Geologic and Tectonic History of the Western Snake River Plain, Idaho and Oregon

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ABSTRACT

The western Snake River Plain is a Neogene-aged intracontinental rift basin, about 70 km wide and 300 km long, trending northwest across the southern Idaho batholith. Its southeastern end merges with the northeast-trending eastern plain, a structural downwarp associated with extension along the track of the Yellowstone hot spot. Orientation of the western plain rift is parallel to several regional northwest-trending crustal discontinuities, such as the Olympic-Wallowa lineament and the Brothers fault zone, suggesting that the rift failed along zones of lithospheric weakness, as the lithosphere was softened by the passing hot spot. Crustal refraction data and gravity show that the rift is not simply underlain by granitic rock, despite its appearance of having broken and extended the southern end of the Idaho batholith. Instead, the crust beneath 1 to 2 km of basin fill is mostly of mafic composition down to the top of the mantle, about 42 km deep beneath the plain. North and south of the plain, the upper crust has velocities more typical of granitic rock. South of the plain, beneath the 9-11 Ma Bruneau-Jarbidge eruptive center of silicic volcanics, is a zone of slightly high seismic velocity at a depth of 23 km that could be restite or an underplate of basalt related to formation of the silicic magma.

In this paper we show that some (12-10 Ma) rhyolite flows and domes erupted near the margins of the plain, but that thick rhyolite does not occur in deep wells in the subsurface of the plain northwest of Boise. For this reason, we suspect that much of the area of the plain was an upland and not a large depositional basin during the period of silicic volcanism.

Geochronology of volcanic rocks on both sides indicate major faulting began about 11 Ma and was largely finished by 9 Ma. Since about 9 Ma, slip rates have been low (less than 0.01 mm/year) with the exception of a short (about 10-km) segment of late Quaternary faulting in the Halfway Gulch-Little Jacks Creek area on the south side.

Earliest sediment of the plain is associated with basalt volcanism and high rates of faulting. Interbedded arkose, mudstone, and volcanic ash constitute this earliest sediment mapped as the Chalk Hills Formation. Local basalt lava fields (dated 10-7 Ma) occur at several levels in the Chalk Hills Formation. An active rift environment is envisioned with lakes interconnected at times by a river system.

The faulted and tilted Chalk Hills Formation is dissected by an erosion surface at the basin margins, indicating a regression of lakes to the deeper basins. Depositional records of the regression are generally absent from the margins, but we suggest that the east Boise fan aquifer sediments and deep basin fill might be such a record. Nothing is known of the cause of the regression of the Chalk Hills lake.

A transgressive lacustrine sequence encroached over slightly deformed and eroded Chalk Hills Formation on the plain margins, locally leaving basal coarse sand, or a thin beach pebble layer now iron-oxide cemented. The upper part of this transgression deposited shoreline oolitic sand deposits, indicating increased alkalinity of a closed lake. In the Boise foothills, much of the exposed sediment appears to be this transgressive lacustrine sequence

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where it is mapped as the Terteling Springs Formation, with shoreline sands and small deltas interfingering basinward with lake muds. The lake rose to its highest elevation of about 3,600 feet (1,100 m) in a period of less than a few million years. At that highest level, it overtopped the spill point into ancestral Hell's Canyon and the Columbia-Salmon river drainage. Reliable geochronology constrains the time of overflow between 6.4 and 1.7 Ma and is in need of better resolution. The rise in lake level may have been indirectly caused by regional tectonic movement of the migrating uplift of the Yellowstone hot spot, as an associated Continental Divide migrated about 200 km eastward from the Arco area to its position in Yellowstone National Park over the period 6 Ma to present. In doing so, the catchment area of the Snake River may have increased as much as 50,000 square km. Captured runoff associated with the shifting topographic divide is hypothesized to have caused the level of Lake Idaho to rise to its spill point about 4 million years ago.

Downcutting of the outlet was apparently slow (about 120 m/Ma) during which time sandy sediment eroded from the basin margins and filled the remaining lake basin with interbedded mud and sand of lacustrine delta systems. This sedimentary sequence of a slowly lowering base level constitutes most of the Glenns Ferry Formation and the main sand-bearing aquifer section of the western plain. It is represented in the Boise foothills by a 60-m-thick unit of coarse sand with Gilbert-type foreset bedding called the Pierce Park sand. Subsequently, fluvial systems with gradients necessary to produce braid-plain sandy gravel deposits flowed to the outlet region near Weiser. These gravel deposits should decrease in age and altitude to the northwest, and at Weiser these oldest gravels occur at elevation 2,500 feet.

During the late stages of the draining of Lake Idaho, basalt volcanism resumed in the western plain, focusing along a line of vents that trends obliquely across the plain at about N. 70º W., named here the Kuna-Mountain Home volcanic rift. Both sublacustrine and subaerial volcanoes erupted and built a basalt upland with elevations of highest shields to 3,600 feet over the last 2.2 million years. Aligned vents and fissures of these volcanoes indicate the present orientation of the principal tectonic stress is N. 70º W., contrasting with the N. 45º W. boundary of the plain and the N. 30º W. alignment of vents in the eastern plain. This N. 70º W. alignment is similar to the same vent features of Quaternary basalt fields in eastern Oregon, suggesting that a province of similar tectonic stress orientation includes the western plain and much of eastern Oregon.

Key words: rift, Cenozoic faulting, lacustrine sediments, Quaternary volcanoes

INTRODUCTION

The western part of the Snake River Plain is an intracontinental rift basin about 70 km wide and 300 km long. It is a normal-fault bounded basin with relief due to both tilting toward the center of the basin and evolving normal fault systems. Maximum depth of Neogene sedimentary fill in the basin is 2-3 km. The offset of older volcanic rocks exceeds 4 km in places. We show here that the rift-basin structure evolved mostly within the last 9.5 million years contrary to estimates by others (Mabey, 1982; Malde, 1991) who have loosely stated the basin began forming 16 to 17 million years ago.

The contiguous lowland of the eastern and western Snake River Plain confused geologists for many years who tried to ascribe a common structural origin to the entire arcuate lowland of southern Idaho (called the "smile face" of Idaho as it appears on physiographic maps). Lindgren (1898) and Kirkham (1939) described the plain as an arcuate structural downwarp. They did not recognize the fault boundaries of the western plain. Malde (1959) was first to report the normal fault boundaries of the western plain.

The eastern part of the plain is not a tectonic rift because it is not fault bounded. Instead, it is a downwarp forming a spectacular low topographic corridor across the actively extending northern Basin and Range Province (Parsons and others, 1998). By all measures, the eastern plain is an unusual lowland formed perhaps by a curious interplay of magmatism and extension (Parsons and others, 1998; McQuarrie and Rodgers, 1998).

The western plain can be more simply explained as a basin and range structure whose formation was triggered by the magmatism of the migrating Yellowstone hot spot (Clemens, 1993). Its orientation is the same as the many northwest-trending half grabens that flank the eastern plain and developed in the "wake" of the northeast-migrating hot spot. In contrast to those half-graben systems, however, the western plain is a much larger feature, 70 km wide compared to less than 30 km wide for the Grand Valley or Lemhi Valley grabens. The western plain lies north of the track of the hot spot. In the hot-spot tectonic model proposed by Anders and others (1989) and Pierce and Morgan (1992), one might expect symmetrically disposed half-graben systems formed beyond the parabolic-shaped "wake" as the hot spot passed by. No corresponding major graben system of similar proportions and age occurs south of the track in the northwestern Utah area.

As shown in Figure 1, the western plain cuts obliquely across an older north-trending Oregon-Idaho graben dated 15.5 to 10.5 Ma (Cummings and others, 2000). It also truncates the south end of the west-central Idaho fault belt identified by Hamilton (1963), a relatively young

system of north-trending normal faults. Early in its history the western plain underwent rapid subsidence and became the locus for the major lacustrine system of Lake Idaho, which persisted from about 9.5 to 1.7 Ma. The lake system underwent a substantial lowering about 6 Ma and then refilled. The 7.8-Ma duration of the lake system is long but typical of other rift lakes, such as east African Lake Malawi discussed by Johnson and Ng'ang'a (1990).

The western plain rift basin is similar in dimensions and structure to intracontinental rifts elsewhere in the world, such as those in east Africa, the Baikal Rift in eastern Russia, and the Rio Grande Rift in southwestern United States. The structure is a complex of half grabens and full grabens similar to that reported by Bosworth (1985) in other continental rift settings. The volcanism of the western plain differs from those rifts by its association with a migrating continental hot spot indicated by a pattern of time-transgressive silicic volcanism.

Practical geological interest in the western plain is largely inspired by the great ground-water resources in the sedimentary fill, and to a lesser extent by the geothermal ground water in the deep volcanic rocks. These resources are essential to the economy of semiarid southwestern Idaho. Recent pumpage has been about 0.3 million acre-feet/year (0.37 cubic km/year; Newton, 1991). Ground-water development has been mostly within the upper 800 feet (250 m) of section and has been spectacularly successful with some wells in sand aquifers producing 3,000 gallons/minute (1,600 cubic m/day; Squires and others, 1992). Many wells, however, are drilled into thick mudstone sections with poor production. The distribution of sand aquifers in the fluvial-lacustrine section is complex, but in just the last few years we are gaining a clearer understanding of the depositional history and gross features of the sedimentary architecture (Squires and others, 1992; Wood, 1994). Toward this end, we have turned the unsuccessful oil and natural gas exploration in the basin to our advantage. We have examined the scattered data on deep holes drilled by petroleum companies mostly in the years between 1950 and 1985 (Wood, 1994). This information and recent studies have improved the ground-water models now being developed for managing the resource and averting conflicts over ground-water use.

In this study, we review geophysical data on the western plain and interpret seismic-reflection data and deep drill-hole data to understand both the tectonic framework of the basin and the sedimentary facies of the basin fill. We incorporate available K-Ar and $^{39}Ar/^{40}Ar$ ages of volcanic rocks and paleomagnetic data for a chronology of events that produced the plain. We elucidate features of the sedimentary fill pertinent to the location of sand aquifers in the predominantly mud sediments of the lake system. From these data and geomorphological considerations, we propose a model for the history of the great lake system that filled the basin and its eventual overflow into ancestral Hell's Canyon. We also discuss Qua-

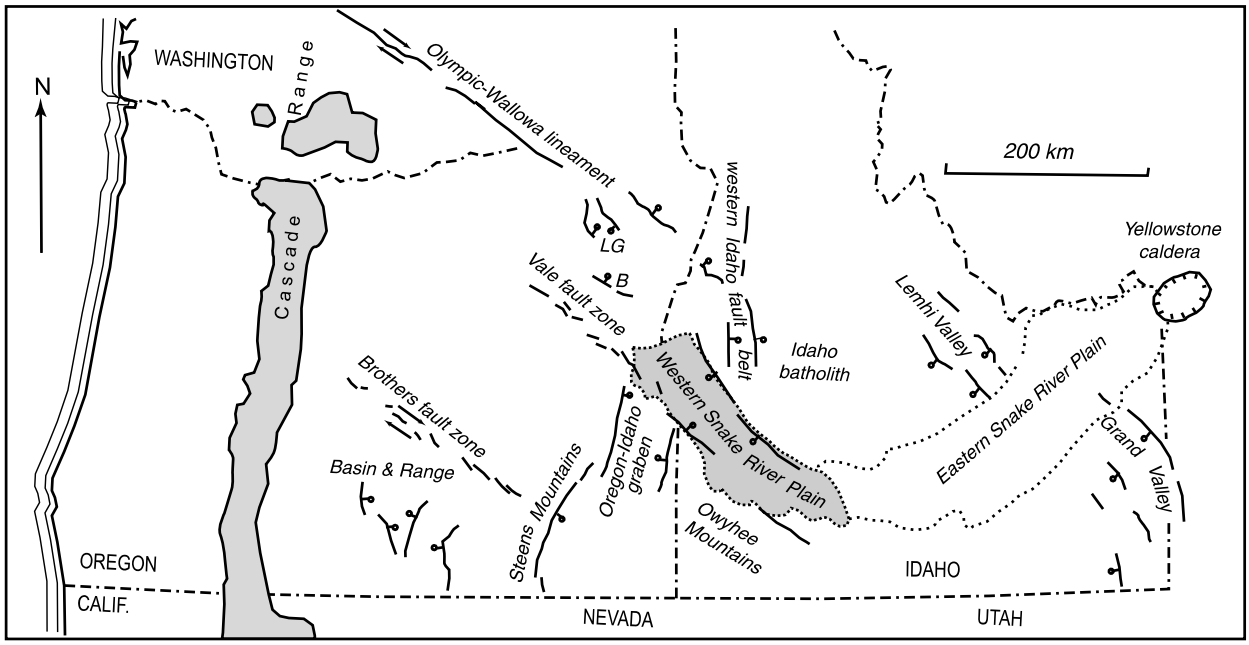

Figure 1. Regional setting of the western Snake River Plain showing related geologic features and emphasizing northwest-trending, late Cenozoic fault structures in Oregon and Idaho. LG—La Grande graben, B—Baker Valley.

ternary geomorphic features of the fluvial system that followed the draining of the lake and tectonic aspects of the Quaternary basalt fields that cover part of the western Snake River Plain.

CRUSTAL STRUCTURE AND TECTONICS

Basic to understanding the origin of the western plain is the following question. What is it about the earth's crust that causes a wide northwest-trending sag and graben just here in southern Idaho? The upper crust extended several kilometers to form this 50- to 70-km-wide rift. The northwest trend of the western plain cuts across north- to south-trending older Miocene extensional structures. The northwest trend, however, is the same as that of several smaller late Cenozoic graben basins northwest of the plain (La Grande Valley, Baker Valley). These basins show late Quaternary faulting on their margins (Pezzopane and Weldon, 1993). Several major lineaments of the northwest United States also have the same northwest-southeast trend (the Olympic-Wallowa lineament, Vale fault zone, and the Brothers fault zone). The lineaments are spaced 70 to 200 km apart through eastern Oregon (Figure 1). Parts of some lineaments are expressed by normal faults. Many of the lineaments show small amounts of late Cenozoic right-lateral movement expressed as *en echelon* normal faults or pull-apart basins. Displacements are but a few kilometers, thought to accommodate differences in extension across the lineament structures (Lawrence, 1976). The pervasive northwest structural trends suggest an orientation for zones of lithosphere weakness that have responded to late Cenozoic stress systems. While inherited zones of weakness might explain the orientation of the western plain, it does not explain the geographic position or width.

The western plain cuts across the southern part of the Mesozoic Idaho batholith, with the Owyhee Mountains segment (Taubeneck, 1971) split off to the south of the main outcrop (Figure 2). The elevation of most of the batholith mountains to the north, as well as the crest of the Owyhee Mountains, is about 8,000 feet (2,440 m). Southwest of the Owyhee Mountains is the Owyhee Plateau, a region of low relief (elevation about 5,500 feet, 1,680 m) extending into northern Nevada and southeastern Oregon. Flows of hot-spot rhyolite and basalt that form the plateau are relatively little deformed by faulting or tilting, and little is known of the underlying crust.

If the western plain is an ordinary graben, it should be underlain by downfaulted granitic rocks of the Idaho batholith. According to Hill and Pakiser's (1967) interpretation of deep crustal refraction data, a significant layer of rock having the velocity of granite (V_P of 5.5-6.4 km/s) does not underlie the plain. Prodehl (1979) reinterpreted the refraction data and concluded similarly that granitic-rock velocities are restricted beneath the western plain. Instead, the plain is underlain below 8-km depth by high-velocity material, V_p = 6.6-6.8 km/s (Figure 3). This contrasts with crust south of the plain (Mountain City to Elko), where 6.1 km/s material extends down to an 18-km depth. We have no refraction data north of the plain; however, the very existence of the extensive Idaho batholith north of the plain indicates granitic crust. The depth to the base of the batholith has not been gravity modeled to our knowledge, although Cowan and others (1986) illustrate it in their cross section to extend to 8 to 10 km. The thickness of most great granite batholiths is probably not more than 15 km, and the erosion of roof rock has probably reduced the depth to the base of the granite of exposed batholiths to less 10 km (Bott and Smithson, 1967; Leake, 1990). By analogy with the Sierra Nevada batholith, granitic rock there extends to a depth of 10 to 15 km and appears to be underlain by a root of low velocity (6.5 km/s) material to a depth of 40 km (Fliedner and others, 1996). In addition, a relatively low-velocity upper mantle material is detected to a 60-km depth. It is reasonable to assume that at one time granite beneath the plain was 8 to 10 km thick, but has been intruded by significant quantities of basaltic rock.

A large positive gravity anomaly is associated with the western plain (about +100 milligals) relative to the bordering batholith regions (Mabey, 1982). The anomaly has two peculiar features (Figure 4). It is composed of a broad positive anomaly paralleling the margins of the plain (N. 40º W.). Superposed on this is a narrow (30-km-wide) feature of about +25 milligals that trends obliquely (N. 70º W.) across the plain, south of Mountain Home, for a distance of 140 km (shown in Figure 4 by the southeastern region enclosed by the -105 milligal contour). The narrow anomaly can be accounted for by the Quaternary basalt field, but the broad anomaly must arise from a deep high-density body. Mabey (1982) modeled a single gravity profile perpendicular to the plain, near Mountain Home, and reproduced the gravity data with a deep body of density 2.9 g/cm^3 from 9 to 18 km and a shallow body of the same density from 3 to 6 km depth. A density of 2.9 g/cm^3 is appropriate for solid basalt or gabbro in contrast to a density of 2.65 g/cm^3 for granitic rocks, and less than 2.3 g/cm^3 for sedimentary rocks. Both the seismic refraction data and gravity data indicate that the sediment and volcanic fill are underlain by material in the intermediate crust having appropriate velocity and density of diabase or gabbro intrusive rocks, and not material typical of granite.

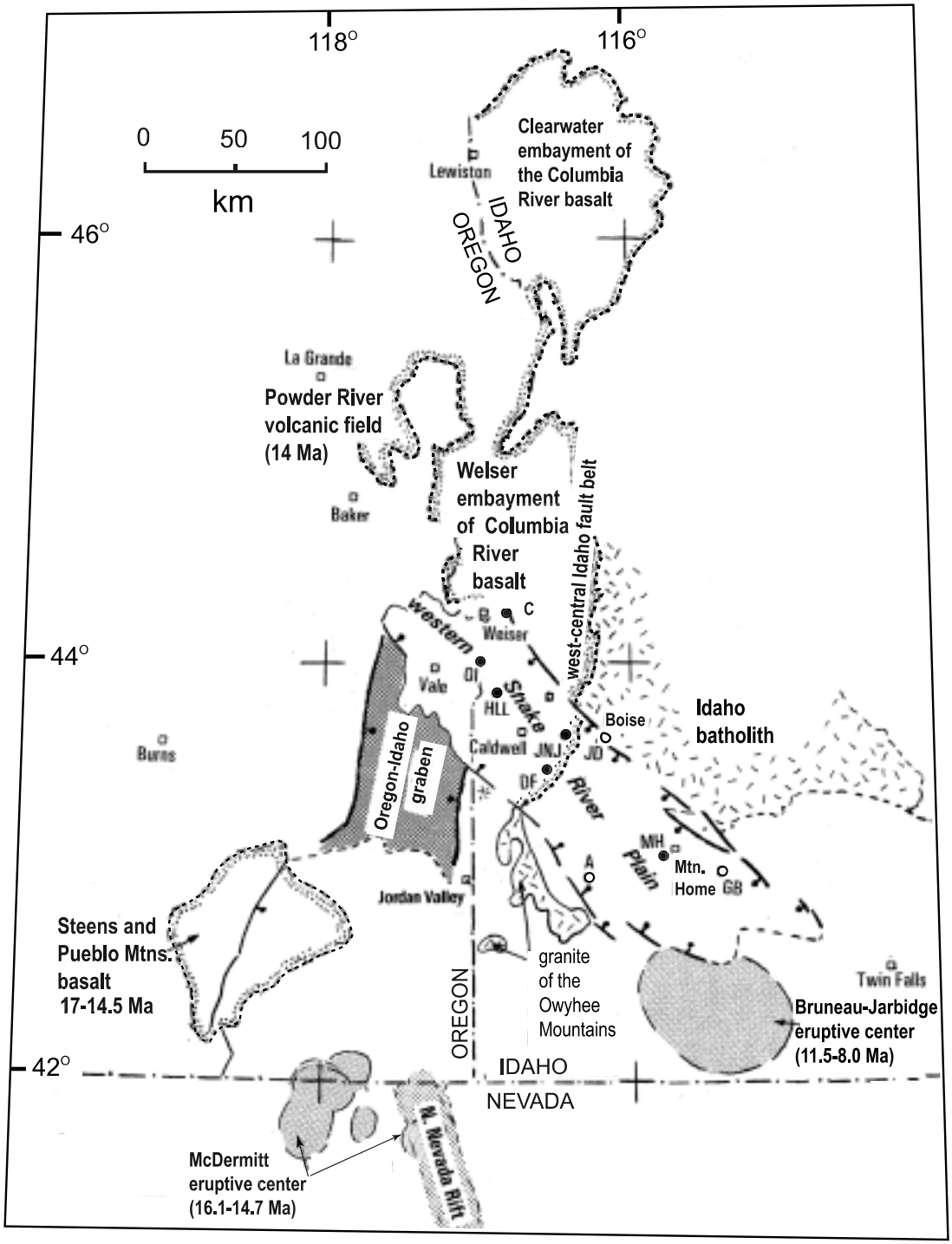

Figure 2. Map showing tectonic features surrounding the western Snake River Plain and locations of deep wells into volcanic basement. Wells only in deep basalt are shown with a solid circle: C—Chrestesen No. A-1; OI—Ore-Ida Foods No. 1; HLL—Highland Land and Livestock No. 1; JNJ—J.N. James No. 1; DF—Deer Flat No. 1; and MH—Mountain Home Air Force Base Geothermal Test. Wells in deep rhyolite are shown with an open circle: JD—Boise Julia Davis Park; A—Anschutz Federal No. 1; and GB—Griffith-Bostic No. 1. Tectonic feature locations and reference sources: northern Nevada rift (Zoback and others, 1994); Oregon-Idaho graben (Cummings and others, 2000); rhyolite-field eruptive centers (Bonnichsen, 1982; Pierce and Morgan, 1992; McCurry and others, 1997); and the Columbia River and Steens Mountain basalt areas (Hooper and Swanson, 1990; Lees, 1994; Hooper and others, 2002a, 2002b.).

A substantial accumulation of Miocene basalt lava lies beneath the sediments of the western plain. We show a hypothesized contact of lavas with underlying intrusive basalt (Figure 3). The deepest drill hole in the plain at Meridian (J.N. James well, 4.3 km deep) penetrates a thick basalt section and bottoms in basalt flows having a geochemical affinity with the Columbia River Basalt Group (Clemens, 1993; Figure 5). No inclusions of deep crustal rocks are known from western plain basalt, so we can only speculate on the geophysical results that either the underlying crust contains insignificant remnant granite greatly intruded by Cenozoic diabase and gabbro, or it is entirely filled with mafic intrusives.

Extension could have thinned the granite, but as we

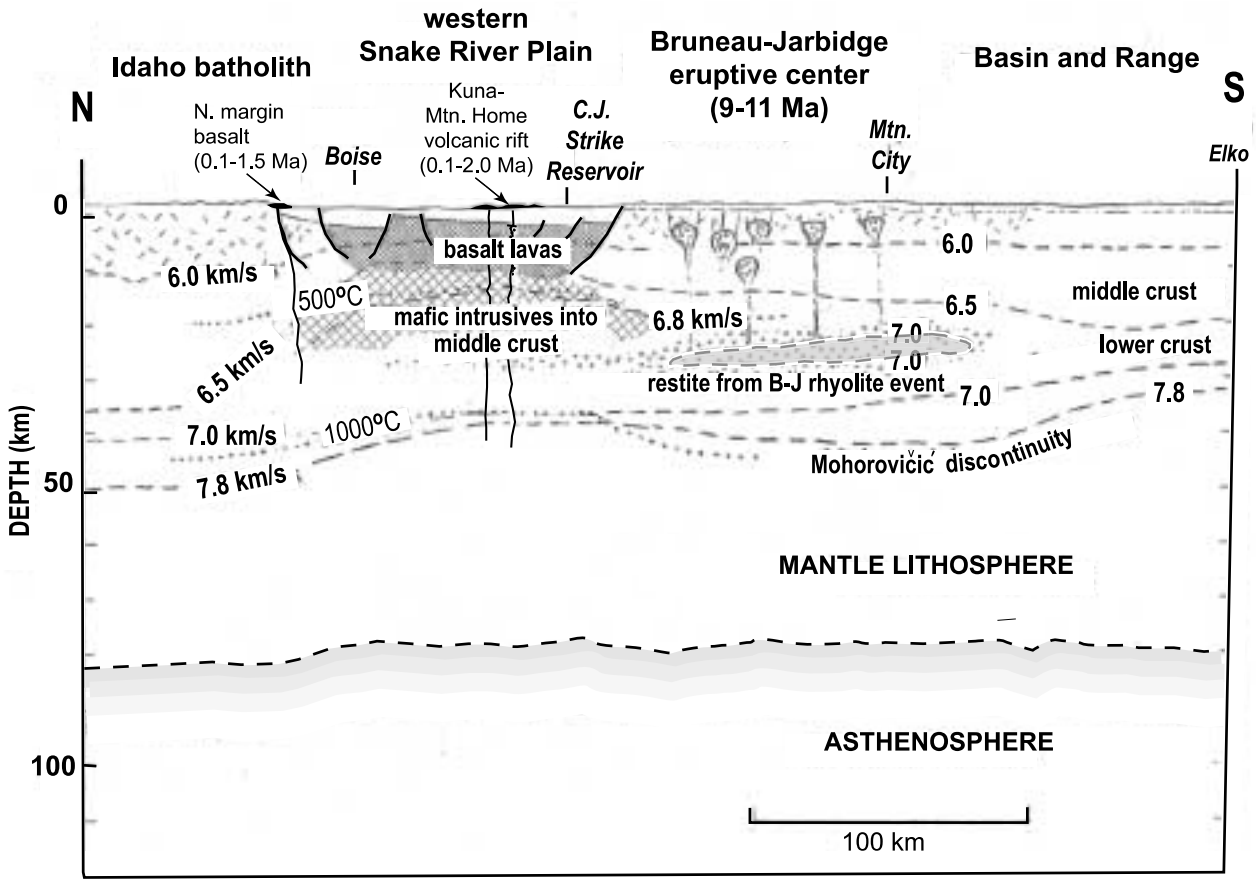

Figure 3 (A)

Figure 3. **(A)** Lithosphere structure interpreted principally from crustal structure beneath the western Snake River Plain and adjacent areas. Interpretation is based on the seismic refraction line of Hill and Pakiser (1967) and the reinterpretation by Prodehl (1979). Location of seismic line shown on map. Noteworthy is the upward bulge of material, with $V_P > 6.6$ km/s beneath the western plain believed to be mafic rock, and the overlying thin (~5 km) layer with velocity between 6.0 and 6.5, believed to be basalt flows or granite intruded by basalt. Prodehl's interpretation shows a high velocity layer ($V_P > 7.0$ km/s) that lies in the deep crust beneath what is now recognized as the Bruneau-Jarbidge rhyolite eruptive center along the track of the hot spot. We suggest the high-velocity layer might be restite remaining from the partial melt and extraction of rhyolite melt. It is important to realize that the crustal refraction velocities shown beneath the plain are obtained from arrivals into a string of detectors, south of Boise, and are an average of the crust between Boise and C.J. Strike. No experiments have explored structure beneath the Idaho batholith. The batholith structure shown is inferred from a section by Hyndman (1978) and Cowan and others (1986) and by analogy to the Sierra Nevada batholith shown by Fliedner and others (1996). Mafic intrusives in the intermediate crust were first suggested by Mabey (1976) from gravity data. Diagrammatic diapirs of silicic melts beneath the B-J area were suggested by Leeman (1989). Dotted-line isotherms are from heat-flow models of Brott and others (1978). One can only guess the depth of the asthenosphere at about 90 km (see Smith and Braile, 1993, Figure 35, for the eastern plain, and by analogy to the Rio Grande Rift, Baldridge and others, 1984, Figure 2).

(B) Prodehl's (1979) reinterpretation of refraction data and crustal structure from Boise to Elko.

(C) Map showing the location of the seismic refraction line. Labeled triangles are shot points, and circles are detector positions.

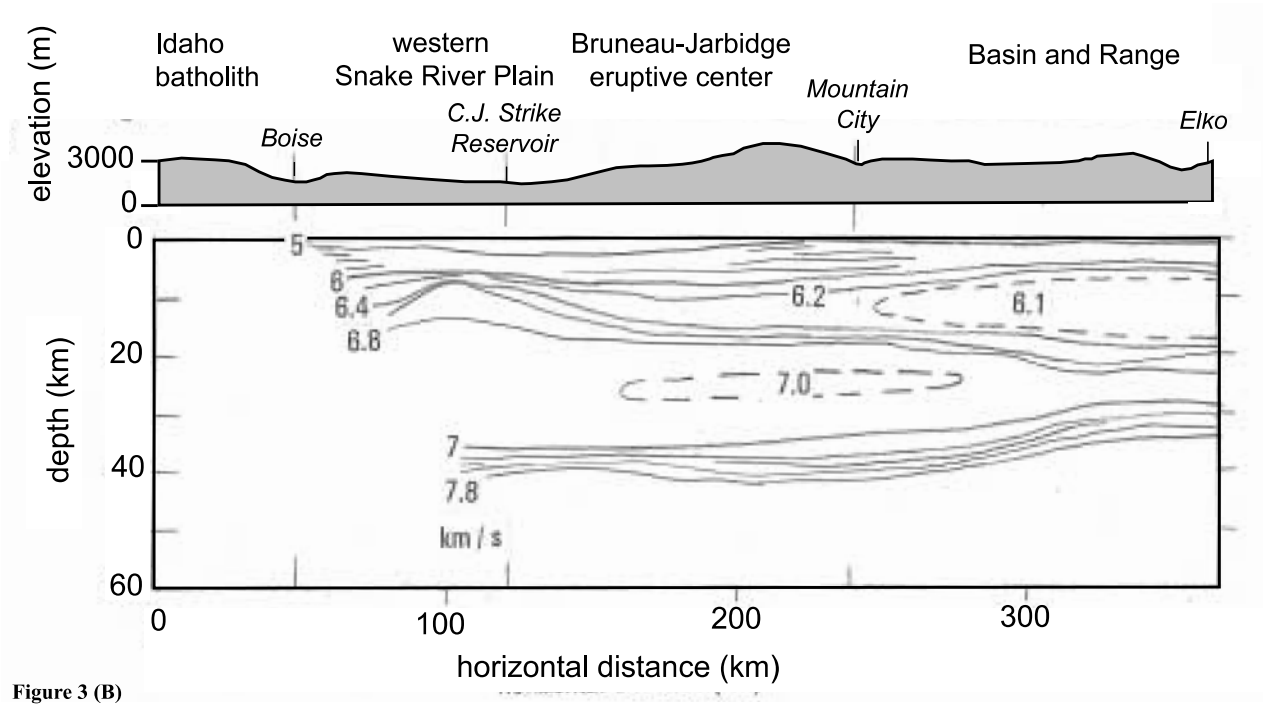

Figure 3 (B)

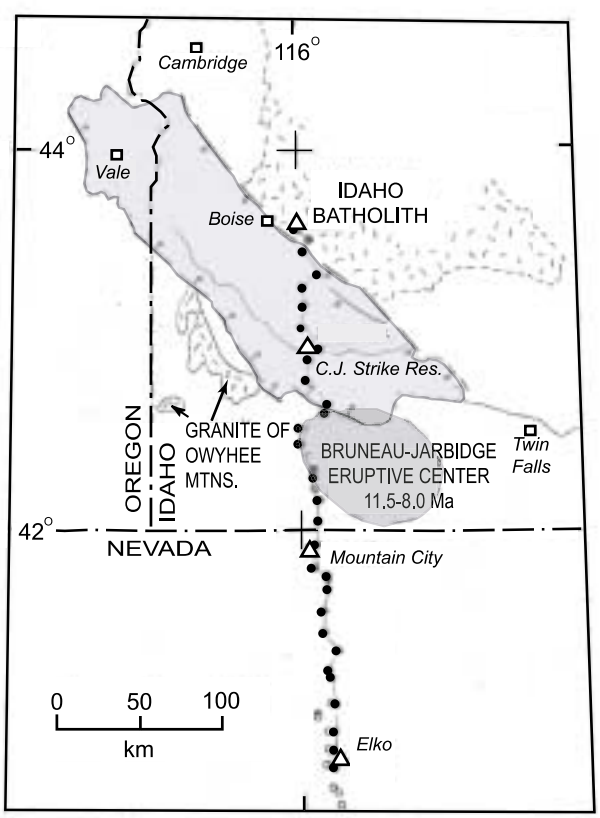

Figure 3 (C)

will show, it is unlikely that this process has greatly altered the thickness of the granite. To evaluate extension, we have added up the vertical dip-slip fault offset of the basement volcanic rock, shown on Figure 5, and assuming 45-degree dipping normal fault planes, we obtain only about 2 km of vertical offset and therefore a corresponding 2 km of horizontal extension of the 60 km width, or about 3 percent extension. Such an evaluation using only fault offset does not take into account the component of basin extension expressed by subsidence due to downwarping. If, instead of faulting, we consider the volume of the basin to have been created by extension of the upper 10 km of crust, we obtain an extension of about 10 percent. This value of extension is comparable to the 7 to 14 percent extension determined for the east African rift basins by Rosendahl and others (1992) and around 10 percent for the Rhine and Baikal rifts (Park, 1988, p. 84). It seems unlikely that extension of 10 percent could thin the granite so that it is undetectable from seismic refraction velocity measurement.

Although the western plain would not be regarded as strongly extended, with about 10 percent extension, the modification of the composition of the middle and lower crust below 8-km depth is considerable. Our preferred explanation is that the middle and lower crust under the western plain has been so invaded by mafic intrusives that it has a seismic velocity (6.6 km/s) similar to that indicated for diabase by Christiansen and Mooney (1995). Dense, high-velocity rock could also form in the middle

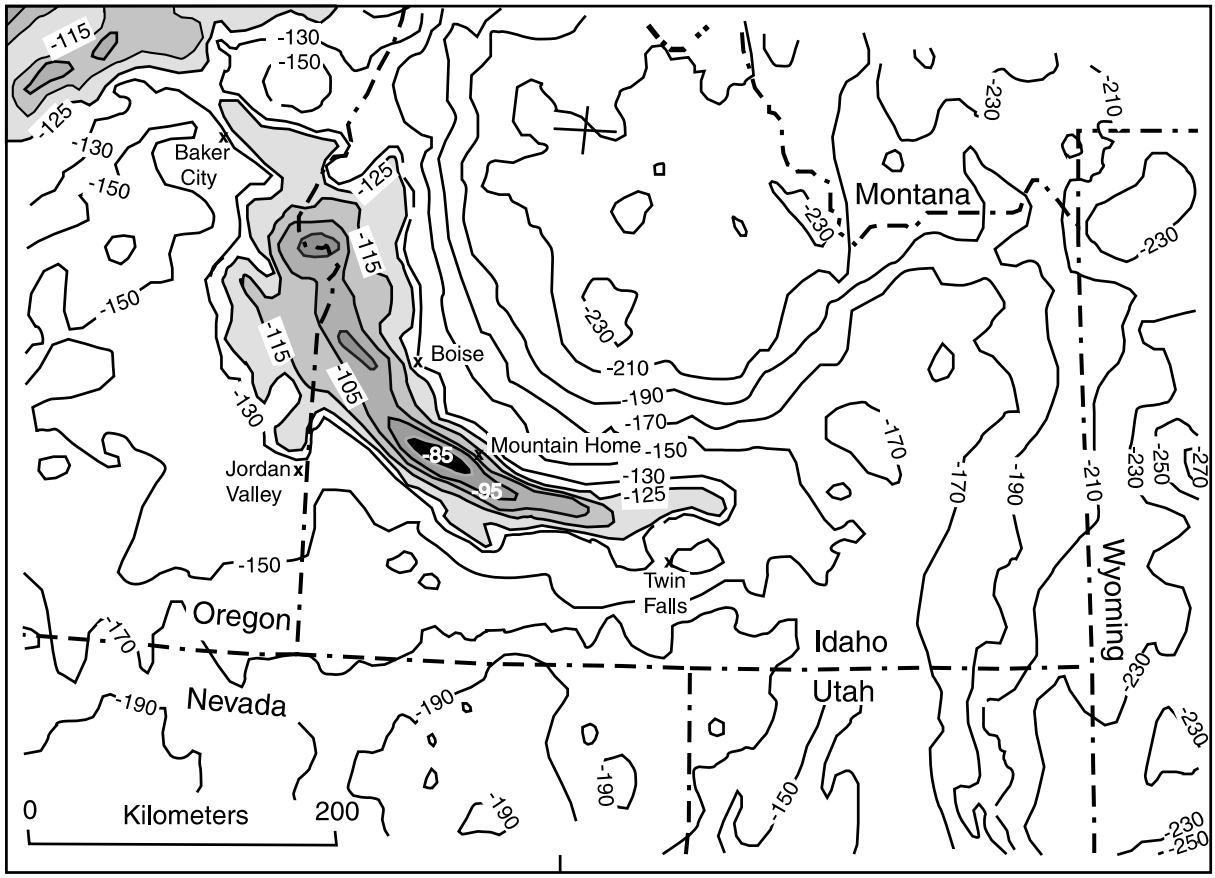

Figure 4. Map of Bouguer gravity anomalies of the western Snake River Plain region. Noteworthy is the high gravity anomaly of the western plain indicating rock of high density beneath the plain. Contour interval is 20 milligals, except for values more positive than -130 milligals. The areas of more than -125 milligals are shaded and contoured every 10 milligals. Map is from the Gravity Anomaly Map Committee (1987). Values are terrain corrected in areas of high relief. Bouguer anomalies are calculated by subtracting the theoretical attraction of rock mass above sea level using a standard crustal density of 2.670 g/cm^3. This attempts to remove the effects of varying topographic relief. Thus Bouguer anomalies result from masses above sea level with densities different from 2.67 g/cm^3, or from any lateral variation of density below sea level. In continental regions, the regional values are negative because topography is usually isostatically compensated by low density crust extending below sea level. Theoretically, corrected gravity will be zero only at sea-level measuring points.

crust by magmatic differentiation or by partial melting and eruption of the silicic component, leaving a residual mafic material; however, the volume of silicic volcanics erupted from the plain area appears to be too small (less than 500 cubic km) to account for a residual volume of high velocity rock in the intermediate crust. Most of the silicic magma of the region erupted from vents south of the western plain.

Crustal velocities in the western plain are different from those in the eastern plain, where such large lateral variations in velocity of the upper and middle crust are not detected. This led Braile and others (1982) and Wendlandt and others (1991) to infer that rising basalt may not have yet significantly modified the middle crustal composition of the eastern plain.

From the reinterpretation of the western-plain seismic refraction profile by Prodehl (1979), one might infer from his velocity model that granitic material lies south of the plain, between depths of 6 to 15 km, and that an anomalous 10-km-thick high-velocity layer of 7.0 km/s lies deep in the crust from 20 to 30 km (Figure 3). The location of that layer coincides with the roots of the large Bruneau-Jarbidge rhyolite eruptive center identified by Bonnichsen (1982) and located along the track of the Yellowstone hot spot (Figure 2). A deep-crustal high-velocity layer in this position supports the idea of a remnant of deep crustal underplate of plume basalt suggested by Leeman (1989). Alternatively, the layer could be the mafic residual of deep crust from which 2,000 cubic km of rhyolite is estimated by Pierce and Morgan (1992) to have erupted from the Bruneau-Jarbidge area.

Thus, rifting by downwarping and normal faulting

along northwest-trending boundary faults initiated extension of the western-plain crust, cutting through the south end of the batholith. It is possible that a pre-existing lithosphere structure along this trend was particularly vulnerable to weakening by heating and extension caused by the passing hot spot 10 to 11 million years ago. The rifting process involved the injection of large amounts of basaltic magma into the middle crust beneath the western plain, and that process may have greatly extended the early rift, more so than is estimated by our examination of the basin volume or faulting of the Miocene basalt basement. Basaltic magma may have been injected into the downfaulted block of granite. Humphreys and others (1999) suggest that much of the subsidence of the eastern plain is caused by the weight of added basalt to the crust, and that explanation invoked by Baldridge and others (1995, p. 455) for other continental rifts, may also apply to the western plain. Again, we point out that the actual track of the hot spot adjacent to the south side of the western plain appears to correlate with a high-velocity (7.0 km/s) body in the deep crust that may contain injected basalt or mafic residue from hot-spot volcanism. In contrast, beneath the western plain the high velocities and inferred injected basalt extend upward to a much shallower depth (8 km). These interpretations of older seismic refraction data indicate that deep-crustal seismic experiments and a reevaluation of gravity data using modern methods would contribute to our knowledge of the Bruneau-Jarbidge eruptive center, the western plain, the batholith, and the effect of the migrating hot spot on the crust.

Because the plain is the product of extension and has clear evidence of high-angle, normal faulting at the surface, it is appropriate to question whether these high-angle faults shoal in dip at depth and merge with low-angle detachment faults. Recently proposed models of extended terrane show such detachments at the brittle ductile transition at depths of 15 to 30 km beneath the Basin and

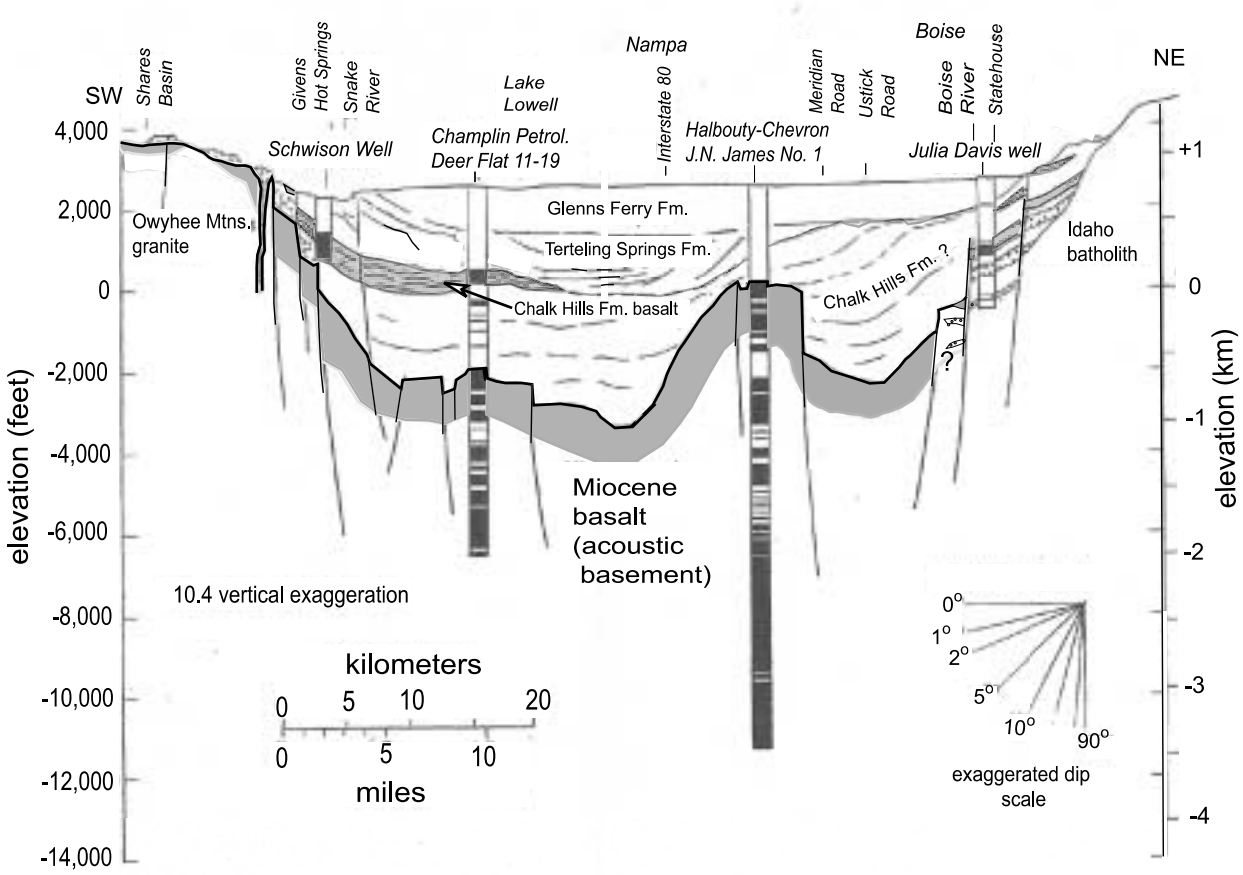

Figure 5. Section across the western plain showing the basalt occurrence (dark pattern) in the deep wells and the configuration of sedimentary fill and associated basalt from seismic reflection data. Shaded outline is the top of Miocene basalt. Stippled pattern is rhyolite that occurs in wells only at the margins of the plain.

Range Province (Wernicke, 1992). Fault block rotation, indicating the curvature of some faults at depth, occurs only along the northeast margin faults at the west end of the Mount Bennett Hills. Wood (1989) shows that the blocks have rotated to northeast dips of 12 to 25 degrees. We have no geophysical observations on the nature of faults deeper than 6 km, but certainly we allow that they could greatly decrease in dip and merge with detachments at depth as they approach the ductile middle crust. The crust is considered ductile below about 13 km because that is the limiting depth of earthquakes in this region.

THE BEGINNING OF THE WESTERN SNAKE RIVER PLAIN BASIN

We address here the time at which the western Snake River Plain first became a depositional basin. Previous workers have loosely assumed it began about 17 to 16 Ma (Mabey, 1982; Malde, 1991), associating the western plain graben with the beginning of large-scale extension in the Basin and Range Province to the south. We believe that the graben basin did not begin forming until 11 Ma. Our presumption is that evidence for the formation of the incipient plain must be (1) northwest-oriented normal faulting and downwarping of the western plain area and (2) accumulation of sediments in a basin.

STRUCTURES BEFORE THE WESTERN PLAIN GRABEN

Before formation of the western plain, extensional-fault and dike structure had a more north-south trend (Figure 1). Extensional faults and dikes were a north-south orientation during the eruption of Columbia River basalt north of the plain in the interval 17 to 14 Ma (Hooper and Swanson, 1990), in the Steens Mountains (Minor and others, 1987), and in the northern Nevada rift from 18 to 14 M (Zoback and others, 1994). The north- to south-trending Weiser embayment of the Columbia River Basalt Group was a basin that accumulated at least 2.1 km of basalt prior to formation of the western plain (well logs from 1978 Phillips Petroleum Chrestesen No. A-1). Columbia River basalts onlap the batholith at the eastern margin of the embayment. It is not known whether this margin is a result of basin downwarping or if it is partly a fault boundary. The Weiser embayment is oriented north to south similar also to the Clearwater embayment (Figure 2). From surface mapping, other workers have reported that the Weiser embayment contains up to 0.7 km of Imnaha Basalt (17.2-16.5 Ma), as much as 0.2 km of Grande Ronde Basalt (16.5-15.6 Ma), and 0.3 km of the younger (< 15 Ma) but undated Weiser Basalt (Hooper and Swanson, 1990; Fitzgerald, 1982; Hooper and Hawkesworth, 1993).

South of the western plain, Cummings and others (2000) have discovered a north- to south-trending basin they have named the Oregon-Idaho graben (Figures 1 and 2). The bounding faults on this graben formed 15.5 to 10.5 Ma. The basin filled with bimodal (rhyolite and basalt) volcanic assemblages and sediments over 2 km thick. Dike orientations in the basin generally trend north to northeast (Ferns and others, 1993). The basin was later truncated by the northwest faulting that formed the western plain. The east side of the Oregon-Idaho graben and the east side of the Weiser embayment coincide with a north-trending zone of steep gravity gradient, where the Bouguer value increases about 25 milligals from the area on the east underlain by granitic batholith to the basalt-filled basin on the west (illustrated in Figure 4 and in more detail by Figure 2-1 of Wood and Anderson, 1981). Though rhyolite eruptives are more conspicuous at the surface of the Oregon-Idaho graben, the gravity data suggest a thick graben fill of dense basalt that contrasts with lower density granitic rock to the east (illustrated just north of Jordan Valley, in Figure 4).

The spatial arrangement of the Weiser embayment basin across the western plain from the Oregon-Idaho graben tempts one to correlate these two north-northeast trending structures (Figure 2). The ages of basalt fill of the Weiser embayment span the time of the pre-Ore-Ida-graben accumulation of at least 0.6 km of tholeiitic basalt and latite and also the accumulation of 2 km of lava flows and volcaniclastic sedimentary fill in the graben. Pre-graben tholeiitic basalt also occurs in a 1-km-thick section to the adjacent east in the Owyhee Mountain upon which Pansze (1974) obtained a whole-rock K-Ar age of 17 Ma (Ekren and others, 1982; Cummings and others, 2000). Basalt dikes within this Owyhee Mountains section trend north-south (Ekren and others, 1981). Therefore, it seems likely that north- and north-northeast-trending basins were the dominant structural-basin pattern before the western Snake River Plain graben formed.

Several deep wells (greater than 2.7 km deep) in the northwestern part of the western plain penetrate thick sections of basalt and tuffaceous sediment beneath the lacustrine sediment (Clemens, 1993). Figure 2 shows a western group of deep wells, drilled only in basalt at depth, that did not intersect rhyolite. The J.N. James No. 1 well was drilled by the Halbouty-Chevron group to a depth of 4.3 km in the center of the western plain at a site 25 km west of Boise. Beneath 0.7 km of lacustrine sediment, the well contains a 1.59-km-thick section of basalt lying upon a 1.37-km-thick sequence of mostly basaltic tuff,

and a bottom section of 0.2-km-thick basalt (Figure 5). The chemistry of five samples of bottom basalt cuttings from 4.08 to 4.11 km in depth were analyzed for major, trace, and REE elements (Clemens, 1993) and fall within the range of Columbia River basalts, being most like the olivine tholeiites of the Powder River volcanic field (Figure. 2) that erupted 14 Ma in the La Grande and Baker grabens (Hooper and Conrey, 1989), but also having some characteristics of the Imnaha Basalt (Peter Hooper, written commun., 1993; Clemens, 1993). We have not yet analyzed the other basalts in the well. The most one can conclude is that no significant rhyolite occurs in this well and that the lower basalt is similar to the basalt in the Weiser embayment.

The Highland Land and Livestock No. 1 well (3.64 km deep) and Ore-Ida No. 1 well (3.06 km deep) drilled through similar sections of interbedded basalt and sediment to a total depth beneath the upper 1.1 km and 1.4 km of lacustrine sediment, respectively. Samples from the Ore-Ida well were submitted for K-Ar dating by participating companies. The resulting dates are known, but laboratory details are not available. Basalt cuttings from a depth of 2.18 km in the Ore-Ida well yielded a whole-rock age of 16.2 ± 1.8 Ma, and core from a depth of 2.50 km a whole-rock age of 9.0 ± 1.8 Ma. The stratigraphic contradiction in the ages suggests that either the apparently older sample has gained radiogenic argon or the older age is correct. The apparently younger and deeper sample could have lost argon by the hydrothermal alteration of the basalt, and therefore the sample could be older. Nevertheless, the 9.0 Ma age can be regarded as a minimum age for this basalt.

THE INCEPTION OF SILICIC VOLCANISM AND UPLIFT

It is widely advocated that the Yellowstone hot spot was first manifested in eruptions of the Imnaha Basalt about 17.5 Ma along north and north-northwest-trending fissures in the Hell's Canyon area (Hooper and Swanson, 1990), as eruptions of the Steens Basalt (16.1 Ma), as eruptions and caldera collapse of several rhyolite centers near McDermott, Nevada, and as basalt dike emplacement along the N-NW-trending northern Nevada rift about 17 Ma (Pierce and Morgan, 1992; Zoback and others, 1994). During the inception of silicic volcanism (17-14 Ma), the rhyolite erupted from vents scattered over a broad region encompassing most of southwest Idaho south of the western plain and adjacent parts of Nevada and Oregon. By 11 Ma, silicic volcanic vents were centered mostly in the Bruneau-Jarbidge region (Bonnichsen and others, 1989) but were also distributed to the northwest along the future site of the western plain graben (Figure 6). The actual vent areas for many of the larger volume tuffs and flows have not been located, but many emanated from the Bruneau-Jarbidge region. Rhyolite vents are evident along the margins of the western plain. Jenks and others (1993) mapped a thick rhyolite breccia with unbrecciated dikes and sills, near Little and Big Jacks Creek, which they called the rhyolite of Horse Basin. They suggest it erupted from a buried NW-trending fissure near the edge of the plain. A faulted 1-km-diameter dome occurs south of Givens Hot Springs at the southwest edge of the plain and edge of the Owyhee Mountains (S.H. Wood, unpub. mapping). Clemens and Wood (1993) and Clemens (1993) have obtained K-Ar ages on near-source rhyolite lava flows of 11.8 to 11.0 Ma near and west of Boise on the north margin. It appears that eruptions of domes and small flows may have accompanied the beginning of the active phase of northwest-trending faulting, but that rhyolite volcanism had largely ceased by the time the basin began to form.

Most of the 12-10 Ma rhyolite accumulations on the margins of the western plain are without significant sedimentary interbeds (Wood and Gardner, 1984). The only known exception is from recent drilling in the subsurface beneath north Boise (City of Boise, Julia Davis Park well; Figure 5) where two rhyolite flows are separated by 130 m of coarse arkosic sand and batholith-derived gravel. The lowest flow is underlain by 60 m of similar coarse sediment at the bottom of the 0.98-km-deep well (P.N. Naylor, written commun. 1998). This sediment occurrence might be explained by local downfaulting associated with the eruptions adjacent to the batholith mountains. Perhaps this was the site of early basin initiation accompanying the rhyolite eruptions, but lack of a fine-grained lacustrine facies in the deep section precludes association with a deep basin. Likewise, the 2.3-km section of rhyolite in the Anschutz-Federal well on the south margin of the plain is without significant sedimentary interbeds (McIntyre, 1979). Elsewhere around the western plain, the lack of sediment in the nearly continuous pile of rhyolite eruptives suggests that during time of silicic volcanism the area was an upland. One would expect to find caldera basins; however, Ekren and others (1984) were impressed by the lack of conspicuous caldera features associated with silicic volcanism south of the plain in the region east of the Oregon border.

We believe it is significant that the deep wells in the center of the western plain, discussed above, have not drilled sections of rhyolite, despite abundant and thick rhyolite on the northeast and southwest margins. The lack of rhyolite beneath the plain is indeed surprising, and we can only conclude that the center of the present plain was formerly an upland of older basalt.

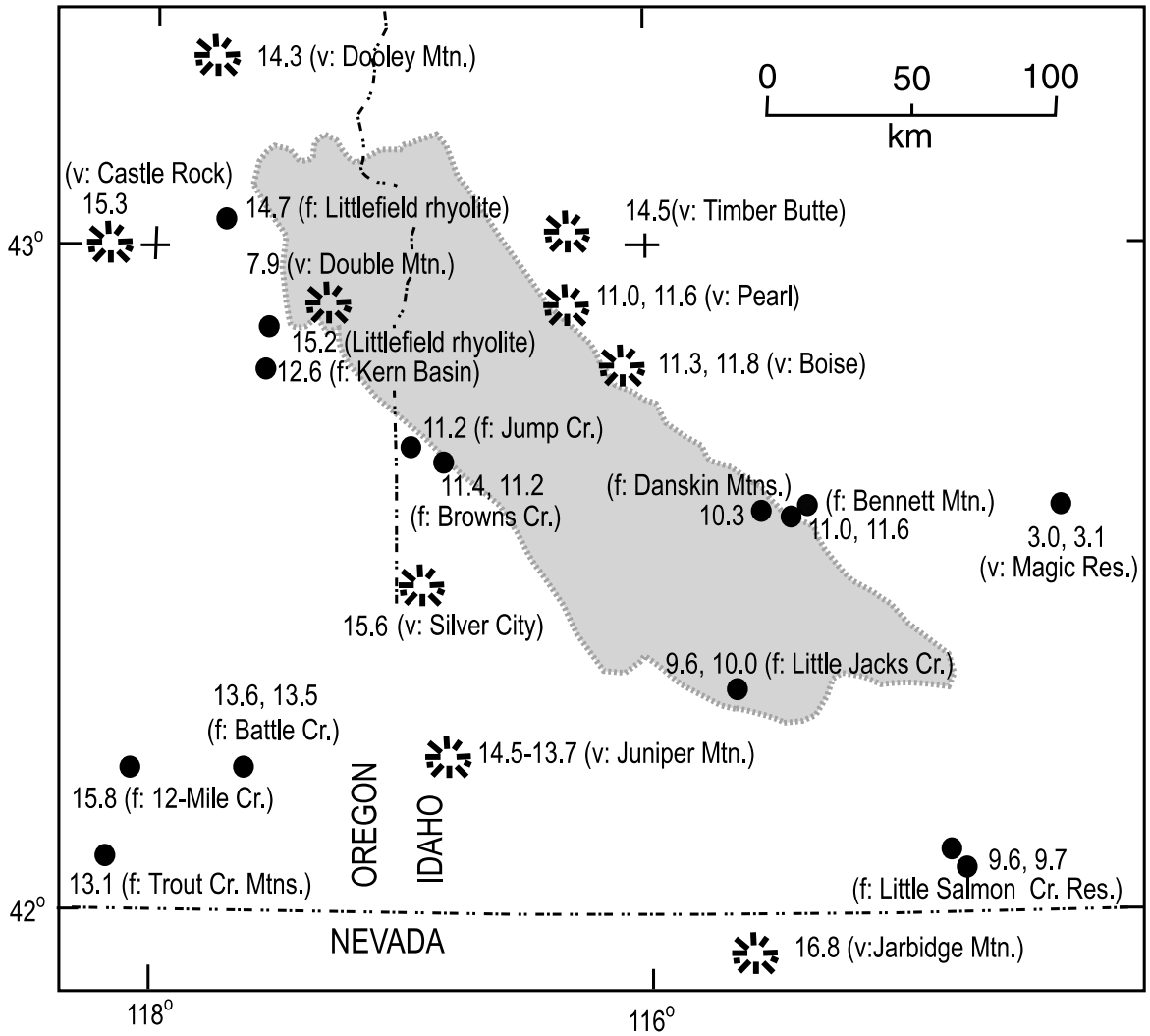

Figure 6. K-Ar ages (Ma) of rhyolite vents (v) and tuffs or lava flows (f) adjacent to the western Snake River Plain from Neill (1975) and Armstrong and others (1980) quoted in Ekren and others (1981, 1984), Clemens (1993), and Clemens and Wood (1991, 1993a, 1993b), Walker and others (1974), Dalrymple and others (1967), MacLeod and others (1975), Ferns and Cummings (1992), Rytuba and Vander Meulen (1991), and Ar-Ar ages from Lees (1992), and Manley and McIntosh (1999). All K-Ar ages calculated using 1976 constants except Trout Creek, 12 Mile Creek, Battle Creek, Salmon Creek, Jarbidge, and Silver City, which may be 0.1 to 0.4 Ma older than shown if recalculated.

East of Boise are two wells that drilled deep rhyolite near the margins of the plain (Anschutz-Federal and Griffith-Bostic). Rhyolite beneath sediment and basalt in the Griffith-Bostic well east of Mountain Home shows that the faulting has displaced rhyolite erupted 10.3 Ma to a depth of 2.0 km (Clemens and Wood, 1993). There the well bottomed in rhyolite at least 0.9 km thick, and the total rhyolite thickness is judged to be about 2.3 km (Wood, 1989). The Anschutz-Federal well (45 km southwest of Mountain Home) drilled through 2.2 km of rhyolite and bottomed in granite (McIntyre, 1979; Ekren and others, 1981). The center of the western plain southeast of Boise has not been explored by deep drilling, so we cannot be certain that rhyolite is absent from the subsurface there. The deepest well is 1.3 km deep at Mountain Home Air Force Base (elevation 3,022 feet, 921 m). This well drilled through 0.75 km of basalt beneath 0.55 km of lacustrine sediment (Lewis and Stone, 1988), a section similar to those in the center of the western plain shown in Figure 5.

Pierce and Morgan (1991) emphasized the broad uplift of 0.5-2.0 km that occurs as the hot spot migrates, and it is shown most graphically by their Plate 1 of topography. Others have suggested that simple vertical expansion due to heating the crustal lithosphere can account for uplift of that order, followed by cooling subsidence

after passage of the hot spot (Brott and Blackwell, 1978). Blackwell (1989) estimates the thermal effect to be 200-250 km wide. This estimate can be projected back in time to the western plain region to suggest that the area was on the north edge of a volcanic highland at 14-11 Ma and not a basin. Before eruption of rhyolite in the western plain area, the geologic evidence discussed earlier shows only north-trending basins laying generally west of the batholith that accumulated basalt. Mapping the rhyolite flow directions in the future should help to reconstruct the topographic picture and confirm or refute the presence and nature of the uplift as the hot spot passed by the position of the western plain.

HISTORY OF GRABEN FAULTING IN THE WESTERN PLAIN

The geochronology of the major graben that forms the western plain has been determined at two localities by correlating dated surface volcanic units to their subsurface equivalents (Figure 7). The offset of volcanic units at the south side of the Bennett Mountains show that most of the 2.8 km offset occurred between 11 and 9.5 Ma. Offset units on the north side of the Owyhee Mountains indicate the 2.2 km of offset was largely completed by 9 Ma, but the date for the onset of vertical movement is not well constrained.

A history of faulting in eastern Idaho along northwest-trending basin and range structures associated with hot-spot migration can be compared with that for the western plain. Anders and others (1989) determined that the rapid rate of faulting south of the eastern plain was constrained in time between 2 and 3 Ma, with vertical slip rates of about 1 mm/year (Figure 7). For the western plain, we derive somewhat smaller slip rates of 0.5 mm/year for the two boundary faults on either side of the plain. Basin relief formed by hot-spot-triggered normal faulting seems to evolve over just a few million years, and then vertical slip rates become very slow (less than 0.01 mm/year). From these fault histories, we argue that most of the western-SRP basin relief was formed in a rather short geologic time between 11.0 and 9.5 Ma. This is not to imply that these normal fault systems become totally inactive after the main period of displacement, but only to note that the average long-term slip rates become very low following the main episode of activity.

The only segment of western-SRP faults shown to have late Quaternary activity is the Owyhee Mountain front 55 km southwest of Mountain Home in the Halfway Gulch-Little Jacks Creek area (Beukelman, 1997). The Halfway Gulch fault trends N. 60º W. to N. 75º W.

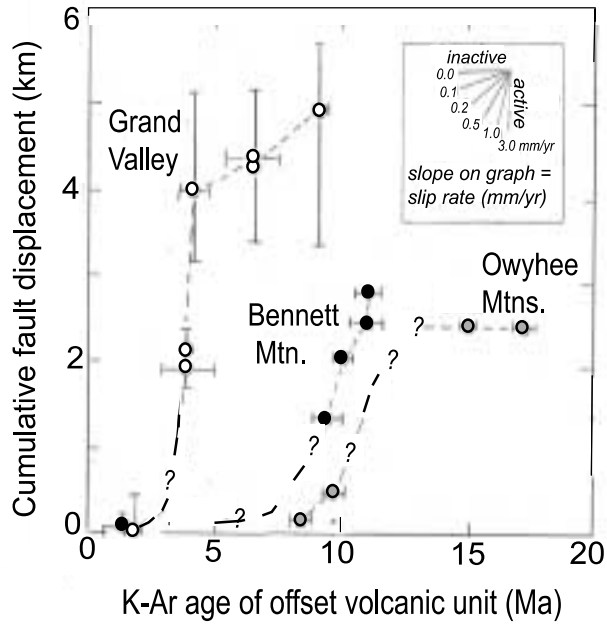

Figure 7. Displacement histories of normal faults bounding the western plain graben (Bennett Mountain is north of and includes the Griffith-Bostic well shown as GB in Figure 2; Owyhee Mountains-Anschutz well is shown as A in Figure 2; modified after Clemens, 1993) and of the Grand Valley fault oriented perpendicular to the eastern plain. Location of Grand Valley fault shown in Figure 1 (after Anders and others, 1989). Illustration shows that basin relief from normal faulting evolves rapidly over a very few million years and that vertical slip rates sharply wane afterwards.

A mappable, late Quaternary scarp occurs along the mountain front for a very short length of only 5.3 km, with a maximum scarp height of 7.7 m. The total length of the system of mappable young faults is less than 12 km, which is unusually short for such large vertical displacement. Scarp-degradation age estimates are about 20,000 years. An associated fault (Water-Tank fault), 7.6 km northeast of the mountain front, strikes N. 35-50º W. and has a 3.6-m Quaternary scarp. A trench-stratigraphic study on the Water-Tank fault shows five episodes of surface rupture within an estimated 26,000 years and estimated average vertical slip rates within these time periods of 0.08 to 0.2 mm/year (Beukelman, 1997). These Quaternary vertical slip rates greatly exceed the long-term slip rates shown in Figure 7 for the Owyhee Mountain front just 20 km to the northwest. This difference in slip rates suggests that episodic reactivation may occur on suitably oriented faults (Beukelman, 1997).

THE EARLIEST SEDIMENT IN THE WESTERN PLAIN: BANBURY BASALT AND THE CHALK HILLS FORMATION

Throughout the western plain margins, a section of basalt flows and pyroclastic layers interbedded with tuffaceous mudstone commonly rests unconformably upon the granite of the Idaho batholith, the late Miocene rhyolite, or the basalt of the Miocene Columbia River Group. At the south margin of the Weiser embayment near Emmett, coarse sand and tuffaceous sediment rest unconformably on Weiser basalt of the Columbia River Basalt Group. Near the plain margins, the lower basaltic material may be interbedded with gravels and sands derived from the batholith. Some of these basaltic rocks have been called the Banbury Basalt. The formation name has been incorrectly applied to any basalt within the sedimentary sequence of the plain. It is unlikely that the Banbury Basalt is a continuous unit or that basalts of identical age occur along the margins and beneath the plain. Basalt eruptive centers probably occurred sporadically in many places and spanned a considerable time after rhyolite volcanism ceased. Bonnichsen and others (1997, p. 401) make a crucial observation that the period of western-plain basalt volcanism following the rhyolite is confined to about 2 million years from 9 to 7 Ma, although we believe that the inception of basalt volcanism extends back to about 10 Ma for reasons indicated in the following discussion.

In some places, sediments may dominate in the basal section, and geologists might call the section the Chalk Hills Formation or the Banbury Basalt with interbedded sediment. We recommend that the term Banbury Basalt be restricted to the basalt field in the vicinity of Banbury Hot Springs, and not be extended to other basalt fields intercalated with sediment of the Chalk Hills Formation. We further recommend that each contiguous basalt field be given a separate name, and each be considered as a member of the Chalk Hills Formation.

In its type area, the Banbury Basalt is about 330 m thick (Malde and Powers, 1961). Armstrong and others (1980) reviewed the age of this basalt section, where it overlies Idavada rhyolite dated 10.1 and 11.0 Ma and encloses a silicic ash K-Ar dated at 10.2 Ma (sample KA 830, whole-rock on coarse ash; Evernden, 1964). Armstrong and others (1980) report two whole-rock K-Ar age determinations of 13.8 ± 1.5 and 8.1 ± 0.7 Ma. These disparate ages were then averaged and reported as 9.4 ± 0.6 Ma, although the validity of averaging such ages is questionable. Ekren and others (1981) report that all the flows they measured had normal magnetic polarity, which suggests the Banbury Basalt erupted during the normal interval between 10.1 and 8.8 Ma.

In the Boise area, a basalt tuff-dominated unit resting on rhyolite or granite is about 200 m thick in drill holes beneath the city, and thinner discontinuous patches occur in the foothills upon the batholith on both the north and south sides of the plain. Clemens (1996) has reported on an undated 4-m-thick rhyolite ash in the lower part of this unit in the Boise foothills.

In the Boise area, the beginning of fluvial-lacustrine sedimentation is dated at before 9.5 Ma by the basalt of Aldape Park intercalated with sediment that crops out in the foothills, and is correlated to the subsurface in geothermal wells. This basalt yields a modified whole-rock K-Ar age of 9.5 ± 0.6 Ma and has a normal magnetic polarity (Clemens and Wood, 1993). A whole-rock K-Ar age of a basalt might be open to question, but the age is within the normal polarity episode from 8.8 to 10.1 Ma, and the other normal episodes are either younger than 8.2 Ma or older than 11.3 Ma, leading us to believe it is a valid age. This basalt layer overlies 150 m of fluvial-lacustrine sediment beneath which is about 200 m of basalt, tuff, and sediment resting upon rhyolite or granitic rocks. Therefore, 150 m of basin sediment was deposited in the Boise area before 9.5 Ma but after emplacement of an earlier basal basalt. We do not have reliable radiometric dates or magnetic polarity on the basalt tuff unit, but it lies on rhyolite dated 11.3 and 11.8 Ma (Clemens and Wood, 1993). We conclude that the 150 m section of sediment beneath the dated basalt marks the beginning of the basin in the western plain between 11.3 and 9.5 Ma.

Sediment that rests on the older basalts, rhyolite, or the Idaho batholith is usually mapped as the Chalk Hills Formation or in some localities as the arkosic sands of the Poison Creek Formation. These formation distinctions are poorly defined and not useful because the arkosic sands are just a fluvial facies that may have a lacustrine equivalent. These sediments are the first clear evidence of the basin in the western plain. Systematic mapping and study of these rocks are yet to be done, but we will review here our present knowledge of the Chalk Hills Formation. The bottommost sediments are usually coarse sand and pebble gravel derived mostly from the Idaho batholith and older volcanics. These sands are interbedded with mudstones which become more prevalent upwards in the section. Within the first 100 m, the section grades upward into tuffaceous muds and clays and many volcanic ash beds, predominantly gray silicic ash and lapilli, but with some basaltic ash beds. Some ash and lapilli beds exceed 20 m in thickness. Pillow basalts occur within these sediments south of Walters Ferry and over an area called the Teapot volcanic complex by Bonnichsen and others (1997) and mapped by Ekren and others(1981) as

basalt of the Murphy area.

We have mapped and measured a 100-m section just south of Walters Ferry in the vicinity of Chalky (sec. 35, T. 2 S., R. 3 W.) and show the complexity of the basal sediments and volcanics of the Chalk Hills Formation in Figure 8. The rhyolite of Browns Creek (11.1 Ma) is overlain by reddish pahoehoe basalt erupted subaerially. The basalt is then overlain by noncalcareous silts with channel arkosic sands in a 50-m sequence that fines upward. Upon this sequence is 30 m of cliff-forming gray silicic ash, mostly silt size, that fines upward. Over the ash are basalt lapilli layers and palagonite tuff. The top of the section here is a 15-m-thick complex of pillow basalt and dikes, called the "Teapot volcanic complex" with a $^{40}Ar/^{39}Ar$ age of 7.95 ± 0.2 Ma (Craig White, written commun., 1997; Bonnichsen and others, 1997). The section beneath the pillow basalt is faulted down to the northeast about 20 m by a northwest-trending normal fault that does not appear to cut the pillow basalt unit.

Swirydczuk and others (1982), Kimmel (1982), and Middleton and others (1985) describe sections of the Chalk Hills Formation on the south margin of the plain. None are measured with respect to the base of the formation. Most are within 50 m of the top of the formation as defined by an oolitic sand and a slight angular unconformity. Kimmel (1982) had some success in tracing ash layers in the formation and obtained nine fission-track ages on glass from the ash layers. Ages ranged from 6.1 to 9.1 Ma with standard deviations of 0.5 to 1.2 Ma. Some question exists regarding the reliability of glass fission track ages, but they are within the range of two whole-rock K-Ar ages on basalt in the lower part of the unit reported by Armstrong (1975) at 8.2 ± 0.7 and 8.6 ± 0.5 Ma.

Perkins and others (1998) examined Kimmel's (1979) stratigraphic section no. 14 (SE 1/4, sec. 19, T. 7 S., R. 4 E.). From trace and major element chemistry, they were able to correlate the ash layers within this Chalk Hills Formation section to regional volcanic ash falls. Ages of these ashes range from 7.49 to 6.4 Ma. The 6.4-Ma ash correlates to the Walcott Tuff on which several whole-rock K-Ar ages, ranging between 6.3 ± 0.3 and 6.5 ± 0.1 Ma, are reported by Morgan (1992). The age of this ash is important because it establishes the youngest known age for Chalk Hills Formation deposition before the formation was partly eroded during a regression of the lake system.

Kimmel's (1979) section no. 14 was examined in this study. The sediments rest upon the subaerial flow basalt of Al Sadie Ranch that rests upon the rhyolite of Horse Basin mapped by Jenks and others (1993). The lower 25 m of sediment is a nonlacustrine sequence of fluvial deposits and paleosols overlain by basalt tuff. The top of the basalt tuff is reworked in part and shows hummocky cross-stratification of shoreline storm waves. The cross-stratified tuff is overlain by 50 m of lacustrine deposits containing the volcanic ashes and fish fossils. From this reexamination and from Perkins and other's (1998) ash chronology, we conclude that the Chalk Hills Formation lake transgressed over this area at some time before 7.49 Ma and that basalt eruptives preceded the lake transgres-

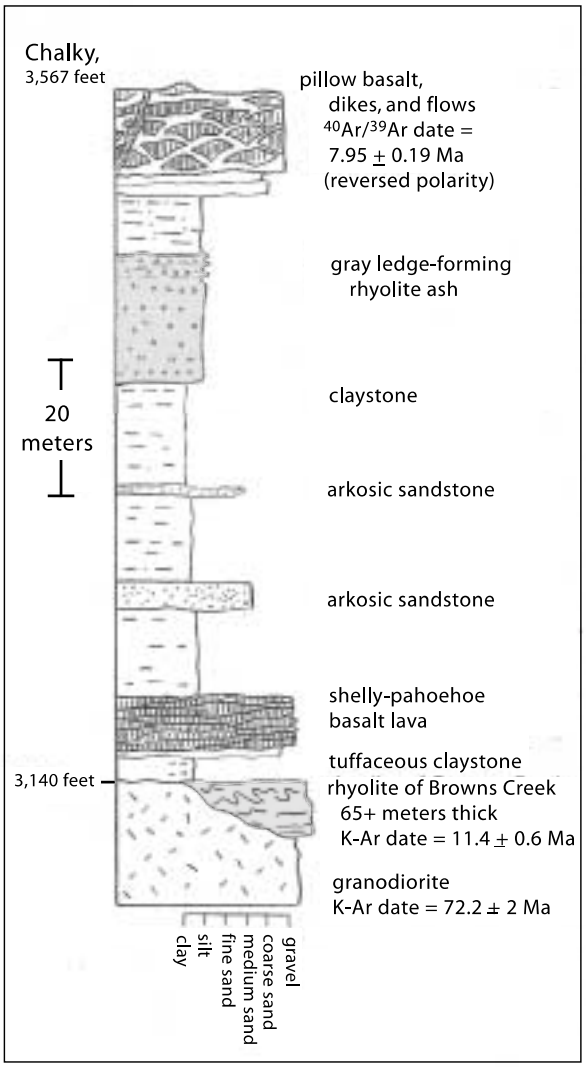

Figure 8. Columnar section of the Chalk Hills Formation at "Chalky" on the south side of the western plain near Walters Ferry (sec. 35, T. 2 S., R. 3 W.). This section was called a reference section (PC-2) for the Poison Creek Formation by Ekren and others (1981), but we regard the Poison Creek as a facies of the Chalk Hills Formation (see text). Section contains two sequences of basalt separated by claystone and a thick rhyolite ash bed. $^{40}Ar/^{39}Ar$ age of 7.95 Ma on the upper basalt is from Craig White (written commun., 1997), and the K-Ar dates on rhyolite and granite are referenced in Ekren and others (1981, 1984). Recalculated age on the Reynolds Creek flow of the Browns Creek rhyolite from Bill Bonnichsen (written commun., 2000).

sion at this locality. The lake then regressed from this site at some time after 6.5 Ma.

The Chalk Hills Formation is relatively thin, about 100 m, along the margins of the plain. It is much thicker beneath the plain (Figure 5). In the type area at the head of Little Valley, Malde and Powers (1962) report the Chalk Hills Formation to be about 90 m thick where it overlies the Banbury Basalt. Ekren and others (1981) report a thickness over 100 m. Sheppard (1991) describes a 100-m-thick section near Oreana that rests on the rhyolite of Little Jacks Creek. The rhyolite of Little Jacks Creek is K-Ar dated at 9.6 ± 1.5 and 10.0 ± 2.0 Ma by Neill (1975; ages recalculated by Bill Bonnichsen (written commun., 2000). The Oreana section contains a 13-m-thick marker gray rhyolite ash, about 35 m above the base, and at least twenty other ash layers interbedded with fine sediment. Sheppard (1991) noted several thin (less than 25 cm) basalt ash layers in the upper part of the formation. The occurrence of one thick (more than 10 m) gray ash within the Chalk Hills Formation and of basaltic ash near the top appears to correlate with our 100-m section (Figure 8) at Chalky near Walters Ferry, about 38 km to the northwest. Thick ash could have erupted from sources to the east; however, at a locality 10 km south of Marsing (SE$^1/_4$ sec. 32, T. 2 N., R. 4 W.) is a 100-m section that contains a 1.5-m layer of rhyolite pumice blocks up to a meter in diameter, 80 m above the base. The layer can only be from a local eruption of a rhyolite dome beneath the lake water (Wood and Wood, 1999), and it shows that at least one rhyolite system continued to erupt near the southeast margin of the western plain during deposition of the formation.

Sheppard (1991) discusses the chemistry of the "Chalk Hills lake," drawing information from an examination of ostracodes by R.M. Forester. The ostracode fauna indicate that at one time the lake-water salinity was greater than 300 mg/l and less than 3,000 mg /l and had a pH about 8 to 9. The water chemistry, though somewhat alkaline, is consistent with lake water supporting the freshwater fish fauna described by Smith and others (1982), not unlike many of the fish-populated Great Basin lakes of today.

Smith and Patterson (1994) show that the water of the "Chalk Hills lake" produced carbonates that are extremely depleted in the heavy oxygen isotope (^{18}O). This suggested to them that the lake was maintained by tributaries of high elevation watersheds and that the waters were little affected by evaporation. They also report that the lake had an unusual mix of fish fauna. Cold-water fish (Salmon and trout species) occur with warm-water species of catfish and sunfish. The lake contained no sculpins or whitefish (cold-water species). It has not yet been resolved if any of the fish were anadromous (Gerald Smith, oral commun., 1995).

Much of the Chalk Hills Formation sediment is noncalcareous mudstone. We are unaware of any significant carbonate facies in the formation. Kimmel (1979) does not describe any calcareous sediment in the many stratigraphic sections he measured. Gypsum partings and veins occur locally near the base of Chalk Hills Formation mudstones. Within thick siltstone layers is a 0.3-m gypsum layer in sec. 23, T. 21 S., R. 46 E., Malheur County, Oregon (Kimmel, 1979, p. 262) and gypsum is associated with volcanic ash layers in Owyhee County, Idaho (sec. 19, T. 7 S., R. 4 E.; Kimmel, 1982). We have noted selenite and satin spar as veins and partings in laminated bentonitic mudstone at the base of the formation north of Marsing (sec. 4 and 5, T. 1 N., R. 4 W.). Kimmel (1979) interpreted these as "displacement gypsum" probably formed by shallow ground-water precipitation in the lake muds. Veins and partings of gypsum in mudrocks suggest that shallow ground water was enriched in sulfate and calcium, and this probably occurred beneath local areas of restricted lake waters that underwent seasonal evaporation. However, we wish to emphasize that the gypsum occurs early in the history of deposition and not in the bulk of the later Chalk Hills Formation.

Kimmel (1982) puzzled (as we have also) over the fall in lake level at the end of Chalk Hills Formation deposition and the subsequent rise in lake level and deposition of the transgressive unit and the Glenns Ferry Formation. He gave several alternate hypotheses for the major lake fluctuation, which we believe reached its lowest elevation at some time after 6.4 Ma. One of these hypotheses involves both tectonic movement and downcutting of the outlet to produce a low lake level, and then basalt volcanism blocking the outlet. Tectonic movement was significant based upon the faulted and slightly tilted nature of most of Chalk Hills Formation and our analysis of rates of faulting (Figure 7). Therefore, it is possible that the basin and its outlet were tectonically lowered. There is no evidence in the sediments of major evaporation in the upper part of the formation to suggest climate change or reduced inflow; however, the stratigraphy of the formation is in need of review to understand the lowering of the lake level. We know nothing of the location of the outlet, but basalt volcanism blocking an outlet in the western plain region is an unlikely cause of the subsequent rise in lake level because Bonnichsen and others (1997) do not find significant basalt volcanism in the western region for the interval between 7 and 2.2 Ma. Rhyolite volcanism, however, was active in what is now the eastern plain region, and that volcanism combined with uplift of the hot-spot region could account for blocking an eastern outlet, if it existed there.

We visualize an environment of a large basin with sporadic basalt volcanism and high rates of basin-relief formation by faulting. At many localities around the plain, basalt volcanic rocks underlie the basal fluvial or lacustrine sediments and are also intercalated with the sediments. From Kimmel's (1982) work, the interconnected Chalk Hills lakes apparently extended from the Bruneau area to the Oregon border, a distance of at least 110 km. From Smith and others' (1982) and Sheppard's (1991) work, the lakes were at times slightly alkaline, but they supported a fresh-water fish fauna. This suggests a system of river interconnections through the evolving topography in the basin, but perhaps not as great a flow-through as the present Snake River discharge. From the above discussion of stratigraphy in the Boise area, the "Chalk Hills lake system" is younger than 10.1 Ma and includes basalt dated 9.4 Ma in the Boise foothills and a pillow basalt complex on the south side of the plain dated at 7.95 ± 0.19 Ma. For the youngest date on the Chalk Hills, we go to the south side of the plain and use Kimmel's (1982) glass fission-track ages between 6.1 and 9.1 Ma, but accepting the uncertainty of these ages. Identification of the Walcott tuff (6.4 Ma) by Perkins and others (1998) in the upper part of the formation establishes a maximum age for the drop in lake level at the end of "Chalk Hills Time." The top of the formation has been removed by erosion marked by a slight unconformity at most localities. Future research should focus on finding a complete upper section in order to better understand events leading to the drop in lake level and the resulting unconformity.

DROP IN LAKE LEVEL AT THE END OF "CHALK HILLS TIME"

A key problem is defining the top of the Chalk Hills Formation. Swirydczuk and others (1979, 1980) show convincingly that a lacustrine shoreline sequence transgresses over beveled, gently tilted strata of the Chalk Hills Formation in the Oreana area. Kimmel (1982) concluded that the Chalk Hills lake lowered or completely drained at the end of the Miocene, and then filled again. Smith and others (1982) suggest that about 1 million years of geologic record is missing in the hiatus between the beveled lake deposits of the Chalk Hills Formation and the transgressive shoreline deposits, and that the missing time is somewhere in the 6 to 4 Ma interval. In some places, the contact is an unconformity with underlying Chalk Hills Formation dipping 4 to 12 degrees basinward, overlain by lake beds dipping less than 4 degrees. At these localities, the upper part of the Chalk Hills Formation has been eroded away. If lakes persisted in the deeper parts of the basin, sediment preserved in the subsurface may contain a more complete sedimentary record. Figure 5 shows the seismically imaged deep sediment that is thought to be the upper part of the Chalk Hills sedimentary record. Because part of the Chalk Hills Formation has been eroded from the margins and we have not, as yet, identified this unconformity in the subsurface, the decline in lake level at the time of the upper Chalk Hills is poorly understood. Apparently, the duration of a lake system could be from 10.1 to about 6 Ma as illustrated in Figure 9.

The transgressive sequence marked by the lowest occurrence of oolitic shoreline sand has been used as the definition of the base of the Glenns Ferry Formation (Malde and Powers, 1962). In the Boise foothills area (Figure 10), lenses of oolitic sand occur over a 120-m-vertical section within the upper part of a shoreline facies that W.L. Burnham and S.H. Wood (written commun., 1992) have named the Terteling Springs Formation. We regard most of the Terteling Springs Formation as a transgressive unit. We have not been able to find a clear indicator in the Boise area of the top of the Chalk Hills Formation. We have found a record of lake-level rise but have not found a record of lake-level drop similar to the unconformity on the south side of the plain.

Possibly, the subsurface alluvial fan deposits of southwest Boise, delineated by Squires and others (1992) and shown in Figure 11, are a record of lake-level fall in Chalk Hills Formation time because the base of the fan rests upon clayey sediments at a subsurface elevation of 2,000 to 2,400 feet (610 to 730 m). The subsurface fan is overlain by lake deposits at about elevation 2,700 feet (820 m). Ed Squires (written commun., 1990) has mapped an outcrop of oolitic sands in the Mayfield area to the east at elevation 3,600 feet overlying the fan deposits. From these relationships, he concludes a lake transgressed over the alluvial fan deposits. If so, then the subsurface alluvial fan deposits are much older than we have previously thought and are contemporaneous with the upper Chalk Hills Formation.

THE TRANSGRESSIVE LACUSTRINE SEQUENCE

A history of the major water-level fluctuations of Lake Idaho is hypothesized in Figure 9. Some event caused the lake level to rise and transgress over the Chalk Hills Formation. The upper part of this transgressive sequence contains oolite lenses marking shoreline regions. Repenning and others (1994, p. 72) thoroughly reviewed

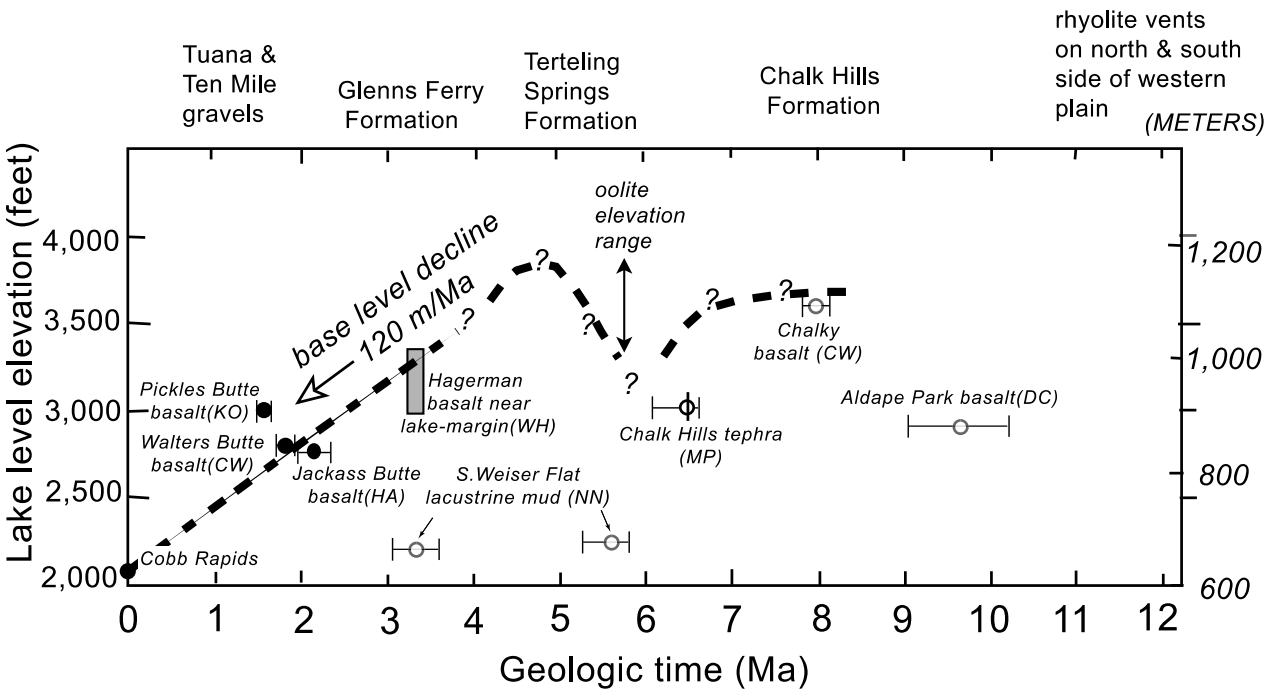

Figure 9. Plot of elevation of lake deposits versus time. This plot does not take into account tectonic deformation that may have altered the elevation of lake deposits. Most localities are on the margin of the plain, which has not been so much affected by tectonic movement or compaction subsidence (Wood, 1994). Points on the graph are dates on lacustrine sediment, or basalt associated with or overlying the lacustrine section: K-Ar dates on basalt are from DC (Clemens and Wood, 1993) and HA (Amini and others, 1984); zircon fission track ages on silicic ash are from NN (Nancy Naeser, published in Thompson, 1991); $^{40}Ar/^{39}Ar$ ages on basalt are from CW (Craig White, written commun., 1997), KO (Othberg and others, 1995), and WH (Hart and others, 1999); and tephrochronology is from MP (Perkins and others, 1998).

previous work on the Glenns Ferry Formation. They proposed that the locally oolitic, rusty-stained sandstone, taken by Malde and Powers (1962) as the base of the Glenns Ferry Formation, be given separate formational recognition. Our mapping on the north side of the plain supports that proposal, for we see at least 120 m of an oolite lens-bearing section up to elevation 3,800 feet (1,160 m). The stratigraphic diagram (Figure 12) shows this relationship in the Boise foothills with oolite lenses up to 3,200 feet (975 m). The lower elevation (3,200 feet) shown in the section is because of the southwest dip of the strata. In principle, this transgressive sequence should have an equivalent open lake facies that may be quite thick. We believe the mudstone facies of the Terteling Springs Formation to be that open lake facies. Consequently, the basinward equivalents of oolite facies on the south side of the plain need to be reexamined because there should be a substantial correlative section of mudstone. Only the 30 m beneath the oolite is described in the literature on this section (Swirydczuk and others, 1980a, 1980b, 1982).

An important implication of a transgressive or rising lake-level sequence is that sand and coarse clastics will be deposited near the shoreline and that deltas will not prograde significantly into the lake basin. Therefore, much of the incoming sand is stored as nearshore sediment, and not deposited in the open lake where only muds are deposited. The rising lake level also suggests that the lake does not have a spillway. A closed basin might have been forming. A steadily filling closed basin perhaps explains the oolites in the upper part of the section. The lake was becoming increasingly alkaline from evaporation as it rose to higher levels, thus favoring precipitation of calcium carbonate. Swirydczuk and others (1980b) compare the lake chemistry at the time of oolite deposition with that of Pyramid Lake, Nevada (pH of 9, and 4,700 mg/l total dissolved solids, but supporting a healthy trout population). Finally, the lake overtopped a spill point and began to slowly drain. As soon as drainage was established, the lake would have become less alkaline.

THE LAKE-LEVEL FLUCTUATION AT 6-4 MA AND A PLAUSIBLE EXPLANATION

The history of lake levels shown in Figure 9 ignores the effects of differential tectonic movement, which with

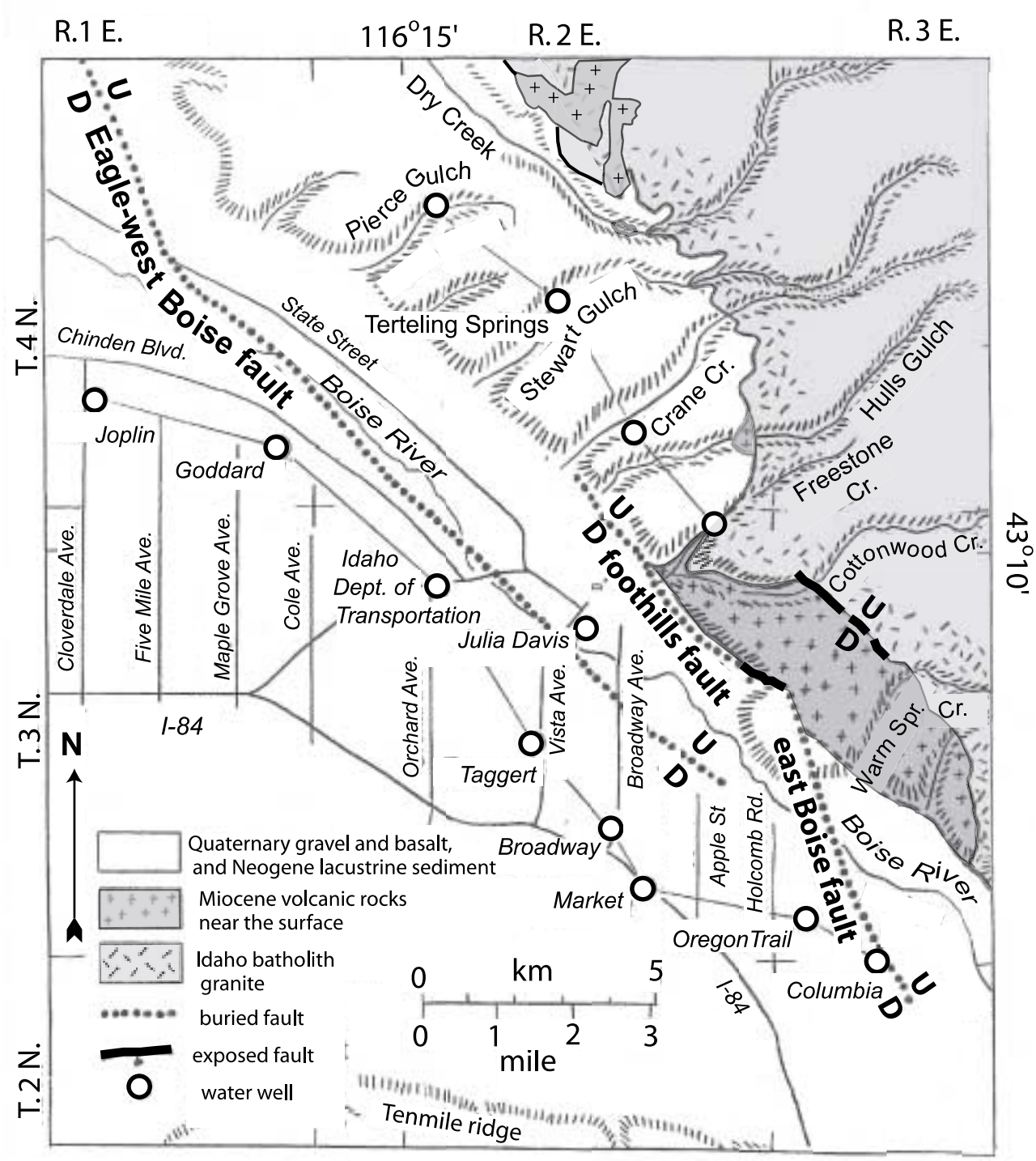

Figure 10. Geologic map of the Boise area showing locations of stratigraphic cross sections of Figure 11 and Figure 12.

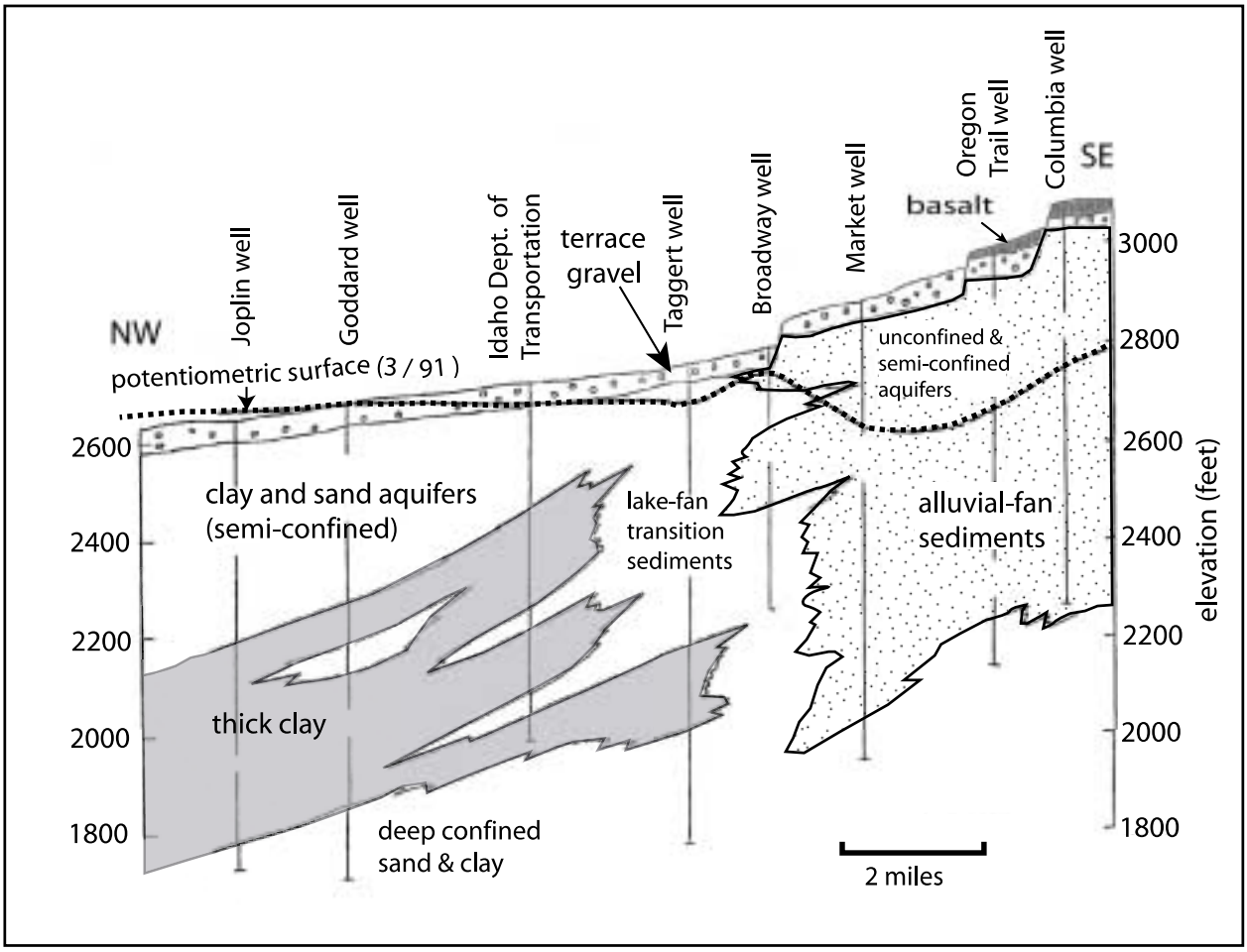

Figure 11. Cross section showing stratigraphy of alluvial fan and lake sediment aquifers beneath Boise (from Squires and others, 1992).

the present knowledge is difficult to unravel from other causes of lake-level fluctuation. We list the possible causes of secular lake-level fluctuations (Table 1) but will pursue only our favored mechanism to keep the discussion short. At the outset, we must state that our study of the Chalk Hills Formation is very limited. The accumulation of the formation is likely a result of continued tectonic foundering of the basin area. The subsequent fall in lake level, about 6 to 7 Ma, is poorly documented and without a satisfactory explanation. The rapid tectonic foundering of the basin, the change to a more arid climate, or the establishment of a lower outlet are all favored possible explanations.

The lake-level rise, marked by the transgressive sequence (oolite section and the upper Terteling Springs Formation), is well documented (Swirydczuk and others, 1980a, 1980b). We propose that the major lake-level fluctuation at about 6 Ma was caused by new stream inflows associated with the migrating Yellowstone hot-spot

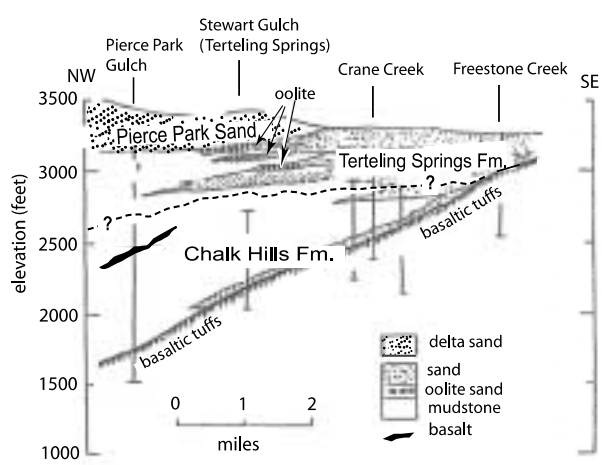

Figure 12. Stratigraphic cross section showing lake margin facies in the Boise foothills and the distribution of oolitic sands. Location of section shown in Figure 10.

Table 1. Causes of lake-level fluctuations.

RISE IN LAKE LEVEL
Increased inflow
Increased runoff
Increased precipitation
Decreased evapotranspiration in catchment
Diversion of major stream into basin by capture or damming
Decreased evaporation from lake surface
Decrease in surface area
A climate change, colder or more humid
Filling of lake basin by sediment
Rise in elevation of the outlet
Landslide, glacier-ice, or lava-flow damming
Relative tectonic warping or faulting of outlet area
DECLINE IN LAKE LEVEL
Decreased inflow
Decreased runoff
Precipitation decrease
Increase in evapotranspiration in catchment region
Diversion of a major tributary stream from basin by capture or damming
Increase in evaporation from lake surface
Warmer or dryer climate
Increase in lake surfacearea
Sediment compaction
Fall in the level of the outlet
Progressive downcutting of outlet
Breaching of a landslide, glacier-ice, or lava-flow dam
Spillover and establishment of a new and lower outlet
Tectonic lowering of outlet by warping or faulting

uplift. An uplifted region about 400 km across with a maximum uplift of 0.5 to 1 km is currently centered on the locus of rhyolite volcanism at Yellowstone (Pierce and Morgan, 1992). The hot-spot model implies a broad northeast-migrating continental uplift region. A migrating uplift also implies an eastward-migrating drainage divide (Figure 13).

Anderson (1947) proposed that the Continental Divide had shifted eastward about 160 km in late Tertiary time. He was intrigued by the peculiar shift in directions of the Salmon River and drainages north of the eastern plain. We realize now that the hot-spot uplift provides a mechanism for his observations. Taylor and Bright (1987) also suspected that the migrating uplift was responsible for drainage changes in the region. Cox (1989) suggested this has been a characteristic of continental hot spots. The Continental Divide may have shifted about 200 km from a position near Arco, where it was associated with the Heise Volcanic field about 6 Ma, to its present position in Yellowstone National Park.

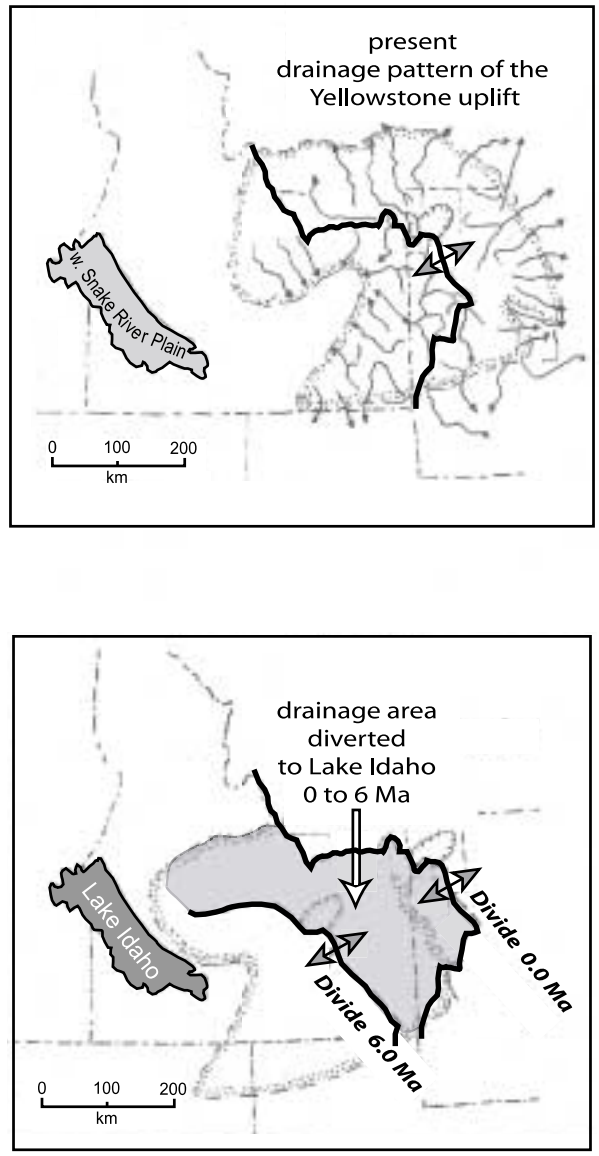

Figure 13. Map showing the area that may have diverted drainage to Lake Idaho, 6 to 0 Ma, on account of eastward-migrating uplift associated with the Yellowstone hot spot.

The idea of a shifting drainage divide could explain a part of the history of lakes in the western plain. About 6 Ma, the lake that was depositing the Chalk Hills Formation declined to a low level. The shifting drainage divide does not offer a simple explanation for the decline to a low level; however, some earlier spillover into another lower elevation basin, a climate change decreasing inflow and increasing evaporation, or an increase in the rate of subsidence could account for the lake-level lowering. Then, about 4.5 Ma the lake began to rise, transgressing over the older Chalk Hills lacustrine deposits. We propose here that the eastward-migrating uplift with

a west slope captured major rivers or allowed major drainage diversions into the western plain basin (Figure 13). We suggest that drainages that had formerly flowed east to the Missouri or Colorado drainages, or into Basin and Range dead ends, now flowed down the west slope of the uplift into Lake Idaho. Such major diversions could have increased the inflow into Lake Idaho causing it to rise to its spill point about 4.0 Ma.

THE SPILLOVER OF LAKE IDAHO INTO HELL'S CANYON AND THE DEPOSITION OF THE GLENNS FERRY FORMATION

Wheeler and Cook (1954) published a key concept in deciphering the history of Lake Idaho. They propose that headward erosion to the south by a tributary to the Salmon-Columbia river system captured the waters of Lake Idaho in the late Cenozoic. Although we differ in the details, the spillover into ancestral Hell's Canyon seems certain. They believed spillover and capture occurred at the Oxbow because of the peculiar bends of the Snake River. The Oxbow and its trend are more satisfactorily explained by diversion of the downcutting river by the resistant northeast-trending mylonite zone in the pre-Tertiary rocks. Wood (1994) believed the capture occurred near the present Cobb Rapids by Weiser. We now believe the spillover point was a divide about 5 km above the present confluence of the Burnt River with the Snake River through a low gap between the present Slaughterhouse Range and Dead Indian Ridge (Figure 14). Our observations are mostly from topographic maps: the area has not been systematically mapped and searched for remnant patches of lake sediment and gravel to test these ideas. We hypothesize a former low divide at about 3,600 feet (1,100 m) in elevation at the headwaters of a steep-gradient, north-draining tributary. The lake, which lay to the south, rose and overtopped this divide. A similar divide exists today at Henley Basin (Figure 14), 8 km to the northeast, a gap similar to that which must have existed at the spillover. This gap is the 3,200-foot (980-m) elevation divide between the north-flowing Rock Creek and the south-flowing Hog Creek. It is a gap through which rising lake water also could have flowed over into ancestral Hell's Canyon, had it not already done so in the gap to the southwest, presumed to have been slightly lower at the time of spillover. That this gap is presently 3,200 feet (989 m) is not really a problem, since in the past several million years, it could have been lowered by erosion 400 feet (120 m; this assumes a reasonable denudation rate of the divide area of 200 feet/Ma or 6 cm/Ka). Note also on Figure 15, that the river gradient steepens in the 20-mile (32-km) reach from the mouth of Rock Creek to Cobb Rapids, suggesting that the smoothed remnant of the former knickpoint created by the spillover has migrated about 10 miles (16 km) above the spillover point.

We hypothesize that most streams on the south side of the Blue Mountains formerly flowed to the south-southeast into the lake. The Burnt River and many nearby creeks have this direction; however, the Burnt River turns sharply northeast in its lowest 4-km reach below Huntington, where we believe it was captured and diverted from its southerly course by the downcutting Snake River.

Repenning and others (1994, p. 71-72) and Van Tassell and others (2001) suggest that the spillover occurred in the early Pliocene, rather than with the onset of the ice ages of the late Pliocene as previously suggested by Othberg (1988) and Wood (1994). Repenning and others (1994) propose that much of the Glenns Ferry Formation deposition occurred after the lake had found the Hell's Canyon outlet, which they believe happened between 3 and 4 million years ago. Othberg (1994) also suggests that the basin may have filled earlier than late Pliocene. He further suggests that the lake drained slowly while flood-plain aggradation continued in the basin. He does associate the beginning of widespread gravel deposition over the plain with a change in stream regimen caused by late Pliocene climates. From his study of gravel terraces and deposits of the Boise Valley, he indicates that basin aggradation gave way to valley cutting in the early Pleistocene.

The time when spillover occurred is not well constrained. Lacking is accurate geochronology tied to traceable marker beds and detailed stratigraphic work in the upper Glenns Ferry Formation. The timing of spillover assumed by other workers rests upon magnetic polarity changes in sedimentary section determined by Neville and others (1979) and Conrad (1980) in scattered stratigraphic sections that were not tied to radiometrically dated ashes or flows or by geologic mapping. We believe the reliable geochronology is as follows: (1) the 6.4-Ma date near the top of the Chalk Hills Formation determined by Perkins and others (1998); (2) the Ar-Ar ages obtained by Hart and others (1999) on basalt at Hagerman and paleomagnetism of sediment determined by Neville and others (1979) and the volcanic ash correlations of Izett (1981) that indicate ages between 3 and 4 Ma for the Hagerman section; and (3) the Ar-Ar age of 1.67 Ma for the subaerially erupted basalt at Pickles Butte obtained by Othberg (1994). These dates reliably set the maximum age for the beginning of the transgressive sequence, the age when the lake shore was near Hagerman, and the age when the lake basin filled with sediment and river gravel

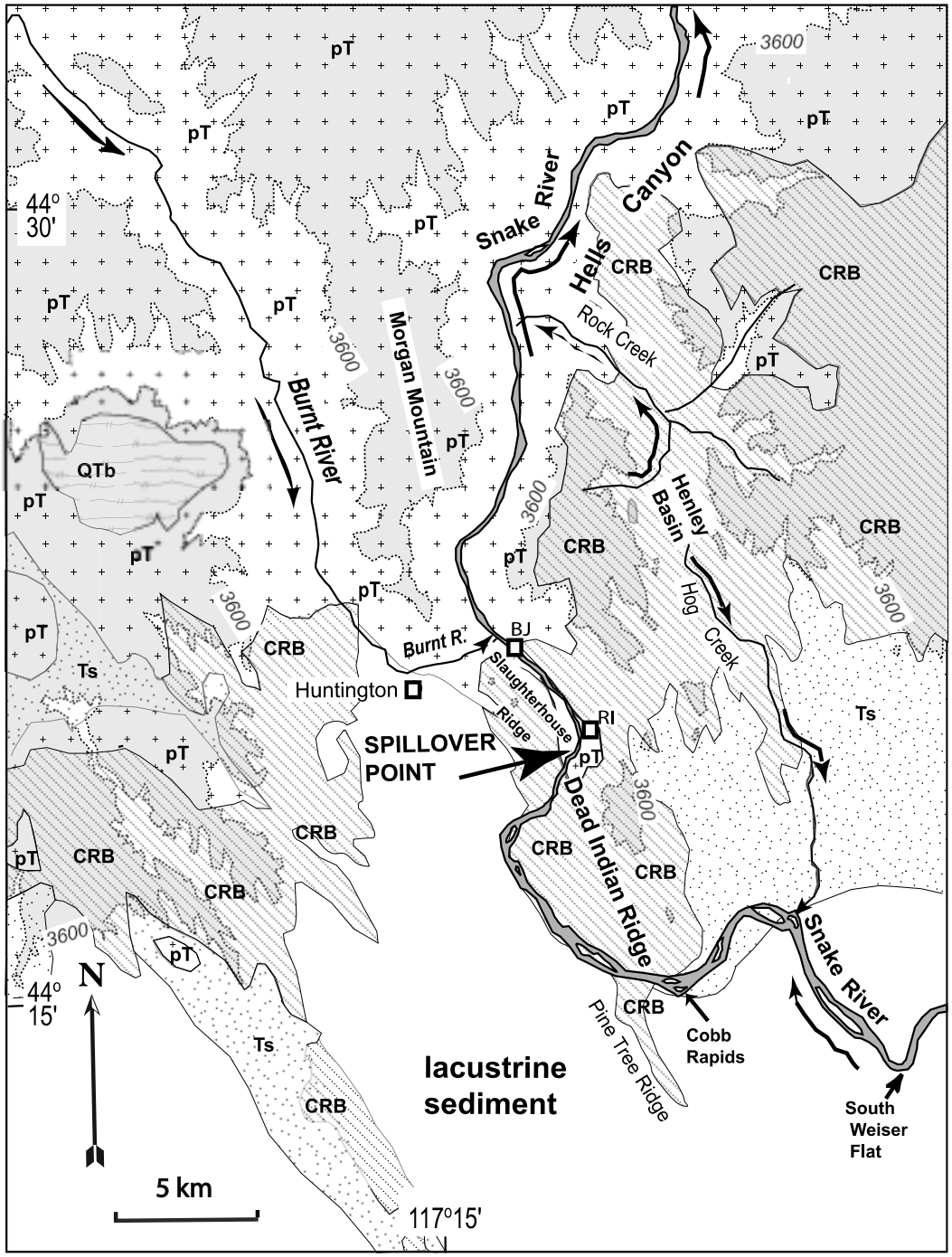

Figure 14. Present topography above 3,600 feet (shaded area) and geology at the upper end of Hell's Canyon showing the proposed spill point of Pliocene Lake Idaho. We hypothesize that the headwaters of a north-flowing tributary of ancestral Hell's Canyon were between Blakes Junction (BJ) and Rock Island Station (RI). The Pliocene divide between the Columbia-Salmon River system and Lake Idaho was between the present Slaughterhouse Ridge and Dead Indian Ridge. That divide was overtopped when the lake rose to about 3,600 feet. Henley Basin is a similar divide that still exists between north-flowing Rock Creek and south-flowing Hot Creek. The knickpoint formed by the capture has subsequently migrated 9 km up to Cobb Rapids (see Figures 15 and 17). As the river canyon lowered, a tributary creek on the west captured the lower part of the south-flowing Burnt River. Geologic units are pT—pre-Tertiary rocks of the Olds Ferry terrane (Vallier, 1998); CRB—Columbia River Basalt Group; Ts—older deformed lacustrine and fluvial sediment, probably equivalent to Sucker Creek and Chalk Hills Formations; QTb—cinder cone and basalt flows of probable Pliocene age; lacustrine sediment—younger sediment of Lake Idaho.

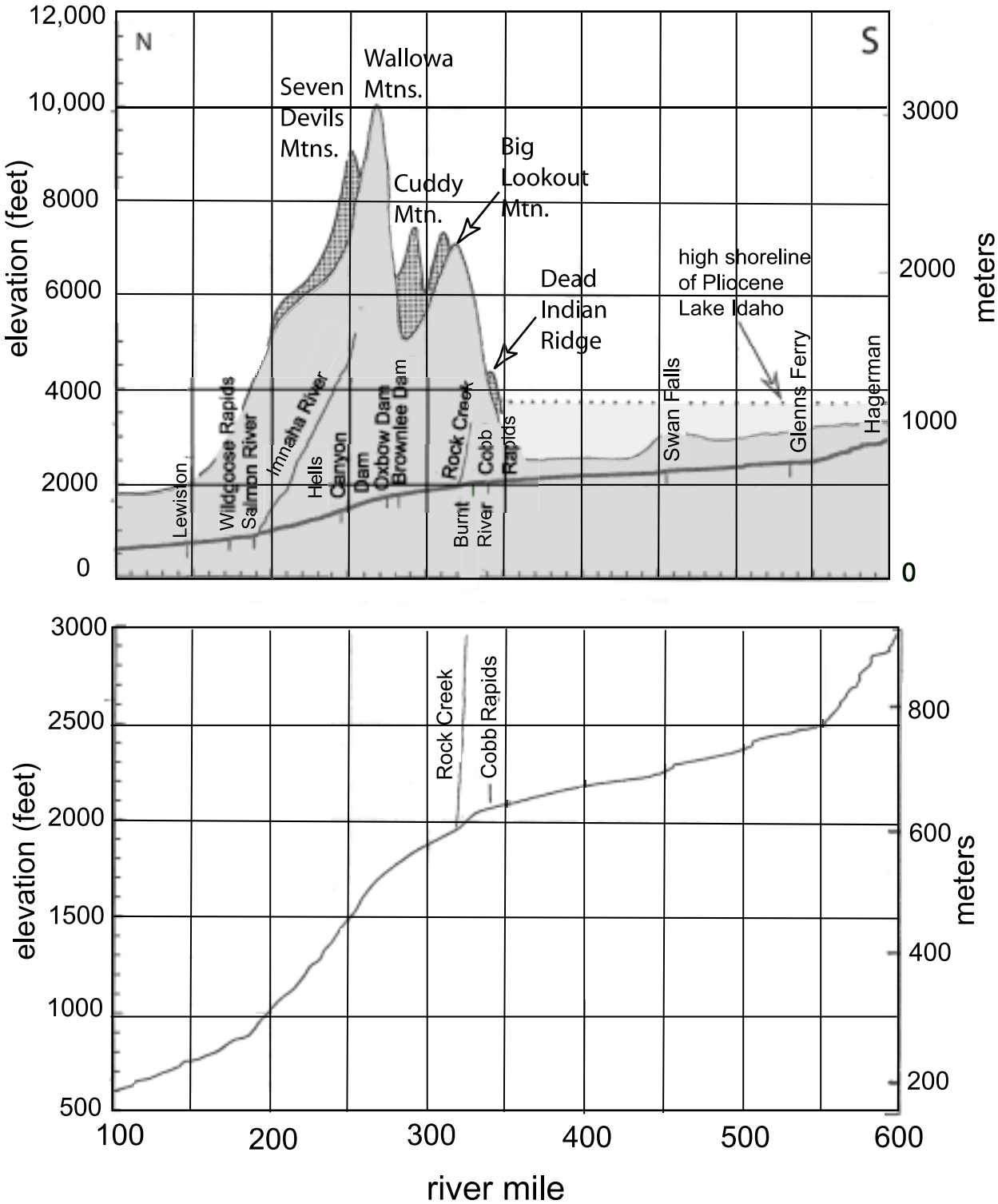

Figure 15. Longitudinal profile of the Snake River showing the change in gradient from Hell's Canyon to the Snake River Plain. Included is the profile of the Imnaha River, which has a source in the high Wallowa Mountains, and Rock Creek, which has headwaters at a divide similar to the proposed spillover location shown in Figure 14. It is thought the ancestral upper Snake River had a similar high-gradient profile at the time it breached the divide into the western plain and began draining Lake Idaho.

deposited over lake sediment near Marsing, respectively. Therefore the rise, spillover, fall, and sediment infilling of the lake occurred at sometime between 6.4 Ma and 1.67 Ma. Van Tassell and others (2001) and Smith and others (2000) provide new geochronological and paleontological evidence suggesting that Lake Idaho connected to the Columbia River drainage at some time between 3.8 and 2 Ma.

It will be important to establish whether the Hagerman section represents a transgressive sequence of lacustrine deposits over beach deposits, deposits of a highstand of the lake, or fluvial deposits over lacustrine deposits. We believe the facies at Hagerman could be the transgressive stage, the highstand, or a facies of the early stages of draining an alkaline lake. Our contention is that the medium and fine sand interbedded with calcareous silt and clay throughout the 250- to 400-foot (75-120 m) sections described by Malde (1972) at Fossil Gulch and Peters Gulch indicate lake conditions of a closed alkaline lake, or one just beginning to drain. Wood (1994) described 1,000 feet (300 m) of very calcareous claystone overlain by 800 feet (240 m) of interbedded fluvial and deltaic sand interbedded with moderately calcareous silt in a geothermal well beneath Caldwell. South of Marsing, the lacustrine sediments above the Chalk Hills Formation and beach gravels are composed of 300 feet (90 m) of noncalcareous mudstone (with gypsum cemented ashes and gypsum veins) conformably overlain by 1,000 feet (300 m) of very calcareous mudstone interbedded with fine and medium sand in the upper part (S.H. Wood, unpub. mapping). We believe the thick calcareous sediments are a record of an alkaline lake of a closed basin.

We constructed the diagram of Figure 9 without knowing that Hearst (1999) had published a similar hypothetical diagram of lake-level elevations over time. She shows the spillover as having occurred about 2.7 to 2.5 Ma, based upon a re-evaluation of the geochronology of Kimmel (1982), Neville and others (1979), and Repenning and others (1994). She accepts the glass fission-track ages of Kimmel (1982) and finds that they concur reasonably with paleontological age estimates of fauna in the Glenns Ferry Formation. Her work has caused us to rethink the ages in our Figure 9. Kimmel's (1982) dates are on volcanic ash within 17 m of the base of the Glenns Ferry Formation. The ages range from 2.2 to 3.3 Ma, and since they are based upon fission tracks in glass and not in zircon, and because they are near the base of the formation, we suspect they may be too young for that part of the section. The ages of 5.5 ± 0.4 Ma and 3.3 ± 1.0 Ma at south Weiser Flat, shown in Figure 7, are fission-track ages on zircons from volcanic ash within lacustrine sediments, but we are uncertain of their correlation to the Glenns Ferry Formation. These ages on ash date a time of deposition of noncalcareous lacustrine mudstone at that site at the northwest end of the western plain basin. Future work should focus on better dates on volcanic ashes in the section explored by Hearst (1999) and the section in the Boise foothills where major sand units prograde over lacustrine muds in the upper part of the section and are an indication of falling lake level after the lake spilled over the outlet into Hell's Canyon.

We realize now that the slow downcutting of the bedrock outlet and deepening of Hell's Canyon are good explanations for features of the final phase of Lake Idaho. Helpful to our understanding were the typical rates of downcutting of bedrock canyons by major rivers (Schumm and others, 1991). Rates are typically about 150 m/Ma. We have added comparative data from other studies that support these relatively slow rates (Table 2). A reconstructed history of lake levels (Figure 9) suggests that the lake level lowered at a rate of about 120 m/Ma. Although climate change may have affected the stream regimen, the rates of downcutting of Hell's Canyon, and the lake levels, we do not think that climate change is necessary to explain the sequence of major events in the late history of Lake Idaho.

A slow lowering of base level would have caused the erosion of sands from the lake margins and delivered sand into the basin by delta progradation, a feature noted in the center of the basin by Wood (1994) but its significance not clearly understood at the time. The abrupt change from deep lake mudstone to prodelta and delta deposits occurs over much of the western plain where it is detected by geophysical logs (Figure 16). We believe this change is a result of the slowly lowering lake level and of sandy sediment prograding into and filling the basin. Wood (1994) calculated from the relief of prodelta slopes that the lake water into which the deltas prograded was at least 255 m deep. Not until the entire western part of the lake was filled did streams flow across the plain to the outlet. At Weiser, the highest gravel deposits are about elevation 2,500 feet (762 m), and we believe these record the time at which the lake basin completely filled with sediment and stream gradients had aggraded across the plain so that braided streams transported gravel across the plain to the outlet area, leaving deposits known as the Tuana Gravel and the Tenmile Gravel (Sadler and Link, 1996; Othberg, 1994). The ages of the gravels must decrease across the plain to the west, but it is believed that the streams were continuous to the outlet near Weiser by about 1.7 Ma (Othberg, 1994). These highest gravels have subsequently been incised by the Snake River and tributaries about 120 m. Cobb Rapids of the Snake River at elevation 2,080 feet (635 m; Figure 17) is the present knickpoint in Miocene basalt as the river changes grade

Table 2. Rates of river incision into bedrock canyons (after Schumm and Ethridge, 1994).

Rate (cm/1,000yr)	Rock Type	Location	Reference
4.0	granite	SE Australia	*Brittlebank, 1900
9.6	basalt	SE Australia	*Brittlebank, 1900
9.5	conglomerate	Arizona	*Rice, 1980
24.8	sandstone	Arizona	*Rice, 1980
30.0	sedimentary rock	Colorado	*Larsen et al., 1975
7.0	metamorphic rock	Colorado	*Scott, 1975
37.0	limestone and basalt	Utah	*Hamblin et al.,1981
2.6	limestone and basalt	Utah	*Hamblin et al., 1981
8.7	granite	Sierra Nevada, west slope, California	*Huber, 1981
50.0	unknown	Dearborn River, Montana	Foley, 1980
5 to 10	granite	Boise River, Idaho	Howard, et al., 1982
30 to 70	Quaternary intracanyon basalt	South Fork Boise River, Idaho	Howard, et al., 1982
15 to 16	Eocene lacustrine sediment	Wind River, Wyoming	Chadwick et al., 1997
11 to 16	granite and metamorphic rock	Middle Fork Salmon River, Idaho	Meyer and Leidecker, 1999

*Schumm and Ethridge's (1994) table contains references to data prior to 1981, which are not referenced in this paper.

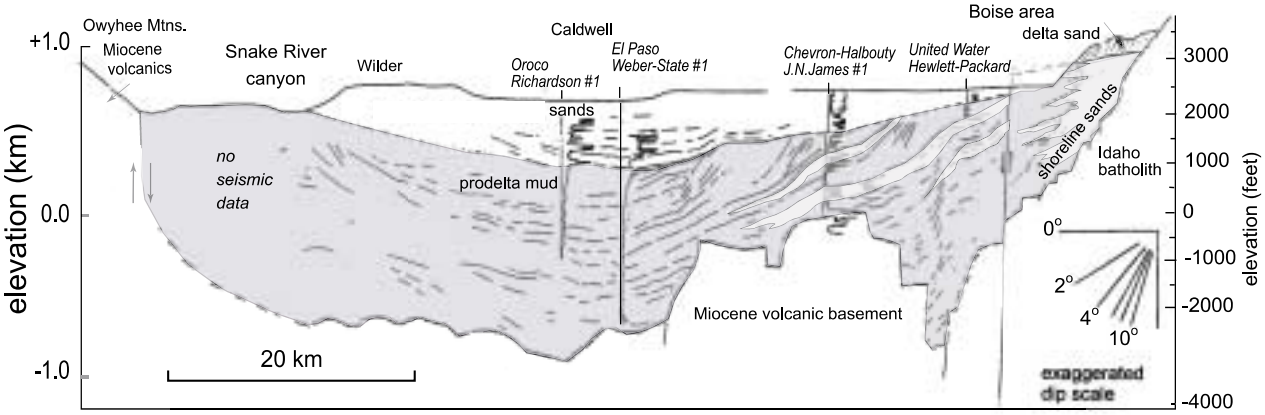

Figure 16. Tracing from seismic reflection section across the western Snake River Plain showing the upper sequence of fluvial-deltaic sediments and character of resistivity logs in wells. Resistivity-log excursions to the right are fresh-water aquifer sands, and the monotonous, near zero, resistivity values are indicative of mudstones. Notice the abrupt change upward from pro-delta muds to the sandy section of the delta systems, delta plain, and fluvial sands in the Caldwell area. Deposition may have been continuous in the center of the plain, but beds appear truncated by erosion to the east, and this surface to the east may be the unconformity at the top of the Chalk Hills Formation. The deltaic deposits in the Richardson well are discussed by Wood (1994). Line of section extends from Terteling Springs in Stewart Gulch shown in Figure 10 westward through the J.N. James well to the Caldwell area shown on Figure 2.

Figure 17. Photograph looking east and upriver of Cobb Rapids of the Snake River near Weiser. Cobb Rapids is the knickpoint of the river profile between the low gradient plain and the high gradient Hell's Canyon reach.

from the plain (1.6 feet/mile, 0.0003) and descends at steeper gradient (4.9 feet/mile, 0.0009) and a convex profile into Hell's Canyon (Figure 15).

LATE PLIOCENE AND QUATERNARY VOLCANISM IN THE WESTERN PLAIN

The resumption of basalt volcanism in the western plain region began around 2.2 Ma. At the western side of the plain, Lees (1994) obtained $^{40}Ar/^{39}Ar$ dates and subdivided the late Cenozoic basalts into two groups. On the older group of olivine basalts, shown by Walker and MacLeod (1991) as Pliocene and Miocene, Hooper and others (2001b) obtained dates ranging from 8 to 13 Ma, which they called the Sourdough and Keeney sequences. The younger group with dates of 1.9 to 0.8 Ma, they named the Kivett sequence. This basalt chronology of Hooper and others (2001b) is roughly consistent with that of Bonnichsen and others (1997) who show a 9-7 Ma basalt episode in the western plain, followed by a 5-Ma-long hiatus in basalt volcanism and then a second episode from about 2.2 Ma to as recently as 100,000 years ago.

Hooper and others (2001b) also dated Malheur Butte, a volcanic neck that is higher (elevation 2,661 feet, 811 m) than Lake Idaho sediment in the surrounding hills (elevation 2,300-2,550 feet, 670-780 m). Fine-grained, blue-gray andesite (about 2 percent plagioclase) yielded a whole-rock $^{40}Ar/^{39}Ar$ date of 0.8 ± 0.7 Ma. We have determined that the andesite is intrusive into an older bentonitic claystone, but we cannot determine from field relationships if it intrudes into or is overlain by the Glenns Ferry Formation. Lees (1994) suspected that flows of Malheur Butte lay buried beneath the lake sediment, but we find it is clearly intrusive into uplifted older sediment.

Basalt volcanism in the western plain beginning about 2.2 Ma erupted a volume of about 300 cubic km (see basalt-isopach map by Whitehead, 1992) from a group of vents with an alignment that is oblique to the orientation of the plain and its boundary faults (Figure 18). We call this basalt field of the western plain the Kuna-Mountain Home volcanic-rift zone. Although small in volume compared to the total late Neogene and Quaternary basalt of the eastern plain (estimated to be 40,000 cubic km

by Kuntz, 1992, p. 231), the western plain basalt field shares characteristic forms of basalt fields of the eastern plain. The vents are marked by numerous shield volcanoes distributed from near Mountain Home to the Kuna-Melba area. Other vents of this age also occur along the north margin of the plain near Boise (Othberg, 1994; Othberg and others, 1995) and within the Idaho batholith north of the plain (Howard and Shervais, 1982; Vetter and Shervais, 1992). Some intracanyon flows from these vents traveled down stream valley, up to 75 km distance, and spilled onto the plain, but their total volume is substantially less than the vents within the plain, probably less than 10 cubic km. A number of basalt vents erupted into the declining stages of Lake Idaho along the south side of the plain (Godchaux and others, 1992), and we used ages on Walters Butte and Pickles Butte to help date the level of the lake in Figure 9. Bonnichsen (written commun., 2000) has pointed out to us that some vents lie off the main line of shields, but we believe the bulk of the volume erupted from the zone shown in Figures 19.

Basalt eruptions from the volcanic rift zone constructed an upland of coalescing large shield volcanoes, generally higher than 3,100 feet (945 m) in elevation, that has confined the Snake River to a southerly course through the plain. Big Foot Butte shield is fully 8 km in diameter, and elevation at the top is 3,535 feet (1,078 m). The Initial Point shield is 10 km in diameter and has built up to an elevation of 3,240 feet (988 m). Most of the lava field erupted upon a surface that is 2,400 to 2,500 feet (730-760 m) in elevation (Wood, 1997). In the area of Little Joe Butte, Whitehead (1992) shows Quaternary basalt down to 1,300 feet (400 m) in elevation, where the basalt is determined to be 1,500 feet (460 m) thick on the basis of resistivity soundings. Two water wells drilled the basalt column down to elevation 1,880 feet (573 m), where they are still in the basalt section (Cinder Butte Farms Well, sec. 27, T. 2 S., R. 4 E.; and Big "D," Inc. Well, sec. 28, T. 2 S., R. 4. E.). The deep basalt in this area appears confined to a zone about 10 km wide over the main vent area and may be partly composed of dikes and sills; however, the driller of the Cinder Cone Butte well described red cinders at elevation 2,048 feet (624 m), and the driller of the Sabin well (sec. 25) described subsurface red lava at the 2,108-foot (647-m) elevation. The red color suggests these are subaerially erupted basalt cinders. In most other wells, basalt rests upon gravel and sand sediment at and above elevation 2,050 feet (625 m). The deep basalts here are poorly understood, particularly the low elevation of the sediment surface upon which most of basalt rests in this local area (i.e., about 2,050 feet, 625 m). We do not believe that the lake deposits of the western plain were incised to that depth before the eruption of basalt lavas. The present Snake River elevation just southeast of this area is 2,340 feet (713 m). Probably, this area of low-elevation basalt-sediment contact has experienced relative downfaulting associated with the basalt volcanism.

An ancestral canyon of the Snake River (filled by basalt) trends northwestward in the subsurface from Swan Falls to near Melba and Bowmont, a feature originally identified by Malde (1987) and called the "Canyon 3 Stage." Malde (1991) estimates the floor of the canyon at elevation 2,150 feet (655 m) on the basis of well data

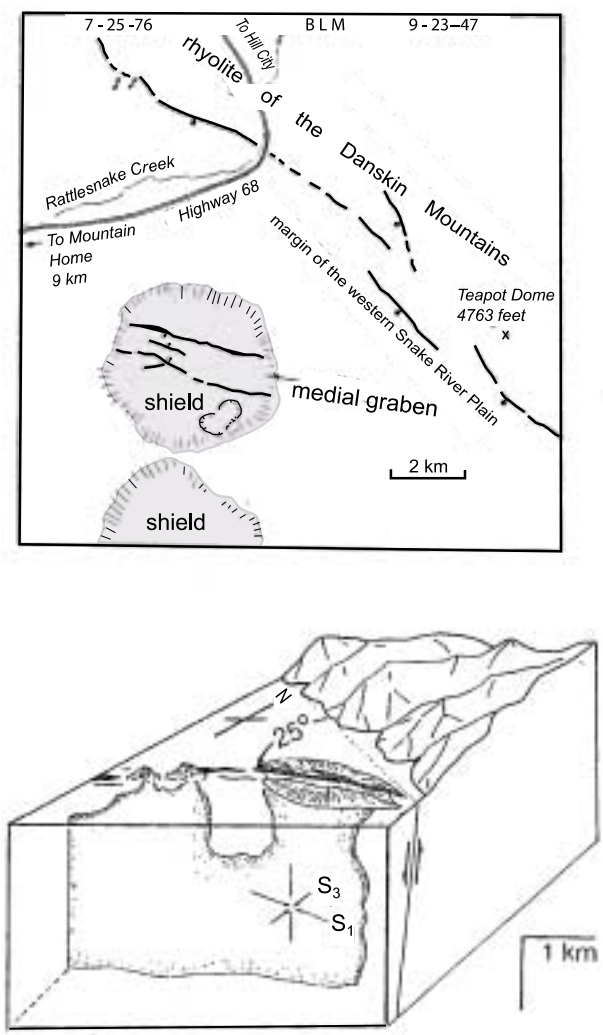

Figure 18. The air-photo tracing shows the northeast margin of the western plain 9 km northeast of Mountain Home. Tracing shows the N. 70° W. trend of fissures in the Pleistocene shield volcano and the N. 45° W. trend of the faults on the margins of the plain. Block diagram shows the inferred direction of principal stress-controlling feeder dike systems for the Kuna-Mountain Home volcanic rift zone. S_1 is the maximum principal stress, and S_3 is the minimum principal stress.

in the Melba area. He believed the canyon went through to the Boise River drainage to the north. We are unable to find the well data cited by Malde. Wood (1981, unpub. mapping) found water wells that drilled basalt down to 2,330 feet (710 m) in elevation near Nampa and U.S. Highway 30, but not down to 2,150 feet (655 m). There is evidence for the incision of the older lake deposits and infilling intracanyon basalt down to 2,330 feet (710 m) in elevation, but not to 2,150 feet (655 m) as suggested by Malde (1991). Determining elevations of the base of gravels overlain by subsurface intracanyon basalt is important to understanding the evolution of the Quaternary deposits and the history of incision of lake deposits (Othberg, 1994). The river at Swan Falls is now at 2,285 feet (697 m) in elevation. It is unlikely the river previously had a lower base level. Base level is established by the bedrock knickpoint at Cobb Rapids (elevation 2,080 feet, 634 m) 100 miles (160 km) downstream (Figures 14 and 15). Incision below 2,300 feet at Swan Falls would require the eroding stream to have had a lesser gradient than now (1.7 feet/mile, 0.0003), or that the Cobb Rapids area has been significantly uplifted in the Quaternary, neither of which seems likely. An elevation range of 2,300 to 2,400 feet (700-730 m) appears to be the deepest level of Quaternary incision in the Swan Falls-Nampa area of the plain.

The western plain basalt field has characteristics identical to the individual volcanic rift zones of the eastern plain, as defined by Kuntz (1992) to be linear arrays of basalt volcanic landforms and structures. The landforms include fissures, spatter ramparts, tephra cones, lava cones, shield volcanoes, and dikes at depth. The structures include noneruptive fissures, faults, and small grabens. The term "rift" needs clarification because we regard the western plain as a tectonic or continental rift, but we define here a volcanic rift zone that cuts obliquely across the western plain. A tectonic rift is a large graben structure not necessarily associated with volcanism. Kuntz (1992) shows that volcanic rift zone alignments are collinear with the strike orientation of active basin-range normal faults on both sides of the eastern plain. The length of the Kuna-Mountain Home volcanic rift zone is about 100 km, similar to typical 80-km lengths of zones in the eastern plain.

Fissures and faults within shield volcanoes of the western plain have the same alignment as the volcanic vents, about N. 70º W., indicating structural control of magma vents related to the regional tectonic stress system. This alignment is about 25 degrees counterclockwise from the bounding fault systems of the western plain (Figure 18). Nakamura and Uyeda (1980), Zoback and Zoback (1991), and Conner and Conway (2000) show such features are reliable indicators of tectonic stress direction whereby the volcano alignment is perpendicular to the least principal stress direction.

Remarkably, the alignments within volcanic rift zones change from the eastern plain to the western plain. Eastern plain zones are mostly aligned N. 30º W., whereas the one zone in the western plain is N. 70º W., and this change in orientation occurs over a distance of 80 km from the Richfield-Burley Butte zone (Shoshone lava field) to the easternmost group of vents near Mountain Home. That general trend N. 70º W. is also characteristic of eastern Oregon Quaternary basalt fields (aligned vents and fissures) and to the distribution of Quaternary basalt fields across the state westward to the Cascade Range (Figure 19). These Quaternary volcanic trends may indicate a province of relatively uniform stress orientation east of the Cascades and including the western plain, that is a different province from the eastern plain region.

CONCLUSIONS

The western Snake River Plain is a normal-fault bounded basin, 70-km wide and 300-km long. The amount of extension that formed the western-plain sedimentary basin is about 10 percent, similar to intracontinental rift basins elsewhere in the world such as Lake Baikal and those in east Africa. Exposed strata on the margins and seismic reflection data show as much as 2 km of basin relief formed by both faulting and by downwarping toward the basin axis. Fault structures are both half and full grabens. The western plain structure contrasts greatly with the northeast-trending eastern plain where extension is not expressed by faulting at the margins but solely by downwarping associated with basalt intrusion in fissure systems oriented perpendicular to the axis of the eastern plain.

Previously published seismic refraction data show that the intermediate and deep crust beneath the western plain is mafic rock, whereas the margins of the plain are granitic rock. The basin sediments are underlain first by basalt lavas which are underlain by middle crustal rock so invaded by mafic intrusives that the original granite is no longer recognizable by seismic refraction-derived velocities. This indicates the plain is not a simple graben of downfaulted granitic crust. The data also show a high velocity layer at a depth of 23 km restricted to the area beneath the Bruneau-Jarbidge eruptive center southeast of the plain. This layer could be restite or an underplate of basalt related to the formation of silicic magma.

The northwest orientation of the western plain basin appears to follow a pre-existing lithosphere weaknesses or megafabrics of the northwestern United States. This

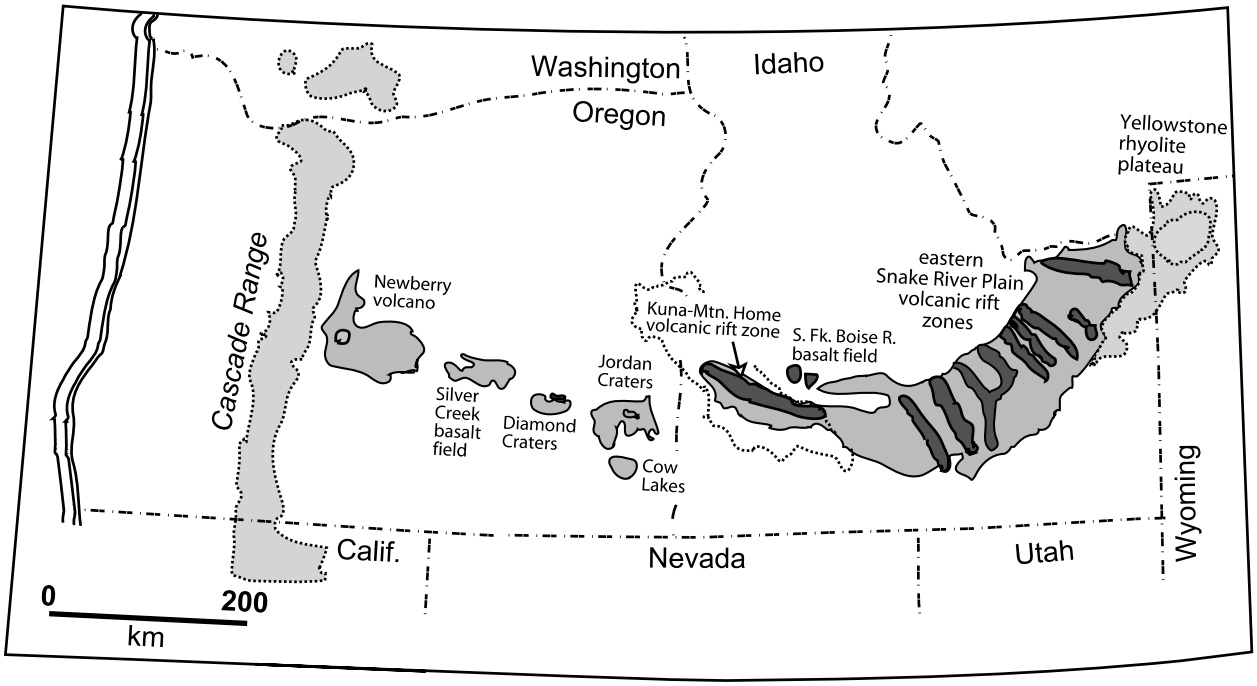

Figure 19. Map showing Quaternary basalt fields of eastern Oregon and southwest Idaho. Darkest shading is the basalt vent area. Note the contrast in the orientation of the volcanic rift zones between the eastern and the western plain. Orientation of the Kuna-Mountain Home volcanic rift zone of the western plain parallels those of eastern Oregon basalt fields, suggesting a similar orientation of tectonic stress direction in the Quaternary but a direction that differs markedly from the eastern Snake River Plain.

orientation is expressed in other structures such as the Olympic-Wallowa lineament, the Brothers fault zone, and the Baker-La Grande grabens.

The geologic and tectonic history of the western Snake River Plain is intimately involved with magmatism as shown by the following sequence of events and features:

(1) Before the basin formed, widespread and voluminous rhyolite volcanism occurred south of the plain, and the early Columbia River and Steens basalt erupted over Mesozoic rocks north and south of the plain, 17-13 Ma. Local eruptions of rhyolite occurred along the fault margins of the plain as the basin began to form about 12-11 Ma.

(2) Timing of the beginning of the basin coincides with the lodging of the hot spot at the Bruneau-Jarbidge rhyolite eruptive center, 11.5-8 Ma, southeast of the western plain. We suggest that heating of the lithosphere initiated the structure and also caused basalt magma intrusion as the basin extended.

(3) Deep wells in the center of the western plain drilled through sediments to basalt flows and basalt tuff exceeding 2.5 km in thickness, and have not drilled through rhyolite. Rhyolite sequences without significant sedimentary interbeds, as much as 2-km thick, occur on the margins of the plain. Lack of rhyolite in deep drill holes suggests that much of the plain area was an upland during the early eruptions of rhyolite from the Bruneau-Jarbidge center.

(4) Normal faulting that formed the western plain basin began about 11 Ma and amounted to over 2 km of offset by 9 Ma. Since that time, long-term average rates of vertical slip have been low (<0.01 mm/year), except for an active fault segment 55 km southwest of Mountain Home. Broad downwarping toward the center of the basin and compaction subsidence of the thick sedimentary fill have since lowered the center of the basin about 0.3 km with respect to the margins.

(5) Earliest sedimentation in the western plain is accompanied by eruptions of local basalt fields dated 10 to 7 Ma. The sediments are mapped as the Chalk Hills Formation. We find that the Poison Creek Formation of arkosic sand and the Banbury Basalt are local features and should be considered as facies and local basalt fields within the Chalk Hills Formation.

(6) The lake systems that deposited the Chalk Hills Formation declined in lake level at some time between 6 and 4 Ma, resulting in erosion of the Chalk Hills Formation from the margins of the lake. Reasons for the decline in lake level are not known.

(7) On the basis of fault history, we assume that elevations of lake-level features preserved as shoreline fea-

tures on the margins of the lake have changed little in the past 6 million years. This assumption allows reconstruction of the history of Lake Idaho as the basin refilled, overflowed, and the outlet was downcut.

(8) Between 6 and 4 Ma, lake levels rose and deposited a transgressive sequence over an unconformity surface on the Chalk Hills Formation. The shoreline transgressive sequence is identified by pebbly sands and oolitic-sand deposits in its upper part. We interpret the oolite occurrence as a product of increasing alkalinity in the lake water, as lake levels rose in a closed basin. We speculate that the lake-level rise was caused by captured drainages associated with the eastward-migrating hot-spot uplift.

(9) The lake overtopped a spill point into ancestral Hell's Canyon at a point between Huntington and Weiser. The spill point appears to be near Dead Indian Ridge at about elevation 3,600 feet (1,100 m). The time when the lake began to drain is poorly constrained, but we speculate it was about 4 Ma.

(10) Overtopping of the spill point connected the Snake River to the Columbia-Salmon River system and to the sea. Recent work by Van Tassel and others (2001) suggests the connection was established between 3.8 and 2 Ma. An implication is that once the lake began to drain, the alkalinity of lake water should have decreased, and this perhaps explains the lack of calcareous muds in the uppermost part of the Lake Idaho stratigraphic section.

(11) In the subsurface of the northwestern part of the western plain, an abrubt transition occurs upward in the section from thick mudstone to overlying deltaic sands. We interpret these sands to be the result of lowering lake level, erosion of lake-margin sands, and delivery of sand to the unfilled basin. The resulting upper section of deltaic sands and fluvial deposits is as much as 400 m thick, and these interbedded sands and muds constitute the main freshwater aquifer section of the western plain.

(12) Ages and elevations of basalt deposited over flood-plain sediment and gravel near the margins of the western plain indicate erosion of the outlet resulting in lake-level decline at an average rate of 120 m/Ma over the past 4 Ma.

(13) The process of filling of the remaining lake basin from southeast to the northwest outlet implies that flood-plain and stream-channel deposits should become younger to the northwest. The lake basin filled with sediment by about 1.6 Ma, and braided stream systems depositing cobble gravel flowed to the outlet.

(14) The resumption of basalt volcanism in the western plain began about 2.2 Ma. Much of the volcanism is expressed as a field of shield volcanoes that trends obliquely across the plain from Mountain Home to Kuna with an orientation of N. 70º W. Fissures and faults within the shields have similar orientations, suggesting the vents are controlled by a regional stress system such that the maximum principal stress is similarly oriented and that the least principal stress is aligned perpendicularly at N. 20º E.

(19) Late Quaternary fault activity occurred on a segment of the southeastern boundary fault system, 55 km southwest of Mountain Home, that trends N. 30º W. to N. 70ºW. The N. 70º W. orientation of some active faults is parallel to the Quaternary Kuna-Mountain Home volcanic rift zone, which suggests that episodic reactivation may occur on faults oriented suitably with respect to the present system of tectonic stress.

Examining the history of the western Snake River Plain brings up a number of questions for future research. Is the basalt crust beneath the western plain a product of early injection of basalt magma, similar to processes thought to be presently occurring in the eastern plain? Is the deep-crustal high-velocity zone beneath the Bruneau-Jarbidge center related to the rhyolite volcanism? These two questions might be answered by deep-crustal seismic exploration and a reevaluation of gravity data. Additionally, such a project might give new data on the way in which hot-spot magmatism has altered the lithosphere.

Other aspects of the western plain that should be a focus of research are as follows: The paleotopography and implied uplift of the region during eruption of rhyolite might be reconstructed if flow directions of rhyolite were mapped. Chronology of the "Chalk Hills" lake is poorly constrained, particularly the time of the lake level decline. Better stratigraphic study and chronology are needed of this formation to understand the processes that affected the lake. Further, the chronology of the lake level rise and overtopping of the spill point is poorly known. Detailed stratigraphy and geochronology combined with an interpretation of lacustrine-sediment geochemistry may significantly contribute to knowledge of the history of Lake Idaho.

We have suggested that the principal tectonic stress orientation indicated by Quaternary basalt vents and active faulting is about N. 70º W., and that a similar orientation is expressed by basalt fields in eastern Oregon. More detailed study of the vents throughout the region is needed to verify this. Also borehole breakouts as indicators of tectonic-stress orientation in future deep wells should be studied with compass-oriented borehole imaging or 4-arm caliper logs.

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