
Old Lava
ONE MILE
Vertical Scale of Profiles = x2
Position of Angaha - Lat. 15°33'42" S: Long. 175°37'46" W.
NIUAFOOU ISLAND
SOUTH PACIFIC
AFTER T. A. JAGGAR, 1930.
A
B
C
D



PARASITIC CONES OF THE ARTEMISIO VOLCANO

" " " MONTE ALBANO

CRATERS

Heights in meters

Redrawn from V. Sabatini: Vulcano Laziale; Memo. descr. della carta geol. d'Italia, vol x, tav. x, 1900.

Geophysical Studies of Calderas

Much effort has been directed to geophysical studies of calderas, mainly by gravity and magnetic methods, especially in Hawaii and Japan; a few geophysical studies have also been made in volcano-tectonic depressions in the continental United States and in New Zealand. Data are summarized in Table 1.

Krakatoan Calderas

By far the greatest number of calderas for which geophysical data are available are those associated with voluminous explosive eruptions of siliceous magma, i.e. with voluminous pumice-falls and pumice-flows. A wide range of structures has been included in this category, as noted already. Some Japanese calderas, such as Aso, Aira, Kuttyaro, and Akan, resemble the Valles caldera of New Mexico, being largely independent of pre-existing composite volcanoes, though some are so large that it is difficult to be sure. Smaller calderas, such as Towada and Hakone, seem to have originated, as did Crater Lake, Oregon, near the centers of large mature volcanoes at the culmination of long magmatic histories.

The foregoing calderas are difficult to distinguish on a geophysical basis. Yokoyama (1966) has pointed out that some of the smaller ones formed within broader areas of uparched basement-rocks. Several relatively small Japanese calderas, for example, Toya, are surrounded by a regional gravity high (figure 31). Seismic reflection profiles on the flank of the Shikotsu caldera and a broad magnetic anomaly trending along the volcanic belt toward Kuttyaro caldera also seem to reflect large-scale basement-highs. But no comparable gravity highs surround very large calderas, such as Aso, Kuttyaro, and

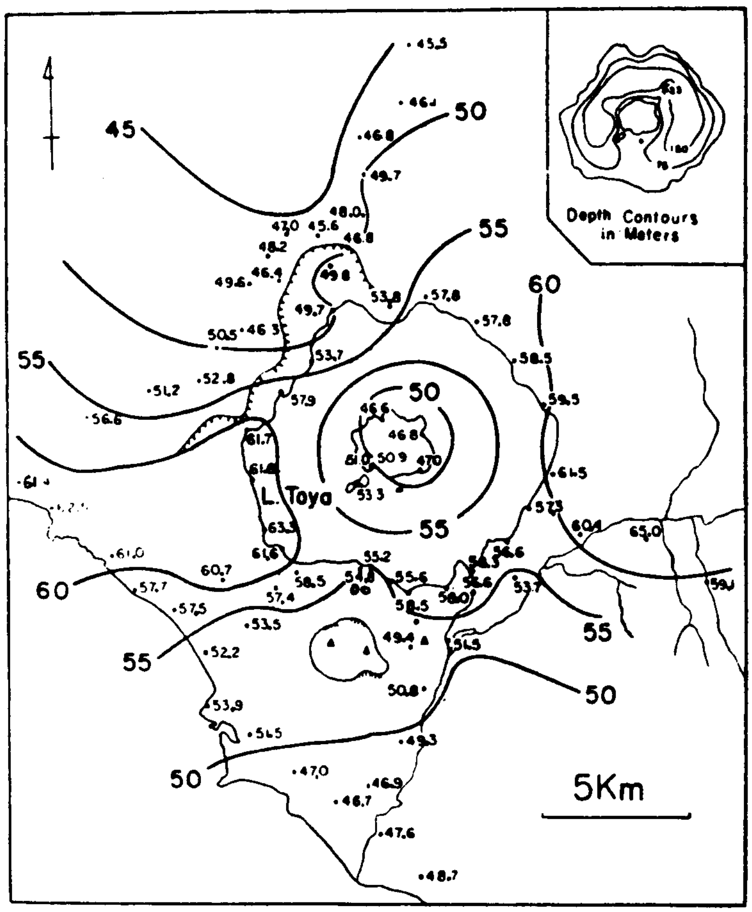


Figure 31. Distribution of the Bouguer anomalies in mgal on and around Toya Caldera (not corrected for topography).

Table 1. Gravity anomalies associated with calderas

Group A. Calderas associated with voluminous eruptions of siliceous tephra.

	Size	Gravity Anomaly	Reference
Towada, Honsyu	8 x 9 km	-10 mgal	Yokoyama (1965)
Hakone, Honsyu	10 x 10 km	-10 mgal	Yokoyama (1965c)
Crater Lake, Oregon	8 x 9.7 km	0 to -5 mgal	Blank (in press)
Akan, Hokkaido	13 x 24 km	-25 mgal	Yokoyama (1958)
Kuttyaro, Hokkaido	20 x 27 km	-16	Yokoyama (1958)
Toya, Hokkaido	10 x 10 km	-14	Yokoyama (1965a)
Aso, Kyusyu	16 x 23 km	-20	Yokoyama (1963)
Aira, Kyusyu	20 x 20 km	-30	Yokoyama (1961)
Shikotsu, Hokkaido	15 x 15 km	-20	Yokoyama (1965b)
Apoyo, Nicaragua	5 x 5.5 km	+30	Richards (unpublished)

Group B. Calderas associated with voluminous eruptions of basaltic lavas.

	Size	Gravity Anomaly	Reference
O-sima, Japan	3 x 4 km	+ 9	Yokoyama (1957)
Masaya, Nicaragua	11 x 6 km	+25	Richards (unpublished)
Mauna Loa, Hawaii	4.5 x 2.6 km	+100	Kinoshita, *et al.* (1963)
Kilauea, Hawaii	4.4 x 3 km	+60	Kinoshita, *et al.* (1963)
Lanai, Hawaiian Islands	3 x 4 km	+40	Krivoy & Lane (1965)
Kauai (Olokele caldera)	16 x 20 km	+90	Krivoy, *et al.*(1965)

Note: Gravity anomalies given for Hawaiian calderas are the maximum Bouguer anomalies for the volcano above an average value at sea level. In most cases the center of the gravity high is eccentric to the caldera and may lie outside the boundary faults.

Aira. This may be due to the masking effect of great masses of low density pyroclastic materials on the outer flanks of these great calderas.

Attempts to interpret the structures of calderas of the Krakatoa and Valles types from gravity and magnetic data have thus far had only limited success. The absence of a strong density or magnetic contrast between the central block and the surrounding rocks makes it difficult or impossible to interpret the attitude of the boundary faults or of large intrusions within the calderas. Yokoyama (1963, 1965) concluded that the boundary faults dip inward at very low angles, on the assumption that they separate light superficial pyroclastic materials on the floor from under-lying denser rocks. However, the contacts may be entirely depositional and have little relation to the major structures of the underlying rocks. If the boundary faults do indeed dip inward at very low angles, it is difficult to see how large-scale subsidences could have taken place.

Calderas of the Krakatoan and Valles type are often said to be characterized by negative gravity anomalies, this feature supposedly distinguishing them from other types of calderas. This conclusion has been based on the negative anomalies measured in each of nine Japanese calderas of the Krakatoan-Valles type. However, Crater Lake, Oregon, and the Nicaraguan caldera, Apoyo, which are of the Krakatoan type, show little, if any negative anomaly or even a strong positive anomaly. Negative anomalies are certainly common among calderas like Krakatoa and Valles, but they appear to result more from superficial low density material on the caldera floors than from major mass deficiencies in the internal structures. Most of the anomalies can be explained by a few hundred or at most a few thousand meters of loose pyroclastic and sedimentary debris on the caldera floors. The largest observed negative anomaly is that of Kuttyaro, where a drill hole through the floor penetrated 1,000 meters of pumice and poorly consolidated materials.

Other calderas, such as Crater Lake and Apoyo, which are believed to have only a thin veneer of recent sediments, show little or no negative anomaly. Accumulation of pyroclastic and sedimentary material is probably greatest within calderas formed by repeated eruptions and by intermittent subsidences, as in many calderas of the Valles type where thick sheets of ignimbrite accumulate while the floors subside.

Magnetic data indicate little contrast in the physical properties of rocks inside and outside calderas of the Krakatoan and Valles types; they also indicate that in general the intra-caldera filling of loose material is relatively shallow. Magnetic surveys have been made of 14 Japanese calderas of the Krakatoan and Valles types, and in many there is little, if any reflection of the boundary faults. Basement structures seem to pass from the surrounding region through the calderas with little offset or difference in intensity. Small but sharp magnetic highs have been detected within some Japanese calderas, such as Shikotsu and Mashu, but they do not correspond to positive gravity anomalies. They have been interpreted as post-caldera intrusions or buried cones too small to be detected by gravity measurements on the lakeshores.

Hawaiian (Kilauean) Calderas

Most calderas associated with basic lavas are on oceanic shield volcanoes, and nearly all studied by geophysical means are in the Hawaiian Islands. O-sima in Japan, and Masaya in Nicaragua, seem to be the only other basaltic calderas for which gravity and magnetic data are available. All calderas of this type show positive gravity anomalies (Table 1).

Large positive gravity anomalies are associated with each of the major volcanoes in the Hawaiian Islands, even though many of them lack summit-calderas. The center of the anomaly on several volcanoes is eccentric with respect to the caldera, and there is no apparent relation between the boundary faults of the calderas and the gravity gradients. Gravity profiles show smooth slopes, not unlike the topographic slopes of the shields (figure 32). It may be somewhat arbitrary, therefore, to infer a direct relationship between the calderas and the positive gravity anomalies.

No large shield volcano is associated with the Masaya caldera, Nicaragua; nevertheless, a positive gravity anomaly coincides almost exactly with the caldera. At O-sima, however, the relatively small positive anomaly coincides with the caldera only near its western edge; elsewhere it shows no apparent control by the surface structure (figure 33).

Magnetic surveys of Hawaiian volcanoes show a strongly magnetized mass beneath the summit-regions and rift zones, but no clear relation to the calderas. O-sima has a similar magnetic dipole centered over the summit-crater of a cone close to the southern margin of the caldera. No significant magnetic anomaly was detected in the Masaya caldera, other than that produced by a large adjacent fault. These limited data suggest that magnetic anomalies are associated with cones and shields but are not distinctive features of calderas.

Macdonald (1965) has pointed out that the mechanism of caldera formation in Hawaii cannot be directly related to the large dense masses responsible for the gravity and magnetic anomalies. Many strong anomalies have no detectable subsidence immediately above their centers, whereas several calderas are substantially offset from the centers of the anomalies. He suggests, however, that subsidence and spreading of a dense, olivine-rich intrusive mass, formed by crystal-accumulation or by upwelling of material from the mantle, may cause widespread distension of the flanks and summit-regions of shield volcanoes.

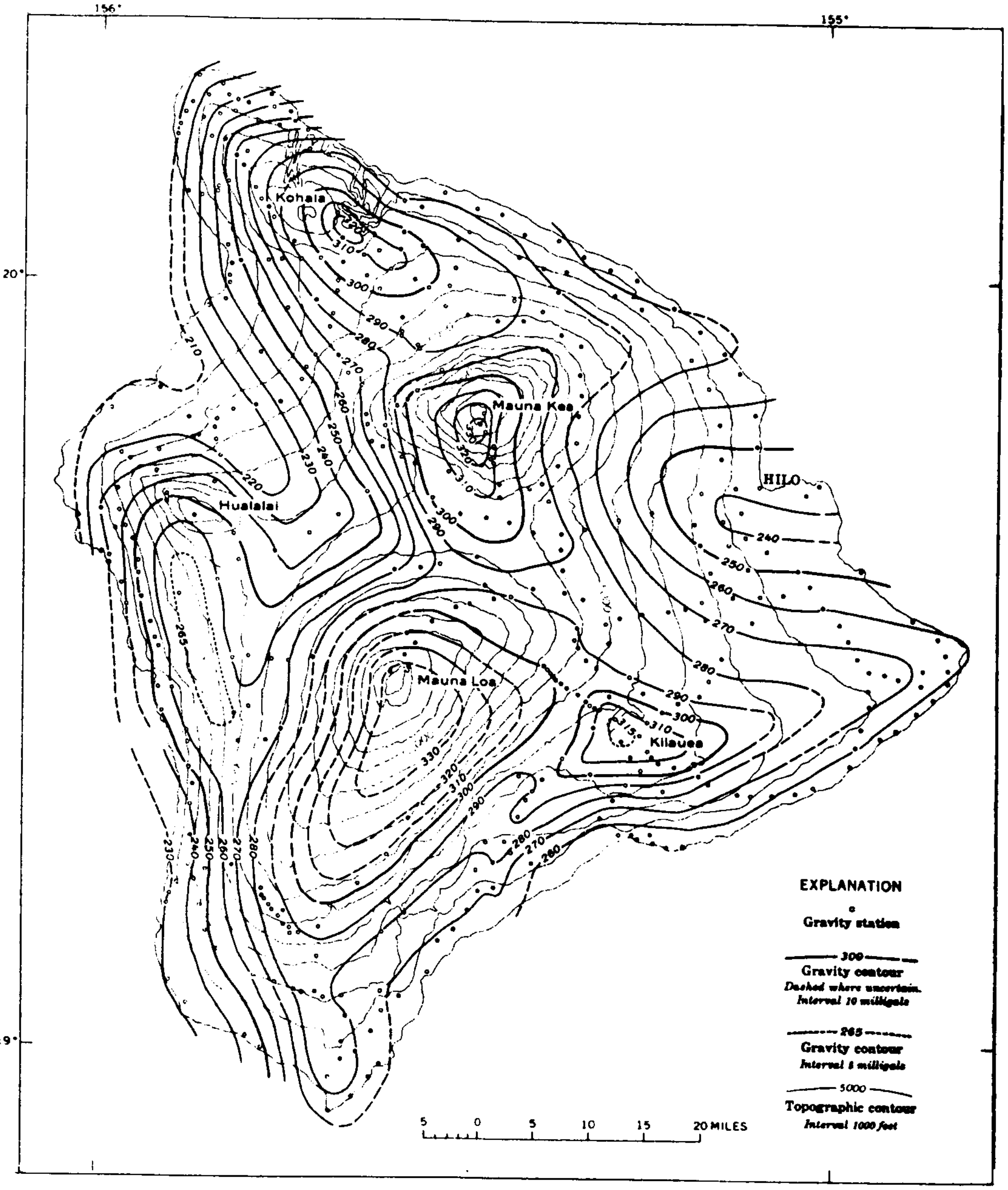


Figure 32. Bouguer gravity-anomaly map of the island of Hawaii.

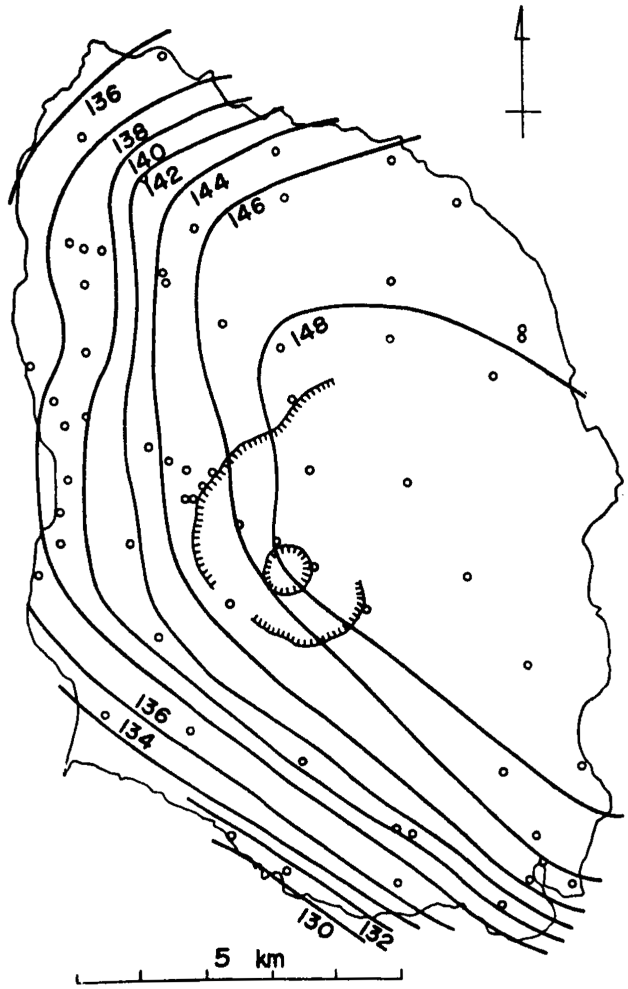


Figure 33. Distribution of the Bouguer anomaly in mgal (adopting the modified Bouguer corrections).

Measurements of level and of tilting around Kilauea caldera have demonstrated that important changes in the shape of the volcano take place during eruptive cycles. Tiltmeters around the flanks of the caldera detect changes of inclination as small as 1 mm. in 5 km., and electronic measurement of distances across the caldera detect horizontal changes as small as 4 to 7 parts per million (Eaton, 1959 and 1962; Decker, et al., 1966). Long before summit-eruptions, the volcano swells, so that there is increased outward tilt of the flanks and lateral spreading of the summit. When magma is drained by eruptions of lava from adjacent rift zones, the volcano deflates to nearly its original form. Uparching of the summit, by widening the fractures surrounding the caldera, permits the floor to subside. Macdonald (1965) suggests, for instance, that collapse of the floor of the pit of Halemaumau within the caldera of Kilauea in 1924, and the basining of the caldera floor in 1868 and 1894 resulted from distension of the summit region during eruptive cycles.

Calderas associated with mixed effusive and explosive eruptions (Glencoe type)

Quaternary calderas of the Glencoe type are extremely rare; the only one that seems to have been studied geophysically is the Newberry caldera in central Oregon, but results of gravity and magnetic surveys now being made under the direction of H. R. Blank are not yet available.

Volcano-tectonic depressions (Toba type)

A few geophysical studies have been made of large depressions of the Toba type. For example, Long Valley, California, an elliptical collapse-depression related to the eruption of the Pleistocene Bishop Tuff (ignimbrite), was recently studied by Pakiser (1962). He found a negative anomaly with a maximum gravity relief of 78 mgal., and a small but prominent high near the center of the gravity low. Magnetic profiles showed a high over the valley, with a sharp central peak corresponding to the small gravity high. Pakiser calculated that an igneous body large enough to produce the magnetic high must lie at a depth of about 1,000 meters. And he suggested that the combined magnetic and gravity data, related by Gauss's theorem, could be produced by Cenozoic deposits with a volume of 1,840 cubic kilometers and a density 0.4 g. per cm^3. less than that of the surrounding pre-Tertiary rocks.

Mono Basin, a few kilometers north of Long Valley, although not related to a single, voluminous ignimbrite-eruption, seems to have been formed by subsidence during repeated eruptions of pumice and lava. It shows a negative gravity anomaly of about 50 mgal. In Pakiser's opinion, the basin occupies a structural depression about 5 to 7 kilometers deep. More recent studies by Christensen and Gilbert (personal communication) indicate, however, that the depth of the Cenozoic deposits within the basin is very much less.

The Taupo volcanic depression in the North Island of New Zealand is probably the best studied volcano-tectonic depression in the world. Gravity and magnetic surveys have been carried out with special attention to the hydrothermal region around Wairakei and Lake Taupo. The Taupo depression is marked by an elongated gravity low that branches from a much larger gravity low covering most of the eastern side of the North Island. The structure, which trends north-northwest through Lakes Taupo and Rotorua, has a gravity relief of 50 to 60 mgal. The anomaly diminishes in intensity northward, becoming indistinct near the andesitic volcanoes close to the north coast of the island. The gravity and magnetic pattern is very complex on a local scale, with many sharp highs and other irregularities superposed on the over-all pattern. In some places, the gravity highs correspond to strong magnetic highs and are therefore interpreted as shallow intrusive bodies or buried cones; elsewhere, no clear magnetic anomaly is associated with the gravity highs, perhaps because of the intense hydrothermal alteration to which the rocks have been subjected.

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