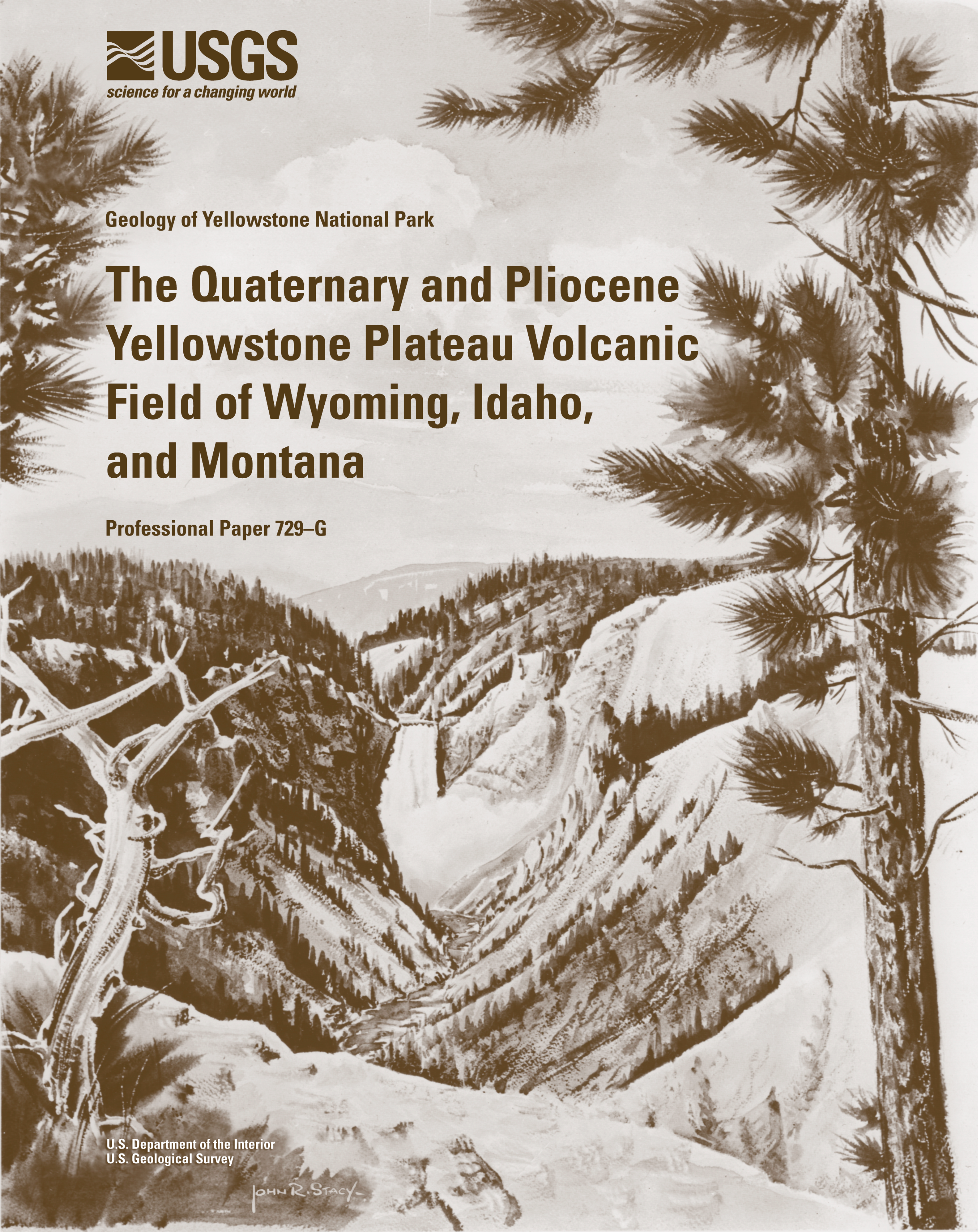

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Geology of Yellowstone National Park

The Quaternary and Pliocene Yellowstone Plateau Volcanic Field of Wyoming, Idaho, and Montana

Professional Paper 729–G

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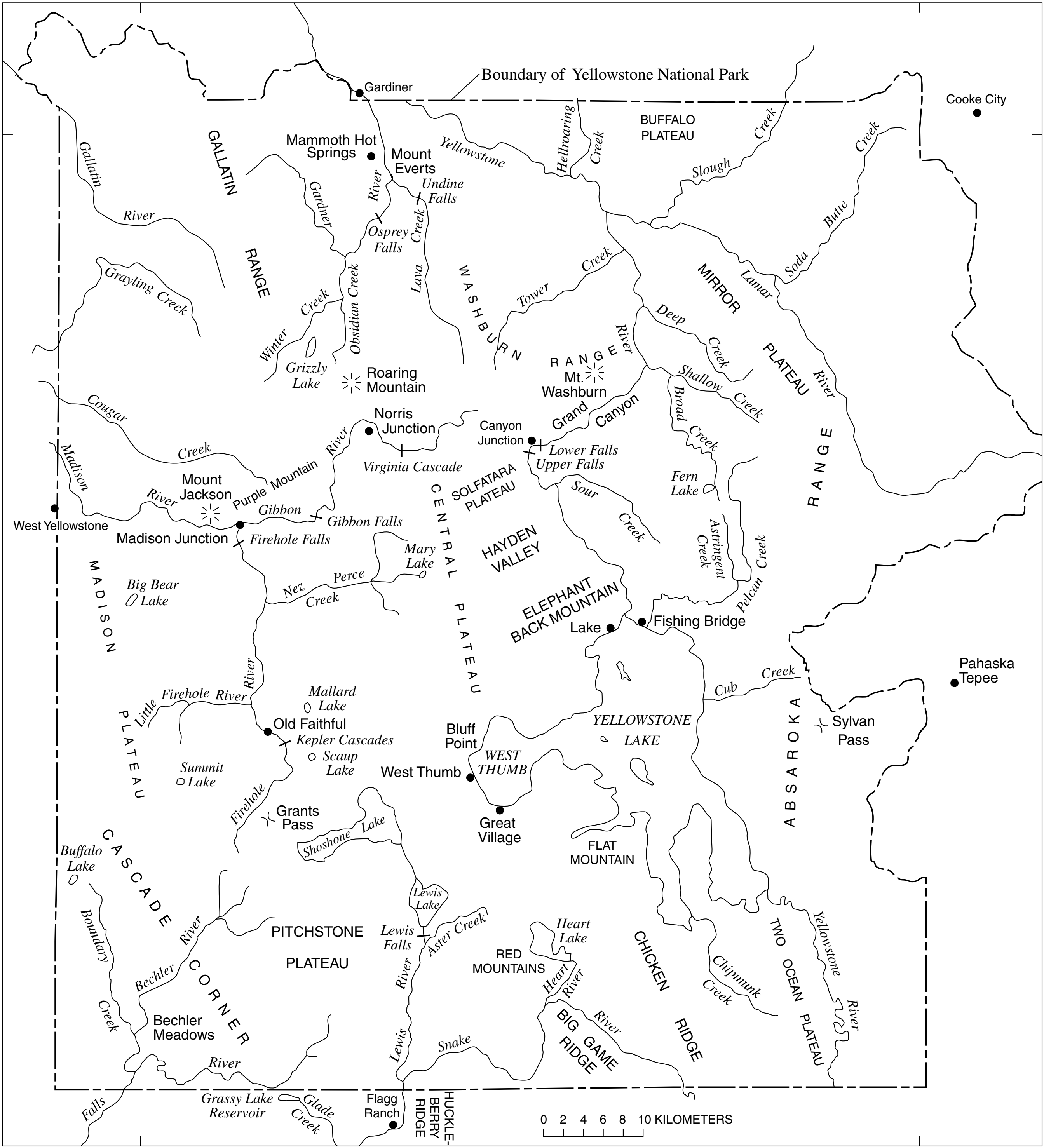
111°
110°
45°
Boundary of Yellowstone National Park
Gardiner
Cooke City
BUFFALO PLATEAU
Mammoth Hot Springs
Mount Everts
GALLATIN RANGE
Gallatin River
Yellowstone River
Hellroaring Creek
Slough Creek
Soda Butte Creek
Gardner River
Undine Falls
Osprey Falls
Lava Creek
WASHBURN RANGE
Tower Creek
MIRROR PLATEAU
Lamar River
Grayling Creek
Winter Creek
Obsidian Creek
Grizzly Lake
Roaring Mountain
Mt. Washburn
Deep Creek
Shallow Creek
Broad Creek
Grand Canyon
Canyon Junction
Lower Falls
Upper Falls
Norris Junction
Virginia Cascade
Cougar Creek
Madison River
Purple Mountain
Mount Jackson
Gibbon River
Gibbon Falls
West Yellowstone
Madison Junction
Firehole Falls
SOLFATARA PLATEAU
CENTRAL PLATEAU
HAYDEN VALLEY
Sour Creek
Fern Lake
Astringent Creek
Pelican Creek
RANGE
Mary Lake
Nez Perce Creek
MADISON PLATEAU
Big Bear Lake
ELEPHANT BACK MOUNTAIN
Lake
Fishing Bridge
Pahaska Tepee
Cub Creek
Firehole River
Little Firehole River
Mallard Lake
Old Faithful
Kepler Cascades
Scaup Lake
Summit Lake
Bluff Point
WEST THUMB
West Thumb
YELLOWSTONE LAKE
Sylvan Pass
ABSAROKA
Grants Pass
Great Village
Shoshone Lake
FLAT MOUNTAIN
Buffalo Lake
CASCADE CORNER
Boundary Creek
Lewis Lake
Lewis Falls
Aster Creek
PITCHSTONE PLATEAU
Heart Lake
RED MOUNTAINS
Heart River
CHICKEN RIDGE
TWO OCEAN PLATEAU
Chipmunk Creek
Yellowstone River
Bechler River
Bechler Meadows
Lewis River
BIG GAME RIDGE
Snake River
Falls River
Grassy Lake Reservoir
Glade Creek
Flagg Ranch
HUCKLEBERRY RIDGE
0 2 4 6 8 10 KILOMETERS

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By Robert L. Christiansen

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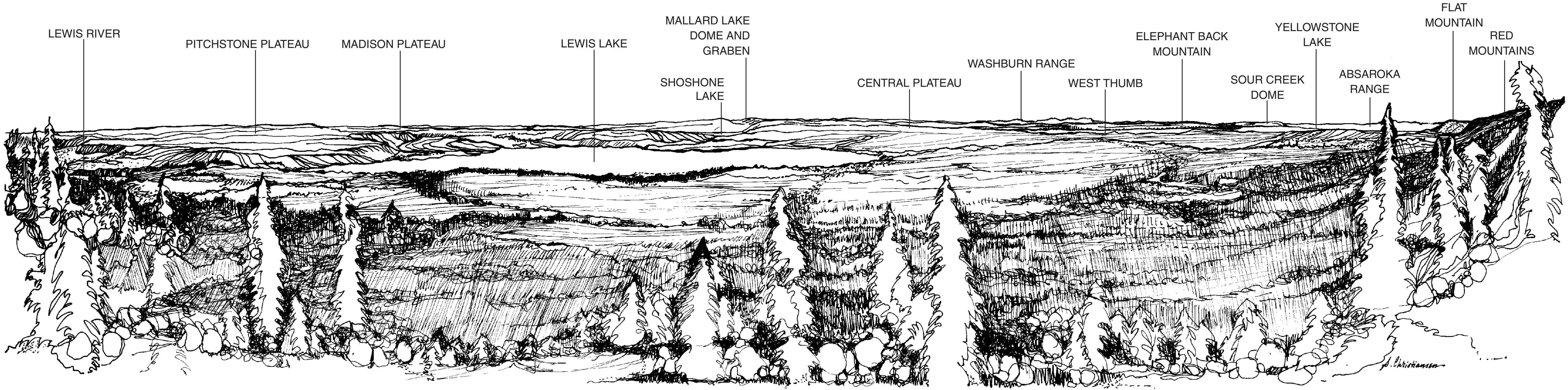
Yellowstone National Park, the oldest of the areas set aside as part of the national park system, lies amidst the Rocky Mountains in northwestern Wyoming and adjacent parts of Montana and Idaho. Embracing large, diverse, and complex geologic features, the park is in an area that is critical to the interpretation of many significant regional geologic problems. In order to provide basic data bearing on these problems, the U.S. Geological Survey in 1965 initiated a broad program of comprehensive geologic and geophysical investigations within the park. This program was carried out with the cooperation of the National Aeronautics and Space Administration, which supported the gathering of geologic information needed in testing and in interpreting results from various remote sensing devices. This professional paper chapter is one of a series of technical geologic reports resulting from these investigations



Frontispiece—Panorama of the Yellowstone caldera from the north side of the Red Mountains, south of Aster Creek. Washburn Range, about 55 km distant, and Absaroka Range, about 45 km distant, delimit the caldera on the northeast and east sides. Drawing by Susan Christiansen.

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THE QUATERNARY AND PLIOCENE YELLOWSTONE PLATEAU VOLCANIC FIELD OF WYOMING, IDAHO, AND MONTANA

By Robert L. Christiansen

ABSTRACT

The rhyolite plateau of Yellowstone National Park and nearby areas is part of the latest Pliocene and Quaternary Yellowstone Plateau volcanic field, which originally covered nearly 17,000 km^2 but has since been disrupted by erosion and by continued volcanism and tectonism. Igneous rocks of the field consist predominantly of rhyolites and subordinately of basalts; there are virtually none of intermediate compositions. The rhyolites comprise numerous lava flows and three major sheets of welded ash-flow tuff separated by unconformities. Stratigraphically these sheets—from oldest to youngest, the Huckleberry Ridge, Mesa Falls, and Lava Creek Tuffs—constitute the Yellowstone Group. The geologic history of the field defines three cycles, in each of which the sequence of volcanic events was similar, climaxed by the eruption of a voluminous sheet of rhyolitic ash-flow tuff and the formation of a large caldera. Preceding and succeeding events included eruptions of rhyolitic lavas and tuffs in and near the source areas of the ash-flow sheets and the eruption of basalts around the margins of major rhyolitic volcanism.

Before the first eruptions, the area was a mountainous terrain built by regional uplift and normal faulting; there was no extensive basin or plateau before about 2 Ma. Volcanism much like that soon to begin in the Yellowstone region was active 50 to 150 km west, in the eastern Snake River Plain.

The first cycle of the Yellowstone Plateau volcanic field began just before 2 Ma. The oldest recognized rocks, erupted between about 2.2 and 2.1 Ma, are basalts in northern and eastern Yellowstone National Park and a rhyolitic lava flow at the south end of Island Park, Idaho. The oldest ash-flow sheet of the Yellowstone Group, the Huckleberry Ridge Tuff, was erupted at 2.1 Ma and was emplaced as a single cooling unit of more than 2,450 km^3 over an area of 15,500 km^2. Collapse of the roof of the Huckleberry Ridge magma chamber formed a caldera more than 75 km long, extending from Island Park into central Yellowstone National Park and probably consisting of three overlapping but distinct collapse zones over separate high-level parts of the magma chamber. Three subsheets of the Huckleberry Ridge Tuff can be related to these three caldera segments. Rhyolitic lava flows west of Island Park are, in part, postcollapse lavas of the first cycle that overflowed the caldera rim; others may be buried within the caldera. Basalts again erupted in northern Yellowstone Park later in the first volcanic cycle.

The second cycle was simpler than the other two. Rocks of this cycle are present just west of Yellowstone National Park and probably are buried beneath the Yellowstone Plateau. Early second-cycle rhyolite flows crop out west of Island Park. The Mesa Falls Tuff, exposed near Island Park, is a cooling unit of more than 280 km^3, erupted at 1.3 Ma within the northwestern part of the first-cycle caldera. This locus of eruption largely restricted the Mesa Falls within the south and east walls of the older caldera. Collapse of the Mesa Falls magma-chamber roof formed another caldera about 16 km in diameter, nested against the northwest wall of the first-cycle caldera. Postcollapse rhyolite domes erupted within and adjacent to this second-cycle caldera, and basaltic lavas erupted intermittently around the margins of the volcanic plateau, particularly southeast of Island Park.

The third cycle perhaps overlapped the second, beginning about 1.2 Ma with eruption of rhyolitic lavas and related tuffs around a growing annular fissure system encircling central Yellowstone National Park. Flows vented periodically along this fissure system for about 600,000 years until ring-fracture development was terminated by rapid and voluminous ash-flow eruptions of the Lava Creek Tuff at 640 ka, probably through the same ring-fracture zone. These ash flows buried more than 7,500 km^2. Collapse occurred along the ring-fracture zone immediately after, and perhaps during, the eruption of the more than 1,000 km^3 of Lava Creek Tuff to form the Yellowstone caldera. Both the Lava Creek Tuff and the Yellowstone caldera display clear evidence for ash-flow eruptions from two separate high-level culminations of the Lava Creek magma chamber, analogous to those related to the three first-cycle caldera segments. The Lava Creek Tuff is a single cooling unit with two distinct subsheets centered around different parts of the caldera. The Yellowstone caldera consists of two ring-fracture zones, each enclosing a cauldron block; these two segments overlap in a single topographic basin 85 by 45 km. Postcollapse resurgence of the two cauldron blocks formed a pair of structural domes broken by axial grabens. Postcollapse rhyolitic

volcanism, which began soon after resurgent doming, has continued to at least 70 ka. Renewed doming of the western cauldron block at 160 ka initiated volcanism of increased intensity between 160 and 70 ka. Basaltic lavas have erupted intermittently throughout the third volcanic cycle on the northeast, north, west, and south margins of the rhyolite plateau. Although the highly active hydrothermal system of Yellowstone National Park is the only current manifestation, volcanism probably has not yet ended.

The erupted magmatic volume of the Yellowstone Group, about 3,700 km^3, accounts for more than half the material erupted in the Yellowstone Plateau volcanic field. The magma bodies that initiated and sustained each cycle must have totaled many times more than even that enormous volume. Although eruptions of the second cycle were less voluminous than the other two, all three produced major volcanic sequences. Each cycle was more or less complete and separate; each lasted about a half-million to a million years. At least the first and third cycles clearly reflect emplacement of large crustal rhyolitic magma bodies; each cycle records the rise of individual magmatic intrusions to high crustal levels, major degassing of the magma by surface volcanism, and eventual consolidation and cooling of the plutons. Intermittent basaltic activity was marginal to each rhyolitic center but essentially independent of the individual rhyolitic cycles. No basalt erupted within the principal volcanic focus of each rhyolitic cycle, but basalt has since erupted through the first- and second-cycle centers. Basalts of the Snake River Plain province now encroach on the Yellowstone Plateau volcanic field as the older parts of the field subside. Basaltic volcanism seems ultimately to represent the fundamental igneous process. The association of basalts and rhyolites is not accidental but represents a genetically unified evolution.

Rhyolites of the Yellowstone Plateau volcanic field are petrologically and geochemically distinct from the common rhyolites of predominantly andesitic calc-alkalic associations. The main basalts of the field are olivine tholeiites, much like the tholeiites of oceanic islands; younger basaltic magmas are progressively more K-rich. The rhyolitic and basaltic series arose from separate parental magmas and are not a single differentiation series.

The Yellowstone Plateau is in a region of active tectonic extension. Late Miocene and younger normal-fault blocks have mainly north to northwest trends. The principal belt of seismic and tectonic activity along the east margin of the active region passes directly through the Yellowstone Plateau volcanic field. Differential uplift and tilting of fault blocks has accompanied volcanism, and the principal volcanic structures were controlled by tectonic structures. The field lies on a northeastward extension of the Snake River Plain volcanic and tectonic axis, and Yellowstone and the eastern Snake River Plain share a unified evolution. The eastern Snake River Plain, a downwarp within the region of tectonic extension, has developed progressively since middle Miocene time. Earlier rhyolitic activity much like that of the Yellowstone Plateau characterized each part of the eastern Snake River Plain successively, the principal rhyolitic focus moving episodically northeastward with time. With the eventual consolidation of granitic plutons, each rhyolitic source area eventually subsided and was flooded by basalts.

The rhyolitic volcanism of the Yellowstone Plateau volcanic field and eastern Snake River Plain represents the successive emplacement of granitic batholiths from southwest to northeast. In each, rhyolitic magma rose to form one or more high-level plutons. Rhyolitic volcanism, particularly the climactic ash flows, degassed each major intrusion. With cooling and consolidation, the upper plutonic levels were successively reintruded by magma from lower levels until the entire body crystallized. Thus, the eastern Snake River Plain-Yellowstone axis, although almost completely flooded by basalts southwest of Island Park, overlies a discordant linear granitic polybatholith. The extended lifetimes of successive rhyolitic systems and their continuous association with, and eventual succession by basalts demonstrate continued intrusion of basaltic magmas at deeper levels to sustain them thermally. Basaltic magmatism and the generation, rise, degassing, and eventual crystallization of rhyolitic magmas are a mechanism of major mass and energy transfer within the lithosphere.

The Yellowstone and eastern Snake River Plain rhyolite-basalt association records a progressive igneous process that begins with the generation of basaltic magmas in the upper mantle. The basaltic magmas represent partial melting of peridotitic upper mantle in a region of extensional pressure release, and they rise in a deeply penetrating extensional stress field. The formation and rise of these magmas greatly augments an already high crustal heat flow. Their emplacement into and rise through a thick continental crust raises geothermal gradients sufficiently to enable localized partial melting of lower-crustal mafic or intermediate granulites by pressure release. The basaltic and rhyolitic magmas have different densities and rise independently in the upper crust; batholithic low-density silicic magma bodies retard or divert the rising basaltic magmas. Ultimate consolidation, extensional fracturing, and subsidence above the batholiths, however, eventually enables basalts to flood the region.

Volcanism almost certainly will recur in the Yellowstone National park region. Three general categories of future volcanic activity are likely: (1) further rhyolitic eruptions could occur within the Yellowstone caldera or on a northern radial zone as the third volcanic cycle decays; (2) basaltic eruptions might occur in several zones of recurrent activity on the margins of the plateau and eventually within the caldera; or (3) a major new magmatic insurgence initiating a fourth cycle and leading toward climactic ash-flow eruptions and caldera formation might occur, and may even be underway at present. Although these three possible modes of future activity cannot be analyzed rigorously, their relative probabilities may be substantially equal for the near geologic future.

INTRODUCTION

The superlative hot springs, geysers, and fumarole fields of Yellowstone National Park are vivid reminders of a recent volcanic past. Volcanism on an immense scale largely shaped the unique landscape of central and western Yellowstone Park, and intimately related tectonism and seismicity continue even now. Furthermore, the volcanism that gave rise to Yellowstone's hydrothermal displays was only part of a long history of late Cenozoic eruptions in southern and eastern Idaho, northwestern Wyoming, and southwestern Montana. The late Cenozoic volcanism of Yellowstone National Park, although long believed to have occurred in late Tertiary time, is now known to have been of latest Pliocene and Pleistocene age. The eruptions formed a complex plateau of voluminous rhyolitic ash-flow tuffs and lavas, but basaltic lavas too have erupted intermittently around the margins of the rhyolite plateau. Contemporaneous and younger basalts flood the Snake River Plain of eastern and southern Idaho, west of the national park and, on its surface, predominate over rhyolitic ash flows and lavas.

PREVIOUS WORK

Scientific knowledge of the region of Yellowstone National Park began only late in the 19th century, but an appreciation of the importance of volcanism in the later geologic history of the region came immediately. In fact, it was realized intuitively to some extent by fur trappers such as Jim Bridger, Warren Ferris, Joe Meek, and Osborne Russell who visited the region during the early and middle parts of that century and were impressed by Yellowstone's hydrothermal spectacles (for example, see Raynolds, 1868, p. 10; Phillips, 1940, p. 257-261; Victor, 1871, p. 75-76; Haines, 1965, p. 97-100). The incredulity with which some of the early accounts were received ultimately led to the first modern explorations and scientific studies of the Yellowstone Plateau region, notably by F. V. Hayden's Geological and Geographical Survey of the Territories (Hayden, 1872; 1873; Bradley, 1873; Peale, 1873). Hayden recognized at the outset of his exploration of the headwaters of the Yellowstone, Gallatin, Madison, and Snake Rivers that this region had been the scene of volcanic activity on a grander scale than had theretofore been recognized in the Rocky Mountain region. Hayden's perceptive first description of the Yellowstone Plateau, viewed from the summit of Mount Washburn, even anticipates the concept of the Yellowstone caldera (Hayden, 1872, p. 81):

> From the summit of Mount Washburn, a bird's-eye view of the entire basin may be obtained, with the mountains surrounding it on every side without any apparent break in the rim * * *. It is probable that during the Pliocene period the entire country drained by the sources of the Yellowstone and the Columbia was the scene of as great volcanic activity as that of any portion of the globe. It might be called one vast crater, made up of thousands of smaller volcanic vents and fissures out of which the fluid interior of the earth, fragments of rock, and volcanic dust were poured in unlimited quantities * * *. Indeed, the hot springs and geysers of this region, at the present time, are nothing more than the closing stages of that wonderful period of volcanic action that began in Tertiary times.

The first geologic work, however, did not resolve the important distinction between the early Tertiary Absaroka volcanism and the late Cenozoic volcanism of the rhyolite plateau. It is now known that these major igneous fields are entirely distinct in time, origin, and tectonic setting, but at the time of the first attempt to summarize the geology of Yellowstone National Park comprehensively (Holmes, 1883a; b), the two fields were regarded as overlapping parts of a single Tertiary volcanic system. The distinction between the older, predominantly andesitic Absaroka volcanism and the younger, predominantly rhyolitic plateau volcanism was first made clear in geological mapping by the U.S. Geological Survey (USGS) under the direction of Arnold Hague, mainly during the years 1883-1889. Until the major remapping of which the present study is a part (U.S. Geological Survey, 1972), the results of Hague's survey remained the best inclusive summary of Yellowstone's geology. The principal results of that early survey were published in two geologic folios and a partly completed memoir (Hague and others, 1896; 1899; Hague, 1904) and were summarized in nontechnical fashion by Hague (1888; 1912). The geology and petrology of the rhyolite plateau during this early survey were studied mainly by Iddings (1888; 1899a; b). Iddings (1896) made particularly clear the major distinctions in petrologic association and age between the Absaroka and plateau volcanic episodes and showed the distinct association of basalts and rhyolites in the latter. Despite this early recognition of major differences in the Absaroka and plateau volcanic episodes, they generally were regarded as successive stages in the evolution of a single major province. This concept was retained even in several more recent discussions (for example, Larsen, 1940; Brown, 1961).

More recent geologic studies of the Yellowstone Plateau and related rocks have focused on more specific topical problems. Iddings' early studies of the mineralogy of lithophysae in the Obsidian Cliff flow (Iddings, 1888; 1891) were followed up by Foshag (1926) and Bowen (1935). Brouwer (1936) discussed the origin of internal structures in rhyolitic lava flows of the plateau region. The origin of mixed-lava complexes of rhyolite and basalt were studied and discussed by Fenner (1938; 1944) and Wilcox (1944). An extensive program of studies on the hydrothermal geology of Yellowstone National Park carried out by the Geophysical Laboratory during the decade of the 1930's (Allen and Day, 1935) included the drilling of two boreholes in the rhyolites of active hydrothermal areas (Fenner, 1936).

Fenner (1937) recognized the similarity between some of the rhyolites of Yellowstone, now recognized as welded ash-flow tuffs, and the "sand flow" of Katmai and the Valley of Ten Thousand Smokes, Alaska. Mansfield and Ross (1935) and Ross (1955) noted that welded tuffs constitute much of the Yellowstone Plateau rhyolite sequence and that they are contiguous with rhyolitic welded tuffs that extend far to the west along the margins of the Snake River Plain. Kennedy (1955, p. 495), in a brief parenthetical note, however, advanced a concept that the voluminous sheets of welded tuff in Yellowstone and elsewhere are the products of collapsed froth flows—presumably very vesicular mobile lavas—rather than of pyroclastic eruptions.

Boyd (1961), in the only comprehensive recent study of the upper Cenozoic rhyolite plateau preceding this one, showed conclusively that the welded tuffs of the region are of pyroclastic-flow origin, providing a considerable body of experimental and theoretical argument to advance the case. Boyd's excellent field study provided a solid foundation for the present study. In many fundamental respects the present work serves to substantiate Boyd's major conclusions; it does show, however, that a more complex series of volcanic events is represented in the record of the Yellowstone Plateau than Boyd recognized.

Hamilton (1959; 1960b; 1963; 1964; 1965) published several papers on the geology of the Yellowstone Plateau and of Island Park—immediately west—on the basis of fieldwork carried out in support of geologic investigations of the Hebgen Lake earthquake of August 17, 1959. He emphasized a close volcanotectonic relation between the Yellowstone Plateau and the Snake River Plain, the presence of large partly lava-filled calderas in the Yellowstone Plateau and at Island Park, and a genetic relation between the rhyolites and basalts of the region. Other recent studies have focused on general stratigraphic and structural questions or on topical problems other than the rhyolite plateau itself. Some that have a direct bearing on the geology of the Yellowstone Plateau volcanic field include those of Howard (1937), Love (1956b; c; 1961), Brown (1961), Hall (1961), Fraser and others (1969), Witkind (1969; 1972), Chadwick (1978), and Hadley (1980).

NATURE OF THIS STUDY

The study reported here is aimed at reconstructing a volcanic history of the Yellowstone Plateau region, interpreting important volcanic mechanisms, and seeking clues to deeper-seated processes that gave rise to the volcanic activity. This study mainly involves areal geologic mapping and the delineation of volcanic stratigraphy and significant structural relations. Fieldwork was carried out jointly with H. R. Blank, Jr. during the summers of 1966-70; I continued fieldwork in 1974 and, with Wes Hildreth, for shorter periods in 1979 and 1981. This fieldwork was done in an area of more than 17,000 km^2 that includes much of the national park but extends well beyond the park boundaries, especially to the west and south. Consequently, knowledge of various parts of the area is uneven; on the whole, the fieldwork can be regarded as constituting a detailed reconnaissance. The primary mapping scale was 1:62,500, but some areas outside the national park were examined only in reconnaissance and were mapped only at 1:250,000 scale. Detailed petrologic studies have been undertaken together with Wes Hildreth, but they are represented in this paper only to the extent that available results seem to bear directly on the main problems of volcanic history and magmotectonic evolution. Stratigraphic, petrologic, and geochemical studies are continuing.

Only a small part of the Yellowstone Plateau is accessible by roads or maintained trails. Most of the plateau is covered by dense forests, particularly of lodgepole pine (*Pinus contorta*). This extensive tree cover, together with a typically thick mantle of glacial deposits and widespread hydrothermal alteration, results in generally poor exposures. In addition, the shallowly rooted lodgepole pines are susceptible to extensive felling by storm winds and grow up rapidly and densely in areas burned over by forest fires. Consequently, many parts of the plateau provide difficult travel by foot or horseback. Partly compensating for these factors, however, mapping in this little-eroded volcanic field is facilitated by the direct topographic expression of many units, particularly the rhyolitic lava flows, and by the common presence of single volcanic-emplacement units, both individual rhyolite flows and distinct sheets of ash-flow tuff or basalt, over large areas. Critical relations are most commonly exposed in cliff sections and in eroded stream valleys.

One of the characteristic features of a rhyolitic field such as Yellowstone is the great variety of lithologies encountered in rocks of similar chemistry and phenocryst content. Iddings (1899a, p. 356-357) remarked on this diversity of appearance as perhaps the most typical feature of the Yellowstone Plateau rhyolites. Even within a single lava flow or ash-flow cooling unit, an almost complete gamut of lithologic types can be found. In general there is as much variation in appearance within any single rhyolitic lava flow as there is within the entire assemblage of flows on the rhyolite plateau. The implication of this fact for geologic mapping should be understood in interpreting maps such as plates 1 and 2 and in following the discussions of this paper. The formational units are defined as nearly as possible on traditional stratigraphic principles (Christiansen and Blank, 1972; Christiansen, 1982), but the units actually mapped in several of these formations—individual rhyolitic lava flows and ash-flow sheets or subsheets—are recognized by complexes of criteria that appear to define lithogenetic entities. Each flow has steep scarplike margins, a bulky lobate morphology, and a surface on which arcuate ridges are generally concave toward the vent; a crystallized interior, largely flow layered, is surrounded by partly crystallized (generally spherulitic) to wholly glassy rhyolite, pumiceous in the outermost portions and forming a flow-breccia at the margins of the flow. Any or all of these criteria may have been used in separating one flow from another during geologic mapping. Identical lithologies occur in many separately mapped flows.

Similarly, as shown by Smith (1960b), the fundamental lithogenetic entities of ash-flow terranes, the cooling units, are defined by zonal lithologic features that are interpreted to indicate that certain of the rocks were emplaced in rapid sequence and subsequently cooled together. Densely welded crystallized interiors of each sheet typically are surrounded by less densely welded to nonwelded exteriors, glassy at the margins; a variety of textures occurs, each typical of certain zones within the cooling unit. The terminology used for these features and for the units that are based on recognition of welding and crystallization zones is that of Smith (1960a; b) and Ross and Smith (1961).

Three major cooling units or composite sheets of ash-flow tuff are present in the Yellowstone Plateau-Island Park region. Each of the three is quite similar to the others in general appearance, phenocryst mineralogy, and chemistry and each has a wide range of lithologies. Distinctions between the three sheets, especially in isolated exposures, often are difficult to make. The distinctions more often depend upon finding a characteristic vertical sequence of lithologic features than upon finding any single distinctive lithology in any one place. It should be obvious, therefore, and must be remembered in evaluating the present work, that some subjectivity is necessarily represented in recognizing the units shown on our geologic maps and in interpretations derived from them. Some mistakes are almost certain to be found in the mapping, particularly in the areas of more generalized reconnaissance.

The plan of this paper does not follow a traditional format but is based on the particular history of the volcanic field. Study of the late Cenozoic volcanism of the Yellowstone Plateau and Island Park shows that the region comprises a single volcanic field with a unified history. Volcanism occurred in three main cycles, in each of which the sequence of events was much the same. Climactic events of each cycle were the eruption of a voluminous rhyolitic ash-flow sheet and collapse of the eruptive source area to form a large caldera. These three climactic episodes were each separated by about 700,000 years, and events of the later volcanic cycles partly obliterated features of the earlier cycles. Therefore, following a general introduction, the stratigraphy, structure, and volcanic history of the youngest of the three volcanic cycles are described, providing a model against which the earlier cycles are then discussed; some relations among the three cycles are then presented. After thus interpreting the volcanic history of the field in an everted chronological sequence, the petrology and geochemistry of the field are considered briefly and certain aspects of the field are described and interpreted from a magmotectonic point of view.

ACKNOWLEDGMENTS

This study of latest Cenozoic volcanism of the Yellowstone Plateau was part of a comprehensive restudy of the geology of Yellowstone National Park that was carried out with the cooperation of the National Park Service and the National Aeronautics and Space Administration. The cooperation of the National Park Service was vital, and former Superintendents Jack Anderson and John S. McLaughlin and their staffs were most helpful during the period of extensive fieldwork from 1966 to 1974. In particular, successive Chief Park Naturalists John M. Good, William W. Dunmire, and Alan Mebane extended many courtesies during that same period.

Parts of these studies that were sponsored by the National Aeronautics and Space Administration were designed to use Yellowstone National Park as a test area for remote-sensing experiments. Simultaneous studies of other geologic topics were carried out in the national park by other Geological Survey parties, whose cooperation greatly facilitated the work presented here. These parties included W. R. Keefer, J. D. Love, E. T. Ruppel, H. W. Smedes, H. J. Prostka, G. M. Richmond, K. L. Pierce, H. W. Waldrop, D. E. White, R. O. Fournier, L. J. P. Muffler, and A. H. Truesdell.

I carried out most of the field mapping jointly with H. R. Blank, Jr. His contributions to this study have been numerous, not the least of which were his companionship in the field and many hours of stimulating and challenging discussions of the outcrops. We were assisted in the field by George M. Fairer in 1966, by John W. Creasy in 1967, by Richard L. Reynolds and Thomas L. Chamberlain in 1968, by Clarence J. Duffy and John E. Eichelberger in 1969, and again by Reynolds in 1970. I was assisted by Lucy Katherine Sidoric in 1974.

I acknowledge in particular the data provided by R. L. Reynolds from a paleomagnetic study that he carried out on volcanic units of the Yellowstone Plateau and related widespread ash beds, subsequent to his participation in our fieldwork. John D. Obradovich provided K-Ar dating for the study. Robert B. Smith has generously shared geophysical data and his insightful interpretations of the Yellowstone region. Bruce R. Doe and William P. Leeman made available isotopic and trace-element data on samples from our Yellowstone suite and from Leeman's extensive studies of the Snake River Plain. Warren Hamilton gave the benefit of his reconnaissance observations in Yellowstone National Park and Island Park at an early stage in our field work and continued to engage me with stimulating discussions of tectonic mechanisms.

I owe much to my colleague Wes Hildreth, with whom I am collaborating in detailed petrologic and geochemical studies of the Yellowstone volcanic suite. Little of our joint study is reported explicitly in this paper, but many of the interpretations discussed here stem from that work or have been tempered through our uncounted discussions in the field and elsewhere. Robert L. Smith has provided me with invaluable advice and guidance for nearly 40 years on this study and others related to it. The manuscript of this paper benefited greatly from the detailed review, criticism, and technical advice of Robert L. Smith, Peter W. Lipman, and Wes Hildreth. A special debt of gratitude is owed to the late

Arthur B. Campbell under whose direction this study was initiated and the field work was largely carried out and who supported the work with much energy and enthusiasm.

OVERVIEW OF THE YELLOWSTONE PLATEAU VOLCANIC FIELD

The following generalized characterization of the Yellowstone Plateau volcanic field and its setting, and a brief summary of its volcanic history, provide a framework for the more detailed but nonchronological discussions of later sections.

GENERAL CHARACTER OF THE FIELD

The Yellowstone Plateau covers an area of about 6,500 km^2 and forms the continental divide between the northern and middle Rocky Mountains although the plateau itself conventionally is considered to be part of the middle Rockies (Fenneman, 1931; Thornbury, 1965). Most of the plateau lies within Yellowstone National Park, but parts extend as much as 13 km farther west (fig. 1). Outliers of the plateau show that it once extended farther onto the Snake River Plain and its margins, as well as outward from the national park along the valleys of the Madison, Gallatin, and Yellowstone Rivers, most of the way across Jackson Hole, and across parts of the western Absaroka Range. The original extent of the volcanic field now represented by the Yellowstone Plateau was nearly 17,000 km^2. The volcanism that formed the Yellowstone Plateau field has occurred over a period of more than 2 million years. The major volcanic events most responsible for present physiographic features of the plateau have occurred within about the last million years, the climactic events—voluminous ash-flow eruptions and caldera collapse—having occurred about 640,000 years ago. The source vents for eruptions of this youngest cycle of volcanism were within the area of Yellowstone National Park. Earlier eruptive episodes of the volcanic field, however, occurred within an area including not only the national park but extending nearly 40 km west across Island Park. Island Park now forms a topographic and geologic transition between the Yellowstone Plateau and the Snake River Plain.

The Yellowstone Plateau is surrounded by mountainous terrain on all sides but the southwest, where it stands above Island Park and the Snake River Plain. Most of the surrounding mountains and intervening valleys have north to northwest trends, but the Centennial Mountains and Centennial Valley, separated from the plateau on the west by only the basin of Henrys Lake, trend westward (fig. 1). The average elevation of the plateau is about 2,400 m. The major surrounding peaks rise to elevations of 3,000-4,000 m, and the Island Park area and eastern Snake River Plain lie at elevations of about 1,600-1,900 m. Boyd (1961) recognized that the plateau comprises two major geologic elements. An outer zone is underlain largely by welded rhyolitic ash-flow tuffs and associated small rhyolitic lava flows and basaltic sheets. The inner zone is underlain mainly by a group of very large rhyolitic lava flows. Boyd (1961, p. 410-412) surmised that the basin filled by these inner rhyolite flows was of volcanotectonic origin and was related to eruption of the voluminous rhyolites, especially the widespread ash-flow tuffs. The present work (see plates 1, 2, 3) essentially confirms Boyd's hypothesis; the nature and origin of this basin, the Yellowstone caldera, are discussed in detail later in this paper. An appreciation of this fundamental relation, however, is helpful at this point in discussing general aspects of the rhyolite plateau.

The geologic recency of volcanism in the Yellowstone Plateau has been noted generally, but most previous accounts have regarded the age as Pliocene (for example, Hague, 1896; Howard, 1937; Boyd, 1961; Love, 1961). Richmond and Hamilton (1960) and Hamilton (1960b) first noted relations between rhyolite flows of the Yellowstone Plateau and late Pleistocene glacial features that suggest a Quaternary age for at least the youngest volcanism. Our stratigraphic work (Christiansen and Blank, 1972) and the K-Ar dating of Obradovich (1992) demonstrate that most of the plateau volcanism occurred during and just before the Pleistocene.

More detailed work has indicated that some of the specific relations cited by Richmond and Hamilton (1960) and Hamilton (1960b) for a late Pleistocene age of some rhyolite flows of the plateau are equivocal. The general ages suggested for most of these flows, however, appear still to be correct. Other features, mainly morphologic aspects of some flows, are described later in this paper to suggest that certain rhyolites were emplaced against glacial ice. Till has been found underlying a rhyolite flow in a single exposure, at Bechler Meadows in southwestern Yellowstone National Park (pl. 1). It is clear that volcanism and glaciation occurred simultaneously or alternately in the Yellowstone Plateau region during much of Pleistocene time (Friedman and others, 1973; Pierce and others, 1976; Richmond, 1976).

The surface of the rhyolite plateau, dating mainly from volcanic events between 640,000 and 70,000 years ago, is young enough so that the present-day topography still reflects volcanic landforms directly. As there are only a few rather small cone-shaped volcanoes in the plateau region, the landforms and drainage patterns are not those commonly noted in young volcanic fields. The main physiographic elements are largely controlled by sheetlike volcanic units, by concentric arcuate surface ridges and steep scarplike margins typical of the large rhyolite flows, by small local constructional volcanoes, and by young volcanic and tectonic structures, particularly fault scarps.

Lakes and waterfalls are plentiful on the plateau and reflect both the youth of the constructional topography and the direct influences of volcanic landforms and rock types. Yellowstone's largest lakes, Shoshone, Lewis, Heart, and

Yellowstone Lakes, as well as the smaller Wolf, Grebe, and Cascade Lakes, all formed where streams draining into or along the margin of the Yellowstone caldera were dammed by rhyolitic lava flows emplaced within the caldera. Many of the larger sediment-filled basins of the park, such as the Upper and Lower Geyser Basins, Little Firehole Meadows, Hayden Valley, and lower Pelican Creek, represent parts of the caldera that were enclosed but not filled by rhyolitic lava flows or volcanic uplifts. The principal streams within the caldera (fig. 2) flow within these large basins and along the scarplike margins of rhyolite flows. The Gibbon River (for about two thirds of its total length), the upper Madison River, and Broad, Astringent, and Aster Creeks flow in courses along the walls of the Yellowstone caldera and are confined by caldera-filling lava flows. Many of the waterfalls are closely related to volcanic structures, particularly the steep margins and terraced surfaces of rhyolite flows where internal flowage layering is nearly vertical. Examples include Lewis Falls and the numerous waterfalls of the Bechler River and its tributaries in southwestern Yellowstone Park (the "Cascade Corner"). Gibbon Falls tumbles across vertically jointed welded tuff at the fault-scarp margin of the Yellowstone caldera; Virginia Cascade is on steeply jointed welded tuff where an uplifted and tilted fault block has raised the gradient of the Gibbon River. Other waterfalls have formed where vertically jointed basalts, welded

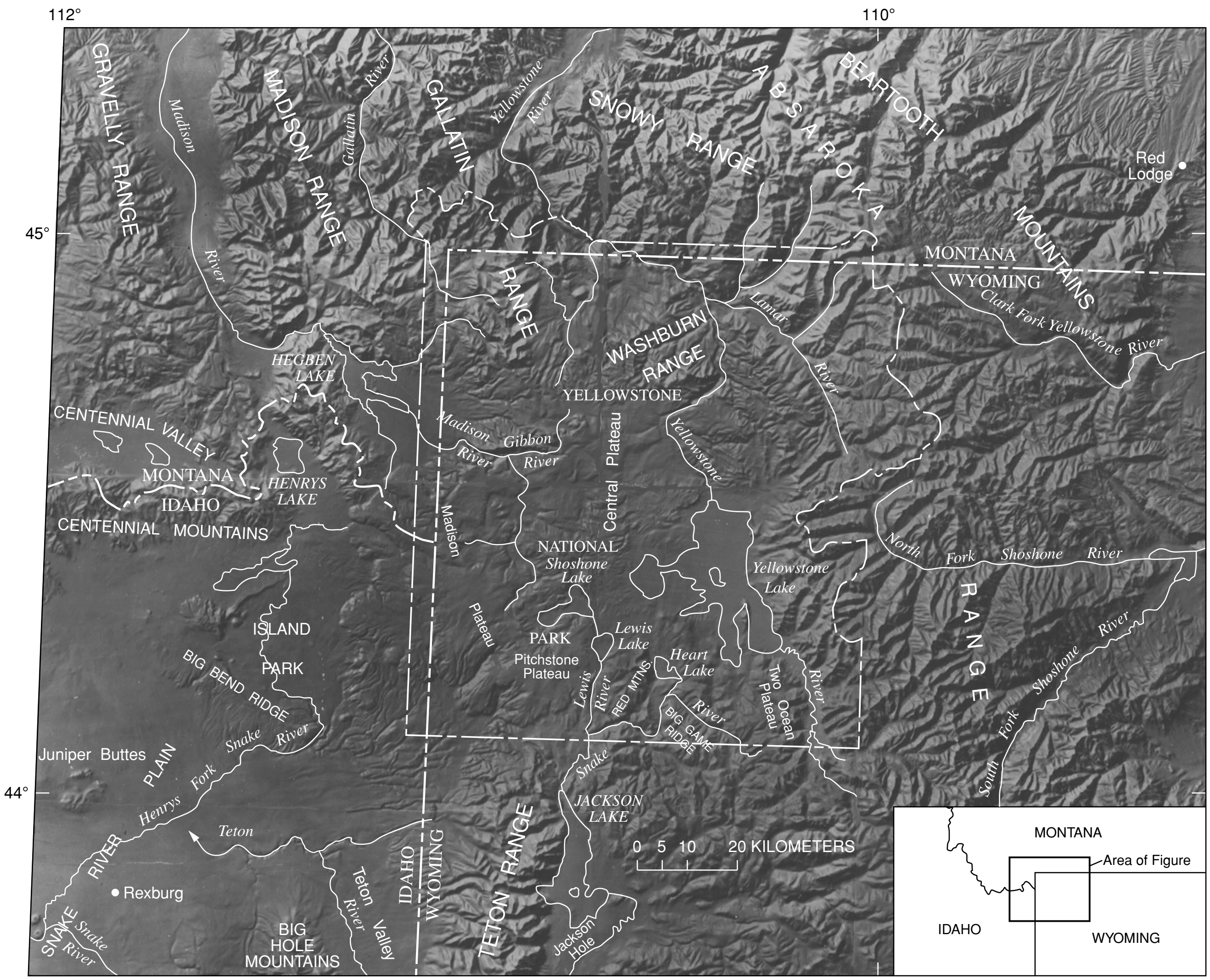


Figure 1.—Index map and physiography of the Yellowstone Plateau region. The modern rhyolite plateau, about 6,500 km^2, covers mainly the area from the Madison Plateau to Yellowstone Lake and from the vicinity of the Washburn Range to north of the Teton Range and Red Mountains. The Yellowstone Plateau volcanic field, originally almost 17,000 km^2, was once continuous onto the Snake River Plain and its margins, along the valleys of the Madison, Gallatin, and Yellowstone Rivers, over most of Jackson Hole, and over parts of the western Absaroka Range.

tuffs, or crystallized rhyolitic lava flows lie above or adjacent to less erosionally resistant rocks such as soft sediments, hydrothermally altered rocks, or the glassy portions of rhyolite flows. Examples of such falls (fig. 2 and pl. 1) include the Upper and Lower Falls in the Grand Canyon of the Yellowstone as well as such smaller falls as Rustic, Undine, Osprey, Crystal, Firehole, and Moose Falls, and Kepler Cascades. Of Yellowstone's principal waterfalls, only Tower Falls and Silver Cord Cascade in the Grand Canyon area (pl. 1) formed like those of Yosemite, where smaller tributary valleys were left hanging by rapid downcutting of a master stream.

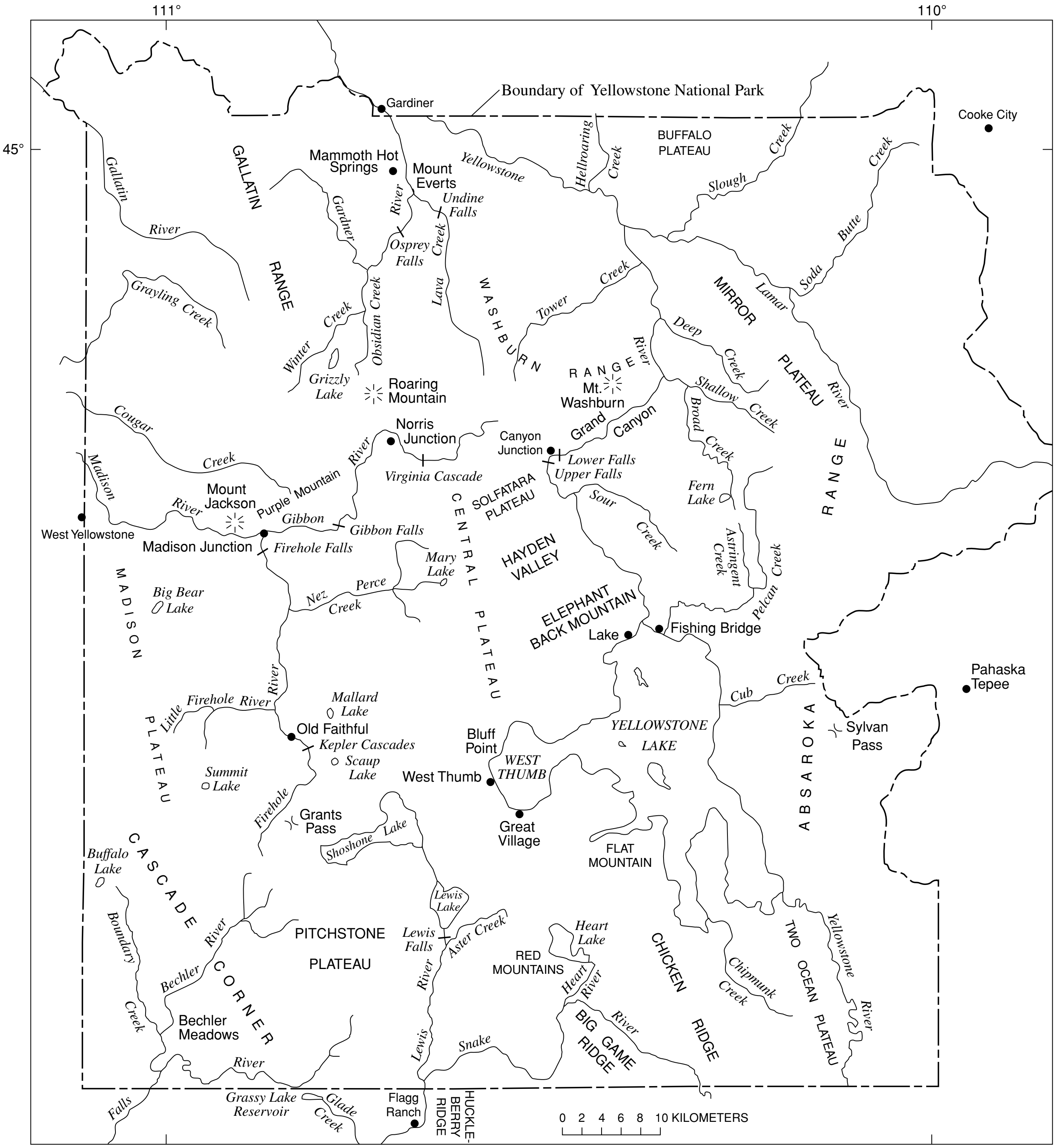


Figure 2.—Principal drainages, waterfalls, and place names of Yellowstone National Park. Major physiographic elements—mountains, plateaus, drainages, and waterfalls—are largely controlled by constructional volcanic topography and by young volcanic and tectonic structures, particularly fault scarps. Other place names referred to in the text are on plates 1-3.

The overwhelming bulk of volcanic rocks of the Yellowstone Plateau field is of rhyolitic composition. (As used in this paper, the term "rhyolite" refers only to a compositional range of volcanic rock, consisting normatively of essential quartz, orthoclase, and subordinate sodic plagioclase; no reference to mode of emplacement is implied.) By volume, rhyolites, including lava flows, ash flows, and fall-out deposits, constitute more than 95 percent of the field (Christiansen, 1984, fig. 3). Most of the rhyolites contain abundant phenocrysts, mainly quartz, sodic sanidine, and oligoclase or sodic andesine, but a few rhyolite flows, including the well known Obsidian Cliff flow (Iddings, 1888; Boyd, 1961), contain few or no phenocrysts. Iron-rich clinopyroxene is the predominant mafic phenocryst mineral; hornblende is absent or only an accessory constituent in most of the rhyolites, and biotite phenocrysts are extremely rare. The rhyolites have high silica contents, generally more than 75 percent.

The Yellowstone Plateau volcanic field is an especially clear example of a compositionally bimodal rhyolite-basalt igneous field. Such fields have been noted in many parts of the world and are common in the upper Cenozoic of the Western United States (Christiansen and Lipman, 1972). Most bimodal fields contain basalts in excess of rhyolites, but all proportions occur. Basalts occur only around the margins of the Yellowstone Plateau, interlayered in the ash-flow plateau that surrounds the central caldera complex. However, at Island Park, an older part of the volcanic field, basalts were erupted within the caldera complex and bury its floor. A limited range of compositions is found in the basalts of the Yellowstone Plateau volcanic field. Most basalts of the field are low-potassium tholeiites, but somewhat more potassic (though still hypersthene-normative) basalts form a subsidiary suite.

REGIONAL SETTING

As noted earlier, the outer part of the Yellowstone Plateau consists mainly of extensive ash-flow sheets with relatively minor interlayered rhyolitic and basaltic lavas. The relations of these volcanic units to the surrounding mountains and valleys show clearly that the main topographic elements of the region had been established before the plateau eruptions. Nevertheless, continued development of the topography surrounding the plateau by block faulting is recorded in the offlaps of successively younger volcanic units. Thus, the Centennial Mountains, the Madison, Gallatin, Snowy, and Absaroka Ranges, Two Ocean Plateau, Big Game Ridge, the Red Mountains, and the Teton Range and their intervening valleys (fig. 1) were each present in essentially their present structural configurations some time before 2 Ma but with lower relief than at present. These relations will be considered further in a later section, but their pertinence here is their control of the form of the rhyolite plateau and of the distributions of its major ash-flow sheets. Thus, in its original form, the ash-flow plateau extended around the flanks of the major mountain ranges, with arms extending along valleys leading away from the central caldera complex. The only broad opening in the surroundings of the plateau was the Snake River Plain to the west and southwest, enabling the ash-flow sheets to extend greatly in those directions. After the first major ash-flow eruptions and formation of the first-stage caldera, the caldera basin partly blocked this direction to younger ash flows.

The Yellowstone region lies at the northeast end of the eastern Snake River Plain, a northeast-trending structural depression about 350 km long (Kirkham, 1931; Malde, 1991). The eastern Snake River Plain, thus, defines a major tectonic axis in the region. The depression is flooded by basaltic lavas, but minor rhyolites occur in the plain and abundant rhyolites and related silicic volcanic rocks are associated with basalts along its margins in a compositionally bimodal assemblage. The basaltic and rhyolitic volcanism migrated from southwest to northeast along the eastern Snake River Plain axis during late Cenozoic time (Christiansen and Blank, 1969; Christiansen and Lipman, 1972; Armstrong and others, 1975; Christiansen and McKee, 1978). The trend of the eastern Snake River Plain is nearly normal to the trend of basin-range fault blocks that form its north and south margins. Local structural relations (for example, Carr and Trimble, 1963, p. 35-39) show that the Snake River Plain structure postdates major development of the basin-range normal faults, but all these tectonic elements as well as the basaltic and bimodal volcanism associated with them date only from middle Miocene and later time (Ruppel, 1964; Armstrong and others, 1975; Christiansen and McKee, 1978).

Seismic activity during historical time in the region discussed here has occurred mainly in a belt that extends parallel to the east margin of the basin-range region, through the Yellowstone Plateau, then turns and branches northward into Montana and westward north of the Snake River Plain (fig. 3*A*; see Sbar and others, 1972; Smith and Sbar, 1974; Smith and others, 1974; Smith, 1978; Smith and others, 1989; Smith and Arabasz, 1991). The destructive Hebgen Lake earthquake of August 17, 1959, was centered within this active seismic zone just northwest of the boundary of Yellowstone National Park (Ryall, 1962; U.S. Geological Survey, 1964; Trimble and Smith, 1975; Doser, 1985; Doser and Smith, 1989). Geophysically and neotectonically the Northern Rocky Mountains are in some ways now more allied to the Basin and Range province than to the middle Rocky Mountains with which they merge physiographically (Gilluly, 1963; Hamilton and Myers, 1966; Blackwell, 1969; Reynolds, 1979; Stickney and Bartholomew, 1987).

The Yellowstone Plateau is at the intersection of the eastern Snake River Plain tectonic axis with the main belt of active seismicity. The Yellowstone Plateau region also lies within an older orogenic belt, the latest Cretaceous and early Tertiary Laramide belt of foreland thrust-fault uplifts and intermontane basins (Foose and others, 1961; Sales, 1968;

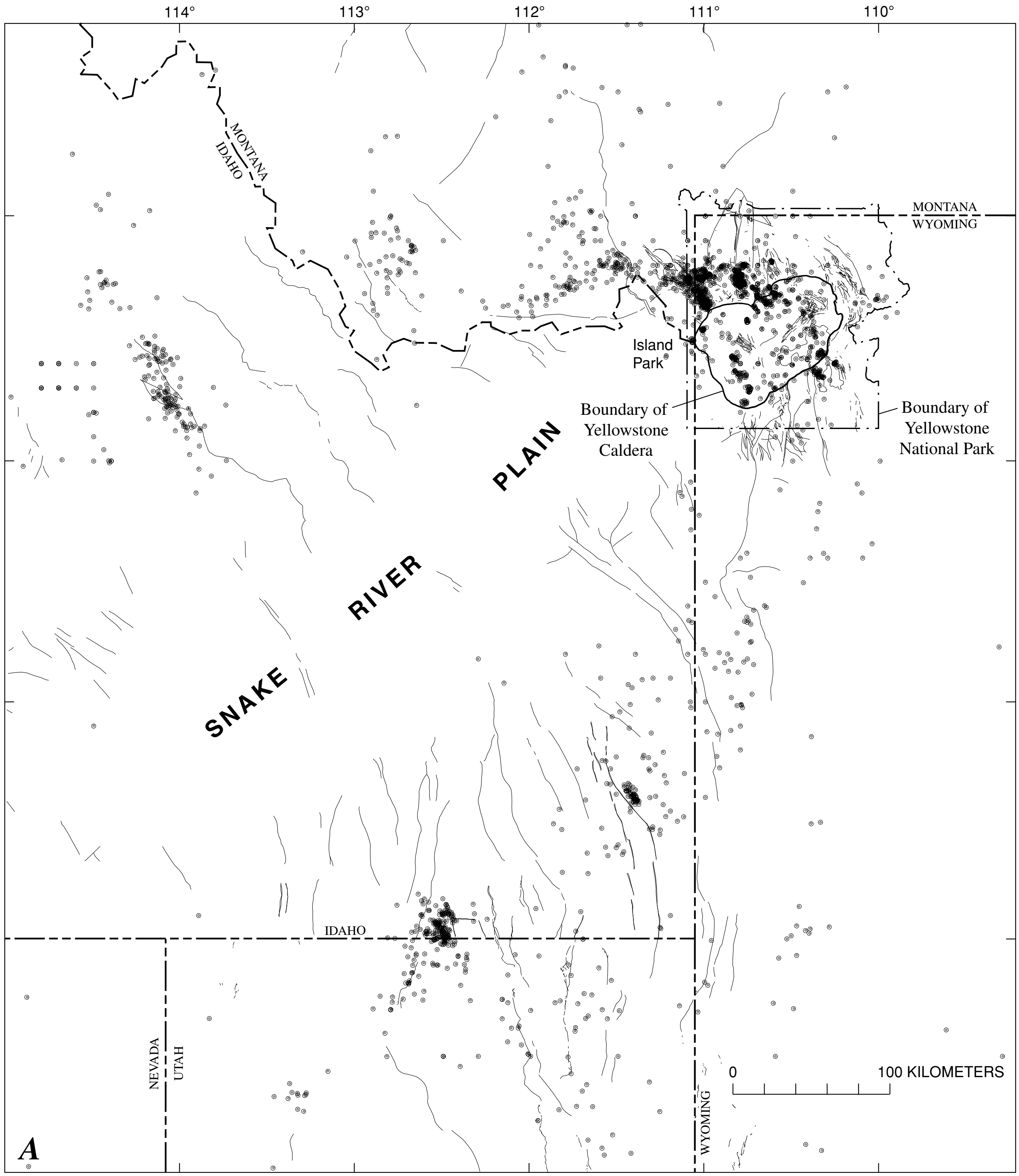
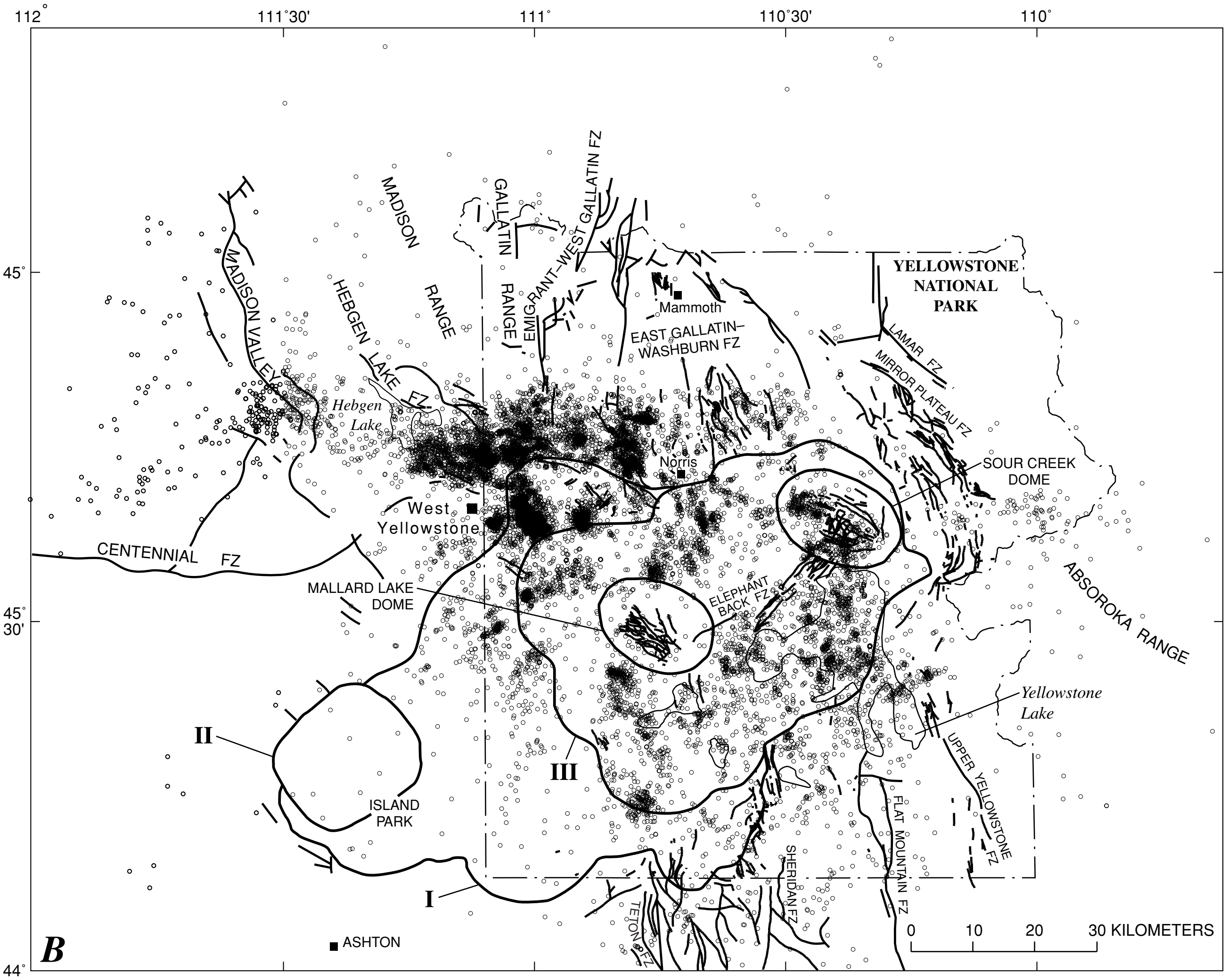


Figure 3.—Major upper Cenozoic faults and earthquake epicenters in the Yellowstone-Snake River Plain region occur mainly in a belt that extends parallel to the east margin of basin-range faulting in southeastern Idaho and western Wyoming, passes through the Yellowstone plateau, then turns and branches northward into Montana and westward north of the Snake River Plain. ***A***, faults and earthquake epicenters from about 1900 to 1985 in the region surrounding the Yellowstone Plateau; after Smith and Arabasz (1991), with additional faults modified after Cohee and others (1962), Malde and others (1963), U.S. Geological Survey (1964; 1972), Pierce and Morgan (1992). ***B***, epicenters in the immediate vicinity of Yellowstone National Park from 1973 to 1996; from Smith and Christiansen (1980), Pitt (1989), and Smith and Arabasz (1991). Fault zones (FZ) named in figure 5 shown by generalized fault distributions; heavy lines labeled I, II, and III indicate first-, second-, and third-cycle calderas, respectively.



Keefer, 1970; Matthews, 1978) which formed much of the middle and southern Rocky Mountains (fig. 4). The Laramide orogenic episode was a direct precursor of the predominantly andesitic Absaroka volcanism of Eocene age in the Yellowstone region (Smedes and Prostka, 1972). Neither the Laramide tectonism nor the Absaroka volcanism, however, seems to have had a primary functional relation to the Yellowstone Plateau volcanism.

SUMMARY OF STRATIGRAPHY AND VOLCANISM

The geology of the Yellowstone Plateau volcanic field is shown in plates 1, 2, and 3, and major late Cenozoic volcanic and tectonic features of the plateau region are summarized in figure 5.

Figure 6 is an attempt to portray the main geologic and topographic features of the Yellowstone region just prior to beginning of the plateau volcanism, a little before 2 Ma. Because the base of the Yellowstone Plateau volcanic field is not exposed over most of its central part or in the Island Park-Snake River Plain area, a true paleogeographic or paleogeologic map cannot be constructed. Relations exposed on the margins of the plateau and estimates of the rates of development of structures based on their relations to the plateau ash-flow sheets have been used to compile this interpretive view. It appears that just before the plateau volcanism, the Yellowstone region was entirely an elevated mountainous terrain formed by differential uplift and tilting of blocks bounded by normal faults. There is little or no evidence for a major basin in the present plateau area. Alignments of subsequent rhyolitic vents and minor younger faults (fig. 5; pl. 1) suggest that there was some kind of connection between the Teton and Madison Ranges although the two blocks have opposite major tilts. Similar relations suggest that the Red Mountains were structurally continuous with the Gallatin Range before formation of the rhyolite plateau. Ruppel (1972) proposed that the Gallatin and Teton Ranges are joined and at depth beneath the rhyolite plateau. Taken together, these reconstructions suggest a pattern of subparallel but locally branching or intersecting fault-

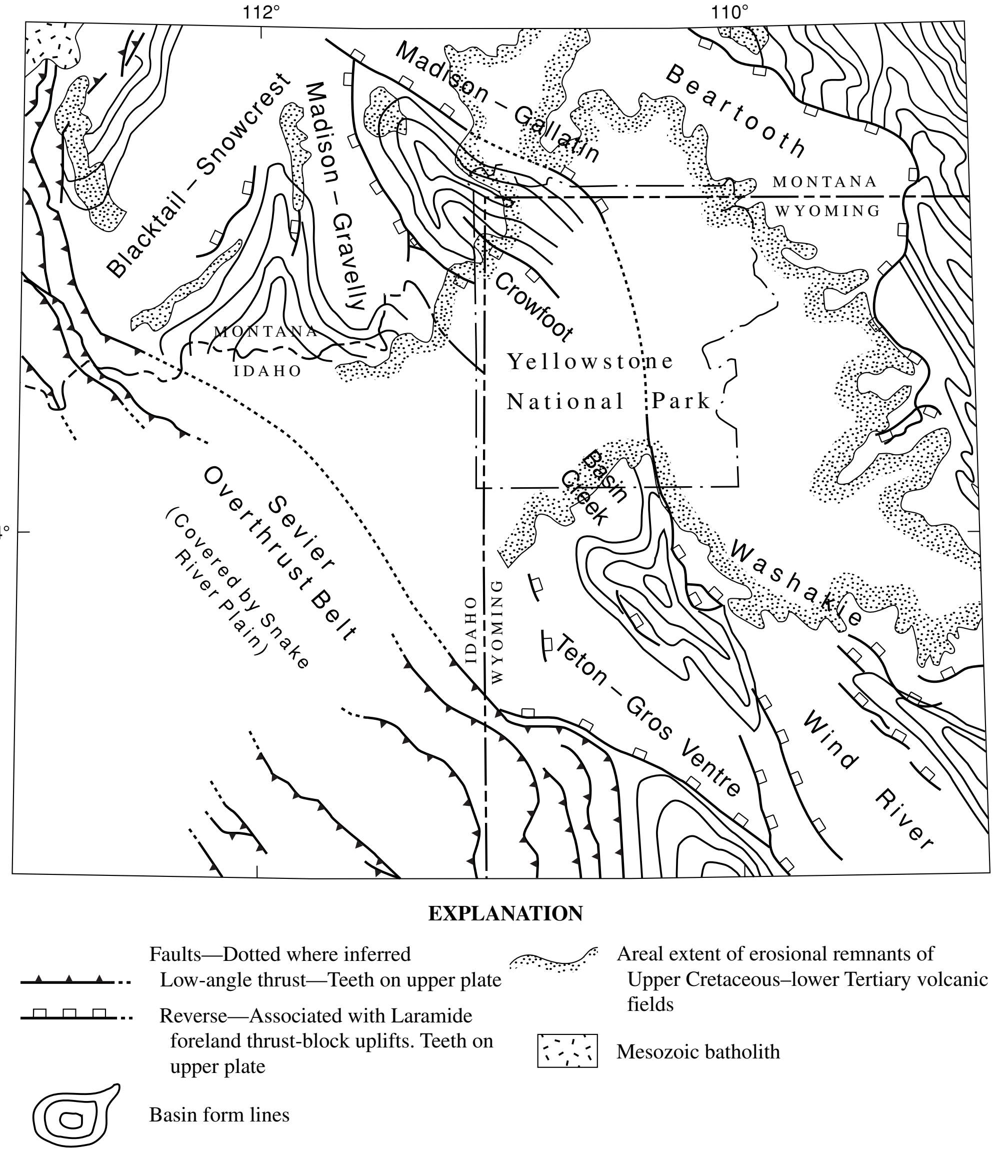
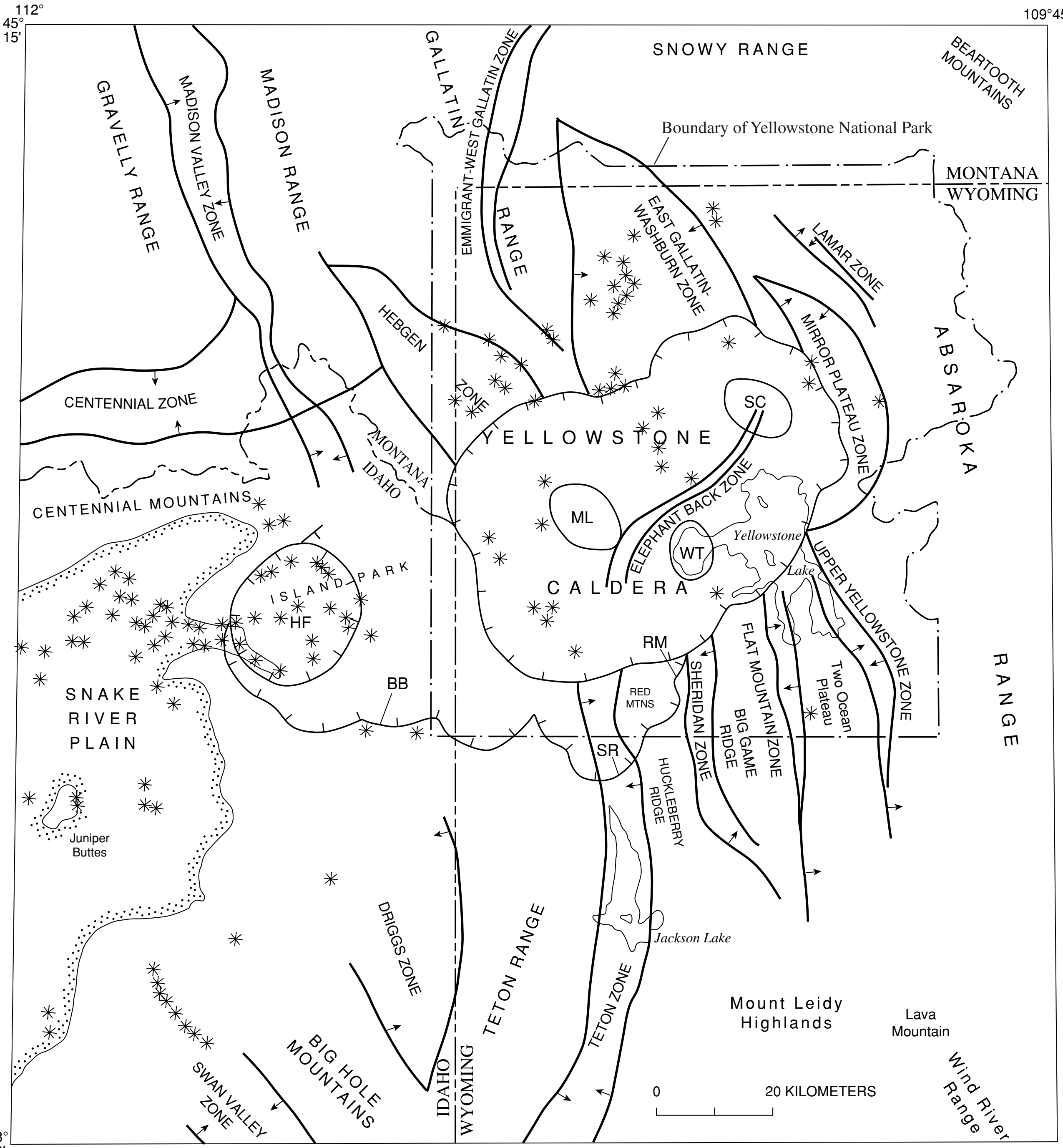


Figure 4.—Upper Cretaceous and lower Tertiary structures and volcanic fields of the Yellowstone Plateau region are the Sevier overthrust belt on the west, the Laramide belt of foreland uplifts and basins in the Middle Rocky Mountains, and both Laramide and early post-Laramide volcanic fields. Uplifts are named. Map modified after Cohee (1962), Foose and others (1961), Ruppel (1972), Love and Keefer (1969; 1975), Scholten (1967), and Scholten and others (1955).

bounded blocks as is common in the basin-range region of the Western United States (Zoback, 1983).

Volcanism similar to that which was to begin shortly in the Yellowstone-Island Park region was active at the east end of the Snake River Plain. A phenocryst-poor generally lithophysal ash-flow sheet related to that volcanism—the tuff of Kilgore of Morgan (1988), Hackett and Morgan (1988), and Morgan (1992)—partly buried the west border of the mountainous region and extended across any preexisting Madison-Teton intersection, partly flooding Jackson Hole. The lithologically similar but older Conant Creek Tuff has a fission-track age of 4.2 Ma and a K-Ar age of 5.9 Ma (Christiansen and Love, 1978; Morgan 1992). The source area of the tuff of Kilgore, the Conant Creek Tuff, and a group of associated voluminous ash-flow tuffs of 4-7 Ma must have been calderas, now buried near the east end of the Snake River Plain (fig. 6; also see Stearns and others, 1939; Mabey, 1978; McBroome, 1981; Embree and others, 1982; Morgan and others, 1984; Morgan, 1988; 1992; Hackett and Morgan, 1988). Although erosion was predominant in much of the Yellowstone region, sediments accumulated locally. In northern Jackson Hole, fluvial gravels and sands called the Bivouac Formation by Love (1956a; b) accumulated in the fault-block basin at the foot of the Teton Range, prob-



EXPLANATION

HF — Caldera rim—Hachures toward collapse. HF, Henrys Fork caldera; BB, Big Bend Ridge caldera segment; SR, Snake River caldera segment; RM, Red Mountains caldera segment; WT, West Thumb caldera.

SC — Resurgent dome—ML, Mallard Lake; SC, Sour Creek

Major late Cenozoic fault zone—Arrows toward structurally lowest part

✳ Known volcanic vents—All others whose general locations are known lie in the same zones as those above.

Figure 5.—Major late Cenozoic volcanic and tectonic features of the Yellowstone Plateau volcanic field. The Yellowstone caldera comprises a western segment, centered on the Mallard Lake resurgent dome, and an eastern segment, centered on the Sour Creek resurgent dome. In part modified after Ruppel (1972), Love (1956d; 1961), Pardee (1950), Ross and Forrester (1947), Ross and others (1955), Love and Christiansen (1985).

ably both before and after emplacement of the tuff of Kilgore and the Conant Creek Tuff. The Heart Lake Conglomerate of the Red Mountains area is believed by Love and Keefer (1969; 1975) to have been derived from a localized tectonic uplift just before eruption of the first major ash-flow sheet of the Yellowstone Plateau as rhyolitic volcanism was beginning in the area. Gravels also were deposited in a broad valley that drained northward from the Washburn Range in the area of Mount Everts, in northern Yellowstone Park. The gravel of Mount Everts, now capped by welded tuffs, contains mainly volcanic clasts that apparently were derived from lower Tertiary volcanic rocks of the Washburn Range and

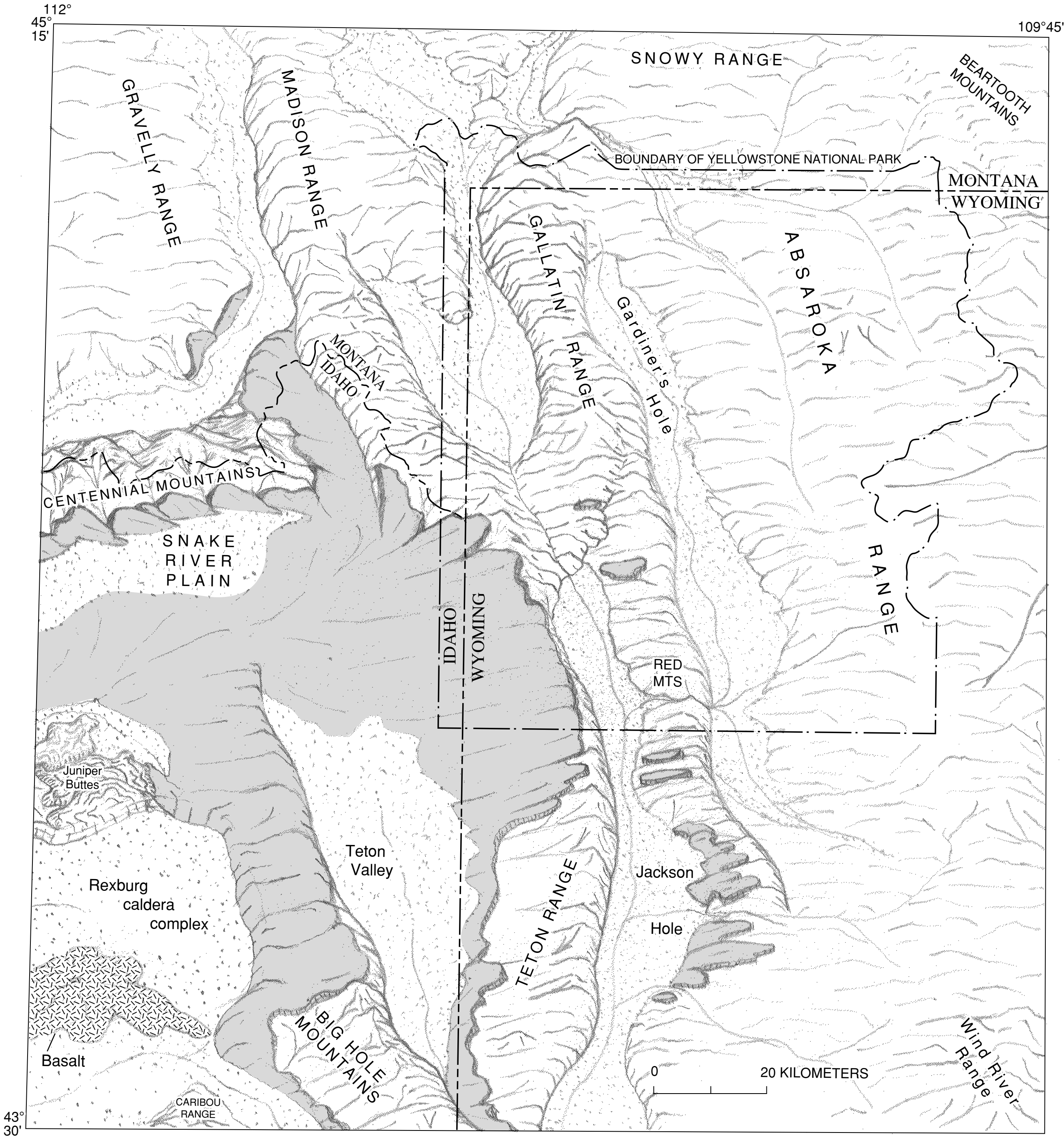


Figure 6.—Interpretive reconstruction of the Yellowstone Plateau region before initial plateau volcanism (a little before 2 Ma). The region was entirely an elevated and faulted mountainous terrain with no basin in the present plateau area. Gray areas are underlain by the tuff of Kilgore, Conant Creek Tuff, and older ash-flow tuffs of the eastern Snake River Plain.

practically none of the Precambrian, Paleozoic, and Mesozoic rocks that are now exposed widely to the north, west, and east. Upper Tertiary gravels also have been preserved in the Madison Valley and in the Swan Valley area of the Snake River west of the Idaho-Wyoming border by a capping of the oldest welded tuff of the Yellowstone Plateau volcanism.

Thus, the main elements of the present topography of areas now on the margin of the Yellowstone Plateau region had been initiated well before the plateau volcanism and were more or less continuous across the present area of the plateau and the Island Park area. Continued block faulting has cut off some areas that were open to emplacement of the oldest ash-flow sheets of the Yellowstone Plateau field, such as Swan Valley and Centennial Valley, by further uplift of the lowest topographic barriers overtopped by those ash flows.

The stratigraphy of the Yellowstone Plateau volcanic field is outlined in table 1. The part of the stratigraphy that is represented within Yellowstone National Park was described by Christiansen and Blank (1972); that of the Island Park area was described by Christiansen (1982). The recognition of three successive cycles of volcanism provides a natural basis for dividing the volcanic stratigraphy. The major ash-flow sheets erupted at the climax of each cycle are the principal stratigraphic units in areas outside the volcanic source region, and together they constitute the Yellowstone Group. Boyd (1961), who named the Yellowstone Tuff, did not recognize that it contained more than one ash-flow sheet. As Boyd mapped and described it, the Yellowstone Tuff contained the oldest and youngest of the three sheets recognized by Christiansen and Blank (1972). We, therefore, retained the name but revised its rank in renaming it the Yellowstone Group. During each of the three volcanic cycles, rhyolitic and basaltic lavas and minor pyroclastic deposits were erupted both before and after the major ash-flow eruptions. Most of the rhyolitic lavas are restricted to the source areas of the major ash-flow sheets, where calderas formed by collapse of the roofs of the magma-chambers that fed each of the three sheets. These flows thus form a caldera-filling complex, but related flows are also present locally in zones marginal to the caldera complexes. The precaldera and postcaldera rhyolites of each cycle and contemporaneous basalts are separated into mappable formations to complete the stratigraphic framework of the volcanic field.

The oldest recognized products of the plateau volcanism consist of basalts assigned to the Junction Butte Basalt, exposed in northeastern Yellowstone National Park (pl. 1), and the rhyolitic lava flow of Snake River Butte, in southern Island Park at the margin of what was shortly to become the first-cycle caldera (pls. 1, 2). Ages of all these units are between 2.2 and 2.1 Ma (Christiansen and Blank, 1972; Obradovich, 1992). The first ash-flow sheet of the Yellowstone Group, the Huckleberry Ridge Tuff, erupted at 2.1 Ma. This extraordinary eruption buried an area of 15,500 km^2 in a period of time so short that no erosion and no appreciable cooling of earlier parts of the deposit occurred before completion of the eruption, certainly within a few hours or days. Eruption of such an enormous volume of magma, about 2,500 km^3, in such a short time resulted in collapse of the magma-chamber roof to form a caldera somewhere between 75 and 95 km long, the first-cycle caldera of the Yellowstone Plateau volcanic field. This feature extended from Big Bend Ridge, west of Island Park, to the Central

Table 1.—Stratigraphy of the Yellowstone Plateau volcanic field.

Age	Volcanic cycle	Precaldera rhyolite	Caldera-forming ash-flow tuff (Yellowstone Group)	Postcaldera rhyolite	Contemporaneous plateau-marginal basalts
Pleistocene	Third			Plateau Rhyolite	Basalts of Snake River Group Osprey Basalt Madison River Basalt Basalt of Geode Creek Swan Lake Flat Basalt Gerrit Basalt Falls River Basalt Basalt of Mariposa Lake
			Lava Creek Tuff		
		Mount Jackson Rhyolite Lewis Canyon Rhyolite			Undine Falls Basalt Basalt of Warm River Basalt of Shotgun Valley
	Second			Island Park Rhyolite	Basalt of the Narrows
			Mesa Falls Tuff		
		Big Bend Ridge Rhyolite[1]			
Pliocene	First			Big Bend Ridge Rhyolite[1]	
			Huckleberry Ridge Tuff		
		Rhyolite of Snake River Butte			Junction Butte Basalt

[1]Part of the Big Bend Ridge Rhyolite comprises postcaldera flows of the first cycle and part comprises precaldera flows of the second cycle.

Plateau (fig. 1; pl. 3). Postcollapse rhyolitic lava flows erupted in the caldera area; remnants are now exposed west of Island Park and others may lie buried beneath the Yellowstone Plateau. Basalts younger than the Huckleberry Ridge Tuff and both older and younger than the succeeding Mesa Falls Tuff are interlayered with sediments near The Narrows of the Yellowstone River's Grand Canyon in northern Yellowstone National Park.

Rocks of the second volcanic cycle are not exposed within Yellowstone National Park but are present within a few kilometers of its west boundary. Early rhyolitic lavas of the second cycle are present near Bishop Mountain, west of Island Park. During or just after eruption of the more than 280 km^3 of the Mesa Falls Tuff from the Island Park area at 1.3 Ma, the source area collapsed to form the second-cycle caldera with a diameter of about 16 km. The west and north rims of the second-cycle caldera are nested against the northwest rim of the first-cycle caldera. Consequently, the Mesa Falls Tuff is distributed north of Island Park and disappears beneath basalts of the Snake River Plain to the west, but most Mesa Falls ash flows that swept southward and eastward were confined within the older, larger caldera and subsequently were buried. Postcaldera rhyolitic domes, designated the Island Park Rhyolite, are exposed within and near the second-cycle caldera. The basalts of Warm River and Shotgun Valley are exposed between the Mesa Falls and Lava Creek Tuffs in the vicinity of Island Park.

The third volcanic cycle began about 1.2 Ma with eruption of the oldest flows of the Mount Jackson Rhyolite. Rhyolitic lavas were erupted from a growing fissure system over a period of about 600,000 years around what would later become the ring-fracture zone of the Yellowstone caldera. The Undine Falls Basalt erupted during this time in northern Yellowstone National Park. The climax of the third cycle of the Yellowstone Plateau volcanic field came with eruption of the voluminous Lava Creek Tuff 640,000 years ago. Eruption of this welded pumice and ash, aggregating more than 1,000 km^3, again probably within a few hours or days, formed an ash-flow plateau that buried an area of more than 7,500 km^2 and, like the first two volcanic cycles, produced windblown ash that accumulated in ponds and hollows on the Great Plains of Nebraska and Kansas to depths of as much as 9 m and that can be found in Pleistocene deposits from California to Iowa, and Canada to Mexico (Powers and others, 1958; Wilcox, 1965; R. L. Reynolds, 1979; Naeser and others, 1973; Izett, 1981; Izett and Wilcox, 1982). At the end of these eruptions the Yellowstone caldera formed by collapse of the magma-chamber roof, producing an elliptical basin 85 km long and 45 km wide. Resurgent doming uplifted the caldera floor. Subsequent volcanism formed the voluminous lavas of the Plateau Rhyolite in and near the Yellowstone caldera, as well as the basalts of Mariposa Lake and Geode Creek and the Falls River, Gerrit, Swan Lake Flat, Madison River, and Osprey Basalts on the margins of the plateau, nearly surrounding the caldera (pl. 1). The youngest known rhyolite flows on the Yellowstone Plateau are about 70,000 years old. Basaltic eruptions of the Snake River Group occurred in the eastern Snake River Plain throughout much of this time and have continued into the Holocene.

VOLCANIC HISTORY OF THE FIELD

Having briefly reviewed the origin of the Yellowstone Plateau volcanic field, the history of each of its three cycles is discussed next, starting with the best understood.

THE THIRD VOLCANIC CYCLE

Because tectonism, erosion, and burial have attenuated the geologic record of the two older volcanic cycles, those cycles are best understood in the context of the better-preserved third cycle. The third eruptive cycle provides a model into which features of the first and second cycles can be placed to reconstruct a coherent history of the field as a whole.

PREERUPTION SETTING

An interpretation of the prevolcanic setting of the field was presented in the preceding section. By the time the third volcanic cycle began, notable changes had taken place in the Yellowstone region. Two major calderas and their related ash-flow plateaus had formed, and the calderas had been partly filled by rhyolitic lavas. Formation of these calderas and some structural sagging around them had disrupted the former continuity of the fault-block mountains that now surround the Yellowstone Plateau. The surrounding topography resembled the present terrain in many respects; the Madison, Gallatin, Absaroka, and Teton Ranges, and neighboring lesser mountainous ridges all were blocked out in more or less their present forms (fig. 7). An ash-flow plateau was centered near the present Madison Plateau and extended into the valleys leading outward from that area. Although the first- and second-cycle calderas were partly filled by rhyolitic tuffs and lavas, they remained as a composite topographic basin within the plateau region. Rhyolite flows probably blocked the caldera basin in the vicinity of the present Madison Plateau. The resulting physiography determined the distributions of ash-flow units of the third volcanic cycle, in particular the first-erupted parts. The area west of Island Park, now covered by basalts of the Snake River Group, was underlain at least partly by rhyolites.

Volcanic rocks of the first and second cycles had been deeply incised by erosion before the principal eruptive events of the third cycle began. This erosion is most vividly apparent in the area of lower Lava Creek and the Gardner River east of Mammoth Hot Springs (pl. 1). In that area, a deep valley cut the Huckleberry Ridge Tuff of the first cycle, leaving mesalike remnants of the welded tuff on Mount Everts and Terrace Mountain. This valley was at least 2¼ km wide and 250 m deep before pre-Lava Creek basalts were emplaced into it. Several tributary valleys drained the south rim of Mount Everts to join the master valley, and these tributary channels are now preserved by younger fillings of Lava Creek Tuff that since have been isolated by erosion of the present Lava Creek Canyon. These tuff-filled paleochannels were perplexing to Boyd (1961), who recognized only a single ash-flow sheet of his Yellowstone Tuff. He interpreted the channels as tuff vents of a late stage that had pierced the main body of Yellowstone Tuff. The actual situation is revealed by the following relations: (1) the tuff that fills the channels is of ash-flow origin, (2) a complete cooling-unit zonation is present in each of the tuff fillings, separate from that of the adjacent Huckleberry Ridge Tuff, (3) the tuff fillings incorporate talus from previously weathered Huckleberry Ridge Tuff along the contact, even at topographic levels where the paleochannel had been cut through the Huckleberry Ridge into underlying Cretaceous sedimentary rocks, and (4) the tuff fillings of the paleochannels are lithologically identical to the Lava Creek Tuff on the south side of Lava Creek Canyon but are distinct from the adjacent Huckleberry Ridge. These field relations have been confirmed by paleomagnetic measurements and K-Ar ages.

EARLY RHYOLITIC AND BASALTIC VOLCANISM

Rhyolitic volcanism of the third volcanic cycle began with the eruption of several large flows assigned stratigraphically to the Mount Jackson Rhyolite (figs. 8, 9). Another group of flows, the Lewis Canyon Rhyolite, erupted more or less simultaneously but is considered separately below. At least eight flows of the Mount Jackson sequence are exposed and one other flow may be inferred (fig. 8; pl. 1), but others might be present beneath the subsequent Yellowstone caldera. Boyd (1961) assigned these lavas to his "Jackson flows."

MOUNT JACKSON RHYOLITE

The best exposed of these rhyolitic lava flows form the lower walls of the Madison Canyon, near the West Entrance to Yellowstone National Park. Mount Jackson and Mount Haynes, named summits on the north and south sides of the canyon, were both carved from this flow, the Mount Haynes flow (fig. 8; pl. 1). The base of this flow is exposed poorly on the north side of the canyon about 3/4 km west of Harlequin Lake and overlies rocks that appear to be pumiceous fallout tuff that was fused by the heat and load of the overlying rhyolite flow (for discussion of such fused tuffs see Smith, 1960a, p. 800; Christiansen and Lipman, 1966, p. 680-684). This fused tuff may represent a pumice and ash cone deposited near the vent of the Mount Haynes flow before eruption of the lava. The Mount Haynes flow has a normal paleomagnetic polarity and yielded a K-Ar age of about 610,000 ka (table 2). The Mount Haynes flow and its fused basal tuff overlie an older rhyolite on the north side of the Madison Canyon, the Harlequin Lake flow. The Harlequin Lake flow has a reverse paleomagnetic polarity and a K-Ar age of about 840,000 years (table 2). Probably a third flow of Mount Jackson Rhyolite, the Big Bear Lake flow, is present in the area southwest of Mount Haynes, outcropping near Jack Straw Basin, Big Bear Lake, and Echo Canyon (fig. 10; pl. 1). The base of neither the Harlequin Lake nor the Big Bear Lake flow is exposed.

The Mount Haynes, Big Bear Lake, and Harlequin Lake flows are overlain and at least partly buried by the Lava Creek Tuff, and the surface topography of the buried flows influenced the form of the Lava Creek ash-flow sheet. Only the highest parts of the Mount Haynes and Big Bear Lake flows probably projected above the ash-flow sheet just after its emplacement. Differential compaction of the welded tuff formed a dip-slope surface down the flanks of the bulbous rhyolite flows and controlled later development of the drainage pattern. Thus, beyond the area of the buried flows the ash-flow plateau has a gently sloping nearly planar surface incised by a joint-modified dendritic drainage net whereas the area underlain by the flows and their ash-flow mantle has a more steeply dipping surface with a strongly rectangular joint- and fault-controlled drainage pattern (fig. 10). Cougar Creek, thus, probably marks the buried north edge of the Mount Haynes flow. The presence of a marked topographic reentrant south of Mount Haynes, intervening between the Mount Haynes and Big Bear Lake flows, suggests that the two flows are distinctly separate. It is possible, however, that the two are merely lobes of a single flow. In either case, the lobate forms of these lava flows, reflected through the mantle of welded tuff, show that the vent areas for all of them were farther southeast in an area that is now part of the Yellowstone caldera. The fused-tuff pyroclastic cone beneath the Mount Haynes flow suggests that its vent was not far from the present caldera wall.

The Lava Creek Tuff shows similar relations to those just described where it overlies another flow of the Mount Jackson Rhyolite around Moose Creek Butte, just west of Yellowstone National Park (fig. 8; pl. 1). There, however, more of the overlying welded tuff has been stripped from the flow by erosion, so that the relations along the marginal flow scarp may be seen clearly (pls. 1, 2). The age of the Moose Creek Butte flow is 1.2 Ma (table 2).

Another flow of the Mount Jackson Rhyolite is exposed northeast of the Yellowstone caldera, near the Grand Canyon of the Yellowstone and Broad Creek (fig. 8; pl. 1). This rhyolite, the Wapiti Lake flow, has been dated at about 1.2 Ma, and it has a reverse paleomagnetic polarity (table 2). The flow overlies Eocene rocks of the Absaroka Volcanic Supergroup, burying a paleovalley cut into these rocks. Both Boyd (1961, p. 395-396) and Holmes (1883a, p. 40) noted that rhyolites of that area were emplaced on a canyon or valley topography with drainage almost along the present lower courses of Deep Creek (Jasper Creek of Holmes) and Broad Creek. Boyd noted that

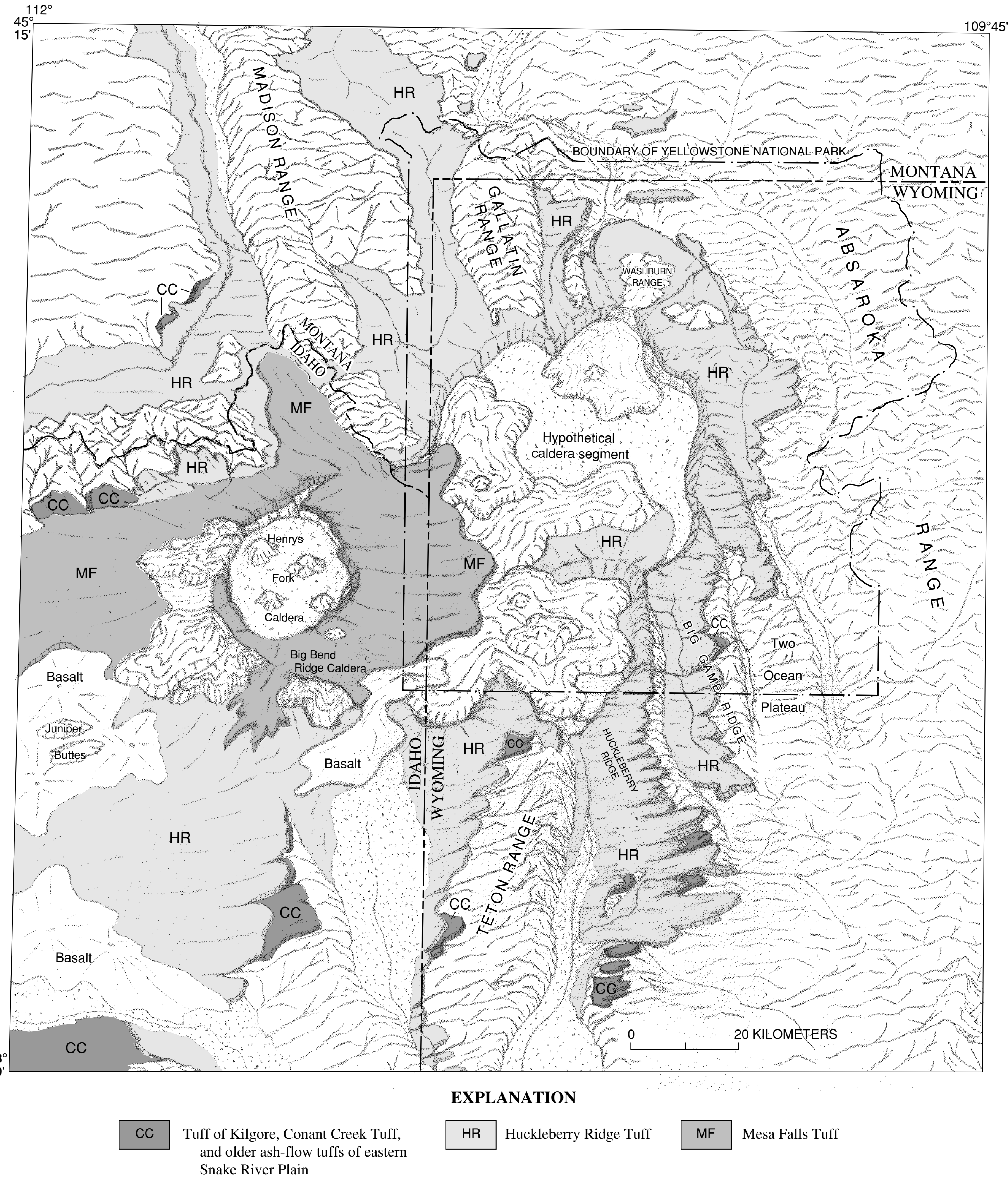
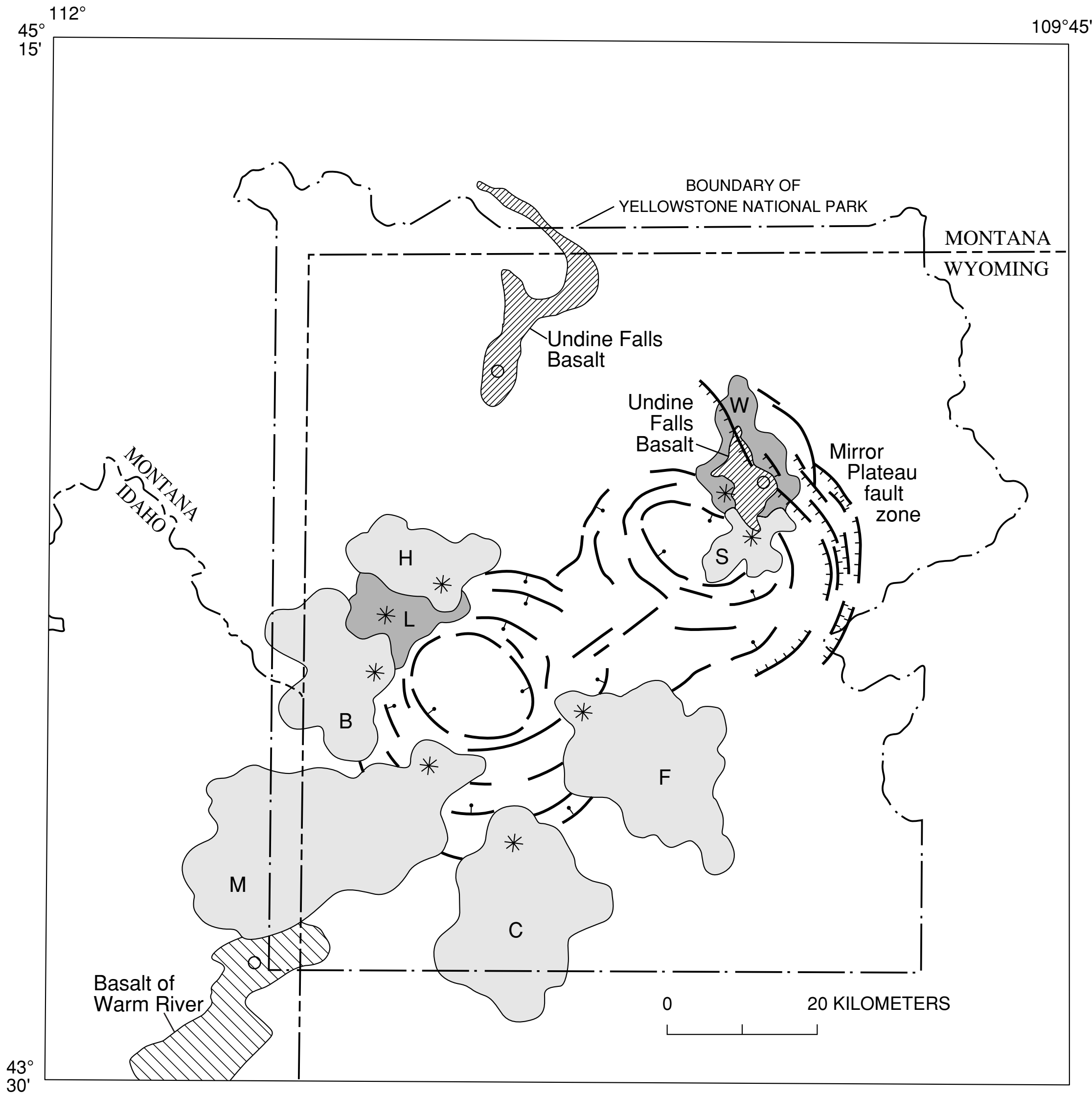


Figure 7.—Interpretive reconstruction of the Yellowstone Plateau region just before the third volcanic cycle (about 1.2 Ma). The Madison, Gallatin, Absaroka, and Teton Ranges and nearby mountain ridges were all present in more or less their present forms. An ash-flow plateau surrounded the present Madison Plateau.



EXPLANATION

Flows of Mount Jackson Rhyolite

- W — Wapiti Lake flow
- S — Possible Stonetop Mountain flow
- F — Flat Mountain flow
- M — Moose Creek Butte flow
- B — Big Bear Lake flow
- L — Harlequin Lake flow
- H — Mount Haynes flow
- C — Lewis Canyon Rhyolite (one or more flows)

- Probable rhyolitic vent
- Probable basaltic vent
- Tectonic faults that function as ring faults near the caldera
- Schematic representation of incipient ring faults—Bar and ball on down-dropped side

Figure 8.—Reconstructed distribution of early basalts and rhyolites of the third volcanic cycle (older than about 0.6 Ma), and incipient ring-fracture zones. The location of the subsequently formed Yellowstone caldera is indicated schematically by hypothetical ring faults.

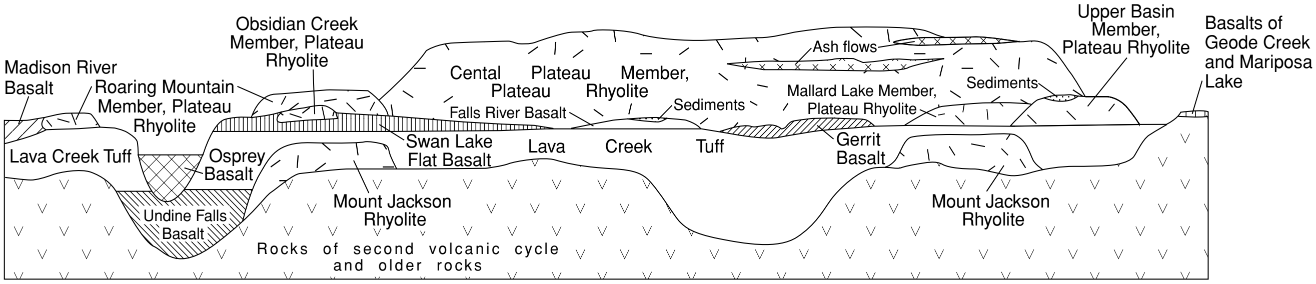


Figure 9.—Stratigraphic diagram of rocks of the third volcanic cycle of the Yellowstone Plateau volcanic field (about 1.2 to 0.1 Ma). The Yellowstone caldera formed immediately after or late during eruption of the Lava Creek Tuff. Basalts and the Obsidian Creek and Roaring Mountain Members of the Plateau Rhyolite occur only outside the caldera.

Table 2.—*Ages and paleomagnetic polarities of rocks from the Yellowstone Plateau volcanic field.*

[1]Except as otherwise noted, polarities based on field measurements with portable fluxgate magnetometer by R. L. Reynolds, H. R. Blank, and R. L. Christiansen.

[2]For basis of "probable ages" see discussions in text of designated units.

[3]All flows measured of these stratigraphic units had the same polarity.

[4]Polarity redetermined in laboratory after alternating-field demagnetization.

[5]K-Ar age from Obradovich (1992), with revisions; weighted mean age shown for multiple analyses of same unit.

[6]Weighted mean age of the group of two dated flows bracketed by lines.

[7]Weighted mean age of the group of five dated flows bracketed by lines.

[8]Weighted mean age of the group of nine dated flows and tuff bracketed by lines.

[9]$^{40}Ar/^{39}Ar$ age from Gansecki and others (1996).

[10]Basaltic lavas of the Snake River Group investigated within the area of plate 2.

[11]Reynolds (1977).

[12]Weighted mean age of the group of three dated flows bracketed by lines.

[13]$^{40}Ar/^{39}Ar$ age determined by M. A. Lanphere (written communication, 2000).

[14]Weighted mean age of the group of two dated flows bracketed by lines.

Unit	Paleomagnetic polarity[1]	Analytical age[2] (Ma)	Probable age[2] (Ma)
Plateau Rhyolite	Normal[3,4]		
Central Plateau Member			
Pitchstone Plateau flow		0.070±0.002[5]	0.071±0.002[6]
Grants Pass flow		0.072±0.003[5]	
Roaring Mountain Member			
Crystal Spring flow		0.080±0.002[5]	
Gibbon River flow		0.090±0.002[5]	
Central Plateau Member			
Solfatara Plateau flow		0.110±0.003[5]	
Hayden Valley flow		0.102±0.004[5]	
West Yellowstone flow		0.108±0.001[5]	0.110±0.001[7]
Bechler River flow		0.117±0.002[5]	
Summit Lake flow		0.112±0.002[5]	
Obsidian Creek Member			
Gibbon Hill dome		0.116±0.008[5]	
Central Plateau Member			
Buffalo Lake flow		0.160±0.003[5]	
Nez Perce Creek flow		0.160±0.002[5]	
Elephant Back flow		0.153±0.002[5]	
West Thumb flow		0.147±0.004[5]	

Table 2.—*Ages and paleomagnetic polarities of rocks from the Yellowstone Plateau volcanic field*—Continued

Unit	Paleomagnetic polarity[1]	Analytical age[2] (Ma)	Probable age[2] (Ma)
Aster Creek flow		0.155±0.003[5]	0.161±0.001[8]
Tuff of Bluff Point		0.162±0.002[5]	
Dry Creek flow		0.162±0.002[5]	
Mary Lake flow		0.165±0.004[5]	
Mallard Lake Member			
Mallard Lake flow		0.151±0.004[5]	
Roaring Mountain Member			
Obsidian Cliff flow		0.183±0.003[5]	
Upper Basin Member			
Scaup Lake flow		0.198±0.008[9]	
Snake River Group, Island Park area[10]	Normal[3]		<0.20
Gerrit Basalt	Normal[3,4]		
Hatchery Butte flow		0.199±0.009[5]	
Osprey Basalt	Normal[3]		
Lamar River flow		0.220±0.041[5]	
Madison River Basalt	Normal[3,4]		>0.13, <0.40
Plateau Rhyolite			
Obsidian Creek Member			
Willow Park dome		0.316±0.005[5]	
Roaring Mountain Member			
Cougar Creek dome		0.399±0.003[5]	
Basalt of Geode Creek	Normal[3]		>0.13, <0.64
Falls River Basalt	Normal[3,4]		>0.11, <0.64
Swan Lake Flat Basalt	Normal[3]		>0.32, <0.64
Plateau Rhyolite			
Upper Basin Member			
Dunraven Road flow		0.486±0.042[10]	
Canyon flow		0.484±0.015[10]	0.481±0.008[12]
Tuff of Sulphur Creek		0.479±0.010[10]	
Biscuit Basin flow		0.516±0.007[10]	
Lava Creek Tuff	Normal[4,11]	0.640±0.002[13]	
Undine Falls Basalt	Normal[3,4]		
Undine Falls, upper flow		0.588±0.026[5]	
Mount Jackson Rhyolite			
Mount Haynes flow	Normal[4]	0.609±0.006[5]	
Big Bear Lake flow	Normal		>0.64, <0.78
Basalt of Warm River	Reverse[3,4]		
Warm River flow		0.759±0.052[5]	
Basalt of Shotgun Valley	Reverse[3]		>0.78, <1.3
Basalt of Bitch Creek	Reverse[3]		>0.78, <1.3
Mount Jackson Rhyolite			
Harlequin Lake flow	Reverse[4]	0.839±0.008[5]	
Lewis Canyon Rhyolite	Reverse[3,4]		
Lewis Canyon flow		0.853±0.007[5]	
Mount Jackson Rhyolite			
Flat Mountain flow		0.929±0.034[5]	
Wapiti Lake flow	Reverse[4]	1.16±0.01[5]	
Moose Creek flow		1.22±0.01[5]	
Sediments and basalts of the Narrows			
Uppermost basalt flow	Normal	1.30±0.35[5]	

Table 2.—*Ages and paleomagnetic polarities of rocks from the Yellowstone Plateau volcanic field*—Continued

[1]Except as otherwise noted, polarities based on field measurements with portable fluxgate magnetometer by R. L. Reynolds, H. R. Blank, and R. L. Christiansen.
[2]For basis of "probable ages" see discussions in text of designated units.
[3]All flows measured of these stratigraphic units had the same polarity.
[4]Polarity redetermined in laboratory after alternating-field demagnetization.
[5]K-Ar age from Obradovich (1992), with revisions; weighted mean age shown for multiple analyses of same unit.
[6]Weighted mean age of the group of two dated flows bracketed by lines.
[7]Weighted mean age of the group of five dated flows bracketed by lines.
[8]Weighted mean age of the group of nine dated flows and tuff bracketed by lines.
[9]$^{40}Ar/^{39}Ar$ age from Gansecki and others (1996).
[10]Basaltic lavas of the Snake River Group investigated within the area of plate 2.
[11]Reynolds (1977).
[12]Weighted mean age of the group of three dated flows bracketed by lines.
[13]$^{40}Ar/^{39}Ar$ age determined by M. A. Lanphere (written communication, 2000).
[14]Weighted mean age of the group of two dated flows bracketed by lines.

Unit	Paleomagnetic polarity[1]	Analytical age[2] (Ma)	Probable age[2] (Ma)
Island Park Rhyolite			
Osborne Butte dome	Reverse	1.28±0.01[5]	
Elk Butte dome	Reverse		
Lookout Butte dome	Reverse[4]		
Warm River Butte dome	Reverse	1.24±0.02[5]	
Mesa Falls Tuff	Reverse[4, 11]	1.292±0.005[13]	
Big Bend Ridge Rhyolite			
Green Canyon flow		1.17±0.01[5]	
Tuff of Lyle Spring	Reverse[4]	1.19±0.01[5]	
Bishop Mountain flow	Reverse[4]	1.20±0.01[5]	
Sediments and basalts of the Narrows			>1.3, <2.1
Lowermost basalt flow	Reverse		
Big Bend Ridge Rhyolite			
Headquarters flow	Normal[4]	1.83±0.01[5]	1.82±0.01[14]
Blue Creek flow	Normal[4]	1.77±0.02[5]	
Huckleberry Ridge Tuff	Subhorizontal[4, 11]	2.053±0.006[13]	
Rhyolite of Snake River Butte			
Snake River Butte flow	Normal (steep to E)[4]	1.99±0.02[5]	
Junction Butte Basalt			
Mount Everts area	Normal[3,4]		
Upper flow		2.01±0.05[5]	
Tower Falls-Junction Butte area	Reverse[3,4]		
Overhanging Cliff flow		2.16±0.04[5]	

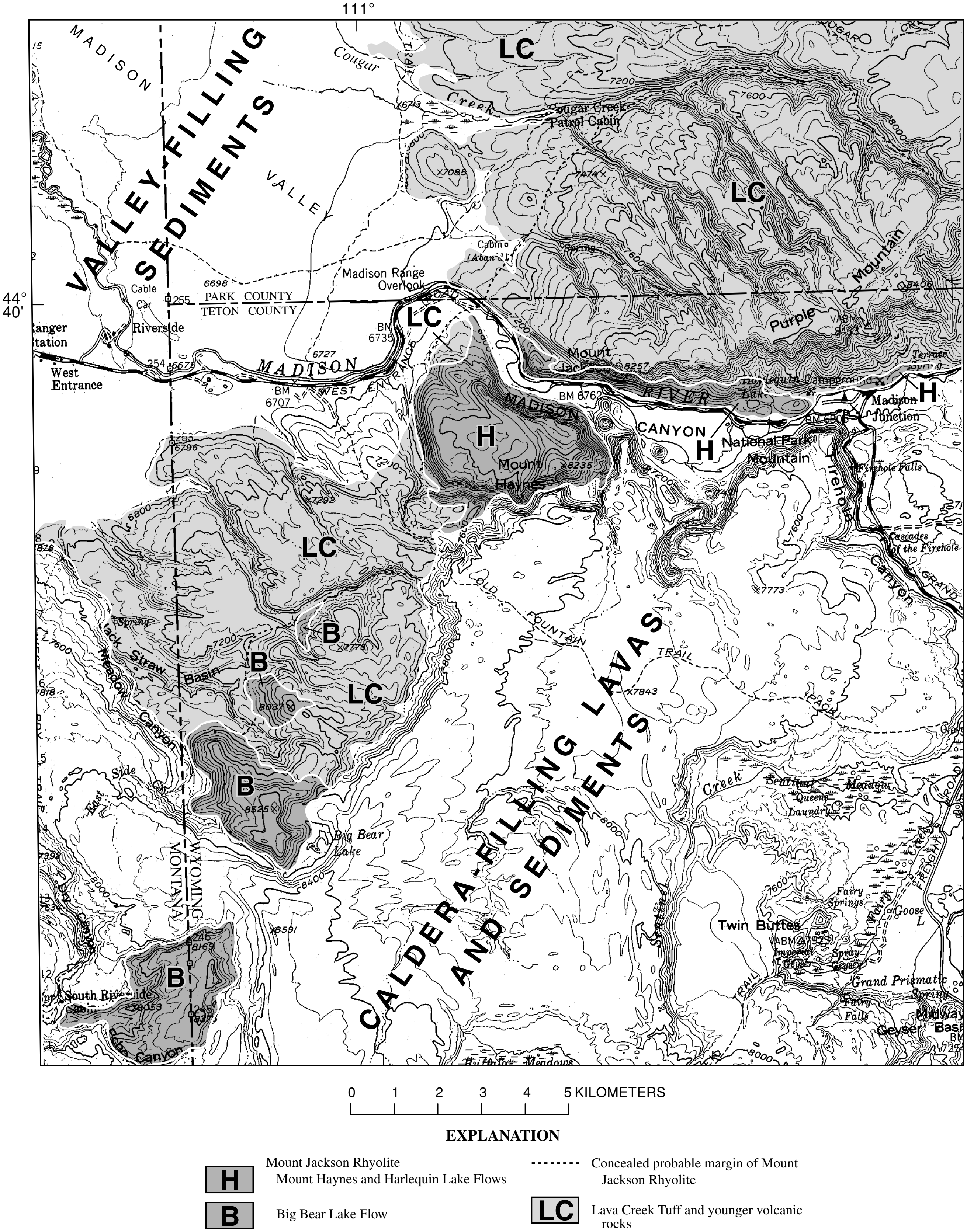


Figure 10.—Topographic expression of principal geologic units in the upper Madison River area, Yellowstone National Park. Base from USGS 1:62,500 topographic map of Madison Junction quadrangle (1958). Contour interval 80 feet.

differential compaction of the welded-tuff sheet would tend to mimic the underlying topography although he did not note the presence of a rhyolitic lava flow at the base of the welded-tuff in the old valley. Geologic mapping shows clearly that the ancestral drainage filled by the Wapiti Lake flow was that of Broad Creek itself and not of the Grand Canyon. Thus, the course of the Yellowstone River above Broad Creek had not yet been established. The vent area of the Wapiti Lake flow is not exposed, but the characteristic lobate form of the rhyolite flow shows that its vent was in the area of Broad Creek near Wapiti Lake, inside the present Yellowstone caldera (fig. 8). A downfaulted part of the Wapiti Lake flow, exposed around upper Broad Creek (pl. 1), was regarded as pre-Huckleberry Ridge by Christiansen and Blank (1972), but that interpretation was negated by K-Ar dating.

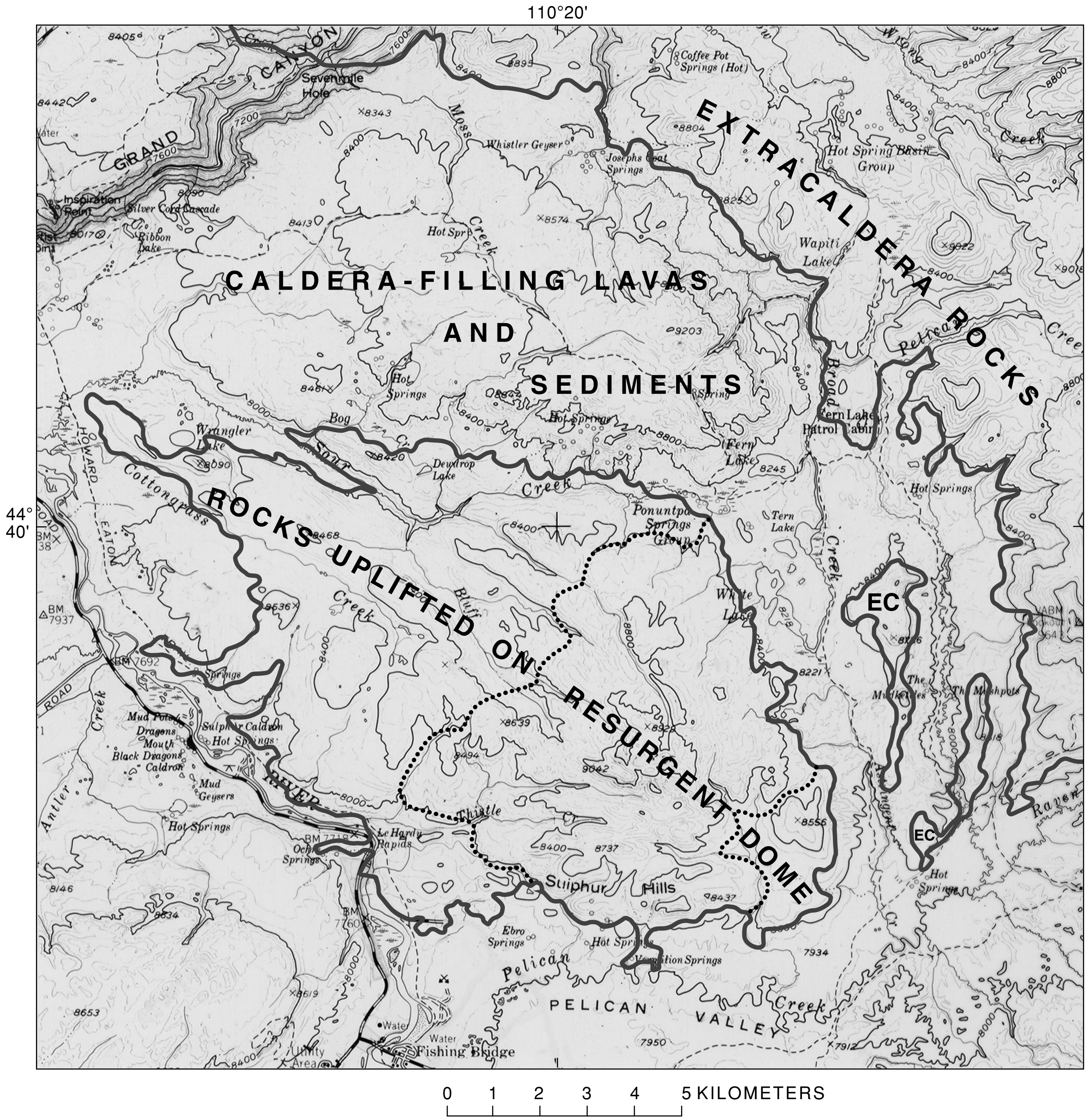


Figure 11.—Topographic expression of principal geologic units in the Stonetop Mountain area, Yellowstone National Park. Dotted line, presumed margin of possible Stonetop Mountain flow where buried by Lava Creek Tuff: EC, extracaldera rocks. Base from USGS 1:62,500 Canyon Village quadrangle (1959); contour interval 80 feet.

A Mount Jackson Rhyolite flow may be buried by the Lava Creek Tuff within the Yellowstone caldera near Stonetop Mountain, north of Yellowstone Lake. No unequivocal outcrop of such a flow is known in this area, but the general nature of the topographic surface and the drainage pattern formed on the Lava Creek Tuff there (fig. 11) are much like those of the Madison Canyon area, where the ash-flow sheet is underlain by the Big Bear Lake, Harlequin Lake, and Mount Haynes flows (fig. 10), and around the Moose Creek Butte flow. The topography of the Stonetop Mountain area has been modified by a system of northwest-trending normal faults, mainly lying farther northwest along Sour Creek and Cottongrass Creek, but the presence of a rather bulbous buried surface is strongly suggested near Stonetop Mountain. Possibly this is the near-vent part of the Wapiti Lake flow (pl. 1; figs. 8 and 11).

At Flat Mountain, southwest of Yellowstone Lake, yet another Mount Jackson Rhyolite flow is exposed beneath the Lava Creek Tuff and is dated at about 930,000 ka (table 2). The only exposures of the Flat Mountain flow (fig. 8) are on the escarpment of the Yellowstone caldera where it was uplifted in a pair of fault blocks (pl. 1). The top of the flow is exposed higher than the Lava Creek Tuff, at the summit of Flat Mountain, and it may never have been buried there. The shape of the flow, inferred from its present exposure and from its absence between the Lava Creek Tuff and an older welded tuff just south of the Flat Mountain summit, shows that the vent must have been within the present area of the Yellowstone caldera.

Most Mount Jackson flows are characterized by about 25-40 percent phenocrysts, varying among the various flows from about 1 mm to about 4 mm average diameter; only the Flat Mountain flow has sparse phenocrysts. Quartz, sanidine, and subordinate oligoclase are the principal minerals; magnetite, clinopyroxene, and rare sphene are minor. Boyd (1961, p. 392-393) reported rare biotite in his Jackson flows but did not note the localities at which it was found. I have not found biotite phenocrysts in the Mount Jackson Rhyolite.

All known flows of the Mount Jackson Rhyolite lie near the margins of the Yellowstone caldera (fig. 8), and all had their vent areas in or near what was subsequently to become the compound ring-fracture zone of the caldera. The flows appear to have erupted over a period of about 600,000 years during an early phase of the third cycle of the volcanic field. They are the first rhyolites known to have erupted on the Yellowstone Plateau after a quiescent period of about a million years following the first volcanic cycle. The relation of their vents to the Yellowstone caldera margin thus implies incipient formation of the ring-fracture zone during this precaldera eruptive phase, which appears to have accompanied the insurgence of a body of rhyolitic magma into a large shallow-level compound magma chamber. Magmatic insurgence, ring-fracturing, and eruption of the Mount Jackson Rhyolite correspond, then, to stage I of the resurgent cauldron cycle of Smith and Bailey (1968a), regional tumescence and generation of ring fractures. The rhyolite flows record the occasional leakage of magma to the surface during this process.

Although the Mount Jackson Rhyolite seems to have been controlled by fractures related to magmatic insurgence, a degree of localization by tectonic structures may be implied by the presence of most of the known flows near places where faults that bound major tectonic blocks intersect the caldera margin. Thus the flows in the Madison Canyon area lie near the projection of the Hebgen fault zone, the Wapiti Lake flow is in the Mirror Plateau fault zone, and the Flat Mountain flow is exposed in the fault blocks of Flat Mountain and Chicken Ridge (pl. 1; fig. 5). These regional normal faults also may have functioned as radial faults complementary to the circumferential ring-fracture zone during precaldera tumescence, rhyolitic extrusions being localized near the intersections of radial and circumferential faults.

LEWIS CANYON RHYOLITE

Additional pre-Lava Creek rhyolitic lavas occur south of any mapped Mount Jackson Rhyolite flows. These rhyolites differ petrographically from the Mount Jackson flows, and Christiansen and Blank (1972) named them the Lewis Canyon Rhyolite. Petrographically and chemically, the Lewis Canyon Rhyolite is much like the early postcaldera Upper Basin Member of the Plateau Rhyolite, containing abundant phenocrysts of quartz, plagioclase, and augite but only sparse sanidine. The plagioclase phenocrysts are andesine, commonly embayed, and show sieved cores. The similarity of the Lewis Canyon Rhyolite to flows of the postcaldera Upper Basin Member, as well as the position of the Lewis Canyon Rhyolite—filling a caldera segment of the first volcanic cycle, as described later—originally suggested to us that it had a role in the first cycle rather than the third (Christiansen and Blank, 1972, p. B10-B11). However, the age of the Lewis Canyon Rhyolite has been determined as about 0.85 Ma by K-Ar dating of sanidine (table 2). Perhaps, being outside the main third-cycle source area, the Lewis Canyon has tapped magma from a level deeper than that reached by other pre-Lava Creek rhyolite flows. Possibly the Lewis Canyon magma represents a local residuum of the first-cycle magma; in some sense, although erupted simultaneously with the third cycle, the Lewis Canyon might be considered a late first-cycle rhyolite.

BASALTS

Relatively minor basaltic volcanism occurred locally during this first stage around the margins of the volcanic system, both to the southwest and to the north. The basalts of Warm River and Shotgun Valley (pl. 2) lie southwest of the Yellowstone Plateau between the Mesa Falls and Lava Creek Tuffs (pls. 2, 3). One flow of the basalt of Warm River has been dated as about 760 ka (table 2), and the reverse paleomagnetic polarity of all the flows is consistent in showing them to be older than 780 ka (Baksi and others, 1992; Spell and McDougall, 1992). Island Park, where the second volcanic cycle was centered, is flanked on the southeast by the basalt of Warm River and on the northwest by the similar basalt of Shotgun Valley; the basalts are shown in table 1 as contemporaneous with the third volcanic cycle. It should be emphasized, however, that plateau-marginal basaltic volcanism has been intermittently active throughout the history of the field, and it is only a matter of descriptive convenience to discuss the basalts with their contemporaneous rhyolitic cycles.

Other basaltic flows that erupted during the time of the Mount Jackson Rhyolite eruptions are included stratigraphically in the Undine Falls Basalt. Flows of this formation in and near its type locality in northern Yellowstone National Park (fig. 8; pl. 1) came from unknown vents north and northwest of the Washburn Range and flowed down a system of deeply eroded valleys approximately along the present courses of the Gardner and Yellowstone Rivers and Obsidian, Lava, and Blacktail Deer Creeks. The basalt flows, which are interlayered with fluvial gravels along these old valleys, have been cut through by the modern drainages and now lie as much as 150 m above the streams. These flows are concordant with the overlying Lava Creek Tuff and all have normal paleomagnetic polarities (table 2); they appear to predate the Lava Creek Tuff only slightly. One flow near Undine Falls has been dated (table 2) as 588±26 ka. The only other basalt assigned to the Undine Falls (fig. 8) occurs in the area of Broad and Shallow Creeks just northeast of the Yellowstone caldera (pl. 1). It lies between the 1.2-Ma Wapiti Lake flow and the 640-ka Lava Creek Tuff. A small cinder cone is preserved at the apparent vent for this basalt, only about a kilometer from the caldera wall.

THE LAVA CREEK TUFF

The Lava Creek Tuff erupted about 640,000 years ago (table 2), as interpreted from $^{40}Ar/^{39}Ar$ dating by M.A. Lanphere (2000, written communication). The Lava Creek Tuff was emplaced by the eruption of an enormous volume of rhyolitic pumice from vents within the area of magmatic insurgence that had been outlined during the previous 600,000 years or so by the Mount Jackson Rhyolite. Clear evidence shows that these gigantic pyroclastic eruptions vented within the area of the Yellowstone caldera and suggests that the vents were fissures of the ring-fracture zone that had been generated during the preceding magmatic insurgence.

In most vertical sections, the Lava Creek Tuff forms a single cooling unit, commonly compound as defined by Smith (1960b). That is, textural features related to welding and primary crystallization of the sheet occur in a sequence related to emplacement of the entire sheet before complete cooling of any of its parts, but there are some variations from the simplest pattern that would result from instantaneous emplacement at uniform temperature and simple cooling of the entire sheet without break. Glassy zones, representing chilled margins against the ground and the air, occur at the bottom and top of the sheet, but there is no continuous glassy zone in its center.

In defining the Lava Creek Tuff, Christiansen and Blank (1972) noted that the ash flows that constitute the bulk of the formation can be divided into two principal units by a widespread reversal of the welding-density profile (fig. 12). This reversal commonly is demonstrated by a less densely welded vapor-phase zone between more densely welded devitrified zones. It is recognized as occurring throughout the sheet at a single stratigraphic horizon because it coincides with a phenocryst-poor zone between relatively phenocryst-rich zones. These relations indicate a pause in the sequence of ash-flow eruptions or a change in the nature or composition of the eruptions that allowed partial cooling of several ash flows before they were overlain by hotter flows. The deposits below this cooling reversal are recognized as member A of the Lava Creek Tuff; the deposits including the reversal and above it are member B. These lower and upper members were defined stratigraphically by Christiansen and Blank (1972). An example of a section containing the contact between the two is illustrated in figure 12.

MEMBER A

Member A of the Lava Creek Tuff is mainly exposed in western and north-central Yellowstone National Park and is as thick as about 480 m in the vicinity of Purple Mountain, north of Madison Canyon (pl. 1). The general horizon of the base of the member is exposed only in the Purple Mountain area, and the basal contact itself is everywhere covered. The lower part of member A, about two-thirds of the section at Purple Mountain, contains fewer and smaller phenocrysts than the upper part and is less densely welded. Pumice inclusions in this lower part are generally larger and more abundant than in most of the upper part and are especially conspicuous in outcrop because of the less dense welding (fig. 13). The lower part of member A commonly is pinkish to purplish but varies to gray and pale yellow or orange as well. Boyd (1961, p. 393) called this lower part the Purple Mountain Pumice Breccia, but that name is not used here because it implies a greater degree of stratigraphic distinction from the remainder of the ash-flow sheet than is justified. The lower part of member A contains larger and more abundant lithic inclusions than the more widespread upper part. Most of these inclusions are rhyolitic, both lavas and welded tuffs. The welded-tuff inclusions are petrographically identifiable as belonging to the Yellowstone Group and presumably were derived from Huckleberry Ridge and Mesa Falls Tuffs in the walls of the ash-flow vents, now downdropped south of

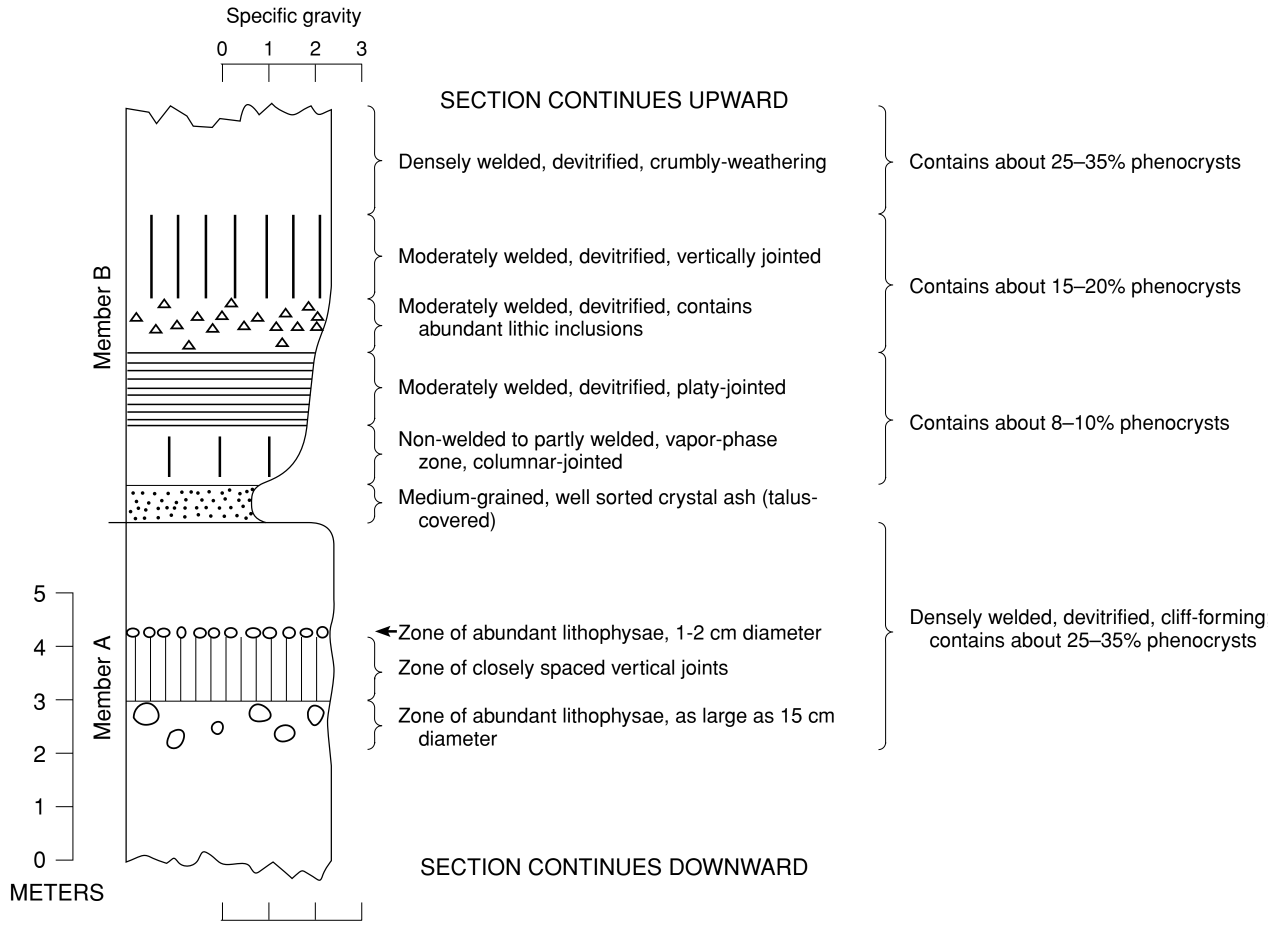
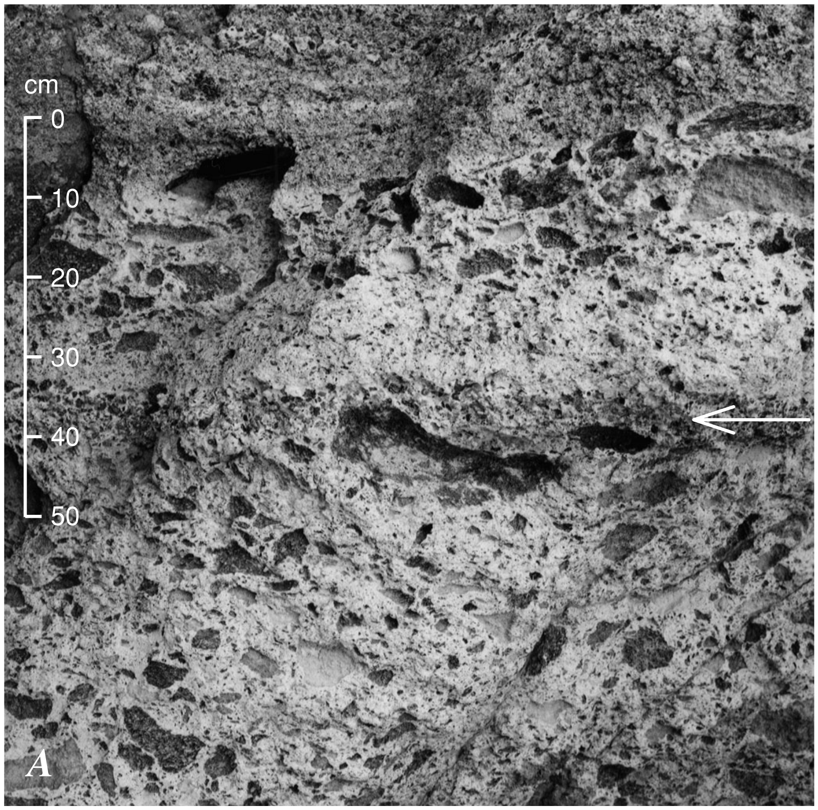


Figure 12.—Partial columnar section of Lava Creek Tuff in the Virginia Cascade area, Yellowstone National Park, showing reversal of the welding-density profile. Member A consists of densely welded devitrified tuff; at the base of member B, medium-grained crystal ash is overlain by nonwelded devitrified tuff, grading up to densely welded. Right edge of column shows vertical variation in specific gravity, as indicated by scale.

Purple Mountain. Not only is the lower part less densely welded than most of the upper part of member A, but the degree of welding of the lower part varies more in vertical sections. Some of this variability may have resulted from the irregular topographic surface of Mount Jackson Rhyolite onto which the lower part was emplaced. Glassy lenses are present locally in the section where certain ash flows were chilled against now shallowly buried ledges of the rhyolite flows. The lower part is known to crop out only in an 11-km long area of the Madison and Gibbon Canyons. Similar rocks exposed around Stonetop Mountain, north of Yellowstone Lake (Boyd, 1961), are parts of member B.

The upper part of member A crops out over a larger area than the lower part, from the South Fork of the Madison River, 7 km west of Yellowstone National Park, to the area of Bluff Creek, 8 km north of Yellowstone Lake (pl. 1). It is densely welded in most outcrops and has abundant large phenocrysts, thus contrasting sharply with the lower part. The fresh rock generally is light gray but weathers to crumbly brown and commonly forms steep bluffs with large rounded joint columns (fig. 14). The contact between the lower and upper parts of member A is exposed only on the face of Purple Mountain, where it is quite sharp. Generally, however, the actual contact is obscured to some degree by talus and slopewash. In a single exposure, high in a gully north of Madison Junction, the contact is perfectly exposed, planar, and appears to reflect scouring of the top of the lower part by the basal ash flows of the upper part. A few angular inclusions of welded tuff, identical in appearance to those occurring just below the contact, occur in the basal meter of the upper part of member A at this contact. Nevertheless, the entire section is densely welded and devitrified through the contact zone. Thus, it appears that the fragments were torn up from the underlying ash flows as they welded in place but before the contact surface had cooled or been eroded by any atmospheric agent.



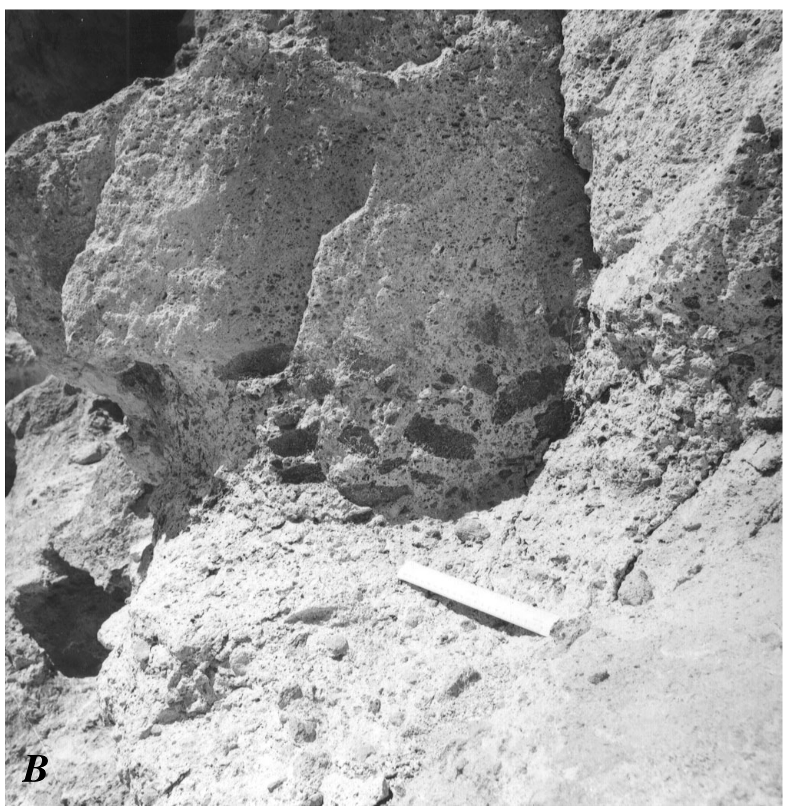


Figure 13.—Lower part of member A, Lava Creek Tuff, at Tuff Cliff, base of Purple Mountain 400 m northeast of Madison Junction, Yellowstone National Park. All the tuff has undergone vapor-phase crystallization. ***A***, Two pumice-rich ash flows separated by sorted layer (arrow); note fine-grained base of upper ash flow. Additional sorted layers at top. ***B***, Contact between two ash flows, showing abundant large pumice clasts concentrated at top of lower ash flow. Scale is 20 cm.

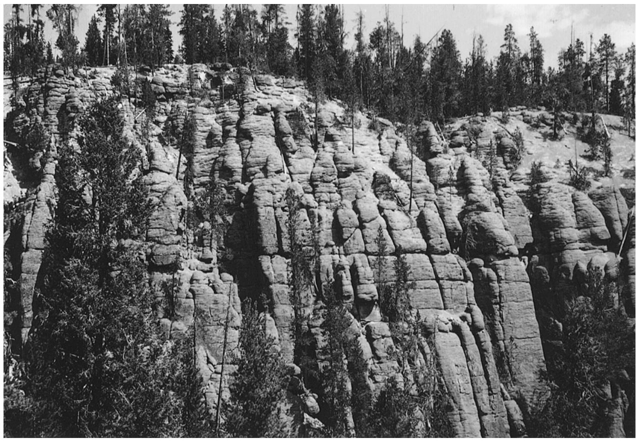


Figure 14.—Typical outcrop of upper part of member A, Lava Creek Tuff, on cliffs just south of Monument Geyser Basin, Yellowstone National Park. The upper part of member A characteristically forms steep bluffs with large rounded joint columns.

BEDDED ASH

The contact zone between members A and B, occurring at the base of a strongly jointed but weakly welded erosionally nonresistant zone, is rarely well exposed. At most localities the contact is mapped by the presence, downslope from the contact, of slabby relatively phenocryst-poor welded tuff from the lower part of member B. The only readily accessible good exposure of the immediate contact zone is in roadcuts along the Virginia Cascade loop road (fig. 12; pl. 1), but even there the actual contact generally is covered by slopewash at the base of the cuts. When the roadcuts were relatively fresh in 1965 and 1966 it was possible to dig out the contact at this locality; about 20-30 cm of loose crystal ash was found between the welded tuffs of the two members. This crystal ash may be part of a ground layer (Walker and others, 1981) related to member B. The few shards present in this ash, which are only partly devitrified, may indicate that a cooling break is present between the members at this locality, but the base of the immediately overlying ash flow, although only partially welded, is completely crystallized to form a vapor-phase zone. The same crystal ash at the base of partially welded ash-flow tuff of member B is exposed locally east of the Yellowstone River near Hayden Valley; I have also recognized it in drill cores from USGS research drill holes Y-9 and Y-12 in Norris Geyser Basin and Y-11 in the Sulphur Cauldron area (pl. 1). At none of these other localities have any still-glassy shards been noted in the crystal ash.

In most exposures of the base of member B where member A is not present, the ash-flow tuff is underlain by bedded pumiceous ash, generally white or very light gray in color but locally weathered to pale yellowish gray. Talus and slopewash commonly obscure this basal deposit. Bedding in the basal ash generally is markedly continuous and planar. Beds vary from laminae less than a millimeter thick to layers nearly a meter thick, but the thicker layers, defined by more or less uniform grain size and mechanical constitution, are almost always found to be internally laminated. Small-scale planar cross lamination occurs locally, and accretionary lapilli are present in some layers. Sparse lithic fragments in the ash beds, commonly 1-5 cm in diameter, are draped over by succeeding layers. Where the basal ash beds overlie uneven surfaces, they mantle irregularities with constant thickness. The bedded ash locally overlies gravels but in many places lies on weathered bedrock surfaces, not changing appreciably in thickness within a small area even at varied heights above the local base of deposition. These features, taken together, are interpreted as indicating the basal pumiceous ash to be an airborne primary fallout deposit, not subaqueously reworked. At most localities the lower half of this fallout deposit, consisting mainly of shards and small pumice fragments, is finer grained than the upper half, which consists of concentrated crystals and small lithic fragments (fig. 15). The thickness of the basal fallout deposit varies from more than 3 m in areas near the Yellowstone caldera to less than a meter at many localities around the margins of the rhyolite plateau; the deposit is absent in some localities.

Because most known exposures of the basal fallout ash occur at places where member A of the Lava Creek is not present and member B overlies older rocks, it is uncertain from stratigraphic evidence alone whether these fallout eruptions were direct precursors of the ash-flow eruptions of member B or whether they mantled much of the surrounding area during or before the eruption of member A. Clinopyroxene is the most abundant mafic phenocryst in the bedded ash, as it is in member B; member A has relatively abundant hornblende and allanite and relatively little clinopyroxene. These relations suggest that at least most of the ash was erupted from the member B magma. At one locality, the south end of the Snake River Bridge at Flagg Ranch about 3 km south of Yellowstone National Park (pl. 1), welded tuff of member B overlies a fallout deposit, the upper part of which is a coarse crystal ash. This deposit in turn overlies an older, glassy partly welded tuff of uncertain affinity.

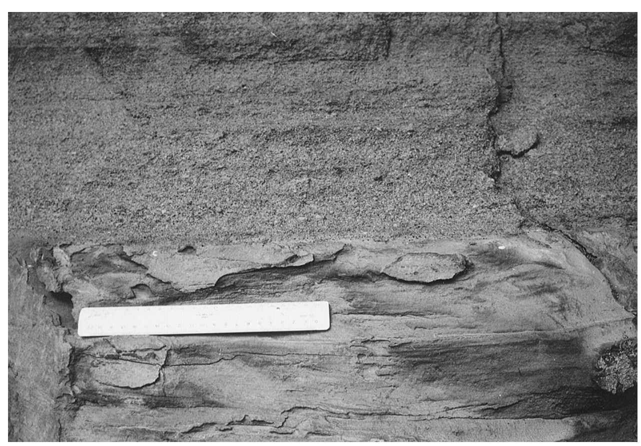


Figure 15.—Fallout ash at base of member B, Lava Creek Tuff, showing sharp contact between finer grained part below and coarser grained part above. Grand Canyon of the Yellowstone, 1.5 km southeast of Tower Falls, Yellowstone National Park. Scale is 20 cm.

Bedded ash that matches the general chemical and mineralogical characteristics of the Yellowstone Group occurs widely in middle Pleistocene deposits of the Western United States. Many of the localities were early correlated with the Pearlette ash of southwestern Kansas (Swineford and Frye, 1946; Powers and others, 1958). Izett and others (1970), Naeser and others (1973), Izett (1981), and Izett and Wilcox (1982) showed that the Pearlette-like ash beds clearly are related to the eruptions of the Yellowstone Group, and that three distinct chemical and mineralogical types can be distinguished among them. In particular, they showed that one of these ash beds corresponds in specific characteristics, including its age, to the Lava Creek Tuff and overlies ash beds correlated with the 760-ka Bishop Tuff, a major ash-flow sheet of eastern California. Sanidine from the Lava Creek Tuff has been dated as 640 ka (table 2). Izett and his coworkers showed that the corresponding distal ash bed is correlative with the Lava Creek Tuff and noted that at a number of localities on the Great Plains of Kansas, Nebraska, and South Dakota the Lava Creek ash bed occurs just above Kansan glacial deposits. Similar ash beds occur in pediment deposits of the Rocky Mountain region that lie stratigraphically just above glacial deposits generally correlated with the Kansan (Scott, 1965; Richmond, 1965). Thus, the early Yarmouth interglaciation of the west-central United states is dated as somewhat older than 640 ka by widespread ash beds derived from the great pyroclastic eruptions of the Lava Creek Tuff. Much of this fallout material may correlate specifically with member B and the bedded ash at its base in the Yellowstone Plateau region. This interpretation seems borne out by the discussion of Izett (1981), who showed that locally there are two distal ash beds related to the Lava Creek Tuff. One is a common and widespread pyroxene-bearing ash; it locally overlies a less commonly preserved ash that contains abundant hornblende and allanite and typically has higher contents of U, Ta, and Rb. Each of these characteristics (compare with chemical data of table 14) relates the lower distal ash bed to member A and the upper, more widespread and voluminous bed to member B or to the bedded ash that occurs at the base of member B in the Yellowstone Plateau area.

MEMBER B

Ash-flow tuff of member B of the Lava Creek Tuff is exposed widely in and around the Yellowstone Plateau and southwestward to the margins of the Snake River Plain. In much of the northwestern sector of the plateau it overlies member A, but elsewhere (except for its basal fallout) it generally overlies a variety of older rocks (pl. 1), most commonly the Absaroka Volcanic Supergroup of Eocene age. As in the ash-flow tuff of member A, there is a lower part containing fewer and smaller phenocrysts than the bulk of the member above, but compared to member A, the relatively phenocryst-poor zone of member B is less thick (a few meters) and grades upward more gradually into the overlying relatively phenocryst-rich densely welded tuff. The lower part of member B typically has a crudely columnar set of smooth, vertical, somewhat conchoidal joints, as well as closely spaced joints parallel to the welding foliation (fig. 16). Outcrops of member B commonly are less crumbly than

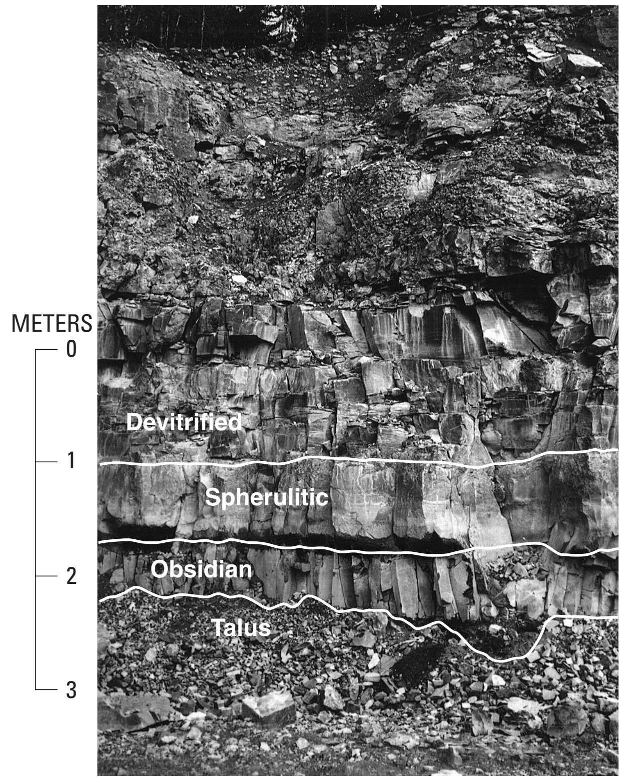


Figure 16.—Outcrop of lower part of member B, Lava Creek Tuff, showing typical joint pattern. Basal columnar zone is obsidian about 0.6 m thick; massive zone above, 1 m thick, is spherulitic; zones of irregular and platy joints above are in devitrified zone, locally lithophysal. Quarry near east end of dam of Grassy Lake Reservoir, just south of Yellowstone National Park.

Figure 17.—Bold outcrop of upper part of member B, Lava Creek Tuff, showing well defined columnar jointing. Upper end of Lewis Canyon, 4.5 km south of Lewis Falls, Yellowstone National Park.

those of member A. At many places member B has rather well-defined columnar jointing, even well above the basal contact (fig. 17). At some localities the outcrops have a bluish or purplish-gray hue in distinction to the common brown weathering of member A.

PETROGRAPHY

All parts of the Lava Creek Tuff, like most rhyolites of the Yellowstone Plateau volcanic field, have phenocrysts of quartz, sanidine (about Or_{50-55}), and subordinate sodic plagioclase (about An_{20}). Minor phenocrysts include magnetite, ilmenite, ferroaugite, fayalite, iron-rich hornblende, zircon, chevkinite, and allanite. Although hornblende is a rare constituent of member B, it is the predominant mafic constituent in glassy portions of both the lower and upper parts of member A—the only rhyolite of the volcanic field of which that is true. Clinopyroxene is comparatively rare in these hornblende-rich rocks of member A, but allanite and chevkinite are especially abundant and easily seen in thin section whereas they generally are noted only in heavy-mineral concentrates of other rhyolites of the field. The relative abundance of hornblende phenocrysts in member A is of only limited use in field mapping or correlation of isolated outcrops because the hornblende almost invariably is altered to formless opaque oxides in devitrified rocks.

DISTRIBUTION AND SOURCE AREAS

The original extent of the Lava Creek ash-flow sheet is indicated in figure 20. Figures 18 and 19 show distributions and thicknesses of members A and B, respectively, of the Lava Creek Tuff. These maps show the major topographic features surrounding the rhyolite plateau that controlled the distribution of the ash-flow sheet and also show that member A was emplaced on a topography of greater relief and has a more limited distribution than member B. Volumes of the two members have been estimated from planimeter measurements on figures 18 and 19. Thicknesses are poorly controlled in member A because its base is exposed only in the Madison Canyon area; furthermore, the member is presumed to be present within the older topographic basin of the first-cycle caldera where it is downfaulted by the younger Yellowstone caldera and buried by rocks that fill it. Member A originally covered 3,240 km^2; assuming a reasonable-looking pre-Lava Creek topography for the Madison Valley near West Yellowstone and an average thickness buried in the Yellowstone caldera somewhat greater than the exposed thickness on the north caldera wall, the estimated volume is about 510 km^3. Both the thickness and the distribution of member B are better controlled than those of member A. Some control within the Yellowstone caldera is provided by exposures near Stonetop Mountain and by USGS research drill hole Y-5 near Rabbit Creek at the edge of Midway Geyser Basin (pl. 1), in which a thickness of 33 m is interpreted for member B. Member B covered 7,300 km^2 and had a volume, calculated similarly to member A, of 490 km^3.

Figures 18, 19, and 20 show that the area of the Yellowstone caldera is the source of the Lava Creek ash flows. The caldera is within the center of distribution of the Lava Creek ash-flow sheet, which wraps around the proximal ends of surrounding ranges and thins outward as it drops in elevation along valleys that lead away from the caldera. Lithologic features of traceable stratigraphic intervals within the sheet provide more direct evidence for the caldera area as the source of the Lava Creek ash flows. Phenocrysts are markedly larger and more abundant near the caldera than farther away, as are lithic inclusions that are distributed widely in specific layers. Some of the lithic inclusions are of rock types found in various places around the rhyolite plateau, but others, particularly rhyolitic welded tuff inclusions of the lower part of member A in the Purple Mountain area, must come from areas that are now buried, probably within the caldera. The tuffs are more densely welded and the basal glass zones thinner near the caldera than in areas closer to the margins of the plateau. These radial variations can be followed discontinuously along roads that lead outward from the caldera from Norris Junction along Obsidian Creek to the Gardner River, and from Lewis Falls along Lewis Canyon and the Snake River to Flagg Ranch (pl. 1).

Besides showing that the source of the Lava Creek Tuff was in the area of the Yellowstone caldera, figures 18 and 19 also show that the source areas of the two members were slightly different. The area covered by member A is west of the center of distribution of member B, and member A forms the rim of the Yellowstone caldera in the Mount Haynes-Purple Mountain area. Radial variations in size and abundance of phenocrysts and xenoliths and in degree of welding show that the source of member A was in the area southeast of Purple Mountain. The lower part of member A is known to occur only near there, and the deep mantling of Mount Jackson Rhyolite flows in that area at elevations above the average surface of the ash-flow sheet nearby can be explained most readily if the Purple Mountain area was near the eruptive vent. Hay (1959), Smith (1960a, p. 804-806), and Sparks and Wilson (1976; Sparks and others, 1978) have emphasized that ash flows are generated by outward motion from the base of a strong vertical eruption column. The features of ash-flow tuffs found at some distance from an eruptive vent record almost entirely the mechanism of emplacement by flowage, but near-vent features may be accounted for by proximity to the vertical eruption column. Thus a topographic obstacle that would be a barrier to the ash flows under some circumstances could be surmounted adjacent to the vent by the eruptive column, and ash flows spreading from the base of the column could mantle the flanks of the obstacle. Furthermore, the pronounced layering of welded tuffs in the lower part of member A near Purple Mountain (fig. 13) is typical of little-traveled ash flows emplaced near the eruption column. This line of reasoning suggests that fissures through which member A of the Lava Creek Tuff erupted lay just to the south of the Purple Mountain area. How-

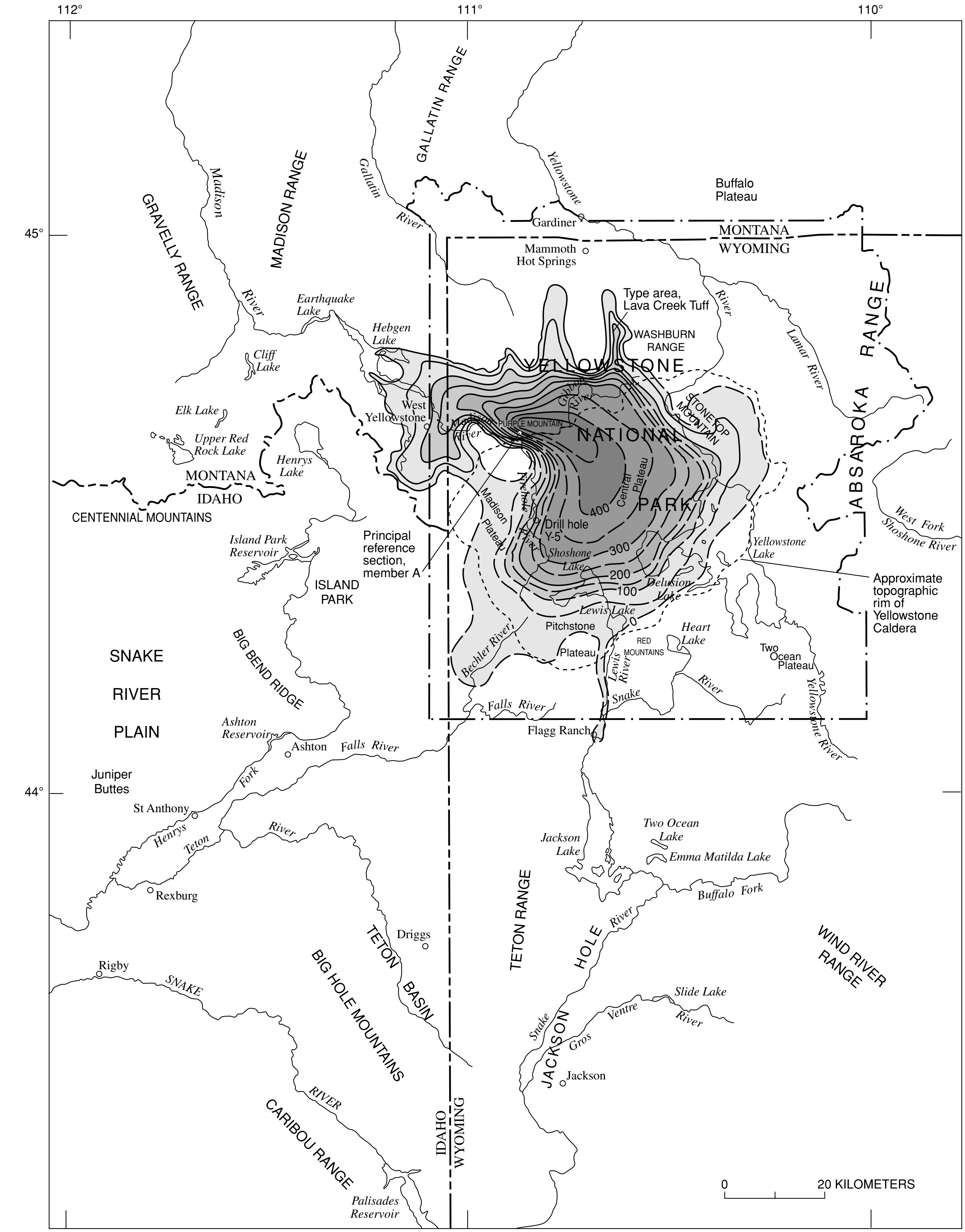


Figure 18.—Areal distribution and thickness of member A, Lava Creek Tuff, in relation to topographic features in the Yellowstone Plateau volcanic field that affected its distribution. Member A was emplaced on erosional topography of high relief from a source area southeast of Purple Mountain. The calculated initial area is 3,240 km^2 and initial volume is 510 km^3. Shading is darker for thicker sections. Isopachs in meters, dashed where inferred.

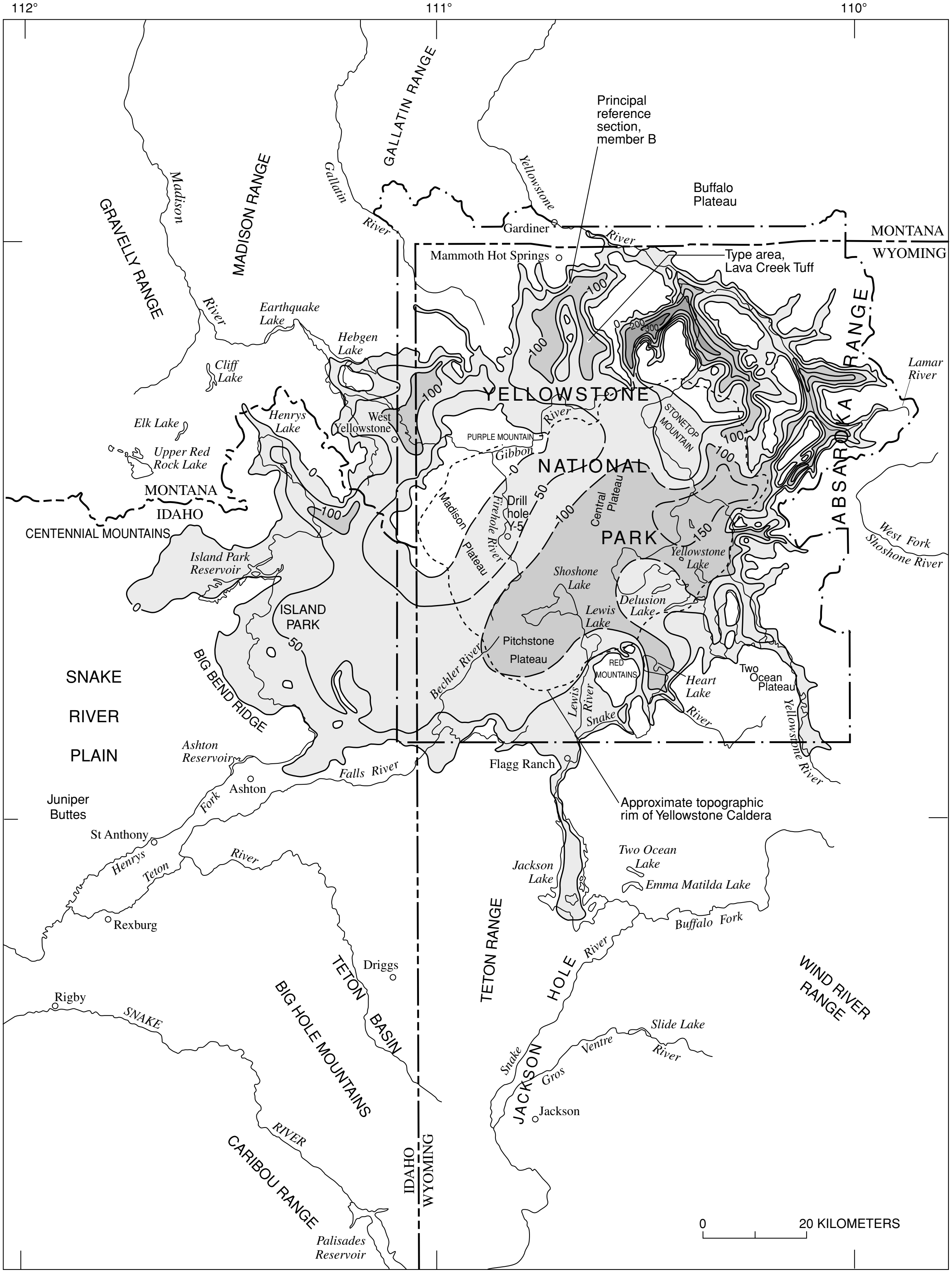


Figure 19.—Areal distribution and thickness of member B, Lava Creek Tuff, in relation to topographic features in the Yellowstone Plateau volcanic field that affected its distribution. Member B was distributed farther than member A on an erosional surface that had been partly filled and evened out by member A. The source of member B was in the eastern part of the Yellowstone caldera. The calculated initial area is 7,300 km^2 and initial volume is 490 km^3. Shading is darker for thicker sections. Isopachs in meters, dashed where inferred.

ever, they must have lain farther within the preexisting south wall of the first-cycle caldera and a probable mass of rhyolite within it near the Madison Plateau, both of which formed barriers to the spread of the ash flows. Member A is not known to be present east of the west flank of the Washburn Range or south of the Central Plateau (pl. 1; fig. 18).

Member B mantles a highly irregular eroded surface of Absaroka volcanic rocks and older strata at the rim of the Yellowstone caldera from west of the Solfatara plateau, eastward along the south flank of the Washburn Range, along Broad and Astringent Creeks, along the east and south sides of the main basin of Yellowstone Lake and westward past Flat Mountain (pl. 1). The area of the Yellowstone caldera within that partial rim lies at the center of distribution of member B, which extends radially outward along paleovalleys and more extensive plateau segments. This distribution and the types of radial variation in phenocryst and lithic-inclusion size and abundance noted earlier indicate that the eastern part of the Yellowstone caldera contains the source area for member B. Also consistent with the proposed eastern source for member B is its general exclusion from the area around Purple Mountain, where high-standing pre-Lava Creek topography prevented its emplacement, despite the presence of a thick mantle of member A in that area, which was close to the eruptive source of A.

ERUPTIVE MECHANISM

Eruption of the Lava Creek Tuff represents stage II of the resurgent cauldron cycle of Smith and Bailey (1968a), the caldera-forming pyroclastic eruptions. It is not known with certainty what triggers voluminous ash-flow eruptions of the type that formed the Lava Creek Tuff, but the succession of stages in the resurgent cauldron cycle suggests one possibility. A large body of silicic magma must be rising intermittently toward the surface during the insurgence represented by stage I of the cycle. The ring-fracture zone may be generated as the magma body reaches a high crustal level; the magma-chamber roof is stretched by continued rise of the magma and occasionally sags with tectonic extension or preclimactic eruptions. The ring fractures presumably form initially in the most highly distended zone, close to the surface; as the roof is further stretched, they probably propagate downwards. Early stages of volcanism, such as the Mount Jackson Rhyolite, may be presumed to record the rise of pods of rhyolitic magma closer to the surface and their escape through the ring-fracture zone. However, the most deeply penetrating ring fractures, perhaps aided by extensional tectonic fracturing, might eventually intersect the magma in the main chamber. At that time there would be an immediate relief of confining pressure, allowing the rapid exsolution of gas from the magma, vesiculation probably at a rate greater than could be accommodated physically in the magma chamber and ring-fracture zone, and explosive escape to start the ash-flow eruption cycle. Once begun, continued pressure relief on the magma by eruption of the upper part of the magma column would tend to sustain the eruption until the supply of exsolving gas that could reach the vents was exhausted or until parts of the magma were reached that were too viscous to flow readily or to vesiculate (see Smith, 1979, figs. 11, 12).

THE YELLOWSTONE CALDERA

The central region of the Yellowstone Plateau is a lava-filled caldera 45 by 85 km across (pl. 1 and frontispiece) that formed as a direct result of the eruption of the Lava Creek Tuff. Although Hayden's casual statement of 1872 was quoted in an earlier section indicating that he thought of this basin as a gigantic volcanic crater, it was not until the work of Boyd (1961, p. 412) and Hamilton (1960b) that the concept of a volcanotectonic depression or caldera in this region was made explicit. Even so, neither Boyd nor Hamilton completely defined or delineated the Yellowstone caldera. The case for the existence of the caldera, however, is compelling. The Lava Creek Tuff surrounds the central region of the rhyolite plateau with scarps that originally stood more than 500 m high (pl. 1; figs. 20, 21). The correctness of the concept of emplacement of great welded-tuff sheets, such as the Lava Creek Tuff, by the ash-flow mechanism has been demonstrated in such studies as Smith's (1960a) and Boyd's (1961) and need not be reviewed here. It is known that such sheets were deposited by pyroclastic flows moving outward at high velocity from near the eruptive fissures. Foliation attitudes in the welded tuffs, which are close to horizontal at most of the scarps surrounding the Yellowstone caldera, show that the ash-flow sheets do not mantle older faults there; they once must have been continuous across those scarps and later must have been truncated by the faults. The former continuity of the ash-flow sheet across the caldera is proven by its presence on two subsequently uplifted blocks within the caldera, in the Stonetop Mountain-Sour Creek area on the east and in the area near the Firehole River penetrated by USGS research drill hole Y-5 (pl. 1, fig. 19).

Although the Yellowstone caldera is a single topographic basin, it comprises two distinct structural entities as shown by its two resurgent domes. Smith and Bailey (1968a), in defining the concept of resurgence, showed convincingly that the structural pattern of resurgent calderas is a general one related to fundamental cauldron mechanisms. The resurgent dome that is formed by postcollapse uplift within a caldera of this type is a cauldron block that foundered as a nearly intact piston bounded by a steeply inclined ring-fracture zone. The fundamental structure of a resurgent caldera is its ring-fracture zone, which probably comprises a narrow zone of main ring faults at the margin of the cauldron block and a concentric set of subsidiary normal faults outside the main ring faults. According to the concept of Smith and Bailey, the ring-fracture zone forms during magmatic insurgence, provides the vent fissures for the climactic ash-flow eruptions, and becomes the accommodating structure for cauldron subsidence. The presence of two ring-fracture zones in the Yellowstone caldera is implied by its two resurgent domes (pl. 1), the Mallard Lake and Sour Creek domes (although the Mallard Lake dome has a more complex history than the Sour Creek, as discussed later). The two domes lie near the centers of two approximately circular segments of the caldera (Christiansen, 1979). These two segments, referred to here by the names of their resurgent cauldron blocks as the Mallard Lake and Sour Creek caldera segments, combine to give the compound caldera a markedly elliptical shape. The two segments are centered in the respective source areas of the

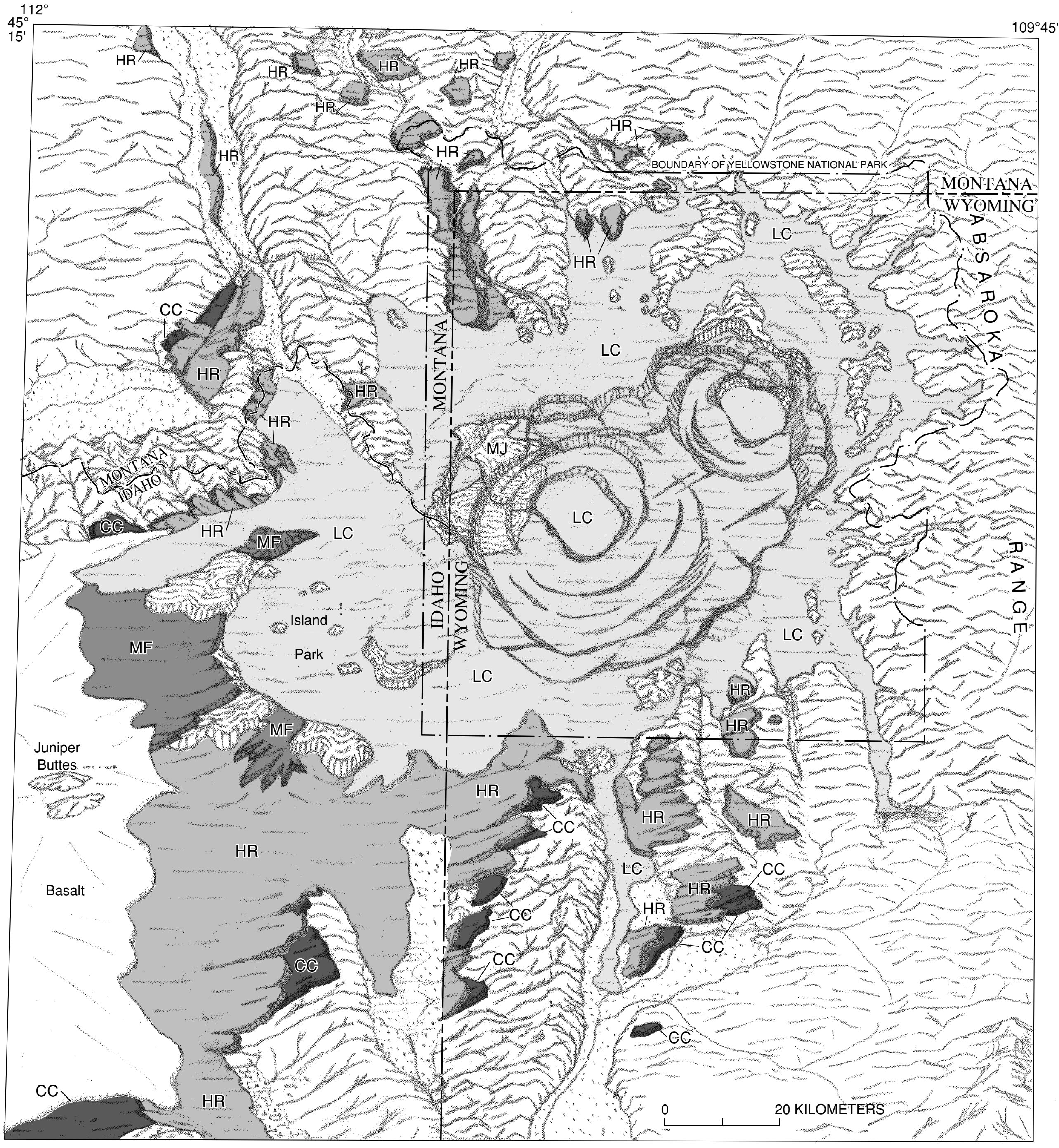


Figure 20.—Interpretive reconstruction of the Lava Creek Tuff ash-flow sheet and the Yellowstone caldera before resurgence. CC, Kilgore Tuff and Conant Creek Tuff; HR, Huckleberry Ridge Tuff; MF, Mesa Falls Tuff; MJ, lava flows of Mount Jackson Rhyolite; LC, Lava Creek Tuff.

two members of the Lava Creek Tuff (figs. 18 and 19) and are presumed to have been formed by subsidence of two separate segments of the roof over the highest parts of the very large Yellowstone magma chamber.

That the Yellowstone caldera formed by roof collapse of the Lava Creek magma chamber as a direct result of the eruption of the ash flows is clear from temporal and spatial relations of the caldera to the ash-flow eruptions. The presence of two ring-fracture zones in the Yellowstone caldera and the serial eruption of two separate subsheets of ash-flow tuff having slight compositional differences—each with an initially erupted portion of relatively phenocryst-poor rhyolitic pumice and ash—indicates that magma intruded to very high crustal levels in two separate but adjoining areas during the insurgence. Essentially simultaneous rise of the two magmatic culminations is indicated by contemporaneous flows of Mount Jackson Rhyolite around the two ring-fracture zones. This contemporaneity of intrusion, the strong petrologic similarity of the two ash-flow subsheets, and the immediate succession of eruption of the two subsheets without intervening time for erosion or even for surface cooling of the earlier one show that the two high-level magma bodies must connect at depth; they probably are two culminations on a single continuous larger magma chamber. A hypothetical alternative—that the member A magma overlay member B magma in a single chamber underlying both ring-fracture zones—is less consistent with both the trace-element data discussed later in this report (table 14) and the oxygen-isotopic data of Hildreth and others (1984). The ranges of trace-element concentrations and of $\delta^{18}O$ are substantially the same in members A and B, confirming that there was not one single highly evolved and ^{18}O-depleted zone at the top of a magma chamber that could be entirely drawn down by the member A eruption before member B began to erupt.

The two high-level culminations of the Lava Creek magma chamber which formed the two caldera segments probably were controlled by preexisting structures of the rhyolite plateau. Explosive eruption of member A, and perhaps subsidence of the Mallard Lake cauldron block, may have provided the trigger to begin the eruption of member B from ring fractures around the Sour Creek block. Later stages of the member B eruption might have been through both ring-fracture zones.

The Yellowstone caldera has the lobate outer walls typical of the slumped margins of almost all calderas. Smith and Bailey (1968a) carefully developed the line of reasoning showing that resurgent calderas are characterized by an inner cauldron block that remains essentially unbroken during collapse and is bounded by a ring-fracture zone. The mechanism of collapse, therefore, must be catastrophic subsidence of the cauldron block along the main ring-fracture zone, accompanied and followed by inward slumping of the oversteepened caldera walls. These lobate slumps and the ring-fracture zone itself generally are buried by caldera-filling avalanche debris, sediments, and lavas later during the cauldron cycle. The steeply dipping principal ring fractures must now lie within the Yellowstone caldera relatively close to the margins of the resurgent domes and must be surrounded by more gently dipping discontinuous concentric fractures that bound lobate blocks of the slumped walls. The presumed

Figure 21.—Aerial view to northwest across part of the Central Plateau, Yellowstone National Park, to the Yellowstone caldera wall at Purple Mountain; the caldera is breached by the Madison Canyon near center of picture. Madison Range in background. Photograph by D. E. White.

positions of the buried ring-fracture zones of the compound Yellowstone caldera are further outlined by the localization of hydrothermal activity (fig. 29) and of the vents for early post-resurgence rhyolite flows (fig. 24). One of the large slumped blocks remains unburied on the north wall of the Mallard Lake caldera segment, at Secret Valley in the vicinity of Gibbon Canyon (pl. 1, fig. 22), where the block lies 300 m below the outer caldera rim. The probable configuration of the Yellowstone caldera just after major failure of the caldera walls is represented in figure 20.

Other faults outside the east wall of the caldera seem to have behaved in effect as outer ring fractures, further lowering the surrounding terrain in progressive steps toward the caldera rim. Love (1961) designated these faults the Mirror Plateau fault system. The largest displacements in the Mirror Plateau fault zone are downward toward the caldera, but more numerous smaller faults of opposite displacement produce discontinuous grabens. The curved traces of these faults, concentric to the Yellowstone caldera (pl. 1), suggest the probable influence of the caldera and its related magma movements on their development. They are not, however, simply parts of a ring-fracture zone that was active only during collapse. Mapping shows that these faults have had continued episodic movement both before and after emplacement of the Lava Creek Tuff across them; the greatest displacements predate the Lava Creek. For example, just east of Mirror Lake a section of Lava Creek Tuff about 60 m thick is exposed in the scarp of one of the Mirror Plateau faults, but the Lava Creek Tuff pinches out on rocks of the Absaroka Volcanic Supergroup on the upthrown block a short distance from the fault. This relation implies an eroded pre-Lava Creek scarp at this locality more than 60 m high. Further displacement has occurred since emplacement of the Lava Creek Tuff; Waldrop (1969) showed that these faults have had renewed movement into late Quaternary time. It is probable that the arcuate faults of the Mirror Plateau system reflect magmatic insurgence, caldera collapse, and postcollapse upward magma movements in an area that did not have a strongly developed preexisting system of tectonic faults (see Eaton and others, 1975).

In complexes of this sort it commonly is difficult to find direct evidence for the timing of caldera collapse in relation to the ash-flow eruptions. Smith and Bailey (1968a) described caldera collapse as stage III of the resurgent caldron cycle, implying that major collapse may often follow the ash-flow eruptions of stage II. They noted, however, that in many calderas related to large volumes of ash-flow tuff, subsidence may accompany eruption. Syneruptive collapse is evident from the thick tuff fills of many deeply eroded older calderas (Lipman, 1984). Threefold evidence from the Yellowstone caldera suggests, however, that at least part of the major collapse there followed the caldera-forming eruptions. First, although the thickness of member B of the Lava Creek Tuff on the Sour Creek resurgent dome is highly irregular, it does not invariably exceed the thickness of member B at the nearby caldera rim. The underlying member A and the older Huckleberry Ridge Tuff both are exposed beneath member B on the northwest end of the dome. Second, where the caldera wall is best exposed, in the Purple Mountain-Gibbon Canyon area, a large slumped block of the

Figure 22.—Slumped block of the Yellowstone caldera wall near Gibbon Canyon, Yellowstone National Park; Secret Valley in middle distance. Aerial view to west. Photograph by L. J. P. Muffler.

caldera wall is preserved, and structural features of the welded tuffs show that at least this slumping occurred after emplacement of member A but while it was still welding and crystallizing. The thickness of the upper part of member A on the slumped block is substantially the same as in the outer wall. In places near the fault at the head of the slumped block, the welded tuff is a pull-apart breccia in which individual fragments are only slightly rotated and spaces between are left unfilled, but the mass as a whole dips toward the caldera. In other places on this slumped block, the compressed pumice inclusions are strongly lineated, plunging directly or obliquely toward the caldera. The most conspicuously deformed welded tuffs near this scarp have folded welding foliation with closely spaced fractures parallel to the axial planes and small gaping tensional joints that form pockets oblique to the foliation (fig. 23); the pockets are lined with vapor-phase minerals perpendicular to the cavity walls. Third, inconclusive evidence from USGS drill hole Y-5 (pl. 1; fig. 19) suggests that at least the last major collapse of the Mallard Lake caldera segment, the source area for member A of the Lava Creek, may not have occurred until after the eruption of member B. Examination of core from that drill hole shows that the originally glassy nonwelded top of the ash-flow sheet is preserved at that site. A zone of relatively few phenocrysts, the base of which is only slightly welded, occurs at a depth of about 33 m; it is the only zone having those characteristics in a cored thickness of 155 m of ash-flow tuff in the hole. This zone resembles the lower part of member B and may represent the contact between members A and B, as implied in figure 19. If this zone truly is the base of member B, the thickness of the member on the Mallard Lake cauldron block would be inconsistent with its floor lying at a low level before emplacement of member B. It is possible that collapse of the Mallard Lake block occurred during the eruption of member A and that the resulting depression was largely filled by late-erupted portions of that member; final collapse of the Mallard Lake and Sour Creek cauldron blocks then might have occurred together at the end of the eruption of member B, resulting in its relative thinness in the area of the drill hole.

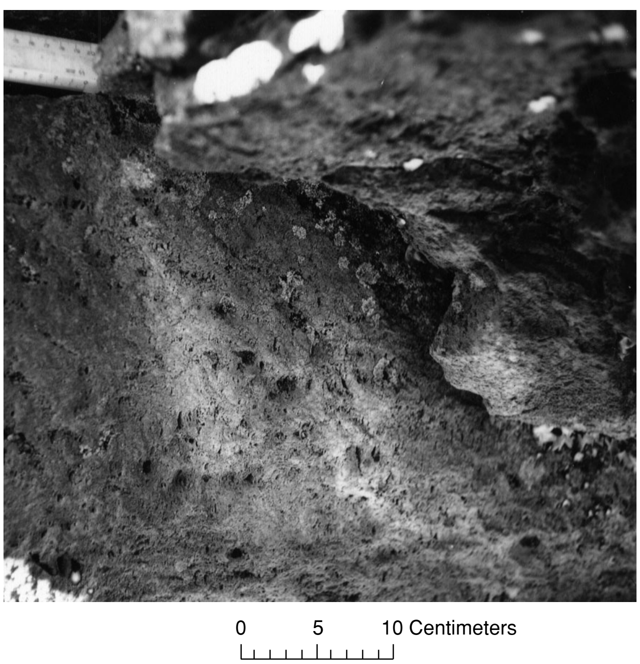


Figure 23.—Exposure of member A, Lava Creek Tuff, on a small vertical joint face at the Yellowstone caldera wall in Gibbon Canyon, Yellowstone National Park. The photograph shows small gaping tension fractures as lenticular vertical dark streaks. Individual open tension fractures, lined with vapor-phase minerals, are elongated perpendicular to horizontal compaction folliation, which is suggested by closely spaced dark open fractures clustered in horizontal bands; lineation of stretched pumices is parallel to plane of exposure.

Some possibly contradictory evidence occurs in the Stonetop Mountain area, on the southeast end of the Sour Creek dome. There, the upper part of member B is more than 150 m thick, commonly contains abundant angular welded-tuff inclusions as much as several meters long, and is only partly welded. The inclusions generally are dark gray, spherulitic, and partly glassy. The source of the abundant xenoliths, commonly as large as 15-20 cm and locally several meters long, is not certain, but the xenoliths look much like the Lava Creek Tuff itself. Possibly this thick section represents partial collapse of the Sour Creek caldera segment during the eruption of member B, with blocks of already welded but uncrystallized caldera wallrock falling back into the vents to be reejected with later-erupted ash flows. Farther northwest along the Sour Creek dome, however, member B is not exceptionally thick and contains few of the dark-gray spherulitic inclusions. If the rocks in the Stonetop Mountain area indicate initial collapse during the member B ash-flow eruptions, the early collapse appears to have been localized in only the southeastern part of the cauldron block.

EARLY POSTCOLLAPSE HISTORY AND RESURGENT DOMING

Collapse of the Yellowstone caldera was a catastrophic geologic event. Within a very short time after collapse, somewhat slower processes related to magmatic resurgence began—doming of the cauldron blocks and the beginning of caldera infilling by rhyolitic lavas. Sediments undoubtedly accumulated in the Yellowstone caldera during a brief preresurgence period, but few exposures are deep enough in the section to expose them. Any sediments that may have accumulated at this stage on the Sour Creek cauldron block have been eroded from the subsequent dome or are masked by an extensive mantle of glacial sediments. Any possible sediments on the Mallard Lake cauldron block would be buried by the rhyolite flow that covers it (pl. 1). Despite the lack of known sediments of this age, some of the early postcollapse rhyolite flows (discussed in the following section) display evidence of having been emplaced in a caldera lake. Any early sediments, or even early rhyolite flows, that may exist within the caldera but are not known would correspond to Smith and Bailey's stage IV of the resurgent cauldron cycle, preresurgence volcanism and sedimentation. Smith and Bailey (1968a, p. 639) noted that stage IV is short and that its record is incomplete in many calderas; volcanism at this stage was recorded in some resurgent calderas but not in others.

Just as the Yellowstone caldera had a more complex collapse history than many others of its type, the history of postcollapse resurgent doming and related volcanism is quite complex. Each of the two initially subsided cauldron blocks is now a domical uplift. The Sour Creek dome, the eastern structure, became resurgent soon after collapse, as is typical of resurgent calderas. By structural analogy, and with incomplete K-Ar dating, it was originally presumed that uplift of the Mallard Lake dome was similarly early (Christiansen and Blank, 1972, p. B12). Subsequent K-Ar dating, however, has shown that the Mallard Lake flow, which is uplifted and faulted on the dome, is much younger than caldera collapse. Reexamination of the field relations has enabled a revision of the age and history of the Mallard Lake dome. The present dome, on which only the Mallard Lake flow is exposed, must be about 160,000 years old and is discussed in a following section. Nevertheless, some evidence suggests that there may also have been early resurgent doming of the Mallard Lake block, and that evidence is discussed in this section after the Sour Creek dome is considered.

Early uplift and doming of the initially subsided cauldron blocks of the Yellowstone caldera, stage V of the resurgent cauldron cycle, occurred soon after collapse. The intervening time could not be resolved by conventional K-Ar dating. The Canyon flow, which onlaps the Sour Creek dome, has, however, been dated by $^{40}Ar/^{39}Ar$ methods as 484±15 ka (Gansecki and others, 1996), thus relaxing the time constraint on early resurgent doming.

In contrast to the rather steep resurgent domes of some calderas, such as the Valles (Smith and others, 1970), Toba (van Bemmelen, 1939), Timber Mountain (Carr, 1964; Christiansen and others, 1977), and Creede (Steven and Ratté, 1965), the exposed part of the Sour Creek dome is gentle, with maximum dips on the flanks of about 15°. The elliptical dome is, nevertheless, about 13 by 2l km across (pl. 1; fig. 5), and its summit is higher than the Lava Creek Tuff at the adjacent caldera rim. The axis of the dome trends northwest toward a zone of faults outside the caldera with very fresh scarps, the Solfatara fault system of Love (1961). Pierce (1973) showed at least some movement on these faults postdating the Pinedale glaciation—the last Pleistocene glaciation of the Rocky Mountains. Thus, some control on the location and shape of the resurgent dome by long-lived tectonic structures is implied.

The Sour Creek resurgent dome (pl. 1) has elements of a fault pattern characteristic of domical uplifts (Cloos, 1939; Wisser, 1960; Smith and Bailey, 1968a). The faults form a complex graben along the northwest-trending structural axis. They have minimum displacement at the ends of the axial graben system, and partial grabens branch outward from the center. This structural pattern, as noted explicitly by Smith and Bailey (1968a) and by Carr and Quinlivan (1968), clearly shows that the faults on the resurgent dome are related to uplift and doming of the cauldron block, not to its collapse. The block, therefore, probably subsided as an essentially intact steep-sided cylinder.

The fault pattern on the southeast end of the Sour Creek dome may have been modified somewhat by the effects of the buried Stonetop Mountain flow (fig. 11) on development of the surficial faults. The fault pattern, furthermore, is complicated by the intersection of younger, northeast-trending faults on the southwest flank, the Elephant Back fault zone, and some very young faults along the margin of the dome along the Yellowstone River near LeHardy Rapids (pl. 1). The Elephant Back fault zone extends between the axial grabens of the Sour Creek and Mallard Lake domes (pl. 1; fig. 5). The trend of the Elephant Back zone is parallel to the axis of the Snake River Plain and to the general alignment of major late Cenozoic calderas from the eastern Snake River Plain through the two segments of the Yellowstone caldera. This parallelism expresses a fundamental tectonic trend, now reflected weakly in the normal faults of the Elephant Back zone. The intersection of this trend with major northwest-trending tectonic zones may have controlled the locations of the two third-cycle caldera segments.

Although, as noted above, the present Mallard Lake dome is younger, a resurgent dome contemporaneous with the Sour Creek dome probably formed in the Mallard Lake cauldron block as well. The only evidence of early doming of the Mallard Lake block comes from USGS research drill hole Y-5 (pl. 1), situated just beyond the margin of the Mallard Lake flow at the northwest edge of the block. The drilling penetrated only about 10 m of probably uppermost Pleistocene sediments before it entered Lava Creek Tuff; the rest of the hole to the bottom at 165 m is in the Lava Creek. Thus, the Mallard Lake flow probably overlies the Lava Creek directly a few hundred meters from the hole, although at least two and perhaps several post-Lava Creek and pre-Mallard Lake rhyolite flows occur beyond the flanks of the dome. This relation suggests that the Lava Creek Tuff had already been uplifted within the center of the caldera before emplacement of the older flows, the oldest of which—the Biscuit Basin flow—is dated as 516±7 ka (Gansecki and others, 1996).

OLDER POSTRESURGENCE RHYOLITIC VOLCANISM

The postcollapse rhyolites of the third volcanic cycle were named the Plateau Rhyolite (Christiansen and Blank, 1972), essentially following Boyd's term "Plateau flows" (Boyd, 1961, p. 403-409). Christiansen and Blank further divided this group of rhyolites into members (fig. 9; table 3), the individual flows of which have a similar mineralogies, geo-

Table 3.—Members of the Plateau Rhyolite in relation to caldera events.

Plateau Rhyolite age group	**Caldera event**	**Intracaldera flows and minor related pyroclastics**	**Extracaldera flows and minor related pyroclastics**	
Younger postcollapse rhyolites		Central Plateau Member	Obsidian Creek Member	Roaring Mountain Member
	Late doming			
		Mallard Lake Member		
Older postresurgence rhyolites		Upper Basin Member		
	Early resurgent doming			

graphic distributions, and age relations to various caldera events. Volcanic eruptions within the Yellowstone caldera spanned at least the time from immediately after resurgent doming, nearly 640,000 years ago, to about 70,000 years ago and may not yet have finished.

UPPER BASIN MEMBER

Flows of the Upper Basin Member of the Plateau Rhyolite occur low in the intracaldera section and have distributions and lithologies different from those of younger intracaldera rhyolites. In particular, they are plagioclase-rich and are exposed in two areas, each adjacent to one of the resurgent domes of the Yellowstone caldera.

In the area of the Upper and Lower Geyser Basins, at least two flows of the Upper Basin Member occur west and southwest of the Mallard Lake dome (fig. 24; pl. 1). The older of these, the Biscuit Basin flow, has been dated at about 516±7 ka (Gansecki and others, 1996) and, as just noted, probably postdates initial uplift of the Mallard Lake dome. Available data are not sufficient to prove that all the exposures mapped as Biscuit Basin flow are in fact parts of a single flow, but their lithologies are similar and all occur within a small area. The Biscuit Basin flow or flows probably were emplaced largely or entirely within a body of water—presumably a caldera lake. Such flows predate the infilling of the caldera by numerous other rhyolite flows and are almost entirely glassy; the glass is a thoroughly hydrated and fractured perlite with only sparse 1- to 2-mm spherical remnants of obsidian. This lithology is not typical of most flows of the rhyolite plateau. The Biscuit Basin flow is overlain by another plagioclase-rich rhyolite, the Scaup Lake flow, that is about 198,000 years old (table 2).

A somewhat more complex sequence occupies a corresponding position on the north and northeast flanks of the Sour Creek dome (pl. 1; fig. 24). In the Grand Canyon of the Yellowstone between Upper and Lower Falls, the base of the exposed postresurgence section is a nonwelded lithic-vitric ash-flow tuff at least 35 m thick, the tuff of Uncle Tom's Trail (Christiansen and Blank, 1975b; Richmond, 1976; 1977). The base is not exposed, but just above Lower Falls it may lie just below the exposure, where gravels are incorporated bodily (not just as individual cobbles) into the basal few meters. These gravels contain andesitic rocks of Absaroka affinity, well-rounded quartzites such as are exposed locally in a conglomerate in the basal part of the Absaroka sequence on the south side of the Washburn Range (Smedes and Prostka, 1972, p. C16), and minor glassy rhyolite of Upper Basin Member petrography. The nonwelded tuff itself contains large angular blocks of intrusive andesite or microdiorite that are identical to rocks exposed near Washburn Hot Springs, about 6½ km to the northeast (Prostka and others, 1975). The blocks are largest and most conspicuous low in the tuff but occur sporadically throughout it. The ash-flow tuff of Uncle Tom's Trail thus may represent material erupted

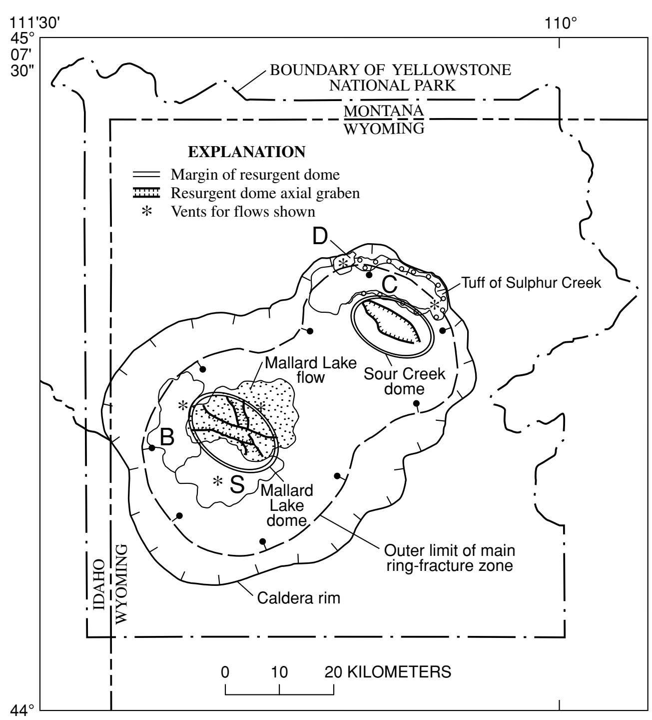


Figure 24.—Reconstructed distribution of flows and vents of the Upper Basin and Mallard Lake Members of the Plateau Rhyolite, Yellowstone Plateau volcanic field. C, Canyon flow; D, Dunraven Road flow; B, Biscuit Basin flow; S, Scaup Lake flow.

as an early part of the Upper Basin Member of the Plateau Rhyolite, into which materials from the Washburn Range were incorporated during eruption and emplacement.

The tuff of Uncle Tom's Trail is overlain by a diamicton, possibly a till (Richmond, 1976), most of whose boulders are lithologically like the blocks in the underlying tuff. Laminated or varved lacustrine silts and other sediments are associated locally with this diamicton, and the sediments are mapped together as the sediments of Lower Falls (Richmond, 1976; 1977; Christiansen and Blank, 1975b). The sources of all the boulder types in the diamicton were exposed along the Yellowstone caldera wall on the south face of the Washburn Range; no clasts of rhyolite have been found in the sediments. Thus, the sediments of Lower Falls probably predate most of the postcollapse rhyolites of this area (although some rhyolitic clasts of Upper Basin petrography do occur in the underlying tuff of Uncle Tom's trail). Small glaciers or debris flows from the caldera wall may have deposited the sediments in the caldera moat.

A complex of rhyolitic flows and tuffs overlies the sediments of Lower Falls, the tuff of Uncle Tom's Trail, and older rocks of the Yellowstone caldera wall. Although Boyd (1961, p. 405-406) discussed this complex as the Canyon flow, it includes more than one lithogenetic unit. The lowest part of the complex is best exposed in the north wall of the Grand Canyon of the Yellowstone about a kilometer north of Sevenmile Hole and east of Sulphur Creek (fig. 25). There, an undeformed sequence of bedded fallout tuffs, the tuff of Sulphur Creek, is about 75 m thick. At the base, the tuff is normal-looking friable glassy ash and pumice, but upwards it grades into densely welded tuff and has a vitrophyric zone overlain by a devitrified zone. Bedding and size-sorting of pumice and phenocrysts, however, are readily apparent throughout this thick welded section. Christiansen and Blank (1975b) interpreted the sequence as an agglutinate—that is, fallout ash that was still hot enough after eruption and deposition to weld (Wentworth and Williams, 1932; compare with Sparks and Wright, 1979). The tuff of Sulphur Creek also is well exposed in the area northeast of Moss Creek and around Broad Creek (pl. 1), but locally in the Broad Creek area it appears to have been deposited in a body of standing water and to have chilled more rapidly than elsewhere. In parts of the Broad Creek area the tuff is not agglutinated but has large vertical pipes of chaotically bedded but generally inward dipping limonite-cemented tuff that appear to rise through the entire thickness of the unit (fig. 26). These structures are interpreted here as fumarolic pipes that formed as the thick tuff cooled subaqueously. At the mouth of Moss Creek and along the walls of the Grand Canyon of the Yellowstone upstream (pl. 1), the agglutinated tuff of Sulphur Creek was plastered over the wall of the Yellowstone caldera, deforming as it accumulated and welded. Large-amplitude slump folds exposed in that area are best seen from across the canyon.

The tuff of Sulphur Creek is overlain directly by a large rhyolitic lava flow (pl. 1; fig. 24) of similar petrography, the Canyon flow as mapped by Christiansen and Blank (1975b). The vent area for the Canyon flow is indicated by the closure of concentric flow-top pressure ridges and steeply inclined foliation about 2½ km northwest of Fern Lake, south-

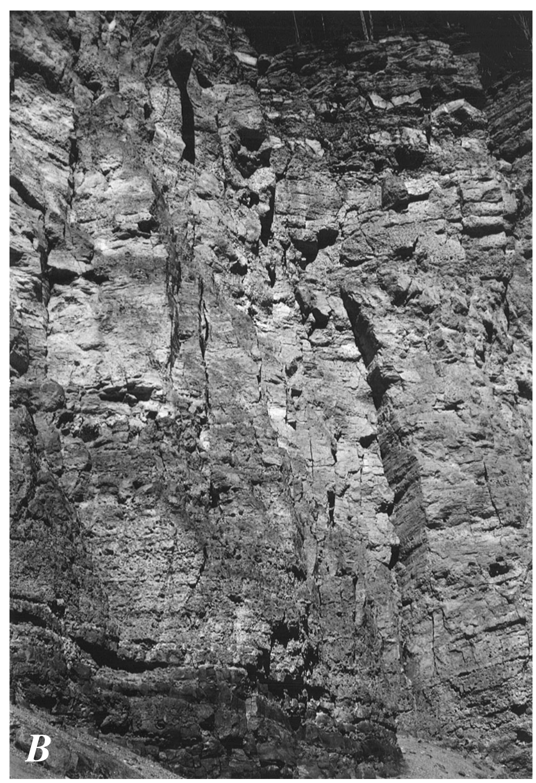


Figure 25.—Agglutinated fallout tuff of Sulphur Creek north of Sevenmile Hole, Yellowstone National Park. ***A***, view to northeast along wall of Grand Canyon of the Yellowstone, showing continuity of horizontal layering. ***B***, closer view showing good bedding and dark vitrophyric basal zone about 3 m thick overlain by lighter devitrified layers.

Figure 26.—Fumarolic pipe in the tuff of Sulphur Creek, exposed at Broad Creek just north of the Wapiti Lake Trail crossing, Yellowstone National Park. The large dark vertical pipe of limonite-cemented tuff—lithic-rich, chaotically bedded, and dipping inward—appears to rise through the entire thickness of the unit.

west of Broad Creek (pl. 1). From that point the flow was constrained by topography to move mainly to the north and northwest, covering an area of more than 100 km^2. Because the vent area of the Canyon flow also coincides with the thickest section of the tuff of Sulphur Creek, which there is intensely agglutinated, devitrified, and lineated downdip, it seems reasonable to presume that the petrographically similar tuff and flow issued from the same vent. The Canyon flow moved over the tuff of Sulphur Creek as it was welding and deforming around the vent and at the caldera wall. The viscous rhyolite flow became infolded with the agglutinate as it deformed, making a complex contact. The mapped contact approximates the true contact as nearly as it could be discerned and mapped, but undoubtedly many intricate contortions are not shown, especially in the intensely altered area of the Grand Canyon below Lower Falls.

Another similar but petrographically distinguishable rhyolitic lava flow overlies the Canyon flow north of the Grand Canyon, south of the Dunraven Pass Road. This flow, the Dunraven Road flow, has a well-preserved flow-breccia base and a pumiceous glassy top.

Units of the Upper Basin Member of the Plateau Rhyolite in the area of the Sour Creek resurgent dome and the Grand Canyon of the Yellowstone were dated with $^{40}Ar/^{39}Ar$ techniques by Gansecki and others (1996). They calculated the weighted mean average tuff of Sulphur Creek, all of the Canyon flow, and the Dunraven Road flow to be 481±8 ka.

Although Boyd (1961, p. 405-406) noted many of the individual relations just described, he was not able to follow them out in detail in his reconnaissance study. He suggested that the Canyon flow (including, as he mapped it, the tuff of Sulphur Creek the Canyon flow itself, and the Dunraven Road flow) might be a froth flow rather than a true lava flow. The froth-flow concept, proposed casually by Kennedy (1955, p. 495), implies a highly vesiculated but not thoroughly disaggregated rhyolitic material that erupted as a froth, then partially collapsed and continued to flow as a lava. The features of the Upper Basin Member in the Grand Canyon-Broad Creek area that led Boyd to interpret a froth flow are consistent with the present interpretation of a thick agglutinate succeeded by two lava flows, the agglutinate and the older flow having deformed together while they were emplaced across the wall of the Yellowstone caldera.

Vents for flows of the Upper Basin Member in the area of the Upper and Lower Geyser Basins are not known with certainty. By inference from outcrop distributions and the fragmentary evidence of flow-top structures, the vents probably were within a few kilometers of the west flank of the Mallard Lake dome (fig. 24). The vents for flows of the Upper Basin Member in the Grand Canyon-Broad Creek area are known (fig. 24). The distribution of the vents indicates that they were controlled by the ring-fracture zones of the two segments of the Yellowstone caldera.

The flows of the Upper Basin Member of the Plateau Rhyolite are mainly two-feldspar rhyolites in which plagioclase phenocrysts equal or exceed sanidine phenocrysts. The plagioclase phenocrysts, notably sieved, are somewhat more calcic (as much as about An_{30}) in some of the Upper Basin flows than in other plagioclase-bearing flows of the Plateau Rhyolite (generally about An_{20}). In some flows of the Upper Basin Member, calcic oligoclase or sodic andesine is the only or the greatly predominant feldspar although the rocks are rhyolites with about 72-76 percent SiO_2 (as recalculated free of water; table 10). Dating of feldspars from most of the Upper Basin flows confirms that they erupted early in the postresurgence phase of volcanism, generally within about 160,000 years of caldera collapse, but the Scaup Lake flow is 198,000 years old (table 2).

YOUNGER POSTCOLLAPSE RHYOLITIC VOLCANISM AND DOMING

Younger intracaldera rhyolites are included stratigraphically in the Mallard Lake and Central Plateau Members of the Plateau Rhyolite.

MALLARD LAKE MEMBER AND DOMING

The Mallard Lake Member of the Plateau Rhyolite comprises a single lava flow, the Mallard Lake flow. This unit was divided stratigraphically from the lavas of the younger Central Plateau Member on the basis of their ages, respectively older and younger, relative to formation of the Mallard Lake dome. Although, as noted in the previous section, the doming of the Mallard Lake flow is not as old as was originally inferred, it remains a unique and important postcollapse event and still appears to divide the rhyolites of the two members. Minor lithologic differences distinguish the Mallard Lake Member from at least the younger flows of the Central Plateau Member.

Glacial erosion of the high-standing Mallard Lake flow has stripped its original glassy and spherulitic shell in all but a few places. In most exposures the flow is crystalline and has conspicuous thin laminae that preserve intricate flow structures. Phenocrysts are moderately abundant (25-30 percent). Quartz and sanidine are common, but plagioclase is sparse. Minor opaque-oxide, clinopyroxene, and orthopyroxene phenocrysts also occur in glassy remnants of the flow carapace. By comparison, most (but not all) rhyolites of the younger Central Plateau Member have sanidine as a single feldspar phenocryst phase and lack orthopyroxene.

The Mallard Lake flow has been uplifted on the Mallard Lake dome. Its northwest and southwest margins preserve a steep marginal scarp and lobate form where they overlie the Upper Basin Member (pl. 1). By contrast, the northeast and southeast sides of the flow are completely buried by younger rhyolites of the Central Plateau Member. Therefore, the eruptive vent of the flow must have been within or beyond the eastern part of its present exposure. The Mallard Lake flow is particularly thick on the southeast side of the resurgent dome, near DeLacy Creek, suggesting that the vent was in that area or farther southeast in the buried ring-fracture zone (fig. 24). Younger faults, an eroded top, and an extensive till mantle prevent detailed observation of initial flow-top structures that might have been used to test this suggestion further.

The Mallard lake dome, which uplifts the Mallard Lake flow and part of the Scaup Lake flow, is elliptical, about 11 by 19 km, and it has a complex northwest-trending graben system along its long axis (fig. 27). The fault pattern, just as on the Sour Creek dome, clearly shows its origin by domical uplift of the underlying block. The northwest end of the graben system shows an outward splaying of faults and a decrease in fault displacement toward the margins of the dome; the southeast end of the graben system is buried but seems to show a similar outward splaying of faults. The graben system bifurcates toward the margin on the northwest and north sides.

The axis of the dome trends toward a major tectonic fault system outside the caldera. The axial graben system of the Mallard Lake dome projects toward the Madison Valley near West Yellowstone. This basin (fig. 5; pl. 1) is bounded by fault zones, including the one reactivated during the destructive Hebgen lake earthquake of 1959 (Witkind, 1964). Crudely linear concentrations of small-earthquake epicenters (fig. 3*B*) coincide with, and extrapolate eastward from, the fault zones bounding the West Yellowstone area.

Uplift of the Mallard Lake dome occurred approximately 160,000 years ago. Sanidines from the Mallard Lake flow on the dome give a weighted-mean K-Ar age of 151±4 ka, and the overlying Elephant Back flow, which postdates the dome, gives a sanidine K-Ar age of 153±2ka; several younger flows have nominal sanidine K-Ar ages ranging from 147 ka to 165 ka (table 2). The weighted mean of these ages is 161±1 ka (table 2), probably indicating formation of the dome about

Figure 27.—Axial graben of the Mallard Lake resurgent dome, Yellowstone National Park. The dome, about 11 by 19 km, uplifts the Mallard Lake flow and part of the Scaup Lake flow. View to northwest across DeLacy Creek.

160,000 years ago within a brief time that is not resolvable by conventional K-Ar dating.

No intracaldera flows are known to postdate the 198,000-year-old Scaup Lake flow and to definitely predate the 160,000-year-old Mallard Lake flow and dome although, of course, such flows could exist beneath other younger flows. Some flows of the Central Plateau Member give nominal ages slightly older than the Mallard Lake flow, but as noted above, they probably are all within the limit of geological resolution of the available K-Ar dating at this age range. As all the older exposed postcaldera flows have abundant plagioclase, the Mallard Lake flow probably is the oldest known rhyolite to have erupted with only sparse plagioclase phenocrysts. This difference in mineralogical composition, the apparent age gap between about 200,000 and about 160,000 years, the doming which accompanied or immediately succeeded the first extrusions of the younger age group, and the voluminous (about 900 km^3) mineralogically similar rhyolites that were erupted following this break in continuity (Christiansen, 1984, fig. 3) all suggest that eruption of the Mallard Lake flow represented the renewed rise of magma to a shallow level in the Yellowstone magma chamber. Conceptually, one may consider the Scaup Lake flow to be the last exposed representative of a crystallizing magma chamber related directly to the Lava Creek Tuff and the Mallard Lake flow to represent a reheating and “resetting” of the Yellowstone magmatic system. This relation will be considered further in a later section.

CENTRAL PLATEAU MEMBER

Flows of the Central Plateau Member of the Plateau Rhyolite occur mainly within the basin of the Yellowstone caldera, but they spill over the rim and bury much of the wall on the west and southwest sides and in small areas near Norris Geyser Basin and Heart Lake. The individual flows of this member are exceptionally large and thick, the largest having maximum diameters as great as 32 km and thicknesses of more than 300 m. Although these very large flows have steep marginal scarps and lobate plans typical of rhyolitic lavas, they are so large that their gently convex profiles appear nearly flat (fig. 28).

Except for their sizes, most of the flows of the Central Plateau Member have the common surface features and flow structures of ordinary rhyolite flows. A few, however, have features that indicate special conditions of emplacement. In particular, the West Yellowstone flow, at the north end of the Madison Plateau, is here interpreted as having been emplaced against a body of glacial ice that must have occupied much of the rhyolite plateau farther east when the flow was extruded. The east margin of the flow (pl. 1)—Little Firehole Meadows, Buffalo Meadows, and the Lower Geyser Basin)—is broadly concave with several large lobate reentrants, a form unlike the frontal scarps of most rhyolite flows or even of the west margin of the West Yellowstone flow itself. Several of these reentrants are interrupted by convex lobate tongues of lava from the West Yellowstone flow that must have been emplaced into the reentrants slightly later than the reentrants themselves were formed. In particular, a large lobe separates the Lower Geyser Basin from Buffalo Meadows (pl. 1). The relative ages of these lobes, where determinable, are progressively younger down-flow from the source. The only reasonable way of explaining these relations is to assume that the initial spread of the lava flows was obstructed by some now-missing body that occupied the reentrants but disappeared progressively with emplacement of the flow, allowing the small lobes to breach the chilled margin and invade the reentrants successively. The presence of glacial ice against this margin would have provided the proper conditions to form the observed features. Along much of this east margin, the flow front is sheathed in perlitic glass, probably reflecting early high-temperature hydration by glacial meltwaters at the chilled margin.

No equally strong argument can be made for other flows of the Central Plateau Member having flowed against ice during emplacement, but large reentrants on the east sides of the Bechler River, Grants Pass, Elephant Back, and Solfatara Plateau flows and possibly the Spring Creek flow could have formed similarly.

Two flows of the Central Plateau Member, the Hayden Valley flow and possibly the Nez Perce Creek flow, appear to have been emplaced partly into water. The Hayden Valley flow is completely glassy and highly perlitic along its east margin where it overlies a sequence of fluvial and lacustrine sediments. These sediments were plowed up locally and incorporated into the base of the flow. They commonly show effects of plastic deformation beneath the flow and injected the flow as numerous clastic dikes; the sediments, therefore, must have been soft and wet when the flow was emplaced on them. Richmond (1976, p. 11-14) also found evidence in sediments associated with the Hayden Valley flow that the flow had erupted into a body of water. The Nez Perce Creek flow locally also has a strongly perlitic breccia phase along roadcuts in Firehole Canyon, and parts of its outer margin also may have flowed into a small body of water near there.

Rhyolitic ash-flow tuffs are present in several areas within the flow sequence of the Central Plateau Member. It was originally thought, before completion of the geologic mapping, that all these areas were likely to be parts of a single ash-flow unit, which was named the Shoshone Lake Tuff Member of the Plateau Rhyolite (Christiansen and Blank, 1972; U.S. Geological Survey, 1972). Because field relations and K-Ar dating have since shown that at least two separate ash-flow sequences occur within the Central Plateau Member, the Shoshone Lake Tuff Member was abandoned by Christiansen and Blank (1974a).

The western ash-flow outcrop areas—west of Shoshone Lake (including the section described by Christiansen and

Figure 28.—Pitchstone Plateau and Teton Range, Wyoming. The Pitchstone Plateau is the nearly flat but broadly domical top of a large flow of the Central Plateau Member of the Plateau Rhyolite. Gently west-dipping timbered slopes on lower flanks of Teton Range are ash-flow sheets of the Yellowstone Group, the tuff of Kilgore, and the Conant Creek Tuff; these slopes indicate the amount of westerly tilting of the range during Pliocene and Quaternary time. Aerial view to south. Photograph by Warren Hamilton.

Blank, 1972, p. B14), in the Bechler River drainage, and perhaps around Little Firehole Meadows—probably are one sequence that postdates the Bechler River flow and predates the Pitchstone Plateau and Grants Pass flows. This unit is shown on more recent geologic maps (pl. 1; Christiansen and Blank, 1974a, b) as the tuff of Cold Mountain Creek and is included within the Central Plateau Member. This pyroclastic sequence is generally nonwelded or only slightly welded. It may date the opening of the Pitchstone Plateau vent and be a direct precursor of the Pitchstone Plateau and Grants Pass flows. The Grants Pass flow lies immediately adjacent to the Pitchstone Plateau flow over the same ash-flow sequence and has a linear vent that trends toward the Pitchstone Plateau vent. The Grants Pass flow probably represents a lateral fissure vent related to the much larger Pitchstone Plateau eruption. Similarly, the Moose Falls flow extends this linear trend outside the caldera to the southeast. The Grants Pass, Pitchstone Plateau, and Moose Falls flows all have sustantially similar ages.

The rhyolitic ash-flow sheet east of Shoshone Lake and around West Thumb is older than the tuff of Cold Mountain Creek and the Bechler River flow. This unit, designated on later maps as the tuff of Bluff Point (pl. 1; Christiansen, 1974b; Christiansen and Blank, 1975a), must have erupted in the general vicinity of the West Thumb of Yellowstone Lake, for it is widely distributed around that area (although largely buried to the east), is most densely welded, has the highest proportion of devitrified welded tuff, and has generally larger and more abundant phenocrysts and xenolithic inclusions near West Thumb than elsewhere. Vitrophyric rhyolitic flow fragments are the most abundant inclusions. For these reasons, it is inferred that the basin of West Thumb is a relatively small caldera within the much larger Yellowstone caldera and that the West Thumb caldera collapsed upon eruption of the tuff of Bluff Point. The northeast and southeast sides of the West Thumb caldera were partly buried by the younger West Thumb and Aster Creek flows; those flows did not quite merge in the northeastern strait of West Thumb. Other flows and sediments also may have partly filled the caldera basin. Nevertheless, West Thumb remains one of the deepest parts of Yellowstone lake. Although small compared to the Yellowstone caldera, the basin of West Thumb is as large in diameter as the well-known caldera of Crater Lake, Oregon.

The tuff of Bluff Point may have erupted onto the surface of the Dry Creek flow while it was still moving. The tuff of Bluff Point in several places, notably along the Lewis River between Shoshone and Lewis Lakes, is reddened and appears to be deformed and infolded into contorted and steeply dipping flow layers in the top of the Dry Creek flow. The Dry Creek flow itself appears to have flowed from a vent now downdropped in the West Thumb caldera. If my interpretation of the very close age relations of these two units is correct, it seems likely that the tuff of Bluff Point was an explosive late-stage eruption from the Dry Creek vent, perhaps resulting from the introduction of water into that vent. This pyroclastic eruption tore fragments of still glassy Dry Creek flow from the vent and caused subsidence of the vent area to form the caldera. The viscous lava may have continued to flow some distance downslope by gravity despite the beheading of its vent, carrying and contorting the tuff of Bluff Point as it moved.

Dating the flows of the Central Plateau Member by the conventional K-Ar method required working near the method's limits of resolution. However, combining the known stratigraphic relations between certain flows of the Central Plateau Member with all the available K-Ar dates suggests that differences in true geologic age between flows can be resolved to within about ±10 ka. The available sanidine ages appear to cluster in groups separated by approximately 45±10 ka, namely at around 161 ka, 110 ka, and 71 ka (table 2). Thus it seems either that the 70,000-year pulse exhausted the available, potentially eruptible magma or that the latest interval has been somewhat longer than previous ones and might even be close to the conditions for recurrence of eruption.

PETROGRAPHY

Rhyolites of the younger postcollapse sequence are generally similar in mineralogy and chemistry. All are high-silica rhyolites of uniform chemical composition, and most flows of the Central Plateau Member contain moderately abundant phenocrysts. The principal phenocrysts in all the late postcollapse flows are quartz and sanidine. Minor sodic oligoclase is also present in the Mallard Lake flow and in a few flows of the younger Central Plateau Member. Ferroaugite and magnetite are present in the glassy parts of all the phenocryst-bearing rhyolites of this sequence. In the two younger groups of Central Plateau flows, fayalite is present and clinopyroxene is more iron-rich than in the oldest group. Phenocrysts of hydrous minerals are not known.

TECTONIC CONTROLS OF YOUNGER POSTCOLLAPSE VOLCANISM

Individual flows and vents of the Central Plateau and Roaring Mountain Members, shown on plate 1, occur in two linear zones. The irregularly shaped Elephant Back flow, as mapped, may actually comprise two or three separate bodies, but no distinguishing features could be found in the field or on aerial photographs to indicate consistent interflow contacts. Even if there should be any additional vents, they would have to be near the one mapped vent. Many of the vents of the Central Plateau Member are known from the forms and flow-top structures of the flows, and the vent areas for most other flows can be inferred within broad limits from similar but incompletely exposed features. Known vents are aligned linearly (pl. 1; fig. 5), and all additional inferred vents must lie within the same linear zones.

Each of the two linear vent zones of the Central Plateau Member trends at both ends toward major tectonic faults outside the caldera. One of these volcanic vent zones lies along the axis of the Madison and Pitchstone Plateaus and essentially connects the Teton fault zone and its northern continuation (Love, 1961; Love and others, 1973; Love and Keefer, 1975) with the Hebgen zone at the edge of the upper Madison Valley. Hamilton (1960b) recognized this volcanotectonic alignment. A small graben extends between Trischman and Douglas Knobs (pl. 1), within and parallel to the alignment of vents on the Madison and Pitchstone Plateaus. The linear vent of the Grants Pass flow and the aligned vents of the Grants Pass, Pitchstone Plateau, and the Moose Falls flows are also on this trend. The second vent zone trends from near the Flat Mountain Arm of Yellowstone Lake, through the West Thumb caldera, and along the axis of the Central Plateau. This vent zone connects recurrently active faults bounding the east face of the Red Mountains (Love and Keefer, 1975) with the fault zone extending north from Norris Geyser Basin to Mammoth Hot Springs. The fault zone along the Norris-Mammoth corridor controlled a major extracaldera zone of volcanism and hydrothermal activity and also connects with the East Gallatin fault system bounding the east side of the Gallatin Range in the northwest corner of the park (fig.1; Ruppel, 1972) and the fault zone bounding Gardners Hole and Swan Lake Flat (pl. 1). Notably, many of the earthquakes that occur within the Yellowstone caldera (fig. 3*B*) are in this eastern zone of vents; a magnitude-6.1 earthquake and an associated swarmlike sequence in June 1975, occurred in the zone (Pitt and others, 1979). The principal flows of the younger postcollapse rhyolites extruded only within those parts of the ring-fracture zone of the Yellowstone caldera that are intersected by these two major fault alignments. The small domes and flows of the Obsidian Creek and Roaring Mountain Members and the Moose Falls flow occur along or near these same fault systems outside the caldera.

Relations described earlier demonstrate that the known early postcollapse rhyolite eruptions postdated resurgent doming and were controlled directly by the ring-fracture zones of the two segments of the Yellowstone caldera. Smith and Bailey (1968a) designated major ring-fracture volcanism as stage VI of the resurgent cauldron cycle. Younger postresurgence rhyolites within the caldera, however, are controlled by the ring-fracture zones only where they intersect linear tectonic structures. An intriguing question arises: Do the younger rhyolites represent a simple continuation of postcollapse volcanism in the caldera, or might they signal a new phase in the eruptive history of the rhyolite plateau? The vent alignments of the younger flows suggest that the major underlying magma body, which sustained the third volcanic cycle of the Yellowstone Plateau volcanic field, had crystallized to some considerable degree and was then broken by renewed extension along tectonic fault zones extending into the partly solidified intrusion. Thus, it seems either that the younger Plateau Rhyolite flows represent the insurgence of new magma through a solidified underlying intrusive body or that they record the upwelling of liquid magma from a deeper level into fractures that have propagated into a partly crystallized roof zone. The latter explanation seems more likely to be correct. In particular, the apparent absence of basalts on the floor of the Yellowstone caldera, in contrast to the abundance of basalts at Island Park and similar older rhyolitic areas farther west, suggests that no throughgoing fractures have penetrated any completely crystallized magma body beneath the Yellowstone Plateau. Furthermore, a consistent secular change in oxygen-isotopic ratios suggests that the Central Plateau magma was present in the Yellowstone magmatic system long before it appeared at the surface as rhyolitic lavas (Hildreth and others, 1984).

Geological and geophysical evidence taken together suggest that magma probably is still at least partly molten beneath the Yellowstone Plateau (Eaton and others, 1975; Lehman and others, 1982; Smith and Braile, 1984; 1994). Therefore, as the upper portions of this rhyolitic magma body cool and crystallize, they may be broken by tectonic faults and reintruded by magma rising from deeper levels. The periodic nature of the youngest stage of volcanism—eruption of the Central Plateau Member flows at intervals of 40,000±10,000 years—appears to reflect the time scale for sufficient cooling of the top of the high-level magma chamber to allow the accumulation of enough stress in the consolidated roof to cause major faulting. If this concept is correct, the erupted lavas must represent partly crystallized magma from levels considerably below the solidified top of the chamber, for they contain moderate amounts of phenocrysts and the phenocrysts of each successive eruptive episode vary slightly toward more evolved compositions. Such episodic rise of magma into the chamber roof represents periodic advective heat transfer in the high-level chamber and implies a continued heat source.

EXTRACALDERA RHYOLITIC VOLCANISM

During the time of postcollapse rhyolitic activity in the Yellowstone caldera, rhyolites of two distinct lithologic types erupted north of the caldera. The more porphyritic of these extracaldera rhyolites constitute the Obsidian Creek Member of the Plateau Rhyolite, and aphyric to very sparsely porphyritic rhyolites form the Roaring Mountain Member.

OBSIDIAN CREEK MEMBER

Only two bodies of the Obsidian Creek Member of the Plateau Rhyolite, the Willow Park dome and Gibbon Hill dome, have been dated, at about 316 ka and 116 ka, respectively (table 2). The Obsidian Creek Member petrographically resembles the Scaup Lake flow of the Upper Basin Member. The Obsidian Creek lavas generally seem to be conformable, or nearly so, on the Lava Creek Tuff outside the Yellowstone caldera. Most of the Obsidian Creek rhyolitic bodies are volcanic domes, but at least two rhyolite-basalt mixed lava complexes are included in the member because of their similar stratigraphic positions and the similar petrography of their rhyolitic components. All known flows of the Obsidian Creek Member occur in a linear zone or corridor that extends from the caldera rim in the vicinity of Norris Geyser Basin northward toward Mammoth Hot Springs. This Norris-Mammoth corridor is a zone of recurrent faulting as shown by fault blocks of Paleozoic and early Tertiary rocks projecting above the Lava Creek Tuff as well as by exposed younger faults (pl. 1). It was also the locus of both early and late postcaldera rhyolitic eruptions and of pre-Lava Creek and post-Lava Creek basaltic vents. The Norris-Mammoth corridor is also the only zone of major hydrothermal activity that extends far outside of the Yellowstone caldera (fig. 29).

The rhyolitic domes that form most of the Obsidian Creek Member of the Plateau Rhyolite are the Paintpot Hill, Gibbon Hill, The Landmark, Appolinaris Spring, and Willow Park domes (pl. 1). Each is about ½-1½ km across. Gibbon Hill, the largest of the domes, rises nearly 250 m above its base. Most of the domes were eroded before the eruption of younger postcaldera rhyolites, but parts of the original glassy shells of each dome except Paintpot Hill are preserved. The Willow Park dome is still largely pumiceous at the surface. All of them have moderately abundant quartz, sanidine, and plagioclase phenocrysts. Another body, thought to be a rhyolite dome by Boyd (1961, p. 400-401, pl. 6) and called the Geyser Creek dome by Christiansen and Blank (1975a), has been reexamined and is considered here (pl. 1) to be part of the Gibbon River flow, a body of the Roaring Mountain Member of the Plateau Rhyolite.

The principal rhyolite-basalt mixed-lava complexes of Yellowstone National Park are the mixed lavas of Gardner River and Grizzly Lake (pl. 1); both are included stratigraphically in the Obsidian Creek Member. The mixed lavas of Gardner River and Grizzly Lake have been described and their origins discussed in several papers. The significance of the intimate association of rhyolite and basalt in these two bodies is reviewed briefly here as it bears on problems of concern in later parts of this paper. Iddings (1899a, p. 430-432), who described both of these complexes, concluded from structural and textural relations that the rhyolitic component of each complex succeeded the basaltic component and partly fused it, resulting in streaky mixtures of the two at their contacts. Iddings noted that this interpretation required the rhyolitic liquid to have been at a temperature above the melting point of the basalt, but, lacking sufficient experimental data, he did not recognize the practical difficulty of that interpretation. Fenner (1938; 1944), in a study of the "Gardiner River complex," concluded that the rhyolite flow moved through a deeply eroded older basalt and that the rhyolite or emanations from it dissolved certain constituents of the preexisting basalts. Fenner recognized that, because the rhyolite was crystallizing as it interacted with the basalt, it could not have been superheated. He attributed its ability to accomplish the partial dissolution to unknown properties of rhyolitic magmas or their emanations. Fenner also noted that the trends of chemical variation in the complex were along straight lines for all elements and that for unexplainable reasons the rhyolitic emanations must have been saturated by the basaltic constituents in exactly the proportions in which they occurred in the basaltic minerals that were dissolved. Wilcox (1944) mapped and studied the petrography of the Gardner River lavas in detail. He concluded that the rhyolitic and basaltic components must both have been magmatic liquids at the time the complex formed. Wilcox showed that the effect of the cooler, crystallizing rhyolitic liquid on the basaltic liquid produced the apparent intrusive relations of the rhyolite into the basalt. He showed that the basalt was chilled against rhyolite at their mutual contacts but never the reverse, although the rhyolite is chilled at contacts on the previous ground surface. Furthermore, quartz and sanidine xenocrysts in the basalts occur at some distance from contacts with rhyolite, showing that the basalt must have been at least partly liquid and must have incorporated rhyolitic materials as well as being partly incorporated into the rhyolitic liquids. The straight-line chemical variation trends, therefore, are simple mixing lines between two initially liquid components. Wilcox assumed that the two liquids were simultaneously erupted lava flows that coalesced by coincidence on the ground surface.

A resolution of these diverse observations and interpretations of the complex was attempted by Hawkes (1945), who reviewed the papers by Fenner and Wilcox from the perspective of his own observations on rhyolite-basalt mixed lavas in Iceland and Scotland. Hawkes proposed that the complex is indeed a result of the intimate mixture of two simultaneous liquids but that rather than representing two separate lavas that met at the surface, the complex is the product of a single composite feeder intrusion with basaltic margins and a rhyolitic core, such as had been described from many localities in the Tertiary of Britain and Iceland. Furthermore, he proposed that the rhyolitic portion of the complex at a higher level, now exposed, intruded the partly solidified but still hot basalt, some of which had already been contaminated by rhyolitic material during its ascent through the composite feeder. Hawkes completed his review by noting that the fundamental question about the mixed lavas of Gardner River, thus, is not so much how the complex itself formed but rather why two magmatic liquids were produced and intruded simultaneously. This problem is returned to later.

Late-stage intrusion of the rhyolite into already hybridized and partly solidified basalt at the presently exposed level should not be a necessary part of this explanation and does not accord with field relations described carefully by Wilcox: the margins of the exposed complex are mainly the rhyolitic component, and hybridized basalt commonly was chilled against hybridized rhyolite. The complex is best explained as a flow from a composite feeder in which rhyolitic and basaltic liquids were mixed to varying degrees. Boyd (1961, p. 403) and Struhsacker (1978) proposed essentially the same idea for both the Gardner River and Grizzly Lake mixed lavas.

ROARING MOUNTAIN MEMBER

A few small rhyolite flows north of the caldera that either completely lack phenocrysts or have only sparse small phenocrysts constitute the Roaring Mountain Member of the Plateau Rhyolite. Although the Roaring Mountain Member was originally assumed to be entirely correlative with the Central Plateau Member (Christiansen and Blank, 1972), K-Ar dating (table 2) has shown a range of ages from about 400 ka (the Cougar Creek flow) to 80 ka (the Crystal Spring flow). The distinctive lithologic characteristic of all the flows of the Roaring Mountain Member—few or no phenocrysts—is evidence that they were emplaced at high temperatures.

The Gibbon River flow was originally regarded (U. S. Geological Survey 1972; Christiansen and Blank, 1975a) as part of the Central Plateau Member. As such, it was the only phenocryst-poor flow of the Central Plateau Member. The vent dome of the Gibbon River flow, near the buried edge of the caldera at the south end of the Norris-Mammoth corridor (pl. 1; fig. 5), has been dated at about 90 ka. Detailed geochemical studies made subsequent to the earlier mapping and referred to later in this paper (table 14), show that the lava of the Gibbon River flow has the distinctive characteristics of lavas of the Roaring Mountain Member, markedly different from those of the Central Plateau Member. The Gibbon River flow is included on plate 1 with the Roaring Mountain Member.

POST-LAVA CREEK BASALTIC VOLCANISM

Basalts have erupted intermittently around the margins of the Yellowstone Plateau since the climactic third-cycle ash-flow eruptions and caldera collapse. Most of the basaltic eruptions occurred in fairly restricted areas that also were the sites of pre-Lava Creek Tuff basaltic eruptions. For example, the Swan Lake Flat Basalt came from several vents in the area north of the Yellowstone caldera, mainly along the zone of recurrent faulting that extends north from Norris Geyser Basin. Basalts also were erupted in this area during the first and second cycles of the volcanic field, as well as early in the third cycle (the Undine Falls Basalt). The Swan Lake Flat Basalt, which erupted early in the postcollapse phase of the third volcanic cycle, is conformable on the Lava Creek Tuff (Christiansen and Blank, 1972). Other early post-Lava Creek basalts, the Falls River Basalt, are conformable on the Lava Creek Tuff in the southwestern part of the Yellowstone Plateau, also an area of earlier basaltic activity.

Basalts in several areas around the margins of the Yellowstone Plateau are confined within old valley topography that was eroded below the earlier post-Lava Creek surface but to levels shallower than the present canyons; other young basalts lie within the Snake River downwarp. These younger basalts include the Madison River and Osprey Basalts in the north half of the volcanic field (pl. 1; see Christiansen and Blank, 1972) and the basalt of the Snake River Group in the western part of Island Park and farther west (pl. 3; see Malde and Powers, 1962; Leeman, 1982b). One flow of Osprey Basalt has a K-Ar age of about 220 ka (table 2).

One other area of post-Lava Creek Tuff basaltic eruptions forms the floor of the Island Park basin. The basaltic lavas of this area were named the Gerrit Basalt by Christiansen (1982), after a railroad siding 5 km east-northeast of Upper Mesa Falls (pl. 2). The Gerrit Basalt overlies the Lava Creek Tuff both conformably and unconformably but appears to be entirely older than the lava flows of the Central Plateau Member of the Plateau Rhyolite. One of the younger flows of the Gerrit has yielded a K-Ar age of about 200 ka (table 2). Since eruption of the Gerrit Basalt, a few basaltic cinder cones and flows have erupted from a linear vent zone that crosses the west rim of Island Park (pl. 2). This zone is related structurally and chemically to the basalts of the Snake River Plain that underlie a large region to the west.

All the plateau-marginal basalts are essentially similar in lithology, mineralogy, and chemistry. In normative composition they are olivine tholeiites (table 16); most have phenocrysts of labradorite and olivine and all have olivine and clinopyroxene as well as plagioclase and iron and titanium oxides in the groundmasses. Minor chemical and mineralogical differences among the various basalts are discussed later. The only major chemical variant among them is an alkalic intermediate-composition lava associated with the basalts of the Snake River Plain in the younger vent zone across the west rim of Island Park near High Point (pl. 2). This lava, called latite by Hamilton (1965), is similar to certain alkalic intermediate flows associated with the Snake River Group farther southwest along the plain (Powers, 1960; Stone, 1966; Leeman and others, 1976). Petrographic and lithologic heterogeneities at the hand-specimen and outcrop scales suggest that it reflects, at least to some degree, the mixing of lavas of different compositions.

In both the northern and southern parts of the Yellowstone Plateau, small mixed-lava bodies, whose age relations to the Lava Creek Tuff are uncertain, are mapped as the basalts of Geode Creek and Mariposa Lake (pl. 1). Although these small bodies are mainly basalt, they are somewhat siliceous (table 16), are compositionally zoned from the margins inward, and have streaky textures and abundant xenocrysts of quartz and sodic plagioclase. These features indicate mixing with rhyolitic material, probably as a magmatic liquid comparable to the more obvious mixed lavas of Gardner River and Grizzly Lake described earlier. Two outcrop areas of the basalt of Geode Creek overlie their eruptive vents, as shown by a dike passing into a flow in the southern outcrop and a fan-shaped joint pattern and an upwelling flow-structure pattern in the northern outcrop.

Most or all of the known vents for plateau-marginal basalts of the third volcanic cycle occur within important tectonic fault zones that have had recurrent late Cenozoic displacement (pl. 1). A few small erosional remnants of Madison River Basalt are present on the caldera wall near Purple Mountain, but no basalts are known to have vented on the floor of the Yellowstone caldera.

HYDROTHERMAL ACTIVITY

Yellowstone is best known as an area of abundant and spectacular hydrothermal activity. Probably no other area of the world is comparable in the variety and intensity of such activity—not even the neovolcanic regions of Iceland, New Zealand, and Kamchatka. In particular the number of actively erupting geysers is greater in Yellowstone than anywhere else (Allen and Day, 1935, p. 171). The current state of knowledge about the Yellowstone hydrothermal system was summarized by Fournier (1989). Smith and Bailey (1968a) designated the final phase of the resurgent caldera cycle as stage VII, terminal solfataric and hot-spring activity. As no magmatic eruptions are known to have occurred within or near the Yellowstone caldera during the last 70,000 years, the region may have entered this stage. It is equally possible, however, that additional rhyolitic eruptions may occur. The Yellowstone caldera, thus, could now be in Smith and Bailey's stage VII, in late stage VI, or could even be in the first stage of a new volcanic cycle.

The distribution of known present and past hydrothermal activity in the Yellowstone Plateau (pl. 1; fig. 29) clearly shows that the activity is structurally controlled. Although most of it occurs within the ring-fracture zones, activity extends locally across the caldera rim into adjacent faulted areas. Extracaldera hydrothermal activity is notable in the northeastern part of the caldera in the vicinity of Hot Springs Basin. Similar areas are present southeast of the caldera at Brimstone Basin and south of the caldera in the Heart Lake Geyser Basin. Most of these areas are near the intersections of

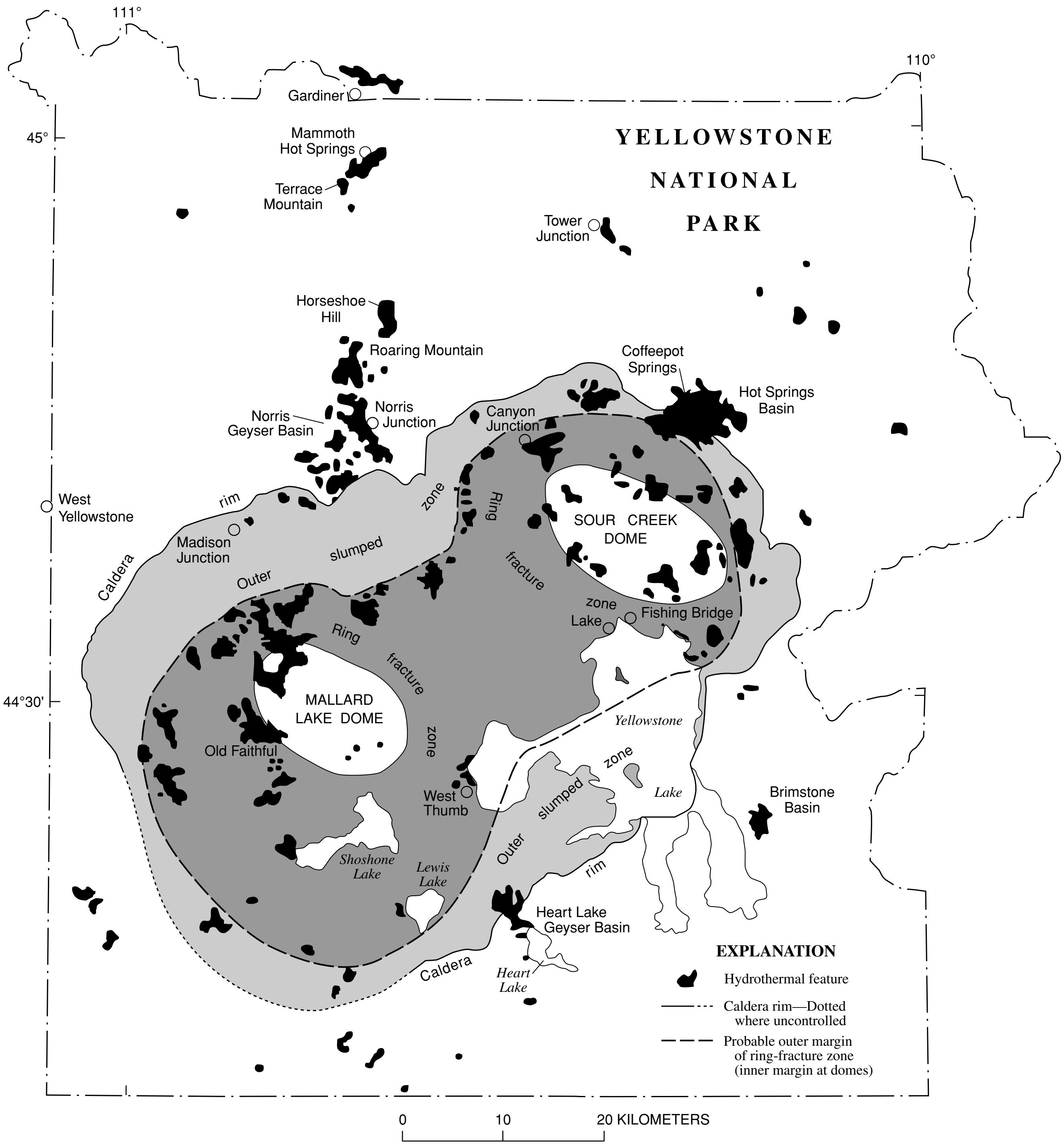


Figure 29.—Distribution of known hydrothermal features in the Yellowstone Plateau region, both active and extinct. The hydrothermal features are structurally controlled. Most major hydrothermal areas are within the ring-fracture zone of the Yellowstone caldera. The only major extracaldera hydrothermal zone extends north from the large boiling springs and geysers of Norris Geyser Basin through areas of fumarolic activity and acid alteration to the moderate-temperature travertine deposits of Mammoth Hot Springs and older travertine terraces near Gardiner. Hydrothermally active and altered areas include siliceous and calcareous hot-spring deposits, areas of acid alteration, hydrothermal-explosion craters, and ice-contact deposits localized and cemented by hot springs (see plate 1).

important zones of tectonic faulting with the caldera rim. Brimstone Basin is about 13 km from the rim; the other areas lie within about 5 km. The only major zone of hydrothermal activity that extends far from the caldera (pl. 1) begins at the rim near Gibbon Canyon and extends northward through Norris Geyser Basin, Roaring Mountain, Horseshoe Hill, Mammoth Hot Springs, and the largely Pleistocene travertine terraces near Gardiner (Fraser and others, 1969; Pierce and others, 1991; Sturchio and others, 1994). This hydrothermal zone along the Norris-Mammoth corridor, about 43 km long, is marked by a variety of activity from the large boiling springs and geysers of Norris Geyser Basin, through areas of extensive fumarolic activity and acid alteration such as Roaring Mountain, to the moderate-temperature travertine deposits of Mammoth and Gardiner. The zone of mappable present and past hydrothermal activity is discontinuous, especially between Horseshoe Hill and Terrace Mountain; boiling springs, extensive acid alteration, and volcanic vents are restricted to the part of the zone south of the Gardner River. As noted previously there are numerous small rhyolitic and basaltic volcanoes, both older and younger than the Lava Creek Tuff, along this part of the zone.

The main controlling factors in the localization and development of the Yellowstone hydrothermal systems, inferred from figure 29 and shown in plate 1, are: (1) the presence of faults and other fractures to enable deep circulation, and (2) the effects of local topography on the hydrologic flow regimen—the most spectacular hot springs and geysers occur within topographic basins.

The most active hydrothermal systems are those within and immediately adjacent to the caldera ring-fracture zones. The Mallard Lake caldera segment is partly encircled by hot springs and areas of active acid alteration in the upper drainage of Nez Perce Creek, the Lower, Midway, and Upper Geyser Basins, Little Firehole Meadows, the Summit Lake area, the area of Lone Star Geyser, and the Shoshone and West Thumb Geyser Basins (pl. 1). The Sour Creek segment is ringed by hot-spring and fumarolic areas in and around Hayden Valley, the upper Grand Canyon of the Yellowstone, Sevenmile Hole, upper Sour and Broad Creeks, the Mushpot-Mudkettle area, Sulphur Hills, Turbid Lake, Mary Bay, and Steamboat Point (pl. 1). Other areas of relatively weak or very local activity occur in the same zones as well as in a few other isolated areas. Such isolated areas are present along fault zones near the Lamar River, near The Narrows of the Grand Canyon of the Yellowstone, and near the Snake River in the southern part of Yellowstone National Park.

Investigators generally agree that the water circulated in the hot-spring systems of Yellowstone and similar regions is mainly of meteoric origin (for example, Hague, 1911; Allen and Day, 1935; White, 1957; Truesdell and others, 1977)); more than 95 percent is meteoric according to isotopic data reviewed by White (1969, p. 280). The source of heat in high-discharge, high-temperature hydrothermal systems like the major ones of the Yellowstone region must be magmatic (White, 1957, p. 1641-1643; Fournier and Pitt, 1985; Fournier, 1989). An exceptionally high flux of ^{3}He from Yellowstone thermal areas, necessarily from deep sources, confirms this conclusion (Craig and others, 1978; 1979; Kennedy and others, 1985; 1987). Such hydrothermal systems, developing as stages in the evolution of volcanic fields, are in effect deeply circulating convective heat-flow systems related to the cooling and crystallization of magma (White, 1957, p. 1642-1643; 1968, p. C101-C103). Density differences between cold and hot water are the primary driving forces of the circulation.

Because of the long-lived hydrothermal activity of many volcanoes, especially silicic ones, it commonly is difficult or impossible to separate the alteration or hydrothermal deposits of the earlier stages from those of the final stages. The hydrothermal system of the Yellowstone Plateau has been active at least since before the Pinedale glaciation, the last major glaciation in the region. Holmes (1883a) long ago pointed out clear evidence that Terrace Mountain, west of Mammoth Hot Springs, is an extinct travertine terrace that was overridden by ice of the last major glaciation (see also Allen and Day, 1935, frontispiece). More recently, U-Th disequilibrium data published by Sturchio and others (1994) indicate a mean age of 373±17 ka for travertine from Terrace Mountain. Thus, the hydrothermal system of the Yellowstone Plateau has been active for much longer than the 47,000-34,000 years (Sturchio and others, 1994) since the beginning of Pinedale glaciation. Zeolite- and opal-cemented kames of Bull Lake age (the last major pre-Pinedale glaciation) occur in numerous areas of present hydrothermal activity around the rhyolite plateau, particularly in the Lower Geyser Basin. These deposits probably were localized and formed by englacial hydrothermal activity (Muffler and others, 1971, 1982a; b). Travertine deposits from the Mammoth Hot Springs-Gardiner area have yielded U-Th disequilibrium ages (Sturchio and others, 1994) ranging from about 7 ka to about 57 ka (as well as the much older deposits of Terrace Mountain noted above). Drill cores from the Biscuit Basin area of Upper Geyser Basin gave U-Th ages of about 15-20 ka (Sturchio and others, 1987). Furthermore, Bargar and Fournier (1988) showed that homogenization temperatures of fluid inclusions from hydrothermal minerals from Yellowstone are always equal to or greater than present hot-spring temperatures nearby, indicating that temperatures have not been below those of the present since at least some time during the latest major glaciation. Chalcedonic sinter (D. E. White, oral commun., 1971) is interlayered with the sediments of Inspiration Point (Christiansen and Blank, 1975b; Richmond, 1976; 1977) on the south rim of the Grand Canyon of the Yellowstone, about 1/2 km west of Artist Point. These sediments appear to have been deposited on the Canyon flow before the Grand Canyon was cut through it; the age of the sediments, therefore, is probably close to 600,000 years. Fournier (1989) concluded that hydrothermal activity began soon after formation of the Yellowstone caldera and has remained vigorous since then.

Total convective heat flow from the principal hydrothermal areas of the Firehole River drainage has been estimated by Fournier and others (1976) as about 2.2 x 10^9 W (5.2 x 10^8 cal/sec). Their method, based on the chloride content of the waters, utilizes temperatures of liquid water in the major underground reservoirs; temperatures are either measured in drill holes (White and others, 1975) or are estimated on the basis of silica contents of the waters (according to the method of Fournier and Rowe, 1966) and by mixing models (Fournier and Truesdell, 1974). At a depth where the hydrothermal fluid is entirely liquid, a mass unit of chloride in the water can be equated with an enthalpy, or heat content, of the liquid. Thus, a measurement of the total flow of anomalous chloride in surface streams draining a hydrothermal basin enables a calculation of the total heat flow from the basin. The average heat flux over the area of the Firehole drainage basin is about 46,000 mW/m^2, more than 700 times the world average heat flow (Fournier and others, 1976; Fournier, 1989). The heat flow from all chloride-bearing springs in Yellowstone National Park northeast of the continental divide, an area containing most major thermal features of the rhyolite plateau except the Shoshone Lake and Heart Lake Geyser Basins, is estimated by this method to be about 5.1 x 10^9 W (Fournier and others, 1976); the total heat flow from all springs and fumaroles in the rhyolite plateau is at least 5.3 x 10^9 W. Essentially all of the hydrothermal areas contributing to this flow are within or adjacent to the Yellowstone caldera or the Norris-Mammoth corridor and lie within an area of about 3,000 km^2. The average flux in this area, therefore, is more than 1,700 mW/m^2, about 27 times the global average and about 20 times the average for the northern Rocky Mountains (Blackwell, 1969). A marine-type survey of conductive heat flow through the bottom sediments of Yellowstone Lake (Morgan and others, 1977) revealed that higher-than-regional values of 100-300 mW/m^2 just outside the caldera rise to 600-700 mW/m^2 at the edge of the ring-fracture zone. Areas of sublacustrine hydrothermal discharge are characterized by values as great as about 1,700 mW/m^2.

White (1968, p. C102) noted that the continued high heat flow from long-lived hydrothermal systems is difficult to explain by a model in which a statically cooling body of magma supplies heat to the deeply circulating hot-water system. In discussing the Central Plateau Member of the Plateau Rhyolite, changing structural control of the plateau flows with time was interpreted to suggest that, as the large body of rhyolitic magma beneath the Yellowstone Plateau cools and crystallizes, its roof is fractured by active faults that extend into the plateau from the surrounding region. Periodic rise of still hot, little-crystallized magma from lower in the body results in increasingly localized rhyolitic vents along the projections of those fault zones. If the fracturing kept pace with crystallization, the mechanism would result in continued major upward advective transfer of heat to the roof of the rhyolitic magma body, thus helping to sustain the circulation of the hydrothermal system. White (1968, p. C102-C103) also noted that convective circulation of magma within the chamber could help to maintain high temperatures at the base of the deep hydrothermal system. Shaw (1965) calculated that the viscosities of rhyolitic magmas should be low enough to enable adequate convection, at least in large bodies. Thus, the volcanic periodicity within the last 160,000 years might also correspond to a form of episodic convection of the magma.

Using the assumptions of White (1957, p. 1642; 1968, p. C102), the heat flow of about 5.3 x 10^9 W estimated for the hydrothermal systems of the Yellowstone Plateau would require the crystallization and cooling from 900° to 500°C of about 0.10 km^3 of rhyolitic magma per year (Fournier and others, 1976; Fournier and Pitt, 1985). If the hydrothermal system had been essentially constant for the last 600,000 years, a total of 60,000 km^3 of rhyolitic magma would have had to crystallize and cool. This calculated figure is much too high, requiring the complete crystallization and cooling to 500°C of a batholith more than 20 km thick under the rhyolite plateau, whereas even now at least part of the magma probably is still liquid (Eaton and others, 1975; Lehman and others, 1982; Smith and Braile, 1984; 1994). Therefore, either the present level of hydrothermal activity in the rhyolite plateau region is higher than the average over the last 600,000 years or a deeper source of energy supplies continuous heat to the rhyolitic magmatic system. The most likely interpretation, based on the volcanic history of the Yellowstone Plateau, is that the present high heat flow results directly from the intrusion to shallow levels of magma from deeper within the Yellowstone chamber during the past 160,000 years, and that rhyolitic magma in the large crustal chamber is maintained at high temperature by the continued supply of energy from below. It is argued subsequently that basaltic magma is the fundamental carrier of thermal energy in the Yellowstone Plateau magmatic system as a whole; the ultimate source of energy required to keep the rhyolitic magma liquid and convecting over such long times must be a continued injection of basaltic magma into the base of the rhyolitic system from a source region in the upper mantle. This conclusion essentially parallels arguments developed for voluminous rhyolitic magmatic systems in general by Smith (1979) and Hildreth (1981) and implies that both hydrothermal and episodic volcanic activity can be expected to remain vigorous indefinitely far in the future (Christiansen, 1984). Again, the high flux of ^{3}He (Craig and others, 1978; 1979) is consistent with a continued mantle contribution to the Yellowstone magmatic system. By any interpretation of the Yellowstone heat-flow data, a very large magmatic system, probably involving the entire thickness of the lithosphere and a substantial sublithospheric thickness, has been supplying heat to the convecting hydrothermal system.

THE FIRST VOLCANIC CYCLE

The youngest of the three volcanic cycles of the Yellowstone Plateau volcanic field has been discussed in the foregoing sections. In the following, that cycle is used as a model to interpret the remainder of the geologic history of the field from its beginning.

EARLY RHYOLITIC AND BASALTIC VOLCANISM

One rhyolitic lava flow is known to underlie the Huckleberry Ridge Tuff, forming Snake River Butte, north of Ashton, Idaho. The flow can be seen to underlie the Huckleberry Ridge Tuff along a railroad cut in the canyon of Warm River (pl. 2). The K-Ar age of the Snake River Butte flow is 1.99±0.02 Ma (table 2). Another flow, on the east side of Broad Creek in the wall of the younger Yellowstone caldera (pl. 1), earlier was also thought to underlie the Huckleberry Ridge (Christiansen and Blank, 1972, p. 10), but subsequent chemical, isotopic, and geochronological work (Hildreth and others, 1983; Obradovich, 1992) indicates that it was mismapped and is actually a downfaulted slice of the early third-cycle Wapiti Lake flow. A stage of regional tumescence and ring-fracture generation during the first cycle cannot be documented in the same way as for the third cycle, but a similar series of events probably led up to the gigantic Huckleberry Ridge ash-flow eruptions. By analogy with the third cycle, the rhyolite of Snake River Butte and perhaps other flows that now are buried may have vented along an incipient ring-fracture zone in the gently uplifted roof of the insurgent first-cycle rhyolitic magma body. The first-cycle ring-fracture zone probably extended from the Island Park area to well within the younger Yellowstone caldera, even east of the Central Plateau.

Because of cover by younger rocks, the amount of basaltic volcanism that occurred early in the first cycle is uncertain. Though the base of the Huckleberry Ridge Tuff is exposed widely around the margins of the field, the base is rarely exposed close to the caldera complex. Early basaltic rocks of the first cycle are exposed in two areas north of the Yellowstone Plateau: in the vicinity of the junction of Tower Creek with the Yellowstone River, and on Mount Everts east of Mammoth Hot Springs (pl. 1). These rocks are designated the Junction Butte Basalt (Christiansen and Blank, 1972). Many basalt flows of this sequence ponded low in a valley system that was cut into hilly topography; these flows are very thick—as much as 60 m—and both overlie and are interlayered with gravels. The most accessible of these flows is well exposed along the road north of Tower Falls, on the west side of The Narrows of the Grand Canyon of the Yellowstone. This flow, the Overhanging Cliff flow, (fig. 30) forms a very steep, high cliff above the road. A basal colonnade about 3 m thick is well developed where the flow overlies gravels and associated silt and sand. The thick overlying entablature has a complex network of crudely developed joint columns that intersect irregularly and form blocks a few centimeters wide; consequently, there is abundant small talus at the base of the cliffs. Two flows beneath the Huckleberry Ridge Tuff on Mount Everts are discontinuously exposed just northwest of Blacktail Pond (pl. l).

The Junction Butte Basalt generally is aphanitic with sparse to moderately abundant phenocrysts of labradorite. Olivine is abundant in the groundmass. The age of the Junction Butte is not much greater than that of the Huckleberry Ridge Tuff; limits of about 2 Ma and 2.4 Ma were inferred on the basis of paleomagnetic and stratigraphic relations to the Huckleberry Ridge by Christiansen and Blank (1972). Ages of about 2.0 Ma and 2.2 Ma have since been obtained by K-Ar dating of whole rocks from two flows (table 2).

THE HUCKLEBERRY RIDGE TUFF

As so much of the main eruptive area of the first volcanic cycle is buried by younger rocks, most of our knowledge of the cycle is inferred from stratigraphic and petrologic relations in the Huckleberry Ridge Tuff, the extensive ash-flow sheet erupted at the climax of that cycle.

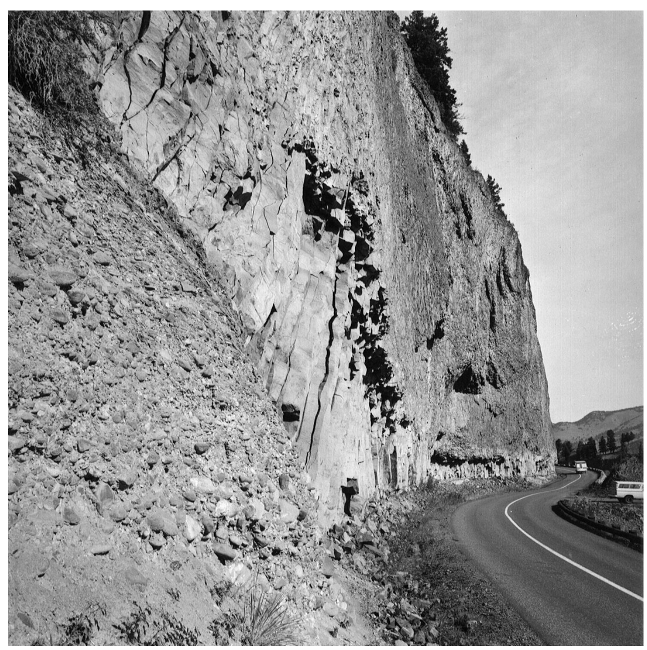


Figure 30.—Overhanging Cliff flow of Junction Butte Basalt at Overhanging Cliff, just north of Tower Falls, Yellowstone National Park. Underlying gravels, exposed on left, are devoid of welded-tuff clasts from the Yellowstone Group in contrast to other gravels nearby. Basal colonnade is about 3 m thick.

BEDDED ASH AND GENERAL RELATIONS

A basal fallout ash, lithologically similar to that at the base of member B of the Lava Creek Tuff, is present locally beneath the Huckleberry Ridge ash flows. The best exposure of this bedded ash (fig. 3l*A*) is on the cliffs of Mount Everts, locally overlying a gravel but also onlapping Cretaceous sandstones and shales. The underlying materials, both gravel and Cretaceous rocks, are reddened to depths of as much as 10 m below the base of the ash, evidently due to the effects of the relatively hot ash on sediments and sedimentary rocks that probably contained some water. The basal part of the ash, like the Lava Creek fallout deposit, is finer grained than the upper part. The upper part is also darker than the basal part, apparently because crystallites of magnetite have exsolved from the glass shards. Most of the beds in this deposit are perfectly tabular and continuous through all contiguous outcrops. The beds drape over undulations in the older topography with constant thickness, even where the ash deposit overrides the old gravel channel onto the underlying bedrock surface. These features, in addition to excellent sorting and lamination and sparse accretionary lapilli beds, indicate the fallout origin of the deposit. A few slightly channeled horizons are overlain by ash layers that pinch out laterally, and one zone a few centimeters thick shows microdune cross laminae. These particular ash layers may be minor ground-surge layers, but they more likely indicate brief episodes of minor reworking of the ash by wind.

The ash-flow tuff that succeeded the fallout ash was so hot that it not only welded densely to its own base but locally welded (or "fused") the top of the underlying ash as much as 15 cm below (fig. 31*B*). Boyd (1961, p. 395, 422-423) described this fused tuff and concluded that it was difficult to explain merely by static heating from higher-temperature ash flows emplaced above. He suggested that some small nonwelded ash flows might be present in the bedded-tuff sequence and that they might have helped maintain a higher temperature during emplacement of the bulk of the fallout. Such ash flows, however, should be recognizable, and none have been observed. The explanation probably lies at least partly in a higher initial emplacement temperature of the fallout ash than Boyd assumed. Clearly the ash mixed with air and cooled to some extent; the top is fine vitric ash, extremely depleted in crystals, and probably reflects settling of volcanic dust from the eruptive column during a brief lull before emplacement of the ash flows at this site, distant from the caldera. However, the bulk of the fallout deposit quickly formed its own insulating blanket. The thick baked zone below the fallout ash and the gradual upward color change due to exsolved magnetite both show that the tuff was not cool

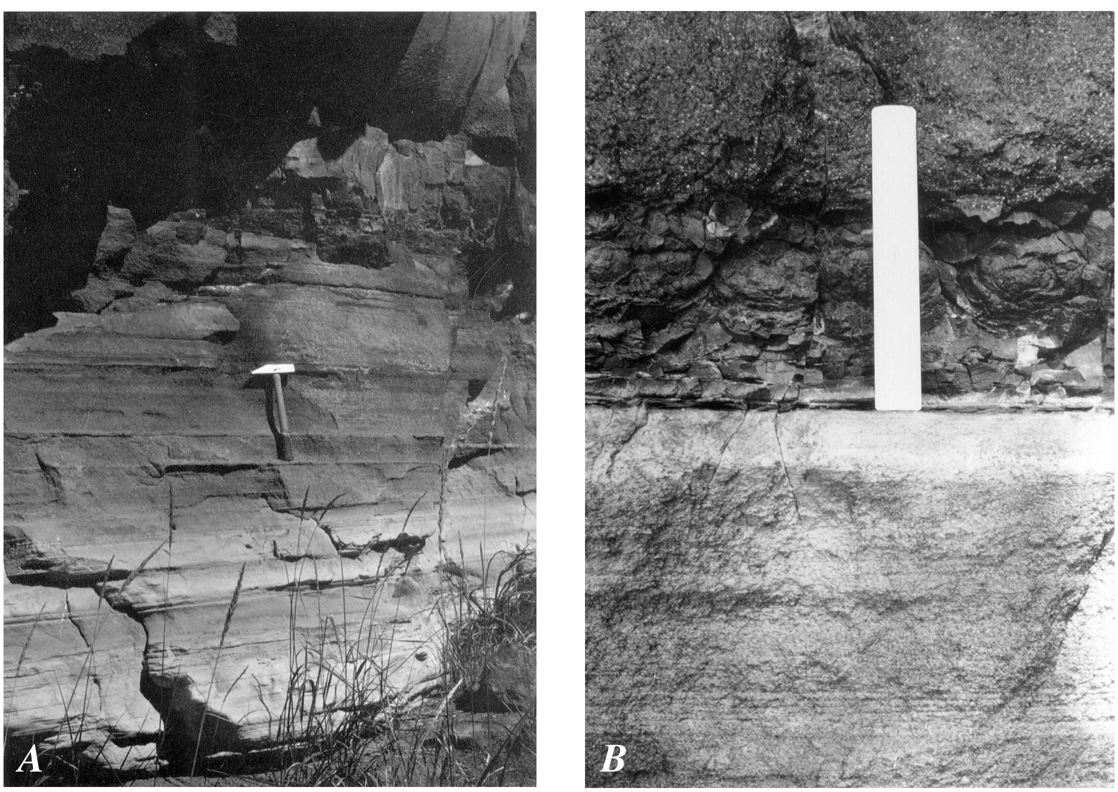


Figure 31.—Basal fallout ash of Huckleberry Ridge Tuff, Mount Everts, Yellowstone National Park. ***A***, continuous planar bedding of fallout tuff; lower, lighter colored part is relatively fine grained, darker upper part is coarser grained. ***B***, fused tuff at top of basal fallout, overlain by vitrophyric welded ash-flow tuff. Fallout-ash flow contact is about 7 cm below top of scale (20 cm long); base of scale is at base of fused zone.

before emplacement of the ash flows. Although the ash was below welding temperature, only a relatively small temperature increase would have been needed to weld the top. The fine vitric ash would have welded readily, and a temperature increase of perhaps 100°C is reasonably surmised, even by static heating from the overlying ash-flow tuff, which was well above welding temperature at the time of its emplacement.

The distribution of the Huckleberry Ridge fallout is quite irregular. The ash is missing beneath many exposures of the basal ash-flow tuff. The Mount Everts section is the thickest known, about $2^1/_2$ m. The ash must have been distributed widely by winds and has been recognized as far east of the Yellowstone region as southwestern Kansas (Naeser and others, 1973; R. L. Reynolds, 1979; Izett, 1981, Izett and Wilcox, 1982). Furthermore, ash transported through the Mississippi River system was deposited as thick ash-rich sediments in the Gulf of Mexico (Dobson and others, 1991). The basal fallout ash of the Huckleberry Ridge Tuff is distinguished from other bedded ashes of the Yellowstone Group by minor chemical and mineralogical differences, notably by the presence in it of pale-pinkish-colored zircon phenocrysts; basal fallout ashes of the Mesa Falls and Lava Creek Tuffs have only clear zircons (Izett, 1981). Naeser and others (1973) reported fission-track age determinations on zircons of the Huckleberry Ridge ash bed in Kansas of about 1.9 Ma, and R. L. Reynolds (1979), who showed that the distinctive horizontal paleomagnetic vector of the Huckleberry Ridge Tuff (table 2) is preserved in the distal ash beds, confirmed the correlation based on mineralogy and glass chemistry.

Ash flows of the Huckleberry Ridge Tuff are divided among three members (Christiansen and Blank, 1972), designated by Christiansen (1979) as members A, B, and C. All three members are present in the type section at Huckleberry Ridge about 2 km northeast of Flagg Ranch, just south of Yellowstone National Park (pl. 1). Members A and B are quite similar although member A has relatively fewer quartz and more plagioclase phenocrysts than member B. Those two members are separated in most sections by only a parting, where the ash-flow tuff is less densely welded than below or above. The contact is much like the one between members A and B of the Lava Creek Tuff; the members form a compound cooling unit in most sections. Member C is lithologically distinctive; it has notably smaller phenocrysts, has a higher ratio of quartz to feldspar phenocrysts than the other two members, and commonly is highly lineated and lithophysal. The contact between members B and C is everywhere sharp, and the top of member B is nonwelded to partially welded at the contact. Member C generally is welded to its base, and in most sections there is no glassy chilled zone at the contact. Thus, although there is a nearly complete break in the degree of welding in many sections, including the type section, the two members are parts of the same cooling unit. In upslope or more distal sections, however, as at Steamboat Mountain northeast of Jackson Lake, glassy bedded tuff is present at the base of member C. Thus, the Huckleberry Ridge Tuff is a composite sheet (Christiansen, 1979; terminology of Smith, 1960b).

MEMBER A

Member A is distributed mainly around the margins of the Yellowstone Plateau but extends to the north limit of Huckleberry Ridge Tuff exposures (pl. 1; fig. 32). Member A is the thickest part of the Huckleberry Ridge Tuff in the valleys of the Madison and Gallatin Rivers and northeastward around Mount Everts, the Blacktail Deer Plateau, and the Mirror Plateau. An initial area of 6,280 km^2 and a volume of more than 820 km^3 is represented by this present distribution.

The percentage of phenocrysts in member A changes markedly upward. This change is evident in the type section on Huckleberry Ridge and in the more easily accessible exposures at Grayling Creek (west of the Gallatin Range) and at Golden Gate (in northern Yellowstone National Park about 5 km southwest of Mammoth Hot Springs). Wherever observed, this change is found to progress upward from phenocryst-rich welded tuff at the base of the member to phenocryst-poor at the top. Although crystallized welded pumice lapilli are moderately abundant in the lower part of the member, they, like the phenocrysts, become sparse upward. As noted by Christiansen (1984), however, where rare welded pumice lapilli do occur in the phenocryst-poor rock, the pumice contains about the same percentage of phenocrysts as in the more crystal-rich lower part of the member. Thus, the observed upward decrease in phenocrysts relative to glass shards does not reflect a zonation of crystal abundance within the magma chamber but must be related to extreme sorting during eruption and emplacement, concentrating crystals (more dense) nearer the source and glass (less dense) farther away. The reason for this extreme sorting during a particular phase of the Huckleberry Ridge eruption is suggested by the extreme paucity of pumice lapilli in the tuff. The erupted magma during that phase probably was highly comminuted by the particularly violent release of contained gases. If the ejecta were nearly completely disrupted to ash-sized material, the pumice fragments would have largely disaggregated to their constituent phenocrysts and glass shards, enabling a high degree of sorting during fallback from the eruptive column and subsequent emplacement by flowage.

The source area for member A must be largely buried by younger rocks, but the same types of criteria used to relate other units of the Yellowstone Group to their specific source areas suggest that the member was erupted somewhere in the central part of the Yellowstone Plateau. That area is within the member's center of distribution, the thickest sections lie radially outward from it, and the densest welding, largest

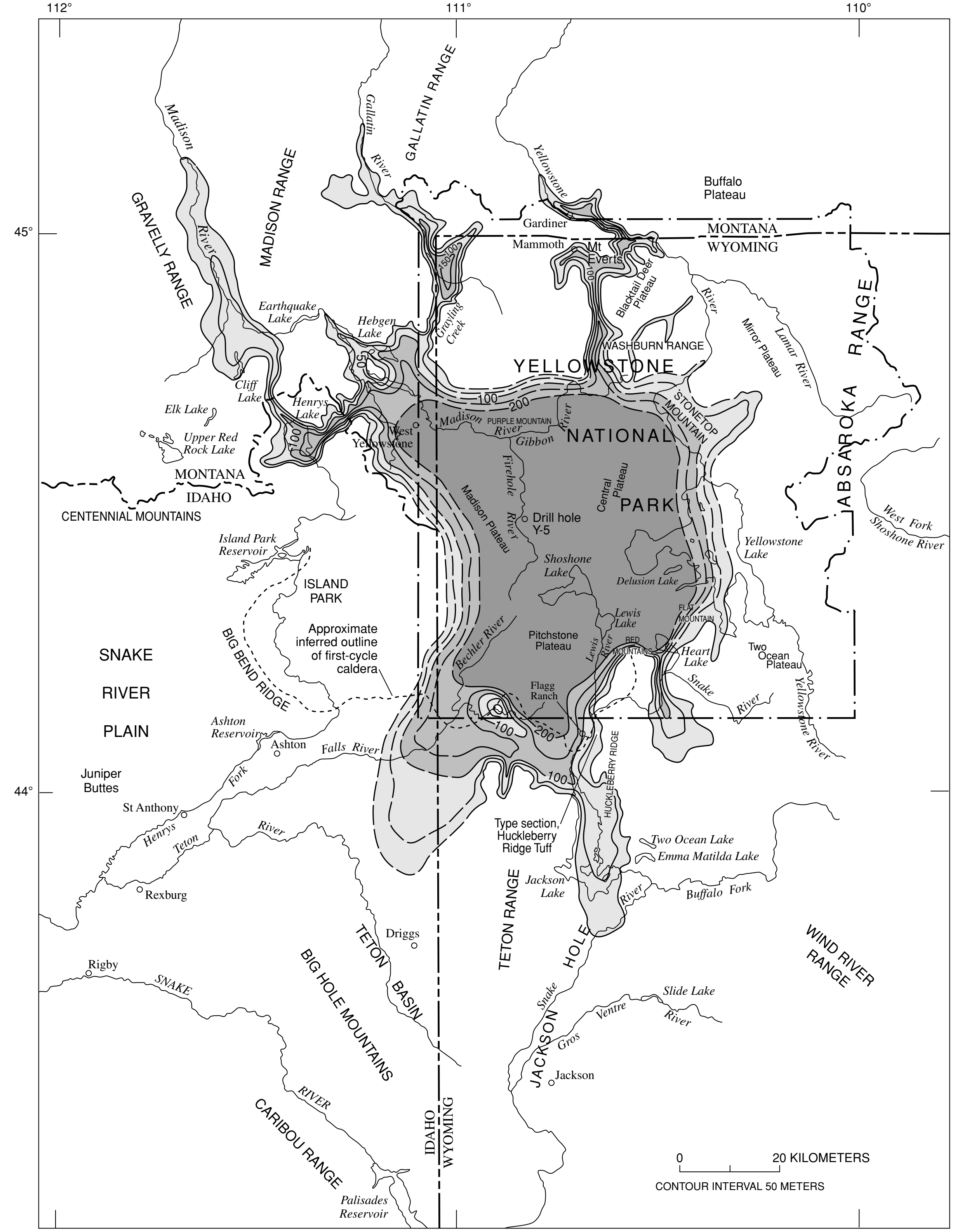


Figure 32.—Areal distribution and thickness of member A, Huckleberry Ridge Tuff, in relation to topographic features in the Yellowstone Plateau volcanic field that affected its distribution. Member A filled erosional topography of high relief, mainly around the margins of the Yellowstone Plateau, and is the thickest part of the exposed section in the valleys of the Madison and Gallatin Rivers and around Mount Everts, the Blacktail Deer Plateau, and the Mirror Plateau. The calculated initial area is 6,280 km^2 and initial volume is 820 km^3. Shading is darker for thicker sections. Isopachs in meters, dashed where inferred.

phenocrysts, largest lithic inclusions, and related features lie around it. The source area for this member probably has been almost completely downdropped within the area of the younger Yellowstone caldera.

MEMBER B

The base of member B of the Huckleberry Ridge Tuff is mapped at a zone of less dense welding than in the tuff either below or above it. Only one such partial cooling break occurs in most sections of members A and B; its consistent stratigraphic level is demonstrated by the vertical phenocryst changes of both members A and B. Member B is relatively phenocryst-poor at its base, similar to the upper part of member A, but this lithology grades upward into progressively more phenocryst-rich rocks. The phenocryst-poor basal part of member B is thinner in most sections than the phenocryst-poor upper part of member A. Welded pumice inclusions in member B contain phenocrysts in proportion to those of the whole rocks, suggesting that in this member, unlike in member A, the variation in the ratio of phenocrysts to glass particles did result from compositional zonation in the erupted magma. The phenocrysts in the relatively phenocryst-rich upper part of member B commonly are even larger than those in member A, large pinkish quartz phenocrysts being especially notable. The uppermost part of the phenocryst-rich tuff of member B contains two pumice types of contrasting compositions. Dark-gray crystallized pumice in the devitrified rocks, which weathers brown, contrasts sharply with white crystallized pumice in the same rocks. Vapor-phase minerals invariably occur in the dark crystallized pumice, including a brownish acicular amphibole that has not been noted in any other rhyolites of this field. Because this is the only two-pumice zone known in either the Huckleberry Ridge or the Lava Creek Tuff, it is an especially useful lithologic marker for mapping.

Member B is present over much of the area of the Huckleberry Ridge Tuff (pls. 1, 2, fig. 33). It is thickest—more than 150 m—around Big Bend Ridge and in nearby areas to the southeast. North of Island Park, around the Madison Valley near West Yellowstone, and into the drainages of Grayling Creek and the Gallatin River, the thickness changes little, remaining about 100-150 m. Member B thins eastward from the Gallatin River area into northern Yellowstone National Park; it also thins eastward from the Island Park area, around the northern Teton Range, to the type section of the Huckleberry Ridge Tuff. Member B has a calculated initial area of 15,400 km^2 and a volume of 1,340 km^3.

In almost all exposures in the northern part of the area, member B lies on member A, only locally onlapping the higher parts of preexisting topography. One such onlap is in the southern Madison Range, about 6 km east-southeast of Targhee Peak (pl. 3; about 3 km south-southeast of Lionhead Peak on the Targhee Pass 1:24,000 topographic quadrangle map). There, member B oversteps member A and forms a vitrophyric welded tuff. The two-pumice zone of member B is particularly noteworthy at this locality; unlike most other areas, the dark pumice inclusions are not welded or devitrified here but are black, somewhat scoriaceous glass. The usually light-colored pumice at this locality forms a dense black glass in which nonhydrated obsidian spheres as much as 1 cm in diameter weather out of a perlitic matrix. South of the Island Park area, member B is commonly the only part present in exposures of the Huckleberry Ridge Tuff.

As member B thickens westward from the northern Teton Range toward Big Bend Ridge, it develops a conspicuous lineation. On Huckleberry Ridge it shows no visible lineation, but in the area around Lake of the Woods, Calf Creek, and North Boone Creek (pl. 1) in the northern Tetons a slight lineation of stretched pumice becomes perceptible. Farther west, the lineation becomes progressively more pronounced, and wavy low-amplitude broad-wavelength folds of the welded-tuff foliation with axes parallel to the pumice lineation become apparent. These features are quite marked along the Cave Falls Road at Porcupine Creek and its tributary Rising Creek. On the south flank of the Island Park area, member B is very densely welded, the broad wavy folds are quite conspicuous, and the pumice lineation is so marked that some outcrops look more like rhyolitic lavas than like typical welded tuffs. One of the most aberrant of these unusual sections is that exposed beneath the Mesa Falls Tuff along the highway north of Ashton, Idaho. The lineation and the axes of the folds are generally circumferential to the first-cycle caldera rim on Big Bend Ridge.

Lateral variations in member B, including the thickness changes and lineations just noted, indicate that the source area of the member was in the Island Park area. The member is present at the south rim, and the strong lineation and folding in member B on the outer flank indicate that the tuff must have been very hot in that area and was deformed to some degree as it was deposited and welded.

From a variety of unusual features, it was suggested earlier (Christiansen, 1982) that part of member B of the Huckleberry Ridge Tuff erupted into a lake that lay on the margin of the Snake River Plain near the foot of Big Bend Ridge. On the ridge, member B is very densely welded, strongly lineated, and entirely devitrified at exposed levels. At the south foot of the ridge, such as along the road across Henrys Fork below Ashton Reservoir (pl. 2), the tuff becomes glassy and spherulitic. The welding generally remains dense, but small open spaces are rather common in layers parallel to the foliation. The foliation becomes intensely and irregularly deformed and locally can have any attitude. Certain layers parallel to foliation are devitrified, especially where the foliation is nearly vertical, although other layers remain glassy. Lineation disappears as a conspicuous feature in these glassy rocks. The dark scoriaceous pumice generally characteristic of the upper part of the member is clayey, greenish, and weathers easily. South of Ashton and in the canyon of the Teton River, fine-grained

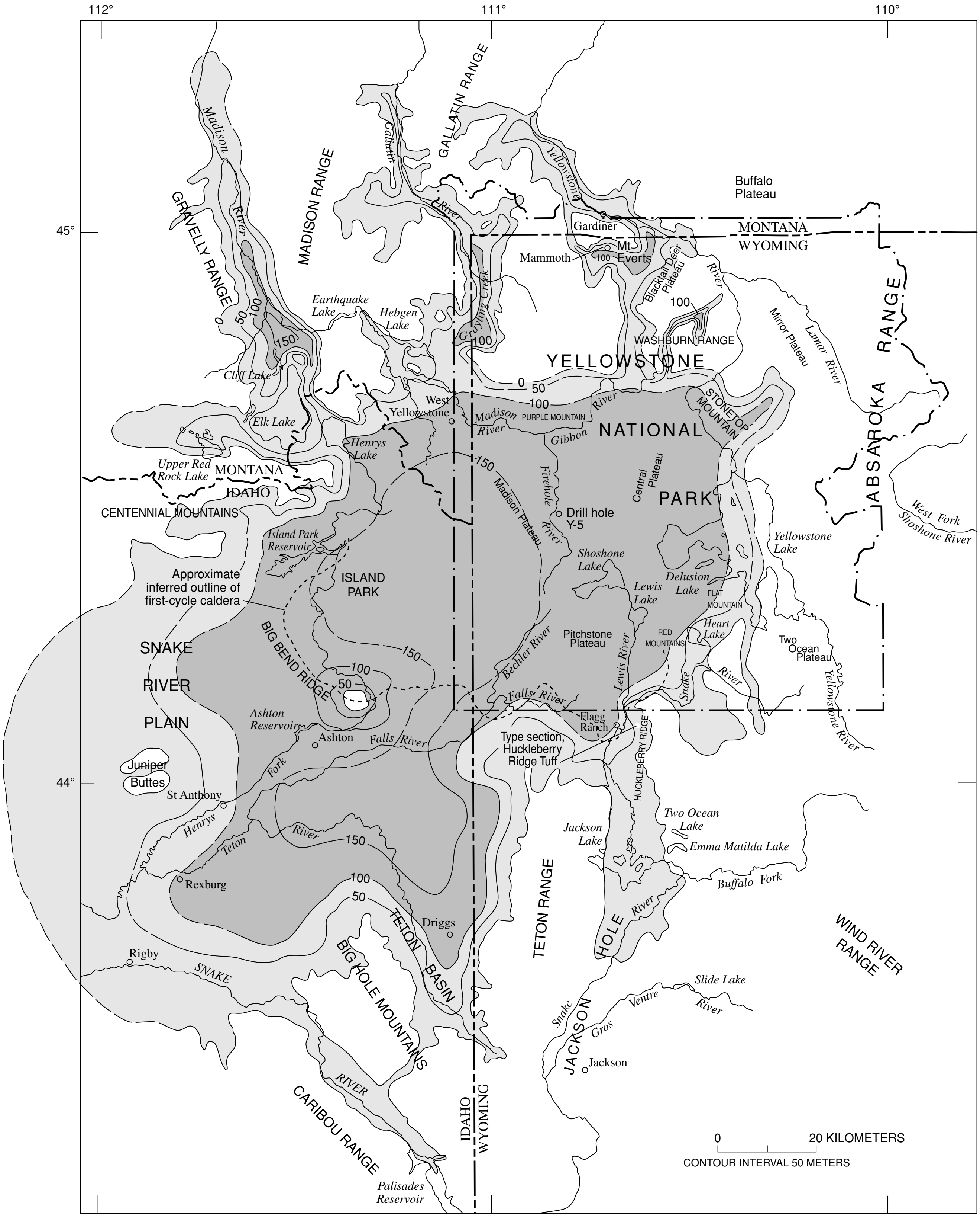


Figure 33.—Areal distribution and thickness of member B, Huckleberry Ridge Tuff, in relation to topographic features in the Yellowstone Plateau volcanic field that affected its distribution. Member B spread widely on surfaces blanketed by older ash flows. The calculated initial area is 15,400 km^2 and initial volume is 1,340 km^3. Shading is darker for thicker sections. Isopachs in meters, dashed where inferred.

clastic sediments that underlie the member along an irregular contact locally invade the base of the ash-flow sheet. The deformational features and anomalous crystallization generally decrease southward toward the south edge of the map area of plate 2. The unit, however, remains predominantly vitrophyric to within 1/2 km north of the Teton River (pl. 3); in an area near there, only small pipelike masses that weather into knobs are devitrified or partially devitrified. Just south of there, the canyon of the Teton River is carved 150 m into more or less normal-looking devitrified welded tuff of member B although it has numerous large cavities lined with lithophysal minerals (Prostka, 1977).

Based upon the features just described, the hypothesis was proposed (Christiansen, 1982) that member B erupted from the area north of Big Bend Ridge, flowed down a steep slope as it welded, and entered a lake, chilling quickly but deforming plastically where underlying sediments on the lake floor slumped or compacted irregularly under the ash flow. The cavities may have formed from trapped steam bubbles, generated by the hot ash flows in contact with lake water. The devitrified bands and pipes could have resulted from release of steam along steeply inclined layering or above pockets of steam trapped below the welded tuff. Because the Huckleberry Ridge Tuff south of the canyon of the Teton River is essentially normal in appearance, it seems likely either that the south shore of the lake was located just south of the present canyon or that the ash flows had completely displaced the lake water by the time they reached that area.

MEMBER C

Member C of the Huckleberry Ridge Tuff is present only south of the Yellowstone Plateau in the area from the northern Teton Range to Flat Mountain and from the Red Mountains to Jackson Hole (pl. 1; fig. 34). It is the smallest distinct ash-flow subsheet of the Huckleberry Ridge Tuff, having an inferred initial area of 3,690 km^2 and a volume of at least 290 km^3, although parts of it probably are buried under the rhyolite plateau. Some of the thicker sections are in the vicinity of the type section of the Huckleberry Ridge Tuff, where the member is nearly 100 m thick. The greatest thickness of the member, however, is in the Red Mountains, where the ash flows appear to have filled a major topographic depression. The true stratigraphic thickness of the member is difficult to determine in the western and northern Red Mountains because the lineation of stretched pumice is extreme and the unit is draped over the edge of the buried depression. Exposed sections of as much as 430 m are present in that area. Wherever the base of the member is exposed, it overlies member B of the Huckleberry Ridge.

Marked lineation of stretched pumice and abundant open-space crystallization, both as spherical lithophysae and as linings of elongate cavities formed by stretched pumices, are characteristic of member C. The lineation is notable throughout the area of the member but is extreme in the Red Mountains and northern Huckleberry Ridge. In the Red Mountains, where the unit is draped over steep slopes and fills a paleodepression, the lineation is directly downdip. Elsewhere, the lineation generally is oriented radially away from the Red Mountains area.

The general distribution, orientations, and degree of development of lineations, the areas of densest welding, and slight lateral variations in phenocryst content indicate that the source area of member C is in the vicinity of the Red Mountains and the south entrance to Yellowstone National Park. Also, in areas relatively near the source, member C is densely welded and lineated to its base. In areas farther away, a nonlineated partially welded zone one to a few meters thick at the base grades upward into lineated more densely welded tuff.

AGE AND REMANENT MAGNETISM

Sanidines separated from each of the three ash-flow members and basal fallout ash of the Huckleberry Ridge Tuff gave concordant K-Ar ages of 2.0 Ma (Obradovich, 1992), just older than the Pliocene-Pleistocene boundary (Berggren, 1972; Obradovich and others, 1982). More recent dating by the $^{40}Ar/^{39}Ar$ laser-fusion technique yields an age of 2.053±0.096 Ma (M.A. Lanphere, 2000, written communication).

Remanent magnetic polarity of the Huckleberry Ridge Tuff is distinctively subhorizontal. Numerous field and laboratory measurements show that the basal vitrophyre of member A at most localities has a stable remanent magnetic vector of low intensity, nearly horizontal inclination, and southwesterly declination (table 2). R. L. Reynolds (1977; 1979), who measured both the welded ash flows and the friable bedded ash of member A, showed that their directions of stable remanent magnetization are closely similar. The fused fallout tuff just below the ash flows of member A on Mount Everts also has the same remanent magnetic direction as the welded ash flows. Both the fallout ash beneath the welded tuff of member A on Mount Everts (fig. 31) and the redeposited ash beds in southwestern Kansas that were derived from the Huckleberry Ridge Tuff (Naeser and others, 1971; Izett, 1981; Izett and Wilcox, 1982) show the same remanent magnetic direction (R. L. Reynolds 1979). Thus, the close similarity of the thermoremanent and depositional remanent magnetization in these deposits indicate that the magnetization records the geomagnetic field direction at the time of the Huckleberry Ridge eruptions.

Similar subhorizontal southwesterly directions of remanent magnetization can be preserved in the devitrified portions of all three members of the Huckleberry Ridge Tuff. These directions, however, even after AC magnetic cleaning, lack the consistency of those of the basal glassy tuffs. Field measurements alone commonly yield anomalous results, frequently indicating normal polarities. The reasons

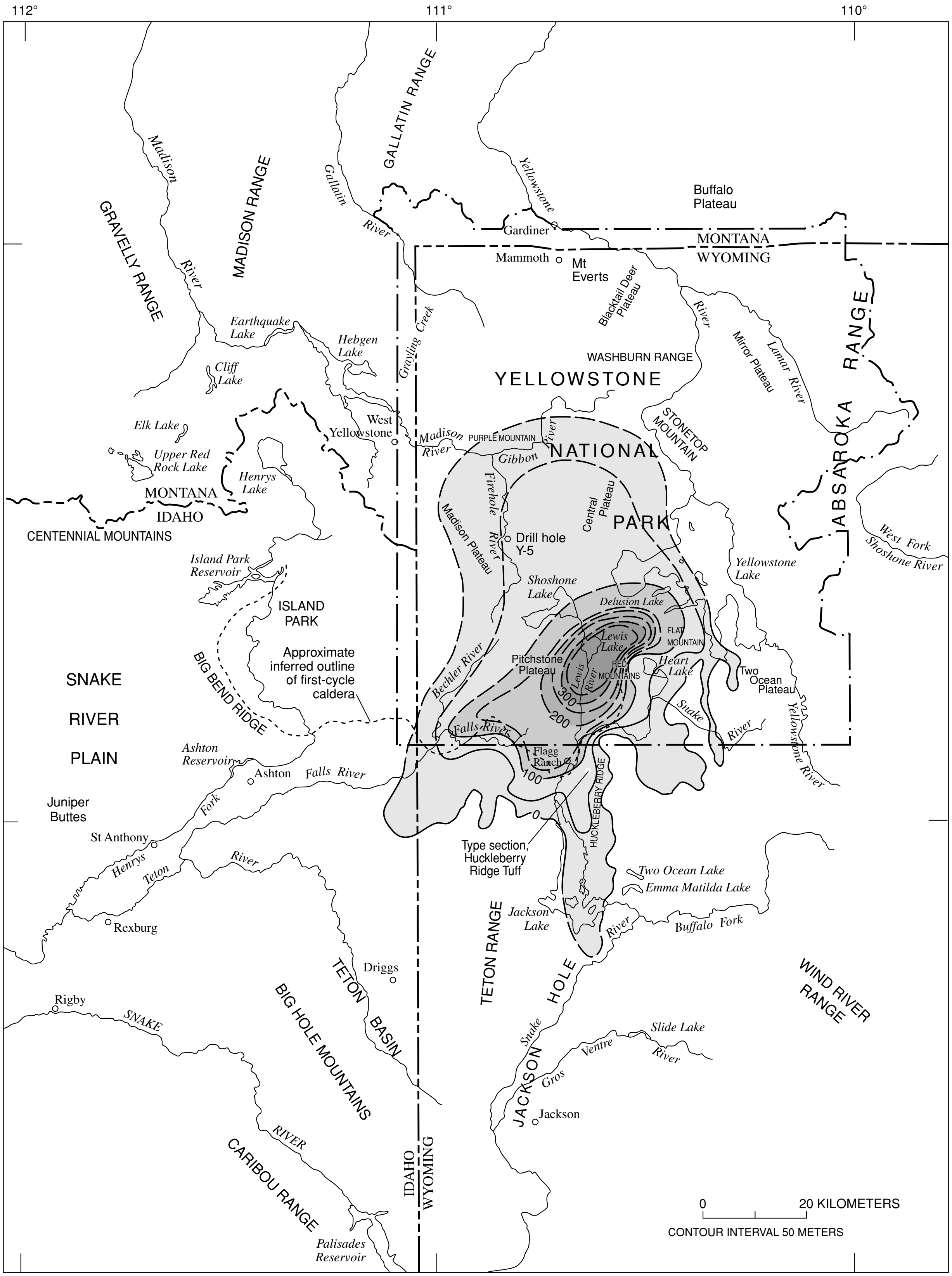


Figure 34.—Areal distribution and thickness of member C, Huckleberry Ridge Tuff, in relation to topographic features in the Yellowstone Plateau volcanic field that affected its distribution. Member C is exposed only near the south margin of the Yellowstone Plateau. The calculated initial area is 3,690 km^2 and initial volume is 290 km^3. Shading is darker for thicker sections. Isopachs in meters, dashed where inferred.

for these inconsistencies in remanent magnetization of the Huckleberry Ridge cooling unit were analyzed by Reynolds (1977) and were shown to be related to the weak transitional geomagnetic field in which the Huckleberry Ridge Tuff cooled. Parts of the unit have been remagnetized during recrystallization of the magnetic phase, particularly with the formation of hematite in a ground-water environment. This chemical magnetization commonly has overprinted the weak thermoremanent magnetization, particularly in devitrified tuffs. These complex relations of remanent magnetization of the Huckleberry Ridge Tuff are noted here only because they have proved to be significant in field correlation and mapping of the Huckleberry Ridge Tuff and all three of its members.

THE FIRST-CYCLE CALDERA

Evidence just reviewed indicates that the three members of the Huckleberry Ridge Tuff—although they were emplaced within an extremely short period of time, have no erosional breaks, and for the most part form a single compound cooling unit—had distinct source areas. An analogy is immediately suggested to the Lava Creek Tuff, each member of which was related to a distinct source area where magma erupted in sequence and which collapsed to form a separate caldera segment. Although physical evidence for three separate segments of a first-cycle caldera is tenuous, it is consistent with a concept, like that developed for the third cycle, in which the three members represent the eruption of three major high-level culminations of a magma body that was connected at depth, the eruption of one area triggering eruption of the next.

Little is seen of any caldera-collapse structure that formed with the eruption of member A of the Huckleberry Ridge Tuff. Analogy with all other units of similar type and size in this volcanic field (and in other fields where the source areas of comparable units are known; Smith, 1960a, p. 819; 1979, fig. 2; Spera and Crisp, 1981) indicates that such a caldera certainly did form. That member A was not erupted from the same vents as member B is shown by the observation that each of these members thins and becomes poorer in phenocrysts in a complementary way as it is traced toward the center of distribution of the other. The collapse structure related to eruption of member A probably is now largely buried within the Yellowstone Plateau. The depression partly filled by member C in the Red Mountains, however, may be part of the member A caldera. Although member C in the Red Mountains may also in part fill a caldera related to its own eruption, such thick sections draped over the wall are not characteristic elsewhere of the rims of the caldera segment related to member C. Thus, the small Red Mountains caldera segment may have existed before the eruption of member C and be part of the collapse area related to member A.

The source area of member B of the Huckleberry Ridge Tuff was in the vicinity of Island Park. Big Bend Ridge, the south rim of the Island Park basin, mainly consists of a thick densely welded section of member B, draped only by a thin carapace of Mesa Falls Tuff. Geologic mapping of the Island Park area (pl. 2 ; Christiansen, 1982) shows that the south wall of the caldera formed during or after the eruption of member B and was a virtual barrier to later emplacement of the Mesa Falls ash flows. The Snake River Butte rhyolitic lava flow was erupted north of this rim before eruption of the Huckleberry Ridge Tuff and associated caldera collapse. Thus, it seems clear that collapse of a magma-chamber roof related to the eruption of member B formed a caldera, part of which is preserved as the south rim of Island Park. This part of the first-cycle caldera was earlier designated as the Big Bend Ridge segment (Christiansen, 1982). Later collapse related to the Mesa Falls Tuff (more fully described in the section on the second volcanic cycle), which occurred in part of the same area, rejuvenated caldera faults along this south rim with small displacements. Only a part of the collapse area related to member B is exposed. Most of the rest of the Big Bend Ridge caldera segment is buried, and its northern and eastern extent are not known.

The marked circumferential lineation and low-amplitude folding of member B on the outer flank of the Big Bend Ridge caldera segment were described in the previous section. Anticlinal axes of the folds on that flank commonly are broken by small ramp faults that cause the upslope part to override the downslope part. These fold axes and associated fractures, which are breached by erosion, display a remarkable concentric pattern around the caldera. They are most conspicuous on the southwest flank, in the lee of a major dune belt, where sandblasting has etched these structural features into relief and where lack of vegetation results in relatively good exposure (fig. 35). The lineation of elongate pumice inclusions parallel to fold axes, and the vapor-phase crystallization in cavities parallel to the lineation and foliation show that the welding, folding, and lineation of these rocks occurred essentially simultaneously and that whatever accounts for the folds and lineation occurred during the eruption, early consolidation, and high-temperature crystallization of member B. The looped pattern of roughly concentric structures shown in figure 35 suggests flowage down the outer flank of the caldera structure during welding of the ash-flow tuffs. The following origin for these features is proposed: the insurgence of rhyolitic magma into a high-level chamber was effectively terminated as an evolving ring-fracture zone tapped the rising magma to erupt member B of the Huckleberry Ridge. Tumescence above the rising magma body had formed a broad domical uplift, the growth of which also was cut short by the eruptions and concurrent or immediately subsequent caldera collapse. The near-source Huckleberry Ridge ash flows probably were very hot and began to weld even as they came to rest on the flanks of this broad dome. The hot, welding ash-flow tuffs appear to have slumped off the flank of this uplift like icing piled too thickly on a warm cake.

About half of a circular fault depression that formed during or after the eruption of member C can be recognized from the south edge of the Red Mountains (where an especially thick section west of Red Creek that draped over an arcuate wall has already been described), along the Snake River below Red Creek to the vicinity of Flagg Ranch, up Glade Creek, through Grassy Lake Reservoir, and across Falls River (pl. 1). Member C, although absent from most sectors of the Huckleberry Ridge outflow sheet, is thick, densely welded, and strongly lineated to its base along this partial rim, which is here designated as the Snake River segment of the first-cycle caldera. Thus, although the ash flows of member C were blocked in other directions by already-formed segments of the first cycle caldera wall, they were not de-

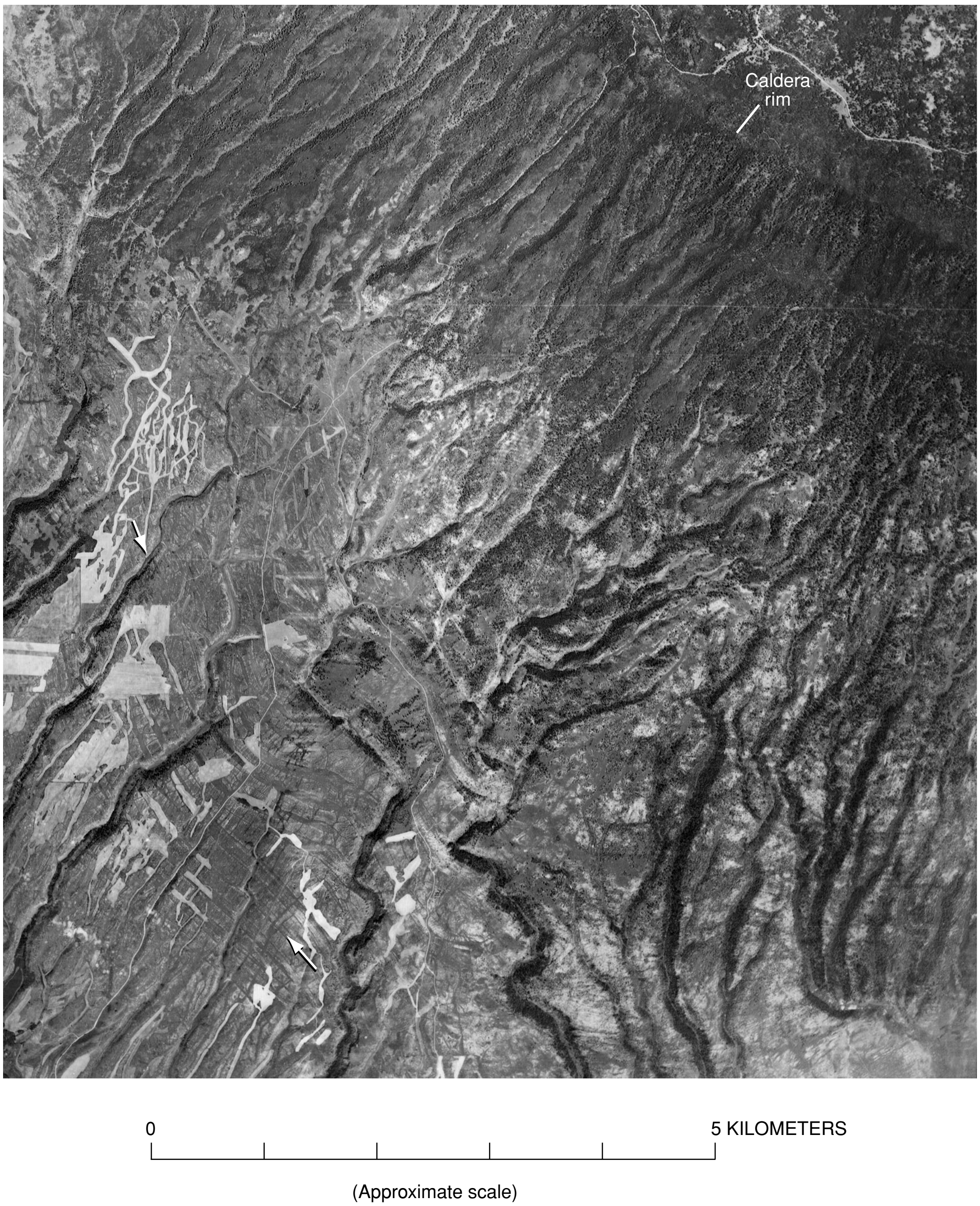


Figure 35.—South outer flank of the Big Bend Ridge caldera segment, showing sandblasted and erosionally breached looped folds and ramp faults that parallel lineation of member A, Huckleberry Ridge Tuff. Arrows indicate examples. North is toward top of photograph. Vertical aerial photograph, U.S. Army Map Service Project No. 121, roll 26, negative No. 4775.

flected by the Snake River segment; the vertical eruption column from which the ash flows of member C were derived probably erupted through fissures near the wall of the Snake River caldera segment. Furthermore, even if—as suggested earlier—the Red Mountains caldera segment is part of the collapsed area related to eruption of member A, the eruptive fissures for Member C must have been situated near the base of the Red Mountains wall in order to explain the character of member C—thick, highly lineated, and draped all the way from the rim to the base of the caldera wall. Thus, the subsidence of the Snake River caldera segment probably resulted directly from the eruption of member C, and the Red Mountains segment subsided further as it was partially filled during that eruptive episode. The Snake River segment was later filled by lavas of the Lewis Canyon Rhyolite and by the Lava Creek Tuff.

Figure 36 is my reconstruction of the first-cycle caldera of the Yellowstone Plateau volcanic field, based on evidence just outlined and on the model of the third-cycle caldera (also see fig. 7). It is pictured as having consisted of three collapse segments which overlapped to form a single topographic basin. Each segment presumably was related initially to a separate ring-fracture zone that formed over a high-level culmination of the magma chamber. Rapid eruption of all three members produced a single composite ash-flow sheet, indicating that collapse of all three caldera segments must have occurred in immediate succession. I infer (as for the third cycle) that, because of the rise of magma to high crustal levels, tumescence, and ring-fracturing of the chamber roofs were essentially simultaneous and that all three high-level intrusions were connected at depth. Catastrophic eruption of the first subsheet (member A) probably triggered eruption of the second, which in turn triggered eruption of the third

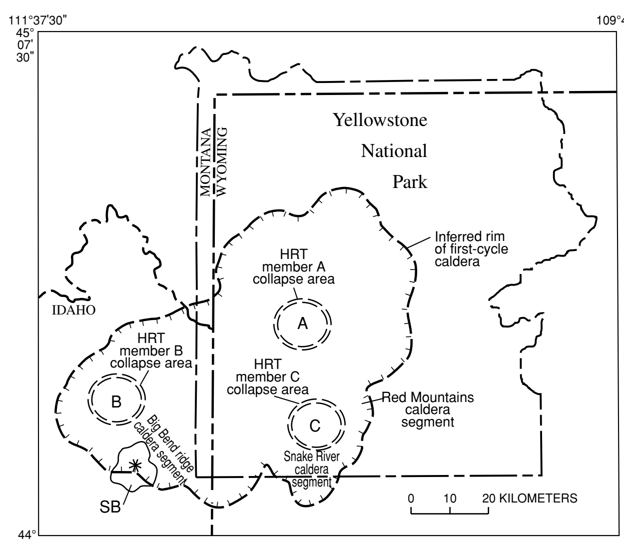


Figure 36.—Reconstructed map of the first-cycle caldera and location of the single known precaldera rhyolitic lava flow (SB) related to it. HRT, Huckleberry Ridge Tuff; SB, Snake River Butte flow; asterisk, probable vent for flow.

subsheet. It is not known whether any of the first-cycle caldera segments was resurgent.

POSTCOLLAPSE RHYOLITIC VOLCANISM

Four rhyolitic lava flows—the Blue Creek, Headquarters, Bishop Mountain and Green Canyon flows—at the north end of Big Bend Ridge were previously designated as the Big Bend Ridge Rhyolite (Christiansen, 1982). In addition, further work shows that Moonshine Mountain and a local ash-flow tuff near Lyle Spring are closely related chemically, isotopically, and mineralogically to the Bishop Mountain flow. (The tuff was shown incorrectly on the geologic map of Christiansen, 1982, as Mesa Falls Tuff.)

As is typical of rhyolitic lavas, the thicknesses of the flows vary greatly; the exposed thickness of what appears to be a single flow is about 375 m on the caldera scarp at the base of Bishop Mountain. Most of the flow surfaces are eroded so that little of their original glassy shells remain, but spherulitic zones are preserved widely. All four of the named flows are shown by mapping (pl. 2) to overlie the Huckleberry Ridge Tuff and to underlie the Mesa Falls Tuff on the rim west of Island Park although Hamilton (1965) mapped the Bishop Mountain and Green Canyon flows as younger than all the ash flows on the rim of what he called the Island Park caldera. The Big Bend Ridge flows contain moderately abundant phenocrysts of quartz, sanidine, and plagioclase. Mafic silicate minerals in the crystallized rocks commonly are altered to iron-oxide minerals; ferroaugite is the predominant mafic mineral in most of the glassy or partly glassy rocks. However, the Bishop Mountain and Green Canyon flows, the tuff of Lyle Spring, and Moonshine Mountain (pl. 2) contain small phenocrysts of biotite, the only rhyolites of the Yellowstone Plateau volcanic field known to carry this phenocryst phase. In addition, as discussed later, this group of rhyolites near Bishop Mountain is chemically distinct from and considerably younger than the Blue Creek and Headquarters flows (tables 2 and 14).

The topographic high points of each flow are near the rim of the Big Bend Ridge caldera segment, and each flow was truncated by subsequent second-cycle collapse. Therefore, these high points do not represent the vent areas, as Hamilton (1965) supposed in outlining his concept of an Island Park caldera. The actual source vents for the flows are not known, but they are presumed to be downdropped within the ring-fracture zone of the second-cycle Henrys Fork caldera. Moonshine Mountain (pl. 2) might be the downfaulted vent dome of the Bishop Mountain flow.

It was difficult to decide from field evidence alone whether the flows of the Big Bend Ridge Rhyolite should be regarded as postcollapse first-cycle lavas or precollapse second-cycle lavas. Detailed studies, however, help to clarify their magmatic affinities. In particular, the ages of about 1.8 Ma on the Blue Creek and Headquarters flows (table 2) and their very low δO^{18} values (Hildreth and others, 1984) suggest a much closer magmatic affinity to the underlying Huckleberry Ridge Tuff than to the overlying Mesa Falls Tuff. These two southern flows of the Big Bend Ridge Rhyolite are regarded in this paper as representing postcollapse lavas of the first volcanic cycle. The younger (about 1.3-Ma) biotite-bearing and chemically more evolved (table 14) Bishop Mountain and Green Canyon flows and tuff of Lyle Spring may represent a partially isolated, highly evolved local magma chamber that erupted at about the time of the second cycle. Other postcollapse rhyolites of the first cycle may lie buried within the younger calderas of the volcanic field.

THE SECOND VOLCANIC CYCLE

The second cycle seems to have been the smallest and simplest of the three major cycles of the Yellowstone Plateau volcanic field, but it nevertheless displays a sequence of events similar to the other cycles.

EARLY RHYOLITIC VOLCANISM

As noted earlier, a local biotitic ash-flow tuff and two of the four post-Huckleberry Ridge rhyolitic lava flows and a small associated ash-flow tuff on the west rim of Island Park near Bishop Mountain erupted at about the time of the second volcanic cycle but may actually represent a highly evolved, at least partly isolated magma chamber. In addition, other intracaldera rhyolites of the second cycle, could lie buried within the Island Park area.

THE MESA FALLS TUFF

The Mesa Falls Tuff, although having less volume and covering less area than either the Huckleberry Ridge or Lava Creek Tuffs, is a major ash-flow sheet. It occurs within an area of nearly 2,700 km^2 in and near Island Park (pl. 2; fig. 37). The thickest known section of the formation, about 150 m, occurs on Thurmon Ridge, the north rim of Island Park. Elsewhere it commonly is about 30-70 m thick. The distribution outlined in figure 37 represents an initial volume of more than 280 km^3, assuming the presence of Mesa Falls Tuff within the first-cycle caldera at about its average thickness. Possibly, however, it is ponded much more deeply there. The Mesa Falls Tuff overlies the Huckleberry Ridge Tuff unconformably in many places, especially in the southern part of the Island Park area (pl. 2). The only intervening units on which the Mesa Falls has been seen to lie locally are the Big Bend Ridge Rhyolite and a wedge of loess in roadcuts and quarry exposures about 5 km north of Ashton (Hamilton, 1965, fig. 3). The Lava Creek Tuff overlies the Mesa Falls unconformably at many places, but southeast and northwest of Island Park the basalts of Warm River and Shotgun Valley, respectively, lie between them.

Lithologically the Mesa Falls Tuff is the most distinctive of the three ash-flow sheets of the Yellowstone Group, except for certain subunits of the Huckleberry Ridge and Lava Creek Tuffs. The Mesa Falls generally forms large rounded boulderlike outcrops, commonly pinkish in color. It has abundant very large phenocrysts, particularly the unbroken sanidines which are as large as 2-3 cm in pumice inclusions. Parts of the other two ash-flow sheets also have conspicuously large sanidine crystals, but they generally are less than 1 cm, and most are only a few millimeters across. Vertically throughout the Mesa Falls, the phenocrysts are large and conspicuous. The phenocrysts include quartz, sanidine, and plagioclase (about An_{15}), ferroaugite, magnetite, ilmenite, fayalite, and accessory hornblende, zircon, chevkinite, and allanite. Large welded pumice inclusions, commonly as much as 10 cm but locally more than 30 cm long, are abundant in the Mesa Falls. Lithic inclusions are not abundant, but where present include welded-tuff lithologies probably derived from the Huckleberry Ridge Tuff. This criterion is sometimes useful in mapping because welded-tuff xenoliths are relatively rare in the Huckleberry Ridge, the unit most easily confused with the Mesa Falls Tuff.

A thick-bedded ash and pumice deposit at the base of the Mesa Falls Tuff is well exposed in roadcuts along the highway on the south outer flank of Island Park and in nearby gravel pits (see, for example, Neace, 1986). This deposit generally is about 5 m thick in these exposures and directly underlies the Mesa Falls ash flows. The deposit is white and very well sorted; generally it has good planar bedding and contains abundant crystals of the same types present in the succeeding ash-flow tuff. As was noted also for both the Huckleberry Ridge and Lava Creek fallout deposits, about the lower half is generally finer grained than the upper part. In the Mesa Falls the lower part contains a marked concentration of crystals over glass. The upper, coarser part of the deposit near Island Park consists predominantly of pumice lapilli about a centimeter long. The bedding is parallel to the underlying slope and mantles irregularities of the buried surface with a nearly uniform thickness of ash and pumice. All the criteria reviewed earlier for a fallout origin of the deposits under the other two ash-flow sheets apply equally to at least most of the Mesa Falls bedded material. However, near the top are a few channeled and cross-stratified units that may have been deposited by ground-surges (Parsons and Doherty, 1974; Doherty, 1976; Neace, 1986; Neace and others, 1986). Also interlayered in this upper part of the bedded ash are a few lenticular ash flows as much as 50 cm thick that are separate from the main body of ash flows above. The glass chemistry of the bedded deposit as well as the characteristics of some of its included crystals are distinctive from those of either the Huckleberry Ridge or Lava Creek bedded ashes (Izett, 1981), and they are recognized eastward in middle Pleistocene deposits of the southern Rocky Moun-

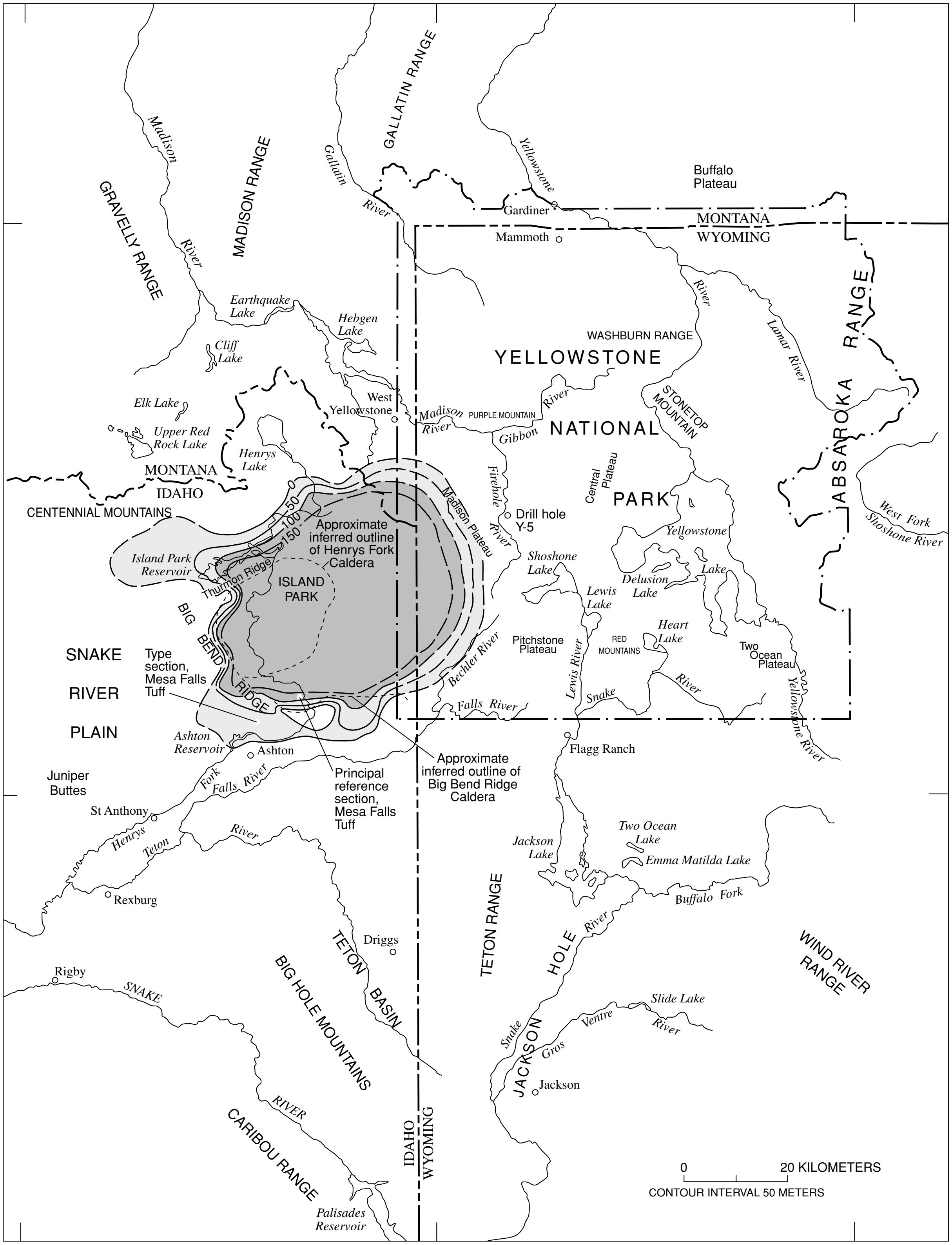


Figure 37.—Areal distribution and thickness of the Mesa Falls Tuff, in relation to topographic features in the Yellowstone Plateau volcanic field that affected its distribution. Exposure is limited to the Island Park area, but buried thickness in the first-cycle caldera may be considerable. The calculated initial area is nearly 2,700 km^2 and initial volume is 280 km^3. Shading is darker for thicker sections. Isopachs in meters, dashed where inferred.

tains and the Great Plains (Izett and Wilcox, 1982). As with the Huckleberry Ridge ash beds, some of this fall material was reworked into thick ash-rich sediments in the Gulf of Mexico (Dobson and others, 1991).

The Mesa Falls Tuff apparently is a simple cooling unit. Its outcrop distribution is restricted to the Island Park area, and densely welded sections occur only within or immediately adjacent to the Island Park basin. Its basal fallout there is thick and coarse-grained. The source area of the Mesa Falls, thus, is in the Island Park area.

Sanidine from the Mesa Falls Tuff has been dated by K-Ar at 1.3 Ma (table 2). Izett and others (1970) noted that a Pearlette-like ash bed, now correlated with the Mesa Falls (Izett, 1981; Izett and Wilcox, 1982), underlies till in northeastern Nebraska.

THE SECOND-CYCLE CALDERA

Eruption of the Mesa Falls Tuff from the Island Park area about 1.3 million years ago caused the roof of the magma chamber from which the eruptions occurred to collapse. This second-cycle caldera, the Henrys Fork caldera, lies largely or entirely within the Big Bend Ridge segment of the first-cycle caldera. The ring-fracture zone of the second-cycle caldera, however, did not correspond exactly with that of the first. The ring-fracture zone of the Henrys Fork caldera, through which the Mesa Falls Tuff presumably erupted, lay against the old caldera wall at Thurmon Ridge, on the north side of Island Park. However, the Mesa Falls can be mapped northward from the south rim of the first-cycle Big Bend Ridge caldera segment almost halfway across the basin of Island Park with fault displacements aggregating less than 100 m—less than the single scarp at the north caldera wall and much less than is required for collapse by the volume of the Mesa Falls Tuff. A circular positive gravity anomaly, which corresponds to the partly basalt-filled Henrys Fork caldera (Blank and Gettings, 1974; Christiansen, 1982), indicates that the main south boundary of the caldera is buried by younger rocks in the middle of the Island Park basin. This inferred caldera boundary defines a circular block about 19 km in diameter (pl. 2). Faults along the south wall of the older Big Bend Ridge caldera segment were rejuvenated when the second-cycle Henrys Fork collapse occurred, but displacements on individual faults during this rejuvenation were only a few tens of meters or less.

Because the Henrys Fork caldera was nested against the north side of a first-cycle caldera segment, the Mesa Falls ash flows were virtually confined by the preexisting south caldera wall. Only fallout ash and pumice and a few thin slightly welded to nonwelded ash flows of the Mesa Falls lie on the south outer flank of the Big Bend Ridge segment. By contrast, the high-standing north wall of the second-cycle Henrys Fork caldera, presumably adjacent to the eruptive fissures, has a thick densely welded section of Mesa Falls ash-flow tuff.

Hamilton (1965) made a reconnaissance study of the Island Park area and outlined a basin which he designated the Island Park caldera. Hamilton's Island Park caldera corresponds in some degree to the Henrys Fork caldera, including the rejuvenated first-cycle caldera faults at the south wall that have only slight renewed displacement. It seems best, however, to describe the features of each cycle in the Island Park area separately and to name them separately as appropriate (Christiansen, 1982). The "Island Park caldera" as Hamilton outlined it is a composite structure. In part it is a compound volcanic-collapse basin that includes parts of both the first- and second-cycle caldera segments, but in part it is a constructional basin, delimited on the east by typical flow-front scarps of large rhyolitic flows of the third cycle (pl. 2). Hamilton regarded certain subsummits on these flows as marking vents whose distributions indicated a buried east ring-fracture boundary for the caldera, but surface flow structures (pl. 1) show clearly that all these flows vented many kilometers farther east. Therefore, Island Park as a whole is a product of all three volcanic cycles but not a caldera in any strict sense (Christiansen, 1982).

POSTCOLLAPSE RHYOLITIC AND BASALTIC VOLCANISM

Five small rhyolite domes within or adjacent to the Henrys Fork caldera have been designated as the Island Park Rhyolite (Christiansen, 1982). The bodies are shown on plate 2 as the Silver Lake, Osborne Butte, Elk Butte, Lookout Butte, and Warm River Butte domes. The ages of these domes relative to other major units are known from general map relations (pl. 2), isotopic ages, and paleomagnetic polarities (table 2) rather than from clear stratigraphic contacts at individual localities. Because the domes stand above the floor of the second-cycle Henrys Fork caldera, they must postdate the Mesa Falls Tuff and the contemporaneous or subsequent caldera collapse. The Lava Creek Tuff that was emplaced later into the Henrys Fork caldera has normal paleomagnetic polarity as do all younger units, but the domes of the Island Park Rhyolite all have reverse polarities (table 2); thus the Island Park Rhyolite must entirely predate the Lava Creek Tuff.

The Island Park Rhyolite is lithologically distinctive. All five bodies consist of extremely phenocryst-rich rhyolite; the phenocrysts generally constitute more than half of the rock. The phenocrysts are very large, commonly 1 to several centimeters. In this respect their close petrographic relation to the Mesa Falls Tuff is apparent. The principal phenocrysts, as in most other rhyolites of this volcanic field, are quartz, sodic sanidine, sodic oligoclase, opaque oxides, ferroaugite, fayalite, and some hornblende.

The rhyolite bodies are typical steep-sided domes with a shell-like flow-structure pattern parallel to their sides and concentric around their vents, mainly endogenous by the criteria of Williams (1932). The domes occur in a linear belt about 30 km long and no more than 7 km wide that trends northwest across Island Park. Most of the domes are within the area of the Henrys Fork caldera, but Warm River Butte lies on the rim of the first-cycle Big Bend Ridge caldera segment. Hamilton (1965) mapped Warm River Butte with the flows of an "Island Park caldera" rim. I, however, correlate it with the Island Park Rhyolite, because it is lithologically like the other Island Park domes, because it does not appear to have been cut by faults during the second-cycle collapse as the other flows on the old rim were, and because it has a similar K-Ar age as other Island Park Rhyolite domes (table 2). Warm River Butte also lies on the same linear trend as the other domes of the Island Park Rhyolite. Although it is possible that other flows and domes of Island Park Rhyolite lie buried within the Henrys Fork caldera, the alignment of the visible bodies suggests a structural control. Their vent zone is parallel to the elongate structural domes in the twin resurgent segments of the Yellowstone caldera. Southeastward they trend toward Driggs in the Teton Basin, a similar-trending fault-bounded valley west of the Teton Range (fig. 5; pl. 3).

The great petrographic similarity of the individual domes and their strong affinity with the Mesa Falls Tuff suggests that they do not fill much of the stratigraphic bracket provided by the Mesa Falls and Lava Creek Tuffs. The Osborne Butte dome is dated at 1.3 Ma and the Warm River Butte dome at 1.2 Ma (table 2); probably all the Island Park Rhyolite domes are only slightly younger than the Mesa Falls Tuff.

Basalts interlayered with sediments in the area of The Narrows of the Grand Canyon of the Yellowstone, on the opposite side of the volcanic field from Island Park (pl. 1), have age and stratigraphic relations that place them in the time of the second volcanic cycle. The sediments and basalts of the The Narrows consist mainly of interlayered gravels and basalts (fig. 38) but also contain some finer grained sediments and several tills and related glacial deposits (Pierce, 1974; 1979, p. F16). These deposits are well exposed in The Narrows of the Grand Canyon. They are about 70 m thick in that area, and they fill a paleovalley that had been cut down through the Huckleberry Ridge Tuff, the Junction Butte Basalt, and older rocks. A fallout ash bed interlayered in these deposits has been identified by its chemistry and age as representing the Mesa Falls eruption (Obradovich and Izett, 1991). The gravels are mainly yellowish in color because of pervasive hydrothermal alteration. They contain abundant clasts of welded tuff of the Yellowstone Group, distinguishing them from gravels interlayered with the older Junction Butte Basalt in the same area (fig. 30). In an earlier paper (Pierce and others, 1970) we mistakenly identified these welded tuff clasts, on the basis of appearance alone, as being from the Lava Creek Tuff . On the basis of more complete subsequent work the clasts are now identified with confidence as being from Huckleberry Ridge Tuff. This identification was confirmed by field measurements of the orientation of remanent magnetization in about 20 clasts by us and by R. L. Reynolds (written commun., 1970). Most of the clasts have directions of natural remanent magnetization nearly in the plane of their foliation as does the Huckleberry Ridge in most places; it is the only welded tuff of this region having remanent magnetization of such low inclination (table 2).

The basalts of The Narrows are very similar in outcrop appearance, petrography, and chemistry to the Junction Butte Basalt. Some of the basaltic flows of The Narrows were ponded deeply near the lower Grand Canyon of the Yellowstone just as the Junction Butte flows were. Well-developed two-tiered columnar jointing is equally characteristic of the basalts of both formations. In fact, the only ways of identifying isolated outcrops of either of these basalts in this area are by the presence or absence of Yellowstone Group welded tuff clasts in associated gravels and, less definitively, by the topographic positions of the flows and of the erosion surfaces on which they rest. Fortunately, some gravel has been found in association with many of the isolated outcrops, but some misidentifications could have been made in our mapping of these basalts in the canyon of the Yellowstone River below its junction with the Lamar.

The age of the sediments and basalts of The Narrows is reasonably well established on the basis both of stratigraphic relations and of $^{40}Ar/^{39}Ar$ and K-Ar age determinations. The unit is younger than the 2.1-Ma Huckleberry Ridge Tuff and

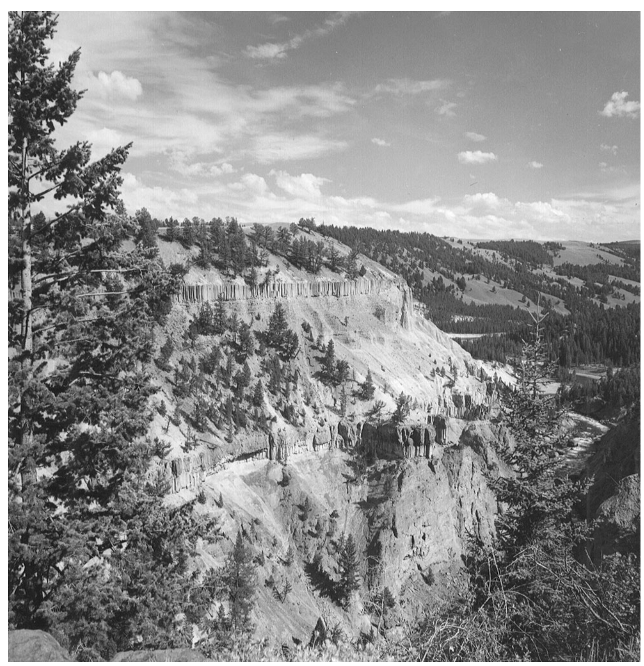


Figure 38.—Sediments and basalts of The Narrows, lower Grand Canyon of the Yellowstone, Yellowstone National Park. Basalt flows with well-developed columnar joints are interlayered with gravels containing clasts of Huckleberry Ridge Tuff. View to southeast from about 3/4 km north of Tower Falls.

older than the 640-Ka Lava Creek Tuff. Sanidine from an ash bed low in the section has been dated by $^{40}Ar/^{39}Ar$ at 1.30±0.01 Ma (Obradovich and Izett, 1991); the basalt flow that caps the sequence is poorly dated by K-Ar at 1.30±0.35 Ma (table 2). Although the forms of the basalt flows show that the accumulation represents repeated cutting and filling of new channels through older gravels, K-Ar ages of the dated units are sufficiently restricted to indicate that the entire formation accumulated at about 1.3 Ma within no more than a few hundred thousand years.

Basaltic lavas younger than the Mesa Falls Tuff and older than the Lava Creek Tuff in Island Park and the nearby area to the southeast have been discussed elsewhere (Christiansen, 1982) as the basalt of Warm River. One flow has been dated as about 760 ka (table 2) and, because all of them have reverse paleomagnetic polarities, they presumably are older than 780 ka (Baksi and others, 1992; Spell and McDougall, 1992). These basalts form a sequence that commonly is about 30 m thick. Contacts with underlying and overlying units are exposed in the area between the settlement of Warm River and the area south of Moose Creek Butte and are also exposed in Island Park near Sheep Falls. Extensive exposures of the basalt occur on both sides of the Warm River in the 5 km stretch above its mouth, but a thick mantle of upper Pleistocene loess covers much of the bedrock and all the important contacts in natural exposures of this area. Near Sheep Falls and Moose Creek Butte the basalt of Warm River overlies Mesa Falls Tuff; the Lava Creek Tuff is the only bedrock unit that overlies the basalt of Warm River.

Other basalts of reverse paleomagnetic polarity are widespread farther south, in the areas from Falls River and Bitch Creek to the Teton River and beyond, but they lie on the older Huckleberry Ridge Tuff. Their ages have not been determined, but at least some of them may correlate with the basalt of Warm River. None of the vents for the basalt of Warm River or the similar basalts farther south is known. Other basalts probably correlative with the Warm River, which also have reverse polarities, occur in the area of Shotgun Valley, northwest of Island Park, where at least some of them vented locally; just north, in the area of Coffee Pot Creek and Henrys Fork, similar basaltic flows are exposed in stream cuts beneath the Lava Creek Tuff (pl. 2).

The basalt of Warm River and its possible correlatives have sparse to moderately abundant phenocrysts of plagioclase. Olivine and clinopyroxene are present together with plagioclase and opaque oxides in the fine-grained matrix.

RELATIONS AMONG THE VOLCANIC CYCLES

Having described the volcanic rocks and structures associated with each of the three cycles of the Yellowstone Plateau volcanic field, it is appropriate to consider some important relations among the cycles.

INDEPENDENCE OF THE CYCLES

From the perspective of the history just presented, the most important units of the Yellowstone Plateau volcanic field are the three voluminous ash-flow sheets. The aggregate volume of these ash flows is more than 3,700 km^3, and the assumptions used in estimating hidden volumes within the calderas were conservative. Furthermore, the fallout volumes would be in addition to that just stated. The minimum volume of magma represented in the three ash-flow sheets of the Yellowstone Group, if an average porosity of about 15 percent were assumed for the welded tuffs, would be at least 3,200 km^3, but more likely the underestimate of the volume of erupted materials is greater than the porosity correction. By either estimate, these tuffs represent an enormous volume of erupted magma accounting for at least half of the estimated magmatic volume of the entire volcanic field (Christiansen, 1984, fig. 3). This great volume of material, erupted in three complex episodes, each of brief duration, played the key roles in the volcanic history of the field. The intrusive bodies that sustained the eruptive activity of each volcanic cycle, however, must have been many times larger, as indicated by the arguments of Smith and Shaw (1975), Smith (1979), and Christiansen (1979). Although all three cycles occurred within the same general region and overlapped to a considerable degree, each cycle was focused in a different major area of eruptive activity. The major volcanic focus of the first cycle extended from Island Park to the present Central Plateau and from the north end of Jackson Hole to somewhere south of the Gallatin Range. This was the greatest of the three cycles, probably more than double the erupted volume of the third cycle. The second cycle, the smallest, was confined to Island Park, within the major volcanic focus of the first cycle but including only a small part of it. The third cycle focused about the inner part of the Yellowstone Plateau, from the Madison Plateau to the Lamar River, from the Red Mountains to Norris Geyser Basin, and along an additional zone extending radially northward to Mammoth Hot Springs.

Each of the cycles lasted between about a half-million and a million years. Each cycle represents a more or less complete sequence of volcanic events, following at least approximately the stages of the resurgent caldera cycle of Smith and Bailey (1968a), except that resurgent doming can be documented only for the third cycle. At least the second-cycle caldera probably was not resurgent, for a dome as high as the caldera rim would not now be completely buried; furthermore, the positive gravity anomaly associated with it suggests that the caldera is filled by a considerable thickness of basalt. It seems necessary to assume that each of the volcanic cycles was related to the evolution of a large body of rhyolitic magma beneath the region of the major volcanic focus; the rise of individual plutons to high crustal levels; major degassing of these high-level magma bodies by surface volcanism, principally by the major pyroclastic eruptions but continuingly for a considerable time by numerous central eruptions; and eventual consolidation and cooling of the magma at depth. The time scale for this entire process seems to have been on the order of a million years, but it probably varied with the size and depth of the major intrusive bodies. Basaltic volcanism was ongoing around the flanks of each major focus of rhyolitic activity and clearly is important in the life cycle of the volcanic field as a whole, but it has been essentially independent of the individual rhyolitic cycles.

The sole possible exception to the picture of essentially complete and independent cycles of rhyolitic volcanism was the second cycle. Since it occurred within part of the major focus of the first cycle and appears to have been smaller and simpler than the others, it seems pertinent to inquire whether the second cycle was truly independent or whether it was merely a rejuvenation of activity related to the first-cycle magma body. About 700,000 years elapsed between the major ash flows of the first and second cycles, and about 650,000 years between the second and third; the second-cycle volcanic focus did not merely reoccupy the older caldera but was nested within it and broke across parts of it; and the activity of the second cycle, although simpler than the other two, seems to have gone through all the essential stages represented by the other cycles. Although there probably was no second-cycle stage of resurgent doming, Smith and Bailey (1968a) noted that the fundamental differences between large rhyolite-related calderas that are resurgent and those that are not are poorly understood. The second cycle clearly does record the rise of a localized large magma body to near the surface, major eruptions from it, caldera collapse, and postcollapse intracaldera volcanism. It may, however, record renewed rise of a residual portion of the first-cycle magma body after most of it had crystallized (Christiansen, 1979).

RELATIONS BETWEEN RHYOLITE AND BASALT

Almost all the igneous rocks present in the Yellowstone Plateau volcanic field are rhyolites and basalts. Virtually the only volcanic rocks of intermediate composition known from the field are mixed lavas of the rhyolite-basalt complexes such as the Gardner River mixed lavas. Rhyolite, including the major ash-flow sheets, is by far the predominant rock type. A reasonable estimate of the volume of the known rhyolites, before erosion, is about 4,730 km^3; a rough estimate of the total volume of all rhyolites of the volcanic field is about 6,500 km^3 (Christiansen, 1984, fig. 3). The volume of basalts in the volcanic field as a whole (pl. 3) was about 250 km^3. There is no reason to believe that, if unknown, unexposed lavas and tuffs could be added to the estimate, the relative proportions would change appreciably.

Hamilton (1963; 1964) based his petrogenetic concepts for the Yellowstone and Island Park regions on the idea that basalt and rhyolite erupted more or less simultaneously throughout the lifetimes of the volcanic systems, both magma types always being available for eruption from the regions of the principal volcanic foci. Although the present work confirms that both rhyolites and basalts erupted intermittently throughout the volcanic history of the field, their roles in the volcanic history were quite different. The rhyolites, both the ash-flow tuffs and the voluminous lavas, erupted in distinct cyclical episodes that seem to be related to deeper-seated magmatic events, as just noted. Some basalts probably erupted before any rhyolites, but during the lifetime of each volcanic cycle, basalts erupted only around the margins of the principal volcanic focus. In this respect, the history of the Yellowstone Plateau field parallels that of the Jemez Mountains of New Mexico (Smith and Bailey, 1968b; Smith and others, 1970). The basaltic eruptions that accompanied rhyolitic activity were frequent but of relatively minor total volume in comparison to the rhyolites. Eventually with the completion of each rhyolitic cycle and the extinction of rhyolitic eruptions from its main focal area, basalts did erupt through the old rhyolitic foci. Island Park, including both its first- and second-cycle caldera segments, has passed through this stage. No basalts, however, are known to be present within the Yellowstone caldera except for a few small exposures of Madison River Basalt on the north wall near Purple Mountain. These basalts erupted from vents outside the caldera basin and flowed down the eroded wall.

Hamilton (1964) noted the presence of abundant basalt in moraines of Bull Lake age on the south side of the Madison Valley near West Yellowstone and proposed that they indicate the presence of voluminous basalts buried beneath the Plateau Rhyolite within the Yellowstone caldera. Most of these basaltic boulders and cobbles appear to be Madison River Basalt similar to that which once formed extensive outcrops on the outer flank of the Yellowstone plateau near the Madison Canyon and to the north. Similar basalts are buried on the outer caldera flank by the West Yellowstone flow. Pierce and others (1976) confirmed by directional scour features that the basaltic erratics in the moraines near West Yellowstone were derived from areas now buried by the West Yellowstone flow and other rhyolite flows of the Central Plateau Member. The caldera wall, however, is buried by these flows 13 km or more east and southeast of the moraines (pl. 1). Thus, the basalts were not necessarily from within the caldera.

Hamilton (1965) also inferred that basalts and rhyolites are interlayered in Island Park. His inference was based on his interpretation of the stratigraphy along the east side of the canyon of the Henrys Fork at Upper Mesa Falls, where the section consists of two ash-flow sheets that are directly superposed, the Mesa Falls and Lava Creek Tuffs, and the Gerrit Basalt, which occurs both concordantly on the Lava Creek at the rim of the canyon and also unconformably on an erosional terrace carved through the vapor-phase zone onto the devitrified zone of the Mesa Falls Tuff, about halfway down the present canyon (pl. 2). Hamilton thought that the youngest terrace-sited basalt of the Mesa Falls area occurred within the ash-flow section.

(Hamilton (written commun., 1972) has revisited the locality and agrees with the present reinterpretation.) The basalt of Warm River is present within the first-cycle Big Bend Ridge caldera segment between the Mesa Falls and Lava Creek Tuffs, but the location of its vent is not known and the basalt is considerably younger than the first-cycle rhyolites that erupted within the Big Bend Ridge segment.

Not only does the geologic record at Island Park show that basalts ultimately erupt through the former focal areas of voluminous rhyolitic volcanism, but the geologic evolution of the eastern Snake River Plain suggests that basalts may ultimately equal or exceed the volume of rhyolite in the Yellowstone Plateau volcanic field. This argument is developed more fully in the final chapter of this paper, but it has been suggested previously (Christiansen and Blank, 1969; Armstrong and others, 1975; Christiansen and McKee, 1978; Leeman, 1982a; Embree and others, 1983) that the axis of the eastern Snake River Plain is the locus of a line of extinct and buried rhyolitic volcanic complexes much like that of the Yellowstone Plateau; the axis has since progressively subsided and been flooded by basalts of great volume. The rhyolitic complexes were active successively from southwest to northeast. Basaltic volcanism was active around the margins of each volcanic complex during its lifetime, as in the Yellowstone Plateau field, but ultimately the basalts of the Snake River Group invaded and buried each of the rhyolitic complexes.

The early appearance of basalt in the Yellowstone Plateau volcanic field, the constancy of plateau-marginal basaltic activity around the principal foci of the rhyolitic cycles, the eventual invasion of the extinct rhyolitic areas by basalts, and the apparent ultimate predominance of basalts over rhyolites in the volcanic evolution of the Snake River Plain-Yellowstone Plateau system as a whole clearly demonstrate that basaltic magmogenesis and basaltic volcanism constitute the fundamental igneous process in the evolution of the field.

Having stated this fundamental primacy of basalt, I also note specifically that the rhyolite-basalt association is an essential, not an accidental association of diverse magmas in time and space. Basalts bear the same general relations to each of the three volcanic cycles and to the older, now largely buried rhyolitic complexes of the eastern Snake River Plain. Furthermore, as noted in the next chapter, the basalts themselves may record a progressive evolution related to the overall evolutionary stage of each volcanic cycle. The rhyolite-basalt association of this system is not a rare or isolated occurrence. This petrologic association has been noted in many igneous fields of the world, both modern and ancient. Examples include Iceland, where Bunsen (1851) over a century ago made a major point of the association, the Tertiary volcanic province of Britain (Richey, 1961), the Karroo, Nuanetsi, and related provinces of southern Africa (Cox and others, 1965; Stillman, 1970), the upper Cenozoic of several areas of the Western United States (Christiansen and Lipman, 1972; Leeman, 1982a), and many others. The specific nature of the rhyolite-basalt relations of such provinces will be discussed in a later section, but here both its fundamental nature and its generality are noted.

PETROLOGY

The work reported in this paper has been primarily a field study of the Yellowstone Plateau region. Detailed petrologic and geochemical studies have been reported elsewhere (Doe and others, 1982; Hildreth and others, 1984; 1991), and as of this writing, others are underway. Here only general descriptions are given, along with some observations and interpretations that bear on hypotheses of magmotectonic evolution.

THE RHYOLITES

Most rhyolites of the Yellowstone Plateau region are remarkably uniform in mineralogy and major-element chemistry. Most of the rhyolites have high silica contents, characteristically exceeding 76 percent SiO_2. Most of them, as noted in preceding descriptions of individual units, contain phenocrysts of quartz, sanidine±sodic plagioclase, ferroaugite, magnetite, ilmenite, fayalite, and accessory zircon, chevkinite, and allanite. A few individual units have slightly different phenocryst assemblages. Most phenocrysts in the rhyolites of the Yellowstone Plateau field are virtually unzoned or only weakly zoned, with specific exceptions noted below.

PHENOCRYSTS

Alkali feldspar phenocrysts in the Yellowstone rhyolites are predominantly sanidine of compositions from about Or_{60} to Or_{50}, although a total range from Or_{56} to Or_{43} is shown by Boyd (1961, p. 399) and from Or_{66} to Or_{50} by Leeman and Phelps (1981). Sanidine analyses discussed by Hildreth and others (1988) range from Or_{61} to Or_{50} in the Lava Creek Tuff and from Or_{66} to Or_{49} in flows of the Upper Basin Member (Plateau Rhyolite). Sanidine in the Mesa Falls Tuff is somewhat less sodic than in most of the other rhyolites; Basu and Vitaliano (1976) noted that the Mesa Falls sanidine is unzoned and uniformly approximates Or_{62} throughout the unit. The sanidine in many of the rhyolites is cryptoperthitic, but Boyd (1961, p. 400) showed clearly that homogeneous feldspars in the basal and upper glass zones of a section of welded tuff (the principal reference section of member B of the Lava Creek Tuff) grade into unmixed feldspars in the devitrified and vapor-phase zones. Thus, at least most of this incipient unmixing of sanidines is related to post-emplacement cooling of the volcanic units.

Plagioclase phenocryst compositions in ash-flow tuffs of the Yellowstone Group and the Plateau Rhyolite generally are between about An_{10} and An_{20}. In some rhyolites of the field, sanidine phenocrysts occur with euhedral cores of albite-oligoclase; in most of these rocks, distinct plagioclase phenocrysts are sparse, and in some of them plagioclase occurs only in the cores of such composite grains. This association seems to record a reaction relation in which magmas with stable two-feldspar assemblages differentiated into liquids with compositions in a one-feldspar stability field. As already noted, the older postresurgence rhyolitic lavas of the third cycle, particularly the Upper Basin Member of the Plateau Rhyolite, are plagioclase-rich, some of them containing no sanidine. The Lewis Canyon Rhyolite is lithologically similar to the Upper Basin Member. Plagioclase in these plagioclase-rich flows is more calcic, averaging about An_{20} to An_{30}. The observed range of plagioclase compositions in lavas of the Upper Basin Member is An_{18} to An_{43} (Hildreth and others, 1988). The plagioclase in many of these plagioclase-rich rhyolites, furthermore, is zoned and highly sieved. Comparison to dendritic plagioclases formed by magma quenching and rapid growth (for example, see Hibbard, 1981) suggests that this texture in the Upper Basin flows may have resulted from rapid growth in a magma chamber that was greatly disturbed by eruption of the Lava Creek Tuff and collapse of the Yellowstone caldera. This interpretation is consistent with the drastic lowering of $\delta^{18}O$ in rhyolite of the Upper Basin member that Hildreth and others (1984) interpret as indicating drastically disturbed equilibrium and interaction with meteoric water in the post-Lava Creek magma chamber. The relation of the Lewis Canyon Rhyolite to the Huckleberry Ridge magma chamber is problematical because K-Ar dating of the Lewis Canyon indicates it to be much younger than the caldera segment in which it occurs.

Mafic phenocrysts in the Yellowstone Plateau rhyolites generally are so sparse as to be only accessory minerals. The mafic phenocryst assemblages in these rhyolites comprise mainly high-iron, low-aluminum minerals, or minerals high in rare-earth elements. As noted earlier, ferroaugite, Fe-Ti oxides, and fayalite are predominant in most of the rhyolites. Ferrohypersthene occurs in place of fayalite in lavas of the Upper Basin Member of the Plateau Rhyolite (Hildreth and others, 1988). Phenocrystic biotite occurs sparsely in a few units at the west edge of the volcanic field but is absent elsewhere. Hornblende, common in the rhyolites of many calc-alkalic volcanic fields, is an accessory constituent in some of the Yellowstone Plateau rhyolites, but only in Member A of the Lava Creek Tuff is it one of the predominant mafic phenocryst minerals.

CHEMISTRY AND MAGMATIC DIFFERENTIATION

Chemically the predominant rhyolites of the field are remarkably uniform. Major-element analyses of Yellowstone rhyolites are listed in tables 4 through 13. Listed in these tables are new analyses, given completely in table 4, as well as most published modern chemical analyses of unaltered (or in a few instances, slightly altered) rhyolites from the Yellowstone Plateau volcanic field. As the sole purpose of tables 5 through 13 is the comparison of major-element analyses that might represent essentially magmatic compositions, all the data have been recalculated to 100 percent without H_2O, CO_2, S, Cl, or F, which can vary with numerous postmagmatic factors and which were not uniformly analyzed in all samples; in addition, all Fe was calculated as FeO. The CIPW norms, also shown only for comparisons and not as petrologic models, were calculated from the recalculated analyses.

The compositionally zoned ash-flow sheets of the Yellowstone Group represent nearly the entire compositional range of the rhyolites (tables 5, 6, 9, and 14). In a gross sense, the individual subsheets—the stratigraphically simple Mesa Falls Tuff and the individual members of the other tuffs—are chemically zoned, with slightly more evolved rhyolite near the base than near the top. In only the upper part of member B of the Huckleberry Ridge Tuff, however, is the major-element variation notable. Most of the Yellowstone rhyolites are high-silica types; all have more than about 72 percent SiO_2 but the principal ones have more than 75 percent (recalculated water-free). The only exceptions to the limited compositional range of high-silica rhyolites occur in upper member B of the Huckleberry Ridge Tuff (table 5), the pre-Lava Creek Lewis Canyon Rhyolite (table 8), and some of the early postresurgence Upper Basin Member rhyolites of the third cycle (see table 10). These are the least silicic rhyolites of the volcanic field; Fenner (1936, p. 264) called rhyolites of the Biscuit Basin flow “dacite” although his best analysis was of a somewhat altered rock. (The low K_2O of Fenner’s analysis, table 10, reflects this alteration.) Silica contents and normative feldspar compositions of the unaltered Biscuit Basin rocks generally qualify them as rhyolites. Rhyolites of the Upper Basin Member do, however, represent the compositional extreme of all the silicic rocks of the Yellowstone Plateau field; besides having distinctively lower SiO_2, they have higher CaO and ΣFeO than the other rhyolites.

Some of the chemical data tabulated here must be used with caution in characterizing magmatic compositions. Mechanical sorting of crystals relative to glass affected many of the pyroclastic rocks. As already noted, minor hydrothermal alteration may cause a problem with a few of the data. Also, certain elements like chlorine and fluorine are significantly affected by devitrification (for example analyses 5 and 6 in table 4). Sodium commonly is leached and potassium may be increased by ion exchange in hydrated glasses (for example, analyses 3 and 7, table 4). Nevertheless, the data indicate some distinctive major-element characteristics of the Yellowstone magmas. Even compared to other highly evolved rhyolites, the Yellowstone suite tends to be high in potassium—especially in the ratio of K/Na (tables 4-13)—and in fluorine—especially in the ratio of F/Cl in undevitrified and unaltered rocks (tables 4 and 14).

Variation within the rhyolite suite, generally small in terms of mineralogy and major-element chemistry, is much more clearly revealed in trace-element compositions and certain isotopic ratios. Some trace-element analyses are listed in table 14. Values shown in the table are representative of a considerably larger set of data acquired for a more comprehensive

Table 4.—*New chemical analyses of rocks from the Yellowstone Plateau volcanic field.*

	1	**2**	**3**	**4**	**5**	**6**	**7**	**8**
Field No.	7YC-288A	8YC-460A	7YC-246	6YC-13D	6YC-13H	6YC-38	8YC-411	8YC-477B
Lab No.	D102475	D102483	D102473	D101806	D102467	D101807	D102480	D102485
SiO_2	75.83	72.29	71.41	75.98	75.57	77.40	74.12	74.18
Al_2O_3	12.42	13.40	13.63	12.33	12.69	12.04	12.13	12.12
Fe_2O_3	.94	.76	1.59	.55	1.45	1.08	1.40	.63
FeO	.63	.97	.54	1.08	.27	.05	.17	.69
MgO	.00	.20	.07	.00	.01	.00	.04	.08
CaO	.52	.58	.81	.49	.27	.26	.43	.67
Na_2O	3.70	2.77	3.18	3.69	3.36	3.23	2.40	3.10
K_2O	5.08	5.40	4.82	5.10	5.18	4.89	5.19	5.01
H_2O+	.25	2.42	2.78	.13	.52	.39	3.09	2.52
H_2O-	.05	.44	.33	.00	.06	.30	.25	.11
TiO_2	.10	.16	.22	.13	.13	.10	.11	.30
P_2O_5	.00	.02	.02	.01	.02	.01	.02	.10
MnO	.04	.04	.05	.04	.04	.02	.03	.04
CO_2	.01	.01	.00	.00	.02	.01	.01	.00
Cl	.06	.11	.04	.12	.03	.01	.17	.06
F	.15	.18	.15	.14	.02	.01	.23	.11
S	.01	.01	.01	.01	.01	.01	.00	.01
Subtotal	99.79	99.76	99.65	99.79	99.65	99.80	99.80	99.73
Less O	.07	.11	.07	.09	.02	.00	.14	.06
Total	99.72	99.65	99.58	99.70	99.63	99.80	99.66	99.67

	9	**10**	**11**	**12**	**13**	**14**	**15**	**16**
Field No.	6YC-106	8YC-441	7YC-271	6YC-113	8YC-374C	6YC-73	6YC-169	6YC-93
Lab No.	D102470	D102481	D102474	D101812	D102478	D101811	D102472	D102469
SiO_2	69.86	72.95	73.58	76.44	76.14	76.51	76.39	76.74
Al_2O_3	13.11	12.50	12.45	12.09	12.18	12.05	11.94	12.27
Fe_2O_3	1.17	1.53	1.26	.46	.44	.47	.47	1.26
FeO	2.14	.22	.27	1.00	.99	1.11	1.17	.05
MgO	.30	.19	.04	.03	.00	.01	.01	.04
CaO	1.51	.84	.43	.45	.44	.42	.44	.42
Na_2O	3.27	3.20	3.09	3.55	3.46	3.65	3.54	3.73
K_2O	4.69	5.10	5.04	5.16	5.27	5.10	5.09	4.92
H_2O+	2.67	2.37	2.67	.26	.33	.23	.22	.14
H_2O-	.13	.10	.28	.02	.05	.02	.05	.04
TiO_2	.45	.26	.16	.14	.16	.13	.14	.10
P_2O_5	.09	.12	.01	.02	.01	.01	.01	.01
MnO	.07	.04	.06	.04	.04	.04	.04	.03
CO_2	.00	.01	.01	.02	.00	.01	.01	.00
Cl	.05	.03	.07	.08	.08	.08	.07	.06
F	.11	.10	.15	.17	.15	.18	.16	.13
S	---	---	.01	---	.01	---	.01	.00
Subtotal	99.62	99.57	99.58	99.93	99.15	100.02	99.76	99.94
Less O	.06	.05	.08	.09	.08	.10	.09	.06
Total	99.56	99.52	99.50	99.84	99.67	99.92	99.67	99.88

Table 4.—*New chemical analyses of rocks from the Yellowstone Plateau volcanic field.*—Continued

	17	**18**	**19**	**20**	**21**	**22**	**23**	**24**
Field No.	65YR-7	6YC-43B	6YC-64	6YC-153	8YC-396A	6YC-144	7YC-295	7YC-297
Lab No.	D101805	D101808	D101810	D102471	D102479	D101814	D102476	D102477
SiO_2	76.59	76.59	76.71	76.68	76.39	46.89	50.06	48.65
Al_2O_3	12.25	12.42	12.29	12.38	12.30	16.79	18.66	16.48
Fe_2O_3	.50	.30	.41	.73	.44	3.37	3.06	1.99
FeO	.72	.90	.81	.50	.83	8.91	6.52	9.95
MgO	.00	.02	.01	.00	.06	7.88	5.26	6.88
CaO	.38	.56	.42	.32	.34	10.06	10.48	9.77
Na_2O	4.01	3.35	3.87	3.75	3.79	2.64	3.13	2.99
K_2O	4.78	5.20	4.91	5.02	4.88	.27	.61	.55
H_2O+	.17	.15	.15	.04	.19	.45	.04	.21
H_2O-	.01	.00	.02	.02	.05	.15	.07	.02
TiO_2	.07	.11	.08	.08	.09	1.80	1.49	1.88
P_2O_5	.01	.01	.01	.01	.01	.30	.20	.38
MnO	.03	.02	.03	.02	.03	.19	.15	.18
CO_2	.01	.01	.00	.01	.00	.01	.01	.00
Cl	.11	.08	.09	.01	.10	.00	.00	.01
F	.26	.14	.19	.15	.18	.03	.03	.03
S	---	---	---	.01	.01	---	.02	.01
Subtotal	99.90	99.86	100.00	99.73	99.69	99.74	99.79	100.04
Less O	.14	.08	.10	.06	.10	.01	.01	.01
Total	99.76	99.78	99.90	99.67	99.59	99.73	99.78	100.03

	25	**26**	**27**	**28**	**29**	**30**	**31**
Field No.	8YC-482	6YC-51	8YC-480	8YC-471C	6YC-44	6YC-139	8YC-446
Lab No.	D102487	D101809	D102486	D102484	D102468	D101813	D102482
SiO_2	49.00	50.49	53.82	49.77	48.64	51.38	46.31
Al_2O_3	16.64	16.03	15.73	15.54	16.41	15.55	16.68
Fe_2O_3	3.09	1.78	2.82	3.33	1.26	1.71	2.16
FeO	8.70	8.61	7.23	7.56	11.37	9.66	11.09
MgO	6.97	7.49	5.39	6.71	6.18	6.05	7.44
CaO	9.66	10.35	8.17	9.75	9.40	8.70	9.44
Na_2O	2.95	2.81	2.83	2.74	2.97	2.93	2.99
K_2O	.53	.45	1.44	.49	.57	1.11	.59
H_2O+	.02	.09	.13	.61	.14	.45	.05
H_2O-	.03	.02	.18	1.03	.01	.07	.09
TiO_2	1.78	1.54	1.64	1.71	2.39	1.86	2.28
P_2O_5	.22	.22	.20	.23	.31	.23	.52
MnO	.17	.16	.15	.16	.20	.17	.20
CO_2	.00	.01	.00	.11	.01	.00	.02
Cl	.00	.00	.01	.00	.01	.01	.00
F	.03	.03	.05	.04	.04	.05	.06
S	.02	---	.02	.01	.01	---	.01
Subtotal	99.81	100.08	99.81	99.79	99.92	99.95	99.93
Less O	.01	.01	.02	.02	.02	.02	.03
Total	99.80	100.07	99.79	99.77	99.90	99.93	99.90

Table 4.—*New chemical analyses of rocks from the Yellowstone Plateau volcanic field.*—Continued

Huckleberry Ridge Tuff

1. Densely fused fallout tuff just below base of ash-flow tuff of member A. From bluff on south rim of Mt. Everts about 1.4 km northwest of Undine Falls. Sample consists of black spheroidal obsidian cores about 2 cm in diameter, separated from their slightly perlitic matrix. Sparse small phenocrysts of quartz, sanidine, oligoclase, and accessory minerals enclosed in welded glass shards.

Mesa Falls Tuff

2. Pumice from glassy nonwelded base of ash-flow sheet just above fallout tuff. From west side of large roadcut in U.S. Highway 20 about 6.8 km north of Ashton, Idaho. Analyzed sample from a single 15-cm lump of coarsely vesicular white pumice. Abundant phenocrysts of quartz, sanidine (as much as 1.5 cm in diameter), oligoclase, and accessory minerals.

Lewis Canyon Rhyolite

3. Scoriaceous glassy rhyolite from upper flow breccia of flow in Lewis Canyon about 8 km north of South Entrance to Yellowstone National Park. Dark gray slightly perlitic glass with abundant phenocrysts of deeply embayed and sieved plagioclase, less abundant quartz, sanidine, and ferroaugite, and accessory minerals.

Lava Creek Tuff

4. Obsidian from glassy densely welded base of member B. From quarry at east end of dam on Grassy Lake Reservoir. Rock is black dense obsidian with rare small rhyolitic lithic inclusions. Sparse small phenocrysts of quartz, sanidine, oligoclase, and accessory minerals.
5. Devitrified densely welded tuff of member B. From about 3 m above base of unit at same locality as number 4. Light-gray platy-jointed rock with some lithophysal minerals and slight limonitic staining on some horizontal surfaces. Sparse small phenocrysts like number 4.
6. Devitrified densely welded tuff of member B. From bank along road about 0.8 km southwest of Norris Geyser Basin. Pinkish phenocryst-rich rock with abundant large quartz, sanidine, and oligoclase, as well as accessory minerals.
7. Pumice from glassy nonwelded top of member B. From near top of Sheepeater Cliffs 2.6 km north-northeast of Bunsen Peak. Nearly white coarsely vesicular pumice lumps separated from ash-flow matrix. Abundant large phenocrysts of quartz, sanidine, and oligoclase, as well as accessory minerals.

Plateau Rhyolite

Upper Basin Member

8. Vitrophyric fused fallout tuff of Sulphur Creek (agglutinate), about 11 m above base. From base of cliff about 0.9 km north of Sevenmile Hole, Grand Canyon of the Yellowstone. Black slightly perlitic glass with moderately abundant phenocrysts of quartz and oligoclase, as well as accessory minerals, in welded shard and pumice matrix.
9. Perlitic glassy rhyolite of Biscuit Basin flow. From low ridge about 300 m west of Morning Glory Pool, Upper Geyser Basin. Dark gray perlitic vitrophyre with phenocrysts of plagioclase, quartz, sanidine, ferroaugite, and accessory minerals.
10. Glassy rhyolite of Scaup Lake flow. From roadcut about 0.9 km west of Scaup Lake. Gray and red slightly perlitic glass with some small nonhydrated cores. Abundant phenocrysts of oligoclase, quartz, sanidine and accessory minerals.

Central Plateau Member

11. Glassy slightly welded ash-flow tuff of Bluff Point. From Shoshone Lake Trail just west of trailhead at Dogshead Creek, north of Lewis Lake. Pale brownish glassy matrix with inclusions of slightly compressed white silky pumice, small angular inclusions of black vitrophyre, and moderately abundant phenocrysts of quartz, sanidine, oligoclase, and accessory minerals.
12. Glassy rhyolite of Spring Creek flow. From just west of trail to Divide Lookout, 0.8 km southwest of Norris Pass. Highly fractured black obsidian. Moderate phenocrysts of quartz, sanidine, and accessory minerals.
13. Glassy rhyolite of Elephant Back flow. From Elephant Back about 2 km northwest of Fishing Bridge. Fractured black obsidian vitrophyre with slight pitchstone luster bordering fractures. Moderate phenocrysts of quartz, sanidine, and accessory minerals.
14. Glassy rhyolite of Summit Lake flow. From top of bluffs on west side of Black Sand Basin (Upper Geyser Basin). Fractured black obsidian vitrophyre. Moderate phenocrysts of quartz, sanidine, and accessory minerals.
15. Glassy rhyolite of West Yellowstone flow. From logging road on west side of Madison Plateau about 3 km south-southwest of West Yellowstone. Fractured black obsidian vitrophyre with slight pitchstone luster around fractures. Moderate phenocrysts of quartz, sanidine, and accessory minerals.

Roaring Mountain Member

16. Glassy rhyolite of Gibbon River flow. From upper part of cliffs on east side of Gibbon Canyon about 1 km southeast of Beryl Spring. Reddish-brown obsidian vitrophyre with moderate phenocrysts of quartz, sanidine, and accessory minerals.
17. Glassy rhyolite of Gibbon River flow. From low plateau about 2 km southeast of Norris Geyser Basin. Black obsidian with sparse small phenocrysts of quartz, sanidine, ferroaugite, and magnetite, rare 1 mm spherulites.
18. Glassy rhyolite of Cougar Creek flow. From south side of outcrop area of the flow, south of Cougar Creek. Black obsidian with small phenocrysts of quartz, sanidine, and accessory minerals.
19. Glassy rhyolite of Obsidian Cliff flow. From 0.6 km east of Lake of the Woods. Black obsidian with no phenocrysts.

Table 4.—*New chemical analyses of rocks from the Yellowstone Plateau volcanic field.*—Continued

20. Crystalline rhyolite of Obsidian Cliff flow. From 0.3 km east of Crystal Spring. Gray conspicuously flow-laminated rhyolite with cavities filled by euhedral sanidine, tridymite, and magnetite. No phenocrysts.
21. Glassy rhyolite of Crystal Spring flow. From 350 m northeast of Obsidian Lake. Black obsidian with no phenocrysts but notably abundant crystallites.

Basalt of Warm River

22. Medium-gray dense basalt from Fish Creek Road about 1.3 km northeast of Warm River. Sparse phenocrysts less than 1 mm of embayed olivine (with thin iddingsite rims) and plagioclase; microphenocrysts of embayed olivine (with iddingsite rims); subophitic to ophitic, plagioclase, clinopyroxene, and opaque oxides.

Undine Falls Basalt

23. Light-gray dense basalt from hill 6805 above lower Rescue Creek. Abundant 1-cm phenocrysts of deeply embayed and sieved plagioclase, and less abundant 1-3 mm olivine (some cumulophyric with plagioclase); microphenocrysts of olivine and clinopyroxene; intergranular plagioclase, clinopyroxene and opaque oxides.
24. Medium-gray dense basalt from Blacktail Deer Creek at about elevation 6350. Sparse $^1/_2$-1 mm olivine and plagioclase phenocrysts (some cumulophyric); microphenocrysts of olivine; subophitic to intergranular plagioclase, clinopyroxene, and opaque oxides.
25. Light-gray slightly vesicular basalt from bluff west of Oxbow Creek about 3.7 km northwest of Phantom Lake. Abundant phenocrysts of embayed plagioclase, typically as large as 5 mm but locally to more than 3 cm long; less abundant 1-mm olivine phenocrysts; microphenocrysts (less than 1 mm) of plagioclase and olivine; ophitic plagioclase, clinopyroxene, and opaque oxides.

Swan Lake Flat Basalt

26. Light-gray dense basalt from low bluffs east of Obsidian Creek, about 1.5 km south of Gardner River. Sparse 1-cm plagioclase phenocrysts, abundant 1-mm and smaller phenocrysts and microphenocrysts of olivine and plagioclase; intergranular plagioclase, clinopyroxene, and opaque oxides.

Basalt of Geode Creek

27. Medium-gray aphanitic basalt with irregular lighter-gray mottling and streaking, from small knob 230 m north of bend in Mammoth-Tower Junction Road about 1.8 km southwest of mouth of Hellroaring Creek. Contains moderately abundant small ($^1/_2$ mm) embayed and sieved plagioclase and euhedral olivine phenocrysts, sparse very large embayed plagioclase (as large as 5 cm), sparse quartz xenocrysts jacketed by augite; outcrop has rare granitic xenoliths (not in analyzed sample); very fine-grained granular plagioclase, clinopyroxene, and opaque oxides.

Osprey Basalt

28. Dark-brownish-gray dense aphanitic basalt from roadcut 150 m west of parking area for Undine Falls. Rare zeolite- and calcite-filled vesicles. Very sparse highly embayed and sieved plagioclase phenocrysts; sparse microphenocrysts of olivine; intersertal plagioclase, clinopyroxene, opaque oxides, and brown glass.

Madison River Basalt

29. Light gray dense basalt from small knob (elevation 7447) north of Cougar Creek 3.2 km east-northeast of Patrol Cabin. Small ($^1/_2$-1 mm) phenocrysts of plagioclase and olivine, some cumulophyric; ophitic plagioclase, clinopyroxene, and opaque oxides.
30. Medium-gray dense basalt from south of Cougar Creek, in gully just east of Cougar Creek rhyolite flow. Sparse 1-mm phenocrysts of plagioclase; microphenocrysts of plagioclase and olivine, commonly rimmed by clinopyroxene; sparse zeolite-filled vesicles; intergranular plagioclase, clinopyroxene, and opaque oxides.

Basalt of Snake River Group

31. Light-gray slightly vesicular coarse-grained basalt from western summit of ridge about 1.5 km east of Antelope Flat. Small ($^1/_4$-1 mm) phenocrysts of plagioclase and olivine; ophitic plagioclase, clinopyroxene, and opaque oxides.

Table 5.—*Chemical analyses of Huckleberry Ridge Tuff, Yellowstone Group.*

Member	**Member A**					**Member A(?)**				**Member B**			**Member B(?)**	
Lithology	Fused basal fallout	Basal welded-tuff vitrophyre			Devitrified	Devitrified			Vitrophyre	Devitrified			Devitrified	
Locality	Mt. Everts		Grayling Creek			Grayling Creek		Tepee Creek	Henrys Lake Mts.	Whits Lakes	Tepee Creek	Campanula Creek	Red Canyon	Grayling Creek
Source	Table 4	Boyd (1961)	Witkind (1969)			Witkind (1969)			Hamilton & Leopold (1962)	Witkind (1969)			Witkind (1969)	
Sample No.	7YC-288A	YP54A2	WG98	WG147	WG146	WG94	WG95	WG102	YS14	WG92	WG100	WG135	WG29	WG149
Analyses in weight percent, recalculated to 100 percent, less volatiles, all Fe as FeO														
SiO_2	76.47	75.19	75.76	76.27	75.23	76.70	76.79	76.29	76.06	75.36	74.24	76.51	75.84	75.62
Al_2O_3	12.52	12.47	12.80	12.73	13.06	12.58	12.65	12.83	12.68	13.18	13.25	12.65	13.01	12.96
ΣFeO	1.49	1.71	1.69	1.84	1.94	1.56	1.65	1.79	1.67	1.77	2.55	1.66	1.88	1.84
MgO	.00	.13	.17	.20	.17	.18	.24	.20	.07	.14	.46	.15	.03	.20
CaO	.52	.73	.52	.36	.47	.29	.39	.40	.59	.38	.99	.43	.35	.54
Na_2O	3.73	4.12	3.51	3.54	3.85	3.55	3.44	3.44	3.62	3.72	3.54	3.54	3.43	3.60
K_2O	5.12	5.42	5.37	4.85	5.06	4.97	4.65	4.85	5.13	5.23	4.65	4.86	5.24	5.04
TiO_2	.10	.14	.13	.15	.17	.12	.13	.15	.12	.16	.24	.13	.13	.15
P_2O_5	.00	.00	.03	.03	.01	.01	——	.01	.02	.01	.02	.01	——	.02
MnO	.04	.04	.03	.03	.04	.02	.05	.04	.04	.04	.06	.06	.08	.03
ZrO_2	——	.04	——	——	——	——	——	——	——	——	——	——	——	——
CIPW norms from recalculated analyses														
q	32.90	28.64	32.20	34.70	30.72	34.92	36.44	35.19	32.70	31.29	31.10	34.87	33.49	32.56
c	——	——	.35	1.08	.43	.85	1.23	1.22	.13	.72	.64	.82	1.05	.66
zr	——	.06	——	——	——	——	——	——	——	——	——	——	——	——
or	30.27	32.01	31.72	28.65	29.91	29.38	27.50	28.66	30.32	30.92	27.49	28.71	30.99	29.79
ab	31.57	33.97	29.69	29.92	32.56	30.05	29.11	29.07	30.65	31.50	29.96	29.97	29.02	30.47
an	2.30	——	2.36	1.61	2.24	1.39	1.96	1.94	2.81	1.83	4.79	2.04	1.75	2.52
ns	——	.21	——	——	——	——	——	——	——	——	——	——	——	——
wo	.13	1.51	——	——	——	——	——	——	——	——	——	——	——	——
en	——	.34	.41	.50	.43	.46	.61	.50	.18	.35	1.13	.38	.08	.49
fs	2.64	2.99	2.94	3.18	3.36	2.71	2.91	3.11	2.94	3.06	4.39	2.94	3.38	3.18
il	.19	.27	.26	.29	.33	.23	.25	.29	.23	.31	.46	.25	.25	.29
ap	——	——	.07	.07	.02	.02	——	.02	.05	.02	.05	.02	——	.05
Salic	97.04	94.69	96.32	95.96	95.86	96.59	96.24	96.08	96.61	96.26	93.98	96.41	96.30	96.00
Femic	2.96	5.31	3.68	4.04	4.14	3.42	3.77	3.92	3.40	3.74	6.03	3.59	3.71	4.01
di	.27	3.18	——	——	——	——	——	——	——	——	——	——	——	——
wo	.13	1.51	——	——	——	——	——	——	——	——	——	——	——	——
en	.00	.17	——	——	——	——	——	——	——	——	——	——	——	——
fs	.14	1.50	——	——	——	——	——	——	——	——	——	——	——	——
hy	2.50	1.66	3.35	3.68	3.79	3.17	3.52	3.61	3.12	3.41	5.52	3.32	3.46	3.67
en	.00	.17	.41	.50	.43	.46	.61	.50	.18	.35	1.13	.38	.08	.49
fs	2.50	1.49	2.94	3.18	3.36	2.71	2.91	3.11	2.94	3.06	4.39	2.94	3.38	3.18

Table 6.—*Chemical analyses of Mesa Falls Tuff, Yellowstone Group.*

Lithology	Glassy pumice	Devitrified		
Locality	Ashton Highway		Upper Mesa Falls	
Source	Table 4	Hamilton (1965)		
Sample No.	8YC-460A	YS34	H3571	H3573
Analyses recalculated to 100 percent, less volatiles, all Fe as FeO				
SiO_2	74.90	76.04	75.28	76.30
Al_2O_3	13.88	12.75	12.79	12.57
ΣFeO	1.71	1.67	2.00	1.67
MgO	.21	.07	.13	.08
CaO	.60	.59	.85	.57
Na_2O	2.87	3.29	3.37	3.52
K_2O	5.60	5.35	5.28	5.04
TiO_2	.17	.18	.23	.17
P_2O_5	.02	.02	.02	.03
MnO	.04	.04	.04	.05
CIPW norms from recalculated analyses				
q	33.91	33.86	31.96	33.98
c	2.06	.53	.04	.36
or	33.06	31.59	31.23	29.77
ab	24.29	27.81	28.53	29.78
an	2.85	2.81	4.06	2.65
en	.52	.18	.33	.20
fs	2.95	2.84	3.37	2.87
il	.32	.34	.44	.33
ap	.05	.05	.05	.07
Salic	96.17	96.60	95.82	96.54
Femic	3.84	3.41	4.19	3.47
hy	3.47	3.02	3.70	3.07
en	.52	.18	.33	.20
fs	2.95	2.84	3.37	2.87

Table 7.—*Chemical analyses of first- and second-cycle rhyolitic lavas (Big Bend Ridge Rhyolite and Island Park Rhyolite).*

Formation	Big Bend Ridge Rhyolite	Island Park Rhyolite
Flow	Bishop Mountain flow	Warm River Butte flow
Lithology	Devitrified	
Locality	Bishop Mountain	Warm River Butte
Source	Hamilton (1965)	
Sample No.	YS79	YS66A
Analyses recalculated to 100 percent, less volatiles, all Fe as FeO		
SiO_2	77.61	77.51
Al_2O_3	12.24	11.94
ΣFeO	1.07	1.39
MgO	.08	.08
CaO	.48	.53
Na_2O	3.31	3.25
K_2O	5.00	5.09
TiO_2	.15	.17
P_2O_5	.01	.01
MnO	.03	.02
CIPW norms from recalculated analyses		
q	37.26	36.84
c	.52	.14
or	29.57	30.07
ab	28.03	27.52
an	2.33	2.57
en	.20	.20
fs	1.77	2.31
il	.29	.32
ap	.02	.02
Salic	97.71	97.14
Femic	2.28	2.85
hy	1.97	2.51
en	.20	.20
fs	1.77	2.31

Table 8.—*Chemical analyses of Mount Jackson Rhyolite and Lewis Canyon Rhyolite.*

Formation	Mount Jackson Rhyolite	Lewis Canyon Rhyolite
Flow	Mount Haynes flow	Lewis Canyon flow
Lithology	Crystallized	Glassy
Locality	Madison Canyon	Lewis Canyon
Source	Hamilton (1963)	Table 4
Sample No.	YS24	7YC-246
Analyses recalculated to 100 percent, less volatiles, all Fe as FeO		
SiO_2	77.76	74.25
Al_2O_3	12.27	14.17
ΣFeO	1.26	2.05
MgO	.08	.07
CaO	.22	.84
Na_2O	3.62	3.31
K_2O	4.63	5.01
TiO_2	.10	.23
P_2O_5	.02	.02
MnO	.04	.05
CIPW norms from recalculated analyses		
q	37.44	32.40
c	.95	1.83
or	27.34	29.61
ab	30.64	27.98
an	.97	4.04
en	.20	.18
fs	2.22	3.48
il	.19	.43
ap	.05	.05
Salic	97.34	95.86
Femic	2.66	4.15
hy	2.42	3.66
en	.20	.18
fs	2.22	3.48

Table 9.—*Chemical analyses of Lava Creek Tuff, Yellowstone Group.*

Member	Member A			Member B								
Lithology	Devitrified			Obsidian	Devitrified							Glassy pumice
Locality	Cougar Creek	S. Fork Madison River	Norris, core C-2, 67.4 m	Grassy Lake Reservoir		Norris Basin	Hebgen Lake	Island Park	Macks Inn	Madison Canyon	Norris, core C-2, 21.4 m	Top of Sheepeater Cliffs
Source	Hamilton (1963)		Fenner (1936)	Table 4			Witkind (1969)	Hamilton (1965)		Hamilton (1963)	Fenner (1936)	Table 4
Sample No.	YS4	YS8	YP750	6YC-13D	6YC-13H	6YC-38	WG58	YS57F	YS55	YS20A	YP669	8YC-411
Analyses recalculated to 100 percent, less volatiles, all Fe as FeO												
SiO_2	76.88	77.89	77.36	76.48	76.45	78.20	75.28	77.15	76.29	76.54	79.06	77.29
Al_2O_3	12.58	12.60	12.24	12.41	12.84	12.17	13.14	12.24	12.65	12.77	11.76	12.65
ΣFeO	1.50	.90	1.23	1.59	1.59	1.03	1.79	1.46	1.53	1.64	.94	1.49
MgO	.15	.07	.12	.00	.01	.00	.11	.03	.03	.16	.06	.04
CaO	.27	.24	.28	.49	.27	.26	.49	.51	.45	.34	.29	.45
Na_2O	3.35	3.33	3.00	3.71	3.40	3.26	3.84	3.18	3.62	3.32	2.60	2.50
K_2O	5.07	4.84	5.53	5.13	5.24	4.94	5.15	5.22	5.24	5.03	5.01	5.41
TiO_2	.14	.10	.13	.13	.13	.10	.16	.18	.16	.14	.11	.11
P_2O_5	.02	.02	.07	.01	.02	.01	.01	.01	.01	.02	.13	.02
MnO	.03	.02	.01	.04	.04	.02	.03	.02	.02	.04	——	.03
ZrO_2	——	——	.03	——	——	——	——	——	——	——	.03	——
CIPW norms from recalculated analyses												
q	36.08	38.78	37.24	33.02	34.81	38.97	30.65	36.48	33.03	35.77	43.73	39.87
c	1.13	1.50	.99	——	1.13	1.00	.38	.45	.22	1.30	1.85	1.91
zr	——	——	.05	——	——	——	——	——			.05	
or	29.97	28.58	32.67	30.34	30.97	29.20	30.45	30.86	30.94	29.72	29.60	31.98
ab	28.32	28.14	25.35	31.43	28.76	27.62	32.49	26.87	30.64	28.08	21.96	21.18
an	1.23	1.07	.94	2.03	1.22	1.24	2.34	2.49	2.19	1.57	.60	2.09
wo	——	——	——	.15	——	——	——	——				
en	.38	.18	.30	——	.03	——	.28	.08	.08	.40	.15	.10
fs	2.58	1.52	2.06	2.77	2.78	1.77	3.07	2.42	2.58	2.85	1.54	2.61
il	.27	.19	.25	.25	.25	.19	.31	.35	.31	.27	.21	.22
ap	.05	.05	.17	.02	.05	.02	.02	.02	.02	.05	.31	.05
Salic	96.73	98.07	97.24	96.82	96.89	98.03	96.31	97.15	97.02	96.44	97.79	97.03
Femic	3.28	1.94	2.78	3.19	3.11	1.98	3.68	2.87	2.99	3.57	2.21	2.98
di	——	——	——	.32	——	——	——	——	——	——	——	——
wo	——	——	——	.15	——	——	——	——	——	——	——	——
en	——	——	——	.00	——	——	——	——	——	——	——	——
fs	——	——	——	.17	——	——	——	——	——	——	——	——
hy	2.96	1.70	2.36	2.60	2.81	1.77	3.35	2.50	2.66	3.25	1.69	2.71
en	.38	.18	.30	.00	.03	.00	.28	.08	.08	.40	.15	.10
fs	2.58	1.52	2.06	2.60	2.78	1.77	3.07	2.42	2.58	2.85	1.54	2.61

Table 10.—*Chemical analyses of Upper Basin and Obsidian Creek Members, Plateau Rhyolite.*

Member	Upper Basin					Obsidian Creek
Flow or Unit	Tuff of Sulphur Creek	Biscuit Basin flow			Scaup Lake flow	Rhyolite, mixed lavas of Gardner River
Lithology	Glassy	Glassy			Glassy	Crystallized
Locality	Sulphur Creek	Upper Geyser Basin	Midway Geyser Basin	Upper Geyser Basin	Scaup Lake	Gardner River
Source	Table 4	Table 4	Boyd (1961)	Fenner (1936)	Table 4	Fenner (1938)
Sample No.	8YC-477B	6YC-106	YP54A3	YP279	8YC-441	YP914
Analyses recalculated to 100 percent, less volatiles, all Fe as FeO						
SiO_2	76.59	72.36	76.32	71.90	75.36	75.84
Al_2O_3	12.51	13.58	12.22	14.02	12.91	12.72
ΣFeO	1.30	3.31	1.54	3.62	1.65	1.73
MgO	.08	.31	.11	.57	.20	.40
CaO	.69	1.56	.70	2.03	.87	.83
Na_2O	3.20	3.39	3.45	4.25	3.31	3.45
K_2O	5.17	4.86	5.31	2.95	5.27	4.81
TiO_2	.31	.47	.27	.50	.27	.16
P_2O_5	.10	.09	——	.09	.12	.03
MnO	.04	.07	.04	.05	.04	.03
ZrO_2	——	——	.04	.02	——	——
CIPW norms from recalculated analyses						
q	35.97	28.04	33.42	28.22	32.96	33.72
c	.64	.13	——	.35	.49	.40
z	——	——	.06	.03	——	——
or	30.57	28.71	31.37	17.42	31.13	28.44
ab	27.08	28.66	29.19	36.00	27.97	29.18
an	2.76	7.15	2.18	9.49	3.50	3.93
wo	——	——	.55	——	——	——
en	.21	.77	.28	1.42	.49	1.00
fs	1.95	5.44	2.46	5.92	2.66	2.96
il	.59	.89	.51	.96	.51	.31
ap	.25	.22	——	.20	.29	.07
Salic	97.02	92.69	96.22	91.51	96.05	95.67
Femic	3.00	7.32	3.80	8.50	3.95	4.34
di	——	——	1.15	——	——	——
wo	——	——	.55	——	——	——
en	——	——	.06	——	——	——
fs	——	——	.54	——	——	——
hy	2.16	6.21	2.14	7.34	3.15	3.96
en	.21	.77	.22	1.42	.49	1.00
fs	1.95	5.44	1.92	5.92	2.66	2.96

Table 11.—*Chemical analysis of tuff of Bluff Point, Central Plateau Member, Plateau Rhyolite.*

Lithology	Glassy
Locality	Dogshead Creek
Source	Table 4
Sample No.	7YC-271
Analysis recalculated to 100 percent, less volatiles, all Fe as FeO	
SiO_2	76.44
Al_2O_3	12.93
ΣFeO	1.46
MgO	.04
CaO	.45
Na_2O	3.21
K_2O	5.24
TiO_2	.17
P_2O_5	.01
MnO	.06
CIPW norm from recalculated analysis	
q	35.59
c	1.20
or	30.94
ab	27.16
an	2.15
en	.10
fs	2.52
il	.32
ap	.03
Salic	97.04
Femic	2.97
hy	2.62
en	.10
fs	2.52

Table 12.—*Chemical analyses of rhyolite flows of Central Plateau Member, Plateau Rhyolite.*

Flow	West Yellowstone					Summit Lake	Elephant Back	Spring Creek
Lithology	Glassy					Glassy	Glassy	Glassy
Locality	S. of West Yellowst.	Reas Pass	S. of West Yellowst.	Reas Pass		W. of Upper Geyser Basin	Elephant Back	Divide Lookout Trail
Source	Table 4	Hamilton (1963)			Boyd (1961)	Table 4	Table 4	Table 4
Sample No.	6YC-169	YS26A	YS7A2	YS27	YP54A1	6YC-73	8YC-374C	6YC-113
Analyses recalculated to 100 percent, less volatiles, all Fe as FeO								
SiO_2	77.01	77.30	76.53	77.20	76.56	76.93	76.84	76.95
Al_2O_3	12.04	12.15	12.12	11.98	11.68	12.12	12.29	12.17
ΣFeO	1.61	1.48	1.94	1.66	1.62	1.54	1.40	1.42
MgO	.01	.06	.18	.13	.33	.01	.00	.03
CaO	.44	.35	.58	.38	.49	.42	.44	.45
Na_2O	3.57	3.37	3.43	3.42	4.00	3.67	3.49	3.57
K_2O	5.13	5.11	4.95	5.03	5.09	5.13	5.32	5.19
TiO_2	.14	.12	.20	.14	.15	.13	.16	.14
P_2O_5	.01	.02	.02	.01	——	.01	.01	.02
MnO	.04	.04	.06	.04	.05	.04	.04	.04
ZrO_2	——	——	——	——	.04	——	——	——
CIPW norms from recalculated analyses								
q	34.51	36.20	34.66	35.73	32.71	33.97	34.17	34.27
c	——	0.50	0.12	0.23	——	——	0.01	——
z	——	——	——	——	0.06	——	——	——
or	30.32	30.17	29.23	29.74	30.05	30.30	31.43	30.70
ab	31.20	28.51	29.05	28.96	31.75	31.06	29.55	30.24
an	1.67	1.59	2.72	1.83	——	1.44	2.14	1.83
ns	——	——	——	——	0.49	——	——	——
wo	0.19	——	——	——	1.01	0.25	——	0.12
en	0.03	0.15	0.45	0.33	0.82	0.03	——	0.08
fs	2.79	2.59	3.34	2.89	2.82	2.69	2.38	2.45
il	0.27	0.23	0.36	0.27	0.28	0.25	0.31	0.27
ap	0.02	0.05	0.05	0.02	——	0.02	0.02	0.05
Salic	96.70	96.97	95.78	96.49	94.57	96.77	97.30	97.04
Femic	3.30	3.02	4.20	3.51	5.42	3.24	2.71	2.97
di	0.42	——	——	——	2.08	0.54	——	0.26
wo	0.19	——	——	——	1.01	0.25	——	0.12
en	0.01	——	——	——	0.24	0.01	——	0.01
fs	0.22	——	——	——	0.83	0.28	——	0.13
hy	2.59	2.74	3.79	3.22	2.57	2.43	2.38	2.39
en	0.02	0.15	0.45	0.33	0.58	0.02	0.00	0.07
fs	2.57	2.59	3.34	2.89	1.99	2.41	2.38	2.32

Table 13.—*Chemical analyses of Roaring Mountain Member, Plateau Rhyolite.*

Flow	Gibbon River		Obsidian Cliff			Cougar Creek		Crystal Spring
Lithology	Glassy		Obsidian		Crystalline	Obsidian		Obsidian
Locality	North of Gibbon Hill	Gibbon Canyon	Obsidian Cliff	Lake of the Woods	Crystal Spring	Cougar Creek		East of Crystal Spring
Source	Table 4		Bowen (1935)	Table 4		Table 4	Hamilton (1963)	Table 4
Sample No.	65YR-7	6YC-93	[not given]	6YC-64	6YC-153	6YC-43B	YS3	8YC-396A
Analyses recalculated to 100 percent, less volatiles, all Fe as FeO								
SiO_2	77.14	77.17	76.95	77.09	77.13	77.01	77.27	77.07
Al_2O_3	12.34	12.34	12.12	12.35	12.45	12.49	12.34	12.41
ΣFeO	1.18	1.19	1.32	1.18	1.16	1.18	1.30	1.24
MgO	.00	.04	.10	.01	.00	.02	.21	.06
CaO	0.38	.42	.57	.42	.32	.56	.58	.34
Na_2O	4.04	3.75	3.80	3.89	3.77	3.37	3.03	3.82
K_2O	4.81	4.95	4.94	4.93	5.05	5.23	5.06	4.92
TiO_2	.07	.10	.08	.08	.08	.11	.13	.09
P_2O_5	.01	.01	.09	.01	.01	.01	.04	.01
MnO	.03	.03	.02	.03	.02	.02	.03	.03
ZrO_2	——	——	.01	——	——	——	——	——
CIPW norms from recalculated analyses								
q	33.58	34.54	33.95	33.81	34.27	35.28	37.82	34.20
c	——	.07	——	——	.22	.29	.92	.19
zr	——	——	.02	——	——	——	——	——
or	28.45	29.24	29.20	29.16	29.84	30.90	29.88	29.09
ab	34.17	31.74	32.14	32.91	31.92	28.50	25.68	32.36
an	1.32	2.03	1.42	1.68	1.53	2.73	2.60	1.64
wo	.22	——	.35	.15	——	——	——	——
en	——	.10	.25	.03	——	.05	.53	.15
fs	2.10	2.08	2.32	2.10	2.04	2.02	2.24	2.18
il	.13	.19	.15	.15	.15	.21	.25	.17
ap	.02	.02	.21	.02	.02	.02	.10	.02
Salic	97.52	97.62	96.73	97.56	97.78	97.70	96.90	97.48
Femic	2.47	2.39	3.28	2.45	2.21	2.30	3.12	2.52
di	.46	——	.73	.33	——	——	——	——
wo	.22	——	.35	.15	——	——	——	——
en	.00	——	.04	.01	——	——	——	——
fs	.24	——	.34	.17	——	——	——	——
hy	1.86	2.18	2.19	1.95	2.04	2.07	2.77	2.33
en	.00	.10	.21	.02	.00	.05	.53	.15
fs	1.86	2.08	1.98	1.93	2.04	2.02	2.24	2.18

Table 14.—*Minor and trace-element analyses of rhyolites of the Yellowstone Plateau volcanic field, in parts per million.*

Cycle	Rhyolites of first volcanic cycle								Rhyolites of second volcanic cycle				
Formation	Rhyolite of Snake River Butte	Huckleberry Ridge Tuff						Big Bend Ridge Rhyolite		Mesa Falls Tuff		Island Park Rhyolite	
Member, flow, or unit	Snake River Butte flow	Fallout	Member A	Member B			Member C	Blue Creek flow	Bishop Mountain flow			Lookout Butte dome	Elk Butte dome
Lithology	perlite	fused obsidian	vitrophyre	light pumice	dark pumice	glassy, partly welded	devitrified pumice	obsidian cores	devitrified	pumice		pumice	glassy
Locality	Snake River Butte	Mount Everts	Teton Pass Road	Lionhead Peak	Ashton	Terrace Mountain	Lizard Creek	Blue Creek	Bishop Mountain	North of Ashton		Lookout Butte	Elk Butte
Sample No.	78YH-13	7YC-288A	0YC-575B	74MR-135A	74IP-149B	65YR-23	7YC-194	78YH-10A	74IP-93	7YC-182	8YC-460A	78YH-8	0YC-559
Cl	620	490	720	270	1,360	——	<5	380	——	——	1,100	600	——
F	——	1,300	470	250	1,000	——	70	——	——	——	1,800	——	——
Be	——	5.0	3.9	4.5	2.7	——	2.7	——	——	——	——	——	——
Zr	330	320	220	310	750	450	380	390	300	410	310	330	270
Hf	8.7	8.3	9.5	9.6	16	10	9.7	9.9	6.4	9.9	7.1	7.9	6.4
Nb	——	56	58	55	51	——	29	——	——	——	——	——	——
Ta	3.1	4.9	4.0	4.4	3.0	2.2	2.1	3.4	7.0	4.2	3.5	3.5	4.0
P	130	140	110	120	480	——	150	87	——	——	90	130	830
Ti	1,200	1,100	900	1,100	2,700	——	1,100	1,300	——	——	960	1,300	540
Mn	360	290	220	330	750	——	250	400	——	——	310	410	150
Zn	70	83	82	82	110	80	80	73	43	80	51	56	49
Sc	2.7	1.6	1.9	1.8	6.0	2.9	2.8	3.1	1.6	3.3	2.4	2.9	1.8
Y	48	64	55	63	57	——	31	61	——	——	——	66	——
La	86	94	100	110	92	68	61	92	58	110	92	79	84
Ce	150	180	200	210	160	130	120	160	110	190	160	130	150
Nd	61	71	81	89	74	54	54	64	55	79	61	56	61
Sm	12	16	17	17	16	14	11	13	15	19	12	12	12
Eu	1.5	.61	1.1	0.84	3.3	2.3	2.3	.92	.21	.85	.94	1.2	.44
Gd	9.4	13	11	13	11	8.6	7.9	9.1	14	13	10	8.9	9.3
Tb	1.3	2.5	2.2	2.5	2.1	1.6	1.5	1.4	2.5	2.5	1.8	1.3	1.8
Ho	1.6	2.7	2.9	2.8	2.4	1.8	1.9	1.5	3.5	3.3	1.6	1.5	2.1
Tm	.74	1.2	1.0	1.1	.92	.70	.64	.87	1.3	1.2	.83	.83	.94
Yb	6.1	8.0	7.1	7.9	6.9	4.4	4.5	6.3	8.0	7.6	5.8	6.3	6.7
Lu	.81	1.1	.96	1.0	.92	.63	.62	.86	1.1	1.1	.78	.86	.91
Li	——	59	41	59	23	——	32	——	——	——	——	——	——
Th	24	32	29	32	19	19	19	26	49	35	29	26	31
U	5.4	8.6	6.9	7.2	4.1	6.5	5.4	6.0	16	7.2	6.3	5.5	7.3
Sr	46	<5	<5	<5	55	——	100	27	——	——	26	36	——
Ba	1,200	210	720	440	2,900	1,900	1,500	650	120	450	490	770	190
Rb	170	220	170	210	120	150	140	190	330	180	180	170	190
Cs	3.2	4.0	3.3	3.3	1.6	5.2	2.9	3.5	5.8	3.9	3.0	2.9	3.2

Table 14.—*Minor and trace-element analyses of rhyolites of the Yellowstone Plateau volcanic field, in parts per million.*—Continued

Cycle, formation	Plateau Rhyolite of third volcanic cycle								
Member	Upper Basin		Obsidian Creek	Roaring Mountain		Mallard Lake	Central Plateau		
Flow or unit	Dunraven Road flow	Biscuit Basin flow	Apollinaris Spring dome	Gibbon River flow	Crystal Spring flow	Mallard Lake flow	West Thumb flow	Solfatara Plateau flow	Pitchstone Plateau flow
Lithology	pumice	perlite	glassy	obsidian	obsidian	devitrified	vitrophyre	vitrophyre	pumice
Locality	Dunraven Pass Road	Upper Geyser Basin	Hill 8232	North of Gibbon Hill	East of Crystal Spring	Craig Pass	DeLacy Creek	Solfatara Plateau	Phantom Fumarole
Sample No.	9YC-530	6YC-106	0YC-609	65YR-7	8YC-396A	9YC-539	8YC-376	3YC-15	0YC-567
Cl	550	400	510	980	7,400	——	530	620	390
F	1,100	970	400	2,200	1,900	——	1,500	1,500	1,500
Be	3.5	3.8	3.8	7.6	5.4	——	4.2	4.4	5.1
Zr	190	310	340	150	170	310	260	320	270
Hf	6.7	11	11	8.4	7.2	7.8	8.0	13	9.6
Nb	45	52	47	64	51	——	50	65	67
Ta	2.8	3.3	2.9	5.1	3.5	4.0	4.3	5.0	4.9
P	140	460	140	100	110	<220	120	110	110
Ti	900	2,900	1,100	600	780	900	1,100	1,000	900
Mn	270	560	360	220	220	<150	290	340	340
Zn	51	84	96	76	66	42	58	88	83
Sc	3.6	6.9	2.2	.91	1.3	2.0	2.2	1.5	1.2
Y	13	35	28	61	59	——	36	54	46
La	61	74	70	43	61	63	82	98	94
Ce	120	140	140	96	120	130	170	200	180
Nd	47	58	60	44	45	39	63	68	76
Sm	10	12	12	13	12	9.3	14	16	16
Eu	.64	1.7	1.2	.07	.16	.67	.68	.66	.49
Gd	7.9	9.1	10	11	9.7	6.3	9.8	11	12
Tb	1.5	1.9	1.8	2.8	2.1	1.0	2.1	2.6	2.5
Ho	1.5	1.5	1.4	2.9	1.9	1.5	2.0	2.5	2.3
Tm	.84	1.2	.87	1.4	1.2	.55	1.2	1.4	1.4
Yb	4.8	5.8	5.2	9.1	6.7	3.5	6.4	7.9	8.1
Lu	.67	.81	.69	1.2	.94	.51	.89	1.0	1.1
Li	53	30	45	90	68	——	44	47	46
Th	21	22	20	36	30	27	29	33	28
U	4.8	5.5	4.2	10	6.4	7.3	7.9	7.5	7.3
Sr	<5	85	8	1.8	<5	——	<5	<5	<5
Ba	690	960	810	34	<300	380	360	230	<300
Rb	160	150	170	300	200	190	190	230	200
Cs	3.4	3.7	3.3	6.8	4.7	2.5	4.0	3.9	4.2

Table 14.—*Minor and trace-element analyses of rhyolites of the Yellowstone Plateau volcanic field, in parts per million.*—Continued

Cycle, episode	Precaldera rhyolites and caldera-forming ash flows of third volcanic cycle									
Formation	Mount Jackson Rhyolite		Lava Creek Tuff							
Member, flow, or unit	Wapiti Lake flow	Harlequin Lake flow	Member A, lower	Member A, upper			Member B			
Lithology	perlite	glassy	devitrified		vitrophyre			glassy, nonwelded		vitrophyre
Locality	Grand Canyon	Madison Canyon	Purple Mountain		Cascade Lake	Mount Jackson	Grand Canyon	Reas Pass	Snake River	Cub Creek
Sample No.	0YC-601	0YC-634	74Y-130C	79Y-194A	8YC-371A	74Y-180B	8YC-408E	7YC-311A	7YC-185	65YR-27A
Cl	880	520	——	——	1,130	580	810	980	490	——
F	1,600	730	——	——	1,600	990	1,400	1,400	630	——
Be	4.6	4.1	——	——	5.7	3.4	5.0	5.1	4.4	——
Zr	190	280	310	200	200	290	180	240	290	420
Hf	6.8	8.4	7.6	6.7	7.7	7.9	7.1	8.4	9.2	10
Nb	56	57	——	——	75	51	62	66	65	——
Ta	4.2	3.7	6.9	7.0	5.0	3.6	5.1	5.0	4.6	3.5
P	140	150	<220	——	100	130	120	110	140	——
Ti	900	1,300	420	——	840	1,000	840	960	1,100	——
Mn	190	290	<150	——	290	280	250	290	330	——
Zn	53	68	59	48	72	67	68	75	76	72
Sc	1.1	2.2	1.3	1.3	1.4	1.7	1.4	1.4	1.5	1.9
Y	40	40	——	——	65	32	46	52	51	——
La	57	91	28	39	74	94	75	89	100	100
Ce	110	170	110	110	140	190	140	170	200	190
Nd	42	68	20	23	57	68	53	66	74	71
Sm	10	14	5.9	6.1	12	13	12	15	15	14
Eu	.26	1.0	.11	.35	.29	1.1	.30	.41	.79	1.3
Gd	7.7	11	5.1	6.6	10	10	11	12	13	10
Tb	1.8	2.1	1.2	1.1	2.3	1.9	2.1	2.3	2.2	1.8
Ho	1.7	2.2	2.1	1.4	1.9	2.1	2.4	2.7	2.4	1.7
Tm	1.0	1.0	.97	.70	1.4	.87	1.1	1.1	1.0	0.78
Yb	6.2	6.4	7.0	5.6	7.7	6.5	7.7	8.1	7.1	6.0
Lu	.86	.90	.98	.79	1.1	.83	1.0	1.1	.97	.82
Li	48	34	——	——	56	56	53	55	49	——
Th	25	28	36	27	31	25	30	31	29	25
U	6.5	6.0	14	7.9	7.3	5.2	8.2	7.3	6.2	4.4
Sr	18	23	2.7	——	<5	<5	<5	<5	<5	36
Ba	210	560	31	300	<300	570	<400	<400	380	660
Rb	150	190	290	210	220	130	220	220	180	150
Cs	3.5	5.4	3.5	1.7	4.8	2.9	4.8	3.9	3.2	2.6

study of the petrology of the Yellowstone Plateau volcanic field. Additional trace-element analyses were published by Doe and others (1982) and Hildreth and others (1991).

Trace-element chemistry clearly demonstrates that most of the Yellowstone rhyolites are highly differentiated. Apparent in these rhyolites are relatively high contents of incompatible trace elements such as Nb, Ta, U, Th, Rb, Cs, Zr, Hf, Zn, Y, and rare-earth elements and very low contents of Sr (corresponding to their low Ca contents) and Sc. Such compositions are indicative of a generally advanced state of magmatic evolution of the Yellowstone rhyolites as a whole. The general similarity of trace-element compositions of most of the voluminous rhyolites, spanning the age range of the volcanic field, indicates that large volumes of magma have differentiated along similar paths.

The trace-element data of table 14 show an extensive range in all three major ash-flow sheets of the Yellowstone Group although the ranges of major elements are much less than in such zoned andesite-rhyodacite sheets as that of Crater Lake, Ore. (Williams, 1942; Bacon, 1983) or—for the Mesa Falls and Lava Creek sheets—in such rhyolitic ash-flow sheets as those of Timber Mountain, Nev. (Lipman and others, 1966; Byers and others, 1976; Christiansen and others, 1977) and the San Juan Mountains, Colo. (Ratté and Steven, 1964; Lipman, 1975).

The results of numerous recent studies suggest that probably most voluminous ash-flow sheets are vertically zoned. Typically, the zonations include upward increase in primary phenocryst content of the erupted materials, upward decrease in SiO_2 contents of the glasses, and related vertical changes in phenocryst and matrix compositions. Petrologic and geologic relations, at least in many areas, show clearly that such zoned ash-flow sheets represent the sequential withdrawal of magmas with a vertical pre-eruptive zonation in their magma chambers, the first-erupted materials representing the more differentiated upper parts of the zoned magma bodies (compare to Lipman and others, 1966; Smith and Bailey, 1968b; Smith, 1979; Hildreth, 1979; 1981). A variety of origins has been proposed for voluminous vertically zoned magmas. Lipman and others (1966) argued for differentiation through gravitational separation of crystals in the magma chamber, and such mechanisms seem to have been important in producing magma bodies zoned from intermediate compositions to rhyolite or from rhyodacite or lower-silica rhyolite to high-silica rhyolite. Smith (1979), Hildreth (1979; 1981), and Macdonald and others (1992), however, discussed patterns of relative trace-element enrichments and depletions in zoned magmas that suggest a significant part of the differentiation to have occurred in the liquid state rather than by crystal-liquid separation.

The range of trace-element zonation in the Yellowstone ash-flow sheets is quite marked (table 14; fig. 39). In several respects the pattern of compositional variation resembles those discussed by Smith (1979), Hildreth (1979; 1981), and Macdonald and others (1992) as indicating liquid-state differentiation. For example, besides apparent roofward enrichment in most of the large-ion lithophile elements in the Yellowstone magma chambers, each had very large negative Eu anomalies and showed relative depletion of the light compared to the heavy rare-earth elements (fig. 39). Other elements and element ratios, such as Zr/Hf, which would be expected to covary in simple ways if differentiation were by fractional crystallization alone, were notably fractionated.

Besides the zonations recorded in the Yellowstone ash flows themselves, Hildreth and others (1984; 1991) also demonstrated that some of the immediate post-caldera lavas, particularly the Upper Basin Member of the Plateau Rhyolite, record an even greater extent of zonation within the magma chambers than was erupted almost instantaneously in the voluminous pyroclastic eruptions. Taken together with the ash flows of the Lava Creek Tuff, the Upper Basin Member reveals a minimum zonal range in the Lava Creek magma,

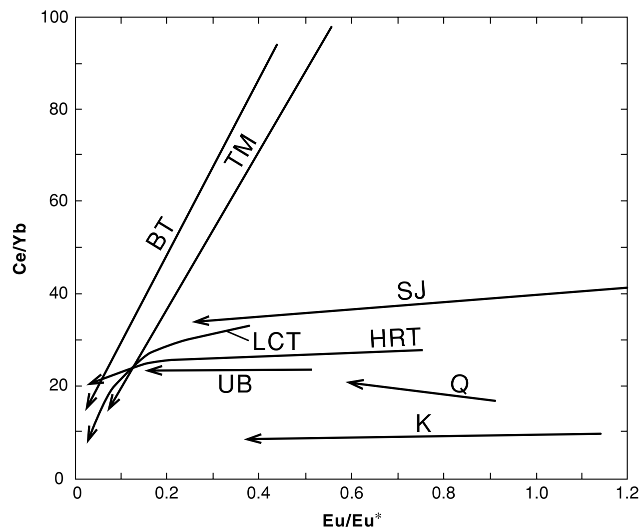


Figure 39.—Variation of rare-earth elements in rhyolites of the Yellowstone Plateau volcanic field. Fractionation of light relative to heavy rare-earth elements is represented by the ratio of Ce (a light rare-earth element) to Yb (a heavy rare-earth element); magnitude of the Eu anomaly is represented by the ratio Eu/Eu*, which is the ratio of analyzed Eu (normalized to a standard value for chondritic meteorites) to an extrapolated (Eu*) value determined by a straight line between chondrite-normalized values of the next lighter and heavier analyzed rare-earth elements. Lines represent variation curves for various units or volcanic fields, arrows pointing in the direction of increasing chemical evolution. Lines for Yellowstone (table 14; E. W. Hildreth and R. L. Christiansen, unpub. data, 1993): HRT Huckleberry Ridge Tuff; LCT, Lava Creek Tuff; UB, Upper Basin Member of Plateau Rhyolite. Lines labeled with symbols for other volcanic fields are for comparison: BT, Bishop Tuff, Calif. (Hildreth, 1979); TM, Timber Mountain and Paintbrush Tuffs, Nev. (Lipman and others, 1982); SJ, San Juan Mountains volcanic field, Colo. (Lipman and others, 1982); Q, 1932 eruption of Quizapu, Chile (Fierstein and others, 1989); K, 1912 eruption of Katmai region and Valley of Ten Thousand Smokes, Alaska (Hildreth, 1983).

not only greater than 4 percent SiO_2 but also, for example, of at least 200 ppm rubidium, 1,000 ppm barium, 0.45 in the value of the negative europium anomaly, and 820°-925°C in temperature as determined by microprobe analysis of coexisting Fe-Ti oxide phenocrysts.

From the sparse data now available, it remains uncertain just what role the Lewis Canyon Rhyolite represents in the magmatic evolution of the Yellowstone Plateau field. As noted earlier, although these lavas resemble flows of the Upper Basin Member and occur in the first-cycle Snake River caldera segment, they have K-Ar ages equivalent to units of the third volcanic cycle. The lavas may actually represent the late appearance of relatively undifferentiated first-cycle magma, and thus, may be analogous to the third-cycle Upper Basin Member in recording compositions deep within a zoned first-cycle magma chamber.

In addition to vertical pre-eruptive zonations in the magma chambers associated with each of the three great ash-flow sheets of the Yellowstone Plateau field, individual members of at least the Lava Creek Tuff show internal vertical zonations. The data in table 14 indicate that the base of member B was more differentiated than the last-erupted portion of either member. Such member-specific variations probably are best understood as being related to the origin of each member in a separate high-level portion of the large magma chamber of the third volcanic cycle. The last-erupted, least differentiated part of member B may represent a volume of magma from beneath those high-level culminations, and the composition of the Upper Basin Member may approach the composition of the much larger unerupted body of rhyolitic magma that lay yet deeper.

Not only are the trace elements zoned in the ash-flow sheets, but important isotopic ratios are zoned as well. The data of Doe and others (1982) and Hildreth and others (1991) show that initial $^{87}Sr/^{86}Sr$ is zoned in the Huckleberry Ridge (from 0.7098 to 0.7268) and Lava Creek Tuffs (from 0.7091 to 0.7109). The value of ε_{Nd} ranges in the Huckleberry Ridge from -8.6 to -18.2 and in the Lava Creek from -7.8 to -9.4. $^{206}Pb/^{204}Pb$ ranges considerably within the Huckleberry Ridge (from 16.669 to 18.051). Hildreth and others (1984; 1991) showed that the Nd, Sr, Pb, and O isotopes vary greatly throughout the entire rhyolitic suite of the Yellowstone Plateau volcanic field, indicating assimilation of crustal roof rocks and perhaps hydrothermal brines by the Yellowstone magmas at high crustal levels as well as isotopic recovery toward normal magmatic values by continued inputs of less contaminated magmas. Pre-eruptive zonation of all these isotopic systems in the compositionally zoned magma chambers is strongly indicated.

The chemical compositions of most of the rhyolitic lava flows, both phenocryst-rich and phenocryst-poor, are quite similar to one another and to the ash-flow tuffs, most of them ranging from about 75 to nearly 78 percent SiO_2 (excluding the Lewis Canyon Rhyolite and Upper Basin Member flows). It thus seems that the most silicic compositions of rhyolites in the volcanic field, both ash flows and lavas, represent the products of differentiation by a combination of processes, perhaps including liquid-state differentiation, in large magma chambers at high crustal levels.

COMPARISON TO OTHER SILICIC IGNEOUS ROCKS

The rhyolites of the Yellowstone Plateau region differ in chemical composition, as shown by the data in tables 4-14, from many of the common rhyolites and granites that typically are associated with andesites and dacites (or granodiorites) in many calc-alkalic igneous provinces. Some examples of these differences are illustrated in table 15. The high SiO_2 contents characteristic of Yellowstone Plateau rhyolites, especially those greater than 76 percent, would themselves be unusual in most calc-alkalic suites. Rhyolites and rhyodacites common in predominantly andesitic suites typically range from about 68 to 74 percent SiO_2 although high-silica rhyolites do occur in some such suites. Compared to most rhyolites of andesite-related suites, not only is SiO_2 high in the Yellowstone rhyolites, but Al_2O_3 tends to be low, CaO and MgO are quite low, and the ratios of ΣFe/Mg and of Ca/Mg are typically high. Similarly, the high potassium, high fluorine, and high concentrations of many incompatible trace elements would be uncommon in the rhyolites of andesite-dacite-rhyolite suites.

The distinctive major-element characteristics of the Yellowstone Plateau rhyolites were noted by Hamilton (1960a), who pointed out that they thus differ from the typical rhyolites of calc-alkalic, predominantly intermediate-composition suites but are similar to the granitic components of many differentiated or bimodal gabbroic complexes such as the Bushveldt and Skaergaard (table 15). This observation of Hamilton's appears to be fundamentally correct, and it is further notable that rhyolites generally low in MgO and CaO but relatively high in SiO_2 and ΣFeO and most incompatible elements are common associates of rhyolite-basalt associations, extrusive as well as intrusive. Examples (some given in table 15) include Icelandic rhyolites (Carmichael, 1964), the Nuanetsi rhyolites (Cox and others, 1965; Stillman, 1970), and rhyolites of the British Tertiary province (Bailey and others, 1924).

THE BASALTS

Although basalts of the Yellowstone Plateau volcanic field vary moderately in chemistry, their mineralogy varies little. Amounts of the constituent minerals, of course, vary from flow to flow and slightly within flows, but their types remain practically constant. Plagioclase phenocrysts, the most common phenocryst type, are present in more than half of the basalts observed. These phenocrysts, commonly a few millimeters to a centimeter across, are slightly rounded and embayed. In a few basalts, however, very large plagioclase phenocrysts—as much as several centimeters across—are deeply sieved by

Table 15.—*Chemical analyses of rhyolites and granites from selected igneous provinces. Recalculated to 100 percent, less volatiles, all Fe as FeO*

	1	2	3	4	5	6	7	8	9
	Yellowstone Plateau	Thingmuli, Iceland	San Juan Mountains, Colo.	Nuanetsi, Africa	Skaergaard, Greenland	San Juan Mountains, Colo.	Talasea, New Britain	Taupo Zone, New Zealand	Sierra Nevada, Calif.
Major elements, recalculated to 100 percent, less volatiles, all Fe as FeO									
SiO_2	77.10	75.67	76.51	73.16	73.26	71.73	75.40	74.12	74.49
Al_2O_3	12.35	12.67	12.63	11.35	13.25	15.22	12.59	13.74	13.77
ΣFeO	1.18	2.28	.98	4.87	3.66	2.19	2.30	2.11	1.43
MgO	.01	.02	.09	.19	.16	.44	.24	.31	.32
CaO	.42	.91	.74	1.70	1.81	1.94	1.25	1.41	1.30
Na_2O	3.89	4.45	3.71	3.09	4.05	3.17	4.02	4.94	3.46
K_2O	4.93	3.69	5.11	5.04	3.28	4.90	3.84	2.97	4.94
TiO_2	.08	.23	.17	.48	.46	.30	.27	.26	.18
P_2O_5	.03	.04	.00	.05	.04	.06	.02	.04	.06
MnO	.03	.04	.00	.05	.04	.05	.07	.10	.05
ΣFeO/MgO	118	114	11	26	23	5.0	9.6	6.8	4.5
CaO/MgO	42	46	8	9	11	4.4	5.2	4.5	4.1
Trace elements in parts per million									
Zr	220	385	190	900	1,000	330	150	195	—
Nb	46	25	80	120	—	—	—	6	10
Mo	4	4	5	—	—	—	—	—	—
V	—	—	10	11	20	—	35	2	10
Ni	—	1	—	—	5	—	40	—	1
Zn	—	95	—	—	—	—	40	—	0
Cu	2	9	3	—	20	8	20	34	2
Li	—	—	—	—	5	—	—	36	—
Sc	—	—	—	—	—	6	—	10	7
Ce	—	—	—	—	—	—	50	45	—
La	81	—	90	—	200	—	20	33	50
Nd	—	—	—	—	—	—	20	19	—
Y	82	70	20	110	200	30	20	33	20
Yb	8	—	—	—	—	—	—	3	1
U	7	—	—	—	—	—	—	2	—
Th	29	—	—	—	—	—	10	10	—
Sr	3	90	32	80	150	140	200	158	400
Ba	48	1,000	960	2,000	1,500	240	645	630	1,000
Rb	239	130	—	—	300	—	55	99	

Sources of Data:

Rhyolites and granophyres associated principally with basalts or gabbros.
1. Obsidian Cliff flow, rhyolitic obsidian (tables 13 and 14, no. 6YC-64).
2. Rhyolitic intrusion, and rhyolitic lavas (Carmichael, 1964, Table 9, no. 21; Carmichael and others, 1974, table 3-1).
3. Rhyolitic ash-flow sheet of younger, rhyolite-basalt association (P. W. Lipman, written commun., 1974).
4. Rhyolite ignimbrite, Karoo igneous province (Cox and others, 1965, Tables 13 and 14, no. LM389).
5. Acid granophyre, Tinden sill (Wager and Brown, 1967, Table 9, no. 5259; Wager and Mitchell, 1951, table D, no. 3058).

Rhyolites associated principally with andesites, and granite of calc-alkalic association
6. Rhyolitic ash-flow sheet of older, andesite-rhyodacite-rhyolite association (Lipman, 1975, Table 6, no. 17).
7. Rhyolite obsidian (Lowder and Carmichael, 1970, Tables 2 and 3, no. 343).
8. Taupo pumice, Taupo lapilli layer (Ewart, 1963, Table 6, no. 3; Ewart and others, 1968, tables 3 and 4, no. 1).
9. Quartz monzonite of Cathedral Peak type (Bateman and others, 1963, Table 3, no. 25).

basaltic glass. Plagioclase compositions commonly range from about An_{65} to An_{70}. Olivine phenocrysts, though somewhat less abundant, occur in many flows, commonly in cumulophyric clusters with plagioclase. The phenocrysts, like the groundmass olivines, generally are near Fo_{80} in composition. Olivine phenocrysts generally are at least slightly rounded or embayed. Plagioclase and olivine are the only minerals noted as phenocrysts in these rocks except for minor spinel included in the olivines and a few crystals of magnetite in cumulophyric clusters with olivine or olivine and plagioclase.

Typically, the basalts appear to be relatively free of xenolithic or xenocrystic inclusions. In every unit examined closely, however, at least a few samples do contain minor xenocrystic quartz or plagioclase of more sodic composition than the typical phenocrysts or groundmass plagioclase. Rare xenoliths of Precambrian granitic gneisses occur in flows of various ages in northern Yellowstone National Park.

Chemically the basalts are somewhat more varied (tables 4, 16) than their gross mineralogical similarities would seem to suggest. All have rather low potassium contents compared

Table 16.—*Chemical analyses of basalts and related rocks of the Yellowstone Plateau volcanic field and adjacent parts of the Snake River Plain. Analyses whose source is indicated as "new" are previously unpublished.*

Age or group	Pre-Huckleberry Ridge Tuff					Post-Huckleberry Ridge Tuff, pre-Mesa Falls Tuff		
Formation or unit	Junction Butte Basalt				Basalt of Soap Creek	Basalt of The Narrows, flow 1		Bas. of The Narrows, flow 2
Locality	Mt. Everts	Blacktail Pond	Overhanging Cliff	Junction Butte	Soap Creek	The Narrows		The Narrows
Source	New				New	New		New
Sample No.	0YC-639	74Y-153	0YC-635	8YC-466	81YH-97	0YC-632A	8YC-475	0YC-632B
Analyses recalculated to 100 percent, less volatiles, Fe_2O_3/FeO calculated as 0.15								
SiO_2	49.51	49.99	51.55	56.79	50.89	51.10	51.51	50.88
Al_2O_3	15.19	15.96	15.33	15.19	15.54	16.29	15.76	16.31
Fe_2O_3	1.70	1.57	1.57	1.26	1.49	1.38	1.38	1.33
FeO	11.33	10.49	10.48	8.41	9.90	9.20	9.23	8.84
MgO	6.68	6.27	5.13	3.94	6.91	6.18	6.29	6.08
CaO	9.53	9.79	9.19	7.17	9.88	10.38	9.95	11.35
Na_2O	2.71	2.69	2.79	3.09	2.76	2.62	2.77	2.54
K_2O	0.59	0.57	1.14	1.91	0.61	0.80	1.08	0.69
TiO_2	2.28	2.23	2.27	1.68	1.61	1.65	1.64	1.58
P_2O_5	0.28	0.27	0.35	0.41	0.23	0.22	0.22	0.22
MnO	0.19	0.18	0.19	0.16	0.17	0.17	0.15	0.19
CIPW norms from recalculated analyses								
q	——	——	1.73	8.36	——	.015	——	——
c	——	——	——	——	——	——	——	——
or	3.49	3.37	6.74	11.29	3.60	4.73	6.38	4.08
ab	22.93	22.76	23.61	26.15	23.35	22.17	23.44	21.49
an	27.54	29.79	25.94	21.94	28.21	30.33	27.38	31.06
di	14.80	14.05	14.43	9.12	15.88	16.25	16.91	19.60
wo	7.48	7.10	7.25	4.57	8.06	8.24	8.58	9.94
en	3.73	3.59	3.36	2.05	4.23	4.29	4.49	5.22
fs	3.59	3.36	3.82	2.50	3.59	3.72	3.84	4.44
hy	19.57	21.84	20.15	17.19	20.97	20.73	18.95	18.28
en	9.98	11.28	9.42	7.76	11.34	11.10	10.22	9.88
fs	9.59	10.56	10.73	9.43	9.63	9.63	8.73	8.40
ol	4.22	1.07	——	——	2.22	——	1.29	0.06
fo	2.05	0.53	——	——	1.15	——	0.66	0.03
fa	2.17	0.54	——	——	1.07	——	0.63	0.03
mt	2.46	2.28	2.28	1.83	2.16	2.00	2.00	1.93
il	4.33	4.24	4.31	3.19	3.06	3.13	3.12	3.00
ap	0.66	0.64	0.83	0.97	0.54	0.52	0.52	0.52
Salic	53.96	55.92	58.02	67.74	55.16	57.38	57.20	56.63
Femic	46.04	44.12	42.00	32.30	44.83	42.63	42.79	43.39

Table 16.—*Chemical analyses of basalts and related rocks of the Yellowstone Plateau volcanic field and adjacent parts of the Snake River Plain. Analyses whose source is indicated as "new" are previously unpublished.*—Continued

Age or group	Post-Mesa Falls Tuff, pre-Lava Creek Tuff								
Formation or unit	Basalt of The Narrows, flow 4		Basalt of Warm River			Basalt of Coffeepot Spring	Basalt of Shotgun Valley		Basalt of Teton River
Locality	The Narrows		Sheep Falls	Fish Creek Road	Warm River	Coffeepot Campgrnd.	Shotgun Valley		Teton River
Source	New		New	Table 4	New	New	New		New
Sample No.	0YC-636	0YC-638	81YH-84	6YC-144	3YC-11	74IP-64	74IP-66	74IP-67	74IP-77
Analyses recalculated to 100 percent, less volatiles, Fe_2O_3/FeO calculated as 0.15									
SiO_2	49.84	49.97	46.12	47.40	47.52	48.67	46.35	47.85	48.89
Al_2O_3	17.32	17.32	15.20	16.97	16.18	16.02	15.45	16.12	14.85
Fe_2O_3	1.41	1.48	1.94	1.60	1.60	1.56	1.89	1.58	1.51
FeO	9.38	9.84	12.93	10.64	10.66	10.39	12.57	10.54	10.09
MgO	5.20	4.07	6.05	7.97	8.36	7.50	6.14	8.22	9.69
CaO	9.35	9.60	9.44	10.17	10.06	10.08	9.76	10.58	10.20
Na_2O	3.50	3.59	2.83	2.67	2.82	2.62	3.13	2.40	2.11
K_2O	1.28	1.32	0.88	0.27	0.43	0.45	0.74	0.33	0.48
TiO_2	2.10	2.17	3.30	1.82	1.84	2.11	2.59	1.88	1.68
P_2O_5	0.41	0.43	1.07	0.30	0.34	0.43	1.14	0.32	0.30
MnO	0.23	0.21	0.24	0.19	0.18	0.18	0.23	0.18	0.18
CIPW norms from recalculated analyses									
q	——	——	——	——	——	——	——	——	——
c	——	——	——	——	——	——	——	——	——
or	7.56	7.80	5.20	1.60	2.54	2.66	4.37	1.95	2.84
ab	29.62	30.38	23.95	22.59	23.86	22.17	26.49	20.31	17.85
an	27.77	27.25	26.17	33.52	30.22	30.62	25.92	32.24	29.63
di	13.20	14.69	11.36	12.29	14.31	13.60	12.51	14.88	15.48
wo	6.66	7.34	5.71	6.25	7.29	6.92	6.28	7.58	7.94
en	3.26	3.13	2.64	3.38	4.02	3.75	2.89	4.18	4.67
fs	3.28	4.22	3.01	2.65	3.00	2.94	3.34	3.12	2.87
hy	2.22	1.68	8.22	7.55	3.68	15.64	2.45	11.96	19.19
en	1.11	0.72	3.85	4.23	2.11	8.77	1.14	6.85	11.88
fs	1.11	0.96	4.37	3.32	1.57	6.87	1.31	5.11	7.31
ol	12.68	10.95	13.54	15.98	18.78	8.05	17.95	12.06	8.91
fo	6.02	4.41	6.01	8.57	10.30	4.32	7.89	6.62	5.31
fa	6.66	6.54	7.53	7.41	8.48	3.73	10.05	5.44	3.60
mt	2.04	2.15	2.81	2.32	2.32	2.26	2.74	2.29	2.19
il	3.99	4.12	6.27	3.46	3.49	4.01	4.92	3.57	3.19
ap	0.97	1.02	2.53	0.71	0.81	1.02	2.70	0.76	0.71
Salic	64.95	65.43	55.32	57.71	56.62	55.45	56.78	54.50	50.32
Femic	35.10	34.61	44.73	42.31	43.39	44.58	43.27	45.52	49.67

Table 16.—*Chemical analyses of basalts and related rocks of the Yellowstone Plateau volcanic field and adjacent parts of the Snake River Plain. Analyses whose source is indicated as "new" are previously unpublished.*—Continued

Age or group	Post-Mesa Falls Tuff, pre-Lava Creek Tuff							
Formation or unit	Undine Falls Basalt							
Locality	Deckard Flats	Blacktail Deer Creek	Undine Falls	Blacktail Deer Plateau	Lava Creek	Deckard Flats	Oxbow Creek	Rescue Creek
Source	New			Table 4	New		Table 4	
Sample No.	0YC-597B	8YC-485	0YC-640A	7YC-297	6YC-162B	0YC-597C	8YC-482	7YC-295
Analyses recalculated to 100 percent, less volatiles, Fe_2O_3/FeO calculated as 0.15								
SiO_2	48.49	48.53	48.80	48.82	48.99	49.08	49.22	50.34
Al_2O_3	16.36	15.94	16.17	16.54	16.40	16.16	16.71	18.77
Fe_2O_3	1.57	1.60	1.59	1.56	1.52	1.54	1.52	1.23
FeO	10.49	10.69	10.59	10.38	10.14	10.26	10.16	8.22
MgO	7.34	6.23	6.93	6.90	7.53	7.40	7.00	5.29
CaO	10.20	11.03	10.03	9.80	9.87	10.14	9.70	10.54
Na_2O	2.74	2.92	2.94	3.00	2.82	2.69	2.96	3.15
K_2O	0.51	0.61	0.55	0.55	0.51	0.51	0.53	0.61
TiO_2	1.88	1.97	1.96	1.89	1.83	1.83	1.79	1.50
P_2O_5	0.23	0.27	0.27	0.38	0.24	0.23	0.22	0.20
MnO	0.18	0.19	0.18	0.18	0.17	0.17	0.17	0.15
CIPW norms from recalculated analyses								
q	—	—	—	—	—	—	—	—
c	—	—	—	—	—	—	—	—
or	3.01	3.60	3.25	3.25	3.01	3.01	3.13	3.60
ab	23.18	24.71	24.88	25.39	23.86	22.76	25.05	26.65
an	30.83	28.58	29.30	30.04	30.58	30.51	30.74	35.27
di	15.03	20.16	15.41	13.25	13.79	15.03	13.12	12.95
wo	7.63	10.18	7.81	6.72	7.02	7.64	6.66	6.56
en	4.04	5.01	4.04	3.49	3.80	4.09	3.50	3.36
fs	3.36	4.97	3.56	3.04	2.97	3.30	2.96	3.03
hy	9.12	3.10	8.47	9.47	11.30	13.01	11.08	10.21
en	4.98	1.56	4.50	5.06	6.34	7.20	6.00	5.37
fs	4.14	1.54	3.97	4.41	4.96	5.81	5.08	4.84
ol	12.44	13.14	12.07	11.86	11.22	9.45	10.75	6.20
fo	6.49	6.27	6.12	6.05	6.03	5.00	5.56	3.11
fa	5.95	6.87	5.95	5.81	5.19	4.45	5.19	3.09
mt	2.28	2.32	2.31	2.26	2.20	2.23	2.20	1.78
il	3.57	3.74	3.72	3.59	3.48	3.48	3.40	2.85
ap	0.54	0.64	0.64	0.90	0.57	0.54	0.52	0.47
Salic	57.02	56.89	57.43	58.68	57.45	56.28	58.92	65.52
Femic	42.98	43.10	42.62	41.33	42.56	43.74	41.07	34.46

Table 16.—*Chemical analyses of basalts and related rocks of the Yellowstone Plateau volcanic field and adjacent parts of the Snake River Plain. Analyses whose source is indicated as "new" are previously unpublished.*—Continued

Age or group	Post-Lava Creek Tuff, higher-K suite						
Formation or unit	Falls River Basalt						Basalt of Grizzly Lake
Locality	Gibson Meadow			Marysville, Ida.	Bitch Creek	Falls River	Grizzly Lake
Source	New						New
Sample No.	81YH-89	6YC-145	6YC-14	78YH-31	3YC-5A	74IP-72	81Y-227C
	Analyses recalculated to 100 percent, less volatiles, Fe_2O_3/FeO calculated as 0.15						
SiO_2	46.54	46.83	46.84	47.98	48.14	49.61	48.93
Al_2O_3	15.05	14.74	14.88	15.02	16.35	15.22	15.57
Fe_2O_3	1.87	1.86	1.84	1.65	1.61	1.56	1.67
FeO	12.49	12.40	12.27	11.03	10.73	10.39	11.15
MgO	7.40	7.39	7.39	8.07	7.70	8.25	6.36
CaO	10.33	10.33	10.39	9.91	9.97	9.64	9.35
Na_2O	2.48	2.58	2.59	2.57	2.67	2.31	3.08
K_2O	0.44	0.50	0.48	0.67	0.48	0.55	0.75
TiO_2	2.64	2.61	2.57	2.36	1.86	1.95	2.54
P_2O_5	0.54	0.55	0.54	0.55	0.28	0.34	0.40
MnO	0.22	0.21	0.21	0.19	0.22	0.18	0.18
	CIPW norms from recalculated analyses						
q	——	——	——	——	——	——	——
c	——	——	——	——	——	——	——
or	2.60	2.95	2.84	3.96	2.84	3.25	4.43
ab	20.98	21.83	21.92	21.75	22.59	19.55	26.06
an	28.63	27.13	27.56	27.47	31.21	29.54	26.44
di	15.76	16.86	16.89	14.84	13.50	13.16	14.32
wo	7.97	8.55	8.54	7.56	6.86	6.71	7.24
en	4.00	4.43	4.32	4.13	3.65	3.74	3.61
fs	3.79	3.88	4.03	3.15	2.99	2.71	3.47
hy	9.05	8.89	7.52	12.03	9.95	24.38	10.97
en	4.65	4.74	3.89	6.82	5.47	14.13	5.59
fs	4.40	4.15	3.63	5.21	4.48	10.25	5.38
ol	13.99	12.71	14.48	11.80	13.42	3.38	9.58
fo	6.85	6.47	7.14	6.41	7.05	1.88	4.65
fa	7.14	6.24	7.34	5.39	6.37	1.50	4.93
mt	2.71	3.41	2.67	2.39	2.33	2.26	2.42
il	5.01	4.96	4.88	4.48	3.53	3.70	4.82
ap	1.28	1.30	1.28	1.30	0.66	0.81	0.95
Salic	52.21	51.91	52.32	53.18	56.64	52.34	56.93
Femic	47.80	48.13	47.72	46.84	43.39	47.69	43.06

Table 16.—*Chemical analyses of basalts and related rocks of the Yellowstone Plateau volcanic field and adjacent parts of the Snake River Plain. Analyses whose source is indicated as "new" are previously unpublished.*—Continued

Age or group	Post-Lava Creek Tuff, higher-K suite							
Formation or unit	Madison River Basalt							
Locality	West Entrance Road	Cougar Creek	Madison River		Cougar Creek			Madison River
Source	New	Hamilton (1963)	New	Hamilton (1963)	Table 4	New	Table 4	Hamilton (1963)
Sample No.	74Y-137	YS6	74Y-178	YS20C	6YC-44	0YC-643	6YC-139	YS5
Analyses recalculated to 100 percent, less volatiles, Fe_2O_3/FeO calculated as 0.15								
SiO_2	46.75	46.97	48.04	48.22	48.77	50.77	51.73	52.13
Al_2O_3	15.22	15.79	15.34	15.74	16.45	14.75	15.66	15.69
Fe_2O_3	1.95	1.87	1.81	1.75	1.66	1.77	1.49	1.45
FeO	13.02	12.48	12.09	11.65	11.05	11.77	9.93	9.63
MgO	6.74	6.84	6.75	6.82	6.20	5.31	6.09	6.07
CaO	9.78	9.35	9.71	9.22	9.42	8.67	8.76	8.61
Na_2O	2.96	3.21	2.85	3.30	2.98	3.03	2.95	3.04
K_2O	0.45	0.46	0.58	0.56	0.57	0.89	1.12	1.12
TiO_2	2.57	2.42	2.34	2.21	2.40	2.47	1.87	1.82
P_2O_5	0.33	0.39	0.28	0.33	0.31	0.37	0.23	0.28
MnO	0.22	0.22	0.21	0.21	0.20	0.20	0.17	0.17
CIPW norms from recalculated analyses								
q	——	——	——	——	——	0.38	0.01	0.42
c	——	——	——	——	——	——	——	——
or	2.66	2.72	3.43	3.31	3.37	5.26	6.62	6.62
ab	25.05	27.16	24.12	27.92	25.22	25.64	24.96	25.72
an	26.91	27.32	27.35	26.48	29.83	24.02	26.18	25.86
di	16.14	13.69	15.65	14.14	12.31	13.81	13.03	12.39
wo	8.12	6.90	7.91	7.14	6.21	6.92	6.59	6.27
en	3.82	3.32	3.96	3.53	3.06	3.07	3.32	3.20
fs	4.20	3.47	3.78	3.47	3.04	3.82	3.12	2.92
hy	2.52	0.92	9.81	3.87	13.05	22.77	22.96	22.80
en	1.20	0.45	5.02	1.95	6.55	10.15	11.85	11.92
fs	1.32	0.47	4.79	1.92	6.50	12.62	11.11	10.88
ol	18.25	20.00	11.25	16.79	8.56	——	——	——
fo	8.25	9.30	5.49	8.06	4.09	——	——	——
fa	10.00	10.70	5.76	8.73	4.47	——	——	——
mt	2.83	2.71	3.32	2.54	2.41	2.57	2.16	2.10
il	4.88	4.60	4.44	4.20	4.56	4.69	3.55	3.46
ap	0.78	0.92	0.66	0.78	0.73	0.88	0.54	0.66
Salic	54.62	57.20	54.90	57.71	58.42	55.30	57.77	58.62
Femic	45.40	42.84	45.13	42.32	41.62	44.72	42.24	41.41

Table 16.—*Chemical analyses of basalts and related rocks of the Yellowstone Plateau volcanic field and adjacent parts of the Snake River Plain. Analyses whose source is indicated as "new" are previously unpublished.*—Continued

Age or group	Post-Lava Creek Tuff												
Formation or unit	Gerrit Basalt												
Locality	Box Canyon	Eccles Butte	Pole Bridge Campgrnd.	Buffalo River	Upper Mesa Falls	South of Eccles Butte	Box Canyon	Upper Mesa Falls	West of Osborne Bridge	Silver Lake	Upper Mesa Falls		
Source	New										Hamilton (1965)	New	
Sample No.	74IP-57	74IP-42	74IP-88	74IP-147	74IP-61E	74IP-85	78YH-38	74IP-61C	74IP-100	74IP-116	YS57G	74IP-61G	74IP-61D
	Analyses recalculated to 100 percent, less volatiles, Fe_2O_3/FeO calculated as 0.15												
SiO_2	46.79	46.84	46.87	46.92	47.01	47.05	47.06	47.16	47.24	47.35	47.43	47.45	47.48
Al_2O_3	16.26	16.73	16.10	15.74	14.37	16.18	15.65	14.92	16.28	15.92	16.23	15.99	15.79
Fe_2O_3	1.76	1.57	1.46	1.81	1.48	1.50	1.80	1.44	1.58	1.57	1.57	1.46	1.40
FeO	11.70	10.45	9.71	12.08	9.88	9.98	12.01	9.61	10.52	10.48	10.48	9.72	9.33
MgO	7.99	8.97	9.21	7.41	12.66	9.62	7.43	11.82	8.62	8.44	8.50	9.20	10.26
CaO	10.04	10.95	12.45	10.13	10.35	11.29	9.83	10.81	11.36	10.71	11.07	11.36	11.37
Na_2O	2.61	2.19	2.19	2.68	2.01	2.24	2.79	2.02	2.30	2.44	2.46	2.26	2.14
K_2O	0.30	0.13	0.18	0.40	0.40	0.24	0.60	0.33	0.17	0.50	0.17	0.39	0.37
TiO_2	2.03	1.67	1.42	2.25	1.41	1.45	2.26	1.39	1.54	1.97	1.69	1.67	1.43
P_2O_5	0.32	0.31	0.23	0.38	0.26	0.26	0.37	0.31	0.18	0.42	0.20	0.31	0.25
MnO	0.20	0.19	0.18	0.21	0.18	0.19	0.20	0.18	0.20	0.19	0.20	0.18	0.17
	CIPW norms from recalculated analyses												
q	—	—	—	—	—	—	—	—	—	—	—	—	—
c	—	—	—	—	—	—	—	—	—	—	—	—	—
or	1.77	0.77	1.06	2.36	2.36	1.42	3.55	1.95	1.00	2.95	1.00	2.30	2.19
ab	22.08	18.53	18.53	22.68	17.01	18.95	23.61	17.09	19.46	20.65	20.82	19.12	18.11
an	31.76	35.43	33.57	29.74	29.01	33.38	28.41	30.67	33.60	31.01	32.74	32.33	32.39
di	13.13	13.78	21.77	14.89	16.69	17.06	14.81	16.95	17.68	15.86	17.09	17.93	18.17
wo	6.66	7.04	11.15	7.53	8.62	8.74	7.49	8.74	9.01	8.09	8.71	9.19	9.35
en	3.48	3.98	6.49	3.79	5.48	5.11	3.79	5.50	4.99	4.53	4.83	5.39	5.71
fs	2.99	2.76	4.13	3.57	2.59	3.21	3.53	2.71	3.68	3.24	3.55	3.35	3.11
hy	6.52	9.72	0.38	6.19	6.29	4.81	3.76	7.28	6.24	6.88	6.31	5.98	5.67
en	3.51	5.74	0.23	3.19	4.27	2.95	1.95	4.88	3.59	4.01	3.64	3.69	3.67
fs	3.01	3.98	0.15	3.00	2.02	1.86	1.81	2.40	2.65	2.87	2.67	2.29	2.00
ol	17.59	15.60	19.35	16.38	23.24	18.85	18.10	20.61	16.39	15.65	16.09	16.33	18.14
fo	9.05	8.84	11.37	8.04	15.27	11.14	8.94	13.36	9.04	8.75	8.90	9.70	11.33
fa	8.54	6.76	7.98	8.34	7.97	7.71	9.16	7.25	7.35	6.90	7.19	6.63	6.81
mt	2.55	2.28	2.12	2.62	2.15	2.17	2.61	2.09	2.29	2.28	2.28	2.12	2.03
il	3.86	3.17	2.70	4.27	2.68	2.75	4.29	2.64	2.93	3.74	3.21	3.17	2.72
ap	0.76	0.73	0.54	0.90	0.62	0.62	0.88	0.73	0.43	0.99	0.47	0.73	0.59
Salic	55.61	54.73	53.16	54.78	48.38	53.75	55.57	49.71	54.06	54.61	54.56	53.75	52.69
Femic	44.41	45.28	46.86	45.25	51.67	46.26	44.45	50.30	45.96	45.40	45.45	46.26	47.32

Table 16.—Chemical analyses of basalts and related rocks of the Yellowstone Plateau volcanic field and adjacent parts of the Snake River Plain. Analyses whose source is indicated as "new" are previously unpublished.—Continued

Age or group	Post-Lava Creek Tuff											
Formation or unit	Gerrit Basalt											
Locality	Antelope Flat	Upper Mesa Falls				Hill 6427	Upper Mesa Falls	Ripley Butte	Upper Mesa Falls	Eccles Butte	Pineview	South of Osborne Butte
Source	New		Hamilton (1965)	New								
Sample No.	74IP-107	6YC-140B	YS57B	74IP-61A	74IP-61B	6YC-142	74IP-39	74IP-45	74IP-61F	74IP-44	74IP-86	74IP-81
Analyses recalculated to 100 percent, less volatiles, Fe_2O_3/FeO calculated as 0.15												
SiO_2	47.64	47.65	47.79	47.86	47.88	47.94	47.95	47.97	47.97	48.15	48.64	48.87
Al_2O_3	15.88	15.58	16.13	15.85	15.56	16.65	16.02	16.09	15.46	16.79	14.97	15.04
Fe_2O_3	1.58	1.46	1.45	1.42	1.44	1.40	1.52	1.48	1.52	1.43	1.66	1.68
FeO	10.52	9.72	9.68	9.45	9.63	9.30	10.16	9.89	10.11	9.56	11.06	11.20
MgO	7.77	9.56	8.71	9.04	9.32	8.93	8.45	9.11	8.34	8.59	7.08	7.58
CaO	10.75	11.26	11.32	11.81	11.64	11.84	11.51	11.06	11.34	11.66	9.92	10.13
Na_2O	2.52	2.21	2.27	2.20	2.14	2.17	2.27	2.29	2.32	2.28	2.51	2.47
K_2O	0.60	0.43	0.37	0.40	0.41	0.23	0.26	0.24	0.51	0.17	0.73	0.66
TiO_2	2.12	1.64	1.78	1.52	1.53	1.20	1.45	1.45	1.90	1.06	2.56	1.56
P_2O_5	0.43	0.31	0.30	0.26	0.27	0.16	0.21	0.22	0.35	0.11	0.69	0.61
MnO	0.19	0.18	0.20	0.18	0.18	0.18	0.20	0.19	0.19	0.19	0.19	0.20
CIPW norms from recalculated analyses												
q	—	—	—	—	—	—	—	—	—	—	—	—
c	—	—	—	—	—	—	—	—	—	—	—	—
or	3.55	2.54	2.19	2.36	2.42	1.36	1.54	1.42	3.01	1.00	4.31	3.90
ab	21.32	18.70	19.21	18.62	18.11	18.36	19.21	19.38	19.63	19.29	21.24	20.90
an	30.25	31.32	32.73	32.19	31.64	35.01	32.75	32.91	30.26	35.08	27.42	28.00
di	16.64	18.32	17.50	20.12	19.82	18.50	18.81	16.75	19.38	18.03	14.23	15.06
wo	8.47	9.40	8.96	10.31	10.16	9.47	9.59	8.57	9.90	9.21	7.22	7.63
en	4.62	5.58	5.19	6.05	5.98	5.50	5.33	4.94	5.58	5.19	3.78	3.91
fs	3.55	3.34	3.35	3.76	3.68	3.53	3.89	3.24	3.90	3.63	3.23	3.52
hy	7.00	7.17	8.84	6.25	7.75	7.60	8.77	9.91	8.33	7.90	19.13	16.77
en	3.96	4.48	5.37	3.85	4.80	4.63	5.07	5.98	4.90	4.65	10.32	8.83
fs	3.04	2.69	3.47	2.40	2.95	2.97	3.70	3.93	3.43	3.25	8.81	7.94
ol	13.94	16.00	13.35	14.90	14.64	14.49	13.47	14.22	12.76	14.35	4.81	8.56
fo	7.55	9.64	7.80	8.84	8.72	8.49	7.46	8.25	7.21	8.10	2.48	4.30
fa	6.39	6.36	5.55	6.06	5.92	6.00	6.01	5.97	5.55	6.25	2.33	4.26
mt	2.29	2.12	2.10	2.06	2.09	2.03	2.20	2.15	2.20	2.07	2.41	2.44
il	4.03	3.12	3.38	2.89	2.91	2.28	2.75	2.75	3.61	2.01	4.86	2.96
ap	1.02	0.73	0.71	0.62	0.64	0.38	0.50	0.52	0.83	0.26	1.63	1.44
Salic	55.12	52.56	54.13	53.17	52.17	54.73	53.50	53.71	52.90	55.37	52.97	52.80
Femic	44.92	47.46	45.88	46.84	47.85	45.28	46.50	46.30	47.11	44.62	47.07	47.23

Table 16.—*Chemical analyses of basalts and related rocks of the Yellowstone Plateau volcanic field and adjacent parts of the Snake River Plain. Analyses whose source is indicated as "new" are previously unpublished.*—Continued

Age or group	Post-Lava Creek Tuff, lower-K suite					
Formation or unit	Swan Lake Flat Basalt					
Locality	Horseshoe Hill	West of Bunsen Peak	Near hill 7717	Gardner River		Lava Creek
Source	New					
Sample No.	74Y-163	74Y-158	74Y-156B	74Y-168	74Y-160	8YC-470
Analyses recalculated to 100 percent, less volatiles, Fe_2O_3/FeO calculated as 0.15						
SiO_2	49.35	49.60	49.63	49.71	49.72	49.76
Al_2O_3	15.28	15.46	16.31	15.20	15.00	16.18
Fe_2O_3	1.47	1.43	1.40	1.45	1.48	1.43
FeO	9.80	9.55	9.34	9.65	9.89	9.56
MgO	8.04	8.37	7.53	7.84	7.73	7.19
CaO	10.96	10.74	10.27	11.07	10.87	10.25
Na_2O	2.42	2.47	2.93	2.51	2.52	2.88
K_2O	0.34	0.35	0.58	0.39	0.41	0.63
TiO_2	1.92	1.65	1.62	1.78	1.95	1.70
P_2O_5	0.24	0.20	0.22	0.23	0.24	0.23
MnO	0.17	0.17	0.17	0.17	0.18	0.17
CIPW norms from recalculated analyses						
q	——	——	——	——	——	——
c	——	——	——	——	——	——
or	2.01	2.07	3.43	2.30	2.42	3.72
ab	20.48	20.90	24.79	21.24	21.32	24.37
an	29.83	30.06	29.64	29.06	28.41	29.36
di	18.78	17.89	16.28	19.93	19.62	16.41
wo	9.59	9.15	8.30	10.17	10.00	8.35
en	5.42	5.24	4.62	5.70	5.56	4.54
fs	3.77	3.50	3.36	4.06	4.06	3.52
hy	16.60	16.24	8.96	15.46	16.73	10.90
en	9.79	9.73	5.19	9.03	9.67	6.14
fs	6.81	6.51	3.77	6.43	7.06	4.76
ol	5.95	7.16	11.30	5.99	5.09	9.40
fo	3.37	4.12	6.27	3.36	2.82	5.07
fa	2.58	3.04	5.03	2.63	2.27	4.33
mt	2.13	2.07	2.03	2.10	2.15	2.07
il	3.65	3.13	3.08	3.38	3.70	3.23
ap	0.57	0.47	0.52	0.54	0.57	0.54
Salic	52.32	53.03	57.86	52.60	52.15	57.45
Femic	47.68	46.96	42.17	47.40	47.86	42.55

Table 16.—*Chemical analyses of basalts and related rocks of the Yellowstone Plateau volcanic field and adjacent parts of the Snake River Plain. Analyses whose source is indicated as "new" are previously unpublished.*—Continued

Age or group	Post-Lava Creek Tuff, lower-K suite							
Formation or unit	Swan Lake Flat Basalt							
Locality	Gardner River				Willow Park	Gardner River		Willow Park
Source	Fenner (1938)			New	Table 4	New		
Sample No.	YP1025	YP904	YP1024	0YC-607	6YC-51	6YC-133	74Y-166	0YC-613
Analyses recalculated to 100 percent, less volatiles, Fe_2O_3/FeO calculated as 0.15								
SiO_2	49.86	50.06	50.07	50.50	50.55	50.67	50.74	52.53
Al_2O_3	15.35	15.58	15.32	15.53	16.05	15.38	15.93	15.54
Fe_2O_3	1.39	1.39	1.39	1.39	1.35	1.40	1.37	1.29
FeO	9.25	9.29	9.25	9.29	9.01	9.33	9.15	8.58
MgO	7.93	7.55	7.62	7.55	7.50	7.33	6.75	7.08
CaO	11.03	10.92	11.11	10.62	10.36	10.56	10.72	9.54
Na_2O	2.45	2.36	2.50	2.58	2.81	2.73	2.72	2.71
K_2O	0.42	0.47	0.44	0.47	0.45	0.51	0.58	0.90
TiO_2	1.91	1.95	1.81	1.68	1.54	1.71	1.68	1.49
P_2O_5	0.25	0.28	0.37	0.22	0.22	0.21	0.21	0.19
MnO	0.16	0.15	0.13	0.16	0.16	0.16	0.15	0.15
CIPW norms from recalculated analyses								
q	—	—	—	—	—	—	—	1.16
c	—	—	—	—	—	—	—	—
or	2.48	2.78	2.60	2.78	2.66	3.01	3.43	5.32
ab	20.73	19.97	21.15	21.83	23.78	23.10	23.02	22.93
an	29.65	30.53	29.28	29.41	29.85	28.20	29.54	27.58
di	19.14	17.83	19.14	17.88	16.45	18.68	18.27	15.15
wo	9.79	9.11	9.78	9.12	8.40	9.52	9.30	7.73
en	5.64	5.16	5.54	5.11	4.72	5.27	5.03	4.34
fs	3.71	3.56	3.82	3.65	3.33	3.89	3.94	3.08
hy	17.48	21.10	18.33	19.41	16.63	17.63	17.53	22.73
en	10.55	12.49	10.85	11.32	9.75	10.15	9.83	13.30
fs	6.93	8.61	7.48	8.09	6.88	7.48	7.70	9.43
ol	4.31	1.42	3.19	2.97	5.25	3.59	2.55	—
fo	2.50	0.81	1.81	1.66	2.95	1.98	1.37	—
fa	1.81	0.61	1.38	1.31	2.30	1.61	1.18	—
mt	2.02	2.02	2.02	2.02	1.96	2.03	1.99	1.87
il	3.63	3.70	3.44	3.19	2.93	3.25	3.19	2.83
ap	0.59	0.66	0.88	0.52	0.52	0.50	0.50	0.45
Salic	52.86	53.28	53.03	54.02	56.29	54.31	55.99	56.99
Femic	47.17	46.73	47.00	45.99	43.74	45.68	44.03	43.03

Table 16.—*Chemical analyses of basalts and related rocks of the Yellowstone Plateau volcanic field and adjacent parts of the Snake River Plain. Analyses whose source is indicated as "new" are previously unpublished.*—Continued

Age or group	Post-Lava Creek Tuff, lower-K suite								
Formation or unit	Mixed lavas of Gardner River				Basalt of Geode Creek				Basalt of Mariposa Lake
Locality	Gardner River				South of road		North of road	South of road	Mariposa Lake
Source	New	Fenner (1938)	New		New		Table 4	New	New
Sample No.	74Y-167	YP969	6YC-134	6YC-132	74Y-150C	74Y-150A	8YC-480	74Y-150B	P-466
Analyses recalculated to 100 percent, less volatiles, Fe_2O_3/FeO calculated as 0.15									
SiO_2	49.99	50.71	53.43	54.25	53.84	54.22	54.22	56.16	53.90
Al_2O_3	15.89	15.96	15.35	15.17	15.51	15.98	15.85	15.46	16.35
Fe_2O_3	1.39	1.35	1.25	1.23	1.33	1.28	1.30	1.19	1.01
FeO	9.25	9.01	8.33	8.20	8.84	8.50	8.67	7.95	6.75
MgO	7.64	7.31	6.69	6.43	6.08	5.37	5.43	5.43	7.98
CaO	10.76	10.35	9.28	9.12	8.38	8.66	8.23	7.72	9.55
Na_2O	2.64	2.72	2.92	2.76	2.73	2.71	2.85	2.73	2.36
K_2O	0.38	0.51	0.99	1.12	1.34	1.35	1.45	1.61	1.12
TiO_2	1.69	1.69	1.44	1.41	1.57	1.59	1.65	1.38	0.69
P_2O_5	0.20	0.25	0.18	0.17	0.21	0.20	0.20	0.21	0.15
MnO	0.16	0.15	0.14	0.14	0.16	0.14	0.15	0.15	0.13
CIPW norms from recalculated analyses									
q	——	——	1.88	3.79	3.60	4.81	4.16	7.32	2.48
c	——	——	——	——	——	——	——	——	——
or	2.25	3.01	5.85	6.62	7.92	7.98	8.57	9.51	6.62
ab	22.34	23.02	24.71	23.35	23.10	22.93	24.12	23.10	19.97
an	30.38	29.83	25.85	25.70	26.11	27.45	26.17	25.17	30.71
di	17.74	16.27	15.57	15.11	11.59	11.72	11.01	9.67	12.73
wo	9.06	8.30	7.94	7.70	5.88	5.93	5.57	4.91	6.55
en	5.10	4.66	4.41	4.24	3.09	3.03	2.84	2.56	4.02
fs	3.58	3.31	3.22	3.17	2.62	2.76	2.60	2.20	2.16
hy	15.92	19.37	21.18	20.57	22.27	19.78	20.49	20.37	24.37
en	9.36	11.33	12.25	11.77	12.05	10.35	10.69	10.96	15.86
fs	6.56	8.04	8.93	8.80	10.22	9.43	9.80	9.41	8.51
ol	5.67	2.77	——	——	——	——	——	——	——
fo	3.20	1.55	——	——	——	——	——	——	——
fa	2.47	1.22	——	——	——	——	——	——	——
mt	2.02	1.96	1.81	1.78	1.93	1.86	1.88	1.73	1.46
il	3.21	3.21	2.74	2.68	2.98	3.02	3.13	2.62	1.31
ap	0.47	0.59	0.43	0.40	0.50	0.47	0.47	0.50	0.36
Salic	54.97	55.86	58.29	59.46	60.73	63.17	63.02	65.10	59.78
Femic	45.03	44.17	41.73	40.54	39.27	36.85	36.98	34.89	40.23

Table 16.—*Chemical analyses of basalts and related rocks of the Yellowstone Plateau volcanic field and adjacent parts of the Snake River Plain. Analyses whose source is indicated as "new" are previously unpublished.*—Continued

Age or group	Post-Lava Creek Tuff, lower-K suite									
Formation or unit	Osprey Basalt									
Locality	East of Bunsen Peak (type section)							Blackt. Deer Plateau	Undine Falls	Lamar Canyon
Source	New								Table 4	New
Sample No.	74Y-159F	74Y-159C	74Y-159E	74Y-159A	74Y-159B	74Y-159G	74Y-159D	9YC-514	8YC-471C	9YC-511
Analyses recalculated to 100 percent, less volatiles, Fe_2O_3/FeO calculated as 0.15										
SiO_2	48.81	49.03	49.04	49.05	49.09	49.26	50.05	50.35	50.89	52.88
Al_2O_3	16.17	15.87	15.91	16.79	16.33	15.78	15.57	15.34	15.89	14.81
Fe_2O_3	1.42	1.55	1.45	1.48	1.43	1.41	1.53	1.37	1.43	1.41
FeO	9.45	10.34	9.66	9.84	9.51	9.40	10.18	9.12	9.51	9.42
MgO	8.30	7.39	8.05	7.05	8.20	8.19	7.16	8.18	6.86	5.83
CaO	11.42	10.51	11.28	10.09	10.69	11.26	9.83	10.59	9.97	9.69
Na_2O	2.31	2.59	2.32	2.99	2.57	2.45	2.73	2.55	2.80	2.81
K_2O	0.14	0.37	0.20	0.51	0.23	0.27	0.63	0.54	0.50	0.79
TiO_2	1.63	1.91	1.74	1.79	1.60	1.63	1.90	1.59	1.75	1.94
P_2O_5	0.18	0.25	0.19	0.24	0.19	0.18	0.25	0.19	0.24	0.24
MnO	0.17	0.18	0.17	0.16	0.16	0.16	0.17	0.16	0.16	0.17
CIPW norms from recalculated analyses										
q	——	——	——	——	——	——	——	——	——	3.25
c	——	——	——	——	——	——	——	——	——	——
or	0.83	2.19	1.18	3.01	1.36	1.60	3.72	3.19	2.95	4.67
ab	19.55	21.92	19.63	25.30	21.75	20.73	23.10	21.58	23.69	23.78
an	33.34	30.58	32.41	30.89	32.34	31.26	28.37	28.82	29.31	25.46
di	18.08	16.37	18.25	14.46	15.89	19.13	15.42	18.35	15.27	17.36
wo	9.24	8.32	9.32	7.35	8.12	9.78	7.83	9.39	7.76	8.79
en	5.30	4.45	5.26	3.93	4.62	5.59	4.17	5.42	4.17	4.48
fs	3.54	3.60	3.67	3.18	3.15	3.76	3.42	3.54	3.34	4.09
hy	14.54	14.92	16.19	8.15	13.71	13.53	17.84	16.97	21.61	19.20
en	8.72	8.25	9.54	4.51	8.15	8.09	9.81	10.27	12.00	10.04
fs	5.82	6.67	6.65	3.64	5.56	5.44	8.03	6.70	9.61	9.16
ol	8.09	7.57	6.51	12.08	9.39	8.18	5.14	5.64	1.22	——
fo	4.66	4.00	3.68	6.39	5.36	4.70	2.70	3.28	0.65	——
fa	3.43	3.57	2.83	5.69	4.03	3.48	2.44	2.36	0.57	——
mt	2.06	2.25	2.10	2.15	2.07	2.04	2.22	1.99	2.07	2.04
il	3.10	3.63	3.31	3.40	3.04	3.10	3.61	3.02	3.32	3.68
ap	0.43	0.59	0.45	0.57	0.45	0.43	0.59	0.45	0.57	0.57
Salic	53.72	54.69	53.22	59.20	55.45	53.59	55.19	53.59	55.95	57.16
Femic	46.30	45.33	46.81	40.81	44.55	46.41	44.82	46.42	44.06	42.85

Table 16.—*Chemical analyses of basalts and related rocks of the Yellowstone Plateau volcanic field and adjacent parts of the Snake River Plain. Analyses whose source is indicated as "new" are previously unpublished.*—Continued

Age or group	Basalts of Snake River Plain										
Formation or unit	Snake River Group and related basalts										
Locality	Antelope Flat		Elk Wallow	Bishop Mtn. Road	Ririe damsite	Antelope Flat		Southwest of Bishop Mtn.	Sand Creek Reservoir	Antelope Flat	Little Butte
Source	New					Table 4	New		Hamilton (1965)	New	
Sample No.	74IP-112	74IP-124	74IP-128	74IP-200	9SR-42A	8YC-446	74IP-109	74IP-111	YS88C	74IP-115	74IP-127
Analyses recalculated to 100 percent, less volatiles, Fe_2O_3/FeO calculated as 0.15											
SiO_2	45.21	45.31	45.37	45.74	46.38	46.47	47.18	47.26	47.63	47.81	53.84
Al_2O_3	14.97	15.27	15.02	15.04	17.81	16.74	16.26	17.09	14.63	13.94	16.38
Fe_2O_3	2.06	2.03	2.01	2.05	1.75	1.73	1.84	1.52	1.91	2.06	1.67
FeO	13.71	13.55	13.43	13.69	11.64	11.52	12.25	10.12	12.74	13.71	11.12
MgO	6.88	6.58	6.92	6.49	4.56	7.46	4.41	8.41	6.32	4.63	1.98
CaO	9.59	9.62	9.79	9.28	10.56	9.47	8.84	10.46	8.98	8.08	5.54
Na_2O	2.65	2.75	2.57	2.78	2.92	3.00	3.28	2.56	2.64	3.05	3.87
K_2O	0.64	0.53	0.51	0.65	0.33	0.59	1.23	0.36	0.84	1.49	2.79
TiO_2	3.24	3.31	3.28	3.20	3.21	2.29	2.98	1.80	3.47	3.79	1.68
P_2O_5	0.81	0.80	0.86	0.83	0.65	0.52	1.47	0.25	0.62	1.21	0.85
MnO	0.24	0.24	0.24	0.25	0.19	0.20	0.24	0.17	0.22	0.24	0.30
CIPW norms from recalculated analyses											
q	——	——	——	——	——	——	——	——	——	——	1.18
c	——	——	——	——	——	——	——	——	——	——	——
or	3.78	3.13	3.01	3.84	1.95	3.49	7.27	2.13	4.96	8.81	16.49
ab	22.42	23.27	21.75	23.52	24.71	25.39	27.75	21.66	22.34	25.81	32.75
an	27.06	27.76	27.94	26.64	34.51	30.47	26.01	34.08	25.59	19.94	19.08
di	12.63	12.24	12.44	11.63	11.38	10.81	6.91	13.24	12.36	10.27	2.46
wo	6.36	6.16	6.27	5.84	5.69	5.48	3.44	6.76	6.23	5.11	1.19
en	3.01	2.88	3.02	2.69	2.45	2.84	1.39	3.81	2.99	2.04	0.29
fs	3.26	3.20	3.15	3.10	3.24	2.49	2.08	2.67	3.14	3.12	0.98
hy	5.23	5.64	8.66	7.55	7.74	0.26	9.54	5.27	17.94	16.86	20.50
en	2.51	2.67	4.23	3.51	3.34	0.14	3.81	3.10	8.76	6.66	4.65
fs	2.72	2.97	4.43	4.04	4.40	0.12	5.73	2.17	9.18	10.20	15.85
ol	17.86	16.87	15.07	15.86	9.57	21.53	10.76	17.43	6.03	5.33	——
fo	8.14	7.59	7.00	6.99	3.90	10.95	4.05	9.84	2.80	1.98	——
fa	9.72	9.28	8.07	8.87	5.67	10.58	6.71	7.59	3.23	3.35	——
mt	2.99	2.94	2.91	2.97	2.54	2.51	2.67	2.20	2.77	2.99	2.42
il	6.15	6.29	6.23	6.08	6.10	4.35	5.66	3.42	6.59	7.20	3.19
ap	1.92	1.89	2.04	1.97	1.54	1.23	3.48	0.59	1.47	2.87	2.01
Salic	53.26	54.16	52.70	54.00	61.17	59.35	61.03	57.87	52.89	54.56	69.50
Femic	46.78	45.87	47.35	46.06	38.87	40.69	39.02	42.15	47.16	45.52	30.58

Table 16.—*Chemical analyses of basalts and related rocks of the Yellowstone Plateau volcanic field and adjacent parts of the Snake River Plain. Analyses whose source is indicated as "new" are previously unpublished.*—Continued

Age or group	Basalts and related rocks of Snake River Plain					
Formation or unit	Trachyandesite of High Point					
Locality	High Point	Antelope Flat		High Point	East of High Point	Antelope Flat
Source	New			Hamilton (1965)	New	
Sample No.	74IP-125	74IP-202	74IP-118	YS77	74IP-126	81YH-80
Analyses recalculated to 100 percent, less volatiles, Fe_2O_3/FeO calculated as 0.15						
SiO_2	59.80	61.46	62.03	62.50	63.43	64.35
Al_2O_3	15.69	17.32	15.71	16.65	16.09	16.19
Fe_2O_3	1.28	0.99	1.15	1.02	0.95	0.85
FeO	8.54	6.62	7.66	6.83	6.32	5.64
MgO	1.09	0.47	0.72	0.48	0.53	0.38
CaO	4.02	2.85	3.07	2.53	2.84	2.52
Na_2O	4.09	4.55	4.09	4.52	4.33	4.47
K_2O	3.77	4.83	4.16	4.35	4.53	4.80
TiO_2	1.03	0.55	0.86	0.73	0.63	0.53
P_2O_5	0.40	0.14	0.31	0.19	0.16	0.13
MnO	0.28	0.22	0.24	0.19	0.19	0.14
CIPW norms from recalculated analyses						
q	7.22	5.30	10.23	9.02	10.29	10.63
c	——	——	——	0.36	——	——
or	22.28	28.54	24.58	25.71	26.77	28.37
ab	34.61	38.50	34.61	38.25	36.64	37.82
an	13.32	12.57	12.22	11.31	11.09	9.93
di	3.48	0.57	0.86	——	1.71	1.51
wo	1.68	0.27	0.41	——	0.82	0.72
en	0.30	0.03	0.06	——	0.10	0.07
fs	1.50	0.27	0.39	——	0.79	0.72
hy	14.35	11.71	13.49	12.05	10.56	9.19
en	2.41	1.14	1.74	1.20	1.22	0.87
fs	11.94	10.57	11.75	10.85	9.34	8.32
ol	——	——	——	——	——	——
fo	——	——	——	——	——	——
fa	——	——	——	——	——	——
mt	1.86	1.44	1.67	1.48	1.38	1.23
il	1.96	1.04	1.63	1.39	1.20	1.01
ap	0.95	0.33	0.73	0.45	0.38	0.31
Salic	77.43	84.91	81.64	84.65	84.79	86.75
Femic	22.60	15.09	18.38	15.37	15.23	13.25

to most late Cenozoic basalts of the Western United States (Leeman and Rodgers, 1970) and are normatively classifiable as tholeiites, defined as having hypersthene in the CIPW norm (as calculated for a Fe_2O_3/FeO ratio of 0.15). They plot, however, on both sides of the alkali/silica variation line that separates Hawaiian tholeiites and alkali basalts (Macdonald and Katsura, 1964), and some of the Yellowstone analyses lie above the corresponding discrimination line of Irvine and Baragar (1971), based on a worldwide sample of basalts (Fig. 40). The basalts tend to form a compositional continuum from relatively low-potassium tholeiites that also have low Fe/Mg ratios, low titanium and phosphorous, and high calcium to relatively high-potassium tholeiites that are opposite in the corresponding characteristics. Most plateau-marginal basalts of the Yellowstone region are closer to the lower-potassium type, as are some basalts marginal to the Snake River Plain that are associated with rhyolites. The basalts of the Snake River Plain axis and some of the basalts of the Yellowstone Plateau region are relatively higher-potassium types although still hypersthene-normative and, thus, tholeiitic. The Snake River basalts generally have lower silica and higher alkalis, titanium, and phosphorous than even the relatively high-potassium units among the plateau-marginal basalts of Yellowstone.

In addition to the range of variation that exists between relatively potassium-poor and potassium-rich basaltic units of the Yellowstone Plateau, there is considerable range in the compositions of lavas within several of the major basaltic units. The ranges of variation are indicated in figure 41. Full discussion of models for origin of the chemical variation within and among these basaltic units is beyond the scope of this paper. It is noted here, however, that simple crystal fractionation does not appear to explain all of the observed compositional variation. In order to account for the total range by fractional separation of crystals, even within any of the individual basaltic units that show a smooth compositional variation, would require the separation of abundant magnetite in addition to olivine and plagioclase phenocrysts. Magnetite phenocrysts are not abundant in any of these rocks and are lacking in many.

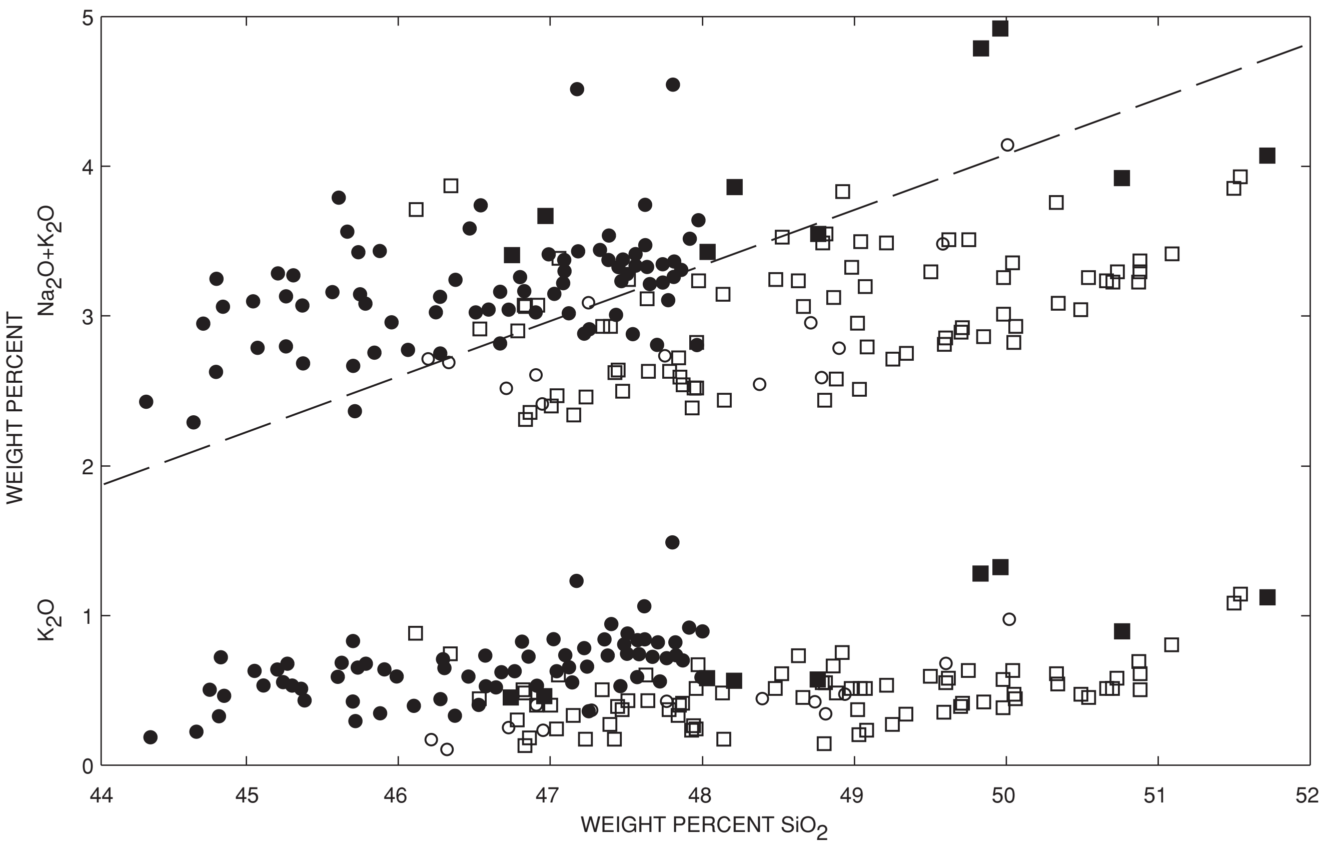


Figure 40.—Variation of total alkalis and of K_2O with SiO_2 in basalts of Yellowstone Plateau volcanic field (squares) and of Snake River Plain (circles). Solid symbols represent higher-K suites (axial basalts of the Snake River Group, the Madison River Basalt, and upper basalt flow of The Narrows); open symbols represent lower-K suites (all other basalts of the Yellowstone Plateau volcanic field and basalts marginal to the Snake River Plain). Dashed line divides compositional fields of Hawaiian tholeiites and alkali basalts (Macdonald and Katsura, 1964), shown only for reference and not as a discriminator of this data set. Data sources: this paper (table 16) for the Yellowstone region; Stone (1967), Tilley and Thompson (1970), Leeman and Vitaliano (1976), and Trimble and Carr (1976) for the Snake River Plain region.

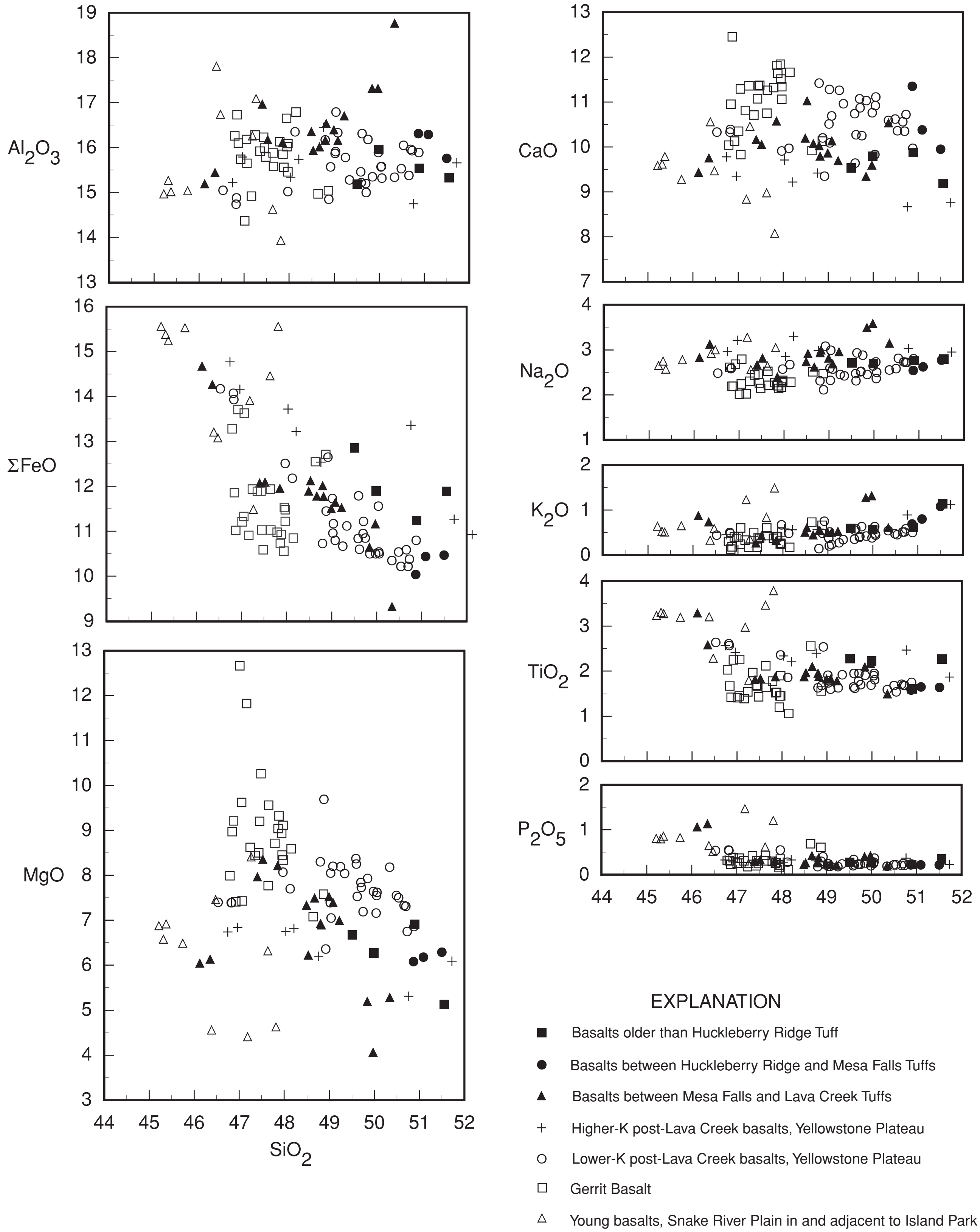


Figure 41.—Silica-variation diagram of basalts of the Yellowstone Plateau volcanic field, showing gradational ranges in compositions of lavas in weight percent. Includes all analyses from table 16 below 52 percent SiO_2, thus excluding rhyolite-contaminated basalts. ΣFeO, total Fe as FeO.

Some of the least evolved tholeiites of the Yellowstone region are chemically similar to some oceanic tholeiites. Although the least evolved basalts of this group have contents of K, Ti, P, and Fe/Mg nearly comparable to those of some mid-ocean ridge basalts, in most respects they are more nearly comparable to oceanic-island tholeiites (table 17). Most continental tholeiites have somewhat higher potassium contents. By contrast, the high contents of phosphorous, fluorine, and especially titanium of the relatively higher-potassium basalts of the Yellowstone-Snake River Plain region are notable (table 17).

The lower-K tholeiites of the Yellowstone Plateau field, compared to typical continental basalts, other than a few major flood-basalt provinces, have low Sr and notably low contents of large-ion lithophile elements such as U, Th, Pb, and Rb (Doe and others, 1982). Some of these basalts are comparable to mid-ocean ridge tholeiites in these trace-element characteristics; in none of the basalts of the Yellowstone Plateau, however, is Ba as low as in typical abyssal tholeiites (table 17). By comparison to the lowest-K tholeiites of the field, the successively higher-K basalts and the basalts of the Snake River Group have higher contents of Zr, rare-earth elements, Pb, Ba, and Rb.

Isotopes of lead, strontium, and neodymium from the basalts of Yellowstone were discussed by Doe and others (1982) and Hildreth and others (1991). The initial ratios of $^{87}Sr/^{86}Sr$ in basalts of the Yellowstone Plateau volcanic field range from 0.7038 to 0.7089, generally higher than typical oceanic values but distinctly lower than values of 0.7103±0.0015 associated with the rhyolites. Many of them lie within a fairly narrow range of values, 0.7060±0.0005. These relatively high ratios of $^{87}Sr/^{86}Sr$ are typical of basalts in a large region of the northwestern United States, including Yellowstone, the Snake River Plain, and the northern Nevada-southeastern Oregon basaltic plateaus (Leeman, 1982a). Values of ε_{Nd} range from -2.4 to -7.9. The generally low ε_{Nd} and high initial $^{87}Sr/^{86}Sr$ might be taken to suggest some degree of parentage from old lithospheric mantle, but significant crustal contributions are also probable. Most of the lead in the basalts probably is of crustal origin (Hildreth and others, 1991).

SEPARATE PARENTAGE OF THE RHYOLITIC AND BASALTIC MAGMAS

The petrologic data presented here for rhyolites of the Yellowstone Plateau volcanic field enable some discussion of the origin of the rhyolitic magmas. Perhaps the most direct conclusion from both field and petrologic relations is the negative one that the rhyolites, although closely associated with basalts, clearly are not products of fractional crystallization of the basalts. Such a conclusion is now conventional for bimodal rhyolite-basalt associations such as that of the Yellowstone Plateau volcanic field. Some consideration of the evidence is appropriate, however, because interpretations of such associations have varied greatly in the past.

First, the virtual lack of igneous rocks with compositions intermediate between the basalts and rhyolites and the great volumetric predominance of rhyolites over basalts in the Yellowstone Plateau field should be emphasized. If the rhyolites had originated by fractional crystallization from basaltic magmas, the resulting magmas of intermediate composition must not have erupted at all, forming only intrusive rocks. The alternative possibility that such intermediate magmas erupted only in the early stages of volcanism as lavas that are now completely buried is highly improbable, given the paucity of comparable xenoliths in the exposed volcanic rocks and the many exposures of the basal part of the section outside the caldera area. Necessarily, this conclusion would have to apply equally to all three major eruptive cycles. By extension, the same conclusions would apply to other rhyolite-basalt eruptive associations of the region. Not only is the crystal-fractionation hypothesis unlikely as an abstract argument, but the existence elsewhere of intrusive analogues to the Yellowstone-Snake River Plain volcanic association, some of them discussed in a later section on the rhyolitic magmas as batholiths, argues persuasively against it.

The presence of rhyolite-basalt mixed lava complexes in the field, although uncommon, further argues against an origin of the rhyolites by crystallization differentiation of the basalts. These complexes show that both rhyolitic and basaltic magmas were simultaneously available for eruption at several sites. Formation of these complexes may have involved a certain amount of chance in the opening of volcanic conduits and vents at times when both magmas were available in particular small areas; nevertheless the existence of the complexes suggests that magmas of intermediate composition probably are lacking beneath the Yellowstone Plateau field. These mixed-lava complexes are more consistent with the simultaneous formation and rise of contrasting rhyolitic and basaltic liquids in the region below the volcanic field.

The data on strontium, neodymium, and lead isotopes, reported by Doe and others (1982) and Hildreth and others (1991) and shown in part in figure 42, are also consistent with separate origins of the rhyolitic and basaltic magmas. Relative to the common leads of many igneous rocks, those of the Yellowstone region, both rhyolites and basalts, contain relatively lower amounts of the radiogenic isotopes ^{206}Pb, ^{207}Pb, and ^{208}Pb. Nevertheless, the rhyolites are systematically more enriched in $^{207}Pb/^{204}Pb$ relative to $^{206}Pb/^{204}Pb$ than the basalts (fig. 42), lower in $^{143}Nd/^{144}Nd$, and uniformly higher in $^{87}Sr/^{86}Sr$ (although their very low strontium contents make them susceptible to contamination). These values possibly reflect, at least in part, derivation of the rhyolitic magmas from source materials which contained less radiogenic neodymium and more radiogenic lead and strontium than the principal source materials of the basaltic magmas. Chemical differentiation by fractional crystallization

Table 17.—*Chemical analyses of basalts from selected igneous provinces. Recalculated to 100 percent, less volatiles, all Fe as FeO*

	1	2	3	4	5	6	7	8	9	10	11	12
	Yellowstone Plateau		Snake River Plain	East Pacific Rise	Medicine Lake Highland	Talasea, New Britain	Columbia River Plateau	Oahu, Hawaii	Thingmuli, Iceland	Lesotho, Basutoland	Oahu, Hawaii	Mojave Desert, Calif.
	Recalculated to 100 percent, less volatiles, all Fe as FeO											
SiO_2	48.88	48.06	47.72	50.11	49.30	51.90	50.00	49.26	50.04	51.16	47.11	47.87
Al_2O_3	16.19	15.04	14.66	16.93	19.04	16.01	14.75	15.86	12.94	14.37	15.68	16.59
ΣFeO	10.74	12.54	14.48	8.73	9.18	9.56	11.77	11.04	15.04	11.31	11.90	10.09
MgO	8.31	8.09	6.33	8.30	7.47	6.77	7.18	7.86	5.16	7.65	8.28	7.43
CaO	11.44	9.92	9.00	11.28	10.11	11.81	10.91	10.42	10.00	10.61	8.96	9.65
Na_2O	2.32	2.57	2.64	2.69	3.06	2.43	2.93	2.21	2.91	2.65	3.04	3.94
K_2O	.14	.67	.84	.24	.44	.44	.45	.29	.47	.76	1.24	1.51
TiO_2	1.63	2.36	3.48	1.41	1.17	.80	1.52	3.02	2.78	1.16	3.02	2.24
P_2O_5	.18	.56	.62	.12	.14	.11	.27	.60	.42	.17	.60	.51
MnO	.17	.19	.22	.19	.09	.17	.22	.17	.24	.16	.17	.17
$\Sigma FeO/MgO$	1.3	1.5	2.3	1.1	1.2	1.4	1.5	1.4	2.9	1.5	1.4	1.4

Sources of Data:

1. Osprey Basalt (Table 16, no. 74Y-159F)
2. Falls River Basalt(Table 16, no. 78YH-31)
3. Snake River Basalt, west of Island Park (Hamilton, 1965, Table 15, no. YS88c)
4. Abyssal tholeiite (Engel and others, 1965, Table 1, no. AD3)
5. Modoc Basalt, high-Al_2O_3 basalt (Anderson, 1941, Table 1, no. 3)
6. Basalt (Lowder and Carmichael, 1970, Table 2, no. 311)
7. Picture Gorge Basalt, lower flows at Picture Gorge (Waters, 1961, Table 2, no. 3)
8. Tholeiite, lower member, Waianae volcanic series (Macdonald and Katsura, 1964, Table 1, no. C-5)
9. Tholeiite (Carmichael, 1964, Table 2, no. 5)
10. Karoo Basalt (Cox and Hornung, 1966, Table 2, no. B21)
11. Alkali basalt, upper member, Waianae volcanic series (Macdonald and Katsura, 1964, Table 2, no. C-30)
12. Mt. Pisgah flow, trachybasalt (Smith and Carmichael, 1969, Table 6, no. 255)

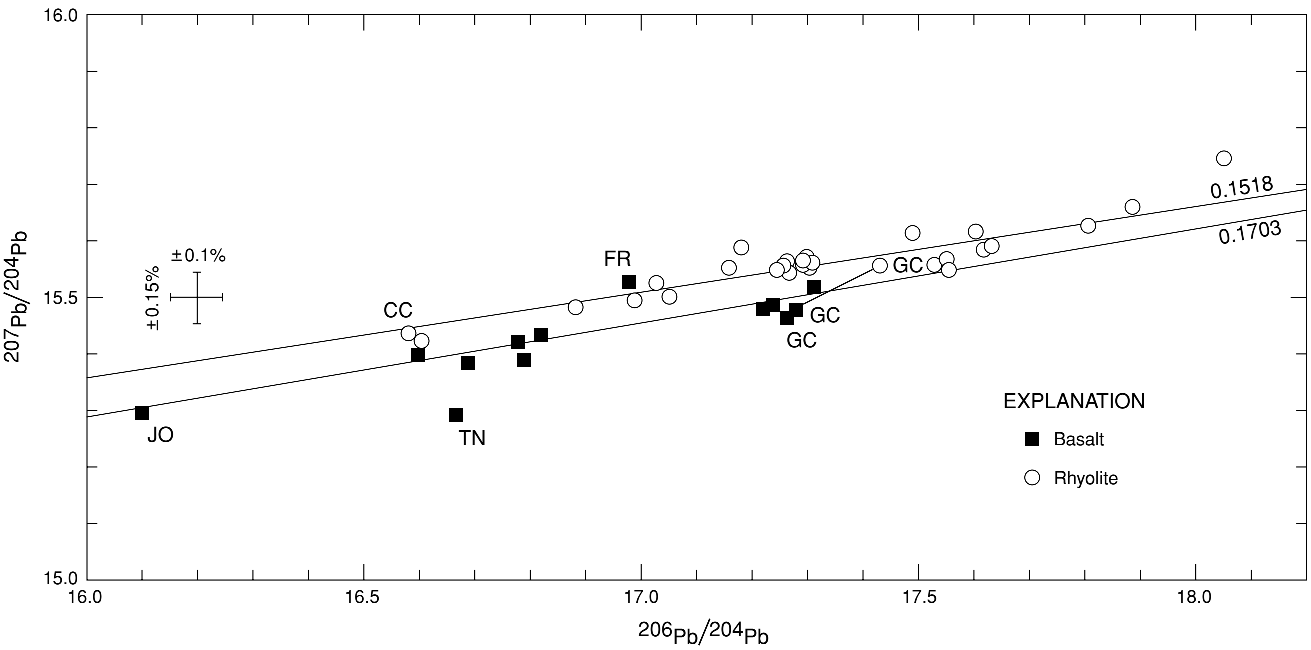


Figure 42.—Lead-isotopic ratios ($^{207}Pb/^{204}Pb$ plotted against $^{206}Pb/^{204}Pb$) from data of Doe and others (1982). JO, Junction Butte Basalt, Overhanging Cliff flow; TN, basalt of The Narrows, upper flow; FR, Falls River Basalt; GC, three samples from mixed rhyolite and basalt of Gardner River, connected by short line; CC, Cougar Creek dome. Lines labeled with values for their slopes are least-squares best-fit lines to the data points for rhyolites (0.1518) and basalts (0.1703); cross shows analytical uncertainties for best analyses.

is incapable of producing directly such differences in the ratios of heavy isotopes, and the amount of incorporation of upper crustal materials that would be required to produce them from uniform initial values by bulk contamination would show its effects clearly in the trace-element concentrations of the rocks.

The lead and strontium isotopic data of Doe and others (1982) provide further evidence on the origin of the mixed-lava complexes. The isotopic ratios of rhyolite and basalt of the mixed lavas of Gardner River differ from one another but are similar to those of other rhyolites and basalts of the field. Furthermore, the Pb-isotopic composition of the rock of intermediate composition analyzed from the mixed-lava complex of Gardner River lies on a straight line between the compositions of the end-member rhyolite and basalt (fig. 42). Thus, the rhyolite-basalt complexes seem to have formed by preeruptive mixing of basaltic and rhyolitic magmas that had independent origins.

Relations discussed in this section are in contrast to corresponding relations in some calc-alkalic volcanic fields where large volumes of rhyolite may have more direct genetic relations to voluminous rocks of intermediate composition. A pertinent example of the latter is the San Juan volcanic field of southwestern Colorado (Larsen and Cross, 1956; Ratté and Steven, 1964; Lipman and others, 1970; 1972; Lipman, 1975). In that field, the least evolved magma of a voluminous sequence was of potassic-andesite composition. The corresponding volcanic rocks are abundant in the San Juan volcanic field, and all compositions intermediate between it and the voluminous rhyodacites and rhyolites are present. The volumes of the erupted materials do not correspond ideally to the proportions of a differentiation sequence, and various mixing and contamination processes clearly affected the suite. Nevertheless, the data are at least within reasonable limits to permit the assumption of crystallization differentiation as a significant process in the formation of the rhyolitic magmas. Furthermore, a later rhyolite-basalt association, also present in the San Juan Mountains, is petrologically similar to that of Yellowstone, and its eruption accompanied and followed a major change in the tectonic setting of volcanism to regionally distributed extension more like that of Yellowstone (Lipman and others, 1970; Lipman and others, 1972). A few rhyolite-basalt mixed-lava complexes are associated with this rhyolite-basalt sequence of the San Juan Mountains but no such complexes occur in the older predominantly andesitic volcanic association (Lipman, and others, l970; Steven and Lipman, 1976; Lipman and others, 1978).

This comparison of relations in the Yellowstone-Snake River province with voluminous, predominantly intermediate-composition associations such as the San Juan Mountains complements a previously noted point: the rhyolites of these two types of igneous associations, although alike in many ways and in some instances similar in composition, typically differ from one another chemically (table 15; fig. 39).

In review, relations noted in this section are: (1) the lack of intermediate-composition rocks in the Yellowstone Plateau volcanic field (other than those of mixed-lava origin), (2) the great volumetric predominance of erupted rhyolites over basalts, (3) consistent differences in the geologic relations and major- and minor-element compositions of the Yellowstone Plateau rhyolites compared to those of voluminous, continuously differentiated (or mixed) andesite-rhyolite associations, (4) isotopic evidence for different magmatic parentage of the rhyolitic and basaltic magmas, and (5) the repeated simultaneous presence of rhyolitic and basaltic magmas in the field as a whole.

The foregoing factors point clearly to an origin of the rhyolitic magmas of the Yellowstone Plateau volcanic field by a process other than fractional crystallization from the mafic magmas represented by the basalts of the field. It is most likely that the rhyolites reflect large-scale melting of lower-crustal materials. This topic will be returned to in a section on origin and evolution of the Yellowstone magmas after a discussion of the tectonic setting of the Yellowstone Plateau.

MAGMOTECTONIC EVOLUTION

This final section interprets some relations between the Yellowstone Plateau volcanic field and surrounding regions, the origins of the Yellowstone magmas, some aspects of crustal and mantle evolution, and the probable volcanic future of the Yellowstone region. Thus, much of what follows is quite speculative.

REGIONAL LATE CENOZOIC TECTONICS

The magmatic history of the Yellowstone Plateau volcanic field must be interpreted in relation to the late Cenozoic volcanism and tectonics of the Snake River Plain, which it adjoins, and to the contemporaneous tectonics of parts of the Northern and Middle Rocky Mountains and the northern part of the Basin and Range province, within which these two volcanic regions lie. Considered first are the regional tectonics of the Rocky Mountain and basin-range regions as they have evolved in Miocene and later time. Summaries of the late Cenozoic tectonic history of the larger region have been drawn on in the following discussion (Gilluly, 1963; Hamilton and Myers, 1966; Christiansen and Lipman, 1972; Stewart, 1978; Eaton and others, 1978; Christiansen and McKee, 1978; Davis, 1980; Eaton, 1980; Christiansen and Yeats, 1992). Tectonic interpretations of the region that more immediately surrounds the Yellowstone Plateau volcanic field include those of Pardee (1950), Myers and Hamilton (1964), Fraser and others (1964), Ruppel (1972), Love and others (1973), Love and Keefer (1975), Smith and others (1977), Anders and others (1989), Pierce and Morgan (1992), and Smith and Braile (1994).

REGIONAL FAULTING AND SEISMICITY

The Yellowstone Plateau volcanic field and the eastern Snake River Plain are flanked by linear mountain ranges separated by parallel valleys. The ranges and valleys generally trend about northward in the southern and eastern parts of the region, but farther north and west they arc toward the northwest (fig. 44). These parallel linear topographic features are fault blocks, most of them tilted. The bounding faults, where specific information is available, are normal. Fault displacements are mainly dip-slip, but some may have relatively minor strike-slip components. Such a pattern of predominant normal faults separating large crustal blocks typifies the Basin and Range physiographic and tectonic province south of the Snake River Plain. Although Ruppel (1972; 1982) considered the linear ranges of Idaho north of the Snake River Plain to be vertically uplifted blocks bounded by monoclinal drape folds and additionally (Ruppel, 1964) interpreted the presence of a set of north-trending strike-slip faults across those ranges, most investigators attribute a basin-range origin to these ranges (for example, Baldwin, 1951; Scott and others, 1985). The magnitude-7.3 Borah Peak earthquake of 1983 produced displacement of more than a meter on the 49°-dipping range-front normal fault of the Lost River Range (Crone and Machette, 1984; Barrientos and others, 1987). Among others, Pardee (1950), Gilluly (1963), and M. W. Reynolds (1979) have called attention to the generally similar pattern that has characterized late Cenozoic development of the northern Rocky Mountain region as well as the Basin and Range province.

The extensional origin of basin-range structure is now generally recognized. Also widely noted is the association of at least some strike-slip deformation with the predominant block-tilting and normal or listric faulting of the region (for example, Locke and others, 1940; King, 1959, p. 157-158; Longwell, 1960; Shawe, 1965; Hamilton and Myers, 1966; Stewart, 1967; Stewart and others, 1968; Stewart, 1978; Davis, 1980; Zoback, 1983; 1989). Although there is not yet a general consensus on major lithospheric mechanisms for the late Cenozoic regionally distributed extensional tectonic pattern, one approach lies in the relations of extensional faulting to plate tectonics as interpreted from the sea floor of the eastern Pacific (McKenzie and Morgan, 1969; Atwater, 1970). Atwater related Miocene and younger regional extension of the Western United States to the predominantly right-lateral transform tectonics of coastal California, the entire system reflecting the response of a wide marginal "soft" zone of the North America lithospheric plate to the relative motion between it and the Pacific plate; this North America-Pacific interaction has developed through late Cenozoic time with the disappearance of most of a previously intervening plate between the East Pacific Rise and the North American continental margin. Christiansen and Lipman (1972) and Christiansen and McKee (1978) adopted this concept in reviews of the tectonic relations of volcanism in the Western United States during late Cenozoic time.

Although tectonic extension has affected much of the Western United States at various times since the Late Cretaceous to early Eocene Laramide orogeny (Christiansen and Yeats, 1992), the widely distributed basin-range extension of late Cenozoic time is most germane to understanding the Yellowstone-Snake River Plain system. The age of initiation of this late Cenozoic extension varies from place to place. In the northern Great Basin and northern Rocky Mountains region, this structural pattern emerged during middle Miocene time; region-wide structures of basin-range type have not been documented for times before the Miocene (Christiansen and McKee, 1978). Recent work has shown that basin-range extension began about 17 million years ago but that the modern basins and ranges of the Great Basin region are generally no older than about 10-12 Ma (Zoback and others, 1981; Christiansen and Yeats, 1992).

Evidence for just when during Miocene time tectonic extension began in the region surrounding Yellowstone and the Snake River Plain has been debated. Some investigators state or imply that such activity started near the beginning of the Miocene. Love (1956c; 1977), for example, indicated that the lower(?) and middle Miocene Colter Formation of Jackson Hole may have accumulated in a normal-fault-bounded basin ancestral to modern Jackson Hole. It seems more likely from the available evidence that rocks of early and middle Miocene age in the Jackson Hole area were still filling remnants of an older basin with a trend inherited from the Laramide (latest Cretaceous and early Tertiary) deformation of the Middle Rocky Mountains (Love and others, 1973). By contrast, the Teewinot Formation (Love, 1956c; Love and others, 1973; 1978, p. 34-35) of late Miocene age clearly did accumulate in a predecessor of Jackson Hole, with the same bounding features, overall size, and orientation as the present valley (although with much lower relief than at present; Love, 1977). Robinson (1960; 1961) also showed that upper Miocene and overlying deposits in western Montana are related to renewed faulting, block tilting, and regional uplift after an interval of quiescence. Ruppel (1964; 1982, p. 14-15) noted the difficulty in dating the beginning of movement on the range-bounding faults of eastern Idaho but concluded that the faults probably began to form in the Miocene, perhaps during middle Miocene time. Scott and others (1985), however, presented evidence that the present ranges in Idaho north of the eastern Snake River Plain date only from 4-7 Ma.

Some investigators have proposed that initial basin-range extension occurred successively later from the central Great Basin eastward to the present belt of active seismicity at the margin of the region (for example, Scott and others, 1971; Allmendinger, 1982; Kellogg and Marvin, 1988; Rodgers and others, 1990; Smith and Braile, 1994), but this is demonstrably not true. Widely distributed basin deposits of generally late Miocene age, mapped locally as Salt Lake Formation, Starlight Formation, or Teewinot Formation, demonstrate that the present basin areas had been blocked out by extension at least as early as about 10 Ma as far eastward as Jackson Hole (Christiansen and McKee, 1978; Christiansen and Yeats, 1992). What can be documented is that a wave of enhanced rates of extension did migrate eastward south of the eastern Snake River Plain from about 14-12 Ma near its southwest end (Compton, 1983) to about 4-2 Ma near its northeast end (Anders and others, 1989; Pierce and Morgan, 1992). This wave of enhanced extension rates also migrated outward from the margins of the plain to about 50-100 km away at present.

Although the initial time of extensional faulting was during the middle Miocene in the Northern and Middle Rocky Mountains region, such movement clearly has continued in the Yellowstone region and in an arcuate belt around the eastern Snake River Plain essentially without break to the present. Faulting in and around the Yellowstone Plateau has continued since formation of the Yellowstone caldera 640,000 years ago and its filling, up to 70,000 years ago, by rhyolitic lavas. A zone of seismicity and of faults with large bedrock scarps and significant Holocene displacement forms a parabolic arc around the north, east, and south sides of the eastern Snake River Plain, including the Yellowstone Plateau (Anders and others, 1989; Smith and Arabasz, 1991; Pierce and Morgan, 1994; Smith and Braile, 1994). The destructive Borah Peak earthquake of 1983 (magnitude 7.3) and Hebgen Lake earthquake of 1959 (magnitude 7.5; Doser and Smith, 1989) were the largest historic events in this zone. Only these two historic earthquakes produced ground displacements on fault scarps (Fraser, 1964; Barrientos and others, 1987) although very young prehistoric scarps are present widely in the surrounding region (Pardee, 1950; Love, 1961; Myers and Hamilton, 1964; Witkind, 1975a; b; c; Pierce and Morgan, 1992). Aftershocks of the Hebgen Lake earthquake were numerous (Murphy and Brazee, 1964). Earthquakes since the 1959 Hebgen Lake earthquake, monitored both by long-term networks and by several detailed short-term surveys, show virtually continuous seismic activity (Pitt, 1989; Smith and Arabasz, 1991; see fig. 3*B*). A significant proportion of this seismicity is concentrated in the northwest-trending zone that includes the main shock of 1959, many of its aftershocks (Trimble and Smith, 1975), and the principal faults related to that earthquake (Witkind, 1964). Numerous other earthquakes, including many within the Yellowstone caldera (for example, Pitt and others, 1979), are scattered more widely in the area of Yellowstone National Park (fig. 3*B*). Seismicity is relatively sparse in the area immediately south of the Yellowstone caldera to Jackson Hole (Smith and Arabasz, 1991)

RATES OF FAULT DISPLACEMENT

Because the several widespread, well-dated ash-flow sheets of the Yellowstone Group and similar older units flank the major late Cenozoic fault blocks around the Yellowstone

Plateau, rates of uplift of certain of the fault blocks during latest Cenozoic time can be documented (fig. 43). It is apparent from the distributions and thickness variations of the sheets that, before emplacement of the first late Cenozoic ash-flow sheets, each of the major fault blocks occupied more or less its present location but had lower relief than now (see fig. 6).

The best-documented example of a major fault block is the Teton Range (see fig. 28). The late Cenozoic ash-flow sheets dip westward more steeply on the west flank of the range than near the crest (fig. 43). The Conant Creek Tuff and tuff of Kilgore of 6.0 and 4.3 Ma respectively (Obradovich and Christiansen, in press; Morgan, 1992), erupted from volcanic sources now largely buried by younger lavas of the Snake River Plain (fig. 6) and were emplaced as ash-flow sheets along the west side of the Tetons, crossed their lower north end, and extended southward into Jackson Hole (Christiansen and Love, 1978). These ash-flow sheets in Jackson Hole, now largely buried, are exposed on small fault blocks that have been uplifted near and within Jackson Hole, most notably on Signal Mountain. Love (1956a) and Pierce and Morgan (1992, p. 10) noted that the Kilgore (or the lithologically similar Conant Creek) on Signal Mountain dips somewhat more steeply than the Huckleberry Ridge Tuff that is exposed farther west on the mountain. Gravel lies between these two welded tuffs on Signal Mountain and thickens downdip, further suggesting a discordance between the two sheets there. The Conant Creek Tuff is present at elevations as high as about 2,500 m in the upper Conant Creek drainage but must be downfaulted out of sight in northern Jackson Hole, north of Jackson Lake. Dip projections there suggest a possible structural relief of more than 1,400 m across the Teton fault (fig. 43).

The distribution of the 2.1-Ma Huckleberry Ridge Tuff around the west, north, and east sides of the Teton Range is very similar to that of the tuff of Kilgore and Conant Creek Tuff, except that the Huckleberry Ridge offlaps their highest parts and overlaps their lower parts (pl. 1). The Huckleberry Ridge Tuff does not dip so steeply westward on the west side of the range as do the Conant Creek Tuff and tuff of Kilgore (fig. 43). Structural relief on the Huckleberry Ridge Tuff across the Teton fault zone at the north end of Jackson Hole is about 950 m (fig. 43). The 640-ka Lava Creek Tuff overlaps faulted and eroded edges of the Huckleberry Ridge Tuff in the northern Tetons but dips only very gently off the core of the range. The Lava Creek rises to an elevation of about 2,360 m in the northernmost part of the range and is present as an island in the floodplain of the Snake River just north of Jackson Lake (pl. 1) at an elevation of about 2,130 m. The structural relief on the Lava Creek Tuff across the Teton fault zone in this area, therefore, is about 230 m. With these figures, the rate of differential uplift of the northern Teton Range above northern Jackson Hole during latest Cenozoic time can be calculated. Between about 4.3 and 2.1 Ma the rate averaged 0.2 mm/yr; between about 2.1 and 0.64 Ma

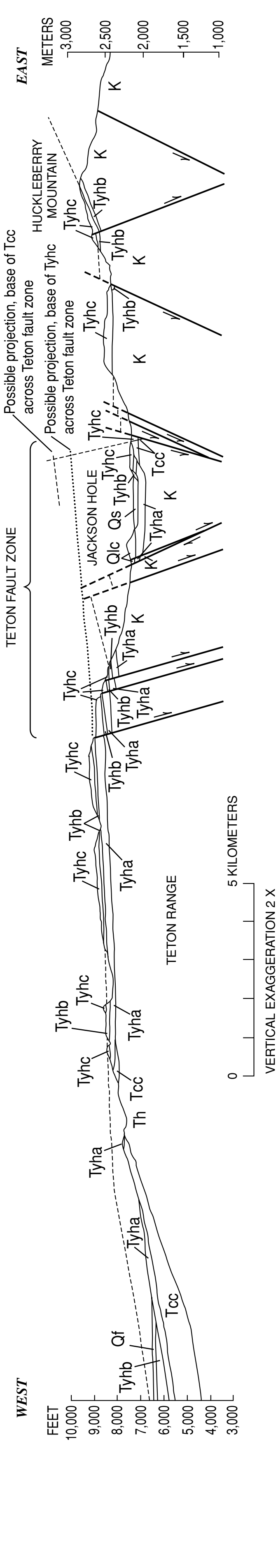


Figure 43.—Cross section through the northern Teton Range, Jackson Hole, and Huckleberry Mountain, along the north-bounding line of T. 47 N., and its eastward extrapolation. Section extends from the west border of the Grassy Lake Reservoir quadrangle to Rodent Creek (see plate 1). Letter symbols: Qs, surficial deposits; Qf, Falls River Basalt; Qlc, Lewis Canyon Rhyolite; Tyhc, Huckleberry Ridge Tuff member C, Huckleberry Ridge Tuff; Tyhb, member B, Huckleberry Ridge Tuff; Tyha, member A, Huckleberry Ridge Tuff; Tcc, Conant Creek Tuff and tuff of Kilgore; Th, Hominy Peak Formation; K, Cretaceous sedimentary rocks.

the rate was about 0.5 mm/yr; the average rate since 0.6 Ma has been about 0.4 mm/yr. Thus, uplift of the range may have accelerated during the Pleistocene. The average rate of uplift for the past 4.3 m.y. has been 0.3 mm/yr; at that rate, the total offset of 2,300 m on this part of the fault zone could have taken place in about 8 m.y., or roughly the time during which the Teewinot Formation and younger deposits have accumulated in the fault-block basin of Jackson Hole.

Pierce and Morgan (1992, fig. 6), using similar data, calculated a much higher rate, somewhat greater than 1 mm/yr, for aggregate displacement across the Teton fault farther south, near Signal Mountain (see pl. 3), and Byrd and others (1994) concluded that much of the total displacement between the central Tetons and Jackson Hole has accumulated since 2.1 Ma.

FAULT BLOCKS ADJACENT TO THE YELLOWSTONE PLATEAU

Faults of late Quaternary age, postdating the Lava Creek Tuff, occur rather widely around the margins of the Yellowstone Plateau, mainly as parts of zones that bound major blocks extending outward from the plateau. The principal fault zones on which Quaternary displacements are mapped are shown on plates 1 and 3. All these faults are normal (although Ruppel, 1972, regarded some of them as vertical). They include fault zones on the east side of the Teton Range, the east side of the Red Mountains and Huckleberry and Wildcat Ridges, east of Chicken and Big Game Ridges, in the upper valley of the Yellowstone River and the Southeast Arm of Yellowstone Lake, in the Lake Butte-Mirror Plateau-Amethyst Mountain area, in the Washburn Range-Lava Creek-Arrow Canyon area, on the east side of the Gallatin Range, and in the Madison Valley in the area of Hebgen Lake and West Yellowstone (also see fig. 5). Most of the faults of these zones, both those predating and those postdating the Lava Creek Tuff, trend between north and northwest. The zones have generally linear trends except for the arcuate faults of the Mirror Plateau area, which probably were considerably influenced by movements of the Yellowstone magma body, as discussed earlier (also see Eaton and others, 1975).

Most of the major faults of the regional system in the area south of the Yellowstone Plateau are downthrown to the east and bound the east sides of west-tilted fault blocks (for example, the Teton Range, the Red Mountains, Chicken Ridge). Smaller normal faults of these same zones have displacements either the same as or opposed to those of the major faults. Faults of the upper Yellowstone River-Southeast Arm form a distinct graben, but like others south of the Yellowstone Plateau they bound the east side of a west-tilted block (Two Ocean Plateau). Comparison of plate 3 and figure 5 shows that each of the major tilted fault blocks south of the plateau, starting with the Teton Range, is of lower height than those to the west. The Absaroka Mountains lie east of the easternmost fault zone of this system and have been only very gently deformed in late Cenozoic time (see, for example, Fisher and Ketner, 1968). There was little or no differential uplift of late Cenozoic age between the Absaroka Range and the adjacent Big Horn Basin to the east; only broad uplift of the entire Absaroka-Big Horn region is apparent. Thus each of the major east-dipping normal fault zones of the area south of the Yellowstone Plateau has less aggregate displacement than the next zone to the west.

The significance of these eastward-decreasing heights of successive fault blocks is uncertain. Possibly, successive fault blocks east of the Teton Range have each been active at approximately the same rates but for successively shorter periods of time. More likely, rates of uplift were successively less for the more easterly major fault blocks during the same span of time, representing a decline in uplift and extension toward the east edge of the region of late Cenozoic tectonic activity.

Love and Keefer (1975) interpret the Red Mountains block as having subsided somewhat since uplift of the whole area south of the Yellowstone Plateau. Their analysis, however (Love and Keefer, 1975, p. D53), assumes absolute uplift or subsidence immediately adjacent to each of the major fault zones rather than a differential uplifting and tilting of the entire assemblage of blocks that might better be assumed.

Normal faults north of the Yellowstone Plateau form a less distinct series of linear zones and intervening tilted blocks than those to the south. Ruppel (1972), in his discussion of the late Cenozoic normal faults of the northern area, regarded the Gallatin Range as a horst bounded by vertical faults. A pattern of Quaternary faults in the northern area (pls. 1 and 3) seems to define one zone (the east Gallatin fault zone) bounding the east side of the Gallatin Range; one zone east of Gardners Hole and parallel to Obsidian Creek; another broad and somewhat diffuse zone farther east that includes areas near Lava Creek, Arrow Canyon, and the western Washburn Range; and the arcuate zone that extends northwest and southeast from the Mirror Plateau area. Most of the largest faults of these sets that offset the Lava Creek Tuff, like major faults south of the plateau, have displacements down to the east and bound small blocks that are tilted westward. Other faults, mainly smaller, have opposed displacements.

Ruppel (1972, p. A51) inferred, on the basis of discontinuous topographic lineaments, that the Gallatin Range is a northward structural extension of the Teton Range, the intervening part of the block being buried by the Yellowstone Plateau west of the Upper and Lower Geyser Basins. If so, this fault-block system may branch; the post-Lava Creek fault zone east of the Gallatin Range swings southeastward to merge with faults that extend generally southward from the Gardners Hole-Swan Lake Flat-Roaring Mountain area. The merged fault zone extends beneath caldera-filling lavas of the rhyolite plateau in the vicinity of Norris Geyser Basin and Gibbon Meadows. In particular, the fault zone that follows the course of Straight Creek and bounds the southwest side of Gibbon Meadows (pl. 1) parallels two other faults, one that extends northwestward from Norris Geyser Basin across Grizzly Lake and Winter Creek and another that extends from Norris Geyser Basin generally along Obsidian Creek. All three of these fault zones change from northwestward to northward trends away from the margin of the Yellowstone caldera. The pre-Lava Creek ancestry and continued post-Lava Creek control of volcanism of the faults north of Norris Geyser Basin (the Norris-Mammoth corridor) are demonstrated by partly buried fault blocks of pre-Quaternary rocks and a linear zone of volcanic vents, including various source vents of the Swan Lake Flat Basalt, the Gardner River mixed lavas, three rhyolite domes of the Obsidian Creek Member, the sources of the Crystal Spring and Obsdian Cliff flows, and the Roaring Mountain hydrothermal area. This band continues across the Yellowstone caldera from the rhyolitic vents just south of Norris Geyser Basin, through the Central Plateau, to flows of the Flat Mountain Arm area (fig. 5).

The continuity of these features suggests strongly that fault zones on the east side of the Gallatin Range (fig. 2) and near Gardners Hole (pl. 1) are structural continuations of faults east of the Red Mountains-Huckleberry Ridge-Wildcat Ridge block, which lies northeast of Jackson Hole (pl. 3). The line of rhyolitic vents and fractures across the Pitchstone and Madison Plateaus, which continues the trend of the Teton fault zone into the structurally depressed block of the Madison Valley near West Yellowstone (fig. 5), suggests at least a branching relation between the buried extensions of the Teton and Madison Ranges, as noted by Hamilton (1960b). Ruppel's suggested alignment (1972) of the Teton and Gallatin Ranges would then add a pattern of branching and bridging between linear fault blocks (see fig. 6), a pattern found rather commonly in the basin-range region although Ruppel (1972, p. A55-A57) regarded it as unique.

These relations suggest that, although the Quaternary normal faults of northern Yellowstone Park do not define such easily identifiable blocks as those south of the Yellowstone Plateau and do not outline such a clear sequence of serially decreasing height eastward, they do represent structural continuations, perhaps with some splaying or "horsetailing," of the major fault zones mapped south of the plateau. The master faults of crustal scale are perhaps those east of the Teton and Madison Range blocks, east of the Red Mountains-Gallatin Range group of blocks, and those at the east edge of the system as a whole along the upper Yellowstone River and the western Washburn Range. This pattern is consistent with the position of the Yellowstone Plateau with respect to regional seismicity (fig. 3). The seismic belt extending into the Yellowstone Plateau from the south branches into a north-trending zone that gradually diminishes across Montana (fig. 48) and a northwest- to west-trending zone through the Hebgen fault system and on westward into Idaho (Smith and Sbar, 1974; Trimble and Smith, 1975; Stickney and Bartholomew, 1987; Smith and Arabasz, 1991).

RELATIONS BETWEEN STRUCTURE AND VOLCANISM

It should be reemphasized here that certain major extensional structures evidently controlled the locations of the principal volcanic features of the Yellowstone Plateau volcanic field. It was noted previously that the locations of many individual volcanic vents apparently were controlled of by major normal-fault zones, including those of the Pitchstone and Madison Plateaus and of the Central Plateau and Norris-Mammoth corridor. In addition, the main subsided blocks of the three caldera complexes of the volcanic field, where known, all lie on the intersections of major extensional fault zones or downdropped blocks with the volcanic axis of the eastern Snake River Plain-Yellowstone Plateau. Both the Big Bend Ridge segment of the first-cycle caldera and the second-cycle Henrys' Fork caldera lie on the intersection of this axis with the Madison Valley fault zone west of the Madison Range (along which are conspicuous prehistoric but young fault scarps; Pardee, 1950; Witkind, 1972; Pierce and Morgan, 1992) and the trend of Quaternary faults in Teton Basin, west of the Teton Range (pl. 3). The Snake River segment of the first-cycle caldera and the Mallard Lake segment of the third-cycle Yellowstone caldera lie on the intersection of the major volcanic axis with Jackson Hole and its possible partial extensions in the basin of West Yellowstone and Hebgen Lake, including faults related to the 1959 earthquake and the coincident major belt of continuing earthquakes (fig. 3*B*). The resurgent dome of the Mallard Lake caldera segment is aligned parallel to the northern part of this fault zone, projecting directly toward the principal linear belt of 1959. The Sour Creek segment of the Yellowstone caldera lies on the projected intersection of the major volcanic axis with the easternmost major linear fault zones—those of the upper Yellowstone Valley-Yellowstone Lake area and of the western Washburn Range-Arrow Canyon-Lava Creek area.

Just as regional structures controlled volcanic structures, the tectonic pattern in turn has been influenced locally by the structural imprint of the volcanic field. An apparent decrease in displacement on the largest fault zones (notably the Teton and Gallatin systems) with approach to the Yellowstone caldera probably reflects in some measure their partial cover by younger volcanic rocks on the plateau in contrast to cumulative displacements farther from the plateau that include visible effects of older movements. Nevertheless, individual faults do lose displacement and the end abruptly at the caldera margin. In addition, there are many more prominent north-trending normal faults within a zone about 25-45 km wide around the Yellowstone Plateau than there are within the same longitudinal band farther north or south (pl. 3; fig. 5). These factors, together with the seismicity pattern (Smith and Arabasz, 1991), appear to show that the Yellowstone caldera and related older volcanic structures occur within a zone of concentrated crustal stress and have acted as a local boundary against which much displacement on the regional fault system has been taken up. Furthermore, broad uplift of the region over the successive Yellowstone magma chambers has raised the general structural level toward its margins but has resulted in local subsidence of individual tectonic fault blocks as they enter the zone of uplift. The net effect is of increased average elevation but of decreased local relief at the immediate boundaries on the Yellowstone Plateau.

EVOLUTION OF THE SNAKE RIVER PLAIN-YELLOWSTONE VOLCANIC AXIS

Only in the last few decades has much detailed work been published on the late Cenozoic geology of the Snake River Plain, and especially of the margins of the eastern plain (for example, see Carr and Trimble, 1963; Prostka and Hackman, 1974; Trimble, 1976; Trimble and Carr, 1976; Prostka and Embree, 1978; McBroome, 1981; Trimble, 1982; Bonnichsen and Citron, 1982; Compton, 1983; Embree and others, 1983; Morgan and others, 1984; Morgan, 1988; Kellogg and Marvin, 1988; Morgan and Bonnichsen, 1989; Anders and others, 1989; Morgan, 1992). Presently available data indicate that the eastern Snake River Plain has developed progressively through late Cenozoic time and shares a unified tectonomagmatic history with the Yellowstone Plateau volcanic field (Christiansen and Lipman, 1972; Armstrong and others, 1975; Christiansen and McKee, 1978; Leeman, 1983a; Pierce and Morgan, 1992; Smith and Braile, 1994). Various parts of the combined region now represent different evolutionary stages of this unified development, and the eastern Snake River Plain and the Yellowstone Plateau each can shed some light on the origin of the other.

GEOLOGIC RELATIONS OF THE SNAKE RIVER PLAIN

The Snake River Plain is an elongate structural basin, largely filled by basaltic lavas and sediments. Two segments of the plain are distinct: a western part trends westward and northwestward from about Twin Falls past Boise to near the Oregon border, and an eastern part trends northeastward from west of Twin Falls toward the Yellowstone Plateau. The axis of the eastern Snake River Plain is continuous with the main volcanic axis of the Yellowstone Plateau volcanic field (fig. 44; pl. 3). The origin of the structural basin is complex. Kirkham (1931) proposed that the Snake River Plain is a structural downwarp; he based his conclusions on inward dips of Cenozoic volcanic and sedimentary units on both margins of the plain. Malde (1959), however, made a convincing case for the western Snake River Plain being a graben that developed progressively during late Cenozoic time, partly filling with lavas and sediments as it deepened. Carr and Trimble (1963)—and later, Trimble (1976) and Trimble and Carr (1976)—noting the straight lines of truncation of basin-range fault blocks on both margins and the presence of at least some mappable faults along the southern line of truncation, suggested that the eastern Snake River Plain too may be mainly a graben. Allmendinger (1982), however, noted that these and other similar faults entirely postdate volcanic rocks erupted from the adjacent parts of the plain and probably reflect local adjustments to broad subsidence rather than rift faulting. Hamilton and Myers (1966) adopted the graben concept for the western plain and the downwarp concept for the eastern plain. Inward dips of late Cenozoic volcanic and sedimentary deposits and inward tilts of at least some of the bounding ranges of the eastern Snake River Plain do appear to indicate regional downwarp along the axis of the plain, perhaps modified by relatively minor northeast-trending faulting at the zone of maximum flexure.

Truncation of Miocene and younger basin-range structures by the Snake River Plain (Carr and Trimble, 1963; Ruppel, 1964; Trimble, 1976; Trimble and Carr, 1976) shows that it is entirely a late Cenozoic feature. Distribution patterns of Cenozoic floras suggested to Axelrod (1968) an Oligocene ancestry for the plain, but there is no independent structural or stratigraphic evidence for such an early date of formation for the feature.

The origin of the downwarped eastern plain must have been extensional. This conclusion is substantiated by minor graben structures within the plain, the temporal and spatial association with extensional basin-range structures at its margins, the presence of numerous basaltic feeder dikes along rift zones to produce the flows that now cover much of the plain (Kuntz, 1977), and the en echelon arrangement of major gravity anomalies of the western plain (Hill, 1963). Hamilton and Myers (1966) suggested that the Snake River Plain is a tensional rift structure whose eastern segment has opened progressively northeastward as a result of northward or northwestward movement of the region that lies to the north. Other workers, including Christiansen and McKee (1978) and Pierce and Morgan (1992), noted that regional tectonic extension affected areas as far east as Yellowstone and Jackson Hole by 9-10 Ma.

Exposed rocks of the western Snake River Plain include thick sedimentary sequences and interlayered basaltic lavas (Malde and Powers, 1962; Malde and others, 1963; Malde, 1972; 1991). By comparison, only moderate thicknesses of sediments are exposed at the surface of the largely basalt-covered eastern Snake River Plain (Stearns and others, 1938) or are revealed in the subsurface by drilling (Walker, 1964; Nace and others, 1975; Doherty and others, 1979). A few small bodies of rhyolite are exposed within the plain, principally Big Southern Butte and East Twin Butte near the center of the eastern plain (Stearns and others, 1938) and the Juniper Buttes near the east end (Kuntz, 1979). Rhyolites and related silicic rocks, predominantly welded ash-flow tuffs

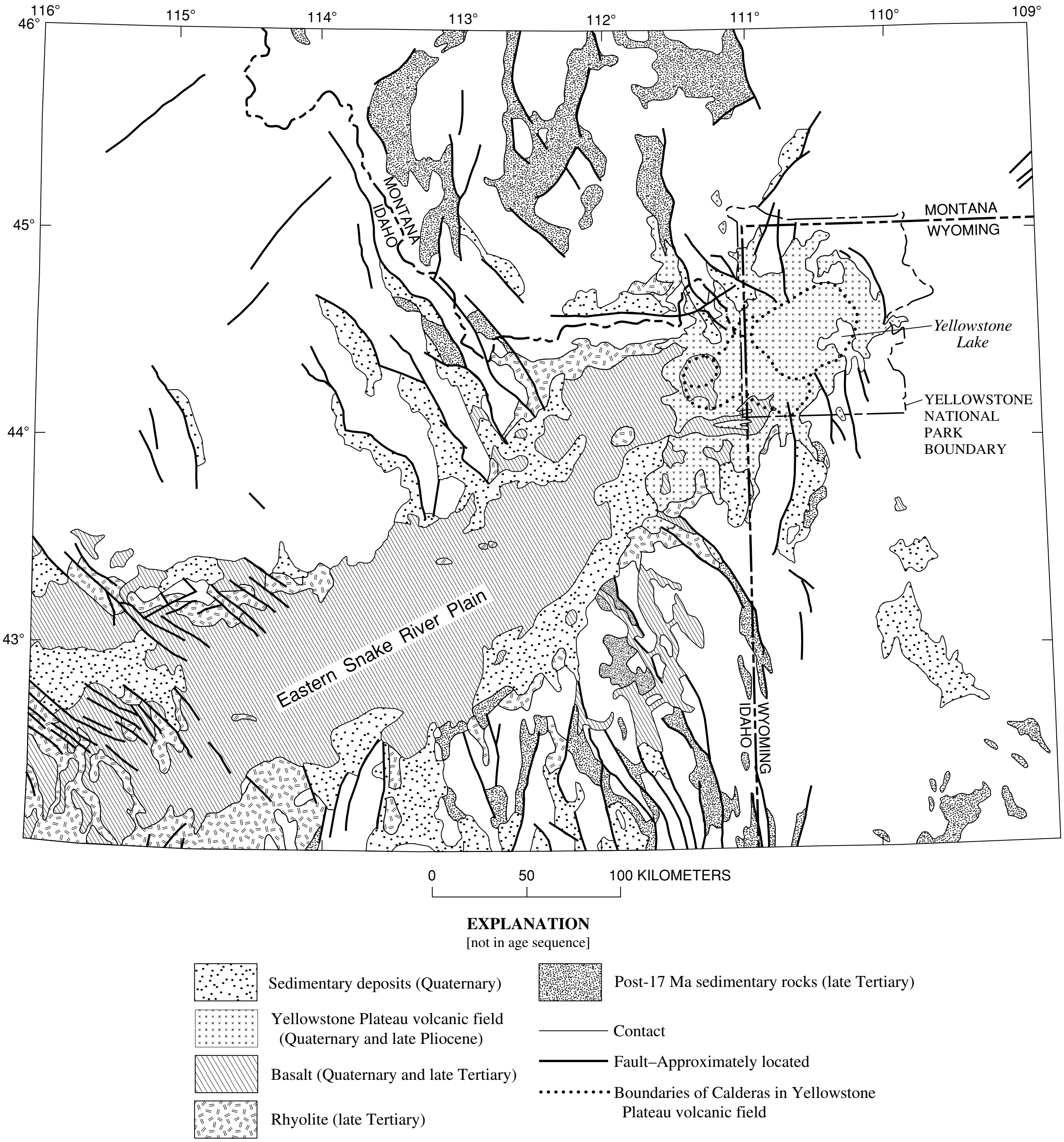


Figure 44.—Generalized map of upper Cenozoic volcanic and tectonic features of the eastern Snake River Plain, the Yellowstone Plateau volcanic field, and surrounding region. The axis of the eastern Snake River Plain is continuous with the Yellowstone Plateau volcanic field Sources of data: this paper, Mansfield (1927; 1952), Stearns and others (1938; 1939), Ross and Forrester (1947), Pardee (1950), Ross and others (1955), Ross (1961), Cohee and others (1962), Malde and Powers (1962), Malde and others (1963), Carr and Trimble (1963), Walker (1964), Malde (1965), King (1969), Mabey and Oriel (1970), Trimble (1976), Trimble and Carr (1976), Reynolds (1979), Stickney and Bartholomew (1987), Anders and others (1989), Pierce and Morgan (1992).

and bedded fallout and reworked tuffs, are abundant along the margins of the Snake River Plain, particularly the eastern segment (Mansfield and Ross, 1935; Mansfield, 1952; Mapel and Hail, 1959; Malde and Powers, 1962; Carr and Trimble, 1963; Prostka and Hackman, 1974; Trimble, 1976; Trimble and Carr, 1976; McBroome, 1981; Trimble, 1982; Embree and others, 1982; Bonnichsen and Citron, 1982; Williams and others, 1982; Allmendinger, 1982; Morgan and others, 1984; Morgan, 1988; Anders and others, 1989; Morgan, 1992). These rhyolitic rocks of the plains-marginal region are locally interlayered with or overlain by basalts, but, as in the Yellowstone region, virtually no igneous rocks of intermediate compositions are associated with them.

A LINEARLY PROPAGATING VOLCANIC SYSTEM

Published studies of the margins of the eastern Snake River Plain enable several important generalizations about the volcanic rocks. The most voluminous volcanic rocks of the plains-marginal suite are welded ash-flow tuffs. Their areal distributions, thickness variations, crystal and lithic contents, and cooling and crystallization zones show that their source areas are now buried by the basalts that cover the axis of the plain; the ash-flow sheets blanket the proximal ends of the marginal ranges but thin to glassy wedgeouts as traced along the valleys away from the plain. Furthermore, the center of distribution of each successively younger sheet or sequence of cogenetic sheets that has been identified along the margins of the plain generally is farther northeast than that of the next older sheet. This relation was first suggested to Christiansen and Blank (1969) by studies in the Yellowstone-Island Park region and by the incomplete stratigraphic and paleontologic data then available for the plains-marginal rocks farther southwest. Additional supporting data came from reconnaissance stratigraphic studies, and the conclusion was confirmed and amplified more recently by the mapping of several workers and by the potassium-argon dating of Armstrong and others (1975). Several authors have suggested that this northeastward age progression began about 15-16 million years ago, but Christiansen and Yeats (1992) noted that a clear linear progression did not begin until about 12.5 Ma. Ages of silicic ash-flow sheets along the west end of the south margin of the eastern Snake River Plain were reported as between about 9 and 12 Ma by Armstrong and others (1975) and Bonnichsen and Citron (1982). Armstrong's dates for most of the rhyolitic volcanic rocks of the south-central margin of the eastern Snake River Plain are between about 8 and 10 Ma. McBroome (1981), Morgan and others (1984), and Morgan (1988) determined fission-track ages between about 7 and 4 Ma for widespread ash-flow sheets in the northeastern plain, west of the Island Park area. The ages of major ash-flow sheets surrounding the Island Park-Yellowstone area are from 2.1 to 0.6 Ma, as discussed in this paper.

Morgan and coworkers showed in a series of papers (McBroome, 1981; Embree and others, 1982; Morgan and others, 1984; Morgan, 1988; Morgan and Bonnichsen, 1989; Morgan, 1992) that the group of three ash-flow sheets of 6.5 to 4.3 Ma, exposed both north and south of the eastern Snake River Plain between about Big Southern Butte and Island Park, erupted to form a complex of calderas. These calderas are now nearly buried by the basalts of the plain but are revealed in part by gravity and aeromagnetic anomalies (Mabey, 1978). Small marginal sectors of these calderas are exposed south of the plain (for example, Prostka and Embree, 1978). This sequence of apparently cogenetic ash-flow sheets and partly nested and overlapping calderas defines the framework of the Heise volcanic field, a complex rhyolitic volcanic field analogous to the Yellowstone Plateau field, including its caldera complex and the ash-flow sheets of the Yellowstone Group, each with a lifetime of about 2 m.y.

These two analogous rhyolitic systems suggest that the voluminous rhyolitic volcanism of the eastern Snake River Plain formed a series of such volcanic fields, each active for a few million years and each grouped in an area northeast of the previous one (Pierce and Morgan, 1992). Thus, rather than a smooth progression of individual rhyolitic centers propagating northeastward, a series of major long-lived magmatic systems succeeded one another northeastward along the axis of the plain.

Several parallels between the geology of the eastern Snake River Plain—especially its margins—and that of the Yellowstone Plateau volcanic field are thus striking. Both volcanic regions lie along a single continuous structural axis; both have evolved during a period of regional tectonic extension; both consist mainly of rhyolitic and basaltic volcanic rocks and practically lack rocks of intermediate composition; the principal rhyolitic rocks of each region are voluminous ash-flow sheets and related fallout ashes that were erupted from source areas in the vicinity of the principal structural axis; basaltic eruptions on the flanks of the principal rhyolitic source areas alternated with the major ash-flow eruptions; and the source area of each major group of ash-flow sheets now exposed along the margins of the two regions tends to lie to the northeast of the next older source area.

Another similarity between the two regions is less obvious but perhaps of equal importance. The basalts of the Snake River Plain, like those of the Yellowstone Plateau volcanic field, are olivine tholeiites (Stone, 1966; 1967; Tilley and Thompson, 1970; Leeman, 1982b); but also, as at Yellowstone, there seem to be two end-member compositional types among the basalts. Most available chemical analyses and most published descriptions of basalts of the Snake River Plain are from the lavas that flood the axis of the plain. Although these lavas are somewhat less silicic than those of Yellowstone, in other respects they are similar to the relatively potassium-rich basaltic type of the Yellowstone region. However, a few of the analyses cited by Stone (1967) and by Trimble and Carr (1976), made on samples taken from the margins of the plain, where the basalt is interlayered with rhyolitic units, span a slightly greater range of silica contents and are much like the predominant plateau-marginal basalts of the Yellowstone Plateau volcanic field in having low potassium contents and low Fe/Mg ratios relative to those along the axis of the plain. These two types of basalt from the Snake River Plain region—the axial flood basalts and plains-marginal basalts interlayered with rhyolites—are shown in figure 40. The two Snake River Plain types, which lie on extended variation curves of the two types from the Yellowstone Plateau field, are effectively separated by the reference line for Hawaiian tholeiites and alkali basalts in figure 40 (Macdonald and Katsura, 1964). As in the two Yellowstone basaltic suites, the predominant, more potassium-rich basalts of the Snake River Plain are not, however, alkali basalts; the variation diagram merely shows the presence of two similar compositional suites in each of the two contiguous regions.

The geologic parallels between the Yellowstone Plateau region and the eastern Snake River Plain together with the systematic northeastward decrease in age of major rhyolitic magmatic systems along their continuous axis, suggest the following sequence, in which the Yellowstone plateau, Island Park, and the eastern Snake River Plain represent successive stages in the evolution of a unitary magmotectonic province. (1) A major focus of volcanism has migrated more than 450 km northeastward since about 12.5 Ma along the axis of the eastern Snake River Plain, from west of Twin Falls to the Yellowstone Plateau, an average rate of migration of about 3.5-4 cm/yr (although some effect of regional extension is included in this figure; Rodgers and others, 1990; Smith and Braile, 1994). The earliest volcanic activity in each new focal area may have been the eruption of relatively low-potassium tholeiitic basalt. (2) Subsequent volcanism in each area involved voluminous rhyolitic ash-flow eruptions, caldera formation, and probably the eruption of postcollapse rhyolitic lavas. During major rhyolitic volcanism, eruption of tholeiitic basalts continued around the flanks of each volcanic system but probably not in the source areas of the voluminous rhyolitic ash flows. The third volcanic cycle of the Yellowstone Plateau field is in this stage. (3) After the major magma body that sustained the rhyolitic cycle at each volcanic focus solidified to a granitic pluton, basaltic lavas were erupted through extensional tectonic fractures in the granites. Basaltic volcanism, thus, has now occurred within the calderas of the first and second cycles of the Yellowstone Plateau field at Island Park. (4) Basaltic volcanism continues along the Snake River-Yellowstone axis as the trail of extinct major rhyolitic volcanism undergoes progressive subsidence, perhaps in part by thermal contraction (Brott and others, 1981). The younger basalts along the subsided and flooded axis of the Snake River Plain are relatively richer in K, Ti, P, and Fe/Mg and poorer in Ca than the corresponding earlier plains-marginal basalts. Somewhat similar basalts have erupted in the western part of the Yellowstone Plateau volcanic field. Thus, there may eventually be subsidence and flooding of the Yellowstone Plateau by relatively high-potassium tholeiites.

The point was noted earlier that, rather than a smooth progression of individual rhyolitic centers, a series of major long-lived magmatic systems evolved successively northeastward along the axis of the Snake River Plain. Each of these major systems, having lithospheric dimensions, would have greatly affected the regional properties of the crust and upper mantle above any continuous asthenospheric focus that might have sustained continued northeastward propagation relative to the lithosphere. Such major lithospheric systems would have tended to thermally buffer the large-scale processes responsible for establishing and sustaining them. Over a few million years, however, continued northeastward propagation of the sublithospheric focus, whose thermal regime would have been more continuous and less easily perturbed than the lithospheric system, would finally break the continuity of the shallower system and its thermal root zone, resulting in the establishment of a new system farther to the northeast. The Yellowstone Plateau volcanic field is the present expression of such a magmatic system.

As each volcanic field became active successively farther northeastward, a period of greatly enhanced extension rates affected areas of the basin-range region adjacent to the plain on the south (Anders and others, 1989; Pierce and Morgan, 1992). The propagating volcanic system, however, remained confined to a narrow zone in the northern part of the already existing region of large tectonic extension (Christiansen and McKee, 1978). Net extension is considerably less north of the eastern Snake River Plain than to the south. Only after about 6.5 Ma did the present major ranges rise along the north margin of the plain (Scott and others, 1985).

ORIGIN AND EVOLUTION OF THE YELLOWSTONE MAGMAS

The attempt is made here to generalize the pertinent observations and to draw some conclusions about the origin of rhyolitic and basaltic magmas of the Yellowstone Plateau volcanic field and about other deep-seated processes related to them. Beginning with considerations of the volcanic history of the region and other voluminous rhyolitic fields, the discussion proceeds to conclusions about the emplacement and cooling of large bodies of rhyolitic magma at shallow to intermediate crustal levels. Later these considerations are extended to problems of magmogenesis at the lowest crustal levels and in the upper mantle. The conclusions, of course, are progressively more speculative as they concern these progressively more deep-seated processes.

EMPLACEMENT AND PHYSICAL EVOLUTION OF THE RHYOLITIC MAGMAS

Rhyolitic magmas are directly responsible for the predominant volcanic features of the Yellowstone Plateau region. Both their volcanic and their intrusive relations are considered here.

VOLUMINOUS RHYOLITIC VOLCANISM AND THE CAULDRON CYCLE

It was once a commonplace view, expressed explicitly by Daly (1933, p. 34-41), that rhyolites represent only a minor fraction of the magmas erupted to the earth's surface. The recognition during the past half century of voluminous rhyolitic tuffs, especially ash flows, throughout the world has gradually led to the now commonly accepted view that rhyolitic pyroclastic materials represent much greater eruptive volumes than do rhyolitic lavas. Bedded tuffs and tuffaceous sediments represent vast quantities of erupted rhyolitic magma (Ross, 1955). Many ash-flow fields, representing the products of geologically brief eruptive episodes, cover thousands of square kilometers and have volumes of tens to thousands of cubic kilometers (Smith, 1960a, p. 819; 1979, fig. 1). Thus, there no longer appears to be any justification for Daly's assertion.

Furthermore, Smith (1960a, p. 819-820; 1979, fig. 2) showed that all of the most voluminous ash-flow sheets are related to major subsidence features in their source areas. Several studies of large ash-flow fields since Smith's review have strengthened that conclusion (for example, Aramaki and Ui, 1966; Steven and Lipman, 1976; Rhodes, 1976; Bailey, 1976; Christiansen and others, 1977; Christiansen, 1979; Lipman, 1984). Smith and Bailey (1968a) showed that many, though by no means all, of the calderas related to voluminous epicontinental rhyolitic ash-flow eruptions are resurgent—characterized by early postcollapse domical uplifts of the caldera floors.

The Yellowstone Plateau volcanic field clearly fits this pattern. The three major ash-flow sheets of the field, which aggregate more than 3,700 km^3, came from known source areas within the volcanic plateau. All three of the major source areas collapsed during or immediately after the eruptions to form gigantic calderas. At least the Yellowstone caldera, the youngest of the three and a complex structure having two cauldron blocks, is resurgent. Although there is no evidence regarding postcollapse resurgence in the oldest caldera, and the second was not resurgent, the three ash-flow sheets and their related calderas have been shown to relate to three discrete volcanic cycles in which the general pattern of events was quite similar. This pattern of events, except for resurgent doming, is virtually the same as that defined by Smith and Bailey (1968a) for the resurgent cauldron cycle and documented by them for several fields of voluminous rhyolitic volcanism. Some of these fields are predominantly andesitic and calc-alkalic, in which the rhyolitic magmas seem to have been intimately related to magmas of intermediate composition. Others are compositionally bimodal fields like that of the Yellowstone region.

A general pattern of volcanic evolution, thus, seems to be a recurring plan in voluminous rhyolitic fields (Christiansen, 1979). Whether or not a resurgent dome uplifts the cauldron block shortly after collapse, the following generalized stages are represented in most such fields: (1) Broad gentle uplift of an area somewhat larger than that which ultimately will become the major rhyolitic source area is accompanied by circumferential and radial fracturing of the uplifted area. Commonly, this broad uplift also is accompanied by extrusions of intermediate to silicic lavas controlled by these fractures. The general location of the area of tumescence as well as specific volanic structures that form during this and later stages may be controlled to a greater or lesser extent by tectonic structures; in particular, regional faults commonly function as radial fault systems for this tumescence. (2) At some critical stage, voluminous rhyolitic ash and pumice are erupted rapidly through the ring-fracture zone that was generated during stage 1 and are emplaced mainly as ash flows; collapse of the magma-chamber roof in the volcanic source area occurs along these same ring fractures to produce a large caldera during or immediately after the ash-flow eruptions. (3) Postcollapse rhyolitic volcanism may occur largely within or adjacent to the source area of the ash flows; it is controlled by the ring-fracture zone and other caldera structures or by their interaction with concurrently active tectonic faults. Resurgent doming, if it occurs, is generally a phenomenon of the early part of this third stage. Although hydrothermal activity at the surface accompanies all stages of this generalized cycle, at the end it becomes the predominant or the only surface manifestation of igneous activity.

This cycle may be repeated one or more times, and the cycle may revert at any point in its development to an earlier stage (Christiansen, 1979). Once started, however, the cycle seems to describe a progressive course of evolution of volcanic fields in which very large volumes of rhyolite are erupted. The generality of this concept, following ideas of Smith (1960a) and Smith and Bailey (1968a), must indicate that it relates to a major crustal process that can occur equally in calc-alkalic igneous fields that form by the evolution of andesitic to rhyodacitic magmas, in silicic alkalic fields where alkali rhyolites or their relatives were derived from differentiation along a trachyte stem, or in bimodal fields in which rhyolitic and basaltic magmas seem to be intimately related but not by crystal-liquid differentiation. The common process is not dependent on the manner in which those magmas formed at greater depths.

The voluminous rhyolitic fields seem to be almost strictly continental features. The only somewhat similar fields in the present oceanic regions of the Earth are in the basalt-rhyolite province of Iceland. Evidence has been cited to show that Iceland has a thicker crust than typical oceanic regions, probably basaltic but about 15-20 km thick rather than 5-10 km (Bath, 1960; Hermance and Grillot, 1970; Palmason, 1971).

THE RHYOLITIC MAGMAS AS BATHOLITHS

Several lines of argument lead to the conclusion that the volume of rhyolitic lavas and pyroclastic debris erupted in voluminous rhyolitic or rhyodacitic fields is only a small part of the total volume of magma emplaced high in the crust. The course of magmatic differentiation in some fields, requiring magmatic volumes much greater than those of the ash-flow sheets, points directly to this conclusion (Lipman and others, 1966; Smith, 1979). The existence of large granitic intrusive bodies with structural features indicative of their previous connection to large calderas—for example, Buddington (1959, p. 680-685), Duffield (1968), Eggler (1968), Bussell and others (1976), and Pitcher (1978)—lends further credence to the concept. Smith and Shaw (1975) pointed out the generality of these volume relations; they as well as others (for example, Smith, 1979; Christiansen, 1979) discussed their geologic significance.

The magnitude of the hydrothermal system of the Yellowstone Plateau further confirms that the volcanic field is underlain by a body of rhyolitic magma much larger than the volume of rhyolites erupted during the third volcanic cycle of the region. It was noted earlier that the principal hydrothermal discharge of the plateau occurs within the Yellowstone caldera and along one major fault zone radial to it. The locations of the hydrothermal features clearly were controlled by caldera structures and their intersections with radially oriented tectonic fault zones. White (1957; 1968), Fournier and others (1976), and Fournier (1989) showed that such hydrothermal features constitute a deeply circulating convective heat-flow system, the magnitude and longevity of which require a large source of heat that can be accounted for only by a body of crystallizing magma at fairly shallow depth under the plateau.

Eaton and others (1975), Smith and Christiansen (1980), and Smith and Braile (1984; 1994) interpreted the combined evidence of geology, heat flow, gravimetry, aeromagnetics, seismicity, and seismic attenuation as showing that a rhyolitic magma body, possibly still partly molten and larger in area than the Yellowstone caldera, lies at shallow depth (about 5-6 km); this conclusion was further substantiated by a series of detailed active-seismic experiments (Lehman and others, 1982; Smith and Braile, 1984).

Several lines of geophysical evidence, although individually inconclusive regarding the actual presence of magma in the Yellowstone system, are consistent with the picture of a large plutonic body that probably remains at least partly molten. It was noted earlier that the present rate of heat flow represented by this hydrothermal system cannot have operated for the last 600,000 years unless magma continues to supply heat at the top of a large magmatic body, probably by successive reintrusion from lower levels as well as by magmatic convection. Patterns of seismicity vary between the caldera and the surrounding zone, especially in the general occurrence of earthquakes to depths no greater than about 6 km beneath the caldera as compared to 10 km or greater outside (Smith and others, 1977; Smith and Braile, 1984; 1994). Precise leveling of a circuit around and across the Yellowstone caldera showed that its central axis, between the resurgent domes, is undergoing alternate uplift and subsidence of large magnitudes and high rates. Maximum uplift of more than 700 mm was measured between successive relevelings about 50 years apart (Pelton and Smith, 1979; 1982). Similar uplift continued to 1984, stopped in 1984-1985, and reversed to subsidence after 1985 (Dzurisin and Yamashita, 1987; Dzurisin and others, 1990; 1994). Uplift resumed in about 1996 (Wicks and others, 1998). Deformation of shoreline terraces around Yellowstone Lake demonstrate that similar alternations of caldera uplift and subsidence have been active for at least the past 12,000 years (Locke and Meyer, 1994; Pierce and others, 1997). Various studies have concluded that movements of magma or hydrothermal fluids as well as and regional tectonic extension may be involved in the ongoing deformation (Vasco and others, 1990; Dzurisin and others, 1990;1994; Savage and others, 1993; Smith and Braile, 1994; Locke and Meyer, 1994; Wicks and others, 1998). A teleseismic study by Iyer and coworkers (Iyer, 1975; 1978; Iyer and others, 1981) showed that a zone of anomalous delays in teleseismic P waves underlies the caldera; azimuthal distribution of the anomalies with respect to direction of origin of the earthquakes demonstrates that a zone of reduced P-wave velocities extends well below the base of the crust. Modeling suggests a 15 percent reduction in velocities within the crust and a 5 percent reduction to depths of 250 km (Iyer and others, 1981). Smith and others (1977) analyzed aeromagnetic data to show that depth to the Curie temperature peripheral to the Yellowstone caldera is about 10 km. Bhattacharyya and Leu (1975) showed that the aeromagnetic low associated with the Yellowstone caldera can be modeled to suggest a Curie depth as shallow as 6 km within parts of the caldera. Smith and Braile (1994) compared the magnetic and gravity fields of the caldera to come to a similar conclusion that a hot body lies at shallow depths, especially beneath the caldera ring-fracture zone. Magnetotelluric soundings suggest that the Yellowstone caldera is underlain at depths as shallow as 5 km by electrically conductive material, having resistivities less than 10 ohm·m (Stanley and others, 1977).

The absence of basalts from the area of the Yellowstone caldera, except for minor bodies on the north wall whose vents lie well outside the ring-fracture zone, contrasts with numerous basaltic eruptions before and after the Lava Creek Tuff eruptions from all sides surrounding the caldera; this evidence suggests that basaltic magma has remained active in the system as a whole throughout its evolution but that its ascent beneath the caldera remains blocked by the less dense and nonrigid rhyolitic magma body. This distribution of basalts further indicates that the body of unconsolidated rhyolitic magma is of batholithic dimensions. The approximate cooling time for such a magma chamber is suggested by the sequence of events in the Island Park area, where basalts probably were not erupted within the 30-km-diameter major rhyolitic source area for slightly more than a million years after the end of rhyolitic volcanism of the second cycle. Thus, about a million years was required to crystallize and cool a granitic batholith sufficiently so that it could be fractured to enable the vertical transmission of basaltic magmas. Until such a large body of rhyolitic magma consolidates by crystallization and cooling, basaltic magmas rising beneath it are deflected laterally through faults or mix with the lower parts of the rhyolitic magma body. Locally on the margins of the system some mixed or partly mixed-lava complexes, such as that of Gardner River, may extrude to the surface although their rhyolitic component is not part of the batholith-sized body beneath the caldera (Hildreth and others, 1984; 1991). The basaltic magmas, rising from the mantle, must continue to supply heat to the shallower rhyolitic magma system in order to account for the longevity of the system and for the continued high hydrothermal output of energy (Christiansen, 1984; Dzurisin and others, 1990).

Beneath the Yellowstone Plateau volcanic field, a granitic polybatholith has been forming at moderate to high crustal levels during about the past 2 million years. The first major magmatic insurgence, related to the Huckleberry Ridge Tuff, probably produced three high-level culminations, each probably having a diameter greater than 15 km. The petrographic, chemical, and isotopic similarity and the immediate succession of eruption of the three members of the Huckleberry Ridge, in less time than any erosion or significant chilling could occur, indicates that the three high-level culminations were closely interconnected; the three culminations were at the top of a single batholithic magma chamber. Within less than a million years, the bulk of this magma had crystallized. Geologic relations indicate, however, that at least a part of the magma remained largely liquid and that this residual magmatic body, comparable in size to one of the three high-level culminations of the first cycle, produced a second major volcanic cycle (Christiansen, 1979). Thus, in a sense, the second cycle represents the ultimate degassing of rhyolitic magma produced in the first major insurgence of the Yellowstone Plateau magmatic system. The third cycle represents a separate major insurgence that began at about the same time as that culminating episode of the first insurgence, but the third cycle did not overlap the residual second-cycle chamber. This major new magmatic insurgence formed a large magma body with two high-level culminations having diameters of more than 30 km each. As during the first volcanic cycle, the simultaneous generation of ring fractures and extrusion of precaldera rhyolite and the rapid-fire successive eruption of the two members of the Lava Creek Tuff, as well as systematic petrographic, chemical, and isotopic relations, indicate the interconnection and interdependence of the two high-level culminations. They probably rose at the top of a magma chamber nearly 100 km long. For both the first cycle and the third, systematic perturbations of isotopic ratios in rhyolites erupted from the entire active area were simultaneous, demonstrating the magmatic unity of the batholithic magma chambers (Hildreth and others, 1984; 1991). This included the isotopes of oxygen, the most abundant element in the magma.

Thus, the Yellowstone Plateau volcanic field almost certainly overlies a crystallizing batholith as Daly (1911) long ago suggested. Furthermore, I conclude that the field overlies a large granitic polybatholith that represents at least two major magmatic insurgences from the level of rhyolitic magma generation. Six high-level intrusive culminations, on the order of 15 to 40 km across, were produced by the rise of these magmas. The tendency, as described earlier, for younger postcollapse rhyolites to become more localized along extensions of tectonic fault trends further suggests that as crystallization of the higher-level parts of this intrusive complex proceeds, the lower-lying liquid parts continue to rise episodically and to passively intrude the top. Thus, Yellowstone's huge hydrothermal circulation system transfers heat from the entire advecting batholithic magma body, which continues to rise, to intrude its own forerunners, and to cool and crystallize at high crustal levels.

As suggested above, heat supplied by basaltic magmas maintains the rhyolitic magma for substantial periods of time. This model of repeated reintrusion of magma within a polybatholithic complex, with energy continually supplied from the mantle, conforms with both the geologic history of the volcanic field and the heat requirements of the massive convective hydrothermal circulation system. Such a long-lived deep-rooted magmatic system, sustained by continued thermal input from the mantle through numerous basaltic-magma injections, is necessary in order to form a crustal magma body as voluminous and highly evolved as that which lies beneath the Yellowstone Plateau volcanic field. Only such a voluminous and long-lived system has the capability of producing pyroclastic eruptions on the scale of those at Yellowstone (Hildreth, 1981; Christiansen, 1984).

The volcanic history of the eastern Snake River Plain, as outlined earlier in this chapter, suggests by analogy that the axis of the plain overlies a complex elongate trail of granitic batholiths like that of the Yellowstone Plateau volcanic field. Subsequent basaltic dike complexes must have cut this batholithic complex as it cooled and contracted to produce the Snake River downwarp (Brott and others, 1978; 1981). A horizontal section of the eastern Snake River Plain and Yellowstone Plateau region at a depth of perhaps 5-8 km would be expected to reveal a northeast-trending belt of granitic ring complexes and central plutons intruded by discordant batholiths. Much of that plutonic complex would be cut complexly by normal faults and basaltic dikes, whose directions are controlled by the regional tectonic stress field (Zoback and Zoback, 1989; Smith and Arabasz, 1991).

Other large rhyolitic ash-flow fields have been interpreted as overlying granitic batholiths or polybatholithic complexes. The Rotorua-Taupo region of New Zealand, which has long been known to have produced voluminous ash-flow sheets (or ignimbrites) (Cole, 1981; 1990; Wilson and others, 1984), has been regarded as overlying a batholithic complex (Williams, 1941, p. 317). The San Juan Mountains of southwestern Colorado encompass a voluminous andesite-rhyolite field, the evolution of which was climaxed by large rhyolitic ash flows and great caldera complexes that must overlie a batholithic complex (Larsen and Cross, 1956; Steven and Ratté, 1965; Lipman and others, 1970; Steven and

Lipman, 1976). Byers and others (1976) and Christiansen and others (1977) reached a similar conclusion for the Timber Mountain-Oasis Valley caldera complex in the rhyolite-basalt volcanic field of southwestern Nevada. Hamilton and Myers (1967) and Lipman (1984) postulated similar relations between voluminous rhyolitic volcanic fields and batholiths in general.

The generalized three-stage volcanic cycle of voluminous rhyolitic fields helps illuminate the batholithic underpinnings of the Yellowstone Plateau volcanic field. That threefold volcanic evolution represents the following magmatic events:

1. Initial formation or additions to a large upper-crustal magma body results in one or more high-level culminations. Differentiation may occur to varying degrees within these bodies as they are emplaced at high crustal levels. Parts of the magma may rise in advance of the rest to form semi-independent smaller magma chambers that can erupt as relatively small rhyolitic volcanoes controlled by a growing fracture system in the roof of the larger plutonic body. The main magma bodies rise to levels where their evolution eventually is interrupted by some critical combination of load pressure, downward propagating fractures from the stretching surface above, volatile saturation and separation of a gas phase, and related factors.
2. Rapid degassing of the magma body occurs as the critical combination of factors is reached. Exsolution of a gas phase and rapid vesiculation of the main magma body drives a magmatic froth upward through the fracture zone that breaks the stretched chamber roof, expanding and erupting through this fissure system as pyroclastic debris. The drastic pressure release caused by eruptive unloading of the main magma body starts a chain reaction of unloading and further vesiculation leading to the voluminous outpouring of rhyolitic pumice and the formation of great ash flows, blanketing the surface surrounding the source area. This major degassing continues until a state of virtual equilibrium is reached between the combined partial vapor pressures of the magma, its dynamic properties such as viscosity, and the tectonic and load pressures imposed on the magma body. Much of the magma remains unerupted after cessation of the major ash-flow eruptions, probably approximating a factor of ten more than was erupted (Smith and Shaw, 1975; Smith, 1979). Withdrawal of support of the magma-chamber roof by the eruption of a considerable part of the initial magma volume, however, results in roof collapse along the system of fractures formed during initial magmatic insurgence and utilized by the ash-flow eruptions. If this gravitational collapse occurs during the ash-flow eruptions, it can provide some of the energy required to displace magma from the chamber to the surface, but this factor alone is insufficient to account for the entire energy of eruption. Because the Yellowstone Plateau field formed in a region of tectonic extension, no considerable tectonic pressure in excess of lithostatic would have been available to supply the energy. Clearly, expansion of the magmatic gas phase was the predominant factor in lifting magma to the surface as well as impelling it into the great eruptive columns that sustained each series of ash-flow eruptions (Blake, 1984). Thus, these major pyroclastic eruptions can be thought of essentially as degassing events.
3. Eventual cooling and crystallization of the magma body at a high crustal level commonly is accompanied by occasional further rhyolitic eruptions of lesser magnitude than the critical degassing episode; these lesser eruptions are controlled by the ring-fracture zone and by other local or regional factors, especially structural intersections and the regional tectonic stress regime. A positive magma pressure generally is reestablished quickly after major degassing, and the magma body continues to rise buoyantly. Depending on local conditions, early postcollapse rise of the magma body may cause a central intrusion and resurgent doming of the collapsed magma-chamber roof. Bailey (1976) suggested that resurgence may involve an isostatic response of the crustal and magmatic system to drastic eruptive unloading. Continued later-stage rise may ultimately form a large discordant pluton that cuts the ring complex and central intrusion related to earlier stages. Volcanism gradually decreases with cooling and crystallization of the major granitic body and eventually ends largely or solely with hydrothermal activity. Ultimate crystallization of the entire magmatic system forms a large high-level granitic complex beneath the area of extinct volcanism. Complete or partial repetitions of the volcanic stages in some fields must express either the successive rise of several magma bodies or the interruption of normal progressive development by renewed rise of magma at some point in the cycle. The world's voluminous rhyolitic fields, thus, reflect the generation, rise, degassing, and eventual crystallization of very large silicic magma bodies; these fields must manifest a major mechanism of energy transfer within parts of the continental lithosphere.

This concept, that great rhyolitic ash-flow fields represents critical degassing to restore a disturbed equilibrium in large rising magma bodies, seems to me to contradict the concept of Hamilton and Myers (1967) that batholiths are essentially surficial features, or great "mantled lava flows" emplaced above downbowed prevolcanic surfaces. That massive silicic batholiths are much thinner than their horizontal dimensions and are only tenuously (though complexly) connected to their magmatic root zones, as implied by Hamilton and Myers, is certainly true; nevertheless, the volcanic history of the Yellowstone Plateau field and similar rhyolitic fields, including the foundering of massive intact crustal blocks to form huge calderas, argues against the idea that the Yellowstone batholith is floored by a former ground surface and roofed only by its own volcanic cover as argued by Hamilton and Myers (1967).

Numerous subvolcanic and plutonic analogues might be suggested to illustrate the concept of the upper-crustal batholithic foundation of the eastern Snake River Plain-Yellowstone region. Several more or less linear belts have been described in which igneous complexes, ranging in affinity from calc-alkalic to strongly alkalic and in tectonic setting from orogenic belts to extensional rifts, include both ring structures related to caldera formation and discordant batholithic bodies related to late-stage continued rise of the silicic magmas. Most of the linear continental igneous belts of generally bimodal or mafic-alkalic affinities occur in postorogenic or nonorogenic regions and, at least in some instances, can be related directly to extensional zones. The middle Jurassic White Mountain Plutonic-Volcanic Suite of New Hampshire (Billings, 1928; 1945; 1956; Creasy, 1974; Foland and Faul, 1977) fits this model except, perhaps, in lacking a clear extensional tectonic pattern; the White Mountain batholith (fig. 45) may afford a plutonic analogue for the Yellowstone Plateau system.

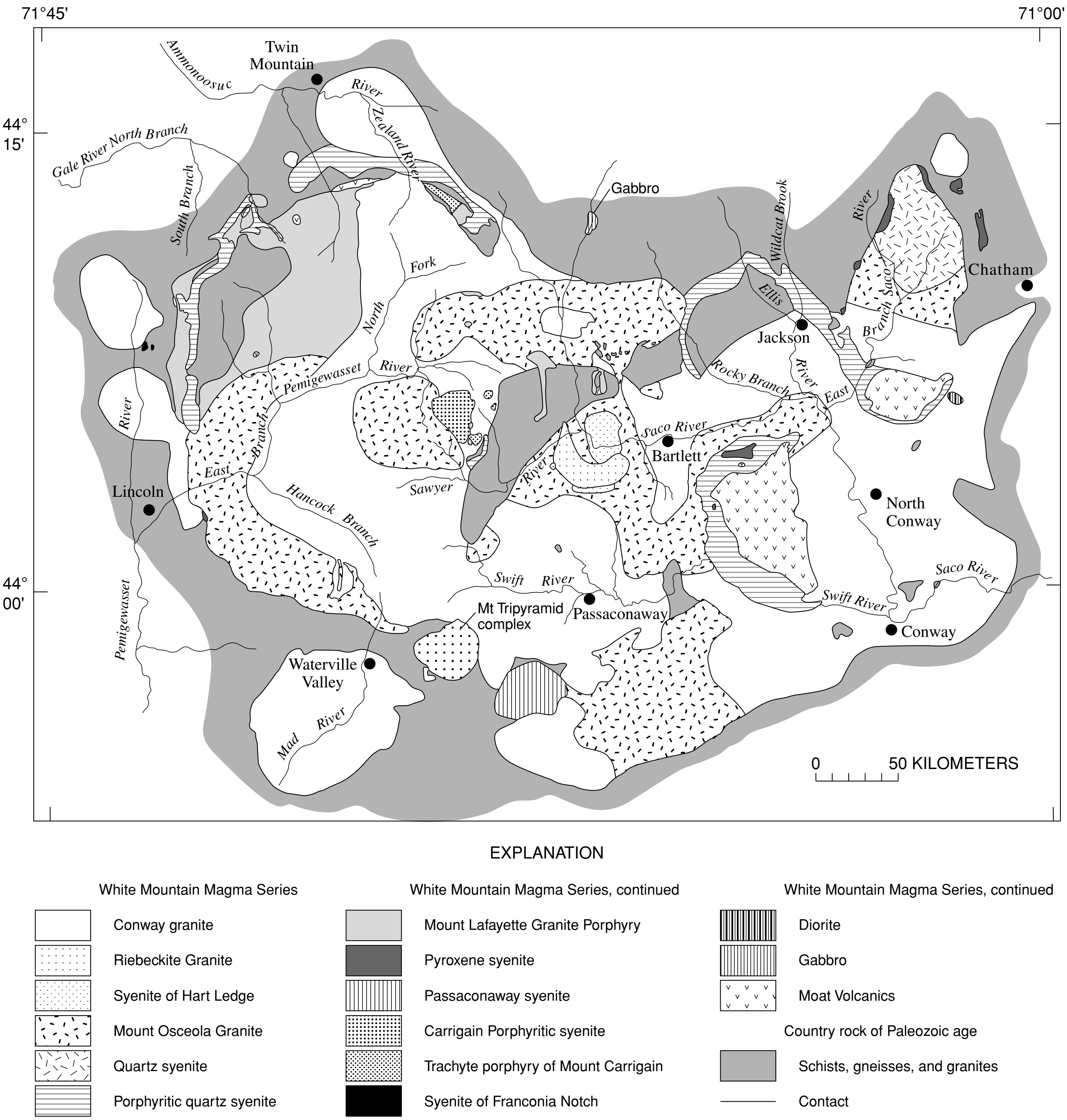


Figure 45.—Generalized geologic map of the Jurassic White Mountains batholith, New Hampshire (modified from Creasey, 1974). The White Mountains batholith is a possible analogue for the plutonic underpinnings of the Yellowstone Plateau volcanic field.

The Permian Oslo graben of Norway (Oftedahl, 1952; 1953; 1978a; 1978b) has more alkalic and more syenitic igneous rocks than those of the Yellowstone-Snake River Plain system, but it may be another analogue (fig. 46) although it lies within a well defined rift system.

Another possible analogue for the Yellowstone-Snake River Plain province may be the upper Paleozoic complexes of northern Queensland, Australia (fig. 47), especially the rhyolitic to bimodal Permian units and their associated structures. Branch (1966) considered all of the Carboniferous to

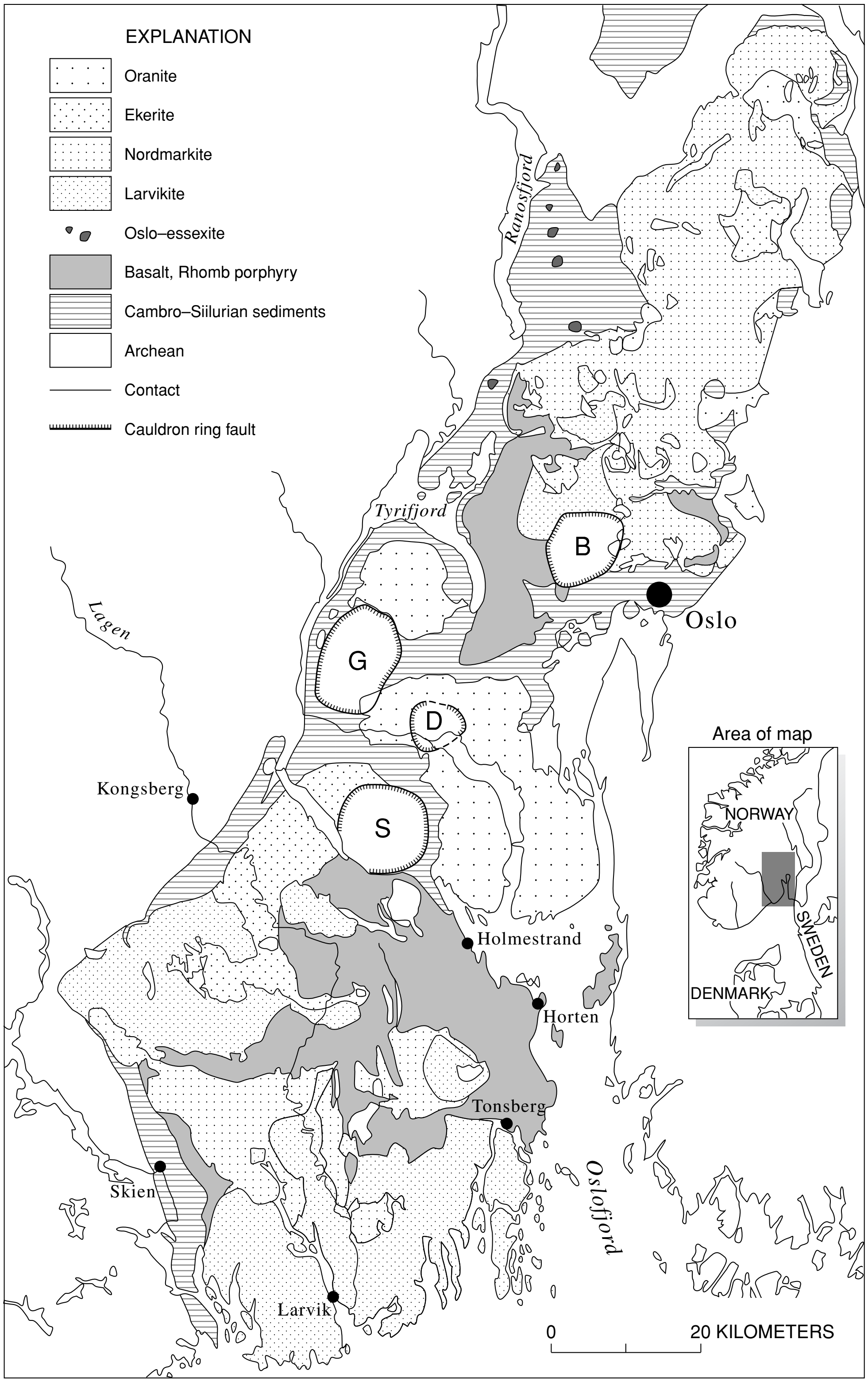


Figure 46.—Generalized geologic map of the Permian Oslo graben, Norway (modifield from Oftedahl, 1953). The Oslo graben, although more alkalic and more syenitic than the Yellowstone-Snake River Plain magmatic system, may be closely similar in magmotectonic setting at a deeper level of exposure. B, D, G, and S are cauldron-subsidence structures.

Early Permian volcanic sections preserved in north-central Queensland to lie within volcanic subsidence structures; many of these features clearly are volcanic cauldrons or subvolcanic ring structures, but some are quite large graben-like structures. Oversby and others (1980) later pointed out that these large linear blocks are not all necessarily related to specific eruptive events but are more complex synvolcanic subsided blocks. They regarded the blocks as first-order volcano-tectonic subsidence features that include the individual cauldrons, which they considered to be second-order volcanic subsidence structures. Mackenzie (1987; 1993; Mackenzie and others, 1993) further emphasized the extensional tectonic setting of the volcanism and the range of styles of associated subsidence—graben-like, sag-like, and ring-faulted. The volcanic history and structural pattern of the Yellowstone-Snake River Plain region suggests the further possibility that some of the large linear blocks of northern Queensland, largely filled with volcanic deposits, may be extensional tectonic features analogous to the downdropped fault blocks of the region around Yellowstone and the eastern Snake River Plain, in which are preserved sections of welded tuffs erupted from the associated ring and cauldron complexes and emplaced as ash flows into adjacent tectonic fault basins. Some of the larger subsided blocks might have been synchronous with volcanism but related to regional tectonic stresses rather than to subsidence resulting directly from the transfer of magma from the crust to the surface. The associated cauldron complexes would represent the subvolcanic parts of calderas related to individual volcanic cycles climaxed by major ash-flow eruptions. Furthermore, some of the associated granites might be cogenetic late-stage discordant plutons such as I have suggested to be associated with the polybatholithic complex beneath Yellowstone.

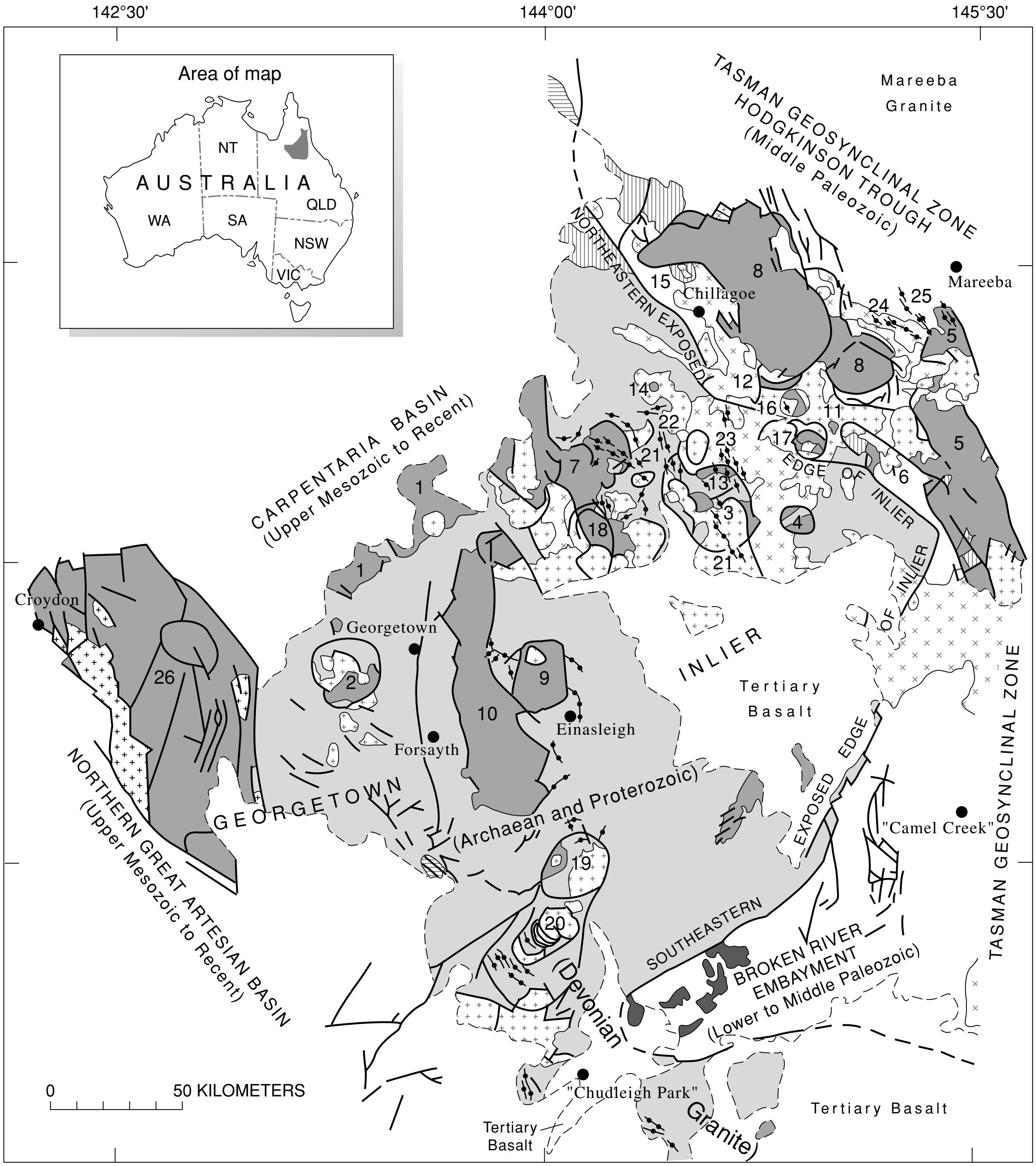
MAGMOGENESIS OF THE RHYOLITE-BASALT ASSOCIATION

This paper has attempted to show that basalts and rhyolites of the Yellowstone Plateau volcanic field and of the Snake River Plain and its margins form an essential (nonaccidental) genetic association. Many lines of evidence, summarized earlier, make it clear, however, that the rhyolites and basalts are not related to one another by fractional crystallization. Questions long debated regarding bimodal volcanic fields such as Yellowstone include: How and where are the basaltic and rhyolitic magmas of such a voluminous bimodal field formed? And, what is the genetic link between the basaltic and rhyolitic magmas? Despite the volumetric predominance of rhyolites in the Yellowstone Plateau volcanic field, the history and geophysical characteristics of the Snake River Plain region indicate that basalts ultimately may be comparable in volume. It is now rather commonly accepted that the development of voluminous rhyolite-basalt igneous fields such as the Yellowstone-Snake River Plain region is a largely continental phenomenon related fundamentally to the rise of basaltic magmas from the upper mantle.

Two alternative mechanisms for generation of rhyolite-basalt bimodal systems unrelated to fractional crystallization have been proposed. Some investigators suggest that the basaltic and rhyolitic magmas are from separate or unmixed liquid fractions derived from a common parental material, others that the two major magma types are each derived from primary melts of different source regions. The evidence from Yellowstone seems clearly to favor the latter hypothesis, with separate source regions in the crust and upper mantle, but first some aspects of the case for fractional melting and the formation of separate liquids from a single source region are considered.

Hypotheses of the formation of parental magmas by separation of immiscible liquids were once widely held but dropped from favor with the lack of any convincing evidence for the importance of immiscibility in natural silicate liquids (Bowen, 1928, p. 7-19). More recently, liquid-immiscibility hypotheses have reappeared in petrologic literature, and the existence of quenched immiscible liquids in lunar basalts (Roedder and Weiblen, 1970) has lent support to the idea.

Hamilton and Wilshire (1965) and Hamilton (1965) suggested that the bimodal rhyolite-basalt assemblage of the Yellowstone-Island Park region and other similar occurrences result from the separation of immiscible "olivine-basaltic" and rhyolitic magmas from a primary tholeiitic magma. Hamilton's argument for this origin of the rhyolites of Yellowstone and Island Park rested largely on his conclusion that rhyolitic and basaltic eruptions had alternated frequently in the later volcanic history of both the Island Park and Yellowstone areas. Subsequent work, as pointed out in this paper (see pls. 2 and 3; also, Christiansen, 1982), negates that conclusion. Hamilton also suggested that a mixture of the olivine basalts and rhyolites of the Island Park area would be equivalent compositionally to an average tholeiite of the Columbia River Plateau. It is now widely recognized, however, that the basalts of Island Park as well as those of Yellowstone and the Snake River Plain that were erupted during periods of nearby rhyolitic eruption, are themselves tholeiites. Some of these basalts have very low alkali contents but are as siliceous as some tholeiites of the Columbia River Plateau (tables 16, 17). To form the basalts and rhyolites of the Yellowstone Plateau volcanic field by separation of immiscible liquids in anything like the relative volumes represented would require a primary magma with a composition more like a rhyodacite. The compositions of the low-potassium Yellowstone and Island Park tholeiites are remarkably similar to those of some oceanic-island tholeiites (table 17) and evolved from a primary magma series (Hildreth and others, 1991). Furthermore, Yoder (1971) noted that for rhyolitic and basaltic liquids to be layered in a single crustal magma chamber as envisioned by Hamilton (1965) requires several hundred degrees of superheat in the rhyolites throughout the time of coexistence; this condition is unlikely as most of the rhyolites, including the voluminous ash flows, had abundant phenocrysts when they erupted.



Presnall's analysis (1969) of the partial fusion process, along with some experimental evidence, gives conceptual support to the hypothesis that two homogeneous liquid phases might be formed by fractional melting of a single parent material in the mantle (Yoder, 1973). The required mechanism would seem to include generation of silicic magma by very small degrees of partial melting, its rise in small batches to a region where it could accumulate into large magma bodies, and then generation of basaltic magmas. This mechanism, however, fits in poorly with the Yellowstone pattern, in which (1) very large batches of rhyolitic magma seem to have been generated repeatedly within the same general area over a period of 2 million years, and (2) basaltic magmas were, from the beginning of activity, erupting around and within this same area.

The contrasting isotopic compositions of neodymium, strontium, and lead in the basaltic and rhyolitic rocks of Yellowstone (Doe and others, 1982; Hildreth and others,

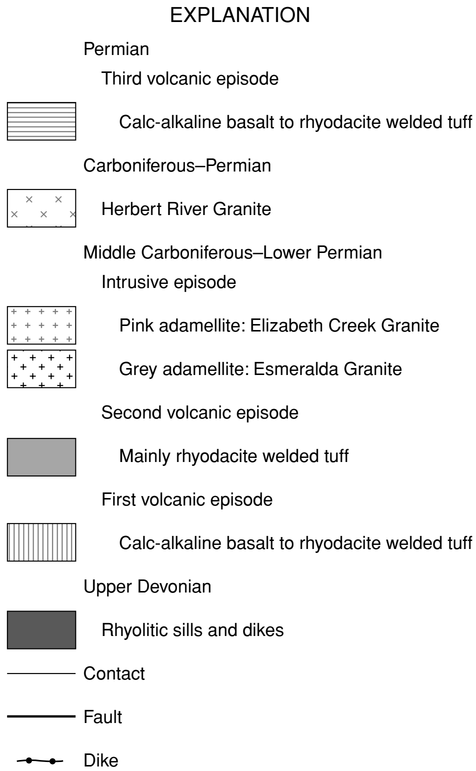


Figure 47.—Generalized geologic map of the Georgetown Inlier, Queensland, Australia (modified from Branch, 1966). The younger units of the Georgetown inlier are bimodal in composition and were erupted in an extensional tectonic system, suggesting possible deeper-level analogues for the relation of the Yellowstone-Snake River Plain system to basin-range tectonics. Numbers correspond to subsidence structures described and discussed by Branch (1966), Oversby and others (1980), Mackenzie (1987; 1993), and Mackenzie and others (1993).

1991) are consistent with generation of the contrasted magmas from separate parental materials. The strontium-isotopic evidence must be applied cautiously because the very low strontium contents of the rhyolitic magmas makes them especially susceptible to the isotopic effects of interaction with upper-crustal materials having high $^{87}Sr/^{86}Sr$ ratios. Three additional points bear on this evidence. First, the $^{87}Sr/^{86}Sr$ ratio of the Yellowstone rhyolites (typically about 0.7103±0.0015) is too low to allow them to be considered as melts of the sialic upper crust, which in this region is mainly older than about 2.5 Ga. and generally has present $^{87}Sr/^{86}Sr$ ratios of 0.740 and higher (for example, Reed and Zartman, 1973; Doe and others, 1982). Second, although some of the variation in Sr-isotopic ratios of the Yellowstone rhyolites probably does reflect additions of upper-crustal strontium (Hildreth and others, 1984; 1991), the ratios generally are much more uniform than would be expected if their deviation from the typical basalt ratios of about 0.706 were wholly the result of crustal assimilation. Third, as noted earlier, because the strontium-isotopic ratios of the Gardner River mixed lavas fit the same patterns as other rhyolites and basalts, these ratios are difficult to explain by bulk assimilation.

Similar arguments apply to neodymium and lead isotopes; the isotopic ratios of the rhyolites are too far from upper-crustal values and too nearly uniform to reflect genesis from upper-crustal rocks although they clearly show the effects of assimilation of upper-crustal materials.

Although the Yellowstone rhyolitic magmas were not partial melts of the ancient sialic crust as now represented by exposed rocks, there is strong support for including the involvement of a thick crust in any hypothesis for generation of the magmas of Yellowstone and other such voluminous bimodal fields. Although these fields are mainly continental, in many fundamental respects they resemble oceanic volcanism. (1) Like the basaltic volcanism of the ocean basins, continental basalt-rhyolite fields may involve either tholeiitic or alkali-basalt magmas or both, and the compositions of these continental basalts can be more or less matched among common oceanic basalts (table 17). (2) Average total eruptive rates, especially in regions with tholeiitic affinities, are higher in both oceanic basalt and continental bimodal settings than in many other volcanic systems. (3) Tectonic conditions during volcanism in both settings are nonorogenic or extensional, as in the spreading oceanic ridges and in the intraplate linear volcanic island and seamount chains. The linear evolution of the Snake River Plain-Yellowstone province is comparable in rate of volcanic propagation and in average volume rates of magma production to the linear Hawaiian volcanic chain (compare with Jackson and others, 1972) except that silicic magmas account for a considerable proportion of the Snake River-Yellowstone system but not of the Hawaiian system. Voluminous rhyolites are absent from nearly all oceanic occurrences; the single notable exception is that of Iceland, and there the rhyolites are associated with a perhaps exceptionally thick oceanic crust.

These considerations lead to the concept of voluminous bimodal rhyolite-basalt volcanism as a generally continental equivalent of major basaltic volcanism in the ocean basins. The same considerations, along with compositional and isotopic data reviewed previously, point to the lower crust as the likely source of the rhyolitic magmas of the Yellowstone field. Evidence reviewed by Leeman and Manton (1971), Leeman (1979; 1982b), and Doe and others (1982) indicates that a mafic to intermediate lower crust, consisting largely of pyroxene granulites, underlies the Snake River Plain-Yellowstone region. These granulites, sampled locally as inclusions in basalts of the Snake River Plain, have generally high values of $^{87}Sr/^{86}Sr$ (0.7125-0.727), low $^{206}Pb/^{204}Pb$ (13.5-16.1), and varied but generally high $^{208}Pb/^{204}Pb$ (34.7-59.1). The lead and strontium data together were interpreted by Doe and others (1982) to suggest significant contributions to the basaltic magmas from the lithospheric upper mantle and derivation of the rhyolitic magmas from granulitic lower crust although isotopic patterns in both the rhyolites and the basalts show significant effects of upper-crustal interactions as well.

The nature of some of these crustal interactions was further elaborated by Hildreth and others (1984; 1991) by considering neodymium and oxygen isotopes as well as the lead and strontium systems. They showed that all the volcanic rocks of the Yellowstone Plateau volcanic field have been affected to some degree by the assimilation of either wall and roof rocks or hydrothermal brines (or both). The large rhyolitic magma chambers were drastically affected during and immediately following the climactic eruptions of the first and third volcanic cycles, but internal mixing in or later magmatic additions to these batholith-sized magma bodies tended to restore them toward initial values.

More fundamentally, Hildreth and others (1991) showed that the relatively limited neodymium, strontium, and lead values of the rhyolites are most consistent with derivation from lower-crustal rocks that were extensively hybridized by basaltic magmas from the upper mantle, probably contemporaneously with evolution of the Yellowstone Plateau magmatic system.

In summary, the generation and evolution of the magmas of the eastern Snake River Plain-Yellowstone Plateau region stems fundamentally from the formation of basaltic magmas by partial melting of the upper mantle. As in many other such regions, this magma generation is associated with regional extension. The common association of oceanic basaltic volcanism and of continental basaltic or rhyolite-basalt volcanism with tectonic extension suggests generation of many of the basaltic magmas by the reduction of pressure in the upper mantle during tectonic extension. It might be further suggested that if it were not for the presence of a crust thicker than about 15 km in the region of magma generation, the volcanism of voluminous bimodal volcanic regions would have been essentially the same as that of various oceanic volcanic zones; the Snake River Plain-Yellowstone province would resemble the Hawaiian chain. This formation and rise of basaltic magma is a mechanism of increased local heat flow from the mantle. The radiogenic heat production of continental crust is high relative to that of the mantle, and the thick crust in effect insulates its own lower part. The rise of major amounts of basaltic magma into and through a thick continental crust—and perhaps in certain other areas of relatively thick crust and high heat flow such as Iceland—may raise geothermal gradients sufficiently to allow partial melting in the crust by the pressure reduction accompanying tectonic extension. Lower crust, probably consisting of pyroxene granulites of mafic to intermediate composition, may be partially melted to form a primary rhyolitic magma. The process may be similar to that proposed by Bailey (1964) for the generation of alkalic magmas beneath zones of uplift and rifting. Bailey (1964; 1970) also pointed out that the physical conditions existing in the lower crust in such a setting might be expected to cause an influx of volatiles to the site of magma generation, further aiding in the process of partial melting.

By such mechanisms the rhyolitic magmas do not necessarily form in direct thermal equilibrium with the basaltic magmas or require superheat during their formation, but they nevertheless owe their origins directly to the formation and rise of basaltic magma into the crust. The basaltic and rhyolitic magmas may rise and differentiate independently in the crust; in such instances, basalt-rhyolite mixed-lava complexes may be the result of no more than an accidental encounter of separate magmas finding their ways to the surface through available tectonic fractures. In voluminous and long-lived systems, however, the basaltic component must play a more important role in providing a continued supply of thermal energy from the mantle to maintain the high-level rhyolitic magmas in a partially liquid state for geologically long times, enabling them to accumulate in large shallow chambers and to undergo extreme differentiation processes that lead to the enrichment of volatile constituents and their release in voluminous pyroclastic eruptions (Smith 1979; Christiansen, 1979; Hildreth, 1979; 1981; Christiansen, 1984). However, the suggestion of Gibson and Walker (1963) that the immediate presence of a basaltic component may always be necessary for viscous rhyolitic magmas to rise in near-surface fissures or to be triggered into eruption is not borne out at Yellowstone, where rhyolitic lava flows with no visible basaltic component are voluminous.

REGIONAL EVOLUTION OF THE CRUST AND UPPER MANTLE

The origin and evolution of such a massive magmatic system as that of the Yellowstone Plateau is but one manifestation of processes that involve the entire crustal and upper-mantle column and perhaps parts of the deeper mantle.

PLATE-TECTONIC FRAMEWORK

A rationale for placing the extensional late Cenozoic tectonics of the Western United States into the picture of global plate-tectonic movements was provided by Atwater (1970) and Atwater and Molnar (1973). More recently, aspects of this framework have been amplified and refined by Stock and Molnar (1988) and Atwater (1989). Lipman and others (1972), Christiansen and Lipman (l972), Christiansen and McKee (1978), and Wernicke and others (1987) attempted to show that this framework is useful in understanding much of the Cenozoic volcanic and tectonic history of the Western United States.

In brief, the Cenozoic Western United States is the leading edge of a migrating North America lithospheric plate. Before about 30 Ma, the plate was bounded on the west by a trench along which the Farallon and related plates, constituting the eastern Pacific Ocean floor, were undergoing subduction into the upper mantle beneath North America (McKenzie and Morgan, 1969; Atwater, 1970). An ancestral East Pacific Rise farther west produced new crust as the Farallon plate and the more westerly Pacific plate diverged from it. Volcanic arcs lay parallel to the continental margin of the Pacific Northwest and Mexico, and a complex volcanic system, predominantly andesitic and calc-alkalic, extended far toward the continental interior in the area between these arcs (Christiansen and Yeats, 1992). These volcanic systems affected most of a very large region that had undergone major compressive deformation and igneous activity during Mesozoic and earliest Tertiary time (Lipman and others, 1972; Elston and Bornhorst, 1979; Wernicke and others, 1987). During the Eocene this continental-margin system extended as far east as the Yellowstone region and produced the Absaroka Volcanic Supergroup. After the Eocene, however, calc-alkalic volcanism at this latitude occurred only in a relatively narrow volcanic arc that was restricted to areas nearer the Pacific margin. The East Pacific Rise and the continental-margin trench approached each other throughout early and middle Cenozoic time. At about 29 or 30 Ma they intersected. As they did, subduction of the Farallon plate in the area of intersection ceased, that segment of the trench closed, and a new kinematic pattern affected the margin of the North America plate, reflecting now the relative motion of the America and Pacific rather than the America and Farallon plates (McKenzie and Morgan, 1969; Atwater, 1970). This right-lateral relative motion is represented most conspicuously by the San Andreas fault zone of California and related faults. With further convergence of the East Pacific Rise and the continental margin, a progressively longer contact between the Pacific and North America plates developed and both the rise and the trench ceased activity along this contact.

The cumulative relative motion between the Pacific and North America plates is greater than that represented by the San Andreas fault system alone. Atwater (1970) and Atwater and Molnar (1973) proposed that the remainder of the motion is distributed over a wide “soft” marginal zone of the North America plate. Christiansen and Lipman (1972), Snyder and others (1976), Wernicke and others (1987), and Christiansen and Yeats (1992) showed that changes in the types of tectonic and igneous activity occurred throughout the area of Mesozoic and earlier Cenozoic orogenic and igneous activity in the Western United States during late Cenozoic time, perhaps shifting in locale to some degree as the zone of Pacific-American contact continued to lengthen. Whereas volcanism before this transition tended to be predominantly intermediate in composition, afterwards it tended to be either predominantly basaltic, mafic and alkalic, or bimodally basaltic and rhyolitic. Following Christiansen and Lipman, (1972), this younger volcanic regime is referred to as “fundamentally basaltic” (although, as pointed out by Hildreth, 1981, in a more profound sense perhaps all terrestrial magmatism is fundamentally basaltic).

After middle Miocene time, the remnant of the magmatic arc system became progressively restricted northward as the Pacific-America plate interface widened and is now represented by the volcanic Cascades of the Pacific Northwest. The history of tectonic extension is more complex; before about 17 Ma extension characterized only specific, generally narrow zones, and from then until 14-10 Ma it produced mainly wide tectonic basins of relatively low relief (Christiansen and Yeats, 1992). The early tertiary extension may have been affected by both back-arc and post-orogenic stress regimes (Zoback and others, 1981). As extension became more widely distributed to form essentially the present-day basin-range system, volcanism changed from predominantly andesitic and generally calc-alkalic to fundamentally basaltic. By about 17 Ma, much of the region east of the continental-margin volcanic arc was characterized by basaltic and bimodal rhyolite-basalt volcanism (Scholz and others, 1971; Christiansen and Lipman, 1972; Noble, 1972; Snyder and others, 1976; Christiansen and McKee, 1978; Christiansen and Yeats, 1992). By 10 Ma, a wide region south of the Snake River Plain and the High Lava Plains of Oregon was undergoing tectonic extension in the modern basin-range mode (Christiansen and McKee, 1978; Zoback and others, 1981; Christiansen and Yeats, 1992).

MacLeod and others (1976) showed that another propagating volcanic system began at about the same time as the Snake River Plain-Yellowstone system in about the same area, the region surrounding the Oregon-Idaho-Nevada boundary. That system, however, extended northwestward across Oregon and now reaches the Newberry volcano at the Cascade front. A line that connects the Newberry volcano, back along the trace of this propagating volcanic system to the region west of the Snake River Plain and then northeastward along the eastern Snake River Plain to the Yellowstone Plateau lies within a diffuse but important boundary zone. To the south along this entire line, the late Cenozoic tectonic pattern is a strongly and typically developed basin-range fault system that continues southward into the Great Basin, where the faults reflect major crustal extension—probably aggregating at least 250 km (Hamilton and Myers, 1966; Stewart, 1971; Proffett, 1977; Hamilton, 1978; Davis, 1980; Wernicke and others, 1988). Numerous earthquakes occur in this region, most of them in diffuse belts toward the margins (fig. 48). The region north of the Newberry-to-Yellowstone boundary zone also contains extensional normal faults and some strike-slip faults, but the faults are fewer, their aggregate extension is much less than to the south, and they generally cease to be the dominant elements of topography within several tens of kilometers north of the boundary. Seismicity declines markedly but continuously northward from near this boundary zone across the Northern Rocky Mountains and Columbia Plateau (fig. 48). This boundary zone itself is essentially aseismic. Not only is this boundary the trace of two diverging volcanic systems that propagated from between about 12 and 10 Ma to the present, but it also has been the locus of continuing basaltic volcanism throughout the Quaternary.

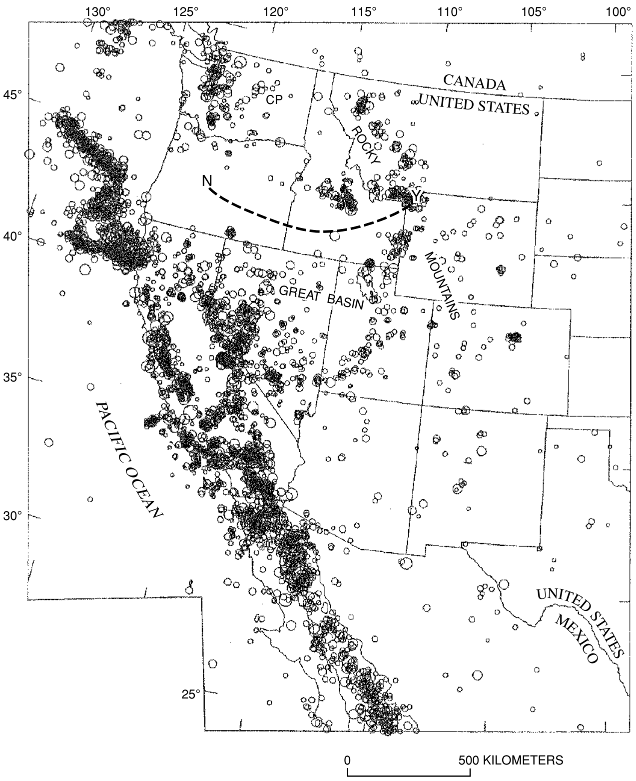


Figure 48.—Epicenter map showing regional historical and recent seismicity of the western contiguous United States (after Dewey and others, 1989, using data from Engdahl and Rinehart, 1988). The Newberry to Yellowstone boundary zone (centered along dashed line N—Y) is virtually aseismic. Relative to the Great Basin region, seismicity declines northward from near the boundary zone across the Northern Rocky Mountains and Columbia Plateau (CP).

Christiansen and McKee (1978) suggested that this diffuse boundary is, in effect, a "soft edge" to Atwater's soft plate margin; it is a boundary against which much of the oblique extension of the main basin-range region is taken up and at which the region of main basin-range deformation tends to become detached from the more nearly rigid bulk of the North America plate. To be clear, this is not envisioned as a strike-slip fault but, rather, as a diffuse zone of accommodation to greatly different amounts and rates of extension to the south and north. The aseismic nature of the boundary zone and its belt of recurrently active Quaternary volcanism may together indicate that temperatures are high and that basaltic magma exists at moderate depth along much of the zone.

MODELS FOR FUNDAMENTALLY BASALTIC MAGMATISM

The basaltic magmatism that has accompanied late Cenozoic extensional tectonism in the Western United States results from partial melting of the upper mantle and probably is an intimate part of the processes that cause high regional heat flow (Lachenbruch and Sass, 1977), a thinner crust and lithosphere, and anomalously low mantle seismic velocities in the basin-range region. These anomalous crustal and upper-mantle properties coincide with the region of late Cenozoic basaltic and bimodal rhyolite-basalt volcanism.

The plate-tectonic framework of late Cenozoic volcanism and tectonic extension leads directly to a basic question: Are the fundamentally basaltic volcanism of the Western United States in general, and of the Yellowstone Plateau and eastern Snake River Plain in particular, direct consequences of the plate interactions as they guide tectonic development of the region, or does the volcanism result from processes of deeper origin that may even be the causes of plate motion? Wilson (1963) and Morgan (1971; 1972) proposed a worldwide framework of convection plumes that rise from the base of the mantle to produce "hotspots" near the global surface where partial melting may produce voluminous magmatism and where the tractive force of the spreading plumes at the base of the lithosphere may tend to break the lithosphere and drive the resulting plates apart. Morgan proposed that Yellowstone marks the site of such a deep mantle convection plume and that the eastern Snake River Plain records the trace of the North America plate over the resulting hotspot. Several points have been raised in favor of this concept. First, as a global group and to a suitable approximation, the hotspots do appear to have remained fixed relative to one another as the lithospheric plates move in relation to them (Minster and others, 1974; Minster and Jordan, 1978; Gripp and Gordon, 1990; Acton and Gordon, 1994). Second, the mantle-plume concept suggests a mechanism that allows long-lived localized melting anomalies like Hawaii to draw on continuously renewed mantle from beneath the depleted asthenosphere as a source material for partial melting. And, third, the mantle-plume hypothesis correlates the average rate of volcanic propagation of certain linear volcanic chains very well with the relative motions determined for the lithospheric plates. In the particular case of Yellowstone and the eastern Snake River Plain, the orientation and linear volcanic propagation rate of the Snake River Plain-Yellowstone axis accords, within the resolution of the data and allowing for the effects of subsequent crustal extension, with the late Cenozoic relative movement of the Pacific and North America plates and the movement of the Pacific plate with respect to a fixed Hawaiian hotspot (Morgan, 1972; Smith and Sbar, 1974; Rodgers and others, 1990; Smith and Braile, 1994).

The concept of a convective plume rising from the lower mantle has, thus, proven attractive and has been applied to the Yellowstone-eastern Snake River Plain system by many investigators, notably including Morgan (1971; 1972), Matthews and Anderson (1973), Smith and Sbar (1974), Suppe and others (1975), Duncan (1982), Westaway (1989), Pierce and Morgan (1992), and Smith and Braile (1994). Several of these authors even credit a Yellowstone mantle plume with being one of the main driving forces of continental tectonics in the Western United States during the late Cenozoic. Hadley and others (1976) also explicitly applied the related concept of a chemical plume to the Yellowstone hotspot—a buoyantly rising mass within the sublithospheric mantle that is anomalously enriched in incompatible and heat-producing elements. Despite the prevailing view, I wish to examine critically the evidence cited in favor of applying the deep-mantle plume model to Yellowstone and to discuss a possible alternative explanation for the region's "hotspot" magmatism.

To begin with, the somewhat negative if obvious point should be made that the consistency of certain data with a hypothesis does not necessarily constitute a sufficient proof of the hypothesis. Thus, if all of the evidence cited above in favor of the mantle-plume hypothesis can be accounted for by alternative explanations, the evidence neither proves nor disproves any of the hypotheses. The evidence for relatively fixed positions of a group of proposed deep-mantle plumes, for example, may allow for relative movements between these sites of as much as 1-2 cm/yr (Minster and others, 1974; Molnar and Stock, 1987), a significant rate in the plate-tectonic movement picture of the earth. Secondly, the existence of more or less fixed melting anomalies or "hotspots," the velocity model of plate movements in relation to these hotspots, and a mechanism for continuously replenishing the source of mantle material for partial melting are equally consistent with other hypotheses such as McDougall's asthenospheric counterflow model (1971) or Shaw and Jackson's shear-melting and thermal-feedback model (1974) for the Hawaiian melting anomaly. None of the geophysical parameters globally concomitant with hotspots uniquely requires that the associated mantle convection originates in the lower mantle (Moriceau and others, 1991; Anderson and others, 1992).

The following deficiencies in application of the primary deep mantle-plume model to the volcanic and tectonic history of the Yellowstone melting anomaly may be noted:

1. The deep mantle-plume model carries no specific mechanism to explain, not merely volcanic propagation along the eastern Snake River Plain-Yellowstone axis, but also continuing basaltic volcanism that characterizes the Snake River Plain along both its western and eastern branches, in distinction, for example, to the pattern of activity in the Hawaiian volcanic chain.
2. The model provides no inherent explanation for the beginning of volcanism near the southwest end of the eastern Snake River Plain in the middle Miocene during a time of major tectonic reorganization from limited linear belts of extension associated with andesite-rhyolite volcanism to regionally distributed extension and fundamentally basaltic volcanism throughout much of the basin-range region, an area more than 2,000 km across. Some models, for example those of Pierce and Morgan (1992) and Parsons and others (1994), relate all of this early regional extension and magmatism to a diffuse plume head, but others, for example Geist and Richards (1993), require the plume to have been focused at the north edge of the region between 17 and 14 Ma in order to account for the voluminous flood-basalt volcanism of the Columbia River Plateau.
3. Effects attributed to the Yellowstone hotspot arose, apparently spontaneously, at about 17 Ma (Pierce and Morgan, 1992; Parsons and others, 1994), within a complexly generated part of the North American continental lithosphere (Christiansen and McKee, 1978); systematic northeastward propagation along the eastern Snake River Plain axis did not begin until about 12.5 Ma (Christiansen and Yeats, 1992). Duncan (1982) proposed that the Yellowstone melting anomaly was in fact continuous from an oceanic spreading center in Paleocene and Eocene time and was overridden by the North America plate along a curving path to generate magmas of the Columbia River Basalt Group and the volcanism of the western Snake River Plain before starting its linear path up the eastern Snake River Plain. Duncan's evidence for ridge-centered hotspot volcanism to form the Paleocene to Eocene basalts of the Oregon and Washington Coast Ranges can be evaluated independently of its identity with the Yellowstone melting anomaly; that identity seems unlikely. There is no record of volcanism on the North America plate above this hotspot between about 49 and 17 Ma; it presumably was shielded by the subducting slab beneath the ancestral Cascades (Duncan, 1982; Hill and others, 1992). In Duncan's model, after having taken 32 m.y. to move (relative to the present surface) 400 km eastward from the Washington Coast Range to the eastern Columbia Plateau (1.3 cm/yr), the hotspot must have turned and migrated 400 km southeastward down the western Snake River Plain between 17 and 15 Ma (20 cm/yr) before turning again and starting northeastward up the Snake River Plain at about 3.5 cm/yr; no such plate path or pair of velocity discontinuities is evident in the oceanic magnetic record relative to the independent hotspot frame of reference.
4. The propagating Yellowstone melting anomaly seems to be one of a pair that evolved at about the same time from the initial widespread basaltic and bimodal magmatism of 17-14 Ma in the region of southeastern Washington, eastern Oregon, western Idaho, and northern Nevada. The other volcanic belt, also a bimodal rhyolite-basalt system, has propagated northwestward across eastern and central Oregon at about the same rate as the propagation of the Yellowstone-eastern Snake River Plain volcanism (MacLeod and others, 1976). The two melting anomalies are basically similar and would seem to have been formed by the closely related processes. Yet, if the propagating Yellowstone melting anomaly were a direct reflection of a deep mantle convection plume, then the propagating Oregon melting anomaly would require a different process of formation. Pierce and Morgan (1992) interpreted such a dif-

ference for this eastern Oregon belt by comparing only its propagating part (less than10 Ma) to the entire post-17-Ma history of the Yellowstone hotspot (although its propagating part too dates only from 12-10 Ma) and by noting its lower magmatic productivity. Nevertheless, its magmatic productivity—though less than Yellowstone's—was significant, including voluminous ash-flow fields of 9.3-6.5 Ma (Walker, 1979), and its symmetry with and similarity in rate of propagation to the Yellowstone hotspot strongly suggest a linked origin of these two systems.

5. The region south of the two zones of propagating rhyolitic volcanism has undergone large amounts of basin-range extension, probably 200-300 km (Wernicke, 1992), whereas that to the north has undergone appreciably less extension, which ends within several tens of kilometers to the north (Lawrence, 1976; Christiansen and McKee, 1978; Christiansen and Yeats, 1992). The primary deep-mantle plume model can explain this only as coincidence.

6. Although the Snake River Plain-Yellowstone axis conforms to the kinematic constraints of North America plate motion (fig. 49), as Smith and Braile (1994) demonstrated in detail, it also coincides with a unique crustal tectonic setting. The hotspot began to propagate at the west edge of the Archaean continental crust of the Wyoming province (Reed, 1993). Eaton and others (1975) and Mabey and others (1978) showed that a regional aeromagnetic anomaly pattern along the axis of the eastern Snake River Plain continues both southwestward into Nevada and northeastward (interrupted only by a magnetic low associated with the Yellowstone magma

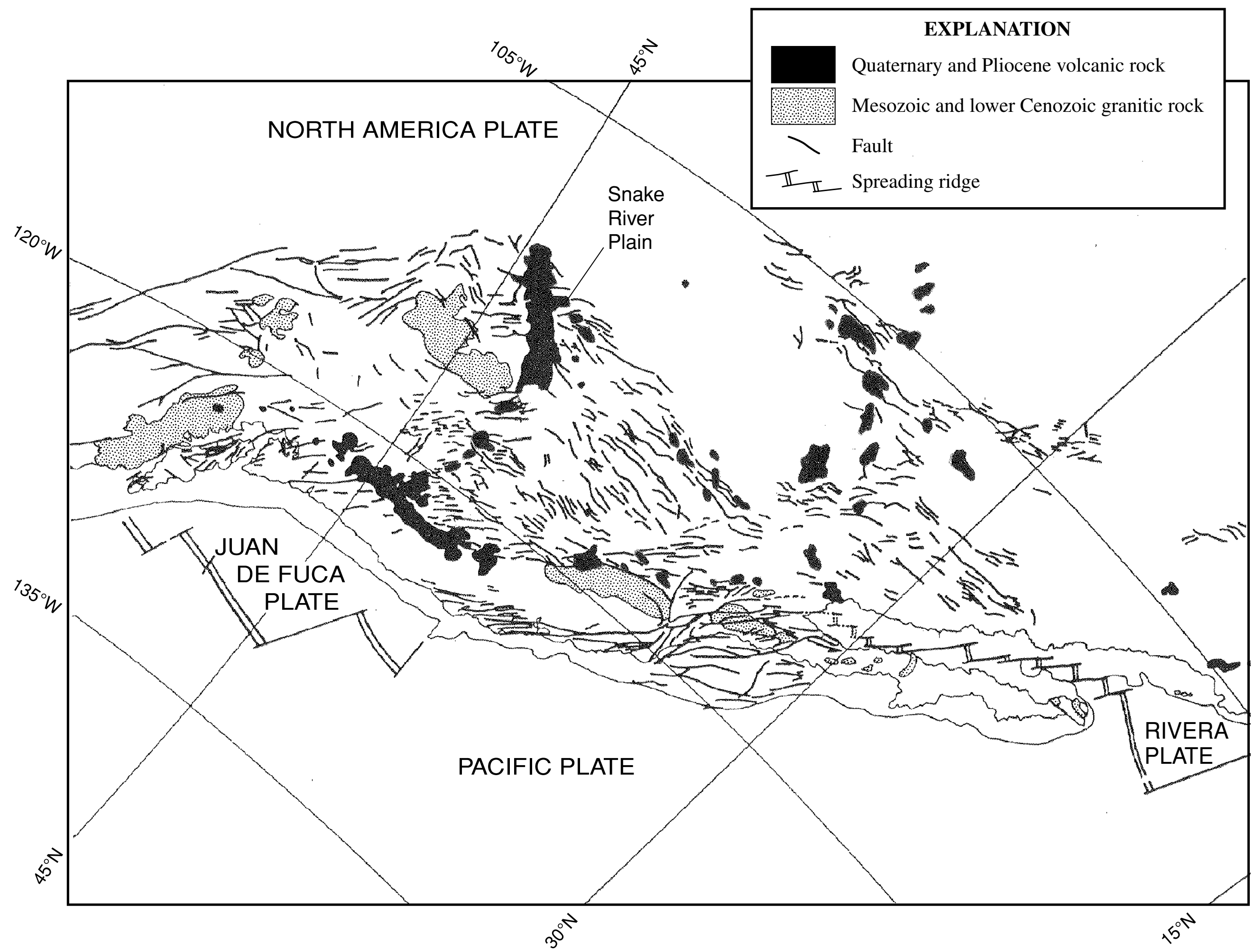


Figure 49.—Selected tectonic elements of Western North America in relation to right-lateral relative motion of the Pacific and North America lithospheric plates. The map is plotted on a Mercator projection with its pole at the pole of relative rotation of these two plates so that structures oriented horizontally in the projection and related to plate motion have strike slip; oblique structures in the projection that are related to plate motion have oblique deformation, with components of both strike slip and either extension or contraction. After Atwater (1970) with volcanic features added from Luedke and Smith (1984).

chamber) across Montana to the Canadian border. This anomaly pattern northeast of Yellowstone cuts across Laramide structures in a region practically undeformed in late Cenozoic time but parallels the major regional Precambrian structural trend; thus, it appears that the Snake River Plain-Yellowstone axis follows a regional structural zone of Precambrian origin (not necessarily a specific fault, as critiqued by Pierce and Morgan, 1992).

The foregoing considerations lead me to outline the following qualitative conceptual model that treats the late Cenozoic basaltic and basalt-rhyolite volcanism of the Western United States, and of the Yellowstone system in particular, as a direct result of late Cenozoic intraplate extension, which in turn is intimately related to late Cenozoic interactions of the lithospheric plates. This model may serve as a future basis for comparison to the widely discussed deep-mantle convective plume concept.

In the model proposed here, as the zone of direct interaction between the Pacific and North America plates grew at the expense of oceanic ridge and trench in early and middle Miocene time, a relaxation of stress affected the adjacent region of the North America plate (Zoback and Zoback, 1989) that previously had been compressively deformed and heated by magmatism during late Mesozoic and earliest Tertiary time. Specific more or less linear zones through this region had also undergone large extensions and voluminous magmatism in Eocene and later mid-Tertiary time. This previously deformed and heated region first responded to region-wide extensional stress by stretching, thinning, and uplifting. By about 17 Ma the remnant arc was restricted to a relatively small part of the continental margin, and extensional thinning affected not only a zone directly inland from the coastal strike-slip system but also the northern basin-range and Columbia Plateau regions behind the arc.

As the remnants of the Farallon plate shrank, stress relief and associated thinning of the previously deformed and heated lithosphere produced regional tectonic extension, rise of the underlying upper mantle, and widespread partial melting in that mantle. Buoyant rise of these melts in turn caused widespread basaltic intrusion into the lower crust, increased regional heat flow, markedly decreased the rigidity of the lithosphere, and further enhanced lithospheric extension and thinning and the development of anomalously low upper-mantle seismic velocities. This chain of heating events seems to have begun at about 17-18 Ma along the axis of the region from the southeastern Great Basin to the Columbia Plateau with a north-northwest strike, reflecting a uniform regional west-southwest minimum horizontal stress orientation (Zoback and Thompson, 1978; Christiansen and McKee, 1978; Christiansen and Yeats, 1992; Zoback and others, 1994; Parsons and others, 1994). The occurrence of the mid-Miocene rift zone along the center of an extended and uplifted region of earlier deformation and volcanism might be interpreted as indicating the greatest efficiency of extensional stress relief in causing partial melting along the mid-line of a region of intense earlier mid-Tertiary magmatism.

Within less than 3 million years after initiation of this extensional tectonic system at about 17-18 Ma, volcanism of fundamentally basaltic character extended outward from the initial rift axis to blanket a large region of eastern Washington and Oregon, western Idaho, and northern Nevada (Christiansen and McKee, 1978; Draper, 1991; Christiansen and Yeats, 1992). With the increasing width of a zone of lithospheric extension and thinning by about 10 Ma, virtually to the west and east margins of the present Great Basin region, the upper crust became segmented into blocks, with extensional faults aligned in directions consistent with a continuously adjusting stress field (Zoback and others, 1981; Zoback and Zoback, 1989) that may reflect the influence of Pacific-North America right-lateral relative motion on parts of the extending region. In some places the normal faults that produce this later segmentation probably are guided by preexisting structures. The melting caused by regionally distributed tectonic stress relief probably resulted in low percentages of partial fusion and the production mainly of an alkali-olivine basaltic suite (Leeman and Rodgers, 1970). Only near the margins of the region of great extension have tholeiites erupted (Christiansen and Lipman, 1972; Christiansen and McKee, 1978; Draper, 1991), perhaps indicating locally shallower melting or higher percentages of partial melting in zones of concentrated extension near those boundaries. These combined tectonic and magmatic processes have produced buoyant rise of upper-mantle material beneath the extending region and a feedback relation between lithospheric extension and upper-mantle convection. With time, perhaps as heightened regional heat flow has reduced the rigidity of the lithosphere within the center of the active zone and that lithosphere has approached a stable configuration, active extension, seismicity (fig. 48), and the production of basaltic magmas have tended to become increasingly concentrated toward the margins of the region (Christiansen and McKee, 1978). Along the Yellowstone hotspot track, a zone of enhanced extension has migrated both along with the hotspot and outward from the margins of the eastern Snake River Plain (Anders and others, 1989; Pierce and Morgan, 1992)

In this model, tectonic stress relief and resultant partial melting in the upper mantle occur up to a diffuse north boundary of the basin-range region. As the formerly wider zones of large extension and crustal basaltic intrusions concentrated outward and narrowed toward the east and west margins of the region of extension, their trailing margins projected along this north boundary as eastward- and westward-sweeping waves of enhanced extension and lithospheric magmatism. Along the axis of the Snake River Plain, this propagating wave was localized and guided by an ancient structural zone that was oriented almost exactly in the direction of plate motion (fig. 49). This propagating wave of basaltic magmatism initiated the Yellowstone hotspot by crustal melting. The seismic belt along the east margin of the basin-range region marks the border between a broad region of high heat flow and the cooler, more or less rigid plate beyond. The intersection of this progressively restricted and focused border zone with the structurally controlled Snake River Plain axis has localized and concentrated voluminous magmatism along the north boundary zone of the basin-range extensional region.

If the Yellowstone melting anomaly in this conceptual model was initially localized by a preexisting structural boundary, it was enhanced because this controlling structure was fortuitously oriented in the direction of motion of the lithospheric plate relative to the underlying mantle (compare with Smith and Sbar, 1974; Smith and Braile, 1994). For this reason, it is proposed shear melting at the base of the lithosphere accelerated the local partial melting process, heating the lithosphere and initiating a thermal feedback cycle, as proposed by Shaw and Jackson (1974) for Hawaii. That is, as large volumes of melt were produced, they rose buoyantly, reduced pressure in the upper mantle, and resulted in further melting of the mantle. This feedback process has sustained the most productive magmogenetic system in the Western United States during late Cenozoic time. These factors of enhanced lithospheric extension, shear melting, and thermal feedback have combined to make this a self-sustaining melting anomaly comparable in magnitude to the Hawaiian system.

The Yellowstone melting anomaly produces a relatively high percentage of partial melting in the affected part of the upper mantle so that the primary magmas are low-potassium tholeiitic types with trace-element patterns comparable to the tholeiites of high-volume oceanic-island systems. Because of the thick continental crust above the zone of basaltic magma generation, however, a higher percentage of the magma may remain low in the crust as intrusive bodies than in an oceanic setting. As the wave of enhanced magmatism has passed northeastward along the axis of the Snake River-Yellowstone system, decreased degrees of partial melting, the rise of relatively undepleted mantle, and increased lithospheric interactions may produce the successively higher-potassium basalts with higher Fe/Mg, Ti, P, and incompatible trace elements, typical of the Snake River basaltic magmas.

In the propagating zone of concentrated extension, enhanced partial melting of the lower crust resulting from the increased regional heat flow and from basaltic intrusions into the lower crust has produced rhyolitic magmas which, being less dense than the basaltic magmas, have risen higher in the crust along the Snake River Plain axis and have accumulated into batholith-sized bodies at favorable structural intersections. As long as continued high rates of basaltic magma production in the mantle sustained these silicic magmatic systems, they produced cycles of voluminous rhyolitic volcanism at the surface. However, as plate motion slowly displaces the lithosphere, it eventually migrates away from the self-sustaining focus of melting. Eventual cooling of the granitic batholiths may result in volume decrease and subsidence along the volcanic axis (Brott and others, 1981; Smith and Braile, 1994); continued regional extension, however, favors episodic rise of smaller volumes of basaltic magma, resulting in occasional eruption of relatively more potassium-rich basalts along the subsiding axis of the plain, loading it and causing further subsidence. The lack of a well-marked graben along this rift may indicate less rigid deformation of this hot, still partly molten zone than of surrounding regions. Thus, if a slice beneath the eastern Snake River Plain at several kilometers depth would be expected to reveal a chain of ring intrusions invaded by a discordant batholithic complex, as suggested in the previous chapter, a slice at perhaps 15 or 20 km might reveal an intrusive complex of differentiated tholeiites. Hamilton's suggestion (1959) that Yellowstone is the surface expression of a lopolith may indeed be conceptually useful although with reference to a greater depth and more complex processes than Hamilton proposed.

The few seismic experiments to date designed specifically to investigate the roots of the Yellowstone magmatic system (Iyer, 1975; 1978; 1984; Evoy, 1978; Iyer and others, 1981; Daniel and Boore, 1982; Humphreys and Dueker, 1994a; b) have shown that anomalous geophysical properties are associated with the system to a depth of about 200 km. Plume-like convection related to the Yellowstone hotspot should result from upward and lateral inward flow of hotter, relatively undepleted mantle material to replace that depleted by partial melting and added to the extending lithosphere.

To summarize this discussion of crustal and mantle evolution and to simplify its rather long speculative reach, the plutonic and upper-mantle components of the Yellowstone magmatic system constitute by far the bulk of its volume. Considered in its entirety, this system extends downward from the volcanic field and its subvolcanic rhyolitic intrusions, through a batholith-sized body of rhyolitic magma—complexly mixed with more mafic magmas in its lower regions—on through a basaltic intrusive regime within and around which the lower crust has been partially melted, into an upwelling zone of basaltic magma generation in the upper mantle. Within and beneath this zone of melting a residual mantle is at least partially displaced by lateral inflow of undepleted mantle. Whether this convective system in the upper mantle is sustained by plume-like convection from the lower mantle—even the core-mantle boundary—remains even more highly speculative and is unresolved.

THE FUTURE OF YELLOWSTONE VOLCANISM

It seems appropriate to conclude this speculative final chapter on magmotectonic evolution of the Yellowstone Plateau volcanic field with a brief appraisal of the prospects for its volcanic future. The problem is considered here from three points of view. First, the Yellowstone Plateau possibly represents the later stages of a resurgent cauldron cycle with possibilities for further, but generally declining rhyolitic volcanism in and near the caldera as this cycle wanes. Second, basaltic volcanism might accompany or succeed the closing stages of the rhyolite cycle, first outside but ultimately within the Yellowstone caldera. Finally, the possibility is quite realistic that yet another volcanic cycle could begin with a new insurgence of rhyolitic magma or might even have been underway for about the past 160,000 years. Each of these three potential states poses the possibility of future volcanic eruptions in the Yellowstone National Park area. Clearly, no statement of any of these possibilities can be made confidently, but each of the three possible states seems to be about equally probable.

It was suggested earlier that the Yellowstone Plateau volcanic field may now be in late stage VI or stage VII of the resurgent cauldron cycle. No rhyolitic eruptions have occurred for about 70,000 years within the third-cycle source area of the volcanic field, and rhyolitic volcanism related to this cycle may have waned. The very high heat flow represented by the Yellowstone hydrothermal system, however, as well as individually inconclusive geologic, gravity, aeromagnetic, geodetic, and seismic evidence, together suggest that rhyolitic magma probably is still crystallizing and perhaps still rising below the plateau region. Most areas of hydrothermal activity are localized by parts of the ring-fracture zones of the Yellowstone caldera where rhyolitic eruptions have not occurred for several hundred thousand years. Consequently, the intense hydrothermal activity of the Firehole River, Shoshone Lake, and Yellowstone Lake-Grand Canyon areas does not in itself appear to herald further volcanic eruptions. By contrast, the area of intense hydrothermal activity at and near Norris Geyser Basin is adjacent to the Gibbon River flow—one of the youngest flows of the rhyolite plateau—and lies at the end of the Norris-Mammoth corridor, an extracaldera belt of marked recurrent hydrothermal activity, faulting, seismicity, and young rhyolitic and basaltic lavas. It seems likely that Norris Geyser Basin and the Roaring Mountain area could overlie localized shallow bodies of rhyolitic magma. Upward movement of these magma bodies, perhaps triggered by tectonic movements, could produce centralized eruptions of rhyolitic ash, pumice, and lava.

The long-term volcanic history of Yellowstone and the eastern Snake River Plain clearly indicates that, at some time, further basaltic eruptions will occur in the Yellowstone National Park region. However, plateau-marginal basaltic eruptions in any given area have been relatively infrequent except for a few times in the history of the rhyolite plateau, such as shortly after eruption of the Lava Creek Tuff. The probability of a basaltic eruption in the near future in any one of the zones of recurrent basaltic volcanism marginal to the Yellowstone caldera is significant but small; no reasonable way of quantifying it is apparent. The evidence for still-crystallizing magma beneath the Yellowstone caldera makes basaltic eruption within the caldera unlikely for at least tens of thousands of years.

Ever since the outlines of a multistage ash-flow and caldera history for the Yellowstone-Island Park region first emerged during fieldwork, it has been necessary to address nagging questions about the possibility of further voluminous ash-flow eruptions and caldera formation. Such events would present a major human disaster, not only for Yellowstone National Park but also for a region of thousands of square kilometers around it. Calamitous effects, including the destruction of crops, property, and perhaps many lives would be likely as far away as the central Great Plains, judging from the known distribution of volcanic ashes related to the three major eruptions of the Yellowstone Group. The chemistry of the youngest lavas of the Plateau Rhyolite suggests that they are evolved about as far as the last precursory rhyolitic lavas of the third-cycle climax (tables 8 and 12-14; Christiansen, 1984, fig. 6). About 700,000 years intervened between the climactic episodes of the first two volcanic cycles and about 650,000 years between the second two; 640,000 years have elapsed since the latest major ash-flow eruptions. Do these factors suggest a periodicity, and that another cycle could be imminent? Again, it is not now possible to meaningfully define quantitative probabilities for such an occurrence within the foreseeable future. It was earlier suggested that doming and the extrusion of the Mallard Lake and Central Plateau lavas probably represent the rise and reintrusion of magma from deeper levels in a magma chamber whose upper part has partly crystallized. Alternatively, however, these lavas could represent to a major new insurgence of magma from deeper levels. The latest major phase of rhyolitic activity began about 160,000 years ago with doming and has produced voluminous rhyolites along the Madison-Pitchstone and Central Plateau zones during three principal eruptive pulses. Vertical deformation at high rates continues to be measured geodetically within the caldera (Pelton and Smith, 1979; 1982; Dzurisin and Yamashita, 1987; Dzurisin and others, 1990; 1994). This doming and volcanism, the existence of young arcuate faults in the Mirror Plateau area, a seismic-attenuation "shadow" in the northeast part of the caldera (Lehman and others, 1982), and the present high convective heat flow could reflect a new cycle of activity.

There are some guides as to where any new major ash-flow eruption and caldera collapse might be centered. Lehman and others (1982) and Smith and Braile (1983; 1994) suggest that magma lies at the shallowest levels in the northeastern part of the Yellowstone Plateau and perhaps in the southwestern part. Each of the three major volcanic source areas of the Yellowstone Plateau volcanic field has formed at the intersection of the Snake River-Yellowstone axis with a major tectonic fault zone that is radial to the volcanic source region (figs. 5, 44). It seems likely that these intersecting fault systems were seismically active at the times of new major magma insurgences; therefore, one guide should be the presently active seismic zones. Rhyolitic lavas have preceded each major ash-flow and caldera episode; therefore, another guide would be the areas of youngest rhyolitic volcanism. In these respects, the Mallard Lake-Central Plateau stage of doming and eruption, relative to the third volcanic cycle, could be analogous to the Bishop Mountain and Green Canyon flows and to the structural setting of the second volcanic cycle relative to the first. It is reasonable to postulate voluminous ash-flow eruption and caldera formation from a new ring-fracture zone in the vicinity of the Firehole River drainage basin. How real and how distant such prospects might be cannot now be quantitatively assessed.

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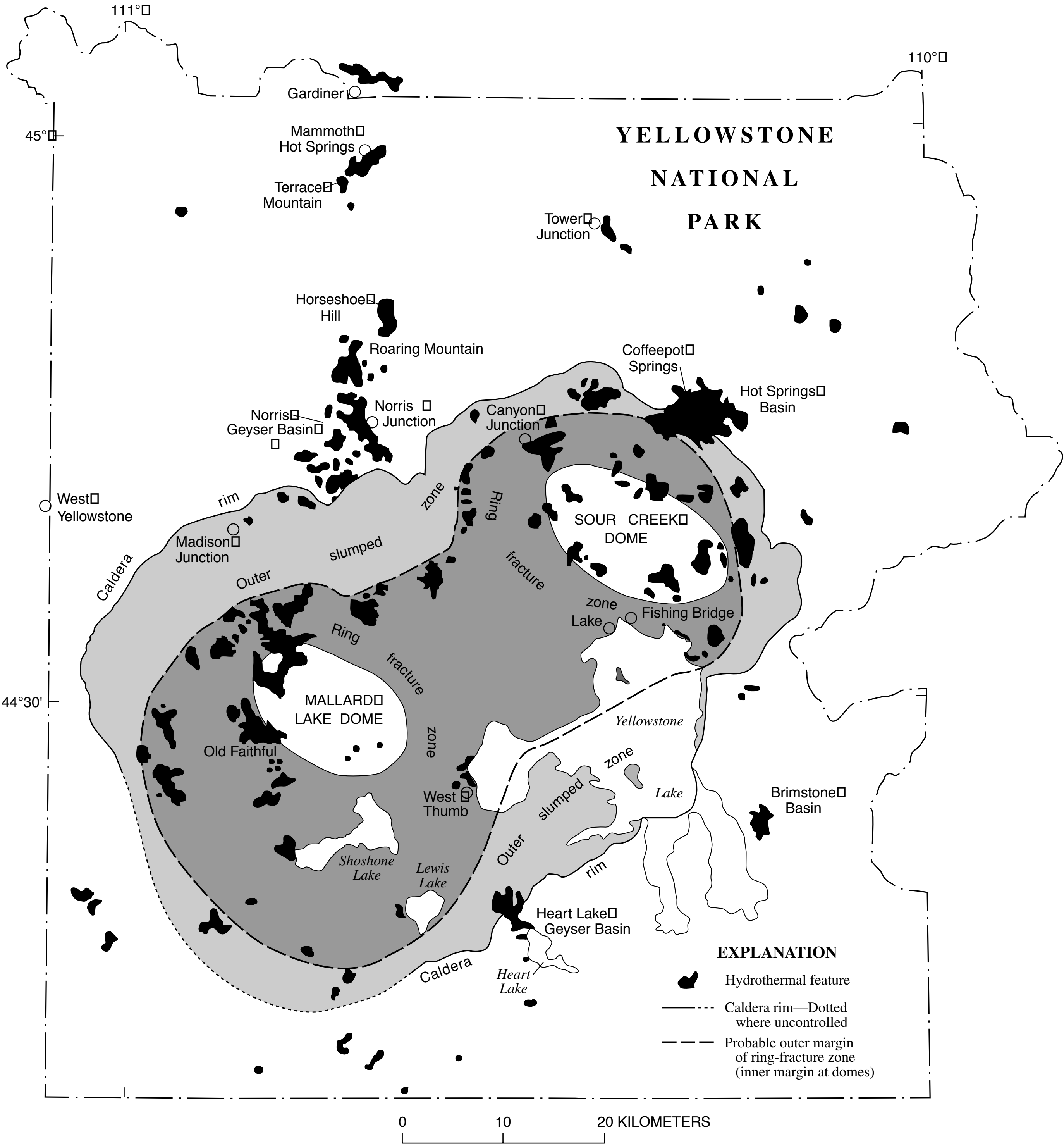
111°
110°
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44°30'
YELLOWSTONE
NATIONAL
PARK
Gardiner
Mammoth Hot Springs
Terrace Mountain
Tower Junction
Horseshoe Hill
Roaring Mountain
Coffeepot Springs
Hot Springs Basin
Norris Junction
Norris Geyser Basin
Canyon Junction
West Yellowstone
Madison Junction
Caldera rim
Outer slumped zone
Ring fracture zone
SOUR CREEK DOME
Fishing Bridge
Lake
MALLARD LAKE DOME
Old Faithful
West Thumb
Yellowstone Lake
Brimstone Basin
Shoshone Lake
Lewis Lake
Heart Lake Geyser Basin
Heart Lake
EXPLANATION
Hydrothermal feature
Caldera rim—Dotted where uncontrolled
Probable outer margin of ring-fracture zone (inner margin at domes)
0
10
20 KILOMETERS