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Magma Beneath Yellowstone National Park

New geological and geophysical data indicate the presence of a batholith, in part still molten.

Gordon P. Eaton, Robert L. Christiansen, H. M. Iyer, Andrew M. Pitt, Don R. Mabey, H. Richard Blank, Jr., Isidore Zietz, Mark E. Gettings

Volcanism is a surficial expression of magmatic systems that continue to deeper levels in the earth. In particular, the formation of very large calderas where rapid voluminous pyroclastic eruptions have caused the volcanic source areas to subside implies that large magma chambers are emplaced at high levels in the earth's crust. The chambers are then partially emptied by gas-driven eruptions, and their roofs subside catastrophically after losing the support of the erupted magma (*1, 2*). In this article we present both geological and geophysical evidence for the existence of a large, shallow body of silicic magma, in part still molten, beneath the Quaternary caldera of the Yellowstone rhyolite plateau, as postulated long ago by Daly (*3*). This body underlies an area more than 85 km long and 55 km wide, and probably is only a few kilometers below the surface and a few kilometers thick. It appears to lie atop an even larger volume of mechanically and thermally disturbed crustal and mantle rocks containing pods of basaltic and silicic magma that extend some 50 km into the mantle. When all of the silicic magma has finally crystallized, it will constitute a granitic batholith.

The evidence presented here was gathered in the U.S. Geological Survey's study of Yellowstone National Park, under the auspices of both an older interdisciplinary Yellowstone research program (*4*) and the present geothermal research program. The data and ideas discussed represent the collective fruits of independent work by the various authors, and we have identified data sources and specific areas of responsibility in the references (*5*). Our purpose in this article is to comment on the interrelations of the various types of data and to offer some preliminary and tentative interpretations. The principal emphases are on how the data support the possible occurrence of silicic magma and on the spatial and temporal relations of the postulated magma to its surroundings.

Late Cenozoic Volcanism: Rhyolite Plateau and Eastern Snake River Plain

The spectacular geysers, hot springs, and fumaroles of Yellowstone National Park are products of a history of intense volcanic activity, mainly of Pleistocene age (*6, 7*). This Quaternary volcanism comprised three cycles of activity, each of which was climaxed by a devastating pyroclastic eruption that produced enormous ash flows and caused collapse of the volcanic source area and formation of a large caldera. Basalts and rhyolites were erupted during each cycle, with rhyolites overwhelmingly predominating. Rocks of intermediate composition are virtually absent. The oldest cycle climaxed about 1.9 million years ago (*7*), forming a caldera that extended from Island Park, 30 km west of Yellowstone National Park (Fig. 1), to perhaps as far east as the central part of Yellowstone. The second cycle was confined to the Island Park area and climaxed about 1.2 million years ago.

The third volcanic cycle, which was responsible for the building of the rhyolite plateau extending eastward from the Madison Plateau to well beyond the Yellowstone River (Fig. 2), began about 1.2 million years ago. Rhyolite flows erupted around the margins of the rhyolite plateau from vents in a fracture system encircling its central region. Younger structural features suggest that these early flows erupted through a pair of growing ring-fracture zones, the centers of which lay along a northeast-trending axis 70 km long. This pattern of early rhyolitic volcanism and generation of ring-fracture zones indicates that a large body of silicic magma was intruding the shallow crust to form a pair of magmatic cupolas [compare with the classical model of Smith and Bailey (*2*)]. Basaltic and subordinate rhyolitic volcanism were active in the plateau region before this central insurgence of silicic magma and have continued intermittently around the margins. However, no later basaltic volcanism has occurred within the area directly above the large silicic magma body.

Approximately 600,000 years ago this pattern of slow ring-fracture generation and episodic extrusion of rhyolitic lavas ended with a catastrophic eruption of rhyolitic pumice and ash totaling more than 900 km^3. Stratigraphic evidence shows that the eruption began from the southwestern of two ring-fracture zones and, before these first ash flows had cooled significantly, was followed by a similar eruption from a paired northeastern ring-fracture zone. These voluminous pyroclastic eruptions removed magmatic support from the apices of the magma chamber and caused the roof to collapse along the two ring-fracture zones. The resulting Yel-

The authors are members of the research staff of the U.S. Geological Survey. Eaton, who coordinates geophysical activities in the Survey's Geothermal Research Program, and Mabey (both exploration geophysicists) are at the Denver Federal Center, Denver, Colorado 80225; Christiansen (a geologist) and Iyer and Pitt (seismologists) are at the USGS, Menlo Park, California 94025; Blank and Gettings (both of the University of Oregon Center for Volcanology) are currently on assignment at the USGS, American Embassy, Jidda, Saudi Arabia; and Zietz (a geophysicist) is at the USGS National Center, Reston, Virginia 22092. Compilation, much of the interpretation and synthesis, and the preparation of the manuscript were primarily the responsibility of Eaton and Christiansen.

lowstone caldera is 70 km long and as much as 45 km wide (Fig. 2).

After the collapse, a resurgent dome (*2*) formed within the eastern caldera segment within a period too short to measure by potassium-argon methods; formation of the dome was followed by the eruption of early postcollapse rhyolitic lavas and tuffs from vents on both ring-fracture zones. Petrographically similar postcollapse rhyolites continued to erupt from the western ring-fracture zone for a period of about 350,000 years. A new magmatic insurgence appears to have begun about 150,000 years ago with doming of the western cauldron block, just east of Old Faithful (Fig. 2). The youngest flows of the Yellowstone region were erupted after this doming, principally in the Madison, Pitchstone, and Central Plateau areas, between 150,000 and 70,000 years ago. Rhyolitic eruptions related to this phase of magmatic activity may not have ended permanently.

The present very active and hot hydrothermal systems probably all date from the youngest magmatic insurgence. It is conceivable that this youngest phase of volcanic and intense hydrothermal activity is actually the early part of a fourth volcanic cycle rather than a continuation of the third. In any case, a major, still-molten magma body probably underlies large parts of the Yellowstone plateau.

The northeast-trending axis of the two segments of the Yellowstone caldera lies along an extended axis that also includes the calderas of the first and second volcanic cycles. This axis, in turn, lies along a northeastward extrapolation of the axis of the eastern Snake River Plain (Fig. 1). Geologic evidence and K-Ar dating of volcanic units along the margins of the plain (*8*) indicate that the eastern Snake River Plain and the Yellowstone plateau volcanic field are closely related and share a similar volcanic and tectonic history that extends back about 15 million years. The evidence indicates that the major focus of volcanism (*9*) has migrated northeastward from the southwestern end of the eastern Snake River Plain to the present area of the Yellowstone rhyolite plateau.

The earliest activity at each newly developing focus of volcanism was the eruption of tholeiitic basalts and small rhyolitic flows and tuffs. Subsequent rhyolitic volcanism climaxed with voluminous ash-flow eruptions, caldera formation, and postcollapse eruption of rhyolitic lavas that recorded the emplacement of large shallow silicic magma bodies. During rhyolitic volcanism at each focus, basaltic volcanism was limited to the flanks of the volcanic system, but after solidification of each silicic batholith, basalts broke through the older volcanic foci and have continued to erupt virtually to the present.

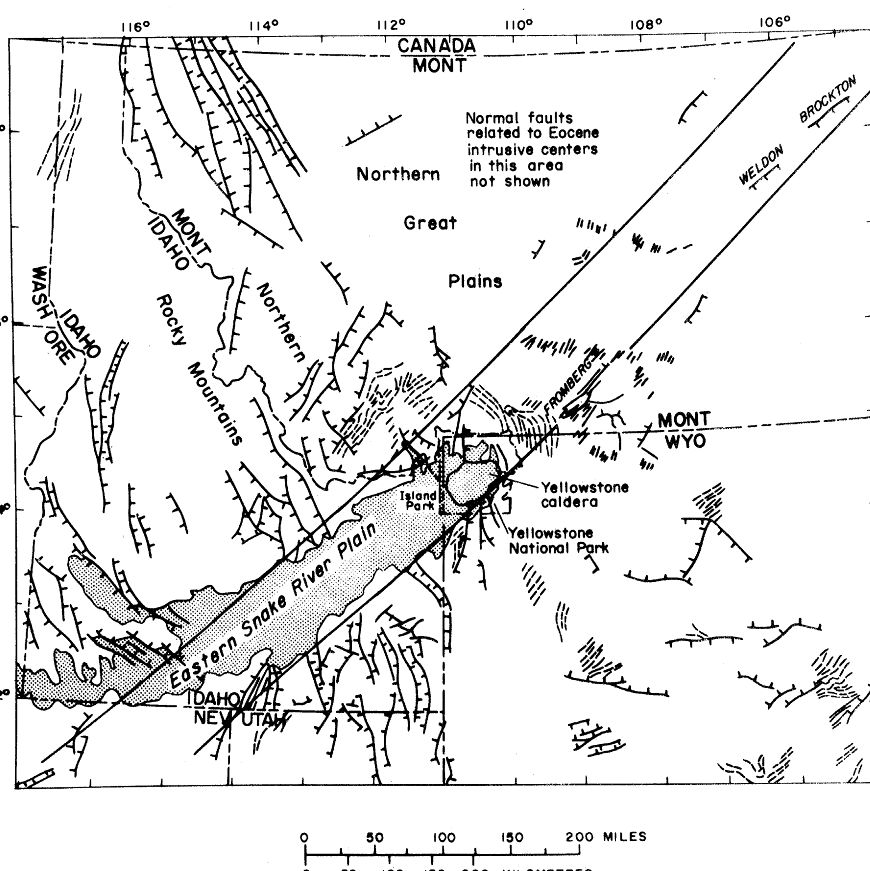

Fig. 1. Regional tectonic map showing relation of Yellowstone National Park region to the eastern Snake River Plain, northern Rocky Mountains, and northern Great Plains. The words *Island Park* mark the site of the caldera of that name. Heavy parallel lines represent boundaries of the regional aeromagnetic anomaly zone shown in Fig. 7. Light, dashed lines show direction of structure in Precambrian basement rocks. Modified from Cohee *et al.* (*30*) and King (*31*).

Geophysical Data

Three fundamentally different geophysical methods (gravimetry, magnetometry, and seismology) were employed to test the hypothesis that silicic magma is present beneath the Yellowstone region. Each provides a unique kind of insight, but the seismic data, utilized in four different forms—seismicity, signal character and attenuation, teleseismic compressional-wave (P-wave) delays, and local absence of shear-wave (S-wave) transmission—are perhaps most conclusive. The region beneath the Yellowstone caldera, extending downward from a depth of a few kilometers, consists of material that (i) has a lower P-wave velocity than its surroundings and, at least in its uppermost part, is less dense and less magnetic than its sourroundings; (ii) is generally incapable of sustaining brittle fracture and does not everywhere transmit S-waves; and (iii) causes anomalous energy losses in elastic waves traversing it. That it also has a very high temperature is inferred from the abundance of fumaroles and boiling springs and the continuing accumulation of siliceous sinter at the surface (*10*).

Gravity

A complete Bouguer gravity map (Fig. 3a) of the Yellowstone National Park region (*11*) is dominated by a large gravity low with a maximum closure of more than 50 milligals and a northeast-southwest elongation. Early definition of this anomaly and its relation to its regional surroundings were provided by Bonini and coworkers (*12*). It defines an areally large, low-density mass which we interpret as a relatively shallow, coherent body of magma beneath the Yellowstone plateau. The gravity data alone do not prove the existence of magma.

In interpreting an earlier set of less complete gravity data from the park, Pakiser and Baldwin (*13*) suggested that the low might be due to any one of the following: (i) a disk-shaped accumulation of rhyolite with gently tapered sides, 3.0 or more kilometers thick; (ii) a local thickening of the low-density silicic upper part of the earth's crust; (iii) a silicic batholith; or (iv) a magma chamber. Malahoff and Moberly

(*14*) later postulated a Quaternary breccia- and rhyolite-filled subsidence basin within basaltic basement rocks as the source of the anomaly.

The caldera rim, mapped on the basis of geologic and topographic criteria, lies close to the locally steep gravity gradient which encloses the principal gravity low in Fig. 3a, except in the northeastern part of the park, where the gradient extends well beyond the caldera rim. The top of this gradient lies almost everywhere along the closed –210 mgal contour or no more than 5 to 10 mgal below it. Throughout most of the map area, this contour skirts the edge of outcrops of the pre-Tertiary basement rocks. The principal exception is in the north-central part of Fig. 3a, where scattered small outcrops of basement rocks lie inside this contour in a belt trending north through Mammoth. (This belt is referred to below as the Norris-Mammoth corridor.) Lower Tertiary volcanic rocks also generally lie outside the –210 mgal contour to the northwest, south, and southeast, but not at the northeast end of the caldera or near Mammoth. Almost all areas of young hydrothermal activity in the park and all of the postcaldera Quaternary rhyolite vents lie within or very near the –210 mgal contour. Apparently, the –210 mgal contour approximately delimits the intrusive source of the rhyolite and related hydrothermal fluids. On the basis of the gravity data alone, this intrusive may or may not be molten.

The steep gravity gradient marks the steeply dipping margins of the low-density mass. The zone of low gravity extending from Norris to Mammoth coincides with a zone of small buried rhyolitic and basaltic intrusions; numerous rhyolite and basalt vents lie along this zone. Part of the elongate gravity low in this area, however, is associated with low-density rocks in a downdropped fault block. North of Mammoth there is a closed local low, poorly controlled on its west side, which could reflect a small shallow satellitic intrusive mass, but more likely is due to a locally thicker sedimentary section.

The fact that the steep gradient lies well beyond the caldera rim at the northeast end of the gravity low, where its closure is greatest, is of critical importance, for it makes it impossible to attribute the main source of the gravity low to a basin of Quaternary breccia or rhyolite fill, although it might be due in part to a structural depression filled with Mesozoic sedimentary rocks (like those exposed east of Mammoth) hidden beneath the Tertiary volcanic rocks. Thick Quaternary rhyolite flows and sediments (Fig. 2) probably contribute a component to the central part of the gravity low, but because they do not extend into this area, they cannot be the principal source of the anomaly.

Calculations based on the steepness of the gradient indicate that the maximum depth to the top of the causative body is 5.5 to 6.0 km; therefore, the anomaly cannot be due to a thickening at the base of the silicic crust. Thus, of the interpretations suggested earlier, only two appear possible: the part of the gravity low not related to either near-surface caldera fill or rhyolite flows reflects either a crystalline or a molten silicic batholith.

On the caldera side of the Lamar River (Fig. 2), which is followed closely by the –210 mgal contour, the Tertiary and Quaternary volcanic rocks are broken by a series of arcuate normal faults that are concentric with respect to the gravity gradient around more than 90° of arc (Figs. 2 and 3a). The youngest movement on these faults is postglacial (*15*), but they were active before caldera collapse as well. Many other faults with a similar trend in the area do not display postglacial movement. This difference in the youthfulness of movement suggests that continuing adjustments have occurred relatively recently in the Cenozoic cover overlying the northeast part of the source of the gravity low. We interpret these adjustments as responses of the roof to movements in a magma chamber.

The "floor" of the gravity low displays an interesting variation. The principal low is divided into two parts by a gentle north-northwest-trending gravity "ridge," two-fifths of the way from the southwest to the northeast end. This gravity ridge coincides with the area between the two geologically defined ring-fracture zones and probably

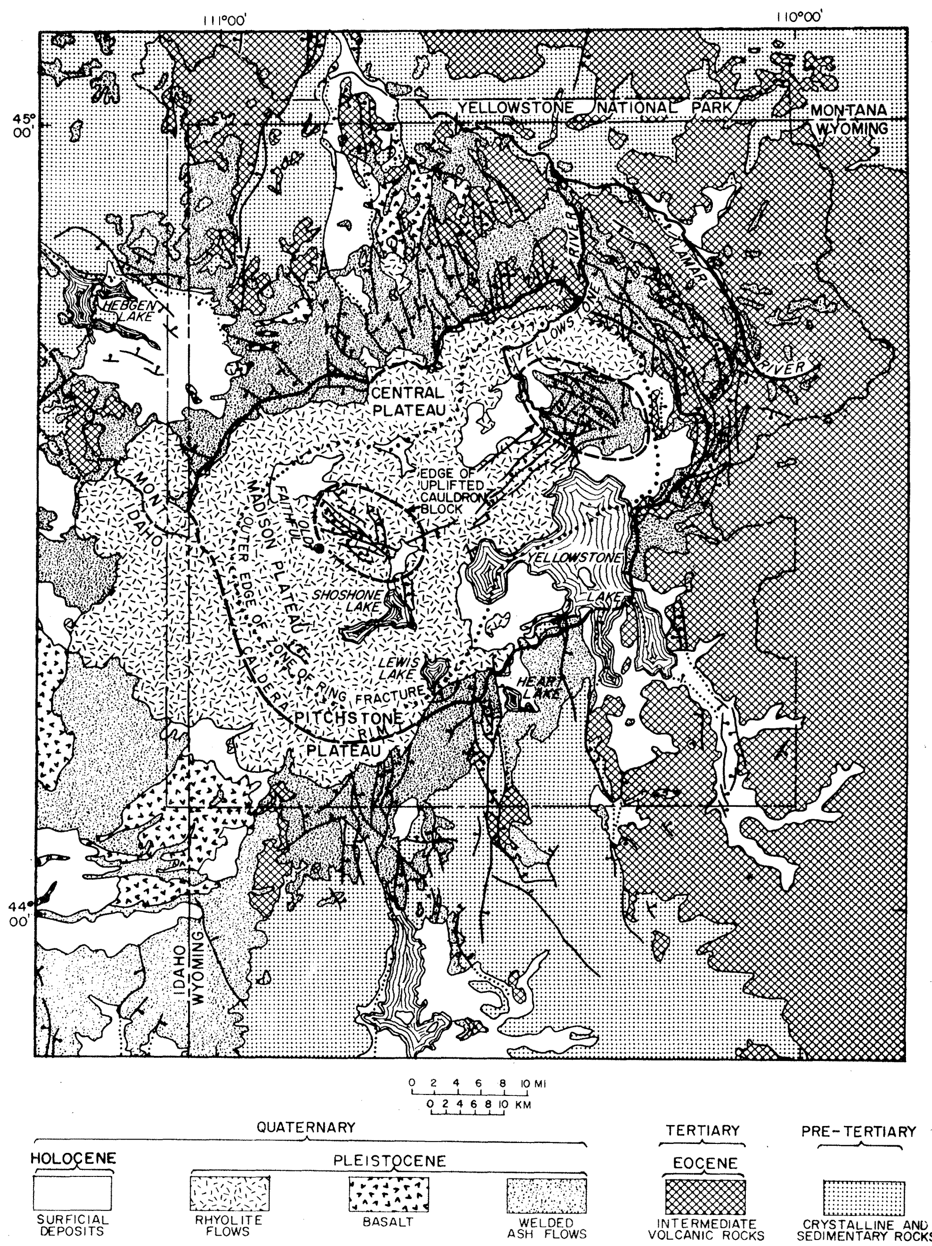

Fig. 2. Geologic map of Yellowstone plateau region showing upper Cenozoic structural features. Caldera rim is dashed where buried. Faults are shown by solid lines with bar and ball on downthrown side.

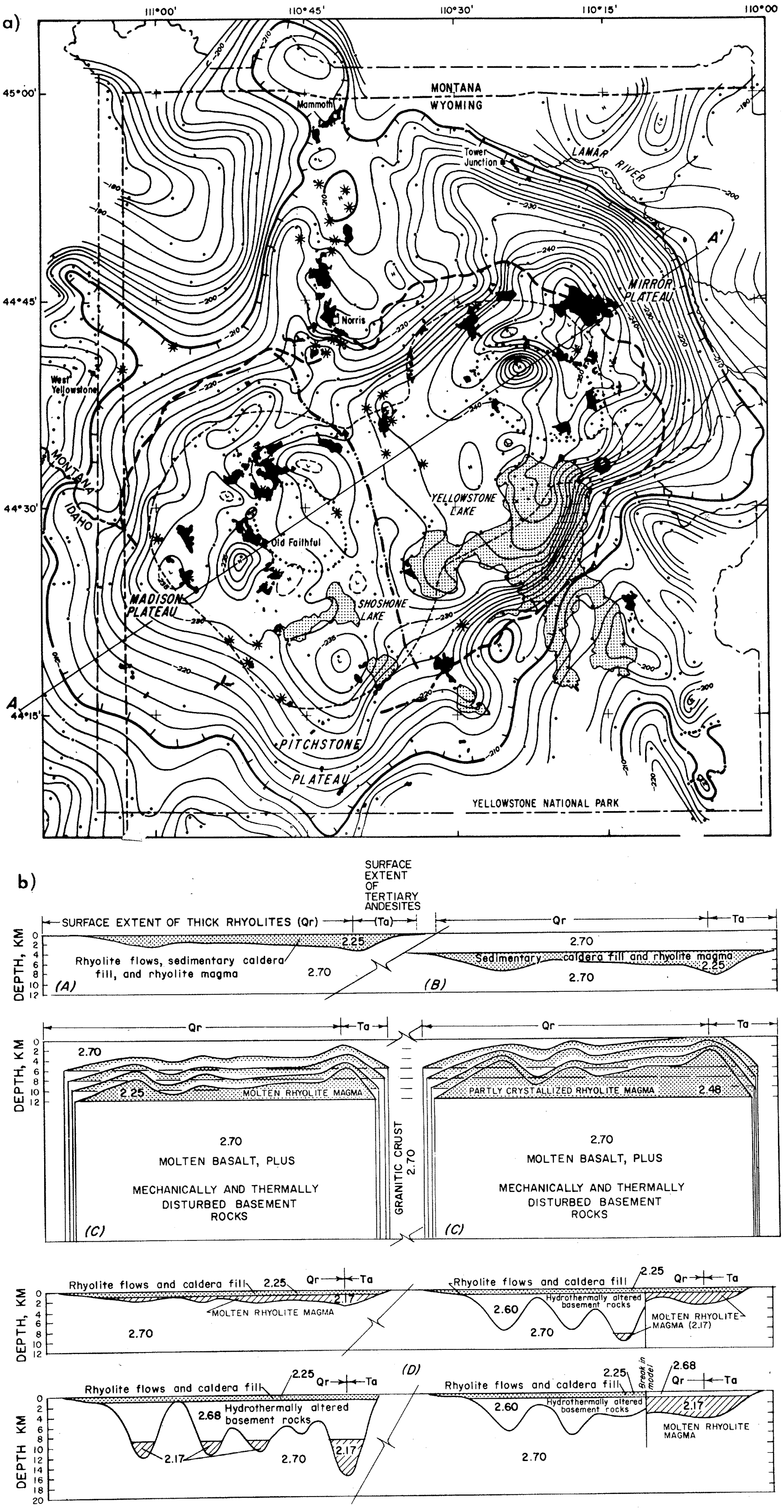

Fig. 3. (a) Complete Bouguer gravity map of Yellowstone plateau region (*11*) and (b) northeast-southwest cross sections through 14 three-dimensional interpretation models whose gravity fields very closely match the observed field. Contour interval, 2.5 mgal. See Fig. 4 for an explanation of the geologic symbols. Models are grouped according to type of source (A, B, C, and D), as discussed in the text. Basaltic magma and disturbed basement rocks are depicted as underlying the rhyolite source body in models of type C in order to accommodate seismic delay data (see Fig. 6). They are also required in models A, B, and D, but are not shown in order to keep the vertical dimension of the figure to a minimum. Abbreviations in (b): ***Qr***, Quaternary rhyolites; ***Ta***, Tertiary andesites.

reflects a broad septum or pendant of denser country rock between two intrusive cupolas. The data are insufficient to warrant quantitative analysis. In the southwestern low, four closed gentle minima with intervening highs lie within the ring-fracture zone, but outside the resurgent dome. These lows may be due to local variations in the thickness of the caldera fill. The largest single area of hydrothermal acid alteration in the park occurs in the area of the lowest gravity minimum, near the northeast end of the caldera. The local gravity closure there may reflect a mass of very high level magma or a crystallized cupola.

We prepared 14 preliminary three-dimensional gravity models (Fig. 3b) whose fields very closely match the observed field. The modeling program used (*16*) requires that either the upper or lower surface of the gravitating body be plane and horizontal. A density of 2.70 g/cm^3 was chosen for the prevolcanic basement rocks, based on their composition. Values of 2.25 and 2.18 g/cm^3 were picked for rhyolite magma (*17*) and an arbitrary value of 2.48 g/cm^3 was used for partly crystallized magma. In some models the near-surface caldera fill and rhyolite flows were assumed to have a density of 2.25 g/cm^3, based on a geologically weighted average of measurements on hand specimens. In others, the fill was assigned this density and the flows a density of 2.70 g/cm^3. In still others, both were assumed to have a density of 2.70 g/cm^3. We do not regard these values as particularly realistic; they merely provided us with modeling convenience. Because the observed gravity field is without question a composite arising from several different materials with contrasting densities (such as rhyolite flows, caldera fill, magma, and granite), a single-body or single-density model such as we have used in models A, B, and C probably represents a substantial departure from reality. Nevertheless, the modeling provides useful constraints.

The models, as shown in Fig. 3b, schematically represent four different source concepts: (model A) a composite body (2.25 g/cm^3) of surface rhyolites, sedimentary caldera fill, and rhyolite magma, with a plane top at the surface of the ground; (model B) a lens (2.25 g/cm^3) of sedimentary caldera fill and magma only, with a planar upper surface at depth beneath denser flows (2.70 g/cm^3); (model C) relatively light rhyolitic magma, or its partly crystalline equivalent (2.25 or 2.48 g/cm^3), beneath denser flows and caldera fill (2.70 g/cm³) and with a planar base resting on a vertical column which has no density contrast with its surroundings (2.70 g/cm³); and (model D) a composite body of rhyolite magma (2.17 g/cm³), overlain by a foundered slab of altered basement rocks (2.68 or 2.60 g/cm³) and capped by light rhyolite flows and sedimentary caldera fill (2.25 g/cm³). Model C was tested for a combination of four basal depths (6, 8, 10, and 12 km) and model D for varying distributions, thicknesses, and densities of the slab of basement rocks. Very briefly, the following generalizations can be made: (i) Regardless of the geologic origin of the source, it is thin relative to its horizontal dimensions. Ignoring values near the edges of the causative body, the minimum thickness for the model in all trials was 0.5 km, and the maximum thickness 9.5 km. (ii) Although rhyolite flows, sedimentary caldera fill, or intensely altered basement rocks more than several kilometers thick could produce the amplitude of the anomaly, they cannot account for its horizontal extent. (iii) A continuous distribution of magma under the whole of the caldera is not required in models of type D. (iv) The plane-based models whose elements show the least root-mean-square difference from elements of the observed field are those with bases at 8 or 6 km.

Magnetics

Figure 4 is a residual aeromagnetic map prepared by subtracting the earth's main magnetic field from the map of observed, total field, magnetic intensity data. It has been shaded in order to make the pattern of intensity variations somewhat clearer. Basically, this map shows an area of relatively low magnetic intensity inside the caldera, surrounded by a belt of generally higher magnetic intensity. The lowest magnetic intensities outside the caldera are associated with Phanerozoic sedimentary rocks. The highest magnetic intensities occur over Tertiary igneous rocks and Precambrian basement rocks, especially those at relatively high elevations. Variations in intensity over the Tertiary volcanic rocks across the northern and eastern part of the map probably result from (i) differences in elevation, (ii) a more magnetic vent facies in the east as opposed to the wider distribution of a relatively less magnetic alluvial facies in the west, and (iii) variations in the polarity of remanent magnetization of rocks at or near the surface.

In general, the areas inside the caldera and in the north-trending corridor from Norris to Mammoth are marked by local, scattered, irregular negative magnetic anomalies that bear a close geographic relation to the areas of present and past hydrothermal activity (*18*). These anomalies may originate in large volumes of rock in the relatively shallow subsurface in which ferrimagnetic minerals have been destroyed by hydrothermal fluids. The hydrothermal areas, in turn, are controlled by the ring-fracture zones and by a linear fault zone (the Norris-Mammoth corridor) related to rhyolite intrusion and extrusion.

The southwestern end of the caldera has an irregular, encircling band of positive magnetic anomalies between the ring-fracture zone and the top of the bounding gravity gradient. These highs lie in large part, but not entirely, over plateau rhyolite flows, some of which have high, measured, normal remanent magnetizations, with Koenigsberger ratios (of remanent to induced magnetization) exceeding 1.0. There do not appear to be any systematic relations to topographic elevation or to individual flow boundaries, and in many places the flows evidently produce little or no disturbance of the field. Similar positive magnetic anomalies lie not far outside the area of the steep gravity gradient to the west and southeast. Depth computations were made for 11 of these anomalies, including the one due north of the northeastern resurgent dome which is essentially the magnetic expression of Mount Washburn, a large massif carved from precaldera Tertiary andesites. Calculated depths to the top of the causative bodies for the other ten anomalies range from 70 to 1400 m, averaging 500 to 600 m. We believe these anomalies to be due to one of two sources: (i) topographic relief on structurally elevated buried Tertiary andesites along the

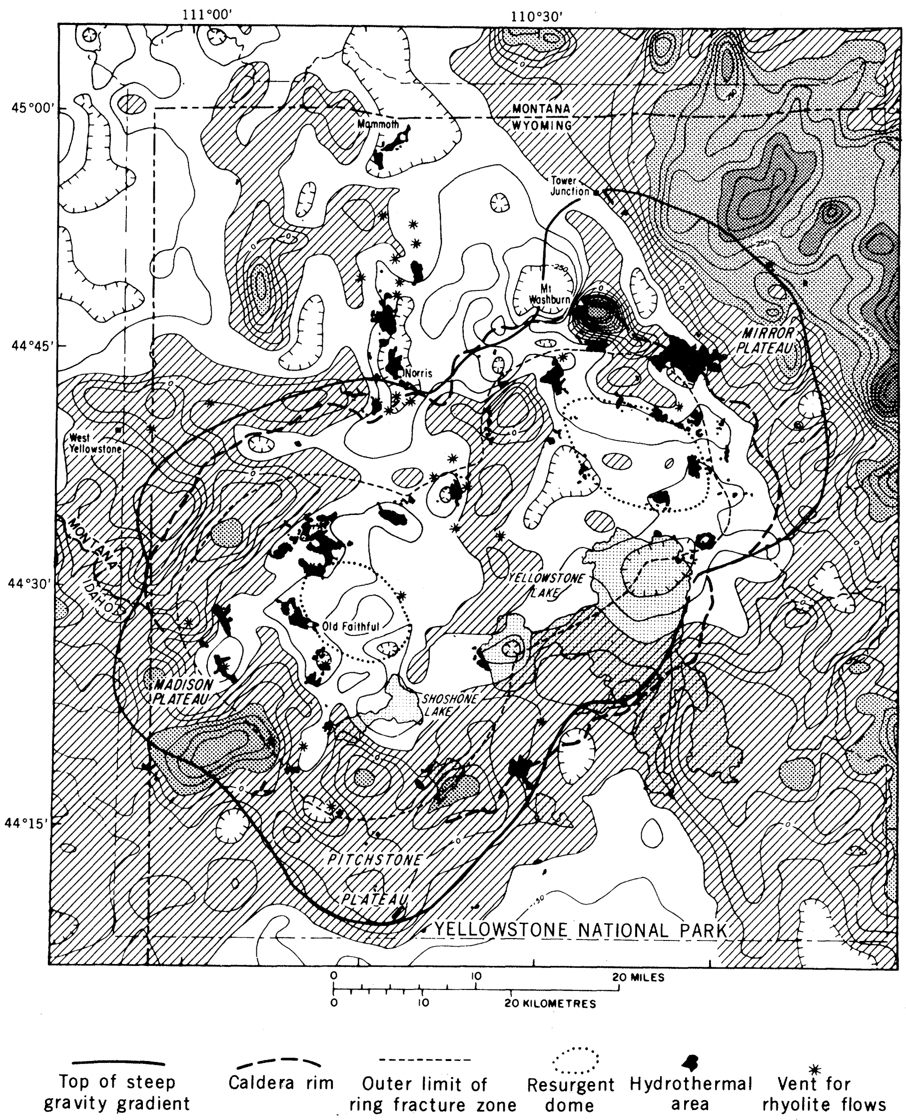

Fig. 4. Residual aeromagnetic map of Yellowstone plateau region (*32*); contour interval 50 gammas (1 gamma = 10^{-5} oersted). Intensity patterns are as follows: dark stipple, 300 to 500 gammas; moderate stipple, 100 to 300 gammas; cross-ruling, –100 to 100 gammas; blank, –300 to –100 gammas. Light stippling marks the major lakes of the park. Flight elevation 3.7 to 4.1 km; flight-line spacing 1.6 km. International Geomagnetic Reference Field removed; data reduced to pole. Prepared from data digitized at 2-km intervals by Smith (*33*).

covered caldera rim, analogous in form and magnetic expression to Mount Washburn, or (ii) younger intrusive rocks in the vicinity of the margin of the caldera and the inferred subjacent magma body. Smith *et al.* (*19*) applied the latter interpretation to our magnetic data in their recently published study.

We suspect that the broad magnetic low in the interior region of the caldera, in addition to locally reflecting discrete bodies of altered near-surface rocks, may be due to a large volume of subsurface material above its Curie temperature. Smith *et al.* (*19*) made an effort to calculate the depth to the Curie isotherm from these data and derived a value of 10 ± 3 km. A somewhat more elaborate, but still preliminary, calculation by two of our colleagues gave a shallower depth inside the caldera and depths of as much as 20 km outside (*20*).

Seismicity

The U.S. Geological Survey has operated photographically recording short-period seismograph stations in the Yellowstone region since 1963. At the end of 1972, the network was upgraded with the installation of a telemetered network with central recording on 16-mm film at Mammoth. At present, the network consists of 12 vertical and 2 horizontal seismometers.

Figure 5 shows epicenters located from late 1964 to 1974. The largest number of earthquakes has occurred immediately east of Hebgen Lake, and activity decreases to the south and east; no earthquake activity has occurred in the northeastern part of the park. The relatively large number of events in the northwestern part of the area in part reflects the concentration of stations in that area, but the events of largest magnitude (and therefore greatest strain release) also occur just east of Hebgen Lake, which indicates that the earthquake distribution is not biased severely by the distribution of recording stations. All well-determined focal depths are less than 15 km, but Smith *et al.* (*19*) noted that within the northwestern part of the caldera itself, the maximum focal depth is only 5 km. The earthquakes commonly occur in swarms of up to several hundred events within a period of a few hours to a few days in a region less than 3 km across.

The general distribution of the epicenters in Fig. 5 is similar to that shown by Smith *et al.* (*19*), although because of a much longer period of observation, we have recorded a greater number of events. The zone of abundant activity extending eastward to east-southeastward from Hebgen Lake ends sharply at the west edge of the Norris-Mammoth corridor. Little activity was recorded within the caldera itself and none in the northeastern sector, where the local amplitude of the gravity low is greatest. There are many young faults in this area. The absence of seismicity in the area of demonstrably young faults may be due to (i) an observational time period inadequate to detect episodic swarms, (ii) thinness of the capping precaldera rocks above the postulated magma chamber, leading to extremely shallow foci and events that are difficult to recognize, or (iii) strain release by creep due to high temperatures at the base of these rocks above a high-level magma chamber. The absence of seismicity in other areas outside the caldera where fault swarms have been mapped geologically is attributed to cessation of tectonic activity sometime in the past.

Some earthquakes within the caldera are spatially associated with the sinuous axis of the weak gravity ridge between the major subsided areas of the two ring-fracture zones. Earthquakes immediately to the north, but south of station N, are associated with one of the youngest rhyolite vents in the region. Most other earthquakes within the caldera occur in areas of major hydrothermal activity.

Seismic waves from local earthquakes that travel through the central part of the Yellowstone plateau show significant attenuation and change of character. P-waves passing through this part of the caldera have longer periods and more emergent first arrivals (less sharply defined onsets) than P-waves traveling other paths. In addition, S-waves are absent on some records and difficult to distinguish on others. All of these characteristics signify a local geologic body with properties different from those of its surroundings. Some of them suggest the presence of material in the fluid state. It has been noted elsewhere (*21*) that these phenomena may indicate the presence of molten material, and this is the interpretation we apply here. The approximate area of this anomalous zone is shown by the stipple pattern in Fig. 5. Not all seismic waves passing through the zone exhibit these changes in character. The changes are not likely to be due to source effects as they have been observed for a variety of travel paths, including reversed travel paths. Some travel paths between event and station may be circuitous, skirting the anomalous zone altogether, even though the zone may lie directly between them; thus, the zone is not regarded as closely controlled. Depths to the top and bottom of the attenuation zone have not yet been determined.

Smith *et al.* (*19*) presented a map of 11

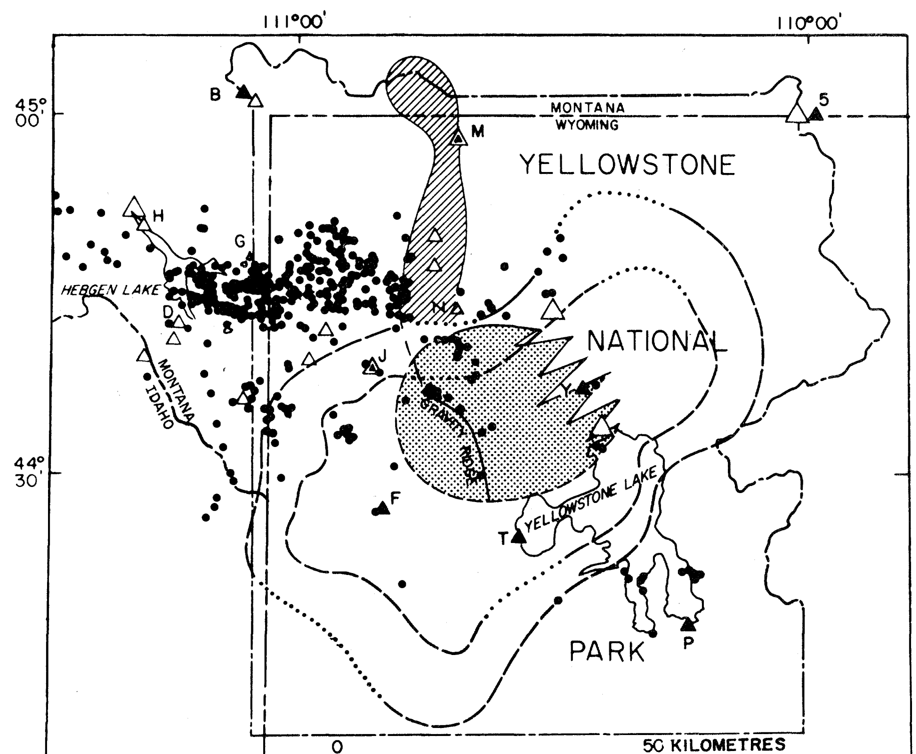

Fig. 5. Seismicity and seismic attenuation in Yellowstone plateau region. Triangles identified by letters indicate seismograph stations operated from 1963 to the present (open triangles, 1963 to 1972; solid triangles, 1972 to the present). Circles denote earthquake epicenters. The stippled area defines a zone of observed attenuation for compressional seismic waves from local earthquakes; shear waves are absent on some records for paths through this zone. Dashed curves denote top and bottom of local gravity gradient; dotted curves mark interrupted portions. Cross-ruling defines the Norris-Mammoth corridor.

fault-plane solutions for the park region. Northwest of the caldera, north-south regional extension is indicated, but in a zone immediately peripheral to the caldera the solutions indicate radial compression, which is suggestive of possible dilation of a magma chamber. Although the data are only permissive, they lend support to our hypothesis that molten magma still exists beneath the park.

Teleseismic P-Wave Residuals

Teleseismic P-waves recorded by the Geological Survey's seismic network in Yellowstone show large delays (Fig. 6a). About 60 teleseisms, primarily from northwest, west, and southeast directions, and 5 man-made nuclear explosions from the north, were used to study the variation of P-wave residuals. A theoretical travel time (*22*) was subtracted from the observed travel time to compute P-wave residuals for each event-station combination. Because of operational problems and high seismic noise levels in some areas, the number of available P-wave arrival times varied from station to station. About 50 readings each were available at stations D, B, S, M, F, and G; 20 each at stations J, N, P (for some azimuths), and H; and fewer than 10 each at stations P (for other azimuths) and Y.

Individual station differences appeared when the P-wave residual data at each station were divided into four groups, based on the azimuth of events (north, northwest, west-southwest, and southeast). For example, station F, in the vicinity of Old Faithful, showed about 4.0-second residuals for all azimuths, whereas station S, near the northeast corner of Yellowstone National Park, showed only 2.6-second residuals. The residuals at stations N, Y, and T were similar to those at F, but the other locations showed a pattern of azimuthal variation. In order to bring out clearly the relative variation of residuals within the region and to eliminate source effects and path effects outside the region, station S was chosen as a reference datum. Residuals at S were subtracted from the residuals at the other stations for each event. These resulting values are referred to as relative residuals and their average for the four azimuths are shown in Fig. 6a. Where fewer than three values were used to compute the average, the uncertainty is indicated by an asterisk after the value. A more complete account of these data and their interpretation has been presented elsewhere (*23*).

The most obvious feature of the pattern of relative residuals is the large positive delay at stations F, N, Y, and T for all azimuths. Note that the relative delays reach values as high as 2.2 seconds at T. All the stations in the northwest and north section of the diagram show a large delay (0.8 to 1.1 second) for the southeast azimuth only. The value of 0.9 second at station M identifies a ray path passing upward across the northeastern part of the delay-causing body. This observation is significant because no source azimuths lie to the northeast, making the definition of the northeast boundary of the body difficult.

A preliminary interpretation of this pattern, which is consistent with the volcanic history of the park and with the other geophysical data, is that a substantial volume of low-velocity material exists under the caldera. The upper part of the material is probably fairly close to the surface (that is, within a few kilometers) in the central region, which includes stations N, Y, and F, where the delays are high for all azimuths. That the low-velocity material penetrates the entire crust and deep into the upper mantle can be inferred from the 1-second southeast delays at stations D, B, G, and H. The teleseismic waves (which emerge at the stations at approximately 25° to the vertical) must penetrate the low-velocity material about 100 km under the caldera. This fact, coupled with the large magnitude of the delays, virtually eliminates the possibility of explaining the delays by surficial phenomena, such as the presence of sediments, or by a varying depth to the Moho (the discontinuity that separates the earth's crust from the mantle).

Although it would be premature to propose a definite model for this relatively

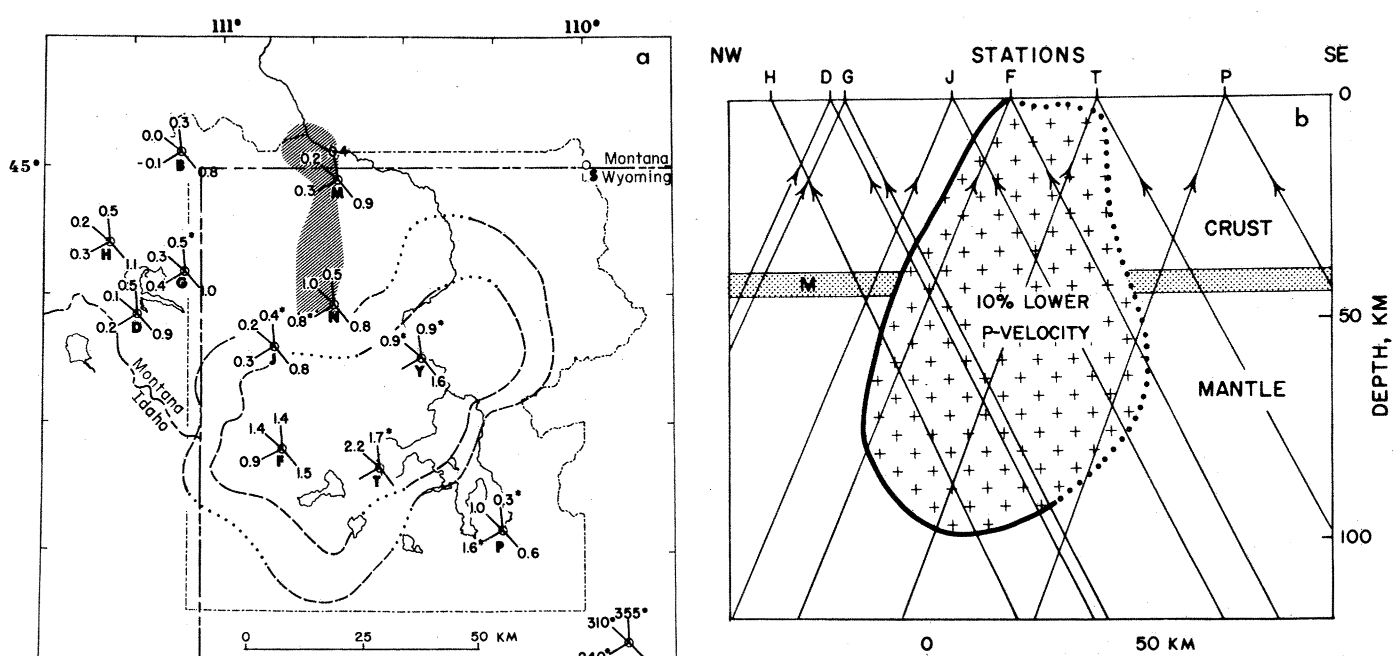

Fig. 6. (a) Map showing seismic network and average teleseismic P-wave residuals (in seconds) relative to station S, for distant events along azimuths north, northwest, west-southwest, and southeast of the network. Individual residuals are based on 20 to 50 readings, except those marked by *. Long dashed lines define the top and bottom of the steep gravity gradient; dotted curves mark interrupted portions. Cross-ruling defines the Norris-Mammoth corridor. Figures in the lower right corner indicate azimuths of arrivals at network measured clockwise from the north. (b) Northwest-trending vertical cross section of the hypothetical low-velocity volume creating the pattern of observed P-wave residuals. The heavy solid and dotted lines locate the boundary of the low-velocity material with varying degrees of assurance. Lines with arrows show schematic ray paths for events from northwest and southeast azimuths. The width of the shaded band *M* (Moho) indicates the uncertainty in crustal thickness (*34*).

low-velocity body, one can nevertheless determine its general size and shape. For example, assuming that P-wave velocities inside the body are about 10 percent lower than those outside, we obtained for a northwest vertical cross section the geometry shown in Fig. 6b. Only data from stations P, T, F, J, G, D, and H were used in the modeling. The only boundary of the anomalous body that is delineated with certainty is that between F and J. For events from the north, the delay decreases from F to N, indicating that the body does not extend far in that direction.

If the velocity within the delay-causing volume is assumed to be 20 percent lower instead of 10 percent, the resulting shape and size of the model volume make it more difficult to explain the high delays observed for deeply penetrating southeasterly rays at stations D, G, B, and H. We estimate that the velocity decrease within the anomaly-causing body must be in the range of 5 to 15 percent, and it probably has a nonuniform distribution. The implications of such a decrease are important both in general geological terms and in relation to the interpretation of our other data. There is little or no adequate experimental information on the effect of elevated temperatures on P-wave velocities in molten lava or magmas, volcanic glasses, rock-melt mixtures, and their crystalline equivalents. The study of Murase and McBirney (*17*) provides a general idea of the effects of elevated temperatures, but because of the limitation of their measurements to samples of but a few compositions, and to atmospheric pressures only, it is not possible to construct an unequivocal argument. Interpolation of their experimental data indicates that P-wave velocities in intermediate and mafic rocks decrease by approximately 60 percent when the temperature increases from ~20° to 1400°C. A 10 percent decrease in velocity, similar to that which we have calculated, occurs in the range from ~20° to 550°-650°C. A decrease of like magnitude also occurs at approximately 1000°C with increases in temperature of as little as 40° to 50°C. Thus it is not possible to determine whether the anomalous body (in its entirety) consists of magma, partial melt, some form of hot "glass," or crystalline rock at greatly elevated temperatures.

It is notable that the difference in velocity between the anomalous body and its surroundings below a depth of 10 to 20 km (depending on the model chosen) is not reflected in the gravity data. Experimental measurements (*17*) indicate that elevating the temperatures of lavas at atmospheric pressure from ~20° to 1400°C results in a decrease in density of only 4 to 7 percent. At 1000°C, an increase in temperature sufficient to lower P-wave velocities by 10 percent results in a decrease of density of less than 0.5 percent. Thus, the seeming absence of a detectable mass anomaly below a level of 10 to 20 km is not cause to assume that magma does not extend to deeper levels. We interpret the delay-producing body as a combination of a shallow silicic magma body; a throughgoing, deep, possibly partly molten "root" containing pods or streaks of both basaltic and silicic magma; and the mechanically and thermally disturbed crustal and mantle rocks surrounding them.

It has been proposed (*24*) that the Yellowstone region lies above a convection plume rising from a hot lower mantle. Such plumes have been visualized as pipes 100 to 200 km in diameter that extend to the base of the upper mantle. The horizontal dimensions of the anomalous low-ve-

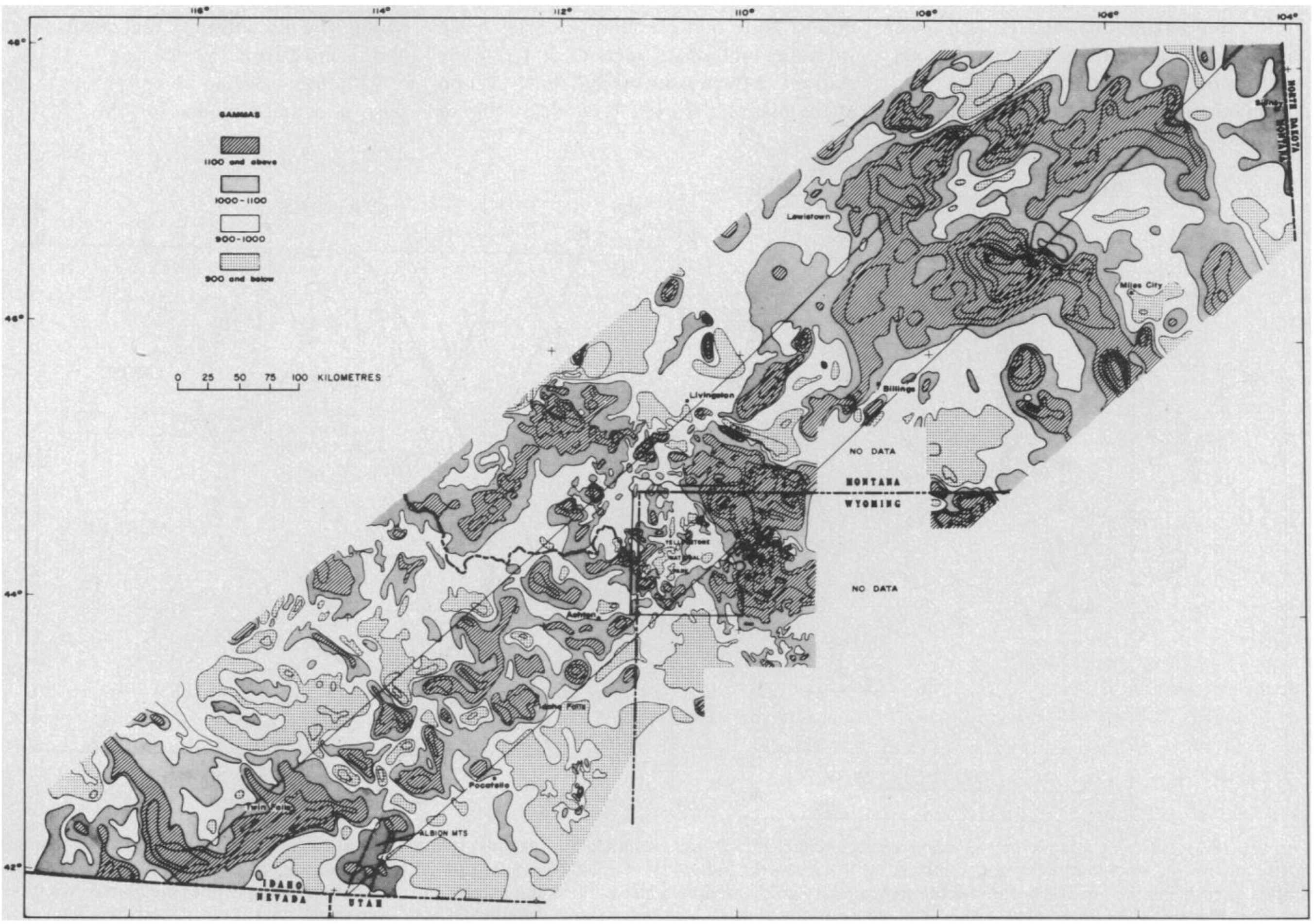

Fig. 7. Regional residual aeromagnetic map of portions of the Snake River Plain, northern Rocky Mountains, and northern Great Plains provinces. Long smooth curves define magnetic boundaries of the Snake River Plain and their extrapolation into the northern Great Plains. They are also shown on Fig. 1.

locity material under Yellowstone seem to fit those of a plume, but additional experiments are required to distinguish a magma chamber and its root, extending to a zone of magma generation, from a deep mantle plume. A test to distinguish the two requires an estimation of the depth of mantle penetration of the low-velocity material. With our present seismic array, any maximum depth estimate is limited to about 100 km. Data from an extended array using permanent and portable instruments are available but remain to be analyzed.

Regional Geophysical Setting—Some Speculations

We have argued on geologic grounds that the Yellowstone volcanic activity represents the latest embryonic development of an extended segment of the Snake River Plain. How are the geophysical data related to this idea?

Regionally, the eastern Snake River Plain is marked by an extensive, coincident gravity high (*11, 25*), clearly related to the relatively low topography of the plain and probably reflecting some sort of isostatic compensation. In testing this concept, Mabey (*26*) found excellent agreement between the amplitude and configuration of the anomaly observed there and those of a hypothetical anomaly computed simply by multiplying the average elevation in areas 64 km in radius around each station by the Bouguer correction coefficient. Turning to the Yellowstone region, examination of the topography shows that elevation increases abruptly as the plateau is approached from the north, but when averaged over circles of radius 64 km the regional elevation is remarkably uniform along the park portion of the profile and immediately to the south. Comparison of these elevation data with the gravity values suggests that on a regional basis the park area has an average elevation approximately 400 m too high for its present gravity field. Thus, the relation between elevation and gravity of the rhyolite plateau is dissimilar to that of the plain. Similarity could be achieved, however, by 400 to 700 m of subsidence relative to regions to the north and south and an increase in density at depth, which would convert the average residual gravity low over the plateau to values like those observed in the eastern Snake River Plain, approximately 100 mgal higher. Crystallization of a large mass of molten material beneath the park would accomplish both. A 4-km-thick mass solidifying with a decrease in volume of 10 percent (a value perhaps too high) would produce a subsidence of approximately 400 m and would increase the Bouguer anomaly values by about 100 mgal. Flooding of the subsided area by basalt would augment both processes.

The magnetic expression of rocks of the Yellowstone area is seen in its regional context in Fig. 7. From this broad perspective, the area lies on a northeast-trending magnetic lineament, or belt, composed of magnetic highs that extend from north-central Nevada into Canada.

Southwest of the park, most of the magnetic anomalies in this belt occur over Cenozoic basalt, and some seem to be clearly related to the basalt. However, near the southern border of Idaho, one of the positive anomalies in the belt occurs over exposed Precambrian rocks in the Albion Mountains—a point of some significance, as will be seen below.

In several places southwest of Yellowstone, two sets of anomaly lineations can be distinguished within the belt, one parallel to its long axis and the other normal to it. The latter set approximately parallels Basin and Range structural trends. Likewise, relatively short wavelength local magnetic highs occur parallel to the southeast and southwest sides of the broad low of the Yellowstone plateau. They are thought to express a fundamental structural skeleton within the Snake River Plain and the Yellowstone plateau. They may reflect structural intersections of the type that localized older volcanic foci along the Snake River Plain–Yellowstone axis. Zietz *et al.* (*27*) noted the presence of these two lineament directions in much of the northwestern United States and stated that "Their regional distribution, overall magnetic character, and geologic evidence suggest that they [the magnetic lineaments] are major structural features in the basement rocks. . . ."

The character of the magnetic belt changes northeast of Yellowstone, but some large positive magnetic anomalies parallel to the magnetic trend of the eastern Snake River Plain continue into the Northern Great Plains of central and eastern Montana. A broad complex positive anomaly of high amplitude intervenes immediately northeast of the Yellowstone plateau, associated with elevated older Tertiary volcanic rocks and Precambrian crystalline rocks. The arbitrary boundaries drawn for the magnetic belt along the Snake River Plain have been extended northeast of the park into the Great Plains in Fig. 7 and enclose northeast-trending elongate anomalies there. The southeast boundary is coincident with the Fromberg fault and lies parallel to, and just southeast of, the Weldon and Brockton-Froid fault zones (Fig. 1). The last of these fault zones offsets Pleistocene deposits of Wisconsin age (*28*), a surprising phenomenon for an area in the stable interior of the continent. As a group, these faults mark a major boundary at which regional northwesterly magnetic trends on the southeast are separated from northeasterly magnetic trends on the northwest. Stratigraphic data for the Phanerozoic rocks of the region indicate that movement on many basement faults has been recurrent in Paleozoic, Mesozoic, and Cenozoic time (*27*).

Although two of the largest highs in the belt under discussion have amplitudes greater than any of the positive anomalies southwest of the park, the general level of magnetic intensity and the width and trend of the belt are approximately the same in both places. Because the belt is parallel to the regional structural grain of exposed Precambrian rocks throughout much of the Rocky Mountains and Colorado Plateaus, we believe it is related to a fundamental, reactivated Precambrian structure, much as has been suggested for young northeast-trending structures of great length on the Colorado Plateaus (*29*) and throughout much of the rest of the northwestern United States (*27*).

Digital modeling of local magnetic anomalies of the Snake River Plain indicates they can be accounted for theoretically in terms of the magnetic properties and thicknesses of the Cenozoic basalts of the plain. Obviously, a different origin must be invoked for the individual anomalies along the extrapolated belt in the northern Great Plains. We believe that they are related to lithologic units in the Precambrian basement. The Albion Mountains anomaly supports this argument. Whatever their lithologic nature, we believe that such units may well have influenced the emplacement of the Cenozoic rhyolites and basalts of the Snake River Plain, in addition to those of the Yellowstone region.

Summary

The Yellowstone plateau volcanic field is less than 2 million years old, lies in a region of intense tectonic and hydrothermal activity, and probably has the potential for further volcanic activity. The youngest of three volcanic cycles in the field climaxed 600,000 years ago with a voluminous ash-flow eruption and the collapse of two contiguous cauldron blocks. Doming 150,000 years ago, followed by voluminous rhyolitic extrusions as recently as 70,000 years ago, and high convective heat flow at present indicate that the latest phase of volcanism may represent a new magmatic insurgence. These observations, coupled with (i) localized postglacial arcuate faulting beyond the northeast margin of the

Yellowstone caldera, (ii) a major gravity low with steep bounding gradients and an amplitude regionally atypical for the elevation of the plateau, (iii) an aeromagnetic low reflecting extensive hydrothermal alteration and possibly indicating the presence of shallow material above its Curie temperature, (iv) only minor shallow seismicity within the caldera (in contrast to a high level of activity in some areas immediately outside), (v) attenuation and change of character of seismic waves crossing the caldera area, and (vi) a strong azimuthal pattern of teleseismic P-wave delays, strongly suggest that a body composed at least partly of magma underlies the region of the rhyolite plateau, including the Tertiary volcanics immediately to its northeast.

The Yellowstone field represents the active end of a system of similar volcanic foci that has migrated progressively northeastward for 15 million years along the trace of the eastern Snake River Plain (*8*). Regional aeromagnetic patterns suggest that this course was guided by the structure of the Precambrian basement. If, as suggested by several investigators (*24*), the Yellowstone magma body marks a contemporary deep mantle plume, this plume, in its motion relative to the North American plate, would appear to be "navigating" along a fundamental structure in the relatively shallow and brittle lithosphere overhead. The concept that a northeastward-propagating major crustal fracture controls the migration path of the major foci of volcanisim is at least equally favored by existing data, as Smith *et al.* (*19*) noted.

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Sex Differentials in the Academic Reward System

What changes have there been since the implementation of federal antibias regulations?

Alan E. Bayer and Helen S. Astin

In the last half decade several major laws and regulations concerning sex discrimination in college and university faculties have become effective (*1*). The first is Executive Order 11246, as amended by 11375, which prohibits discrimination in employment (including hiring, upgrading, salaries, fringe benefits, training, and other conditions of employment) on the basis of sex and certain other factors and requires an affirmative action plan (including numerical goals and timetables) of all federal contractors. A second is Title VII of the Civil Rights Act of 1964 as amended by the Equal Employment Opportunity Act of 1972, which likewise prohibits discrimination in academic employment. A third, the Equal Pay Act of 1963 as amended by the Education Amendments of 1972 (Higher Education Act), specifically prohibits discrimination in salaries and covers nonacademic as well as academic employees. Title IX of the Education Amendments of 1972 reaffirms the compliance regulations of the earlier orders, extends coverage to part-time employees, and requires that equal pensions for men and women employees shall be determined.

In this article we present estimates on a national scale of current sex differentials in academic employment and of the extent to which equity has been approached since antibias regulations have been in effect. Many studies of sex discrimination in aca-