

Chapter 9

Quaternary geology and structural history of the Snake River Plain, Idaho and Oregon

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INTRODUCTION

The Snake River Plain is a major late Cenozoic tectonic feature in the North American crust. The plain consists of distinctive western and eastern parts, which differ in structural trend and geology: a northwest-trending sedimentary basin in the west, and a northeast-trending volcanic plain in the east (Fig. 1). Structurally, the western Snake River Plain separates the Cretaceous Idaho batholith of west-central Idaho from outliers of the batholith in southwestern Idaho, and the eastern Snake River Plain interrupts the continuity of mountain ranges and valleys of the Basin and Range Province. The Snake River Plain is entirely covered by Cenozoic volcanic and sedimentary deposits, of which the exposed material is largely of Quaternary age.

The Snake River Plain is an area of subdued, locally almost featureless, relief surrounded by mountains and highlands. The plain is as much as 70 km wide in the western part, 90 to 100 km wide in the eastern part, and about 600 km long. Its surface rises with a gradually increasing gradient from altitudes of less than 700 m at the western end to about 2,000 m at the eastern end.

The geology of substantial parts of the Snake River Plain has been studied in moderate detail, a chronologic framework based on radiometric ages of the volcanic rocks is fairly well established, and considerable geophysical data pertinent to the crustal structure has been gathered—including seismic profiles, gravity and aeromagnetic surveys, magnetotelluric and resistivity soundings, and measurements of thermal gradients and heat flow. Furthermore, the subsurface has been probed by drilling several deep exploratory holes, of which one reached a depth of 4.3 km. Other subsurface information has been obtained from thousands of water wells, most of them being less than 300 m deep.

Despite information obtained from geologic, chronologic, and geophysical studies, the origin of the Snake River Plain is inadequately understood. No unified explanation of why the plain exists has been proposed. For many years, perhaps because of physiographic continuity and perhaps because so much of the Snake River Plain is covered by distinctive olivine basalt, the plain has been referred to as a single geologic province—as if to imply that its structure and geology reflect one set of forces and

processes from end to end. In the last decade, however, a concept has emerged that the western and eastern parts of the Snake River Plain represent different entities, structurally and geologically. The two parts of the plain also are now thought to differ appreciably in age.

The western Snake River Plain is a structural basin, bounded by high-angle faults and filled with sedimentary deposits as much as 1.7 km thick. The sediments are underlain by a thick slab of basalt and by still deeper tuffaceous sediments and silicic and intermediate volcanic rocks. The basalt slab probably was emplaced at a time of rifting, which may have begun as early as 17 Ma. Judging from the distribution of certain ash-flow tuffs, parts of the western plain also may have been sites of silicic eruptions at about 11 Ma.

The sedimentary deposits began to accumulate soon after these silicic volcanic rocks erupted but decreased in age to the southeast as the rift continued to grow headward. Where the head of the western Snake River Plain is overlapped by basalt of the eastern plain, the sedimentary deposits have ages of about 3 Ma; but sediments were deposited farther west until about 2 Ma. Quaternary deposits in the western Snake River Plain mainly are represented by a veneer of basalt on the basin sediments, by lake beds impounded by lava dams in several ancestral canyons of the Snake River, and by various pediments and alluvial terraces. The Snake River in the western Snake River Plain is incised to depths of 150 to 300 m, forming steep-walled canyons in the basaltic rocks and strongly gullied badlands in the sediments.

The eastern Snake River Plain is a bimodal volcanic province, characterized by rhyolitic volcanic rocks covered by basaltic lava flows. The rhyolite extends to depths of at least 3 km, and the overlying basalt is about 1 km thick. The rhyolite and basalt are thought to be products of a mantle plume, or hot spot (Vink and others, 1985), which produced a belt of volcanic terrain extending from southeastern Oregon and adjacent Nevada to the Yellowstone Plateau, as the North American plate drifted southwestward between 17.5 and 0.6 Ma. In southwestern Idaho, the volcanic terrain produced by the hot spot is a broad swath of

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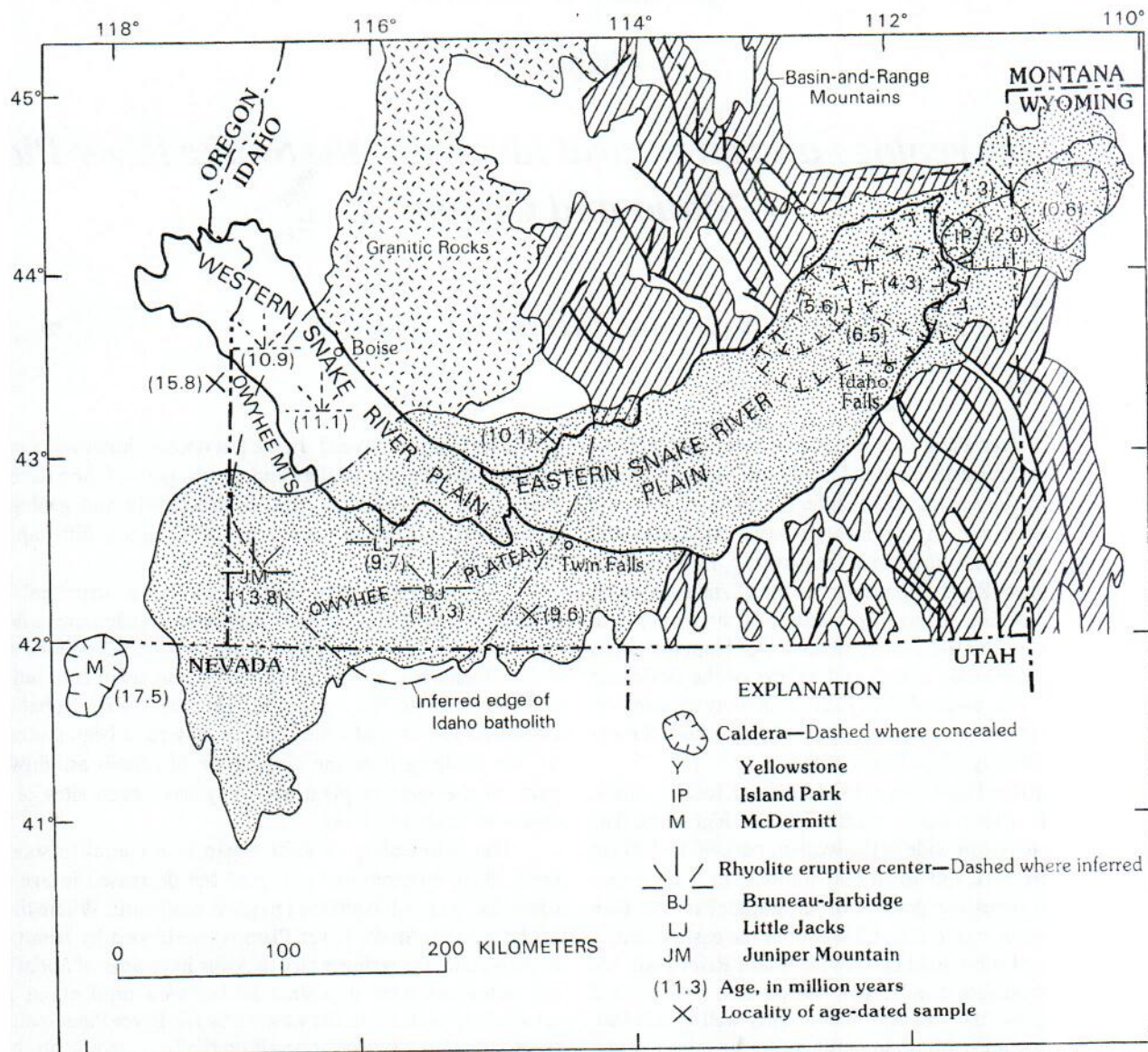


Figure 1. Map of principal features associated with the western and eastern Snake River Plain. Area stippled in blue represents rhyolite and basalt of the eastern Snake River Plain trend, as described in the text. Heavy lines are Basin and Range faults. Adapted from Taubeneck (1971), King and Beikman (1974), Ekren and others (1984), Morgan and others (1984), and Christiansen (1984).

rhyolite thinly veneered with basalt, which together form an upland called the Owyhee Plateau. The terrain produced by the hot spot has been named informally the "eastern Snake River Plain trend" (Ekren and others, 1984). This trend coincides with part of a major northeast-trending magnetic lineament, called the Humboldt zone, which is inferred to be a regional flaw in the Precambrian crust (Mabey and others, 1978). Rhyolite of the eastern Snake River Plain trend in the Owyhee Plateau ranges in age from 13.8 Ma at the southwest to 9.6 Ma at the northeast. The overlying basalt also decreases in age to the northeast and ulti-

mately is buried by the basalt of the eastern Snake River Plain. Exposed basalt in the eastern Snake River Plain is largely, or entirely, of Quaternary age. The youthful age of the basalt of the eastern plain reflects widespread eruptions along volcanic rift zones perpendicular to the axis of the plain. Such zones are aligned with high-angle faults at the fronts of mountain ranges to the north and south. The eastern Snake River Plain is undissected, except where incised by the Snake River downstream from Twin Falls.

Where the Owyhee Plateau intersects the western Snake

River Plain, roughly between longitude 116° and 115° W, down-faulted volcanic rocks of the plateau are overlapped by sedimentary deposits of the western basin. The geologic relations imply that the sedimentary basin of the western Snake River Plain grew headward into the volcanic terrain that had been produced by passage over the hot spot.

Because of the marked geologic contrasts between the western Snake River Plain and the eastern Snake River Plain, these regions are described separately in the review that follows. Although such contrasts are expressed by many Quaternary features, differences between the western and eastern parts of the Snake River Plain are partly a legacy of late Tertiary structural events. Thus, the description of each part includes summaries of its structural features. Structural aspects of the Snake River Plain are considered in light of recent reviews of its geology and geophysics, which have been partly conceived in the context of current regional studies of tectonic forces, intraplate motions, and volcanic history. Especially noteworthy are reviews of the geophysics and tectonics of the Snake River Plain (Mabey, 1982), the petrochemistry of its volcanic rocks (Leeman, 1982a), seismic evidence for the structure of its crust and upper mantle (Iyer, 1984a), and subsurface features near Boise (Wood and Anderson, 1981; Wood, 1984). The discussion begins by describing the binding seam of the Snake River Plain—the Snake River itself.

QUATERNARY AGE OF THE SNAKE RIVER

Present course of the river

The course of the Snake River as seen today is a Quaternary feature (Fig. 2). From its origin near Jackson Hole, Wyoming, the river emerges from the mountains of southeastern Idaho and takes a path along the southern edge of the Quaternary basalt that forms the surface of the eastern Snake River Plain. The principal upper tributary of the Snake, the Henrys Fork, follows a similar path. The basalt is slightly higher topographically than the margins of the eastern plain, and buildup of the basalt clearly has determined the river's course.

At Milner Dam, the river becomes incised in a narrow canyon 50 m deep, which abruptly deepens to 150 m and becomes wider at Shoshone Falls, near the city of Twin Falls. As explained below, when describing displacements of the ancestral Snake River in the eastern Snake River Plain, the narrow canyon between Milner Dam and Shoshone Falls is a youthful stretch of the Snake River, which was established when a former canyon of the river about 10 km north was filled with late Quaternary basalt. The present canyon below Shoshone Falls is partly a relic of the former canyon, although it was enlarged and extended headward by passage of the late Pleistocene Bonneville Flood (this catastrophic flood is discussed at the end of this section). The canyon from Shoshone Falls to King Hill marks the boundary between Quaternary basalt of the eastern Snake River Plain and an upland of older basalt and sedimentary deposits south and west of the Snake River. Thus, the course of the river in this area was determined by lava flows that form the northern rim of the canyon.

After leaving the eastern Snake River Plain, the river enters the sedimentary basin of the western plain. Here, the course of the river is limited on the south and southwest by an upland veneered with gravel of late Pliocene and early Quaternary age, and on the north and northeast by an upland of Quaternary basalt. That is, as in the eastern Snake River Plain, the river generally follows the southern edge of the basalt. The basalt had a decisive role in causing several minor changes in the early Quaternary course of the Snake River (see discussion of Bruneau Formation).

In the westernmost part of the western Snake River Plain, the river turns north and enters a mountainous stretch that leads to Hells Canyon, one of the deepest gorges in North America. The river then turns west and joins the Columbia, thereby making the Snake River a branch of the Columbia River system.

Capture of the Snake River by the Columbia River

The physiographic connection of the Snake River with the Columbia River is a result of capture of the Snake River, supposedly at an anomalous bend at the head of Hells Canyon called the Oxbow (Wheeler and Cook, 1954). The circumstances of the capture and its timing with respect to the stratigraphy and history of the Snake River Plain are still incompletely understood, but available information implies that capture took place close to the onset of Quaternary time, at about 2 Ma.

Wheeler and Cook (1954) concluded that capture of the Snake River by the Columbia was accomplished by spillover of "Idaho Lake," the name then given to the site of deposition of the sedimentary deposits in the western Snake River Plain. The height of spillover was estimated to be at an altitude of 1,005 m, which was said to coincide with "high-level outwash gravel at several localities in southwest Idaho and eastern Oregon" (Cook and Larrison, 1954). The possible relations of such gravel to the history of the Snake River and Hells Canyon have not been worked out but currently are being studied by K. L. Othberg, Idaho Geological Survey. Some of the high-level gravel in the western Snake River Plain is known by the names Tuana Gravel and Tenmile Gravel (Fig. 2).

The Tuana and Tenmile Gravels are the first obvious signs in the geologic record of the vigorous Snake River that we know today. That is, when compared with underlying fine-grained deposits, these gravels clearly represent a high-energy drainage system, graded to an unobstructed outlet—presumably Hells Canyon. The physiographic relations and ages of the Tuana and Tenmile Gravels are thus considered to be important evidence for deducing the time of capture of the Snake River. Whether the Tuana and Tenmile Gravels are contemporaneous is unknown, but both appear to be graded to a base level below the inferred level of spillover at the head of Hells Canyon.

Tuana Gravel is identified east of longitude 116° W, on benches about 245 m above the Snake River (Malde and others, 1963). The gravel came from sources to the south, near the Idaho-Nevada border, and beveled Pliocene deposits of the Glens Ferry Formation (see the discussion of general

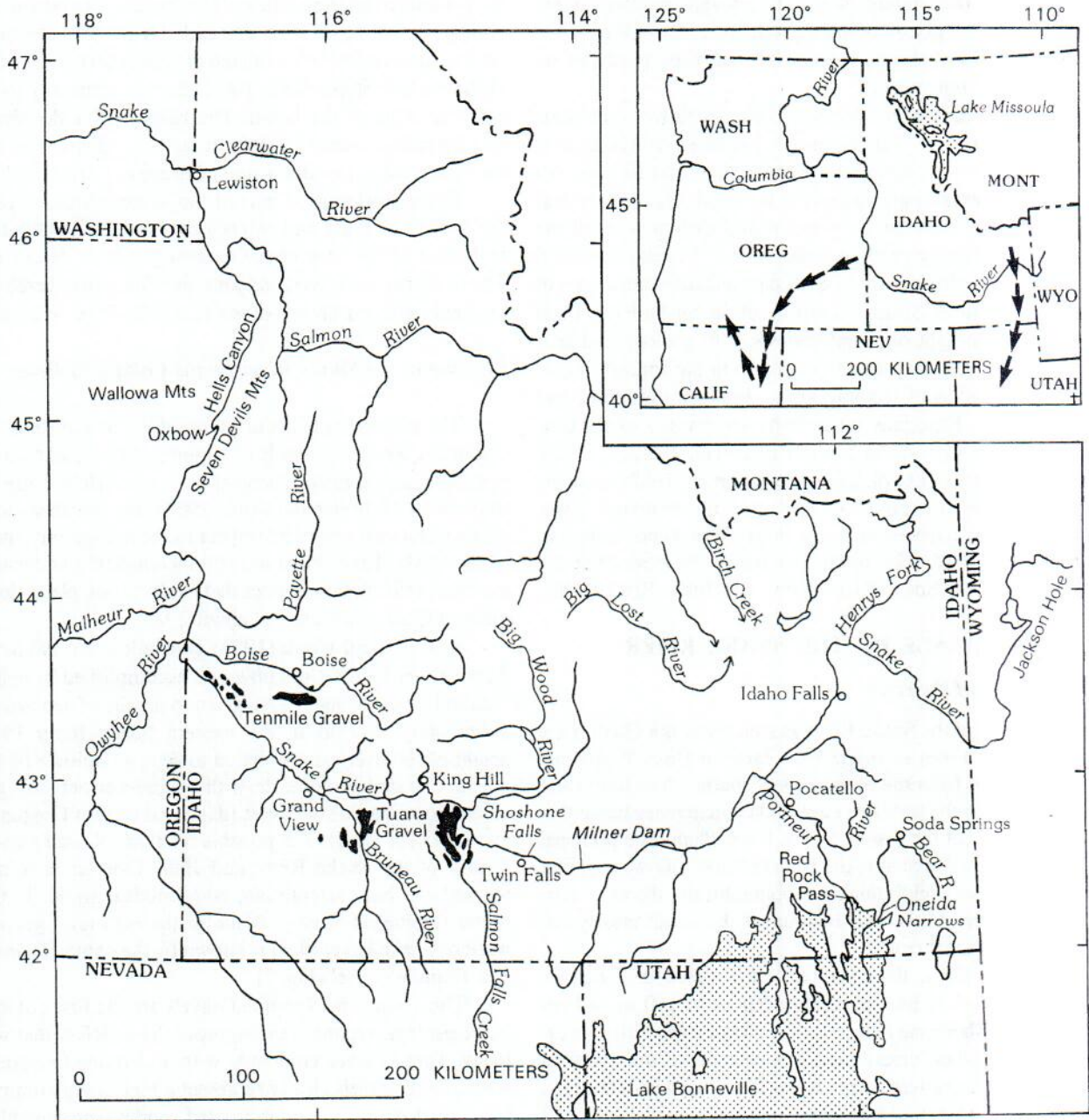


Figure 2. Map of the Snake River drainage system and the northern part of Pleistocene Lake Bonneville at its last maximum stand. Tuana Gravel and Tenmile Gravel in uplands of the western Snake River Plain were deposited at about 2 Ma, just after the Snake River was captured by the Columbia River, probably at the Oxbow. Inset map shows Pleistocene Lake Missoula and routes of Pliocene drainage (blue arrows) inferred by Taylor (1985) from biogeographic evidence.

features of the western Snake River Plain). Its distribution suggests deposition as alluvial fans along two ancestral drainages—ancestral Salmon Falls Creek on the east and the ancestral Bruneau River on the west. The dissected toes of these fans range in altitude from 990 m for the fan of Salmon Falls Creek to 975 m for the fan of the Bruneau River, or 15 to 30 m below the inferred height of spillover at the head of Hells Canyon.

The Tenmile Gravel, which spreads westward from the mountain front near Boise (Savage, 1958), also bevels the Glenss Ferry Formation (Wood and Anderson, 1981). The base of the Tenmile Gravel at its westernmost remnants is at an altitude of 800 m, or 120 m above the Snake River, although the topographic relations are disturbed by normal faults of small displacement (Wood and Anderson, 1981).

The Tuana and Tenmile Gravels are both older than the Bruneau Formation, which was deposited in the western Snake River Plain after the ancestral Snake River began to be entrenched to its present depth. The oldest K-Ar age for the Bruneau thought to be reliable is 1.92 ± 0.16 Ma (Amini and others, 1984, sample 79H9). In addition, the Tenmile Gravel is tentatively dated as older than 2 Ma, judging from its relation to basalts that flowed down the Boise River canyon east of Boise (Howard and others, 1982; Othberg, 1986). By this chronology, given the premise that the Tuana and Tenmile Gravels represent the action of a vigorous stream, the ancestral Snake River had an unobstructed outlet before about 2 Ma. Very likely, this outlet was the present course of the Snake River through Hells Canyon, and downcutting there at the time of the Tuana and Tenmile Gravels already was underway. No evidence has been put forward that such a course could have been established at a later time.

A limiting older date for capture of the Snake River can be deduced from the biostratigraphy and age of the Glenss Ferry Formation. The youngest part of the Glenss Ferry is preserved in a section of floodplain deposits 110 m thick at Jackass Butte, 14 km northwest of Grand View (Fig. 2), as well as in other related sections farther northwest. These deposits yield a limited fish fauna of "small, essentially modern fishes" (Smith and others, 1982) and are considered to be younger than a rich fish fauna associated with several regressive beaches in lake deposits of the Glenss Ferry Formation a short distance south. The regressive beaches, in turn, imply a drop in lake level of 120 m, perhaps because an outlet for the Snake River through Hells Canyon was being lowered by erosion (Kimmel, 1982; G. R. Smith and others, 1982). This conjecture has not been tested by further study, but it seems less plausible than linking the Tuana and Tenmile Gravels with an eroding threshold. The sedimentology of the Jackass Butte deposits, like several other sections in the Glenss Ferry Formation, implies an environment subject to overbank flooding (Middleton and others, 1985), not a stream influenced by a newly found outlet. Hence, tectonic control and other factors that could have lowered the Glenss Ferry lake are matters worth investigating.

The Glenss Ferry deposits in Jackass Butte are a little more

than 2 m.y. old. According to Repenning (1988), specimens of *Synaptomys vetus*, a fossil lemming collected by Conrad (1980) from localities 47 to 75 m below the top of the section, lived in North America between 2.6 and 2.0 Ma. Conrad also found a thin, normally magnetized zone of sediment 30 m below the top of the section, which he correlated with the Early Reunion Normal Polarity Subchron (2.14 to 2.12 Ma). These findings agree with an age of 2.18 ± 0.18 Ma for a basaltic lava flow conformable on the sedimentary deposits of Jackass Butte (Amini and others, 1984, sample 80H1). By this evidence, if the youngest part of the Glenss Ferry Formation is indeed older than the threshold at the head of Hells Canyon, capture of the Snake River by the Columbia is not older than about 2 Ma.

Biogeographic evidence of former drainage patterns

From physiographic relations, Wheeler and Cook (1954) speculated that the Snake River before its capture by the Columbia reached the Pacific Ocean by flowing through southeastern Oregon, northern Nevada, and northern California (Fig. 2). The time of diversion into Hells Canyon was considered to have been late Pleistocene because mammals in the Seven Devils Mountains are virtually identical to those in the Wallowa Mountains, across a barrier supposedly formed by Hells Canyon (Cook and Larison, 1954). In contrast, differences between the fauna of eastern Oregon and the "Owyhee desert" south of the Snake River suggested an earlier barrier to migration—that is, the former Snake River.

Biogeographic evidence of an aquatic connection between the western Snake River Plain and northern California, via southeastern Oregon, was found in the distribution of the Pliocene and Pleistocene freshwater clam, *Pisidium ultramontanum*, and in the related distribution of several relict fishes (Taylor, 1960). Since then, closely spaced linkages between the western Snake River Plain and northern California have been documented by tracing the distribution of several more species of fossil and relict mollusks and fishes (summarized by Taylor, 1985).

As a result of block faulting and volcanism, the inferred course of the former Snake River in Oregon crosses a region now characterized by internal drainage. The mollusks and fishes that define the former drainage, however, are species that lived in lakes and large streams. Thus, the habitat required by the mollusks and fishes implies a fairly stable aquatic connection with northern California. Such stability suggests that the western Snake River Plain during the Pliocene supplied a more or less continuous flow of water to the basins and connecting streams farther west. In other words, the western Snake River Plain before the time of capture by the Columbia River was an area of exterior drainage, not a closed basin. A through-flowing system of fresh water in the western Snake River Plain also is implied by a scarcity of chemical precipitates in the Pliocene sedimentary deposits (Middleton and others, 1985), apart from a few near-shore beds of oolite.

At the time when an aquatic connection was maintained

between the western Snake River Plain and California, biogeographic evidence suggests the presence of a drainage divide between the western and eastern Snake River Plain. That is, the Pliocene molluscan faunas in southwestern Idaho and in southeastern Idaho and nearby Utah indicate divergent speciation of "numerous related but distinct species" (Taylor, 1985). The fishes also differ (G. R. Smith, 1981). Hence, the Pliocene drainage of southeastern Idaho probably was southward, to the Bonneville basin, and the headwaters of the eastern Snake River Plain appear to be a Quaternary addition to the drainage basin (Fig. 2).

Biogeographic evidence for the time when the Snake River became part of the Columbia River drainage is found in the distribution of the snail, *Fisherola nuttalli* (Taylor, 1985). According to Taylor, *Fisherola* occurs first as a fossil in the Bruneau Formation. He believes it could have migrated up the swift current of Hells Canyon to the western Snake River Plain because of its streamlined, limpetlike shell.

Streams that enter the eastern Snake River Plain from the north encounter higher ground formed by lava flows of basalt and fail to reach the Snake River. Rather, they end in playas or shallow lakes at the northern margin of the plain (Nace and others, 1975). Biogeographic evidence for a former connection with the Snake River is found only in Birch Creek (Fig. 2), where three species of mollusks known in the main stem survive (Taylor, 1985). These mollusks must be relict descendants of a population that existed before the flow of Birch Creek was blocked by eruption of basalt on the eastern Snake River Plain.

Features of the Bonneville Flood

The Bonneville Flood is the name given to catastrophic outflow from Pleistocene Lake Bonneville (Malde, 1968), which overtopped its rim at Red Rock Pass in southeastern Idaho about 15,000 years ago (Scott and others, 1983) and discharged a vast volume of water down the Snake River (Fig. 2). The flood descended the Portneuf Valley (Scott and others, 1982), built a boulder fan on the Snake River Plain at the mouth of the Portneuf (Trimble and Carr, 1961), swept down the Snake River, in places forming scabland channels on the nearby upland (Carr and Trimble, 1963; Malde, 1968), descended Hells Canyon to the Lewiston Basin (Webster and others, 1982), and ultimately joined the Columbia River. At Lewiston, gravel deposited by the Bonneville Flood was subsequently overlain by rhythmically bedded layers of gravel, sand, and silt laid down as backwater deposits by floods from Pleistocene Lake Missoula (Waitt, 1985).

The path of the Bonneville Flood downstream to Twin Falls is marked mainly by erosion, although rounded boulders of basalt with diameters larger than 3 m are found in several places. Farther west, the flood path along the Snake River canyon is strewn with bouldery debris, picturesquely called Melon Gravel (Malde, 1968), which occurs in gravel bars that reach heights 100 m above the Snake River. A canyon segment enlarged by flood erosion near the city of Twin Falls was a major source for the gravel. Shoshone Falls, near the head of the enlarged segment,

is a scenic remnant of the flood. The next 200 km of canyon below Shoshone Falls contain an estimated 2.4 km³ of gravel, and a wide stretch of canyon 275 km downstream contains another 0.2 km³ (Malde, 1968).

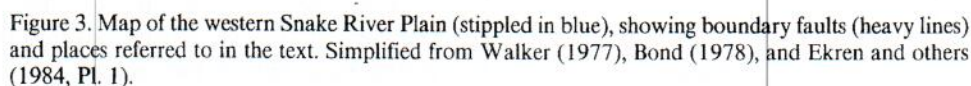
Discharge of the Bonneville Flood at a constricted reach south of Boise formerly was estimated as 425,000 m³/s (C. T. Jenkins in Malde, 1968). Based on new evidence of the flood height, the most probable discharge at the constriction now is estimated to have been 935,000 m³/s, using the step-backwater method (Jarrett and Malde, 1987). For comparison, Jarrett and Malde cited peak discharges of 70,000 m³/s for the 1927 flood on the Mississippi River and 385,000 m³/s for a flood on the Amazon River in 1953.

The volume discharge by Lake Bonneville during the Bonneville Flood equals the volume between the Bonneville shoreline (altitude 1,552 m) and the Provo shoreline (1,444 m), or a volume of 4,700 km³, when the altitudes are adjusted for hydroisostatic rebound (Currey and Oviatt, 1985). At a constant discharge of 935,000 m³/s, a flood of this volume would have lasted eight weeks, although attenuated flow surely would have made the duration even longer. The flood duration could be estimated more closely by computing a dam-break model for Red Rock Pass.

WESTERN SNAKE RIVER PLAIN

General features

The most distinctive geologic features of the western Snake River Plain are its sedimentary deposits, locally interbedded with basalt, which together fill a structural basin bounded by older rocks (Fig. 3). The deposits are well exposed in gullied terrain and badlands, generally south of the Snake River, and in walls of the Snake River canyon, many of which are more than 150 m high. The northeastern margin of the basin is a well-defined, nearly linear fault zone, which forms an abrupt, mountainous boundary (Malde, 1959). The southwestern margin is more irregular, being marked by the steep, faulted front of the Owyhee Mountains (Ekren and others, 1981) and, farther east, by the more subdued, steplike outline of the Owyhee Plateau, which has been formed by a set of en echelon faults (Malde and others, 1963). The western end of the plain is an upland of Miocene and Pliocene volcanic rocks, broken into several northwest-trending fault blocks. The eastern end is marked by basalt of the eastern Snake River Plain. Except at the upper reach of the basin, where sedimentary deposits overlap downfaulted volcanic rocks, the distribution of the sedimentary deposits closely coincides with the outline of the structural basin. As defined by the extent of the sedimentary deposits, the western Snake River Plain is about 260 km long. Its width narrows headward in a series of steps from a maximum of 70 km in the central part to a width of 35 km at the eastern end. At the eastern end of the basin, however, the distance between outcrops of volcanic rocks overlapped by the sedimentary deposits is only about 20 km. Altitudes within the western



As determined by seismic reflections from underlying basalt, the sedimentary deposits near Caldwell have a maximum thickness of 1.7 km, or a depth nearly 1 km below sea level (Wood and Anderson, 1981; Wood, 1984). Considerable thicknesses of sediment also are found in drill holes farther northwest and southeast (Wood and Anderson, 1981; Youngquist and Kiilsgaard, 1951). The deposits represent lacustrine and fluvial continental environments (Middleton and others, 1985), and the deeper layers clearly have subsided to depths below sea level. Minor thicknesses of basalt are interbedded in the sedimentary deposits in several parts of the basin. Based on outcrops in dissected areas along the Snake River, the deposits are divided into several dis-

In the J. N. James 1 exploratory well just west of Boise (Fig. 3), the sedimentary deposits rest on 2,100 m of basalt (interbedded with some sandstone and rhyolitic tuff), which in turn is

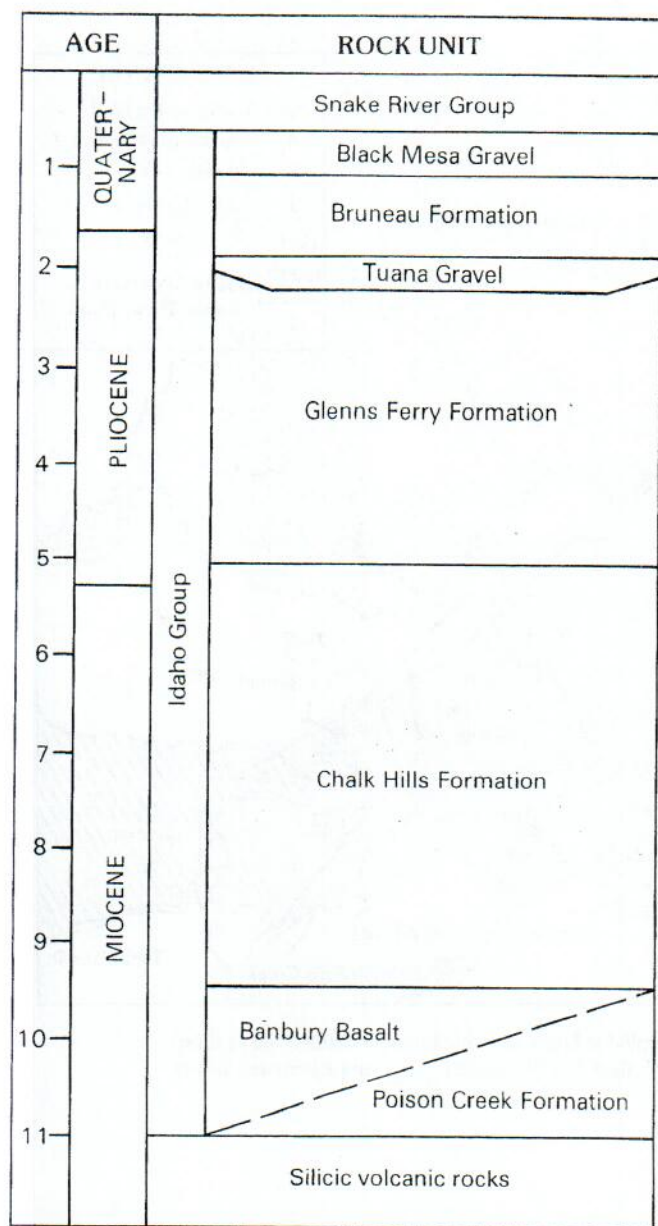


Figure 4. Sequence of upper Cenozoic rocks that underlie the western Snake River Plain, modified from Malde and Powers (1962) and Kimmel (1982). Numerical scale is in millions of years. Ages of epoch boundaries are from Berggren and others (1985).

underlain by 815 m of tuffaceous sediments and (still deeper) by 435 m of volcanic rocks of silicic and intermediate composition (Wood and Anderson, 1981). The seismic reflection profile reported by Wood and Anderson shows that the basalt extends across the full width of the western Snake River Plain. In deep wells northwest of the seismic profile, however, particularly in the Ore-Ida 1 well, the basaltic section has more fine-grained sediment than basalt (Wood and Anderson, 1981). Wood and Ander-

son correlate the tuffaceous deposits below the basalt with a unit in the Owyhee Mountains that has a K-Ar age of about 16 Ma, and they suggest that intermediate volcanic rocks in the lowest 240 m of the well may be as old as Oligocene or Eocene.

The pre-Quaternary sedimentary deposits decrease in age toward the eastern end of the basin. Thus, in dissected areas along the southern margin of the basin, the Poison Creek Formation is exposed no farther east than Reynolds Creek, the Chalk Hills Formation is exposed no farther east than the Bruneau River, and the exposed pre-Quaternary deposits farther east are entirely the Glens Ferry Formation. Although the Glens Ferry Formation is as young as about 2 Ma in the central part of the western Snake River Plain, its youngest deposits at the head of the sedimentary basin have ages of about 3 Ma, based on K-Ar dating (Evernden and others, 1964) and paleomagnetic stratigraphy (Neville and others, 1979). Probably, therefore, the sedimentary basin reached its maximum headward extent by 3 Ma.

The sedimentary basin also becomes progressively more shallow toward its head, so that the Glens Ferry Formation thins eastward and pinches out between Banbury Basalt and overlying basalt of the eastern Snake River Plain about 20 km east of Hagerman, as shown in a test well drilled near Wendell (Whitehead and Lindholm, 1985).

As described in the section on the Quaternary age of the Snake River, the Glens Ferry Formation is beveled by two upland deposits of gravel, the Tuana Gravel on the east and the Tenmile Gravel on the west, which have estimated ages of about 2 Ma. Deposits of the western Snake River Plain that are incised below the upland formed by the Tuana and Tenmile Gravels consist of basalt and lake beds of the Bruneau Formation, basalt of the Snake River Group, gravel on terraces and pediments along the Snake River and the Boise River, and other local surficial deposits (Fig. 4). The deposits of principal interest are discussed separately in the sections that follow. A noteworthy feature of possible structural significance is that vents for basalt of Bruneau and younger age generally decrease in age to the northwest.

The pre-Quaternary deposits dip inward at gentle angles toward the axis of the western Snake River Plain, typically less than 2°, although dips in some deposits at the southern and northern edges of the basin locally exceed 5° (Kimmel, 1982; S. H. Wood, Boise State University, written communication, 1986). Bedding in deposits near the center of the basin in most places is virtually horizontal. Such dips cannot entirely account for the depth of deposits in the center of the basin. Rather, the basin deposits surely have subsided on faults.

Structural relief produced by displacements in the fault zone along the northeastern margin of the western Snake River Plain is at least 2,900 m near Mountain Home, as measured by the difference in altitude between the top of silicic volcanic rocks in the Bostic 1-A well (Arney and others, 1984) and eroded silicic volcanic rocks in the Mount Bennett Hills, 14 km northeast (Fig. 3). This amount of vertical offset is about the same as that estimated from displacements seen in rocks exposed along the fault zone

(Malde, 1959). The fault zone at the southwestern margin of the western Snake River Plain possibly is less abrupt, but the structural relief between granite in the bottom of the Anschutz Federal 1 drill hole and an outcrop 13 km southwest is 3,560 m (McIntyre, 1979; Ekren and others, 1981, section A-A').

Faulting also is reported within the sedimentary basin. Thus, near Vale, Oregon, a concealed northwest-trending fault zone is identified from gravity data and from the pattern of circulation of geothermal water (Lillie and Couch, 1979). Near Boise, deeply buried northwest-trending faults that produced horst and graben structures in basalt under the sedimentary deposits are inferred from seismic evidence (Wood, 1984), and faulting in the uppermost sedimentary deposits is expressed by scarps as much as 30 m high in Tenmile Gravel (Wood and Anderson, 1981). Tuana Gravel in the upper part of the sedimentary basin, however, is not disturbed by faulting. Furthermore, Quaternary deposits of the western Snake River Plain are not known to be structurally deformed, except for local disturbances near volcanic vents. Altitudes of the Quaternary deposits are thus considered to be trustworthy for correlating certain stratigraphic units and for deducing former physiographic relations.

Rocks adjacent to the western Snake River Plain establish limits for its geologic age. On the northern side, west of the Idaho batholith, basalt and interbedded sedimentary deposits are assigned to the Columbia River Basalt Group, dated as 17 to 14 Ma (Fitzgerald, 1982). On the southern side, granite in the core of the Owyhee Mountains is correlated with the Idaho batholith (Taubeneck, 1971; Ekren and others, 1981). Still more distant outliers of granite occur south of the Owyhee Plateau in northernmost Nevada (see inferred edge of the batholith on Fig. 1). The Owyhee Mountains also contain remnants of Eocene and Oligocene volcanic rocks, as well as a roof pendant of pre-Cretaceous meta-sedimentary rocks (Ekren and others, 1984). The main part of the Owyhee Mountains, however, as described by Ekren and others (1984), consists of lava flows of olivine basalt and latite that have K-Ar ages of about 17 Ma, overlain and intruded by silicic volcanic rocks that are mineralized with precious metals, and that have ages averaging 16 Ma. Ekren and others (1984) inferred from the field relations of one rhyolite tuff that the adjacent western Snake River Plain was a structural lowland by 16 Ma.

Rocks adjacent to the upper reach of the western Snake River Plain consist of rhyolitic tuff and overlying basaltic lava flows. The rhyolite is widely called the Idavada Volcanics, and the basalt generally is known as Banbury Basalt (Malde and Powers, 1962; Ekren and others, 1984). To the north, the Idavada and Banbury make up most of the Mount Bennett Hills. To the south, they form the Owyhee Plateau.

The rhyolite and basalt of the Mount Bennett Hills and the Owyhee Plateau are considered here to be part of the eastern Snake River Plain trend. The Owyhee Plateau, however, is much higher in altitude than might be expected from its continuity with the eastern Snake River Plain, and Banbury Basalt on the plateau is anomalously thin when compared with the thickness of basalt farther east, being usually less than 100 m thick. These features of

the Owyhee Plateau are discussed in the following section on structure and origin.

On the Owyhee Plateau, near the southwestern end, rhyolite tuffs of the Idavada Volcanics centered at Juniper Mountain (Fig. 1) have ages of 12.0 to 13.8 Ma (Ekren and others, 1984). Volcanism farther east on the plateau is represented by rhyolite tuff that erupted at 9.7 Ma near Little Jacks Creek (the Little Jacks Tuff of Ekren and others, 1984), by the Bruneau-Jarbridge eruptive center, with an age of 11.3 Ma (Bonnichsen, 1982), and still farther east by Idavada Volcanics and overlying Banbury Basalt that form the structural southern boundary of the upper reach of the western Snake River Plain. The Idavada of this area has ages of 9.6 to 10 Ma, and the estimated age of the Banbury is 9.4 Ma (Armstrong and others, 1980, sample SRP 70-11). East of Salmon Falls Creek, another basalt on the upland south of the Snake River intertongues with sedimentary deposits of the Glens Ferry Formation (Malde and Powers, 1972). The age of the basalt is estimated as probably 4 to 5 Ma (Armstrong and others, 1980), but the fission-track age for zircon in silicic volcanic ash below the basalt is 3.75 ± 0.36 Ma (Izett, 1981, Peters Gulch ash bed). This basalt and the underlying Banbury Basalt are exposed below Quaternary basalt of the eastern Snake River Plain along the northern side of the Snake River canyon upstream to Twin Falls (Malde and others, 1963).

In the Mount Bennett Hills, the Idavada Volcanics are about 10 to 11 m.y. old (Ekren and others, 1984), and a bed of silicic volcanic ash interbedded in Banbury Basalt has an age of 10 Ma (Evernden and others, 1964, sample KA 830).

Besides the silicic volcanic rocks discussed above, Ekren and others (1984) infer that certain ash-flow tuffs at the southern edge of the western Snake River Plain may have erupted at two sites within the western plain at about 11 Ma (Fig. 1). Field relations show that these tuffs are older than the Poison Creek Formation, but they are inexplicably missing under the sedimentary deposits penetrated by the J. N. James 1 well (Wood and Anderson, 1981), which was drilled near one of the inferred eruptive centers.

Structure and origin

The geologic features reviewed above indicate that the western Snake River Plain is a structural basin perhaps as old as 16 Ma, which grew headward (toward the southeast) until 3 Ma. Geophysical findings outlined below, although still limited in scope, imply that the structural basin is an area of crustal extension and that its present low altitude relative to uplands and mountains to the north and south reflects isostatic compensation for a large underlying mass of basalt and an unusually thick lower crust (Mabey, 1982). The crustal rifting inferred from these findings is consistent with the role of the western Snake River Plain in various regional tectonic models (reviewed by Barrash and Venkatakrishnan, 1982).

A further feature of the headward growth of the western Snake River Plain is that it intersects and overlaps downfaulted rhyolite and basalt, which seem to be part of the volcanic assem-

blage that marks the eastern Snake River Plain. That is, the sedimentary basin of the western Snake River Plain grew headward into the path of the hot spot that produced the eastern Snake River Plain trend (Fig. 1). In this light, the hypothesis of Armstrong and others (1975), that volcanic rocks in the eastern Snake River Plain eventually will be succeeded by sedimentary deposits as the western Snake River Plain grows headward, lacks a structural basis. Rather, the sedimentary basin of the western Snake River Plain has grown to the southeast, not the northeast, and its growth probably stopped at about 3 Ma.

A seismic refraction profile between central Nevada and Boise (Hill and Pakiser, 1967) shows that the lower crust thickens abruptly under the Owyhee Plateau and the western Snake River Plain (Fig. 5). Moreover, the upper crust of the Basin and Range Province to the south, which consists of a crystalline layer 20 km thick having a velocity of 6.0 km/s, disappears under the southern edge of the Owyhee Plateau. Beneath the plateau and the western Snake River Plain, the crystalline layer is replaced by a layer half as thick, which has a velocity of 5.2 km/s. This comparatively thin upper crustal layer probably consists of volcanic rocks and associated sedimentary deposits. The total thickness of the crust under the Owyhee Plateau and the western Snake River Plain is 40 to 45 km, or about 12 km thicker than the crust in the Basin and Range Province. Because of the anomalously thick lower crust and the apparent absence of a crystalline upper crust, Hill and Pakiser (1967) postulated that an original granitic crust was pulled apart by tensional stress and that resulting fractures were filled with basalt injected from the mantle.

A thick lower crust under the western Snake River Plain also is shown by gravity surveys (Bankey and others, 1985). The western Snake River Plain is a region of high gravity, most conspicuously marked by three elongated, northwest-trending gravity highs with amplitudes of 70 to 100 milligals (Fig. 6). Crustal models derived from the gravity data fall within fairly narrow limits, for which the latest model is given by Mabey (1982). Because of findings from magnetic surveys, which are discussed below, Mabey's model involves combining a dense slab of basalt with a thinned upper crust. The model shows a high-density mass across the full width of the plain, lying beneath about 2 km of surficial volcanic and sedimentary rocks, and extending to a depth of 5 km.

The aeromagnetic map of Idaho (Zietz and others, 1978) shows a well-defined linear magnetic high along the southwestern margin of the western Snake River Plain and a corresponding, but less conspicuous, magnetic low along the northeastern margin (Fig. 7). Mabey (1982) interpreted these features as the boundaries of a layer of basalt a few kilometers deep, which is strongly magnetized in the direction of the present magnetic field. Mabey (1982) also suggested that the edge of the buried basalt is further marked by a magnetic high that continues southward just east of longitude 116° W and then turns eastward. Bonnicksen (1982), on the contrary, interpreted the curving trend of the magnetic high and a less distinct magnetic low to the northeast as the outline of the Bruneau-Jarvis volcanic center (Fig. 1). If the

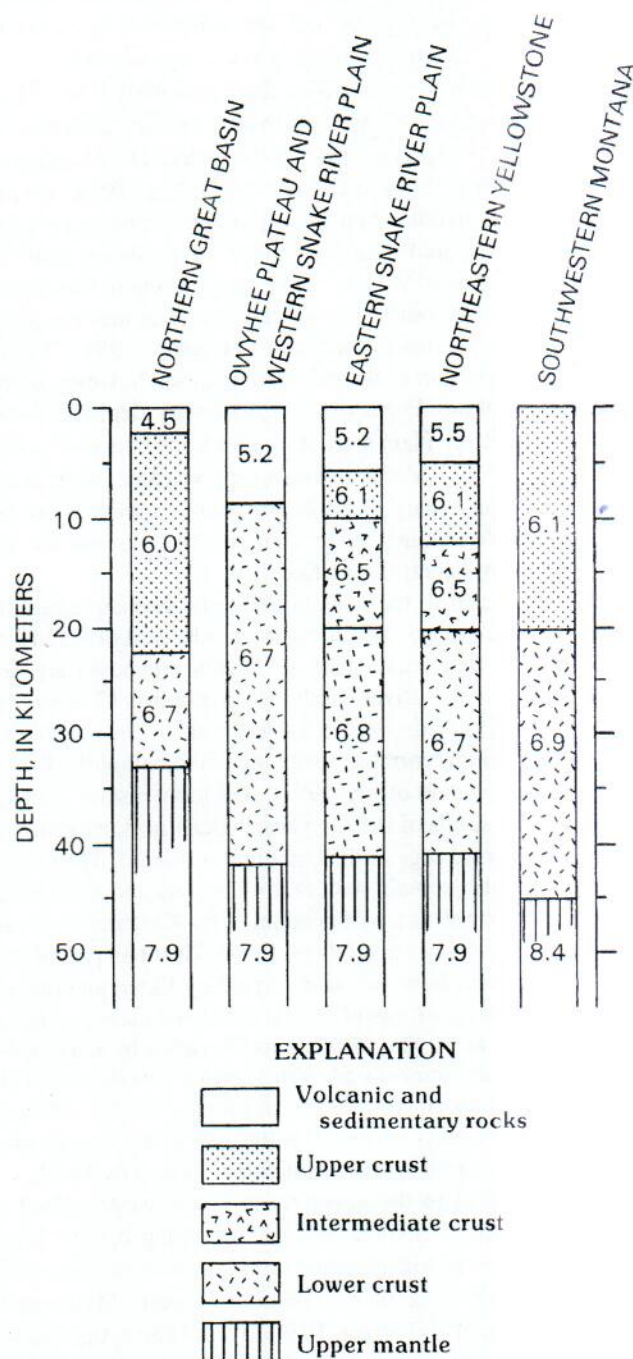


Figure 5. Crustal models of the Snake River Plain and adjacent areas, deduced from seismic refraction profiles. Numbers refer to P-wave velocities in kilometers per second. From Hill and Pakiser (1967) and Iyer (1984a).

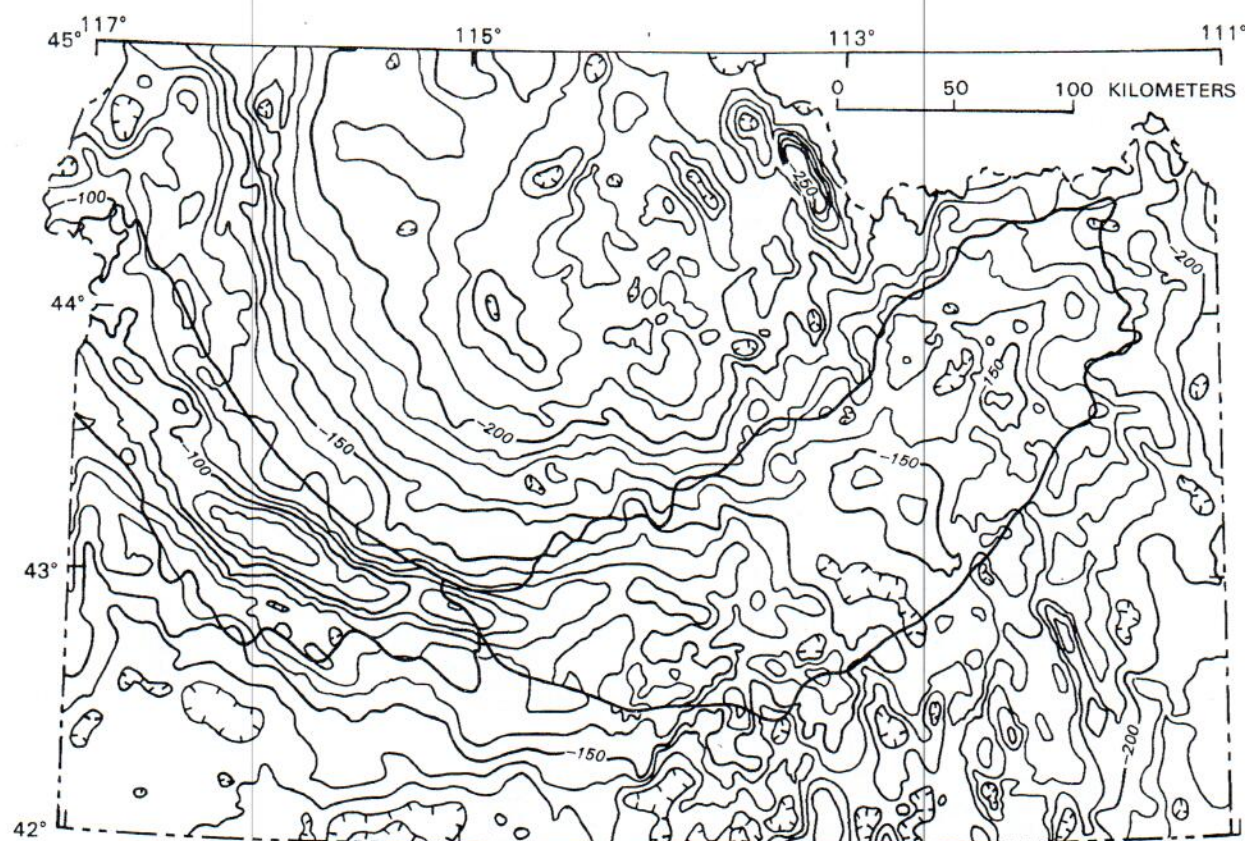


Figure 6. Bouguer gravity contour map of southern Idaho, simplified from Bankey and others (1985). Contour interval is 10 milligals. Hachures indicate closed gravity lows. Blue lines indicate boundaries of western and eastern Snake River Plain.

magnetic high of this area is indeed a consequence of silicic volcanism, different parts of the continuous magnetic high have different origins, and the apparent continuity of the buried slab of basalt is misleading.

A considerable number of heat-flow measurements have been made for the western Snake River Plain (Brott and others, 1978; R. N. Smith, 1981), which indicate values of less than 80 mW/m^2 over much of its width. Temperature gradients also have been measured in several deep exploratory wells (Wood, 1984). The heat flow increases abruptly at the margins of the plain, reaching values greater than 150 mW/m^2 in granitic terrain just north of Boise and at the northern end of the Owyhee Mountains (R. N. Smith, 1981). In general, the heat flow pattern is symmetrical about the axis of the western Snake River Plain, although the values increase most abruptly just beyond the northeastern margin. Brott and others (1978) interpreted the pattern as the residual effect of a large heat source, possibly a large mafic intrusion, which was intruded into the lower crust between 10 and 15 Ma. They present a model in which the inferred heat source is considerably wider than the western Snake River Plain, and in which the plain is represented as a deep trough, filled with

sedimentary and volcanic material of low thermal conductivity. Thus, the anomalously high heat flow at the edges of the western Snake River Plain was attributed by Brott and others to thermal refraction of heat into more conductive granite at the margins. Mabey (1982) pointed out that the pattern of heat flow is consistent with the hypothesis that the western Snake River Plain is a region of crustal extension.

Mabey (1982) summed up the geophysical findings by concluding that, at about 17 Ma, the western Snake River Plain formed as a rift, which was filled with upwelling basalt and with sedimentary debris. Mabey pointed out, however, that granite in the bottom of the Anschutz Federal 1 drill hole is inconsistent with this origin, unless the granite is a slide block or some other anomalous feature. Eocene or Oligocene rocks under the western Snake River Plain, as perhaps found in the lowest part of the J. N. James 1 well, also would be inconsistent with late Cenozoic rifting. Without further drilling, the possible extent of such pre-Miocene rocks under the western Snake River Plain is unknown. Mabey's conclusion that the western Snake River Plain formed by rifting is modified here only by suggesting that the rift continued to grow southeastward until 3 Ma and that the rift thus

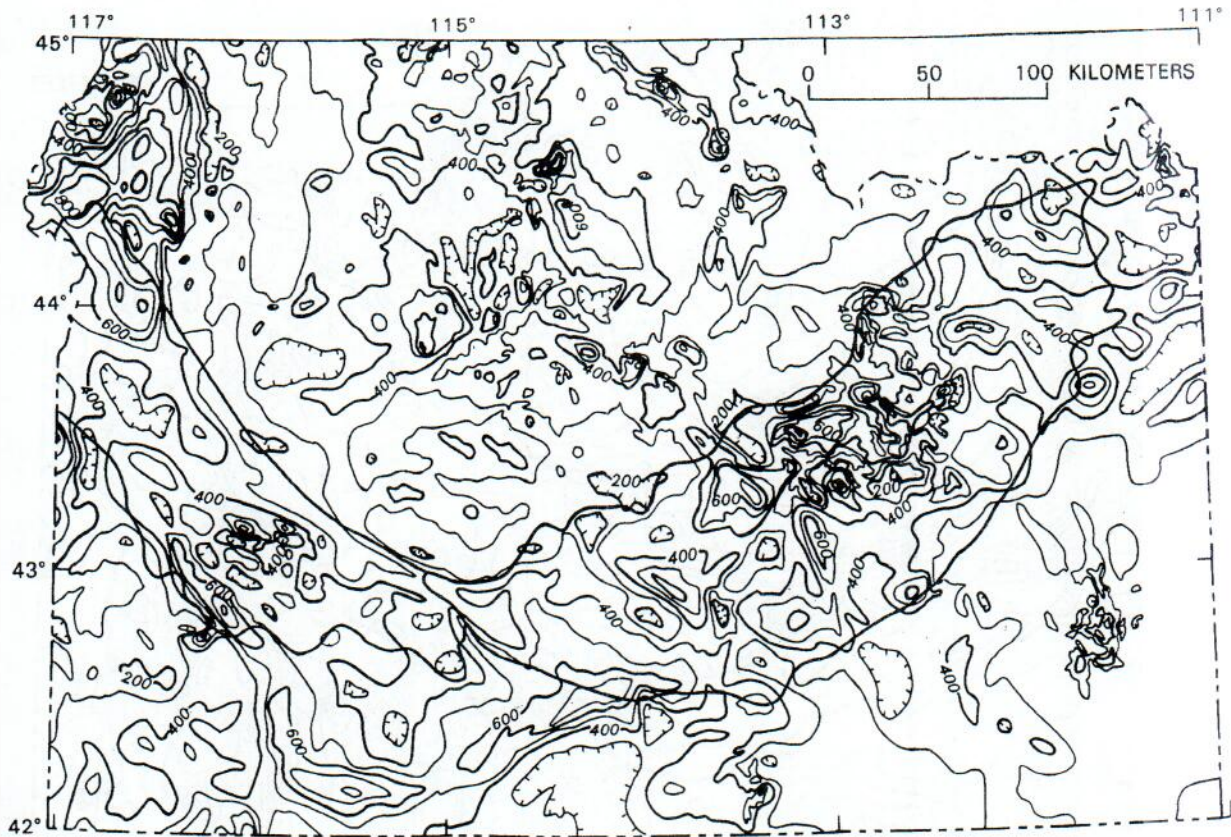


Figure 7. Aeromagnetic contour map of southern Idaho, simplified from Zietz and others (1978). Magnetic intensities are given relative to an arbitrary datum. Contour interval is 100 gammas. Hachures indicate areas of lower magnetic intensity. Blue lines indicate boundaries of western and eastern Snake River Plain.

intersected a terrain of rhyolite and basalt associated with the eastern Snake River Plain trend. As the expanding rift grew headward, rhyolite and basalt of the Owyhee Plateau and the Mount Bennett Hills were dropped into the rift on en echelon faults and were then partly overlapped by sedimentary deposits.

The inferred origin of the western Snake River Plain thus involves some understanding of the volcanic processes that account for the eastern Snake River Plain. As explained in the discussion of the eastern Snake River Plain, these processes produced explosive eruptions of ash-flow tuffs from shallow magma chambers and eventually resulted in lava flows of basalt as the magma chambers solidified. In this way, the eastern Snake River Plain came to be underlain by many coalescing batholiths, and its present altitude is a result of progressive thermal contraction. This explanation of volcanic processes within the eastern Snake River Plain also accounts for the bimodal volcanic assemblage of the Owyhee Plateau—its inferred predecessor. Nonetheless, the Owyhee Plateau differs from the eastern Snake River Plain in its comparatively high altitude and in the relative thinness of its covering basalt. These differences seem to point toward a strong

crust under the Owyhee Plateau, or one less subject to thermal contraction, and to a short history of basalt eruptions.

The seismic refraction profile of Hill and Pakiser (1967) suggests that the crust under the Owyhee Plateau has none of the granite found to the north and south. Because of the spacing of the data, however, the profile may not accurately distinguish granite from the silicic volcanics, and more of the original granitic crust may be present than has been inferred.

The crust of the Owyhee Plateau may be only partly modified because of the character of the silicic volcanism. From study of mineral assemblages and structural relations, Ekren and others (1984) found that magma chambers for two ash-flow tuffs were deep seated, within the lower crust, and that the magma chambers lack overlying calderas. If widely spaced, such sources of magma might leave some of the original crust undisturbed. Widely spaced magma chambers also might have made the Owyhee Plateau less susceptible to subsidence by thermal contraction.

Banbury Basalt on the Owyhee Plateau has been little modified by erosion, as shown by its well-preserved shield volcanoes

(Malde and others, 1963; Bonnichsen, 1982). Thus, the thinness of the Banbury, when compared with the 1 to 2 km of basalt on the eastern Snake River Plain, expresses a relatively small initial volume. A possible reason for the small volume of the Banbury is that the Owyhee Plateau has no recognized volcanic rift zones, which are sites of voluminous basaltic eruptions on the eastern Snake River Plain, where such zones are aligned with basin-and-range extensional faults (see the discussion of the crustal structure of eastern Snake River Plain). Rather, most of the Owyhee Plateau is thought to be within the original boundary of the Idaho batholith (Fig. 1), which has not been affected by basin-and-range faulting.

Bruneau Formation

Because of entrenchment in the western Snake River Plain after the Snake River was captured by the Columbia River, the pre-Quaternary sedimentary deposits were beveled by the Tuana and Tenmile Gravels and were then incised by canyon cutting. The deposits that date from this time of canyon cutting and that accumulated until the Snake River was established in its present canyon, are known as the Bruneau Formation (Malde and Powers, 1962). The Bruneau deposits consist of basaltic lava dams in four ancestral canyons of the Snake River (Fig. 8), lake beds impounded by the lava dams, upland lava flows of basalt (generally north of the present canyon), and alluvial fans and pediment deposits on marginal uplands. The distribution and stratigraphic relations of the lava dams and lake beds define four canyon-filling stages of the Bruneau Formation, ranging in age from about 1.9 to 1 Ma (Table 1). Because several canyon-filling units can be traced into upland areas, either as lava flows or as overlapping lake beds, widespread upland deposits can be correlated with the canyon-filling stages. Thus, the geology and physiographic relations of the Bruneau Formation provide a basis for understanding the latest Pliocene and early Quaternary history of large parts of the western Snake River Plain.

The distribution of lava flows, lake beds, and alluvial fans of the Bruneau Formation from Grand View to Glenns Ferry has been known since 1963 (Malde and others, 1963), and the Bruneau deposits near Glenns Ferry have been mapped in some detail (Malde and Powers, 1972). However, lake beds near Hagerman that previously were assigned to the Bruneau Formation (Malde and Powers, 1972) are now known to have been impounded by the late Pleistocene McKinney Basalt and thus are much younger than the Bruneau (see the discussion of displacements of the ancestral Snake River in the eastern Snake River Plain). The relations described here are based mainly on unpublished studies since 1978, which are partly summarized in a recent guide to the Snake River canyon downstream from Grand View (Malde, 1985). An important part of these studies consisted of K-Ar dating and paleomagnetic measurements by Hassan Amini in 1979 and 1980 (Amini, 1983; Amini and others, 1984). The geology of the Bruneau Formation west of Grand View and south of the Snake River, as understood in 1978, is portrayed on the

geologic map of the western part of Owyhee County (Ekren and others, 1981).

The main features of the Bruneau Formation express the interplay of basaltic volcanoes and the ancestral Snake River. Four successive canyons cut by the river were blocked by canyon-filling basalts, commonly to depths of several hundred meters, and the river thereby was diverted into new channels, each usually south of the earlier canyon. The last diversion of the Snake River, which was caused by eruption of the large volcano called Initial Point, established the river in its present canyon (the fifth) and is considered to be of post-Bruneau age (lava flows younger than the Bruneau Formation are discussed below as basalts of the Snake River Group). Thick, canyon-filling sections of subaqueous basalt, such as pillow lava and palagonite tuff, were deposited where basalt spilled into the impounded water, and such deposits are now well exposed in walls of the present canyon. Some of the subaqueous basalt intertongues upstream with lake beds, especially in the third canyon of the Bruneau Formation.

Two general areas of basaltic activity of Bruneau age can be distinguished, one in uplands near Mountain Home and north of Glenns Ferry, the other forming a line of vents approximately coinciding with the Snake River from Grand View to Walters Ferry (Fig. 8). Two vents in the Mount Bennett Hills also were sources of basalt of Bruneau age, which flowed down canyons to the upland north of Glenns Ferry.

Volcanic activity along the line of vents northwest of Grand View decreased in age to the northwest. In particular, volcanoes contemporaneous with the first canyon stage of the Bruneau Formation are between Grand View and Castle Creek, and those of the third canyon stage reach from Sinker Creek to Walters Ferry. Volcanic activity along this northwesterly trend continued after Bruneau time, but the area of active vents shifted still farther northwestward.

Most basalts of the Bruneau Formation are partly weathered and typically have a brownish color, although the basalts are seldom decomposed. The coloration is produced by iron oxides and secondary clays, commonly described under the names chlorophaeite and palagonite but more generally identified as clays of the montmorillonite group. In upland areas, the surfaces of the lava flows are further modified by development of patterned ground (Malde, 1964), which obliterates most (or all) of the original volcanic landforms. Hence, in both weathering and surficial features, the Bruneau basalts usually are distinguishable from basalts of younger age, which typically are fresh.

Before any of the Bruneau Formation was deposited, downcutting by the Snake River had reached a considerable depth. In the northern wall of the present Snake River canyon just north of Grand View (Fig. 8), 75 m of palagonite tuff in the lower part of the first canyon-filling stage of the Bruneau rests on a canyon floor cut in the Glenns Ferry Formation at an altitude nearly 190 m lower than the nearest Tuana Gravel, 30 km upstream. Near Jackass Butte a short distance downstream, basalt of the first canyon stage is incised 60 m below a lava flow with an

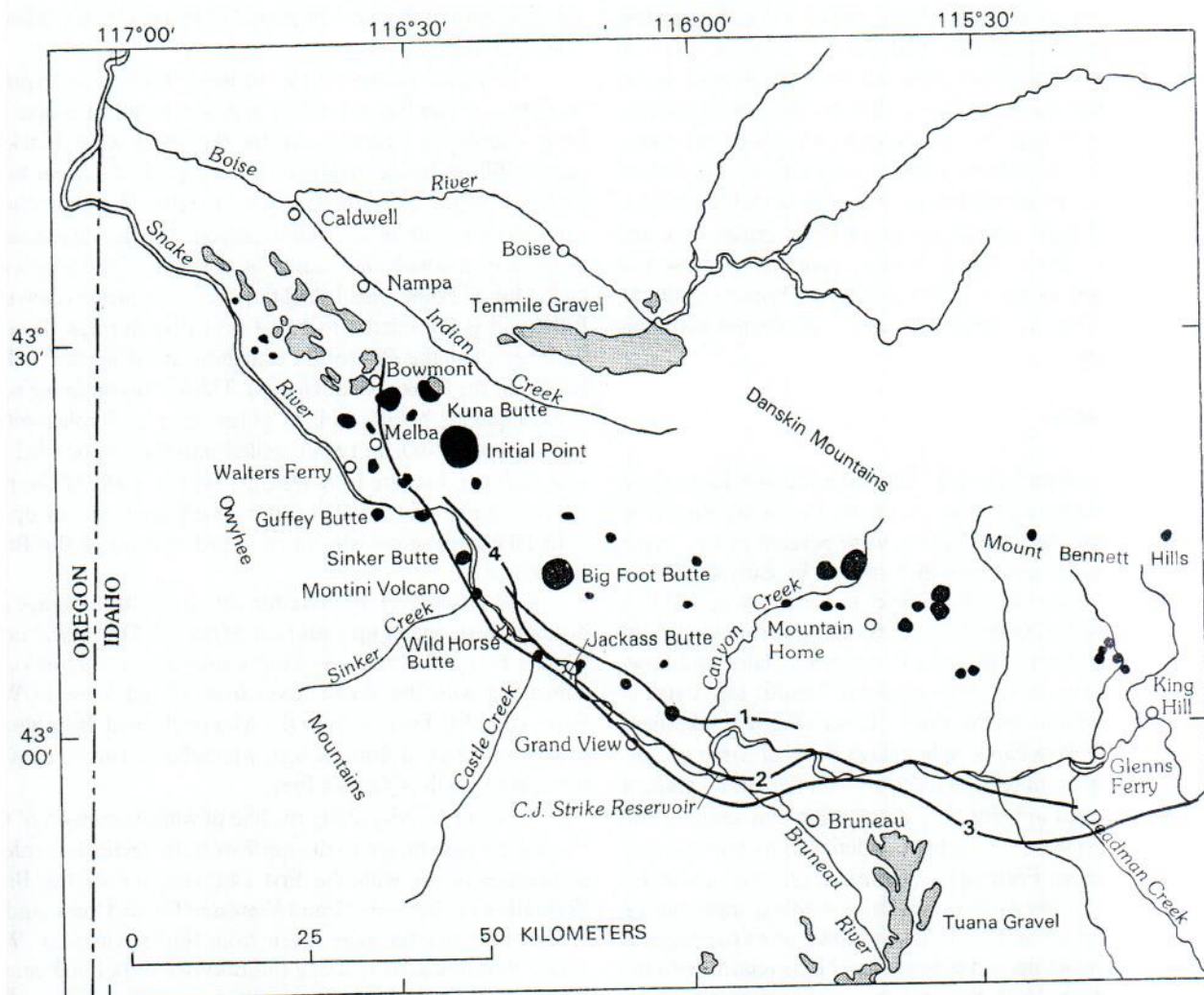


Figure 8. Map showing the Tuana and Tenmile Gravels (shaded areas), the approximate axis of each successive early Quaternary canyon cut by the ancestral Snake River in the western Snake River Plain (blue lines, numbered 1 through 4 from oldest to youngest), basaltic vents of the Bruneau Formation (blue areas), and vents younger than the Bruneau (black areas).

age of 2.18 ± 0.18 Ma (Table 1, sample 80H1), which, as previously explained, is considered to be part of the Glenns Ferry Formation. A lava flow of the first canyon stage in this area has been dated at 2.06 ± 0.24 Ma (Table 1, sample 80H4), but this age seems too old, considering the depth of downcutting below Tuana Gravel and the depth of the first canyon below the still older basalt of the Glenns Ferry Formation. Rather, a more accurate age for the first canyon stage probably is about 1.9 Ma, as determined for the basalt of Wild Horse Butte, 14 km downstream (Table 1, sample 79H9).

Because the first canyon of the Bruneau Formation was filled with resistant basaltic rocks, the Snake River was diverted southward, and a second canyon was cut in easily erodible deposits of the Glenns Ferry Formation. The second canyon is as deep as the present canyon at the longitude of Bruneau, or more than 240 m below Tuana Gravel. The second canyon can be traced

from a place on Deadman Creek south of Glenns Ferry to the mouth of the Bruneau River, and the canyon also is thought to be represented by a lava cascade a short distance downstream from Grand View. Between Deadman Creek and the mouth of the Bruneau River, the second canyon is filled with basalts that flowed south from vents between Mountain Home and the upland north of Glenns Ferry. Much of the canyon-filling basalt is pillow lava, which reaches heights 160 m above the canyon floor, thereby showing that the canyon was at times deeply ponded. Upland equivalents of the canyon-filling basalts form a thin veneer on alluvial fans and pediments north of the ancestral canyon and thus preserve part of the landscape that existed when the basalts erupted. K-Ar ages for these basalts range from 1.2 to 1.5 Ma (Table 1) but probably are erroneously too young. In light of ages of basalt in the next younger canyon of the Bruneau Formation (the third), ages of basalts of the second canyon-filling stage

Table 1. POTASSIUM-ARGON AGES OF LOCAL GEOLOGIC UNITS IN THE WESTERN SNAKE RIVER PLAIN

Groups and formations	Walters Ferry to Grand View			Grand View to Glens Ferry	
	Local geologic units (Modified from Amini, 1983; Malde, 1985)	K-Ar age (Ma) (Amini and others, 1984)	Canyon stage	Local geologic units (Modified from Malde and Powers, 1972)	K-Ar age (Ma) (Evernden and others, 1964; Armstrong and others, 1975)
Snake River Group	Basalt of Tadpole Lake..... Basalt of Kuna Butte. Basalt of Initial Point.	0.47±0.07 (79H3A)	---	Basalt of C. J. Strike Reservoir (N).	
Bruneau Formation	Basalt of Guffey Butte (R)..... Basalt of Big Foot Butte (R)..... Basalt of Promontory Point (N).....	{ 1.04±0.07 (79H10-7) 1.09±0.06 (79H10-1) 0.95±0.04 (79H1A1) 0.78±0.04 (80H15)	4		
				Berry Ranch lava flow..... King Hill lava flow (R).....	1.36 (KA 1188) 1.7±0.04 (70-7(L))
	Basalt of Sinker Butte (R)..... Rosette-spar basalt (R)..... Basalt of Montini volcano (R).....	1.37±0.11 (79H19) 1.52±0.07 (79H8) 1.64±0.18 (79H5-7)	3	Lake beds.	
	Basalt of Swan Falls (R)..... Basalt of Otter Massacre site (R).....	{ 1.36±0.11 (80H17) 1.58±0.13 (80H11) 1.60±0.23 (80H16) 1.67±0.09 (80H19) 1.81±0.11 (80H20) 1.26±0.17 (80H7-1) 1.36±0.09 (79H17)			
	Lava cascade near Grand View.		2	Morrow Reservoir lava flow (R)..... Cold Springs Creek lava flow. Hammett lava flow (R)..... Deadman lava flow (R).....	1.2±0.2 (70-8(L)) { 1.5±0.3 (70-4(L)) 1.2±0.2 (59P271(HP)) 1.35±0.07 (69-59(L))
	Basalt of Wild Horse Butte (R)..... Unnamed basalt unit (R).....	1.92±0.16 (79H9) 2.06±0.24 (80H4)	1		
	Tenmile Gravel.		---	Tuana Gravel.	
Glenns Ferry Formation	Basalt of Jackass Butte (R).....	2.18±0.18 (80H1)	---		
Note: magnetic polarities are shown as either normal (N) or reversed (R). Near Glens Ferry, lake beds of the Bruneau Formation mapped by Malde and Powers (1972) are classified here as belonging to a single unit, part of the third canyon-filling stage.					

of the Bruneau may be 1.7 to 1.8 Ma. These basalts are well exposed, either as rimrock or as deep canyon-filling deposits where the present Snake River canyon and its tributaries intersect the ancestral canyon.

The third canyon of the Bruneau Formation, which trends nearly parallel to the second canyon but a short distance south, is the longest of the Bruneau canyons. Above the mouth of Sinker Creek, deposits of the third canyon stage consist of distinctive lake beds, composed of massive units of clay, silt, and diatomite, as well as beach deposits of gravel reworked from the Tuana. The lake beds form a continuous section 210 m thick in the northeastern wall of the Bruneau River valley, rising nearly to the toe of the Tuana Gravel. Elsewhere, the lake beds are widely exposed in gullied terrain and badlands, commonly being preserved as remnants on upland areas of earlier Bruneau lava flows and alluvial fans. An outcrop of the lake beds near Glens Ferry is the source of tooth fragments of *Mammuthus* (Madden, 1980), which have long been the basis for assigning the Bruneau Formation to the Irvingtonian land mammal age (Malde and Powers, 1962).

The lava dam that accounts for the lake beds of the third canyon-filling stage was constructed by several local volcanoes, which extend northwestward along the trend of the third ancestral canyon from the mouth of Sinker Creek to Walters Ferry (Fig. 8). At the mouth of Sinker Creek, the lake beds intertongue with basaltic tuff of a subaqueous volcano called informally the Montini volcano (Malde, 1987). Exposures of the lava dam downstream along the present canyon also display many subaqueous features, some indicating deeply ponded water. The highest level of ponded water is implied by well-bedded palagonite tuff in Sinker Butte at an altitude of 1,005 m, or 305 m above the Snake River. The floor of the third canyon is nearly 40 m below the Snake River, at an altitude of about 655 m, as indicated by the depth of basalt and gravel in wells at Melba (Stevens, 1962) and at Bowmont (R. L. Whitehead, U.S. Geological Survey written communication, 1986).

Basalts of the third canyon-filling stage of the Bruneau Formation yield K-Ar ages of 1.8 to 1.4 Ma, in addition to many determinations that are considered to be discordant (Table 1).

North of Glens Ferry, ages of 1.4 and 1.7 Ma for lava flows that overlie lake beds of the third canyon-filling stage (Table 1) are more or less consistent with the relation of the lake beds to the Montini volcano.

The downstream reach of the third canyon, north of the present Snake River, has a northerly trend, through a gap in hills capped by Tenmile Gravel, and toward the Boise River (Fig. 8). Well records, together with outcrops along the westward course of the present canyon, confirm that the upland to the east and west is underlain by the Glens Ferry Formation and that all the canyon-filling deposits in this area closely coincide with the trend of the third canyon. These relations imply that the downstream part of the Boise River was the ancestral Snake River in Bruneau time. The presence of concealed canyon-filling deposits of the Bruneau Formation between the Snake River and the Boise River probably influences the location and movement of deep ground water in this area.

The resistant lava dam in the third canyon eventually diverted the ancestral Snake River to the northeast, probably because overflow along the southwestern side of the lava dam was prevented by the proximity of the Owyhee Mountains. The fourth canyon of the Bruneau Formation then began to be cut. At an early stage of downcutting, when the fourth canyon was only a broad swale, the canyon was filled with the basalt of Promontory Point (named informally from a projection in the canyon wall just east of Sinker Butte). The basalt of Promontory Point has a K-Ar age of 0.78 ± 0.04 Ma (Table 1, sample 80H15), but this age may be erroneously too young. Rather, because the basalt of Promontory Point is normally magnetized and because it is overlain by the reversely magnetized basalt of Big Foot Butte, it may correlate with the Jaramillo Normal Polarity Subchron, dated as 0.97 to 0.90 Ma.

After the fourth canyon reached the depth of the present canyon, it was filled with basalt (largely pillow lava) that erupted from Big Foot Butte and Guffey Butte, both having K-Ar ages of about 1 Ma (Table 1). The ancestral Snake River then meandered for a time between Walters Ferry and Nampa, where it spread gravel and wore down some of the upland on which the Tenmile Gravel had been deposited.

Pediment deposits and fan gravel

South of Glens Ferry, a dissected pediment thinly veneered with gravel and sand called the Black Mesa Gravel, the uppermost unit of the Idaho Group (Fig. 4), bevels lake beds of the third canyon-filling stage of the Bruneau Formation. From a toe at an altitude of 930 m, or 170 m above the Snake River, the pediment surface gradually climbs southward to a gentle escarpment in the Glens Ferry Formation, 20 km south. Gravel and sand on the pediment were reworked from underlying shore gravels of the Bruneau Formation and from sandy deposits of the Glens Ferry Formation (Malde and Powers, 1962). The field relations show that Black Mesa Gravel was graded to a base level near the axis of the present Snake River but considerably below

the level formed by Bruneau deposits. That is, the highest lake beds of the third stage of the Bruneau Formation are 40 m above the toe of the Black Mesa Gravel, at an altitude of 970 m, as consistently found at several places from Glens Ferry westward. The distribution of Black Mesa Gravel, however, provides no evidence of canyon cutting by the ancestral Snake River, nor evidence of the existence of any ancestral tributaries.

Fan gravel as much as 150 m thick, possibly correlative with Black Mesa Gravel, occurs at the front of the Owyhee Mountains over a wide area northwest and southeast of the lower reach of Sinker Creek (Malde and Powers, 1962; Ekren and others, 1981). The fan gravel bevels lake beds and basalt of the third canyon-filling stage of the Bruneau Formation. The toe of the fan gravel northwest of Sinker Creek is at an altitude of about 900 m, or 195 m above the Snake River. Thus, stratigraphic and physiographic relations of the fan gravel resemble those of the Black Mesa Gravel.

Basalts of the Snake River Group

Several eruptions of basalt younger than the Bruneau Formation have been recognized in the western Snake River Plain, but their relations to each other are poorly known. These basalts are assigned to the upper Quaternary Snake River Group (Malde and Powers, 1962) because of their freshness and generally well-preserved surface morphology. The basalts are easily traced to their sources in undissected shield volcanoes of the kind widely found in the eastern Snake River Plain (see the section on the character of the basalt eruptions). Judging from unconformable relations to the Bruneau Formation, the basalts are younger than 1 Ma. One of the basalts 5 km southeast of Big Foot Butte, near a place called Tadpole Lake, has an age of 0.47 ± 0.07 Ma (Table 1, sample 79H3A). These basalts are widely distributed from Mountain Home northwestward, but the largest eruptions overlap the northwestern limits of the Bruneau Formation, and several of the vents are still farther northwest (Fig. 8).

The basalt of the Snake River Group of greatest physiographic importance in the western Snake River Plain erupted from the large shield volcano called Initial Point (Fig. 8). Lava from Initial Point flowed southward and westward across the upland where the ancestral Snake River was meandering after the canyon-filling eruptions of Big Foot Butte and Guffey Butte (Malde, 1985). Several other shield volcanoes a short distance west of Initial Point, near Melba, were formed at about the same time. The vents still farther northwest may be approximately the same age. As a result of these eruptions, but probably mainly because of lava flows from Initial Point, the ancestral Snake River was turned westward, away from its former confluence with the Boise River. The present Snake River canyon then was incised along the southern margin of the lava flows of the Snake River Group, in much the same way that canyons of the Bruneau Formation were formed south of the canyon-filling Bruneau basalts. Upstream from the abrupt bend south of Initial Point, where the Snake River turns westward, the present canyon follows the western and southern edge of the basalt of Big Foot Butte.

After the Snake River canyon had nearly reached its present depth, basalt from Kuna Butte (or from a closely related vent near Kuna Butte) spilled into the canyon south of Melba (Malde, 1985). The lava filled the canyon to a depth 90 m above the present river. Later downcutting has exposed the basalt in an intracanyon cliff 70 m high.

Basalt of the Snake River Group also occurs south of the Boise River along the upper reach of Indian Creek (Nace and others, 1957; Savage, 1958). The same basalt is buried at a shallow depth by an alluvial terrace along Indian Creek between Nampa and Caldwell (Wood and Anderson, 1981).

Another basalt of the Snake River Group within the Snake River canyon is found just north of the dam for the C. J. Strike Reservoir (Malde and others, 1963). The basalt spilled southward over the eroded edge of basalts of the first and second canyon-filling stages of the Bruneau Formation, at a tributary canyon of the main canyon (Canyon Creek), and covered a terrace cut on lake beds of the third canyon-filling stage of the Bruneau. The lava-capped terrace is 75 m above the Snake River and extends 10 km downstream to its terminus northeast of Grand View. The source of the basalt of C. J. Strike Reservoir was one or more vents in the upland about 30 km west of Mountain Home. Other vents 5 to 13 km northwest of Mountain Home also were sources of basalt of the Snake River Group and resulted in lava flows that reach the canyon rim to the south (Malde and others, 1963). Thus, a considerable part of the upland west and south of Mountain Home is covered by basalt of the Snake River Group.

Upland alluvial deposits

An upland surface 60 m above the Boise River near Caldwell is underlain by at least 30 m of alluvial sand and gravel, called informally "sediments of the Deer Flat surface" (Wood and Anderson, 1981). Although many water wells have been drilled in this area, the base of these alluvial sediments has not been identified. If the sediments extend considerably below the elevation of the Boise River, they may include alluvium of Bruneau age, related to the time when the ancestral Snake River joined the ancestral Boise River in this area. This conjecture is based on the observation that the Boise River valley at Caldwell is about 70 m above the base of canyon-filling Bruneau deposits at Melba and evidently has been backfilled, either in Bruneau time or later (Malde, 1985). According to Wood and Anderson (1981), the basalt of Indian Creek is interbedded in the sediments of the Deer Flat surface. Along the lower reach of Indian Creek, this basalt is further covered by one of several descending alluvial terraces that were formed by the Boise River (Nace and others, 1957).

The present Snake River canyon near Glens Ferry contains remnants of two terrace gravels, 122 m and 73 m above the river. Both are rich in porphyry, quartzite, and granite derived from the mountains of central Idaho (Malde and Powers, 1972, table of gravel counts), and both have been attributed to times of mountain glaciation (Malde and Powers, 1962).

The higher and older of the terrace gravels, called Sugar Bowl Gravel, is pebbly and has a gradient of 0.4 m/km, which is nearly equal to that of the present Snake River in the western Snake River Plain. The Sugar Bowl Gravel is preserved in wide stretches of the canyon that coincide with tributary junctions, showing that the present drainage system was established when the gravel was being deposited (Malde and Powers, 1962). The Sugar Bowl Gravel is distributed only along a reach of the canyon 45 km long, centered on Glens Ferry. As explained below when describing displacements of the ancestral Snake River, the Sugar Bowl Gravel rests on Madson Basalt, which is the oldest of several canyon-filling basalts of the Snake River Group identified in the western part of the eastern Snake River Plain.

The lower gravel is cobbly and has a comparatively steep gradient of 1.1 m/km, last being recognized in a terrace near Grand View, 15 m above the river (Malde, 1985, mile 30.4 in roadlog). Upstream, near King Hill, the lower gravel is 76 m above the river and is incised by a canyon filled with the youngest of the local canyon-filling basalts—the McKinney Basalt.

Geothermal resources

Excluding Yellowstone National Park, southern Idaho has the largest known resources of geothermal water in the United States. About 85 percent of the energy in geothermal systems with temperatures 90°C or higher, or 450×10^{18} joules, is in the Bruneau–Grand View area (Mabey, 1983). Young and Lewis (1982) reported recent observations about the extent of this geothermal system and reviewed the hydrology and geochemistry.

The size of the geothermal resource in the Bruneau–Grand View area is related to the large volume of the ground-water reservoir, not to exceptionally high temperatures. Chemical geothermometers and measurements of heat flow indicate maximum reservoir temperatures near 100°C. The temperature of water derived from aquifers in volcanic rocks ranges from 40° to 83°C, and the temperature of water from shallow aquifers in sedimentary deposits seldom exceeds 35°C. The geothermal waters are of the sodium carbonate and bicarbonate type, some having more than 1,000 mg/L dissolved solids. A few analyses of deuterium and oxygen-18 in the geothermal water suggest recharge at a time when the climate was 3° to 5°C colder than the present, and estimates of the recharge rate and pore volume indicate minimum residence times in the system of 3,400 to 6,800 years. From such features, a deep convective system is inferred. Young and Lewis proposed a conceptual model in which meteoric water that falls on the Owyhee Plateau circulates northward in semiconfined rhyolite at depths as great as 1,500 m, while being heated by an average heat flow of 92 mW/m².

A geothermal system at Boise is the source of hot water for the Boise Warm Springs Water District, which is the oldest geothermal heating company in the United States (Wood and Burnham, 1987). The first wells were drilled in 1890, and the district currently provides heat for more than 250 houses. The geothermal water also is used to heat several state and county build-

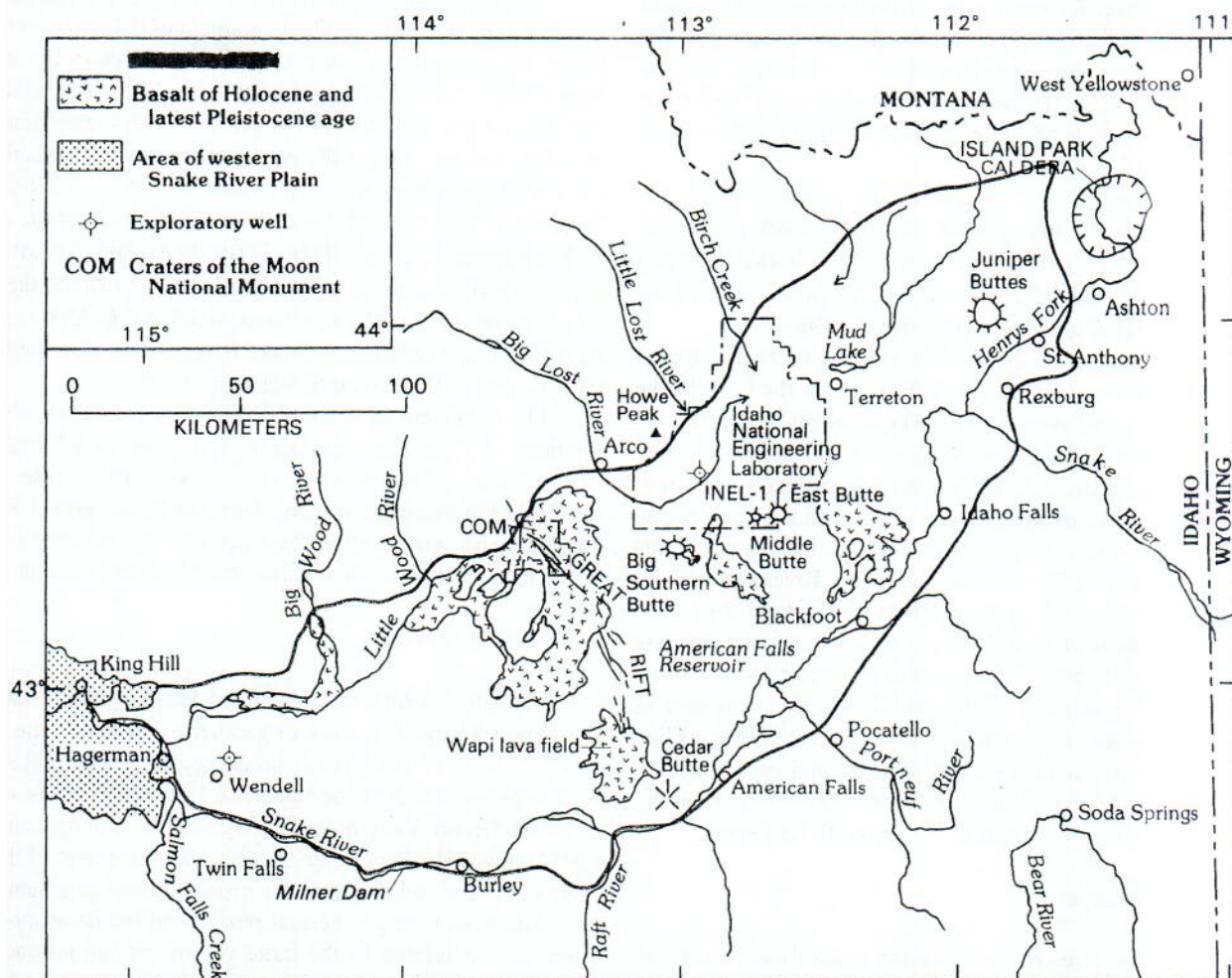


Figure 9. Map of the eastern Snake River Plain (outlined by blue line), showing places referred to in the text. Geologic boundaries are from Bond (1978).

ings. The water is at a nearly constant temperature of 77°C. As described by Wood and Burnham, the water comes from fractured rhyolite in the northwest-trending fault zone that marks the northeastern boundary of the western Snake River Plain. To supply present users during the winter, the water is pumped at a rate of 54 to 63 L/s, but demand in the spring and fall drops to 13 to 32 L/s. Radiocarbon dates for the geothermal water indicate mean residence times in the system of 6,700 to 17,000 years (Mayo and others, 1984).

EASTERN SNAKE RIVER PLAIN

General features

The eastern Snake River Plain is a volcanic province, covered by undissected lava flows of basalt throughout most of its length and underlain by rhyolitic volcanic rocks (Fig. 9). The

western end is marked by the north-flowing reach of the Snake River through Hagerman Valley, where the basalt abuts and overlaps sedimentary deposits of the western Snake River Plain. At the eastern end, at Island Park, the basalt gives way to rhyolitic ash-flow tuffs that continue northeastward to Yellowstone Park. These volcanic rocks, which are associated in places with bordering alluvial and lacustrine deposits, abut the ends and flanks of mountains to the north and south. As thus defined, the eastern Snake River Plain has a fairly uniform width of 90 to 100 km and a length of 350 km. If the Yellowstone region is combined with the eastern Snake River Plain, because of its volcanic affinity, the length of this volcanic province is nearly 450 km. The surface of the eastern plain steadily rises from an altitude of 975 m at Hagerman to 1,950 m northwest of Ashton. Nearby mountain peaks reach altitudes from 1,000 m to nearly 2,000 m higher. Howe Peak, for example, 6.5 km north of the northern edge of the plain, is 1,100 m above the basalt at its foot.

Carboniferous and other Paleozoic sedimentary rocks in mountain ranges north and south of the eastern Snake River Plain are similar in stratigraphy, lithology, and thickness (Skipp and others, 1979). Despite the continuity implied by these similarities, no evidence has been found that Paleozoic rocks are present beneath the volcanic rocks of the eastern Snake River Plain.

Although adjoining mountain ranges end more or less abruptly at the margins of the eastern Snake River Plain, boundary faults have been inferred only locally, using seismic methods (Pankratz and Ackerman, 1982), but such displacements probably are ring fractures related to calderas (Morgan and others, 1984). That is, clear-cut evidence for boundary faults has not been found. Rather, surrounding ash-flow tuffs dip inward toward the axis of the plain, as do adjacent sedimentary deposits of Miocene and Pliocene age (Carr and Trimble, 1963), and the eastern Snake River Plain has long been considered to be a downwarp (Kirkham, 1931).

The rhyolitic volcanic rocks that underlie the eastern Snake River Plain generally are accepted as being the products of a hot spot, located in the mantle beneath a part of the North American plate that has been drifting relatively southwestward (Armstrong and others, 1975; Brott and others, 1981; Iyer, 1984a; Leeman, 1982a; Mabey, 1982). Rhyolitic volcanism above the hot spot began at the McDermitt volcanic center in southeastern Oregon at 17.5 Ma and progressed northeastward along the eastern Snake River Plain trend (Fig. 1), culminating in three explosive caldera-forming eruptions in the Yellowstone area at about 2.0, 1.3, and 0.6 Ma (Christiansen, 1984; Izett, 1981; Luedke and Smith, this volume). During the 17 million years of volcanic activity, therefore, centers of rhyolitic volcanism migrated 580 km northeastward at an average rate of about 3.5 cm/yr, in accord with an estimate for the last 10 million years (Armstrong and others, 1975).

The rhyolites are primarily ash-flow tuffs. In the eastern Snake River Plain, most (or all) of the tuffs are associated with calderas that are concealed by overlying basalt (Morgan and others, 1984). Tuffs related to calderas range in age from 11.3 Ma at the Bruneau-Jarbridge eruptive center on the Owyhee Plateau (Bonnichsen, 1982) to 6.5, 5.6, and 4.3 Ma near the eastern end of the plain (Morgan and others, 1984). The silicic volcanic rocks reach a depth greater than 3 km at the bottom of the INEL-1 exploratory well at the Idaho National Engineering Laboratory (Fig. 9; Morgan and others, 1984), and seismic evidence indicates that the base of the silicic rocks may be as deep as 6 km (Braile and others, 1982).

Basalt that overlies the rhyolite began to erupt soon after any given point had passed over the hot spot. As previously explained, rhyolite in the eastern part of the Owyhee Plateau is overlain by Banbury Basalt dated 9.4 Ma and then by basalt of the Glens Ferry Formation younger than 3.75 Ma. Farther east, these basalts are overlain by basalt that forms the surface of the eastern Snake River Plain, which is largely (or entirely) of Quaternary age. As described below in the section on displacements of the ancestral Snake River, fresh lava flows contemporaneous

with entrenchment of the Snake River west of Twin Falls are classified as units of the Snake River Group and are considered to be late Quaternary (Malde and Powers, 1962). The few published K-Ar ages for the basalt are less than 2.0 Ma (Armstrong and others, 1975, 1980; Champion and others, 1981; Kuntz and others, 1980; Lanphere, 1986; Scott and others, 1982), but several of these ages have large experimental errors. A number of unpublished K-Ar ages for the exposed basalt in a wide area that encloses the Idaho National Engineering Laboratory (Kuntz and others, 1984) indicate that the basalt is not older than 1.2 Ma (M. A. Lanphere, U.S. Geological Survey, written communication, 1986). Furthermore, with few exceptions, exposed lava flows throughout the eastern Snake River Plain have normal magnetic polarity (Allan Cox *in* Malde, 1971b; Kuntz and Dalrymple, 1979; Kuntz and others, 1980; Champion and others, 1981), showing that virtually all of them erupted during the Brunhes Normal Polarity Chron—or in the last 750,000 years. Radiocarbon dating of charred eolian sediment shows that several of the basalts are Holocene (Kuntz and others, 1986a).

Resistivity measurements (Zohdy and Bisdorf, 1980) and drilling indicate that the basalt is from less than 1 km to at most 2 km thick. The basalt thins toward the margins of the plain, partly by abutting older rocks and partly by intertonguing with marginal alluvium and lake beds (Walker, 1964; Scott, 1982; Rightmire, 1984). The basalt also becomes thin in the westernmost part of the eastern Snake River Plain, as shown by drilling and by an abrupt steepening of the water table (see the discussion of the Snake River Plain aquifer).

Basaltic vents are relatively numerous near the axis of the eastern Snake River Plain (Malde and others, 1963; LaPoint, 1977; Kuntz, 1978; Kuntz and others, 1979; Spear and King, 1982). As a consequence, the basalt forms a subdued medial ridge, which is recognizable along most of the length of the eastern plain. Because of the ridge of basalt, streams north of the plain either end in playas near the northern margin, as previously explained, or, in the case of the Little Wood and Big Wood rivers, turn westward and eventually join the Snake River where the basalt ends.

The distribution of basalt vents also is influenced by volcanic rift zones, which cross the eastern Snake River Plain approximately at right angles to its axis (Kuntz, 1978). The longest and best known of these is the Great Rift, 85 km long, reaching from mountains just north of the plain southeastward through Craters of the Moon National Monument to the Wapi lava field west of American Falls (Fig. 9; Kuntz and others, 1986b). Thus, the Great Rift extends across most of the eastern Snake River Plain. At least seven other volcanic rift zones were sources of basalt of Holocene and latest Pleistocene age (Kuntz and others, 1986a, 1986b). The wide distribution of these eruptions, reaching from longitude 114°30' W to 112° W, indicates that basaltic volcanism occurred along most of the length of the eastern plain during the Holocene. Earlier volcanic activity along volcanic rift zones also is recognized, as shown by linear arrays of volcanic landforms and structures (Kuntz and Dalrymple, 1979), and it

seems likely that such zones were sources for many of the widespread Quaternary basalts.

Near the center of the eastern Snake River Plain, northwest of Blackfoot, the basalt is intruded by four prominent rhyolite domes, all probably of Quaternary age (Fig. 9; Kuntz and others, 1984): Big Southern Butte (0.304 ± 0.022 Ma); Middle Butte; an unnamed butte (1.42 ± 0.02 Ma); and East Butte (0.58 ± 0.09 Ma). The domes are aligned parallel to the northeasterly trend of the plain, and their alignment is thought to express structural control (Spear and King, 1982) or control by ring fractures along the margin of a buried caldera (Kuntz and Dalrymple, 1979). The domes are in an area where relatively high crustal temperatures are inferred (B. K. Bhattacharyya *in* Mabey, 1983). Another rhyolite intrusion at Juniper Buttes northwest of St. Anthony has formed a faulted dome (Kuntz, 1979), which may have been uplifted in the early Quaternary (M. A. Kuntz, U.S. Geological Survey, written communication, 1986).

Crustal structure

The origin of the eastern Snake River Plain still is incompletely understood, but its structural history is thought to involve the total thickness of the crust, as modified by a hot spot within the upper mantle. The crustal structure is deduced from seismic studies (Iyer, 1984a), supported by results from gravity and magnetic surveys and by data obtained from other geophysical techniques (Mabey, 1982).

Teleseismic profiles across the eastern Snake River Plain near Rexburg and Blackfoot show a large anomalous body about 100 km wide extending to a depth of more than 200 km (Evans, 1982). The body is distinguished by low compressional wave velocities, which imply partial melting. Evans interpreted this body as a residual feature of a magmatic center, similar to the magmatic body now located under Yellowstone Park.

The anomalous crustal structure of the eastern Snake River Plain is further revealed by seismic refraction profiles (Fig. 5). In the surrounding region, an upper-crustal layer of crystalline rock, with a velocity of 6.1 km/s, is 20 km thick, but this layer is only 2 to 3 km thick within the eastern Snake River Plain. Volcanic rocks 3 to 6 km thick above the crystalline layer in the eastern plain account in part for its thinness, but most of the thinning is caused by an anomalous underlying "intermediate" layer with a velocity of 6.5 km/s. The presence of the intermediate layer suggests thinning of granitic crust from below by mafic intrusions (Iyer, 1984a). Such intrusions could have melted the granitic crust, giving rise to silicic volcanism, and eventually could have reached the surface as lava flows of basalt (Christiansen and McKee, 1978; Leeman, 1982a). Below the intermediate layer, a lower crust of nearly uniform thickness, with a velocity of 6.8 km/s, is continuous between the eastern Snake River Plain and the surrounding region.

Gravity surveys indicate that the contrast in density between the eastern Snake River Plain and the surrounding region is fairly subdued, and Mabey (1978) found nothing in the gravity data to

suggest that the upper crust beneath the eastern plain is distinctly different from the crust to the north and south. Mabey (1978) interpreted the Bouguer gravity map of the eastern Snake River Plain (Fig. 6) as indicating from 4.5 to 7 km of thinning of the upper crust. The thinning is expressed by a broad regional gravity high centered over the eastern plain but extending well beyond the edges of the plain. Superimposed on the regional gravity high is a complex pattern of highs and lows, which undoubtedly reflects differences in thickness and density among the volcanic rocks of the plain itself. The trend of this pattern is across the eastern plain, parallel to elongate gravity anomalies associated with basin-and-range topography to the north and south, although the highs and lows of the plain are twice as broad. The gravity map further shows that some basin-and-range structures end before reaching the plain, some terminate at the edges, and some extend a few kilometers onto the plain. From this evidence, the structural border of the eastern Snake River Plain is more irregular than the fairly straight borders expressed by its physiographic boundary (Mabey, 1978).

The eastern Snake River Plain is a region of relatively high magnetic intensity (Zietz and others, 1978), partly because of its strongly magnetized volcanic rocks and partly because such magnetized rocks are more abundant in the plain than in the surrounding region (Mabey, 1978). Most of the magnetic intensity probably comes from the upper 1 to 2 km of basalt. The intensity is expressed as a complex pattern of highs and lows with considerable amplitude, which (like the gravity pattern) displays an elongate grain perpendicular to the axis of the plain (Fig. 7). Some of the small-scale magnetic anomalies can be linked with exposed volcanic features, but sources for the broad magnetic highs are thought to be deeper (Bhattacharyya and Mabey, 1980). The computed depth to the base of the magnetized crust, which marks the depth to the Curie isotherm (or about 580°C), is fairly shallow: from 12 to 20 km (B. K. Bhattacharyya *in* Mabey, 1983). Thus, the character of the magnetized crust provides further evidence for a shallow source of heat under the eastern Snake River Plain.

The presence of high temperatures at shallow depth under the eastern plain also is implied by magnetotelluric soundings, which provide information about the electrical structure of the crust to a depth of 25 km (Stanley, 1982). High electrical conductivities are associated with high temperatures. The soundings suggest that a highly conductive layer—hence, a temperature greater than 700°C—is reached at depths ranging from 12 to 20 km. This range in depth is consistent with the thermal gradient implied by the computed depth of the Curie isotherm and with the presence of a partially melted body inferred from teleseismic studies.

From numerous measurements of heat flow in southern Idaho, Brott and others (1981) proposed a thermal model in which a heat source at constant temperature under the Snake River Plain migrated relatively northeastward to Yellowstone Park at a rate of 3.5 cm/yr. This rate corresponds with the rate of progress of silicic volcanism inferred from the distribution and ages of the volcanic rocks, as previously explained. The model

predicts a temperature at the melting point of basalt (1,070°C) just below the Yellowstone caldera, the present site of the heat source (Christiansen, 1984). The model further predicts a temperature of 1,000°C at a depth of 25 km under the axis of the plain northwest of Idaho Falls. The model accounts for the systematic southwesterly decrease in elevation of the eastern Snake River Plain as a consequence of thermal contraction, after any given point had passed over the heat source. Such contraction also would produce the inward dipping attitudes of ash-flow tuffs at the margins of the eastern plain. The thermal contraction model, however, does not account for inconsistently high altitudes on the Owyhee Plateau southwest of Twin Falls, where the volcanic rocks seem to be the forerunners of those under the eastern Snake River Plain. For instance, the altitude of a plateau of rhyolite and basalt at the southwest corner of Idaho is about 1,600 m, not the 900 m predicted by the thermal contraction model. In support of the thermal model, precise leveling indicates that the eastern Snake River Plain currently is subsiding relative to points to the north (Reilinger and others, 1977).

The eastern Snake River Plain is a region of low seismicity, even though its eastern end lies across the north-south trend of the Intermountain Seismic Belt (Smith, 1978) and across the trend of a corresponding belt of neotectonic activity (Scott and others, 1985). Shear wave velocities under the eastern Snake River Plain also are lower than in the Basin and Range Province to the south (summarized by Iyer, 1984a). Mabey (1983) inferred from such evidence that strain in the eastern Snake River Plain is accommodated with little brittle fracturing, because of the high temperature of the crust.

Despite low seismicity in the eastern Snake River Plain, the presence of volcanic rift zones and widely distributed Quaternary basalt implies that deep-seated fractures are present. As sources of basalt, the volcanic rift zones must extend deeply beneath the eastern Snake River Plain, and they presumably are signs of tension in the crust. The identified volcanic rift zones are extensions of range-front faults to the north (Kuntz, 1978; Kuntz and Dalrymple, 1979) and possibly are further evidence of the regional extension expressed by basin-and-range structures to the north and south, which have the same trend. Whether movements on the range-front faults are linked with volcanism along the volcanic rift zones cannot be determined from available dates, but a study prompted by the 1983 Borah Peak earthquake identified recurrent late Pleistocene movements on the faults (Scott and others, 1985). On the other hand, the volcanic rift zones may simply express thermal contraction and, hence, may be features of the thermal history of the eastern Snake River Plain (Brott and others, 1981).

Origin of the volcanic plain

Explanations for the origin of the eastern Snake River Plain must account for its bimodal association of rhyolite and basalt and its apparent lack of underlying Paleozoic rocks. Although xenoliths of crystalline rocks occasionally are found in the vol-

canic rocks, no Paleozoic inclusions have been reported. Three explanations for the missing Paleozoic rocks seem worth considering: the eastern Snake River Plain represents an extensional rift; substantial blocks of Paleozoic rocks exist under the eastern plain but have not been carried up as inclusions in the volcanic rocks; and Paleozoic rocks have been eliminated from the eastern Snake River Plain by emplacement of its volcanic rocks, particularly the rhyolites.

A structural origin for the eastern Snake River Plain as an extensional rift was once considered likely from inferred relative movements and stresses within the western North American plate (for example, as modeled by Furlong, 1979). This origin, however, is inconsistent with seismic, gravity, and magnetic data. As reviewed above, this evidence shows a lower crust under the eastern Snake River Plain similar to the lower crust in the surrounding region, not the anomalously thick lower crust that implies rifting in the western Snake River Plain. Furthermore, geophysical and geological data from the eastern Snake River Plain imply a crust modified by volcanic processes, not a trough bounded by high-angle faults. Structurally, the trend of the eastern Snake River Plain is nearly perpendicular to extensional basin-and-range structures to the north and south (Stewart, 1978). No convincing hypothesis has been proposed that might account for both of these opposing structural trends as tensional features.

Possible survival of blocks of Paleozoic rocks under the eastern Snake River Plain is thus far an imponderable hypothesis, lacking any evidence that such rocks are present. If Paleozoic rocks occur beneath the volcanic rocks, they might be expected to be carried upward as inclusions in volcanic conduits, as has happened with xenoliths of crystalline rocks. The presence of large blocks of Paleozoic rocks might be inferred from a suitable analysis of magnetic data, which could show areas of anomalously low magnetic intensity. Such an analysis would require subtracting the magnetic effects of the basalt, which is the source of most of the small-scale differences in magnetic intensity. To the degree that the magnetic patterns reflect differences in the thickness and magnetism of the basalt, such an analysis would further depend on much more knowledge of the distribution and magnetic properties of the basalt than is now available.

A complete lack of Paleozoic rocks beneath the eastern Snake River Plain is a likely consequence of the processes by which its silicic volcanic rocks were emplaced. An analogy is found in the contiguous caldera-forming pyroclastic eruptions of the Yellowstone Plateau volcanic field. Although Precambrian and Paleozoic rocks have been inferred to extend under the Yellowstone volcanic deposits (Ruppel, 1972), their presence would be incompatible with seismic evidence and with current understanding of the evolution of calderas. As in the eastern Snake River Plain, the Yellowstone Plateau volcanic field is underlain by an upper crustal layer with low seismic velocity and by an underlying intermediate layer (Fig. 5; Smith and Braille, 1984). Further, from a detailed study of the caldera-forming eruptions, Christiansen (1984) concluded that the eruptions were preceded

by evolution of granitic magma and that each caldera is underlain by a granitic batholith. His concept of how the calderas evolved is supported by geophysical evidence (Iyer, 1984b) and by studies of many calderas in the western United States (Lipman, 1984). Because three overlapping calderas are recognized in the Yellowstone Plateau volcanic field, a group of coalescing batholiths can be inferred, measuring about 50 by 110 km. These dimensions are comparable to those inferred for several overlapping calderas that are thought to underlie the full width of the upper end of the eastern Snake River Plain (Morgan and others, 1984). Batholiths beneath these calderas would have displaced any Paleozoic rocks that may have been present. Such a process, extended back through time, could account for the apparent lack of Paleozoic rocks beneath the full length of the eastern Snake River Plain. Indeed, emplacement of overlapping batholiths presumably also accounts for the lack of older rocks in the Owyhee Plateau farther southwest, along the eastern Snake River Plain trend.

The volcanic processes that produced the crustal structure of the eastern Snake River Plain, at any given point along the track of the hot spot, are now fairly well understood (Leeman, 1982a). The stages of volcanism can be seen by following the apparent path of the hot spot backwards from Yellowstone Park through the eastern Snake River Plain, and thence through the Owyhee Plateau to southeastern Oregon (Fig. 1). Sites of volcanic activity along the path then can be understood as places where the same volcanic sequence was enacted, in general, at successively earlier times. At each volcanic center, volcanism was preceded by intrusion of basaltic magma into the granitic crust. For this reason, the "fundamental character" of the volcanism has been described as being basaltic (Christiansen and McKee, 1978). Intrusion of basaltic magma increased the regional heat flow, caused crustal thinning, produced rhyolitic magmas by local melting of the granitic crust, and caused regional uplift by thermal expansion. In this way, granitic batholiths were formed that became sources for explosive pyroclastic eruptions of rhyolite, usually followed by collapse of associated calderas (Christiansen, 1984). Occasionally, basaltic lava erupted concurrently with the rhyolite. After a batholith cooled and solidified, basalt could erupt through the rhyolite, thus producing bimodal complexes of rhyolite and basalt, as at Island Park (summarized by Christiansen, 1984). With further cooling, thermal contraction caused former areas of silicic volcanism to subside and produced inward dips toward the path modified by the hot spot (Brott and others, 1981). The rhyolite eventually became completely covered by basalt, a phenomenon now found throughout the eastern Snake River Plain but not yet developed in Yellowstone Park. Basaltic volcanism in the eastern Snake River Plain, however, is not just the final volcanic stage at a given volcanic center. Rather, basalt has erupted widely from fractures that have formed perpendicular to the trend of the eastern plain. The widespread basaltic eruptions include lavas of Holocene age, suggesting that the eastern Snake River Plain still may be underlain by widely scattered sources of basaltic magma. Such magma may account for local melting of the granitic crust or for remelting of silicic volcanic rocks, as expressed by the

Quaternary rhyolite domes northwest of Blackfoot and at Juniper Buttes.

Character of the basalt eruptions

Considerable information has been gathered about the petrography and geochemistry of basalts of the eastern Snake River Plain and about the distribution and form of the eruptive products. Moreover, from initial studies of the ages and volumes of basalts of Holocene and latest Pleistocene age, a beginning has been made toward estimating rates of output of the basalt and, hence, predicting when and where further eruptions are likely to occur.

Analyses of basalts in the eastern Snake River Plain, supplemented by studies of basalts in the western Snake River Plain, show that the basalts are chemically distinctive compared to most other basalts, being low in SiO_2 and total alkalis and high in total iron (Stout and Nicholls, 1977; Leeman, 1982b; Menzies and others, 1984). The basalts are classified as both olivine tholeiites and as alkali olivine basalts. The basalts typically contain up to 20 percent phenocrysts of plagioclase and olivine, and hand specimens of various lava flows in a given area usually can be distinguished from each other by minor differences in texture, crystallinity, mineralogy, and other lithologic properties. Two dominant magmatic types are recognized: a group of relatively primitive, low-potassium lavas (usually referred to as the Snake River Plain type), derived from the continental lithosphere at depths of 60 to 90 km; and a group of variably fractionated or "evolved" lavas, which are attributed to liquid-crystal fractionation and to partial melting of peridotite mantle at depths of 40 to 60 km. In addition, a few lavas represent a contaminated type, also called "hybrid" lavas, which may have interacted with crustal rocks and fractionated at depths of about 30 km. All three magmatic types are represented by lava flows that erupted along the Great Rift (Kuntz and others, 1986b).

The basalt of the eastern Snake River Plain forms a nearly flat constructional surface; local relief over distances of 10 to 20 km rarely exceeds 100 m. Surfaces of individual lava flows are marked by flow ridges, pressure ridges, and pressure plateaus, which make for as much as 10 m of hummocky relief. Basalts of Holocene age are so rough and so little covered by surficial deposits that they are easily seen on Landsat images as well-defined dark areas (Rothery and Lefebvre, 1985). In contrast, the surfaces of late Pleistocene and older basalts typically are more than 25 percent covered by loess and eolian sand (Scott, 1982), although such basalts are otherwise little modified by weathering or erosion. For some of the oldest lava flows, the degree of cover is so great that virtually none of their volcanic features are visible. Drilling shows that similar surficial deposits commonly form thin interbeds between lava flows (Kuntz and others, 1980). No well-defined drainage systems have developed on the basalt, but surface drainage collects in a few small playas, commonly in shallow depressions between adjoining lava flows. Most runoff, however, seeps into the basalt and becomes part of the Snake River Plain

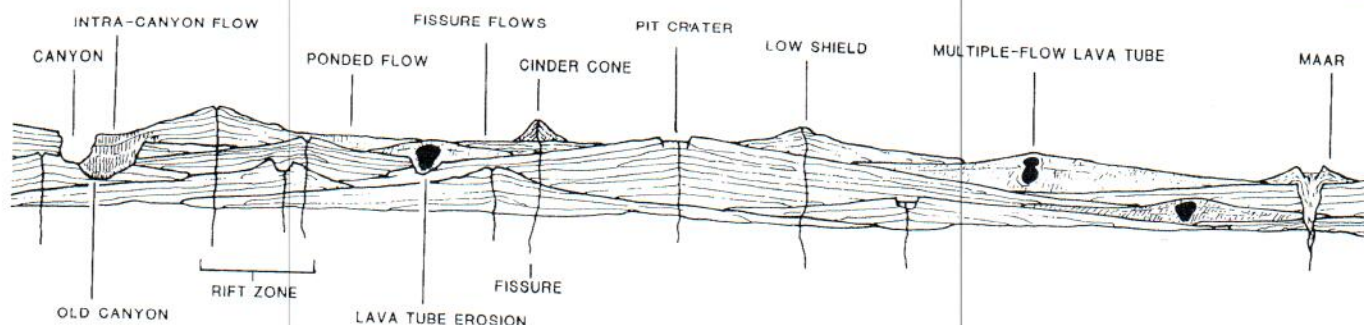


Figure 10. Schematic cross section of constructional basaltic landforms in the eastern Snake River Plain, greatly exaggerated in vertical scale. Reproduced from Greeley (1982).

regional aquifer system (see the discussion of the Snake River Plain aquifer).

The basaltic eruptions can have considerable morphological diversity. Greeley (1982) distinguished three principal landforms: shield volcanoes, fissure flows, and tube-fed flows. In addition, several minor morphological features distinctive of basalt eruptions are fairly common (Greeley, 1982; King, 1982), and these make the eastern Snake River Plain a useful analog for interpreting certain planetary features (R. Greeley in Baker and others, 1987). In a few places, lava that interacted with ground water produced phreatic eruptions (Creighton and King, 1981; Womer and others, 1982). Most descriptions of the various constructional volcanic forms pertain to their surface appearance and topography, because cross sections through the basalt are exposed only along the Snake River canyon downstream from Milner Dam. Lava flows exposed along the canyon are unusual, however, in that they occur as canyon-filling lavas not known to be present elsewhere in the eastern Snake River Plain.

Shield volcanoes are the dominant basaltic landforms of the eastern Snake River Plain, making up perhaps three-fourths of the volume of the basalt (Kuntz, 1978). Shield volcanoes are broad, dome-shaped cones, built by hundreds or thousands of lava flows from a single vent. A typical shield volcano in the eastern Snake River Plain contains 7 km^3 of lava and is 15 km across at the base (Greeley, 1982). Except for the youngest volcanoes, parts of their edges commonly are covered by overlapping eruptions. The shield volcanoes are formed almost entirely from successive highly fluid pahoehoe flows 5 to 10 m thick, which are capable of spreading 10 km or more. Rare flows of more viscous a'a lava, 15 to 20 m thick, also are present (Kuntz and Dalrymple, 1979). Shield volcanoes on the eastern Snake River Plain have slopes of very low gradient, 1 to 2° , and typically rise to elevations only 50 to 150 m above the base. They account for most of the relief on the basalt plain. Greeley called the shield volcanoes of the Snake River Plain "low shield volcanoes," thus distinguishing them from higher shield volcanoes in Hawaii and elsewhere. Figure 10

is a diagrammatic cross section showing how the basalt plain may have been constructed.

The basaltic landforms of the eastern Snake River Plain also can be described as being typical of eruptions along volcanic rift zones. The resulting landforms are analogous to those from volcanic rifts in Hawaii (Kuntz, 1978) and are shown particularly well by basalt that erupted along the Great Rift (Kuntz and others, 1986b). There, fissure eruptions were followed by spatter cones, cinder cones, and lava cones. Eventually, shield volcanoes were built, some being from vents elongated in the direction of the volcanic rift. The Wapi lava field on the Great Rift is such a shield volcano, covering 300 km^2 and containing 6 km^3 of lava—one of the largest shield volcanoes on the eastern Snake River Plain (Kuntz and others, 1986b). The Wapi lava field erupted at about 2,250 years B.P. (Kuntz and others, 1986a). Similar features of eruptions along volcanic rift zones are found in various stages of development throughout the eastern Snake River Plain (Kuntz, 1978).

Studies of volcanic hazards at the Idaho National Engineering Laboratory (Fig. 9) have led to preliminary estimates of the probable frequency and extent of lava flows on the eastern Snake River Plain. For instance, by assuming that 75 percent of the basalt is represented by shield volcanoes, each with an average volume of 7 km^3 , and by calculating the volume of the basalt as $23,600 \text{ km}^3$, Kuntz (1978) inferred 2,500 shield volcanoes. Given that the basalt may be as old as 3.5 Ma, a shield volcano would have formed every 1,400 years. In local areas, however, intervals between eruptions obviously would have been longer. For example, in one area of the Idaho National Engineering Laboratory, K-Ar ages of lava flows from seven vents indicated major eruptive episodes at 450, 225, and 75 ka (Kuntz and others, 1980). The volcanic activity, however, was episodic, rather than periodic, such that intervals between major eruptions were as long as 225,000 years and as short as 150,000 years. In another area, at least five major eruptions from various sources were identified in the last 500,000 years, the average recurrence interval being

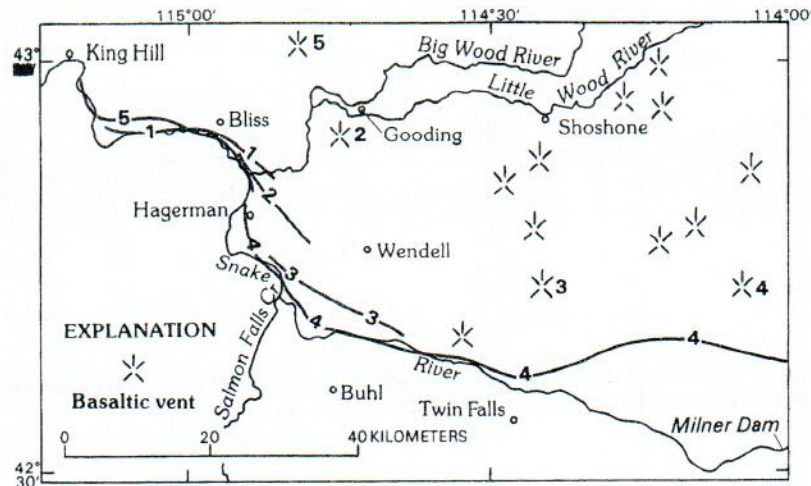


Figure 11. Map showing the approximate axis of each successive Quaternary canyon cut by the ancestral Snake River in the westernmost part of the eastern Snake River Plain (blue lines, numbered 1 through 5 from oldest to youngest). Madson Basalt, probably from an unidentified vent east of Gooding, filled the first canyon. Subsequent canyons were filled with basalts from corresponding late Quaternary basaltic vents, as follows: (2) Malad Member of Thousand Springs Basalt, (3) Thousand Springs Basalt, (4) Sand Springs Basalt, and (5) McKinney Basalt. Adapted from Covington and others (1985).

about 80,000 to 100,000 years (Kuntz and Dalrymple, 1979). Again, K-Ar ages indicated that the lava erupted at irregular intervals; the times between flows were possibly as short as 10,000 to 100,000 years and as long as 300,000 to 400,000 years.

The frequency and volume of some basalt eruptions of Holocene and latest Pleistocene age also have been estimated, particularly along the Great Rift (Fig. 9). Eruptions at Craters of the Moon National Monument on the Great Rift were assigned to eight eruptive periods, ranging from 15,000 to 2,000 years B.P., using radiocarbon dates for 20 of the 60 lava flows (Kuntz and others, 1986a). Then, from measurements of the erupted volumes (Kuntz and others, 1986b), the "magma output rate" was determined to have been $1.5 \text{ km}^3/1,000 \text{ yr}$ between 15,000 and 7,000 years B.P., and $2.8 \text{ km}^3/1,000 \text{ yr}$ from 7,000 to 2,000 years B.P. The period of repose between eruptions ranged from 500 to 3,000 years and averaged about 2,000 years—this interval being the time evidently needed for local driving forces to produce an eruption. For comparison, Kuntz and others (1986b) pointed out that output rates at Craters of the Moon were about 10 times less than those for Hawaiian volcanoes, about 20 times less than the rate for Iceland, and 40 times less than the rate inferred for the Columbia Plateau.

These data indicate that volcanism at Craters of the Moon took place at a fairly uniform rate during the last 15,000 years. This finding, in light of other results from related geological and chemical studies of the lava flows, led Kuntz and others (1986b) to speculate that another eruptive period at Craters of the Moon is likely in the next 1,000 years, when from 5 to 6 km^3 of lava might be erupted.

Displacements of the ancestral Snake River by lava flows

The present canyon of the Snake River downstream from Milner Dam intersects a series of ancestral canyons filled with basaltic lava flows, which erupted from vents near the western end of the eastern Snake River Plain (Malde, 1971b; Covington and others, 1985). Because of these canyon-filling basalts, the river was diverted progressively farther south, each time cutting a new canyon that was subsequently filled with still younger basalt (Fig. 11). In this way, the Snake River was displaced about 30 km south, and the present course of the river around the southern and western limits of the lava flows was established.

The earliest of the canyon-filling lava flows, the Madson Basalt, has a K-Ar age of $0.54 \pm 0.08 \text{ Ma}$ (Armstrong and others, 1975, sample 69-53[L]); and the youngest, the McKinney Basalt, is perhaps 70 to 50 ka, judging from the lack of buried soils in overlying loess (Pierce and others, 1982). All the basalts are fresh, and several of them have normal magnetic polarity.

The canyon-filling lava flows are exposed in the northern wall of the present canyon between Twin Falls and King Hill. In several places near Hagerman, the filled canyons are incised in relatively impermeable sedimentary deposits of the Glens Ferry Formation. Farther upstream, the ancestral canyons are incised in Banbury Basalt or in basalt of the Glens Ferry Formation, both of which also are relatively impermeable.

The three earliest canyons are at least 60 m deep. By the time of the fourth canyon-cutting episode, which is represented by Sand Springs Basalt, canyon cutting had reached the depth of

the present canyon—or 120 m below the canyon rim at Hagerman. McKinney Basalt of the fifth, and last, canyon-filling episode fills a canyon near Bliss 30 m deeper than the present canyon (Malde, 1982).

The canyon-filling basalts differ from most other basaltic rocks in the eastern Snake River Plain in that they consist largely of pillow lava and, in places, fragmental glassy basalt. Such deposits accumulated in deep water upstream from lava dams, and the thickness of a given subaqueous basalt is evidence for the depth of the impounded water. For example, a massive section of McKinney pillow lava is more than 150 m thick in an ancestral canyon just west of Bliss (Malde, 1971b). Downstream from the lava dams, the ancestral canyons were relatively dry, and canyon-filling lava flows were deposited as subaerial columnar basalt. The canyon-filling lava flows are further represented by subaerial basalts that flowed across the upland to reach the ancestral Snake River, often a considerable distance. The longest lava flow, the Sand Springs Basalt, extends 80 km from its vent to the terminus near Hagerman.

Lake beds also were deposited in the water impounded by the lava dams. In particular, scattered outcrops of finely laminated clay extend 40 km upstream from the dam formed by the McKinney Basalt, and the clay reaches altitudes nearly as high as the lava dam (Malde, 1982). Farther upstream, sediment deposited behind a dam formed by the Sand Springs Basalt may be present in the city well at Burley (Fig. 12), where "150 feet of alluvium was deposited in the lake" (Stearns and others, 1938). These lacustrine deposits imply that their respective lava dams leaked as much water as was supplied by the river (Stearns and others, 1938). If not, the impounded water would have overflowed, downcutting would have begun, and the lacustrine deposits would not have reached their observed heights (Malde, 1982).

As discussed below, the canyon-filling lava flows are major points of discharge for the Snake River Plain aquifer, which produces numerous springs in the northern wall of the Snake River canyon between Twin Falls and King Hill. Recent unpublished mapping by H. R. Covington, U.S. Geological Survey, confirms that all the larger springs discharge from canyon-filling basalts, mainly from pillow lava and basaltic sand. These canyon-filling volcanic rocks, because of their thickness and high hydraulic conductivity, have "very high transmissivity and extremely high storage capacity" (Covington and others, 1985).

Besides the hydrologic importance of the canyon-filling lava flows, their history of interplay with the ancestral Snake River has had a marked effect on development of the regional physiography. The lava plain formed by the lava flows is topographically high with respect to the broad canyon floor in the sedimentary basin west of King Hill, and it has been so since downcutting to the west began. That is, the general altitude of the lava plain was maintained—or somewhat increased—each time that the river was diverted by a canyon-filling eruption. The river then was forced to begin a new cycle of downcutting, while downcutting continued unimpeded to the west. At King Hill, the Snake River was 120 m above its present level when the Madson Basalt

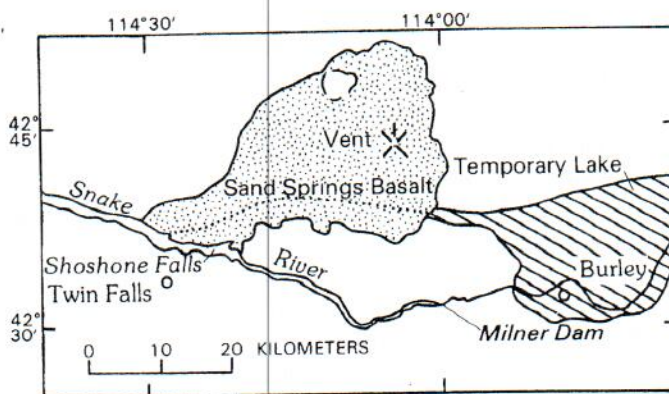


Figure 12. Map of the Snake River between Burley and Twin Falls, Idaho, showing diversion of the former Snake River (dotted line) by eruption of Sand Springs Basalt. A temporary lake impounded by the basalt spilled westward and cut a new canyon that heads at Milner Dam. The new canyon joins the wide, prebasalt canyon at Twin Falls. The ragged canyon that extends 12 km upstream from Twin Falls was enlarged by the late Pleistocene Bonneville Flood. Modified from Covington and others (1985).

erupted, as measured by a terrace of Sugar Bowl Gravel just above the terminus of the Madson. By the time of the McKinney Basalt, which terminates at the level of the Snake River near King Hill, the river farther west had reached its present depth of downcutting. Meanwhile, the process of downcutting continues above King Hill, and the Snake River from King Hill to Twin Falls still has an average gradient of nearly 2 m/km—one of the steepest gradients of any large river in the United States. In contrast, the gradient of the Snake River in the next 140 km below King Hill is about 0.4 m/km.

Marginal lacustrine and alluvial deposits

Three principal bodies of Quaternary lake beds and alluvium were deposited in topographically low areas at the margins of the eastern Snake River Plain (Fig. 9): the Raft Formation, present in the lower Raft River Valley and along the Snake River for about 35 km upstream; the American Falls Lake Beds, exposed around the shores of the American Falls Reservoir and for about 10 km downstream; and deposits of Lake Terretion, which formerly covered a wide area near Mud Lake (Scott, 1982).

The Raft Formation consists of fine-grained alluvium, lacustrine deposits, and possibly loess (Carr and Trimble, 1963). Calcareous concretions are locally abundant, and a few gravel lenses and local layers of basaltic tuff and silicic volcanic ash also are present. The formation has an exposed thickness of 45 m near the mouth of the Raft River, if 25 m of loessial silt at the top is excluded (Scott, 1982). Drilling indicates that the formation may be as much as 360 m thick in the lower part of the Raft River Valley (Williams and others, 1982) and 48 m thick in a well

northeast of American Falls (Carr and Trimble, 1963). The beds dip gently northwestward and are thought to have accumulated in response to a rising base level caused by buildup of basalt "some distance downstream from the mouth of the Raft River" (Williams and others, 1982). Fossil mollusks indicate a probable late Pleistocene age (D. W. Taylor in Carr and Trimble, 1963). An older age, however, is indicated by a bed of silicic volcanic ash assigned to the Raft Formation at a locality 10 km east of the mouth of the Raft River (Izett, 1981, bed 6). This volcanic ash may correlate with an ash bed at Thornton Beach near San Francisco, now informally named the Rockland ash bed (Sarna-Wojcicki and others, 1985), which is dated at 0.4 Ma.

Fine-grained deposits around the American Falls Reservoir are related, in part, to a lava dam formed by Cedar Butte, 17 km southwest of American Falls (summarized by Carr and Trimble, 1963). Cedar Butte impounded a lake 65 km long and nearly 20 km wide, which extended upstream to Blackfoot. Carr and Trimble (1963) thought that this lake accounted for most of the American Falls Lake Beds, which have an exposed thickness of 15 m. Recent study, however, shows that only the upper 2 to 6 m is lacustrine. Scott and others (1982) reviewed current understanding of the stratigraphy, paleoenvironment, and geologic history of these deposits. The lacustrine part of the section consists of diatomaceous clay, laminated silty clay, and (at the top) peat. Alluvium composed of evenly bedded and cross-bedded silt, sand, clay, and fine gravel under the lacustrine beds may have been deposited because an early lava flow from Cedar Butte (or from some other nearby vent) raised the base level of the Snake River. A thin, basal gravel contains abundant vertebrate fossils, including *Bison latifrons* (Hopkins and others, 1969), and the lacustrine beds yield pollen, spores, mollusks, diatoms, and remains of plants, which indicate a climate and vegetation similar to the present (Bright, 1982). The K-Ar age for basalt exposed at Cedar Butte, which is correlated with the lacustrine beds, is 72 ± 14 ka (Scott and others, 1982). Peat at the top of the section is dated by radiocarbon as greater than 51,900 years B.P. (Bright, 1982). The peat provides a maximum age for fossil mollusks and vertebrates (Carr and Trimble, 1963; Lundelius and others, 1983) found in younger alluvial deposits (Pierce and Scott, 1982).

Mud Lake just north of Terreton is the principal remnant of a shallow lake, Lake Terreton, which formerly covered a wide area in the northeastern part of the eastern Snake River Plain. Stearns and others (1939) described the Lake Terreton deposits as "lake beds near Terreton" and attributed them to runoff from streams on the north, which were impounded by lava flows to the south. The broad depression occupied by Lake Terreton is an undissected closed basin. Topographic relief in the basin is subdued, and a small change in the level of Lake Terreton would have made a large change in its areal extent. Scott (1982) pointed out that the lake probably overflowed eastward into the Snake River at a threshold now concealed by younger basalt. As mapped by Scott, the lake beds cover 600 km² but are concealed in many places by stabilized linear dunes (see the discussion below of loess and eolian sand). Well records show that the lake

beds rest on an irregular basalt surface and have a maximum thickness of 42 m, consisting of silt with lesser amounts of sand and clay (Nace and others, 1975). Well records further show that lake beds on the northern side of the basin merge with deltaic deposits and alluvium (E. H. Walker, 1964). The lake beds are thought to have been deposits during the Pinedale glaciation, when runoff would have been relatively large (Kuntz and others, 1979; Scott, 1982). Stearns and others (1939) suggested that an east-west scarp in lake beds north of Mud Lake represents 15 m of vertical displacement on a high-angle fault, but evidence for such a fault has not been found by exploratory drilling (Malde, 1971a).

Loess and eolian sand

Basalt in much of the eastern Snake River Plain is covered by loess more than a meter thick, and a still larger area has scattered patches of loess at least half a meter thick (Lewis and Fosberg, 1982; Scott, 1982). The Holocene basalt, however, has no loessial cover. Where the basalt is more than 70 percent covered, the loess is commonly 1 to 4 m thick, and where it is virtually all covered, the loess is as much as 30 m thick. In uplands south and east of the Snake River, the thickness of the loess is commonly 7 to 15 m and in places exceeds 60 m. Loess also is found in hilly uplands along rivers that drain mountainous areas in southeastern Idaho and adjacent Wyoming (Glenn and others, 1983). The absence of loess on Holocene basalt indicates that the loess is Pleistocene, and the thick deposits overlie comparatively old basalt. That is, the presence and thickness of loess are factors related to the ages of the basalt. Particle sizes in the loess show no consistent pattern on the Snake River Plain, but the particles decrease in size east and southeast of the Snake River with increasing distance from the river (Lewis and Fosberg, 1982). This trend suggests that the local flood plain of the Snake River was a major source of the loess. Even so, because some loess is found as far west as the Bruneau River, more western sources also must have been present.

Deep exposures and auger holes in loess south and east of the Snake River show one or more buried soils that contain more carbonate, clay, and soluble salts than are found in the surface soil (Lewis and Fosberg, 1982). One of the buried soils, "better developed than the surface soil," is considered to be a widespread stratigraphic marker, separating upper and lower units of loess along a transect reaching from West Yellowstone to the Raft River Valley (Pierce and others, 1982). By correlating this buried soil with a buried soil developed on nearby glacial deposits dated at 140 to 150 ka and by noting the absence of the buried soil in loess above Cedar Butte Basalt, Pierce and others (1982) concluded that the lower loess unit was deposited by about 140 to 150 ka, and that the upper loess unit ranges in age from 70 to 10 ka. This chronology provides a basis for estimating the minimum ages of basalts covered by these loessial units.

Eolian sand occurs on the eastern Snake River Plain as stabilized dunes and as a few hills of active dunes (Scott, 1982). The stabilized dunes form two linear areas: one, a broad band 15

km wide, stretches northeastward 80 km from the Big Lost River playa along the northern margin of the plain; the other, a band of irregular width, extends 140 km along the Snake River from Milner Dam to Idaho Falls (Fig. 9). Stabilized dunes also occur at Juniper Buttes and along a short stretch northeast of St. Anthony. All these areas are marked by long, straight, parallel ridges variously called linear, longitudinal, or seif dunes, which are 0.7 to 7.0 m high and 20 to 60 m wide. Although vegetated with grass and shrubs, the dunes are subject to deflation where the ground cover is disturbed, and some areas have blowing and drifting sand (Scott, 1982). Dunes of the northern band overlie deposits of Lake Terretion, and those along the Snake River seem to be partly younger than the loess. These relations imply that some of the stabilized dunes are Holocene. The distribution of the dunes, however, indicates that they were derived from terminal playas of streams on the north and from the flood plain of the Snake River on the south, both being areas most abundantly supplied with sand by glacial outwash. Hence, some of the stabilized dunes are probably of late Pleistocene or older age.

Active dunes on the eastern Snake River Plain are confined to an area around Juniper Buttes, where hills of blowing and drifting sand and migrating transverse and U-shaped dunes rise from a floor of stabilized dunes. Features of the dune field have been mapped in detail by Kuntz (1979) and have been reviewed by Smith (1982). The dune field splits around Juniper Buttes in the form of a tuning fork, making an area of active and stabilized dunes 20 to 30 km long and 10 to 15 km wide. One dune on the south flank of Juniper Buttes is about 100 m high. A geomorphic study of dune movement determined that slip faces on the U-shaped dunes had advanced northeastward 13 to 28 m between 1941 and 1972. A study of vegetation on the dunes, however, suggested a much faster rate of advance, 3 m per year, and implied that the dunes may become stable only in another 1,000 years.

SNAKE RIVER PLAIN AQUIFER

Basalt of the eastern Snake River Plain makes of the Snake River Plain aquifer, one of the world's most productive ground-water systems (Mundorff and others, 1964). The configuration of the water table has been mapped and described by Lindholm and others (1986). Typically, the water table is less than 150 m below the surface, although the depth to water locally exceeds 300 m. Saturated basalt extends to depths of 1 km or more in the central part of the eastern Snake River Plain (Whitehead, 1986). The water table slopes southwestward at gradients between 1 and 2 m/km over large areas, but the gradient is 4 m/km just east of the Great Rift and between 5 and 6 m/km in the westernmost part of the eastern plain. The comparatively steep gradient at the Great Rift probably reflects the influence of

basaltic dikes in fractures perpendicular to the direction of ground-water flow, which would reduce the transmissivity of the basalt (Lindholm and others, 1986). On the other hand, the steepness of the water table at the western end coincides with a thinning of the basalt and, hence, with a thinning of the aquifer (Whitehead and Lindholm, 1985).

The Snake River Plain aquifer is intercepted by the Snake River canyon between Milner Dam and King Hill, where extraordinary amounts of water are discharged as spectacular springs, especially near Hagerman. This reach of the canyon has 11 of the 65 springs in the United States whose average discharge exceeds $2.8 \text{ m}^3/\text{s}$ (Meinzer, 1927). Groups of springs have average discharges of 7 to $34 \text{ m}^3/\text{s}$ (summarized by Whitehead and Covington, 1987). During the peak of the irrigation season, July and August, virtually the entire flow of the Snake River from Milner Dam to the Bruneau River is supplied by these springs. Before 1910, aggregate discharge from the springs was slightly less than $120 \text{ m}^3/\text{s}$. Thereafter, large-scale flood-and-furrow irrigation with water diverted from the Snake River resulted in augmented recharge, so that discharge from the springs increased to $164 \text{ m}^3/\text{s}$ in the early 1930s and $192 \text{ m}^3/\text{s}$ in the early 1950s (Kjellstrom, 1984). By 1980, discharge had decreased to $170 \text{ m}^3/\text{s}$, mainly because more water was pumped from the aquifer, less surface water was diverted for irrigation, and farmland was being irrigated more efficiently (Whitehead and Covington, 1987). According to Garabedian (1986), total ground-water discharge to the Snake River from the eastern plain in 1980 was about $280 \text{ m}^3/\text{s}$, and another $64 \text{ m}^3/\text{s}$ of ground water was pumped for irrigation. Recharge was by direct precipitation on the plain ($30 \text{ m}^3/\text{s}$), underflow from tributary basins ($48 \text{ m}^3/\text{s}$), seepage from irrigation canals and the Snake River ($54 \text{ m}^3/\text{s}$), and infiltration of surface water used for irrigation ($200 \text{ m}^3/\text{s}$). The difference between recharge and discharge represents ground-water storage and estimation errors.

The ground water mainly flows through highly permeable zones in the basaltic lava flows and is a calcium-sodium-bicarbonate type, having generally less than 325 mg/L dissolved solids (Wood and Low, 1986). Modeling of the aquifer indicates transmissivities mostly between 28,000 and 121,000 m^2/day , with a maximum of about 450,000 m^2/day (Garabedian, 1986). The wide range in values reflects differences in the hydraulic conductivities and thicknesses of dense and open-textured basalts. Near the Idaho National Engineering Laboratory, the velocity of ground-water flow ranges from 1.5 to 6 m/day (Jones, 1963).

Further hydrologic studies of the Snake River Plain aquifer are underway to refine knowledge of the ground-water flow system, to determine the effects of conjunctive use of ground water and surface water, and to describe the water quality and geochemistry (Lindholm, 1986). From these studies, the response of the aquifer to various ground-water management alternatives is expected to be evaluated.

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