10 March 1961, Volume 133, Number 3454 SCIENCE

CURRENT PROBLEMS IN RESEARCH

Volcanology

Volcanoes furnish some of our best clues to the nature of the earth's interior.

Gordon A. Macdonald

A volcano is commonly defined as a place where molten rock (magma) and gases issue at the surface of the earth, or as the hill or mountain built up around that vent by the escaping material. But volcanology, the science of volcanoes, goes beyond these surficial structures and the processes that take place at and around them and concerns itself also with the structures and processes within the earth that give rise to the surface phenomena. Indeed, volcanoes are one of our principal windows to the earth's interior; the composition and condition of materials issuing at volcanoes supply our best direct evidence as to the nature of the rocks and the processes that go on within the earth below the very thin rind that is known to us from direct observation. Less than 0.3 percent of the radius of the earth is directly visible to geologists, even in the deepest drill holes and the most deeply eroded mountain belts. All the rest must be studied by indirect methods, such as seismology, and by inference from things observed in the visible surficial film and at volcanoes. There is today a tendency to include all magma chemistry within volcanology, perhaps partly because of the growing realization that many of the deep-seated, "plutonic" rocks traditionally considered to be igneous—that is, to have consolidated from a molten state—actually have never been molten. There can, however, be no question that the molten lavas poured out at volcanoes represent real magmas.

At any rate, today volcanology includes the study not only of volcanoes per se but of all even remotely related things in the earth beneath.

In a sense, volcanology is a very old science. The ancients had an intense, partly superstitious, interest in volcanoes. The great Roman natural historian, Pliny the Elder, lost his life on an expedition undertaken in part to study the great eruption of Vesuvius in the year A.D. 79. (Eruptions of that type are now called Plinian eruptions.) Nevertheless, modern volcanology is very new. Classical volcanology was almost wholly descriptive. A great mass of information was collected on the physical nature of lavas and the structure of volcanic mountains, but the relatively small amount of interpretation of these facts was almost wholly speculative. Within the last half century the emphasis in volcanology has shifted to interpretation. We are now trying to deduce the processes that must have gone on within the earth to produce the results we see at the surface.

That this interpretive phase was so long delayed is the result of two general factors. First, volcanic eruptions are relatively infrequent and the physical difficulties of studying them at close range are great, and therefore it has taken a good deal of time to accumulate sufficient information on which to base interpretations. Even more, however, the delay has been the result of the border-line nature of volcanology, which depends very heavily on other sciences, particularly physics and chemistry. Interpretive volcanology of necessity had to await the development of physical and chemical methods and theory.

Current Problems in Volcanology

The major problems in volcanology are today, as they always have been, the origin of volcanic heat, the locus and means of origin of magma, the mechanics of rise of magma to the earth's surface, the processes that give rise to the considerable range in composition of magmas that reach the surface, the subsurface structure of volcano systems, and the origin of the water that forms the major portion of the volcanic gases.

There are, of course, a host of other, lesser problems, only a few of which can be mentioned here. As examples I might cite the mechanics of formation of large sunken-in craters, or calderas, such as that of Crater Lake, in Oregon; the mechanics of lava flow and the effects of flows on objects they encounter; the physical properties of lavas and magmas; the mechanics of volcanic explosion and the formation of volcanic ash; and the mode of eruption and emplacement of certain sheets of rock, commonly of vast volume and extent, that partake of the characteristics of both lava flows and volcanic ash.

In another class fall the practical applications of volcanology: the utilization of volcanic heat, either directly or as steam, for heating and—vastly more important—for generation of power; the appraisal of risk from volcanic activity to persons and property; the prediction of time and type of eruption; the control of lava flows and volcanic mud flows. Ignored by some scientists, these practical problems are of tremendous importance to the millions of persons living on and near active volcanoes and form a link be-

The author is a volcanologist at the Institute of Geophysics, University of Hawaii, Honolulu.

Table 1. Compositions of basaltic and peridotitic rocks (percent by weight). (1) Average tholeiitic basalt of the Deccan region, India (Holmes and Harwood, 1928); (2) average tholeiitic basalt of the Hawaiian Islands (Kuno *et al.*, 1957); (3) average alkali basalt, Hawaiian Islands; (4) average high-alumina basalt, northeastern California; (5) eclogite, Glenelg, Scotland (Yoder and Tilley, 1959); (6) average of silicate phase of stony meteorites (Brown and Patterson, 1948); (7) average dunite (Wager, 1958).

Rock	1	2	3	4	5	6	7
SiO_2	49.7	50.9	46.7	50.9	50.0	46.2	41.9
Al_2O_3	13.0	13.2	14.4	17.5	13.4	3.4	1.1
Iron as FeO	13.2	10.9	11.9	8.5	13.7	17.0	7.3
MgO	5.7	8.0	8.8	8.3	6.5	27.6	46.2
CaO	10.1	10.6	10.6	10.2	11.0	2.9	1.1
Na_2O	2.3	2.2	3.0	2.9	2.4	1.1	0.1
K_2O	0.5	0.4	0.9	0.5	0.4	0.2	0.005
TiO_2	2.6	2.8	3.0	1.0	1.6	0.2	0.1
P_2O_5	0.3	0.3	0.4	0.2	0.1	0.4	0.1
MnO	0.2	0.1	0.1	0.2	0.2	0.4	0.1

tween the science of volcanology and volcano technology.

There is not space here to discuss all of the problems of volcanology, or even to discuss any of them really adequately. I shall attempt merely to present a brief review of some of the more basic ones, and to indicate something of the complex interplay of various scientific disciplines within the field of modern volcanology.

Origin of Magma

The earth's crust is a thin shell, about 20 to 60 kilometers thick beneath the continents and as little as 3 kilometers thick beneath the oceans. Beneath the crust lies a much thicker shell, known as the mantle, that extends to the boundary of the core at a depth of 2900 kilometers. The molten rock, or magma, that is erupted at volcanoes is generally believed to originate in the outer part of the mantle, but we actually have little direct evidence of just where or how it is formed.

For a quarter of a century we have recognized earthquakes stemming from a zone about 40 kilometers deep below the Hawaiian volcanoes (about 35 kilometers below the top of the mantle in that region), and it has been presumed that the magma originates there. In recent years, with the installation of more sensitive seismographs, there has been found associated with the earthquakes a rhythmic trembling of the ground, known as volcanic tremor. This tremor has long been known to arise also in a zone near the surface, where it quite definitely accompanies the movement of molten rock in the volcanic conduits. Thus, recognition of volcanic tremor in association with earthquakes from the 40-kilometer zone tends to confirm the belief that moving magma is present at that level. In the absence of better evidence, we are assuming that Hawaiian magmas come from that depth. Seismological evidence also suggests that magma is formed at or near the top of the mantle, at a depth of 50 to 60 kilometers, beneath the Kamchatkan volcanoes.

Near the surface, temperature increases downward within the earth at an average rate of about 1°C per 30 meters. It seems quite certain that this rate of increase does not continue to great depths, but just what the rate is beyond the outermost few kilometers is highly speculative. No direct measurements are available, and estimates are strongly influenced by such theoretical considerations as the assumed mode of formation of the earth, the origin of the earth's heat, and the composition and physical properties of the mantle and core. Most estimates, however, place the temperature at a depth of 200 kilometers, between 1400° and 1750°C. One of the latest estimates (*1*) envisages a steep temperature gradient near the surface, reaching 1100° or 1200°C at 100-kilometer depth and about 1500°C at 200 kilometers and then flattening to about the adiabatic gradient through the rest of the mantle. This and other recent estimates are shown in Fig. 1. These estimates are, of course, intended as general averages, and some degree of departure from them, both upward and downward, is to be expected locally.

Nearly all recent estimates place the temperature in the outer part of the mantle above the melting point of basaltic lava under surface conditions. Yet seismological evidence appears to indicate unequivocally that the earth is essentially solid down to the boundary of the core. Old ideas of a thin, solid crust overlying a general zone of liquid rock have had to be abandoned. The general solidity of the mantle is, no doubt, the result of rise of melting temperature with increasing pressure. Verhoogen's (*1*) estimate of the increase of melting temperature of basalt at increasing depth within the earth is also shown in Fig. 1. It is noteworthy that at about 200 kilometers the melting-point curve very closely approaches the curves of estimated temperature. Presumably, however, since the mantle is largely solid, the temperature curves do not in general reach the melting curve. Nevertheless, it would appear that, locally, either a comparatively small rise in temperature or reduction of melting point through lowering of pressure or introduction of fluxes might result in melting and formation of magma.

None of the estimates (Fig. 1) bring the temperature at 40-kilometer depth close to even the surface melting point of basalt or the observed temperatures of eruption of Hawaiian lavas (1100° to 1200°C). It seems, therefore, that if our deductions on the depth of origin of Hawaiian magma are correct, its formation cannot be accounted for by simple release of pressure. A local rise in temperature appears to be demanded. Such a rise is not unlikely. All over the earth volcanic areas commonly have thermal gradients considerably higher than the general average. This indicates a local rise of the isogeotherms (surfaces of equal temperature within the earth) that may be not only the result but the cause of volcanism.

The problem of the origin of volcanic heat is closely tied up with that of origin of the heat of the earth as a whole. Verhoogen (*2*) has pointed out that the amount of heat brought to the earth's surface by volcanic action is only a small fraction of the heat being conducted to the earth's surface and radiated into space, and that therefore the amount of heat represented by volcanism is relatively minor. Though this is true, there still remains the problem of how heat becomes concentrated locally to produce volcanic activity. Is it brought up from deep in the mantle by convection currents? Is it in some way generated by friction? At present we really have no adequate answer.

The amount of energy liberated as heat during some eruptions is very large. Thus, during the 1952 eruption of Kilauea the heat given off was approximately 4.3×10^{16} calories (*3*). In that particular eruption energy was released almost wholly in the form of heat, but in others, large amounts are liberated in explosion and in earthquakes. Yokoyama (*4*) estimates that the total energy released in volcanic eruptions ranges from about 10^{16} to 10^{25} ergs. Almost certainly most of this energy originated at depth and was transported upward by rising magma.

Nowdays the heat of the earth as a whole is generally regarded as resulting largely from compression of the original earth-forming materials and, to a lesser extent, from liberation of heat through radioactivity. The suggestion has been made repeatedly that volcanism is the result of a local greater-than-average concentration of radioactive material. However, field checks with Geiger counters on still-active hot lava flows and within the volcanic gas cloud during an eruption reveal no increase in radioactivity over the general background level. Locally, a minor amount of heating takes place at the earth's surface as a result of burning of volcanic gases, and formerly it was supposed that an important amount of heating took place in the magma as it approached the surface, by oxidation and interreaction of included gases and other exothermic reaction (*5*), such as that between iron oxide and steam. It now appears, however, that such heating is probably relatively unimportant (*6*). Liberation of latent heat of crystallization also helps maintain the temperature of a magma during late stages, but it is probably a minor factor even then, and can have no effect until the magma cools enough for crystallization to commence.

Most of the heat of a volcanic eruption must be brought up from depth with the rising magma. This conclusion is supported by the fact that the temperature of magma erupting at the surface is little if at all higher than the melting temperature of the rock even under surface conditions, let alone under the higher pressures existing at the depth of magma generation. Small amounts of heat resulting from exothermic reactions in the rising magma may help make up for the losses of heat by conduction and radiation and by cooling resulting from the expansion of included gas bubbles, but heat from this source must be very minor.

In whatever manner the local rise in temperature is brought about, there is no question that melting does occur and magmas are formed. The molten rock pours out at the surface for all to see. An additional problem arises, however, because the composition of the melts reaching the surface is very different from the composition generally attributed to the mantle material. By far the most abundant volcanic rock is basalt, and it appears probable that other volcanic rocks have been derived in large part through modification of basalt magma. Basalt itself varies somewhat in composition, and three principal intergradational varieties are now recognized (Table 1). Of these, by far

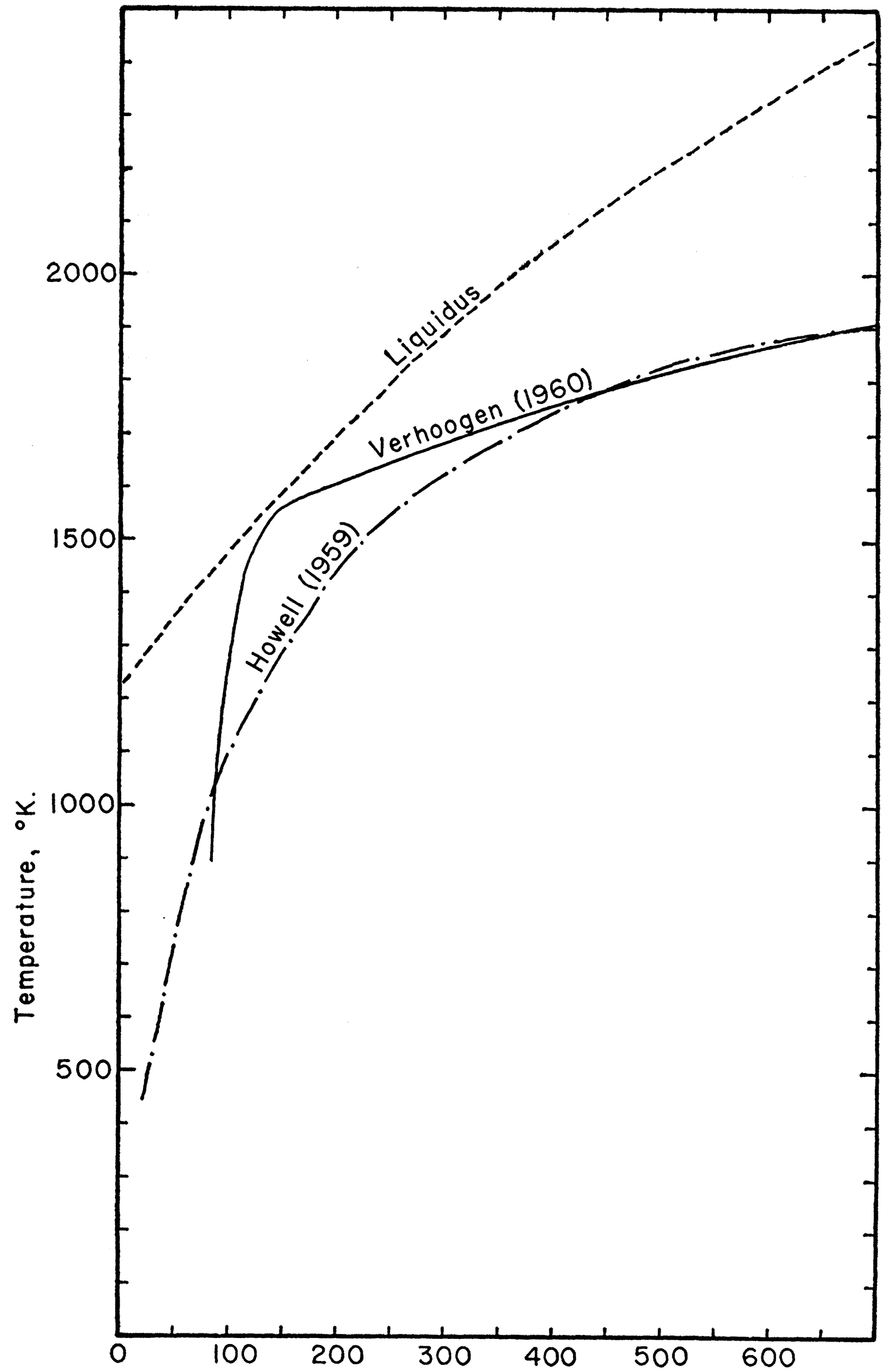

Fig. 1. Curves showing possible temperature distribution in the earth. The upper, dashed line, marked "Liquidus," indicates the approximate temperatures at which basalt would become completely molten, according to Verhoogen. [After J. Verhoogen and B. F. Howell, Jr.]

the most abundant is that known as tholeiitic basalt, which forms the major portion of the great Hawaiian volcanic mountains, with a volume approaching 100,000 cubic miles, as well as the huge accumulations of the Columbia River plain ($\approx$100,000 cubic miles), the Deccan region of India (>100,000 cubic miles), and other, similar masses. The other types of basalt may possibly arise independently of the tholeiitic basalt through melting within the mantle (*7*), or they may be derived from it.

Most geologists and geophysicists consider that the outer part of the mantle consists of a rock known as peridotite, which contains abundant olivine (peridot) and is rich in iron and magnesium. It should be remembered that we have never seen any samples that we are certain came from the mantle. Fragments of peridotite brought to the surface in rising lava have been considered to represent mantle rock (*8*), but the fragments show all gradations into gabbro (a coarse-grained rock of the same composition as basalt) and resemble, in mineral composition and texture, rocks formed by sinking of crystals in certain great gabbro masses that have been intruded into, and solidified within, the earth's crust and have later been exposed to view by erosion. It appears equally probable that the peridotite fragments were derived from similar intrusive masses transected by the rising magma.

The generally accepted theory of peridotite composition of the mantle is based on two general lines of evidence: the physical properties of the mantle material indicated by seismological and other geophysical evidence and the composition of stony meteorites. Whether the latter represent samples of the sort of material from which the earth was formed or fragments of a disrupted planet, it seems probable that they supply a clue to the composition of the earth's interior. Stony meteorites approximate peridotite in chemical composition (Table 1), and peridotite has the density and elastic properties demanded by geophysical data on the mantle. But if basalt magma originates from melting of peridotite, the melting is incomplete. Peridotite, like other rocks composed of several different minerals, does not have a single melting point but has, rather, a melting range of several hundred degrees. As the temperature rises into the melting range, the most easily melted constituents melt first, and if the liquid is then removed from the remaining solid it has a composition different from that of the total original solid. Basalt magma may be formed from peridotite in that way.

Another possibility is that the outer part of the mantle is not actually peridotite at all but, instead, is a high-pressure polymorph of basalt—a rock known as eclogite, which has the same chemical composition as basalt but is composed of different minerals and has a high density about equal to that of peridotite. At depths of 30 to 60 kilometers, such as we find for the upper surface of the mantle under much of the continents, a phase change of this sort seems quite possible. Beneath much of the oceans, however, the depth to the top of the mantle is only about 5 kilometers, and pressure is comparatively low. Nevertheless, Lovering (*9*) and Kennedy (*10*), among others, believe that even at this shallow depth the phase change takes place. If the outer mantle is indeed eclogite (or equivalent amphibolite or pyroxenite), formation of basaltic magma involves a total, rather than a partial, melting of the mantle rock. Such total melting appears to be a possibility because of the narrow melting range of basalt—in general less than 150°C (*11*).

Geological evidence points to the rapid generation, at many times and places throughout the history of the earth, of large volumes of basaltic magma of rather uniform composition (*12*). With only one principal exception, these are of tholeiitic composition, and even in that one area (the Thulean province of the northeast Atlantic Ocean) tholeiitic basalt is abundantly associated with the alkali basalt. In other areas, also, it appears that later lavas are alkali basalt, possibly derived from the tholeiitic magma by chemical processes (differentiation) in the magma chamber. The uniformity of the initial tholeiitic basalt is perhaps more readily accounted for by total melting of a rather uniform eclogite shell with the chemical composition of tholeiitic basalt than by partial melting of peridotite.

It is hoped that speculation on the composition of the outer part of the mantle will soon be eliminated by the evidence of actual samples taken from it in a deep hole (the "Mohole"), to be drilled through the crust and into the mantle somewhere on the deep sea floor. Establishment of the composition of the mantle will likewise serve to limit and direct petrologic speculation on the origin of basalt magma.

Subsurface Structure of Volcanoes

The structure of the upper few thousand feet within volcanic cones is fairly well known from the study of natural sections exposed in them by deep erosion. Below this, however, we again enter a region about which our knowledge is distinctly speculative.

The broadly rounded mountains known as shield volcanoes are built by numerous outpourings of fluid basaltic lava with very little accompanying explosive activity. These eruptions are fed principally by magma rising through cracks in the volcanic edifice, and the cracks are concentrated in certain zones, known as rift zones, that extend radially outward from the apex of the mountain. The eruptions are "fissure eruptions," characterized by lava spurting from a crack over a length of several hundreds or thousands of feet, or even several miles. On the great Hawaiian shield volcanoes, Kilauea and Mauna Loa, there is a rough alternation between eruptions at or near the summit of the mountain and those part way down the flank of the volcano, but all are fissure eruptions. The rift zones extend to the greatest depths exposed by erosion, about 4000 feet, and there consist of hundreds of closely spaced thin dikes, each dike a fissure filled with congealed magma. Presumably the rift zones extend on down beyond visible levels for several miles.

In contrast, the characteristic cones of continental volcanic regions, including such famous examples as Mt. Shasta in California, Mt. Mayon in the Philippines, and Fujisan in Japan, are built primarily by eruptions from a central pipelike conduit, which likewise extends downward as far as can be seen in areas exposed by erosion. Many dikes are present, and flank eruptions occasionally occur, but the dominant activity is eruption from the central vent. Explosive activity is more abundant, and indeed some volcanic cones are composed almost wholly of exploded (pyroclastic) material, though most of the large ones are composite cones composed of interbedded pyroclastic material and lava flows. The greater explosiveness of these volcanoes is partly the result of greater viscosity of the erupting magma, but it is commonly, if not always, also due to a greater amount of gas in the magma as it reaches the surface. Why these volcanoes have a central pipe conduit, instead of the fissure-type conduits of shield volcanoes, is not known, but this

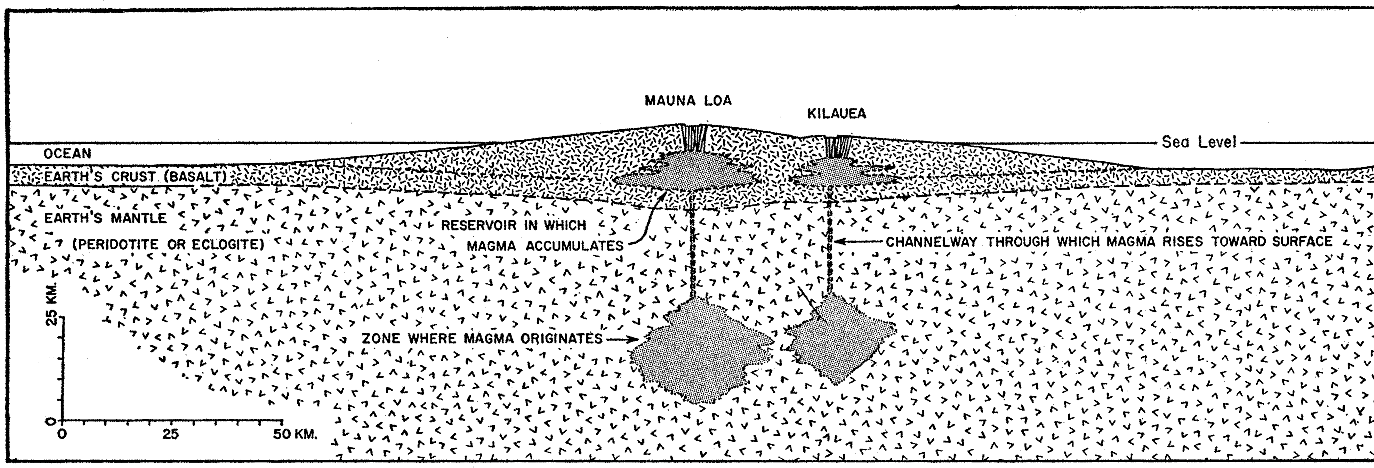

Fig. 2. Hypothetical cross section through Mauna Loa and Kilauea volcanoes, Hawaii, showing the area in which magma is believed to originate and the shallow chamber in which it is stored temporarily before it is erupted to the surface.

may well be the result of a fluxing action by the gas-rich top of the magma body and of coring out of the vent by outrushing gas during explosive eruptions.

Whether the surficial conduit is a fissure zone or a central pipe, it appears almost certain that many, perhaps all, major volcanoes are underlain at a comparatively shallow level by a chamber in which magma is stored temporarily before it is erupted to the surface. The evidence for such a chamber is the tumescence and detumescence of the volcanic mountain that commonly precede and accompany eruption—swelling and shrinking that may cause the mountaintop to move up and down several feet, and that seem explicable only by the inflation and deflation of an underlying chamber. True, the chamber may be far from simple and may be a plexus of compartments partly filled with hot, mushy, magmatic material rather than a well-defined inflatable bladder, but some sort of inflatable structure there must be.

The swelling and shrinking of the mountain is measured partly by ordinary surveying methods and partly by tiltmeters—instruments that measure the change in the angle of inclination of the volcano's side. Some tiltmeters are capable of measuring directly an angle as small as 0.01 second of arc.

Evidence for the depth of the magma chamber has been recognized at a few volcanoes. At Vesuvius, erupting lavas bring to the surface many fragments of dolomite of Triassic age, more or less altered by heat and volatiles from the magma, which are thought to be fragments from the roof of an enlarging magma chamber (*13*). The thickness of the sedimentary sequence in the region is well known, and the strata extend nearly horizontally beneath the volcano, making it possible to place the dolomite, and by implication the top of the magma chamber, approximately 6 kilometers below the general land surface. In western Scotland a quite different sort of evidence has been found. There, at volcanic centers of Tertiary age, thin intrusive sheets of igneous rock occupy upward-diverging conical fractures that appear to have resulted from upward thrust by an underlying magma body. Analysis of the attitudes of these cone sheets leads to the conclusion (*14*) that the tops of the magma bodies lay at depths of about 4 to 7 kilometers. Recent analysis of the pattern of tumescence and detumescence at Kilauea volcano has led J. P. Eaton to conclude that the inflatable chamber there lies at a depth of about 5 kilometers below the summit of the mountain—possibly even above the level of the surrounding ocean floor. In Japan, the changes in magnetism that accompany eruptions of Mihara volcano can be explained on the basis of temperature changes in a spherical mass (presumably a magma body) at a depth of 5 kilometers (*15*). The improbable shape of the magma chamber was assumed merely for the purpose of calculation.

Thus, at quite a few volcanoes, evidence suggests magma chambers at a depth of about 5 kilometers. On the other hand, Rittmann thinks the magma chamber at Ischia may be as shallow as 1 kilometer, and other areas of very closely spaced volcanic vents suggest a feeding chamber—possibly a broad sheet, or sill—of magma at very shallow depth. Locally, as in Tahiti (*16*) and in the West Maui Mountains of Hawaii (*17*), bodies (stocks) of coarse-grained igneous rock exposed by erosion indicate that sizable masses of magma worked their way to within 1 or 2 kilometers of the surface. However, these appear to be offshoots (cupolas) from the main reservoir, or even independent small reservoirs, rather than the main magma chamber itself.

Earthquake shear waves will not travel through liquids. In Kamchatka, the absence of shear waves of certain earthquakes that follow paths of transmission beneath the Klyuchevsky group of volcanoes has led Gorshkov (*18*) to believe that a magma-filled chamber lies at a depth of 50 to 60 kilometers, at the boundary between the earth's crust and mantle or within the outer part of the mantle. This finding may, however, be related to the seat of magma generation rather than to the shallow magma chamber that is responsible for the tumescence of the volcano.

Even less is known about the form of the magma chamber than about the depth. Daly (*19*) long ago suggested that the magma chamber beneath Kilauea might be a laccolith (a lens-shaped body of intrusive igneous rock that bends upward the layers of rock overlying it). It appears that this suggestion may have been a good one, though it was based not on any real evidence but on the intuition of an exceptionally capable geologist. Gorshkov (*18*), on seismological evidence, considers the magma chamber beneath the Klyuchevsky volcanoes to be a flat lens possibly 10 to 12 kilometers thick and with a diameter of the order of 30 kilometers. Only at one place, however, do I know of what may be direct evidence of the shape of the magma chamber that underlay a volcano. In the Messum area, in southwest Africa, deep erosion has exposed a lens-shaped mass of intrusive rock that underlay the ancient volcano at a depth of 4 or 5 kilometers and almost surely served it as a feeding chamber (*20*). Underlying rock layers sagged downward beneath the body, which therefore is classed by the investigators as a "lopolith." The sagging may, however, have taken place very late in the history of the volcano. At any rate, a generally lenticular intrusive body capable of distention and contraction seems to be our best current picture of the shallow magma chamber. The concept is shown diagramatically, but to natural scale, in Figure 2.

Volcanic Engulfments

Calderas are simply unusually large volcanic craters. It appears, though this is not necessarily implicit in the term, that all calderas have been formed by insinking of the summit of a volcanic mountain as a result of removal of support from beneath. The formation of some calderas has been accompanied by the outpouring, on the flanks of the mountain, of large volumes of fluid lava. The birth of others is attended by the explosive ejection of great clouds of pyroclastic material, especially pumice, and huge avalanches of incandescent ash that rush down the mountainside. Volcano-tectonic depressions are still larger basins, also formed by collapse in volcanic regions. As in the case of the second type of caldera mentioned above, the formation of these depressions is accompanied by the eruption of voluminous flows of incandescent ash, commonly still so hot when they come to rest that the glassy fragments become welded together to produce dense rocks that resemble lava flows. The volume of the welded-ash flows (ignimbrites) in the rhyolite plateau of the North Island of New Zealand (the eruption of this plateau attended the subsidence of the basin of Lake Taupo), is about 2000 cubic miles (*21*).

It is a natural first conclusion that the removal of support that brought about the collapse was the result of the copious eruptions of lava or ash. Careful consideration of the volumes involved indicates, however, that this cannot be more than part of the answer. Commonly, if not always, the volume of the depression is much greater than that of the erupted material. Thus, the depression in Kilauea caldera in 1823 had a volume of 539,500,000 cubic meters, whereas the volume of the accompanying lava flow was only about 13,800,000 cubic meters. Similarly, Williams (*12*) has shown that, whereas the volume of the mountaintop engulfed to form the caldera occupied by Crater Lake, in Oregon, was approximately 17 cubic miles, not more than 7.5 cubic miles of material was thrown out in the accompanying eruptions. Material, presumably magma, must indeed have been removed from beneath the depressions to cause the lack of support, but much of it must simply have been shifted within the earth rather than erupted to the surface. Just where it went we do not know. The shiftings may be related to the much greater migrations of material that must take place within the earth during mountain building, isostatic adjustments to changes of surface load, and other up or down movements of large portions of the earth's surface.

Instead of the eruptions bringing about the collapse, it may to some degree be the other way around. The subterranean removal of material, producing the potential void that permits the collapse, may also bring about a marked reduction of pressure on magma rich in dissolved gases. The reduction in pressure may result in rapid separation of the gas and frothing of the magma, with expansion to several times its former volume, which brings about, in turn, explosive eruption of pumice and voluminous outpourings of pumiceous ash at the surface.

Volcanic Water and Other Gases

Water is the chief component of volcanic gases at the earth's surface, generally constituting more than 75 percent, by volume, of all gas collections at volcanic vents. Other common gases include carbon dioxide, carbon monoxide, sulfur dioxide, sulfur, sulfur trioxide, hydrogen sulfide, hydrochloric acid, and ammonium chloride, and in lesser abundance hydrogen, hydrogen fluoride, boric acid, methane, nitrogen, and argon. Not all of these gases are always present, and their relative abundance varies considerably, but water is nearly always predominant and often overwhelmingly so.

The amount of gas present in erupting magma has never been determined accurately, because of the physical difficulties in collecting suitable samples, and estimates vary widely. A limit can probably be placed on the possible amount by the solubility of gases in magma, as determined in the laboratory. Thus, at pressures prevailing at depths of 30 to 40 kilometers, granitic magma can contain up to about 10 percent of water by weight (*22*). The amount that can be dissolved in basaltic magma is less well known but may be about half as much. It is unlikely, however, that magmas at depth are often saturated. Many eruptions produce gas clouds of tremendous volume, and sometimes, as possibly in the 1906 eruption of Vesuvius, the gas may exceed in amount the liquid and solid material ejected. Generally, however, when the amount of gas is estimated carefully and recalculated in terms of percentage by weight, it represents only a very small proportion of the magma. Estimates during recent eruptions of Hawaiian volcanoes range around 1 percent by weight. Similar proportions of gas to lava have been estimated for recent eruptions of Nyamuragira volcano in central Africa, Hekla volcano in Iceland, and Paricutin volcano in Mexico. In spite of the generally small proportion, by weight, of gas to total erupted material, it is probable that throughout geologic ages all of the water on the earth's surface has been produced by volcanoes.

What is the origin of the water? It has been maintained by some volcanologists, notably by T. A. Jaggar, that it is formed, as such, only close to the surface, by the oxidation of deep-seated hydrogen. However, it is at least as likely that the hydrogen in volcanic gas collections is the result of the equilibrium reaction

$$H_2S + 2H_2O \rightleftharpoons SO_2 + 3H_2$$

which shifts toward the right at high temperature and low pressure (*4*). Most volcanologists regard the water as very largely derived from the mantle, along with the other constituents of the magma. There is theoretical reason to believe that all magmas contain at least a little water, even at depth, but this does not mean that all of the water given off by volcanoes originates at depth. Some water may be assimilated from fragments of sedimentary rock picked up by the ascending magma. Some may also be taken into the magma from ground water in surrounding rocks. Though all surface water is probably ultimately of volcanic origin, much of the water released in any one volcanic eruption may have been recycled, even repeatedly. It is hoped that determination of the abundance of isotopes, especially of oxygen, may indicate how much volcanic water originates at depth and how much is of surface derivation.

One of the principal difficulties in the study of volcanic gases has been the reaction that takes place between the gases in the container after collection but before analysis. J. J. Naughton, of the University of Hawaii, is now developing a method by which the gases are separated by means of an absorption column in the field at the time of collection. It is hoped that this will make possible the determination of the actual composition and interrelationships of the gases at the time they arrive at the surface. From this determination, thermodynamic calculations will indicate something of the condition the gases, including water, must have been in under various earlier temperature-pressure relationships within the earth. Thus far, the results seem promising (*23*).

References and Notes

1. J. Verhoogen, *Am. Scientist* **48**, 134 (1960).
2. ——, *Phys. and Chem. Earth* **1**, 17 (1956).
3. G. A. Macdonald, *Bull. volcanol.* **22**, 30 (1959).
4. I. Yokoyama, *Bull. Earthquake Research Inst. Tokyo Univ.* **35** (1955), 75 (1957).
5. T. A. Jaggar, *Volcano Letter Hawaii* **26**, 1 (1925).
6. A. J. Ellis, *Am. J. Sci.* **255**, 416 (1957); G. C. Kennedy, *ibid.* **246**, 529 (1948).
7. H. Kuno, *Bull. volcanol.* **20**, 37 (1959).
8. C. S. Ross, M. D. Foster, A. T. Myers, *Am. Mineralogist* **39**, 693 (1954).
9. J. F. Lovering, *Trans. Am. Geophys. Union* **39**, 947 (1958).
10. G. C. Kennedy, *Am. Scientist* **47**, 491 (1959).
11. H. S. Yoder, Jr., and C. E. Tilley, *Carnegie Inst. Wash. Yearbook* **56**, 156 (1957).
12. H. Williams, *Quart. J. Geol. Soc. London* **109**, 311 (1954).
13. A. Rittmann, *Vulkane und ihre Tätigkeit* (Enke, Stuttgart, Germany, 1936), p. 160.
14. J. E. Richey, *Scotland: The Tertiary Volcanic Districts* (His Majesty's Stationery Office, Edinburgh, 1935), p. 115.
15. T. Rikitake and I. Yokoyama, *J. Geophys. Research* **60**, 165 (1955).
16. H. Williams, *Bishop Museum Bull.* (1933), p. 105.
17. H. T. Stearns and G. A. Macdonald, *Hydrography Bull. Hawaii Div.* **7** (1942).
18. G. S. Gorshkov, *Rept. Acad. Sci. U.S.S.R. Geol. Ser.* **106**, 703 (1956).
19. R. A. Daly, *Proc. Am. Acad. Arts Sci* **47**, 47 (1911).
20. H. Korn and H. Martin, *Trans. Geol. Soc. S. Africa* **57**, 83 (1954).
21. R. L. Smith, *Bull. Geol. Soc. Am.* **71**, 795 (1960).
22. R. W. Goranson, *Am. J. Sci.* **22**, 481 (1931).
23. This article is Hawaii Institute of Geophysics Contribution No. 19.

High School Backgrounds of Science Doctorates

A survey reveals the influence of class size and region of origin, as well as ability, in Ph.D. production.

Lindsey R. Harmon

In the current resurgence of interest in the high school curriculum, major emphasis has been placed on the improvement of science teaching. An increasing concern has been felt for many years regarding the deficiencies of the high school courses of study pursued by most students, from the standpoint of preparation for possible pursuit of scientific studies in college and graduate school. The aim of the present article is to shed some light on this question through an examination of the high school backgrounds of a representative sample of recent science doctorates—specifically, the whole 1958 crop of doctorates from American universities.

This study was made possible by the existence of a file of third-level research degrees from all United States universities from 1936 to the present, maintained currently by the Office of Scientific Personnel of the National Academy of Sciences–National Research Council and supported by grants from the National Science Foundation and the U.S. Office of Education. This file includes doctorates in all fields; in the present study it will be useful to compare the findings on science doctorates with those on doctorate-holders in other fields. Currently, each candidate for a third-level degree fills out a simple one-page questionnaire as he approaches graduation; these completed questionnaires are collected by the deans of the graduate schools and forwarded to the Office of Scientific Personnel. One item on this questionnaire is the name and address of the high school from which the new doctorate-holder graduated. These high school addresses made the present study possible.

The initial use of these high school names was that made by Samuel Strauss, lately of the District of Columbia public school system, who had conducted a small-scale study on his own of the doctorate-level graduates of two nearby universities. He had had a good response from the high schools and sought a wider sample, based on the the Doctorate Records file of the Office of Scientific Personnel. His request for funds from the National Institutes of Health was supported by the Office of Scientific Personnel and backed up by a parallel request to the National Science Foundation from that office itself. Both requests were granted. Strauss undertook a study of the 1957 graduates, and the Office of Scientific Personnel made a study of the 1958 graduates, along practically identical lines. This article is based on the 1958 results.

Last spring, a questionnaire form was prepared for each holder of a 1958 doctorate, to be mailed to his former high school. All forms for a given high school were assembled and sent, together with a letter, to the principal, informing him of the relative standing of his high school in the state and nation with respect to the number of graduates in the 1958 doctorate "crop." For each of its graduates who held a

The author is director of research, Office of Scientific Personnel, National Academy of Sciences–National Research Council, Washington, D.C.