

**Topographic signature, space-time evolution,
and physical properties of the Yellowstone-
Snake River Plain volcanic system: the
Yellowstone hotspot**





Frontispiece. Photograph of Steamboat Geyser, Norris Junction, Yellowstone National Park, in full eruption on July, 6, 1984 (photograph by Robert B. Smith).

Topographic signature, space-time evolution, and physical properties of the Yellowstone-Snake River Plain volcanic system: the Yellowstone hotspot

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Abstract

The track of the Yellowstone hotspot—an 800 km linear age progression of silicic volcanic centers that runs through the Snake River Plain to the Yellowstone caldera—provides direct evidence for mantle plume-plate interaction as the mechanism responsible for the Yellowstone-Snake River Plain (YSRP) 16 million-year volcanic system. Caldera-forming rhyolitic volcanism, active crustal deformation, extremely high heat flow (30 times the continental average), and intensive earthquake activity are manifestations of the hotspot currently active at Yellowstone National Park. Anomalous low P-wave velocities in the upper crust of the Yellowstone caldera are interpreted as solidified but still hot granitic rocks, partial melts, hydrothermal fluids, and sediments. Unprecedented deformation of the Yellowstone caldera of as much as 1 m of uplift from 1923 to 1984, followed by subsidence of as

much as ~12 cm from 1985 to 1991, clearly reflects a giant caldera at unrest.

The regional signature of the Yellowstone hotspot is highlighted by an anomalous, 600 m high, approximately 600 km wide, topographic bulge centered on the caldera. We suggest that this feature reflects long-wavelength tumescence of the hotspot. Yellowstone is also the center of a +10 m to +12 m geoid anomaly—the largest in North America—which extends about 500 km laterally from the caldera, similar in width to the geoid anomalies of oceanic hotspots and swells. The 16 million year trace of the Yellowstone hotspot, the seismically quiescent Snake River Plain, is surrounded by “bow-wave” or parabolic-shaped regions of earthquakes and high topography. The systematic topographic decay southwest along the Snake River Plain, totaling 1,300 m, fits a

model of lithospheric cooling and subsidence that is consistent with the passage of the North American plate across a mantle heat source. The rate of silicic-volcanic age progression of the YSRP (4.5 cm/yr) includes a component of southwest motion of the North American plate, modeled at ~2.5 cm/yr, and a component of concomitant crustal extension estimated to be 1 to 2 cm/yr. The YSRP also exhibits anomalous crustal structure, which we believe is inherited from magmatic and thermal processes associated with the Yellowstone hotspot. This includes a thin, (2 - 5 km thick) surface layer composed of basalt and rhyolites and an unusually high-velocity (6.5 km/s), mid-crustal mafic layer that we suggest reflects extinct "Yellowstone" magma systems, which have replaced much of the normal granitic upper crust. Direct evidence for a mantle connection for the YSRP system is from anoma-

lously low P-wave velocities which extend from the crust to depths of approximately 200 km. These properties and the kinematics of the YSRP are consistent with an analytic model for plume-plate interaction that produces a "bow-wave" or parabolic pattern of upper mantle flow southwest from the hotspot, similar to the systematic patterns of regional topography and seismicity. Our unified model for the origin of the YSRP is consistent with the geologic evidence: basaltic magmas ascend from a mantle plume to interact with a silicic-rich continental crust, producing partial melts of rhyolitic composition and the characteristic caldera-forming volcanism of Yellowstone. Cooling and contraction of the lithosphere follows the passage of the plate over the hotspot, with continuing episodic eruptions of mantle-derived basalts along the SRP.

Introduction

Our knowledge of mantle hotspots has been derived largely from observations of active volcanoes located primarily in oceanic settings. The Yellowstone-Snake River Plain (YSRP) volcanic system of Idaho, Wyoming, and Montana, including the Yellowstone silicic volcanic field and the related bimodal basaltic-rhyolitic volcanic field of the eastern Snake River Plain (SRP), affords an opportunity to investigate the dynamics and properties of one of the Earth's largest silicic-dominated volcanic systems. We will develop a unifying model for the origin of the YSRP that reflects mantle plume-plate interaction consistent with observed geologic and geophysical data. We will also show how the Yellowstone hotspot has affected more than 20% of the western United States during its 16 million year evolution (Figure 1).

The currently active element of the YSRP is a large caldera complex located at Yellowstone National Park, Wyoming. However, unlike most of the Earth's volcanic and tectonic features generally found at or near plate boundaries, Yellowstone National Park (here loosely called Yellowstone) is located ~2,000 km from the western margin of the North American plate. Yellowstone is also located at the boundary of two major tectonic provinces, the Basin and Range and the Rocky Mountains. Here,

the immense scale of its Pleistocene volcanism, extremely high heat flow (~30 times the continental average), intense seismicity, widespread hydrothermal activity, and unprecedented crustal deformation makes Yellowstone a natural laboratory for investigating volcanic and related tectonic processes (**Frontispiece**).

The 1983, $M_s = 7.3$, Borah Peak, Idaho, earthquake has recently focused attention on the influence of the Yellowstone hotspot on the seismotectonics of the YSRP (Smith and others, 1985). To examine this relationship we plotted the regional topography, locations of silicic volcanic centers (Figure 2), earthquakes, and late Cenozoic normal faults in a region we believe has been affected by the hotspot (Figures 3 and 4). The most important features of Figures 3 and 4 are: (1) a northeast progression of decreasingly younger silicic volcanic centers along the YSRP, (2) "bow-wave" or parabolic-shaped areas of high topography and seismicity surrounding the aseismic SRP, and (3) systematic patterns of similar-aged normal faulting subparallel to the SRP which follow the general patterns of topography and seismicity. In this paper, we will relate these features to a model for plume-plate interaction and show how the geophysical and geodetic data of the YSRP are consistent with a mantle hotspot source.

What is a hotspot?

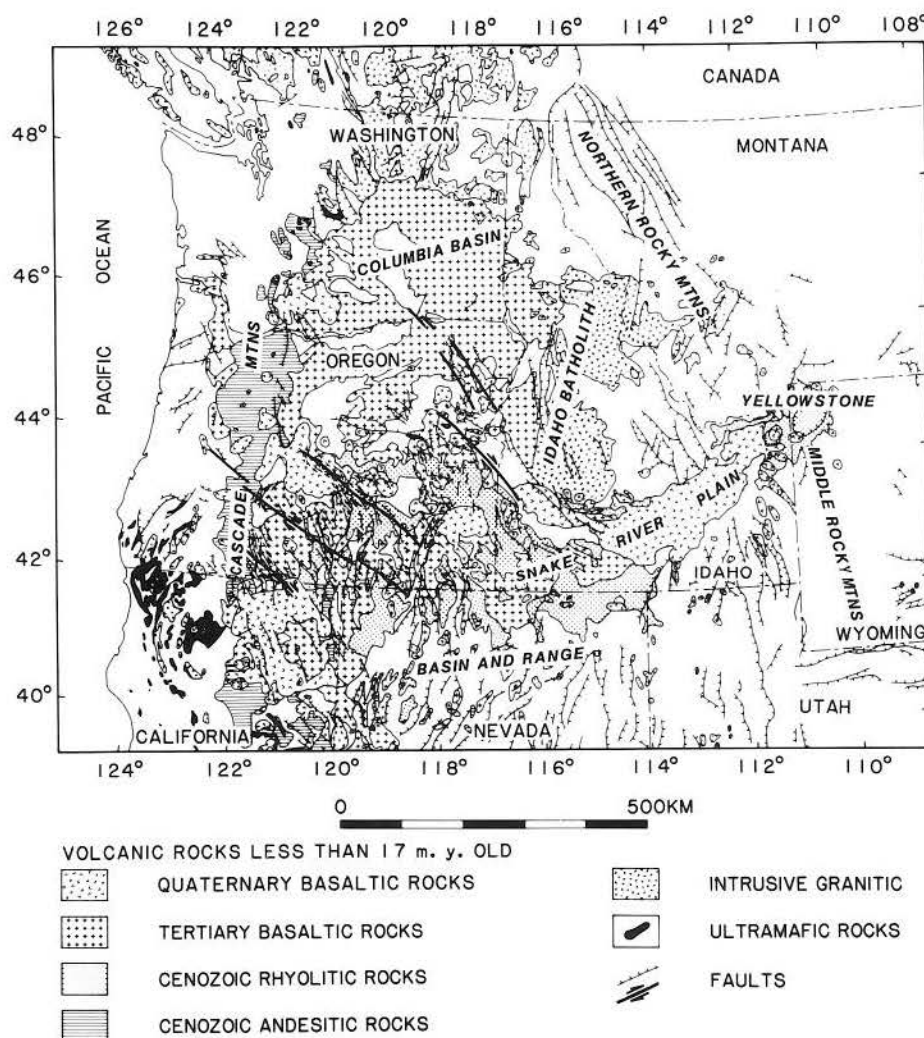
We associate the term "hotspot" with a geologic feature of long-lived active volcanism and high heat flux originating from the asthenosphere and resulting in regional topographic uplift and undulations of the geoid. These features have been interpreted for various oceanic swells and linear tracks of oceanic volcanism as evidence of upwelling mantle plumes. Our use of the term hotspot also reflects the generalized definition of Crough (1978) of a region of mid-plate or anomalous ridge crest volcanism that is either persistent or accompanied by a broad topographic swell. These definitions encompass such oceanic features as the Hawaiian chain, Cape Verde Rise, the Azores, and the Galapagos spreading center, all of which share properties with the YSRP.

As discussed by Thompson and Gibson (1991), we expect that the hotspot location (defined by surface manifestations of deep-seated processes) is within the plane of the axis of an underlying mantle plume. We thus define the Yellowstone hotspot as a large area of voluminous silicic volcanism that has evidence for mantle heat and a basaltic magmatic source. Further discussions of hotspots and plumes are contained in recent reviews by Richards and others (1989), Sleep (1990), and Duncan and Richards (1991).

The Yellowstone-Snake River Plain volcanic system

The heart of Yellowstone National Park's majestic mountain scenery is a large, Pleistocene, silicic volcanic field that is distinguished by three large calderas with a total eruptive volume of $\sim 8,500 \text{ km}^3$ (Christiansen and Blank, 1972; Christiansen, 1984, 1993). This area is best known for its spectacular display of geysers, fume-roles, and hot springs (Frontispiece). However, the accumulated effect of Yellowstone's volcanism is a $\sim 2,500 \text{ m}$ high, relatively flat and undulating mountainous terrain, termed the Yellowstone Plateau, which is surrounded by high peaks of the Rocky Mountains, some exceeding $4,000 \text{ m}$ elevation.

To the southwest, the Snake River Plain occupies an 800 km long and 80 km wide topographic depression that descends to elevations of $\sim 1,200 \text{ m}$ in southwestern Idaho (Figure 3).



SOURCES: VOLCANIC ROCKS—STEWART AND CARLSON (1978)
NORMAL FAULTS AND INTRUSIVE ROCKS — KING (1969)
STRIKE SLIP FAULTS — LAWRENCE (1976)

Figure 1. Generalized tectonic map of the northwestern United States showing the Yellowstone and Snake River Plain region.



Figure 2. Photograph of Big Southern Butte, an example of a siliceous volcanic cone (in the far distance), and Middle Butte, a basaltic cone (in the distance) on the Snake River Plain near Arco, Idaho (photograph by Lawrence W. Braile).

The Snake River Plain, covered primarily by late Cenozoic basalt flows, is underlain by silicic volcanic rocks (Figure 2) similar in composition to those found in Yellowstone. The volcanic rocks age progressively to the southwest, from 0.6 Ma at Yellowstone to as old as 16 Ma in southwestern Idaho and northern Nevada (Armstrong and others, 1975). The systematic increase in the age of the silicic volcanic centers and the decrease in topography along the SRP are the best and most important surface manifestations of the Yellowstone hotspot. We will show corroborating evidence from geophysical and geodetic data for this hypothesized mantle source.

Early naturalists in Yellowstone

We point out the role of Yellowstone, not only as one of the world's greatest scenic wonders, but as the birthplace of the U. S. National Park system. Ferdinand V. Hayden (1872, 1873) was the first scientist to describe Yellowstone during his exploration of "Coulter's Hell", as it was known by fur trappers in the early 19th century for its rumored expanse of geysers and hot springs in a largely unexplored region of the Rocky Mountains. Hayden's perspective and championing of Yellowstone's geologic wonders and spectacular mountain scenery lead to the enactment of legislation in 1872 establishing the first national park in the world, Yellowstone National Park, and helped establish the beginnings of the global environmental ethic.

Impressed with Yellowstone's magnificent high mountain scenery and evidence of youthful volcanism, Hayden summarized an insightful interpretation of its origin that is still valid today:

From the summit of Mt. 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. This basin has been called by some travelers the vast crater of an ancient volcano. 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 rocks, and volcanic dust were poured in unlimited quantities.

Hayden (1872)

Hayden's experiences and insight into the significance of the scientific opportunities at Yellowstone were amplified by the views of Lt. Gustavus C. Doane, the military leader of a 1870 expedition into Yellowstone and cited from his personal journal entry on September 24, 1870. Doane's statement emphasizes the importance of Yellowstone as a multidisciplinary natural laboratory:

As a country for sight seers, it is without parallel. As a field for scientific research it prom-

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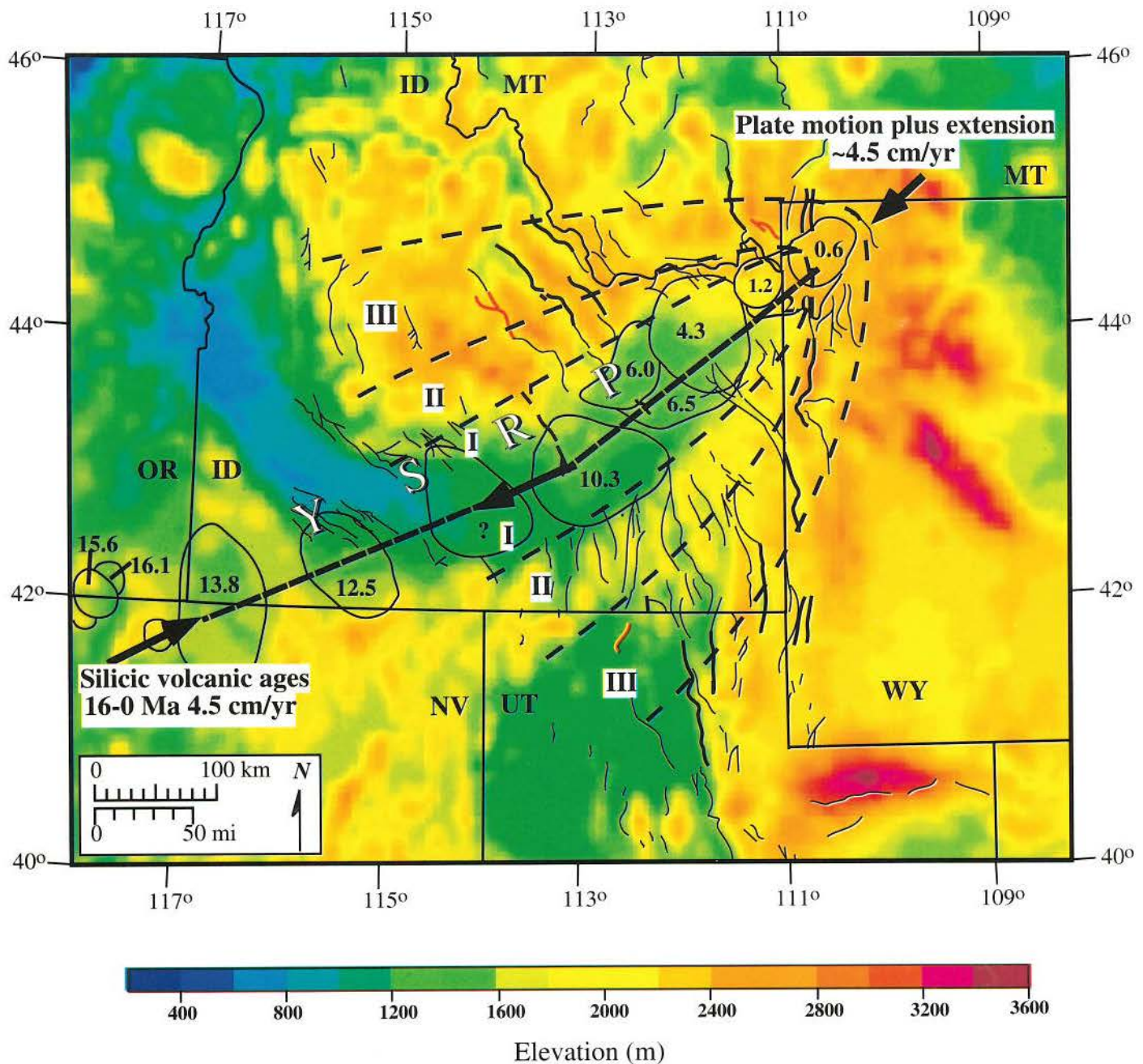


Figure 3. Topography, late Cenozoic faulting, and volcanic features of the Yellowstone-Snake River Plain. The digitized topography was contoured by color with the warmest colors (reds, purples) corresponding to the highest topography (exceeding 3,500 m), grading to lowest elevations, (~1,200 m) shown by the cooler colors of green, blue and, magenta. The locations of silicic volcanic centers are from the maps of Christiansen (1984, 1993) for Yellowstone and Island Park; and for the Snake River Plain, from Mel Kuntz (U. S. Geological Survey, 1985, personal communication), Morgan and others (1984), and Pierce and Morgan (1990, 1992). The dashed line corresponds to the estimated center line of the YSRP. The plate motion and silicic volcanism progression vectors are from this paper (see Table 1). Selected late Cenozoic normal faults from compilation of Smith and Arabasz (1991). Light lines = late Cenozoic to late Quaternary, bold lines = faults $\leq 14,000$ ^{14}C years, and red lines = faults with historic rupture. This figure is modified and updated from earlier versions by Smith and others (1985, 1990) and Smith (1989, 1990).

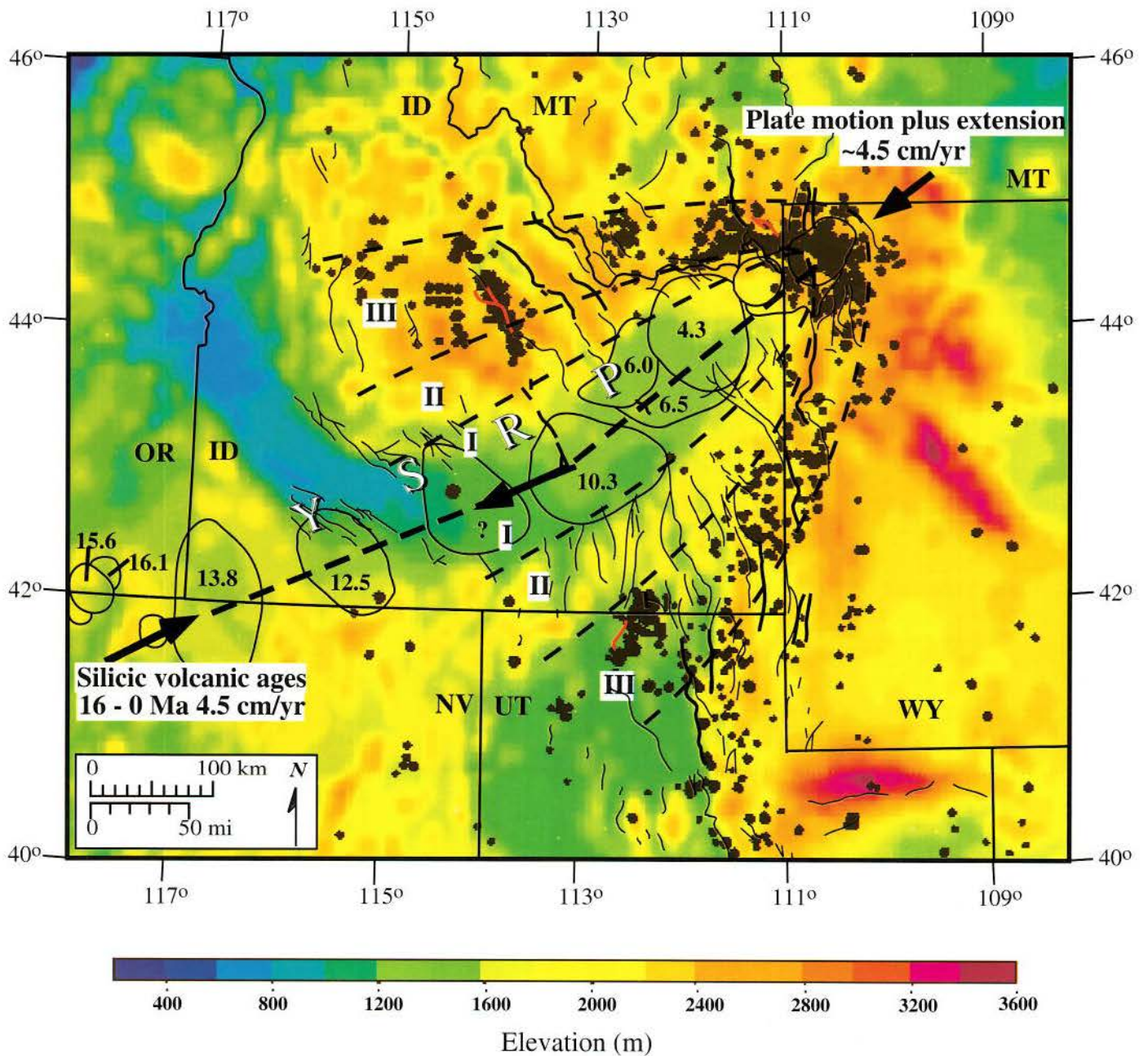


Figure 4. Topographic and seismicity signature of the Yellowstone hotspot, showing the “bow-wave” or parabolic-shaped seismotectonic domains (defined in text) I, II and III. Earthquake epicenters are shown as black filled circles and are scaled in size to magnitude; $2.5 \leq M \leq 7.5$, for the period, ~1900 to 1985. The earthquake data are from earthquake compilations of the Intermountain region by the University of Utah (Eddington and others, 1987; Smith and Arabasz, 1991) and Engdahl and Rinehart (1988).

ises great results, in the branches of Geology, Mineralogy, Botany, Zoology, and Ornithology. It is probably the greatest laboratory that nature furnishes on the surface of the globe.

Doane (1870)¹

We show throughout this paper that the YSRP system is indeed a natural laboratory and provides an unusual window into the workings of the Earth's interior revealing information on such important processes as mantle derived volcanism, magmatic transport and heatflow, and surficial manifestation of plume-plate interactions. We will address these topics by first summarizing the late Cenozoic volcanic history of the YSRP, then describing the distinctive physical properties of this system such as seismic velocity structure, heat flow, gravity and geoid

fields, topographic features, and finally, concluding with a discussion of a mantle plume-plate model which accounts for many of the observations.

We include a discussion of the geophysical and geological evidence regarding the space-time evolution and physical properties of the YSRP volcanic system as it pertains to an origin associated with a mantle hotspot. We conclude with a discussion emphasizing the corroboration of diverse geophysical findings of the YSRP with the results of volcanological, geomorphic, and petrological studies and comment on the unusually good agreement between these data and a unified model for plume-plate interaction as the mechanism responsible for the Yellowstone hotspot.

Track of the Yellowstone hotspot—the Snake River Plain

The volcanic history of the Snake River Plain is not as well known as that of Yellowstone, primarily because most of the late Cenozoic silicic volcanic rocks and related calderas are buried beneath younger basalts (Figure 2). Nonetheless, the first definitive evidence for a systematic time-progressive volcanic origin of this region was recognized by Armstrong and others (1975), who mapped and dated silicic volcanic rocks by the K-Ar method along the Snake River Plain. Morgan and others (1984) employed paleomagnetic methods and identified several large calderas in the northern part of the SRP with dimensions similar to those of Yellowstone (Figure 3). Leeman (1982) summarized the Snake River Plain volcanic system based on petrology and trace-element analysis of volcanic rocks.

Discussions of the YSRP must also include its possible relationship to the Miocene Columbia Plateau basalt field of western Oregon and Washington, which has been attributed by several authors to a mantle plume. For example, Thompson (1977) and Draper (1991) suggested that the short-lived magmatism of the Columbia Plateau volcanic field was fed by upper mantle dikes from a "Yellowstone" plume located in southwestern Idaho, 16 to 17 Ma. While there is no directional age progression of the Columbia Plateau basalts, the immense scale of this volcanic field suggests that a mantle source cannot

be ruled out; however, there is little geophysical data for the Columbia Plateau or its transition to the Snake River Plain. We therefore will not attempt to address the possible relationships between these provinces, but we offer, in a later section, reasons why we believe that the YSRP is not directly related to this flood basalt province.

Age progression of silicic volcanic centers

The best evidence for systematic motion of the North American plate across a mantle magmatic source is the progressive increase in age of silicic volcanic rocks southwest along the YSRP (Figure 5). The youngest silicic volcanic centers, in the Yellowstone volcanic field, are less than 2.0 Ma. These are followed by a sequence of silicic centers at about 6 Ma, southwest of Yellowstone. A third group, about 10 Ma, is centered near Pocatello, Idaho. The oldest mapped silicic rocks of the SRP, about 16 Ma, are distributed across a 150 km wide zone in southwestern Idaho and northern Nevada, the suspected origin point of the YSRP (Figure 3). See Table 1 for a compilation of data relating to plate motions of the YSRP. Appendices I and II contain the data used in the linear regression models shown in Figure 5.

¹ Cited in Bonney and Bonney (1970, p. 387).

A 3.5 cm/yr systematic northwestward decrease in age of silicic volcanic rocks of the YSRP was first described by Armstrong and others (1975). However, these authors did not speculate on a mechanism. Since Armstrong's pioneering work on the Snake River Plain, his results have been cited by various authors as evidence for the track of the North American plate across a mantle plume. Smith (1990) and Smith and others (1990) used Armstrong's data and calculated a 4 cm/yr rate for the silicic volcanic age progression, which is similar to the rate of 4.5 cm/yr given by Rodgers and others (1990).

We have also calculated the silicic volcanic age-progression rate by fitting a linear regression of silicic volcanic ages and locations along the YSRP (Figure 5). Our analysis shows an average rate of silicic volcanic age progression of 4.5 cm/yr for the past 20 million years, similar to that of several other studies cited here. However, we think that the data distinguish two regimes: (1) a rate of 3.3 cm/yr for the 0 to 8 Ma period, and (2) a rate of 6.1 cm/yr for the 16 to 8 Ma period. Variations in the rates of age progression may be due in part to varying episodes of superimposed Basin and Range extension, to the combined scatter of the reported dates by various authors, and to spatial variations in the reported locations and widths of the silicic centers. Nonetheless, the 0 to 16 Ma averages for all reported age rates fall within 4 ± 1 cm/yr.

Plate tectonic models

The first recognition of plate motion at Yellowstone was by Morgan (1972), who calculated a southwest-trending vector for the

North American plate of 1 to 2 cm/yr from a hypothesized plume beneath Yellowstone with a track along

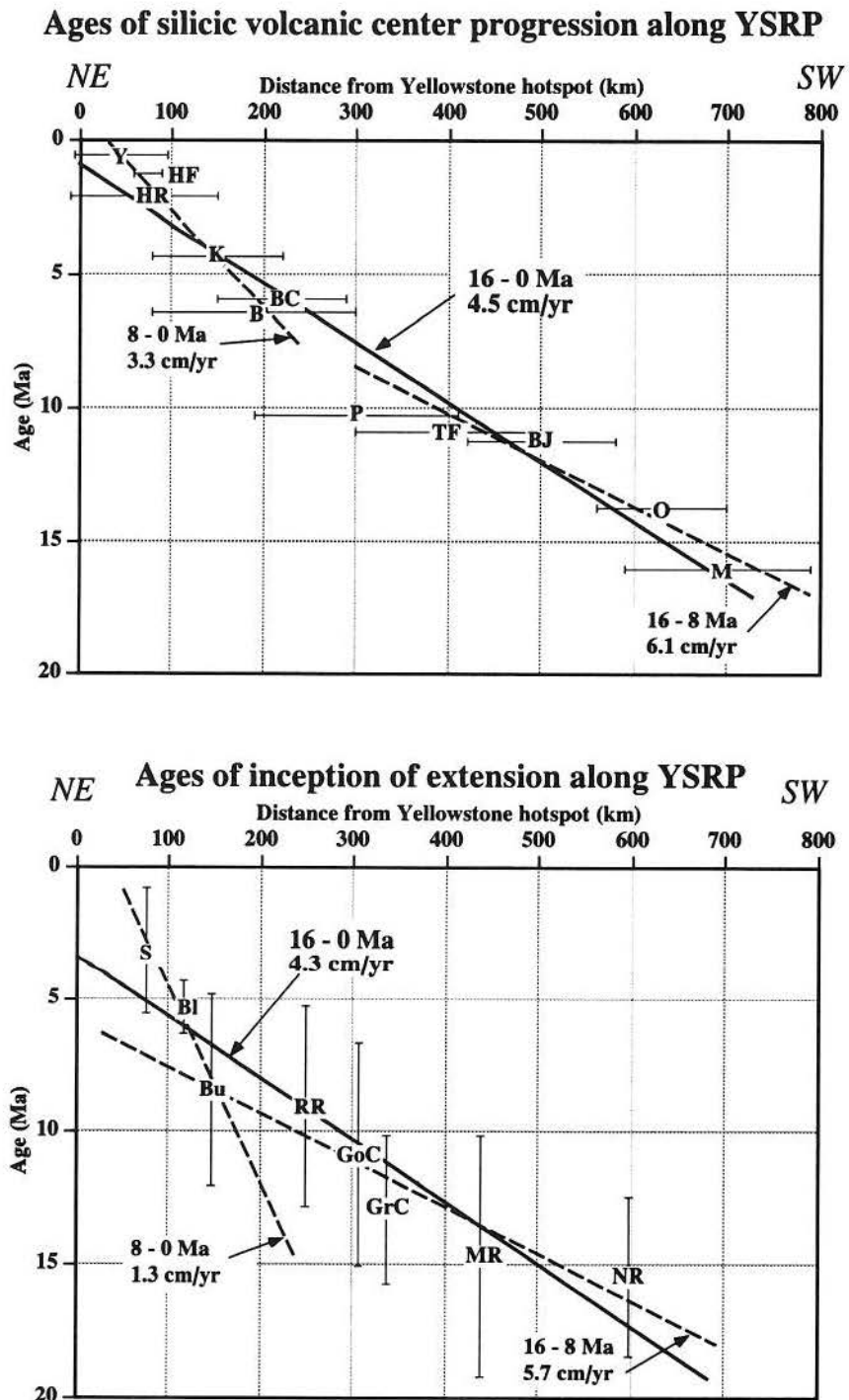


Figure 5. (Top) calculated rates of silicic volcanic age progression. (Bottom) rates of inception of crustal extension (data from Rodgers and others, 1990) for the YSRP. See Appendix I and II for references and lists of data used to construct the regression models. Solid lines are fits to all of the data. Brackets give an indication of errors due to the size of calderas (top) and the uncertainties in ages (bottom). Bold dashed lines are fits to data in two periods: 0-8 Ma and 10-16 Ma.

the Snake River Plain.² A plate motion vector at Yellowstone of 2.5 cm/yr and at an azimuth of 245° was calculated by Smith and Sbar (1974) for the relative motion between the North American plate at Yellowstone with respect to the Juan de Fuca and Pacific plates. They also calculated a rate of 4.5 cm/yr and a 245° azimuth for plate motion at Yellowstone, constrained by a 6 cm/yr rate of interplate motion between the North American and Pacific plates along the San Andreas fault. Using an average displacement rate of 2.4 cm/yr along the San Andreas fault (Wallace, 1990) and the same data as Smith and Sbar (1974), we calculated a plate motion rate at Yellowstone of 3.4 cm/yr. These rates compare well with values determined from a global multi-plate inversion of 2 to 3.1 cm/yr evaluated at Yellowstone (Table 1). The differences in the various plate motion vectors come from errors in the calculated Euler poles produced by variations in such data as oceanic paleomagnetic anomalies, slip vectors from earthquakes, and the various assumptions of net-rotation and net-torque models.

The most recent models of plate motion at Yellowstone are from DeMets and others (1990), who determined rates of North American plate motion using variations of the multi-plate model, NUVEL-1, and incorporated fixed hotspot models of Gripp and Gordon (1989, 1990). Their calculations yielded plate velocities at Yellowstone of 0.9 cm/yr to 2.5 cm/yr. We have used a rate of ~2 cm/yr that is averaged for the absolute plate motion at Yellowstone by excluding the extreme values of multiplate NUVEL-1 models (Table 1).

Rate of inception of extension

The rates of plate motion implied by the silicic volcanic age progression (Armstrong and others, 1975) include the accompanying extension of the SRP and the surrounding Basin and Range crust. Richard L. Armstrong first suggested to one of us (RBS) that he believed the apparent progression of the inception of normal faulting adjacent to the Snake River Plain reflected regional extension related to the age progression of silicic volcanism. Scott and others (1985) and Pierce and Morgan (1990) noted similar progressions of the inception of normal faulting along the SRP, whereas Rodgers and others (1990) used the ages of inception of sedimentation in basins on the south side of the SRP as

evidence for inception of regional extension (Figure 5).

We have also examined the rates of inception of crustal extension by modeling the ages of inception of sedimentation given by Rodgers and others (1990). We excluded the Teton fault because new evidence suggests that this feature may have evolved as early as the Miocene (Byrd and Smith, 1991). From our calculations (Figure 5), we suggest that the ages of inception of extension reflect the same two regimes as noted in the silicic volcanic age progressions. In the first phase, from 16 to 8 Ma, the rate of inception of extension of 5.7 cm/yr compares with the rate of 6.1 cm/yr from the age progression of silicic volcanic centers (Figure 5). Whereas for the younger regime of 8 to 0 Ma, the rate of extension inception of 1.3 cm/yr is smaller than the silicic age rate of 3.3 cm/yr. Our 16 to 0 Ma rate was 4.3 cm/yr which is remarkably similar to the silicic volcanic age progression rate of 4.5 cm/yr.

How the Pacific and North American plate motions have affected the SRP evolution during its 16 million-year history is discussed by Pollitz (1988). He made calculations of rates based upon the silicic volcanic ages and moreover noted two regimes: (1) 20 to 10 Ma with a rate of 3.1 cm/yr at an azimuth of 267°, and (2) 9 to 0 Ma, marked by a change in azimuth to 234° with a rate of 2.9 cm/yr (Table 1). Our model (Figure 5) also reveals the change at 8 to 10 Ma that we believe marks a distinct decrease in the relative rates of either plate motion or extension rate. Pollitz (1988) attributed the change in plate direction at ~9 Ma to either a plate reorientation or to a change in the magnitude of forces acting on the northwest margin of the Pacific plate. The 10 to 8 Ma change is also near the time of inferred clockwise rotation (estimated to be about 45°) of the least principal stress direction from ENE-WSW to WNW-ESE in the northern Basin and Range province at approximately 10 Ma (Zoback and others, 1981). Thus, the change in rate and the inception of extension at 8 to 10 Ma on the YSRP may be associated with late Cenozoic intraplate adjustment within the western U. S.

An important factor not incorporated into previous estimates of Yellowstone plate motion is the effect of concomitant crustal extension associated with the evolution of the Basin and Range province. We estimated the long-term extension rate of the SRP by

²Note that these azimuths are for the direction of plate motion and are opposite to the directions implied by the progression of silicic ages.

Table 1. Table of calculated and observed rates of plate motion for the North American plate relative to the Yellowstone hotspot.

No.	Feature	Type of measurement	Rate cm/yr	Azm. (°)
<i>North American plate motion modeled at Yellowstone</i>				
1.	Yellowstone hotspot	Multiplate model	1 to 2	SW
2.	Yellowstone hotspot	North American-Pacific interplate model	4.5	246°
			2.4	246°
3.	Yellowstone hotspot	Multiplate model	2.4	214°
4.	Yellowstone hotspot	North American-Pacific plate model	~ 2	225
5.	Trend of the YSRP	Intraplate model	2.8 3.0	234° 245°
6.	Yellowstone hotspot	NUVEL1-HS3 multiplate model with hotspots (Gripp and Gordon, 1989)	0.9	258°
7.	Yellowstone hotspot	NUVEL1-HS2 multiplate model with hotspots (Gripp and Gordon, 1990)	2.3	238°
8.	Yellowstone hotspot	NUVEL1-NNR (no net rotation) multiplate model with hotspots (Argus and Gordon, 1991)	1.9	238°
9.	Yellowstone hotspot	N. America-Pacific interplate model	2.5	240°
<i>Age progression of silicic volcanic rock along YSRP volcanic system</i>				
10.	YSRP	Age determinations of dated silicic volcanic rocks	3.5	SW
11.	YSRP	Map projection of dated silicic volcanic rocks	4.0	065°
12.	YSRP	Map projection of dated silicic volcanic rocks	4.5	056°
13.	YSRP	Map projection of dated silicic volcanic rocks	2.9 7.0	054° 070°-075°
14.	YSRP	Map projection of dated silicic volcanic rocks	2.9 7.0	054° 073°
<i>Inception of normal faulting and extension adjacent to YSRP</i>				
15.	Northern Basin and Range	Intraplate motion across the Basin and Range at lat. 41° N.	1.7	E-W
16.	Southeast of SRP	Inception of normal faulting	1.0	—
17.	Southeast of SRP	Inception of normal faulting	3.7	056°
18.	Southeast of SRP	Inception of normal faulting	1	—
19.	North of SRP	Inception of late Quaternary normal faulting	1	—
20.	Basin-Range	Inception of normal faulting	1	—
21.	South side of SRP	Late Quaternary faults	~1-2	SW

Table 1. *Continued.*

Comments	Reference
Generalized interplate motions	Morgan, 1972
Constrained by San Andreas motion at 6 cm/yr	Smith and Sbar, 1974
Constrained by San Andreas motion of 3.4 cm/yr	Wallace, 1990
Improved multiplate inversion model	Minster and Jordan, 1978
Multiplate model 50 - 0 Ma	Engelbreton and others, 1985
9 - 0 Ma 20 - 10 Ma	Pollitz, 1988
Multiplate inversion using hotspot framework	Calculated by C. DeMets for this paper.
Multiplate inversion using hotspot framework	Calculated by C. DeMets for this paper.
Multiplate inversion using hotspot framework	Calculated by C. DeMets for this paper.
Relative motion with respect to Juan de Fuca plate	This paper
Dating silicic and basaltic volcanic rocks of the YSRP	Armstrong and others, 1975
Re-evaluation of silicic volcanic ages of Armstrong and others	Smith and others, 1990
Re-evaluation of silicic age rate and related extension	Rodgers and others, 1991
Re-evaluation of silicic age rates 10 - 0 Ma 16 - 10 Ma	Pierce and Morgan, 1992
Re-evaluation of silicic age rates 10 - 0 Ma 17 - 10 Ma	This paper
Paleomagnetic rotations in the northern Basin and Range	Magill and Cox, 1981
Estimated extension rate from dike intrusion and inception of faulting	Anders and others, 1989
Evaluation of extension inception in basins adjacent to the southern SRP	Rodgers and others, 1990
Inception of late Cenozoic normal faulting adjacent to southern boundary of SRP	Pierce and Morgan, 1992
Estimated extension rate from Quaternary faulting, NW side of SRP	Knuepfer and others, 1990
Estimated extension rate from dike intrusion and inception of faulting	Zoback and others, 1981
Estimated from distribution of normal faults along and adjacent to the YSRP	This paper

subtracting the age-progression rate determined from the ages of silicic volcanic rocks (which we assume reflects the combined effects of extension and plate motion) from the modeled plate motion rates. As shown in Figure 5, this suggests, for the period 9 to 0 Ma, an extension rate of ~2 cm/yr. Distinguishing extension from plate motion is critical to analytical models of plate-plume interaction (discussed in later sections) as well as for differentiating magmatic-related deformation from crustal extension as inferred from geologic data.

Topographic expression of the YSRP

Perhaps the most striking surface feature of the YSRP is the systematic decrease in average elevation along the Snake River Plain from ~2,500 m at the Yellowstone Plateau, to an average elevation of ~1,200 m in southwestern Idaho about 800 km to the southwest (Figure 3). The SRP depression is about 80 km wide and generally corresponds to the extent of basaltic volcanism. However, we point out a ~15° change in direction at about 250 km from Yellowstone (Figure 3), which appears to be related to the large-scale plate reorientation at 8 to 10 Ma discussed in the previous section.

Arcuate topographic highs extend southwest from Yellowstone in two branches to form a "bow-wave" or parabolic-shaped pattern of high topography surrounding the

depression (Figure 3). The systematic topographic features of the SRP were first recognized by Myers and Hamilton (1964), who pointed out the arcuate pattern of surrounding earthquakes and suggested they were the result of general topographic collapse. Suppe and others (1975) hypothesized that the topographic doming of the Yellowstone Plateau was due to a mantle plume. The topographic rims surrounding the SRP were also discussed by Smith and others (1985) and Smith (1989) in their assessments of the seismotectonics of the YSRP and by Anders and others (1989, 1992) who ascribed the systematic parabolic-shaped topography of the YSRP to the Yellowstone hotspot. Pierce and Morgan (1990, 1992) also associated the topography with a crescent-shaped uplift, warping around Yellowstone, which they attributed to a mantle hotspot.

Lithospheric deformation associated with large-scale mantle processes generally has an observable

response in river drainage systems. To examine this possibility, we plotted the rivers and streams of the YSRP and surrounding region (Figure 6). The resulting pattern reveals the high topographic divides surrounding the SRP, which separate the Snake River and Salmon River drainages to the northwest and the Great Basin and ancestral Snake River drainages to the southeast. The highest topography is associated with the topographic divides at 1,500 to 2,000 m in elevation (also shown in Figure 3). The bordering topographic shoulders extend southwestward from Yellowstone for 400 km, wrapping around the SRP in the characteristic parabolic pattern. The width between the drainage divides increases to ~200 km, 300 km southwest of Yellowstone.

We also included on Figure 6 the locations of the larger, $M_s > 5.7$, earthquakes of the region, which demonstrate an important observation, namely that the background seismicity is generally located in the

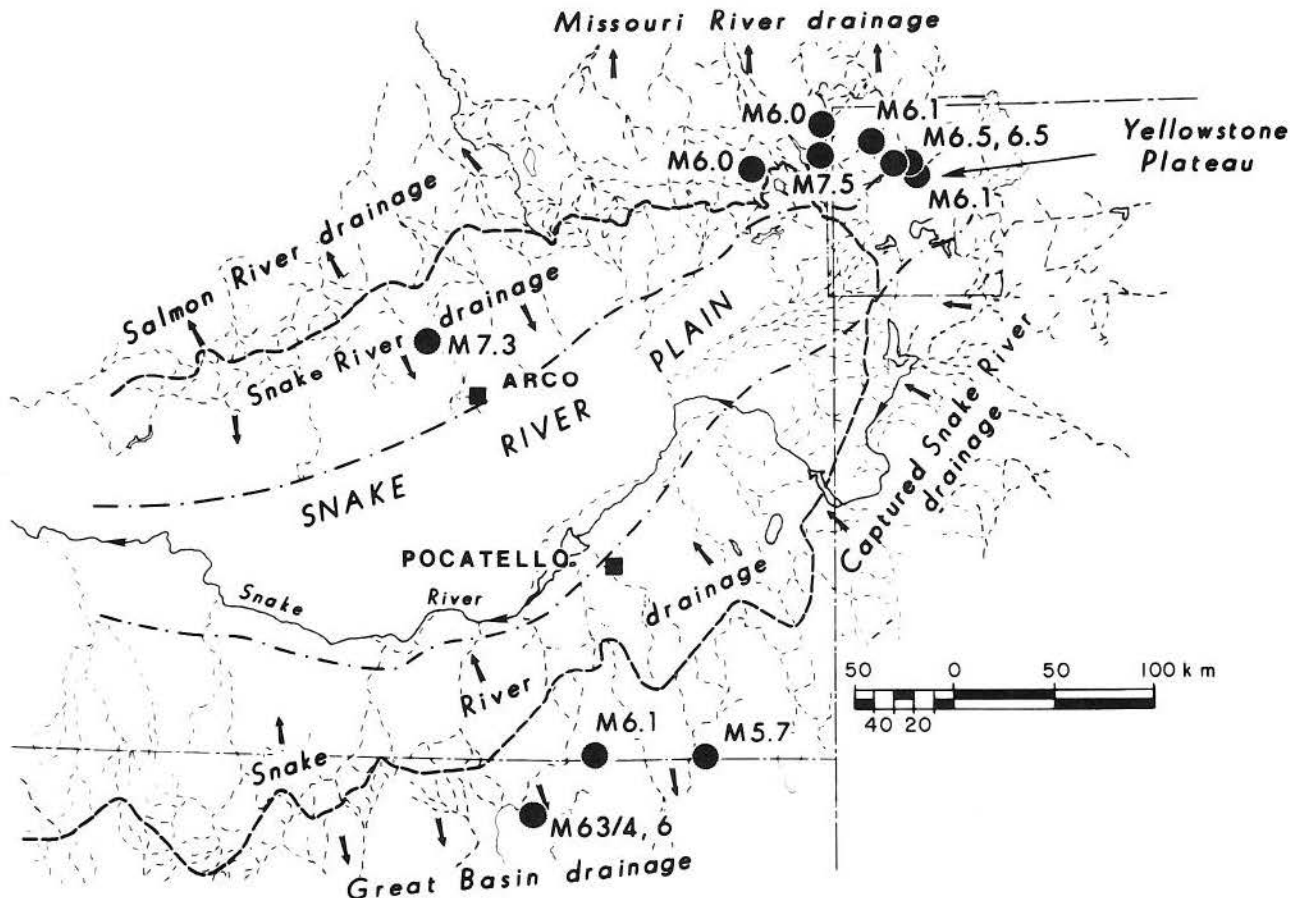


Figure 6. Drainage patterns of the upper Snake River basin and surrounding regions of the YSRP. Light dashed lines correspond to rivers and streams, bold dashed lines are the main drainage divides. Epicenters of the larger ($M_s \geq 5.7$) historic earthquakes from Table 3 are shown.

areas of the SRP drainage divides (except for the 1983 Borah Peak nucleation point, which is 25 km southeast, but its surface rupture extends northwest beyond the drainage divide). This correlation suggests a common mechanism for the elevated topography and seismicity (Smith and others, 1985).

The residual topography of the Yellowstone Plateau and the Snake River Plain (Figure 7) also reveals anomalous properties. For Yellowstone, the topographic residual was determined by radially averaging digital elevation data out to distances of 400 km. We selected the Yellowstone data in two azimuthal windows: (1) to the northeast and (2) to the southwest. This reveals a distinct ~600 km wide topographic anomaly of up to 600 m in height that we suggest reflects the region now affected by the Yellowstone hotspot (Figure 7). We also note that the

area of the high topography is more than ten times larger than the ~2,500 km² Yellowstone caldera and encompasses the nearby Hebgen Lake fault zone to the west, the Teton Range to the south, and the Absaroka and Beartooth ranges to the east—the general definition of the Yellowstone Plateau.

Subsidence of the SRP is revealed by a 300 km long, residual topographic profile (Figure 7). The SRP topographic residual anomaly, produced by removing a long-wavelength linear trend from the digital topographic profile across the SRP, exhibits a pronounced 200 - 300 m depression surrounded by topographically high shoulders with relief in excess of 400 m. The surrounding uplift shoulder and central subsidence anomaly have a dimension of ~650 km, about the same width as the regional topographic uplift anomaly of the Yellowstone Plateau.

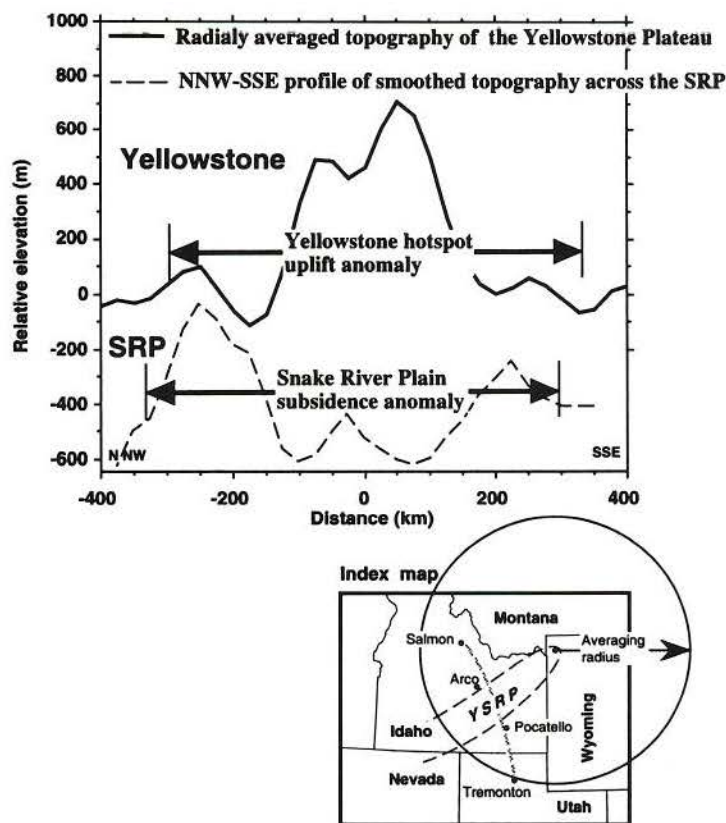


Figure 7. Topographic anomalies of the Yellowstone Plateau and the Snake River Plain affected by the Yellowstone hotspot. The Yellowstone topographic anomaly was determined by radially averaging (smoothing) the digitized topography in Figure 3 out to distances of 400 km. The Snake River Plain residual topography was computed by removing a linear regional trend and was plotted along a general north-northwest-trending profile from Salmon, Idaho (north of the SRP), across the SRP (zero distance), to Tremonton, Utah (south of the SRP).

Comparison of YSRP topography with oceanic mantle hotspots

The long-wavelength topographic response of the Yellowstone hotspot (Figure 7) can be compared with those of oceanic hotspots such as the Cape Verde, Galapagos, Azores, and Hawaiian islands (Crough, 1983). Topographic highs associated with these features range from 1 to 2 km above the abyssal plain and their widths range from 700 km to 1,500 km (Table 2). We caution that there may not be a direct one-to-one comparison between the topography of Yellowstone and the oceanic features because of differing thicknesses and compositions of oceanic versus continental crust, age of crust, and thermal regimes. Nonetheless, the 600 km wide topographic anomaly of Yellowstone (Figure 7) (and the ~1000 km wide geoid anomaly to be discussed in a later section) is similar to those associated with oceanic swells and hotspots and provides corroborative evidence for a mantle hotspot origin of Yellowstone.

We note the striking similarity between the anomalous Yellowstone topography (Figure 7) and the excess lithospheric topography modeled by Hill and others (1992) for a rising mantle plume interacting with a stationary continental lithosphere,

i.e., without shearing the top of the plume near the base of the overriding plate. They suggested up to 500 m of excess topographic uplift across 1,000 km wide zones and describe the YSRP as a "plume tail province." This width compares with that of Yellowstone; however, we assume that there is a significant component of shearing of mantle material accompanying the plume-plate interaction of the North American plate, plus a component of concomitant plate extension. We discuss our plume-plate model for Yellowstone in a later section.

Patterns of late Cenozoic faulting

The style of late Cenozoic normal faulting of the YSRP and its related deformational patterns (Figure 3) is discussed using a compilation of late Tertiary to Holocene normal faults taken from Smith and Arabasz (1991). The systematic pattern of late Cenozoic faulting paralleling the YSRP has been noted by Smith and others (1985), Scott and others (1985), and Pierce and Morgan (1990, 1992) as evidence related to the track of the Yellowstone hotspot. Here we differentiate areas of faulting of similar age and seismicity, which we designate as Zones II and III (Figures 3 and 4). Zone I corresponds to the generally undeformed and seismically quiescent SRP volcanic field. The areas of similar-aged faulting are differentiated on the basis of recency of slip, not on direction of fault motion, and are generally oblique to orthogonal to individual faults trends. Zones encompassing faults of common age are generally parallel to the boundaries of the SRP (Figures 3 and 4).

Zone II extends outward ~80 km from the boundaries of the SRP and is characterized by faults whose most recent displacements are generally older than post-glacial, i.e., generally greater than ~14,000 ^{14}C years (marked by thin lines in Figures 3 and 4).

Zone II was suggested by Smith and others (1985) as possibly related to a thermal shoulder of the Yellowstone hotspot. Zone III encompasses the active earthquake zones surrounding the SRP and appears to have been much more active throughout Holocene and historic time than Zone II. Faults that have ruptured in historic time are generally within Zone III and are indicated by bold red lines on Figures 3 and 4. Although Pierce and Morgan (1990, 1992) subdivided the region of faulting into six divisions based upon age, height of adjacent triangular facets, etc., we could not distinguish the finer divisions of fault activation that they proposed.

Regional seismicity

In this section, we summarize the seismicity of the YSRP and surrounding region (Figures 4, 6 and 8). We also point out important observations from the Intermountain Seismic Belt (ISB) that are relevant to the patterns of earthquakes associated with the Yellowstone hotspot. Our seismicity data

Table 2. Topographic and geoid characteristics of the Yellowstone-Snake River Plain volcanic system.

Long Wave Length Topography	Width (km)	Height/ Subsidence (km)
Yellowstone Plateau		
Radially averaged topography (Figure 6)	600 km	~650 m
Snake River Plain		
Width of volcanic province, Zone I, 300 km southwest of Yellowstone (Figure 3)	~ 80 km	- 200 m ¹
Width of Zones I and II, 300 km south west of Yellowstone (Figure 3)	250 km	-
Maximum width; Zones I, II and III 500 km southwest of Yellowstone (Figure 3)	600 km	-
Oceanic Swells ²		
Hawaiian	1100 km	2 km
Cape Verde Rise	1500 km	2 km
Azores	1500 km	2 km
Galapagos	900 km	1 km
Yellowstone Hotspot Geoid Anomaly		
GEOID90 model ³ (Figure 7)	~1,000 km	+10 to +12 m anomaly

¹At 300 km distance from Yellowstone. The Snake River Plain has subsided an additional 200 m with respect to the averaged topography of the Yellowstone Plateau.

²From Crough (1983).

³From Milbert (1991).

are taken from regional seismograph network compilations summarized by Smith and Arabasz (1991). They noted that the networks in the Intermountain region consist generally of widely spaced seismic stations, averaging 30 to 50 km separation. This spacing precludes accurate focal depth determinations, except in the case of Yellowstone, where focused monitoring, using a dense array of seismographs with about 15 km spacing, provides sufficiently accurate focal depths to correlate with structure.

The coherence of the dual-branched pattern of seismicity surrounding the YSRP (Zone III in Figures 3, 4 and 8) has become apparent with evolving compilations of seismicity of the Intermountain region, starting with Smith and Sbar (1974). The regional continuity of the ISB led Smith and Arabasz (1991) to distinguish three parts referred to as the southern, central, and northern ISB (Figure 8). The central region, which we believe has been affected by the Yellowstone hotspot, extends from latitude 42° 15' to 45° 25' N and, in part, follows the Basin and Range-Middle Rocky Mountains transition. However, this zone is complicated by a west-trending branch defining an arc that "wraps around" the eastern Snake River Plain (Smith and others, 1985; Anders and others, 1989).

Thirty moderate to large earthquakes ($5.5 < M_s < 7.5$) since 1900 (Table 3) characterize the historical seismicity of the YSRP. This area is also distinguished by the occurrence of the two largest, scarp-forming earthquakes in the Intermountain region in historical time: (1) the $M_s = 7.5$, 1959 Hebgen Lake, Montana, and (2) the $M_s = 7.3$, 1983 Borah Peak, Idaho, earthquakes (also see Figure 6). In contrast, the SRP is seismically quiescent at the magni-

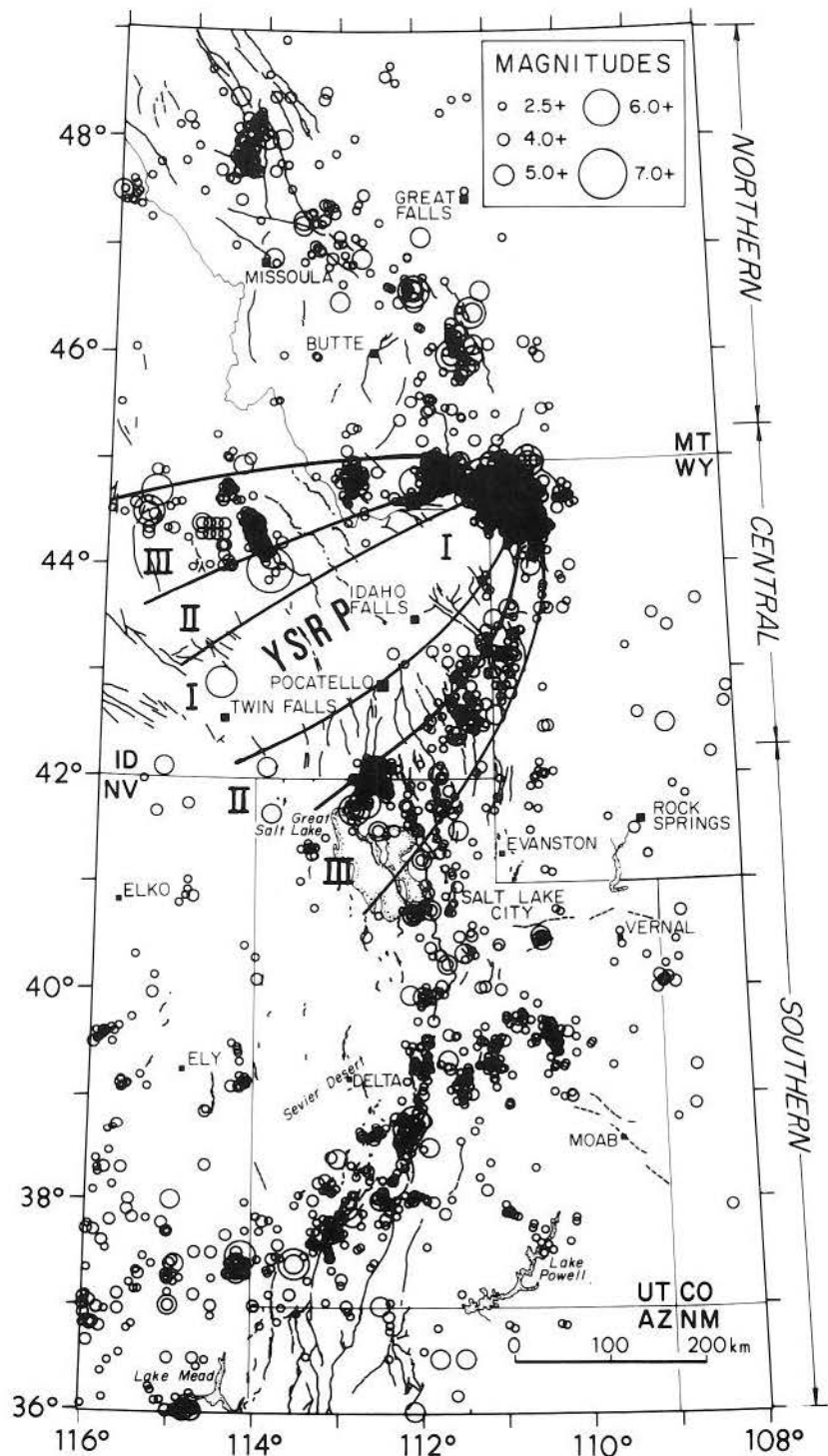


Figure 8. Earthquake epicenters of the Intermountain region, 1900-1985, outlining the Intermountain seismic belt (ISB) and the YSRP volcanic system (epicenters taken from Smith and Arabasz, 1991). The fault data were taken from published information on late Tertiary to Holocene normal faults in the Intermountain region, summarized by Smith and Arabasz (1991). Quaternary faults in Yellowstone taken from maps by Christiansen (1984, 1993). Roman numerals correspond to the seismotectonic domains of Figures 3 and 4.

tude 2+ level (Smith and Arabasz, 1991; S. M. Jackson and others, unpublished data).

Beginning at the southern margin of the YSRP (Figures 6 and 8), earthquakes near the Utah-Idaho border occur near a change in direction between the southern and central parts of the ISB (Smith and Arabasz, 1991). North of the Utah border there is a marked discordance between the northeast-trending seismicity belt and northwest-trending zones of late Cenozoic normal faulting that we believe reflects the influence of the Yellowstone hotspot. Seismicity studies by Piety and others (1986) of the region between southeastern Idaho and western Wyoming

showed little correlation between earthquakes and Quaternary faults. Seismicity of the Idaho-Wyoming border area, encompassing the Teton region, reveals a scattered northeast trend, approximately paralleling the SRP, that continues into the southern Yellowstone Plateau but with a notable gap in historical seismicity along the Teton fault (Figure 8).

Seismicity continues north across about a 100 km wide, north-trending zone, intersecting the caldera in the vicinity of Yellowstone Lake. Within Yellowstone, widespread earthquake swarms characterize the caldera with a distinct change in regional trend from north-south at the southern Yellowstone Pla-

Table 3. Thirty earthquakes in the Yellowstone-Snake River Plain region of magnitude 5.5 and greater, 1900 through 1985 (area defined from Lat. 41°N to 46°N, Long. 110°W. to 116°W). Earthquakes accompanied by surface faulting have their origin time and location in bold print. Table summarized from Smith and Arabasz (1991).

No.	Date (GMT)		Time ¹ (GMT) hr mn	Lat. (°N)	Long. (°W)	Magnitude ²		Region
						M _{conv}	M _w	
1.	1905	Nov 11	21:26	42.9	114.5	(5.5 ±)	—	South central Idaho (Shoshone)
2.	1909	Oct 06	02:50	41.8	112.7	(6 ±)	—	NW Utah (Hansel Valley)
3.	1914	May 13	17:15	41.2	112.0	(5.5 ±)	—	Ogden, Utah
4.	1925	Jun 28	01:21	46.00	111.50	6.75	6.6	Clarkston Valley, Mont.
5.	1930	Jun 12	09:15	42.6	111.0	5.8	—	Grover, Wyo.
6.	1934	Mar 12	15:05	41.77	112.67	6.6	6.6	Hansel Valley, Utah
7.	1934	Mar 12	18:20	41.57	112.75	6	5.9	Hansel Valley, Utah, aftershock
8.	1934	Apr 07	02:16	41.5	111.5 ?	5.5	—	Hansel Valley, Utah, aftershock?
9.	1934	Apr 14	21:26	41.73	112.60	5.6	—	Hansel Valley, Utah, aftershock
10.	1934	May 06	08:09	41.96	112.82	5.5	—	Hansel Valley, Utah, aftershock
11.	1944	Jul 12	19:30	44.41	115.06	6.1	—	Central Idaho (Seafoam)
12.	1945	Feb 14	03:01	44.61	115.09	6.0	—	Central Idaho (Clayton ?)
13.	1947	Nov 23	09:46	44.92	111.53	6.25	6.1	Virginia City, Mont.
14.	1959	Aug 18	06:37	44.88	111.11	6.3	6.3	Hebgen Lake, Mont., double event
15.	1959	Aug 18	06:37	44.84	111.03	7.5	7.3	Hebgen Lake, Mont., double event
16.	1959	Aug 18	07:56	45.00	110.70	6.5	—	Hebgen Lake, Mont., aftershock
17.	1959	Aug 18	08:41	45.08	111.80	6	—	Hebgen Lake, Mont., aftershock
18.	1959	Aug 18	11:03	44.94	111.80	5.5 +	—	Hebgen Lake, Mont., aftershock
19.	1959	Aug 18	15:26	44.85	110.70	6.3	6.3	Hebgen Lake, Mont., aftershock
20.	1959	Aug 19	04:04	44.76	111.62	6	6.0	Hebgen Lake, Mont., aftershock
21.	1962	Aug 30	13:35	41.92	111.63	5.7	5.6	Cache Valley (Logan), Utah
22.	1964	Oct 21	07:38	44.86	111.60	5.8	5.6	Hebgen Lake, Mont., aftershock
23.	1975	Mar 28	02:31	42.06	112.53	6.0	6.2	Pocatello Valley (Idaho-Utah border)
24.	1975	Jun 30	18:54	44.69	110.62	6.1	—	Yellowstone Park, Wyo.
25.	1976	Dec 08	14:40	44.76	110.80	5.5	—	Yellowstone Park, Wyo., aftershock
26.	1983	Oct 28	14:06	43.97	113.92	7.3	6.9	Borah Peak, Idaho
27.	1983	Oct 28	19:51	44.05	113.92	5.8	5.4	Borah Peak, Idaho, aftershock
28.	1983	Oct 29	23:29	44.24	114.06	5.8	5.5	Borah Peak, Idaho, aftershock
29.	1983	Oct 29	23:39	44.24	114.11	5.5	—	Borah Peak, Idaho, aftershock
30.	1984	Aug 22	09:46	44.37	114.06	5.8	5.6	Borah Peak, Idaho, aftershock

¹The local time (Mountain Standard Time) for the earthquakes in this table is found by subtracting seven hours from Greenwich Mean Time. In some cases (War Time, Daylight Savings Time), the difference is six hours. Non-instrumental and instrumental earthquake locations are listed with one- and two-decimal-point accuracy, respectively.

²Abbreviations for earthquake magnitude: M_{conv} = conventional magnitude, M_L (local magnitude), M_S (surface-wave magnitude), and ^mb (body-wave magnitude); M_w = moment magnitude.

teau to northwest on the northwest margin of the caldera.

West of the Yellowstone caldera, a pronounced east-west band of earthquakes passes through the Hebgen Lake, Montana region, including the epicenter of the $M_s = 7.5$, 1959 Hebgen Lake earthquake that occurred about 30 km west of the caldera. This earthquake produced a 26 km long, west- to northwest-striking rupture with a maximum vertical displacement variously cited as 5.5 to 6.7 m (see Smith and Arabasz, 1991). Earthquakes in this area are part of the divergent belt of seismicity that extends more than 400 km along the north side of the SRP in a west-southwest direction into central Idaho. Smith and Sbar (1974) characterized this as an independent seismic zone and Stickney and Bartholomew (1987) called it the "Centennial tectonic belt". However, Smith and others (1985) included the area within a "V-shaped" pattern of seismicity surrounding the YSRP and suggested that this zone of seismicity was related to the Yellowstone hotspot.

Along the northwest side of the SRP, **Figures 4 and 8** reveal a northwest alignment of epicenters that corresponds to the mainshock and aftershocks of the $M_s = 7.3$, 1983 Borah Peak, Idaho, earthquake. This large scarp-forming earthquake (**Table 3**) was accompanied by a 34 km long fault with up to 2.7 m of coseismic vertical slip. It occurred 60 km northwest of the Snake River Plain in an area characterized by late Quaternary Basin and Range faulting (Scott and others, 1985) and widespread hot springs, but low background seismicity.

We note the relative seismic quiescence of the SRP and the abrupt increase in earthquake activity about 50 km to 100 km from its margins (see **Figures 4 and 8**). The low seismicity of the SRP with pronounced seismically active "shoulders" is enigmatic, and led Smith and others (1985), Anders and others (1989), and Anders and Sleep (1992) to suggest thermal and magmatic mechanisms associated with the Yellowstone hotspot. Parsons and Thompson (1991) postulated that magmatic overpressuring in the SRP reduced the normal stresses on incipient fractures, thereby precluding earthquakes. However, their mechanism does not account for the lack of seismicity in the adjacent aseismic area, Zone II, which extends up to 100 km beyond the boundary of the SRP volcanic field and which shows little evidence of volcanic features or high heat flow. Nonetheless, if the surrounding belt of seismicity, Zone III,

is a signature of the regional stress field associated with the Yellowstone hotspot, then its effects on stress extends 100 to 200 km beyond the boundary of the SRP volcanic field.

Geoid anomaly

Just as the Earth's topographic field responds to crustal sources, the long-wavelength gravity field generally reflects deeper, lithospheric sources. The isostatic properties of the YSRP can be seen by examining the geoid model, GEOID90, from Milbert (1991) for North America (**Figure 9**). The geoid is considered to represent the degree of isostatic compensation and is an equipotential surface reflecting an amalgam of topographic relief and density variation. Most of the local geoid features are due to topographic variation. Density variations form a secondary, but important source of the geoid anomaly.

The most significant geoid anomaly of the entire North American continent is a very broad (about 1,000 km wide +10 to +12 m) circular geoid high centered at Yellowstone (**Figure 9**). We suggest that the major component of the Yellowstone geoid anomaly is the approximately 600 m high excess topography (**Figure 6**) of the Yellowstone Plateau, with a superimposed long-wavelength geoid signal produced by a large upper-mantle low-density body associated with the mantle hotspot. The radius of this anomaly (marked by the outer dashed line in **Figure 9**) is similar to the radii of geoid anomalies associated with oceanic swells and rises which range from ~800 to 1,500 km and average 10 to 15 m in amplitude (Crough, 1983).

The 1,000 km wide Yellowstone geoid anomaly, however, is not a direct indicator of the hotspot dimensions. The gravitational signal has been upward-continued to the Earth's surface from 100 to 200 km depths, thus indicating a smaller dimension at the source. We suggest that the geoid anomaly reflects the effects of buoyancy and lateral flow of the upper mantle, spreading out from a plate-plume interaction at Yellowstone with source dimensions of ~150 km. This model will be discussed in detail in a later section. The affect of the overriding plate across a rising heat source is also indicated in the lateral deflections of the +8 m to +12 m contours, streaming to the southwest adjacent to the SRP. This pattern is similar to the parabolic-shaped patterns of topography and seismicity surrounding the YSRP (**Figures 3 and 4**).

Plate tectonic framework of the YSRP

The significance of the YSRP within the plate framework of North America can be examined by plotting the location of the Yellowstone hotspot and several of the major tectonic features of the western

U. S. at systematically older ages of 0, 5, 10, and 15 Ma (Figure 10), corresponding to the 16 million year evolution of the YSRP. These figures were produced by assuming a rigid-body framework for tectonic sub-provinces of western North America (for example, the Sierra Nevada, Great Valley, Cascade

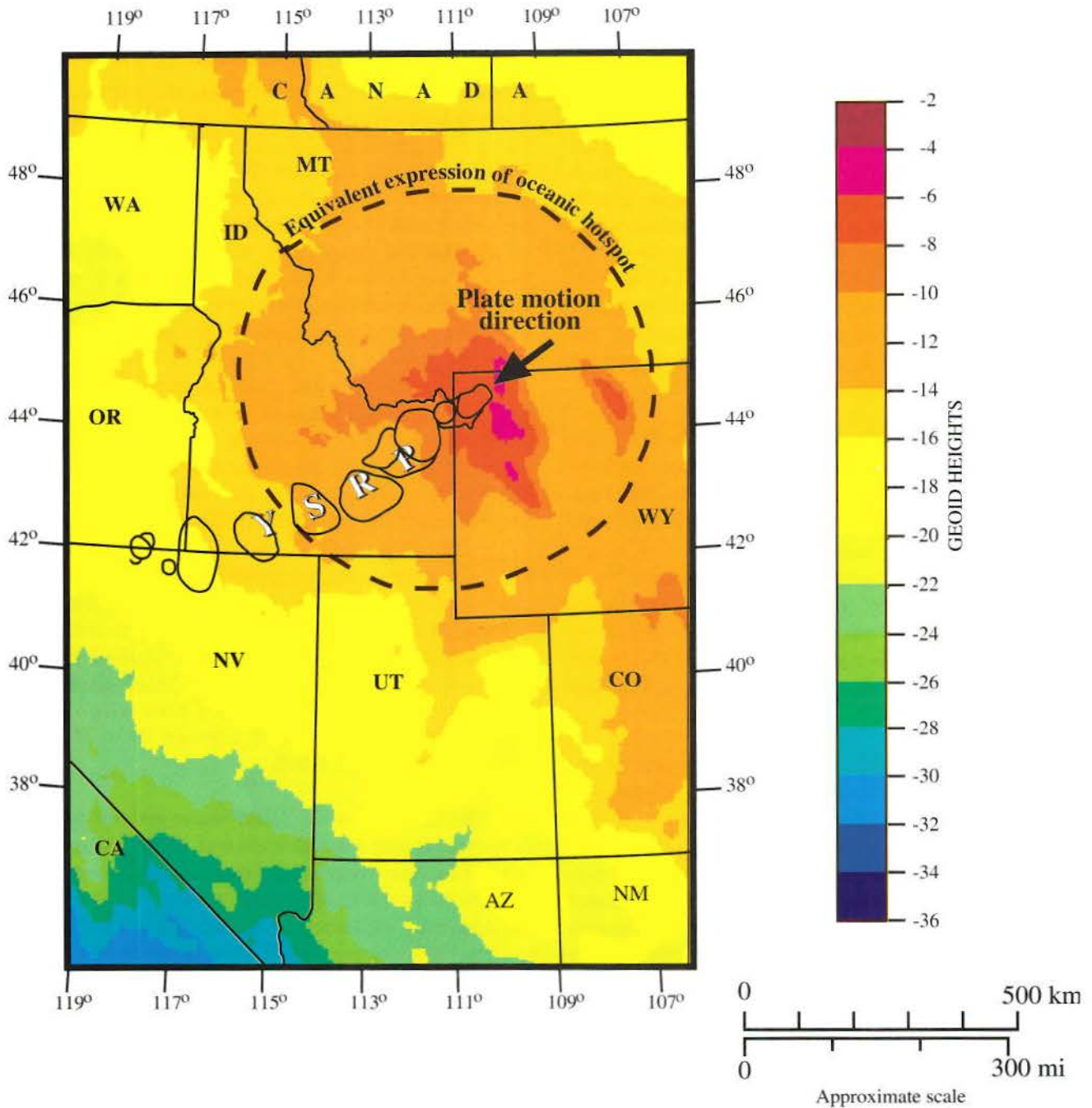


Figure 9. Map of geoid heights for the western United States from the GEOID90 model of Milbert (1991). Geoid heights are contoured at 2 m intervals. YSRP calderas are shown for reference. Dashed line marks the outline of geoid height anomaly of approximately 10 m, which is similar to geoid anomalies for oceanic hotspots and is considered a combined topographic-density signature of the Yellowstone hotspot.

Topographic signature, space-time evolution, and physical properties of the Yellowstone-Snake River Plain volcanic system: the Yellowstone hotspot

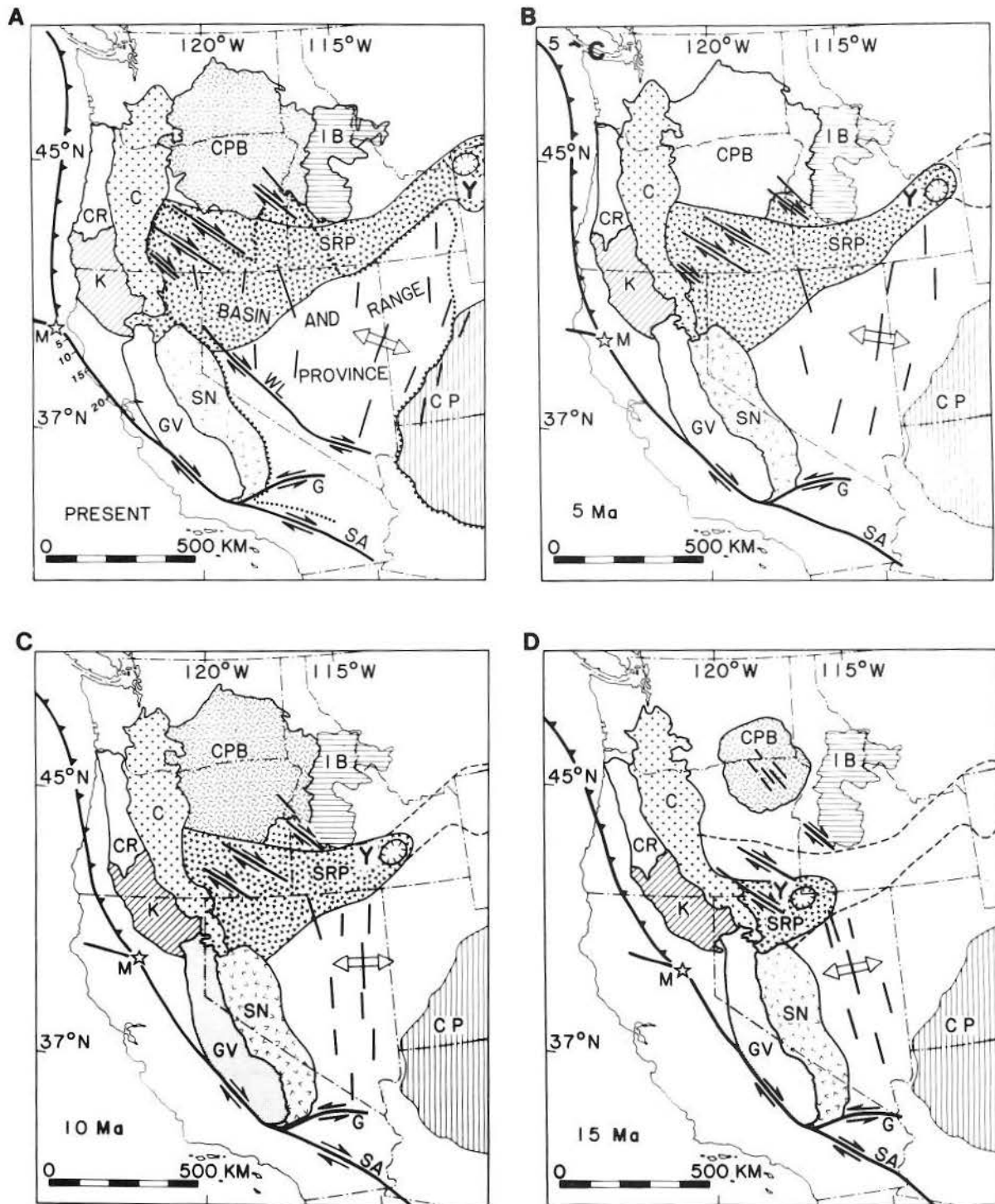


Figure 10. Maps showing plate reconstructions and the related tectonic framework of the western United States. Reconstructions were made at 5 Ma, 10 Ma, and 15 Ma taking into account the relative motion of the North American plate of ~ 2.5 cm/yr southwest and a regional east-west extension rate of the Basin and Range province of ~ 1.7 cm/yr (from Magill and Cox, 1981). The bimodal rhyolitic-basaltic volcanic rocks of the YSRP are shown as the same pattern as those of the High Lava Plains of southern Oregon. SRP = Snake River Plain, Y = Yellowstone hotspot, M = Mendocino triple junction, CP = Colorado Plateau, SA = San Andreas fault, GV = Great Valley, G = Garlock fault, SN = Sierra Nevada, K = Klamath Mountains, C = Cascade Range, CR = Coast Ranges, CPB = Columbia Plateau basalt field, IB = Idaho batholith.

Range, etc.) with a superimposed east-west extension rate of ~ 1.7 cm/yr and for the northern Basin and Range given by Magill and Cox (1981). We then projected the North American plate motion backwards in time at ~ 2.5 cm/yr in a southwest direction, with respect to the location of the Yellowstone hotspot, subtracting the component of east-west regional extension accompanying Basin and Range evolution.

The primary features of the intraplate reconstructions are as follows (Figure 10):

0 Ma—The Yellowstone hotspot is at its present location and the Basin and Range province is fully extended (100 % of its original width).

5 Ma—The Yellowstone hotspot volcanic center was located approximately 75 km southwest of its current position, at Island Park, and the Basin and Range was extended approximately 70% of its original width.

10 Ma—The Yellowstone hotspot was located near the center of the eastern Snake River Plain near Pocatello, Idaho. The northern Basin and Range was about half of its present width, and the western North American plate boundary was approximately 200 km east of its current position.

15 Ma—This time is near the age of the oldest silicic volcanic rocks of the Snake River Plain, at about 16 Ma, and at a position about 700 km southwest of Yellowstone in southwestern Idaho and northern Nevada. This location is also where most authors place the inception of YSRP volcanism based upon the oldest YSRP silicic volcanic rocks. This was also near the time of initiation of widespread extension in the Basin and Range province at 17 to 20 Ma (Zoback and others, 1981).

We have also examined the hypothesis that the Yellowstone hotspot originated earlier by continuing the intraplate reconstructions to 20 and 23 Ma. At 20 Ma, the reconstruction would place the Yellowstone hotspot in northern California, where there is little evidence for volcanic rocks of this age. The Yellowstone hotspot may have passed beneath the Sierra Nevada crust, approximately 20 km thicker than Basin and Range crust (Coney and Harms, 1984), without producing surface volcanism. Presumably much more heat would be required to penetrate the thick and cool Sierra Nevada crust, and possibly this thicker crust was not heated sufficiently to bring magmatic material to the surface.

Speculating further, at about 23 Ma our model places the Yellowstone hotspot near the subducting margin of the west coast of North America and near the Mendocino triple junction (within the error of ± 100 km of our calculations). A location for the Yellowstone hotspot at approximately 50 Ma at the western plate margin was suggested by Engebretson and others (1985), who constructed models of relative motions between the North American and Pacific plates for the past 180 million years. Their estimate did not consider the extension of the Basin and Range province that in 16 million years could have increased its width by 100% and thus could also be in error by 100 km or more (Engebretson and others, 1985). Nonetheless, both of these models indicate the possibility of a hotspot location at or near the plate boundary in middle Tertiary time.

Despite the uncertainties in these reconstructions, they offer insight into the general space-time behavior of the YSRP volcanic system for at least the last 16 million years and support the origin of the YSRP as an interaction between the North American plate (moving southwest) and a fixed mantle magmatic source.

The Yellowstone hotspot today

As described in the previous section, the present location of the YSRP hotspot is thought to be beneath Yellowstone. In this section we discuss the volcanic history and physical properties of the crust and mantle of Yellowstone assessed primarily by geophysical and geodetic surveys. The resulting models

reveal an active volcanic system controlled by vigorous magmatic and heat sources originating in the upper mantle, which are manifested in the upper-crust by marked thermal, earthquake, gravity, and crustal deformation signals.

Volcanic history of Yellowstone

The most comprehensive treatment of Yellowstone's volcanic history is from Christiansen and Blank (1972), Christiansen (1984), and maps and data from Christiansen's (1993) forthcoming U. S. Geological Survey Professional Paper. Keefer (1971) summarized Yellowstone's complete geologic history since Precambrian time, including an overview of its volcanic history. A detailed review of Yellowstone's volcanic history is given by Christiansen (1984), and a recent account of the isotopic content of Yellowstone volcanic rocks is given by Hildreth and others (1991). Christiansen (1982) also discussed the volcanic history of the Henry's Fork transitional area between Yellowstone and the Snake River Plain. We also recommend summaries of the geology and geophysics of the Yellowstone volcanic system by Smith and Christiansen (1980) and a National Research Council report on the objectives for deep scientific drilling in Yellowstone National Park (Yellowstone National Park Task Group to the Continental Drilling Committee, 1987) for further reading.

The earliest scientific account of Yellowstone's volcanism and earthquakes was given by Hayden (1872, 1873), based upon his late 19th century, pioneering field expeditions into Yellowstone. Hayden recognized that the Yellowstone region was the center of a giant volcanic system and remarked on the occurrence of numerous small- to moderate-magnitude earthquakes:

While we were encamped on the northeast side of the lake [Yellowstone Lake] on the night of the 20th of July [1871], we experienced several severe shocks of an earthquake, and these were felt by two other parties, fifteen to twenty-five miles distant, on different sides of the lake. We were informed by mountain-men that these earthquake shocks are not uncommon, and at some seasons of the year very severe, and this fact is given by the Indians as the reason why they never visit that portion of the country. I have no doubt that if this part of the country should ever be settled and careful observations made, it will be found that earthquake shocks are of very common occurrence.

Hayden (1872)

So persistent were the tremors on Yellowstone Lake that Hayden named this camp "Earthquake Camp," an aptly named site, for we will show later in

this section that Yellowstone Lake has persisted as one of the most notable areas of earthquake swarms in the entire Yellowstone region.

Geologic studies by various U. S. Geological Survey and university scientists continued through the early 20th century to the present. Interest in the seismotectonics of the region was heightened by the occurrence of the 1959, $M_s = 7.5$, Hebgen Lake, Montana, earthquake, the largest earthquake in the Intermountain region in historic time. Following this earthquake, investigations by Hamilton (1959) distinguished the Yellowstone caldera from the nearby Island Park caldera and further resolved the differences in volcanism between Yellowstone and the Snake River Plain. However, it has been the detailed volcanological studies of Yellowstone by geologists of the U. S. Geological Survey that led to an elucidation of the cyclic nature and the immense scale of Pleistocene volcanism (e.g., see Christiansen and Blank, 1972; Christiansen, 1984, 1993; Hildreth and others, 1991).

The history of the Yellowstone volcanic system is dominated by three catastrophic, caldera-forming eruptions (Figure 11), which produced more than ~4,800 km³ of ash flows at 2.0 Ma, 1.2 Ma, and 0.6 Ma (Christiansen, 1984, 1993). The first cycle of Yellowstone's silicic volcanism, Huckleberry Ridge Tuff, occurred at 2.0 Ma (denoted as I in Figure 11) and produced a large caldera that is now mostly covered by younger volcanic flows and is thought to have had an eruptive volume of 2,500 km³. The second-cycle, Henry's Fork caldera (denoted as II in Figure 11) was restricted to the Island Park area, southwest of the Yellowstone Plateau, and produced 280 km³ of ash-flow tuff. Following the 0.6 Ma caldera-forming eruption of 1,000 km³ (denoted as III), drainage of a large magma chamber released roof support, causing it to collapse and produce the 45 km wide by 75 km long "Yellowstone caldera." In the last 150,000 years, there have been 40 or more silicic volcanic eruptions on the Yellowstone Plateau and several basaltic eruptions on the margin of the plateau, with the most recent magmatic activity at 70 Ka (Christiansen, 1993). The cumulative effects of post-caldera eruptions have produced the gently rolling and forested topography of the Yellowstone Plateau.

Following caldera collapse at 0.6 Ma, magmatic resurgence formed two structural domes: (1) the

Sour Creek dome, in the northeast caldera, began its uplift following the main eruption at 0.6 Ma; and (2) the Mallard Lake dome (Figure 12), in the southern caldera, was initiated at 0.15 Ma (Christiansen, 1984). Both of these domes are thought to have been produced by uplift above magmatic sources and are considered to be active (Christiansen, 1984, 1993). Following the explosive eruption of 0.6 Ma, the roof of the magma chamber was dropped several hundred meters, producing the subsided floor of the Yellowstone caldera that was later covered by post-caldera flows and sediments (Christiansen, 1993). Glacial moraines and alluvial fans are offset across northwest-striking normal faults along the crests of both domes, and the circular caldera rim-boundary fault has accommodated caldera-wide subsidence and extension.

On a regional scale, the Quaternary tectonic story of the Yellowstone Plateau is dominated by normal faulting and crustal extension associated with general Basin and Range epeirogeny. The main Quaternary faults of the Yellowstone Plateau are the Gallatin, Madison, and Teton fault zones of late Tertiary age that strike north-south at the northern and southern margins of the plateau (Figure 13). Buried extensions of these faults may exist in areas covered by the Quaternary silicic volcanic rocks.

The youngest mapped faults of the Yellowstone caldera (Figure 13) are Holocene in age and have several meters of offset (Meyer and Locke, 1986). These faults extend along the west side of the South Arm of Yellowstone Lake in an area of pronounced seis-

micity, to be discussed later. These faults are part of a regional right-stepping fault zone that extends south of the Yellowstone Plateau in an *en echelon* series of north-south striking faults to the Teton fault. On the northwest side of the Yellowstone Plateau, the seismically active Hebgen Lake fault zone extends about 30 km east-southeast toward the caldera boundary, but its relationship to the rim boundary fault is not known. The relationship of the north-south striking Gallatin fault, that extends south of Mammoth and terminates at the northwest caldera, is poorly defined because of limited access and extensive cover.

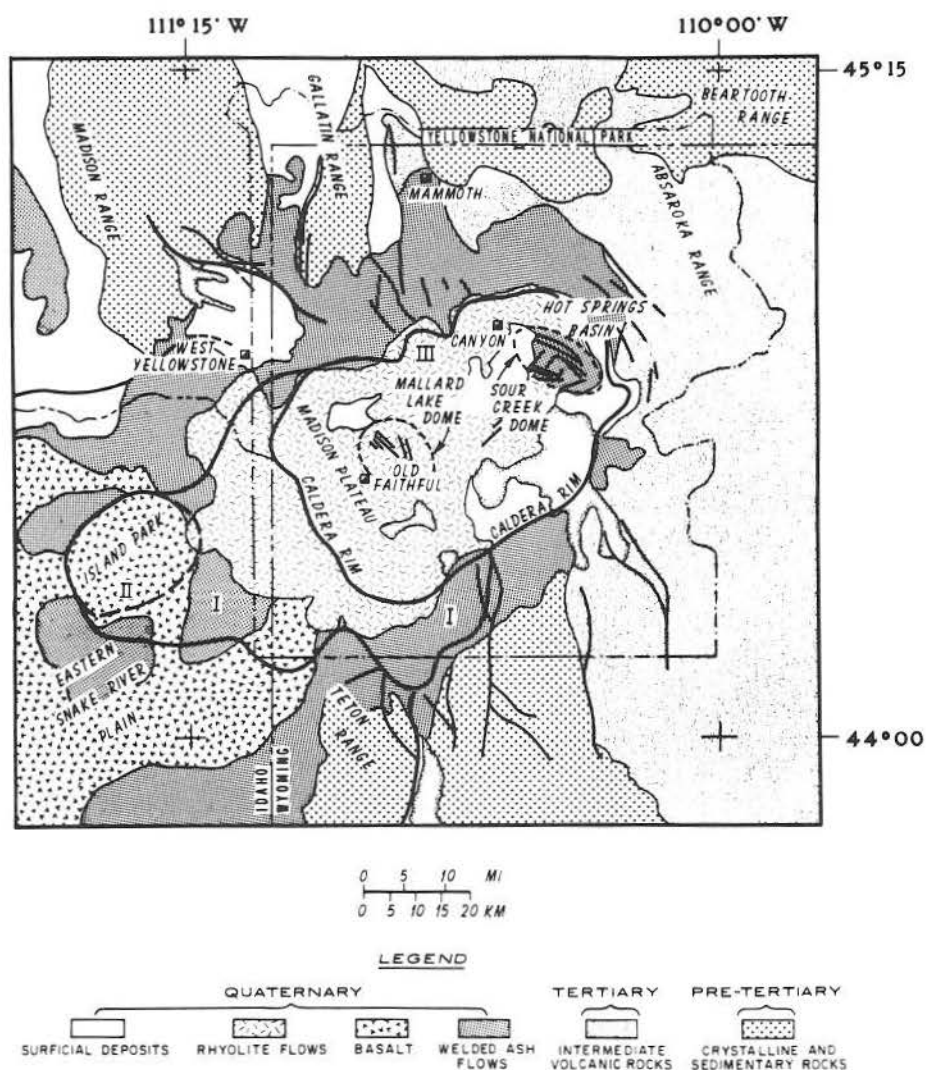


Figure 11. Geologic map of the Yellowstone Plateau showing the locations of Yellowstone calderas, Quaternary volcanic flows, and late Cenozoic normal faults (modified from Christiansen, 1984). Yellowstone calderas are shown by age: I = 2.0 Ma, II = 1.2 Ma, and III = 0.6 Ma.



Figure 12. Photograph of the Mallard Lake resurgent dome, southeastern Yellowstone caldera, looking northeastward across the Yellowstone Plateau and showing the approximate 400 m topographic dome and the northwest-striking central graben. The dome is highlighted by a fresh snow cover. Aerial photo taken in February, 1989, by Rick Hutchinson, National Park Service.

On Figure 13 we have also plotted the locations of volcanic vents that fed the post-collapse caldera flows. The vents form two distinct groups: (1) a northwest alignment in the southwest Yellowstone caldera, and (2) a similar north to northwest alignment in the northwest caldera that projects into a series of west- to northwest-striking Quaternary faults near Norris Junction. On the basis of contemporary seismicity, we believe that the vents southeast of Norris Junction are related to an active fault system that may continue southeast beneath the central caldera.

Global significance of Yellowstone's explosive volcanism

To portray the importance of Yellowstone's volcanism, we have plotted the eruptive volumes of several well-known volcanic eruptions of global significance along with those of Yellowstone (Figure 14). The smallest eruption in this comparison was in 1980 at Mt. St. Helens, Washington. While this dacitic volcano produced substantial destruction and ash that was deposited in areas of several surrounding states, its eruptive volume is thought to be only about 0.7 km^3 . The 1991 eruption of Mt. Pinatubo, The Philippines, killed more than 600 per-

sons and expelled 10 km^3 of ash. Another major volcanic eruption of North America was the 6.8 Ka Mazama volcano, which ejected 75 km^3 of ash and was responsible for producing the high mountain lake at Crater Lake National Park, Oregon. Another large historic eruption was that of Krakatoa, southwest Pacific, which expelled more than 18 km^3 tephra in 1883. The even larger Tambora, Indonesia, eruption of 1815 produced 150 km^3 tephra.

However, none of these eruptions comes even close to the smallest of Yellowstone's Pleistocene, caldera-forming, explosive eruptions. In Figure 14, Yellowstone's climactic eruptions (Figure 11) expelled ash flows with volumes of: (1) $2,500 \text{ km}^3$ as the 2.0 Ma Huckleberry Ridge Tuff, (2) 280 km^3 as the 1.2 Ma Mesa Falls Tuff, and (3) $1,000 \text{ km}^3$ as the 0.6 Ma Lava Creek Tuff (Christiansen, 1984, 1993). Continuous small-scale rhyolitic and basaltic eruptions between the climactic events contributed an additional $3,500 \text{ km}^3$ of volcanic material for a total of about $8,000 \text{ km}^3$. These volumes make Yellowstone one the world's largest, if not the largest, known center of active silicic volcanism. We suggest that the Pleistocene volcanism of Yellowstone likely compares in volume with the eruptions from the older late Tertiary calderas of the Snake River Plain volcanic system shown in Figure 3.

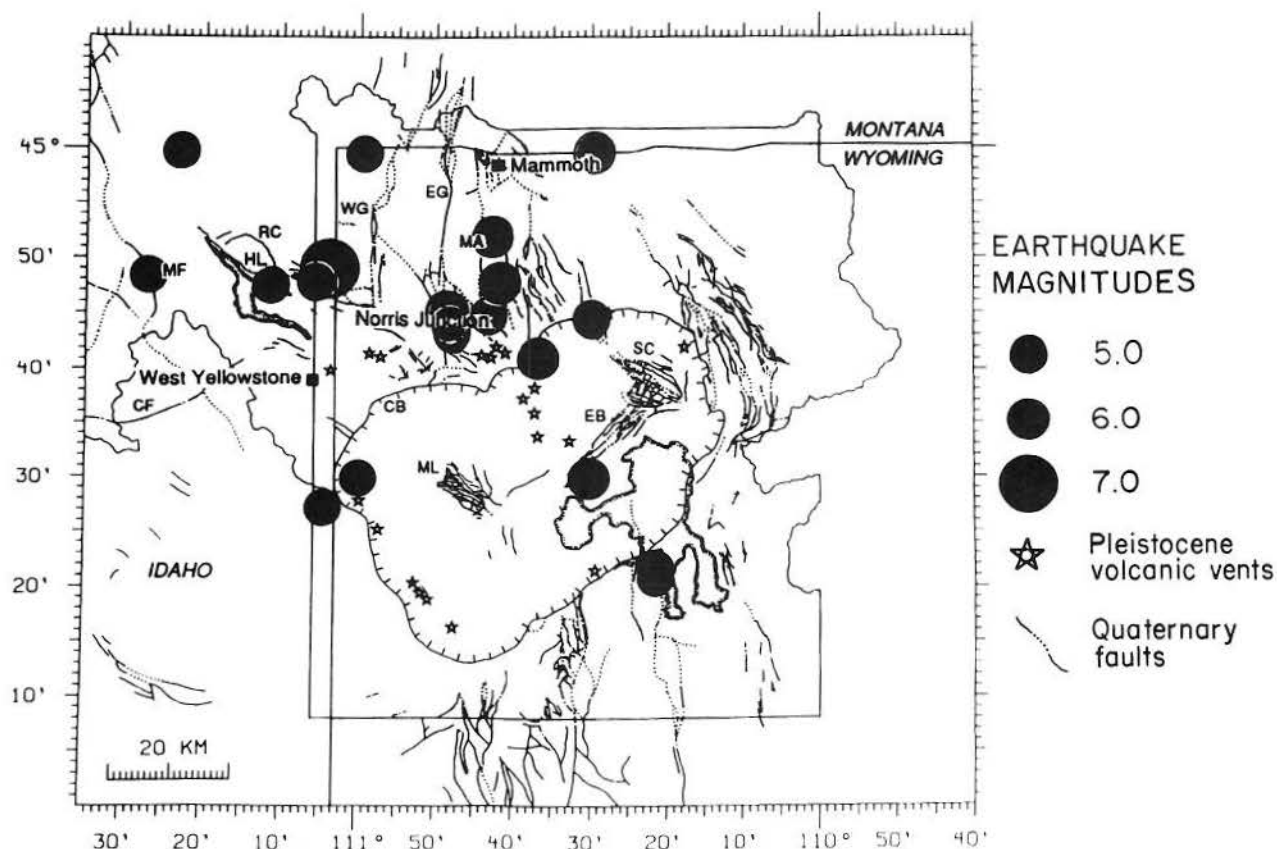


Figure 13. Map of Quaternary faults, Pleistocene volcanic vents, and larger, historic earthquakes ($M > 5.5$) of the Yellowstone Plateau. Volcanic vents are from U. S. Geological Survey 15-minute geologic quadrangle maps. Faults were digitized from the maps of Christiansen (1993). Fault zones are: RC = Red Canyon, HL = Hebgen Lake, WG = West Gallatin, EG = East Gallatin, CB = caldera boundary, EB = Elephant Back, and MA = Mammoth corridor. ML = Mallard Lake dome and SC = Sour Creek dome.

To further portray the scale of Yellowstone's silicic volcanism, we plotted the areas covered by ash deposits from four large eruptions in the western U. S. (Figure 15). These include: (1) the 0.74 Ma Bishop tuff with 500 km^3 of ash from the Long Valley caldera, California (Sarna-Wojcicki and Davis, 1993); (2) the 0.6 Ma Lava Creek Tuff of Yellowstone (Sarna-Wojcicki and Davis, 1993); (3) the 2 Ma Huckleberry Ridge Tuff of Yellowstone (Christiansen, 1984); and (4) the 1980 Mt. St. Helens volcano, Washington (Dzurisin and Newhall, 1988). The significance of these eruptions is evident by the immense areas of ash coverage, from thicknesses of a few

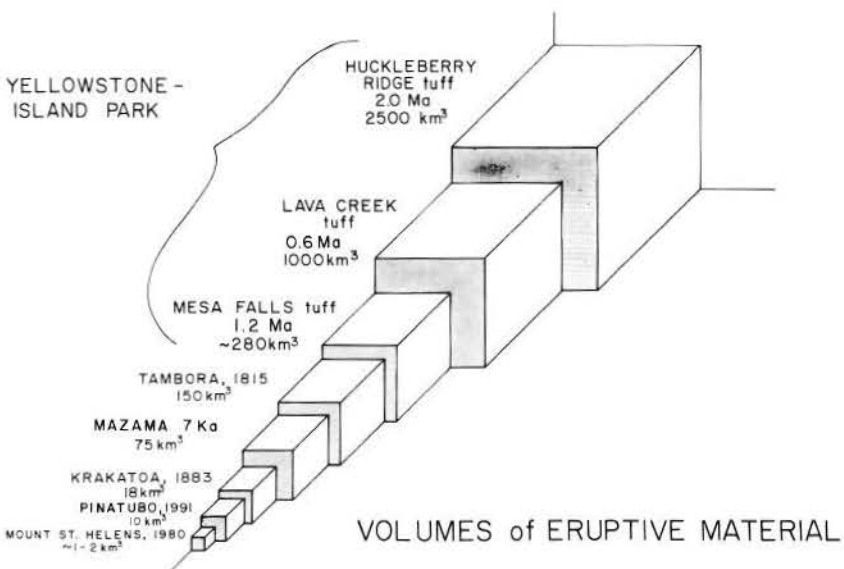


Figure 14. Volumes of Yellowstone's explosive volcanic eruptions compared with other well-known volcanic eruptions of the world. Data in part taken from Dzurisin and Newhall (1988) and Decker and Decker (1989).

millimeters to several meters, that blanketed much of western North America. For example, the Bishop tuff covered most of the southwestern U. S. and northern Mexico. Ash from Yellowstone's Huckleberry Ridge and Lava Creek tuffs, however, covered the entire western U. S., much of the Midwest, parts of the eastern Pacific, and northern Mexico. Compared to the ash fall coverage of Mt. St. Helens, which had such a dramatic affect on the landscape and the population of western Washington and Oregon, the scale of Yellowstone's prehistoric caldera-forming eruptions is immense. And while the Mt. Pinatubo and Mt. St. Helens eruptions had disastrous effects, they are very small compared to Yellowstone's past and potential for future catastrophic, caldera-forming eruptions.

Geophysical properties of Yellowstone

In this section we focus on the geophysical and geodetic properties of Yellowstone and show how heat flow, seismic velocity structure, seismicity, and crustal deformation are consistent with crustal heat flow from magmatic sources beneath the Yellowstone caldera. Previous geophysical summaries by Smith and others (1974, 1977), Eaton and others (1975), and Smith and Braile (1984) provide an interesting perspective on these issues that have changed with the acquisition of much new data since their publication.

Heat flow

There is a single physical property that sets Yellowstone apart from virtually all other continental regions, namely its extremely high heat flow of

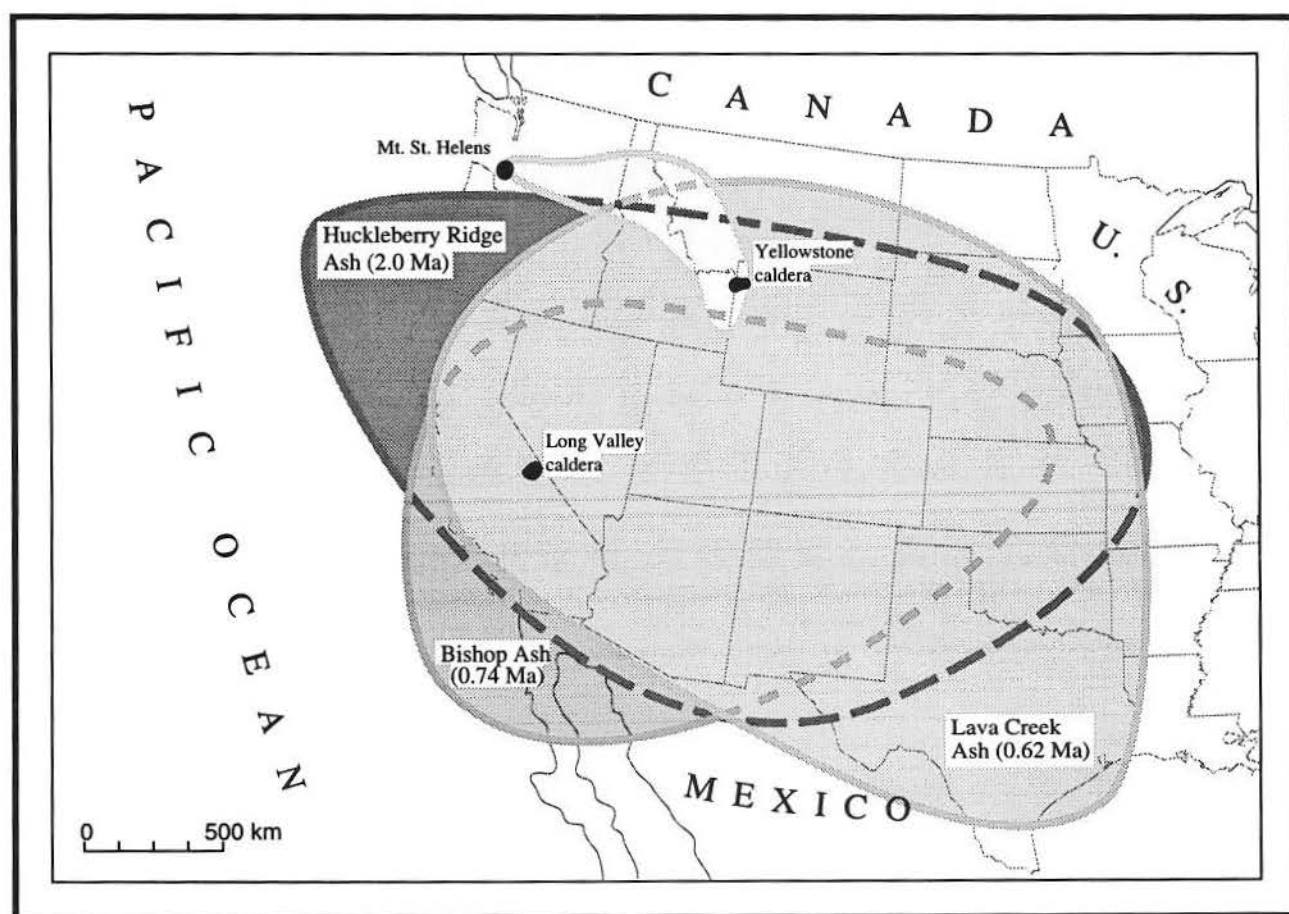


Figure 15. Map showing areas covered by airborne ash flows, tuffs, and pumices of four large Cenozoic volcanic eruptions of the western U. S.; the 0.6 Ma Yellowstone eruption, the 2.0 Ma Yellowstone eruption, the 0.7 Ma Long Valley caldera eruption, and the 1980 Mt. St. Helens eruption. Data from Sarna-Wojcicki and Davis (1993) and Sarna-Wojcicki and others (1982) for Mt. St. Helens.

~2,000 mWm⁻², which is about 30 times the continental average (Figure 16). Not only is heat an essential property of the youthful volcanism and hydrothermal activity of Yellowstone, we believe it drives such important mechanisms as thermal buoyancy, which affects the topography and the distribution of earthquakes of the YSRP system.

The coherence of a distinct thermal province associated with the YSRP became apparent with evolving compilations of heat-flow measurements (Blackwell, 1978; Lachenbruch and Sass, 1978; Blackwell and others, 1991). The regional heat flow of the Snake River Plain ranges from 40 to 100 mWm⁻² (Brott and others, 1981; Blackwell and others, 1991). In contrast, the heat flux north and east of Yellowstone in the Beartooth and Absaroka mountains is 50 to 80 mWm⁻² (Heasler and Hinckley, 1985). Thus there is a clear distinction between the high heat flow of Yellowstone (2,000 mW/m²) and those of the Basin and Range province (75 to 90 mWm⁻²), transitional values of ~100 mWm⁻² within the Snake River Plain, and much lower values of 50 - 80 mWm⁻² north and east of the YSRP.

Only a few borehole measurements of temperature have been made in the Yellowstone Plateau and those revealed temperatures exceeding 240°C and higher at depths less than 300 m (White and others, 1975; David Blackwell, personal communication, 1989). These measurements document a shallow, high-temperature regime but also reveal that conductive heat-flow measurements of Yellowstone are contaminated by pervasive ground-water circulation, primarily associated with the widespread hydrothermal systems.

The convective heat-flow component of Yellowstone, however, can be estimated using chloride content of hot water drained through hot springs and geysers and col-

lected in streams and rivers. Fournier (1989) and Fournier and Pitt (1985) estimated a total convective flux for the entire Yellowstone Plateau of 4×10^{16} cal/yr (using geochemical methods), which corresponds to a heat-flow value of about 2,000 mWm⁻². Unfortunately, the chemical method of heat-flow measurements cannot differentiate between individual heat sources, such as localized magma bodies, because the measurements were from rivers that collect chloride from large drainage basins of the Yellowstone Plateau.

The only conductive heat-flow measurements in the Yellowstone caldera were made by Morgan and others (1977) in Yellowstone Lake using marine-type thermal probes in shallow sub-bottom sediments (Figure 16). Their mapped values ranged from 125 to 300 mWm⁻² in the southern arms of Yellowstone Lake and rapidly increased northward across the caldera boundary, with an average value of 1,600 mWm⁻² for the main lake basin. This dramatic change in thermal flux is compelling evidence for a

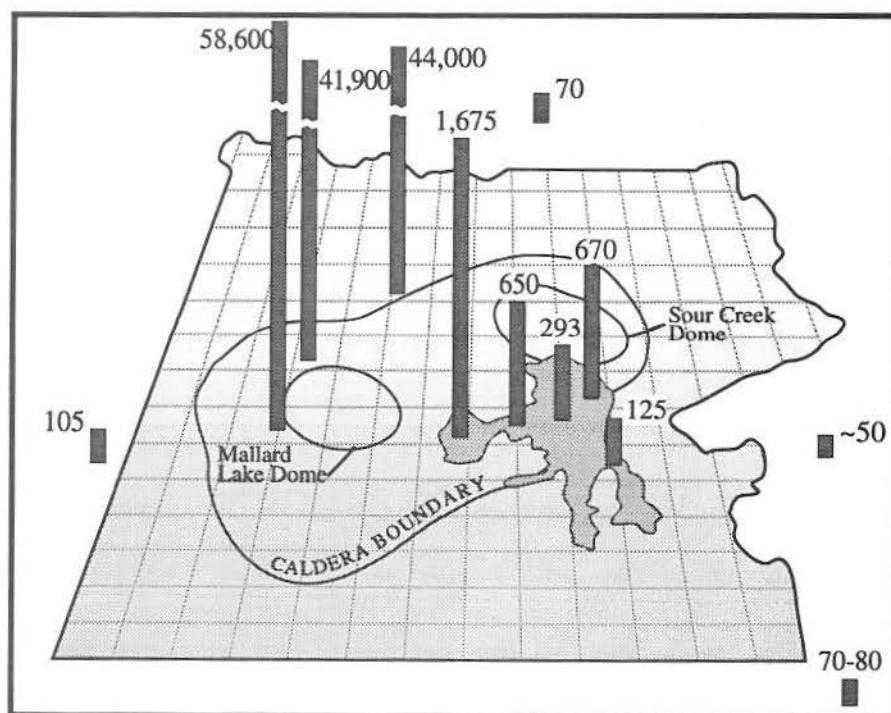


Figure 16. Heat-flow distribution of the Yellowstone Plateau (in units of mWm⁻²). Heat-flow values in Yellowstone Lake were determined by thermal probe measurements and are considered the most reliable indicators of the conductive heat flux of the Yellowstone caldera (Morgan and others, 1977). Heat-flow values for the western caldera were estimated from geochemical determinations by Fournier (1989) and mainly reflect the convective component of heat flow. Heat-flow values of the surrounding region are from Blackwell (1989) and Heasler and Hinckley (1985).

shallow heat source beneath the northern and western parts of Yellowstone Lake. Fournier (1989) showed that the $2,000 \text{ mWm}^{-2}$ heat flux is consistent with the release of heat by crystallization and cooling of about 0.1 km^3 of rhyolitic magma per year from 900°C to 500°C . We believe that this estimate is a reasonable value of magmatic recharge rate for Yellowstone and is therefore good evidence that magmas, Partial melts, and extensive hydrothermal systems are present beneath the caldera.

The significance of Yellowstone's heat flow on a global scale can also be seen by comparing its $2,000 \text{ mWm}^{-2}$ flux with the background heat flow of the North American continent of 40 to 60 mWm^{-2} (Table 4). The thermal energy release of Yellowstone is of the same order as such large volcanic features as the entire Cascade Range, the Columbia Plateau, and the ignimbrite fields of the Basin and Range province. Moreover, it is 2 to 3 times greater than that of the Valles, New Mexico, and Long Valley, California, calderas. Blackwell (1978) estimated that fully 5% of the total thermal energy of the entire U. S. Cordillera is released at Yellowstone, which occupies less than 0.1 % of its area.

Regional gravity field of the YSRP

The YSRP Bouguer gravity field mimics the regional patterns of topography (compare with Figure 3), i.e., gravity lows correspond to topographic highs, and vice versa, and consequently we have not included it in this paper. On large-scale compilations, the gravity signature of the YSRP is a pronounced arcuate positive gravity anomaly that systematically decreases northeastward along the SRP from values of -110 mGal in southwestern Idaho, to -190 mGal at the transitional Island Park area, then rapidly decreases to a regional -200 mGal value and a pronounced -60 mGal closed low at Yellowstone.

Interpretations of the SRP crustal structure were made from a 300 km-long northwest-southeast gravity profile (Figure 17) across the SRP in conjunction with seismic velocity determinations from the 1978 YSRP seismic refraction experiment (Sparlin and others, 1982). The SRP model revealed an 80 mGal positive gravity anomaly centered over the volcanic province that suggests significant density and compositional variations compared to the surrounding thermally undisturbed region. The combined SRP

Table 4. Comparison of continental heat production with that of Yellowstone. Data taken in part from Blackwell (1978), Dzurisin and Newhall (1988), and the Yellowstone National Park Task Group to the Continental Drilling Committee (1987).

Feature	Area (km^2)	Time span (Ma)	Total energy (cal/s)	Total heat flow (mWm^{-2})
General volcanism				
Cascade Range	17,500	20	4.1×10^9	20
Andes	--	10	--	6 - 10
Columbia Plateau	250,000	3	3×10^9	55
Basin-Range ignimbrites	180,000	20	2.6×10^8	6
San Juan caldera, Colorado	25,000	3	2.7×10^8	46
Magmatic intrusion-conduction				
Andes				~16
Yellowstone caldera	2,500	15	3×10^8	~400
Hydrothermal convection				
Western United States	2,300,000	----	$> 2.4 \times 10^8$	0.4
Yellowstone caldera	2,500		1.2×10^9	1,400 - 2,000
Stored heat energy				
U. S. Cordillera	2,000,000	40-80	$\sim 4 \times 10^{10}$ watts	~80
Other U. S. calderas				
Long Valley, California	450	0.7 m. y.	0.7×10^8	630
Valles, New Mexico	150	1.1 m. y.	0.2×10^8	500

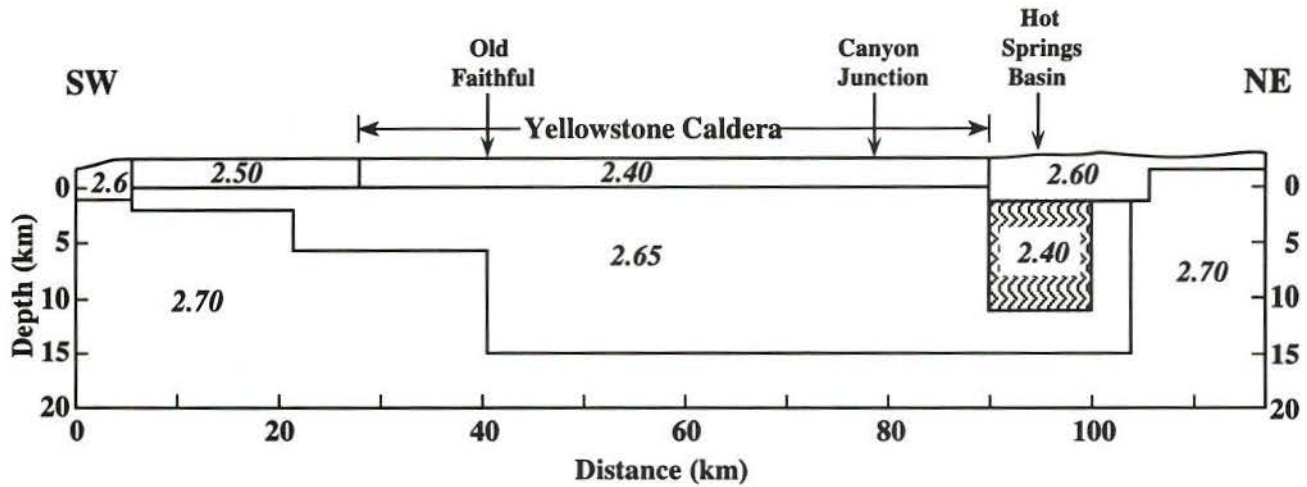


Figure 19. Density model of the upper crust of the Yellowstone Plateau. The gravity model was determined by analyzing the two-dimensional velocity and density data of Lehman and others (1982) using three-dimensional time-term analyses of seismic refraction data acquired in the 1978 seismic experiment. Numbers are densities in $\text{gm}\cdot\text{cm}^{-3}$.

Curie depths of the Yellowstone caldera

The unusually high heat flow and lateral variations in velocity structure of Yellowstone suggest significant variations in thermal sources. To examine this property and to estimate the locations of excess heat in the crust, we investigated the temperature field by determining the depths of the Curie isotherm, the temperature at which rocks lose their magnetization, here assumed to be about 575°C . Smith and others (1974) first determined a one-dimensional Curie depth model of the Yellowstone Plateau by spectral analyses of the magnetic field and inferred the depth to the bottom of magnetic crust as 10 ± 3 km. This value corresponds to a thermal gradient of $45^{\circ}\text{C}/\text{km}$ and a conductive heat flow of about 113 mWm^{-2} . This Curie depth was corroborated by Bhattacharya and Leu (1975) who determined a two-dimensional Curie isotherm map of Yellowstone that averaged 5 to 6 km beneath the central caldera, a circular zone 6 to 8 km deep outside the caldera, and 10 km deep and greater northwest of the caldera. These values correspond to a conductive thermal flux of 170 mWm^{-2} for the central caldera and an average heatflow of 152 mWm^{-2} for the remaining parts of caldera.

We employed an extension of the Curie depth method by Barnoff (1977) to investigate the three-dimensional distribution of heat sources at Yellowstone (Figure 20). This involved a comparison of the observed Bouguer gravity field, $\bar{G}$, with that of the pseudogravity field deduced from the transformation of the Earth's magnetic field due to the magne-

tized crust. The magnetic ($\bar{M}$) and Bouguer gravity ($\bar{G}$) fields of Yellowstone were digitized at 2.1 km intervals from the Bouguer gravity map of Yellowstone (Figure 18) by Blank and Gettings (1974) and from an unpublished magnetic map of Yellowstone (U. S. Geological Survey aeromagnetic map of Yellowstone, 1973). We then converted the magnetic field to an equivalent-stratum gravity field called the pseudogravity field and subtracted it from the observed gravity field (Figure 20A). This gives a residual gravity field that is a reflection of mass at temperatures exceeding the Curie isotherm ($\sim 575^{\circ}\text{C}$), i.e., "hot" rocks. The residual pseudogravity anomalies in Figure 20B reflect the distribution of rocks that exceed the Curie temperature and reveal three features: (1) the caldera is characterized by small residual anomalies of +5 to +15 mGal surrounded by a distinctive circular pattern of positive anomalies of +20 mGal along the southwest caldera rim; (2) a separate and large +30 mGal anomaly over the northeast caldera; and (3) a transitional gravity gradient from the caldera at its north and south margins.

We have not modeled these data, but the southwestern caldera anomaly has a half-width of 6 to 8 km, implying shallow heat sources in the upper crust. These anomalies coincidentally correspond to the area of the youngest (150 Ka to 70 Ka) post-collapse caldera flows of Yellowstone (Christiansen, 1984, 1993) and could represent ring dikes or cupolas of shallow low-density material. The northeastern anomaly is more circular, suggesting a spherical or a

Topographic signature, space-time evolution, and physical properties of the Yellowstone-Snake River Plain volcanic system: the Yellowstone hotspot

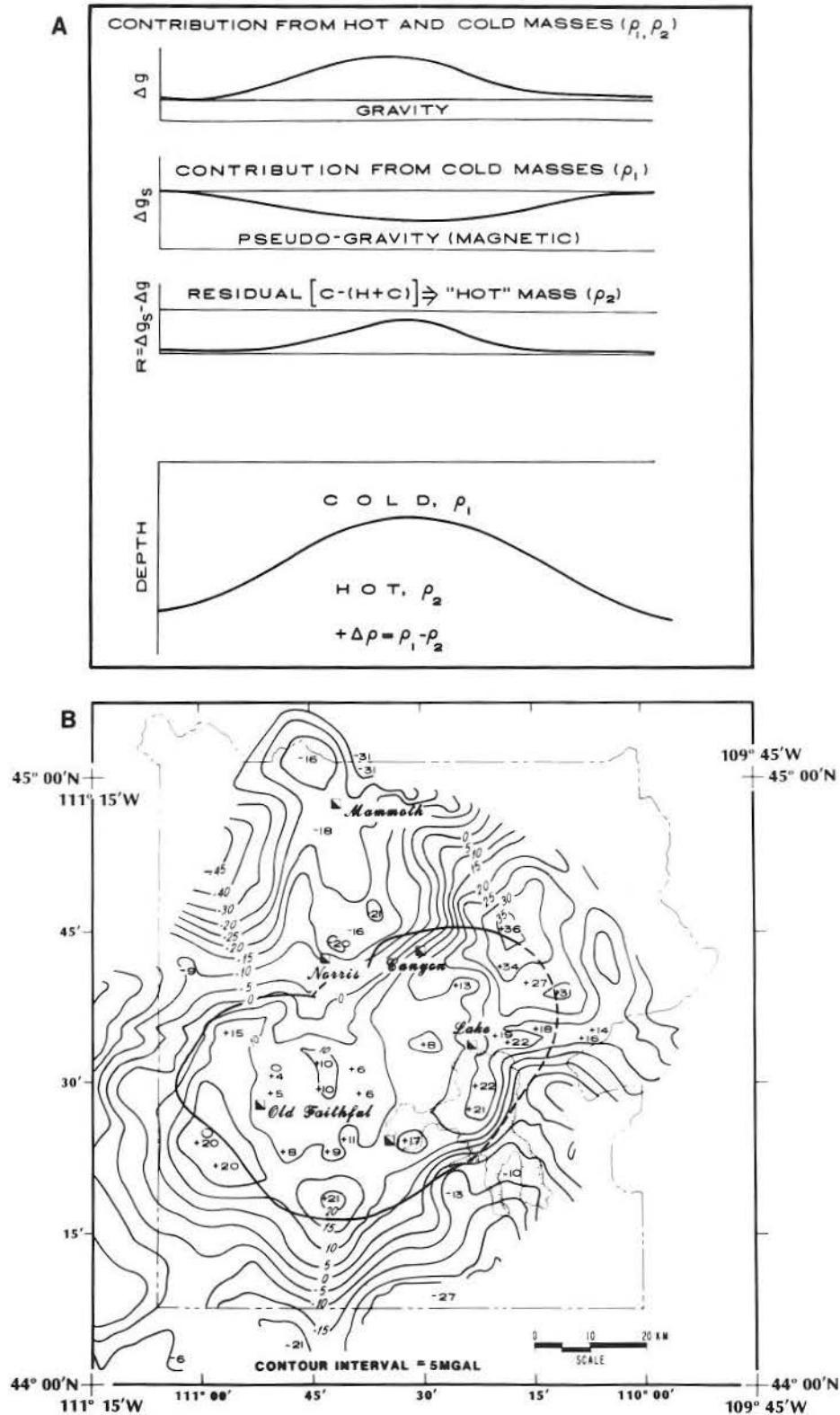


Figure 20. Curie depth-gravity response of high-temperature crustal rocks at Yellowstone using the pseudogravity field method. A, Cartoon of method; and B, Curie-depth residual gravity field (pseudogravity minus observed gravity) of the Yellowstone Plateau. Positive residual anomalies correspond to crustal rocks at temperatures in excess of the Curie isotherm of $\sim 575^\circ\text{C}$.

vertical-cylindrical shaped source. The half-width is about 6 km for a spherical source and suggests the presence of a shallow upper-crustal source. While we cannot determine the temperature distribution associated with these inferred bodies, we suggest that their association with the young Yellowstone Plateau rhyolite flows on the southwest rim and other anomalous geophysical properties, to be discussed in the next sections, reflect remnant magma/partial melt-zones or hydrothermal fluid-filled bodies.

Crustal structure

The YSRP has been the focus of two major seismic refraction experiments in 1978 and 1980 (Braile and others, 1982; Smith and others, 1982). The 1978 project targeted the large-scale tectonic and volcanic features of the YSRP. The 1980 project focused on the upper crust of the volcanically covered SRP and evaluated anomalous features revealed in the 1978 project.

The first observations of the Yellowstone velocity structure were summarized by Schilly and others (1982). They recognized large P-wave delays (up to 1.5 sec) compared to zero traveltimes delays in a uniform upper-crustal velocity model of 5.95 km/s (Figure 21). Lehman and others (1982), using a three-dimensional time-term analysis of these data, de-

duced a two-dimensional upper-crustal low-velocity body of 4.0 - 4.8 km/s compared to a thermally undisturbed, 6.05 km/s, Precambrian upper crust outside the caldera.

The upper-crustal, three-dimensional P-wave structure of Yellowstone was investigated by Brokaw (1985) using the combined 1978 and 1980 refraction seismic experiments. This structure is shown by a northeast-southwest profile from Island Park to Yellowstone that extends to depths of 10 km (Figure 22) and reveals: (1) a general caldera-wide upper crustal velocity of 5.4 km/s, which is 0.6 km/s slower than the surrounding thermally undisturbed crust of the Rocky Mountains; (2) a low-velocity zone of 4.8 km/s in a restricted area in the northeast caldera; and (3) a transitional zone between the Yellowstone caldera and Island Park with P-wave velocities of 5.4 to 5.6 km/s, intermediate between those of the caldera and the Snake River Plain to the southwest. The intermediate-depth crustal structure was not as well determined by the refraction data, but is based primarily upon wide-angle reflections recorded in the 1980 experiment. A thin remnant of the 6.15 km/s layer, observed to be 15 km thick, was modeled by Chiang and Braile (1984) for the Yellowstone to Snake River Plain transition.

The three-dimensional P-wave structure of the upper crustal structure of Yellowstone has also been studied by tomographic inversion of traveltimes from local earthquakes recorded by the Yellowstone seismograph network. Benz and Smith (1984) first inverted the traveltimes of P waves from local earthquakes recorded in Yellowstone to epicentral distances of up to 100 km with an average station spacing of approximately 15 km recorded on 25 stations. Their study revealed two areas of anomalously low velocities, with up to a 10% velocity reduction compared to a homogeneous model for areas surrounding the YSRP. They occur at depths between 3 and 15 km and are located in the northeast caldera and in the southwest caldera. Nagy and Smith (1988) inverted the P-wave structure of Yellowstone using the expanded Yellowstone seismograph network and traveltimes arrivals from ~500

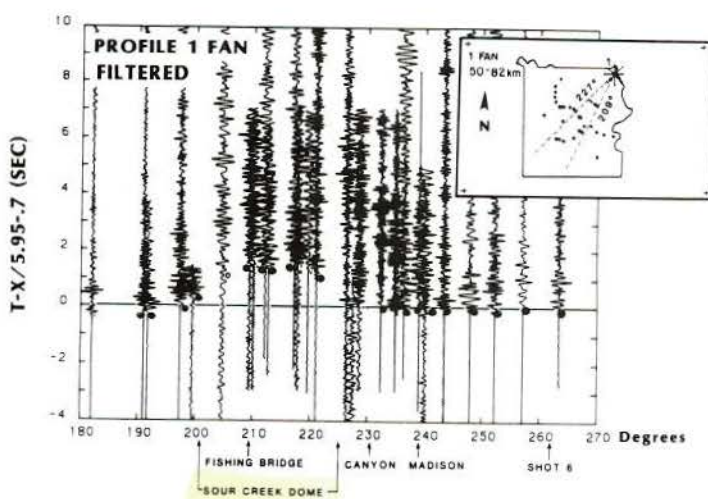


Figure 21. Pg-traveltime arrivals across the northern Yellowstone Plateau plotted versus azimuth, showing the traveltime delays of up to 1.8 sec. The delays were interpreted to be the result of a major upper-crustal low-velocity zone by Schilly and others (1982). Data were reduced to a flat datum using a reducing velocity of 5.95 km/s.

Topographic signature, space-time evolution, and physical properties of the Yellowstone-Snake River Plain volcanic system: the Yellowstone hotspot

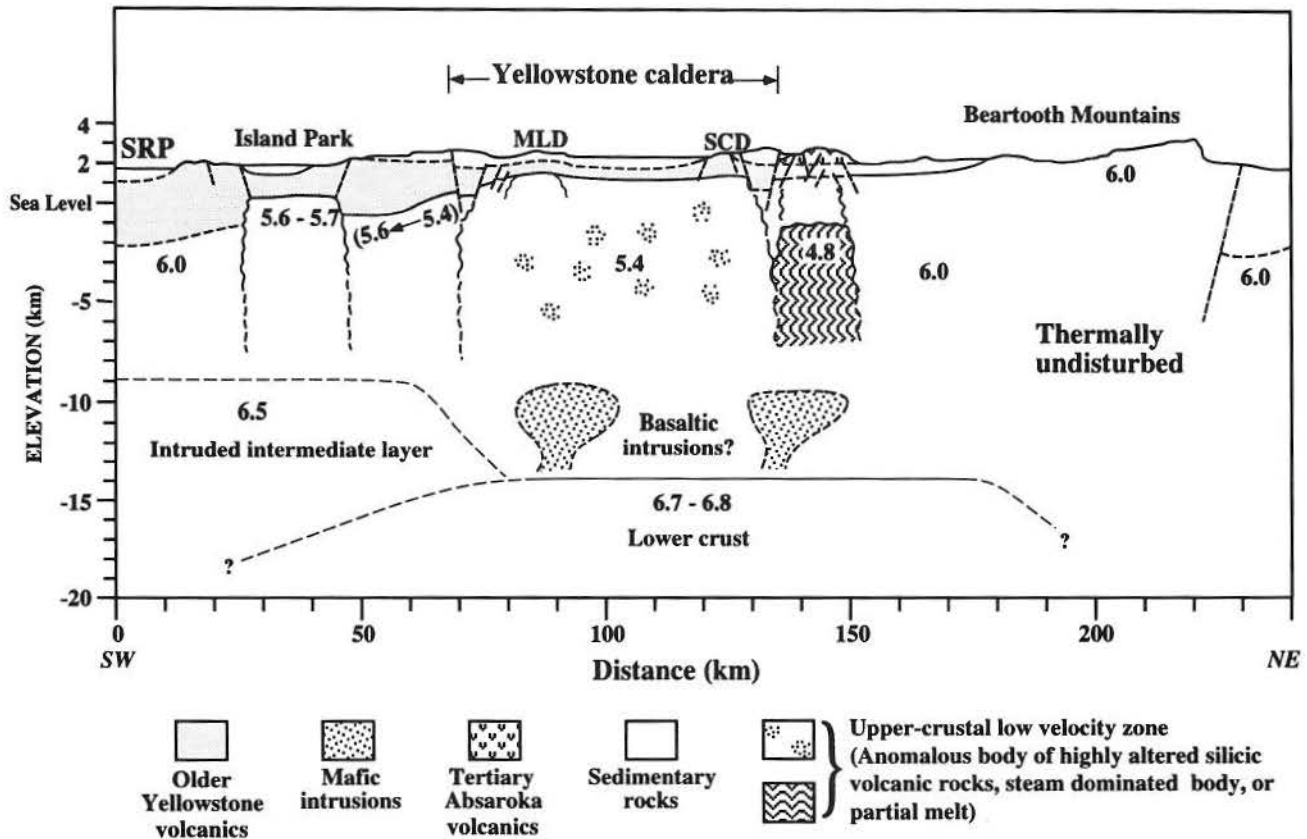


Figure 22. Cross section (northwest-southeast) of the upper-crustal P-wave velocity structure of the Yellowstone Plateau; numbers are km/s (from Brokaw, 1985). Interpretation based upon two-dimensional ray trace modeling of data from the 1978 and 1980 YSRP seismic experiments. MLD = Mallard Lake dome and SCD = Sour Creek dome.

local earthquakes. They determined a 15% decrease in velocity at depths of 3 to 15 km in the northeastern caldera and general caldera-wide 3-5% velocity reduction in upper crust. These results compare to Kissling's (1988) study of Yellowstone's velocity structure at depths from 3 and 12 km with up to a 6% velocity reduction for the caldera.

Seismicity

The historical seismicity of Yellowstone is characterized by a large, scarp-forming earthquake, the 1959, Hebgen Lake, Montana, $M_s = 7.5$ event located on the northwest margin of the caldera; its extensive aftershock sequence; and widespread earthquake swarms within the Yellowstone caldera. Here we describe the spatial and depth distribution of the earthquakes of the Yellowstone Plateau, outlining what is known about the association of seismicity with tectonic and volcanic features (Figure 23).

Intense swarms of shallow earthquakes and occasional moderate-sized events as large as the 1975, $M_L = 6.1$, Norris Junction earthquake characterize the well defined 1973-1990 seismicity of the Yellowstone region (Figure 23). The general pattern of background seismicity in the northwestern caldera is oriented northwest, similar to the orientation of nodal planes of focal mechanisms for the 1975 Norris Junction event and its aftershock sequence (Pitt and others, 1979). These features also parallel the northwest-trending, post-caldera volcanic vents (Figure 13) and suggest that the same stress field is responsible for both the tectonic and magmatic features.

However, seismicity notably varies across the caldera. Within the caldera, inner-ring fracture-zone earthquakes have not exceeded magnitude $M_L = 4.5$, and epicenters are generally scattered, except for

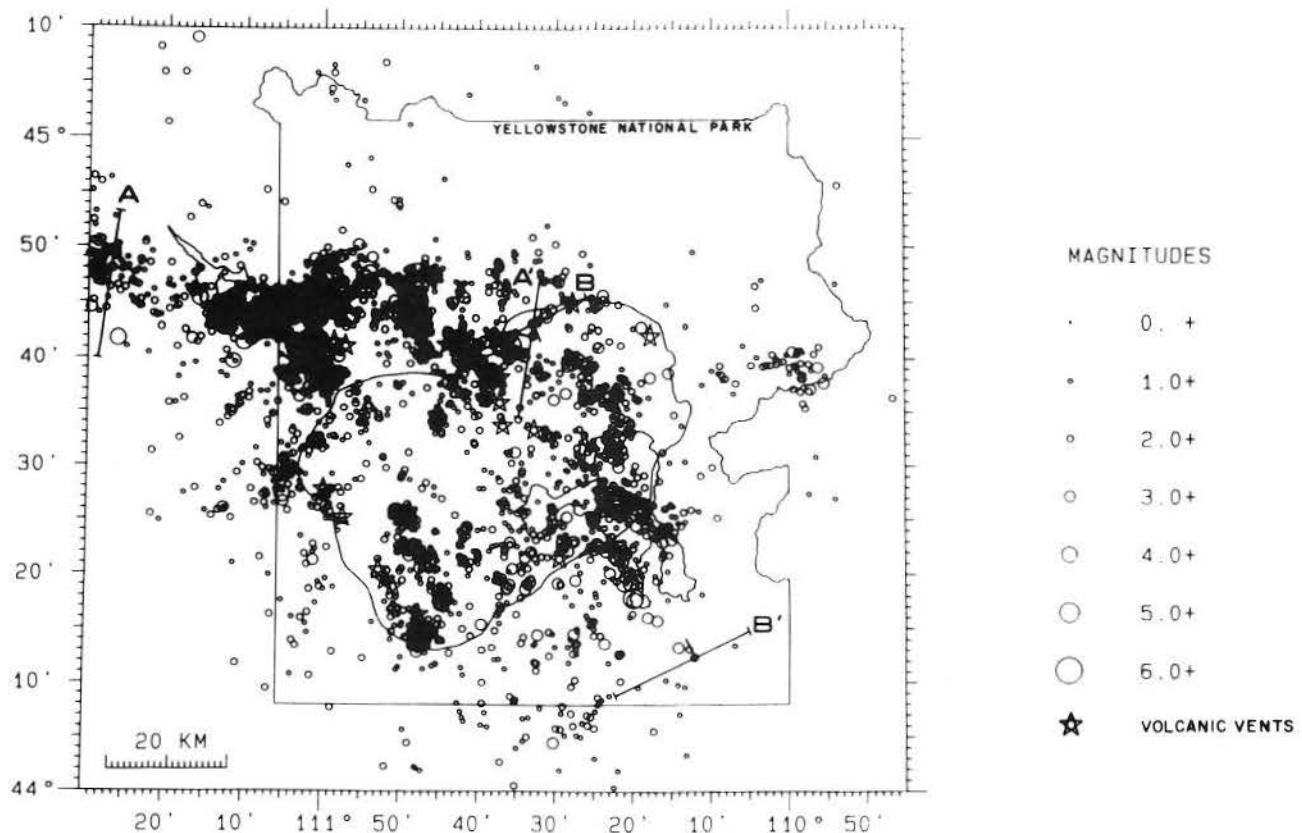


Figure 23. Seismicity map of the Yellowstone Plateau. Earthquakes of magnitude $M_L=1.5$ or greater are from compilations by the U. S. Geological Survey for 1973-1981 (Pitt, 1989) and by the University of Utah for the period 1984-1990. A-A' and B-B' define the end points of the cross section shown in Figure 25.

alignments spatially associated with volcanic vents. In several examples, there is a good correlation of seismicity with the onset of new hydrothermal activity (Pitt and Hutchinson, 1982) suggesting seismic slip on the boundaries of small upper-crustal blocks, which may reflect a combination of deformation caused by local magma/hydrothermal fluid transport and by the regional tectonic stress field.

The temporal distribution of earthquakes in Yellowstone (Figure 24) demonstrates the importance of the long-lasting earthquake swarms. Hayden (1872) first commented on the occurrence of extensive earthquakes (swarms) during his exploration of Yellowstone in 1871, described in an earlier section. Trimble and Smith (1975) and Pitt (1989) noted the occurrence of earthquake swarms with magnitudes as large as $M_L = 4.5$. Pitt and Hutchinson (1982) documented a major swarm with focal depths less than 5 km deep in 1978, between Yellowstone Lake and Canyon Junction, which lasted for several months and coincided with major changes in hydrothermal features.

The 1985 northwestern caldera swarm was the largest and most persistent in the historic record (Figure 24). Nagy and Smith (1988) documented a 6-month period of intense earthquake activity that peaked in frequency of occurrence with earthquakes of magnitude as large as $M_L \sim 4.5$ in November and December, 1985. Focal depths for these events ranged from 3 to 10 km, and the epicenters extended along a 15 km long northwest-trending zone next to the caldera boundary. Focal mechanisms for this swarm were dominated by strike-slip solutions with a common left-lateral slip nodal plane striking to the northwest and aligned with the trend of the swarm epicenters. Notably, the trend of this swarm projected northwest, orthogonal to the caldera boundary, suggesting a radial fracture extending from the caldera, perhaps either a propagating fracture extending from the Hebgen Lake fault toward the caldera or a magma-filled fracture extending radially outward from the caldera.

On the southeastern side of the caldera, epicenters associated with background seismicity are

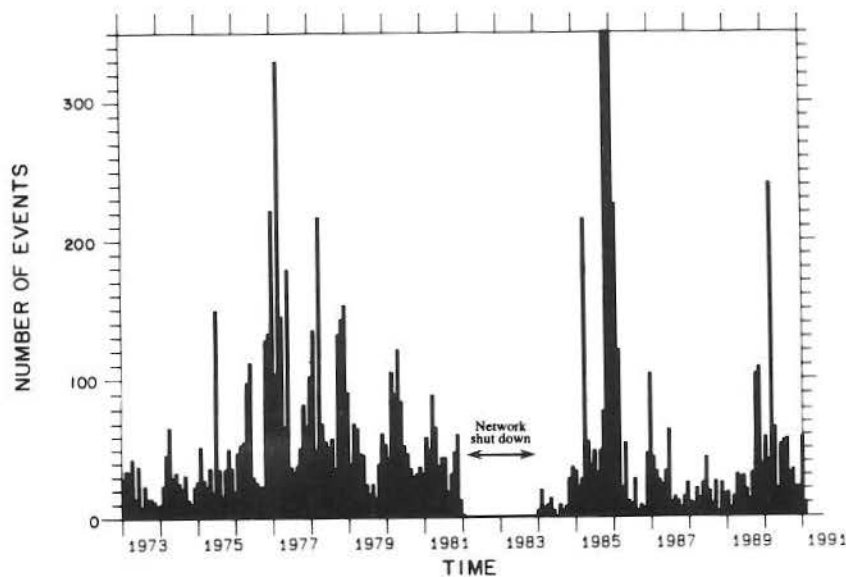


Figure 24. Temporal variation of recorded earthquakes in Yellowstone. Numbers of earthquakes are plotted in one-month intervals between 1973 and 1990. Note the sporadic and intense earthquake swarms. Source of data is the same as in Figure 23.

aligned with north-south-trending zones, including epicenters near a normal fault along the west side of the Yellowstone Lake that exhibits several meters of Holocene slip (Meyer and Locke, 1986). The alignment of epicenters in the southern Yellowstone Plateau may reflect a buried extension of the Teton fault (Figures 13 and 23).

Focal depth variations and temperatures of the Yellowstone caldera

Lateral variations in focal depths of earthquakes of the Yellowstone caldera are thought to reflect variations in depth to the brittle-ductile transition (Figure 25). Maximum focal depths outside the caldera are generally less than 15 to 20 km, but decrease to less than 5 km beneath the caldera. This distinctive shallowing suggests a thin seismogenic brittle layer above a major heat source. Within the caldera, the crust appears to be in a quasi-plastic state at depths exceeding 5 km and at temperatures greater than 350° to 450°C, incapable of supporting shear stresses (Sibson, 1982; Smith and Bruhn, 1984).

If we assume that the 80th percentile of focal depths of the Yellowstone caldera reflects the depth to the 350°C isotherm, as hypothesized by Smith and Bruhn (1984) for extensional regimes, we can estimate the temperature gradient. The maximum focal depths occur at about 10 km immediately outside and west of the caldera and correspond to a thermal

gradient of 35°C/km. Inside the caldera, the average 80th-percentile depth at ~5 km corresponds to a gradient of 70°C/km (Figure 25) and a corresponding conductive heat flow value is ~200 mWm⁻². These data suggest that the conductive heat flow of the caldera is ~200 mWm⁻², which is ~ 10 times smaller than the total heat flux of the caldera (~2,000 mWm⁻²) and thus requires a convective heat transfer of ~ 1,800 mWm⁻².

The ratio of conductive to convective heat flux provides a measure of the relative magnitude of thermal convective heat transport as specified by the Nusselt number, which is defined as the ratio of the convective to conductive heat flow.

For Yellowstone, a Nusselt number of 10 compares with Nusselt numbers, calculated by the same method by Hill (1992) for the 0.7 Ma Long Valley caldera, California, of 6 to 8.

Crustal deformation

While subsidence of the Snake River Plain over its 16 million year history reflects the long-term deformation pattern of that region, the Yellowstone Plateau has been the site of unprecedented crustal deformation in historic time. Pelton and Smith (1982) first documented the historical uplift of the Yellowstone caldera of up to 76 cm based upon repeated 1st-order precision leveling in 1975-76-77 at benchmarks originally surveyed in 1923 (Figure 26). From the mid-1970s to 1984, leveling surveys revealed an additional 25 cm of uplift. However, by 1985, the deformation reversed to subsidence which exceeded about 12 cm by 1990 (Dzurisin and Yamashita, 1987; Meertens and others, 1992).

Vasco and others (1990) modeled the uplift data (Figure 27) by dividing the crust into a number of discrete blocks, each of which may undergo a fractional dilatational volume change, ($\Delta\text{volume}/\text{volume}$). Their three-dimensional model extended from the surface to 9 km depth. For the uplift data, a maximum volume change of 0.22 km³, in the 3 km to 6 km depth range was determined in the southern

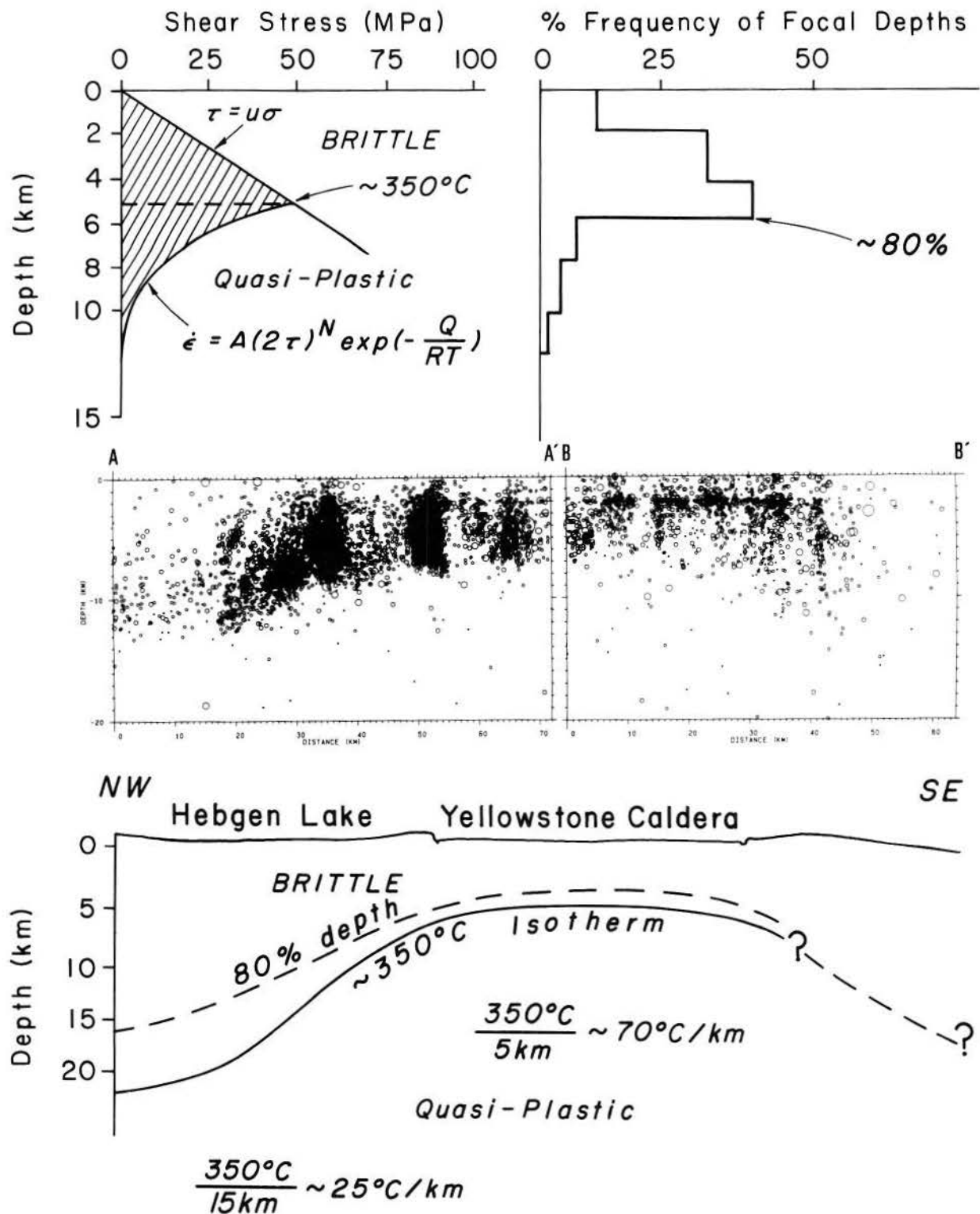


Figure 25. Cross sections of rheology properties, earthquake foci and idealized rheological model for the Yellowstone Plateau. Focal depths correspond to profiles A-A' and B-B' in Figure 23, but extended to all shocks of $M_L > 0$. The brittle-ductile transition was calculated from rheological parameters with a quartz rheology, a regional strain rate of $\sim 10^{-16}\text{s}^{-1}$, and for a high conductive heat flow, exceeding 100 mWm^{-2} .

Topographic signature, space-time evolution, and physical properties of the Yellowstone-Snake River Plain volcanic system: the Yellowstone hotspot

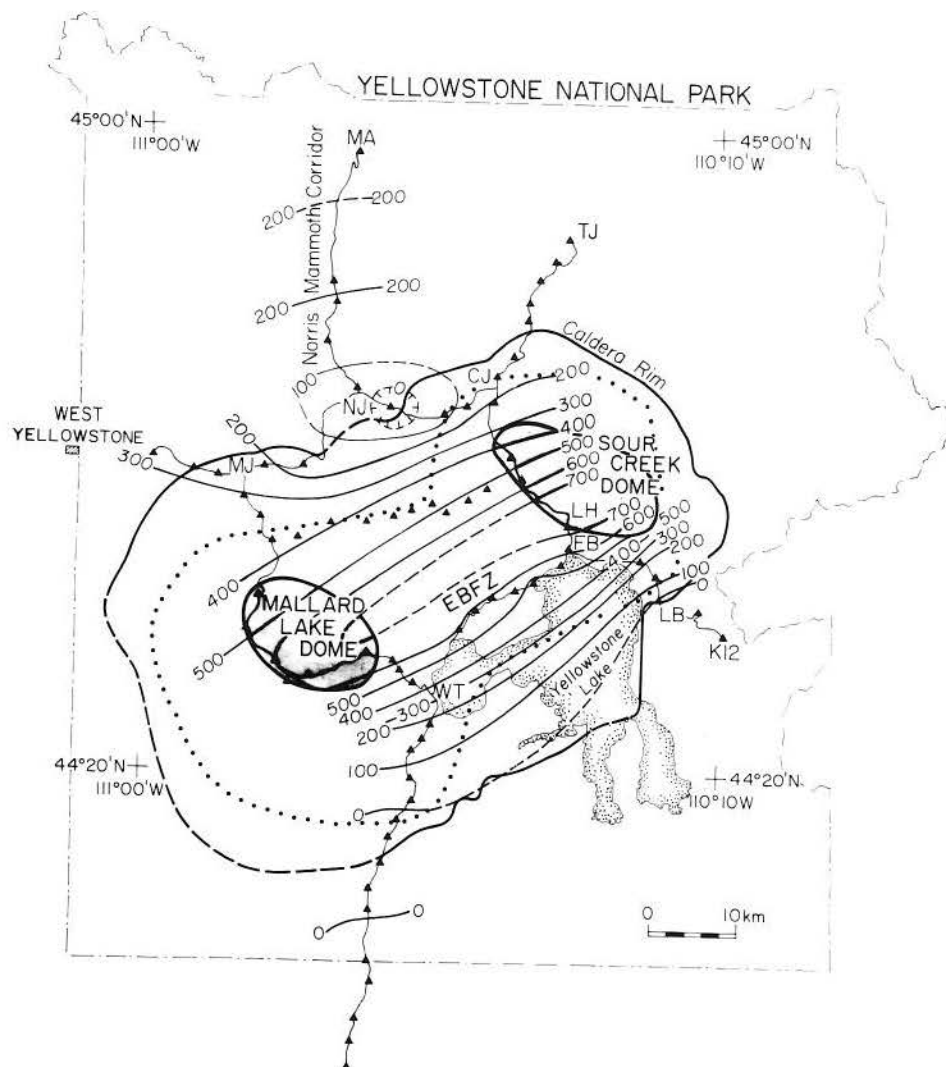


Figure 26. Map of 1923 to 1977 surface uplift of the Yellowstone caldera (modified from Pelton and Smith, 1982). Contours are in millimeters relative to a reference base station, K12, on the east side of the caldera that was assumed least affected by volcanic or tectonic activity of Yellowstone. Dotted line shows inner edge of caldera-rim fracture zone. EBFZ-Elephant Back fault zone, MA-Mammoth Hot Springs, TJ-Tower Junction; WT-West Thumb; MJ-Madison Junction; NJ-Norris Junction; CJ-Canyon Junction; FB-Fishing Bridge; and LB-Lake Butte.

caldera and a second source of 0.18 km^3 volume increase was located in the northeast caldera (Figure 27). The total volume change for the 1923 to 1977 data totaled approximately 0.73 km^3 , which corresponds to an inflation rate of $0.01 \text{ km}^3/\text{yr}$ to $0.03 \text{ km}^3/\text{yr}$. Vasco and others (1990) attributed the volume change to migration of magmas and/or hydrothermal fluids into the upper crust.

Causative mechanisms for the caldera-wide uplift, summarized by Pelton and Smith (1982), included magmatic, tectonic, and glacial-isostatic sources, but these authors suggested that the most

likely source of the 1923 to 1977 uplift was transport of magma. Dzurisin and others (1990) concluded that basaltic intrusions into the middle or upper crust or pressurization of a deep hydrothermal system (by magmatic gas or brine released by crystallization of a rhyolitic melt) were also plausible sources for the uplift.

Following the 50-year episode of historical uplift, deformation abruptly changed to subsidence of up to 6 cm between 1984 and 1985 (Dzurisin and others, 1990), and subsidence has continued to 1991. Meertens and others (1992) reported subsidence of

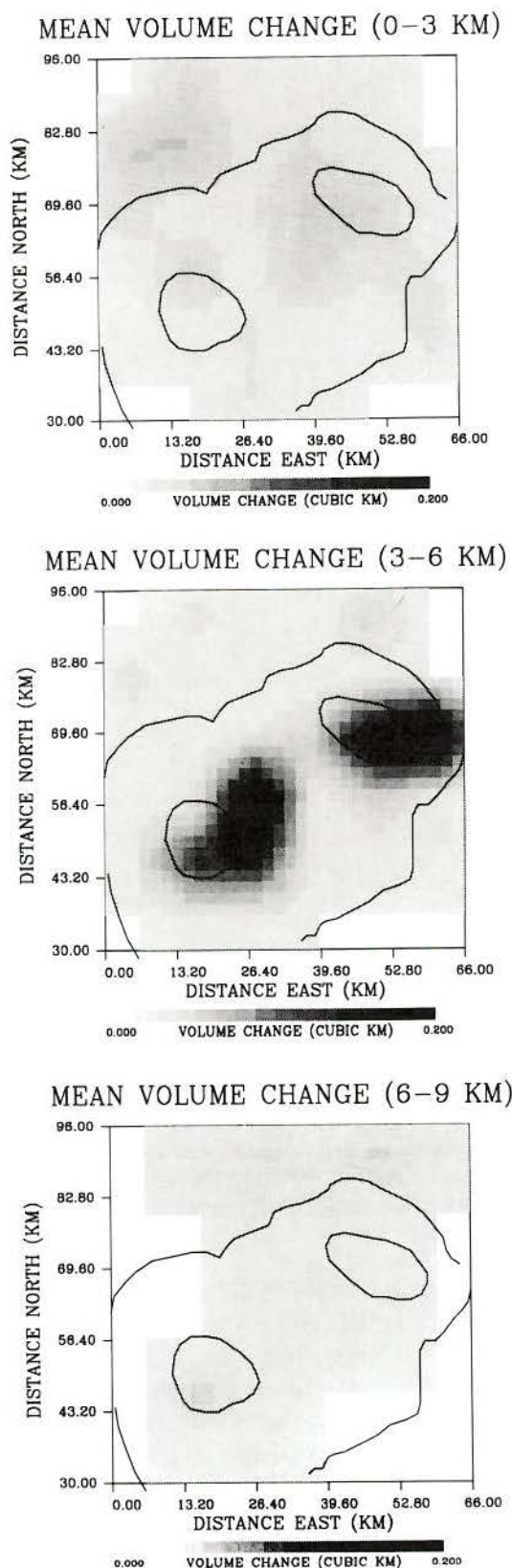


Figure 27. Locations and volumes of sources modeled for the 1923 to 1977 uplift of the Yellowstone caldera at 0-3 km, 3-6 km and 6-9 km depth ranges. Gray-scale contoured to volumetric increases in km^3 (from Vasco and others, 1990).

up to 7 cm for the period 1987 to 1991 over the entire caldera using GPS (Global Positioning Satellites) surveys corresponding to the same general area that experienced uplift from 1923 to 1977 (Figure 28). The spatial correlation between the area of uplift and the area of subsidence suggests related mechanisms. The uplift phase of the Yellowstone caldera is postulated to have resulted from intrusion of magma and/or hydrothermal fluid into the 3 to 6 km-deep upper crust. These fluids may have, in turn, produced excess hydrothermal fluids in a shallow overlying layer, which then degassed and/or experienced a reduced rate of input, producing the observed subsidence.

Evidence of a widespread magmatic connection between the Yellowstone caldera and the Hebgen Lake fault was recently suggested by Savage and others (1993), who modeled the strain field of the Hebgen Lake trilateration network for the period 1973 to 1987. They calculated a uniaxial extension of approximately $0.27 \mu\text{strain/yr}$ at an azimuth of 015° . Although these data can be explained by dislocation on a southward dipping normal fault at Hebgen Lake, their data imply significant deformation north of the surface projection of the Hebgen Lake fault. Savage and others (1993) suggested an alternate mechanism related to magmatism, namely inflation of a vertical dike and accompanying rift that extends west-northwest about 100 km from the Yellowstone caldera to the Hebgen Lake fault zone. This is a provocative interpretation, as it provides a tie between tectonic and magmatic mechanisms across a large area of the Yellowstone Plateau.

Nonetheless, we believe that the migration of magma and/or hydrothermal fluids into the middle and upper crust of the Yellowstone caldera is the most plausible explanation for the modern crustal deformation. This mechanism must have been important over much of the 2-million-year volcanic history of Yellowstone, and magma intrusion must have been the principal mechanism in the development of

the resurgent domes. The modeled volume changes of Yellowstone's historic crustal deformation are commensurate with intrusion of hydrothermal fluids and rhyolitic/basaltic melts into the upper crust as a likely mechanism.

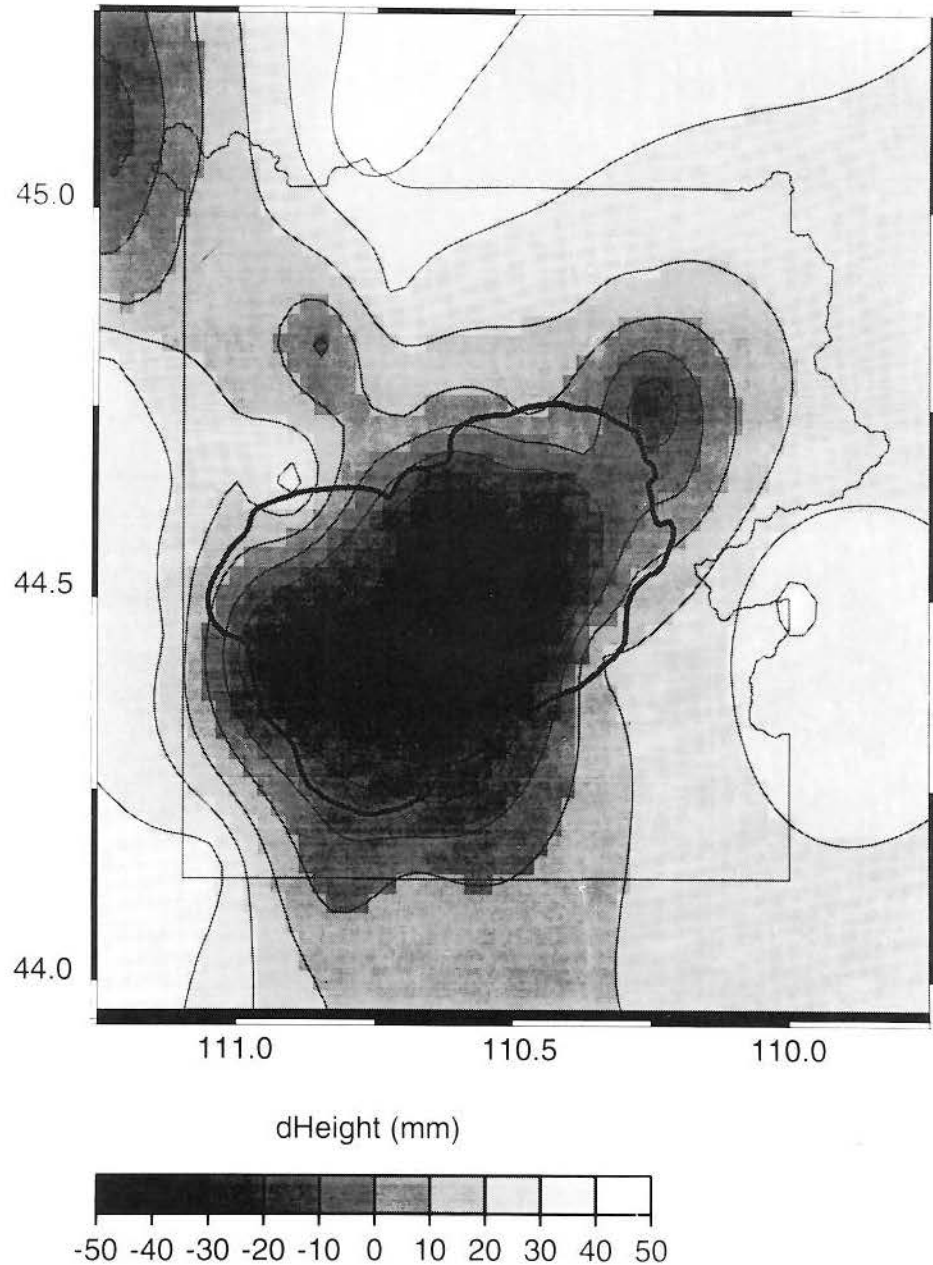


Figure 28. Map of 1987 to 1991 Yellowstone caldera subsidence from Global Positioning Satellite (GPS) measurements (from Meertens and others, 1992). Height changes, $d\text{Height}$, are gray-scale contoured in mm. Base level taken by averaging measurements of GPS stations outside the caldera that are assumed to be least influenced by the Yellowstone hotspot. Caldera boundary is shown by heavy black line.

Mantle origin of the YSRP

An enormous amount of heat and a continuous supply of magma are required for the 16-million-year history of YSRP silicic volcanism as well as the $2,000 \text{ mWm}^{-2}$ heatflow of Yellowstone. We believe this requires a mantle source and suggest that mantle plume-plate interaction (the Yellowstone hotspot) is responsible for the time-spatial characteristics of the

YSRP. In this section, we build upon the picture of a seismically and thermally active Yellowstone caldera as the active element of the YSRP and examine the evidence for mantle coupling by a plume-plate model. We also comment on alternate mechanisms that may account for some of the observations.

Lithospheric structure of the YSRP

Perhaps the most compelling evidence for a plume-plate model for the YSRP is the progression in silicic volcanism ages discussed in an earlier section of this paper. The P-wave crustal structure of the YSRP also reveals a similar systematic variation in thicknesses, velocities, and compositions of intermediate- and upper-crustal layers that corresponds to the variation in ages of the silicic volcanic rocks. In Figure 29, the upper crustal structure of the YSRP is plotted along with the averaged surface elevation. The granitic upper crust of the North American continent, which is normally 10 to 15 km thick with a P-wave velocity of 5.9 - 6.1 km/s (Braile and others, 1989), is absent or thin along the entire YSRP. We believe it has been largely replaced by silicic volcanic rocks and mafic intrusions related to the Yellowstone hotspot. At the Yellowstone

caldera, a heterogeneous upper crust extends to depths of about 10 km.

Along the SRP, a southwest decrease in topography also reflects a systematic change in crustal structure. The near-surface basaltic layer thins from 2 km in southwestern Idaho to zero at Island Park and is apparently absent at Yellowstone. Correspondingly, the deeper, 2 km-thick, silicic volcanic layer extends from southwestern Idaho 800 km northeastward to Yellowstone beneath the younger basalt wedge. The high-velocity (6.5 km/s) intermediate crustal layer is interpreted as a solidified mafic remnant of the crustal magma sources. Neither the Moho nor the seismic velocity structure of the lower crust of the SRP is as well determined as the velocity structure of the upper crust, but these features appear relatively homogeneous and resemble the lower crust of the surrounding thermally undisturbed regions.

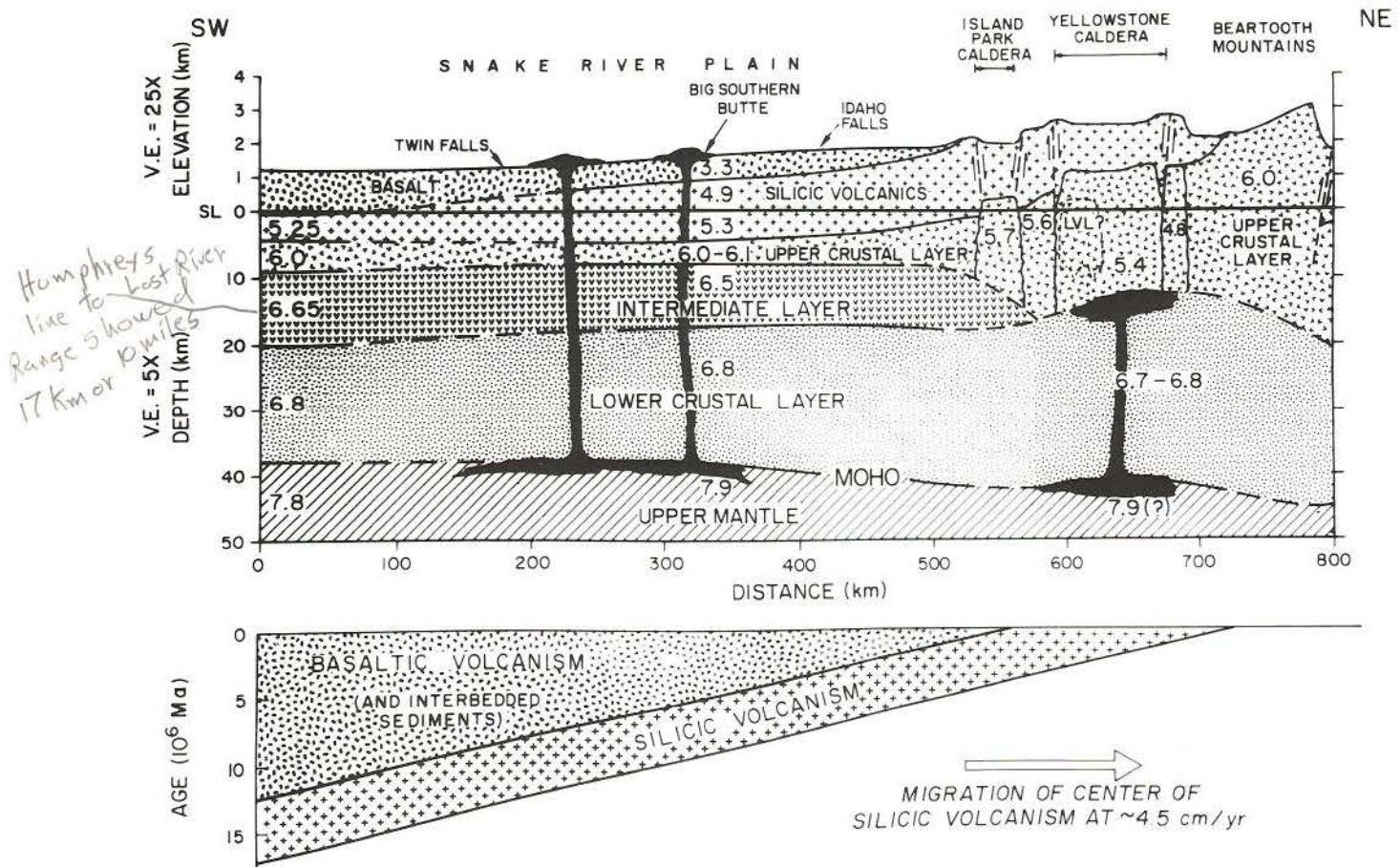


Figure 29. Generalized crustal structure and topography along the path of the Yellowstone hotspot from southwestern Idaho and northern Nevada to the Yellowstone Plateau. Figure compiled from the data and models of Hill and Pakiser (1966), Braile and others (1982), Smith and others (1982), Elbring (1984), and Brokaw (1985). P-wave velocities (numbers within patterned areas) are in km/s. Basaltic rocks thin and silicic volcanic rocks thicken with decrease in the ages of the silicic centers shown in Figure 3. Idealized locations of mantle-derived basaltic intrusions noted by black pattern. V.E.=vertical exaggeration.

Mantle velocity structure

The most definitive information on upper mantle structure comes from observations of seismic traveltimes from earthquakes, using tomographic inversion. These studies used P-wave seismic traveltimes from earthquakes recorded at teleseismic distances (greater than 5,000 km) to provide information on deeper structure of the lithosphere and suggest an anomalously low-velocity crust and upper mantle structure of the entire YSRP.

The first detailed teleseismic P-wave study of the upper mantle of Yellowstone, by Iyer and others (1981), revealed caldera-wide delays as large as 1.5 sec. They modeled the traveltime anomalies as the result of velocity decreases of up to 10% in the upper crust, 5% in the lower crust, and 6% in the upper mantle, with velocity anomalies evidenced as deep as 165 km. Evoy (1978) used the same teleseismic traveltime data simultaneously with the long-wavelength Bouguer gravity field for a combined velocity-density distribution model to depths of about 100 km. His model identified density anomalies of $\pm 0.1 \text{ gm-cm}^{-3}$ for the heterogeneous upper crust and -0.05 to -0.10 gm-cm^{-3} reduction for the lower crust and upper mantle.

The most recent study of the upper mantle structure is from teleseismic P-wave data acquired by regional seismic networks in the YSRP (Dueker and Humphreys, 1989). These authors modeled teleseismic P-wave delays along three 300 km-long profiles across the YSRP from depths of 40 km to 200 km, which are characterized by low velocities, with up to a 3% reduction compared with the surrounding thermally undisturbed Rocky Mountains (Figure 30). Dueker and Humphrey's (1993) model also includes a zone of P-wave reduction (up to a 3%) in the crust beneath Yellowstone that is underlain by a pervasive upper-mantle low-velocity body with up to a 3% reduction at depths of 40 km to 100 km. Low-velocity mantle extends southwest beneath the SRP to depths of 200 km, indicating a deepening and broadening mantle anomaly, whereas the crustal anomaly decreases in its magnitude and extent (Figure 30). Anomalously low crust and upper mantle P-wave velocities along the entire YSRP are consistent with models that suggest of partial melting and excess fluids in the lithosphere.

Geochemical evidence

He^3/He^4 ratios determined from samples acquired in the Yellowstone's hydrothermal systems by Craig and others (1978) and Kennedy and others (1985) are also consistent with a mantle source. This conclusion was corroborated by anomalous isotopic ratios of Nd and radiogenic isotopes of Pb and Sr in the YSRP, which are also consistent with an undepleted mantle source (Doe and others, 1982; Leeman, 1982). Additional evidence for mantle coupling comes from an assessment of Nd, Sr, Pb, and O isotopic compositions of basalts and post-caldera rhyolite flows of the Yellowstone Plateau (Hildreth and others, 1991). Their data reveal variations in the isotopic compositions of these elements that were interpreted to reflect sources from subcaldera upper-crustal regions and sources independent of the caldera magma system. These authors universally argue that the isotopic composition of Yellowstone rhyolites requires deep-crustal sources that have been pervasively altered by mantle-derived basalts.

Stress-field directions

Earthquake focal mechanisms, fault slip directions, and orientations of volcanic dikes and vents are the main sources of information on stress directions for the YSRP. Because the dominant mode of contemporary deformation in the ISB is extension, the direction of maximum extensional strain, corresponding to the T-axes of focal mechanisms, is assumed to be in the same direction as σ_{hmin} (Figure 31). However, Peyton (1991) made a detailed study of the stress field of Yellowstone by inversion of focal mechanisms and showed that vectors orthogonal to mapped faults, dikes, and rifts are not necessarily in the direction of σ_{hmin} . We have thus only included reliable stress direction indicators that have been evaluated in published studies. The stress field data in Figure 31 are from Eddington and others (1987) and from a summary by Zoback and Zoback (1989).

Beginning with the south side of the YSRP (Figure 31), the directions of σ_{hmin} are generally south-east- to east-trending throughout southeastern Idaho and western Wyoming, an area of moderate seismicity. Northward into the Yellowstone region, the directions of σ_{hmin} from inversion of focal mechanisms are northeast (Peyton, 1991). Northwest of the Snake River Plain, σ_{hmin} orientations are also northeast.

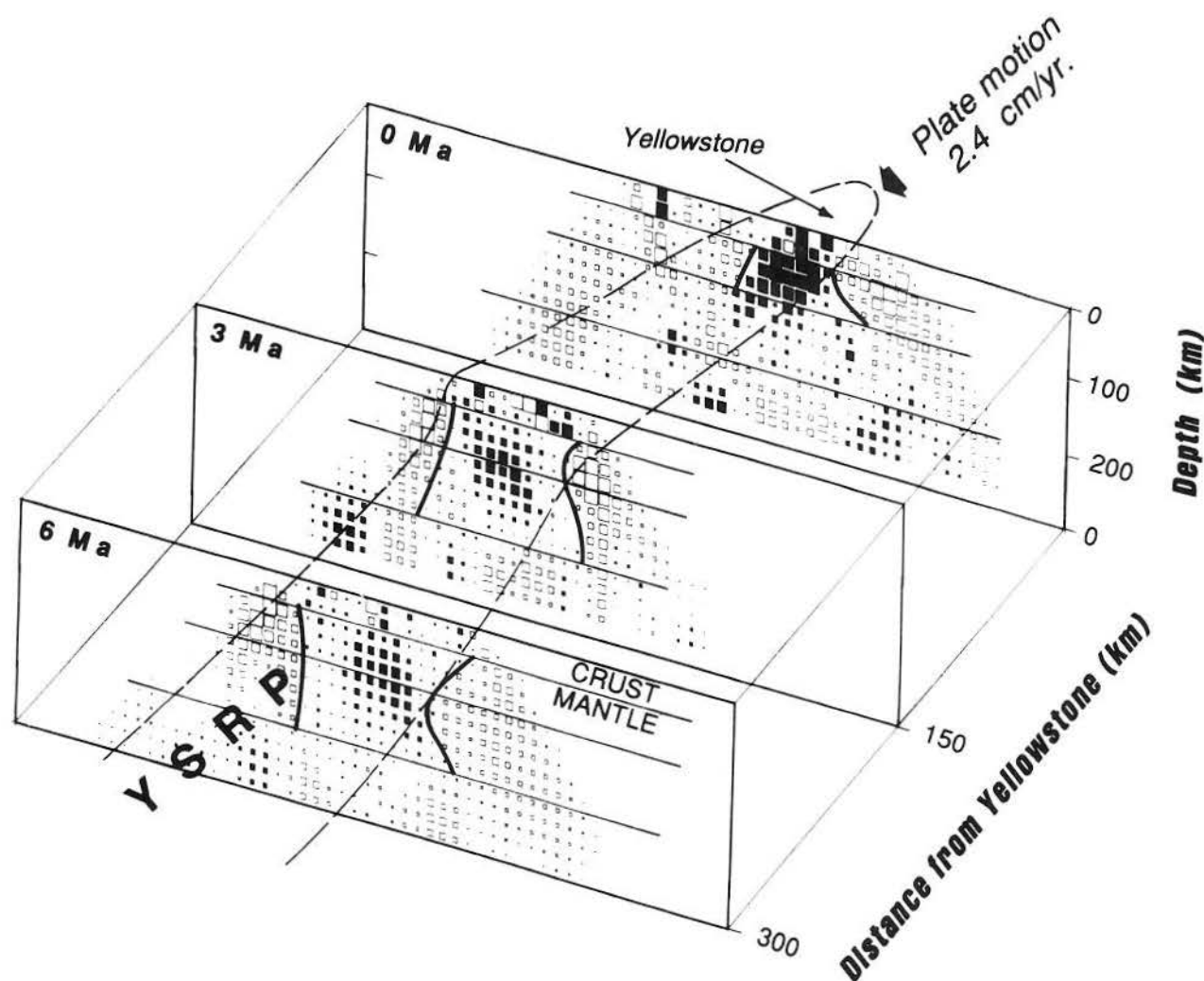


Figure 30. P-wave velocity profiles of the crust and upper mantle of the YSRP from the surface to depths of approximately 300 km determined from inversion of traveltimes of teleseismic earthquakes (from Dueker and Humphreys, 1989, 1993). The velocity-depth profiles correspond to 0, 150, and 300 km distances from the Yellowstone Plateau. Velocity differences are scaled from a laterally homogeneous model and are shown by square symbols that are scaled in size to the percent velocity perturbation. The largest symbols are $\pm 3\%$, dark = slow velocities and open = fast velocities.

On the basis of focal-mechanism studies west of Yellowstone National Park, Smith and others (1977) and Stickney and Bartholomew (1987) identified a localized stress regime with a north-trending σ_{hmin} . This region is distinguishable from the northern ISB in the vicinity of the Montana-Idaho border where σ_{hmin} trends northeast, the same as south of Yellowstone (Figure 31). We suggest that this restricted regime of north-south σ_{hmin} is a perturbation of the stress field perhaps associated with lithospheric bulging. It may also be associated with an east-west zone of magmatic related-rifting, which is hypothesized by Savage and others (1993) to extend 100 km

westward from the northern Yellowstone caldera to Hebgen Lake.

An important feature of the regional YSRP stress field is the 45° rotation of σ_{hmin} from northeast on the northwest side of the eastern SRP, to east-west southeast of the SRP (Figure 31). The older Basin and Range normal faults northwest of the SRP and vents and fissures located on the SRP are consistent with northeast-trending σ_{hmin} . This suggests that structures within the SRP were influenced by the same stress field as that of Yellowstone. However, a change in direction of σ_{hmin} to east-west is evident

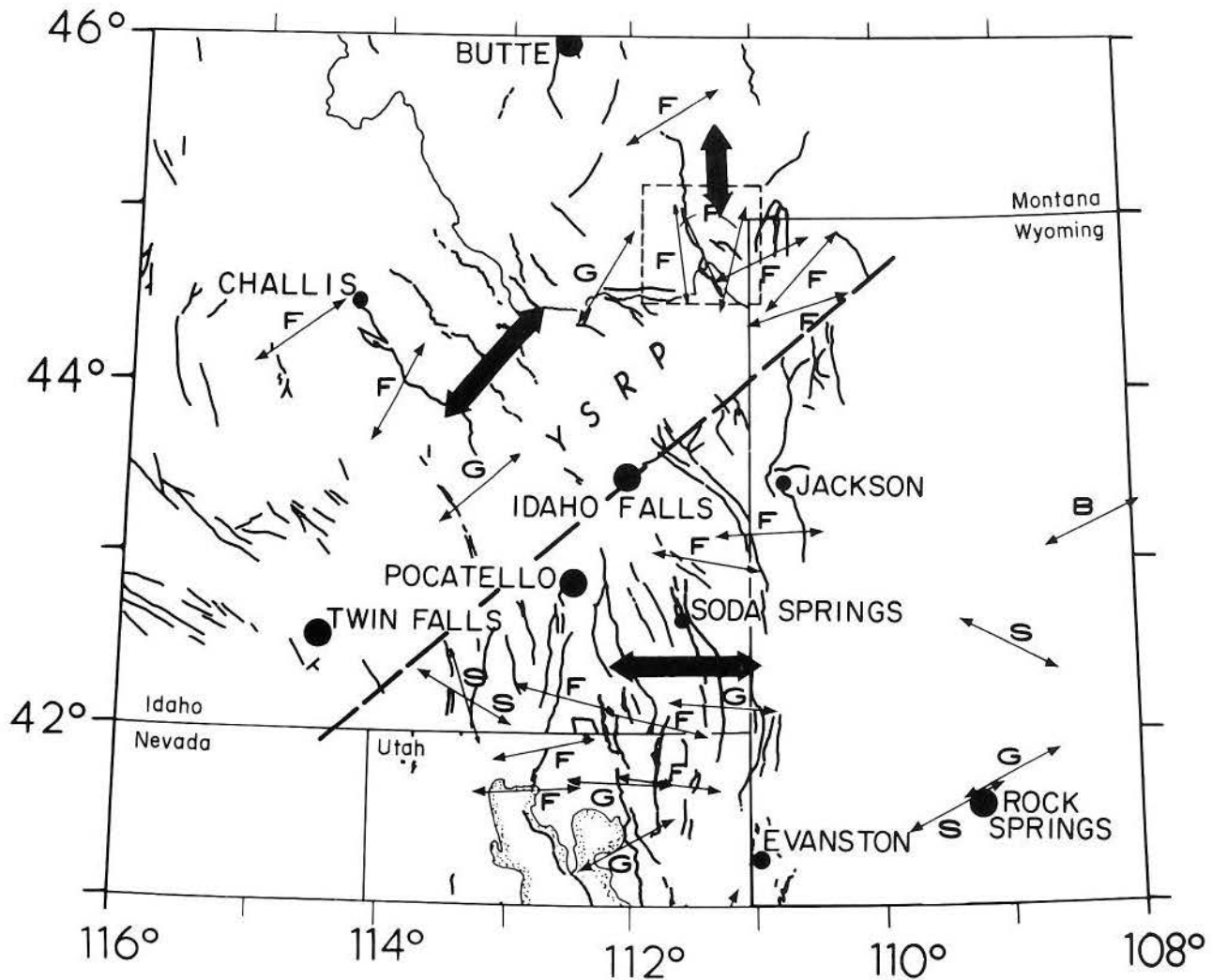


Figure 31. Map showing the orientations of crustal extension of the YSRP, assumed to be in the direction of minimum horizontal compressive stress, σ_{hmin} (modified from Smith and Arabasz, 1991). Data are from compilations by Eddington and others (1987) and Zoback and Zoback (1989), and include selected Cenozoic faults shown in Figure 3. Stress indicators: F = focal mechanism; G = geologic; and S = in situ stress.

from the alignment of normal faults on the southeast side of the caldera and the *en echelon* zone of apparently right-stepping, north-trending normal faults extending to the Teton fault. Along the southeastern edge of the SRP, the stress field appears to change trend to east-west (Figure 31) and continues into southeastern Idaho, where the seismicity belt becomes discordant with northwest-striking Quaternary faults near this stress-field boundary.

Topographic decay of the YSRP

Spatial variations in topography and the elevations of crustal boundaries attributed to thermal mechanisms are shown in Figure 32. The systematic

topographic decay along the SRP was first noted by Brott and others (1978, 1981), who fitted curves corresponding to the averaged topography in response to the passage of a crustal heat source. They postulated that the overriding plate produced cooling and thermal contraction of the surface and the corresponding 1.5 to 2 km topographic subsidence of the eastern Snake River Plain.

We have also fitted the averaged surface elevations along the YSRP (Figure 32), the boundary between the basaltic and rhyolite layers, and the boundary between the rhyolite and the 6.0 km/s basement layer. The surface topographic decay fits an expression, $z = 3.07 - 0.52 \sqrt{t}$, (where t is the age of the silicic volcanism in millions of years). This

expression is the common scaling law for fitting topography behind oceanic hotspots and mid-ocean ridges (Sclater and Francheteau, 1970; Parsons and Sclater, 1977). Moreover, the boundaries between the rhyolite and the granitic crustal layers also fit a decay curve of $z = 1.22 - 1.53 \sqrt{t}$. This curve reflects a faster decay than the surface topography and may be the result of additional crustal intrusion by basalts as well as isostatic adjustment for the additional load

of the high-density mid-crustal layer, sediments, and basaltic eruptions in the SRP.

Mantle plume-plate interaction of the Yellowstone hotspot

Smith and Sbar (1974) first suggested that the influence of convective upper-mantle flow on the

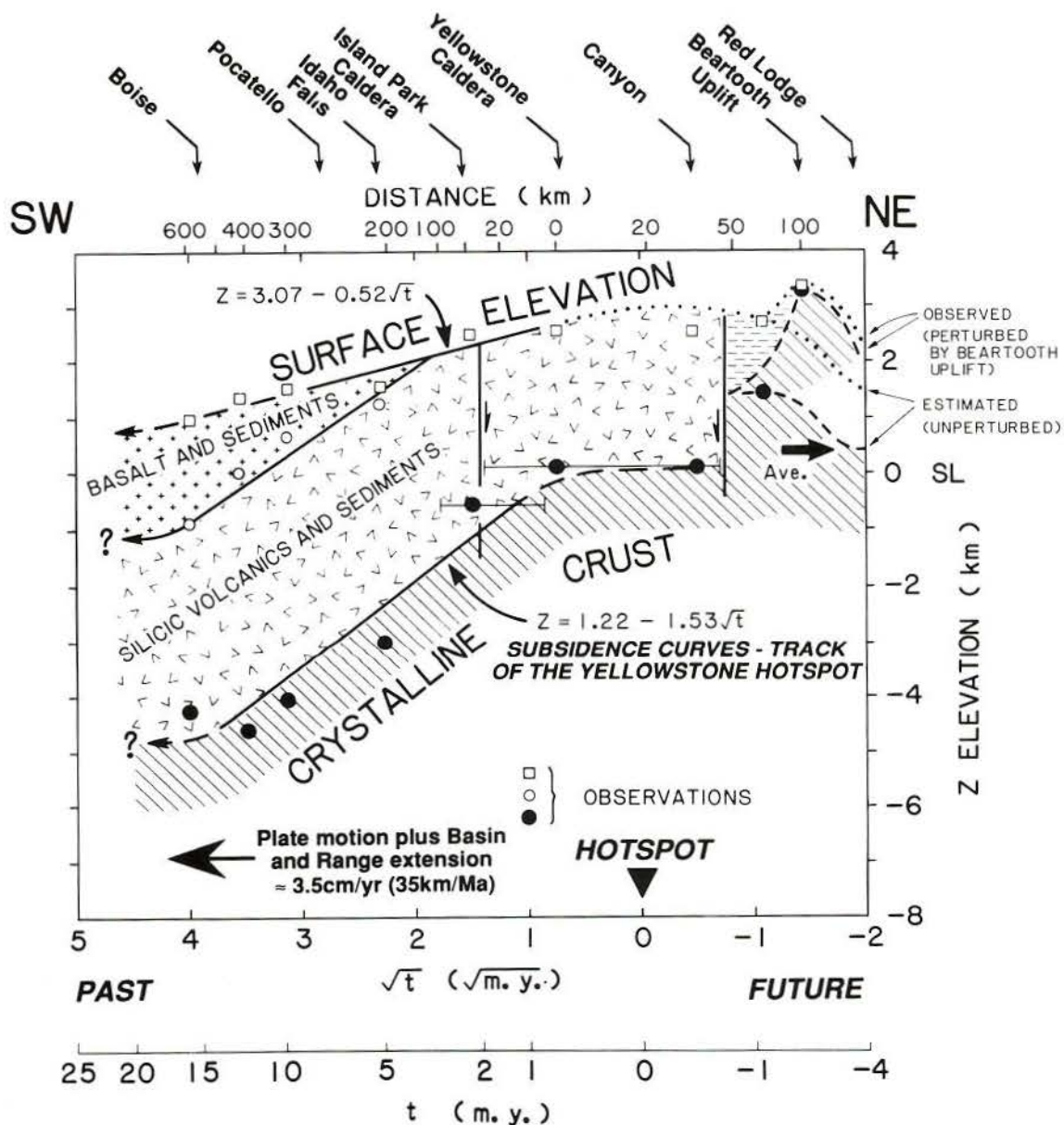


Figure 32. Plot of surface topography and elevations of the boundaries between the silicic and basaltic rocks and the crystalline granitic crust along the YSRP, showing the systematic elevation decay proportional to the $\sqrt{t}$. t is the age of the silicic volcanic rocks in millions of years from Armstrong and others (1975).

overlying North American plate may contribute to differential intraplate velocities across the Yellowstone region and hence to the tectonic processes manifest at the surface. More recently, Westaway (1989) described an analytic model for interaction of return flow associated with an hypothesized Yellowstone plume with the overlying North American plate to explain the dual-branched pattern of seismicity and similar patterns of faulting surrounding the Snake River Plain. Pierce and Morgan (1990, 1992) also described idealized models of plume ascension and lateral spreading at the base of the lithosphere to account for the parabolic patterns of seismicity of the YSRP.

To analytically evaluate the surface manifestation of a mantle plume-plate interaction at Yellowstone, we collaborated with Neil Ribe, Yale University, who developed a three-dimensional fluid-flow modeling code incorporating the interaction of an upwelling mantle plume with a southwestward moving lithospheric plate (Ribe and Christensen, 1992a,b). **Figure 33** shows the preliminary plume-plate model for the Yellowstone hotspot, described as a rectangular box filled with a fluid whose viscosity varies as a function of depth, temperature, and strain rate. We assumed conditions and properties deduced for the YSRP including a cooler, 80 km thick lithosphere, determined from surface-wave analyses for the northern Basin and Range province (Priestly and others, 1980) and represented by higher viscosity material near the top of the model. The motion of the lithosphere was simulated by imposing a horizontal velocity of 3.5 cm/yr (including a 2.5 cm/yr North American plate velocity and a 1 cm/yr concomitant extension rate) on a low-viscosity layer.

A thermal profile is specified as an upstream boundary condition and all sides of the model except for the top are open to permit flow through the boundaries. The imposed surface velocity drives large-scale shear flow in the box that contains a buoyant thermal plume, which was produced by imposing a Gaussian-shaped temperature anomaly on the bottom of the model (**Figure 33**). The rising plume spreads out against the base of the overlying lithosphere and at the same time is swept downstream by the shear flow. For this model, the buoyancy flux was perturbed until the "stagnation line" (Sleep, 1990) matched the outer margin of Zone III, i.e., the outer margin of the seismicity of the

YSRP. This yielded a buoyancy flux of 0.38 Mg/sec (**Figure 33**). A vertical cross section (AA') of the resulting temperature field (**Figure 33**) shows the upper mantle plume at a depth of 80 km.

While **Figure 33** represents a preliminary model, it provides some important insights into the expected patterns of observable quantities such as topography, seismicity, and stresses. Material supplied by the plume flows in from the right and is separated by a sharp front of roughly parabolic shape, corresponding to the stagnation streamline. Note how the stagnation line has the same shape as the stream-line features observed along the YSRP, including the dual-branched parabolic-shaped seismicity and topography (**Figures 3 and 4**).

While we have not performed a quantitative analysis of this model, the correspondence of the distinctive patterns of seismicity and topographic highs surrounding the SRP with downwelling mantle flow may reflect surface loads buoyed up above the outer wake of the plume as it impinges on the stable mantle. This model fits a narrow (approximately 150 km-wide) hot core at Yellowstone that cools to the southwest behind the hotspot, producing the systematic topographic decrease along the SRP attributed to thermal decay (**Figure 32**).

We realize our model for the Yellowstone hotspot does not explicitly require a mantle plume, but a model for plume-plate interaction matches the general properties of the observed topography and seismicity that distinguish the YSRP. Further work is required to incorporate such features as concomitant crustal extension and explain how the plume would initially develop. These considerations are beyond the scope of this paper.

A unified model for the space-time evolution of the YSRP

We complete our discussion by remarking on the excellent correlation between the space-time evolution of Yellowstone's silicic volcanism, its signature in the topographic field, and its regional geophysical properties. We believe that the observations and data discussed throughout this paper are consistent with a unified working hypothesis for a hotspot origin of Yellowstone's volcanic system.

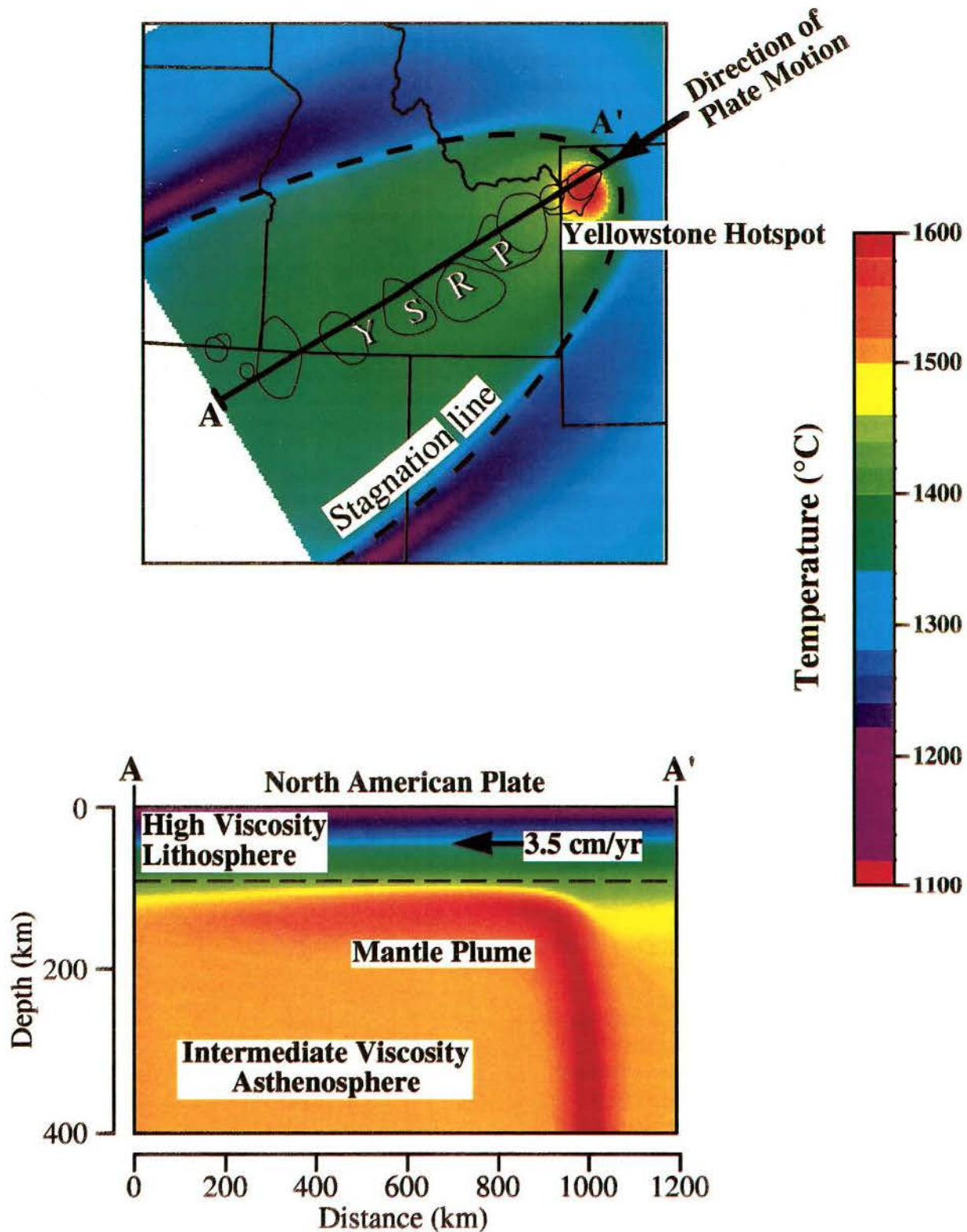


Figure 33. Model of Yellowstone hotspot origin from a mantle plume-plate interaction. The model is characterized by its temperature field contoured in °C. Top: planform view showing the "bow-wave" shaped flow lines taken from a horizontal cross section at the 100 km depth. The dark line corresponds to the approximated stagnation line of Sleep (1990), which separates up-flowing from down-flowing mantle and also corresponds to the outer edge of anomalous YSRP seismicity. Bottom: vertical cross section of the interaction of the plume with the overriding lithospheric plate. Mantle plume-plate model courtesy of Neil Ribe of Yale University.

Topographic signature, space-time evolution, and physical properties of the Yellowstone-Snake River Plain volcanic system: the Yellowstone hotspot

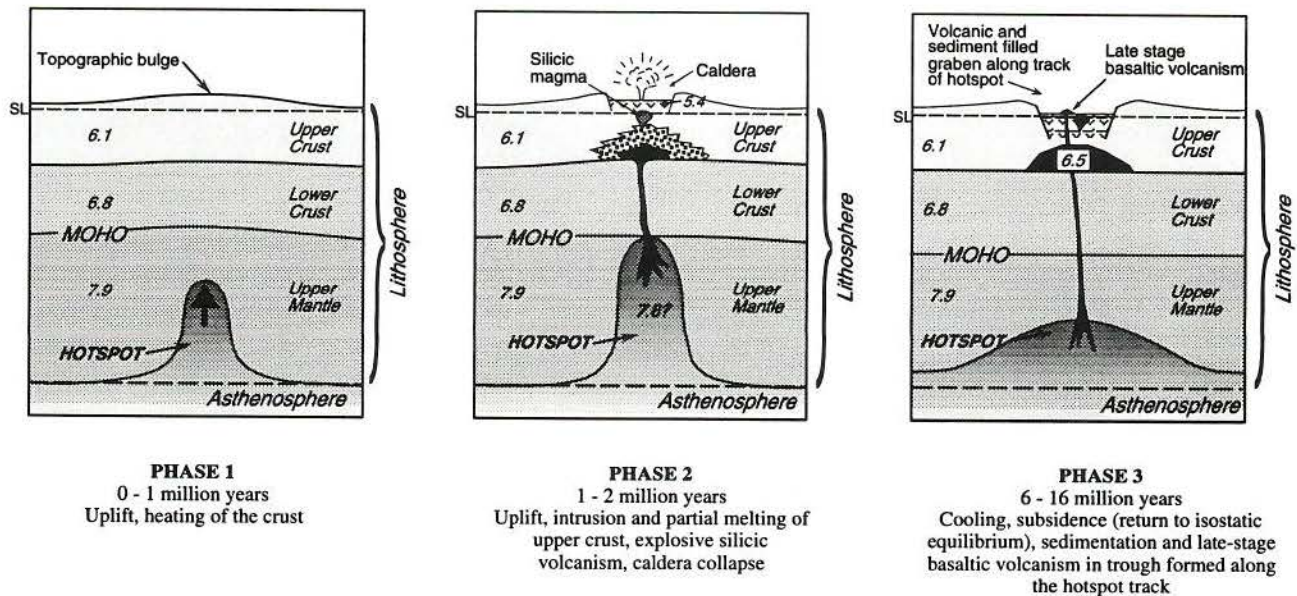


Figure 34. Models of lithospheric evolution along the track of the Yellowstone hotspot, showing the generalized P-wave velocity structure and topography: (Phase 1) the pre-caldera phase—regional topographic tumescence associated with initial hotspot interaction with the lithosphere, (Phase 2) the “Yellowstone” phase—explosive, silicic volcanism and caldera formation with accompanying caldera collapse, and (Phase 3) the “Snake River Plain” phase—cooling and thermal contraction of the lithosphere with episodic eruptions of mantle-derived basalts.

Together, **Figures 34** and **35** summarize our model for the 16-million-year space-time evolution of the YSRP and its origin as related to the interaction between a plume and the southwestward moving North American plate. The systematic variations in lithospheric structure of the YSRP are seen by plotting three time-slice cross-sections at: 0, 1 to 2, and 6 to 16 Ma (**Figure 34**). Physical properties such as topography, heat flow, and gravity are shown as a function of distance and age of silicic volcanism along the YSRP, plus our conceptual model for the hotspot (**Figure 35**). A hypothetical working model for the lithospheric evolution of the YSRP, which corroborates the scenario of basaltic and rhyolitic magma generation proposed by Leeman (1982) for the SRP and for Yellowstone by Christiansen (1984; 1993), includes these observations (**Figures 33** and **34**). Our model is consistent with those of Hildreth (1981), Glazner and Ussler (1989), and Johnson (1991), who describe how silicic magmas originate from mid-crustal melts accompanying the buoyant ascent of plume generated basalts.

Pre-caldera phase

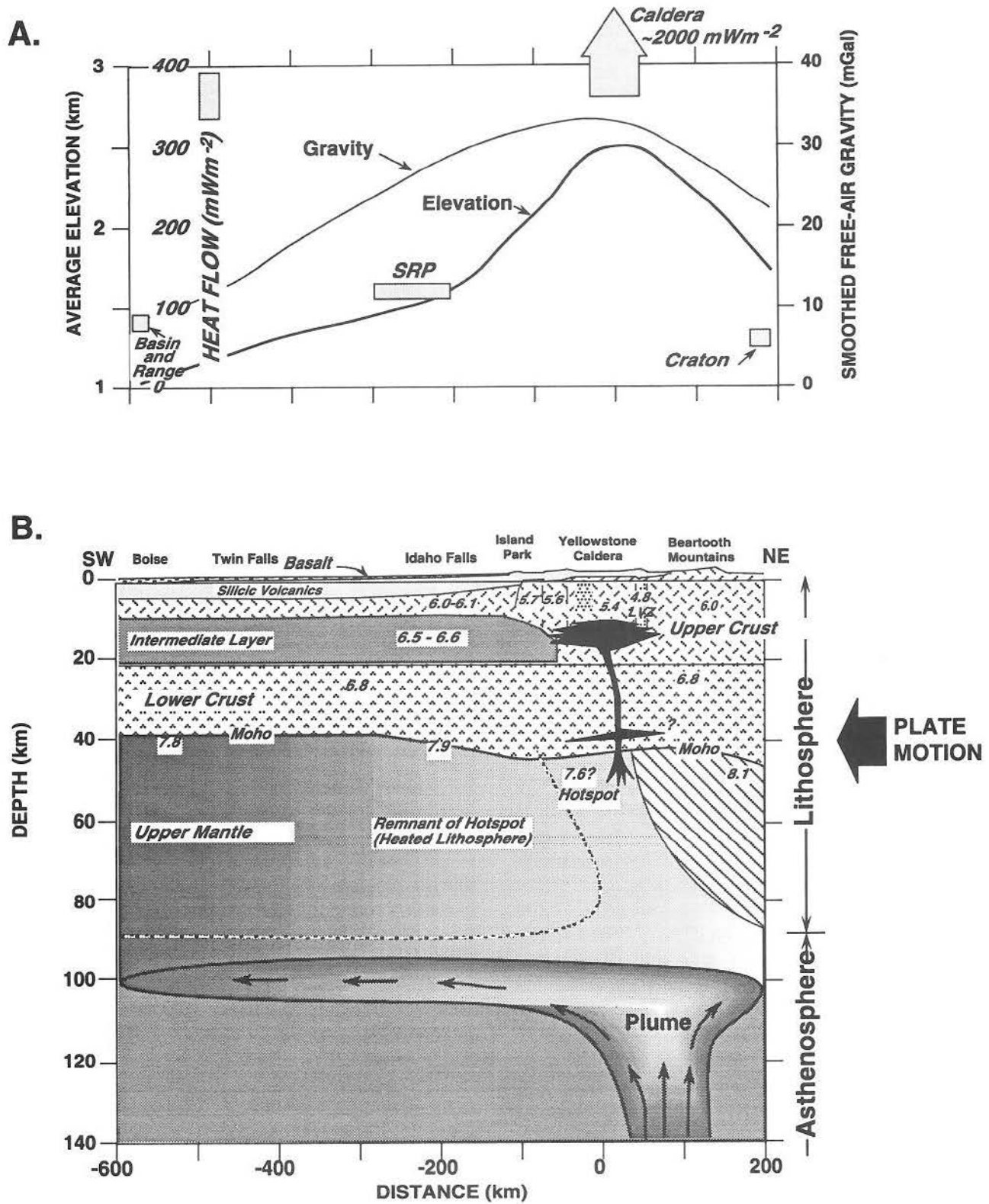
In the first, or pre-caldera phase, parental basaltic magmas originate in an upper mantle “keel” (defined by Leeman, 1982) that is thought to provide a coupling between the hotspot and crust. The keel

corresponds to the trailing portion of our plume model. The combined effects of heating and partial melting reduce P-wave velocities to ~ 7.8 km/s, compared to 8.0 - 8.1 km/s upper mantle velocities determined beneath the stable craton (Braile and others, 1989). At this stage, ascent of basaltic magmas into the upper mantle begins to produce long-wavelength lithospheric tumescence.

Eventually, a heat flux of $\sim 2,000$ mWm⁻² is available in the crust to produce melting. This corresponds to accelerated tumescence of the lithosphere. Mafic magma bodies likely stagnate below the crust because of density variations (**Figures 22**, **33**, and **34**), then ascend through or into the lower crust (Hildreth and others, 1991) in response to buoyancy and extensional pressure release at the base of the upper crust. These bodies act as localized heat sources, which in turn produce partial melting by pressure release, yielding the characteristic rhyolitic magmas of the YSRP.

Yellowstone phase

Beginning the “Yellowstone” phase, the rise of rhyolitic magmas forms batholith-scale bodies in the upper crust (**Figure 21**), perhaps as a single batholith with two near-surface cupolas manifest by the two resurgent domes in Yellowstone (Christiansen,



1984). The upper-crustal batholith in turn produces ring-fracture zones, a process that is accentuated by the superimposed northeast Basin and Range regional extension. Seismicity is accelerated at shallow depths, where heat flow is still sufficiently low to permit brittle failure. As rhyolitic magmas ascend, thermal expansion and magma transport produce crustal uplift on a caldera-wide scale. The batholithic basement is evidenced by the low P-wave velocities (4.8 to 5.6 km/s) and concomitant low-density zones that are interpreted as rhyolitic and basaltic magmas, solidified but hot granitic rocks, partial melts, and hydrothermal fluids. These are in contrast to 6.0 km/s for the thermally undisturbed and colder basement of the surrounding Rocky Mountains. The low-velocity bodies extend to depths of about 10 km and are primarily responsible for the negative gravity anomaly of the Yellowstone Plateau and generally reflect the location of Yellowstone's batholith. Localized crustal deformation, like that measured historically in Yellowstone, reflects ephemeral episodes of uplift and subsidence associated with the migration of magma and hydrothermal fluids in the upper crust. Earthquakes occur along crustal zones of weakness (Figure 23), such as pre-existing fault zones or alignments of volcanic vents with high temperatures, limiting the depth of brittle failure within the caldera to less than ~5 km (Figure 25).

During the climactic stage of the "Yellowstone" phase of silicic volcanism, violent caldera-scale eruptions are produced when shallow magma chambers intersect pre-existing faults or other zone of weakness. Explosive volcanism is likely exacerbated by ruptures accompanying large earthquakes that reduce the confining pressure and produce rapid gas exsolution at a rate greater than can be physically accommodated. This results in violent pyroclastic, caldera-forming eruptions with concomitant loss of roof support and caldera roof collapse.

As the siliceous magmas cool, mafic magmas are erupted at the margins of the caldera, which continue to subside. Residual magma bodies and subsidence of the caldera continue to feed the resurgent domes. Magmatic and hydrothermal fluids beneath or near Yellowstone's domes fit our models of quasi-plastic flow and source volumes responsible for Yellowstone's modern crustal deformation (Figures 25 and 27), as well as the localized gravity and seismic-velocity anomalies. Ultimately, cooling and

fracturing of the Yellowstone batholith allows the parental basalts to flood the surface.

Snake River Plain phase

The last phase of YSRP volcanism is hypothesized as the "Snake River Plain" stage (Figures 17 and 34). In this stage, thermal contraction, surface loading from episodic basaltic magmatism, and subsurface loading associated with the high-density mid-crustal body produce the notable topographic decay of the eastern SRP. Heat flow is reduced to background values of ~100 mWm⁻². The velocity structure of the upper crust in this late stage reflects the characteristic bimodal volcanism manifested by a near-surface basalt layer (4.9 km/s) underlain by a rhyolite layer (5.3 km/s) (Figure 29). The anomalous, high-velocity (6.5 km/sec), high-density intermediate crustal layer is interpreted as a mafic remnant of the crustal differentiation process and is primarily responsible for the positive gravity anomaly of the SRP.

We note that upper-mantle teleseismic P-wave and density models do not resolve physical properties to depths greater than ~250 km. Thus our model (Figure 25) does not necessarily specify the depth of origin of the Yellowstone hotspot, it simply accounts for a mantle source of magmatism and heat.

Alternate hypotheses for the origin of Yellowstone

Several hypotheses have been variously suggested for the origin of the YSRP, varying from mantle plume models to lithospheric fracturing. Eaton and others (1975) attributed the mechanism for the YSRP to mantle-derived magmas, which ascended along a northeast-striking, pre-existing zone of weakness identified by subtle trends in regional magnetic anomalies. Morgan (1972) and Smith and Sbar (1974) proposed plume models to account for the progressive volcanism of the YSRP. Smith (1977) proposed a regional model of a dual-branched—northeast (the YSRP) and northwest (into the High Lava Plains of Oregon)—rift emanating from a mantle bulge, continental triple junction, in north-central Nevada. The rifting model also resembles the concept of "unzipping" along a lithospheric fracture (see Furlong, 1979), with buoyant ascent of mantle magmas to account for the systematic variations of

silicic volcanism for the YSRP. However, the models of Smith (1977) and Draper (1991) require northwest progression of silicic volcanic centers into the southern Oregon volcanic province, but this silicic zone of volcanism does not have a well-defined age progression of silicic centers nor the volume of volcanic rocks that would be expected.

Christiansen and McKee (1978) suggested that the silicic volcanic age progression of the YSRP originated along a transform fault boundary with right-lateral offset. However, there is little evidence for strike-slip offset along the boundary of the SRP. Furthermore, the direction of regional northwest Basin and Range extension, i.e., σ_{hmin} , has persisted since about 17 Ma and is inconsistent with the proposed strike-slip offset. Christiansen and McKee (1978), while agreeing with the idea of a melting anomaly at Yellowstone, argued against a deep mantle plume and suggested that concentrations of radioactive minerals in the upper mantle could produce the required heat to drive the YSRP. Our plume-plate model cannot differentiate a shallow or deep (core-mantle boundary) plume. On the other hand, the idea of concentrations of radioactive minerals in the mantle is speculative and untested.

Several authors have attributed the origin of the Miocene Columbia River flood basalts to the initial development of the Yellowstone plume in southeastern Oregon. For example, Thompson (1977) suggested shallow-rooted mantle diapiric upwelling for the origin of the YSRP, beginning as early as Oligocene, which he suggested was also the source of the basaltic volcanism of the Columbia Plateau. White and McKenzie (1989) characterize flood basalts, such as in the Columbia Plateau, as a passive response to rifting or extension above the head of a mantle plume that spreads out beneath the lithosphere. This model requires a dimension larger than 200 km, indeed as large as 1,000 km, and we believe there is little evidence now for such a large-scale feature. Thompson and Gibson (1991) recently suggested that the Miocene volcanism of the Columbia Plateau was the result of a lithospheric "thinspot" that may have originated when the Yellowstone "plume" was located beneath southwestern Oregon and western Idaho at ~ 17-15 Ma. However, these authors could

not link the narrow and linear volcanic pattern of the YSRP to the hypothesized thinspot. Draper (1991) suggested that the Miocene basalts of southeastern Oregon originated from a shallow plume source initiated at 18 Ma. He failed to note that the Columbia Plateau magmatism declined at the time when silicic volcanism of the Oregon lava plains began to propagate.

Duncan (1982), Thompson and Gibson (1991), and Pierce and Morgan (1990, 1992) link the origin of the voluminous Columbia river basalts with the initiation of the Yellowstone plume about 16 Ma. However, plate reconstructions are inconsistent with the hotspot track implied by these models. For this reason, Thompson and Gibson (1991) suggested that a pre-existing lithospheric "thinspot" at the location of the Columbia Basin may have allowed the rising Yellowstone plume head to spread laterally as it reached the base of the lithosphere and produce the Columbia River basalt eruptions away from the axis of the plume.

Despite the unknown, deep source of the Yellowstone hotspot, the uncertainty about the nature of the initial interaction of the hotspot with the North American plate before 16 Ma, and the unclear relationship between the Yellowstone hotspot and the Columbia River basalts, we believe that the mantle plume-plate model described in this paper and schematically illustrated in Figures 34 and 35 currently represents the best working model for the late Cenozoic evolution of the YSRP tectono-magmatic system. Although there is much to learn about the exact mechanisms of plume interactions with the continental lithosphere, the model explains the main characteristics of the YSRP system including: (1) a source of intense heat, mantle-derived basalts, and siliceous partial melts; (2) the systematic age progressions of silicic volcanism, topographic bulging and subsidence of the Yellowstone caldera along with topographic decay of the SRP; (3) plate reconstructions for which we can infer the former positions of the Yellowstone hotspot; (4) systematic variations in crustal velocity structure along the YSRP; and (5) the parabolic pattern of seismicity and topography observed in the YSRP region.

Concluding remarks

We conclude by discussing the stress field and the possibility of magmatic sources at Yellowstone and speculating on the future of volcanism and earthquakes in this region.

Extensional stress field at Yellowstone

We have discussed the evidence that the historic uplift of the Yellowstone caldera has a magmatic or hydrothermal origin. These observations suggest an important question about the long-term processes of Yellowstone; namely, what is the regional pattern of crustal deformation excluding the influence of the hotspot? To examine one possible scenario, Meertens and others (1989) constructed three-dimensional, finite-element models of the Yellowstone-Hebgen Lake region assuming an elastic layer over a fluid half-space with the Yellowstone caldera as a soft inclusion that was loaded by tectonic and volcanic mechanisms. Their modeling suggested that the occurrence of the 1959, $M_s = 7.5$ Hebgen Lake earthquake may have stressed the Yellowstone caldera, causing uplift within a region dominated by crustal extension. This mechanism is consistent with the modeling of the 1987 to 1991 caldera subsidence (Meertens and others, 1992), which requires both a component of caldera subsidence at 1 to 2 cm/yr and the superposition of regional northeast extension.

Additional evidence for long-term crustal extension is given by Peyton's (1991) study of the stress field by inversion of focal mechanisms. Her study revealed a dominant north-northeast direction of σ_{hmin} , which is also consistent with the strain directions implied by trilateration networks in the same region (see Dzurisin and others, 1990; Savage and others, 1993) and by the orthogonal direction to the alignment of volcanic vents associated with the post-caldera rhyolite flows (see Figure 13).

These observations suggest the Yellowstone Plateau has been dominated by northeast crustal extension that has been active since at least the time of the volcanic vent emplacements 0.6 Ma. On the basis of evidence for long-term regional extension from these geologic and geophysical indicators, we thus believe that the northeast-trending extensional stress field has persisted for at least the last 10 million years and

that the Yellowstone Plateau has been largely influenced by Basin and Range extension that continues today.

However, the stress field of Yellowstone must be episodically interrupted by tectonic events (large earthquakes) and magmatic events (eruptive and non-eruptive emplacements of magmas and hydrothermal fluids), which produce caldera-wide uplift, such as that observed historically. Nonetheless, we believe that these short-term perturbations are followed by relaxation and continued extension with resulting caldera subsidence.

Present-day magmatic fluids at Yellowstone

For Yellowstone, one of the key issues implied by the Quaternary silicic volcanic record (Christiansen, 1984; 1993) is the extent and location of magma sources. Geophysical and geodetic surveys by their nature, provide inferential information on physical properties and hence on locations of magmatic systems. However, these models do not uniquely define the locations of the hypothesized magmatic systems, but rather provide constraints on plausible models. We suggest that the $\sim 2,000 \text{ mWm}^{-2}$ thermal flux is the key parameter that characterizes the evolution of Yellowstone's volcanic system, because heat flow of this magnitude must originate from the upper mantle, implying a long-lived magmatic process.

Direct evidence for shallow magmatic/hydrothermal fluids at Yellowstone is the unprecedented 1923-1984 period of uplift totaling about 1 m followed by subsidence of about 25 cm from 1985 to 1991. The synchronicity of caldera-wide uplift and the rapid change to subsidence across the entire caldera is the best evidence for a shallow, connected source zone. Vasco and others (1990) suggested that low-viscosity fluids such as hydrothermal brines or gases associated with deeper magmatic sources may be transported through an upper-crustal conduit system along the entire 75 km long Yellowstone caldera or connected between the two cupolas underlying the resurgent domes. This idea is consistent with modeling of the crustal deformation data,

which places the source of 1923 to 1977 uplift in the 3 to 6 km depth range near the resurgent domes, whereas the 1987 to 1991 subsidence was modeled in the 3 to 6 km depth zone along the entire caldera. A magma or partial melt may have invaded the deeper zone producing crustal uplift, then generating hydrothermal fluids and gases.

In an earlier section, we showed the distribution of excess mass that exceeded the Curie temperature. The Curie depth anomaly on the southwest side of the caldera appears to reflect bodies with upper crustal sources, implying hot material in the upper crust at 6 to 8 km depths. This area also corresponds to the youngest (150 Ka to 70 Ka) post-collapse caldera flows of Yellowstone (Christiansen, 1984) and the circular nature of this anomaly around the caldera rim suggests that it could represent ring dikes or bodies that follow the inner-caldera rim fracture zone. Together with the upper crustal low-velocity body in the northeast caldera, these observations suggest the presence of magmatic and hydrothermal or highly altered rock bodies in the northeast caldera, near the Sour Creek dome, around the southwest caldera rim, and near the Mallard Lake dome. The continuous nature of the 1974 to 1991 subsidence of the entire central caldera also argues for a connection between the northern and southern caldera, perhaps zones of hydrothermal fluid flow.

Future of volcanism at Yellowstone

We close by speculating on the future of tectonic and magmatic activity of the Yellowstone Plateau. Smith and Christiansen (1980) and Christiansen (1984) pointed out that there is little evidence that Yellowstone's volcanism has ceased and, therefore, that volcanic eruptions are quite likely in the future. The record of unprecedented modern crustal deformation of the Yellowstone caldera and its Quaternary record of rather continuous silicic volcanism, interrupted by the caldera-forming explosions, reflects an intense and long-lived heat source that drives the Yellowstone system. While the anomalous crustal properties do not indicate what form the volcanism might take, Smith and Christiansen (1980) suggest it could range from small phreatic and hydrothermal explosions throughout the caldera, to eruptions of small rhyolite and basalt flows at the margins of the caldera, to medium-sized eruptions of rhyolitic lava within the caldera, and finally, to cata-

strophic explosive eruptions of ash. While we do not know where we are in the present cycle, Christiansen (1993) suggests that we may be in the end of the third cycle or in the beginning of a new fourth cycle.

Note that the total volume of source material modeled for the 1923 to 1976 caldera uplift episode is on the order of about 1 km³, which is about the same volume as that expelled in the historic 1980, Mt. St. Helens explosive eruption. Thus the storage chamber(s) for the historic uplift of the Yellowstone caldera reflects a sufficiently large source to produce a sizable volcanic or phreatic eruption. However, we point out that the historic crustal deformation and inferred source models do not necessarily imply imminent volcanic eruptions at Yellowstone. Nonetheless, the Quaternary volcanic record, the extremely high heat flow, extensive hydrothermal activity, active crustal deformation, and the historic record of large earthquakes warrant systematic and continued monitoring of the Yellowstone region.

We feel that any renewal of volcanism at Yellowstone would be preceded by increased earthquake activity, increased rates of crustal deformation, and increased gas emissions—all phenomena that can be monitored with modern surveillance equipment. Surveillance for long- and short-term volcanic and earthquake precursors and related hazards should include systematic geodetic measurements (GPS, leveling, and gravity), earthquake monitoring, and monitoring of gases and hydrothermal fluids.

We also suggest that large earthquakes, $M_s \leq 7.5$, could occur on the periphery of the Yellowstone Plateau with intense earthquake swarms and $M_s \leq 6.5$ earthquakes around the periphery of the caldera and shallow earthquakes within the caldera. The modern record of seismicity and hydrothermal activity suggests that Yellowstone's geysers, hot springs, and fumaroles will continue to evolve, but may also be extinguished by related volcano-tectonic processes and strong ground shaking accompanying earthquakes. There is no reason to believe that uplift and subsidence of the caldera on the scale measured in historic time will not continue. Over longer periods—tens of millions of years—we suggest that lithospheric tumescence, magmatism, and seismicity associated with the Yellowstone hotspot will progress to the northeast, affecting southern Montana.

We close by quoting T. A. Jaggar (1922), the founder of the Hawaiian Volcano Observatory, who visited Yellowstone in 1920 by horseback and whose keen observations noted the dynamic processes so well manifested in Yellowstone:

Anyone who has spent summers with pack-train in a place like Yellowstone comes to know the land to be leaping....The mountains are falling all the time and by millions of tons. Some-

thing underground is shoving them up....How much is the ground tilting? Jaggar (1922)

How prophetic are Jaggar's words. We now recognize Yellowstone as one of the world's largest and most active calderas and one that we believe owes its existence to the interaction between a mantle plume and the overriding North American plate. Indeed, Yellowstone deserves recognition as the "The Yellowstone National Hotspot."

Acknowledgments

This paper presents the findings of multiple investigations made by the University of Utah in Yellowstone and studies by Purdue University in the Snake River Plain with related work from other researchers. We are particularly grateful for cooperative ties made with scientists of the U. S. Geological Survey; in particular, for the fruitful collaborations with R. L. Christiansen, who generously assisted in integrating our geophysical and geodetic data with his volcanological findings.

Funds for the crustal deformation surveys were provided by the National Science Foundation, research grants EAR 86-8618341 and EAR 89-04473 to the University of Utah. The U. S. Geological Survey, Volcano Hazards and Geothermal Research program, grant 1434-92-A-0975 supports the recording and maintenance of the Yellowstone Seismograph Network at the University of Utah Seismograph Stations. Partial support for this research was also furnished by NSF grants, EAR 77-23357 and EAR 84-01330, and by U. S. Geological Survey grant #14-08-0001-G674 to Purdue University.

We gratefully acknowledge the permission and excellent cooperation of the National Park Service and the personnel of Yellowstone National Park and their long-term support and assistance. We especially thank Superintendent Robert Barbee and members of his staff: John Varley, Rick Hutchinson, Wayne Hamilton, and John Lounsbury. Many other Yellowstone National Park rangers, naturalists, and staff members have also assisted us, and we thank them all.

Several students and staff from the University of Utah have worked on projects in Yellowstone and

surrounding areas since 1971 and contributed to this paper. Natalie Kelsey materially assisted with the preparation of the paper, especially for drafting of the figures using computer graphics, for editing, and for general assistance. Sue Nava assisted with the preparation of the earthquake data. Anthony Lowry assisted with the color plotting. Walter Nagy was particularly helpful with preparation and interpretation of the earthquake data and the development of the color graphics. We are indebted to many other University of Utah graduate students who have worked on Yellowstone and related projects including: Al Trimble, Terry Daw, Jeff Evoy, Bob Otis, Jim Bailey, Mike Schilly, Jay Lehman, Harley Benz, Diane Doser, Mark Brokaw, Steve Clawson, Walter Nagy, Lynn Peyton, Bill Hardman, and John Byrd. Dan Trentman assisted with computer matters such as data bases, earthquake interpretation, and general computations. Erwin McPherson, Walt Arabasz, Linda Hall, and Sue Nava contributed to technical support and management of Yellowstone Seismograph Network through their affiliation with the University of Utah Seismograph Stations. We also thank graduate students Greg Elbring, Mark Baker, Keith Peregrine, Carl Daudt, and Mark Sparlin from Purdue University for their assistance on the Snake River Plain seismic experiments. Donald Browne pointed out the original references pertaining to the quote of Lt. Gustavus C. Doane.

Neil Ribe contributed the computer models of plume-plate interactions. Chuck Meertens was involved with the Global Positioning Satellite measurements and modeling of crustal deformation. Dave Blackwell cooperated in the heat flow measurements of Yellowstone Lake, and with thermal interpretations of the Snake River Plain, and made a pre-

print of unpublished work available. Dan Dzurisin provided information on crustal deformation from published and unpublished leveling data. In addition, we thank Robert Fournier, Mitch Pitt, Ken Pierce, and Dave Love of the U. S. Geological Survey for contributing their knowledge and data on Yellowstone. Mark Anders and Ken Pierce kindly provided preprints of unpublished work. Charles

DeMets computed plate motions for various NUVEL-1 models. Gene Humphreys and Ken Dueker provided unpublished data on the lithospheric velocity structure of the YSRP. David Fountain, Tom Hanks, David Blackwell, and Robert Fournier provided critical and constructive reviews of the manuscript, which are much appreciated.

Appendices

I. Ages of silicic volcanic fields of the Yellowstone-Snake River Plain region, with distances from Yellowstone.

Data from Armstrong and others (1975), Morgan and others (1984), Rodgers and others (1990), Pierce and Morgan (1990, 1992), and Mel Kuntz (U. S. Geological Survey, personal communication, 1985).

Volcanic field	Caldera	Volcanic unit	Age (Ma)	Distance (km)	Error (km)	Symbol
Yellowstone, Wyo.	Yellowstone	Lava Creek Tuff	0.6	45	50	Y
Yellowstone, Wyo.	Henry's Fork	Mesa Falls Tuff	1.3	75	15	HF
Yellowstone, Wyo.	Huckleberry Ridge	Huckleberry Ridge Tuff	2.1	70	80	HR
Heise, Ida.	Kilgore	Kilgore Tuff	4.3	150	70	K
Heise, Ida.	Blue Creek	Walcott Tuff	6.0	220	70	BC
Heise, Ida.	Blacktail Creek	Blacktail Ck. Tuff	6.5	190	110	B
Picabo, Ida.	Blackfoot	Arbon Valley Tuff	10.3	300	110	P
Twin Falls, Ida.	Not specifically identified	multiple tuffs	10.9 (average)	400	100	TF
Bruneau-Jarbridge, Ida.	Not specifically identified	Cougar Point tuff	11.3	500	80	BJ
Owyhee, Ida.-Nev.	Juniper Mtn	multiple tuffs	13.8	630	70	O
McDermitt, Nev.	multiple	multiple tuffs	16.1	690	100	M

II. Inception of extension along Yellowstone-Snake River Plain system, with distances from Yellowstone

Data taken from Rodgers and others (1990, GSA Repository Item # 90-28, Supplementary data), excluding data from the Teton fault. Average age and error estimate (this paper).

Location of extensional basin	Inception age (Ma)	Cessation age (Ma)	Average (Ma)	Error (Ma)	Distance (km)	Symbol
Swan Valley, Ida.	2	4.4	3.2	2.4	80	S
Blackfoot Mtn, Ida.	4.9	5.9	5.4	1	120	BI
Buckskin Mtn, Ida.	6.7	10.3	8.5	3.6	150	Bu
Raft River., Ida.	7.2	11	9.1	3.8	250	RR
Goose Ck., Ida-Utah.	8.8	13	10.9	4.2	310	GoC
Grouse Ck., Ida-Utah	11.6	14.4	13	2.8	340	GrC
Mary's River., Nev.	12.5	17	14.75	4.5	440	MR
Northern Nev.	14	17	15.5	3	600	NR

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