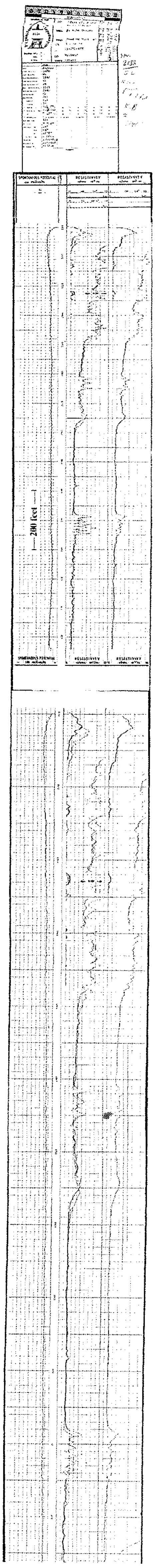
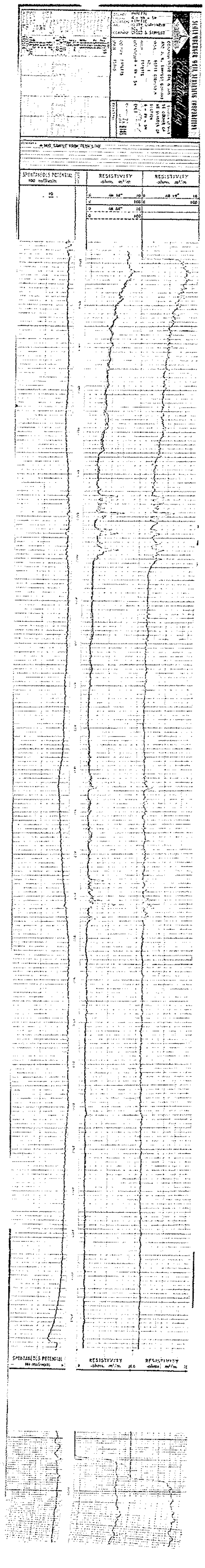
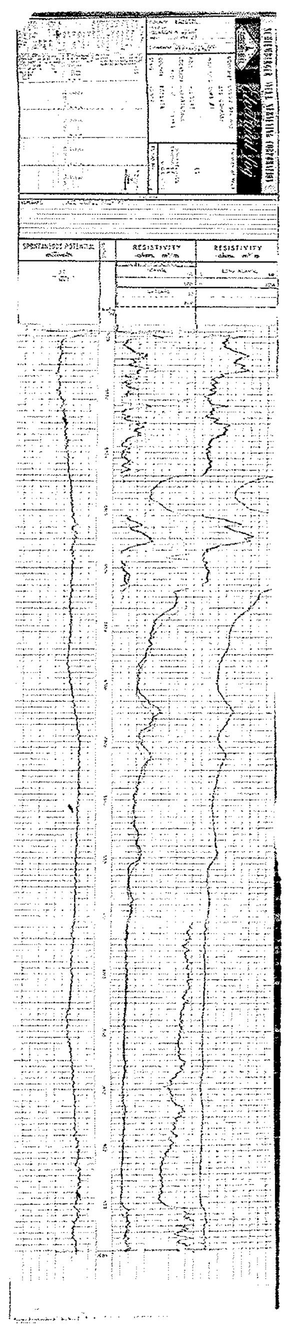
RB1



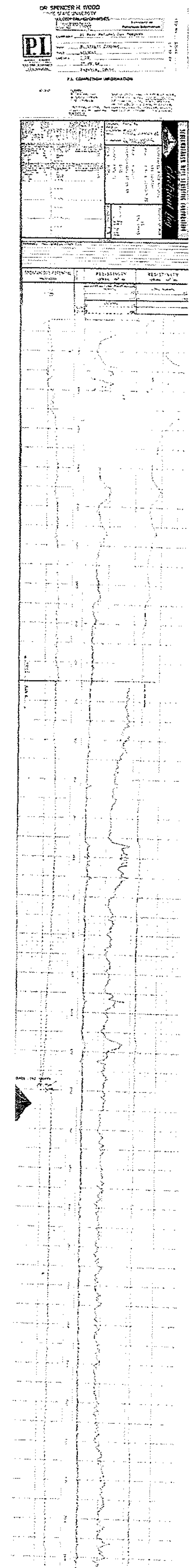
BC1



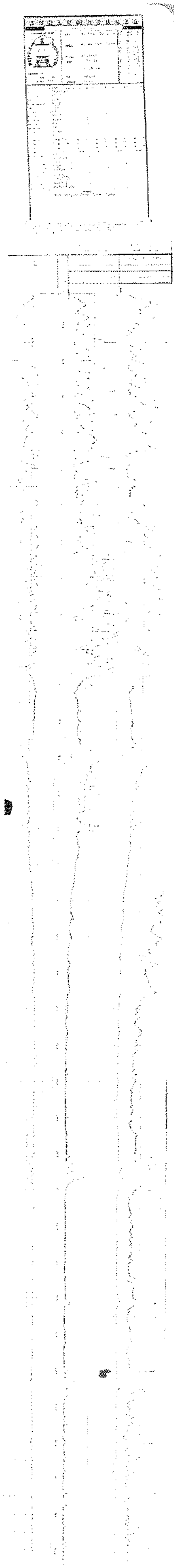
TD1



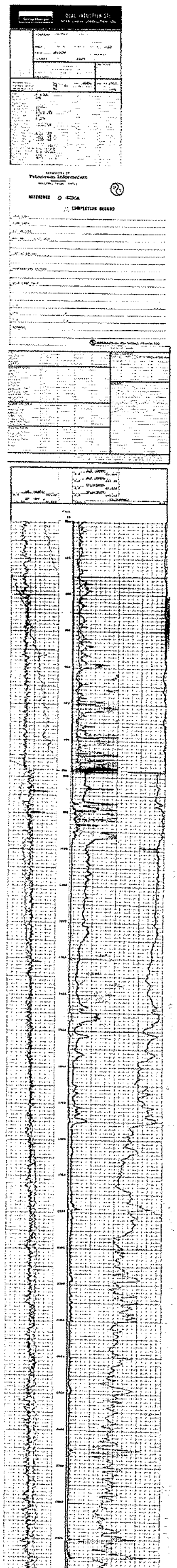
VJ1



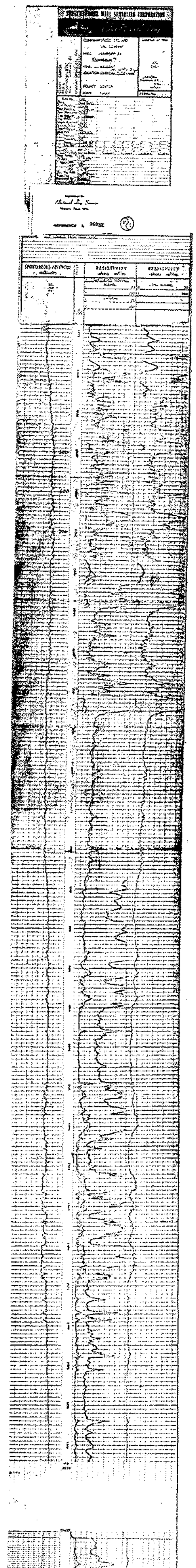
WS1



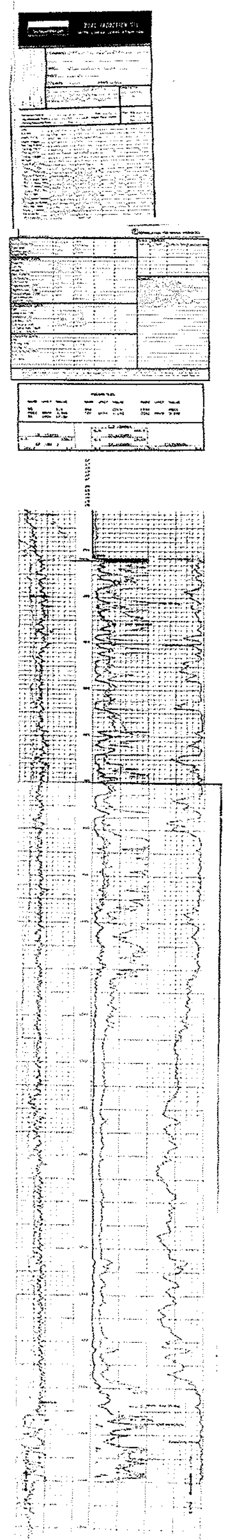
HL



R



DF



Qualitative Interpretation of Magnetic Anomalies and Progress Report on Geologic Mapping in the Foothills North of Eagle, Ada and Gem Counties, Idaho

Report prepared for HydroLogic, Inc. (May 20, 2007)

by

Spencer H. Wood
421 E. Crestline Dr.
Boise, ID 83702
swood@boisestate.edu

MAGNETOMETER SURVEY

Summary: The survey clearly identifies a down-to-SW fault along Willow Creek Road (Fig.1), and width of the anomaly suggests the sediment section over volcanics is at least 3,000 feet deep (a surprisingly deep estimate, which if important, needs verification by more detailed survey and interpretation). The fault aligns with the NW extension of the previously identified West Boise – Eagle fault. Another major down-to-SW fault is identified near the mouth of Big Gulch, and crude estimate of thickness of sediment section over faulted volcanics is 2,300 to 4,600. Better definition of faulting on the M3 properties could be obtained by a long magnetometer line (14 km), along ridgetop roads between Willow Creek and Big Gulch, with close supervision of data quality.

Measurements: Data was acquired under the supervision of Dr. Paul Donaldson, mostly by graduate student Carlyle Miller.

The first data set was acquired in about June, 2006 as lines along Little Gulch and Big Gulch. The data set is generally of good quality with a noise level of ± 5 nanoTeslas (nT).

A second set of data was acquired later in the summer of 2006 along Willow Creek and Chaparral Road ("north line"), along the Farmers Union Canal ("south line"), an extension of the Big Gulch Line to the southwest, and a short line near Highway 16 between Big Gulch and Willow Creek (Figs. 2 and 3). This data set is quite noisy with a level of ± 15 nT, and many unexplained excursions exceeding 80 nT. It is likely that the magnetometer malfunctioned intermittently. The instrument was subsequently sent to Geometrics for repair by Lee Liberty, and a number of problems were repaired. However, the majority of measurements cluster along a trend, so that the larger anomalies were measured. Nevertheless, it is recommended that anomalies detected by the second data set be verified by running an additional line between Willow Creek and Big Gulch, along the ridge top roads through the northern M3 properties and as far SW and NE, as is possible on other lands. The fluctuation in altitude along such a line, greater than 20 m along such should be noted, but it is not likely to significantly affect the measurement of anomalies arising from buried volcanics over 300 m deep (see for example Figs. 5 and 6), which shows that elevation (depth) difference from 300 to 400 m causes a diminishment of 80 to 60 nT, or 20 nT. These anomalies also have an associated long wavelength, greater than 600 m, and should be easily distinguished from elevation effects. However, for shallow anomalies, shallower than 300 m depth, the effect is greater, and elevation effect over a few hundred meters distance could be mistaken for a shallow anomaly if not noted or corrected.

The instrument used is a Geometrics G-858 cesium magnetometer with two sensors spaced one meter apart, on a vertical staff, the lower sensor about 1 meter off of the ground. This instrument measures the total magnetic field of the earth on each sensor to a precision of 1 nanoTesla (nT. The field values ranged from 53,600 to 53,900 nT. The survey was made by walking and continuously recording. Locations were obtained by simultaneous recording from a GPS unit with an accuracy of ± 8 m. No base stations measurements were made, but tie lines on the Big Gulch and Little Gulch lines gave agreement measurements, indicating no drift or large secular variation in the first set of data. It is possible that gradual diurnal variation of up to 60 nT may occur over a period of 12 hours. I was not provided with detailed field notes showing times of measurement – so one cannot evaluate that source of error – when splicing data together taken at very different times of day.

Source of magnetic anomalies: Magnetic anomalies arise from faulted rock sequences or topographic variation in rocks having a substantial magnetic susceptibility (k), i.e. greater than 0.0001 SI units. (note: SI values are 4π times cgs units, and both are dimensionless). Anomalies may also arise from spatial variation in k of rocks. Basaltic volcanic rocks range from 0.001 to 0.05, dependent largely upon the percent and grain size of the mineral magnetite. For example, a rock with 7 per cent magnetite will have a volume susceptibility of 0.038 SI units (Breiner, 1999). In addition to the magnetism induced in rocks by the prevailing magnetic field, basaltic rocks in particular may have a frozen remnant magnetism set by crystallization of magnetite when the lava cooled. Polarity of the frozen magnetism may be of different or opposite polarity from the induced magnetism, on account of reversed polarity of the earth's magnetic field at the time of crystallization. Remnant effects are difficult to predict without detailed sampling and modeling. For this qualitative interpretation, it is assumed that anomalous magnetism is entirely caused by induced magnetism of the present earth's field.

Qualitative interpretation of anomalies of this survey is based on an assumption that most of the magnetic features are faulted volcanic rocks beneath the Idaho Group sediments. Mapping the in the Pearl area by Clemens (1993) indicates that above the granitic rocks is about 100 m thick section of basalt, locally overlain by as much as 120 m of rhyolite, which is in turn overlain by another 25 m of basalt. Overlying this volcanic section are the Idaho Group sediments which are at least 200 m thick over most of the area. The faulted basalt and rhyolite would produce magnetic anomalies. The deeper granite may produce broad anomalies of longer wavelength and lower amplitude than the volcanics. Very low k of sediments indicate they to not contribute to the field. A reasonable model is 100 to 200 meter thick slab of volcanic rock, overlain by sediments and faulted by normal faults.

Models of faulted volcanic rocks at depth:

Analytical models of the total-field magnetic anomalies for a 2-dimensional faulted slab are published int Telford et. al. (1990). The model is for a cross section of a plate of thickness, t, extending infinitely in the 3rd dimension. The formula on p. 100, Eqn 3.59b was calculated to visualize the effect of depth, thickness and susceptibility, for offset volcanic rocks. Figures 5 and 6 show a simplified version, of just the upthrown block, assuming an offset of 100 m, of a 100 meter thick section of volcanic rocks of susceptibility of 0.003 SI.

$F(x) = +2\,k\,t\,F_e \{ 1/(d^2 + x^2)\} \{d \sin 2I \sin\beta - x(\cos^2 I \sin^2\beta\ - \sin^2 I)\}$.

In terms for EXCEL:

F(x) = + 2*(*k*)*(*t*)****Fe****((1/((*d*^2)+A5^2))*(*d**(SIN(2*RADIANS(*I*)))* (SIN(RADIANS(***strike***))) - A5*(((COS(RADIANS(*I*)))^2)*((SIN(RADIANS(***strike***)))^2) - ((SIN(RADIANS(*I*)))^2)))).

Where
k = magnetic susceptibility, SI units
t = thickness of slab of magnetic rock, in meters.

F(x) = total field measured by the magnetometer.
F_e =Approximate total field of the earth in the area, which induces the anomalous field in susceptible rocks. i.e., the background value, estimated to ± 1000 nT (used a value of 54,000 nT) for model calculations,

d = depth to top of slab
x = horizontal distance from fault edge (parameter A5 in EXCEL code above) (assumed vertical fault plane)
I = inclination of the earth's magnetic field, which for this area is 60 degrees.
β = strike angle between magnetic north, and the strike of the fault. For a fault with strike of N45W, and down to SW, use 45 degrees. For a fault N45W and down to NE use 225 degrees. Telford et. al. (1990) are not clear on their conventions, and to make their formula reproduce their Fig. 3.22b on page 103, I had to change the sign of the equation above from – to a +.

In all formulas for 2D bodies of uniform cross section, the amplitude varies as $2\,k\,t\,F_e$, so that for any given fault depth and offset, at a given orientation to the earth's magnetic north, the anomaly varies linearly with the thickness of the magnetic slab (t) and the susceptibility (k).

Figures 5 and 6 show results of calculation for NW-SE trending faults with vertical planes. The model is appropriate, since faults in the area generally trend NW-SE. The anomaly shape is the same, and changes sign, if faulting sense is opposite (i.e., up to southwest, or up to northeast)

Figure 5 shows that a down-to-SW fault, with 100 m of displacement on the volcanic section will have a positive bulge in the field of 100 to120 nT, if top of upthrown block is 200-m deep, the width of anomaly will be about 600 m. If top of the block is 600-m deep, the anomaly will diminish to 20 to 40 nT, and width will be about 1200 m.

Figure 6, is the anomaly of a down-to-NE fault, and is just the negative value of Figure 4. This calculation for a 700-m deep anomaly show a width of about 1700 m, and an anomaly of 20 to 40 nT for the 100 m thick volcanic section, offset 100 m.

To re-iterate again, the strength of the modeled anomaly is linearly related to the magnetic susceptibility and thickness of the section of volcanic rocks, and diminishes with depth as 1/d.

Specialized and costly software exists for more exact modeling of anomalies (QUICKMAG-PRO $2850 from Rockware), however it was not immediately available for this study. However, I just became aware of a freeware GEOMODEL, also from Rockware, that allows 2.5 D modeling, and that will be done subsequent to this report.

Simple estimates of depth to anomalous features: Because distance breadth of the anomaly increases with increasing depth of the anomalous feature, the width is an indicator of depth to rock feature or susceptibility giving rise to the change in the earth's field. A rule of thumb is that depth is about ½ the width of the anomaly. For example, if an anomaly is 1200 m wide, the depth is about 600 m. Such depth estimates are crude and may be off by 50 per cent (Breiner, 1999).

Another simple measure is to measure the map distance, x_z, of the straight portion of the slope of one of the limbs of an anomalous bulge or depression in field values. "Straight portion" is the distance between the inflection point (i.e.curvature changes from concave upward to convex upward, for positive anomalies). In this estimate, the depth, z, is simply calculated from:

$$z = C\ x_z$$

where C varies from 0.5 to 1.5 (Breiner, 1999).

Discussion of observed anomalies along survey lines. All survey lines trend approximately SW-NE, so that they will yield profiles perpendicular to strike of SW-NW trending faults, believed to be the dominant structures in the area (Fig. 1). Their location will be discussed in terms of UTM km easting coordinate for easy reference on profiles and map. Because the lines are oriented mostly SW-NE, the easting km coordinate separation is multiplied by 1.4 to determine the width of anomaly along the SW-NE oriented survey line. Anomalies are labeled by circled letters on Figure 1, and the following sections discuss the profile lines in terms of those lettered anomalies.

North Line (Chaparral Road – Willow Creek Valley), 9 km long

A: A +20 nT positive anomaly, with a (600-m x 1.4) = 840-m width can be discerned from the rather noisy data at the NE end of the line, centered at 546.15 east (Fig. 2). Anomaly is not well defined, and large excursions of 100 nT in the data are probably instrument malfunction – indicating the data should be re-run if this anomaly is of interest.

B: Field values rise to the SW to point 544.0, 30 to 40 nT, and then fall off sharply to the SE to point **C**. Interpretation of point B is uncertain. Elevation drop of 33 ft between point is not enough to explain rise in values.

C: Most profound anomaly along this 9-km line is a negative anomaly centered at UTM km 543.0, of -60 to -80 nT, and a width of (1.2 km x 1.4) = 1700 m. The negative value indicates a down-to-SE fault, with more than 100 m displacement, and at an estimated depth to volcanic section on the upthrown block of (½ x 1700 m) = 850 m (2,800 ft).

D: Values rise irregularly from edge of the point **C** anomaly to the end of the line, 60 nT, over a distance of 2.5 km. Data is so noisy (±20 nT) that shape of any individual anomalies, if they exist, cannot be discerned.

Big Gulch Line and the SW extension, 12-km long.

E: A positive +50 nT anomaly, centered at 550.5 east, has a width of (1.4 x 3 km) = 4.2 km (Fig. 2). The same anomaly is larger on the Little Gulch line (Fig. 3), and the tie line (Fig.4), where its magnitude is + 100 to + 180 nT, and breadth is 3 to 4 km. The positive value and width suggest a down-to-SW faulted volcanic section, perhaps 1.5 to 2 km (4,800 to 6,000 ft) deep. Magnitude of the anomaly suggests large displacement (>200 m). The depth estimate seems large, and it is possible that the diminishing NE limb may be a thinning of the volcanic section, and the anomaly width due to faulting is less.

F: A well defined low of -20 nT, with a width of (1.4 x 2 km) = 2800 m. Possibly a down to NE fault at depth, but associated with anomaly at E, may indicate complicated geometry of faulted volcanics.

G. Slight rise and decline with amplitude of about 20 nT, and width of less than (1 km x 1.4) = 1.4 km may indicate a down to SW fault of small displacement (< 100 m).

H: A + 80 nT anomaly centered at 544.5 east with a width of at least (1.4 x 2 km) = 2.8 km, suggests a large down to SW fault, with top of upthrown block about 1.4 km (4,600 ft) deep; however, the SW definition is very noisy due to malfunction of the magnetometer when that SE extension data of this line was obtained (i.e., data SW of 545.0 east). The "south line" point **M,** indicates a 2,300 to 3,300 ft sediment section over the volcanics here, but in both cases the data is very noisy, and needs to be re-run for more precise estimates.

Little Gulch line

I: The largest and best defined anomaly of the survey is this +100 nT, (1.4x 2km) 2.8 km wide positive anomaly, centered at 550.8 east (Fig 3). The positive value suggests a large-displacement (>200 m), down-to-SW fault, with top of faulted volcanics on the upthrown block of about 1.4-km (4,600-ft depth. As discussed above under point E, depth estimate seems much larger than expected, and none of the lines fully defines the full NE limb of the anomaly. NE extension of line could allow better modeling, and also tie to the known volcanic section in upper Little Gulch area drilled by Conolley's wells about 1000 ft deep, 4 km west of here, and the exposed volcanic contact about 4 km to the NE (Fig. 1).

J: A negative anomaly of about -20 nT and a width of (1.2 x 1.2 km) = 1.4 km is centered at 547.3 east, suggesting a small displacement fault at a depth of 700 m (2,300 feet).

K: The rise from anomaly J, at end of line, may be the NE limb of the larger positive anomaly detected by the SW end of the "south line". That SW continuation is shown on this profile, and defines a + 80 nT anomaly centered at about 544.8 east.

South Line (Farmers' Union Canal)

L: Values rise at the east end of the line at 551.7, and are likely the west limb of the anomaly I on the Little Gulch line (Fig 3).

M: Values rise continuously to the west along this line, and then jump up sharply at 547.2 (Fig. 3). The sharp jump may be a splice of data taken at different times, and a result of diurnal variation and of no geologic significance. The peaking of values at 545.5 suggest a large displacement, down-to-SW fault, shown by a positive anomaly of at least + 100 nT, and a width of (2 to 3 km x 0.7) = 1.4 to 2 km. Because this is an EW line, the width of the anomaly, measured perpendicular to a NW-SE strike fault, is reduced by multiplying by 0.7. This same feature is also defined on the Big Gulch line at point **H** (Fig.2) confirming its NW-SE trend, and together suggest a volcanic-section feature at least ½ x 1.4 to 2 km (2,300 to 3,300 ft) deep beneath the sedimentary section.

Willow Creek Road Line ("perpendicular tie line")

N: Data on this line is displayed on both the northing coordinate and the easting coordinate, to better define this large (+200 nT) anomaly centered at 550.4 east (Fig. 4). This is the same feature detected on the Little Gulch line at point **I** (Fig. 3), and probably also at lesser magnitude as point **E** on the Big Gulch line (Fig. 2), and defines a large down-to-SE fault with NW-SE strike.

GEOLOGIC MAPPING PROGRESS & DISCUSSION OF FAULTS

Mapped geologic features are compiled on Fig. 1. Thorough geologic mapping is shown in the northeast corner of the area from Clemens (1990). Mappable faults shown in Fig. 1 are from a number of unpublished mapping projects by S.H.Wood.

Oolitic sands: Occurrence of oolitic sands has been mapped as a part of this project, because they are one of the few cemented rock outcrops in the sedimentary section. Oolitic sands are carbonate coated grains of lake shore deposits, and their origin and significance is discussed in Swirydczuk et al (1980), Wood and Clemens (2002), Wood (2004), Wood and Squires (2007). Significant to this study is the observation that these carbonate-cemented sands occur over a stratigraphic interval, limited to a few hundred feet, in the upper part of the Terteling Springs Formation, on the north side of the western Snake River Plain. Therefore these outcrops are considered a rough geologic marker bed, and the elevation of their occurrence a rough indicator of the tectonic tilting or faulting across the area. Elevations of oolite outcrops are posted on the map (Fig. 1). In the SE Pearl area, their elevation is 3,100 to 3,200 feet. Along Willow Creek, near Lynn's Ranch they occur at 2,900 to 3,000 feet, and along the Old Freezeout Hill Road they occur at 2,700 to 2,750. From these elevations, it is concluded that the vertical-fault offset combined with tectonic tilt of the sedimentary section is about (3,200 - 2,700), 500 ft, from the Pearl area (NE corner of map) to Old Freezeout Hill Road.

Oolitic sands and the Terteling Springs Formation: The oolite deposits are also an identifying feature of sand facies of the Terteling Springs Formation, a facies that lies within several miles of the contact with the Idaho batholith, and grades basinward to the SW to the mudstone faceies (Wood and Clemens, 2004). Therefore, the thick sand section observed along the Old Freezeout Hill Road is Terteling Springs Formation, and is not the overlying Pierce Gulch Sand.

Pierce Gulch Sand: Apparently the Pierce Gulch Sand, with a significant thickness in the western part of the M3 properties, is in a downfaulted section exposed along the bluffs of the Payette River Valley, west of the Old Freezeout Hill Road. I believe the strata in the bluffs in Section 28, T2W, R6N is the Pierce Gulch Sand, because a characteristic white volcanic ash bed (locally 2-ft thick) occurs at elevation 2,610 ft within the dominantly coarse sand deposits. A similar volcanic ash bed occurs in the bluffs of sand sediment along the north side of the Payette River near Birding Island, 18 miles to the northwest. However mapping of the section exposed along the Payette River is in the very preliminary stages, and these correlations are tentative.

Faults: Exposures of strata in the M3 area are scarce; however faint stratification lines are apparent on Google-earth imagery and BLM aerial photography. The faults indicated by the magnetometer survey do not have obvious surface expression, except for one small fault exposed in the roadcut along Willow Creek Road in the hill just north of Big Gulch (Fig.1). That fault is associated with the major down-to-SW fault along Willow Creek Road detected by the magnetometer survey (points **E, I, L,** and **N**), and also known in the southwestern part of the map area from drillers logs and geophysical logs (Fig. 1). Width of the magnetic anomaly suggests a thick sediment section over the volcanics here; however these depth estimates of 4,800 ft are crude. If one uses the formula $z = Cx_z$, and

$x_z = (1.4 \times 1.5 \text{ km}) = 2000$ m, and a constant for C of 0.5, the depth of sediment over volcanics computes to 1000 m (3,300 ft). The fault is a NW extension of the West Boise – Eagle fault system identified by Squires (1992).

It is puzzling that the down-to-SW fault exposed in the Highway 16 roadcut at Freezeout Hill (Fig. 1) does not have a strong expression on the magnetometer survey (as noted above, that data on the "north line" is very poor in quality on account of instrument malfunction, and another traverse is needed to understand this puzzle). Because the exposed fault shows strata that cannot be matched up in the 60 ft vertical exposure, the offset on this down-to-SW fault must exceed 60 ft. However I believe that the sands at this exposure are entirely the Terteling Springs Formation sand facies. The fault projects SE to the oolite bed exposures on the north side of Willow Creek, near Lynn's Ranch, but the oolite beds are not faulted; therefore, if it is a significant fault, it must project to the SW or to the NE of these beds.

The magnetometer survey of Big Gulch and the "south line" shows a major down-to-SE fault at the western M3 property, labeled points **H, K,** and **M** (Fig. 1). Thickness of sediment over the volcanic section is crudely estimated at 2,300 ft based on width of magnetic anomaly. This fault could be better defined by a survey along the ridgetop roads between Big Gulch and Willow Creek, west of Highway 16 (not M3 property).

The fault indicated by point **C** on the "north line" has an unexpected down-to-NE magnetic signature (negative anomaly) (Figs 1 and 2). Data is very noisy along this line, and the data between points **C** and **D**, while suggestive of complicated faulting, need to be repeated before a more confident interpretation can be made.

REFERENCES

Breiner, S., 1999. Applications manual for portable magnetometers. Geometrics, Inc., San Jose, California. 58 p.
internet:ftp://geom.**geometrics**.com/pub/mag/Literature/m-ampm-05Apr06.pdf

Clemens, D.M., 1993. Tectonics and silicic volcanic stratigraphy of the western Snake River Plain, southwestern Idaho. M.S. Thesis, Arizona State University, Tempe, 209 p., 10 plates.

Swirydczuk, K., Wilkinson, B.H., and Smith, G.R., 1980. The Pliocene Glenns Ferry oolite-II – Sedimentology of oolitic lacustrine terrace deposits: Journal of Sedimentary Petrology, v. 50, p. 1237-1248.

Telford, W.M., Geldart, L.P., and Sheriff, R.E., 1990. Applied Geophysics (2nd Edition). Cambridge University Press, Cambridge, UK, 770 p.

Wood, S.H., 2004. Geology across and under the western Snake River Plain, Idaho; Owyhee Mountains to the Boise Foothills. *in* Haller, K.M. and Wood, S.H., Geological field trips in southern Idaho, eastern Oregon, and northern Nevada. U.S. Geological Survey Open-File Report 2004-1222.
internet: http://pubs.usgs.gov/of/2004/1222/Ch7.pdf.

Wood, S.H., and Clemens, D.M., 2002.Geologic and tectonic history of the western Snake River Plain, Idaho and Oregon. *in* Bonnichsen, Bill.,White, C.M., and McCurry, M., eds., Tectonic and magmatic evolution of the Snake River Plain Volcanic Province; Idaho Geological Survey Bulletin 30, p. 69-103.

Wood, S.H., and Squires, Ed., 2007. Geology and Hydrogeology of Boise, Idaho (abstract). 88th Annual Meeting of the American Association for the Advancement of Science (Pacific Section). June 17-21, 2007, Boise, Idaho.

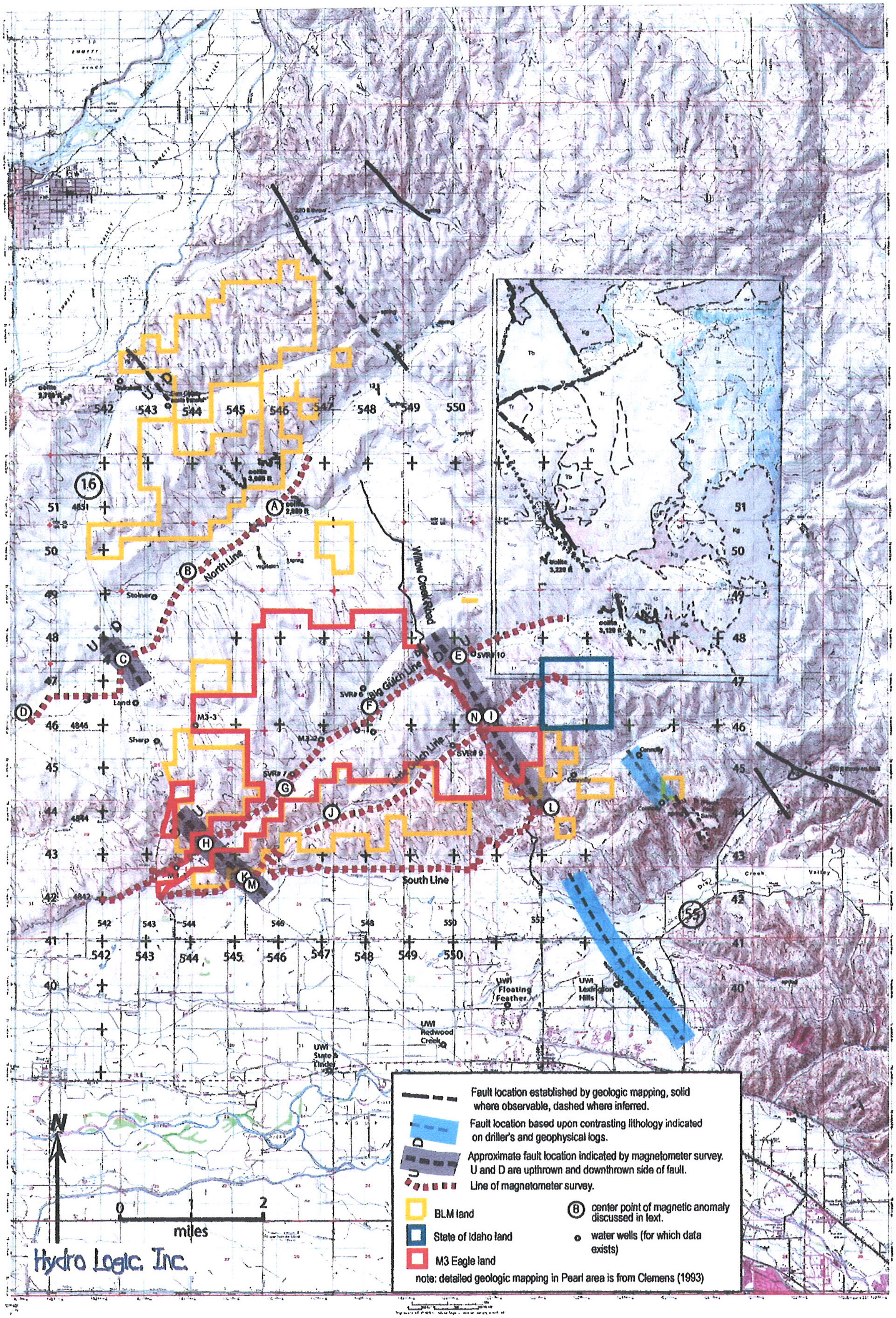


Figure 1. Map showing M3 property, geologic mapping in progress, and location of magnetometer lines. Center point of magnetic anomalies are circled - refer to text for discussion.

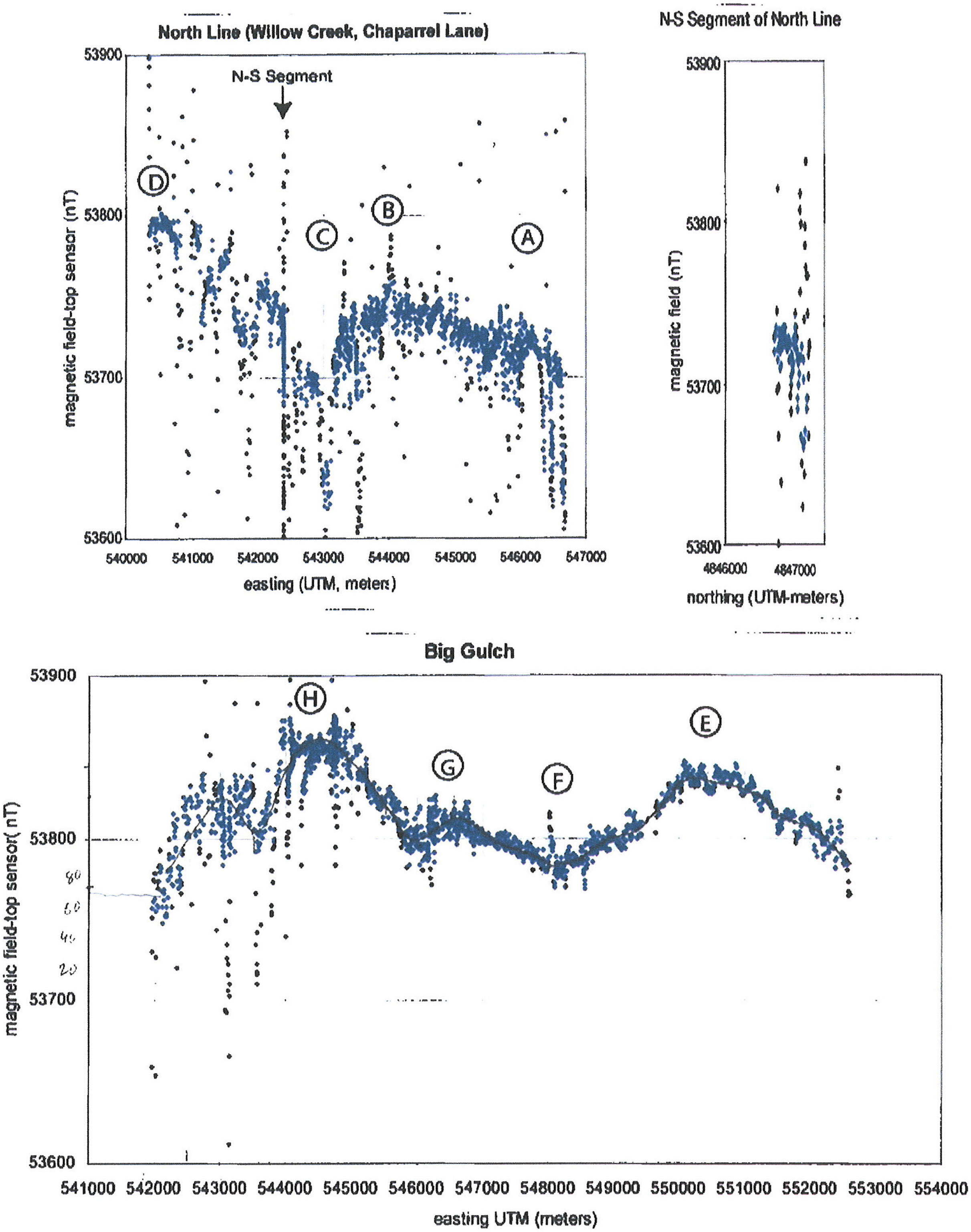
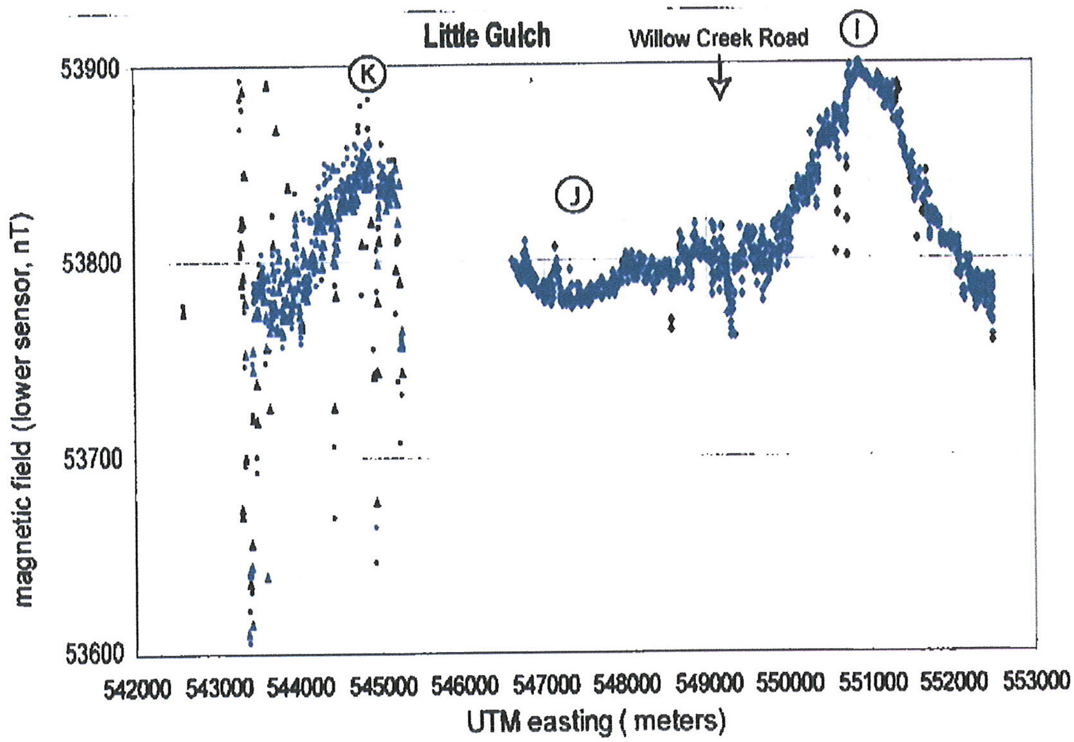


Figure 2. Magnetometer lines along Willow Creek - Chaparral Lane and Big Gulch



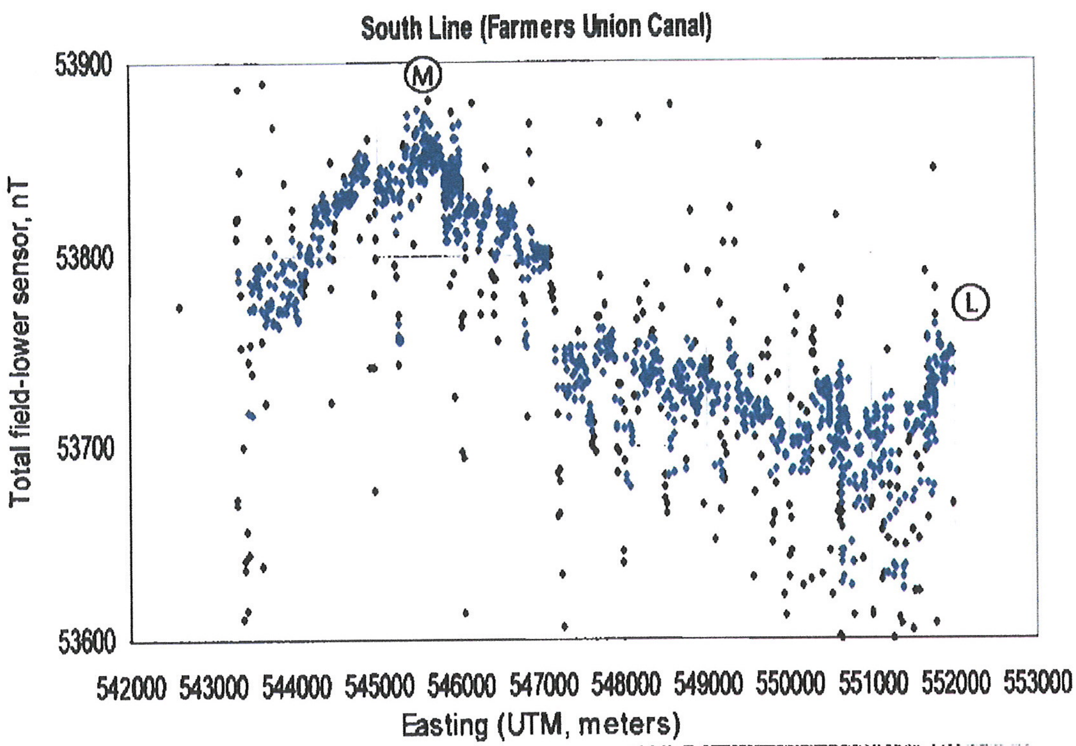


Figure 3. Magnetometer lines along Big Gulch and the Farmers' Union Canal.

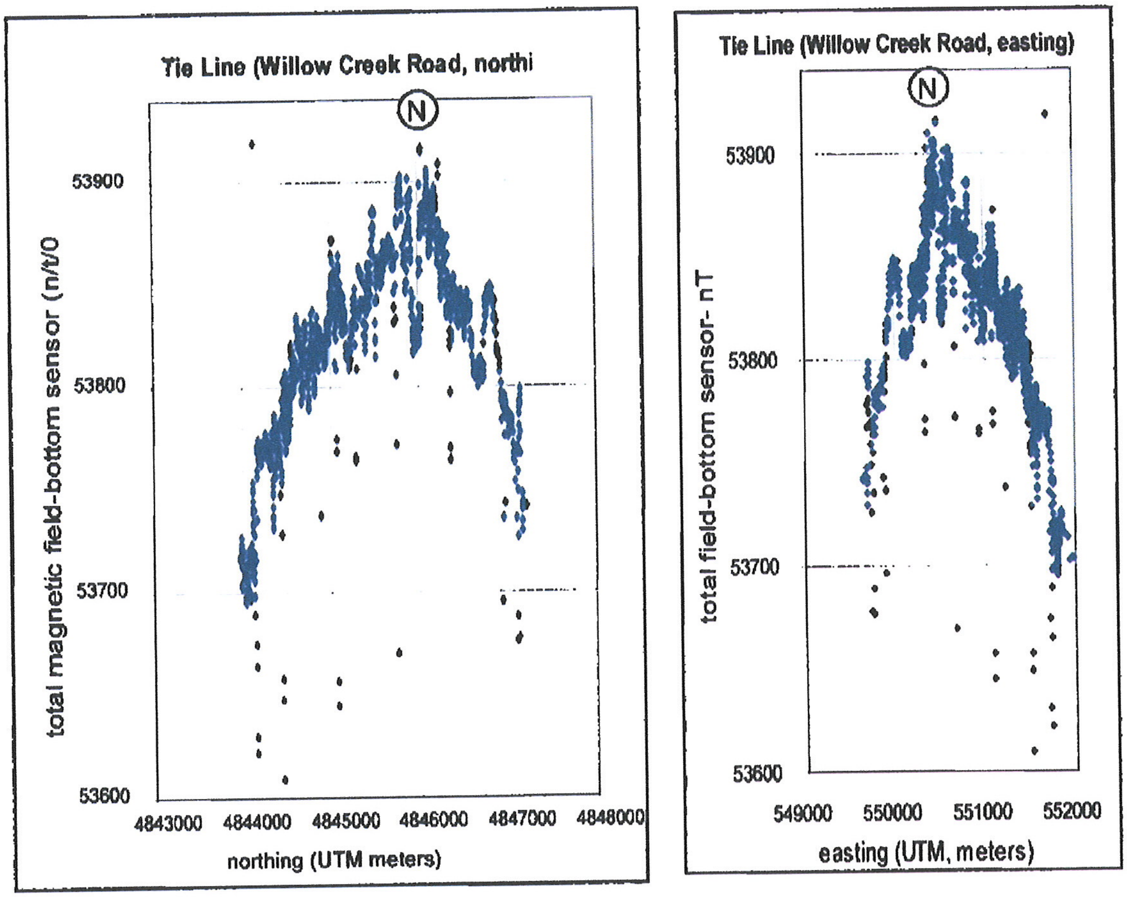


Figure 4. Magnetometer line along Willow Creek Road.

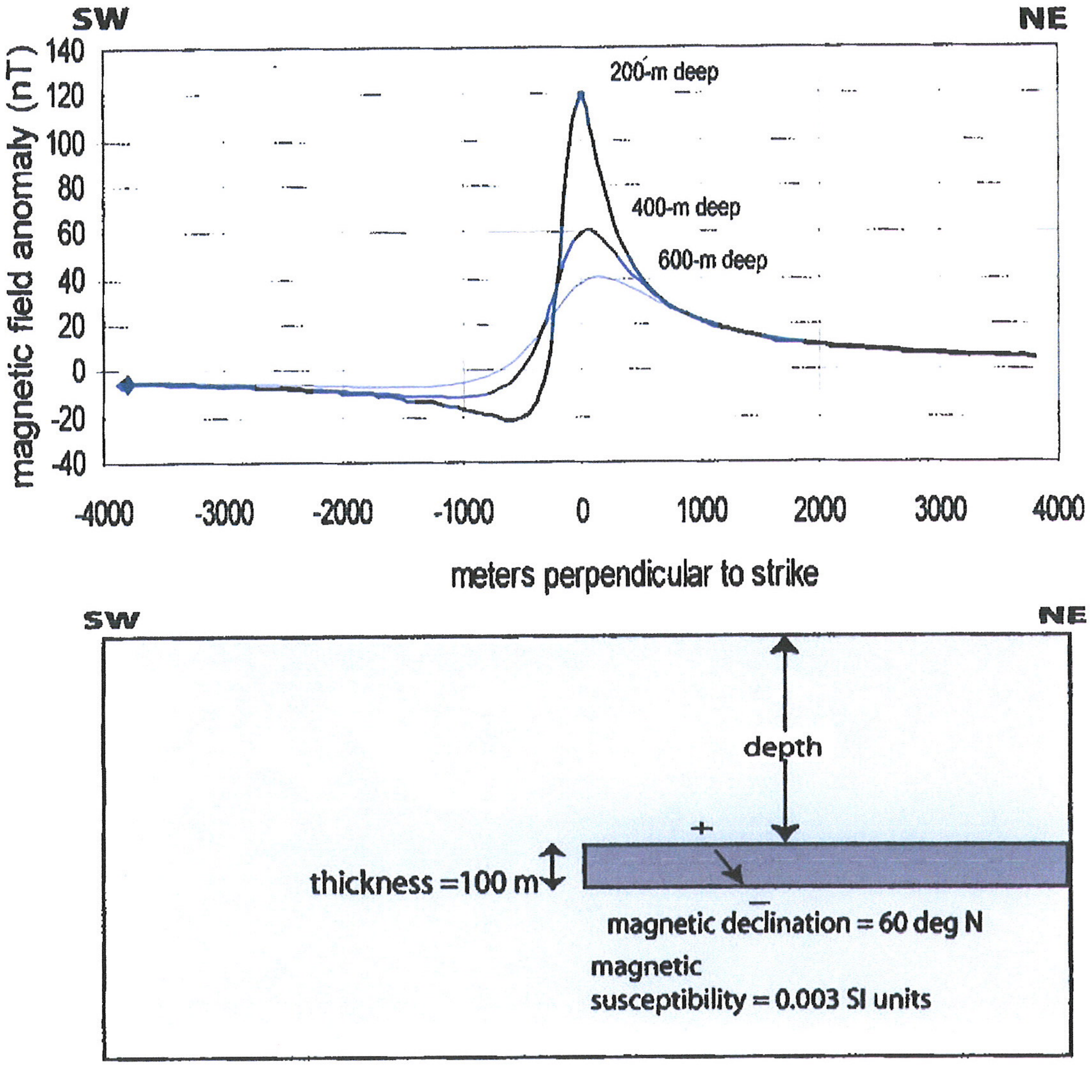


Figure 5. Model calculation of a down-to-SW faulted volcanic slab, fault with a NW-SE strike - simplified to show only the upthrown block.

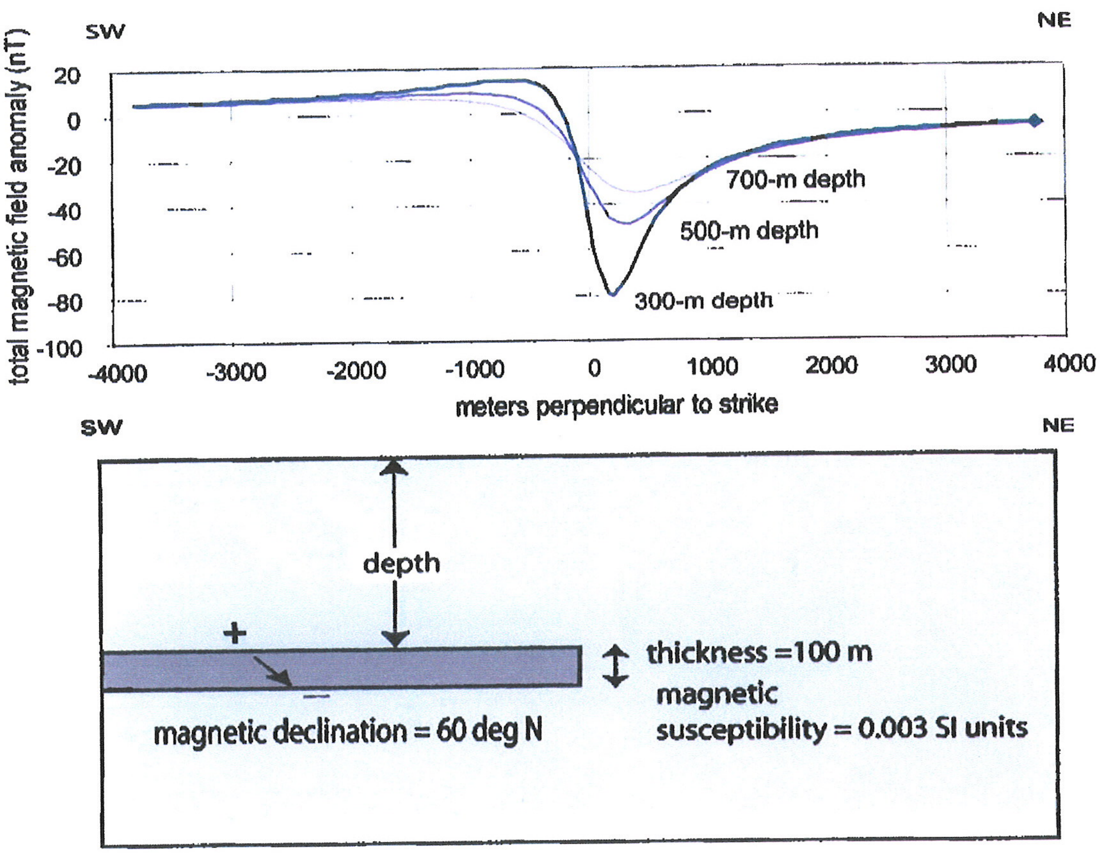


Figure 6. Model calculation of a down-to-NE faulted volcanic slab, fault with a NW-SE strike - simplified to show only the upthrown block.

USGS
science for a changing world

Geological Field Trips in Southern Idaho, Eastern Oregon, and Northern Nevada

Edited by Kathleen M. Haller and Spencer H. Wood

Open-File Report 2004-1222

U.S. Department of the Interior
U.S. Geological Survey

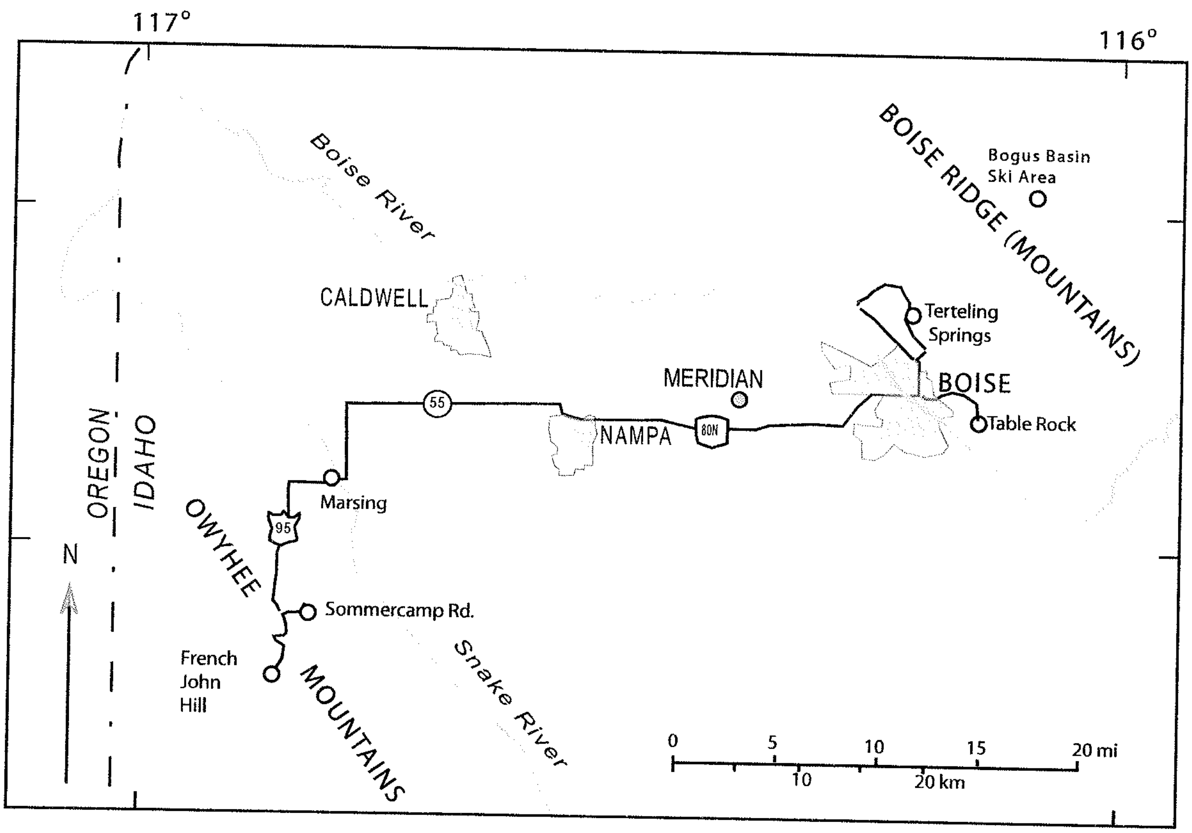


Figure 1. Field-trip route from Boise to the Owyhee Mountains and back to points of interest in the Boise foothills.

Geology Across and Under the Western Snake River Plain, Idaho: Owyhee Mountains to the Boise Foothills

By Spencer H. Wood

Introduction

This 1-day field trip is a transect across the western Snake River Plain (fig. 1). The western plain is a continental-rift structure, 300 km long and 70 km wide. It is bounded and internally faulted by northwest-trending normal faults. The western Snake River Plain has a different orientation and structure than the eastern plain. The eastern plain is a curious downwarp related to magmatism and extension along the track of the Yellowstone hot spot (fig. 2). The faulted basin of the western plain began forming about 12 m.y. ago, and much of the relief was completed by 9 Ma. This timing corresponds with the passage of the hot spot located to the south about 11 Ma. Wood and Clemens (2002) suggest that softening of the lithosphere by the passing hot spot triggered extension and basin formation. The hot spot passage was accompanied by voluminous rhyolite volcanism to the south and by eruptions of rhyolite at or near the margins of the western plain (Bonnichsen and others, 2004; Perkins and Nash, 2002; Pierce and Morgan, 1992). Northwest of the western plain and in southeastern Oregon voluminous eruptions of the Columbia River and Steens Mountains flood basalts occurred between 16.1 and 15.0 Ma (Hooper and others, 2002a, 2002b; Camp and others, 2003). Earliest Columbia River basalts are as old as 17.5 Ma (Baksi, 2004)

Understanding of the sedimentary record builds upon earlier work in the central and southern part of the western plain by Malde and Powers (1962) who defined many of the stratigraphic units. Work of Squires (1992) improved our knowledge of the subsurface near Boise. The sediments record

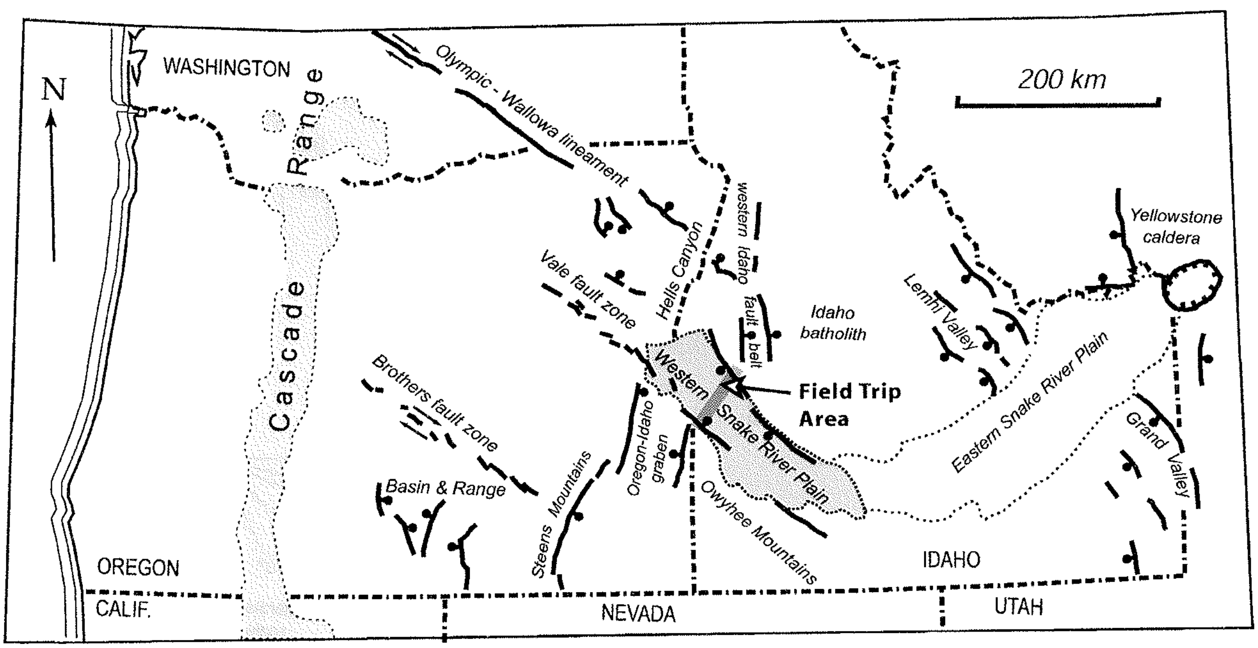


Figure 2. Western and eastern Snake River Plain and late Cenozoic geologic features of the northwestern United States. The western plain was a lake basin in the late Miocene to late Pliocene, usually referred to as "Lake Idaho" (after Wood and Clemens, 2002).

Department of Geosciences, Boise State University, Boise, Idaho 83725,
swood@boisestate.edu

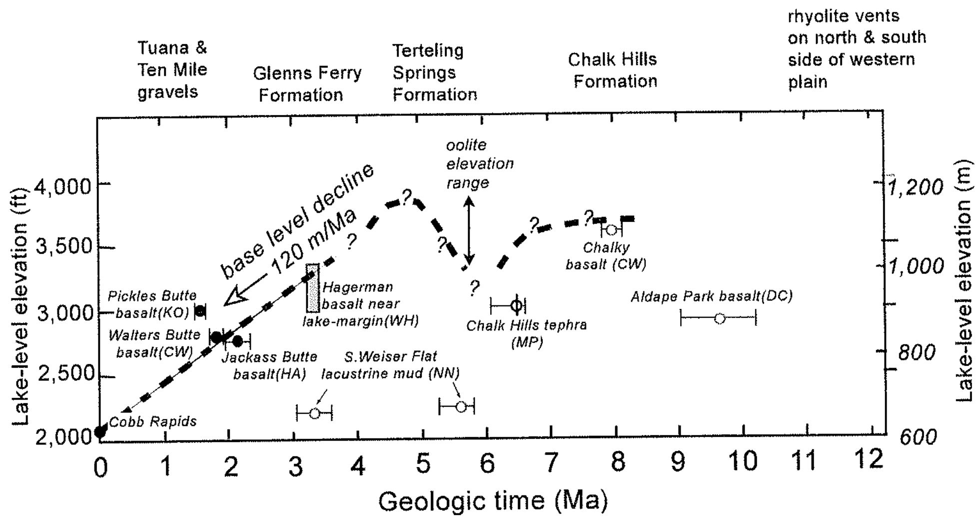


Figure 3. Plot of elevation of the lake deposits versus time. This plot does not take into account tectonic deformation that may have altered the elevation of lake deposits. Most localities are on the margin of the plain, which has not been so much affected by tectonic movement or compaction subsidence (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: DC (Clemens and Wood, 1993), HA (Amini and others, 1984), Zircon fission-track ages on silicic ash are NN (Nancy Naeser, published in Thompson, 1991); $^{40}Ar/^{39}Ar$ ages on basalt are: CW (White and others, 2004), and KO (Othberg and others, 1995), WH (Hart and others, 1999), and tephrochronology from MP (Perkins and others, 1998).

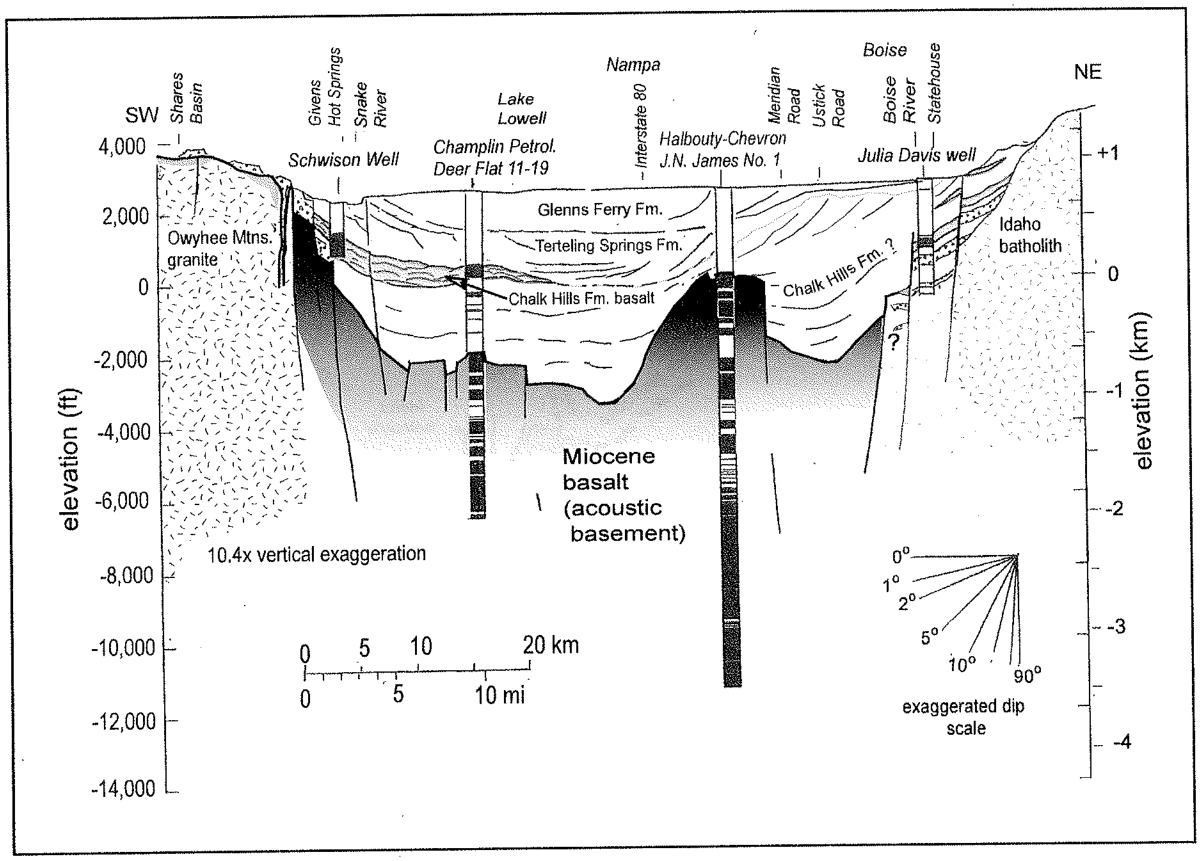


Figure 4. Section across the western plain showing major stratigraphic units, based upon seismic-reflection and well data. Dark patterns on well columns are basalt intervals: dotted pattern is rhyolite: no pattern is sediment.

two major episodes of large lakes that filled the basin (figs. 3 and 4). The Chalk Hills Formation records the first deposition of sand, lacustrine muds, and intercalated volcanics. Subsequently the level of the "Chalk Hills lake" declined, or perhaps the lake completely drained. Sediments and volcanics of the Chalk Hills Formation are deformed by tilting and faulting. The lake system then refilled and transgressed over eroded Chalk Hills Formation. The transgressive lake sediments grade upward and shoreward to calcareous muds and oolites; this sequence is called the Glenns Ferry Formation on the south side of the plain, and the Terteling Springs Formation on the north side (figs. 3 and 4). This last lake system is popularly known as "Lake Idaho". The calcareous sediments indicate increased alkalinity of a lake within a closed basin. The lake then overtopped its spill point into ancestral Hells Canyon and, as it lowered, the drained basin filled with mostly sandy delta-plain units that are important aquifers.

The field trip starts in a rhyolite field in the Owyhee Mountains: the 11-Ma Jump Creek Rhyolite (fig. 5). We then look at the high gravels overlying the rhyolite at elevation 1,100 m. Descending onto the plain, we will examine the basal lacustrine sediments along Sommercamp Road, mapped as the Chalk Hills Formation. The lower Chalk Hills Formation contains numerous volcanic ash beds and an unusual pumice-block layer. Several basalt fields occur in the subsurface and in exposures along the margins. Along the Owyhee Mountains front, the transgressive sequence overlying the deformed and eroded Chalk Hills Formation is readily identified by an angular unconformity with a locally occurring ledge of nearly horizontal, brown, coarse, pebbly sandstone laying upon the

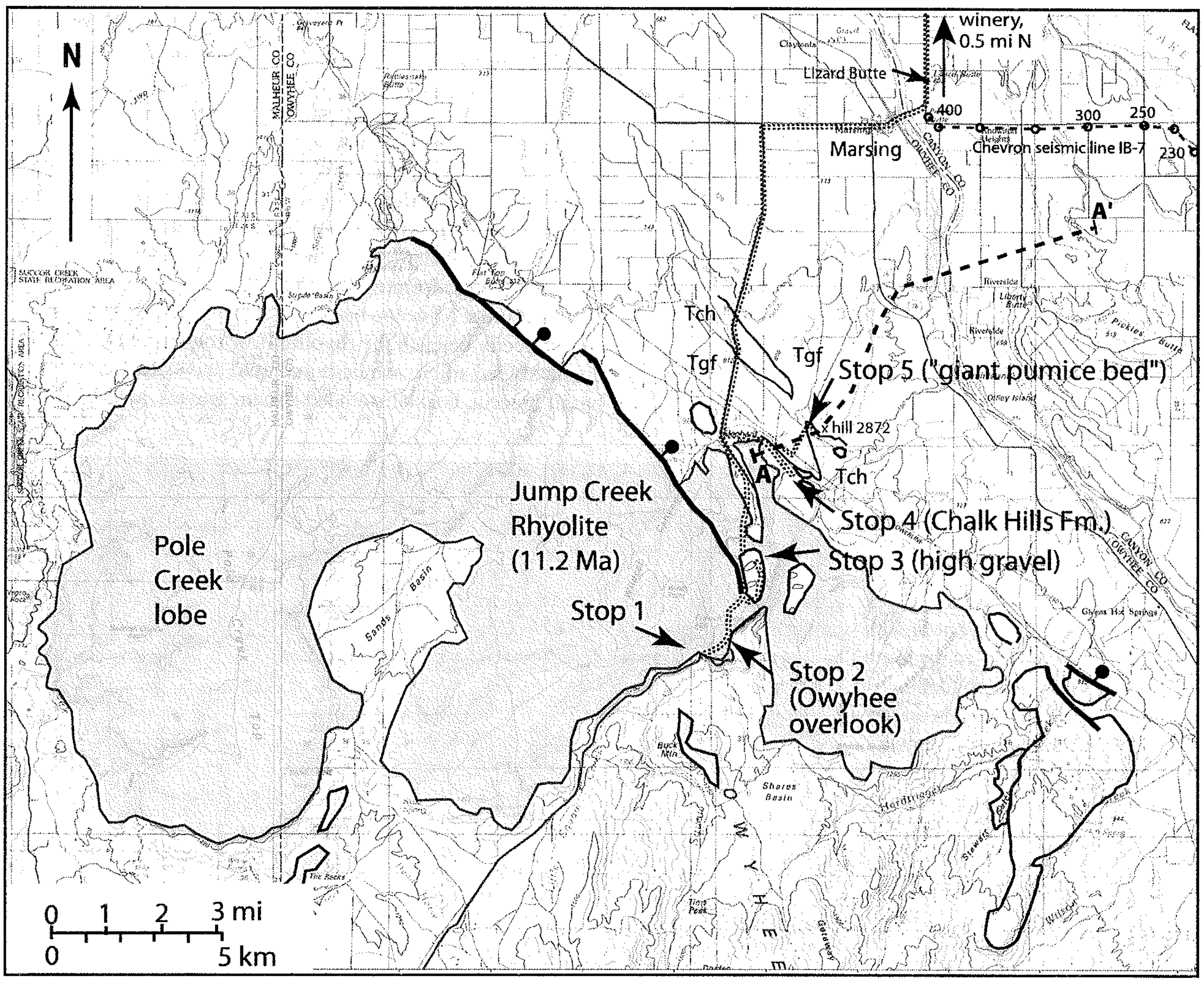


Figure 5. Map showing field-trip route along the Owyhee Mountains front near Marsing, Idaho, distribution of the 11-Ma Jump Creek Rhyolite, and associated geologic features along the edge of the western Snake River Plain. Location of cross-section A-A′ (fig. 16), seismic line (fig. 12), field-trip route (shaded line), and Stops 1-5 are shown.

tilted mudstone. This sandstone ledge is overlain by mudstone of the Glenns Ferry Formation.

Over most of the plain, we rely on seismic data and deep wells for information on the subsurface. From Marsing, we will drive across the plain through the towns of Nampa and Meridian to Boise. Squires and others (2003) describe a 15-km-long seismic line from Meridian to Boise from which we can interpret the episodes of lake filling and sedimentation as seen on the plain margins.

The angular unconformity between the Chalk Hills Formation and the Glenns Ferry Formation is not exposed along the Boise foothills; however, a section containing oolite bars is interpreted as the upper part of the transgressive sequence named the Terteling Springs Formation. Overlying the transgressive sequence is a massive Gilbert-style delta sequence of coarse sand that is interpreted as the response to regression of the lake after the lake overtopped the spill point at Hells Canyon. Downcutting of the outlet resulted in a slowly declining lake level and delta progradation over the basin. The basin filled and rivers flowed across the plain. Along the field-trip route, we observe valleys of the Snake and Boise Rivers that are incised about 150 m into the lake sediments.

FIELD-TRIP ROAD LOG

*Mileages for this first leg of the trip are based upon the **green milepost signs**, posted every mile on the northbound side of the highway.*

Milepost	Inc.	
13.8	0.0	We travel north on U.S Highway 95 toward Boise. Starting point for mileage is at the south end of a large road cut at the crest of the highway at French John Hill, 14 mi south of Marsing, Idaho. Parking area is on the north side of highway, or continue south about 0.1 mi to large pullout on south side, where one can make a U-turn with good visibility.

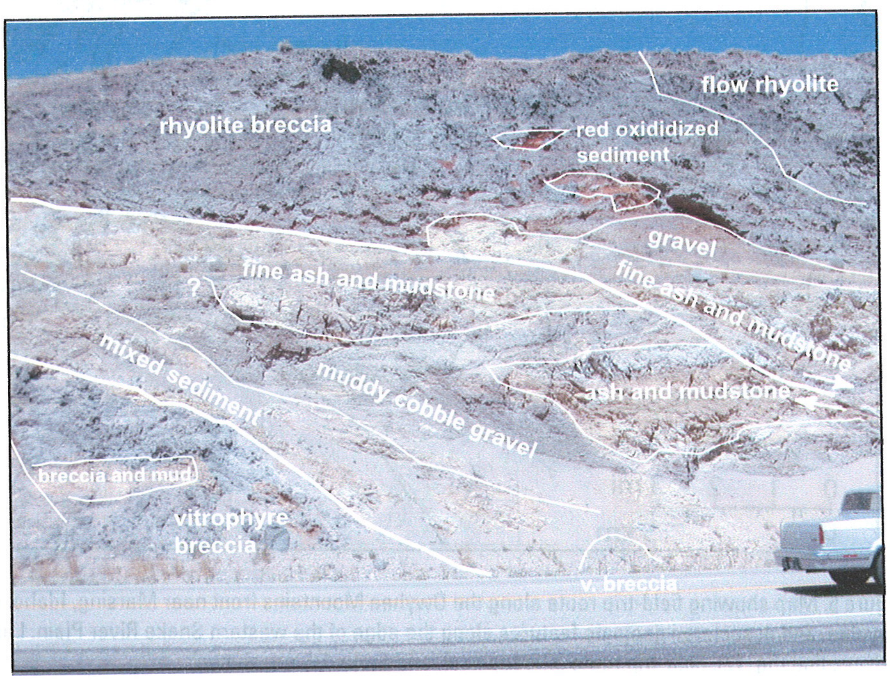


Figure 6. Deformed sediment and rhyolite at the base of the Jump Creek Rhyolite, Stop 1, west side of road cut on U.S Highway 95 (milepost 13.8).

Stop 1. Road Cut with Deformed Sediments and Pyroclastics at the Base of the Jump Creek Rhyolite, French John Hill, Owyhee Mountains

Road cuts here expose a jumble of rhyolite vitrophyre breccia and older sediments beneath the flow-banded and sheeted stoney rhyolite flows on the skyline (fig. 6). The Jump Creek Rhyolite covers 275 km^2 and is one of the younger rhyolites erupted along the margin of the western Snake River Plain (fig. 5). Neill (1975) obtained a K-Ar age of 11.2±0.2 Ma on sanidine. The rhyolite has many lobate flows, one upon another, and in this area is up to 250-m thick. Volume is estimated at 70 km^3. Bonnichsen and others (2004) consider this a rhyolite field made up of many segments. They report $^{40}Ar/^{39}Ar$ and K-Ar ages ranging from 10.6 to 11.7 Ma.

Rocks of the Jump Creek Rhyolite are identified by abundant (12–23 percent), conspicuous (up to 4 mm) phenocrysts of plagioclase. Sanidine and quartz vary in abundance up to 4 percent and are up to 2 mm in size. Microphenocrysts of ferrohypersthene and clinopyroxene are present (Ekren and others, 1981, 1982). A single analysis of this rock, on a water-free basis, shows 71 percent silica making it truly a rhyolite (analysis published in Ekren and others, 1984).

In the valley southeast of U.S. Highway 95, Squaw Creek has cut through the rhyolite into a 130-m-thick section of ashy stratified sediments mapped by Ekren and others (1981) as the Sucker Creek Formation. In the lower part of this section is an 8-m-thick, ledge-forming ash-flow tuff. The section rests upon 500 m of older Miocene basalt.

Directly beneath the rhyolite flow rock is about 20 m of dismembered silicic-ash beds and mud and ashy sediments, basalt breccia, beds of rounded stream gravels, and rhyolite

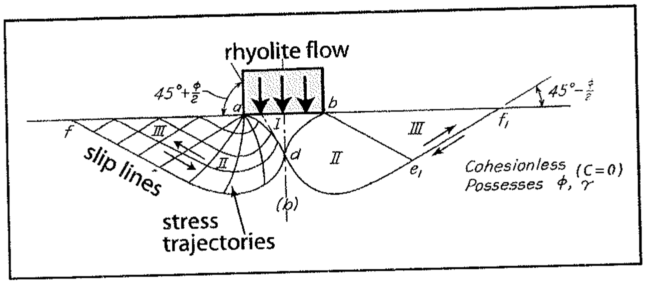


Figure 7. Diagram showing model of bearing-failure slip lines for fairly dense or stiff soil with frictional strength and no cohesion. This type of failure, with low-angle slip lines may have occurred as the thick flows of the Jump Creek Rhyolite piled up over wet sediments of the underlying Sucker Creek Formation (drawing after Terzaghi and Peck, 1967).

breccia. Such structural complexity originally suggested to me that this is a near-vent area of the rhyolite. Low-angle faulting cuts through the section exposed in the road cut which suggested sliding of near-vent volcanic topography (fig. 6). However, the low-angle faults are more likely analogous to bearing failures produced by loading the area with the thick rhyolite flows (fig. 7). The complex structures here are similar to larger scale features in the Absaroka volcanics described by Decker (1990) and which he attributed to liquefaction of saturated epiclastic and pyroclastic rocks in response to loading by lava flows. In this area, the overlying rhyolite lava was up to 200-m thick and imposition of such a great load over a short period of time (perhaps days to decades) could well have liquefied and deformed these materials.

Ekren and others (1984) suggested the Jump Creek Rhyolite may have been a rheomorphic tuff, but Bonnichsen and Kauffman (1987) have shown that thick breccias and features seen here are typical of large rhyolite flows. This area warrants a detailed description and study as few areas of the large hot-spot related rhyolite flows are so well exposed as here in Squaw Creek Canyon.

The older Miocene volcanic rocks of the Owyhee Range host major silver-gold deposits (*e.g.*, the Delamar Mine), about 30 mi due south of here (Halsor and others, 1988). The volcanic rocks rest upon Cretaceous (62 to 70 Ma) granitic rocks exposed at several places in the mountains southeast of here, which Taubeneck (1971) called the southern extension of the Idaho batholith.

Mile-post	Inc.	
		Walk north 0.5 km through the road cut to see the deformed sediments and rhyolite. Continue walking north, and the vans will be moved to the Owyhee Mountains Viewpoint, to pick you up.
14.1	0.3	Park for Stop 2.

Stop 2. Owyhee Country Viewpoint

Large parking area on the east side of road. The sign explains that the origin of the name "Owyhee" is not an Indian name as it would seem. It is an antiquated spelling of Hawaii, as used by Captain Cook. In 1819, Donald MacKenzie of the Canadian Northwest Company sent a group of trappers into these mountains. Among the party were several native Hawaiians who were never seen again, and so the mountains are named for those lost "Owyheeans".

The hill to the west of here is called French John Hill, named for "French" John Carrey who built a road in the early 1870s parallel to the present U.S. Highway 95 (Boone, 1988). From the view point, one looks across the deep rhyolite gorge of Squaw Creek and the north flank of the Owyhee Mountains to the western Snake River Plain. Across the plain are the Idaho batholith mountains north of Boise. Rimrock basalt along the north side of the Snake River Canyon overlies gravels, which are underlain by lake beds of the Pliocene Glenns Ferry Formation. Pickles Butte is a basalt vent area $^{40}Ar/^{39}Ar$ dated by Othberg (1994) at 1.58±0.085 Ma, which is the minimum age usually quoted for the complete withdrawal of Lake Idaho from this area. Basalts in this area, including Lizard Butte, are the westernmost extent of young basalt that overlies lakebeds.

About 14,500 yr ago, floodwaters of the Bonneville flood roared 60-m (200–ft) deep through the Snake River canyon below, but were confined to the canyon below 755-m (2,480-ft) elevation (O'Conner, 1993).

Mile-post	Inc.	
15.0	0.9	Turn east at milepost 15 on a paved road, that turns to gravel within 0.1 mi. Proceed for 0.3 mi, and continue straight at the Y to the old Highway 95 grade, and follow the abandoned highway north for a total of 1.1 mi from the turnoff at milepost 15. The road cuts are in the high gravel deposit of Stop 3.

Stop 3. High Gravels of the Owyhee Mountains and an Overlook of the Chalk Hills Formation-Glenns Ferry Formation Contact

From this abandoned grade of U.S. Highway 95, there is a good view over the western plain and the area of the next stop, Stop 4 (fig. 8). The road cuts are in an alluvial-fan gravel perched high in the Owyhee Mountains. Ekren and others (1981) mapped these high gravels (figs. 5 and 9) but did not comment on their significance. The gravels clearly overlie the Jump Creek Rhyolite, but their relationship to the lake deposits is unclear. The elevation here is 1,100 m (3,600 ft), which is the highest elevation of most lake deposits around the plain margin. It is likely this is an old alluvial fan sequence

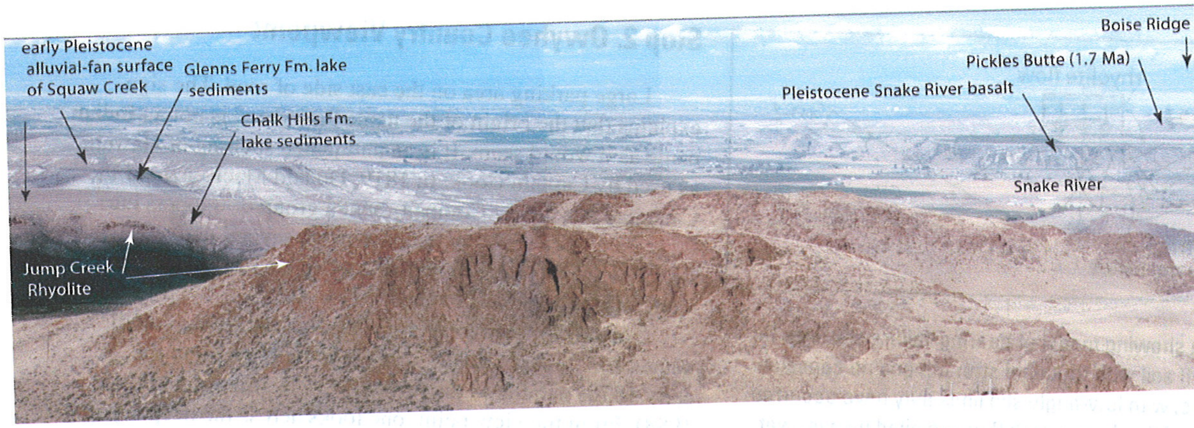


Figure 8. Overlook of the western Snake River Plain, looking north from Stop 3.

that graded to Lake Idaho, but we do not know the age of the gravels or the amount of vertical faulting that displaced the gravel downward to the north. We will see a thinner, but very similar, gravel overlying the top of the Chalk Hills Formation in the Sommercamp Road cut at the next stop (see fig. 10).

The road is blocked at a distance of 1.3 mi, so turn around, park, examine the gravels and the view (fig. 8), and then return by the same route back to the new U.S. Highway 95.

Mile-post	Inc.	
15.0	1.1	Turn north toward Boise on U.S Highway 95 and proceed down the grade.
16.8	1.8	Road cut to the west side of the road exposes the contact between vitrophyre and overlying stoney rock of the Jump Creek Rhyolite. The vitrophyre contains abundant spherulites and stretched lithophysae. These features in the glassy part of the flow are attributed to pockets of higher vapor content that cause crystallization and devitrification emanating from the vapor-filled voids. These stoney-rhyolite spherulites form in the hot glass (Lofgren, 1971; Cas and Wright, 1988).
16.9	0.1	Road cut on west side of road exposes a northwest-trending fault, with a 1.5-m-thick fault breccia. Bentonitic claystone of the lower Chalk Hills Formation is faulted down to the northeast against the Jump Creek Rhyolite.
18.4	1.5	Turn east from Highway 95, just past the weigh station, on to Sommercamp Road. Travel east on Sommercamp Road for 0.9 mi to the top of the grade and road cut.

Figure 9. High gravels at Stop 3 (see fig. 5 for location). Mudstone overlain and interbedded with cross-bedded coarse sand and subangular gravel. Rock pick handle is 30 cm, for scale.

Stop 4. Chalk Hills Formation Along Sommercamp Road

We will walk down the grade and examine the road cut of deltaic sediments and then return to vans at the top of the grade on Sommercamp Road.

The Chalk Hills Formation represents sediments of the earliest large lake in the western Snake River Plain rift. The age of the formation is poorly constrained. In their review of radiometric ages on the formation, Wood and Clemens (2002) place its age between 10 and 6 Ma. The road cut along Sommercamp Road exposes a section of foreset and topset beds of a delta (fig. 10). These sediments are near the base of the formation and rest upon a section of clay-altered vitrophyre of the Jump Creek Rhyolite (fig. 11).

The delta sediments in the road cut are unconformably overlain by a gently-west-dipping, grey cobble gravel, similar to that seen at Stop 3. The gravel is faulted at the west end of

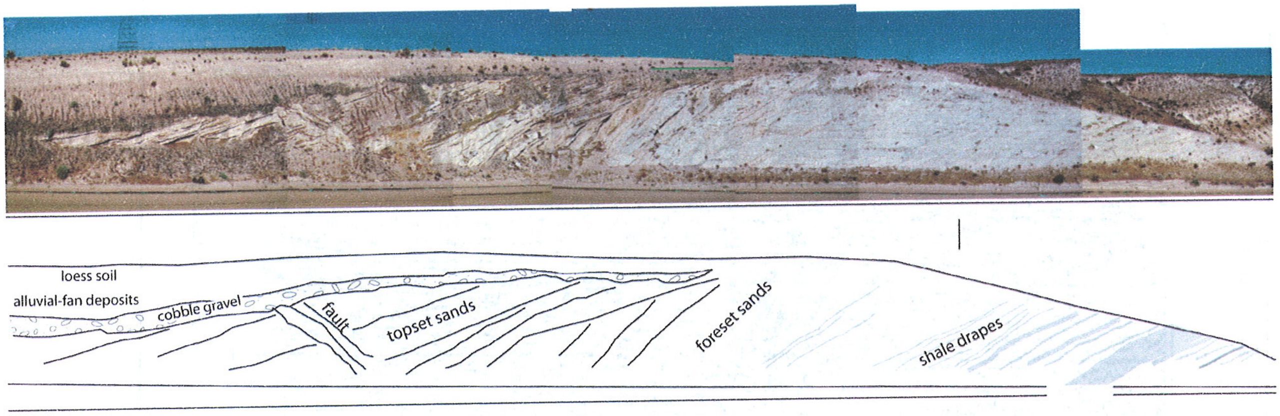


Figure 10. Delta foreset and topset beds of the Chalk Hills Formation along road cut of Sommercamp Road at Stop 4. The basalt ash marker bed in figure 11 is about 50 m west of photo along road cut.

the road cut and is overlain by unfaulted alluvial fan deposits and a loess soil. The Glenns Ferry Formation is missing in this road-cut exposure. One kilometer north of this road-cut exposure, 10 m of lacustrine sediment of the Glenns Ferry Formation unconformably overlies the Chalk Hills Formation with an angular discordance of about 8°.

The Chalk Hills Formation is about 110 m thick in this area (fig. 11). We will be looking at the sediment and volcanic features at two outcrop areas, here along Sommercamp Road, and another in a gulch 1.6 km (1.0 mi) northeast of here. The formation here is mostly claystone with an arkosic sandstone bed near the base, abundant intercalated silicic-ash layers, and delta sands that are mostly coarse silicic ash. Topset beds have well-preserved ripple marks. The foreset beds are mostly coarse sand ash with mud drapes, typically 0.1–0.5 m thick. Mud drapes form on the foreset sand beds in the time intervals between sand avalanche and deposition on the foreset slopes, or when a particular depositional delta lobe is temporarily abandoned by a distributary.

Exposed along Snake River Plain margins and also detected on seismic sections and wells beneath the plain, 10 km (6 mi) north of here (figs. 12 and 5) are intercalated local basalt fields within the Chalk Hills Formation. Many of these basalt fields erupted into water (Bonnichsen and others, 1997). At this locality, basalt occurs as one or two layers of scoriaceous ash and lapilli beds about 0.5-m thick at the west end of the road cut and indicated in the upper part of the stratigraphic section (fig. 11).

At the top of the Sommercamp Road grade, take the dirt road to the north through the barbed-wire gate and follow the track that parallels the power transmission line to the southeast. Travel 0.3 mi to a road that turns left (north) and follows the wooden-pole power line to the northeast. Travel 0.7 mi to the edge of the mesa and park.

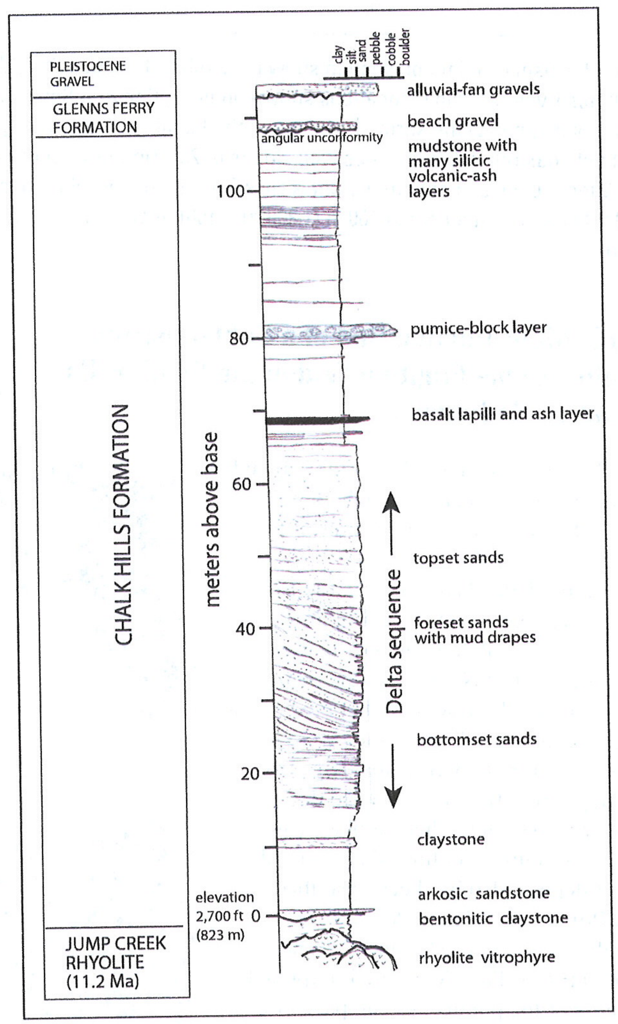


Figure 11. Graphic stratigraphic column of the Chalk Hills Formation at Stops 4 and 5.

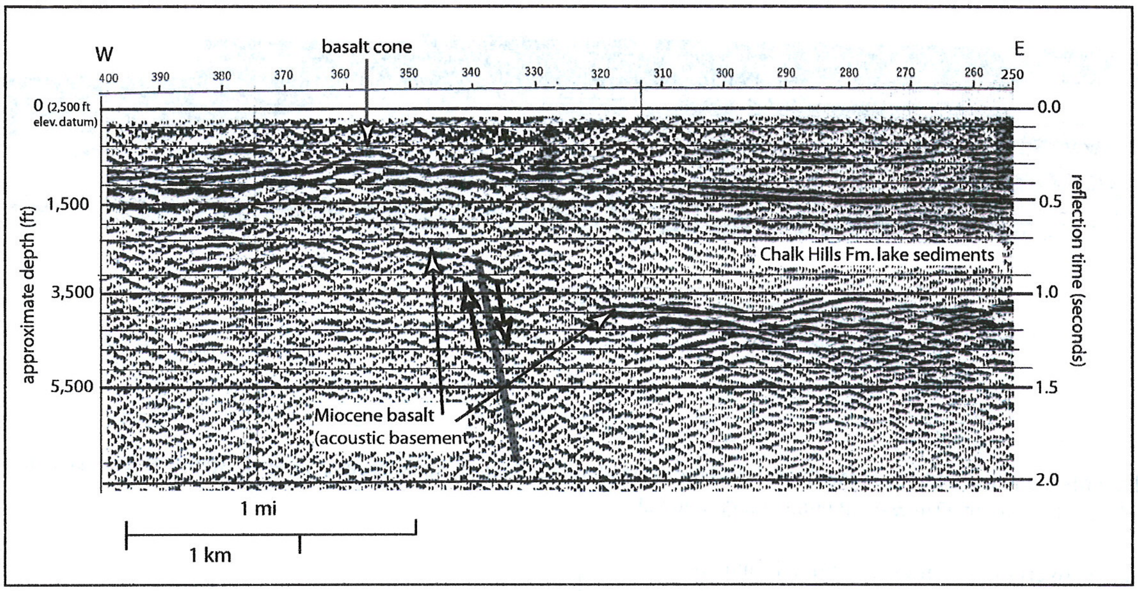


Figure 12. Seismic-reflection section showing a subsurface basalt field in the Chalk Hills Formation, between Marsing and Lake Lowell Reservoir (shotpoint locations shown on fig. 5, Chevron Seismic Line IB-7, shot points 250–400). The basalt field is 305–460 m (1,000–1,500 ft) below the surface. The top of the basalt appears as high-amplitude reflections from the sediment-basalt interface. The basalt field has relief due to cones with slopes of 23°. On the west end of the section, at a depth of 670 m (2,200 ft), is the surface of the older Miocene basalt beneath the Chalk Hills Formation. This older basalt surface is faulted down to the northeast to a depth of 1,070 m (3,500 ft,) or a displacement of 400 m (1,300 ft). Faulting of the basalt field within Chalk Hills Formation appears to be less than 90 m (300 ft).

Stop 5. Giant-Pumice Bed of a Sublacustrine Rhyolite-Dome Eruption Within the Chalk Hills Formation Sediments

There is a rough road over the rim and down to the valley below, on which we will walk 300 m northeast to exposures of the "giant-pumice bed".

A remarkable layer of large (up to 1 m) pumice blocks, about 1.5-m thick, is exposed in a gulch in the SW quarter of section 32, T. 2 N., R. 5 E. The pumice blocks are radially fractured, but the broken pieces are intact, indicating they were deposited before they fractured (fig. 13). Post-deposition fractures indicate the blocks were still hot when emplaced and then cracked upon cooling. This unusual type of deposit also has been described at La Primavera Volcano, Mexico, by Clough and others (1982) and Cas and Wright (1988). The blocks are interpreted as pieces of the pumiceous carapace of a rhyolite dome that erupted beneath a lake and then floated and shoaled while still hot and intact (fig. 14). Also in this gulch, is the continuation of the section above the basalt ash seen in the road cut of Stop 4. Here the section contains many silicic volcanic ash layers within claystone, and is locally overlain by

Figure 13. Large fractured pumice block in pumice-block layer within the upper Chalk Hills Formation at Stop 5.

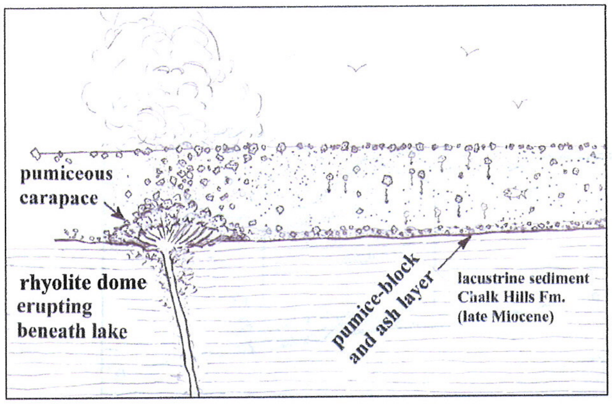


Figure 14. Sketch of a pumice dome eruption beneath the lake. It seems unlikely the large blocks waterlogged and sank, because they must have been hot and intact when emplaced. More than likely, the large blocks were steaming hot and still intact and then fractured in place upon further cooling.

a 1-m-thick horizontal ledge of pebbly sandstone and 10 m of massive lacustrine mudstone of the Glenns Ferry Formation (figs. 15 and 11).

The Glenns Ferry Formation dips a few degrees to the north, and the section is much thicker north of here (fig. 16). About 0.5 km (0.3 mi) south of this valley along the dirt road that climbs the north mesa is a sequence of silicic volcanic ashes, one of which is about 1-m thick and can be traced for several miles to the north (fig. 16). The lowest part of the Glenns Ferry Formation contains veins of selenite gypsum suggesting local lagoons along the lake shore that dried intermittently. However, none of the formation here is calcareous. It does become very calcareous in its upper part, about 3 km north of here (fig. 16). Much of the lacustrine mud in a 670-m (2,200-ft) deep well beneath Caldwell, 22 km north of here, is also calcareous (Wood, 1994).

Optional 0.25-mi walk is a 180 ft climb to the mesa to the east (hill 2872), for a good view and photo of the angular unconformity. In this valley, at the bottom of the mesa to the north is the contact of the Glenns Ferry Formation over the Chalk Hills Formation; however, the prominent rusty-colored sandy gravel visible on the south side of the valley is missing here. The lower 7 m of the Glenns Ferry Formation contains selenite gypsum veins (satin spar), and the 70-m-thick section here is a monotonous mudstone with several volcanic-ash beds.

Mile-post	Inc.	
		Return to vans at top of the south mesa and retrace the route back west to U.S. Highway 95.
18.4		Turn right (north) on to the new U.S. Highway 95.
19.8	1.4	The angular unconformity is in the road cut on the west side of the highway. A horizontal ledge of brownish-orange pebbly sandstone of the basal Glenns Ferry Formation overlies the slightly tilted claystone and ash layers of the Chalk Hills Formation.
24.0	4.2	Junction of U.S. Highway 95 and State Route 55. Turn right (east) on to State Route 55 toward Marsing.

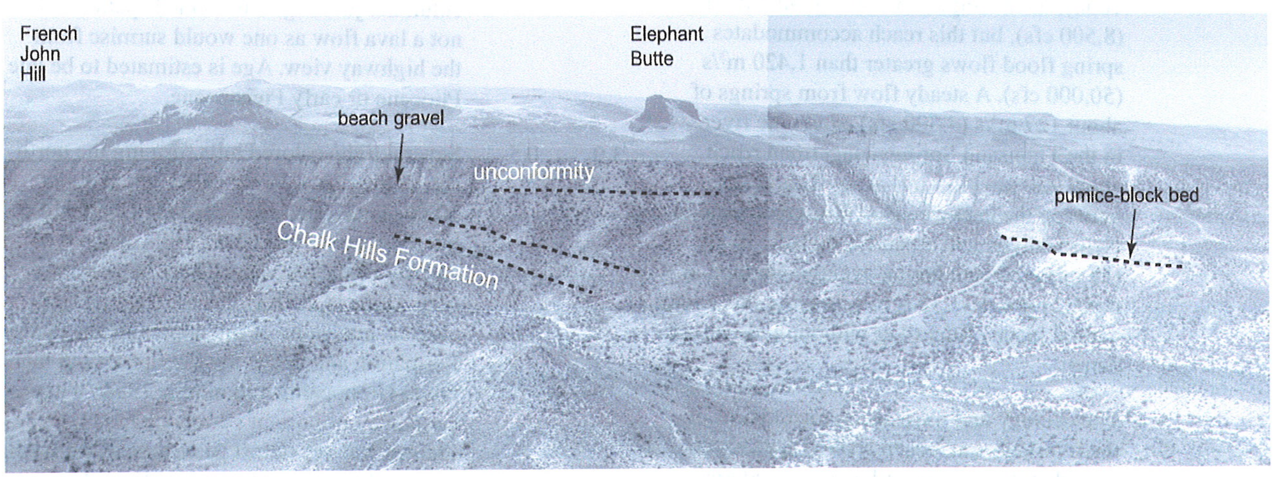


Figure 15. Angular unconformity between the tilted Chalk Hills Formation and the Glenns Ferry Formation. Base of the Glenns Ferry Formation is an 1-m-thick discontinuous sandstone, interpreted as a beach sand. Overlying the sandstone is 10 m of mudstone. Mudstone thickens to the north as shown in figure 19. View is looking south from mesa north of Stop 5.

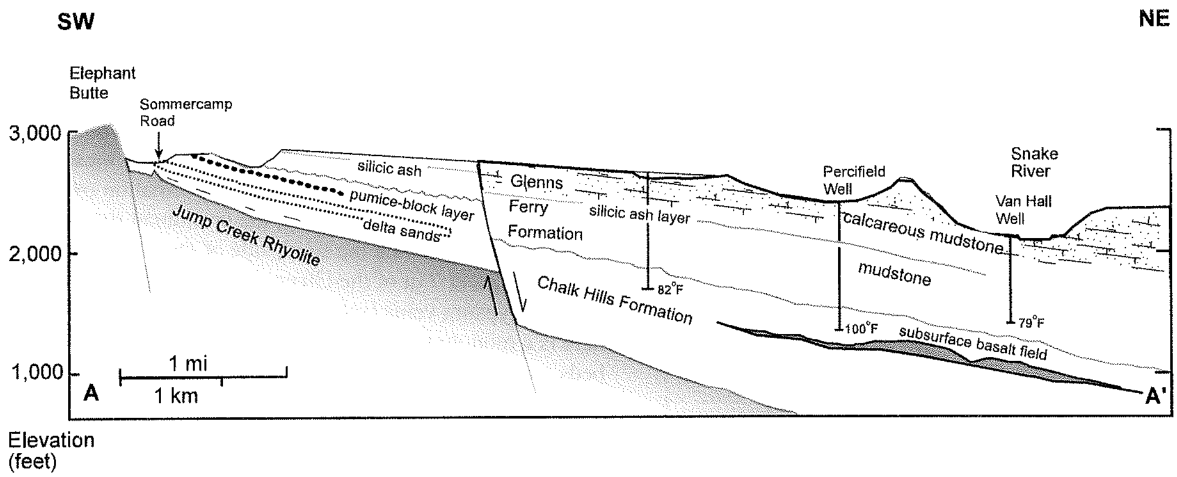


Figure 16. Stratigraphic cross section showing the position of calcareous mudstones of the upper Glenns Ferry Formation. Section is from Sommercamp Road area to the Marsing Hills. Location of section line A-A' is shown on figure 5.

Mileposts on State Route 55 start at 0.0 at this junction

Mile-post	Inc.	
0.0	0.0	Junction U.S. Highway 95 and State Route 55, traveling east on State Route 55.
1.7	1.7	Marsing
2.7	1.0	Snake River bridge at Marsing. The river at Marsing flows in a natural low-gradient reach. The valley of the river is about 120 m (400 ft) below the surrounding plain. When the Bonneville flood roared through here 14,500 yr ago, the raging river was about 60-m (200-ft) deep over Marsing (O'Conner, 1993). Today, the average flows are about 240 m^3/s (8,500 cfs), but this reach accommodates spring flood flows greater than 1,420 m^3/s (50,000 cfs). A steady flow from springs of about 127 m^3/s (4,500 cfs) enters the river in the Thousand Springs-Hagerman reach, about 250 km (155 mi) away. This reach between the backwaters of Brownlee Dam in Hells Canyon and Swan Falls Dam, 56 km (35 mi) upstream has an average gradient of 0.00025 m/km (1.32 ft/mi), making this a long natural reach of the river without hydroelectric dams.

The ecology and physical characteristics of the riverbank have been studied recently to provide information for legal proceedings between the U.S. Department of Fish and Wildlife and the State of Idaho over claims for Federal water rights to protect wildlife refuges on the islands (Ostercamp and others, 2001; Ostercamp, 1998; and Dixon and Johnson, 1999).

3.4	0.7	Lizard Butte east of the highway is the eroded remnant of a complex basalt volcano having many features of hydrovolcanism. The volcano erupted through the wet lacustrine sediment of the Glenns Ferry Formation. The deposits include deformed stream gravels, steeply dipping bomb-and-lapilli beds of scoria and surge deposits. The cap rock is a hard basalt agglutinate with chunks of white sediment, which is interpreted as a welded basalt spatter (Craig White, personal communication, 2003). Although the cap rock displays columnar jointing, it is welded spatter and not a lava flow as one would surmise from the highway view. Age is estimated to be late Pliocene or early Pleistocene.
3.9	0.5	Several light-colored hills forming the north rim of the Snake River valley are at about 10:00. These also are called the Chalk Hills, but they are not the Chalk Hills of the type locality of the Miocene Chalk Hills Formation. These "Chalk Hills" are sand and calcareous mudstone of the upper Glenns Ferry Formation and are of the floodplain and marsh facies (Reppening and others, 1994). They contain an important vertebrate fauna of early Pleistocene age (latest Blancan-earliest Irvingtonian), known as the Froman Ferry fauna. Important fossils, collected by local resident George R. Scott, are microtine rodents, early Pleistocene horse, puma, and an archaic

rabbit. Sediments are of reversed magnetic polarity (Van Domelen and Rieck, 1992), and estimated age is 1.5-1.7 Ma (Reppening and others, 1994).

4.8 0.9 **STOP FOR PICNIC LUNCH.** Ste. Chappelle Winery is the first post-prohibition winery in Idaho. Experimental vineyards were first planted in 1972, and the winery established in 1976 by the Symms Fruit Ranch. Well-drained loess soils, the microclimate of south-facing slopes of the Snake River valley, and the late growing season combine to produce Riesling, Chardonnay, Merlot, and other varietal grapes. There are now eight wineries in this area of which Ste. Chappell is the largest, shipping 200,000 cases per year. Figure 17 is a view from the winery.

6.8 2.0 Going north and then due east on State Route 55, we leave the Snake River valley and travel across the broad western plain.

9.2 3.6 Between Huston and the Lake Lowell turnoff is the "Lower Deer Flat channel", a 3-km-wide, northwest-trending, gravel-mantled, middle-Pleistocene channel incised about 30 m into an upland surface of Glenns Ferry Formation capped by the Tenmile Gravel (Othberg, 1994). This probably is an old channel of the Snake River that was blocked to the southeast by the basalt fields of Kuna Butte and other large shield volcanoes. Eruptions of these lavas forced the river south to its present course on the south side of the plain.

16.2 7.0 Continue straight east through the stoplight at the intersection of State Route 55 and Caldwell Boulevard. This easterly continuation is named Karcher Boulevard, and there are no more mileposts. You will cross over the railroad tracks and the Interstate heading toward the Amalgamated Sugar Refinery.

18.0 1.8 Just past the sugar beet refinery, turn right (south) on Northside Boulevard and then left (east) to enter Interstate 84, heading east to Boise.

Mileages are now based upon mileposts of Interstate 84.

Mile-post	Inc.	
38.0	3.0	Just before Exit 38, the Interstate Highway descends over a basalt-mantled surface (one of the Amity or Deer Flat surfaces). This lava can be traced to Caldwell where it was $^{40}Ar/^{39}Ar$ dated by Othberg (1994) at 0.799±0.095 Ma and has a reversed magnetic polarity. From here east, the Interstate is on the Sunrise Terrace surface.
46.0	8.0	From Exit 46 (Eagle Road) the Interstate heads east, along a route for which we have a high-resolution seismic line all the way to Boise (about 8 mi east). Because of urbanization and traffic in this area, we used a route along the Union Pacific Railroad right-of-way, one mile south of the Interstate.
49.0	3.0	Take signs to CITY CENTER, as you negotiate the "Flying Y" complex of freeway overpasses.
49.5	3.5	About 0.5 mi beyond the Interstate "Flying Y" intersections, the highway descends to the Whitney Terrace surface.
51.8	2.3	Four lanes descend from the Whitney terrace onto the modern flood plain of the Boise River and downtown Boise. The Interstate Connector merges with city streets at Front Street. Continue east on Front Street.

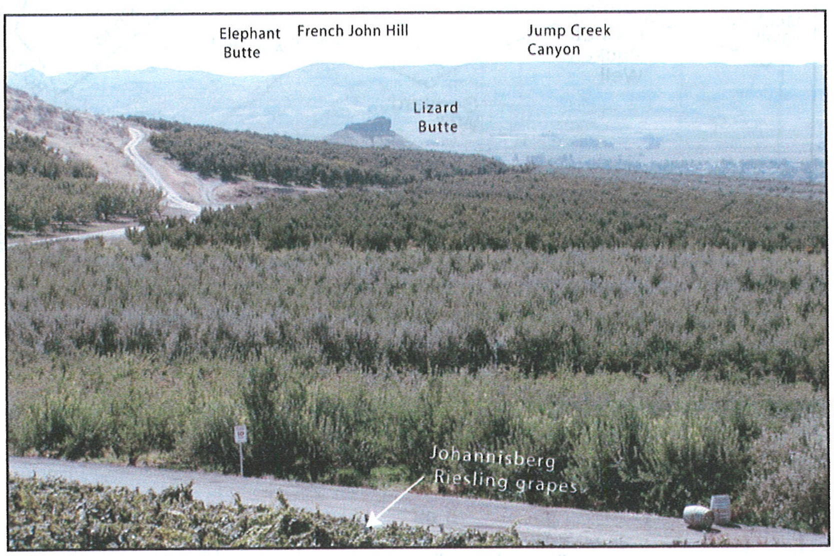


Figure 17. View to the south across the Snake River valley to the Owyhee Mountains from the Ste. Chappelle vineyards (Lunch Stop).

53.5	1.7	At the Capitol Boulevard stoplight, turn left (north-northeast), and the State Capitol Building should be in sight about 5 city blocks ahead.
53.8	0.3	At the Idaho State Capitol Building, turn left one half block in front of the capitol, and then right on 8th Street to the northwest corner of the State Capitol. The 1920 State Capitol building is constructed of Miocene Boise Sandstone from the Table Rock Quarries (Stop 9).

Reset mileages at the 8th Street and State Street intersection.

WESTERN BOISE FOOTHILLS LEG OF FIELD TRIP

Mileage Cum.	Inc.	
0.0	0.0	This leg of the field trip starts at 8th and State Street, the northwest corner of the State Capitol grounds (fig. 18). Continue 3 blocks northeast on 8th Street and turn left (northwest) on Hays Street.
0.0	0.2	Hays Street and 8th Street. Travel 0.6 mi northwest on Hays Street.

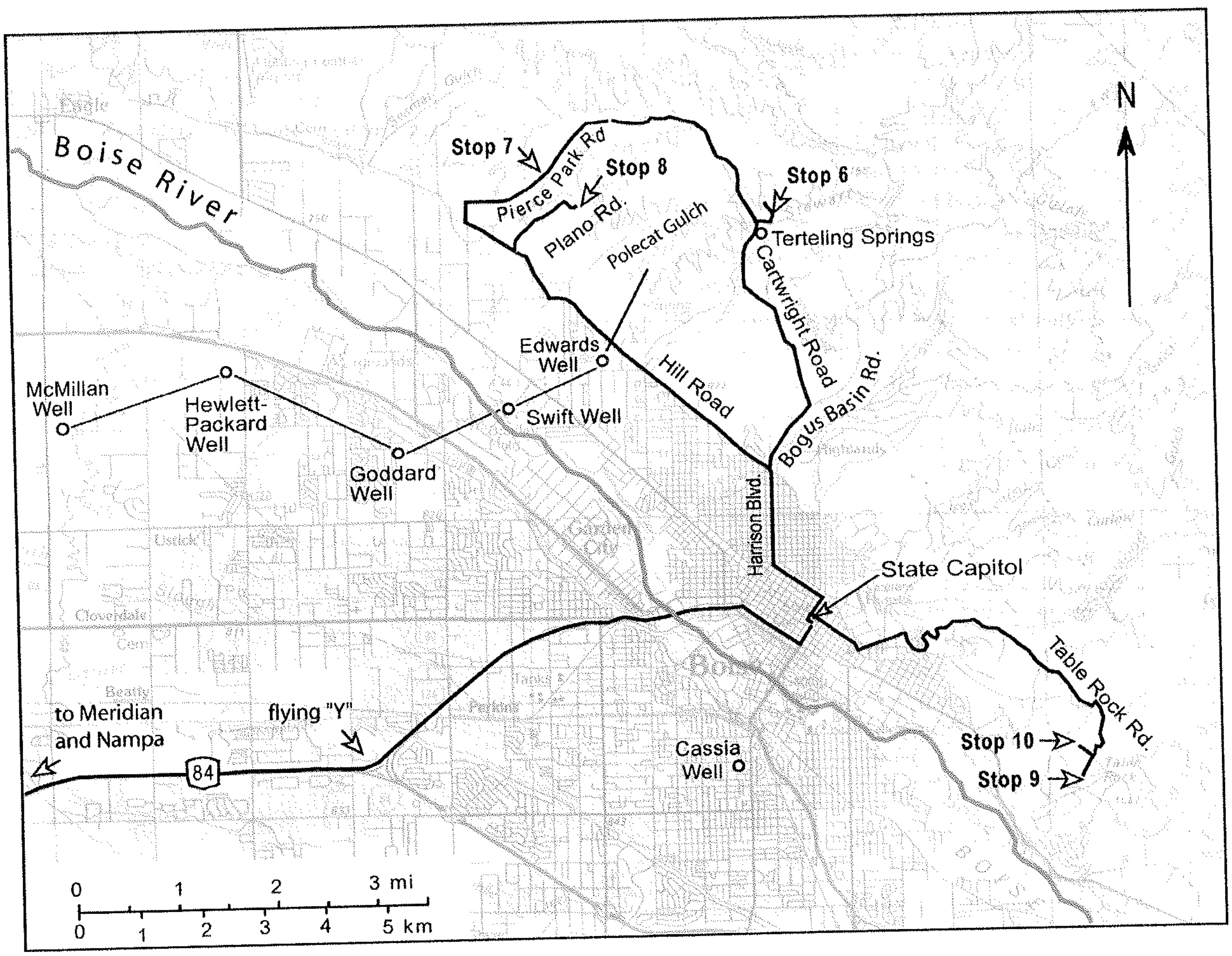


Figure 18. Map showing route of trip through the Boise foothills.

0.8 0.6 Bear right (north) and merge with Harrison Boulevard. Continue straight ahead (north) to the Hill Road intersection.

1.8 1.0 Hill Road stoplight. Continue straight ahead. Harrison Boulevard turns into Bogus Basin Road.

2.1 1.0 In the road cuts behind the buildings, on the left (west), is fine- and medium-grained sand with hummocky cross stratification (HCS) (Gallegos and others, 1987). HCS is produced by currents near the wave base, induced by storm waves. This type of cross stratification indicates near shore sedimentation, and under oceanic conditions within a 10–20-m depth (Walker and Plint, 1992).

2.6 0.5 Turn left (northwest) onto Cartwright Road. This turn is beneath the green-lawn-covered hill of the J.R. Simplot Mansion. Sediment along the road is mostly nearshore muds and sand of the Terteling Springs Formation.

3.3 0.7 Crest of hill. Ahead are hills with the conspicuous contact between mudstone of the Terteling Springs Formation and the overlying coarse Gilbert-style delta sands of the Pierce Gulch Sand (fig. 19). The mudstone is grass covered because of good moisture retention of the fine soil, whereas the Pierce Gulch Sand has only dark bitterbrush clumps with roots that reach deeper into the ground for moisture on dry sandy slopes. This vegetation contrast is seen on most south-facing slopes of these lithologies. North-facing slopes in the foothills have a substantial mantle of loess soil, 1–4-m thick, which supports bitterbrush and sage thickets.

4.6 1.3 Sediments dipping 28° W. in the road cut are inferred to be the Chalk Hills Formation. Sediments of the overlying Terteling Springs Formation, at the gate to the Owyhee Motorcycle Club, 0.2 mi to the northeast, are dipping only 12°. Generally, it is inferred that an unexposed angular unconformity exists between these units. This is the only place where existence of an angular unconformity can be demonstrated within the lacustrine sequence in the foothills. Possibly, it compares with the unconformity at Stop 5 along the Owyhee Mountains front.

4.8 0.2 Park for Stop 6.

Stop 6. Oolites and Fossil Clams of Lake Idaho, Owyhee Motorcyle Club in Stewart Gulch

Park at the entrance to the Owyhee Motorcycle Club. View the foreset beds at the gate, go through the gate, and walk around the hill to the northeast, observe the sediments in the excavated cut at the bleachers on the racetrack, and then proceed about 200 m up the gulch to the oolites. Coarse sand foreset beds of a Gilbert-style delta are overlain by Terteling Springs Formation mudstone at the entrance gate of the motorcycle club (figs. 20 and 21). This contact is interpreted as a “flooding surface” caused by a rise in lake level. Alternatively, one might argue it is the result of lobe switching of delta distributaries, but mass-failure deposits into deep water seen around the hill to the east suggest deepening of waters. The foreset-bedded sand unit at the entrance gate crops out again as cemented sandstone in the canyon of Dry Creek, about 1 mi north of here. In Dry Creek valley, this sand is one layer of foreset-bedded sandstone about 25-m (80-ft) thick. It is believed that Terteling Springs, just below the road here, is

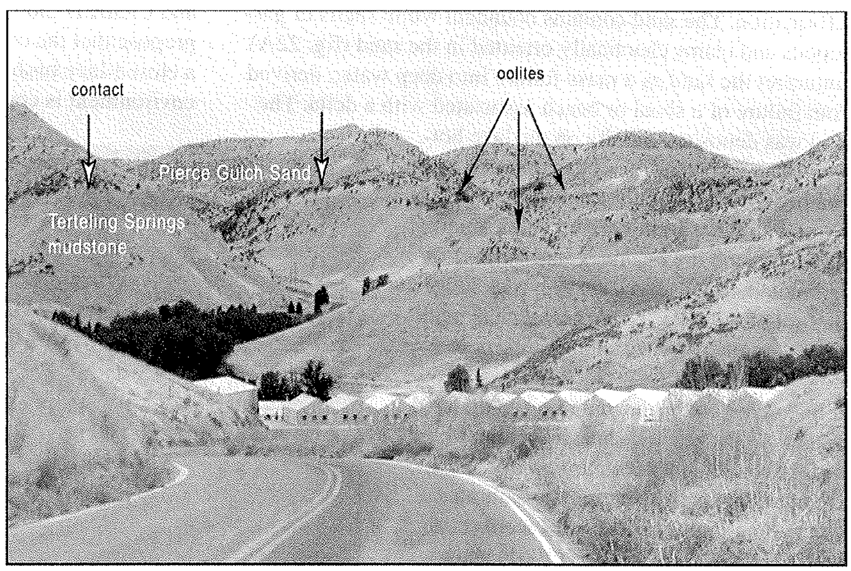


Figure 19. Contact of the Pierce Gulch Sand overlying mudstone and oolite bars of the Terteling Springs Formation. View is to the northwest from the Cartwright Road across Stewart Gulch. Greenhouses in the foreground are heated by geothermal water.

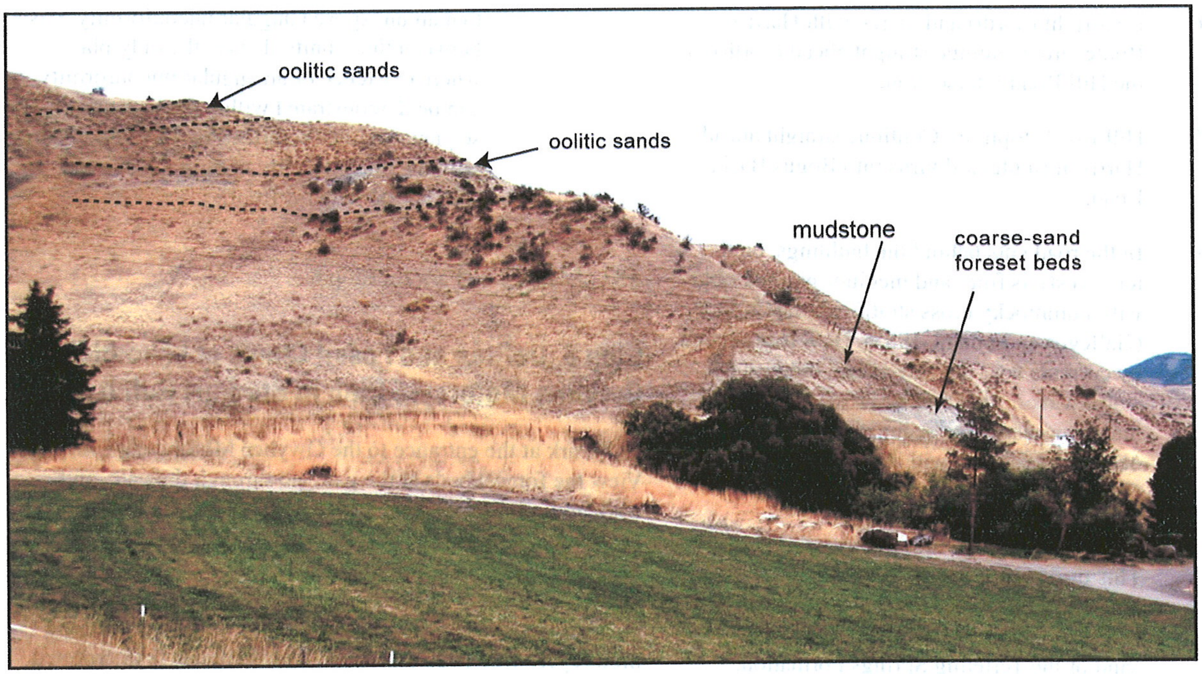


Figure 20. Coarse sand of a Gilbert delta overlain by mudstone and the oolite bars at Terteling Springs in Stewart Gulch. View is to the northeast from Cartwright Road.

discharge from this permeable sand unit intercalated within mudstones of the Terteling Springs Formation.

Walk through the gates and continue left around the hill about 200 m past the bleachers to the beer and pop stand. In the excavated cut behind the stand is a bed of coarse orange sand, over 3-m thick, and fine beds contorted by soft sediment deformation. The sand contains abundant white shells of gastropods and clams chaotically oriented in the sand (fig. 22A). I interpret the sand as a mass-failure into deep water, derived from failure of a shoal or beach associated with a delta. The sand was deposited rapidly on the lake bottom causing soft sediment deformation of the fine beds.

Continue walking up the gulch about 100 m beyond the racetrack and through an unmaintained fence to a smooth rock ledge at the foot of the hills. This rock ledge is made up of carbonate-coated sand grains, commonly called "oolites" (fig. 22B). Formation of lacustrine oolites occurs in wave-agitated waters of beaches and shoals. Origin of ooids is still debated, but precipitation of concentric microlayers from carbonate-saturated waters onto grains of fine sand seems clear here. Davaud and Giradclos (2001) show that biofilms act as a catalyst or substrate for submicron-sized calcite crystals forming on oolites in temperate, freshwater Lake Geneva (Switzerland), at depths of 1–5 m.

Several lenses of oolitic sand occur here over a stratigraphic interval of 120 m (fig. 23). On the south side of the western plain, these oolite occurrences have been interpreted as a transgressive unit at the base of the Glenns Ferry Formation (Malde and Powers, 1962; Swirydczuk and others, 1979, 1980a, 1980b, Reppening and others, 1994). Wood and Clemens (2002) agree they are a transgressive unit, and propose that the oolites are an indication that Lake Idaho was a closed-lake basin of increasing alkalinity. The closed-basin environment is confirmed by a ^{13}C isotopic analysis of del^{13}C

Figure 21. Closeup of foreset-bedded coarse sand overlain by mudstone of the Terteling Springs Formation.

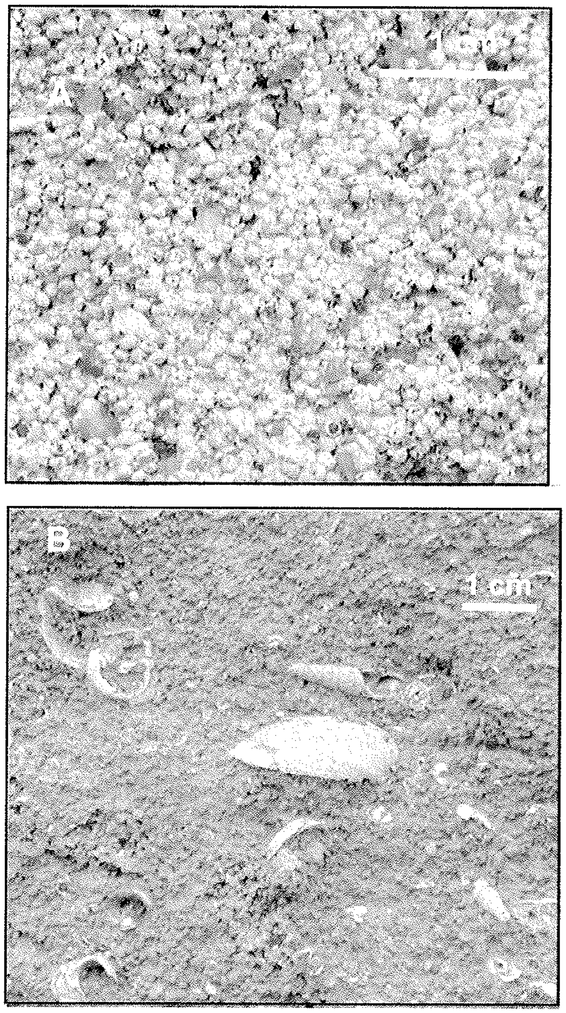


Figure 22. (A) Carbonate-coated grains (oolites), mostly about 0.8 mm in size from the outcrop at the Owyhee Motorcycle Club. (B) Fossil clams and snails in a coarse sand at the Owyhee Motorcycle Club. Shells are scattered without orientation indicating this may be a mass-wasting deposit collapsed from a delta front and not a beach deposit.

= +2.0 for carbonate in oolites collected from drill cuttings from a depth of 40 m in the Cassia Street water well in central Boise (fig. 18), about 8 km (5 mi) south of here (Cavanagh, 2000). Worldwide, the del^{13}C of closed-lake carbonates range from -2 to +5, whereas open-lake carbonate ranges from -5 to -15 (Talbot, 1990).

Mileage

Cum.	Inc.	
		Walk back to the vans parked at the entrance to the motorcycle club. Continue driving northwest on Cartwright Road.
5.8	1.0	Crest of the hill. The road traverses the contact of the Pierce Gulch Sand over the Terteling Springs Formation.
6.3	0.5	Pierce Park Road. Bear left (southwest) on Pierce Park Road.
6.7	0.5	Many springs occur at the contact of the base of the permeable Pierce Gulch Sand upon mudstone along this road, as indicated by the abundant black locust, cottonwood, and willow trees and blackberry thickets.
7.6	0.9	Mudstone of the Terteling Springs Formation with a 2-cm-thick white silicic volcanic ash. Near here, visible when the grader cleans the road cut, is a white sand hummock within the mudstone, probably storm-wave reworked sands that avalanched from a delta edge.
7.9	0.3	Pull off on the right side of road and park at the entrance to the sand quarry in the Pierce Gulch Sand. Walk down the road (south) to the road cut in Terteling Springs Formation.

Stop 7. Pierce Park Road: Pierce Gulch Sand Over Mudstone of the Terteling Spring Formation

The quarry exposes 20-m-thick foreset beds of the Pierce Gulch Sand. The sand is a Gilbert–style delta of coarse sand. Walk down the road (about 150 m) just past the high road cut in Terteling Springs Formation mudstone, on the west side of road, to see the 20-cm-thick, white silicic volcanic ash near the top of the mudstone unit. The coarsest grains at the base

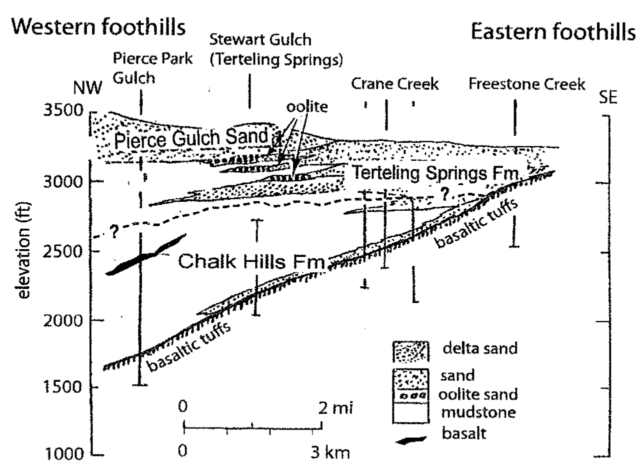


Figure 23. Stratigraphic diagram showing relation of the Chalk Hills Formation with the overlying Terteling Springs Formation and the Pierce Gulch Sand (from Wood and Clemens, 2002).

of this ash layer consist of 0.15- to 0.2-mm shards of white pumice and very light-gray glass. This size suggests it came from a volcanic source within several hundred kilometers. This relatively thick ash occurs in many places in the western foothills and is a good marker near the top of the Terteling Springs mudstone.

Mileage

Cum.	Inc.	
		After this stop, return to vans and continue down Pierce Park Road.
9.1	1.2	Pierce Park Road and Hill Road. Turn left (east) onto Hill Road.
9.7	0.6	North Plano Road and Hill Road. Turn left (north) on North Plano Road, and continue northeast. We will travel 1.4 mi up this road.
	0.1	End of pavement. Continue on North Plano Road. It is uncertain whether this is still a private road, despite the abundance of "no-trespassing notices" on the side of the road. The road serves several residences at the top of the hill. I think it now is a public county road.
	0.8	Turn around and park at the switchback, which is a posted entrance to private sand quarries. Walk back about 150 m to the road cut on the south side of the road.

Stop 8. North Plano Road: Base of the Pierce Gulch Sand

The road cut on this dirt road into the foothills is the only good exposure of the base of the Pierce Gulch Sand over mudstone of the Terteling Springs Formation. The contact is so easily mapped on air photos by contrast in vegetation and soil that one might believe it is an unconformity. However, at this locality the deposition appears continuous, and the boundary between these two different lithologic units is simply that of a coarse-sand delta prograding and downlapping over prodelta mudstone. On a large scale, when viewed from a distance, the lower sand crossbeds shallow downward in slope and appear tangential to the contact.

Geophysical logs of water wells west of here show a major delta unit in the upper section (fig. 24). At the McMillan Well, 10 km (6 mi) west of here, the delta sand thickens to 210 m (700 ft). The delta unit is built out into a deeper lake basin and is thicker there because of "greater accommodation depth".

In this road cut, a 1.5-m-thick bed of coarse sand overlies mudstone, and the sand is overlain by yet another mudstone. The main thick body of coarse sand lies just a few meters above the road cut (fig. 25). Mudstone fingers and lenses penetrate the sand, suggesting the muds were semisoft when overlain by the sand. Small sand dikes also wedge upward about 0.3 m into the overlying mudstone. The boundaries are quite sharp and appear conformable, suggesting that sand prograded rapidly over the mud bottom of the lake.

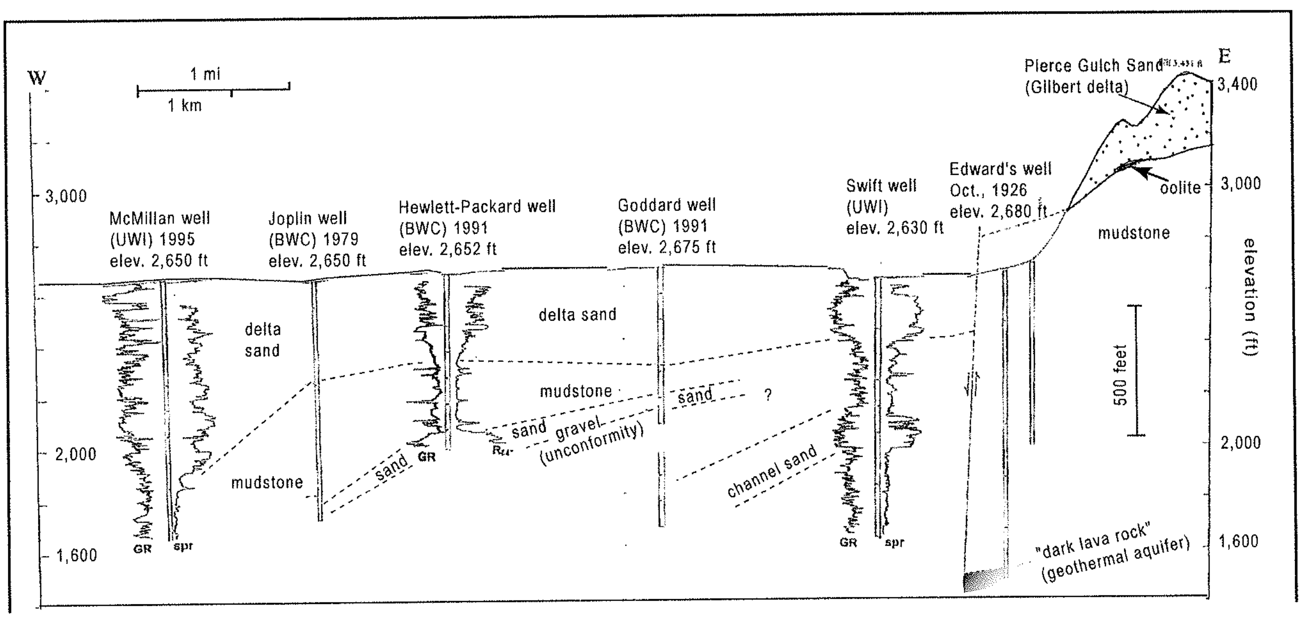


Figure 24. Stratigraphic section and well logs from Polecat Gulch in the foothills to west Boise, showing the Pierce Gulch Sand and equivalent delta facies in the subsurface. Well locations shown in figure 18. GR, natural gamma log; spr, single-point resistance log; and R_{64}, 64-inch normal-resistivity log. Well logs courtesy of United Water Idaho, Inc. (UWI), and its predecessor, Boise Water Company (BWC).

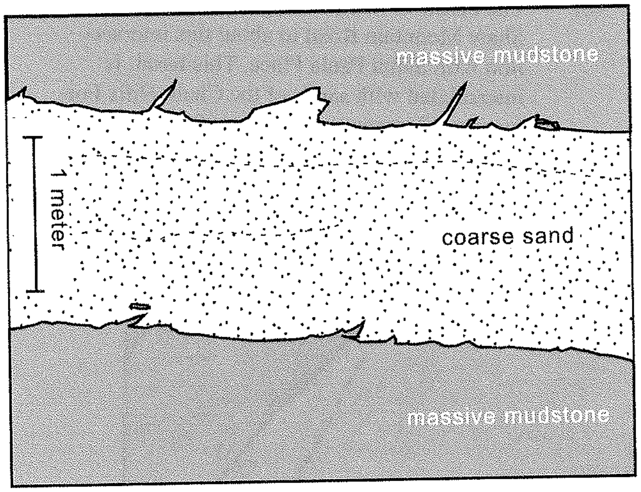


Figure 25. Tracing of photo showing an bed of coarse sand interbedded within mudstone, just beneath the base of the massive Gilbert-style delta of the Pierce Gulch Sand at North Plano Road. The sand bed is interpreted to be a sublacustrine avalanche out on to the muddy lake bottom, formed as the main foreset sand body was advancing basinward.

Mileage Cum.	Inc.	
		After viewing this stop, drive 1.4 mi back to Hill Road.
9.7	1.4	Return to the intersection of Hill Road and North Plano Road and turn left (east) on Hill Road.
11.1	1.4	Hill road makes a 90° turn to the left (east).
11.9	0.8	To the right (south) is the road to Edward's Greenhouse, a highly successful geothermally-heated greenhouse operation raising garden plants and decorative flower baskets of all sorts for the Boise market. Location of the Edward's well is shown in figure 18. The geothermal system here is of historical importance in Idaho groundwater law, because in 1931, the Idaho Supreme Court issued a landmark decree (Silkey v. Tiegs) on the doctrine of "the illegality of mining of an aquifer". The court interpreted the declining pressure and artesian-well flows of hot water to be the result of discharge exceeding recharge of an aquifer. In those situations, the rights of the junior appropriators (those owners of the later wells) to produce water are curtailed. The wells are about 350-m (1,150-ft) deep, into a "dark volcanic rock", probably a rhyolite, and produce 47° C (117° F) water (Young and others, 1988).
12.4	0.53	6th Street and Hill Road stoplight, continue southeast on Hill Road.
13.8	1.4	Bogus Basin Road and Hill Road and Bogus Basin Road (Harrison Boulevard) intersection. Turn right (south) on Harrison Boulevard, and travel 1.0 mi to where Harrison merges with the east-bound lane of Hays Street, follow east on Hays Street for 0.8 mi to 9th Street.
15.6	1.8	9th Street and Hays Street. Turn right (south) on 9th Street.
15.8	0.2	State Street and 9th Street. Turn left (east) on State Street.
15.9	0.1	Idaho State Capitol Building, 8th and State Street.

Reset mileages at the 8th Street and State Street intersection.

EASTERN BOISE FOOTHILLS LEG OF FIELD TRIP

Mileage Cum.	Inc.	
0.0	0.0	This leg of the trip starts at the intersection of 8th Street and State Street, the northwest corner of the State Capitol Building grounds. Travel east on State Street.
0.1	0.1	Intersection of Fort Street and State Street. Bear right (east) onto Fort Street.
0.4	0.3	Turn left (north) on Reserve Street.
0.6	0.2	Intersection of Mountain Cove Road and Reserve Street, continue north on Reserve Street.
		The dike on the northwest (left) side of the street was built in 1998 to increase the volume of the sediment retention reservoir where Cottonwood Creek flows from the foothills. The September 1996 foothills fire focused awareness on potential debris flows out of the foothills gulches, particularly after major fires in the watersheds.
0.9	0.3	Main road bears right and uphill (east) and changes name to Shaw Mountain Road at this intersection with San Felipe Way. Three hundred feet northwest along a gated gravel

road is the Boise Geothermal, Inc. Well No. 1, drilled to a depth of 634 m (2,080 ft). Idaho batholith granite is at a depth of 512 m (1,680 ft) in this well (Burnham and Wood, 1983).

1.0 0.1 The basalt of Aldape Heights crops out along Shaw Mountain Road to about this intersection with Santa Paula Place. This basalt is intercalated with sands of the Chalk Hills Formation. Clemens and Wood (1993) obtained a whole-rock K-Ar age of 9.5±0.6 Ma. They

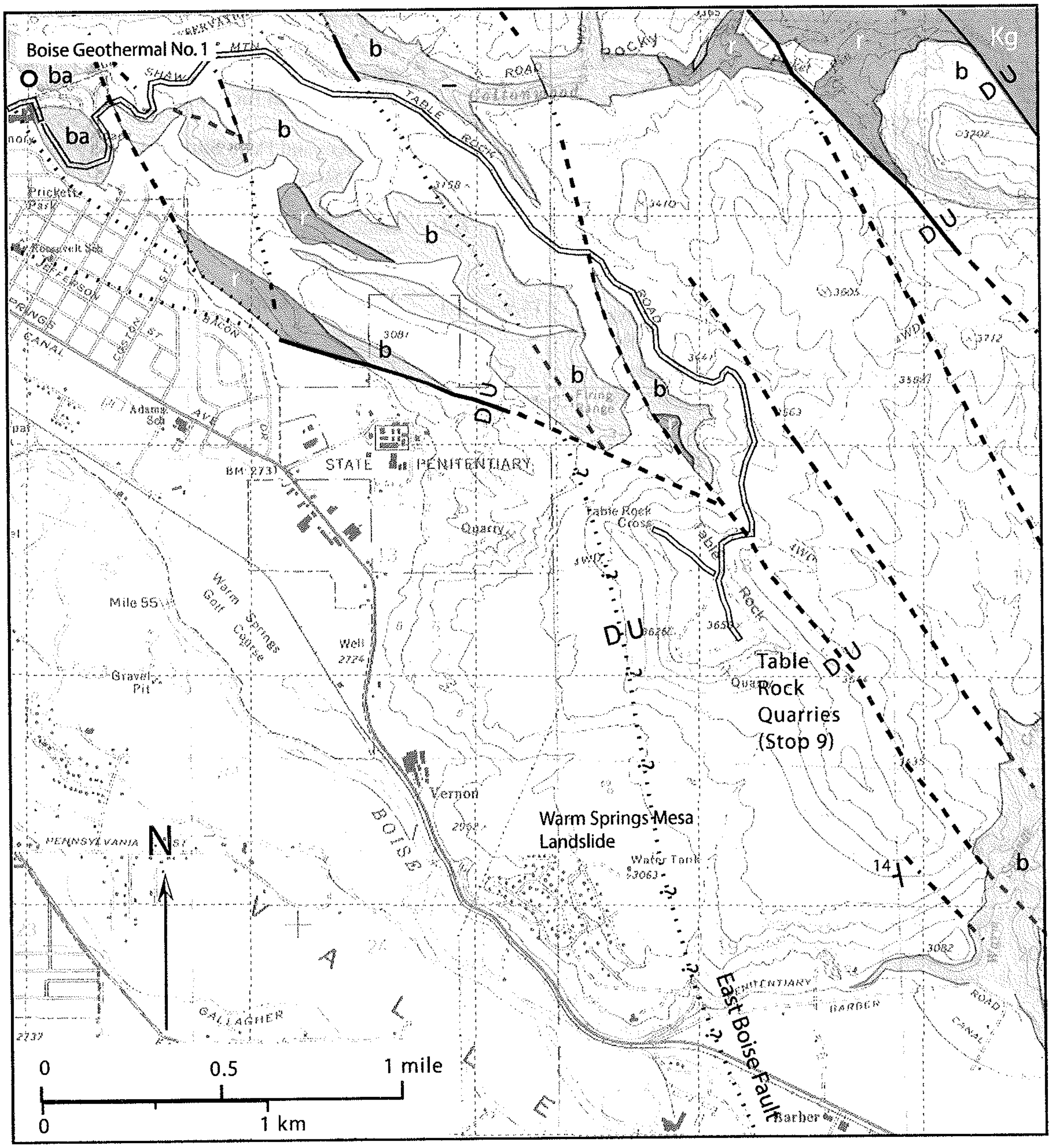


Figure 26. Geologic map showing the route to Table Rock. The Table Rock area is a downfaulted block, with respect to the foothills on the north. Because it is structurally lower, it contains one of the more complete stratigraphic sections in the foothills of 210 m (700 ft) of fluvial and lacustrine sediment over the Miocene rhyolite and basalt. Symbols on map: Kg, granodiorite of the Idaho batholith; r, Quarry View Park and Cottonwood Creek rhyolite (11.8±0.6 and 11.3±0.3 Ma, respectively); b, basalt and basaltic tuffs; ba, basalt of Aldape Park; and unshaded areas are fluvial and lacustrine sediment (after Clemens and Wood, 1993).

show that this rock petrochemically matches basalt that occurs in the Capitol Mall Geothermal wells by the Statehouse, at a depth of about 223 m (730 ft), indicating vertical separation due to dip of beds and offset along faults is about 260 m (850 ft) between the outcrop and the wells by the Statehouse.

2.0 1.0 Turn right (east) on Table Rock Road at this intersection of Shaw Mountain Road and Table Rock Road.

4.0 2.0 Bear right (south) on Table Rock Road at this intersection with E. Wildhorse Lane (private lane).

4.2 0.2 Gate unlocked between 1 hour before sunrise and 1 hour after sunset. This point is in a saddle between hills and is the location of the normal fault (fig. 26) that down drops the Table Rock section of sediments to the southeast relative to the foothills to the north.

4.5 0.3 Park for Stop 9.

Stop 9. Table Rock Quarries

Park here and procede 0.25 mi to the southeast along a deeply-rutted road to the rim of the mesa and the quarries for the "Boise Sandstone".

Table Rock is a mesa capped by silica-cemented sandstone, forming a prominent landmark above east Boise. The mesa is comprised of younger lake and stream sediments faulted down relative to the northern foothills, about 300 m (1,000 ft), thus preserving a 200-m (700-ft) section of sedimentary strata (Wood and Burnham, 1987; Clemens and Wood, 1993) (fig. 26). The 15-m-thick sandstone layer at the top (fig. 27) has been quarried for over 100 yr and widely marketed as a dimension stone. It is the stone of which the Idaho State Capitol Building was constructed in 1920. This "Boise Sandstone" also is a standard for many rock mechanics and petroleum reservoir experiments (Wong and others, 1997). Porosity of 0.27 and a permeability of 910 mD (9.6×10^{-4} cm/s, hydraulic conductivity) are reported by Kovscek and others (1995).

The massive character of this sandstone makes it a good stone for sculptors and for dimension stone. I have puzzled over how a 15-m-thick layer of sand, of such uniform grain size ($D_{10} = 0.2$, $D_{50} = 0.35$, $D_{95} = 0.7$ mm), was deposited. Gallegos and others (1987) suggested it was a gravity-driven mass flow beneath the lake based on the character of its basal scour surface. Above the massive sandstone is 5 m of bedded sandstone with 0.5 mm coated grains (ooids) and rip-up slabs of coarse sandstone mixed with ooids (fig. 27 and 28). Above the bedded sand is thin-bedded mudstone and uncemented sand and sandy gravel.

Silica cementation of the sand occurred as silica-saturated geothermal waters percolated through permeable beds and cooled allowing the dissolved silica to precipitate. Silica-cemented sandstone around the western plain commonly occurs within several kilometers of fault systems. The faults were conduits for upward flow of silica-bearing hot water from depths of a few kilometers.

From the quarries is the best view of the Pleistocene terrace sequence of the Boise River Valley (fig. 29). Othberg (1994) established a chronology for the four terraces observed

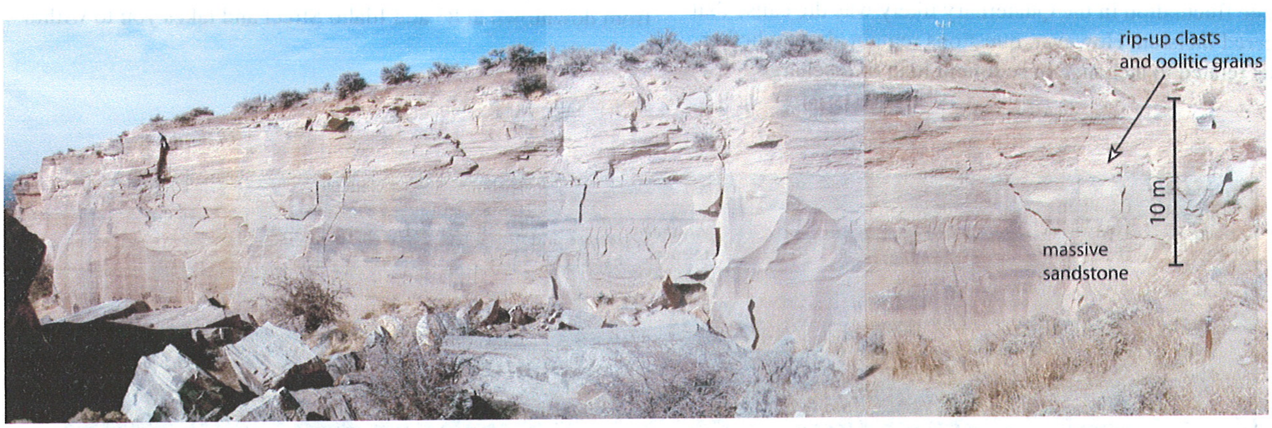


Figure 27. Massive medium-grained sandstone at the Table Rock Quarries. This 15-m-thick layer is the "Boise Sandstone" sold throughout the United States as a dimension stone during the early 20th Century. The layer can be traced laterally for at least 800 m. Sedimentary architecture of this sand has not been studied, and its origin remains uncertain. It is overlain by bedded sands that contain up to 30 percent coated grains, that look very much like carbonate coated oolites, but do not effervesce. The sands also contain rip-up clasts of oolite-bearing coarse sand (fig. 28). The upper bedded sands display low-angle, tangential cross stratification. Above the sandstone is mudstone and pebbly sands.

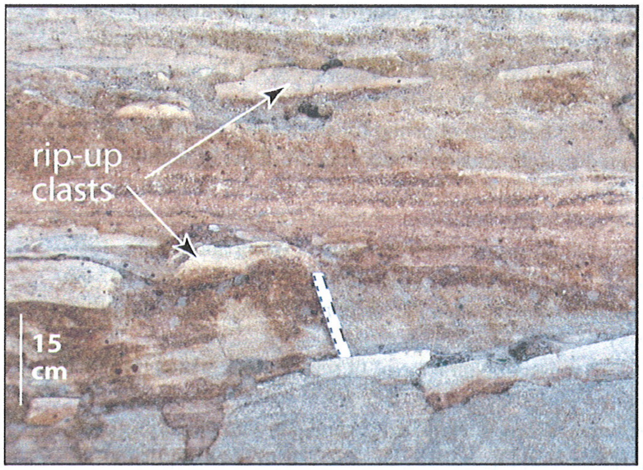


Figure 28. Rip-up clasts (light colored) of cemented coarse sand with 0.5 mm coated grains within bedded coarse and small-pebble sandstone. Locality shown in figure 27 at the Table Rock Quarries.

here by dating the basalt layers that lay upon the terrace surfaces. He also recognized the much older (but undated) Bonneville gravels on the southeastern skyline of figure 29. The terraces are typically overlain by 15 to 30 m of gravel on a strath surface of eroded lacustrine sand or mudstone (Squires, 1992). Downward erosion of the Boise Valley leaving this sequence of abandoned flood plains was in response to lowering of base level and related downcutting by the Snake River. During the late Pliocene and the Quaternary, the Snake River was downcutting the entrance to upper Hells Canyon as shown by the "base-level decline" of figure 3. This is a good place to address the question of why rivers incise episodically leaving broad terrace remnants. Othberg (1994) indicated that climate fluctuation in the Quaternary likely was the cause that triggered incision. However, questions remain about whether it was related to a change in sediment load or discharge characteristics associated with glaciation in the headwaters of the Boise River or related to overall climate change. Pazzaglia and Brandon (2001) present a model for coastal streams where they associate fill and flood plain deposition with the increased load accompanying deglaciation. In this model, strath incision occurs during periods of alpine glacial advance. Their study of coastal streams, however, also is complicated by sea-level fluctuation. At this time, the factors tht controlled or triggered downcutting and terrace formation in this region are unknown.

Mileage Cum.	Inc.	
		After viewing the quarries return to the parked vans and proceed west toward the group of radio towers and continue on to the parking area at the large white cross.
4.7	0.2	Park for Stop 10.

Stop 10. Table Rock Cross Viewpoint: View to the West Over Boise and the Western Plain

This final stop of the trip looks west over the city of Boise (fig. 30). Directly below us are the grounds of the 1860 Idaho Territorial Penitentiary and the Boise Warm Springs geothermal area. The prison buildings were constructed by prisoners from sandstone quarried from the hill below. The area now is a public park and houses the Idaho Museum of Mining and Geology. From this area, one can embark on short hikes on well-marked trails into the foothills. City bus service along Warm Springs Avenue leaves every 30 minutes from downtown at 8th and Idaho Street and takes you to within

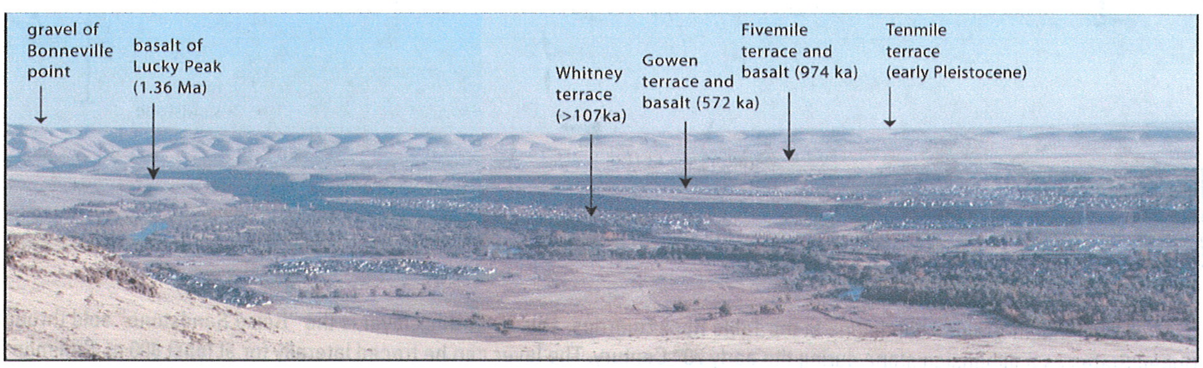


Figure 29. View to the southeast from the Table Rock Quarries. Othberg (1994) obtained Pleistocene ages on the several basalt units that flowed over the successively lower braided floodplains of the Boise River. The early Pleistocene Tenmile terrace surface forms the skyline. This surface is 150 m (500 ft) above the modern Boise River.

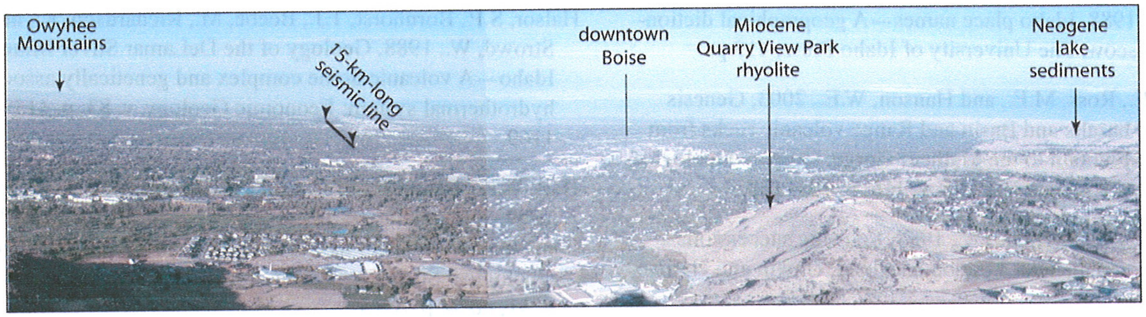


Figure 30. View to the west over the Boise Valley and the western Snake River Plain. Building area at the bottom of the hill, beneath this viewpoint, are the grounds of the 1860 Idaho Territorial Penitentiary and the Boise Warm Springs area.

300 m (0.2 mi) of the park on weekdays; however, there is currently no weekend service.

The outcrop of the Quarry View Park rhyolite is described by Wood and Burnham (1987) and shown by the arrow in fig. 30. Clemens and Wood (1993) report a K-Ar age of the rhyolite of 11.8±0.6 Ma on andesine plagioclase separates. The rhyolite is overlain by basalt, which is in turn overlain by a ledge of cemented conglomeritic sandstone. The overlying lacustrine and fluvial sediments dip generally west-southwest 4–10°. Several northwest-trending normal faults trace through the foothills and beneath downtown Boise (Liberty, 1998). Despite our high elevation at the Table Rock viewpoint, 1,108 m (3,636 ft), we are on a down-thrown block relative to the hill below containing the Miocene rhyolite (figs. 26 and 30). Throw on the fault between the rhyolite and the flat area below is about 210 m (700 ft)(Wood and Burnham, 1987).

All of the state office buildings, many commercial buildings in downtown Boise, and many of the older homes in northeast Boise are heated by hot water from wells. The original two hot water wells were drilled 123-m (404-ft) deep in the warm springs area by the Penitentiary in the early 1900s. They initially had an artesian flow of 351 l/s (550 gallons/minute) of 77° C (170° F) water, but pressure has declined, and the wells must be pumped. The Warm Springs Water District is the oldest geothermal heating district in the United States. The State of Idaho, the U.S. Veterans Hospital, and the city of Boise since the 1980s, produce water from wells 600–920-m (2,000–3,000-ft) deep beneath downtown Boise and along the edge of the foothills. A reinjection well was drilled in 1994 near the State Historical Museum, 5 blocks south of the Convention Center to a depth of 975 m (3,200 ft). Location was based on a seismic reflection survey by Liberty (1998) that imaged the faulted basalt above the rhyolite. Flow from the rhyolite aquifers below 670 m (2,200 ft) was about 57 l/s (900 gallons per minute) of 77° C (170° F) water, and these aquifers have a shutin artesian hot-water-column head of 13 m (43 ft) above ground level. It is the most successful production well, but due to its location at the tail end of the building-heat-circulation system, it is used as a reinjection well. The geothermal aquifer system is in the same fractured rhyolite rock exposed on the upthrown fault block in the foothills below. Production is just now increasing for the city of Boise system to serve new buildings, but it is uncertain whether aquifer pressure and temperature will sustain continued exploitation in this downtown area. The area south of the Boise River has never been explored by deep wells. It is likely the downfaulted aquifer is deeper and hotter immediately to the south.

This is the end of the trip, and vans will return back to the Convention center.

References

Amini, H., Hehnert, H.H., and Obradovich, J.D., 1984, K-Ar ages of late Cenozoic basalts from the western Snake River Plain, Idaho: Isochron/West, v. 41, p. 7–11.

Baksi, A.K., 2004, Discussion and reply: Ages of the Steens and Columbia River flood basalts and their relationship to extension related calc-alkalic volcanism in eastern Oregon: Geological Society of America Bulletin, v. 116, p. 247–250.

Bonnichsen, B., McCurry, M., and Godchaux, M.M., 2004, Miocene Snake River Plain rhyolites of the Owyhee front, Owyhee County, Idaho: this volume.

Bonnichsen, B., White, C.M., and Godchaux, M.M., 1997, Basaltic volcanism, western Snake River Plain, *in* Link, P.K., and Kowallis, B.J., eds., Proterozoic to recent stratigraphy, tectonics and volcanism in Utah, Nevada, southern Idaho and central Mexico: Brigham Young University Geology Studies, v. 42, Part 1, p. 399–422.

Bonnichsen, B., and Kauffman, D.F., 1987, Physical features of rhyolite lava flows in the Snake River Plain volcanic province, southwestern Idaho, *in* Fink, J.H., ed., The emplacement of silicic domes and lava flows: Geological Society of America Special Paper 212, p. 119–145.

Burnham, W.L., and Wood, S.H., 1983, Field Trip Guide, Boise Front Geothermal Area: 12th Annual Rocky Mountain Groundwater Conference, Boise, Idaho, 36 p.

Boone, L. 1988, Idaho place names—A geographical dictionary: Moscow, The University of Idaho Press, 413 p.

Camp, V.E., Ross, M.E., and Hanson, W.E., 2003, Genesis of flood basalts and Basin and Range volcanic rocks from Steens Mountain to the Malheur Gorge, Oregon: Geological Society of America Bulletin, v. 115, p. 105–128.

Cas, R.A.F., and Wright, J.V., 1988, Volcanic successions—Modern and ancient: London, England, Unwin and Hyman, 528 p.

Clemens, D.M., and Wood, S.H., 1993, Radiometric dating, volcanic stratigraphy, and sedimentation in the Boise foothills, northeastern margin of the western Snake River Plain, Ada County, Idaho: Isochron/West, v. 59, p. 3–10.

Clough, B.J., Wright, L.V., and Walker, G.P.L., 1982, An unusual bed of giant pumice in Mexico: Nature, v. 289, p. 49–50.

Cavanagh, B.C., 2000, Western Snake River Plain, Idaho—Fluvial-lacustrine sediments, exhumation estimates from mudstone compaction, unconformity identification by buried soil carbonate, hydraulic conductivity estimates from well cuttings: Boise, Idaho, Boise State University, unpublished M.S. thesis, 96 p.

Davaud, E., and Girardclos, S., 2001, Recent freshwater ooids and oncoids from western Lake Geneva (Switzerland)—Indications of a common organically mediated origin: Journal of Sedimentary Research, v. 71/3, p. 423–429.

Decker, P.L., 1990, Style and mechanics of liquefaction-related deformation, lower Absaroka Volcanic Supergroup (Eocene), Wyoming: Geological Society of America Special Paper 240, 71 p.

Dixon, M.D., and Johnson, W.C., 1999, Riparian vegetation along the middle Snake River, Idaho—Zonation, geographical trends and historical changes: Great Basin Naturalist, v. 59, p. 18–34.

Ekren, E.B., McIntyre, D.H., Bennett, E.H., and Marvin, R.F., 1982, Cenozoic stratigraphy of western Owyhee County, Idaho, *in* Bonnichsen, B., and Breckenridge, R.M., eds., Cenozoic geology of Idaho: Idaho Geological Survey Bulletin 26, p. 215–235.

Ekren, E.B., McIntyre, D.H., Bennett, E.H., and Malde, H.E. 1981, Geologic map of Owyhee County, Idaho, west of longitude 116° W: U.S. Geological Survey Map I-1256, scale 1:125,000.

Ekren, E.B., McIntyre, D.H., Bennett, E.H., 1984, High-temperature, large-volume, lava-like ash-flow tuffs without calderas: U.S. Geological Survey Professional Paper 1272, 76 p.

Gallegos, D., Johnson, P., Wood, S.H., and Snyder, W., 1987, Depositional facies along the Boise front: Northwest Geology, v. 16, p. 47–59.

Halsor, S.P., Bornhorst, T.J., Beebe, M., Richardson, K., and Strowd, W., 1988, Geology of the DeLamar Silver Mine, Idaho—A volcanic dome complex and genetically associated hydrothermal system: Economic Geology, v. 83, p. 1159–1169.

Hart, W.K., Brueseke, M.E., Renne, P.R. and McDonald, H.G., 1999, Chronostratigraphy of the Pliocene Glenns Ferry Formation, Hagerman Fossil Beds National Monument, Idaho: Geological Society of America Abstracts with Programs, v. 31, no. 4, p. A-15.

Hooper, P.R., Binger, G.B., and Lees, K.R., 2002a, Age of the Steens and Columbia River flood basalts and their relationship to extension-related calc-alkalic volcanism in eastern Oregon: Geological Society of America Bulletin, v. 114, p. 43–50.

_______, 2002b, Erratum—Ages of the Steens and Columbia River flood basalts and their relationship to extension-related calc-alkalic volcanism in eastern Oregon:Geological Society of America Bulletin, v. 114, p. 923-924.

Kovscek, A.R., Patzek, T.W., and Radke, C.J., 1995, A mechanistic population balance model for transient and steady-state foam flow in Boise sandstone: Chemical Engineering Science, v. 50, p. 3783–3799.

Liberty, L., 1998, Seismic reflection imaging of a geothermal aquifer in an urban setting: Geophysics, v. 63, p. 1285–1295.

Lofgren, G., 1971, Spherulitic textures in glassy and crystalline rocks: Journal of Geophysical Research, v. 76, p. 5635–5648.

Malde, H.E., and Powers, H.A., 1962, Upper Cenozoic stratigraphy of the western Snake River Plain, Idaho: Geological Society of America Bulletin, v. 73, p. 1197–1220.

Neill, W.M., 1975, Geology of the southeastern Owyhee Mountains and environs, Owyhee County, Idaho: Stanford, California, Stanford University, unpublished M.S. thesis, 59 p.

O'Conner, J.E., 1993, Hydrology, hydraulics, and geomorphology of the Bonneville flood: Geological Society of America Special Paper 274, 83 p.

Ostercamp, W.R., Johnson, W.C., and Dixon, M.D., 2001, Biophysical gradients related to channel islands, middle Snake River, Idaho, *in* Dorava, J.M., Montgomery, D.R., Palcsak, B.B., and Fitzpatrick, F.A., eds., Geomorphic processes and river habitat: American Geophysical Union, Water Science and Application Volume 4, p. 73–83.

Ostercamp, W.R., 1998, Processes of fluvial–island formation, with examples from Plum Creek, Colorado and Snake River, Idaho: Wetlands, v. 18, p. 530–535.

Othberg, K.L., 1994, Geology and geomorphology of the Boise Valley and adjoining areas, western Snake River Plain, Idaho: Idaho Geological Survey Bulletin 29, 54 p.

Othberg, K.L., Bonnichsen, B., Swisher, C.C., III, and Godchaux, M.M., 1995, Geochronology and geochemistry of Pleistocene basalts of the western Snake River Plain and Smith Prairie, Idaho: Isochron/West, v. 62, p. 16–29.

Pazzaglia, F.J., and Brandon, M.T., 2001, A fluvial record of long-term steady-state uplift and erosion across the Cascadia forearc high, western Washington State: American Journal of Science, v. 301, p. 385–431.

Perkins, M.E., Brown, F.H., Nash, W.P., McIntosh, W., and Williams, S.K., 1998, Sequence, age, and source of silicic fallout tuffs in middle to late Miocene basins of the Basin and Range province: Geological Society of America Bulletin, v. 110, p. 344–360.

Perkins, M.E., and Nash, B.P., 2002, Explosive silicic volcanism of the Yellowstone hotspot—The ash-fall tuff record: Geological Society of America Bulletin, v. 14, p. 367–381.

Pierce, K.L., and Morgan, L.A., 1992, The track of the Yellowstone hot spot—Volcanism, faulting, and uplift, *in* Link, P.K., Kuntz, M.A., and Platt, L.B., eds., Regional geology of eastern Idaho and western Wyoming: Geological Society of America Memoir 179, p. 1–53.

Repenning, C.A., Weasma, T.R., and Scott, G.R., 1994, The early Pleistocene (latest Blancan-earliest Irvingtonian) Froman Ferry fauna and history of the Glenns Ferry Formation, southwestern Idaho: U.S. Geological Survey Bulletin 2105, 86 p.

Squires, E., 1992, Hydrogeologic framework of the Boise aquifer system, Ada County, Idaho: Boise, Idaho, Boise State University, unpublished M.S. thesis, (also Research Technical Completion Report 14-08-0001-G1559-06, Idaho Water Resources Institute, Moscow, Idaho, University of Idaho), 114 p.

Squires, E., Liberty, L.M., and Wood, S.H., 2003, Hydrostratigraphic characterization and Quaternary/Neogene history of Boise and Meridian, Idaho using drill-cuttings analysis, borehole geophysical logs, and high resolution seismic images: Geological Society of America Abstracts with Programs, v. 35, no. 6, p. 571.

Swirydczuk, K., Wilkinson, B.H., and Smith, G.R., 1979, The Pliocene Glenns Ferry oolite—Lake-margin carbonate deposition in the southwestern Snake River Plain: Journal of Sedimentary Petrology, v. 49, p. 995–1004.

———, 1980a, The Pliocene Glenns Ferry oolite—Lake-margin carbonate deposition in the southwestern Snake River Plain—Reply: Journal of Sedimentary Petrology, v. 50, p. 999–1001.

———, 1980b, The Pliocene Glenns Ferry oolite-II—Sedimentology of oolitic lacustrine terrace deposits: Journal of Sedimentary Petrology, v. 50, p. 1237–1248.

Talbot, M.R., 1990, A review of the paleohydrological interpretation of carbon and oxygen isotopic ratios in primary lacustrine carbonates: Chemical Geology, v. 80, p. 261–279.

Taubeneck, W.H., 1971, Idaho batholith and its southern extension: Geological Society of America Bulletin, v. 82, p. 1899–1928.

Terzaghi, K., and Peck, R.B., 1967, Soil mechanics in engineering practice: New York, Wiley and Sons, 729 p.

Thompson, R.S., 1991, Pliocene environments and climates of the Western United States: Quaternary Science Reviews, v. 10, p. 115–132.

Van Domelen, D.J., and Rieck, H.J., 1992, Paleomagnetic polarity of some vertebrate fossil localities of the Glenns Ferry Formation in the Chalk Hills, near Froman Ferry, western Snake River Plain, southwest Idaho: U.S. Geological Survey Open-File Report 92-542, 6 p.

Young, H.W., Parliman, D.J., and Mariner, R.H., 1988, Chemical and hydrologic data for selected thermal-water wells and non-thermal springs in the Boise area, southwestern Idaho: U.S. Geological Survey Open-File Report 88-471, 35 p.

Walker, R.G., and Plint, A.G., 1992, Wave-and-storm-dominated shallow marine systems, *in* Walker, R.G., and James, N.P., eds., Facies models—Response to sea level changes: Geological Association of Canada, p. 219–238.

White, C.M., Hart, W.K., Bonnichsen, B., and Matthews, D., 2004, Geochemical and Sr-isotopic variations in western Snake River Plain basalts, Idaho, *in* Bonnichsen, B., White, C.M., and McCurry, M., eds., Tectonic and magmatic evolution of the Snake River Plain Volcanic Province: Idaho Geological Survey Bulletin 30, 14 p.

Wong, T.F., David, C., and Zhu, W., 1997, The transition from brittle faulting to cataclastic flow in porous sandstone—Mechanical deformation: Journal of Geophysical Research, v. 102, p. 3009–3025.

Wood, S.H., 1994, Seismic expression and geological significance of a lacustrine delta in Neogene deposits of the western Snake River Plain, Idaho: American Association of Petroleum Geolgists Bulletin, v. 78, p. 102–121.

Wood, S.H., and Clemens, D.L., 2002, Geologic and tectonic history of the western Snake River Plain, Idaho and Oregon, *in* Bonnichsen, B., White, C.M., and McCurry, M., eds., Tectonic and magmatic evolution of the Snake River Plain Volcanic Province: Idaho Geological Survey Bulletin 30, p. 69–103.

Wood, S.H., and Burnham, W.L., 1987, Geological framework of the Boise Warm Springs geothermal area, Idaho, *in* Buess, S.S., ed., Rocky Mountain Section of the Geological Society of America, Centennial Field Guide, v. 2, p. 117–122.

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

Spencer H. Wood[1] and Drew M. Clemens[2]

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

---

Editors' note: The manuscript was submitted in July 1998 and has been revised at the authors' discretion.

[1]Department of Geosciences, Boise State University, Boise, ID 83725
[2]U.S. Army Corps of Engineers, CENAE-EP-HC, 696 Virginia Road, Concord, MA 01742

Wood, S.H., and D.M. Clemens, 2002, Geologic and tectonic history of the western Snake River Plain, Idaho and Oregon, *in* Bill Bonnichsen, C.M. White, and Michael McCurry, eds., Tectonic and Magmatic Evolution of the Snake River Plain Volcanic Province: Idaho Geological Survey Bulletin 30, p. 69-103.

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 braidplain 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. 70° 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. 70° W., contrasting with the N. 45° W. boundary of the plain and the N. 30° W. alignment of vents in the eastern plain. This N. 70° 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-

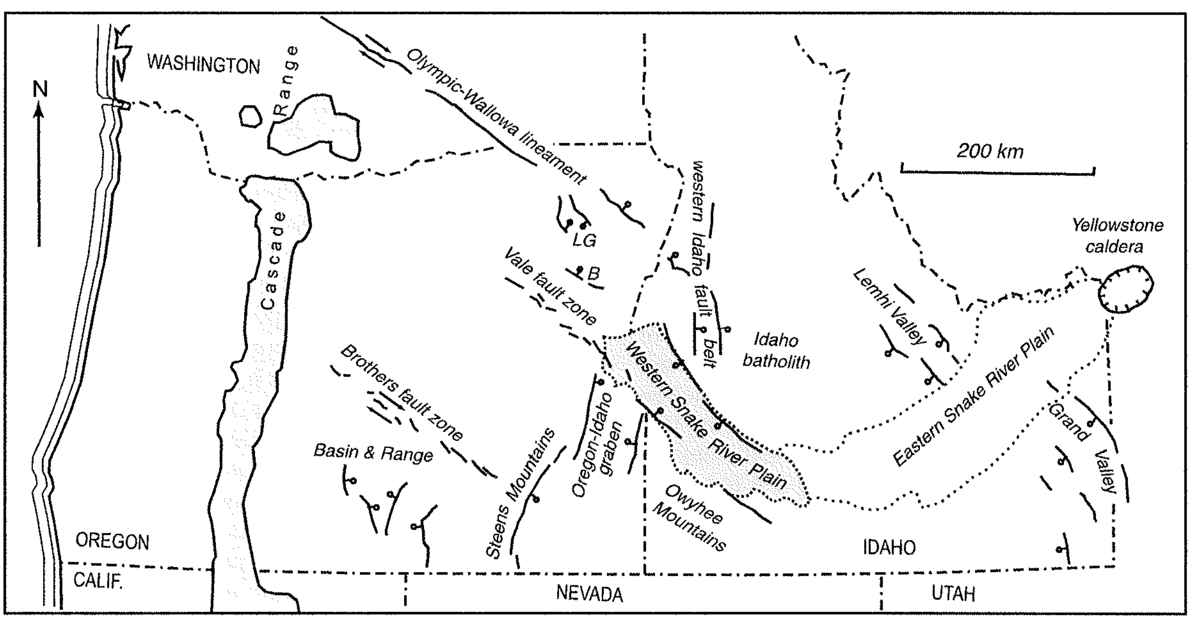


Figure 1. Regional setting of the western Snake River Plain showing related geologic features and emphasizing northwest-trending, late Cenozoic fault structures in Oregon and Idaho. LG—La Grande graben, B—Baker Valley.

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. 40° W.). Superposed on this is a narrow (30-km-wide) feature of about +25 milligals that trends obliquely (N. 70° 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.

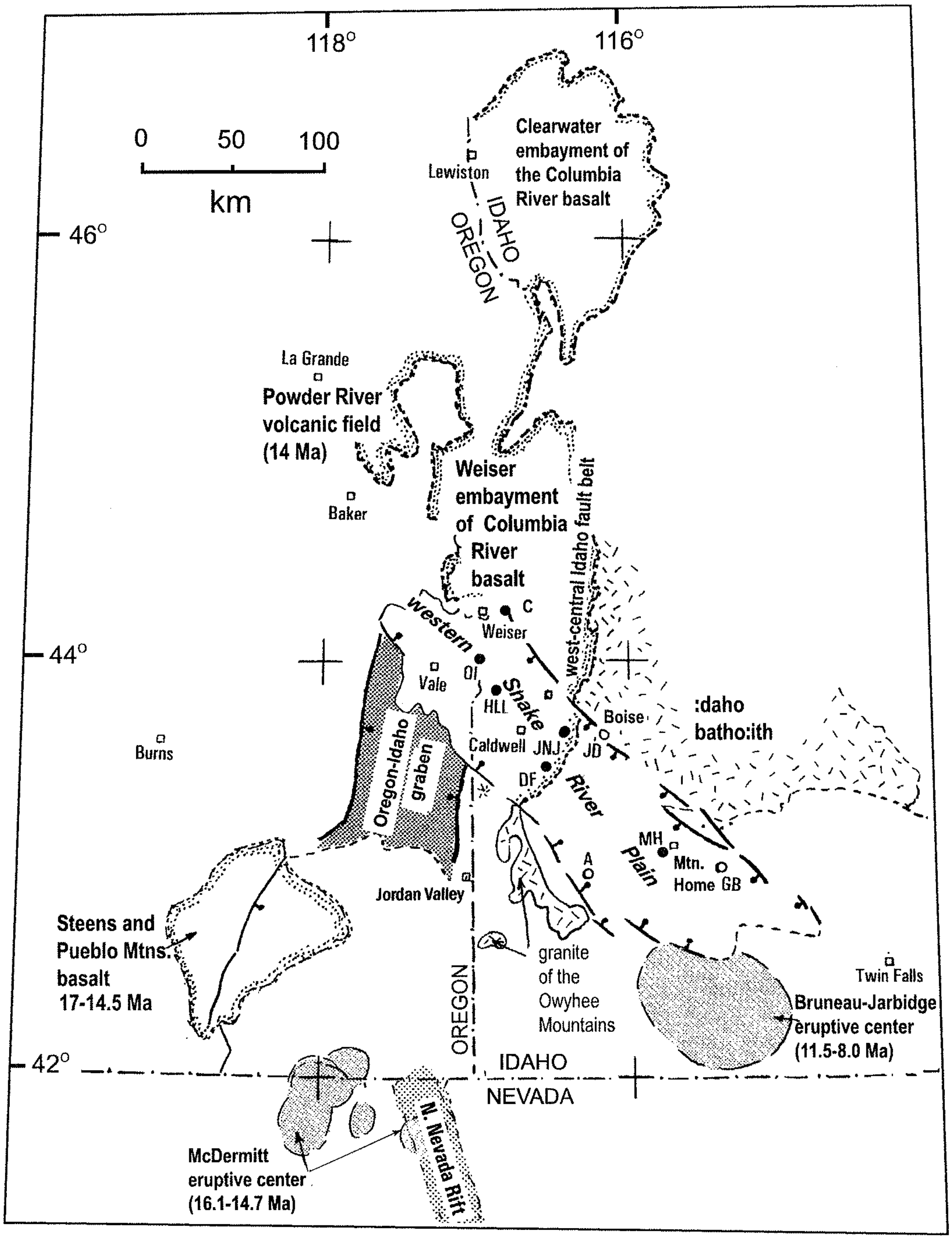


Figure 2. Map showing tectonic features surrounding the western Snake River Plain and locations of deep wells into volcanic basement. Wells only in deep basalt are shown with a solid circle: C—Chrestesen No. A-1; OI—Ore-Ida Foods No. 1; HLL—Highland Land and Livestock No. 1; JNJ—J.N. James No. 1; DF—Deer 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

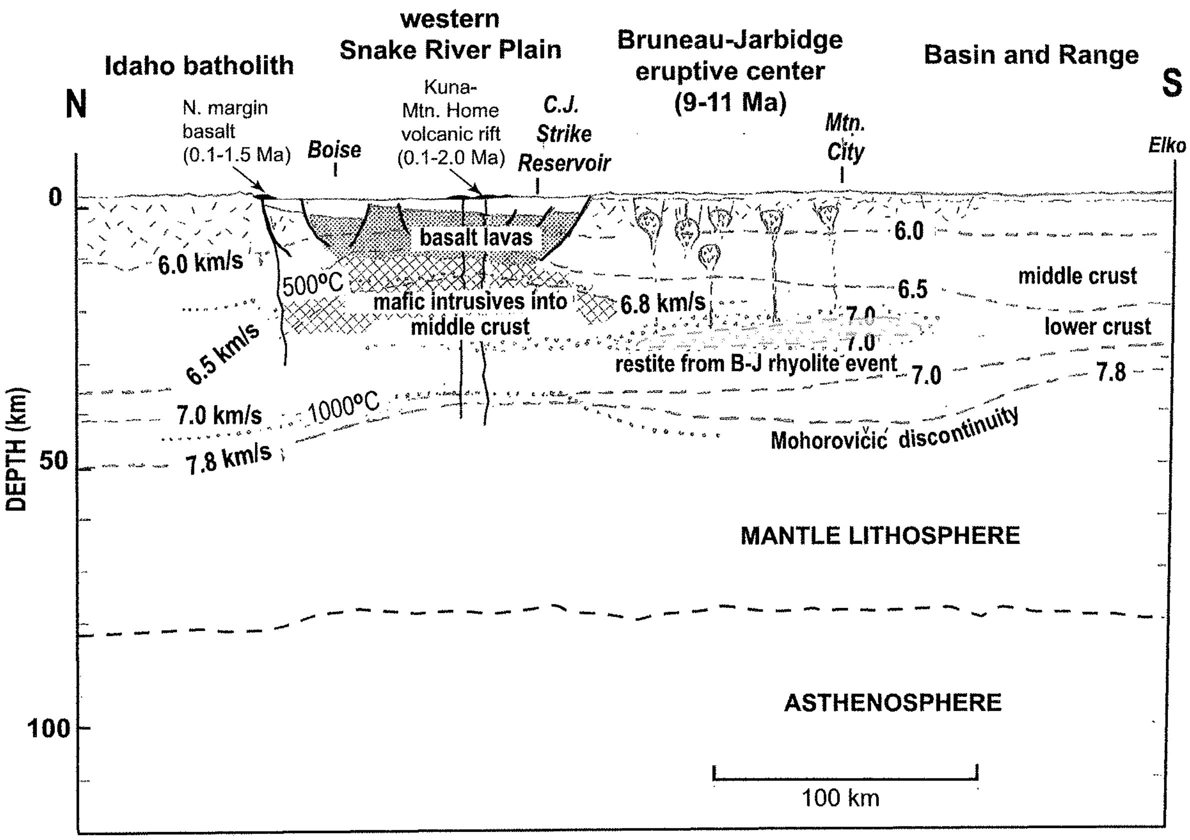


Figure 3 (A)

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.

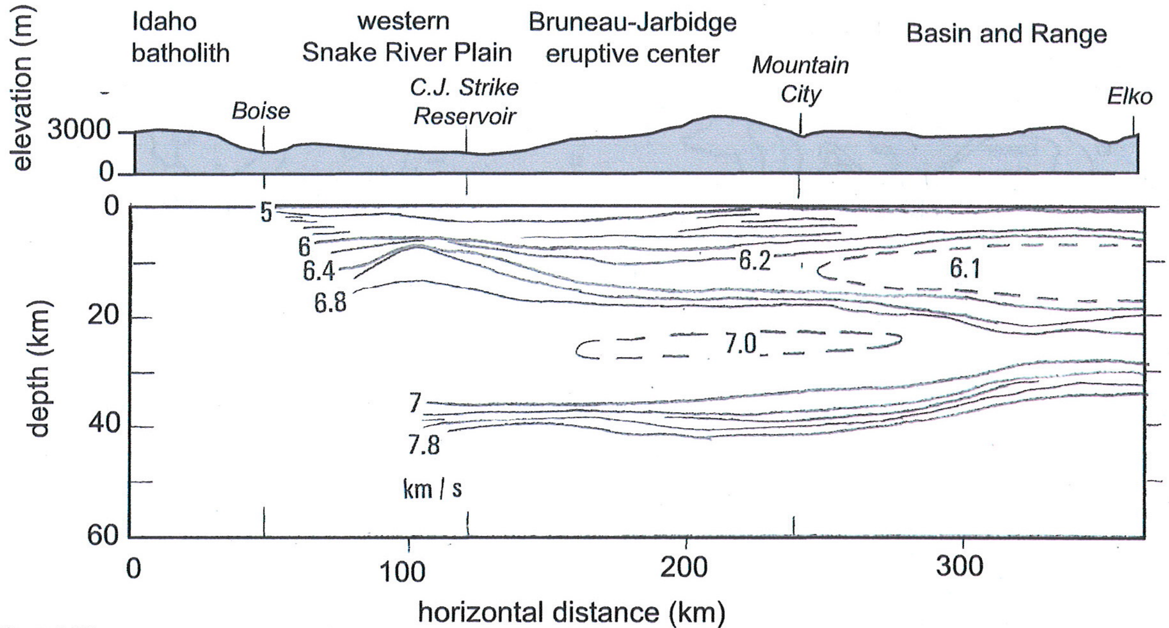


Figure 3 (B)

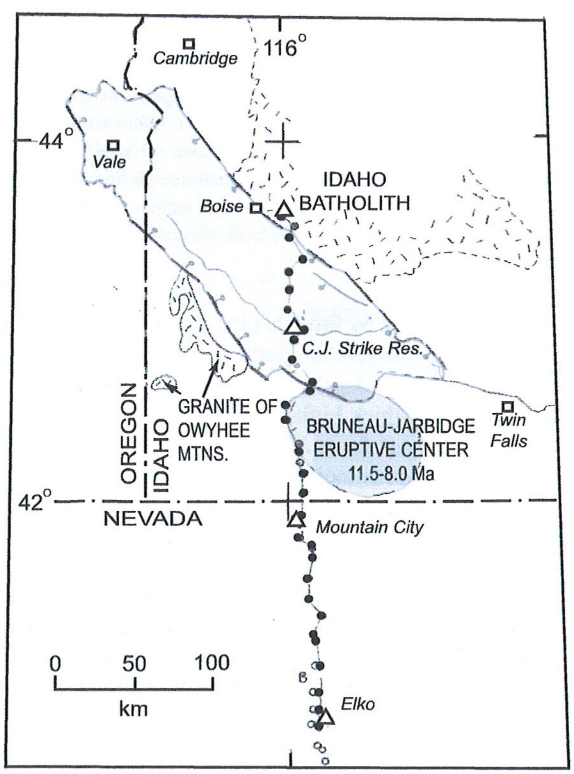


Figure 3 (C)

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

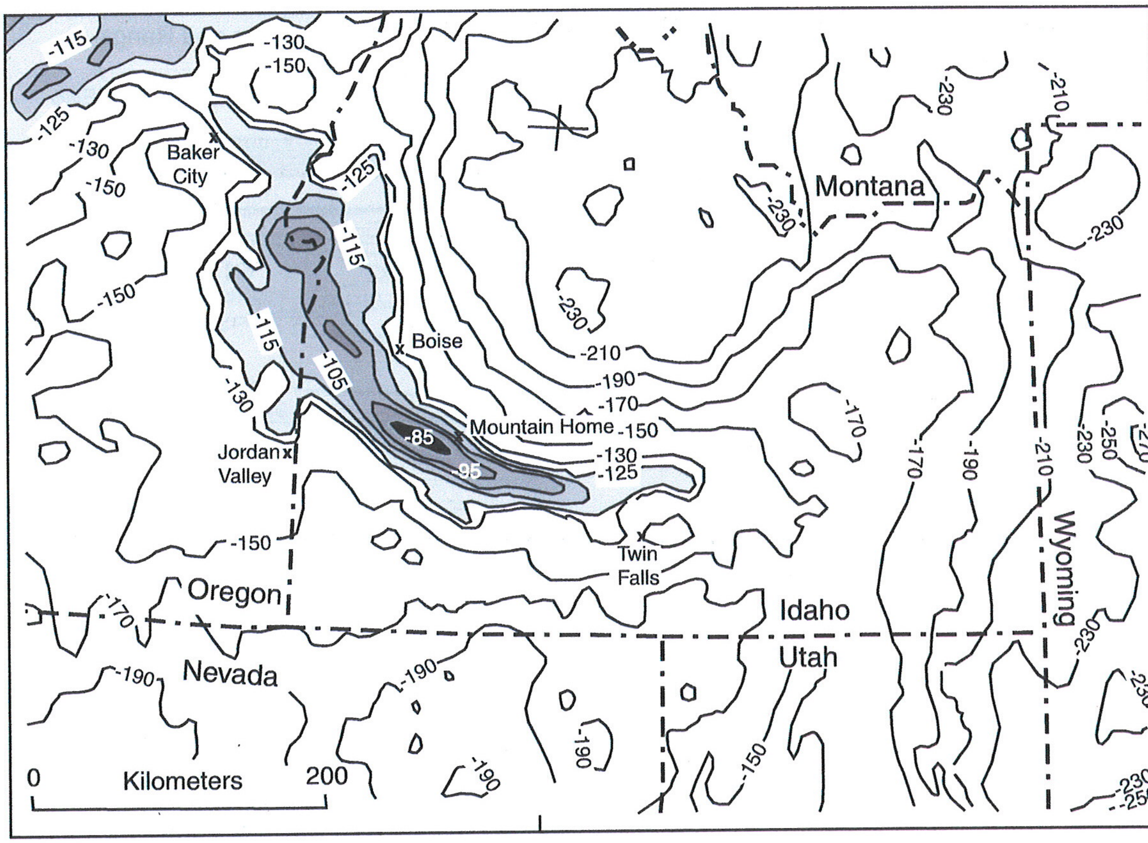


Figure 4. Map of Bouguer gravity anomalies of the western Snake River Plain region. Noteworthy is the high gravity anomaly of the western plain indicating rock of high density beneath the plain. Contour interval is 20 milligals, except for values more positive than -130 milligals. The areas of more than -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

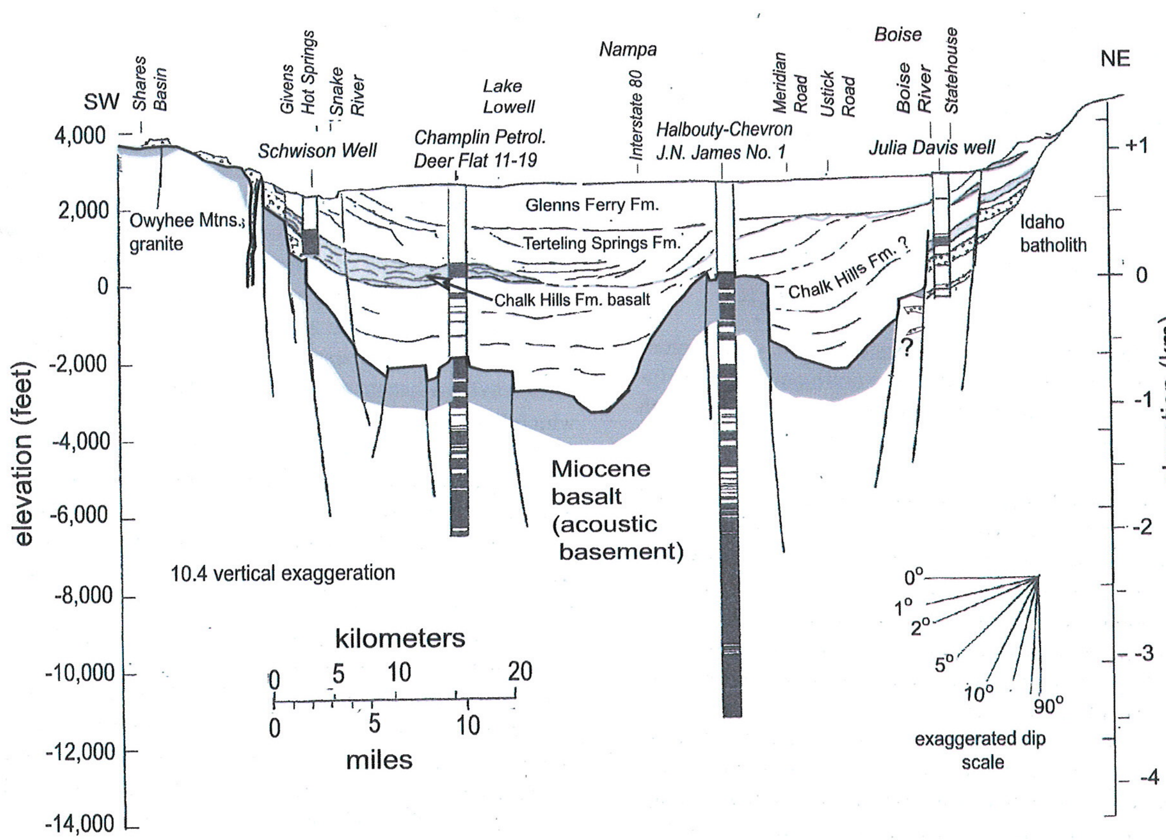


Figure 5. Section across the western plain showing the basalt occurrence (dark pattern) in the deep wells and the configuration of sedimentary fill and associated basalt from seismic reflection data. Shaded outline is the top of Miocene basalt. Stippled pattern is rhyolite that occurs in wells only at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

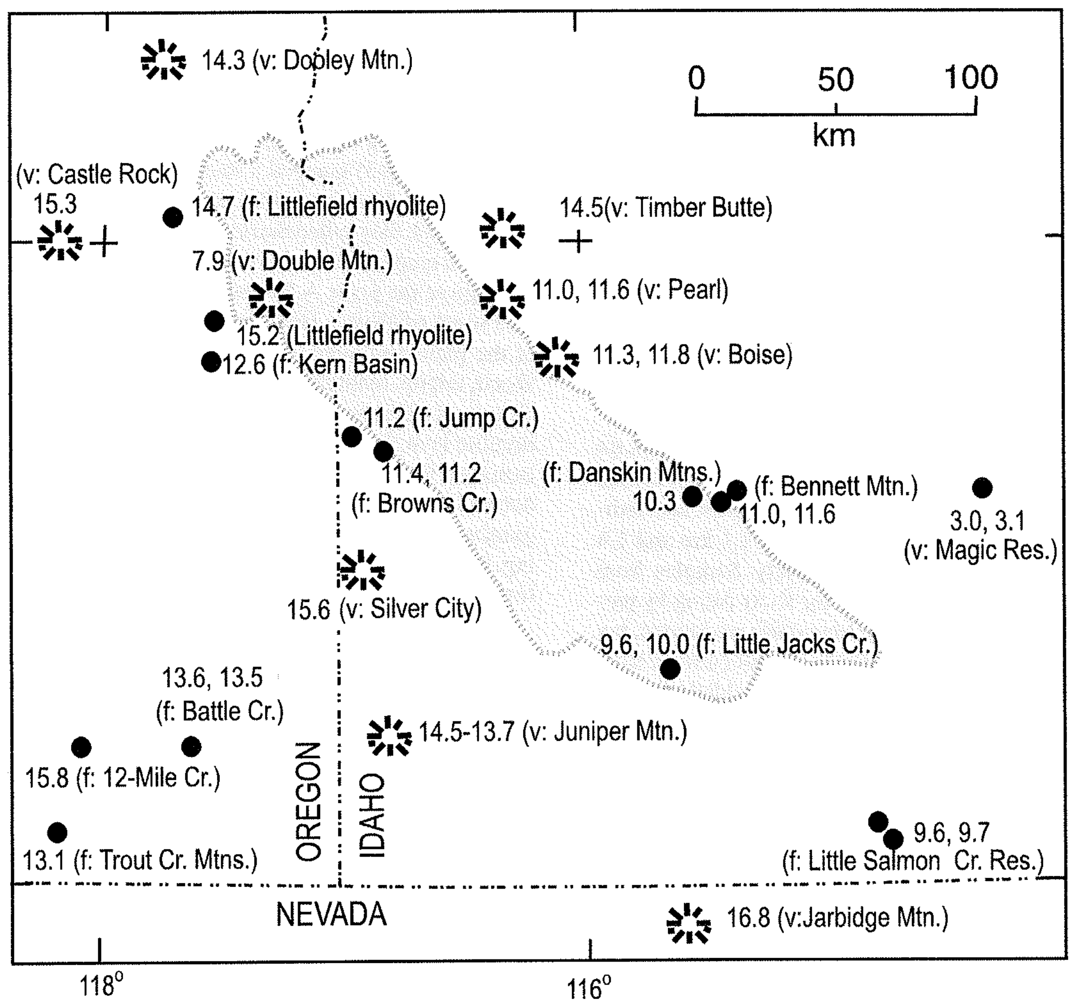


Figure 6. K-Ar ages (Ma) of rhyolite vents (v) and tuffs or lava flows (f) adjacent to the western Snake River Plain from Neill (1975) and Armstrong and others (1980) quoted in Ekren and others (1981, 1984), Clemens (1993), and Clemens and Wood (1991, 1993a, 1993b), Walker and others (1974), Dalrymple 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. 60° W. to N. 75° W.

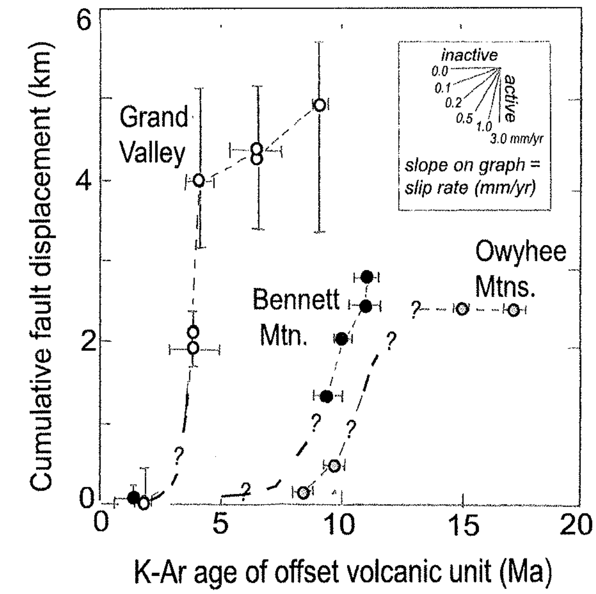


Figure 7. Displacement histories of normal faults bounding the western plain graben (Bennett Mountain is north of and includes the Griffith-Bostic well shown as GB in Figure 2; Owyhee Mountains-Anschutz well is shown as A in Figure 2; modified after Clemens, 1993) and of the Grand Valley fault oriented 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. 35-50° 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-

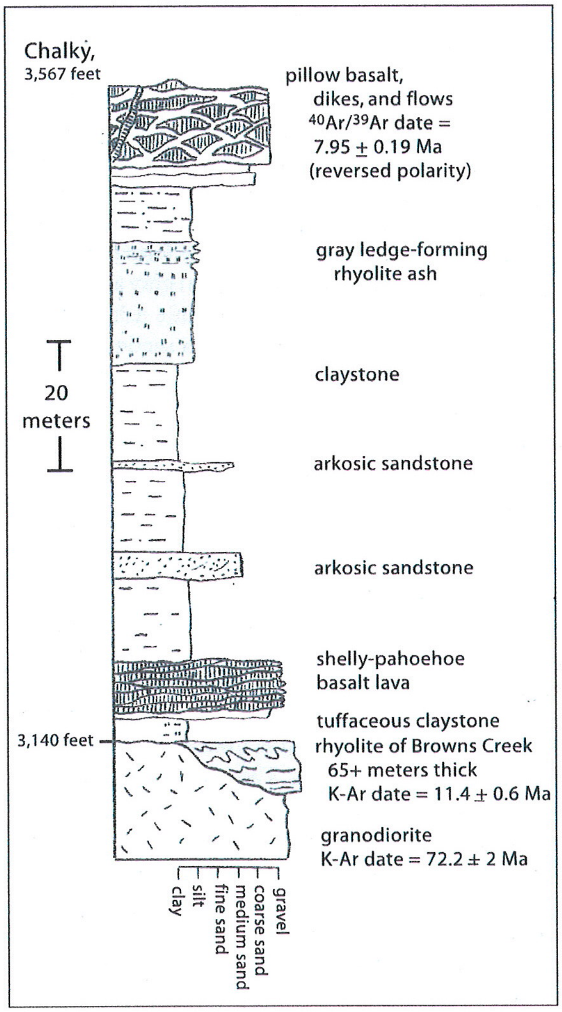


Figure 8. Columnar section of the Chalk Hills Formation at "Chalky" on the south side of the western plain near Walters Ferry (sec. 35, T. 2 S., R. 3 W.). This section was called a reference section (PC-2) for the Poison Creek Formation by Ekren and others (1981), but we regard the Poison Creek as a 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

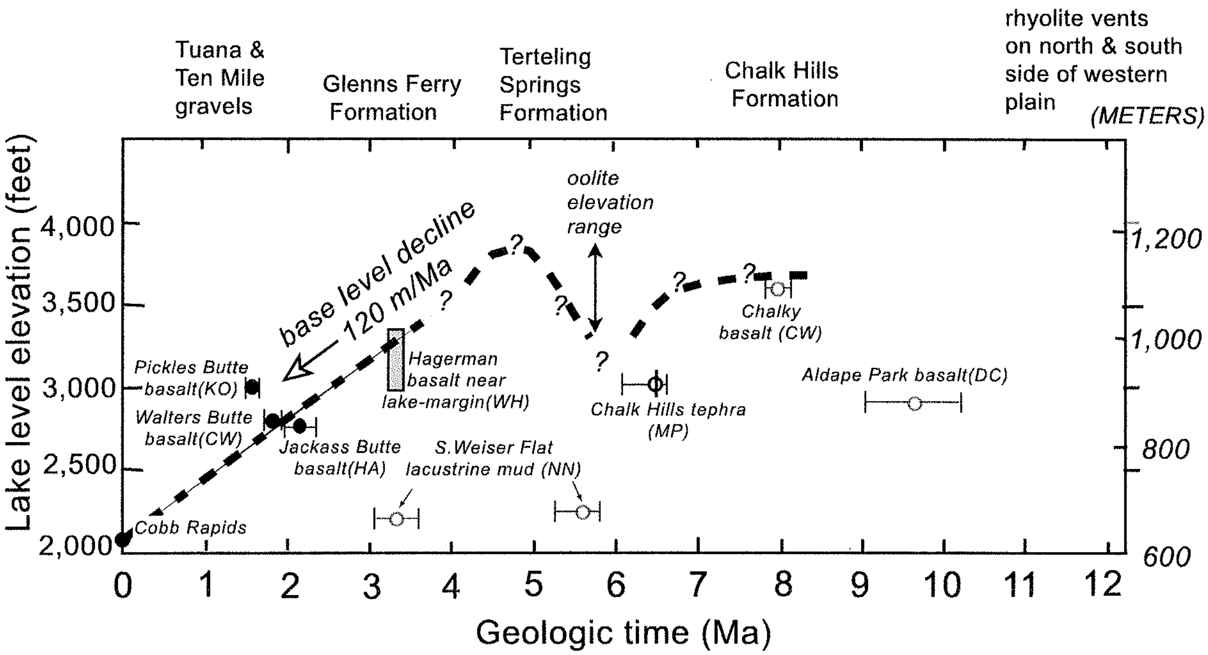


Figure 9. Plot of elevation of lake deposits versus time. This plot does not take into account tectonic deformation that may have altered the elevation of lake deposits. Most localities are on the margin of the plain, which has not been so much affected by tectonic movement or compaction subsidence (Wood, 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

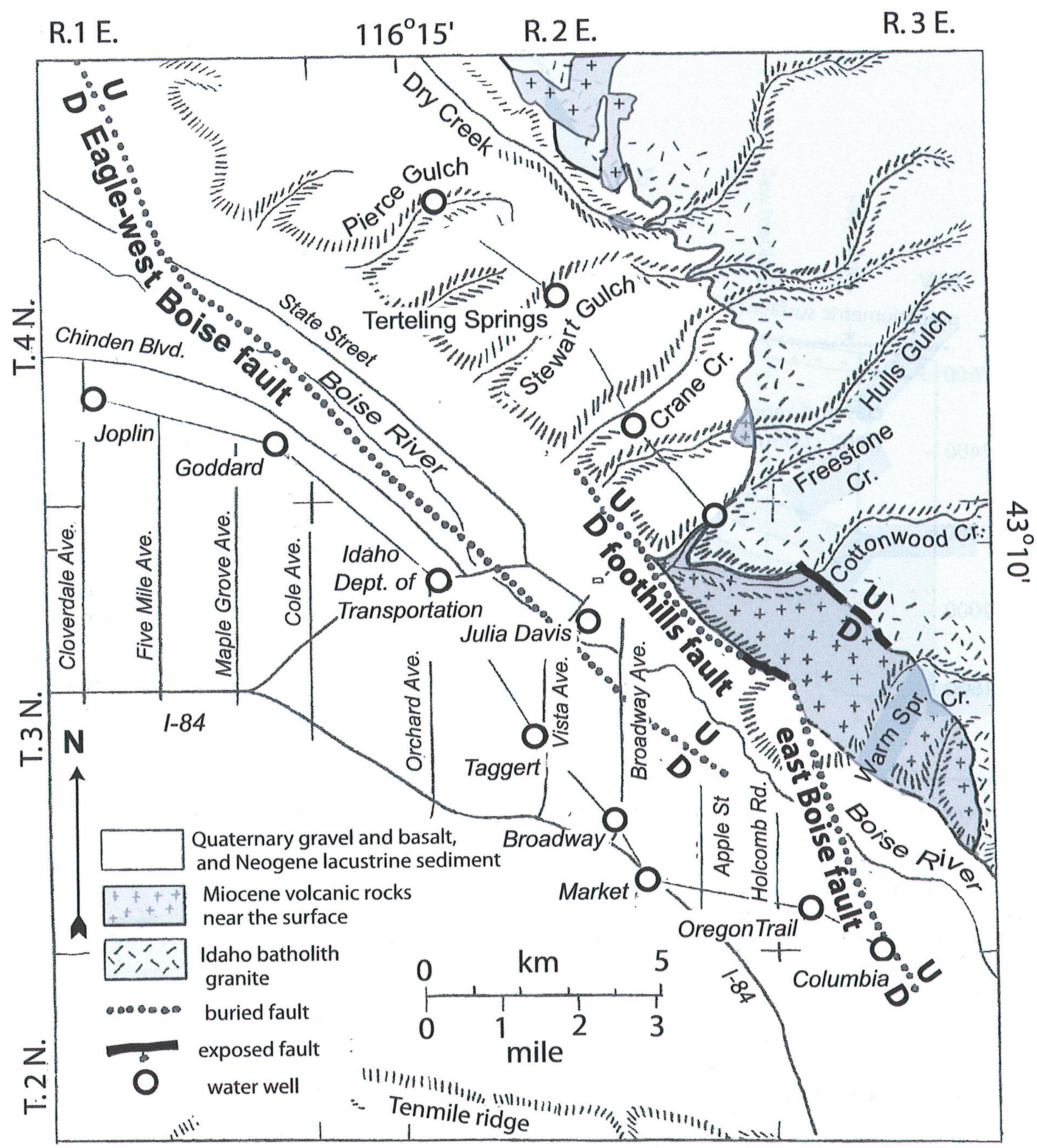


Figure 10. Geologic map of the Boise area showing locations of stratigraphic cross sections of Figure 11 and Figure 12.

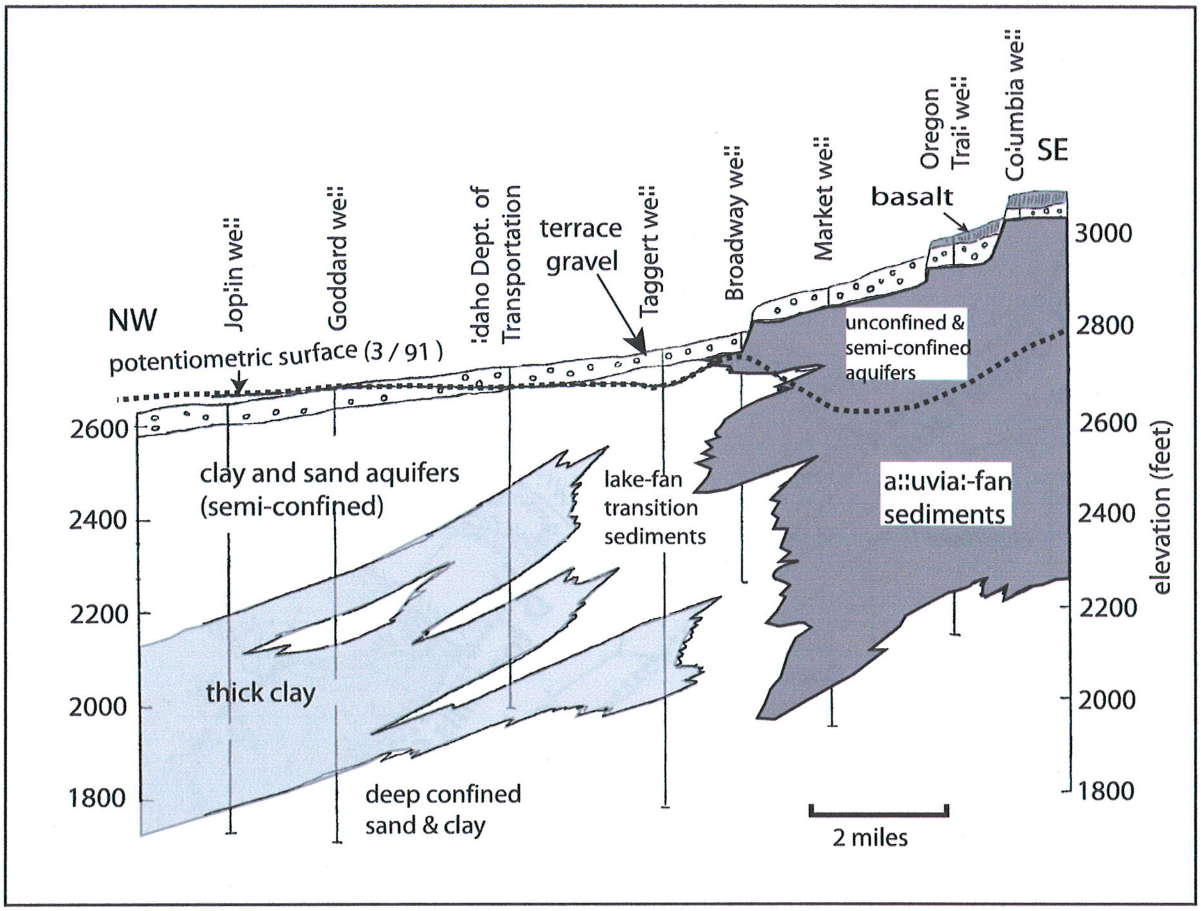


Figure 11. Cross section showing stratigraphy of alluvial fan and lake sediment aquifers beneath Boise (from Squires and others, 1992).

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

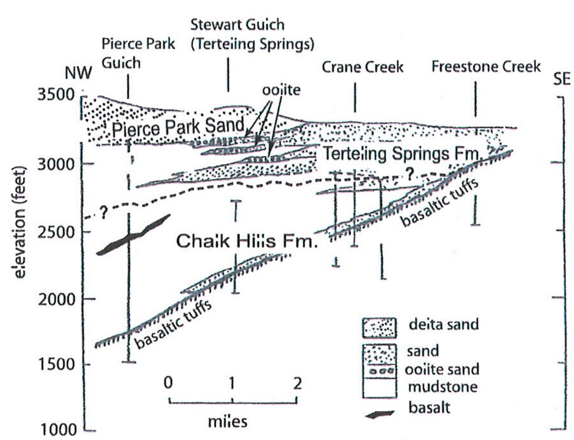


Figure 12. Stratigraphic cross section showing lake margin facies in the Boise foothills and the distribution of oolitic sands. Location of section shown in Figure 10.

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.

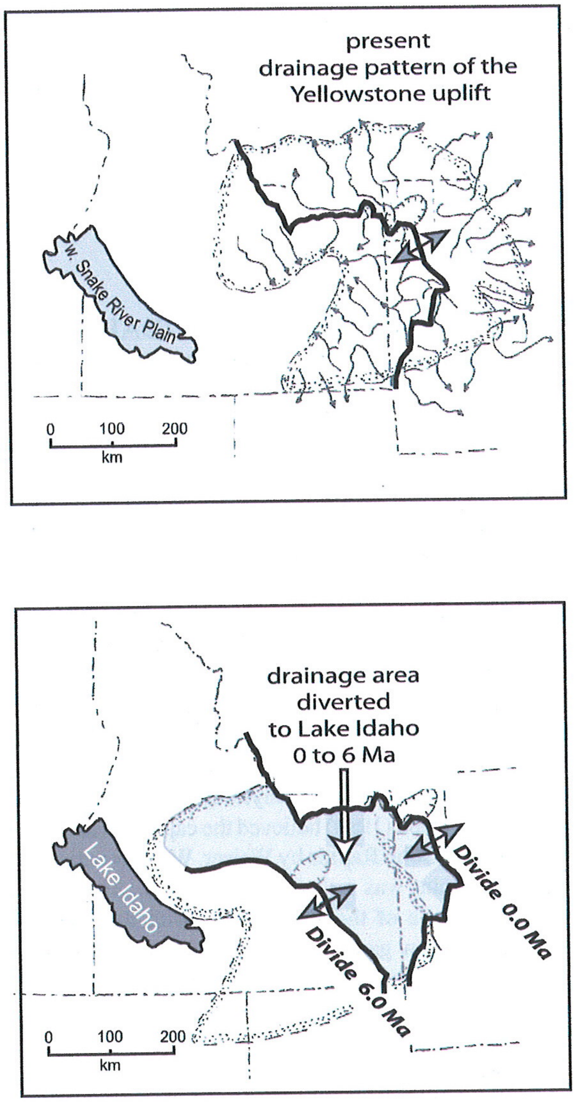


Figure 13. Map showing the area that may have diverted drainage to Lake Idaho, 6 to 0 Ma, on account of eastward-migrating uplift associated with the Yellowstone hot spot.

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

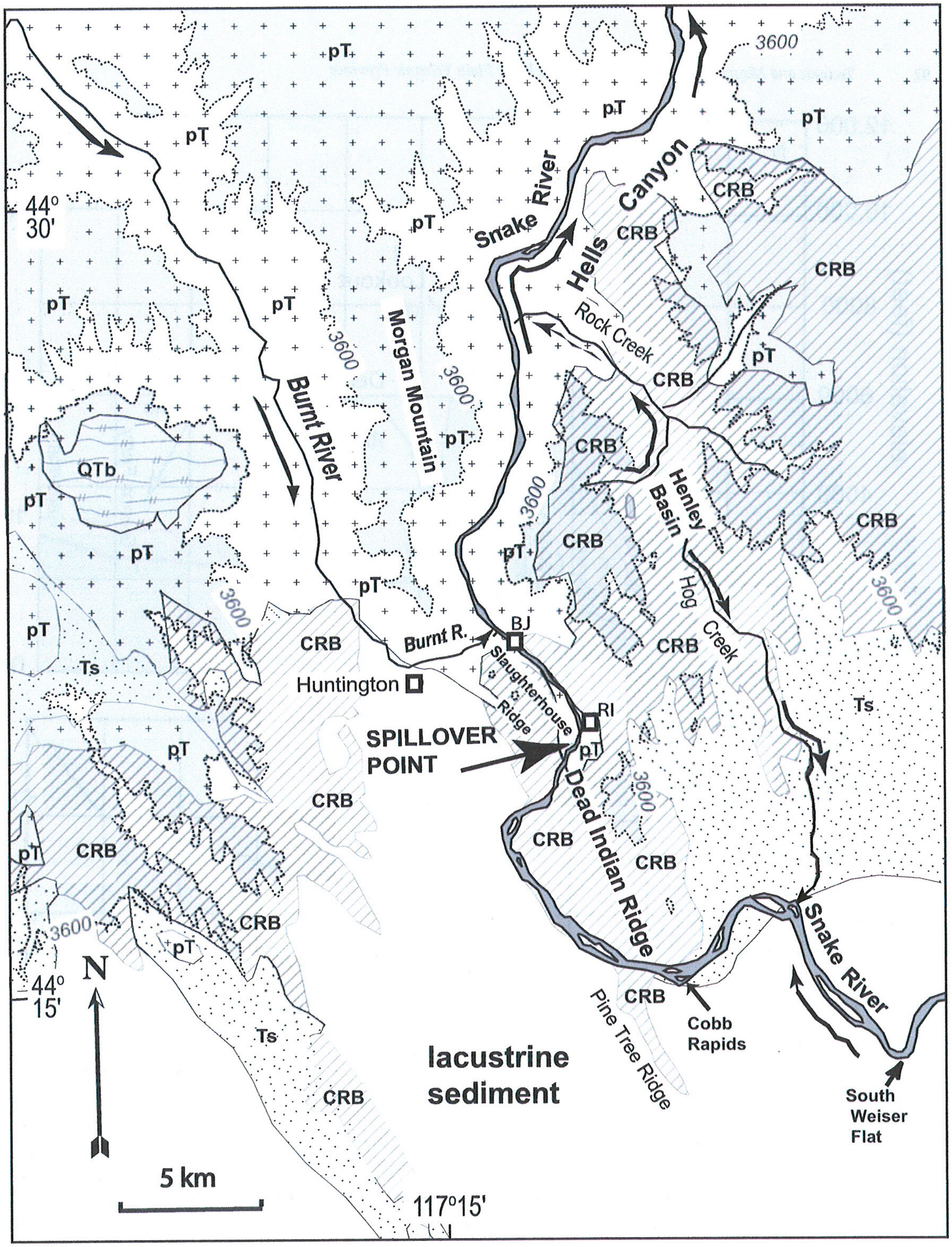


Figure 14. Present topography above 3,600 feet (shaded area) and geology at the upper end of Hell's Canyon showing the proposed spill point of Pliocene Lake Idaho. We hypothesize that the headwaters of a north-flowing tributary of ancestral Hell's Canyon were between Blakes Junction (BJ) and Rock Island 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 Hog 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.

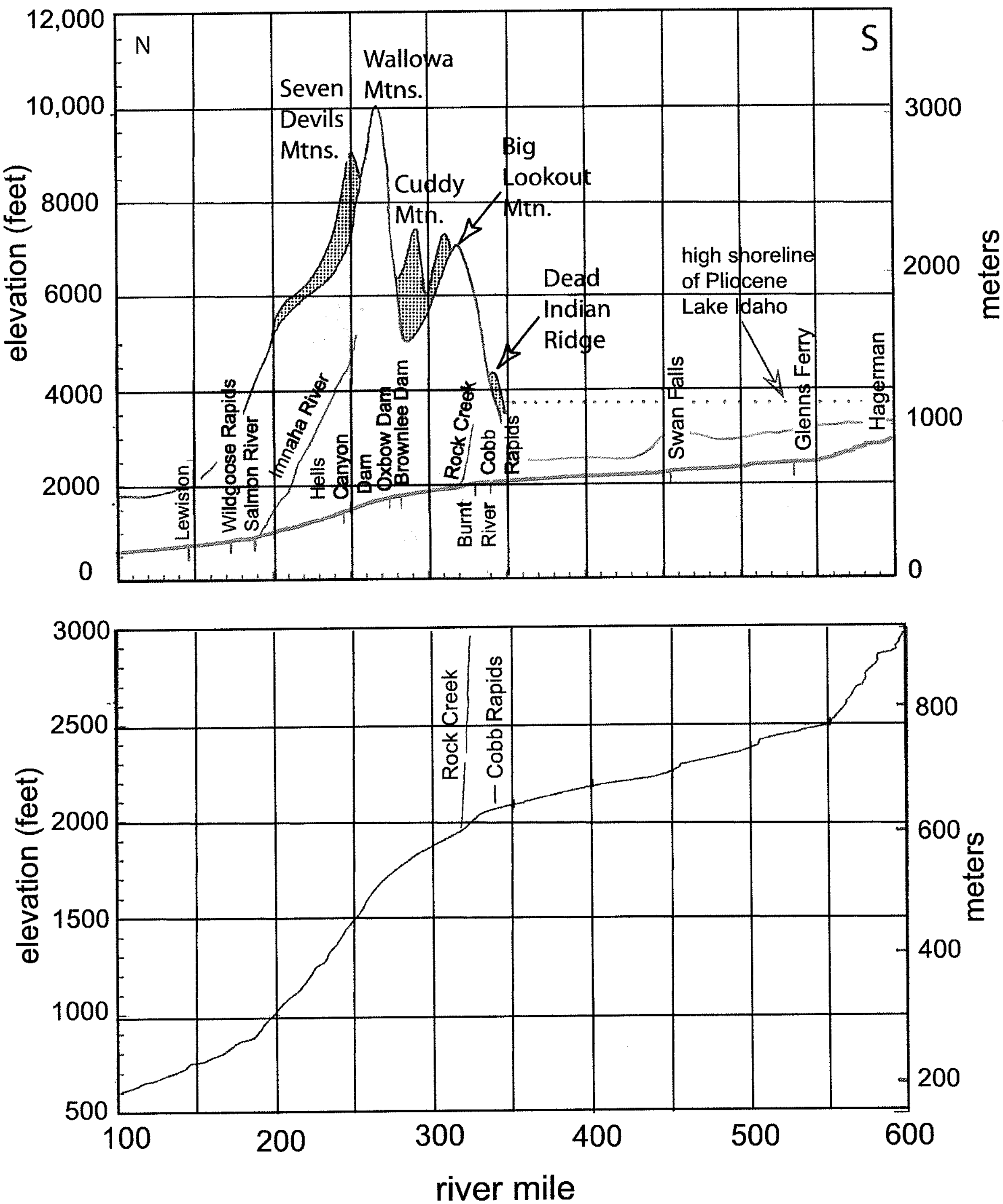


Figure 15. Longitudinal profile of the Snake River showing the change in gradient from Hell's Canyon to the Snake River Plain. Included is the profile of the Imnaha River, which has a source in the high Wallowa Mountains, and Rock Creek, which has headwaters at a divide similar to the proposed spillover 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.

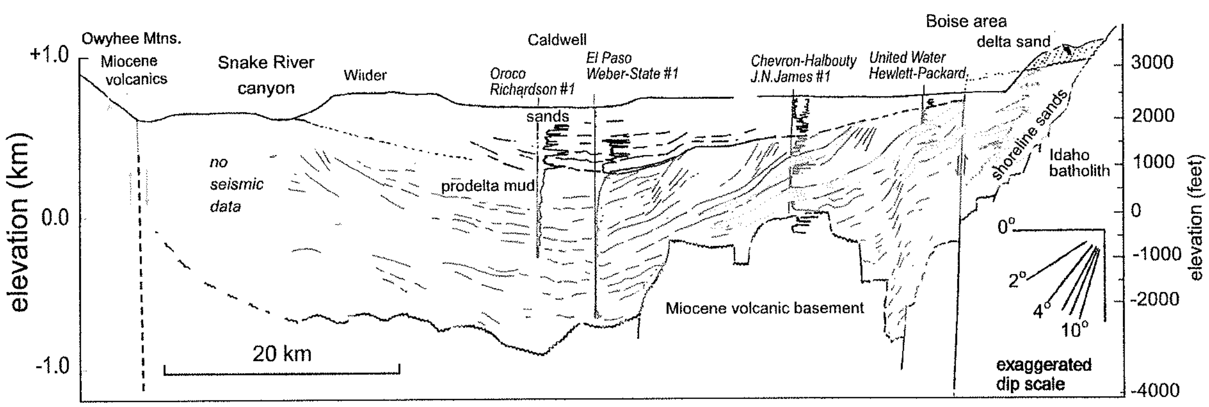


Figure 16. Tracing from seismic reflection section across the western Snake River Plain showing the upper sequence of fluvial-deltaic sediments and character of resistivity logs in wells. Resistivity-log excursions to the right are fresh-water aquifer sands, and the monotonous, near zero, resistivity 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.

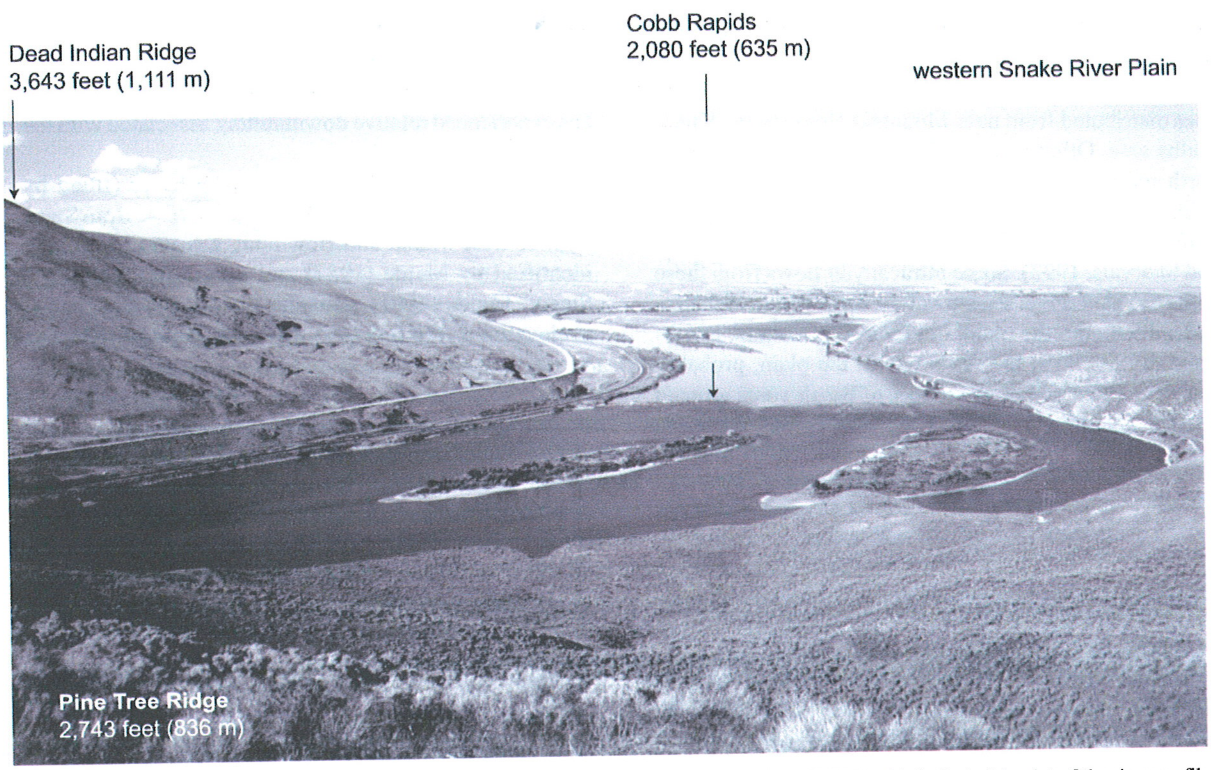


Figure 17. Photograph looking east and upriver of Cobb Rapids of the Snake River near Weiser. Cobb Rapids is the knickpoint of the river profile between the low gradient plain and the high gradient Hell's Canyon reach.

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

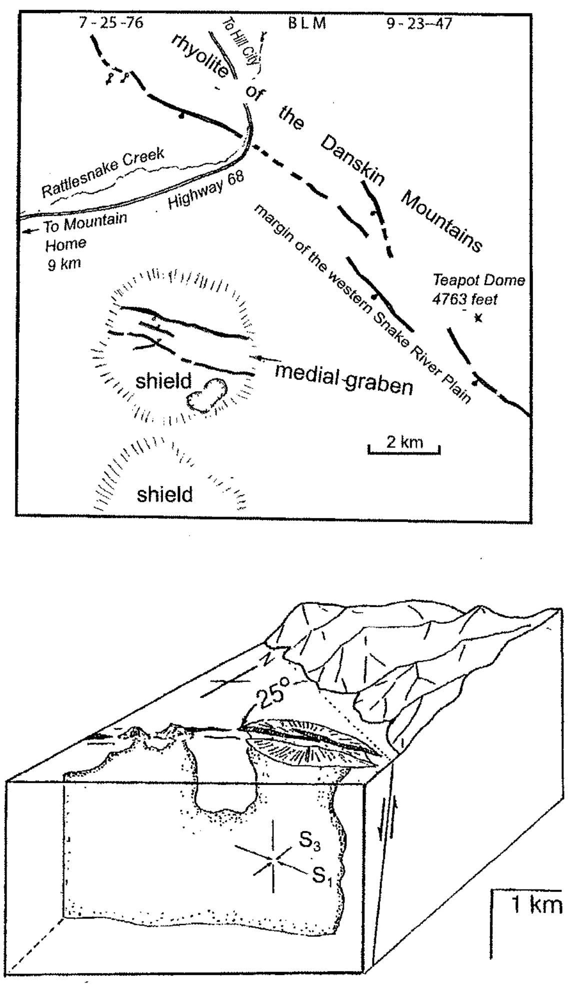


Figure 18. The air-photo tracing shows the northeast margin of the western plain 9 km northeast of Mountain Home. Tracing shows the N. 70° W. trend of fissures in the Pleistocene shield volcano and the N. 45° W. trend of the faults on the margins of the plain. Block diagram shows the inferred direction 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. 70° 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. 30° W., whereas the one zone in the western plain is N. 70° 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. 70° 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

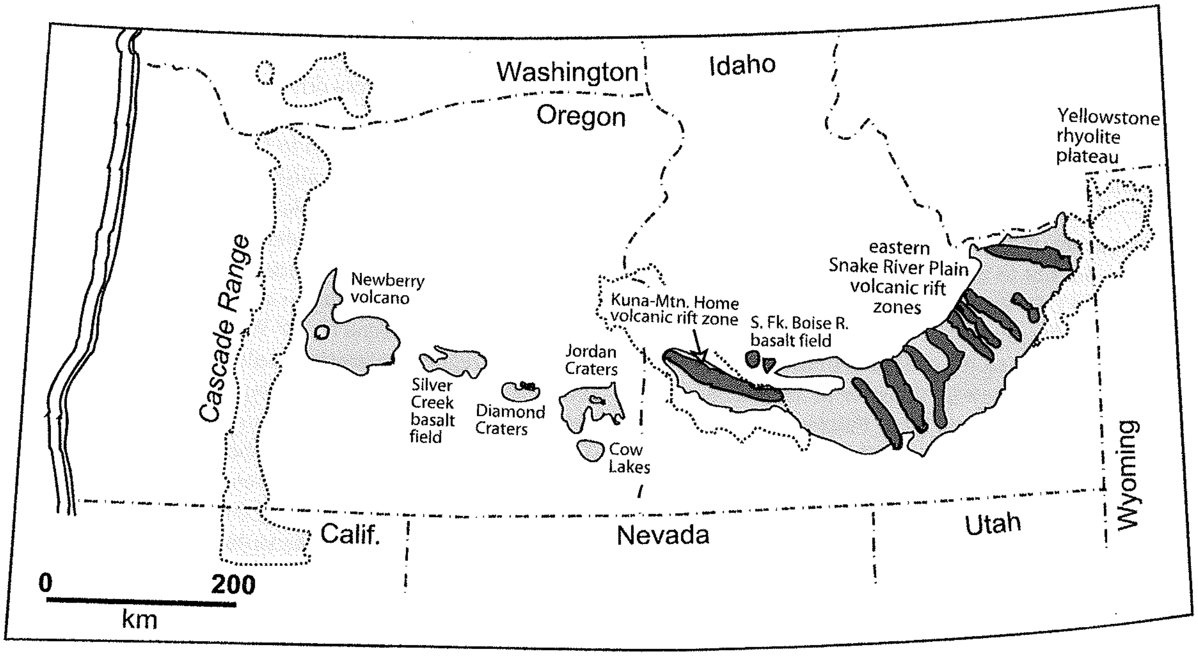


Figure 19. Map showing Quaternary basalt fields of eastern Oregon and southwest Idaho. Darkest shading is the basalt vent area. Note the contrast in the orientation of the volcanic rift zones between the eastern and the western plain. Orientation of the Kuna-Mountain Home volcanic rift zone of the western 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. 70° 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. 20° 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. 30° W. to N. 70°W. The N. 70° 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. 70° 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.

ACKNOWLEDGMENTS

Our studies of the western plain have been supported at various times by the Idaho Department of Water Resources, United Water Idaho, Idaho State Board of Education, Los Alamos National Laboratory, J.R. Simplot Company, and Boise State University faculty research grants. This study contains new information developed for the Treasure Valley Hydrologic Study of the Idaho Department of Water Resources. Our ideas have been shaped through discussion and data shared by many colleagues. We would like to thank Ed Squires, Dan Brown, Terry Scanlon, Patrick Naylor, David Hawk, Lee Liberty, Dora Gallegos, Gregg Beukelman, Kurt Othberg, Craig White, Ken Pierce, Grant Meyer, and the late Will Burnham. A special thanks to Bill Bonnichsen for an insightful review and information on areas he has explored far more than any other person. Many thanks also to Gary Fuis for reviewing and clarifying parts of the manuscript.

REFERENCES

Amini, H., H.H. Hehnert, and J.D. Obradovich, 1984, K-Ar ages of late Cenozoic basalts from the western Snake River Plain, Idaho: Isochron/West, v. 41, p. 7-11.

Anders, M.H., J.W. Geissman, L.A. Piety, and J.T. Sullivan, 1989, Parabolic distribution of circumeastern Snake River Plain seismicity and latest Quaternary faulting: Migratory pattern and association with the Yellowstone hotspot: Journal of Geophysical Research, v. 94, p. 1589-1621.

Anderson, A.L., 1947, Drainage diversion in the northern Rocky Mountains of east-central Idaho: Journal of Geology, v. 55, p. 61-75.

Armstrong, R.L., J.E. Harakal, and W.M. Neill, 1980, K-Ar dating of Snake River Plain (Idaho) volcanic rocks—New results: Isochron/West, v. 27, p. 5-10.

Armstrong, R.L., W.P. Leeman, and H.E. Malde, 1975, K-Ar dating, Quaternary and Neogene volcanic rocks of the Snake River Plain, Idaho: American Journal of Science, v. 275, p. 225-251.

Baldridge, W.S., G.R. Keller, and L.W. Braile, 1995, Continental rifting: A final perspective, *in* K.H. Olsen, ed., Continental Rifts: Evolution, Structure, Tectonics: Elsevier Science, Amsterdam, p. 453-459.

Baldridge, W.S., K.H. Olsen, and J.F. Callender, 1984, Rio Grande Rift: Problems and perspectives, *in* W.S. Baldridge, P.W. Dickerson, R.E. Riecker, and J. Zidek, eds., Rio Grande Rift: New Mexico Geological Society Guidebook, 35th Field Conference, p. 1-12.

Beukelman, G.S., 1997, Evidence of active faulting in the Halfway Gulch-Little Jacks Creek area of the western Snake River Plain, Idaho: Boise State University M.S. Thesis, 170 p.

Blackwell, D.D., 1989, Regional implications of heat flow of the Snake River Plain, northwestern United States: Tectonophysics, v. 164, p. 323-343.

Bonnichsen, Bill, 1982, Rhyolite lava flows in the Bruneau-Jarbidge eruptive center, southwestern Idaho and vicinity, *in* Bill Bonnichsen and R.M. Breckenridge, eds., Cenozoic Geology of Idaho: Idaho Geological Survey Bulletin 26, p. 103-128.

Bonnichsen, Bill, R.L. Christiansen, L.A. Morgan, F.J. Moye, W.R. Hackett, W.P. Leeman, N. Honjo, M.D. Jenks, and M.M. Godchaux, 1989, Excursion 4A: Silicic volcanic rocks in the Snake River Plain-Yellowstone Plateau Province, *in* C.E. Chapin and J. Zidek, eds., Field Excursions to Volcanic Terranes of the Western United States, v. II, Cascades and Intermountain West: New Mexico Bureau of Mines and Mineral Resources Memoir 47, p. 135-182.

Bonnichsen, Bill, C. White, and M.M. Godchaux, 1997, Basaltic volcanism, western Snake River Plain [section within Mike McCurry, Bill Bonnichsen, Craig White, M.M. Godchaux, and S.S. Hughes, 1997, Bimodal basalt-rhyolite magmatism in the central and western Snake River Plain, Idaho and Oregon, *in* P.K. Link and B.J. Kowallis, Proterozoic to Recent Stratigraphy, Tectonics, and Volcanology, Utah, Nevada, Southern Idaho and Central Mexico: Brigham Young University Geology Studies, v. 42, pt. 1, p. 381-422], p. 399-422.

Bosworth, W., 1985, Geometry of propagating continental rifts: Nature, v. 316, p. 625-627.

Bott, M.H.P., and S.B. Smithson, 1967, Gravity investigations of subsurface shape and mass distributions of granite batholiths: Geological Society of America Bulletin, v. 78, p. 859-878.

Braile, L.W., R.B. Smith, J. Ansorge, M.R. Baker, M.A. Sparlin, C. Prodehl, M.M. Schilly, J.H. Healy, S. Mueller, and K.H. Olsen, 1982, The Yellowstone-Snake River Plain seismic profiling experiment: Crustal structure of the eastern Snake River Plain: Journal of Geophysical Research, v. 87, p. 2597-2609.

Brott, C.A., D.D. Blackwell, and J.C. Mitchell, 1978, Tectonic implications of the heat flow of the Snake River Plain, Idaho: Geological Society of America Bulletin, v. 89, p. 1697-1707.

Chadwick, O.A., R.D. Hall, and F.M. Phillips, 1997, Chronology of Pleistocene glacial advances in the central Rocky Mountains: Geological Society of America Bulletin, v. 109, p. 1443-1452.

Christensen, N.I., and W.D. Mooney, 1995, Seismic velocity structure and composition of the continental crust: A global view: Journal of Geophysical Research, v. 100, p. 9761-9788.

Clemens, D.M., 1993, Tectonics and silicic volcanic stratigraphy of the western Snake River Plain, Idaho: Arizona State University M.S. thesis, 209 p.

———, 1996, The Barber ash, a tephrachronology case study in southwestern Idaho, U.S.A.: Compass, v. 71, no. 2, p. 58-68.

Clemens, D.M., and S.H. Wood, 1991, K-Ar age of silicic volcanic rocks within the lower Columbia River Basalt Group at Timber Butte, Boise and Gem counties, west-central Idaho: Isochron/West, v. 57, p. 3-7.

———, 1993a, Radiometric dating, volcanic stratigraphy, and sedimentation in the Boise foothills, northeastern margin of the western Snake River Plain, Ada County, Idaho: Isochron/West, v. 59, p. 3-10.

———, 1993b, Late Cenozoic volcanic stratigraphy and geochronology of the Mount Bennett Hills, Snake River Plain, Idaho: Isochron/West, v. 60, p. 3-14.

Conner, C.B., and M.F. Conway, 2000, Basalt volcanic fields, *in* H. Sigurdsson, ed., Encyclopedia of Volcanoes: Academic Press, New York, p. 331-343.

Conrad, G.S., 1980, Biostratigraphy and mammalian paleontology of the Glenns Ferry Formation from Hammett to Oreana, Idaho: Idaho State University Ph.D. dissertation, 334 p.

Cowan, D.S., and C.J. Potter (principal compilers) with contributions from M.T. Brandson, D.M. Fountain, D.W. Hyndman, S. Y. Johnson, B.T.R. Lewis, K.J. McLain, and D.A. Swanson, 1986, Centennial continent/ocean transect #9, B-3 Juan de Fuca spreading ridge to Montana thrust belt: Geological Society of America, Decade of North American Geology, 12 p., 3 sheets.

Cox, K.G., 1989, The role of mantle plumes in the development of continental drainage systems: Nature, v. 342, p. 873-877.

Cummings, M.L., J.G. Evans, M.L. Ferns, and K.R. Lees, 2000, Stratigraphic and structural evolution of the middle Miocene synvolcanic Oregon-Idaho graben: Geological Society of America Bulletin, v. 112, p. 668-682.

Dalrymple, G.B., A. Cox, R.R. Doell, and C.S. Gromme, 1967, Pliocene geomagnetic polarity epochs: Earth and Planetary Science Letters, v. 2, p. 163-173.

Ekren, E.B., D.H. McIntyre, and E.H. Bennett, 1984, High-temperature, large-volume, lava-like ash-flow tuffs without calderas: U.S. Geological Survey Professional Paper 1272, 76 p.

Ekren, E.B., D.H. McIntyre, E.H. Bennett, and H.E. Malde, 1981, Geologic map of Owyhee County, Idaho, west of longitude 116° W.: U.S. Geological Survey Map I-1256, scale 1:125,000.

Ekren, E.B., D.H. McIntyre, E.H. Bennett, and R.F. Marvin, 1982, Cenozoic stratigraphy of western Owyhee County, Idaho, *in* Bill Bonnichsen and R.M. Breckenridge, eds., Cenozoic Geology of Idaho: Idaho Geological Survey Bulletin 26, p. 215-235.

Evernden, J.F., D.E. Savage, G.H. Curtis, and G.T. James, 1964, Potassium-argon dates and the Cenozoic mammalian chronology of North America: American Journal of Science, v. 262, p. 145-198.

Ferns, M.L., and M.L. Cummings, 1992, Geology and mineral resources map of The Elbow quadrangle, Malheur County, Oregon: Oregon Department of Geology and Mineral Industries Geological Map Series GMS-62, scale 1:24,000.

Ferns, M.L., H.C. Brooks, J.G. Evans, and M.L. Cummings, 1993, Geologic map of the Vale 30 x 60 minute quadrangle, Malheur County, Oregon, and Owyhee County, Idaho: Oregon Department of Geology and Mineral Industries Map GMS-77, scale 1:100,000.

Fitzgerald, J.F., 1982, Geology and basalt stratigraphy of the Weiser embayment, west-cental Idaho, *in* Bill Bonnichsen and R.M. Breckenridge, eds., Cenozoic Geology of Idaho: Idaho Geological Survey Bulletin 26, p. 103-128.

Fliedner, M.M., S. Ruppert, and Southern Sierra Nevada Continental Dynamics Working Group, 1996, Three-dimensional crustal structure of the southern Sierra Nevada from seismic fan profiles and gravity modeling: Geology, v. 24, p. 367-370.

Foley, M.G., 1980, Quaternary diversion and incision, Dearborn River, Montana: Summary: Geological Society of America Bulletin, v. 91, pt. 1, p. 567-577.

Godchaux, M.M., Bill Bonnichsen, and M.D. Jenks, 1992, Types of phreatomagmatic volcanoes in the western Snake River Plain, Idaho, USA: Journal of Volcanology and Geothermal Research, v. 52, p. 1-25.

Gravity Anomaly Map Committee, 1987, Gravity anomaly map of North America, 1:5,000,000: Decade of North American Geology, Continent-Scale Map-002, Geological Society of America.

Izett, G.A., 1981, Volcanic ash beds: Recorders of upper Cenozoic silicic pyroclastic volcanism in the western United States: Journal of Geophysical Research, v. 86, p. 10,200-10,222.

Hamilton, W., 1962, Late Cenozoic structure of west-central Idaho: Geological Society of America Bulletin, v. 73, p. 511-516.

Hart, W.K., M.E. Brueseke, P.R. Renne, and H.G. McDonald, 1999, Chronostratigraphy of the Pliocene Glenns Ferry Formation, Hagerman Fossil Beds National Monument, Idaho (abs.): Geological Society of America Abstracts with Programs, v. 31, no. 4, p. A-15.

Hearst, J., 1999, Depositional environments of the Birch Creek local fauna (Pliocene:Blancan), Owyhee County, Idaho, *in* W.A. Akersten, H.G. McDonald, D.J. Meldrum, and M.E.T. Flint, eds., And Whereas . . . Papers on the Vertebrate Paleontology of Idaho Honoring John A. White, Volume 1: Idaho Museum of Natural History Occasional Paper 36, p. 56-93.

Hill, D.P., and L.C. Pakiser, 1967, Seismic-refraction study of crustal structure between the Nevada Test Site and Boise, Idaho: Geological Society of America Bulletin, v. 78, p. 685-704.

Hooper, P.R., G.B. Binger, and K.R. Lees, 2002a, Ages of the Steens and Columbia River flood basalts and their relationship to extension-related calc-alkalic volcanism in eastern Oregon: Geological Society of America Bulletin, v. 114, p. 43-50.

———, 2002b, Correction: Geological Society of America Bulletin, v. 114, p. 923-924.

Hooper, P.R., and R.M. Conrey, 1989, A model for the tectonic setting of the Columbia River basalt eruptions, *in* S.P. Reidel and P.R. Hooper, eds., Volcanism and Tectonism in the Columbia River Flood-Basalt Province: Geological Society of America Special Paper 239, p. 293-306.

Hooper, P.R., and C.J. Hawkesworth, 1993, Isotopic and geochemical constraints on the origin and evolution of the Columbia River basalt: Journal of Petrology, v. 34, p. 1203-1246.

Hooper, P.R., and D.A. Swanson, 1990, The Columbia River Basalt Group and associated volcanic rocks of the Blue Mountains Province: U.S. Geological Survey Professional Paper 1437, p. 63-99.

Howard, K.A., J.W. Shervais, and E.H. McKee, 1982, Canyon-filling lavas and lava dams on the Boise River, Idaho, and their significance for evaluating downcutting during the last two million years, *in* Bill Bonnichsen and R.M. Breckenridge, eds., Cenozoic Geology of Idaho: Idaho Geological Survey Bulletin 26, p. 629-642.

Humphreys, E.D., K. Dueker, D. Schutt, and R. Saltzer, 1999, Lithosphere and asthenosphere structure and activity in Yellowstone's wake: Geological Society of America Abstracts with Programs, v. 31, no. 4, p. A-17.

Hyndman, D.W., 1978, Major tectonic elements and tectonic problems along the line of section from northeastern Oregon to west-central Montana: Geological Society of America Map and Chart Series, MC-28C.

Jenks, M.D., and Bill Bonnichsen, 1989, Subaqueous basalt eruptions into Pliocene Lake Idaho, *in* V.E. Chamberlain, R.M. Breckenridge, and Bill Bonnichsen, eds., Guidebook to the Geology of Northern and Western Idaho and Surrounding Areas: Idaho Geological Survey Bulletin 28, p. 17-34.

Jenks, M.D., Bill Bonnichsen, and M.M. Godchaux, 1993, Geologic maps of the Grand View-Bruneau area, Owyhee County, Idaho: Idaho Geological Survey Technical Report 93-2, 21 p., scale 1:24,000.

Johnson, T.C., and P. Ng'ang'a, 1990, Reflection on a rift lake, *in* B.J. Katz, ed., Lacustrine Basin Exploration: American Association of Petroleum Geologists Memoir 50, p. 113-136.

Kimmel, P.G., 1979, Stratigraphy and paleoenvironments of the Miocene Chalk Hills Formation and Pliocene Glenns Ferry Formation in the western Snake River Plain, Idaho: University of Michigan Ph.D. dissertation, 331 p.

———, 1982, Stratigraphy, age, and tectonic setting of the western Snake River Plain, Oregon and Idaho, *in* Bill Bonnichsen and R.M. Breckenridge, eds., Cenozoic Geology of Idaho: Idaho Geological Survey Bulletin 26, p. 103-128.

Kirkham, V.R.D., 1931, The Snake River downwarp: Journal of Geology, v. 39, p. 456-482.

Kuntz, M.A., H.R. Covington, and L.J. Schorr, 1992, An overview of basaltic volcanism of the eastern Snake River Plain, Idaho, *in* P.K. Link, M.A. Kuntz, and L.B. Platt, eds., Regional Geology of Eastern Idaho and Western Wyoming: Geological Society of America Memoir 179, p. 227-267.

Lawrence, R.D., 1976, Strike-slip faulting terminates the Basin and Range Province in Oregon: Geological Society of America Bulletin, v. 87, p. 846-850.

Leake, B.E., 1990, Granite magmas: Their sources, initiation, and consequences of emplacement: Journal of the Geological Society of London, v. 147, p. 579-589.

Leeman, W.P., 1989, Origin and development of the Snake River Plain—An overview, *in* R.P. Smith and W.F. Downs, eds., Snake River Plain-Yellowstone Volcanic Province: 28th International Geological Congress Guidebook T305, American Geophysical Union, Washington, D.C., p. 4-12.

Lees, K., 1994, Magmatic and tectonic changes through time in the Neogene volcanic rocks of the Vale area, Oregon, northwestern USA:

Milton Keynes, Great Britain, The Open University Ph.D. dissertation, 284 p.

Lewis, R.E., and M.A.J. Stone, 1988, Geohydrologic data from a 4,403-foot geothermal test hole, Mountain Home Air Force Base, Elmore County, Idaho: U.S. Geological Survey Open-file Report 88-166, 30 p.

Lindgren, W., 1898, Description of the Boise quadrangle, Idaho: U.S. Geological Survey Geologic Atlas, Folio 103, 7 p.

Mabey, D.R., 1976, Interpretation of a gravity profile across the western Snake River Plain, Idaho: Geology, v. 4, p. 53-55.

———, 1982, Geophysics and tectonics of the Snake River Plain, Idaho, *in* Bill Bonnichsen and R.M. Breckenridge, eds., Cenozoic Geology of Idaho: Idaho Geological Survey Bulletin 26, p. 139-154.

MacLeod, N.S., G.S. Walker, and E.H. McKee, 1975, Geothermal significance of eastward increase in age of upper Cenozoic rhyolitic domes in southeastern Oregon, *in* Proceedings, Second United Nations Symposium on the Development and Use of Geothermal Resources, v. 1, p. 465-474.

Malde, H.E., 1959, Fault zone along northern boundary of western Snake River Plain, Idaho: Science, v. 130, p. 272.

———, 1972, Stratigraphy of the Glenns Ferry Formation from Hammett to Hagerman, Idaho: U.S. Geological Survey Bulletin 1331-D, 19 p.

———, 1991, Quaternary geology and structural history of the Snake River Plain, Idaho and Oregon, *in* R.B. Morrison, ed., Quaternary Nonglacial Geology, Conterminous U.S.: Geology of North America, Geological Society of America, v. K-2, p. 251-280.

Malde, H.E., and H.A. Powers, 1962, Upper Cenozoic stratigraphy of the western Snake River Plain, Idaho: Geological Society of America Bulletin, v. 73, p. 1197-1220.

Manley, C.R., and W.C. McIntosh, 1999, The Juniper Mountain volcanic center, Owyhee County, SW Idaho: Age relations, physical volcanology, and magmatic evolution: Geological Society of America Abstracts with Programs, v. 31, no. 4, p. A-24.

McCurry, Mike, Bill Bonnichsen, Craig White, M.M. Godchaux, and S.S. Hughes, 1997, Bimodal basalt-rhyolite magmatism in the central and western Snake River Plain, Idaho and Oregon, *in* P.K. Link and B.J. Kowallis, eds., Proterozoic to Recent Stratigraphy, Tectonics, and Volcanology, Utah, Nevada, Southern Idaho and Central Mexico: Brigham Young University Geology Studies, v. 42, pt. 1, p. 381-422.

McIntyre, D.H., 1979, Preliminary description of Anschutz Federal No. 1 drill hole, Owyhee County, Idaho: U.S. Geological Survey Open-File Report 79-651, 15 p.

McQuarrie, N., and D.W. Rodgers, 1998, Subsidence of a volcanic basin by flexure and lower crustal flow: Tectonics, v. 17, p. 203-220.

Meyer, G.M., and M.E. Leidecker, 1999, Fluvial terraces along the Middle Fork Salmon River, Idaho, and their relation to glaciation, landslide dams, and incision rates: A preliminary analysis and river-mile guide, *in* S.S. Hughes and G.D. Thackray, eds., Guidebook to the Geology of Eastern Idaho: Idaho Museum of Natural History, p. 219-235.

Middleton, L.T., M.L. Porter, and P.G. Kimmel, 1985, Depositional settings of the Chalk Hills and Glenns Ferry Formations west of Bruneau, Idaho, *in* R.M. Flores and S.S. Kaplan, eds., Cenozoic Paleogeography of the West-Central United States: Society of Economic Paleontologists and Mineralogists, Rocky Mountain Section, p. 37-53.

Minor, S.A., J.J. Rytuba, M.J. Grubensky, D.B. Vander Meulen, C.A. Goeldner, and K.J. Tegtmeyer, 1987, Geologic map of the High Steens and Little Blitzen Gorge Wilderness Study Areas, Harney County, Oregon: U.S. Geological Survey Map MF-1876, scale 1:24,000.

Morgan, L.A., 1992, Stratigraphic relations and paleomagnetic and geochemical correlations of ignimbrites of the Heise volcanic field, eastern Snake River Plain, eastern Idaho and western Wyoming, *in* P.K. Link, M.A. Kuntz, and L.B. Platt, eds., Regional Geology of Eastern Idaho and Western Wyoming: Geological Society of America Memoir 179, p. 215-226.

Nakamura, K. and S. Uyeda, 1980, Stress gradient in back-arc regions and plate subduction: Journal of Geophysical Research, v. 85, p. 6419-6428.

Neill, W.M., 1975, Geology of the southeastern Owyhee Mountains and environs, Owyhee County, Idaho: Stanford University M.S. thesis, 59 p.

Neville, C., N.D. Opdyke, E.H. Lindsay, and N.M. Johnson, 1979, Magnetic stratigraphy of Pliocene deposits on the Glenns Ferry Formation, Idaho, and its implications for North American mammalian biostratigraphy: American Journal of Science, v. 279, p. 503-526.

Newton, G.D., 1991, Geohydrology of the regional aquifer system, western Snake River Plain, southwestern Idaho: U.S. Geological Survey Professional Paper 1408-G, 52 p.

Othberg, K.L., 1988, Changeover from basin-filling to incision in the western Snake River Plain: Geological Society of America Abstracts with Programs, v. 20, no. 6, p. 461.

———, 1994, Geology and geomorphology of the Boise Valley and adjoining areas, western Snake River Plain, Idaho: Idaho Geological Survey Bulletin 29, 54 p.

Othberg, K.L., Bill Bonnichsen, C.C. Swisher III, and M.M. Godchaux, 1995, Geochronology and geochemistry of Pleistocene basalts of the western Snake River Plain and Smith Prairie, Idaho: Isochron/West, v. 62, p. 16-29.

Pansze, A.J., Jr., 1975, Geology and ore deposits of the Silver City-Delamar-Flint region, Owyhee County, Idaho: Idaho Bureau of Mines and Geology Pamphlet 161, 79 p.

Park, R.G., 1988, Geological Structures and Moving Plates: Blackie and Sons, Glasgow, 337 p.

Parsons, T., G.A. Thompson, and R.P. Smith, 1998, More than one way to stretch: A tectonic model for extension along the plume track of the Yellowstone hotspot and adjacent Basin and Range Province: Tectonics, v. 17, p. 221-234.

Perkins, M.E., F.H. Brown, W.P. Nash, W. McIntosh, and S.K. Williams, 1998, Sequence, age, and source of silicic fallout tuffs in middle to late Miocene basins of the Basin and Range Province: Geological Society of America Bulletin, v. 110, p. 344-360.

Pezzopane, S.K., and R.J. Weldon, 1993, Tectonic role of active faulting in central Oregon: Tectonics, v. 12, p. 1140-1169.

Pierce, K.L., and L.A. Morgan, 1992, The track of the Yellowstone hot spot: Volcanism, faulting, and uplift, *in* P.K. Link, M.A. Kuntz, and L.B. Platt, eds., Regional Geology of Eastern Idaho and Western Wyoming: Geological Society of America Memoir 179, p. 1-53.

Prodehl, C., 1979, Crustal structure of the western United States: U.S. Geological Survey Professional Paper 1034, 74 p.

Repenning, C.A., T.R. Weasma, and G.R. Scott, 1994, The early Pleistocene (latest Blancan-earliest Irvingtonian) Froman Ferry fauna and history of the Glenns Ferry Formation, southwestern Idaho: U.S. Geological Survey Bulletin 2105, 86 p.

Rosendahl, B.R., E. Kilembe, and K. Kaczmarick, 1992, Comparison of the Tanganyika, Malawi, Rukwa, and Turkana Rift zones from analysis of seismic reflection data: Tectonophysics, v. 213, p. 235-256.

Rytuba, J.J., and D.B. Vander Meulen, 1991, Hot-spring precious-metal systems in the Lake Owyhee volcanic field, Oregon-Idaho, *in* G.L. Raines, R.E. Lisle, R.W. Schafer, and W.H. Wilkinson, eds., Geology and Ore Deposits of the Great Basin: Symposium Proceedings, Geological Society of Nevada, Reno, p. 1085-1096.

Sadler, J.L. and P.K. Link, 1996, The Tuana Gravel: Early Pleistocene response to longitudinal drainage of a late-stage rift basin, western Snake River Plain, Idaho: Northwest Geology, v. 26, p. 46-62.

Schumm, S.A., and F.G. Ethridge, 1994, Origin, evolution, and morphology of fluvial valleys, *in* Incised Valley Systems; Origin and Sedimentary Sequences, Society of Economic Paleontologists and Mineralogists Special Publication 51, Society for Sedimentary Geology, Tulsa, p. 11-27.

Sheppard, R.A., 1991, Zeolite diagenesis of tuffs in the Miocene Chalk Hills Formation, western Snake River Plain, Idaho: U.S. Geological Survey Bulletin 1963, 27 p.

Smith, G.R., N. Morgan, and E. Gustafson, 2000, Fishes of the Pliocene Ringold Formation of Washington and history of the Columbia River drainage: University of Michigan Museum of Paleontology Papers on Paleontology, v. 32, 42 p.

Smith, G.R., K. Swirydczuk, P.G. Kimmel, and B.H. Wilkinson, 1982, Fish biostratigraphy of late Miocene to Pleistocene sediments of the western Snake River Plain, Idaho, *in* Bill Bonnichsen and R.M. Breckenridge, eds., Cenozoic Geology of Idaho: Idaho Geological Survey Bulletin 26, p. 519-541.

Smith, G.R., and W.P. Patterson, 1994, Mio-Pliocene seasonality on the Snake River Plain: Comparison of faunal and oxygen isotopic evidence: Paleogeography, Paleoclimatology, Paleoecology, v. 107, p. 291-302.

Smith, R.B., and L.W. Braile, 1993, Topographic signature, space-time evolution, and physical properties of the Yellowstone-Snake River Plain volcanic system: The Yellowstone hotspot, *in* A.W. Snoke, J.R. Steidtmann, and S.M. Roberts, eds., Geology of Wyoming: Geological Survey of Wyoming Memoir No. 5, p. 694-754.

Squires, E., S.H. Wood, and J.L. Osiensky, 1992, Hydrogeologic framework of the Boise aquifer system, Ada County, Idaho: Idaho Water Resources Research Institute, Research Technical Completion Report 14-08-0001-0G1559-06, University of Idaho, 114 p.

Swirydczuk, K., G.P. Larson, and Gerald R. Smith, 1982, Volcanic ash beds as stratigraphic markers in the Glenns Ferry and Chalk Hills Formations from Adrian, Oregon, to Bruneau, Idaho, *in* Bill Bonnichsen and R.M. Breckenridge, eds., Cenozoic Geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 543-558.

Swirydczuk, K., B.H. Wilkinson, and G.R. Smith, 1979, The Pliocene Glenns Ferry oolite: Lake-margin carbonate deposition in the southwestern Snake River Plain: Journal of Sedimentary Petrology, v. 49, p. 995-1004.

———, 1980a, The Pliocene Glenns Ferry oolite-I: Lake-margin carbonate deposition in the southwestern Snake River Plain—Reply: Journal of Sedimentary Petrology, v. 50, p. 999-1001.

———, 1980b, The Pliocene Glenns Ferry oolite-II: Sedimentology of oolitic lacustrine terrace deposits: Journal of Sedimentary Petrology, v. 50, p. 1237-1248.

Taylor, D.W., and R.C. Bright, 1987, Drainage history of the Bonneville Basin, *in* Cenozoic Geology of Western Utah—Sites for Precious Metal and Petroleum Accumulations: Utah Geological Society Publication 16, p. 239-256.

Taubeneck, W.H., 1971, Idaho batholith and its southern extension: Geological Society of America Bulletin, v. 82, p. 1899-1928.

Thompson, R.S., 1991, Pliocene environments and climates of the western United States: Quaternary Science Reviews, v. 10, p. 115-132.

Vallier, Tracey, 1998, Islands and Rapids: A Geological Story of Hell's Canyon: Confluence Press, Lewiston, Idaho, 151 p.

Van Tassell, J., M. Ferns, V. McConnell, and G.R. Smith, 2001, The mid-Pliocene Imbler fish fossils, Grande Ronde Valley, Union County, Oregon, and the connection between Lake Idaho and the Columbia River: Oregon Geology, v. 63, p. 77-96.

Vetter, S.K., and J.W. Shervais, 1992, Continental basalts of the Boise River group near Smith Prairie, Idaho: Journal of Geophysical Research, v. 97, p. 9043-9061.

Walker, G.W., G.B. Dalrymple, and M.A. Lanphere, 1974, Index to potassium-argon ages of volcanic rocks of Oregon: U.S. Geological Survey Miscellaneous Field Studies, Map MF-569, 2 sheets.

Walker, G.W., and N.S. MacLeod, 1991, Geologic map of Oregon: U.S. Geological Survey, scale 1:500,000.

Wendlandt, R.F., W.S. Baldridge, and E.R. Neumann, 1991, Modification of the lower crust by continental rift magmatism: Geophysical Research Letters, v. 18, p. 1759-1762.

Wernicke, B., 1992, Extensional tectonics in the western United States: Geology of North America, v. G-3, The Cordilleran Orogen, Conterminous U.S.: Geological Society of America, Boulder, Colorado., p. 553-582.

Wheeler, H.E., and E.F. Cook, 1954, Structural and stratigraphic significance of the Snake River capture, Idaho-Oregon: Journal of Geology, v. 62, p. 525-536.

Whitehead, R.L., 1992, Geohydrologic framework of the Snake River Plain, Idaho and eastern Oregon: U.S. Geological Survey Professional Paper 1408-B, 32 p.

Wood, S.H., 1989, Silicic volcanic rocks and structure of the western Mount Bennett Hills and adjacent Snake River Plain, Idaho, *in* R.P. Smith and K.L. Ruebelmann, eds., Snake River Plain-Yellowstone Volcanic Province: 28th International Geological Congress, Guidebook T305, American Geophysical Union, Washington, D.C., p. 69-77.

———, 1994, Seismic expression and geological significance of a lacustrine delta in Neogene deposits of the western Snake River Plain, Idaho: American Association of Petroleum Geologists Bulletin, v. 78, p. 102-121.

———, 1997, Structure contour map of the base of Quaternary basalt in the western Snake River Plain: Idaho Department of Water Resources, Treasure Valley Hydrologic Project, Boise, scale 1:100,000.

Wood, S.H., and J.E. Anderson, 1981, Geology, *in* J.C. Mitchell, ed., Geothermal Investigations in Idaho, Part 11, Geological, Hydrological, Geochemical, and Geophysical Investigations of the Nampa-Caldwell and Adjacent Areas, Southwestern Idaho: Idaho Department of Water Resources Water Information Bulletin 30, p. 9-31.

Wood, S.H., and J.N. Gardner, 1984, Silicic volcanic rocks of the Miocene Idavada Group, Bennett Mountain, Elmore County, southwestern Idaho: Final contract report to the Los Alamos National Laboratory from Boise State University, 39 p., 1 map, scale 1:100,000.

Wood, Steven, and S.H. Wood, 1999, Large pumice-block layer marks sublacustrine rhyolite-dome eruption in the Miocene Chalk Hills Formation, near Marsing, Idaho, western Snake River Plain: Geological Society of America Abstracts with Programs, v. 31, no. 4, p. A-62.

Zoback, M.L., E.H. McKee, R.J. Blakely, and G.A. Thompson, 1994, The northern Nevada rift: Regional tectono-magmatic relations and middle Miocene stress direction: Geological Society of America Bulletin, v. 106, p. 371-382.

Zoback, M.D., and M.L. Zoback, 1991, Tectonic stress field of North America and relative plate motions, *in* D.B. Slemmons, E.R. Engdahl, M.D. Zoback, and D.D. Blackwell, eds., Neotectonics of North America: Geological Society of America Map Volume 1, p. 339-365.

Schumm, S.A., and F.G. Ethridge, 1994, Origin, evolution, and morphology of fluvial valleys, *in* Incised Valley Systems; Origin and Sedimentary Sequences, Society of Economic Paleontologists and Mineralogists Special Publication 51, Society for Sedimentary Geology, Tulsa, p. 11-27.

Sheppard, R.A., 1991, Zeolite diagenesis of tuffs in the Miocene Chalk Hills Formation, western Snake River Plain, Idaho: U.S. Geological Survey Bulletin 1963, 27 p.

Smith, G.R., N. Morgan, and E. Gustafson, 2000, Fishes of the Pliocene Ringold Formation of Washington and history of the Columbia River drainage: University of Michigan Museum of Paleontology Papers on Paleontology, v. 32, 42 p.

Smith, G.R., K. Swirydczuk, P.G. Kimmel, and B.H. Wilkinson, 1982, Fish biostratigraphy of late Miocene to Pleistocene sediments of the western Snake River Plain, Idaho, *in* Bill Bonnichsen and R.M. Breckenridge, eds., Cenozoic Geology of Idaho: Idaho Geological Survey Bulletin 26, p. 519-541.

Smith, G.R., and W.P. Patterson, 1994, Mio-Pliocene seasonality on the Snake River Plain: Comparison of faunal and oxygen isotopic evidence: Paleogeography, Paleoclimatology, Paleoecology, v. 107, p. 291-302.

Smith, R.B., and L.W. Braile, 1993, Topographic signature, space-time evolution, and physical properties of the Yellowstone-Snake River Plain volcanic system: The Yellowstone hotspot, *in* A.W. Snoke, J.R. Steidtmann, and S.M. Roberts, eds., Geology of Wyoming: Geological Survey of Wyoming Memoir No. 5, p. 694-754.

Squires, E., S.H. Wood, and J.L. Osiensky, 1992, Hydrogeologic framework of the Boise aquifer system, Ada County, Idaho: Idaho Water Resources Research Institute, Research Technical Completion Report 14-08-0001-0G1559-06, University of Idaho, 114 p.

Swirydczuk, K., G.P. Larson, and Gerald R. Smith, 1982, Volcanic ash beds as stratigraphic markers in the Glenns Ferry and Chalk Hills Formations from Adrian, Oregon, to Bruneau, Idaho, *in* Bill Bonnichsen and R.M. Breckenridge, eds., Cenozoic Geology of Idaho: Idaho Bureau of Mines and Geology Bulletin 26, p. 543-558.

Swirydczuk, K., B.H. Wilkinson, and G.R. Smith, 1979, The Pliocene Glenns Ferry oolite: Lake-margin carbonate deposition in the southwestern Snake River Plain: Journal of Sedimentary Petrology, v. 49, p. 995-1004.

———, 1980a, The Pliocene Glenns Ferry oolite-I: Lake-margin carbonate deposition in the southwestern Snake River Plain—Reply: Journal of Sedimentary Petrology, v. 50, p. 999-1001.

———, 1980b, The Pliocene Glenns Ferry oolite-II: Sedimentology of oolitic lacustrine terrace deposits: Journal of Sedimentary Petrology, v. 50, p. 1237-1248.

Taylor, D.W., and R.C. Bright, 1987, Drainage history of the Bonneville Basin, *in* Cenozoic Geology of Western Utah—Sites for Precious Metal and Petroleum Accumulations: Utah Geological Society Publication 16, p. 239-256.

Taubeneck, W.H., 1971, Idaho batholith and its southern extension: Geological Society of America Bulletin, v. 82, p. 1899-1928.

Thompson, R.S., 1991, Pliocene environments and climates of the western United States: Quaternary Science Reviews, v. 10, p. 115-132.

Vallier, Tracey, 1998, Islands and Rapids: A Geological Story of Hell's Canyon: Confluence Press, Lewiston, Idaho, 151 p.

Van Tassell, J., M. Ferns, V. McConnell, and G.R. Smith, 2001, The mid-Pliocene Imbler fish fossils, Grande Ronde Valley, Union County, Oregon, and the connection between Lake Idaho and the Columbia River: Oregon Geology, v. 63, p. 77-96.

Vetter, S.K., and J.W. Shervais, 1992, Continental basalts of the Boise River group near Smith Prairie, Idaho: Journal of Geophysical Research, v. 97, p. 9043-9061.

Walker, G.W., G.B. Dalrymple, and M.A. Lanphere, 1974, Index to potassium-argon ages of volcanic rocks of Oregon: U.S. Geological Survey Miscellaneous Field Studies, Map MF-569, 2 sheets.

Walker, G.W., and N.S. MacLeod, 1991, Geologic map of Oregon: U.S. Geological Survey, scale 1:500,000.

Wendlandt, R.F., W.S. Baldridge, and E.R. Neumann, 1991, Modification of the lower crust by continental rift magmatism: Geophysical Research Letters, v. 18, p. 1759-1762.

Wernicke, B., 1992, Extensional tectonics in the western United States: Geology of North America, v. G-3, The Cordilleran Orogen, Conterminous U.S.: Geological Society of America, Boulder, Colorado., p. 553-582.

Wheeler, H.E., and E.F. Cook, 1954, Structural and stratigraphic significance of the Snake River capture, Idaho-Oregon: Journal of Geology, v. 62, p. 525-536.

Whitehead, R.L., 1992, Geohydrologic framework of the Snake River Plain, Idaho and eastern Oregon: U.S. Geological Survey Professional Paper 1408-B, 32 p.

Wood, S.H., 1989, Silicic volcanic rocks and structure of the western Mount Bennett Hills and adjacent Snake River Plain, Idaho, *in* R.P. Smith and K.L. Ruebelmann, eds., Snake River Plain-Yellowstone Volcanic Province: 28th International Geological Congress, Guidebook T305, American Geophysical Union, Washington, D.C., p. 69-77.

———, 1994, Seismic expression and geological significance of a lacustrine delta in Neogene deposits of the western Snake River Plain, Idaho: American Association of Petroleum Geologists Bulletin, v. 78, p. 102-121.

———, 1997, Structure contour map of the base of Quaternary basalt in the western Snake River Plain: Idaho Department of Water Resources, Treasure Valley Hydrologic Project, Boise, scale 1:100,000.

Wood, S.H., and J.E. Anderson, 1981, Geology, *in* J.C. Mitchell, ed., Geothermal Investigations in Idaho, Part 11, Geological, Hydrological, Geochemical, and Geophysical Investigations of the Nampa-Caldwell and Adjacent Areas, Southwestern Idaho: Idaho Department of Water Resources Water Information Bulletin 30, p. 9-31.

Wood, S.H., and J.N. Gardner, 1984, Silicic volcanic rocks of the Miocene Idavada Group, Bennett Mountain, Elmore County, southwestern Idaho: Final contract report to the Los Alamos National Laboratory from Boise State University, 39 p., 1 map, scale 1:100,000.

Wood, Steven, and S.H. Wood, 1999, Large pumice-block layer marks sublacustrine rhyolite-dome eruption in the Miocene Chalk Hills Formation, near Marsing, Idaho, western Snake River Plain: Geological Society of America Abstracts with Programs, v. 31, no. 4, p. A-62.

Zoback, M.L., E.H. McKee, R.J. Blakely, and G.A. Thompson, 1994, The northern Nevada rift: Regional tectono-magmatic relations and middle Miocene stress direction: Geological Society of America Bulletin, v. 106, p. 371-382.

Zoback, M.D., and M.L. Zoback, 1991, Tectonic stress field of North America and relative plate motions, *in* D.B. Slemmons, E.R. Engdahl, M.D. Zoback, and D.D. Blackwell, eds., Neotectonics of North America: Geological Society of America Map Volume 1, p. 339-365.

Stratigraphic Studies of the Boise (Idaho) Aquifer System Using Borehole Geophysical Logs With Emphasis on Facies Identification of Sand Aquifers

Report to the Treasure Valley Hydrologic Study
Idaho Department of Water Resources

By

Edward Squires
Hydro-logic, Inc.
988 S. Longmont Ave.
Boise, Idaho 83706
hlogic@msn.com
208-342-8369

and

Spencer H. Wood
Department of Geoscience and CGISS
Boise State University
1910 University Drive
Boise, Idaho 83725
swood@boisestate.edu
208-426-3629

January 8, 2001

Stratigraphic Studies of the Boise (Idaho) Aquifer System Using Borehole Geophysical Logs With Emphasis on Facies Identification of Sand Aquifers

Report to the Treasure Valley Hydrologic Study, Idaho Department of Water Resources

By

Edward Squires
Hydro-logic, Inc., 988 S. Longmont Ave.
Boise, Idaho 83706
hlogic@msn.com, 208-342-8369

and

Spencer H. Wood
Department of Geoscience and CGISS
Boise State University, 1910 University Drive
Boise, Idaho 83725
swood@boisestate.edu, 208-426-3629

January 8, 2001

Index

Page

Figures

Introduction

The cold-water aquifer system beneath the City of Boise is composed of sandy sediments interbedded with claystone and mudstone that were deposited near the shores of lakes which filled the western Snake River Plain during the late Miocene and Pliocene epochs (10 to 1.7 million years ago) (Figures 1and 2). The sand layers are the deposits of stream channels, beach sands winnowed by wave action, deltas built out into the lake, and possibly density-flows across the lake bottom from collapse of parts of the delta shelf. These depositional environments do not produce broadly distributed sand layers. Instead the sand layers are typically restricted in their horizontal and vertical continuity by interbedded mudstone or lateral termination into mudstone. The difficulty here lies with correlation of sand layers and determination of their shapes. Important is to predict whether sand layers found in wells have some sort of hydraulic connection, and which are not interconnected. By analogy to modern sedimentary environments and subsurface studies by others, our goal is to obtain at least a partial understanding of the three-dimensional geometrical shapes of sand aquifers.

Structural downwarping coupled with normal faulting along the margins of the plain further complicates deciphering the stratigraphic section. For example, a 4-degree dip of strata (370 feet per mile), results in strata identified in one well, being 370 feet deeper in a well one mile away in the down dip direction (Figure 2). Stratigraphic offset along down-to-basin normal faults (up to 800 feet of vertical throw on some faults) pose additional complexity, because in the older strata, faulting was contemporaneous with deposition.

For this report, we have compiled available geophysical logs (natural gamma, single-point resistance, or normal-resistivity logs at a scale of 1 inch equals 200 feet for the wells shown in Figure 3. Most of the logs are owned by United Water Idaho, Inc. and we have been permitted to use other data by the City of Boise, Micron Technology, and the Idaho Department of Transportation. On many of the United Water Idaho (UWID) wells, drill cuttings were examined by geologists, and the grain size of sands are known. On other wells, the driller's descriptions are the only record of lithology, and the grain-size of sands is not consistently reported. We explain our attempts to correlate geologic strata and identify depositional facies for the line of wells from Claremont Subdivision (end of 8th Street in the foothills) to the Cassia Street well, and for the line of wells from " Hill 3431" above Stewart Gulch in the foothills to the McMillan well of west Boise. Correlation from the Micron Technology plant area to Cassia Street are still preliminary, and not included. Correlation from Idaho Street well to the Bethyl Street well and west to the St. Lukes well (at east Meridian) is currently being evaluated using seismic reflection data acquired by the Boise State University CGISS group this summer of 2000.

Methods

Facies interpretations of geophysical logs are discussed in Rider (1996), Berg (1986); and Galloway and Hobday (1996); however, examples from lacustrine basins are few. Most of the examples from Rider's interpretations (shown in Figure 4) are from marine environments; however, his examples of a channel-point bar, delta-border progradation, transgressive shelf, and prograding shelf are probably applicable to the lacustrine setting. The study of the Pliocene lacustrine Ridge Basin in California by Link and Osborne (1978) is helpful, because they interpret logs of a mouth bar complex overlying a delta

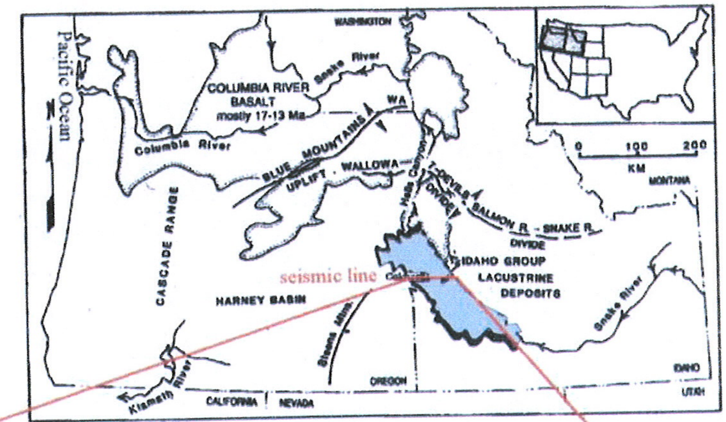


Figure 1. Location of Neogene Lake Idaho deposits in the northwest United States (from Wood, 1994)

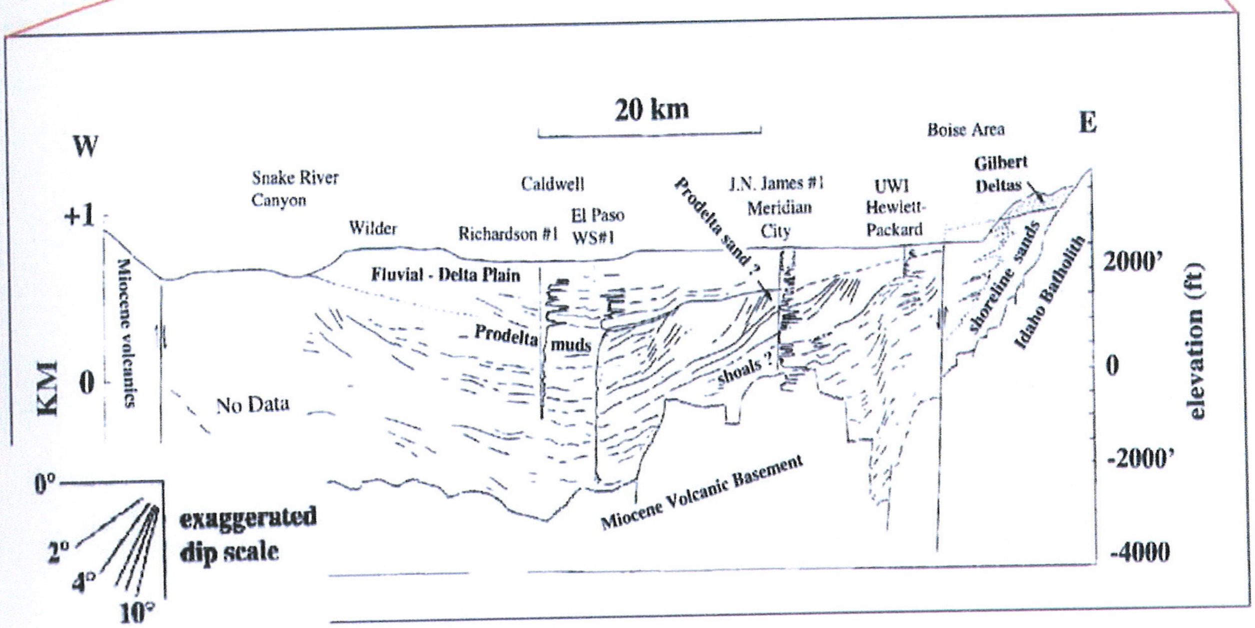


Figure 2. Interpreted cross-section from seismic lines across the western Snake River Plain, showing the geometry of strata that reflect seismic waves within lake and stream deposits and the resistivity response of borehole geophysical logs of deep wells. Drawn from petroleum industry seismic reflections lines which were focused on the deeper strata and did not image the strata shallower than 700 ft depth (i.e., the main cold-water aquifer section).

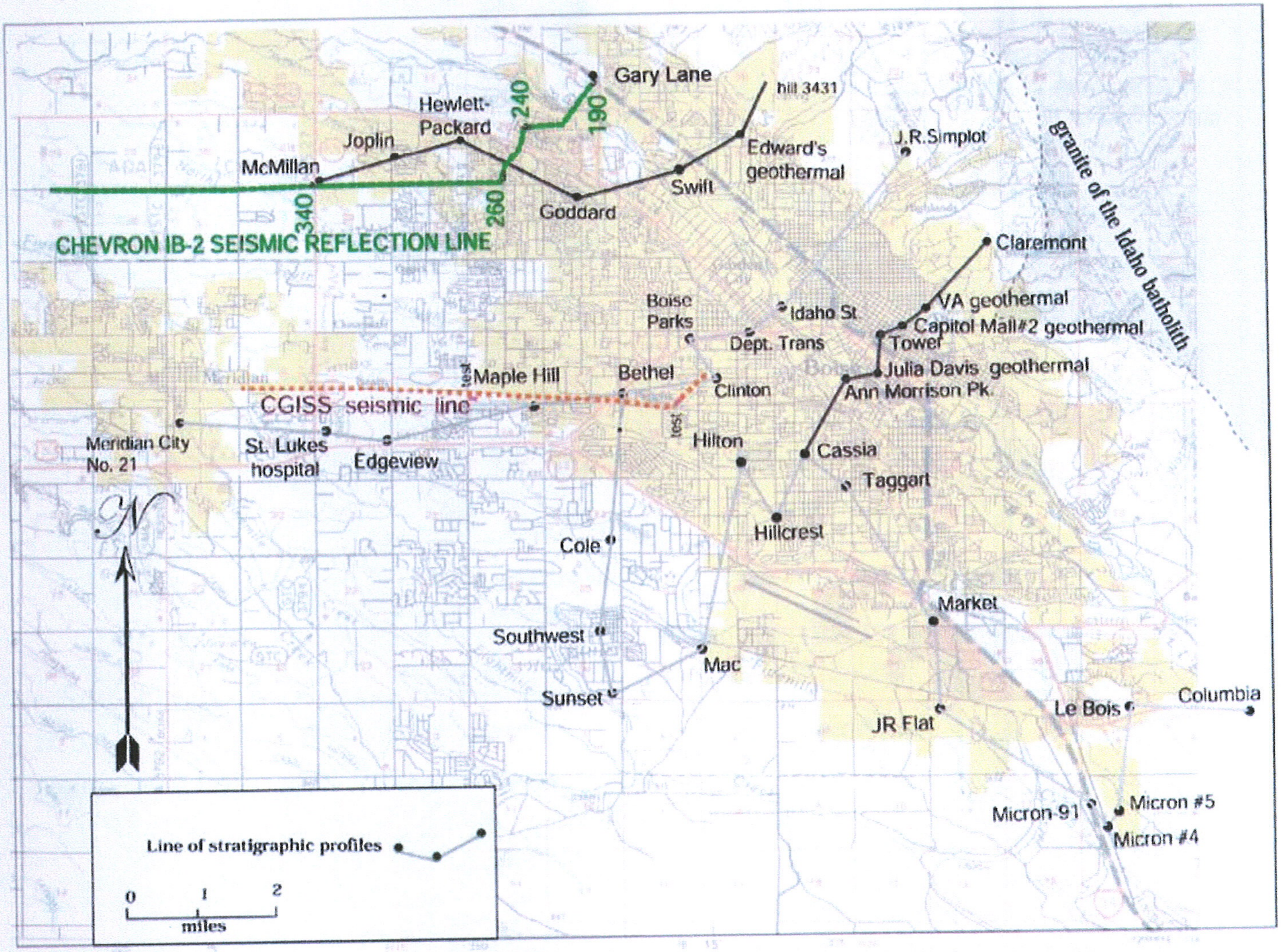


Figure 3. Map showing deep wells in the Boise area and locations of seismic lines. Profiles of Figures 9 and 10 are highlighted lines.

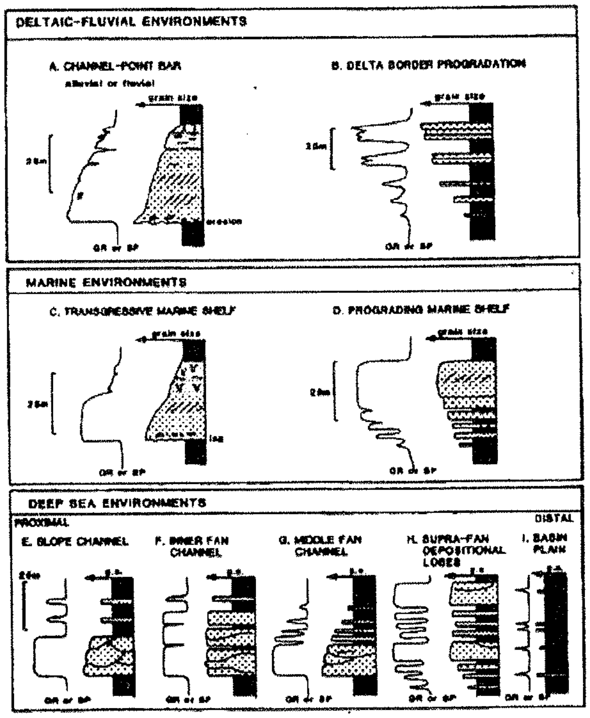


Figure 4. Idealized well log shapes (gamma or SP, or mirror image of resistivity) for sedimentary facies (From Rider, 1996).

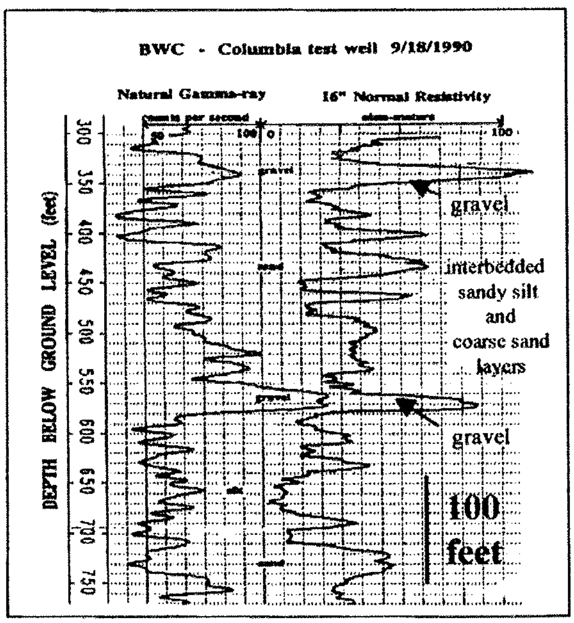


Figure 5. Geophysical log character of the mixed alluvial-fan - braided-stream facies beneath southeast Boise. Gravel layers are typically 20 feet thick and have higher radioactivity (high counts/sec on the natural gamma log) on account of a large percentage of cobbles of high-potassium porphyritic felsite. Gravels show a tendency to be more silty upwards (fining of matrix upward), and have abrupt bases typical of channels. Sands are typically 20 feet thick with abrupt bases and fining upward. Silty beds show a "spiky" character on the natural gamma log. (From Squires and others, 1992).

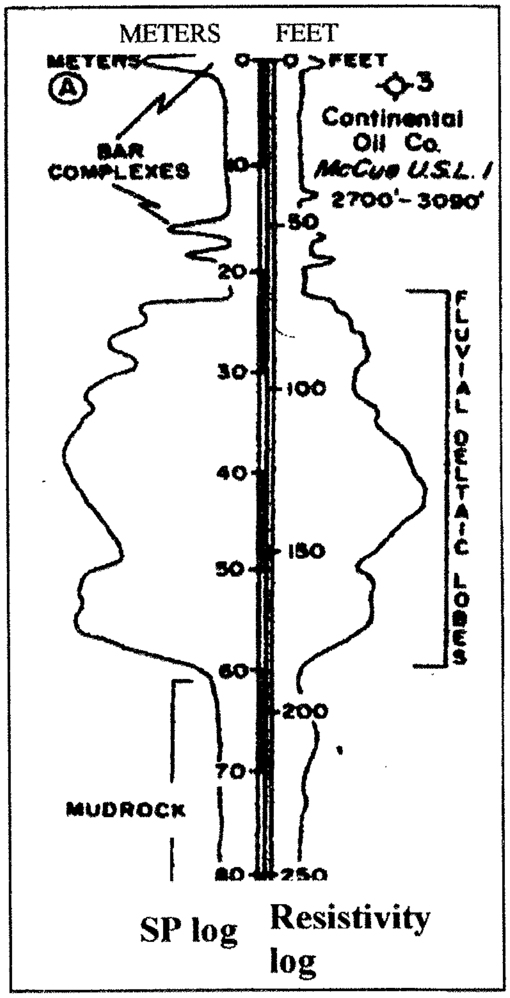


Figure 6. SP and resistivity log of lacustrine prodelta mudrock overlain by delta sand lobes, overlain by a mouth-bar complex: Ridge-Basin, California (from Link and Osborne, 1978).

1)**Funnel-shaped logs, over 20 to 100 m vertical extent**, overlain by abruptly more resistive sands, and correlative over 10's of km. These are interpreted as prodelta coarsening upward mud sequences overlain by channel sands.

2)**Funnel-shaped logs, of 5 to 20 m vertical extent**, and repeated in the section. Several to 10 or more units are repeated over intervals of 30 to 100 m. Units are rarely correlative over a few km. Sands generally medium-grained. These are interpreted as stacked mouth bars of streams entering the lake, or as beach sands.

3) **Monotonous low resistivity, and irregular medium gamma** activity, over vertical extents of 30 to 400 or more meters, interpreted as muds deposited in the open lake environment.

4) **Thin bell shaped units, less than 3 m thick**, and enclosed by thick muds, are interpreted as fining upward density-flow sands, and generally made up of fine well-sorted sand.

5) **Bell-shaped logs, with abrupt bases, 7 to 40 m in vertical extent**. Sands are typically coarse, and some have gravel at base. These are interpreted as fluvial channels, and rarely correlative over a few km.

6) **Spikey logs, with resistive units 3 to 7 m,** and low resistance units typically 3 m in vertical extent. 7 m units with abrupt bases, and high gamma. These are interpreted as alluvial fan deposits, with high-gamma rhyolite gravel with abrupt bases, and alternating silt and muddy sand layers none of which can be individually correlated.

Figure 7. Useful electric and natural gamma-ray signatures of depositional facies in Lake Idaho sediments (from Wood and others, 2000)

lobe (Figure 6). Studies by Ainsworth and others (1999), Flint and others (1988, 1989) of Miocene lake sediments in Thailand illustrate the continuity of bars and other lacustrine sands. They found mouth-bar sands to be coarsening upward (funnel-shaped logs), fine-grained sands with a sheet geometry of variable thickness. Individual sheets of mouth-bar sands, vary in thickness from 0.5 to 5 meters,and could be traced for a distances up to 5 km. Their studies of a Thailand oil field was for wells about 1 km apart accompanied by a 3D-seismic survey, a set of data far more detailed than that usually available in water resource studies.

The shapes of logs we have found useful in interpreting facies in the sedimentary section beneath Boise are summarized in Figure 7. Squires and others (1992) identified the log character of mixed alluvial-fan-and-braided-stream facies in the southeast Boise area (Figure 5).

The great value of borehole-geophysical logs is their ability to detect sand and gravel aquifers, and distinguish aquifers from low permeability mudstone, siltstone, and clay. The best log is one that measures electrical resisitivity in an open, uncased hole full of water or drilling mud (the 16 or 64-inch normal log, or induction logs). This log shows high resistivity in a fresh-water-filled clean sand or gravel aquifer, but the response is greatly reduced by low-resistivity (high-conductance, > 400 μS) groundwater in the aquifer. The single-point resistance log gives a similar response to resistivity in the sand and mudstone section and is equally useful, though not quantitative. Electrical resistivity logs have low values in silt and clay bearing sediments, thereby distinguishing them from aquifer sands and gravels. Natural gamma logs are generally useful, but their similar response in silt-clay and gravels with cobbles of high potassium content make an ambiguous interpretation of gravel or clay, unless accompanied by a resistivity or resistance log.

Background Geology and Problems

In the past, aquifers were typically named for the geologic formations in which they occurred. However, the variety of depositional environments of the lake-stream systems and the changing environments with fluctuating lake levels tells us that the sand units are complex. In previous reports on the western plain (Whitehead, 1992), the aquifer systems are associated with a set of geologic formations originally defined by Malde and Powers (1962). The stratigraphic order and characteristic lithology of the formations is a useful framework, because the changing lithology in some cases can be attributed to basin-wide geologic events or progressions of similar depositional environments across parts of the basin. However, it is unlikely that these formation units reliably relate to hydraulic connectivity of aquifers. In the areas Malde and Powers had geologically mapped, they found a general sequence of granitic rocks overlain by rhyolite, overlain by basalt, overlain by lake and stream sediments. They recognized an early set of lake and stream deposits that had been tilted, faulted and beveled off by erosion which they called the Chalk Hills Formation. Their concept of the Banbury Basalt beneath the Chalk Hills Formation is misleading, because we now know that many local basalt fields erupted during the time lake and stream sediments were accumulating in the basin (Wood and Clemens, in press). The Poison Creek Formation was originally identified on the south side of the plain as sediments dominated by one or several layers of 1-to-7-meter-thick fluvial or coarse-grained delta sand at or near the base of the lake and stream section.

These sands are discontinuous, and the Poison Creek Formation is best regarded as just a local sand facies of the Chalk Hills Formation (Wood and Clemens, in press). The basis for the top of the Chalk Hills Formation is well established as an unconformity with slight angularity on the south side of the western plain. On the south side, it is overlain by beach gravel and in places shoreline carbonate oolitic sands. On the north side of the plain, we have searched for such a straight-forward contact of a widespread angular unconformity overlain by the oolites or gravel, but have it only in one locality in the Boise foothills (discussed later in this section), and are uncertain of its occurrence in the subsurface, as discussed later in this report.

In the Boise foothills and in the subsurface are several gravel and coarse-sand occurrences indicating stream deposition or beach gravel over lake-bed muds. This can result from streams flowing out over the exposed mud surface as the lake lowers. These layers can also result from beach sand facies being spread as waves work the shoreline of a rising lake. We are looking for criteria to identify both situations, but the relationships in outcrop and on the well logs are not yet obvious to us. The gravel and coarse sand layers occur at several horizons within the mudstones, but we are still uncertain of the correlation of many of these layers.

Carbonate oolitic sands lenses occur in a 400-ft thick section of shoreline sands in the foothills which Burnham and Wood (in press) have always considered a part of their definition of the Terteling Springs Formation. The oolite bearing section varies greatly in thickness, and appears to be absent in places. The oolites were deposited as shoreline sand beaches and bars, and therefore occur only as lenses within the sediments. The oolite lithology appears to be a unique lithology associated with lake level rise (Wood and Clemens, in press; Repenning and others, 1994, p. 54). In the subsurface beneath Boise, oolite sands have been documented in the UWID'S Cassia, Taggert, and Cole water wells (locations of these wells is shown on Figure 3) at a relatively shallow depth (<300 ft), and above elevation 2500 ft.

Oolite sand also occurs in the Veterans Administration (VA) Reinjection Well, at a depth of 450-ft deep at elevation 2,300 feet. We mention this here, because the oolite layer is only 90 feet above the basalt of Aldape Heights which Clemens and Wood (1993) have dated by K-Ar as 9.4 million years old. The oolite sand occurrence here is not well understood. Because oolite in the VA well is only 90 feet above the basalt, it may be that most of the Chalk Hills sediments were eroded from this locality near the edge of the foothills, and oolite of the Terteling Springs Formation transgressed over the basalt surface. It is unlikely that the oolite in the VA well is a carbonate facies within the older Chalk Hills Formation, because nowhere else in the basin has carbonate sediment been found in the Chalk Hills Formation. The oolite layer in the VA well has probably been faulted down at least two hundred feet deeper than occurrences in the 3 UWID wells of west central Boise, suggesting a graben structure near the edge of the foothills.

We believe the sediments containing lenses of shoreline oolites to be a transgressive sequence over an unconformity at the top of the Chalk Hills Formation. An angular unconformity is mapped at the base of the oolite in Stewart Gulch (NW ¼, sec. 22, T. 4 N., R. 2 E.), but elsewhere in the foothills the contact appears to be conformable.

Overlying the oolite-bearing section in the western Boise foothill outcrops, is the "Pierce Park Sand", a 150-to-250- foot-thick layer of coarse sand. This thick sand represents a large "Gilbert-type" delta system. Where the oolite section is absent, the

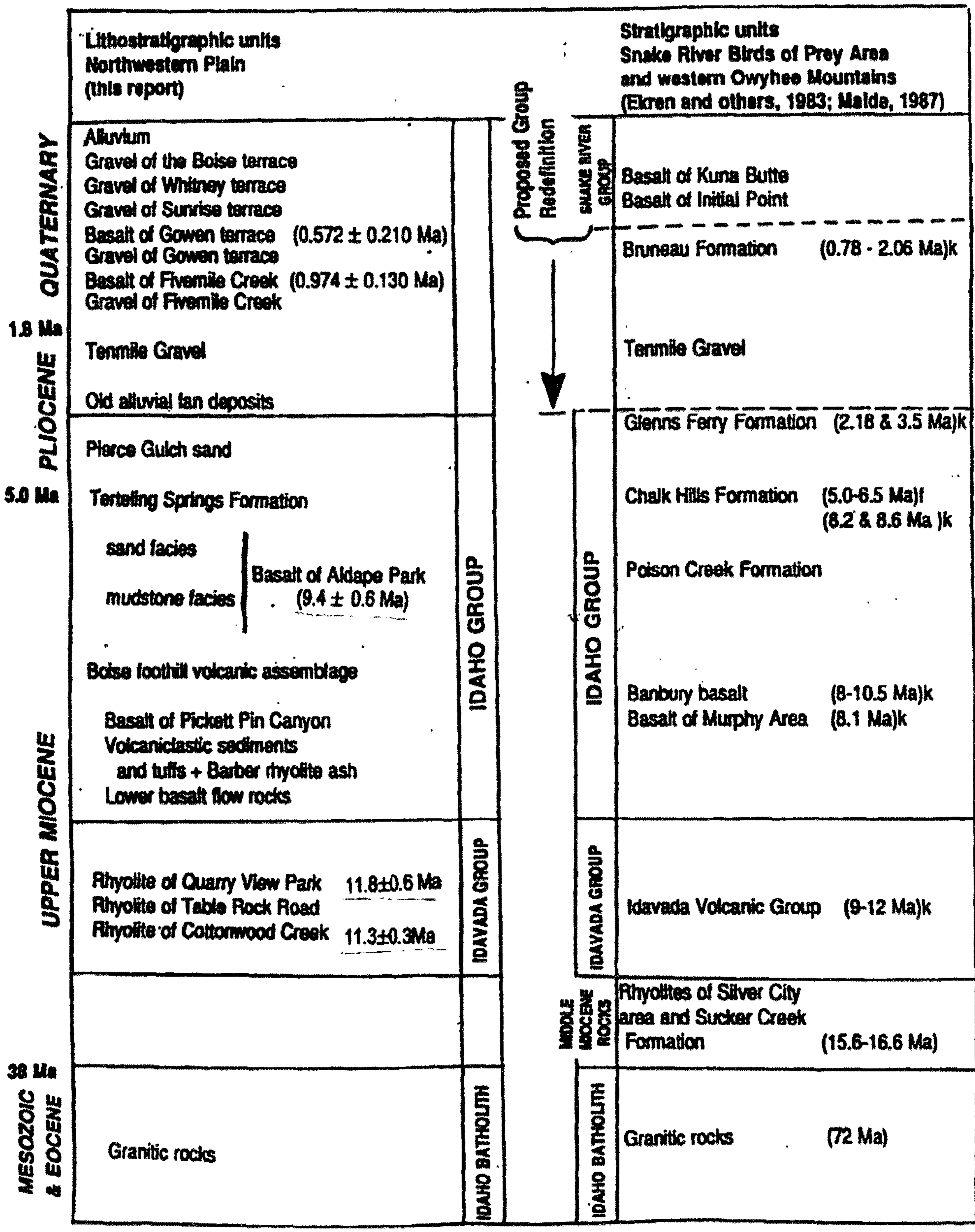


Figure 8. Stratigraphic names in the western Snake River Plain: Comparison of stratigraphic units of the Boise area (from Burnham and Wood, in press) to units mapped on the south side of the plain by other workers.

Pierce Park Sand conformably overlies mudstone. The reason the coarse sand directly overlies mudstone is because the delta prograded basinward over muds of the deep-lake deposits. The delta apparently formed as the level of Lake Idaho began lowering after it reached its spill point into Hells Canyon, probably in the early Pliocene (about 4 or 5 million years ago).

For the first time, in this study, we attempt to correlate the main features mapped in the Boise foothills with the subsurface sedimentary sequence in water wells beneath the city. We make an assumption that the oolite sequence is a singular occurrence in the stratigraphic record. We allow that gravel occurrences in the subsurface could be channel gravel, beach gravel, or river-terrace gravel, and that all gravel occurrences require a beach or a stream environment for deposition. The occurrence of an angular unconformity is not generally detectable in geophysical logs or the sequence of drill cuttings. Logs that detect dip (dip meter, resistivity borehole imager, or acoustic televiewer) have not been run in the Boise area water wells. To detect dip and angular unconformities, the configuration on high-resultion seismic reflection is also useful.

Discussion of the NE-W section from Claremont Subdivision Well to Cassia Street Well (Figure 9)

The dip of the deeper section between the Capitol Mall #2 well and the Julia Davis Well may be as much as 11° to the southwest as determined by a seismic reflection survey by Liberty (1998). The deeper sediments are a mudstone sequence more than 800 feet thick, shown on the borehole geophysical log of the Julia Davis well. Overlying the mudstone, between elevations 2,100 and 2,280 feet, is a sequence of stacked, coarse sand layers, each one 25 to 70 feet thick. Each of the upper 4 sand layers shows a coarsening-upward log signature. The log character of this sand sequence is similar to that shown in the Ridge Basin, California study by Link and Osborne (1978) as a delta-front bar complex (Figure 6). Similar beds occur near the bottom of the Ann Morrison Park well between elevation 2,000 and 2,200 foot elevation.

The overall funnel shape of the Cassia Street logs indicates a fine and medium sand delta above the silty mud at 2,100 foot elevation. These delta sands appear to correlate to sands in the Ann Morrison Park Well, but the funnel-shaped log signature is not as prominent in this well. That correlation implies an apparent dip of about 1.4° in the southwest direction between these two wells, which seems too low a dip, in view of deeper strata dipping in excess of 10 degrees in southwest direction. Another possibility is correlation of the Ann Morrison Park well sands to the deepest sands of the Cassia Street well.

Gravel occurs at the bottom of the Cassia Well at elevation 1,700 feet, but that part of the section was not geophysically logged. No other wells drilled gravel at that depth. The gravel indicates either a gravel beach deposit or a stream channel deposit. The gravel was deposited about 4 miles basinward from the basin margin indicating that the lake environment was not at this locality at the time of deposition: this locality was either a stream bed or a shoreline.

The major sand section of these wells is below an unconformity marked by an upper gravel, generally occurring in wells between elevation 2,400 and 2,550 feet (line colored pink on Figure 9). In the next paragraph we discuss evidence interpreting this

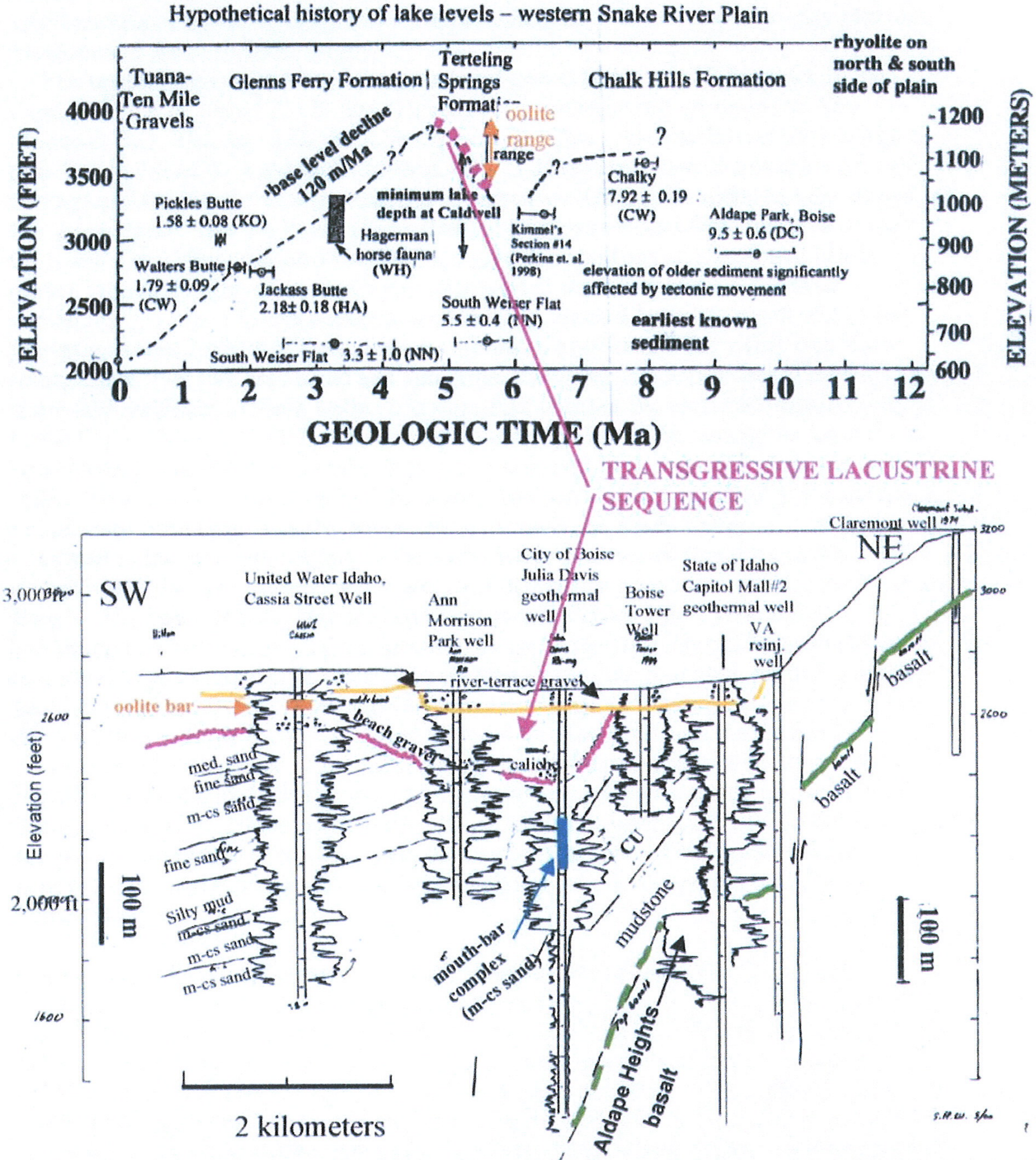


Figure 9. Correlation of strata in wells in a section from the Boise foothills (Claremont well, Boise Heights) to the Cassia Street Well. Basalt of Aldape Heights (9.5 Ma) shown in green. Mouth-bar complex of a sand delta deposited over mudstone shown in blue at the Julia Davis well. CU refers to coarsening upward sand layers, shown by small arrows. Dark pink line is an unconformity at about 2,400-ft elevation, indicated by Carbon-13 isotope analysis of caliche in the Julia Davis Well by Cavanagh (2000, p. 68-72). The unconformity surface is identified by occurrence of gravel in other wells, and is interpreted as the top of the Chalk Hills Formation. Sediment above the unconformity contains oolite, and is interpreted to be a part of the Terteling Springs Formation. Quaternary-aged terrace gravel shown in yellow. Hypothetical lake history corresponding to deposits is from Wood and Clemens (in press).

unconformity as the top of the Chalk Hills Formation. The deeper sand section in these wells on the north side of the Boise River all appear to be good aquifers down to at least 1,600 elevation; however, temperatures approaching 85°F below that level may preclude development for cold-water supply.

The unconformity is marked by a prominent gravel bed about 30 feet thick in the Cassia Well at elevation 2, 550 feet. This gravel correlates with gravel in the Ann Morrison Park Well and with the Julia Davis well. This gravel horizon is shown with a pink line in Figure 9. Cavanagh (2000, p. 68-72) discovered from examination of well cuttings and Carbon-13 isotopic analysis of carbonate that a soil caliche had developed in this gravel layer. This discovery shows that the lake had receded from this area at one time, and a soil formed upon a river-terrace gravel, or perhaps an abandoned beach gravel. Above this gravel, in the Cassia well, a 30-ft thick carbonate oolite sand containing gastropod fossils occurs as a coarsening upward layer at a depth of 110 feet (elevation about 2,670 ft.) These features are taken as evidence that rising lake water transgressed over this gravel and and laid down the upper 200 to 300 feet of mostly near-shore lake sediment of these wells. It is uncertain whether the erosional unconformity is in the Capitol Mall#2 well. It may warrant re-examination of the cuttings to see if it can be identified. No obvious correlative gravel is reported from that well or the recently drilled Tower well, suggesting that this gravel horizon was not deposited to the northeast, or has since been eroded by the downcutting of the Boise River Valley.

Gravel of the active floodplain of the Boise River is about 60 feet thick in the downtown Boise area of these wells, and so is the terrace gravel that mantles the Whitney Bench. The bases of both river gravels are shown in Figure 9 by a yellow line.

Although we do not have a good geochronology of the lake deposits beneath Boise, the major geologic events can be tentatively correlated with the proposed, albeit hypothetical lake history proposed by Wood and Clemens (in press). In this correlation, none of the deposits shown in the section of Figure 9 would be correlated with the Glenns Ferry Formation. The very thick sediment section beneath the transgressive sequence would be regarded as the Chalk Hills Formation. The overlying sequence would be the Terteling Springs Formation. To the north of here (Swift Well to McMillan Well) is a delta sequence that appears to correlate with the Pierce Park sand. The Pierce Park sand is believed to be correlative in time with the declining lake level and the Glenns Ferry Formation.

Discussion of E-W section from Swift Well to the McMillan Well (Figure 10)

The two Swift-well sand aquifers appear to correlate with the Hewlett-Packard well, almost straight across; however, the Chevron Seismic data shows the lower sand in Swift to have an apparent west dip of about 2°, and to lie at least 300 feet below the bottom of the Hewlett Packard well (Figure 10). Well-log character of the deeper aquifer sand in the Swift Well has an abrubt base, typical of a channel sand, and the grain size is medium-fine sand. That lower aquifer appears to correlate with the bottom sand aquifer in the Goddard well. However, information on that bottom aquifer in the Goddard well is limited to the 1968 driller's description (see Squires and others, 1992, p. 94). It is problematic that no major sand units occur in the up-faulted foothills section to the northeast, either in outcrop or in wells, that might correlate with the lower Swift well

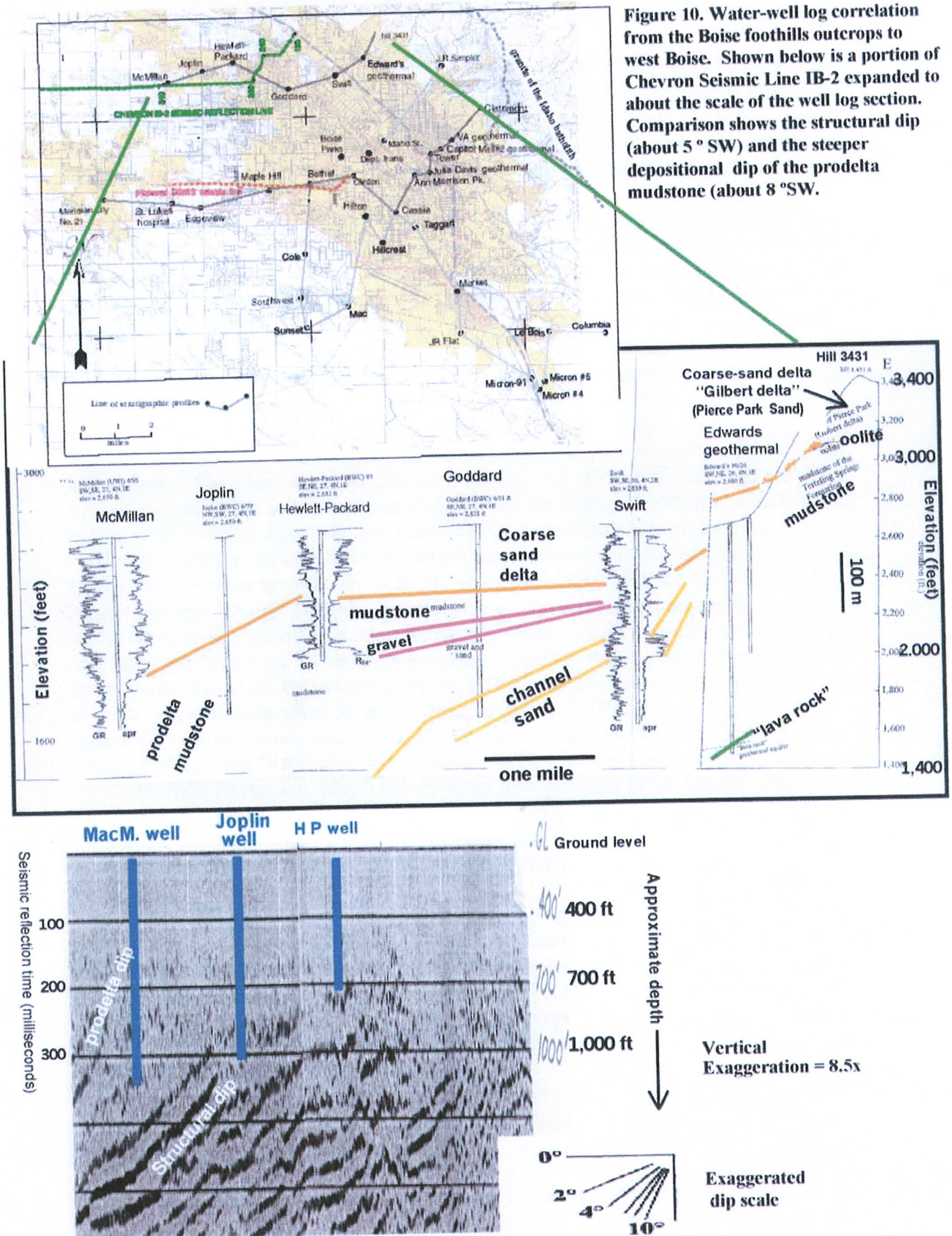


Figure 10. Water-well log correlation from the Boise foothills outcrops to west Boise. Shown below is a portion of Chevron Seismic Line IB-2 expanded to about the scale of the well log section. Comparison shows the structural dip (about 5 ° SW) and the steeper depositional dip of the prodelta mudstone (about 8 °SW.

aquifer. The nearby deep wells in the foothills (Hillside Jr. High, Quail Hollow Golf Course, and wells on the Terteling Stewart-Gulch properties) drilled mostly mudstone according to driller's logs. Possibly the section containing the lower Swift well aquifer sand has been eroded from the upthrown fault block, and for that reason it does not occur in the foothills. Alternatively the lower sand may be a channel deposit that simply was not deposited to the northeast.

One and one-half miles to the northwest of the Swift well, an 80-ft-thick coarse sand occurs in the UWID Gary Lane well below 350 feet of mudstone at a depth of 740 feet. In this well, the grain size is mostly 0.5 to 2 mm, and much coarser than the medium-fine sand reported for the lower Swift well aquifer. Otherwise, it appears to be a similar and possibly correlative sand. Three and one-half miles to the northwest of the Swift well is the Treasure Valley Hydrologic Project Monitoring Well #1 (TVHP#!) in which several medium-fine-grained sand beds occur beneath 300 feet of mudstone at a depth of 770 feet. Individual beds are 2 to 10 feet thick, and the sandy sequence is 120 feet thick (Dittus and others, 1997). Grain-size and depth are similar to the lower Swift well aquifer. The TVHP#1 well was not completed below 340 feet, so the hydraulic connection to the other two wells cannot be studied. All of these occurrences are probably along strike and indicate sand about 700 feet deep and 300 feet beneath a thick mudstone unit, suggesting they may correlate and be hydraulically connected.

Gravel occurs at about 2,100-ft elevation near the bottom of the Hewlett Packard well, but it does not occur in the Swift well. The Hewlett Packard well is situated about 6 miles basinward from the basin margin, therefore the gravel must represent a river or beach out in the basin at the time of deposition. Its absence in the foothills, requires that the gravel was not deposited there, or has been eroded from the section nearer the foothills. It is problematic that no gravel occurs in the Swift Well, but the "pink line" is projected east to show where an unconformity might occur in that well. "Red-colored" gravel is reported by the driller at a corresponding level in the Goddard well (see Squires and others, 1992, p. 94). We suspect the gravel in the Hewlett Packard well marks the top of the Chalk Hills Formation, which we represent by a pink line on the section (Figure 10). The mudstone above the gravel is tentatively correlated with the Terteling Springs Formation, deposited by the rising transgressive lake water. That mudstone, changes upward to the prodelta mud of a delta prograding basinward, discussed in the next paragraph.

In the foothills, and in all wells on the section depicted in Figure 10 is a clear signature of a delta prograding into the Basin. The delta signature on the logs the a funnel shape of side-by-side gamma and resistivity logs (Figure 4 B and D) seen in the Swift and Hewlett Packard wells above 2,350-ft elevation, and in the McMillan well above 1,900-ft elevation. The sand thickness and depth increase progressively to the west. The delta sand is coarse in the foothills, and the driller reported coarse sand in the depth interval 153 to 280 feet in the McMillan well.

Because this is the uppermost delta in the lacustrine sequence, we correlate it to the Pierce Park sand that crops out in the upper part of the foothills section west of Crane Creek. In the foothills, this unit is mostly foreset beds of coarse sand typical of the "Gilbert-type" of delta. Some foreset bed sets are 60 feet thick, and the sand unit as a whole is up to 250 feet thick in the foothills (Burnham and Wood, in press). This delta is then correlated by Wood and Clemens (in press) to the history of Lake Idaho. Since it is the uppermost major delta in the section its deposition over mudstone, is explained as a prograding sand delta in response to the slow lowering of lake levels after Lake Idaho spilled over into Hells Canyon. In the lake history context, this uppermost delta should correspond to the upper part of the Glenns Ferry Formation of Malde and Powers (1962) and Reppening and others (1994) which crops out south and southeast of the Boise area along the Snake River

We feel fairly certain that there is a "long term" hydraulic connection in the sands of the upper delta sequence (Figure 10); however, local lenses of mudstone in that section may prevent short-term detection of well-drawdown responses. It may take months to decades for large drawdowns to propagate through this seemingly continuous section of interbedded sand and thin muds.

Other older sand deltas surely occur in the lacustrine section. For example in Figure 9, the lower sands in the upper 800 feet in the Julia Davis well are mouth-bar sands of a delta sequence over mudstone. However because this section is steeply dipping, this delta is much older than the upper 800 foot section of the McMillan well (Figure 10). In our interpretation the delta sands in the Julia Davis well are part of the much older Chalk Hills Formation.

Comments on structure

We did not re-interpret the structure along the two cross-sections; however the framework published by Squires and others (1992, their Figure 11) generally concurs with observations in this study. They show the Eagle-west Boise fault with 800 feet of offset between the Swift well and Goddard well using data available to them at that time. The 800-ft offset they show is the offset of the volcanic basement. The fault does not necessarily offset of the upper section. The aquifer section below 2000-ft elevation (about 700 ft deep) may be offset, but the Eagle-west Boise fault does not appear to significantly offset the upper delta sequence.

Squires and others (1992) also show the "foothills fault zone", just northeast of the Swift well. In our Figure 10, the base of the upper delta is shown as offset by 300 feet with respect to the foothills outcrops, however this elevation shift could also be due to tilting without significant faulting. It is likely that a major fault or fault system occurs between the Swift well and the Edwards geothermal well and that this fault offsets the deeper section. Lava rock is reported by the driller at elevation 1,500 ft in the 1926 well (Idaho Supreme Court Records, *Silkey vs. Teigs,* 1931) showing that the "volcanic basement is relatively shallow here (Figure 10). None of these geothermal wells have been logged by borehole geophysics, and we do not know whether the geothermal aquifer in this area is basalt or rhyolite.

Conclusions

In this study we have identified gravel layers, oolite beds, thick mudstone sequences, and thick delta sequences that appear to correlate with regional concepts of fluvial and lacustrine deposition as outlined by Wood and Clemens (in press). We are still uncertain how many gravel layers are in the section; however, occurrence of gravel layers out in the basin, several miles basinward from the basin margin is an indication that the lake level lowered several times and streams flowed over exposed lake beds and delivered gravel away from the uplands. Some gravel may be beach gravel, and some may be channel or remnant terrace gravel; however, it had to be originally delivered to the site by a stream.

Our best evidence that a lake transgressed over the gravel is the occurrence of oolite sands in the shallow section of the Cassia, Taggert, and Cole wells (Figure 3 and 9). This section shown in Figure 9 also indicates that there may be topographic relief upon the tilted and eroded older section. This is also our best evidence that the sedimentary sequences shown in this profile (Figure 9) correlate to the Chalk Hills Formation and the overlying Terteling Springs Formation. We also conclude that the Pierce Park sand (correlative with the upper Glenns Ferry Formation) is absent from the section shown in Figure 9, but the Pierce Park Sand is the same sand as the upper delta sequence in the Swift, Hewlett-Packard, and McMillan wells shown in Figure 10.

Deeper sedimentary strata comprising the aquifer system have been tilted basinward (to the southwest, 2 to 11 degrees. The deeper strata are offset by faulting, but the amount of offset is known only where seismic data is available. The extent to which shallow parts of the aquifer system are faulted and tilted has not been determined, because the available petroleum-industry data was not focused on the shallow section (<600 feet) (see seismic section in Figure 10).

Eventually, the aquifer geometry can be determined accurately using high-resolution seismic reflection methods and continued diligence in obtaining borehole geophysical logs and cuttings examination of drilled water wells. We hope that this study provides a framework for future study.

Ackowledgements

The authors would like to thank United Water Idaho, Inc. for granting permission to use their well data. Roger Dittus was a great help in many aspects of this study. Terry Scanlon helped us with data from Idaho Dept. of Transportation, and City of Boise, and Micron wells. The study has benefited from discussions with Christian Petrich and Jon Hutchings of the Idaho Water Resources Research Institute and the Idaho Department of Water Resources. This study was supported by the Treasure Valley Hydrologic Project of the Idaho Department of Water Resources.

References Cited

Ainsworth, R.B., Sanlung, M., and Duivenhvoorden. S. T. C., 1999, Correlation techniques, perforation strategies, and recovery factors: an integrated 3-D reservoir modeling study: Sirikit Field, Thailand: AAPG Bulletin, v. 83, p. 1536-1551.

Berg, R.R., 1986, Reservoir Sandstones: Prentice-Hall, Inc., Englewood Cliffs, New Jersey, 481 p.

Burnham, W.L., and Wood, S.H., in press, Geologic Map of the Boise South 71/2 minute quadrangle: Idaho Geological Survey Technical Report Series: 1:24,000.

Cavanagh, B.C., 2000, Western Snake River Plain, fluvial-lacustrine sedimentation: Exhumation estimates from mudstone compaction, unconformity identification by buried soil carbonate, hydraulic conductivity estimates from well cuttings: M.S. dissertation, Boise State University, Boise, Idaho , 96 p

Clemens, D. M., and Wood, S.H., 1993, Radiometric dating, volcanic stratigraphy, and sedimentation in the Boise foothills, northeastern margin of the western Snake River Plain, Ada County, Idaho: Isochron/West, v. 59 p. 3-10.

Ekren, E. B., D. H. McIntyre, E. H. Bennett and H. E. Malde, 1981, Geologic map of Owyhee County, Idaho, west of Longitude 116° W: U. S. Geological Survey Map I-1256, 1:125,000.

Flint, S., Stewart, D.J., Hyde, T., Gevers, E.C.A., Dubrule, O.R.F., and Van Riessen, D.D., 1988, Aspects of reservoir geology and production behavior of Sirikit Oil Field, Thailand: an integrated study using well and 3-D seismic data: AAPG Bulletin, v. 72, p. 1254-1269.

_______, 1989, Reservoir geology of the Sirikit oilfield, Thailand: lacustrine deltaic sedimentation in a Tertiary intermontane basin: in Whateley, M.K.G., and Pickering, K.T., eds, Deltas, Sites and Traps for Fossil Fuels: Geological Society (of London) Special Publication No. 41, p. 223-237.

Galloway, W.E. and Hobday, D.K., 1996, Terrigenous Clastic Depositional Systems: Applications to Fossil Fuel and Groundwater Resources: Springer-Verlag, New York, 487 p.

Liberty, L., 1998, Seismic reflection imaging of a geothermal aquifer in an urban setting: Geophysics, v. 63, p. 1285-1295.

Link, M.H., and Osborne, R.H., 1978, Lacustrine facies of the Pliocene Ridge Basin Group, California: *in* Matter, A., and Tucker, M.E., Modern and Ancient Lake Sediments, Special Publication No. 2, International Association of Sedimentologists, p. 169-187.

Malde, H.E., 1987, A guide to the Quaternary geology and physiographic history of the Snake River Birds of Prey area, Idaho: Northwest Geology, v. 16, p. 23-46.

_______, 1972, Stratigraphy of the Glenns Ferry Formation from Hammett to Hagerman, Idaho: U. S. Geological Survey Bulletin 1331-D, 19 p.

_______, 1991, Quaternary geology and structural history of the Snake River Plain, Idaho and Oregon, *in* R. B. Morrison, editor, Quaternary nonglacial geology, conterminous U.S., Geology of North America, Geological Society of America, v. K-2, p. 251-280.

Malde, H. E., and H. A. Powers, 1962, Upper Cenozoic stratigraphy of the western Snake River Plain, Idaho: Geological Society of America Bulletin, v. 73, p. 1197-1220.

Rider, M., 1996, The Geological Interpretation of Well Logs (2nd edition):Caithness, UK, Whittles Publishing, 280 p.

Squires, E., Wood, S.H., and Osiensky, J.L., 1992, Hydrogeologic framework of the Boise aquifer system, Ada County, Idaho: Research Technical Completion Report 14-08-0001-0G1559-06, Idaho Water Resources Research Institute, University of Idaho, Moscow, 114 p.

Whitehead, R. L., 1992, Geohydrologic framework of the Snake River Plain, Idaho and eastern Oregon: U.S. Geological Survey Professional Paper 1408-B, 32 p.

Wood, S.H., 1994, Seismic expression and geological significance of a lacustrine delta in Neogene deposits of the western Snake River Plain, Idaho: AAPG Bulletin, v. 78, p. 102-121.

Wood, S.H., Liberty, L., and Squires, E., 2000, Geophysical signatures of lacustrine facies: Experiences from hydrogeologic studies of temperate Neogene Lake Idaho sediments and aquifers, U.S.A., (abs.) 31st International Geological Congress, Rio de Janeiro, Brazil (Abstracts Volume CD)

Wood, S.H., and Clemens, D.M., *in press*, 2000, Geologic and Tectonic history of the western Snake River Plain, Idaho and Oregon: *in* Bill Bonnichsen, M. McCurry, and C. White(editors), Tectonic and Magmatic History of the Snake River Plain Volcanic Province, Idaho Geological Survey Bulletin.

Repenning, C. A., T. R. Weasma, and G. R. Scott, 1994, The early Pleistocene (latest Blancan-earliest Irvingtonian) Froman Ferry fauna and history of the Glenns Ferry Formation, southwestern Idaho: U. S. Geological Survey Bulletin 2105, 86 p.

Contribution to the Treasure Valley Hydrologic Project - 1997

STRUCTURE CONTOUR MAP OF TOP OF THE MUDSTONE FACIES, Western Snake River Plain, Idaho

By
Spencer H. Wood
Geosciences Department
Boise State University, Boise, Idaho 83725
Tel. (208) 385-3629, email: swood@trex.idbsu.edu

August, 1997

Boise Map Sheet, Idaho: Scale = 1:100,000
Elevations in feet with respect to sea level
Contour Interval = 100 feet (contours dashed where uncertain for lack of control wells)

Doty '79 ○ = deep water well used for control, giving name of owner listed on driller's log, and date of completion.

Oroco Oil and Gas Cleveland #1 (T.D. 4038 ft) = deep petroleum or geothermal exploration well, giving operator, lease name, and total depth

= edge of sedimentary basin

= approximate trace where contact is at the surface (i.e. the contact between the fluvial-deltaic section and the underlying mudstone skylights in the Snake River Canyon)

Introduction

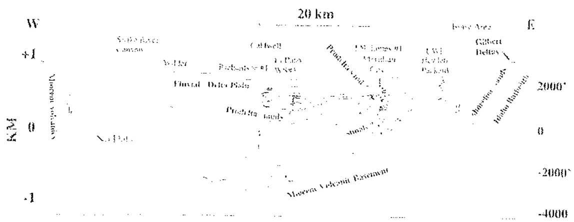
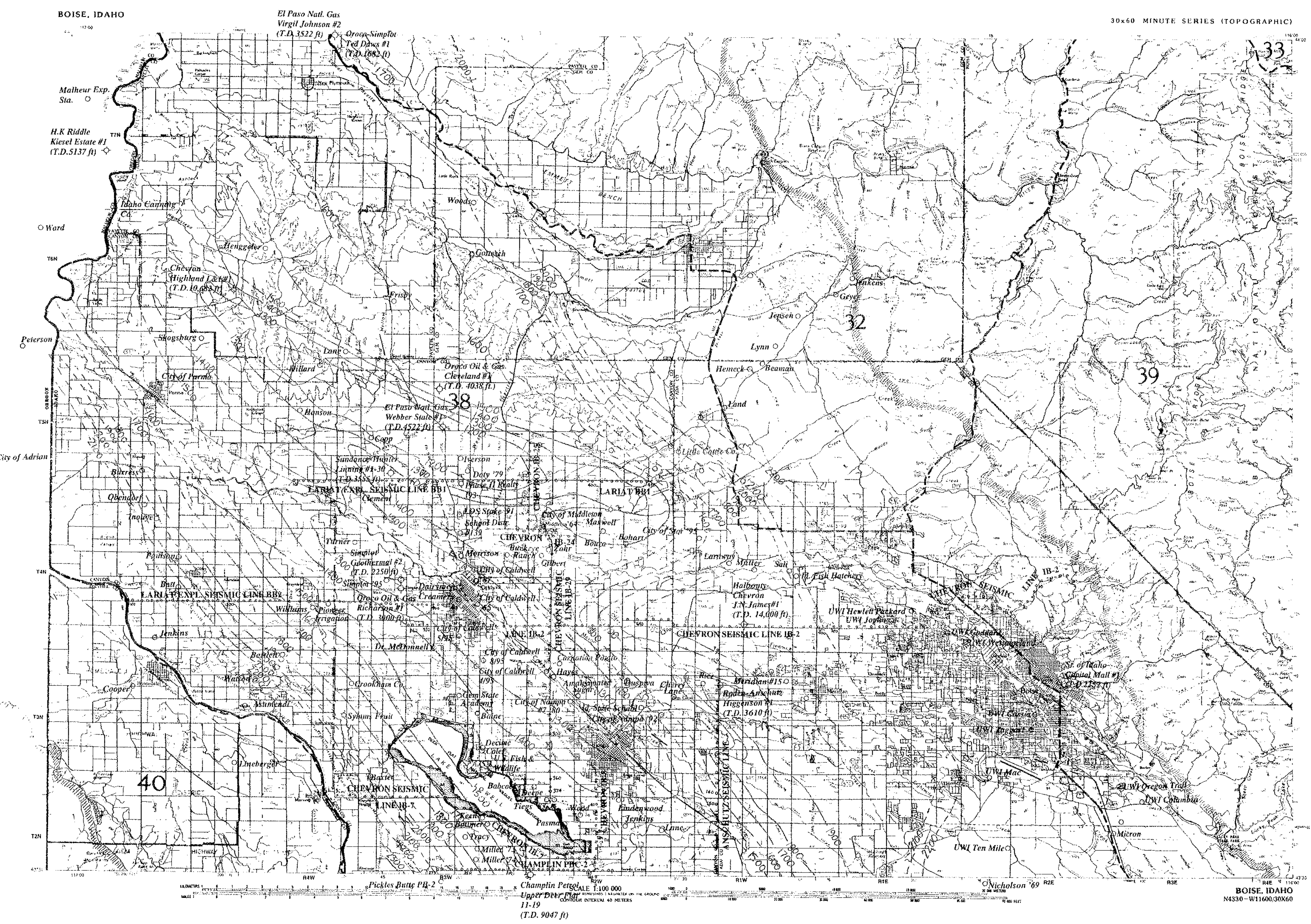


Figure 1.

Data used for structure contours



Seismic Expression and Geological Significance of a Lacustrine Delta in Neogene Deposits of the Western Snake River Plain, Idaho[1]

Spencer H. Wood[2]

ABSTRACT

High-resolution seismic reflection profiles and well data from the western Snake River plain basin are used to identify a buried lacustrine delta system within Neogene Idaho Group sediments near Caldwell, Idaho. The delta system is detected, 305 m (1000 ft) deep, near the center of the basin by progradational clinoform reflections having dips of 2–5°, a slope typical of prodelta surfaces of modern lacustrine delta systems. The prodelta slope relief, corrected for compaction, indicates the delta system prograded northwestward into a lake basin 255 m (837 ft) deep. Resistivity logs in the prodelta mud and clay facies are characterized by gradual upward increase in resistivity and grain size over a thickness of about 100 m (300 ft). Lithology of the prodelta is mostly calcareous claystone, with several layers of fine sand, some of which fine upward, indicating a density-flow mechanism of transport and deposition. Delta-plain and front sediments are mostly very fine-grained, well-sorted sand separated by thin mud layers. These sediments produce several to five cycles of horizontal, high-amplitude reflections with a toplap relationship to prodelta clinoforms. The sands have an abrupt lower contact with prodelta muds and have high resistivity on logs. In this study, permeable lacustrine sands within a predominantly mud and clay section are located by using high-resolution seismic reflection data.

Identification of a delta system in the Idaho Group provides insight into the history of Pliocene "Lake Idaho." The present depth of the delta/prodelta facies contact of 305 m (1000 ft) is 445 to 575 m (1460–1900 ft) below the lake deposits on the margins. Estimated subsidence from compaction is 220 m (656 ft), and the remaining 225 to 325 m (740–1066 ft) is attributed to tectonic downwarping and faulting.

The original lake area had been reduced to one third of the original 13,000 km^2 (5000 mi^2) by the time the delta front prograded to the Caldwell area. The original lake area may have been sufficient to evaporate most of the inflow, and the lake may have only occasionally spilled into other basins. Diminished area for evaporation later in the history of the lake, combined with reduced evaporation accompanying onset of the ice ages, may have caused the lake to rise and overtop a basin sill about 2 Ma, and subsequently deepen Hells Canyon.

INTRODUCTION

Lacustrine deposits in rift basin environments have become major petroleum exploration targets in many parts of the world (Katz, 1990). Likewise, important groundwater resources are exploration targets in lacustrine sediments. Exploration methods are needed to identify facies of permeable strata within thick sections of impermeable mud rocks characteristic of lacustrine deposits because these permeable strata may be hydrocarbon reservoirs or freshwater aquifers. Recognition of clastic depositional systems and permeable facies on seismic reflection sections in the marine environment has been advanced by the work of Mitchum et al. (1977), Sangree and Widmier (1977), and Berg (1982). This study illustrates the seismic expression and well-log character of a fine-sand delta facies within lacustrine clay and mud deposits of the Neogene western Snake River plain, and compares its geometry with lacustrine deltas studied elsewhere in the world.

With the exception of a study of the Paleocene Fort Union Formation of Wyoming by Liro and Pardus (1990) and Quaternary Lake Biwa, Japan (Ikawa,

[1]Manuscript received, April 8, 1991; revised manuscript received, May 13, 1993; final acceptance, August 11, 1993.

[2]Department of Geosciences, Boise State University, Boise, Idaho 83725

This study was made possible through the donation of seismic data to Boise State University by David Hawk. His continued help and encouragement in this study is greatly appreciated. J. K. Applegate provided information on the original acquisition and processing of the data. The author thanks Ed Squires of Boise State University, employees of J. R. Simplot Co., Vaughn Thomas Drilling Co., and COLOG, Inc. for their help in acquiring drilling, logging, and production data, and John Armentrout for a most helpful review of the manuscript. This study was partly supported by Idaho State Board of Education Grant 91-099, Center for Geophysical Investigation of the Shallow Subsurface Contribution 0024.

Table 1. Comparison of Settings of Lacustrine Deltas*

	[Delta Area] Lake Area (km^2)	River Drainage Area (km^2)	Average Annual Discharge (10^9 m^3/yr)	Annual Suspended Load (10^6 m^3/yr)	Annual Bedload (10^6 m^3/yr)	Basin Origin	Latitude
Lake Constance Rhine River (Bodensee) Switzerland	[6+] 600	6122	6.95	2.57	0.04	Würm-age glaciated basin	temperate
Pyramid Lake Truckee River Nevada, USA	[20] 428	4785	0.72 0.26**	0.1	?	tectonic, lake level lowered, eroded and rebuilt delta	temperate 40° N
Lake Laitaure Sweden	[10] 9	684	6.3	0.24	0.04	glaciated	arctic 67°N
Lake Maracaibo Catatumbo River Venezuela	[1200] 12,900	15,000	29	4.8		broad tectonic subsidence	tropical 10°N
Neogene Lake Idaho (Snake River) Idaho, USA	13,000†	180,000†	16††	11††	0.4-1.1††	tectonic, subsidence & faulting	temperate 44°N

*References: Lake Constance = Müller, 1966; Pyramid Lake = Born, 1972; Lake Laitaure = Axelsson, 1967; Lake Maracaibo = Hyne et al., 1979.
**Discharge after diversion of the Truckee River for irrigation. Much of the modern delta was built after diversion in the early 20th Century.
†Original lake area taken from distribution of Idaho Group sediments. Position of the buried delta discussed in this paper suggests the floodplain occupied about two thirds of the basin, and the lake about one third or 4300 km^2. Discharge of the modern Snake River at Weiser, Idaho, 70-year average (from Kjelstrom, 1986).
††Sediment loads measured at Lewiston, Idaho are the only available measurements (from Seitz, 1976). Bedload (>0.2mm caught in bed-load sampler) ranges from 5 to 10% of the suspended load. This reach of the river is below Hells Canyon, and the river is impounded by a series of hydroelectric dams about 100 km above here; therefore, these measurements may not be representative of sediment load in Neogene time.

1991), little has been published that relates seismic stratigraphy of lacustrine deposits to their subsurface geology. Most of the literature is concerned with outcrop-scale features (Fouch and Dean, 1982). Comparison to studies of the seismic expression of marine delta systems, particularly the paper by Berg (1982) was most useful for interpretation in the present study. The few studies of modern lacustrine deltas cited in Table 1 are helpful in understanding facies distributions in these environments, but research that incorporates shallow subsurface studies to understand the facies geometry of modern lacustrine environments is clearly needed.

REGIONAL SETTING

The western Snake River plain basin is a normal-fault–bounded intermontane basin between the northern Rocky Mountains province and the northern Basin and Range extensional province. The plain is underlain by a sequence of up to 2 km (6000 ft) of lacustrine and fluvial sediment of the Idaho Group (Wood and Anderson, 1981). The history of deposition and facies distributions within this section of strata have not been previously studied. High-resolution seismic reflection profiles combined with geophysical logs and well-site study of cuttings provide an opportunity for interpretation of the Neogene subsurface geology near Caldwell, Idaho (Figures 1, 2).

The western plain basin is a northwest-trending graben-form tectonic basin that evolved after voluminous rhyolite volcanism (16–9 Ma) on the south side of the present basin, and after voluminous Columbia River basalt volcanism (17–14 Ma) on the north side (Leeman, 1989; Wood, 1989a). Tectonic setting of the Columbia River flood basalts is reviewed by Hooper and Conrey (1989). Tectonic setting of the rhyolite systems has the appearance of a northeast-migrating continental hot spot (Armstrong et al., 1975; Rodgers et al., 1990; Pierce and Morgan, 1992). Brott et al. (1981) hypothesized that cooling of the lithosphere

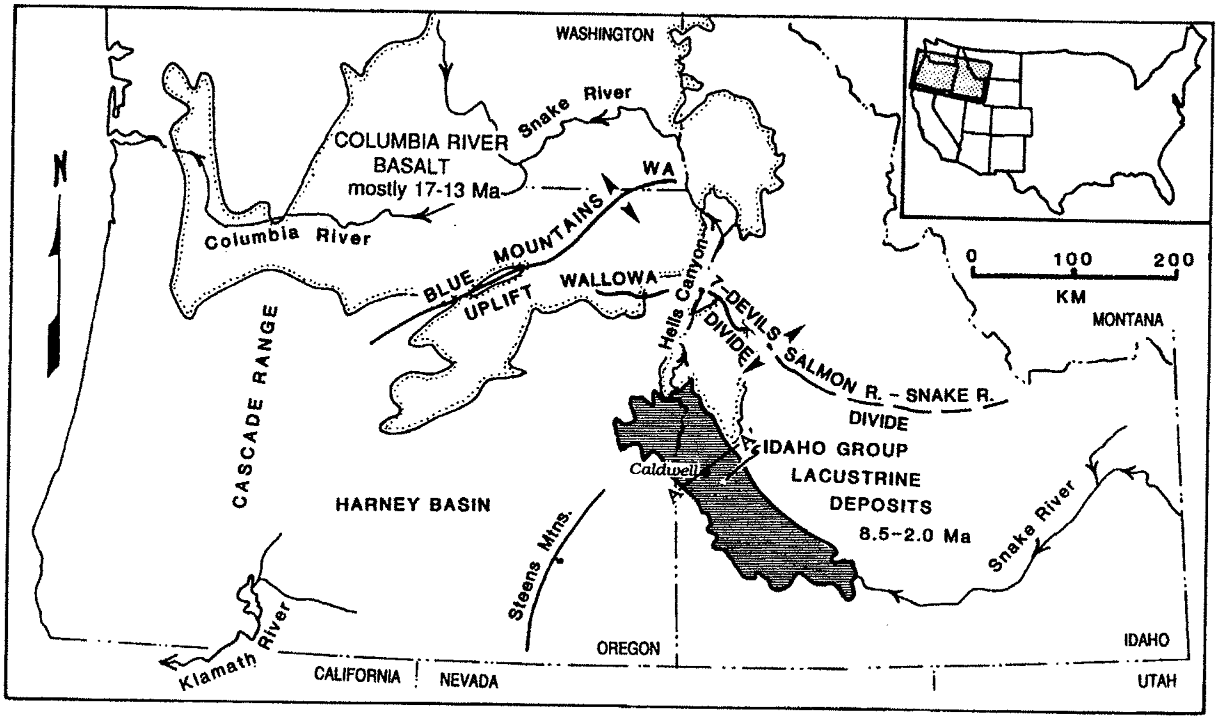


Figure 1—Index map showing distribution of Neogene Idaho Group lacustrine deposits in the western Snake River plain, location of physiographic features and drainages influencing the Neogene lakes that occupied the western plain region, and distribution of Columbia River basalt, shown by dotted line. Accumulation of Columbia River basalt in the early Neogene and subsequent broad uplift of the Blue Mountains and Wallowa–Seven Devils Mountains formed a barrier to northern drainage of the Snake River system. Outlet of the lake system may have been southwest through Harney Basin. Other workers have suggested connection to the Klamath River system of northern California. AA' is location of cross section shown in Figure 3.

after passage of the hot spot produced the downwarp structure of the eastern Snake River plain. Although the eastern and western parts of the Snake River plain form a continuous physiographic lowland, the eastern plain is a southwest-trending downwarp without major fault boundaries, whereas the northwest-trending western plain is clearly a graben form with normal fault boundaries. The fault boundaries parallel other extensional features that formed perpendicularly to the hot spot track.

Middle Miocene volcanic rocks form an acoustic basement about 2 km beneath the deepest part of the western plain basin (Figure 3). Sporadic olivine-tholeiite basalt volcanism has continued through the Neogene and Quaternary from vents of a north 70° west trending volcanic zone that traverses the western plain, and from vents on the margins (Wood, 1989b). Minor amounts of these basalts are intercalated with the sedimentary fill of the western plain. The eastern plain is largely covered with Pliocene and Quaternary basalts, which erupted from N40°W trending volcanic rift zones (Kuntz et al., 1992), with accumulations that are locally 1000 m (3280 ft) thick.

Because of its large size and graben structure the western plain basin has been called a tectonic continental rift. The western plain basin meets the definition of a tectonic rift in that it is a "long, narrow continental trough that is bounded by normal faults, a graben of regional extent" (Bates and Jackson, 1980). The structural geometry is similar to other continental rift basins described by Bosworth (1985). The buried horst block in the center (Figure 3) is interpreted to be an interbasinal ridge or an extension transfer zone (using terminologies of Rosendahl et al., 1986; Morley et al., 1990) between half-graben structures formed during early basin evolution. Areal extent of the horst is indicated by a +20 mGal gravity anomaly outlined by the dotted line in Figure 3. Structural dip of the margin fault blocks and infilling sediment toward the center of the basin suggest the lithosphere has more recently undergone extension and necking that create negative buoyancy of the basin, and broad subsidence of the center of the basin in a manner suggested by Braun and Beaumont (1989). Some components of the dip of the sediments and subsidence of the center of the basin, however, are because of com-

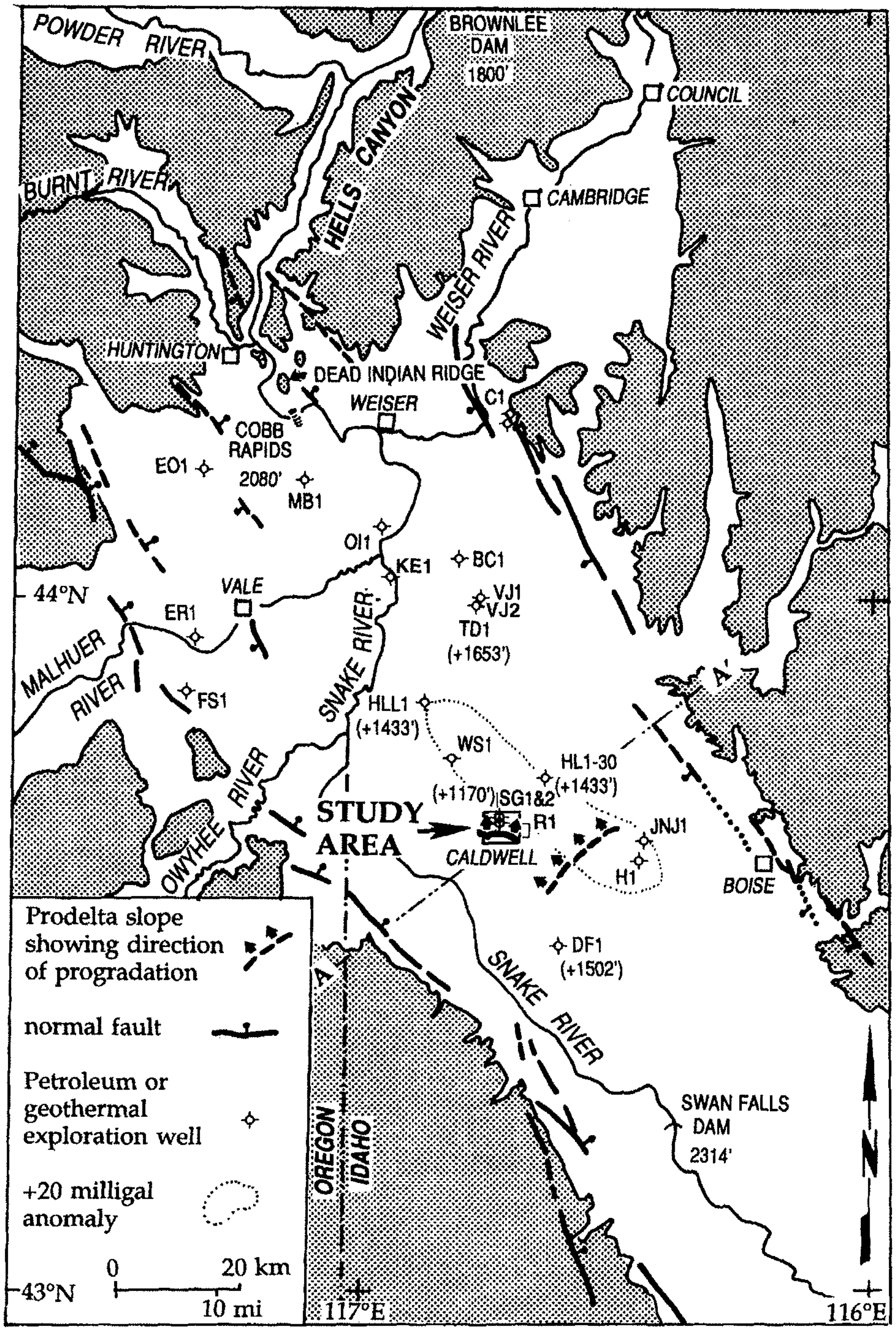


Figure 2—Map of western Snake River plain showing location of study area of Figure 5. Map shows position and progradation direction of the lacustrine delta system in the Pliocene interpreted from this study and from a previous study by Wood and Anderson (1981). Numbers in parentheses are the present elevations (in feet above sea level) of the prodelta/delta-front transition identified on well logs. Names of deep wells examined for this relationship are listed in Table 2. AA' is the line of section shown in Figure 3. Dashed line enclosing northwest-southeast–trending oval area north of Caldwell is a +20 mGal gravity-anomaly expression of the buried horst shown on the cross section of Figure 3.

paction of the thick accumulation of fine sediment in the center of the basin.

The basin contained large freshwater lakes and river systems from about 8.5 to 2.0 Ma and filled with predominantly fine siliciclastic sediment and minor intercalated pyroclastic and flow basalt layers. Numerous earlier papers summarized in Wheeler and Cook (1954) and Jenks and Bonnichsen (1989) alluded to a large Pliocene lake in the western plain, commonly referred to as Lake Idaho, but little is

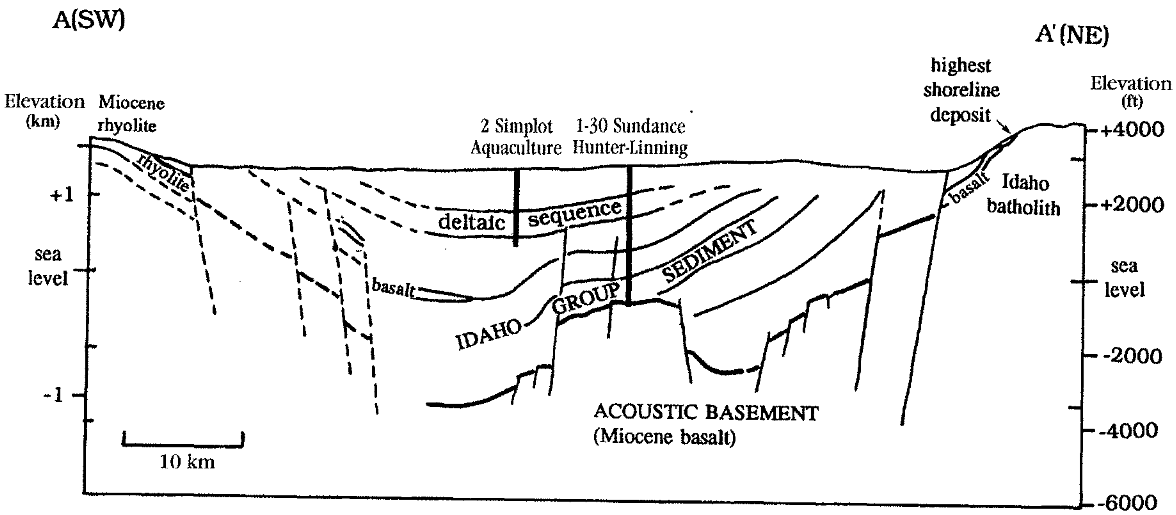


Figure 3—Structural features of the western plain in an exaggerated vertical-scale cross section showing position of the deltaic sequence discussed in this paper. Location of section is shown in Figure 2. Cross section shows that the plain has continued to subside since the progradation of delta sequence. Basis for section is seismic reflection data reported by Wood and Anderson (1981). Dashed contacts where seismic data are not available are inferred from geologic mapping.

known of the history of lake systems. Discontinuous lacustrine shoreline deposits typically occur up to about 975 m (3200 ft). Higher deposits such as 1128 m (3700 ft) in the foothills on the north side of the basin (Gallegos et al., 1987) and 1036 m (3400 ft) on the south side of the basin (Smith et al., 1982) may have been elevated by faulting. Although the oldest lake deposits in the graben are clearly downfaulted one or more kilometers from their remnants in the margin foothills, evidence for the highest lake levels remarkably occurs on both sides of the basin at similar elevations (1000 ± 130 m). These elevations suggest relatively little differential vertical movement of the basin margins since the last high-lake level. Apparently, the graben subsided by downwarping and minor normal faulting, with minor flank uplift during the late Neogene and Quaternary.

At its maximum extent, the sedimentary basin covered about 13,000 km^2, judging from distribution of thick, fine-grained sediments and shoreline deposits (Figure 1). The drainage basin delivering water to the lake may have been similar in size to the present upper Snake River catchment (above Weiser, Idaho), which is 175,000 km^2. Some have speculated that in the Neogene the lake basin drained through southeastern Oregon to the Klamath River system in northern California (Figure 1) (Wheeler and Cook, 1954; Taylor, 1960, 1985; Smith, 1975; Smith et al., 1982). Physical evidence for such an outlet through the Great Basin has not been found; the evidence is mainly paleozoological affinities of fossil mollusks and fish in lacustrine deposits of the western plain (Smith, 1981). Wheeler and Cook (1954) proposed that the downcutting of ancestral Hells Canyon through the Blue Mountains and the Wallowa–Seven Devils structural arches caused capture of the Snake River basin by the Columbia River drainage system. Surficial deposits of fluvial gravel deposited over the lake sediments mark the conversion of the basin from an aggrading lacustrine and fluvial basin to an eroding stream system (Malde, 1991; Othberg, 1992; Othberg and Sanford, 1992). This enormous diversion of Snake River basin waters caused further downcutting of Hells Canyon, now the deepest gorge in North America. The Snake River is now incising its lacustrine deposits in the western plain basin.

STRATIGRAPHY OF BASIN FILL

Stratigraphy of the basin fill has been developed from the study of dissected deposits, mostly on the southwest margin of the plain. A diagram of lithostratigraphic units published by Malde and Powers (1962) is shown in Figure 4. Much of the mud, silt, and clay in the upper 1000 m of the basin fill could be called mudrock, mudstone, siltstone, and claystone, because, in a geological sense, they are mostly soft friable rocks. In a geotechnical sense these materials fall to both sides of the definition of rock [unconfined compressive strengths above 1500 kN/m^2 (200 psi); in other words, materials that can-

AGE (Ma)	SERIES		GROUPS AND FORMATIONS (typical lithology)	GEOCHRONOLOGY	THICKNESS (ft)
	PLEIS-TOCENE		SNAKE RIVER GROUP & BRUNEAU FORMATION (basalt lava, river-terrace and lava-dammed-lacustrine deposits)	0.11 to 2.1 Ma (K-Ar)4, (Ar-Ar)5	0-1000
1.8 -	PLIOCENE	NEOGENE IDAHO GROUP	TEN MILE GRAVEL (fluvial gravel)		0-70
			GLENNS FERRY FORMATION (lacustrine & fluvial (sediment, minor basalt)	1.8 & 1.9 Ma (f) 2.4 Ma (f) 3.5 Ma (K-Ar)1 3.75 Ma (fz)	3000+
5.0 -			CHALK HILLS FORMATION (lacustrine & fluvial sediment, silicic volcanic ash, minor basalt)	5.0 Ma (f) 6.5 Ma (f) 8.4 Ma (K-Ar)1	300+
	UPPER MIOCENE		POISON CREEK FORMATION (tuffaceous sediment, arkosic sand, minor basalt)		400+
			IDAVADA VOLCANIC GROUP (rhyolite flows and tuffs)	9-12 Ma (K-Ar)1&2	0-3000+
			COLUMBIA RIVER BASALT GROUP (NORTH SIDE OF PLAIN) (basalt lavas and minor sediment)	14-17.5 Ma (K-Ar)3	0-9000+
24.0 -			SUCKER CREEK FORMATION & SILVER CITY RHYOLITE (SOUTH SIDE OF PLAIN) (tuffaceous lacustrine and fluvial sediment, rhyolite and basalt)	15-17 Ma (K-Ar)2	0-2300+

Figure 4—Neogene stratigraphy of the western Snake River plain near Caldwell, Idaho, after Malde and Powers (1962), Wood and Anderson (1981), Middleton et al. (1985), and Kimmel (1982). f = fission track (glass) from Kimmel (1982); fz = fission track (zircons in glass) from Izett (1981); (K-Ar)1 = potassium-argon ages from Armstrong et al. (1975; 1980), (K-Ar)2 = ages from Ekren et al. (1981, 1984); (K-Ar)3 = ages from Fitzgerald (1982); (K-Ar)4 from Malde, 1987; (Ar-Ar)5 from Othberg and Stanford (1992).

not be crushed between one's fingers (e.g., Sowers, 1979, p. 3)]. The rocks or sediments generally yield good cuttings from rotary wells, but they will be referred to only by their sediment names unless indurated by cementation or calcite recrystallization to the extent that they are no longer friable. The Miocene Poison Creek Formation is defined as a clastic and volcaniclastic unit lying unconformably on the Idavada rhyolites. The formation is mostly silicic volcanic ash, fine-grained tuffaceous material in massive beds with some beds of locally cemented arkosic sand and gravel, and some basaltic pyroclastic beds. The Chalk Hills Formation is a silt and sand unit with numerous thin layers of fine silicic ash. The Pliocene-aged Glenns Ferry Formation is the thickest and most widespread of the exposed deposits. At its type locality along a 25-mi reach of the Snake River near Glenns Ferry, Malde (1972) described 660 m (2000 ft) of section of flood-plain and lacustrine facies. The fluviatile facies is described as thick, evenly layered beds of drab, very pale brownish-gray sand and some silt. Among the layers of gray and brown sand and silt are some beds of olive silt, dark-olive clay, and paper shale, which are characteristic of the flood-plain facies. The lacustrine facies is described as dominantly massive layers of tan silt that form drab monotonous outcrops having little lithologic variety but marked by faint, diffuse, gray bands arranged parallel to bedding and spaced several feet apart. Neither vertical nor lateral facies distributions of the basin fill have ever been studied on a basin-wide scale. In the Caldwell area of the present study, similar muds, silts, and sands occur to a depth of about 365 m (1200 ft), below which the section is dominantly calcareous claystone with minor sand layers to a depth of at least 670 m (2200 ft). Surprisingly, the section described by Malde (1972) contains only minor claystone and carbonate, whereas in the subsurface to the west, calcareous claystone is the dominant lithology. In the present study, seismic sections combined with examination of cuttings and geophysical logs from an area near Caldwell (Figures 5, 6, 7) give a clearer picture of depositional features than the study of outcrops has

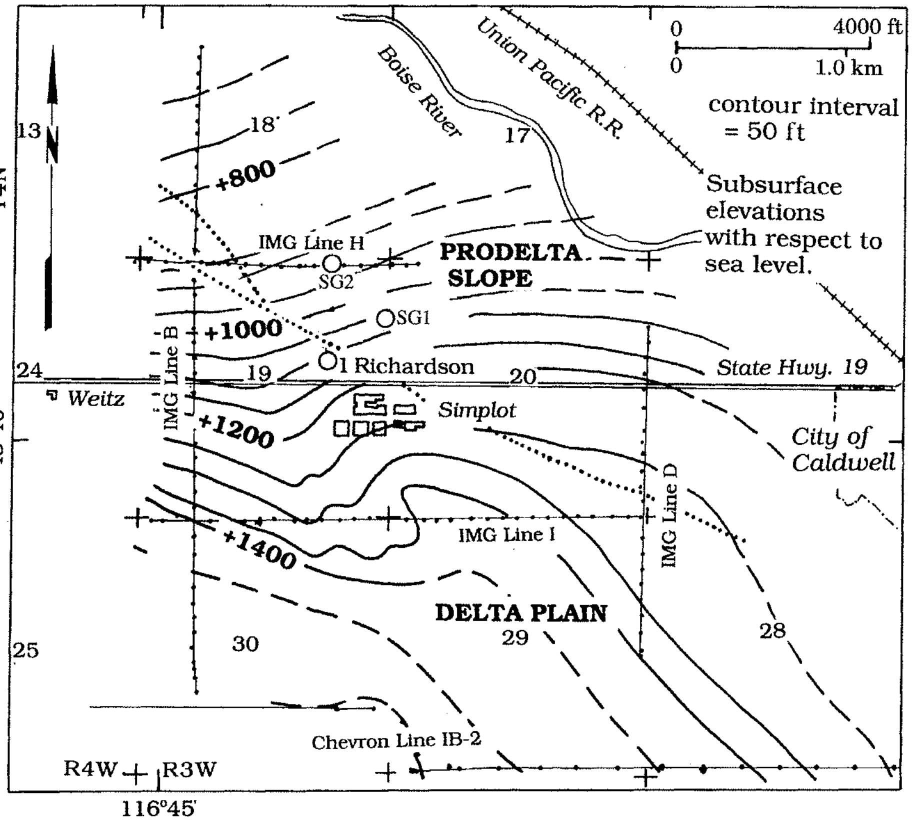


Figure 5—Location map of Caldwell study area showing structure contours of a reflecting horizon. Mapped reflector is shown on the interpreted section at about 0.5 s on IMG Line H (Figure 6b) as a heavy line within the prodelta clay facies. Configuration of the reflector is considered a depositional surface (i.e., not tectonically deformed) passing northward from the delta plain to the prodelta slope environment. Dotted line with a northwest-southeast orientation is the trace of the fault on the contoured surface. Displacement appears to be less than 17 m (50 ft) on the contoured surface. SG1 and SG2 are Simplot Aquaculture 1 and 2 Simplot Geothermal wells. IMG Lines I and D, and Chevron Line IB-2 used to construct this map, are not published in this paper. Base map is the Caldwell, Idaho quadrangle.

given. In particular, a large northwestward-prograding delta system is identified in the subsurface, about 305 m (1000 ft) deep beneath the middle of the plain near Caldwell, Idaho.

PETROLEUM AND GEOTHERMAL EXPLORATION RESULTS

A number of petroleum and geothermal exploration wells (Table 2) have drilled the Neogene lacustrine deposits and the deeper Miocene volcanic and sedimentary rock beneath the western Snake River plain (Figure 2). Earlier results are reviewed by Kirkham (1935) and Newton and Corcoran (1963), and geothermal aspects of recent drilling have been reviewed by Wood and Anderson (1981) and Blackwell (1989). Many early wells had hydrocarbon gas shows in the sandy strata. Several uncontrolled blowouts and gushers either sanded up or depleted gas in the reservoir within a few days (Kirkham, 1935). Controlled formation tests in several wells drilled since 1950 produced low flows (50–350 mcf/day), but these flows also dwindled within a few days or less (Newton and Corcoran, 1963). Sanyal et al. (1980) and GeothermEx, Inc. (1980) report significant gas on the mud log of the Ore-Ida Foods 1 Ore-Ida well from the section above 2195 m (7200 ft). Possible high-gas saturation from the interval 2106 –2137 m (6910–7010 ft) is interpreted from geophysical logs by Sanyal et al. (1980). No production has been established in the basin, and as Deacon and Benson (1971) concluded, test results thus far indicate low-volume reservoirs of gas.

Dry cuttings samples from the deepest wildcat well (Halbouty-Chevron 1 J. N. James) were studied by Geochem Laboratories, Inc. (unpublished data) for prospective source rocks. Ten sample intervals of fine sediment having the highest total organic carbon content ranging from 0.43 to 1.95% were examined. Organic matter type in the kerogen concentrate is predominantly woody material with secondary amounts of herbaceous spore pollen and tertiary inertinite, indicating the sediments are potential source rocks for gas but not for liquid hydrocarbons. Samples from 300 to 640 m (1000–2100 ft) (Pliocene Idaho Group sediments) are considered thermally immature based on the greenish light-yellow to orange-brown coloration of recognizable plant cuticle contained in the kerogen concentrate. Samples from the interval 1170–2650 m (3840–8700 ft) (sediments from below the acoustic basement interface shown in Figure 3) are interpreted as thermally mature based on orange-brown to dark-brown coloration of recognizable plant cuticle kerogen (stage 2+ to 3 alteration). Only two wells (Halbouty-Chevron 1 J. N. James and Champlin Petroleum 1 Deer Flat) have penetrated this deeper section. Questionable solid hydrocarbon material was found in the Halbouty-Chevron 1 J. N. James well, and successful drill-stem testing could not be done on the prospective zones in the Champlin Petroleum 1 Deer Flat well.

Geothermal gradients in the western plain basin are relatively high, ranging from 30 to 40°C/km (16.5–22°F/1000 ft) (Wood and Anderson, 1981; Blackwell, 1989). Several exploratory wells were drilled for geothermal water in the 1980s (Table 2), but significant water production has been 74°C (165°F) or less. The present study is based on data from a geothermal aquaculture project.

In the study area near Caldwell, Idaho (Figure 5), Oroco Oil and Gas Co. drilled the 1 Richardson wildcat well in 1955 to a depth of 926 m (3036 ft). A production test of the lower 50-ft interval initially flowed methane at 50 mcf/day, but rapidly dwindled to a low flow (Newton and Corcoran, 1963). The well currently produces artesian geothermal 40°C (105°F) water that flows at a rate of 3.1 L/s (50 gal/min), and has a surface shut-in pressure of 0.17 MPa (24 psi). The well is presently obstructed below 769 m (2524 ft). A temperature log to that depth shows a temperature of 64°C (148°F), and flow of geothermal water entering the well through several places where the casing was broken by explosives.

To support a geothermal aquaculture operation in the same area, in 1988 the J. R. Simplot Co. completed a 342-m (1120 ft) well (G1 on Figure 5) into a fine sand aquifer, with an artesian flow of 39°C (101°F) water at a rate of 25 L/s (400 gal/min) and a shut-in well-head pressure of 0.22 MPa (32.5 psi). In 1989 a second well (G2 on Figure 5) was drilled to 686 m (2250 ft) to explore the deeper and hotter sands indicated on the original logs of the Oroco 1 Richardson well, but drilling difficulties prevented completion of the well. Examination of rotary cuttings from these wells now allows lithologies to be associated with the reflections on the seismic sections and the character of the geophysical logs.

SEISMIC STRATIGRAPHY

The emphasis of this paper is on the interpretation of a clinoform structure on seismic sections, illustrated in Figure 6, that is now buried 305 m (1000 ft) beneath the plain. The upper 0.7 s (approximately 760 m or 2500 ft) of high-resolution seismic records show three distinct seismic facies units. These units will be described in terms of seismic facies shown best in Figure 6a and then calibrated to geological data from boreholes.

The uppermost unit (unit 1) contains four high-amplitude, continuous reflections, each separated by low-amplitude or reflection-free intervals about 0.04 s (43 m or 141 ft) thick. The basal reflections of this unit are inclined northward from 0.24 to 0.31 s (about 250 to 300 m or 800 to 1000 ft). Strong reflections are continuous over 300 to 500 m (985–1640 ft), and their terminations generally overlap by 150 m (492 ft) with an adjacent, continuous, strong-amplitude cycle. Some strong reflectors show subtle low-angle onlap to the south with underlying reflections.

The next unit (unit 2) is typically four or five adjacent cycles of strong-amplitude reflections, which are continuous for about 1 km. The upper reflections overlap, climb in elevation to the south, and generally terminate to the north and east into the basal reflection-free interval of unit 1. The lower reflections of unit 2 are the upper topset segment of a complex sigmoid-oblique progradational reflection configuration (using terminology from Mitchum et al., 1977, p. 125). This unit has the highest amplitude reflections on the profile, and is about 0.05 s (50 m or 165 ft) thick.

Below unit 2 are the clinoform reflectors of the progradational reflection configuration with a 5° depositional dip to the northwest. This interval is labeled unit 3 on Figure 6a. These clinoforms have mostly oblique toplap relationships with the topset reflectors of unit 2, although some may be crudely sigmoid. The upper clinoforms segments are less than 400 m (1312 ft) long. Amplitude of the clinoform reflections in this unit (unit 3) varies from strong, where the unit is thin in the south, to very weak, where the unit thickens to the north and west. The lower segments of the clinoforms as they flatten are continuous for about 1.2 km (1 mi), and some have relatively high amplitude. The lower segments have a downlap relationship to underlying reflectors of lesser dip.

Below unit 3 are reflectors with low dips to the north, which are probably the distal ends of the progradational configurations. The reflections appear to have high amplitude and ringing of lower frequency wavelets, partly as a result of the natural loss of high-frequency content and partly as a result of the data processing parameters that emphasize resolution in the upper section.

LITHOLOGY AND LOG EXPRESSION OF SEISMIC FACIES UNITS

The Simplot Aquaculture 2 Simplot Geothermal well (G2 on Figure 5) was drilled through the stratigraphic interval of seismic facies units 1, 2, and 3.

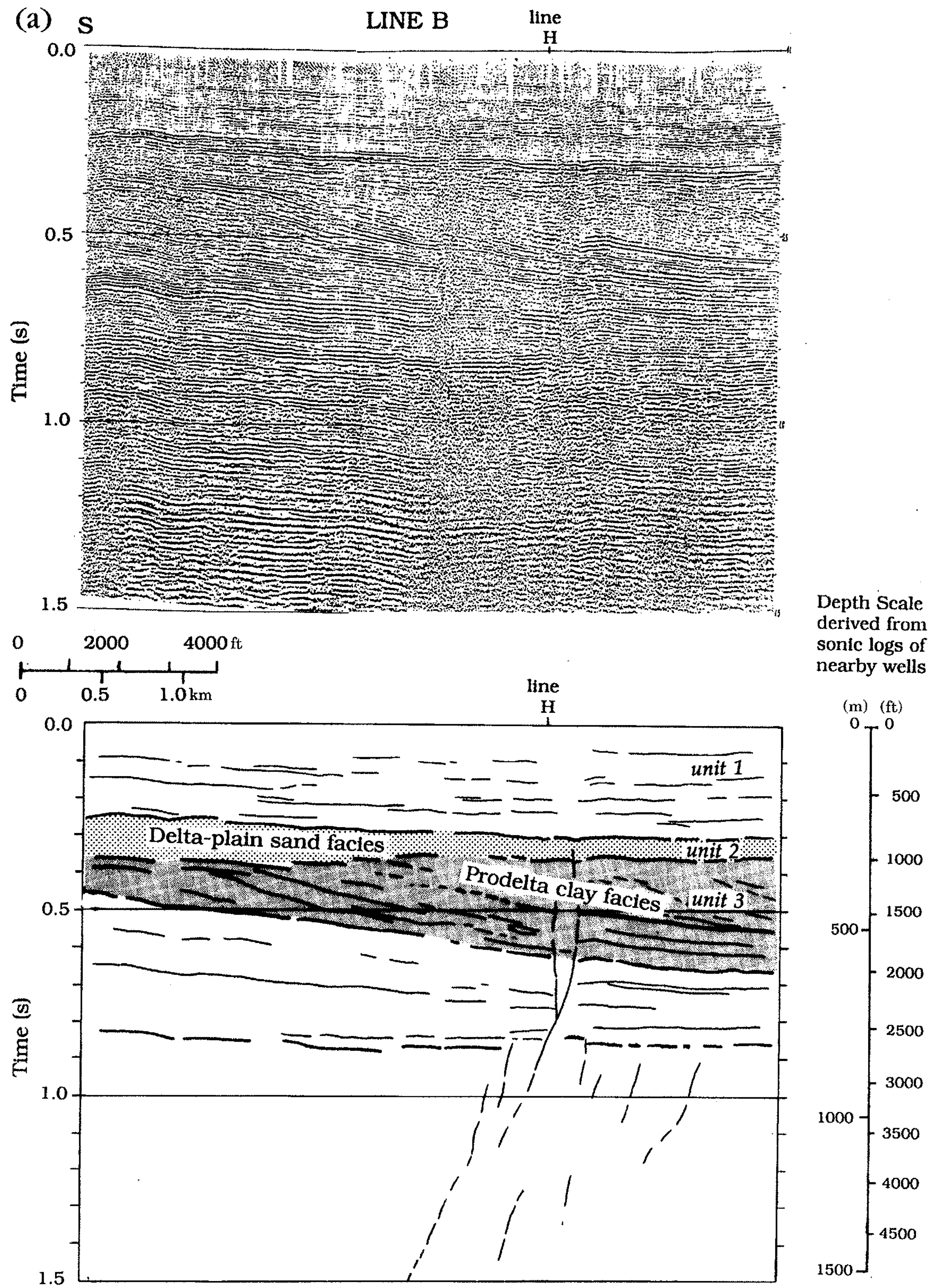


Figure 6—High-resolution seismic sections of IMG lines B and H (unmigrated). Location of lines shown in Figure 5. Datum is 762 m (2500 ft) above sea level. Elevation of ground level along lines is between 707 and 728 m (2320 and 2390 ft).

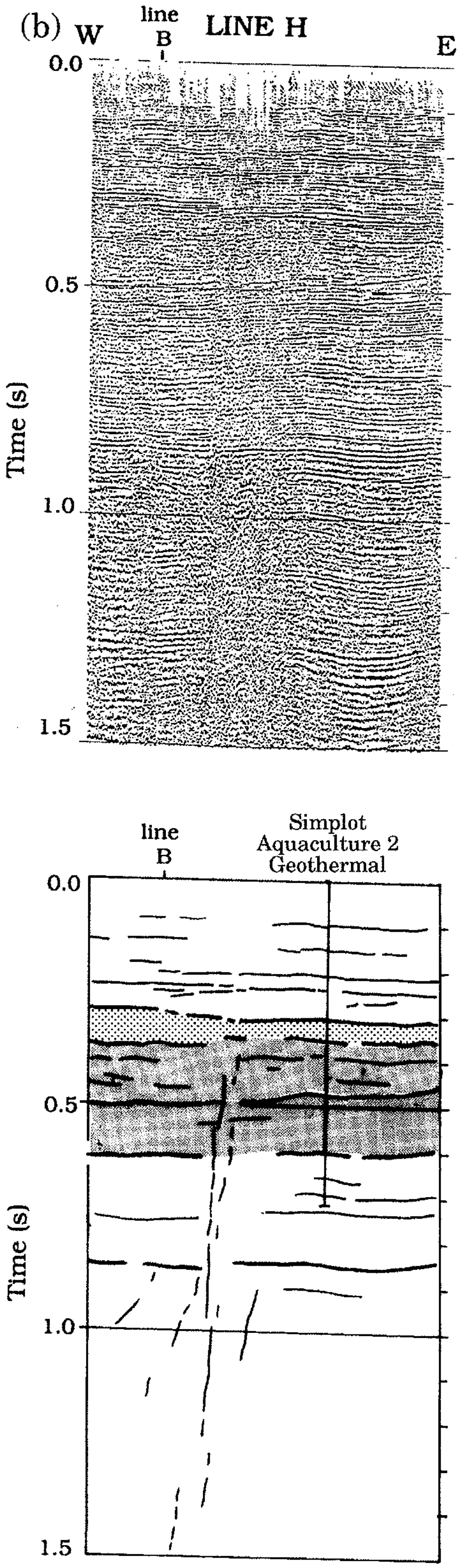


Figure 6–Continued.

The well was drilled with a truck-mounted rotary rig, using bentonite mud to control artesian flow. The mud line did not have a desander so fine sands circulated with the mud. This and other drilling problems resulted in an uneven quality of cuttings; nevertheless, sand units could be detected by a flush of sand in the cuttings. Samples were taken at 3.3 m (10 ft) intervals and washed through a stack of sieves, the finest sieve being #200 (0.074 mm). The relative amounts of sand and grain-size distribution are estimated from sieve analysis. The samples were examined and described with the use of a binocular microscope. The carbonate content of the clay and mud cuttings was estimated using criteria on Table 3, and colors were described using the Munsell system (Munsell Color, 1990). These attempts to quantify examination of cuttings of poorly consolidated sediment will prove useful in future studies comparing outcrop and other well data.

The lithology of the section can best be understood by viewing Figure 7, and will be discussed starting at the bottom of the figure. The deepest sediment from which cuttings have been recovered and examined from the study area is calcareous claystone, which dominates the lithology from 689 to 375 m (2250–1230 ft). This stratigraphic interval corresponds to unit 3 and underlying strata of the seismic profiles (Figure 6). Within this section of calcareous claystone are several fine-grained sand units about 7 m (20 ft) thick, some showing log profiles suggestive of fining-upward beds (Figure 7). Correlation of gamma and resistivity logs from 1 Richardson to 2 Simplot Geothermal wells (Figure 7) is consistent with the down-to-north depositional dip of this stratigraphic interval, as interpreted from the seismic sections.

Resistivity logs in Figure 8 show gradually upward-increasing resistivity from 457 m (1500 ft) to the top of seismic unit 3 at 366 m (1200 ft), indicating diminishing clay content and increasing silt and sand content upward in the sequence. This log character distributed over a thickness of 100 m (328 ft) is considered characteristic of distributary mouth-bar deposits of a prograding marine delta sequence by Coleman and Prior (1982) and Serra (1986). This coarsening-upward character may be too gradational to detect in cuttings, core, or outcrop of a claystone or mudstone; however, it is clearly the log signature of the transition from the prodelta facies to the delta-plain facies.

Overlying the relatively thick section of calcareous claystone is the principal sand facies, corresponding to seismic unit 2. Resistivity logs (Figure 8) show the sand is made up of several 3- to 7-m–thick beds, separated by thin clay beds about 1 m thick. Fine stratification (beds less than 20 cm), if present, cannot be resolved because the 16 in. (0.5 m) normal resistivity log cannot detect such thin beds. Cuttings and sand produced during well development are loose, very well sorted, and fine grained, having

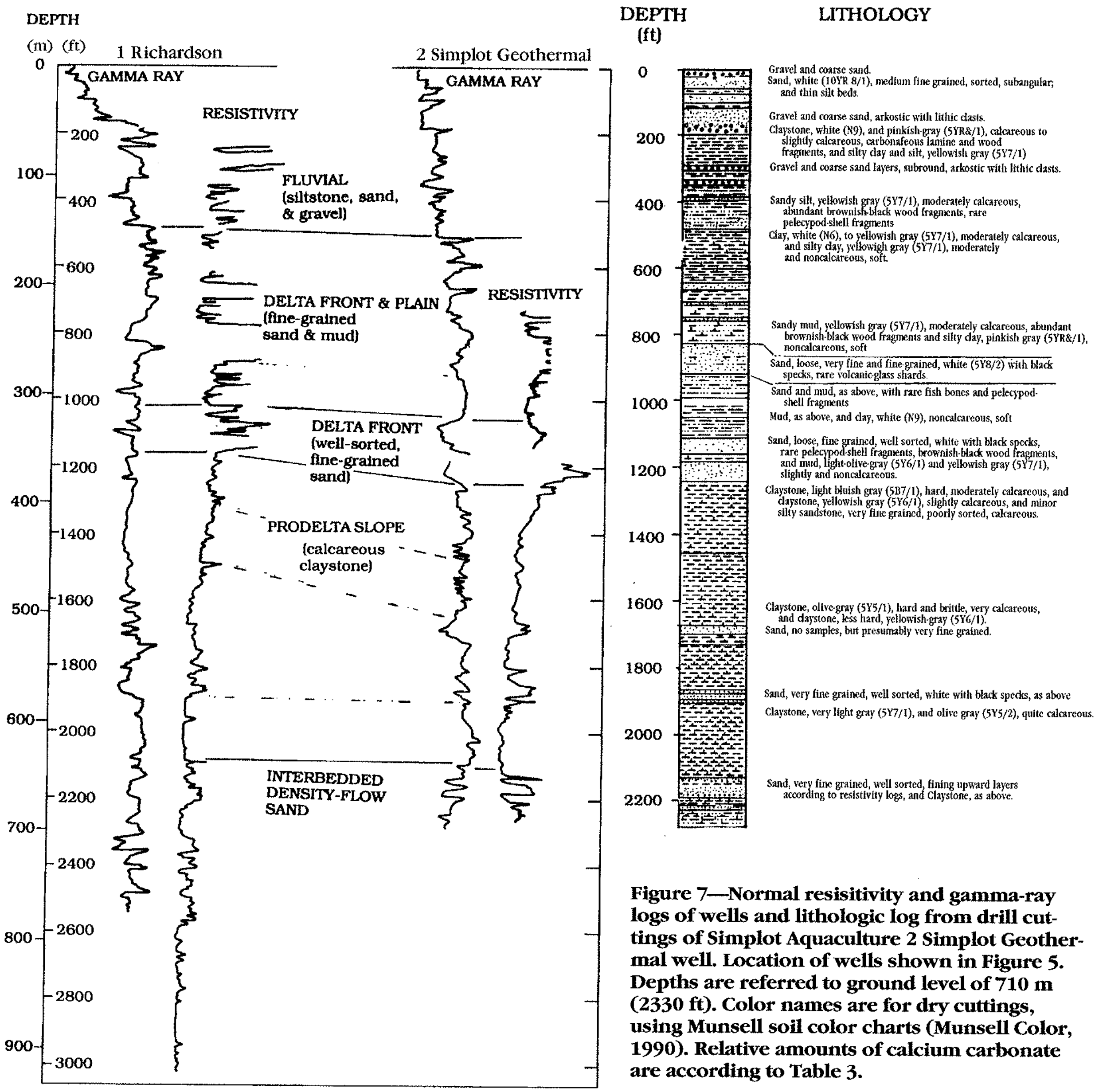


Figure 7—Normal resisitivity and gamma-ray logs of wells and lithologic log from drill cuttings of Simplot Aquaculture 2 Simplot Geothermal well. Location of wells shown in Figure 5. Depths are referred to ground level of 710 m (2330 ft). Color names are for dry cuttings, using Munsell soil color charts (Munsell Color, 1990). Relative amounts of calcium carbonate are according to Table 3.

a mean diameter of 0.1 mm (ϕ = 1, with one standard deviation of 0.5ϕ). Log profiles are interpreted as both massive beds and beds that coarsen (become less silty) upward. The logs show that sands are separated by thin clay beds. The Simplot Aquaculture 1 Simplot Geothermal well was completed with a 6-m screened interval in this sand. Pump testing of this 6-m section indicated a hydraulic conductivity of the screened section of 12 m/day, equivalent to a permeability of 1.0 d.

Between 457 m (1200 ft) and 107 m (350 ft), the stratigraphic interval corresponding to seismic unit 1, are numerous layers of well-sorted fine sand interbedded with moderately to slightly calcareous mud. The sand beds are typically 3 to 12 m (10–40 ft) thick. Woody and carbonaceous material are abundant in silt cuttings. Above 60 m (200 ft), the mudstones are noncalcareous, and the sands are mostly medium grained with some coarse-grained sands and gravels.

Table 2. Deep Petroleum and Geothermal Exploration Wells*

Map Symbol	Operator, Well Name, Date	Location	Total Depth (ft)	Bottom Formation
EO1	Sinclair 1 Eastern Oregon Land Co. (1955)	SW, Sec. 15, T16S, R44W	4888	unknown
MB1	Oroco 1 Mcbride (1955)	SE, Sec. 19, T16S, R46E	4506	unknown
OI1	Ore-Ida Foods/DOE 1 Ore-Ida (1979)	NE, Sec. 3, T18S, R47E	10,054	Miocene(?) basalt
KE1	H. K. Riddle 1 Kiesel Estate (1955)	SW, Sec. 8, T19S, R47E	5137	unknown
ER1	Idaho-Oregon Prod.Co. 1 Elvera-Recla (1950)	Sec. 19, T19S, R44E	4611	unknown
FS1	El Paso Natural Gas 1 Federal Spurrier (1955)	NE, Sec.5, T20S, R44E	7470	unknown
C1	Phillips Petroleum Geothermal A1 Chrestesen (1980)	NW, Sec. 2, T11N, R6W	7978	granite
BC1	Oroco-Simplot 1 Betty Carpenter (1955)	Sec. 4, T8N, R5W	2775	Idaho Group
VJ1	Oroco-Simplot 1 Virgil Johnson (1955)	SE, Sec.27, T8N, R4W	4040	unknown
VJ2	El Paso Natural Gas 2 Virgil Johnson (1956)	SE, Sec. 27, T8N, R4W	3522	Idaho Group
TD1	Oroco-Simplot 1 Ted Daws (1955)	NE, Sec. 34, T8N, R4W	1682	Idaho Group
HLL1	SOCAL 1 Highland Land & Livestock (1972)	Sec. 24, T6N, R5W	10,682	Miocene basalt
WS1	El Paso Natural Gas 1 Webber State (1956)	SW, Sec. 16, T5N, R3W	4522	Idaho Group
HL1	Sundance 1-30 Hunter-Linning (1983)	SW, Sec. 30, T5N, R3W	3555	Miocene basalt
R1	Oroco 1 Richardson (1955)	NW, Sec. 19, T4N, R3W	3000	Idaho Group
SG2	Simplot Aquaculture 2 Simplot Geothermal (1989)	NW, Sec. 19, T4N, R3W	2250	Idaho Group
JNJ1	Halbouty-Chevron 1 J. N. James (1976)	SE, Sec. 27, T4N, R1W	14,000	Tertiary volcanics
H1	Roden-Anschutz 1 Higgenson (1972)	NW, Sec. 9, T3N, R1W	3610	Miocene basalt
DF1	Champlin Petroleum 11-19 Upper Deer Flat (1981)	NW, Sec. 19, T2N, R2W	9047	Miocene basalt

*Locations shown in Figure 2.

INTERPRETATION OF SEISMIC FACIES UNITS

Seismic facies unit 3 is interpreted as the prodelta clay facies of a lacustrine delta system. The 4 to 5° dip of clinoform reflectors within this stratigraphic interval is similar to the slope of modern lacustrine deltas (Figure 9). The logs through this interval show it to be relatively monotonous claystone, very gradually coarsening upward in the upper 100 m, and interrupted in the lower part by interbedded fine sand layers. The seismic section (Figure 6a) shows this unit produces few or no reflections, and is interpreted to be characteristic of the monotonous claystone. The reflections within this unit and the higher amplitude lower segments of the clinoforms are interpreted to be intercalated sand layers on the prodelta slope, and density-flow sand accumulating as downlapping layers at the base of the slope.

Seismic facies unit 2 is interpreted to be the delta-front and delta-plain sand facies. This unit is the topset interval of the progradational reflection configuration. Unit 2 contains the principal sand aquifers in this section because of the relatively thick and very well-sorted sands of the delta front.

In Mitchum et al. (1977, p. 128), seismic facies units 2 and 3 would be considered a single seismic facies unit of a prograding depositional system. The units are discussed separately in this paper because of the great difference in lithology and permeability between the sand of unit 2 and the predominant claystone of unit 3.

Seismic facies unit 1 is interpreted as the delta-plain, fluctuating-shoreline, and river flood-plain facies. The uppermost calcareous muds (about 65 m or 215 ft deep) are an indication of a lacustrine envi-

Table 3. Terminology for Carbonate Content of Well Cuttings Used in Figure 7: Response to 10% HCl, Cold*

Very calcareous: sample reacts violently, floats on top of acid and moves about the surface (90–100% $CaCO_3$).

Calcareous: sample reacts immediately but only moderately and jumps off bottom every 0.3–1 s, moves about in acid between bottom and surface (50–90% $CaCO_3$).

Moderately calcareous: Sample reacts slowly at first, but accelerates to a continuous reaction after a minute or two with some bobbing at the bottom (50–10% $CaCO_3$).

Slightly calcareous: very slow reaction, bubbles evolve one at a time, with several seconds between release of each bubble (<10% $CaCO_3$).

Noncalcareous: sample put into warm HCl to verify that no bubbles are released.

Dolomitic: pure dolomite effervesces only when crushed or abraded and put in cold 10% acid or put into warm acid; if dolomite is suspected, it can be verified by digestion in warmer or stronger acid or by staining methods.

*Modified after Exploration Logging, Inc. (1981).

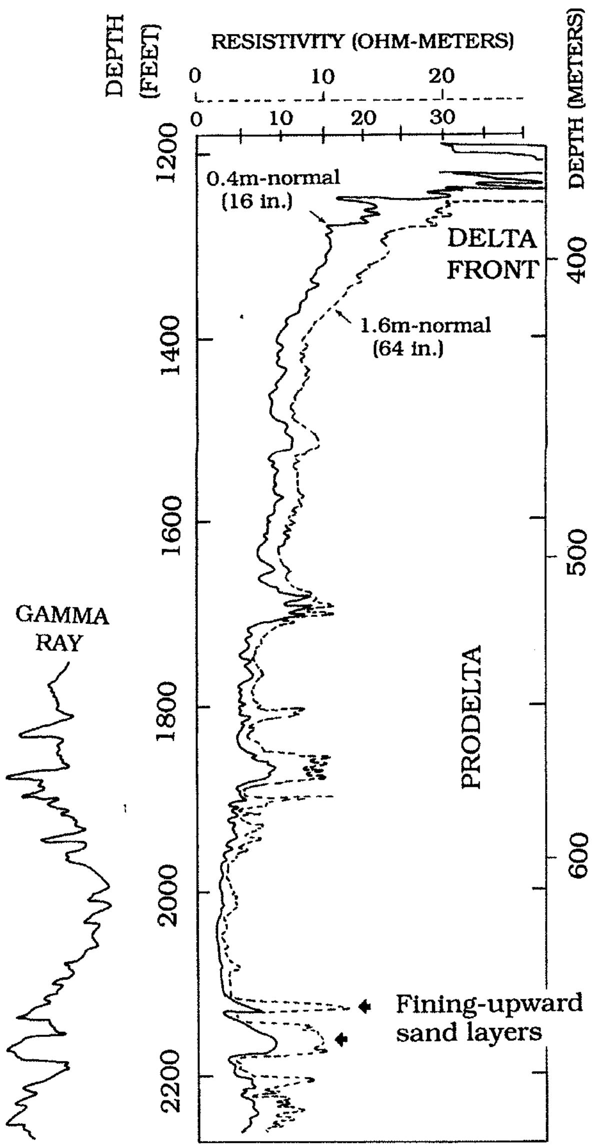


Figure 8—Detailed resistivity log of Simplot Aquaculture 2 Simplot Geothermal well showing fining-upward sand layers in the prodelta facies, the gradual upward-increasing resistivity in the prograding prodelta clays, and the abrupt contact and high resistivity of the delta-front sands.

ronment; however, interbedded gravel and coarse sand indicate fluvial systems, and lakeshore environments may have alternated during deposition of this unit. Abundant woody material in parts of this unit indicates a delta-plain environment, and muds may be interdelta bays and lagoons.

SLOPE GEOMETRY OF LACUSTRINE DELTAS AND CLINOFORM REFLECTIONS

The key to identifying the delta system is recognition of the unit with clinoform reflections as the prodelta slope. To verify that this is indeed the prodelta slope, bathymetry of modern lakes was examined. Two studies of the geometry of modern lacustrine deltas (Born, 1972; Müller, 1966) are sufficiently detailed to use as a model of the features seen on seismic sections and well logs. Profiles of modern deltas of the Truckee River in Pyramid Lake of Nevada, and of the Rhine River into Lake Constance, are shown in Figure 9. These delta bathymetric profiles are nearly identical to clinoform reflections on the seismic sections (Figure 6). Prodelta slopes of these modern lacustrine deltas are 2 to 6°. Seismic sections showing clinoforms of a prograded fluvial-dominated marine prodelta (Figure 10b) closely resemble the slope geometry of lacustrine deltas shown on Figure 9).

A point of confusion still exists among some geologists working with seismic data but using ideas of delta deposition from older literature dating back to Gilbert (1885). Large-scale morphology of bedded delta deposits is vastly more complex than Gilbert's illustration (Figure 10a) for the form of bed-load material deposited by a stream issuing into a body of still water. The illustration describes internal sedimentary structure that occurs in some sand layers at the delta front, but it is not generally applicable to all sand layers. The term "Gilbert delta" still pervades geological literature and is a common misconception for the shape of all delta deposits. The feature Gilbert described should be called "Gilbert bedding" because it is a form of stratification of one type of bed, and is not a characteristic of the delta as a whole. The distinction is appropriate here because the true- scale diagram of Gilbert bedding is nearly identical in appearance to the appearance of the prodelta beds shown on a seismic section with a 10× vertical exaggeration (Figure 10b). This apparent similarity of geometry is a matter of display scale. The foreset beds of Gilbert bedding may constitute a sand unit 2 to 30 m (7–100 ft) thick, with internal stratification dipping 8 to 33°, whereas the seismically displayed prodelta unit is in reality a mud or clay unit several hundred meters thick with internal stratification dipping only 2 to 6°.

FINE-GRAINED DELTAS AND COARSE-GRAINED DELTAS

Recent literature on alluvial deltas distinguishes between deltas that have steep delta-face slopes composed of coarse clastic material, and deltas that have gently sloping delta-face slopes composed

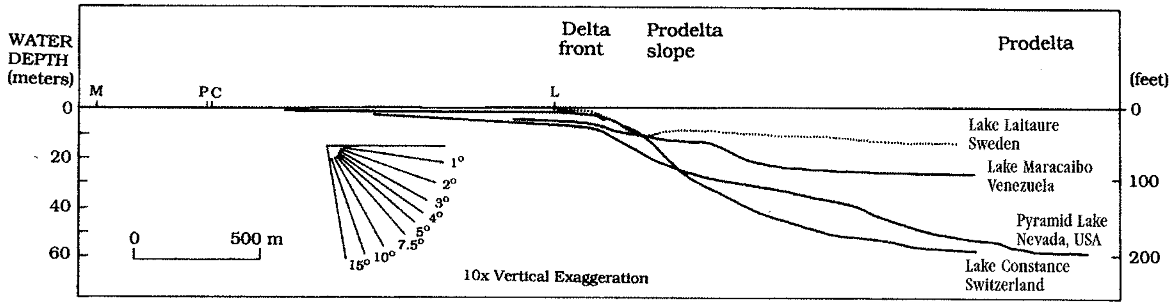


Figure 9—Profiles of modern lacustrine deltas showing the delta front and the prodelta slope. Data sources are Lake Constance (Müller, 1966); Lake Laitaure (Axelsson, 1967); Lake Maracaibo (Hyne et al., 1979); and Pyramid Lake (Born, 1972; Russell, 1885).

mostly of fine sand and mud (Nemec, 1990). In the "coarse-grained delta," gravel and coarse sand beds are deposited at the delta front; these deposits are characteristic of braid deltas and fan deltas (McPherson et al., 1987). Coarse-grained deposits do not occur in the delta-front deposits described in this paper.

The delta described in this paper is a fine-grained delta because only fine sand is deposited at the delta front. Stanley and Surdam (1978) describe a lacustrine-delta sequence of mostly fine sand and mud in the Laney Member of the Green River Formation. In this delta only medium to fine sand sediment, along with suspended silts and clays, was transported to the delta front. Presumably, this delta is the deposit of a large river system where sediment is transported over a long, low-gradient flood plain and delta plain to the lake edge, a considerable distance from the source of coarse clastic material. The sand beds are 30 m thick, with an internal stratification as described by Gilbert (Figure 10a). Therefore, the Gilbert type of internal stratification also occurs in tabular bodies of fine sand; however, examination of fine-grained delta systems in the Green River Formation on a broader scale may show that much of the sand bedding is horizontal and not angle-of-repose bedding at the delta front.

Beyond the delta front is the prodelta slope, formed mostly by the deposition of the suspended load of silt and clay, as illustrated in Figure 11. Slopes of the prodelta beds are generally a few degrees (Figure 9). The prodelta deposits, however, make up most of the volume of deposits of the fine-grained deltas, and also are the most easily identified part of the delta on seismic sections. The zone of gently inclined bedding of the lacustrine prodelta extends at least 300 m (1000 ft) horizontally, and more typically 1500 to 2000 m, (5000–6500 ft) beyond the delta front (see examples in Figure 9). The vertical drop over the slope of Gilbert foreset beds seen in outcrop is often used as an indicator of lake depth; however, the vertical drop of the low-dipping prodelta slope deposits commonly is much greater and is a better indicator of the total depth of the lake basin. As stated by Stanley and Surdam (1978), thickness of the Gilbert-type sequence is an indicator of the depth of water into which bed-load deposits prograded, but not the depth of the lake basin itself.

EVIDENCE FOR DENSITY-FLOW ORIGIN OF SAND IN THE PRODELTA DEPOSIT

Sand layers are relatively rare in the thick sequence of the prodelta calcareous claystone described in this paper; nevertheless, several high-amplitude seismic reflections do occur from the lower part of this unit (Figure 6), and indeed, the resistivity logs show layers of fine sand 3 to 7 m (10–20 ft) thick in the 2 Simplot Geothermal well at depths of 518 m (1700 ft), 579 m (1900 ft), and 640 m (2100 ft) (Figure 7). All of these sand layers have resistivities less than 15 ohm-meters, much less than resistivities of the clean-sand aquifers at 335 m (1100 ft). Although drilling problems prevented collection of good cuttings from the well, the resistivity logs show that the layers at 640 m (2100 ft) are fining upward. This kind of graded bedding with better sorted, coarser grains at the base of the bed, suggests a density-flow mechanism for dispersal of sands into layers several to 10 m (33 ft) thick on the prodelta slope. Sturm and Matter (1978) describe prodelta slope deposits in Lake Brienz, Switzerland. In the bottom sediments about 5 km (3 mi) from the delta front, they found centimeter-scale graded fine sand and silt beds, deposited annually with laminated calcareous mud, and rare thick beds (up to 1.5 m or 5 ft) of sand. Sturm and Matter attribute the annual thin sand layers to low-density turbidity currents during high discharge that travel in channels down the delta slope and spread on the flat lake floor. They attribute the thick layers to high-density turbid-

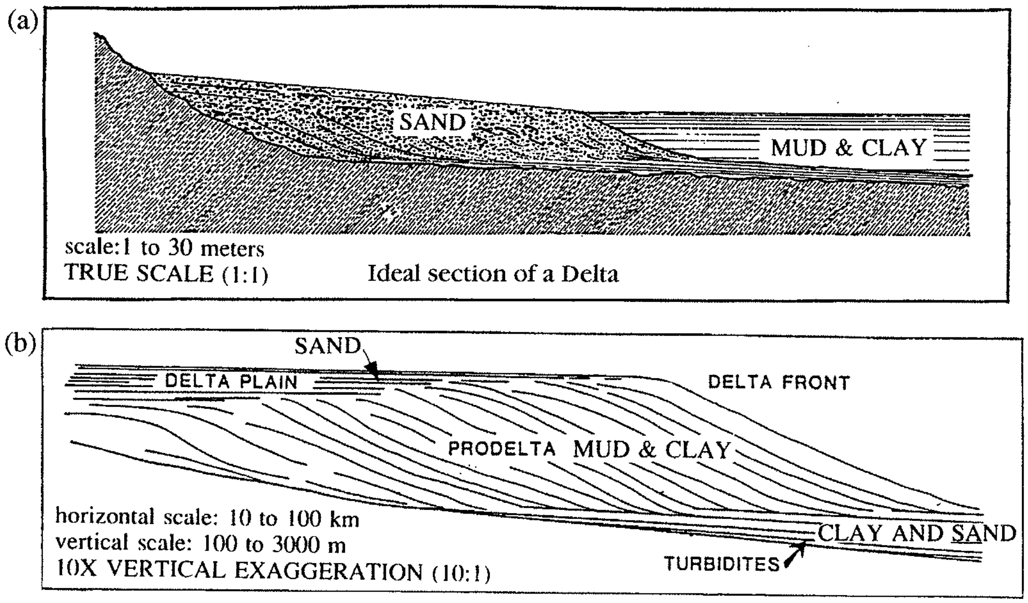


Figure 10—Illustration of the similarity in morphology of relatively small scale bedding of the Gilbert delta compared to the grand scale of Berg's (1982) model of a fluvial-dominated marine delta. (a) Misconception of delta morphology dating shown on G. K. Gilbert's (1885) true-scale illustration of a section through a delta based on observations of coarse-grained deposits of streams that issued into Pleistocene Lake Bonneville, Utah. Thickness of the foreset-bedded sand is typically 1 to 30 m. (b) Berg's (1982) generalized model of a large prograded fluvial-dominated delta from seismic sections. Horizontal scale is 10 to 100 km, and vertical scale is 100+ to 3000 m (O. R. Berg, 1991, personal communication).

ity currents of catastrophic floods that occur once or twice a century. These floods could produce a rapid accumulation of delta-front sands on angle-of-repose slopes that fail to produce turbidity flows and thick sand units beyond the delta front. Sand layers shown by the logs (Figure 7) are thicker than those described by Sturm and Matter (1978); however, they may be of similar origin.

Further evidence of the graded nature of these beds is gained from attempts to develop these layers for artesian flow of hot water. Casing was run to the top of sands at 649 m (2130 ft), and 9 m (30 ft) of well screens set across these sands, resulting in a flow of only 2.5 L/s (40 gal/min) of 40°C (141°F) water detected by downhole flowmeter and temperature logging. The test was an imperfect measure of the artesian flow from the zone because of a leak in the annular packer at 336 m (1200 ft), which allowed pressure of about 0.15 MPa (21 psi) and flow of about 9.5 L/s (150 gal/min) from the upper sands to enter the hole at this higher level. Nevertheless, it appears from the resistivity log and the flowmeter test that these deeper sand layers yield only small flows and have lower permeability relative to shallower sands. Because of their graded stratification, only the thin basal part of the sands may be clean, sorted, and have good permeability.

IMPLICATIONS FOR LAKES IN THE WESTERN SNAKE RIVER PLAIN BASIN

The delta sequence described in this paper records an episode of sediment infilling of Pliocene Lake Idaho. Relief of the prodelta slope on the seismic section (Figure 6a) is 180 m (600 ft). Decompacting this thickness of prodelta mud by an estimated 120% implies an original prodelta slope relief of about 255 m (840 ft). From this, one can conclude that the prodelta was prograding into a lake that was about 255 m (840 ft) deep, and that the delta was prograding northward. Seismic data to the east of the Caldwell study area also show a similar prograding sequence with a slope facing northwestward (Wood and Anderson, 1981); these directions of progradation are shown on Figure 2.

The following discussion attempts to reconcile the relatively low elevation of the delta system preserved in the middle of the basin with the much higher elevation lake deposits exposed around the margin of the basin, generally within 130 m (430 ft) of an elevation of 1000 m (3300 ft). The interface between the prodelta mud and the delta sand facies occurs in shallow water (about 10 m or 33 ft according to modern examples in Figure 9). Therefore, the base of the delta sands can be regarded as about lake level at the time of deposition.

The top of the buried prodelta facies in the Caldwell area is at elevation of 405 m (1330 ft), considerably lower than the present bedrock sill at Cobb Rapids (634 m, 2080 ft), and much lower than the highest lake deposits near the present outlet, which are 850 to 975 m (2790–3200 ft) near Huntington, Oregon. The top of the prodelta is identified on the resistivity logs of four other wells generally in the center of the basin, and the elevation is shown in brackets on the map (Figure 2). These elevations range from 437 to 504 m (1433–1653 ft), up to 100 m (330 ft) higher than at Caldwell. It appears that the deltaic sequence at Caldwell has subsided about 445 to 575 m (1460–1900 ft) from the level of the highest lake deposits since it was deposited. Differential compaction and a small tectonic component may account for the higher elevation in the other wells. The amount of compaction subsidence can be estimated for a horizon that is now 405 m deep, (1330 ft) and underlain by about 853 m (2800 ft) of mudstone. Using the method of Sclater and Christie (1980) programmed for PC-computer calculation by Hsui (1986), the calculation suggests compaction subsidence at Caldwell may be about 220 m (720 ft). The remaining 225 to 325 m (740–1060 ft) can be attributed to tectonic faulting and downwarping of the center of the basin. This tectonic component agrees well with 180 to 370 m (600–1200 ft) of offset of Idaho Group sediment typically observed on the boundary normal faults on the north side of the plain (Figure 3) (Wood and Burnham, 1987). Therefore, the relatively low elevation of the delta deposits can be reconciled with the higher elevation of remnant lake deposits on the basin margins by taking into account compaction and tectonic fault displacement and downwarping.

Position of the delta in the middle of the basin (Figures 1, 2) suggests that at the time of deposition, the upper southeastern two thirds of the basin was delta plain and flood plain, and the lower northwestern third was a deep lake environment. The fact that prodelta deposits slope north and northwest (Figure 2) indicates that a major source of sediment lay to the southeast, just as the Snake River flows today.

The previous discussion provides some insight on the history of the lake and its outlet. During part of the history of the basin, the lake may have had only an ephemeral outlet; much of the inflow may have evaporated. Present inflow to the basin averages 16 $\times 10^9$ m^3/yr. Typical annual evaporation from reservoirs in the Snake River plain region is about 1 m (3 ft), and this interior continental region evaporation may have been similar during the Pliocene. At the maximum areal extent shown in Figure 1, a surface area of 13,000 km^2 (5000 mi^2) evaporation could have exceeded 13 $\times 10^9$ m^3/yr, an amount that nearly equals the inflow, so that in some years most or all of the inflow may have evaporated. The general lack of evaporitic facies in the Glenns Ferry Formation (Kimmel, 1982) is evidence against a closed basin environment. Although gypsum is common in the older Chalk Hills Formation, little is reported in the Glenns Ferry Formation. Deposition of oolitic sands on shorelines and assemblage of fish fossils led Swirydczuk et al. (1980) to suggest a water environment similar to that of the present Pyramid Lake, Nevada. This analogy is further supported by the abundance of calcareous claystone in the prodelta facies (Figure 7) that is similar to the prodelta clay of Pyramid Lake, which was examined by Swain and Meader (1958). Pyramid Lake waters are 5% total dissolved solids and have an alkaline pH of 9. Therefore, evaporation appears to have significantly increased alkalinity and chemical concentration in Lake Idaho over the 0.01% total dissolved solids of inflow streams. Occasional outflow prevented the water from becoming saline.

At the time of deposition of the deltaic sequence, near Caldwell, the upper two-thirds of the basin was a river flood plain and the surface area of the lake may have been reduced to one third of that shown in Figure 1, or about 4000 km^2 (1500 mi^2). As the lake area was reduced by sediment infilling, reduced evaporation must have allowed the lake level to rise in a closed basin and ultimately spill into adjacent basins.

Many workers have cited paleobiological evidence of freshwater mollusks and fish as evidence of a former course of the Snake River through eastern Oregon to the Klamath basin of California (Malde, 1991) shown in Figure 1. No physical evidence for such an outlet channel has been found, although if lakes only occasionally spilled into other closed basins, a deep canyon might not have been cut nor the fluvial sediment preserved. Details of the lake's past spill points and chemical history of its waters remain a fascinating puzzle for future research.

The lake was eventually doomed, with waning rates of tectonic subsidence in the western plain in the late Pliocene (Clemens, 1993), and as sediment displaced the lake area causing a reduced area for evaporation. The lake waters finally overtopped a spillway into the Columbia–Salmon rivers drainage near Huntington, Oregon. Timing of the capture of the waters of the Snake River drainage basin appears to be about 2 Ma (Malde, 1991; Othberg, 1992). It is tempting to correlate the final lake rise with the climatic change that heralded the northern hemisphere ice ages, which, on the basis of the marine oxygen–isotope record, Shackleton and Opdyke (1977) place at 2.5 m.y. Benson (1981) shows that 100-m (330-ft) lake-level rises in the Lake Lahontan basin were caused by paleoclimate fluctuations in the late Pleistocene. Such lake-level rises also likely occurred in the early Pleistocene and overtopped

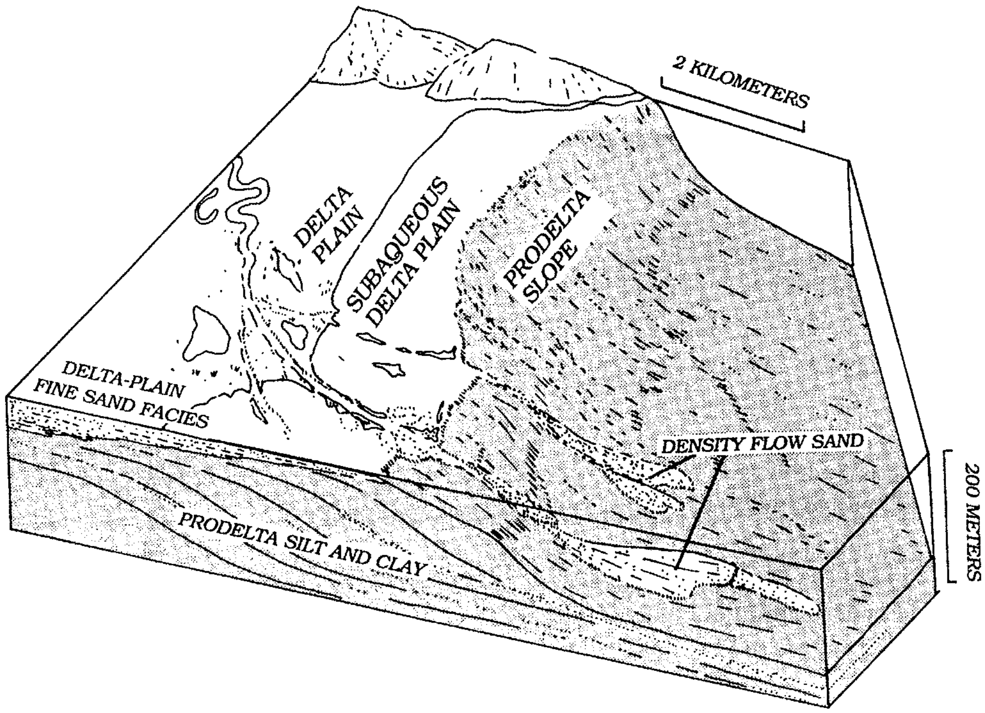


Figure 11—Drawing of a lacustrine delta environment, with a 10× exaggerated vertical scale, showing distribution of sand units within the dominantly mud and clay sequence.

the basin sill. Once the new outlet was established, the river would have eroded the sill and deepened the ancestral Hells Canyon. Present geomorphology suggests that Dead Indian Ridge near Weiser, Idaho, was the sill (Figure 2), and that the outlet has cut down about 215 m (700 ft) to the present river level of 635 m (2080 ft) at Cobb Rapids, where it descends into Hells Canyon.

CONCLUSIONS

Neogene lake sediments of the western Snake River plain provide an example of a lacustrine delta and prodelta system that can be identified in the subsurface by seismic reflection patterns. The key to recognizing the delta system is progradational clinoform reflections of the prodelta mud and clay facies detected on seismic sections oriented normal to the direction of progradation. The clinoforms have a 2 to 6° depositional dip, which is similar to the prodelta slopes of modern lakes. Upper ends of the prodelta clinoforms are overlain in a toplap relationship by strong amplitude, near-horizontal reflections from the delta-front sands. Delta-front sands are expressed by several to 5 cycles of strong amplitude reflections, each continuous for several hundred meters. The prodelta can be identified on resistivity logs as a gradually coarsening upward mud unit typically 100 m (330 ft) thick and abruptly overlain by high-resistivity sands of the delta front. The downward distal ends of the prodelta clinoforms have a downlap relationship to underlying strong amplitude reflections of density-flow sand layers. Density-flow sands 3 to 7 m (10–25 ft) thick are identified on resistivity logs of boreholes by a fining-upward character.

In the delta described here, the delta-front sands are well sorted and fine grained. The prodelta sediment is mostly calcareous mud and clay with intercalated layers of density-flow sands in the lower part.

On an exaggerated vertical-scale seismic section the prodelta reflection patterns have the appearance of the true-scale diagram of a Gilbert delta. It is a misconception that these large delta systems resemble the Gilbert delta with steeply dipping foreset beds. Foreset bedding is a relatively small-scale feature limited to internal bedding of sands, whereas the prodelta slope is less than 6° and is mostly mud and clay.

The delta system described here is the first major depositional system identified for deposits of the Idaho Group, and provides some insight into the history of Pliocene Lake Idaho. The delta facies occurs in the center of the basin and has subsided 445 to 575 m (1460–1900) with respect to sediment exposed on the basin margins. Estimated subsidence from sediment compaction is about 220 m (720 ft), leaving about 225 to 325 m (740–1060 ft) attributable to tectonic downwarping and faulting. The prodelta slope relief, corrected for compaction, indicates a lake basin about 255 m (840 ft) deep at the time of delta deposition in the center of the basin.

Position of the delta in the northwestern part of the basin indicates that the southeastern two thirds of the original lake area of about 13,000 km^2 (5000 mi^2) had filled by the time the delta sequence had reached Caldwell, Idaho, so that the existing lake area was about 4000 km^2 (1500 mi^2). The large surface area of the early lake could have evaporated most of the annual inflow, suggesting the lake only occasionally spilled into other basins. As lake-surface area was reduced by sedimentation, evaporation diminished, and the lake level rose. Onset of the northern hemisphere ice ages further reduced evaporation and perhaps triggered a lake-level rise that overtopped a basin sill near Weiser, Idaho, and spilled into the Columbia–Salmon River drainage about 2 m.y. Present geomorphology near the spillway suggests the Snake River subsequently cut down the spillway about 215 m, (700 ft) and deepened Hells Canyon, now the deepest gorge in North America.

REFERENCES CITED

Armstrong, R. L., W. P. Leeman, and H. E. Malde, 1975, K-Ar dating, Quaternary and Neogene rocks of the Snake River plain, Idaho: American Journal of Science, v. 275, p. 225–251.

Armstrong, R. L., J. R. Harakal, and W. M. Neill, 1980, K-Ar dating of Snake River plain (Idaho) volcanic rocks—new results: Isochron-west, v. 27, p. 5–10.

Axelsson, V., 1967, The Laitaure delta—a study of deltaic morphology and processes: Geografiska Annaler, v. 49, p. 1– 127.

Bates, R. L., and J. A. Jackson, 1980, Glossary of geology, 2nd ed: Washington, D. C., American Geological Institute, 751 p.

Benson, L. V., 1981, Paleoclimatic significance of lake level fluctuations in the Lahontan basin: Quaternary Research, v. 16, 390–402.

Berg, O. R., 1982, Seismic detection and evaluation of delta and turbidite sequences: their application to exploration for the subtle trap: AAPG Bulletin, v. 66., p. 1271–1288.

Blackwell, D. D., 1989, Regional implications of heat flow of the Snake River plain, northwestern United States: Tectonophysics, v. 164, p. 323–343.

Born, S. M., 1972, Late Quaternary history, deltaic sedimentation, and mudlump formation at Pyramid Lake, Nevada: Reno, University of Nevada Center for Water Resources Research Desert Research Institute, 97 p.

Bosworth, W., 1985, Geometry of propagating continental rifts: Nature, v. 316, p. 625–628.

Braun, J., and C. Beaumont, 1989, A physical explanation of the relation between flank uplifts and the breakup unconformity at rifted continental margins: Geology, v. 17, p. 760–764.

Brott, C. A., D. D. Blackwell, and J. P. Ziagos, 1981, Thermal and tectonic implications of heat flow in the eastern Snake River plain, Idaho: Journal of Geophysical Research, v. 86, p. 11,709–11,734.

Clemens, D. M., 1993, Tectonics and silicic volcanic stratigraphy of the western Snake River plain, southwestern Idaho: M.S. thesis, Arizona State University, Tempe, Arizona, 209 p.

Coleman, J. M., and D. B. Prior, 1982, Deltaic environments of deposition, in P. A. Scholle and D. Spearing, eds., Sandstone depositional environments: AAPG Memoir 31, p. 139–178.

Deacon, R. J., and G. T. Benson, 1971, Oil and gas potential of eastern Oregon and Washington and Snake River basin of western Idaho, in I. H. Cram, ed., Future petroleum provinces of the United States—their geology and potential: AAPG Memoir 15, p. 354–359.

Ekren, E. B., D. H. McIntyre, and E. H. Bennett, 1981, Geologic map of Owyhee County, Idaho, west of Longitude 116°W, U. S. Geological Survey Miscellaneous Investigations Map I-1256, scale 1:125,000.

Ekren, E. B., D. H. McIntyre, and E. H. Bennett, 1984, High temperature, large-volume, lava-like ash flows without calderas in southwestern Idaho: U. S. Geological Survey Professional Paper 1272, 76 p.

Exploration Logging, Inc., 1981, Formation evaluation, part 1: geological procedures, section 4: carbonate rocks: Sacramento, California, Exploration Logging, Inc., 43 p.

Fitzgerald, J. F., 1982, Geology and basalt stratigraphy of the Weiser embayment, west-central Idaho, in B. Bonnichsen and R. M. Breckenridge, eds., Cenozoic geology of Idaho: Idaho Geological Survey Bulletin 26, p. 103–128.

Fouch, T. D., and W. E. Dean, 1982, Lacustrine and associated clastic depositional environments, in P. A. Scholle and D. Spearing, eds., Sandstone depositional environments: AAPG Memoir 31, p. 87–114.

Gallegos, D., P. Johnson, S. Wood, and W. S. Snyder, 1987, Depositional facies patterns along the Boise front: Northwest Geology, v. 16, p. 47–59.

GeothermEx, Inc., 1980, Technical report—deep well test and exploration program for Ore-Ida No. 1, Ontario, Oregon, v. 1 (text): U. S. Department of Energy subcontract ET-78-C-07-1725-GTX, 61 p.

Gilbert, G. K., 1885, The topographic features of lake shores: U. S. Geological Survey Fifth Annual Report, p. 69–123.

Hooper, P. R., and R. M. Conrey, 1989, A model for the tectonic setting of the Columbia River basalt eruptions, in S. P. Reidel and P. R. Hooper, eds., Volcanism and tectonism in the Columbia River flood-basalt province: Geological Society of America Special Paper 239, p. 293–305.

Hsui, A. T., 1985, subside, a basin analysis program for IBM PC: Urbana, Illinois, Callidus Software.

Hyne, N. J., W. A. Cooper, and P. A. Dickey, 1979, Stratigraphy of intermontane, lacustrine delta, Catatumbo River, Lake Maracaibo, Venezuela: AAPG Bulletin, v. 63, p. 2042–2057.

Ikawa, T., 1991, Die knozoischen sedimente des Biwa-Sees—dargestellt mit hilfe der reflexionsseismik, in S. Horie, ed., Die geschichte des Biwa-Sees in Japan: Innsbruck, Austria, Universitatsverlag Wagner, p. 40–57.

Izett, G. A., 1981, Volcanic ash beds: recorder of upper Cenozoic silicic pyroclastic volcanism in the western United States: Journal of Geophysical Research, v. 86, p. 10,200–10,222.

Jenks, M. D., and W. Bonnichsen, 1989, Subaqueous basalt eruptions into Pliocene Lake Idaho: Snake River plain, Idaho, in V. E. Chamberlain, R. M. Breckenridge, and B. Bonnichsen, eds., Guidebook to the geology of northern and western Idaho and surrounding area: Idaho Geological Survey Bulletin 28, p. 17–34.

Katz, B. J., ed., 1990, Lacustrine basin exploration: AAPG Memoir 50, 340 p.

Kimmel, P. G., 1982, Stratigraphy, age, and tectonic setting of the Miocene-Pliocene lacustrine sediments of the western Snake River plain, Oregon and Idaho, in B. Bonnichsen and R. M. Breckenridge, eds., Cenozoic geology of Idaho: Idaho Geological Survey Bulletin 26, p. 559–578.

Kirkham, V. J., 1935, Natural gas in Washington, Idaho, eastern Oregon, and northern Utah, in H. A. Ley, ed., Geology of natural gas: AAPG, p. 221–244.

Kjelstrom, L. C., 1986, Flow characteristics of the Snake River and water budget for the Snake River plain, Idaho and eastern Oregon: U. S. Geological Survey Hydrologic Investigations Atlas HA-680, 2 sheets, scale 1:500,000.

Kuntz, M. A., H. R. Covington, and L. J. Schorr, 1992, An overview of basaltic volcanism of the eastern Snake River plain, Idaho, in P. K. Link, M. A. Kuntz, and L. B. Platt, eds., Regional geology of eastern Idaho and western Wyoming: Geological Society of America Memoir 179, p. 1–53.

Leeman, W. P., 1989, Origin and development of the Snake River plain—an overview, in R. P. Smith and K. L. Ruebelmann, eds., Yellowstone–Snake River plain volcanic province: International Geological Congress Guidebook T-305, Washington D. C., American Geophysical Union, p. 69–77.

Liro, L. M., and Y. C. Pardus, 1990, Seismic facies analysis of fluvial-deltaic lacustrine systems—upper Fort Union Formation (Paleocene), Wind River basin, Wyoming, in B. J. Katz, ed., Lacustrine basin exploration, AAPG Memoir 50, p. 225–242.

Malde, H. E., 1972, Stratigraphy of the Glenns Ferry Formation from Hammett to Hagerman, Idaho: U. S. Geological Survey Bulletin 1331-D, 19 p.

Malde, H. E., 1987, A guide to the Quaternary geology and physiographic history of the Snake River Birds of Prey area, Idaho: Northwest Geology, v. 16, p. 23–46.

Malde, H. E., 1991, Quaternary geology and structural history of Snake River plain, Idaho, in R. B. Morrison, ed., Quaternary nonglacial geology, conterminous United States: Boulder, Colorado, Geological Society of America, The Geology of North America v. K-2, p. 251–281.

Malde, H. E., and H. A. Powers, 1962, Upper Cenozoic stratigraphy of the western Snake River plain, Idaho: Geological Society of America Bulletin, v. 73, p. 1197–1220.

McPherson, J. G., G. Shanmugam, and R. J. Moiola, 1987, Fan- deltas and braid-deltas: varieties of coarse-grained deltas: Geological Society of American Bulletin, v. 99, p. 331–340.

Middleton, L. T., M. L. Porter, and P. G. Kimmel, 1985, Depositional settings of the Chalk Hills and Glenns Ferry formations west of Bruneau, Idaho, in R. M. Flores and S. S. Kaplan, eds., Cenozoic paleogeography of the west-central United States: Rocky Mountain Section of the Society of Economic Paleontologists and Mineralogists, p. 37–53.

Mitchum, R. M., P. R. Vail, and J. B. Sangree, 1977, Seismic stratigraphy and global changes in sea level, part 6: stratigraphic interpretation of seismic reflection patterns in depositional sequences, in C. E. Payton, ed., Seismic stratigraphy—applications to hydrocarbon exploration: AAPG Memoir 26, p. 117–133.

Morley, C. K., R. A. Nelson, T. L. Patton, and S. G. Munn, 1990, Transfer zones in the east African rift system and their relevance to hydrocarbon exploration and rifts: AAPG Bulletin, v. 74, p. 1234–1253.

Müller, G., 1966, The new Rhine delta in Lake Constance, in M. L. Shirley, ed., Deltas in their geologic framework: Houston Geological Society, p. 107–125.

Munsell Color, 1990, Munsell soil color charts: Baltimore, Maryland, Munsell Color, Macbeth Division of Kollmorgen Instruments, 7 charts.

Nemec, W., 1990, Deltas—remarks on terminology and classification, in A. Collela and D. B. Prior, eds., Coarse-grained deltas: International Association of Sedimentologists Special Publication 10, p. 3–12.

Newton, V. C., and R. E. Corcoran, 1963, Petroleum geology of the western Snake River basin, Oregon-Idaho: Oregon Department of Geology and Mineral Industries Oil and Gas Investigation 1, 67 p.

Othberg, K. L., 1992, Geology and geomorphology of the Boise Valley and adjoining areas, western Snake River plain, Idaho: Ph.D. thesis, University of Idaho, Moscow, Idaho, 124 p.

Othberg K. L., and L. R. Sanford, 1992, Geologic map of the Boise Valley and adjoining area, western Snake River plain, Idaho: Idaho Geological Survey, Geologic Map Series, scale 1:100,000.

Pierce, K. L., and L. A. Morgan, 1992, The track of the Yellowstone hot spot: volcanism, faulting, and uplift, in P. K. Link, M. A. Kuntz, and L. B. Platt, eds., Regional geology of eastern Idaho and western Wyoming: Geological Society of America Memoir 179, p. 1–53.

Rodgers, D. W., W. R. Hackett, and H. T. Ore, 1990, Extension of the Yellowstone plateau, eastern Snake River plain, and Owyhee plateau: Geology, v. 18, p. 1138–1141.

Rosendahl, B. R., D. J. Reynolds, P. M. Lorber, C. F. Burgess, J. McGill, D. Scott, J. J. Lambiase, and S. J. Derksen, 1986, Structural expressions of rifting: lessons from Lake Tanganyika, Africa, in L. E. Frostick, R. W. Renaut, I. Reid, and J. J. Tiercelin, eds., Sedimentation in the African rifts: Geological Society Special Publication 25, p. 29–43.

Russell, I. C., 1885, Geological history of Lake Lahontan: U. S. Geological Survey Monograph 11, 288 p.

Sangree, J. B., and J. M. Widmier, 1977, Seismic stratigraphy and global changes in sea level, part 9: seismic interpretation of clastic depositional facies, in C. E. Payton, ed., Seismic stratigraphy—applications to hydrocarbon exploration: AAPG Memoir 26, p. 165–184.

Sanyal, S. K., J. B. Gardner, J. B. Koenig, and J. McIntyre, 1980, Well-site evaluation of logs from a geothermal well: Geothermal Resources Council Transactions, v. 4, p. 471–474.

Sclater, J. G., and P. A. F. Christie, 1980, Continental stretching: an explanation of the post-mid-Cretaceous subsidence of the central North Sea Basin: Journal of Geophysical Research, v. 85, p. 3711–3739.

Seitz, H. R., 1976, Suspended- and bedload-sediment transport in the Snake and Clearwater rivers in the vicinity of Lewiston, Idaho: Boise, Idaho, U. S. Geological Survey Fourth Annual Basic-Data Report, 70 p.

Serra, O., 1986, Fundamentals of well-log interpretation (v. 2), the interpretation of logging data: Pau, France, Elf-Aquitaine or Amsterdam, Elsevier, 684 p.

Shackleton, N. J., and N. D. Opdyke, 1977, Oxygen isotope and paleomagnetic evidence for early Northern Hemisphere glaciation: Nature, v. 270, p. 216–219.

Smith, G. R., 1975, Fishes of the Pliocene Glenns Ferry Formation, southwest Idaho: Papers on Paleontology, University of Michigan Museum of Paleontology, v. 14, (Claude W. Hibbard Memorial Volume 5), p. 1–68.

Smith, G. R., 1981, Late Cenozoic freshwater fishes of North America: Annual Reviews of Ecological Systematics, v. 12, p. 163–193.

Smith, G. R., K. Swirydczuk, P. G. Kimmel, and B. H. Wilkinson, 1982, Fish biostratigraphy of late Miocene to Pleistocene sediments of the western Snake River plain, Idaho, in B. Bonnichsen and R. M. Breckenridge, eds., Cenozoic geology of Idaho: Idaho Geological Survey Bulletin 26, p. 519–541.

Sowers, G. F.,1979, Introductory soil mechanics and foundations: New York, MacMillan Publishing Co., 621 p.

Stanley, K. O., and R. C. Surdam, 1978, Sedimentation on the front of Eocene Gilbert-type deltas, Washakie basin, Wyoming: Journal of Sedimentary Petrology, v. 48, p. 557–573.

Sturm, M., and A. Matter, 1978, Turbidites and varves in Lake Brienz (Switzerland): deposition of clastic detritus by density currents: International Association of Sedimentologists Special Publication 2, p. 147–168.

Swain, F. M., and R. W. Meader, 1958, Bottom sediments of southern part of Pyramid Lake, Nevada: Journal of Sedimentary Petrology, v. 28, p. 286–297.

Swirydczuk, K., B. H., Wilkinson, and G. R. Smith, 1980, Pliocene Glenns Ferry oolite lake margin carbonate deposition in southwestern Snake River plain—reply: Journal of Sedimentary Petrology, v. 50, p. 999–1001.

Taylor, D. W., 1960, Distribution of the freshwater clam Pisidium ultramontanum: a zoogeographic inquiry: American Journal of Science, v. 258-A, (Bradley volume), p. 325–334.

Taylor, D. W., 1985, Evolution of freshwater drainages and molluscs in western North America, in C. J. Smiley, A. E. Leviton, and M. Berson, eds., Late Cenozoic history of the Pacific Northwest: interdisciplinary studies of the Clarkia fossil beds of northern Idaho: San Francisco, American Association for the Advancement of Science Pacific Division, p. 265–321.

Wheeler, H. E., and E. F. Cook, 1954, Structural and stratigraphic significance of the Snake River capture, Idaho-Oregon: Journal of Geology, v. 62, p. 525–536.

Wood, S. H., 1989a, Silicic volcanic rocks and structure of the western Mount Bennett Hills and adjacent Snake River plain, Idaho, in R. P. Smith, and K. L. Ruebelmann, eds., Yellowstone-Snake River plain volcanic province: International Geological Congress Guidebook T-305, Washington D. C., American Geophysical Union, p. 69–77.

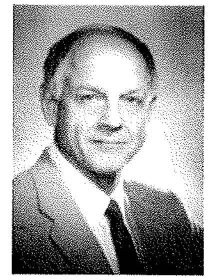
Wood, S. H., 1989b, Quaternary basalt vent-and-fissure alignments, principal stress directions, and extension in the Snake River plain (abs.): Geological Society of America Abstracts with Programs 21, p. 161.

Wood, S. H., and J. E. Anderson, 1981, Geology, in J. C. Mitchell, ed., Geothermal investigations in Idaho, part 11, geological, hydrological, geochemical, and geophysical investigations of the Nampa-Caldwell and adjacent areas, southwestern Idaho: Idaho Department of Water Resources Water Information Bulletin 30, p. 9–31.

Wood, S. H., and W. L. Burnham, 1987, Geologic framework of the Boise Warm Springs geothermal area, Idaho, in S. S. Buess, ed., Geological Society of America Centennial Field Guide, v. 2, p. 117–122.

ABOUT THE AUTHOR

Spencer Wood



Spencer Wood received a geophysical engineering degree from the Colorado School of Mines (Golden, Colorado) in 1964, and worked for Mobil Oil Corporation in Netherlands and Libya until 1968. He completed an M.S. in geophysics and a Ph.D. (1975) in geology at the California Institute of Technology (Pasadena, California), and subsequently taught at Occidental College (Los Angeles, California) and the University of Oregon (Eugene, Oregon). From 1976–1977 he worked on neotectonics for the U. S. Geological Survey National Center for Earthquake Research. Since 1978 he has been on the faculty of Boise State University (Boise, Idaho), where he is a professor conducting research on regional geology of the northwestern United States, earthquakes, volcanic rocks, hydrogeology, and borehole geophysics.

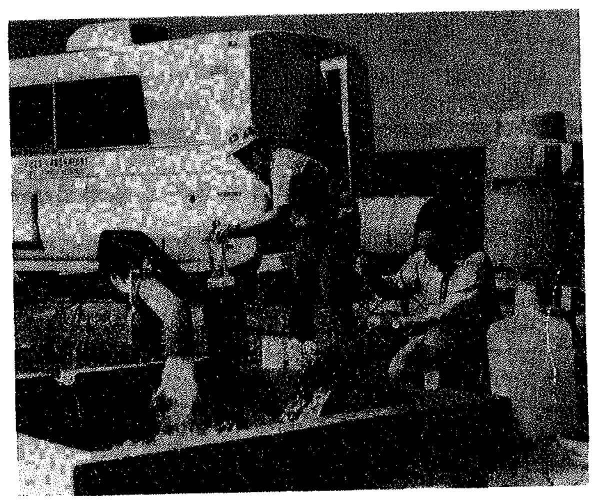
S. Wood
Last Copy.

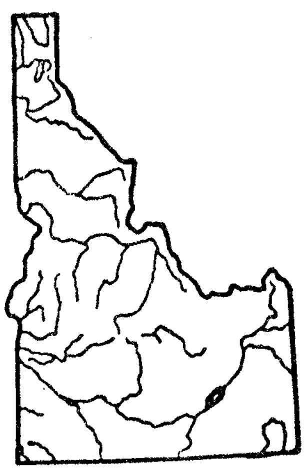
GEOTHERMAL INVESTIGATIONS IN IDAHO

Part 11

GEOLOGICAL, HYDROLOGICAL, GEOCHEMICAL AND GEOPHYSICAL INVESTIGATIONS OF THE NAMPA-CALDWELL AND ADJACENT AREAS, SOUTHWESTERN IDAHO

Well 2N-3W-27bab1, south of Lake Lowell, being sampled for chemical and isotopic constituents. Photo-Alan Mayo





IDAHO DEPARTMENT OF WATER RESOURCES
WATER INFORMATION BULLETIN NO. 30
DECEMBER 1981

WATER INFORMATION BULLETIN NO. 30
GEOTHERMAL INVESTIGATIONS IN IDAHO
PART 11

Geological, Hydrological, Geochemical and Geophysical
Investigations of the
Nampa-Caldwell and Adjacent Areas
Southwestern Idaho

John C. Mitchell
EDITOR

Work performed under U.S. Department of Energy Contract
No. DE-AS07-77ET28407
Modification No. A003
Designation DOE/ET/01834-3
Idaho Department of Water Resources
Statehouse
Boise, Idaho

December 1981

TABLE 1-1
Conversion Factors

To Convert from	To	Multiply by
	DISTANCE	
inches (in)	centimeters (cm)	2.540
feet (ft)	meters (m)	0.305
yards (yd)	meters (m)	0.914
miles (mi)	kilometers (km)	1.609
centimeters (cm)	inches (in)	0.394
meters (m)	feet (ft)	3.281
meters (m)	yards (yd)	1.094
kilometers (km)	miles (mi)	0.621
	AREA	
square miles (sq mi)	square kilometers (sq km)	2.589
sq kilometers (sq km)	square miles (sq mi)	0.386
	VOLUME — MASS	
gallons (gal)	liters (l)	3.785
ounces (oz)	grams (gm)	28.349
liters (l)	gallons (gal)	0.264
grams (gm)	ounces (oz)	0.035
	ENERGY	
British thermal units (BTU)	calories (cal)	1.996
British thermal units (BTU)	joules (j)	1054.35
calories (cal)	British thermal units (BTU)	0.004
calories (cal)	joules (j)	4.186
joules (j)	British thermal units (BTU)	0.0009
joules (j)	calories (cal)	0.239
	GRADIENTS	
degrees Fahrenheit/100 ft (°F/100 ft)	degrees Celcius/kilometers (°C/km)	1.822
degrees Celcius/kilometer (°C/km)	degrees Fahrenheit/100 ft (°F/100 ft)	0.055
	THERMAL CONDUCTIVITY	
$\frac{\text{millicalories}}{\text{centimeter/second °Celcius}}$ $\frac{\text{(mcal)}}{\text{(cm/sec °C)}}$	$\frac{\text{calories}}{\text{centimeter/second °Celcius}}$ $\frac{\text{(cal)}}{\text{(cm/sec °C)}}$	0.001
$\frac{\text{calories}}{\text{centimeter/second °Celcius}}$ $\frac{\text{(ucal)}}{\text{(cm/sec °C)}}$	$\frac{\text{millicalories}}{\text{centimeter/second °Celcius}}$ $\frac{\text{(mcal)}}{\text{(cm/sec °C)}}$	1000
	HEAT FLOW	
$\frac{\text{microcalories}}{\text{centimeter}^2\text{ second}}$ $\frac{\text{(ucal)}}{\text{(cm}^2\text{ sec)}}$	milliwatts/meter^2 (mwatt/m^2)	41.871
milliwatts/meter^2 (mwatt/m^2)	$\frac{\text{microcalories}}{\text{centimeter}^2\text{ second}}$ $\frac{\text{(m)}}{\text{(cm}^2\text{ sec)}}$	.024

CHAPTER 2 - GEOLOGY

By
Spencer H. Wood[1]
and
John E. Anderson[2]

INTRODUCTION

The cities of Nampa and Caldwell lie within the western Snake River Plain. The plain is a northwest-trending physiographic lowland and contains the incised valleys of two major rivers flowing to the northwest. The Snake River rises at the far eastern end of the plain and flows nearly 700 km (435 mi) across the arcuate plain until it turns north and descends through Hells Canyon and into the Columbia. The Boise River heads in the granitic mountains north of the plain and then flows northwestward until it joins the Snake.

The plain is also a great structural basin formed by downwarping and faulting in late Cenozoic time (Figure 2-1). The amount of late Cenozoic vertical movement in the western plain is impressive. The basin-fill sediments and volcanic flows have been drilled to a depth of 4.3 km (2.7 mi), and Mabey (1976) has suggested the total basin fill may be 7 km (4.4 mi) or more.

A variety of ideas and models have recently been advanced since Armstrong and others (1975) published the K-Ar geochronology of silicic volcanics of the plain. Their study shows that the inception of silicic volcanism becomes progressively younger eastward across the plain; and that basalt volcanism persists for a long time after the inception of silicic volcanism. Brott and others (1978) show evidence for increasing values of heat flow eastward across the plain and emphasizes the significance of the increase in elevation eastward across the plain. These features of the plain are modelled by an eastward propagating thermal event which passed through the western plain 8 to 16 million years ago and is currently manifested as dormant volcanism and as the hydrothermal systems of the Yellowstone area.

The model of an eastward propagating thermal event coincident with the eastward development of the plain is reasonably consistent with available data. A similar, but diametrically opposite, pattern of silicic volcanism is recognized as having propagated westward across eastern Oregon also from the area of the western plain (MacLeod and others, 1975), but this system was not accompanied by the formation of a deep structural basin.

The present western Snake River Plain has the appearance of a northwest-trending graben 50 km (31 mi) wide. The western plain differs from the plain to the east in that the upper 700 to 1700 m (435 to 1056 ft) of basin fill is mostly sediment, whereas in the eastern plain the upper section is mostly Quaternary basalt. Although basalt volcanism has been active during the Quaternary of the western plain, it is not nearly so voluminous as in areas to the east.

[1] Department of Geology and Physics, Boise State University.
[2] Idaho Department of Water Resources.

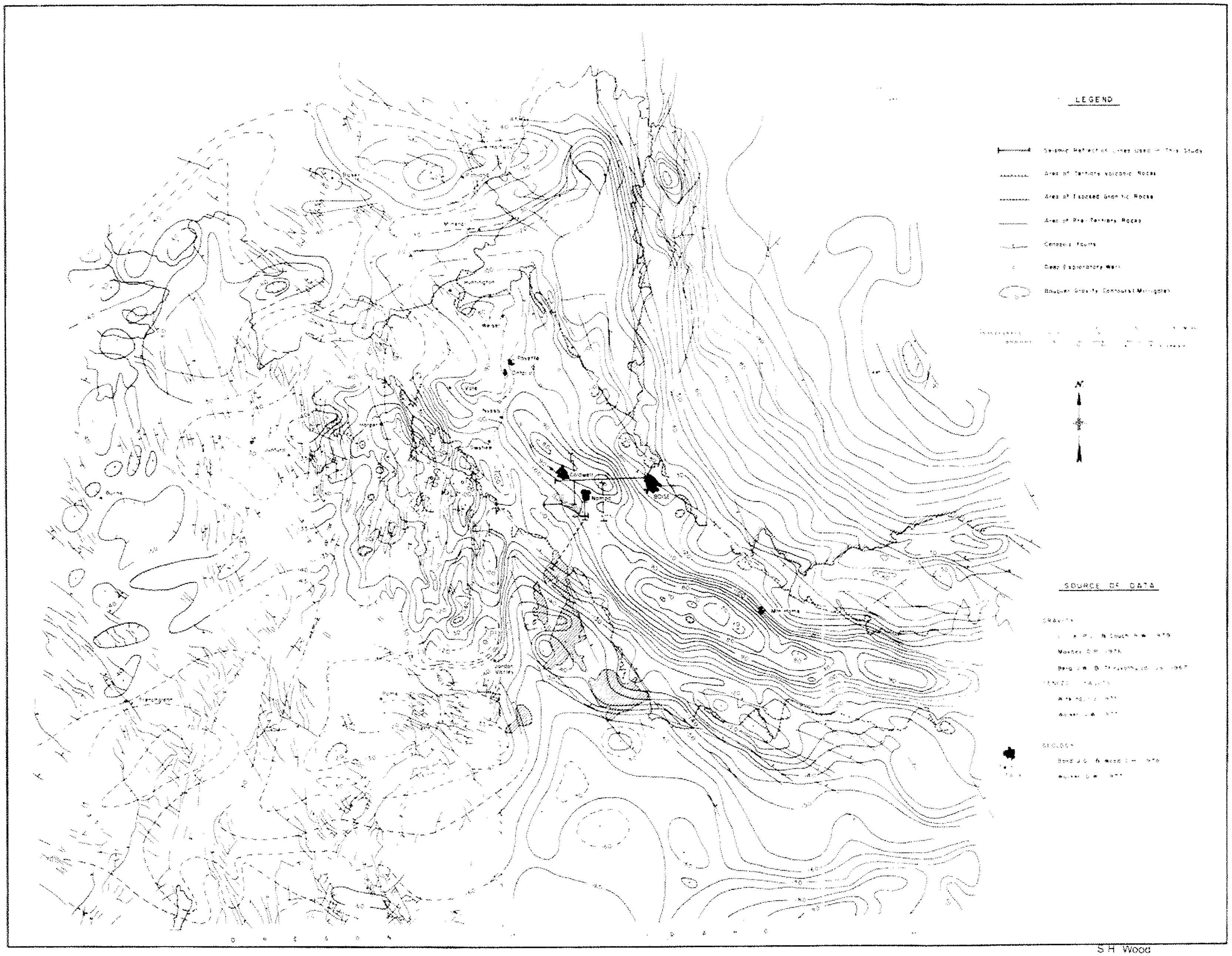


FIGURE 2-1. Map showing generalized geology, major faults, and Bouguer gravity contours in southwestern and west-central Idaho and eastern Oregon.

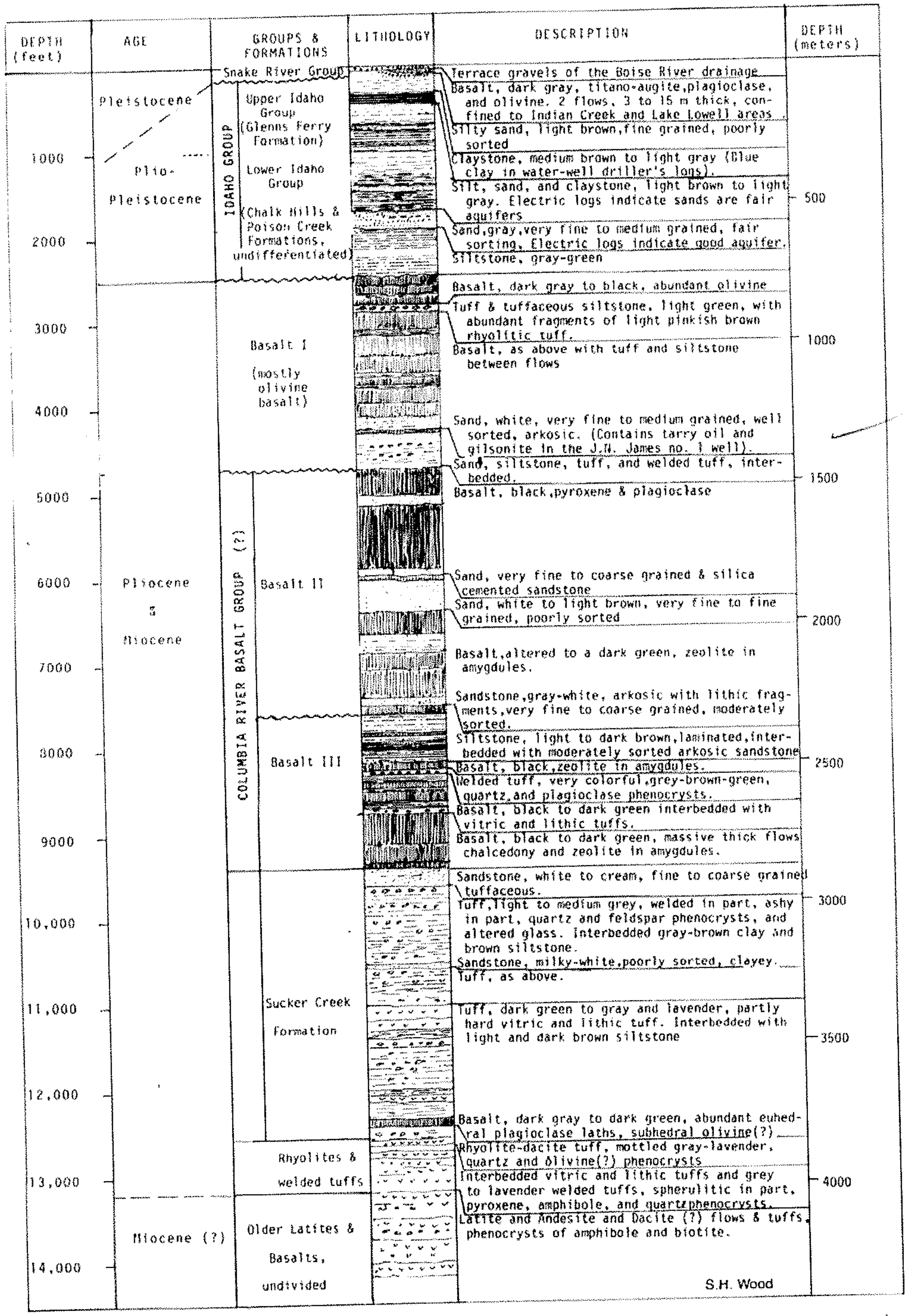
The main basis for this geological report of the Nampa-Caldwell area is a study of available well logs from deep wells in the area (fig. 2-2, and 2-3) and seismic reflection profiles (figure 2-4 and 2-5). Field mapping at a 1:24,000 scale was completed for all of the Nampa, Middleton, Caldwell and Lake Lowell quadrangles and a part of the Melba, Walters Butte and Given's Hot Springs quadrangles (figure 2-6, shown at 1:62,500 scale in pocket). Little geologic mapping had previously been done in the relatively featureless and flat part of the plain. Previous work by Lindgren and Drake (1904) and Savage (1958) did not recognize the extensive faulting of the Quaternary Gravels. Mapping utilized the limited exposures in sand and gravel pits, road cuts and river banks. Mapping of the Snake River Group basalts in the area relied heavily upon well logs filed by water-well drillers. The recent soil survey of Canyon County (Priest and others, 1972), and various terraces and topographic surfaces were also helpful in identifying the surficial deposits. In order to understand the subsurface stratigraphy a limited amount of detailed mapping was done of exposed strata on the margins of the plain.

The following section of this report discusses the stratigraphy of subsurface rocks and of the exposed surficial units. Exposures are quite limited, but localities are discussed in sufficient detail to allow examination of type sections of the mapping units. The structural geology of the Nampa-Caldwell area is discussed more fully in the geophysics chapter.

GENERAL STATEMENT

The thickness of Cenozoic volcanic and sedimentary fill in the western Snake River Plain is at least 4.3 km (2.7 mi) as demonstrated by the section penetrated by the exploratory well drilled near Meridian, Idaho (J.N. James No. 1). Basement rock was not penetrated. Because granitic rocks of the Idaho Batholith and associated metamorphic rocks occur in the mountains north and south of the plain (Taubeneck, 1971; Ekren and others, 1978a, and figure 2-1, it is commonly supposed that these rocks comprise basement beneath the plain. Although Challis volcanics have not been encountered in deep wells, it is possible that Challis volcanics and associated sediments lie deep in the subsurface, beneath the plain. Numerous, northeast-striking, porphyry dikes of intermediate to silicic composition invade the granitic rocks along the north margin of the plain. These porphyries have been dated at 40 to 42 $\pm$ 10 million years by the Lead-alpha method in the Lowman area (Jaffe and others, 1959 quoted in Armstrong, 1974) and may have been feeder dikes to Challis volcanism. The Challis volcanic sequence is at least 1,200 m (3900 ft) thick in the central Idaho Mountains and may be as thick as 3,200 m (10,500 ft) (Cater, et al., 1973). Challis-aged volcanics are also mapped in the eastern Owyhee Mountains by Ekren and others (1981) where they may be greater than 300 m (985 ft) thick.

In this report, the subsurface stratigraphy (figure 2-2) expected in the Nampa-Caldwell area is based upon the lithologic and geophysical logs of the 4.3 km deep Meridian well (J. N. James No. 1) and a limited amount of information from the Champlain Oil Company (Deer Flat No. 1) well recently drilled and abandoned south of Nampa. Interpretation of seismic reflection profiles discussed in Chapter 6 provides considerable stratigraphic information and shows that the section in the J. N. James No. 1 well may not be entirely representative of the western Snake River plain because it is drilled on a structural high with at least 700 m (2100 ft) of relief (figure 2-4). Some rock units similar to those encountered in the well are described from exposures at a number of localities on the margins of the plain by Malde and Powers (1962), Kittleman and others (1965), and Ekren and others (1981). Correlation of subsurface units to exposed sections can be made for some distinctive lithologic



The generalized stratigraphy of Cenozoic rocks beneath the Nampa-Caldwell-Meridian, Idaho, area is based upon the lithologic log of the M.T. Halbouty - Chevron, U.S.A. exploratory well near Meridian (Sec. 27, T4N, R1W), "J.N. James #1." Rock descriptions are from the cuttings log compiled by Hiner (1976): it is not known if any of the volcanic rock descriptions on the log were verified by petrographic examination. Surficial rock and sediment of the Snake River Group in the uppermost 50 m (166 ft) are shown for completeness, but the Snake River Basalts do not occur at the well. Basalts within the Idaho Group do not occur at this well and are not shown, but they are encountered in the Champlain Oil Co., Deer Flat #1, 22 km (6.7 mi) to the southwest. Stratigraphic units are defined by Malde and others (1963), and Ekren and others (1978a). Stratigraphic units reported in the operator's records of this well do not exactly agree with the interpretation here. Correlation of units in this well to exposed rocks on the margin of the plain are discussed in the text.

FIGURE 2-2. Generalized stratigraphy of Cenozoic rocks from J.N. James No. 1 well near Meridian, Idaho.

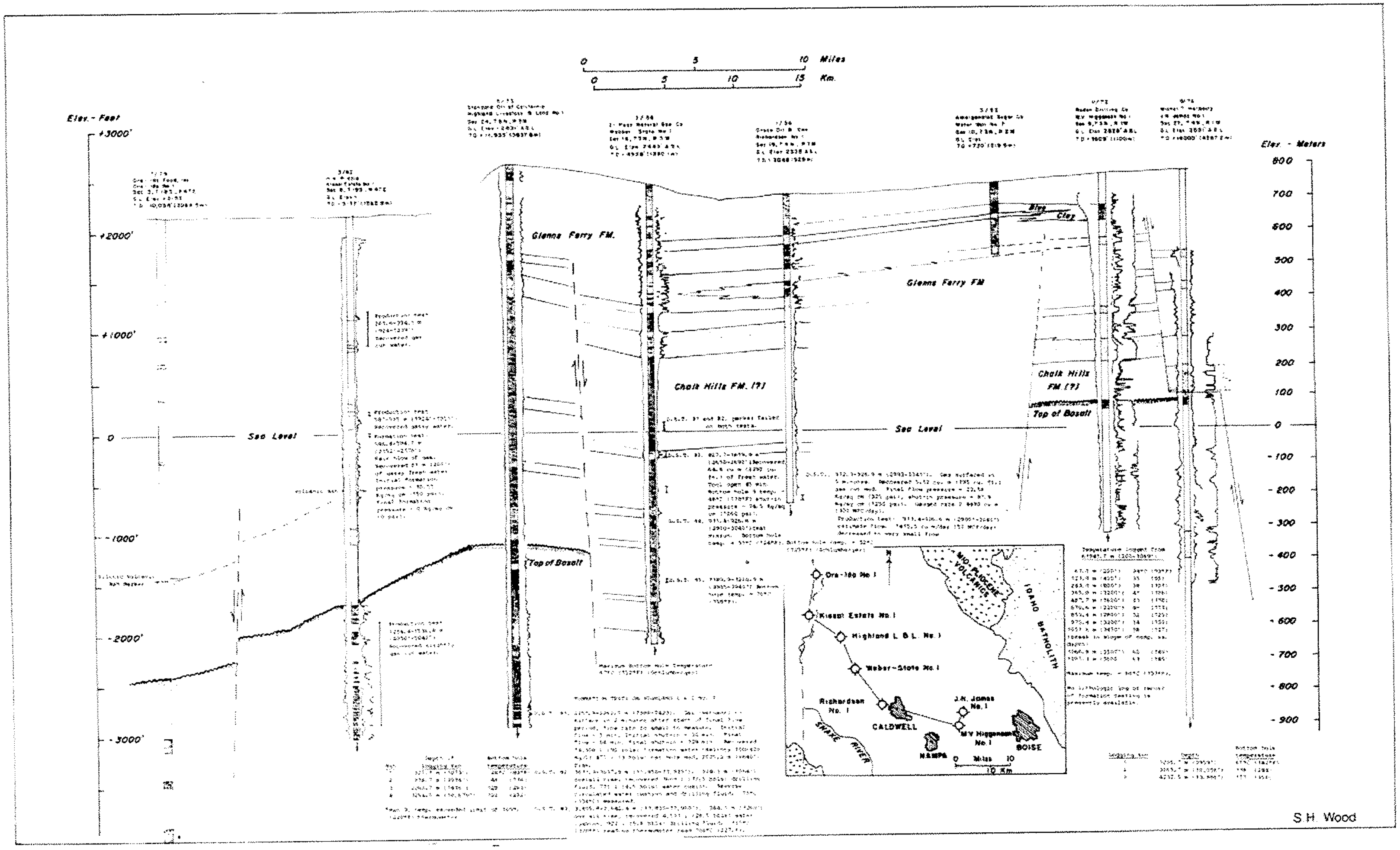
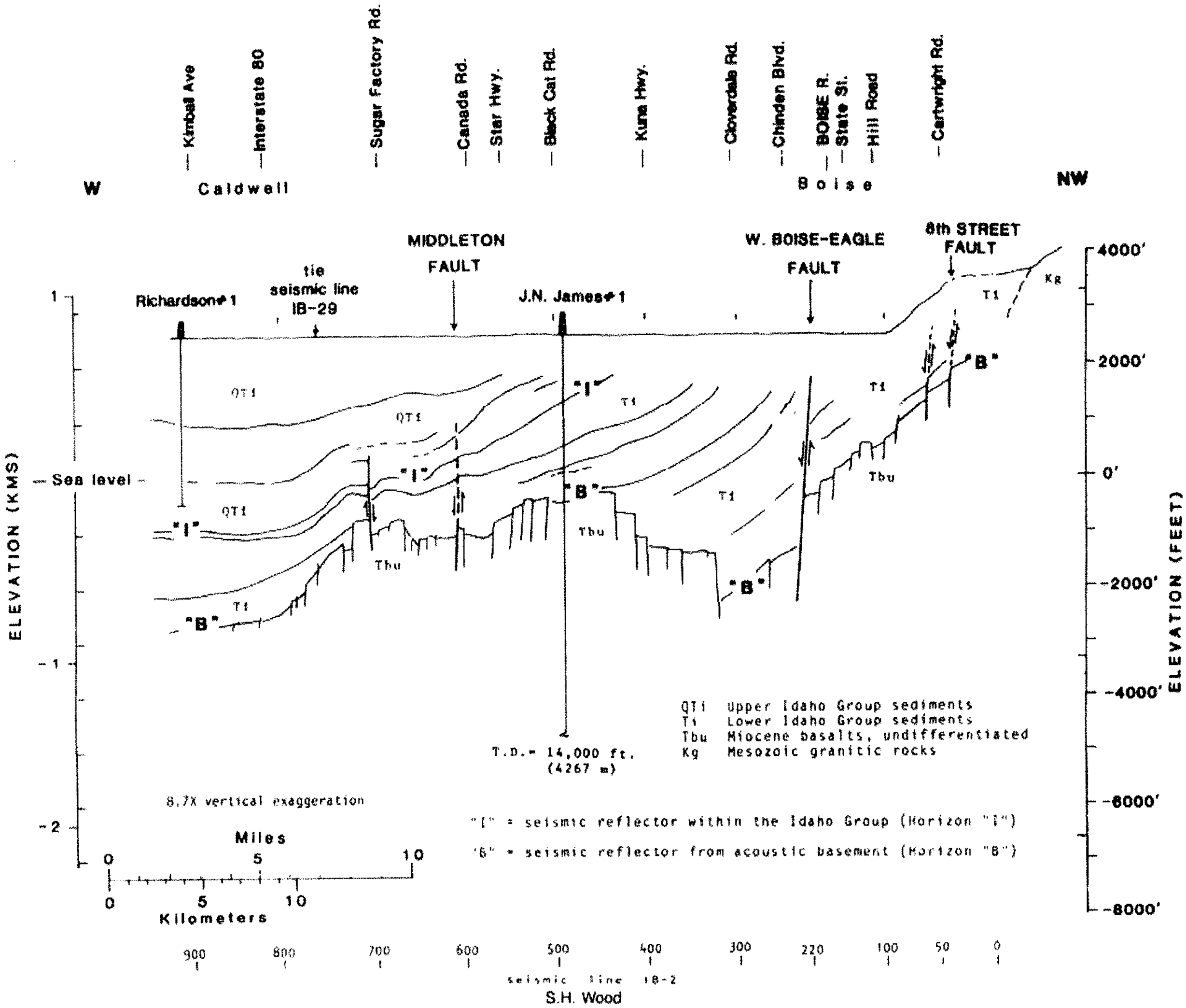
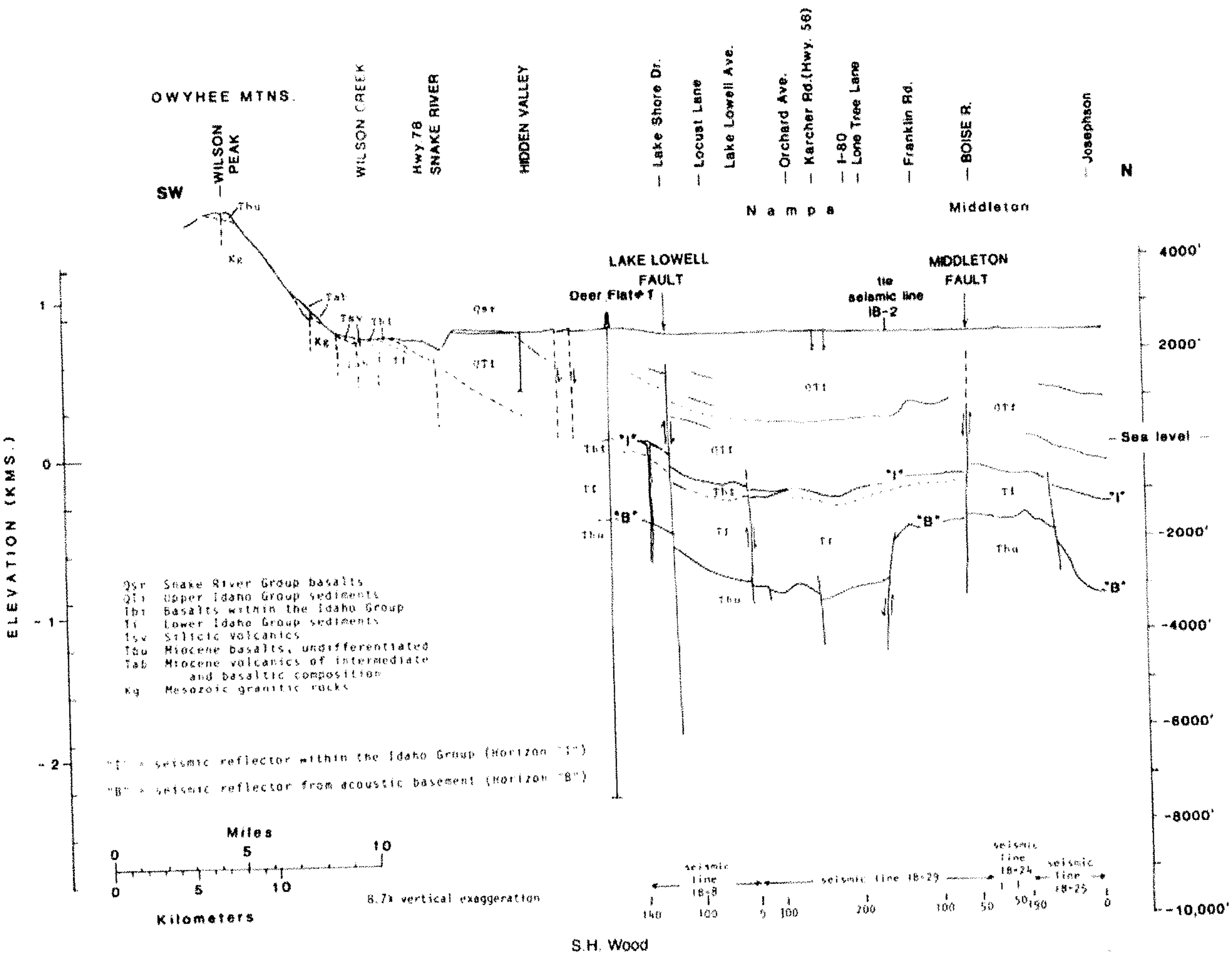


FIGURE 2-3. Stratigraphic cross section of late Cenozoic deposits, western Snake River Plain, Idaho, with electric log correlations, temperature and formation testing data



Structural cross-section prepared from seismic reflection profile IB-2. Line starts in the Boise foothills and runs down Stuart Gulch, v.p. 1-110, and then northwest along Hill Road, v.p. 110-135, along Castle Road, v.p. 135-180, across the Boise River at the Strawberry Glen Bridge, and then westward along Elm Lane to Caldwell, v.p. 260-1025. Location of section is shown on figure 2-1.

FIGURE 2-4. East-west geologic cross-section along seismic reflection line IB-2.



Structural cross-section prepared from seismic reflection profiles IB-25, IB-24, IB-29, and IB-8. Seismic line begins 5.5 mi north of Middleton and continues south along Cemetery Road to Middleton, IB-25, v.p. 1-190, and then west 1.6 km (1 mi) along Highway 44, IB-24, v.p. 75-25, and then south on Middleton Road to the edge of Lake Lowell, line IB 29, v.p. 25-134, and from the south city limits of Nampa south along Highway 45, line IB8 - v.p. 1-140. Continuation of section south to the Owyhee Mountains is based upon preliminary results from the Champlain Oil Deer Flat No. 1 well, unpublished geologic mapping by S.H. Wood, and from Ekren, McIntrye, and Bennett (1978). Location of section is shown on Figure 2-1.

FIGURE 2-5. North-south geologic cross-section along seismic-reflection lines IB-25, IB-29, and IB-8.

units, but a number of units such as the Banbury Basalts, Columbia River Group(?), the Sucker Creek Formation and Owyhee Basalts do not have obvious correlative units exposed on the margin. No detailed stratigraphic work or geologic mapping has been published for the northeast margin of the plain. It is futile to attempt correlations to units that were only loosely defined in mapping by Lindgren (1898), Lindgren and Drake (1904), and Savage (1958 and 1961). Type sections and good stratigraphic descriptions have never been published for the Payette Formation, the Columbia River Basalts and the Idaho Group in this area. Many of the late Cenozoic fluvial, lacustrine and volcanic units are very similar in appearance and correlations have been made on the basis of similar lithology rather than detailed geologic mapping. Correlation is further complicated by numerous facies changes within sedimentary units such as described by Malde and Powers (1962), Malde (1972) and Kimmel (1979) in the Glenns Ferry Formation in the region east of this study area.

The lithologic log from the J.N. James No. 1 well (figure 2-2), the correlation cross section of wells in the western plain (figure 2-3) and cross sections prepared from seismic reflection profiles (figures 2-4 and 2-5) show the stratigraphy as it is presently understood in the Nampa-Caldwell area. Knowledge of rocks older than the massive basalt flows that constitute acoustic basement of seismic reflections is entirely from the J.N. James No. 1 well, and from discussions of exposed stratigraphy by Malde and Powers (1962), Newton and Corcoran (1963), Ekren and others (1978a), Armstrong and others (1976) and Armstrong and others (1980).

This discussion of subsurface stratigraphy should be prefaced by a statement that matching subsurface rock units in the center of the plain to published stratigraphic sequences of exposed rocks on the margins of the plain could not be made, in most cases, with available data. The greatest difficulty in matching the stratigraphy advanced by Armstrong and others (1980) is the lack of a clearly identifiable Idavada Group of silicic volcanics at intermediate depth in the well logs of the J. N. James No. 1 well drilled near Meridian, Idaho. Exposures and deep wells on both the north and south margins of the Snake River Plain strongly suggest that the Idavada Group of silicic volcanics should be at least 330 m (1,000 ft) thick beneath the Plain (Arney and others, 1980; Wood and Vincent, 1980; McIntyre, 1979; Ekren and others, 1978a; Malde and Powers, 1962). A section of basaltic volcanics is encountered from 720 m to 2,895 m (2,360 to 9,500 ft) in the J.N. James well; however, only one 20 m (70 ft) thick unit of rhyolitic tuff is reported in that section (Hiner, 1976). The well bottoms in a thick unit of silicic and intermediate volcanics extending from 3,840 m (12,600 ft) to total depth of 4,267 m (14,000 ft), but these seem too deep to be correlative with the Idavada Group. Acoustic basement is a thick massive basalt section overlain by a number of thinner basalt flows interbedded with sediment. No geochemical data or detailed petrographic studies upon which to subdivide the subsurface basalts are available. Lack of an obvious Idavada Group within the subsurface volcanic section makes it impossible at this time to correlate the section under the Plain with the stratigraphic section of Armstrong and others (1980). For instance, it is important to know if the uppermost basalts in the J.N.James No. 1 well are correlative with the Pliocene Banbury Group defined by Malde and Powers (1962) or if they are the Miocene basalts coequal with the Miocene-to-Pliocene aged Columbia River Group. These two groups of basalt should be separated by the Idavada Group. These questions will not be answered until the subsurface samples are given the same detailed geochemical and petrographic examination as the better studied surface exposures.

STRATIGRAPHY

Older Volcanics of Intermediate Composition

The J.N. James No. 1 well bottomed at 4,267 m (14,000 ft) in 240 m (800 ft) of volcanics variously described as latite, dacite(?), andesite, vitric and lithic tuffs, and minor basalt. Some of the felsic rock bears phenocrysts of amphibole and biotite. This section resembles the Miocene or older rocks described by Ekren and others (1981) in parts of the Owyhee Mountains. Volcanic rocks of intermediate composition are in the Reynolds and Salmon Creek drainages. Dikes bearing amphibole and biotite phenocrysts occur in the War Eagle Mountain area (Pansze, 1973). These rocks could also be Challis-aged volcanics mapped on the south side of the plain by Ekren and others (1981) and on the north side by Malde and others (1963).

Older Rhyolites and Welded Tuffs

The 195 m (640 ft) section from 3,828 m (12,560 ft) to 4,023 m (13,200 ft) contains a group of rhyolitic rocks described as gray-to-lavender welded tuffs that contain spherulites. Other rocks of the group are lithic and vitric tuffs. According to Hiner (1976) some of the rocks bear phenocrysts of pyroxene, amphibole and quartz. Because this unit is overlain by sediments that are considered the Sucker Creek Formation by Warner (1977), these rhyolites in the well are probably related to the Miocene "gold bearing rhyolites" of Malde and Powers (1962). A number of rhyolites underlying the Sucker Creek Formation are mapped in the Owyhee Mountains by Ekren and others (1981). In particular, the Silver City Rhyolites of Bennett and Galbraith, (1975) locally exceed 200 m in thicknesses. K-Ar dating on the Silver City Rhyolites yielded ages of 15.6 to 15.7 + 0.4 million years (Pansze, 1973).

Sucker Creek Formation

A unit of 815 m (2,760 ft) of monotonous brown claystone and siltstone with interbedded welded and non-welded ashy tuffs, and one basalt flow near the base of the unit is encountered from 2,865 m (9,400 ft) to 3,828 m (12,560 ft). Descriptions of the cuttings and the overall character of the unit, and the stratigraphic position are very similar to the type section of the Sucker Creek Formation of Kittleman and others (1965), and Ekren and others (1978a) along the Oregon-Idaho border.

Along the northern margin of the western Snake River Plain, much of the area mapped as Payette Formation by Lindgren (1898, p. 632), Kirkham (1931a or b), and Savage (1958 and 1961) may contain strata correlative with the Sucker Creek Formation; however, mapping and stratigraphic correlations necessary to establish continuity or contemporaniety of the formations has not been attempted to date. Kirkham (1931a or b) defined the Payette Formation as the silty and clayey sediments that interfinger with basalts of the Columbia River Group in the vicinity of Horseshoe Bend and Weiser. McIntyre (1976) found that much of the sediment in the Weiser area identified as Payette Formation by Kirkham (1931a, or b) was, in fact, overlying the main volcanic sequence in this region. A basaltic andesite dike cutting the Weiser area sediments yields a whole rock K-Ar age of 10 $\pm$ 0.6 million years indicating a late Miocene or older age for the sediments (McIntyre, 1976).

Age of the Sucker Creek Formation is late Miocene based upon Barstovian mammalian fossils collected near the type locality and middle Miocene based on flora of Mascall age (Kittleman, et al., 1965, p. 7). A basalt within the Sucker Creek Formation yielded a potassium-argon age of 16.7 million years (Evernden and James, 1964, p. 971). The above age assignments from the Owyhee Canyon area are consistent with four potassium-argon ages on the overlying Owyhee Basalts (Bryan, 1929) obtained by Watkins and Baksi (1974, p. 173) that ranged from 13.1 to 13.6 million years.

Basalt Section in the J.N. James No. 1 Well

Well logs indicate a 2,100 m (7,000 ft) thick section of basalt with lesser amounts of interbedded arkosic and tuffaceous sediments overlie the Sucker Creek Formation. This basalt section is entirely penetrated by J.N. James No. 1 well. Basalt is first encountered at a depth of 728 m (2,390 ft) and the basalt dominated section extends to a depth of 2,865 m (9,400 ft). Deep wells west of the study area, Highland Land and Livestock No. 1 (T.D. = 3,638 m [11,935 ft]) and Ore-Ida No. 1 (T.D. = 3,064 m [10,054 ft]) encounter a greater proportion of fine-grained sediment, much less basalt, and generally thinner basalt units (figure 2-2). Thick flows within the upper 100 to 150 m (300 to 500 ft) of this basaltic section act as the acoustic basement for seismic reflection profiling over most of the western plain.

The following breakdown of the 2,100 m (7,000 ft) section of basalt and sediments is based entirely upon the lithologic log description of Hiner (1976) and geophysical logs of the J.N. James No. 1 well where basalt is first encountered at 728 m (2,390 ft) depth.

BASALT UNIT I — The lowermost unit of interbedded sediment and basalt extends from 2,322 m (7,620 ft) to 2,865 m (9,400 ft), a thickness of 542 m (1,780 ft). The lower 189m (620 ft) is made up of at least two thick basalt flows, described as black to dark green with zeolite and calcite amygdules. These thick flows rest upon Sucker Creek Formation sediments. The upper part of this unit is mostly brown siltstone with minor arkosic sandstone interbedded with thin basalt flows. Chevron geologists considered this unit to be a part of the Sucker Creek Formation (unpublished records on file with the Idaho State Petroleum Engineer, see Hiner, 1976).

BASALT UNIT II — A 884 m (2,900 ft) thick unit of basalt, extends from 1,438 m (4,720 ft) to 2,322 m (7,620 ft). This unit consists of about 60 percent thick flows of pyroxene-plagioclase phyric basalt and about 40 percent moderately-to-poorly-sorted, fine to coarse grained, arkosic sandstone. Absence of visible olivine in the flows is inferred because it is not mentioned in the well log (Hiner, 1976). Identification of such phenocrysts in basalt on field lithologic logs must be regarded with caution for it is unlikely that the drill cuttings were viewed in thin section with a petrographic microscope.

BASALT UNIT III — The uppermost basalt unit from 728 m (2,390 ft) to 1,438 m (4,720 ft) is composed of 520 m (1,700 ft) of dark grey to black, olivine basalt with minor interbeds of arkosic white sand, welded and non-welded tuff, and siltstone and claystone. The lithologic log describes cuttings of a light-pinkish-brown rhyolite tuff from the upper part of this unit. Such a distinctive rock may serve as a marker in other wells. It is also possible that this tuff is the lone representative of the Idavada Group in this area. The lower part of the unit contains 61 m (200 ft) of well sorted sand indicating the possibility of sand aquifers within the basalt. This sand also contains a tarry-black hydrocarbon-like material (Hiner, 1976). Thick flows within this unit constitute the regional acoustic basement mapped as a reflection horizon in figure 2-7 and discussed in Chapter 6.

Basalts in Adjacent Areas

The basalt stratigraphy in this region of Idaho is complex. No correlative subsurface marker beds have thus far been established. Age of the basalts are not known, although it is likely that much of this basalt section is Miocene to early Pliocene in age, and is related to the basalts of the Columbia River Group. Recent studies of the Columbia River Group summarized in Swanson and others (1979) indicate that the Lower-to-Middle-Miocene Imnaha and the widespread Grande Rhonde Basalt Formations extend into the Weiser embayment of the Columbia River Basalt Plateau. These basalt units are down faulted to the west (Newcomb, 1970), and dip to the west and southwest beneath sediments of the lower Idaho Group. They presumably underlie a part of the western Snake River Plain. Feeder dikes of these two oldest formations of the Columbia River Group comprise the majority of dikes of the north-northwest striking Chief Joseph dike swarm of northeastern Oregon and western Idaho (Taubeneck, 1970 and Swanson and others, 1979). Most flows of the Imnaha basalt are coarse grained and plagioclase phyric. Zeolite amygdules and smectite alteration are common and widespread. Imnaha basalt flows are known only in the southeastern part of the Columbia Plateau, on both sides of the Blue Mountain-Seven Devils uplift. Maximum thickness of the Imnaha basalt is reported to be about 500 m (1600 ft) in Hells Canyon (Vallier and Hooper, 1976; and Hooper, 1981).

The Grande Ronde Basalt is now recognized as the most widespread formation of the Columbia River Group and exceeds 500 to 700 m (1640 to 2300 ft) thick on the north flank of the Blue Mountains (Swanson and Wright, 1981). Basalt flows of the Grande Ronde Formation are overwhelmingly aphyric to very sparsely phyric fine-grained tholeiitic basalts (Swanson and others, 1979). Most flows in the Grande Ronde contain rare plagioclase microphenocrysts and plagioclase-clinopyroxene clots visible in hand specimens. Olivine is generally absent as phenocrysts but is commonly present in the ground mass of all but the least magnesium flows.

K-Ar ages on the Grande Ronde Basalt range from 14 to 15.5 million years. The underlying Imnaha Basalt has not been extensively dated. Chemical and petrographic variations within these formations are discussed in Wright and others (1973 and 1979). Magneto-stratigraphy of the group is summarized in Swanson and others (1979).

Geochronology and magneto-stratigraphy of the Basalts of the Owyhee Ridge in eastern most Oregon (Bryan, 1929, and Kittleman and others, 1965) are discussed by Watkins and Baksi (1974). No detectable age difference was found across the 16 flows in this area. Four K-Ar age determinations ranged from 13.1 to 13.9 million years. These ages indicate the Owyhee Basalts are a younger group of flows than the recognized Columbia River Group in the area immediately north of the Snake River Plain. The Owyhee Basalts are described as fine grained with rare phenocrysts of plagioclase in a ground mass of olivine, pyroxene and feldspar, Watkins and Baksi (1974).

Previous workers have applied local formation names of basalts to distant localities without consideration of the basalt petrography, geochemistry, or careful geologic mapping. Formation names such as the Grassy Mountain Formation (Bryan 1929, Kittleman and others, 1965), the Banbury Basalt (Stearns and others, 1938, Malde and Powers, 1962, Armstrong and others, 1975 and 1980, and Ekren and others, 1978a) have led to considerable confusion when applied regionally in the western Snake River Plain. These two formations may have correlatives within the lower Idaho Group, but neither of them are expected in the

J.N. James well section. The seismic reflection profiles across the well site show a regional, complexly-faulted, acoustic basement overlain by Idaho Group sediments with a marked unconformity (figures 2-4 and 2-5). The basalts on the structural high can be considered to be older than other basalts that are interbedded and conformable with the sedimentary section of the Idaho Group that laps upon the structural high.

Idavada Volcanic Group

To be consistent with the stratigraphic system of Malde and Powers (1962), Ekren and others (1981), and Armstrong and others (1980), the Idavada Group of silicic volcanics is discussed here. However, no obvious silicic or intermediate volcanics of the Idavada Group are noted in the upper two subunits of the basalt section with the exception of the light pinkish brown fragments of a felsic tuff in the J.N. James No. 1 well encountered between 853 and 884m (2,800 and 2,900 ft). Likewise, the only silicic volcanics noted in the lithologic log of the Highland Land and Livestock No. 1 well are described as greenish gray and light brown hard felsites between 2,659 and 2,774 m (8,720 and 9,100 ft), and black obsidian(?) at 1,650 m (5,415 ft) depth. The remainder of the volcanic units in the Highland Land and Livestock No. 1 well are described as basalt. In the Ore-Idaho Foods No. 1 well, no hard felsic rocks are described, but hard tuffs are noted in the intervals 1,692-1,700 m (5,550-5,580 ft), 2,560-2,585 m (8,400-8,480 ft) and at 2,972 to 2,978 m (9,750-9,770 ft). None of the above rock units bear a resemblance to the Idavada Group described in the north side of the Plain by Malde and Powers (1962) and Malde and others (1963). Nor do any of the above seem thick enough to be correlatives with the various rhyolites and welded tuffs mapped in the Owyhee Mountains by Ekren and others (1981).

The Idavada silicic volcanics are described by Malde and Powers (1962) as being mostly silicic latite, bearing phenocrysts of andesine, clinopyroxene, hypersthene and magnetite. They are distinguished from the older "gold-bearing" silicic rocks by their lack of hornblende or biotite. The Idavada sequence is made up of both intermediate-to-silicic ash flow and lava-flow units, that commonly have prophyritic black vitric layers and thinly flow-banded lavender and gray felsite layers. Ekren and others (1978b) have suggested that some of the widespread silicic rocks were initially emplaced as very hot ash flow sheets that remelted and remobilized as lava flows. Ekren and others (1981) also indicate that some of the silicic volcanics in the Owyhee Mountains of this group may have originated from vents in the present area of the Snake River Plain.

Ages for the Idavada Volcanic Group are published by Armstrong and others (1980). Thirteen potassium-argon dates range from 14.2 and 13.5 million years for the member units of the formation near Poison Creek to 9.6 million years for the stratigraphically youngest units.

The Idaho Group

The sedimentary deposits and local basaltic vents and lava flows that unconformably overlie the rhyolite rocks (Idavada Group) on the margins of the plain are considered by Malde and Powers (1962) to be the Idaho Group. Seven overlapping formations are discussed in their redefinition of the group. The Idaho Group in the western Snake River Plain is locally overlain by a relatively thin group of fresh, unaltered basalt flows and associated fluviatile and to a lesser extent lacustrine deposits. This latter group, containing numerous unaltered, olivine basalt lavas, is called the Snake River Group.

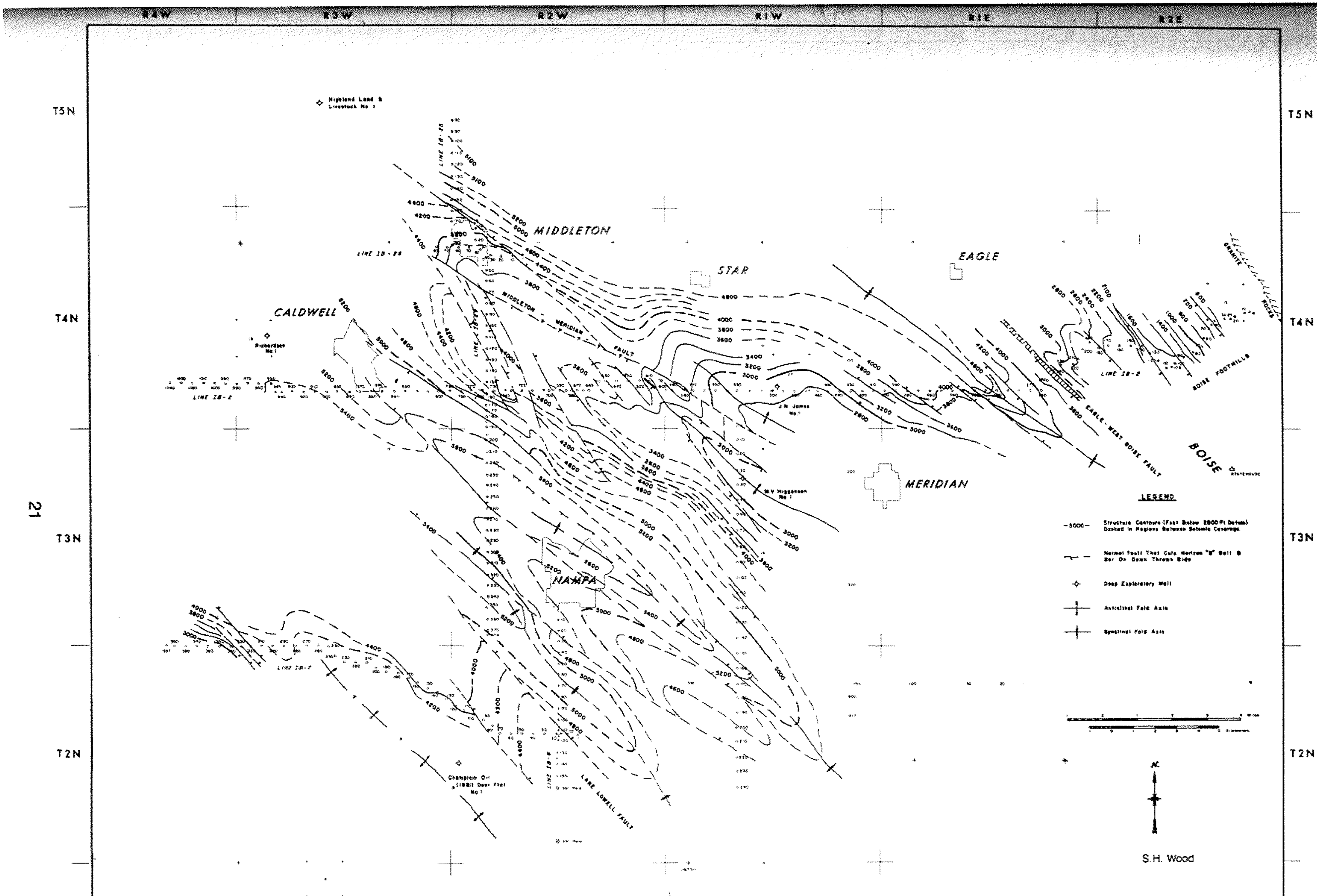


FIGURE 2-7. Structural contour map on acoustic basement (Horizon "B").

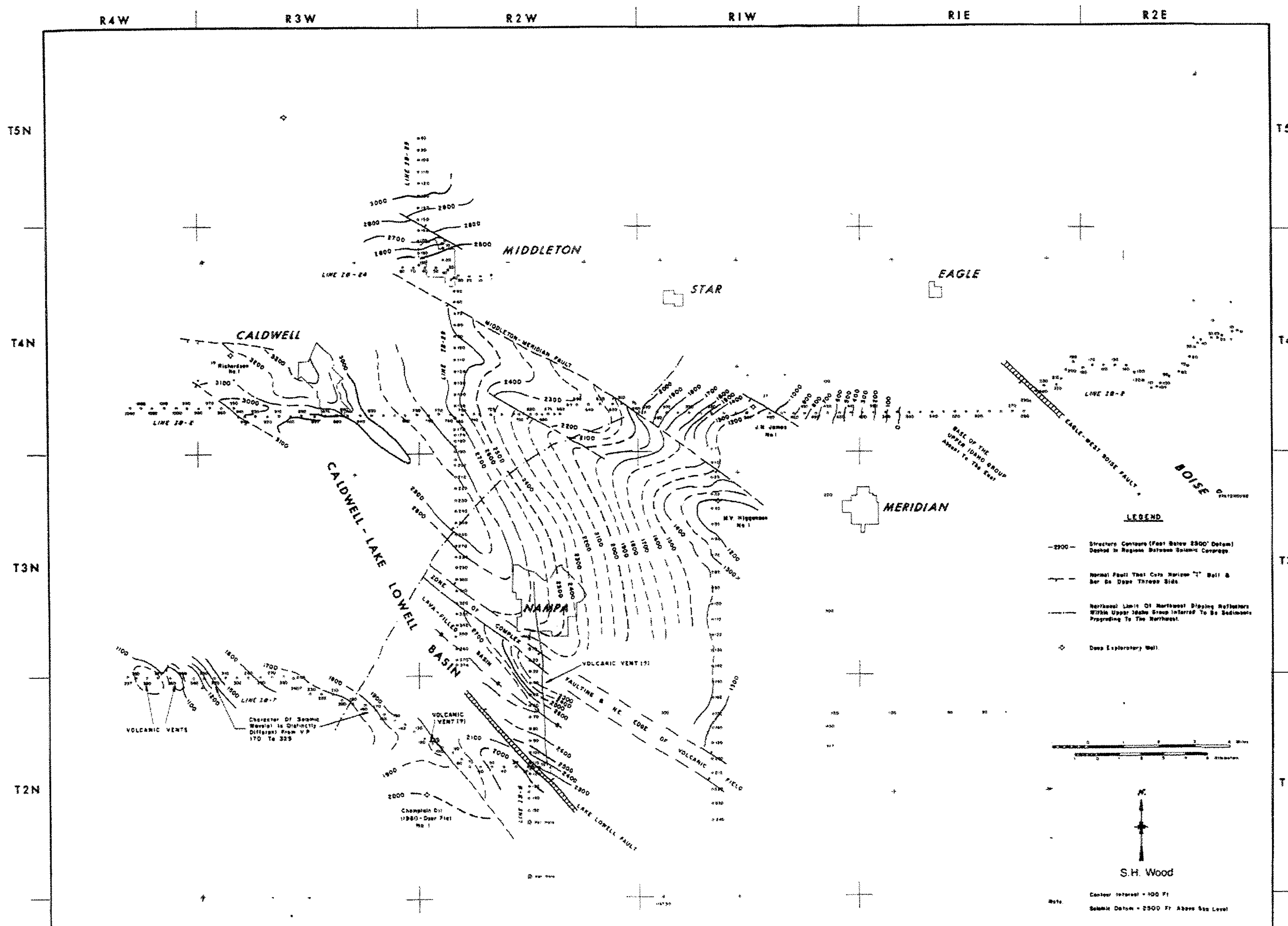


FIGURE 2-8. Structural contour map on horizon within the Idaho Group (Horizon "I").

The stratigraphic relationships within the Idaho Group beneath the western plain are complex; however, geophysical well logs (figure 2-3) and seismic reflection profiles, discussed in this study, produce a generally consistent group of subsurface units beneath the Nampa-Caldwell area. All those deposits lying between the seismically-defined "acoustic basement" and the Snake River Group are considered to be the Idaho Group in this report. Seismic reflection data shows that a maximum of 1710 m (5,600 ft) of Idaho Group sediment underlies the western part of the study area (figure 2-7). A similiar thickness is penetrated by the Ore-Ida Foods No. 1 well at Ontario (figure 2-3). The Idaho Group lies unconformably upon the faulted and eroded surface of basalts that acts as acoustic basement.

In the Nampa-Caldwell area the subsurface Idaho Group can be subdivided into a lower and an upper part. The division is made at the upper surface of a local subsurface basalt field within the group that is mapped from seismic reflection profiles south of Nampa. This surface can be traced as a seismic reflection (Horizon "I", figures 2-4 and 2-5) over most of the study area, but its lithologic character changes from a siltstone-basalt contact south of the zone of complex faulting (figure 2-8) to a sandstone-siltstone contact in the region north of Nampa. The surface rises to the east and northeast so that its depth is less than 330m (1,000 ft) beneath the Meridian area, and it may be eroded and absent from the area east and north of Meridian. The seismic horizon used to subdivide the upper and lower Idaho Group unconformably overlies the structural subsurface high of older basalts in the middle of the Plain (figure 2-5). The seismic reflection profiles also indicate that the lower Idaho Group thickens to the north and east, and that most of the exposed Idaho Group on the north margin of the Plain should be considered Lower Idaho Group. In contrast the upper Idaho Group thickens to the southwest and reaches a maximum of about 915 m (3,000 ft) beneath Caldwell.

LOWER IDAHO GROUP (including Poison Creek Formation, Banbury Basalt, and the Chalk Hills Formation) — Type localities for sediments and basaltic volcanics of the Lower Idaho Group are entirely on the south side of the Snake River Plain along the northern edge of the Owyhee Mountains. Malde and Powers (1962) defined the Poison Creek and Chalk Hills formations, but their age relationship to one another and to related basaltic volcanics is obscure (Ekren and others, 1978a). K-Ar age control within the Banbury Basalt and the Idaho Group is published by Armstrong and others (1980). The Banbury Basalt mapped by Malde and others (1963), on the north side of the Plain is dated at 9.4 $\pm$0.6 million years correcting earlier reported younger ages (Armstrong, 1980). The exposed sediments are generally described as lake and stream deposits of buff, white, brown, and grey sand, silt, clay diatomite, numerous thin beds of vitric ash and some basaltic tuffs. Well-cemented arkosic sandstones within the Poison Creek unit form dipping mesas and tilted buttresses along the north side of the Owyhee Mountains. Basalts intruding and within these beds are described as intergranular-to-ophitic textured olivine basalt, and some are mapped by Ekren and others (1978a) as the Banbury Basalt Formation.

Boise Area — Sediments similar to those of the lower Idaho Group exposed along the edges of the Owyhee Mountains on the south side of the Plain are exposed in the Boise foothills where the lower Idaho Group is a series of arkosic, deltaic and lake margin sand units and buff siltstones about 200 m (650 ft) thick (Wood and Vincent, 1980). On the north side of the Plain in the Boise foothills these sediments rest upon an irregular erosion surface of silicic volcanics of the Idavada Group, or upon weathered, eroded, plagioclase phyric basalt flows and basaltic tuffs that are similar to the Banbury Basalt Formation north of Mountain Home, described by Malde and others (1963). A flow of fine-grained olivine-bearing basalts also occurs within the lower part of the arkosic lake margin sand units, and at places these basalts rest upon an eroded surface of a sedimentary unit of tuffaceous silt and white vitric tuffs.

Most of the Idaho Group exposed in the Boise area might be the Lower Idaho Group. Previous workers have described these sediments as the Glenns Ferry Formation of the Idaho Group (Savage, 1958, Hollenbaugh, 1973, Thomas and Dion, 1974); however cross sections of the Plain prepared from seismic reflection profiles (figure 2-4) show that these foothills units emerge to the northeast from deep beneath the Plain at dips of 7° to 10°, and they may not be a part of the Glenns Ferry Formation. The Glenns Ferry Formation, if it exists as a subsurface unit beneath Boise, must lie unconformably on the Lower Idaho Group and be less than 300 m (1,000 ft) thick. Stratigraphic details of the upper 300 m (1,000 ft) are not known, for we have no seismic reflection data for the upper section. The wide-spread "blue clay" encountered in water wells within the upper 180 m (600 ft) of section may be a part of the Glenns Ferry Formation, and so might the sandy coarse facies that rests upon it. It is important for geohydrologic considerations to recognize that the emerging deep units are not conformable. A geohydrologic framework for the upper 300 m (1,000 ft) of section beneath Boise is proposed by Burnham (1979) and represents the first reasonable geologic explanation of the groundwater system; however, well data is sparse and inconclusive regarding the detailed stratigraphy of the upper section.

Nampa-Caldwell Area — West of Boise, in the Meridian area, the Lower Idaho Group fills a deep basin developed by faulting and downwarping before and during the deposition (figures 2-4, 2-5). The basin is bounded to the southwest by structurally high basalt. Lower Idaho Group sediments filled and overtopped the structure and sediments were transported to the west-southwest into the Lake Lowell-Caldwell Basin. Thick sand units at 600 to 700 m (2,000 to 2,300 ft) depth, encountered in wells drilled upon the structural high probably represent a deltaic facies prograding to the west across the submerged high to the subsiding basin to the south in the Lake Lowell-Caldwell area. These sands should exist as aquifers along the top of the high and to the northeast where they emerge or are truncated by younger sediments. The contrast between the deep and shallow investigation resistivity logs (figure 2-9 and Chapter 3, indicates these sands should be good aquifers with moderate permeability. These sands have never been formation tested, nevertheless, they have resistivities and thicknesses comparable to shallower units in the Upper Idaho Group that yield good flows to wells. It is not known if, in the Lower Idaho Group, these sands persist in the subsurface south of the structural high into the Lake Lowell-Caldwell Basin. Coarse arkosic sandstone of the Lower Idaho Group outcrops along the north margin of the Owyhee Mountains south of Nampa and Caldwell (Ekren, and others, 1981). Some of these sandstones have fair permeability; however, many are also tuffaceous and silty and of low permeability. However, the source of these sands may not be the same as the sand in the subsurface for they may have been derived from sources south of the subsiding plain.

The Lower Idaho Group locally contains subsurface basaltic lava fields and several thin flows. A subsurface basaltic lava field is mapped from seismic reflection profiles in the area immediately southwest of Nampa (figure 2-5). The basalt appears to be thickest in the Lake Lowell basin. The top of the basalt surface ranges in depth from 549 m (1,800 ft) to about 823 m (2,700 ft). Within areas that were topographically low at the time of eruption seismic profiles indicate the Idaho Group Basalts may be as much as 150 m (500 ft) thick.

In the Champlain Oil Company (Deer Flat No. 1) well only a few flows 10 to 20 m (30 to 65 ft) thick are encountered at 822 m (2,040 ft) depth. Much of the basaltic material occurs as interbedded sedimentary tuffaceous silt and sand. To the west in the vicinity of Knowlton Heights, the basalt shallows to a depth of 335 m (1,100 ft) and several volcanic vents appear on the reflection profiles (figure 2-5). Subsurface occurrence of these basalts is shown on figure 2-8, and the horizon (Horizon "I") at which the basalts occur is arbitarily used in this report to subdivide the Idaho Group into an upper and lower part.

UPPER IDAHO GROUP — The Upper Idaho Group, as defined in this report, includes the sedimentary sequence above seismic Horizon "I" (figures 2-2, 2-7, and figure 2-8). During deposition of the upper Idaho Group the axis of the sedimentary basin lay beneath Lake Lowell and Caldwell. The basin axis has a northwest trend and plunges down to the northwest. The deepest part of the basin lies northwest of Caldwell beyond the area for which seismic reflection coverage was obtained (figure 2-8). Sedimentation during Upper Idaho Group time was not affected by the large subsurface structural high in the middle of the basin for it was buried by sediments.

A number of continuous seismic reflections occur within the Upper Idaho Group. These reflections are at depths of 240 to 1000 m (800 to 3,200 ft), and are apparently higher velocity siltstone and claystone. These sand units are encountered in deep wells and are well defined by their high resistivity on electrical logs shown on figure 2-9 and higher acoustic velocity on some logs. These widespread sands may represent fluvial episodes that extended out into the basin during epoisodes of regression of a lacustrine environment.

The base of the upper Idaho Group, as defined by reflector "I" in the subsurface, is probably close to the base of the Glenns Ferry Formation as mapped by Malde and others (1963) and Ekren and others (1981). In the foothills along the base of the Owyhee Mountains between Marsing and Murphy, Idaho, gray siltstone of the Glenns Ferry Formation rests unconformably on sediments and basaltic volcanics of the Chalk Hills and Poison Creek Formations. To the east, in the Bruneau area, the base of the Glenns Ferry Formation is locally marked by algal and oolitic limestone resting on beveled beds of the Chalk Hills Formation with slight angular discordance (Malde and Powers, 1962, p. 1207). Oolitic limestone near the base of the Glenns Ferry Formation has been described at numerous localities in a recently completed study by Kimmel (1979).

On the north margin of the plain in the foothills between Boise and Emmett, oolitic sand and limestone occur in the uppermost strata of a sequence of deltaic sands and siltstone that rest upon older basalt (Wood, 1981, unpublished mapping). Oolitic and algal limestone is a common shoreline facies of large saline lakes (Eardley, 1938), and it is likely that the occurrence of oolitic deposits around margins of the Snake River Plain may mark a limited stratigraphic interval within the Idaho group and show the outline of a large lake that existed during Idaho Group time. Kimmel (1979) has suggested that oolitic limestone and sandstone may locally serve as a definition for the base of the Glenns Ferry Formation. It is not known if oolitic sand occurrences on the margins of the plain are stratigraphically equivalent to the seismically defined boundary between the upper and lower Idaho Group; however, there does appear to be a coincidence between the stratigraphic occurrence of oolites on the margins, locally identified unconformities and the uppermost occurrence of local basalt fields in the strata of the Idaho Group of the western plain.

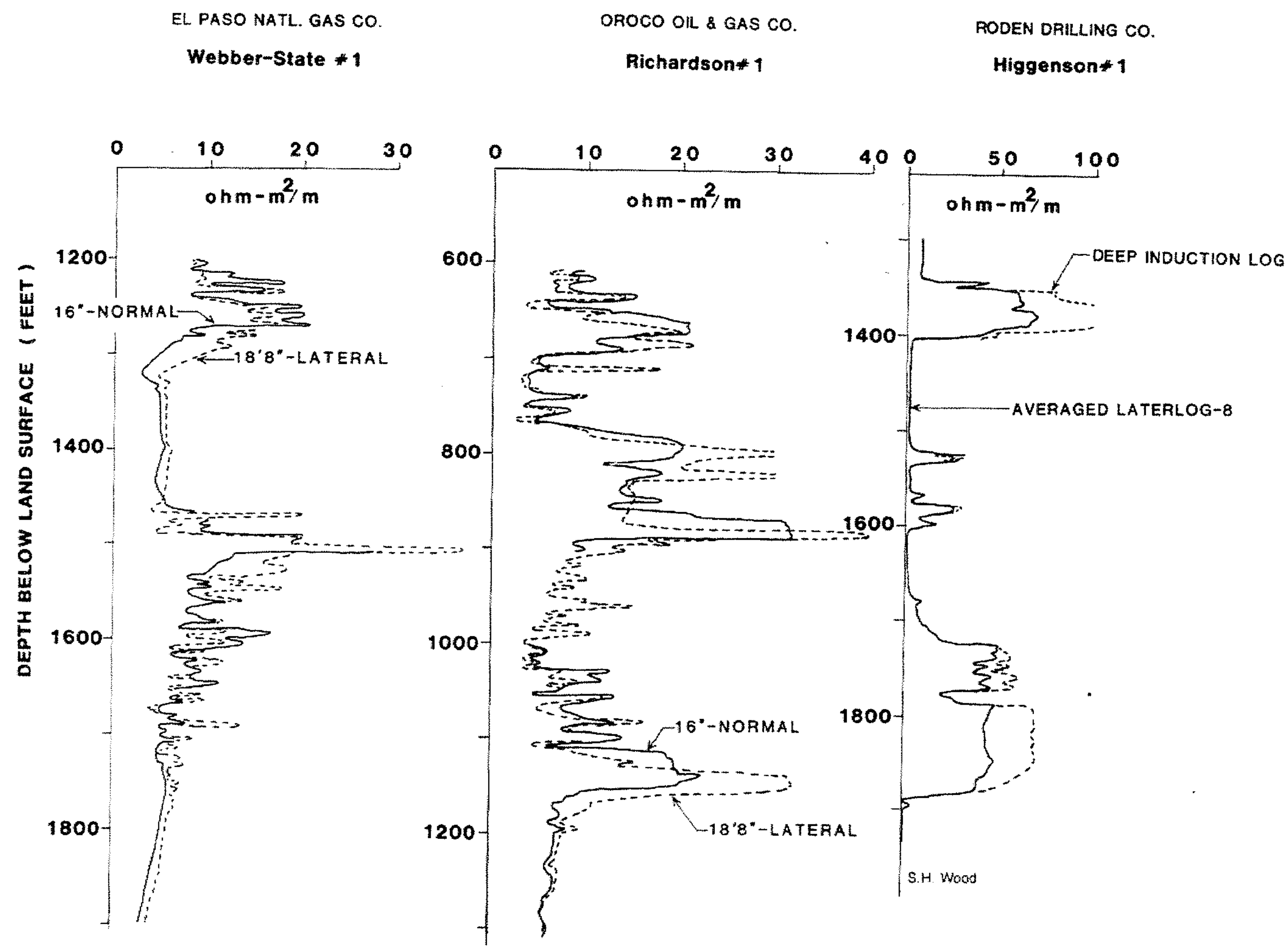


FIGURE 2-9. Resistivity log character of deep aquifer in the Nampa-Caldwell area.

Exposures of the Glenns Ferry Formation mapped along the Snake River Canyon south of Nampa and Caldwell are entirely within the upper Idaho Group (figure 2-6). Here the section consists of 100 m (300 ft) of monotonous, indistinctly-bedded, gray, lacustrine siltstone and fine sandstone. The only variation in this lithology is a 10 m $\pm$ (30 ft $\pm$) thick layer of indurated buff-colored, tuffaceous, poorly sorted, fluvial sandstone that outcrops in Dead Horse Canyon (Sections 17, 18 and 20, T. 2N, R. 3W, figure 2-6). At this locality, and also locally along the basalt-capped white bluffs on the north side of the river, the Glenns Ferry siltstones are conformably overlain by a 20-m (65 ft) thick gravel deposit identified by clast content and stratigraphic setting as the Tuana Gravels of Malde and Powers (1962).

Fission-track ages have been determined on a number of volcanic ash members within the Idaho Group (Kimmel, 1979). Ages of ash layers within the Chalk Hills Formation range from 8.5 $\pm$ 1.2 to 7.0 $\pm$ 0.5 million years. Ash layers bracketing oolitic limestone in the basal Glenns Ferry Formation are 3.2 $\pm$ 0.4 to 2.5 $\pm$ 1.0 million years and 2.4 $\pm$ 0.2 million years. This geochronology and field evidence of an unconformity between the Chalk Hills and Glenns Ferry Formation indicate a hiatus of deposits on the margin of the Plain between about 7.0 $\pm$ 0.5 and 3.2 $\pm$ 0.4 million years.

Tuana Gravels

Well sorted pebble and cobble gravels and coarse sand are exposed along the walls of the Snake River Canyon (Deadhorse Canyon in the SW corner of the Lake Lowell Quadrangle). Stratigraphic position and clast content indicates that these gravels correlate with the Tuana Gravels of Malde and Powers (1962). These gravels contain a few clasts of orange quartzite indicating a Snake River drainage provenance and distinguishing them from the Tenmile Gravels. At this locality Tuana Gravels dip 5 to 10° north and mark an angular unconformity with the overlying, horizontal, Bruneau lake beds. Thickness of gravel is 0 to 15 m (0-50 ft).

The Bruneau Formation

The lacustrine facies and basalt of the Bruneau Formation locally lie unconformably on the north dipping Tuana Gravels and Glenns Ferry Formation in localities along the north side of the Snake River (figure 2-10). The lacustrine deposits are horizontal and contain both gray siltstone and beds of brown tuffaceous sandstone bearing palagonized basalt and scoria. A number of scoria deposits, minor basalt flows and small basalt necks of the Bruneau Formation intrude and overlie the Glenns Ferry Formation. These basalt vents stand out as small hills where the Glenns Ferry Formation has been eroded away in the cutting of the modern Snake River Valley. The Bruneau Formation is mostly confined to the present Snake River Valley area and is probably less than 50 m (160 ft) thick or non-existent in the Nampa-Caldwell area. The formation thickens to the east and is up to 240 m (800 ft) thick in the Bruneau-Grandview area (Malde, 1965). The Bruneau Formation is thought to have originated from a complex series of lava dammed lakes along the canyon of the Pleistocene Snake River (figure 2-10), but many features of its origin have yet to be explained (Malde, 1965).

Age of the Bruneau Formation is middle Pleistocene based upon a single K-Ar determination on basalt of 1.4 million years (Evernden and others, 1964). All lavas have reversed magnetic polarity and are included in the Matuyama Polarity epoch (Malde and Williams, 1975).

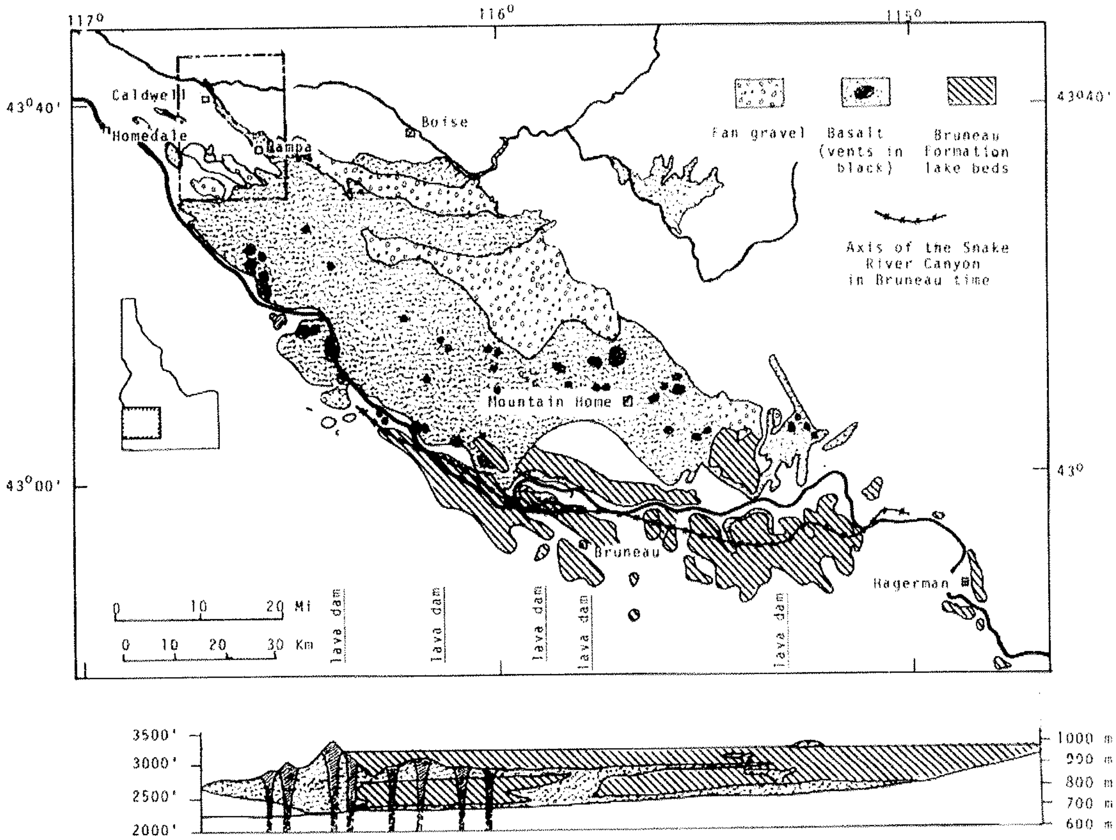


FIGURE 2-10. Occurrences of the Bruneau Formation in the western Snake River Plain. After Malde (1965a).

The Tenmile Gravels

In the Nampa-Caldwell area, the top of the upper Idaho Group is marked by a widespread fluvial gravel deposit that caps many of the hills in the area. The gravel extends eastward across the plain to the mountains east of Boise where it laps against the granitic bedrock. The gravel has a cobble assemblage of granitic rocks and felsic porphyries derived from the adjacent Idaho Batholith terrain. This gravel unit was named the Tenmile Gravels by Savage (1958) because it forms the entirety of Tenmile Ridge south of Boise where it is up to 150 m (500 ft) thick. In the Nampa-Caldwell area, the thickness rarely exceeds 15 m (50 ft). It is widely exploited as a source of gravel for construction and concrete pipe manufacturing. The geometry of the deposit and the cobble assemblage clearly mark the Boise River drainage as the source of the Tenmile Gravels. The Tenmile Gravels are distinct from the Tuana Gravels, because the Tuana Gravels bear a few orange quartzite cobbles and other metamorphics whereas the clasts of the Tenmile Gravels are almost entirely intrusive porphyries and granitic rocks. The stratigraphic relationship of the Tenmile Gravels to other Plio-Pleistocene units such as the Bruneau Formation is uncertain.

Snake River Group

SEDIMENTS OF THE DEER FLAT SURFACE — The higher upland surface both north and south of the Boise River is underlain by at least 30 m (100 ft) of unconsolidated fluvial sand and gravel. The sands are mostly coarse, arkosic and cross-bedded. The gravels are typically in beds 0.5 to 2 m (1.5 to 7 ft) thick and are composed of rounded clasts of intrusives derived from the Boise River drainage. The gravels are indistinguishable from the Tenmile Gravels. The unit is identified by its topographic position on a surface that is about 60 m (200 ft) above the present Boise River. The surface rises generally from 760 m (2,500 ft) on the west side of the map to 823 m (2,700 ft) elevation in the northeast corner of the Middleton 7 1/2 minute quadrangle (figure 2-6). A good exposure of this unit is located in a gravel pit in the SWI/4, Section 3, T. 4N, R. 4W, 3 km (1.9 mi) north of Caldwell. No extensive exposures of this unit are known south of the Boise River. The unit is best identified by the elevation of its surface and by soil development. Priest and others (1972) classify soils upon this surface in the Scism or the Power-Durham series characterized by a moderately well-cemented caliche horizon 0.5 to 1.2 m (1.5 to 4 ft) deep.

The Indian Creek Basalt flows erupted onto surfaces of the Deer Flat Sediments, and are reported in the driller's logs to be interbedded with this unit. The unit is probably middle Pleistocene in age.

INDIAN CREEK BASALT FLOWS — Basalt which outcrops in Section 13, T. 3N, R. 1W by the Nampa State School and also in the north Caldwell area along the Boise River is the same flow unit. The flow is about 0.7 km (0.5 mi) wide. This flow is named the Indian Creek Basalt, for it extends as a continuous flow from beyond the west edge of the map to Caldwell and parallels the course of Indian Creek. Subcrop of this flow has been mapped from the numerous driller's logs in the area. The flow is a gray, titano-augiteplagioclase-olivine basalt, and is typical of basalts of the Snake River Group. The gray color results from an abundance of small lath shaped plagioclase crystals.

One or more deeper flows are identified from the water-well driller's logs at depths of 15 to 30 m (50 to 100 ft). This deeper flow underlies much of the city of Nampa, but the previously discussed shallow flow lies mostly to the north of Nampa. Subcrops of both the flow units are shown with dotted lines on figure 2-6. Both flows probably erupted onto an aggrading surface of sediment and can be considered as interbedded with the fluvial sediments of the Deer Flat Surface. Paleomagnetism of samples from three localities was measured with a field fluxgate magnetometer. Declination was consistently east-northeast, and inclination was much too shallow to clearly identify the direction of paleomagnetism as reversed or normal. The cause of this anomalous magnetization is unknown. Because of the association of the flows with the sediment of Deer Flat Surface, and the presence of an indurated caliche, the age of these basalts is probably middle Pleistocene.

The Indian Creek Basalts constitute the important unconfined, shallow groundwater system of the Nampa-Lake Lowell-Caldwell area. The basalt flow aquifers are tapped by numerous, shallow, domestic and irrigation wells. The groundwater is presently recharged mostly by seepage of imported irrigation water and canal leakage. The water level is quite stable. Because of their shallow situation and high transmissivity, the Indian Creek basalt aquifers can be easily contaminated by surface spills or leakage from buried storage tanks. The extent of these aquifers, is shown in figure 2-6. Wells situated on the hills capped by the older Tenmile Gravels do not encounter shallow aquifers and wells must be drilled deeper to tap the confined aquifers of the upper Idaho Group in order to obtain flows adequate for irrigation.

WHITNEY TERRACE DEPOSIT — Alluvium of the Whitney Terrace of the Boise River (Nace and others, 1957) consists of rounded cobble and pebble gravels, arkosic sands and minor silts. The unit is identified by the relatively flat terrace surface 10 to 20 m (30 to 65 ft) above the present Boise River floodplain. Soils developed upon the terrace gravels have a well developed profile with a brown silty clay loam "B" horizon composed of up to 0.9 m (3 ft) of calcareous nodules extending to a depth of up to 1.3 m (4 ft). These soils are assigned to the Power and Greenleaf series by Priest and others (1972). Soil profiles developed on this surface suggest a pre-Wisconsin age for the Whitney Terrace Deposits, or an age of 100,000 years or more.

QUATERNARY ALLUVIUM AND FAN DEPOSITS — Recent fluvial deposits cover the modern flood plain of the Boise River and Indian Creek. The Boise River flood plain is about 5 km wide in the map area. In the Middleton Quadrangle, ephemeral tributaries have formed small alluvial fans that extend into the Boise River flood plain.

STRUCTURE

Early Pleistocene gravels and older stratigraphic units in the Nampa-Caldwell area have been deformed by normal faulting, gentle tilting of fault blocks and broad downwarping toward the axis of the plain. Normally faulted gravels and sand layers are observed in several quarries in the area (figure 2-6) with displacements of a few decimeters to 10 m (1 to 30 ft). All mapped faults strike in a Northwest-southeast direction with dips ranging from 55° to nearly vertical. The only exception is the North Caldwell Fault exposure with an east-west strike. Linear ridges capped by the Tenmile Gravels trend northwest through the area in the vicinity adjacent to, and southwest of, Lake Lowell. The steeper sides of these asymmetric ridges are interpreted as fault-line scarps of normal faults that have apparent displacements up to 30 m (100 ft). Cross sections constructed from driller's logs (figure 3-1) show offsets up to 50 m (150 ft) of distinctive lithologic units. These offsets are interpreted as normal faults (figure 2-6). Bedding attitudes on exposure of the late Pliocene or early Pleistocene Glenns

Ferry Formation and Tuana Gravels are generally down to the northwest 3° to 10°. These attitudes are taken along the north rim of the Snake River Canyon and are consistent with dips interpreted from seismic reflection profiles (figure 2-5).

Attitudes on the Tenmile Gravel bedding are generally horizontal, but variations in elevation of the base of the gravel indicate that the gravels are faulted and slightly tilted. The elevation of gravels and direction of throw on faults is not in any consistent direction and no broader structures larger than individual fault blocks one or two km (0.6 to 1.2 mi) wide are indicated by deformation of the gravels.

Lack of the Tenmile Gravels northeast of Nampa, in the area covered by Terrace deposits and Quaternary alluvium, is attributed to erosion of the gravels by the Boise River and not to downfaulting. Very few surface faults are shown in the area on the map (figure 2-6) because the younger Boise River alluvial deposits are not faulted. Seismic reflection profiles show that a number of subsurface faults do occur in this area, and these are shown with a different symbol on figure 2-6.

The upper surface of the middle Pleistocene Snake River Basalts varies somewhat, and along the north rim of the Snake River the basalts appear to be slightly tilted, a few degrees, downward toward the basin axis of the plain to the northeast. The basalt surface is probably slightly downwarped by continued subsidence of the plain, but the flows do not appear to be broken by faulting in the map area. The larger tilted and faulted basalt buttes in the map area are probably earlier basalts of the Bruneau Formation (Pickles Butte, and the buttes in Hidden Valley); although, some of these basalts may have been mapped as the Snake River Group on figure 2-6.

The middle Pleistocene Snake River Basalts are not obviously faulted except in the vent areas. Faulting and vent fissures trend northwest across the two major basalt-shield vents shown on the map (Kuna Butte and Powers Butte). Kuna Butte differs from Powers Butte in that its north flank is made up of tilted Tenmile Gravels that were either a pre-existing fault block, or were arched by a shallow intrusion of the basalt. It is not a simple shield volcano such as Powers Butte. The northwest trend of the vent fractures suggests that the basalt erupted from deep sources along fissures opened by the same northeast-southwest oriented extentional stresses that produced the faulting in the slightly older Tenmile Gravels.

SELECTED REFERENCES

Anderson, J.E., 1981. Capital Mall Geothermal Exploratory Well #1· Drilling and Completion Report, Idaho Dept. of Water Resources, 13p.

Applegate, J.K., and Donaldson, P.R., 1977, Characteristics of selected geothermal systems in Idaho, **in** The Earth's Crust, Amer. Geophy. Mono. no. 20, p. 676-692.

Armstrong, R.L., 1974, Geochronometry of the Eocene volcanic plutonic episode in Idaho: Northwest Geology, v 3, p.1-15.

Armstrong, R.L., Leeman, W.P., and Malde, H.E., 1975, K-Ar dating of Quaternary and Neogene volcanic rocks of the Snake River Plain, Idaho: Am. Jour. Sci., v. 275, p.225-251

Armstrong, R.L., Harakal, J.E., and Neill, W.M., 1980, K-Ar dating of Snake River Plain (Idaho) volcanic rocks - new results: Isochron/West, no. 27, p. 5-10.

Arney, B.J., Beyer, J.H., Simon, D.B., Tonani, F.B., and Weiss, R.B., 1980, Hot dry rock geothermal site evaluation, western Snake River Plain, Idaho: Transactions, Geothermal Resources Council, v. 4, p 197-200.

Atwater, Tanya, 1970, Implications of plate tectonics for the Cenozoic tectonic evolution of western North America: Geol. Soc. Am. Bull., v. 81, p. 3513-3536.

Bennett, E.H., and Galbraith, J.H., 1975, Reconnaissance geology and geochemistry of Silver City-South Mountain region, Owyhee County, Idaho: Idaho Bur. Mines and Geology, Pamph. 162, 88 p.

Berg, J.W., and Thiruvathuskal, J.K., 1967, Complete Bouger gravity anomaly map of Oregon. Map GMS-4b. Oregon Dept. Geology and Mineral Industries, Portland, Oregon.

Birch, F., 1950, Flow of heat in the Front Range, Colorado: Geol. Soc. Am. Bull., v. 61, p. 567-630.

Birch, F., and Clark, H., 1940, Thermal conductivity of rocks and its dependence on temperature and composition: Am. Jour. Sc., v. 238, no. 9, p. 613-635.

Birkeland, P.W., Crandell, D.R., and Richmond, G.M., 1971, Status of correlation of Quaternary stratigraphic units of the western conterminous United States: Quaternary Research, v. 1, p. 208-227

Blackwell, D.D., and Chapman, D.S., 1977, Interpretation of geothermal gradient and heat flow data for basin and range geothermal systems: Geothermal Resources Council, Trans. Vol. 1, p 19-20.

Blackwell, D.D., and Steele, J.L., 1977, The terrain effect on terrestrial heat flow, a mathematical solution [submitted for publ.]: Jour. Geophy. Res.

Bond, J.G. and Wood, C.H., 1978, Geologic map of Idaho: Idaho Dept. of Lands, Bur. of Mines and Geol., 1:500,000 scale map, 1 sht.

Bonini, W.E., 1963, Gravity anomalies in Idaho: Idaho Bur. Mines and Geol., pamphlet no. 132, 38 p.

Bonnichsen, W.B., and Travers, W.B., 1975, Rhyolitic volcanism and structural evolution of the Snake River Plain (abs.): Geol. Soc. Am., Rocky Mt. Section, v 7, no. 5, p.589-590.

Bowen, R.G., 1972, Geothermal gradient studies in Oregon: Ore Bin, v. 34, no 4, p. 68-71

_________________, 1973, Geothermal activity in 1972: Ore Bin, v. 35, no. 1, 7p.

Bowen, R.G., and Blackwell, D.D., 1975, The Cow Hollow geothermal anomaly, Malheur County, Oregon: Ore Bin, v. 37, no. 7, p 107-124.

Brook, C.A., Mariner, R.H , Mabey, D.R., Swanson, R.R., Guffanti Marianne, and Muffler, L.J.P., 1979, Hydrothermal convection system with reservoir temperatures 90 C, in Assessment of Geothermal Resources of the United States—1978, U.S Geol Surv Circular 790, p. 18-85.

Brott, C.A., Blackwell, D.D., and Mitchell, J.C., 1976, Geothermal investigations in Idaho, part 8, heat flow in the Snake River Plain Region, southern Idaho: Idaho Dept. of Water Resources, Water Inf. Bull. no. 30, 195 p.

Brott, C.A., Blackwell, D.D., and Mitchell, J C., 1978, Tectonic implications of the heat flow of the western Snake River Plain, Idaho: Geol. Soc. Amer. Bull., v. 89, p.1697-1707

Brott, C.A., (submitted for publication), Snake River aquifer.

Bryan, Kirk, 1929, Geology and dam sites with a report on the Owyhee Irrigation Project, Oregon: U.S. Geol. Surv. Water Supply Paper 597-A, 89 p.

Burnham, W.L., 1979, Southwest community waste management study, groundwater subtask, Ada County, Idaho: Technical Memoranda 308.04g (Nov 1979), Ada County Planning Assoc., Boise, Idaho. 66 p.

Buwalda, J.P., 1923, A preliminary reconnaissance of the gas and oil possibilities of southwestern and south-central Idaho: Id. Bur. of Mines and Geol. Pamph. no.5, 10p.

Carslaw, H.S., and Jaeger, J.C., 1959, Conduction of heat in solids: Clarendon Press, Oxford, 510 p.

Cater, F.W., Pinckney, D.M., Hamilton, W.B., Parker, R.L., Weldin, R.D., Close, T.J. and Zilka, N.T., 1973, Mineral resources of the Idaho Primitive Area and vicinity, Idaho: U.S. Geol. Surv. Bull. 1304, 431 p.

Craig, H., 1961a, Isotopic variation in meteoric waters: Sci. v. 133, p. 1702-03.

Craig, H., 1961b, Standard for reporting concentrations of deuterium and oxygen-18 in natural waters: Sci. v 133, p. 1833.

Day, N.F., 1974, Linears map of Idaho, band 5, MSS-ERTS: Id. Bur. Mines and Geol., Moscow, Idaho.

Decker, E.R., and Smithson, S.B., 1975, Heat flow and gravity interpretation across the Rio Grande Rift in southern New Mexico and west Texas: Jour. of Geophys. Res., v. 80, p. 2542-2552.

Dion, N.P., 1972, Some effects of land use change on the shallow groundwater system in the Boise-Nampa area, Idaho: Idaho Dept. of Water Resurces Water Inf. Bull. No. 26, 47 p.

Domenicao, P.A., and Palciauskas, V.U., 1973, Theoretical analysis of forced convective heat transfer in regional groundwater flow: Geol. Soc. Am. Bull., v. 84, p.3803-3814.

Eardley, 1938, Sediments of the Great Salt Lake, Utah: Bull. Amer. Assn. Petroleum Geologists, v. 22, p. 1305-1411.

Ekren, E.B., McIntyre, D.H., and Malde, H.E., 1981, Geologic map of Owyhee County, Idaho, west of longitude 116°W. U.S. Geol. Survey Map I-1256.

Ekren, E.B., McIntyre, D.H., and Bennett, E.H., 1978a, Preliminary geologic map of the west half of Owyhee County, Idaho: U.S. Geol. Surv. open-file report 78-341, 14 p.

________________, 1978b, Welded ash-flow sheets that reverted to high visocity liquids, Owyhee County, Idaho: Geol. Soc. Am. Abs. with Programs, v. 10, no. 5, p. 215.

Evernden, J.F., Savage, D.E., Curtis, G.H., and James, G.T., 1964, Potassium-argon dates and the Cenozoic mammalian chronology of North America: Am. Jour. Sci. 262, pp. 145-198.

Evernden, J.F., and James, G.T., 1964, Potassium-argon dates and Tertiary floras of North America: Am. Jour. Sci. v.262, p. 145-198.

Fournier, R.O., 1979, Geochemical and hydrologic considerations and the use of enthalpy-chloride diagrams in the prediction of underground conditions in hot-spring systems: Jour. Volcan. and Geothermal Research, v. 5, p. 1-16.

________________, 1977, Chemical geothermometers and mixing models for geothermal systems: Geothermics, v. 5, no. 1-4, p.41-50.

Fournier, R.O. and Potter, R.W., II, 1979, A magnesium correction for the Na-K-Ca chemical geothermometer: Geochem. et. Chosmochim Acta. v. 43, p. 1543-1550.

Fournier, R.O. and Rowe, J.J., 1966, Estimation of underground temperatures from the silica content of water from hot springs and wet steam wells: Am. Jour. Sci., v. 264, p.686-695.

Fournier, R.O., Sorey, M.L., Mariner, R.H., and Truesdell, A.H., 1979, Chemical and isotopic predictions of aquifer temperatures in the geothermal system at Long Valley, California, J. Volcanol. Geotherm. Res., v. 5, p.17-34.

Fournier, R.O., and Truesdell, A.H., 1970, Chemical indicators of subsurface temperature applied to hot waters of Yellowstone National Park, Wy., U.S.A., in Proceedings United Nations Symp. on the Development and Utilization of Geothermal Energy, Pisa, 1970, v. 2, Part 1, Geothermics, Spec. Issue 2; p.529-535.

________________, 1973, An empirical Na-K-Ca geochemical thermometer for natural waters: Geochim, et. Cosmochim, Acta., v. 73, p. 1255-1275.

________________, 1974, Geochemical indicators of subsurface temperature, and fraction of hot water mixed with cold: U.S. Geol. Surv. Jour. of Research, v. 2, no. 3, p.264-270.

Fournier, R.O., White, D.E. and Truesdell, A.H., 1974, Geochemical indicators of subsurface temperature, Part 1. Basic Assumptions: U.S. Geol. Surv. Jour. of Research, v. 2, no. 3, p 264-270.

Gat, G.R., 1971, Comments on the stable isotope method in regional groundwater investigations: Water Resource Research v 7, no. 4, p.980-993.

Grindley, G.W., 1970, Subsurface structures and relation to steam production in the Broadlands Geothermal Field, New Zealand, Geothermics V. 2, part 1, Spec. Issue 2, p. 248-261

Hamilton, W.B. and Myers, B.W., 1966, Cenozoic tectonics of the western United States: Rev. Geophys., v. 4, p.509-549.

Hammer, Sigmund, 1939, Terrain corrections for gravimeter stations: Geophysics vol. 4, pp. 184-194.

Harrison, J.E., Griggs, A.B., and Wells, J.D., 1974, Tectonic features of the Precambrian belt basin and their influences on post-belt structures: U.S. Geol. Surv. Prof. Paper 866, 15 p.

Heiland, C.A., 1968, Geophysical Exploration, Hafner Publishing Co., p. 74-84.

Hem, J.D., 1970. Study and interpretation of the chemical characteristics of natural water· 2nd ed., U.S. Geol. Surv. Water Supply Paper 1475, 363 p.

Hill, D.P., Baldwin, H.L., Jr., and Pakiser, L.C., 1961. Gravity, volcanism and crustal deformation in the Snake River Plains: U.S. Geol. Surv. Prof Paper 424-b, p. 248-250.

Hill, D.P., 1963, Gravity and crustal structure in the western Snake River Plain, Idaho. Jour. Geo. Res. 68, pp. 5807-5818.

_______________, 1972, Crustal and upper mantle structure of the Columbia Plateau from long range seismic refraction measurements. Geol. Soc. Amer. Bull. vol. 83, pp. 1639-1648.

Hill, D.P., and Pakiser, L.C., 1965, Crustal structure between the Nevada test site and Boise, Idaho, from seismic-refraction measurements, in the Earth beneath the continents, American Geophysical Union. Geophysics Monthly 10, pp. 391-419.

Hiner, J.E., 1976, Lithogic and mud log of the M.T. Halbouty, J.N. James No. 1 well, Meridian, Idaho: Unpublished logs of well on file with the Idaho State Petroleum Engineer (W. Pittman, Idaho State Depart. of Lands, P.O. 670, Coeur d'Alene, ID.

Hollenbaugh, K.M. 1973, The evaluation of geologic processes in the Boise Foothills that may be hazardous to urban development: Ada Council of Governments, 525 W. Jefferson, Boise, ID., 88 p

Hooper, P.R., 1981, The role of magnetic polarity and chemical analysis in establishing the stratigraphy, tectonic evolution, and petrogenesis of the Columbia River Basalt: Symposium: "Deccan volcanism and related basalt provinces in other parts of the world:" Geol. Soc. of India (in press).

Hull, D.A., 1975, Geothermal gradient data Vale area, Malheur Co., Oregon: State of Oregon, Dept. of Geol. and Min. Indust., open-file rpt. 0-75-4, 18 p.

Hull, D.A., Blackwell, D.D., Bowen, R.G., and Peterson, N.W., 1977, Heat flow study of the Brothers Fault Zone, Oregon: State of Ore., Dept. of Geol. and Min. Indust., open-file rept. 0-77-3, 43 p.

Hull, D.A., Blackwell, D.D., and Black, G.L., 1978, Geothermal gradient data: State of Ore., Dept. of Geol. and Min. Indust., open-file rept. 0-78-4, 187 p.

Idaho Geothermal Development Projects Annual Report for 1976, 1977. A summary of the developments in the joint program of the Id. Natl. Engr. Lab. (E.G.&G. Idaho, Inc. and ERDA): The Raft River Rural Electric Cooperative.

Jaffe, H.W., Gottfried, D., Waring, C.L., Worthing, H.W., 1959, Lead-alpha age determinations of accessory minerals in igneous rocks (1953-1957): U.S. Geol. Surv. Bull. 1097B, p. 64-148.

Kelly, Jack, 1976, Written Communciation to K. Ames, J.U.B. Engineers, Inc.

Kimmel, P.G., 1979, Stratigraphy and paleoenvironments of the Miocene Chalk Hills Formation and Pliocene Glenns Ferry Formation in the western Snake River Plain, Idaho: Ph.D. Thesis, Univ. of Michigan. 331 p.

Kirkham, V.R.D., 1930, Old erosion surfaces in southwestern Idaho: Jour. of Geol., v. 38, no. 7, p. 652-663.

_______________, 1931a, Revision of the Payette and Idaho Formations: Jour. of Geol., v. 39, p. 193-239.

_______________, 1931b, Snake River downwarp: Jour. of Geol., vol. 39, no. 5, pp. 456-482.

_______________, 1931c, Igneous geology of southwestern Idaho: Jour. of Geol., vol. 39, no. 6, pp. 564-591.

Kittleman, L.R., Green, A.R., Hagood, A.R., Johnson, A.M., McMurray, J.M, Russell, R.G., and Weeden, D.A., 1965, Cenozoic Stratigraphy of the Owyhee Region, southeastern Oregon: Bull. no. 1, Museum of Natural History, University of Oregon, Eugene, OR 45 p.

Kunze, J.F., 1975, Idaho geothermal R&D project report for period January 1, 1975 to March 31, 1975, Idaho National Engineering Laboratory ANCR Report No. 1222, p. 17

Krueger, H.W., 1981, personal communication, Krueger Enterprises, Inc., Geochron Laboratories Division, Cambridge, MA.

Lachenbruch, A.H., and Sass, J.H., 1977, Models of an extending lithosphere and heat flow in the basin and range province, **in** Cenozoic Tectonics and Regional Geophysics in the Western Cordillera: Geol. Soc. Am. Mem. 152, p. 209-250.

LaFehr, T.R., 1962, Gravity survey in the eastern Snake River Plains, Idaho - a progress report: U.S. Geol. Surv. open-file rept. 35.

Larsen, E.S., Jr., Gottfried, D., Jaffe, H.W., and Waring, C.L., 1958, Lead alpha ages of the Mesozoic batholiths of western North America: U.S. Geol. Surv. Bull. 1070-B, p. 35-62.

Lewis, R.E., and Young, H.W., 1980, Thermal springs in the Payette River Basin, west central Idaho: U.S. Geol. Survey Water Resources Investigations open file report 80-1020, 23p.

Lillie, R.J., and Couch, R.W., 1979, Geophysical evidence of fault termination of the basin and Range Province in the vicinity of the Vale, Oregon Geothermal area: Rocky Mountain Assoc. of Geologists Utah Geol. Assoc., 1979 Basin and Range Symposium, p. 176-184.

Lindgren, W., 1898, Description of the Boise Quadrangle, Idaho: U.S. Geol. Surv. Geol. Atlas of the U.S. Folio 45, 7 p.

Lindgren, W., and Drake, N.F., 1904, Description of the Nampa Quadrangle, Idaho, Oregon: U.S. Geol. Surv. Geologic Atlas Folio 103.

Lindholm, J. 1981, Personal Communication: U.S. Geol. Surv. Geophys. Div., Denver, Col.

Mabey, D.R., 1976, Interpretation of a gravity profile across the western Snake River Plain, Idaho: Geology, v.4, 53-55.

Mabey, D.R., Peterson, D.L., and Wilson, C.W., 1974, Preliminary gravity map of southern Idaho: U.S. Geol. Surv. open-file rpt.

MacLeod,, N.S., Walker, G.W., and McKee, E.H., 1975, Geothermal significance of eastward increase in age of late Cenozoic rhyolite domes in southeastern Oregon: 2nd U.N. Symp. on Development and use of Geothermal Resources, San Francisco, May 1975, V. 1, p. 465-474. Also in U.S. Geol. Survey open-file report, 75-348.

Malde, H.E., 1959, Fault zone along the northern boundary of western Snake River Plains: Sci., v. 130, no. 3370, p. 272.

________________, 1965a, Field Trip Guide, Mountain Home to Malad Springs, **in** Guidebook Field Conference E - Northern and Middle Rocky Mountains. (G.M. Richmond, R. Fryxell, J. Montagne, and D.E Trimble, eds., pp. 94-98, INQUA Congress VII, Nebraska Academy of Sciences, Lincoln.

________________, 1965b, Snake River Plain, in the Quaternary of the United States: Princeton, N.J., H.E. Wright and D.G. Frey, Editors, Princeton University Press, pp. 255-263.

________________, 1972, Stratigraphy of the Glenns Ferry Formation from Hammett to Hagerman, Idaho: U.S. Geol. Surv Bull. no. 1331-D, 19 p.

Malde, H.E., and Powers, H.A., 1962, Geologic Map of the Glenns Ferry-Hagerman area, west-central Snake River Plain, Idaho: U.S. Geol. Surv. Map I-696.

________________, 1962, Upper Cenozoic stratigraphy of western Snake River Plain, Idaho: Geological Soc. of America Bull., v. 73, p. 1197-1220.

Malde, H.E., Powers, H.A., and Marshall, C.H., 1963, Reconnaissance geologic map of west-central Snake River Plain, Idaho: U.S. Geological Surv. Misc. Geology Inv. Map I-373. 1:125,000.

Malde, H.E,. and Williams, P.L., 1975, Geology of the western Snake River Plain: Rocky Mountain Section, Geological Soc. of America Meeting, Unpublished field trip guide.

Mariner, R.H., 1981, personal communication, U.S. Geol. Surv., Menlo Park, CA.

Mayo, A.L., 1981, personal communication, Dept. of Geology, Univ. of Idaho, Moscow, Idaho.

McKee, B., 1972, Cascadia: McGraw-Hill Inc., N.Y., N.Y., P 255-270.

McIntyre, D.H., 1976 Reconnaissance geologic map of the Weiser geothermal area, Washington County, Idaho: U.S. Geol. Surv Map MF-745.

________________, 1979, Preliminary description of Anschutz Federal No. 1 drill hole, Owyhee County, Idaho: U.S. Geol. Surv. open-file rpt. 79-651, 15p.

Mitchell, J.C., Johnson, L.L., and Anderson, J.E., 1979, Geothermal investigations in Idaho, part 9, potential for direct heat application of geothermal resources: Idaho Dept. of Water Resources Water Inf. Bull. no. 30, 396 p.

Mohammed, O.M.J., 1970, Hydrology of the Boise Ridge area (M.S. Thesis): College of Mines and Geol., Univ. of Idaho, Moscow, Idaho, 75 p.

Mundorff, M.J., Crosthwaite, E.G., and Kilburn, C., 1964, Ground-water for irrigation in the Snake River Plains, Idaho: U.S. Geol Surv. Water-Supply Paper 1654, 224 p.

Muller, A.B., 1981, Personal communication, Fuel Cycle Risk Analysis, Sandia National Laboritories, Albuquerque, N.M.

Nace, R.L., West, S.W., and Mower, R.W., 1957, Feasibility of ground-water features of the alternate plan for the Mountain Home project, Idaho, U.S. Geol. Surv. Water Supply Paper 1376, 116 p.

Neville, C., Opdyke, N.D., Lindsay, E.H., and Johnson, N.M., 1979, The paleomagnetic stratigraphy of Pliocene deposits of the Glenns Ferry Formation Idaho, and its implication for North American mammalian biostratigraphy: Amer. Jour. Sci., V. 279, P 503-526.

Newcomb, R.C., 1971, Tectonic structure of the main part of the basalt of the Columbia River Group, Washington, Oregon, and Idaho: U.S. Geol Surv. Map I-587

________________, 1972, Quality of the groundwater in basalt of the Columbia River Group, Washington, Oregon, and Idaho: U.S. Geol Surv. Water-Supply Paper 1999-N, 71 p.

Newton, V.C., and Corcoran, R.E., 1963, Petroleum geology of the western Snake River Basin, Oregon-Idaho: Oregon Dept. of Geol. and Mineral Indust. Oil and Gas Inv., no. 1, 67 p.

Olsen, J.R., 1980, A gravity study of the Nampa-Caldwell Area Canyon County, Idaho: Brigham Young Univ. Geol. Studies, v. 27, part 1, p. 101-115.

Pakiser, I.C., 1963, Structure of the crust and upper mantle in the western United States: Jour. of Geophys. Res., v. 68, no. 20, p. 5747-5756.

Pansze, A.J., Jr., 1975, Geology and ore deposits of the Silver City-Delamar-Flint region, Owyhee County, Idaho: Idaho Bur. Mines and Geology Pamph. 161, 80 p.

Parasnis, D.S., 1971, Temperature extrapolation to infinite time in geothermal measurements: Geophysical Prospecting, vol 19, no 4, p. 612-614.

Powers, H A., 1947, Diatomite deposits of southwestern Idaho: Bur. Mines and Geol., Mineral Resources Report: no.4, 27 p.

Priest, T.W., Case, C.W., Witty, J.E., Preece, R.K., Jr., Monroe, G.A., Biggerstaff, H.W., Logan, G.H., Rasmussen, L.M., and Webb, D.H., 1972, Soil Survey of Canyon County area, Idaho: U.S. Dept. of Agriculture, Soil Cons. Service, 118 p., 1:20,000.

Prinz, M., 1970, Idaho rift system, Snake River Plains, Idaho: Geol. Soc. Am. Bull., v. 81, p. 941-948

Ralston, D.R., and Chapman, S.L., 1968, Ground-water resources of the Mt. Home area, Elmore County, Idaho: prepared and published by Idaho Dept. of Reclamation, Water Inf. Bull no. 4, 63 p.

________________, 1970, Ground-water resources of southern Ada and western Elmore Counties, Idaho: Idaho Dept. Water Admin. Water Inf Bull. no. 15, 52 p.

Randall, W., 1974, An analysis of the subsurface structure and stratigraphy of the Salton Sea geothermal anomaly (Ph.D. Thesis): Univ. of Cal., Riverside, Cal., 92 p.

Rightmire, C.T., Young, H.W. and Whitehead, R.L., 1976, Geothermal investigations in Idaho, part 4, Isotopic and geochemical analysis of water from the Bruneau-Grand View and Weiser areas, southwest Idaho: Idaho Dept. Water Resources, Water Inf Bull. no. 30, p.28.

Ross, C.P., 1956, Quicksilver deposits near Weiser, Washington County, Idaho: U.S. Geol. Surv. Bull 1042-D, p.79-104.

Ross, S.H., 1971, Geothermal potential of Idaho: Idaho Bur. of Mines and Geology Pamph. no. 150, 72 p., also **in** Proceedings, United Nations Symp. on the Development and Utilization of Geothermal Resources, Pisa, v. 2, Part 2, Geothermics Spec. Issue 2, p.975-1008.

Roy, R.F., Blackwell, D.D., and Decker, E.R., 1972, Continental heat flow: **in** the Nature of the Solid Earth, E.C. Robertson, editor, McGraw-Hill, N.Y., p. 506-543.

Russell, I.C., 1902, Geology and water resources of the Snake River Plains: U.S. Geol. Surv. Bull. 199, 192 p.

________________, 1903, Preliminary report on artesian basins in southwestern Idaho and southeastern Oregon, U S. Geol. Surv Water Supply Paper 78, 53 p

Sass, J.H., Lachenbruch, A.H., and Munroe, J.R., 1971 Thermal conductivity of rocks from measurements on fragments and its application to heat flow determination: Jour of Geophys. Research, v 76, no 14, p. 3391-3401

Savage, C.N., 1958. Geology and mineral resources of Ada and Canyon Counties, Idaho: Idaho of Bur. Mines and Geol., County Report No. 3, 94 p., 1:125,000. Scale Map.

________________. 1961. Geology and mineral resources of Gem and Payette Counties: Idaho Bur. of Mines and Geol., County Report No. 4, 49 p., 1:125,000. Scale Map.

Smith, R.N., 1977. Geophysics in groundwater exploration, a case history: Unpublished report.

________________. 1980. Heat flow of the western Snake River Plain: Washington State Univ. Dept. of Geol., M.S. Thesis. 141 p.

Stearns, H.T., Crandall, L., and Steward, W.G. 1938. Geology and groundwater resources of the Snake River Plain in southeastern Idaho: U.S. Geological Survey Water Supply Paper 774, 268 p.

Stearns, N D., Stearns, H.T., and Waring, G.A., 1937. Thermal springs in the United States: U S Geol. Surv. Water-Supply Paper 679-B, p. 59-206.

Stevens, P.R., 1962. Effects of irrigation on groundwater in southern Canyon County, Idaho: U.S. Geol. Surv Water-Supply Paper 1585, 74 p.

Stone, G.T., 1967 Petrology of upper Cenozoic basalts of the Snake River Plain (Ph d Thesis): Univ. of Colorado, 392 p.

Svaty, Norman, 1979, personal commuication, Cannon Farms, Inc . Nampa, ID

Swanberg, C.A., and Blackwell, D.D., 1973. Areal distribution and geophysical significance of heat generation in the Idaho Batholith and adjacent intrusions in eastern Oregon and western Montana: Geol Soc. of Amer Bull. v. 64, no. 4, p. 1261-1282.

Swanson, D.A., Wright, T.L., Hooper, P.R., and Bentley, R.D., 1979. revisions in stratigraphic nomenclature of the Columbia River Basalt Group: U.S. Geol. Surv Bull. 1457-G, 59 p.

Swanson, D.A., and Wright, T.L., 1981, Guide to geologic field trip between Lewiston, Idaho and Kimbely, Oregon, emphasizing the Columbia River Basalt Group: **in** John, D.A., and Donnelly, N.J., eds., Guides to some volcanic terranes in Washington, Oregon and Idaho: U.S. Geol. Survey Circular 838, p. 1-28.

Swanson, J.R., 1977. Data File Geotherm: U.S. Geol. Surv. open-file data. Menlo Park, CA.

Taubeneck, W.H., 1970. Dikes of the Columbia River basalt in northeastern Oregon, western Idaho, and southwestern Washington, **in** Gilmour, E.H., and Stradling, D., eds., Proceedings Second Columbia River Basalt Symposium: Cheney, Eastern Washington State College, Press. p.73-96.

Taubeneck, W.H., 1971. Idaho Batholith and its southern extension: Geol. Soc. America Bull., v. 83, p.1899-1929.

Telford, W.M., Geldart, L., Sheriff, R.E., and Keys, D.A., 1976, Applied geophysics, Cambridge Univ. Press, Cambridge, Massachusetts.

Thomas, C.A . and Dion, N.P., 1974. Characteristics of streamflow and groundwater conditions in the Boise River Valley, Idaho: U.S. Geol. Surv. Water Resources Investigations no. 38-74, 56 p.

Thompson, R.N., 1977. Columbia-Snake River, Yellowstone magmatism in the context of western U.S.A. geodynamics: Tectonophysics, v. 39, p. 621-636.

Truesdall, A.H., and Fournier, R.O. 1977. Procedures for estimating the temperature of a hot-water component in a mixed water by using a plot of dissolved silica versus enthalpy: U.S. Geol. Surv. Jour. of Research, v. 5, no. 1, p.49-52.

Truesdall, A.H., and Hulston, J.R., 1980. Isotopic evidence environments of geothermal systems: **in** Handbook of Environmental Isotopic Geochemistry, P Fritz and J. Ch. Fontes, ed. Vol 1 Terrestrial Environment, Elsevier Scientific Publishing Co., N.Y.

Truesdall, A.H., and Nathenson, Manuel, 1977, The effects of subsurface boiling and dilution on the isotopic composition of Yellowstone thermal waters; Jour. of Geophys. Research, v. 82, no. 26, p. 3694-3704

Urban, R.C . and Diment, W.H., 1975. Heat flow on the south flank of the Snake River Rift (abs.): Geol. Soc. Am. Rocky Mt. Section, v 7, no. 5, p. 648. U.S. Geol. Survey, 1971, Aeromagnetic map of southwestern Idaho: U.S. Geol. Survey open-file rept.

Vallier, T.L. and Hooper, P.R., 1976, Geologic Guide to Hells Canyon, Snake River: Field Guide No. 5, Cordilleran Section, Geol. Soc of Am., Dept. of Geol., Washington State University, Pullman, 38 p.

Van Ostrand, C.E., 1938, Temperature in the lava beds of east-central and south-central Oregon: Am. Jour. of Sc., v 35, no 205, p 22-46.

Walker, G.W., 1977, Geologic Map of Oregon east of the 121st meridian. U.S. Geol. Surv. miscellaneous investigations I-902. 1:500,000 scale map, 2 shts.

Waring, G.H., (Revised by Blankenship, R.P., and Bentel, Ray), 1965. Thermal springs of the United States and other countries of the world - a summary: U.S. Geol. Survey Prof Paper 492, 383 p.

Warner, M.M., 1975. Special aspects of Cenozoic history of southern Idaho and their geothermal implications: United Nations symposium on Development of Geothermal Potential, 2nd, Proceedings: Washington, D.C., U S Government Printing Office, p. 653-663.

Warner, M.M., 1977. The Cenozoic of the Snake River Plain of Idaho: 29th annual field conference, 1977 Wyoming Geological Association Guidebook, pp. 313-326.

Washburne, C.W., 1909, Gas and oil prospects near Vale, Oregon and Payette, Idaho: U S Geol. Surv. Bull., 431, pt. 11, Mineral Fuels, p. 35-49.

Watkins, N.D., and Baksi, A.K., 1974. Magnetostratigraphy and oroclinal folding of the Columbia River, Steens, and Owyhee basalts in Oregon, Washington, and Idaho: Am Jour. Sci., v 274, p. 148-189.

Wheeler, H.E., and Cook, E.F., 1954, Structural and stratigraphic significance of the Snake River Capture, Idaho-Oregon: Jour. of Geol., 62, p. 525-536.

White, D.E., Barnes, I., O'Neil, J.R., 1973. Thermal and mineral waters of nonmeteroic origin, California Coast Ranges: Geol Soc. Amer. Bull. v. 84, p.547-560.

Whitehead, R.L., 1981, U.S. Geol. Surv., Water Resources Division, Boise, Idaho, personal communication

Witkind, I.J., 1975, Preliminary map showing known and suspected active faults in Idaho: U.S. Geol Surv open-file rept no. 75-278, 71 p.

Wood, S.H., 1981, Unpublished geologic mapping, Boise Front: Dept. of Geol and Geophys., Boise State Univ

Wood, S.H., Mitchell, J.C., and Anderson, J.E., 1980, Subsurface geology and geothermal prospects in the Nampa-Caldwell area of the western Snake River Plain, Idaho: Transactions, Geothermal Resources Council, v. 4, p. 265-267

Wood, S.H., and Vincent, K., 1980, Geologic map of the Military Reserve Park area, Boise Foothills, Idaho. Boise State University, Dept. of Geol. and Geophys. contribution no. 50.

Woodside, W., and Messmer, J.H., 1961, Thermal conductivity of porous media, 2, consolidated rocks: Jour. of Applied Physics, Vol. 37 no 9, p. 1699-1706.

Wright, T.L., Grolier, M.J. and Swanson, D.A., 1973, Chemical variation related to stratigraphy of the Columbia River Basalt: Geol. Soc. America Bull., v. 84, p. 371-386.

Wright, T.L., Swanson, D.A., Helz, R.T., Byerly, G.R., 1979. Major oxide trace element, and glass chemistry of the Columbia River basalt samples collected between 1971 and 1979: U.S. Geol. Surv open-file report, 79-711, 13 p

Young, H W., and Mitchell, J.C., 1973, Geothermal investigations in Idaho, part 1, geochemistry and geologic setting of selected thermal water: Idaho Dept. of Water Resources Water Inf. Bull. no. 30, 43 p.

Young, H.W. and Whitehead, R.L., 1975a, Geothermal investigations in Idaho, part 2, an evaluation of thermal water in the BruneauGrandview area, southwest Idaho: Idaho Dept. of Water Resources Water Inf. Bull no. 30, 126 p.

Young, H.W., and Whitehead, R.L., 1975b, Geothermal investigations in Idaho, part 3, an evaluation of selected thermal water in Weiser area, Idaho: Idaho Dept. of Water Inf Bull. no. 30, 35 p.

Young, H.W., 1977, Reconnaissance of groundwater resources in the Mt. Home plateau area, southwest Idaho: Water Resources investigations 77-108, U.S. Geol. Surv. open-file rept., 40 p.

Young, H.W., Harenberg, W.A., and Seitz, H.R., 1977, Water Resources of the Weiser River Basin, west-central Idaho: Idaho Dept. of Water Resources Water Inf Bull. no. 44, 188 p.

Young, H W., and Lewis, R.L., 1980, Hydrology and geochemistry of thermal ground water in southwestern Idaho and north-central Nevada: U.S. Geol. Surv. open-file report 80-2043.