
Seismic Expression and Geological Significance of a Lacustrine Delta in Neogene Deposits of the Western Snake River Plain, Idaho¹

Spencer H. Wood²

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 km² (5000 mi²) 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,

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

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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 km². 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

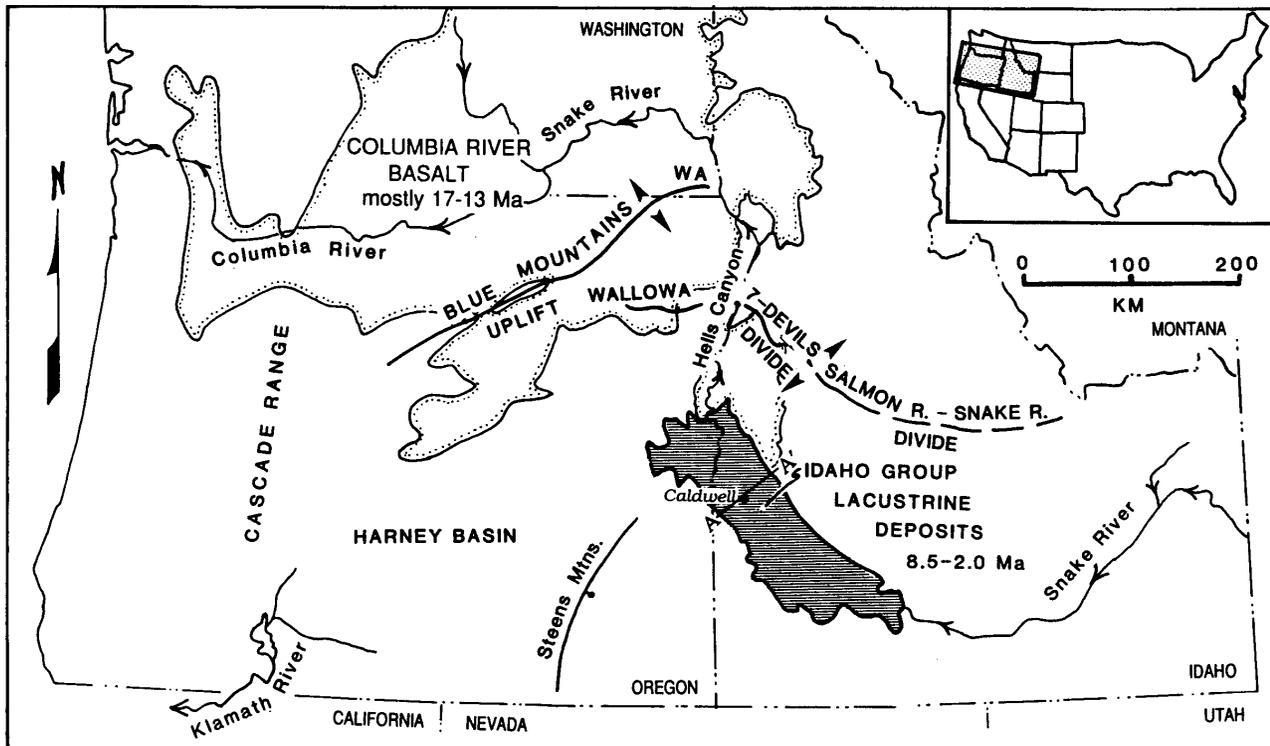


Figure 1—Index map showing distribution of Neogene Idaho Group lacustrine deposits in the western Snake River plain, location of physiographic features and drainages influencing the Neogene lakes that occupied the western plain region, and distribution of Columbia River basalt, shown by dotted line. 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-

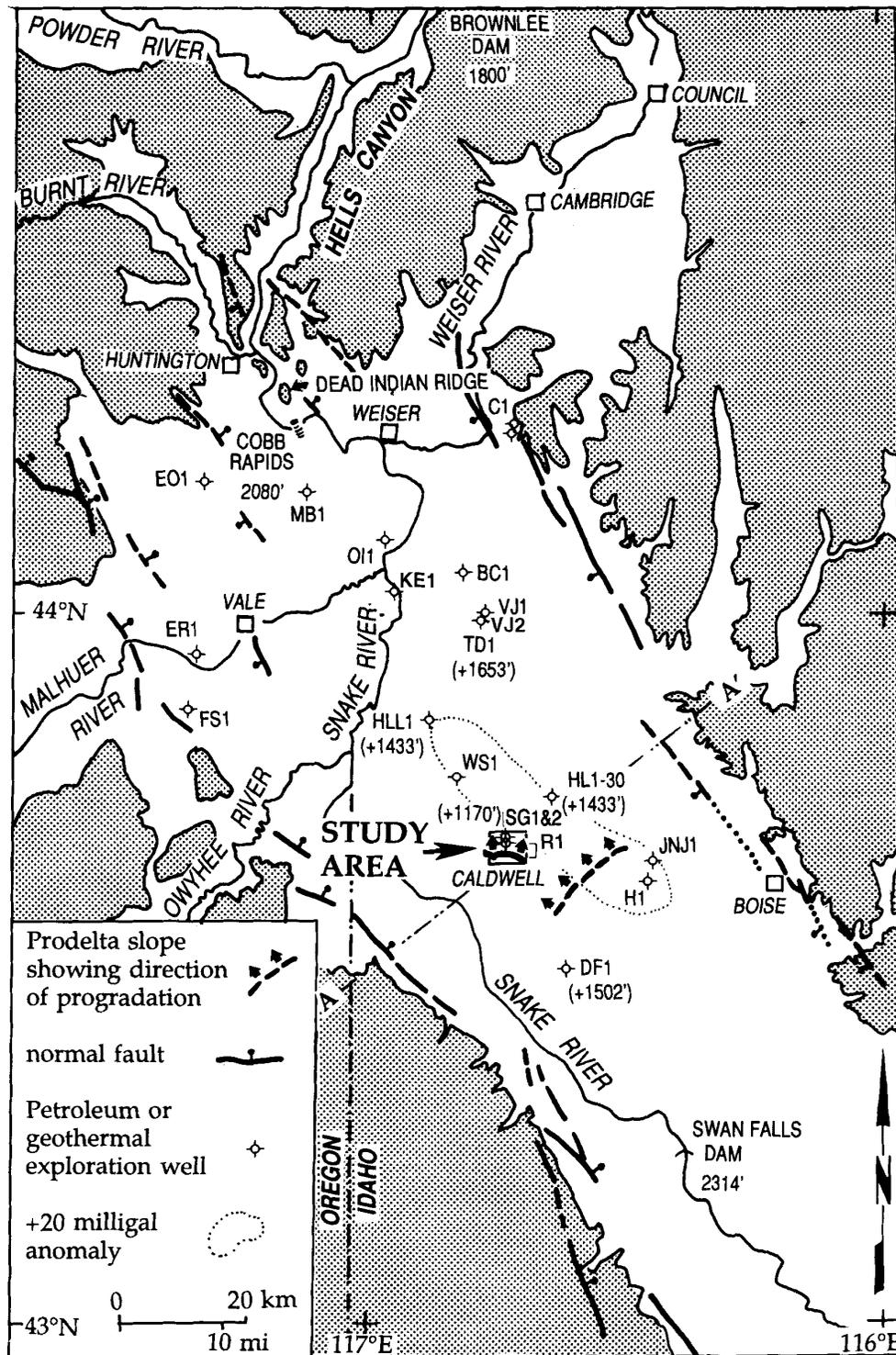


Figure 2—Map of western Snake River plain showing location of study area of Figure 5. Map shows position and progradation direction of the lacustrine delta system in the Pliocene interpreted from this study and from a previous study by Wood and Anderson (1981). Numbers in parentheses are the present 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 Bonnicksen (1989) alluded to a large Pliocene lake in the western plain, commonly referred to as Lake Idaho, but little is

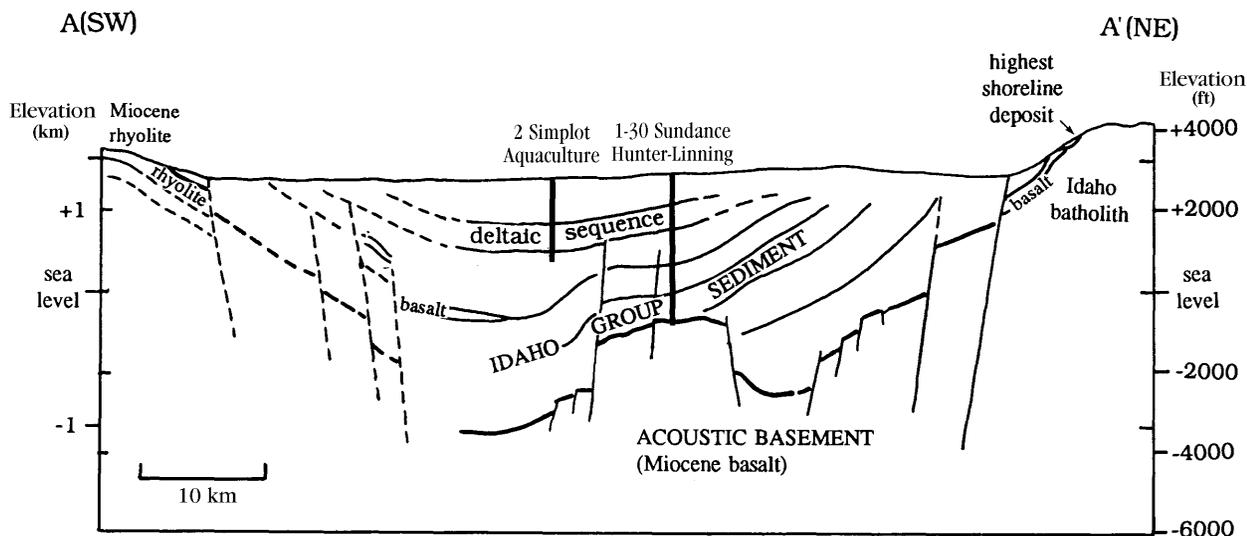


Figure 3—Structural features of the western plain in an exaggerated vertical-scale cross section showing position of the deltaic sequence discussed in this paper. Location of section is shown in Figure 2. Cross section shows that the plain has continued to subside since the progradation of delta sequence. 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 km², 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 km². 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/m² (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	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 -	UPPER MIOCENE	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+
		POISON CREEK FORMATION (tuffaceous sediment, arkosic sand, minor basalt)		400+
		IDA VADA 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+
		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+
24.0 -				

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 volcanoclastic 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 Glens 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 Glens Ferry, Malde (1972) described 660 m (2000 ft) of section of flood-plain and lacustrine facies. The fluvial 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

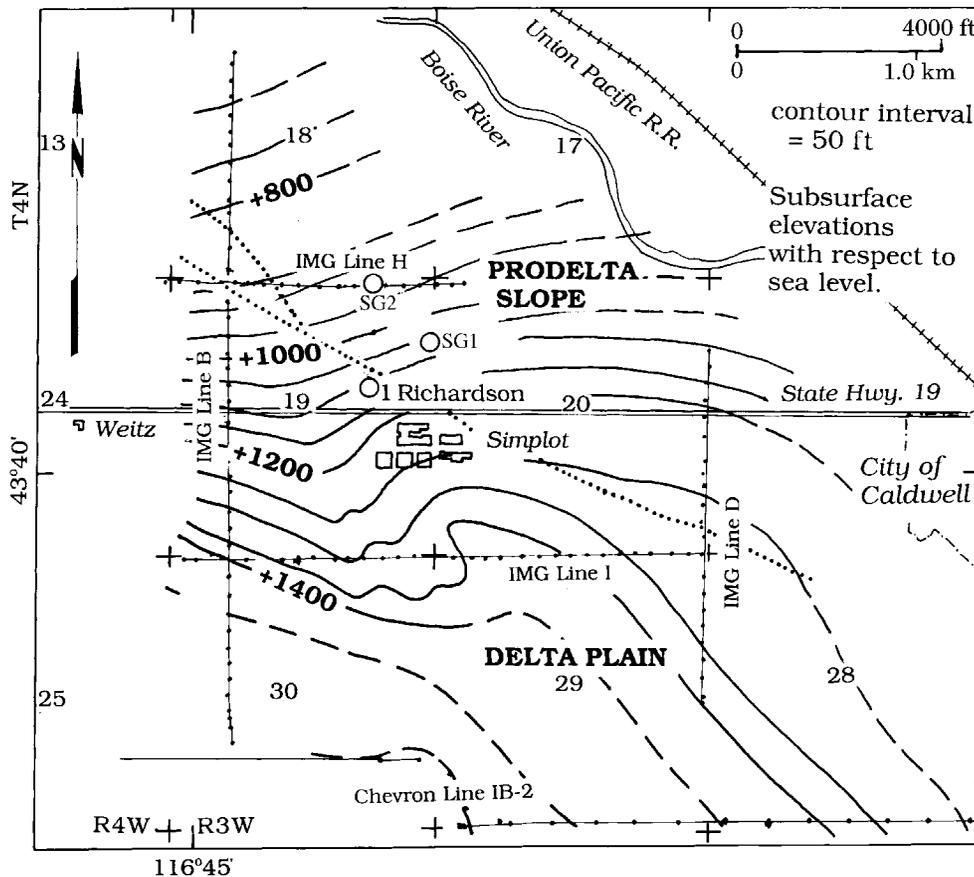


Figure 5—Location map of Caldwell study area showing structure contours of a reflecting horizon. Mapped reflector is shown on the interpreted section at about 0.5 s on IMG Line H (Figure 6b) as a heavy line within the prodelta clay facies. Configuration of the reflector is considered a depositional surface (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.

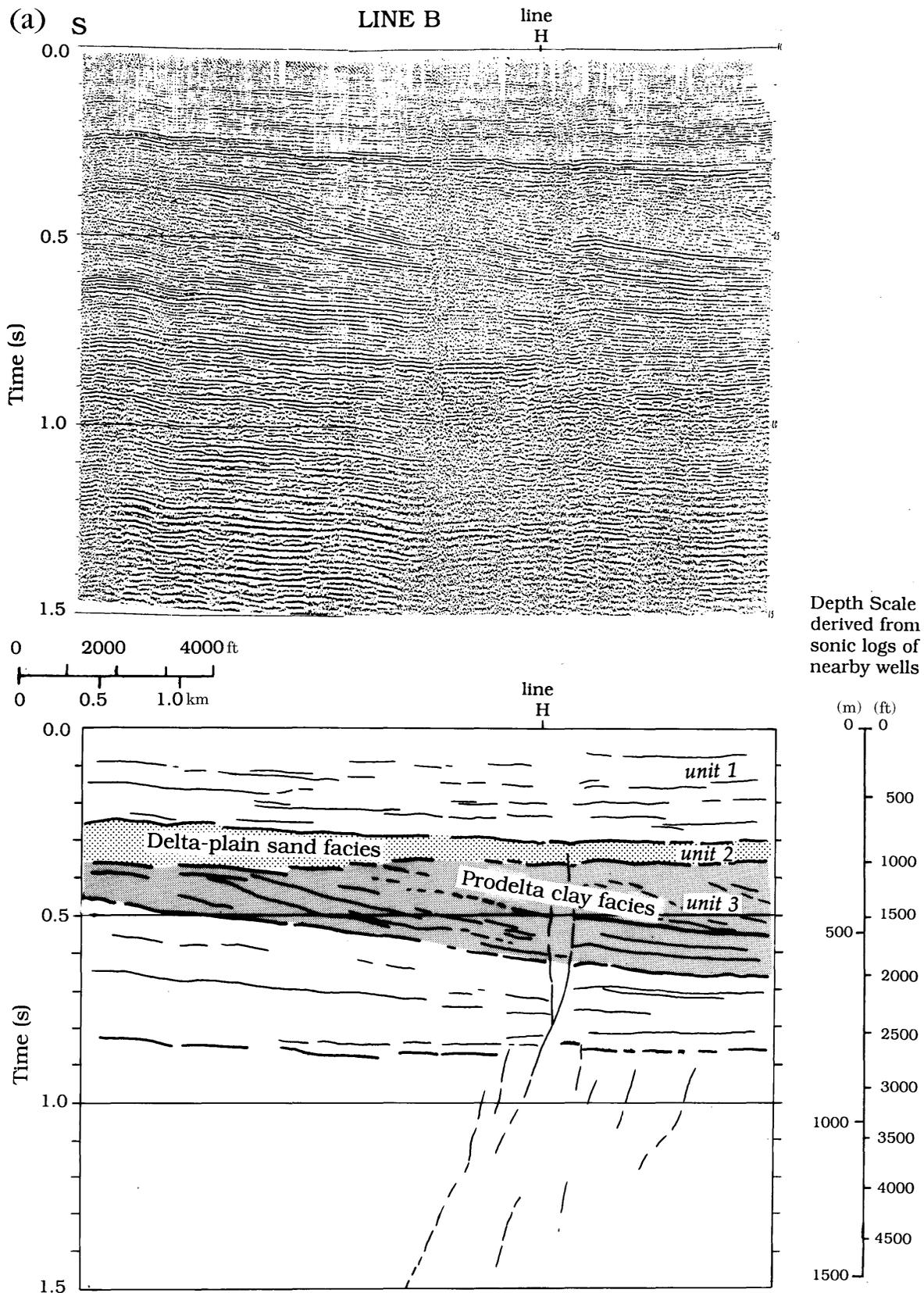


Figure 6—High-resolution seismic sections of IMG lines B and H (unmigrated). Location of lines shown in Figure 5. Datum is 762 m (2500 ft) above sea level. Elevation of ground level along lines is between 707 and 728 m (2320 and 2390 ft).

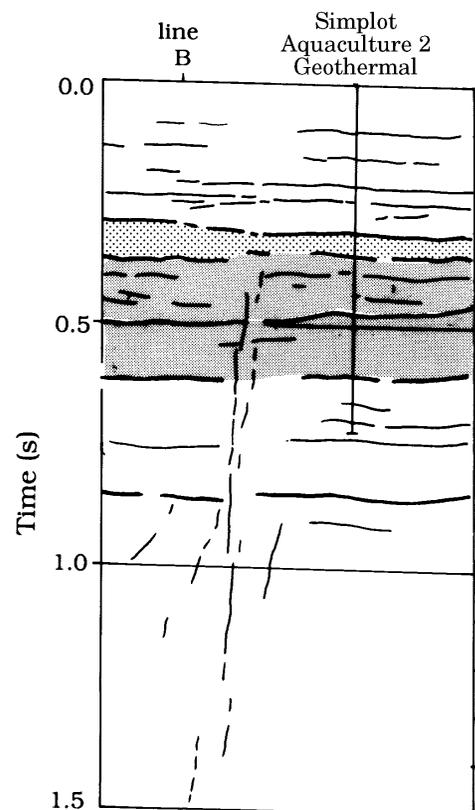
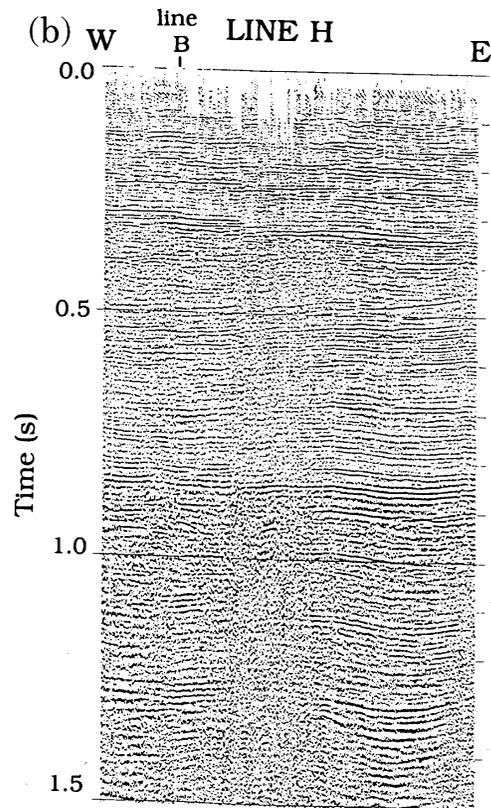


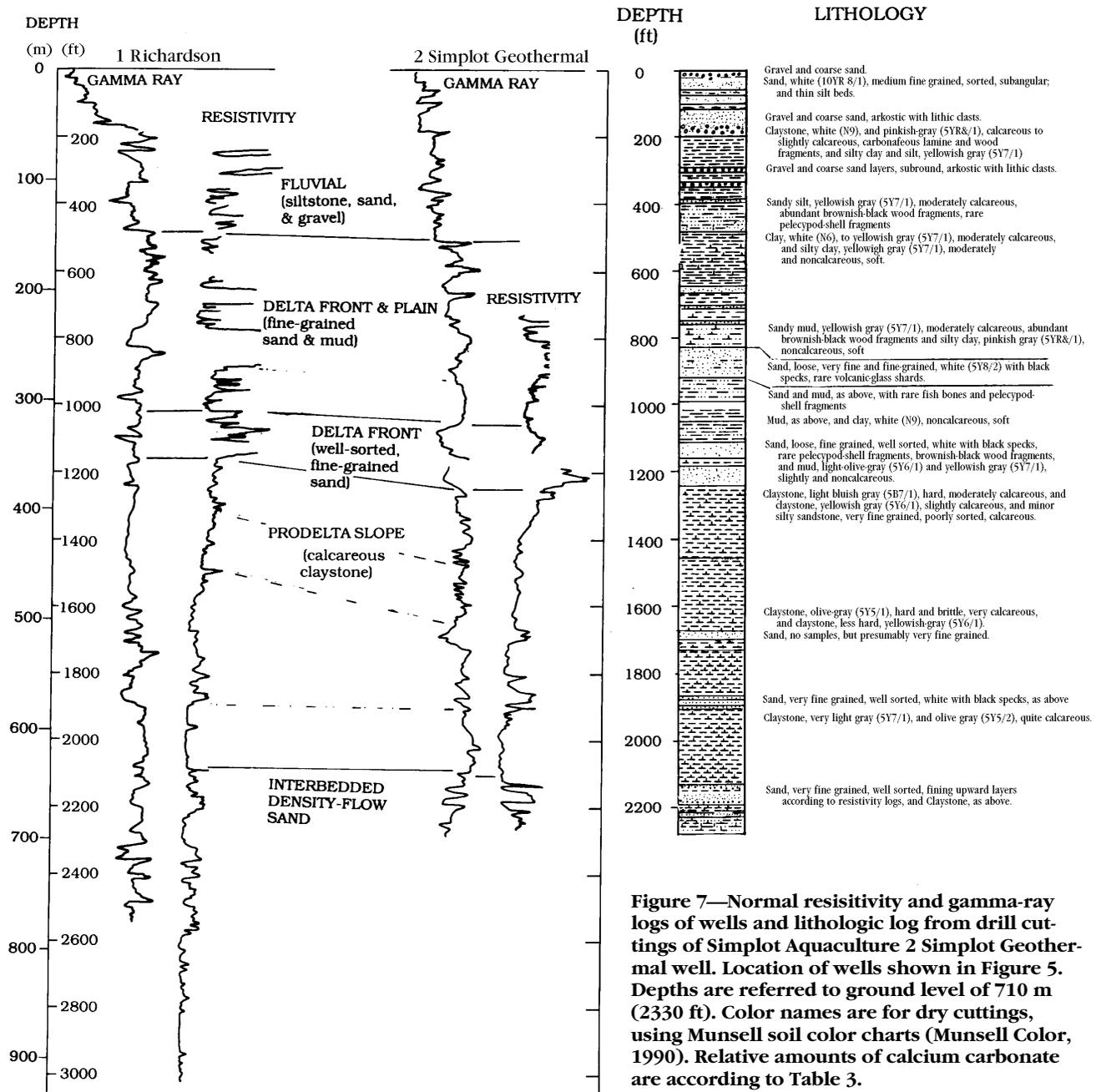
Figure 6—Continued.

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



a mean diameter of 0.1 mm ($\phi = 1$, 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-

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% CaCO₃).

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% CaCO₃).

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% CaCO₃).

Slightly calcareous: very slow reaction, bubbles evolve one at a time, with several seconds between release of each bubble (<10% CaCO₃).

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).

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-

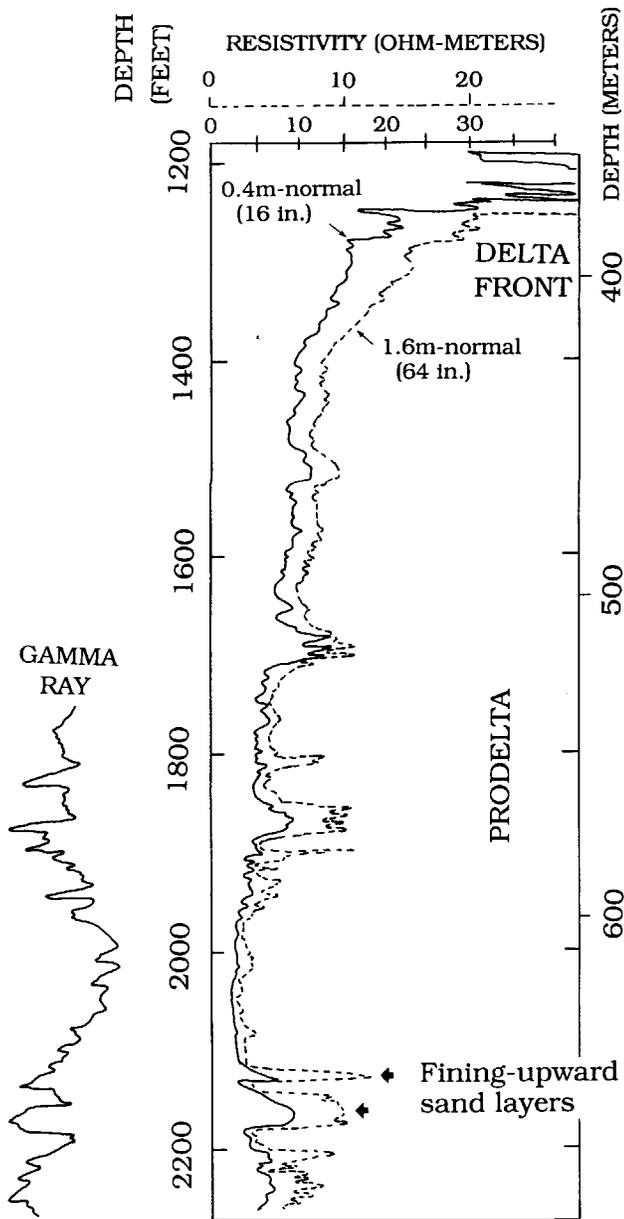


Figure 8—Detailed resistivity log of Simplot Aquaculture 2 Simplot Geothermal well showing fining-upward sand layers in the prodelta facies, the gradual upward-increasing resistivity in the prograding prodelta clays, and the abrupt contact and high resistivity of the delta-front sands.

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 interdeltic 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

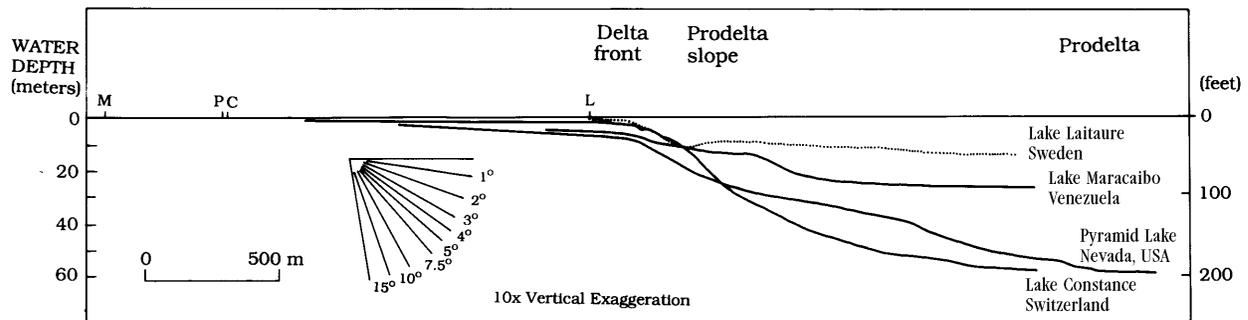


Figure 9—Profiles of modern lacustrine deltas showing the delta front and the prodelta slope. Data sources are Lake Constance (Müller, 1966); Lake Laitaure (Axelsson, 1967); Lake Maracaibo (Hyne et al., 1979); and Pyramid Lake (Born, 1972; Russell, 1885).

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.

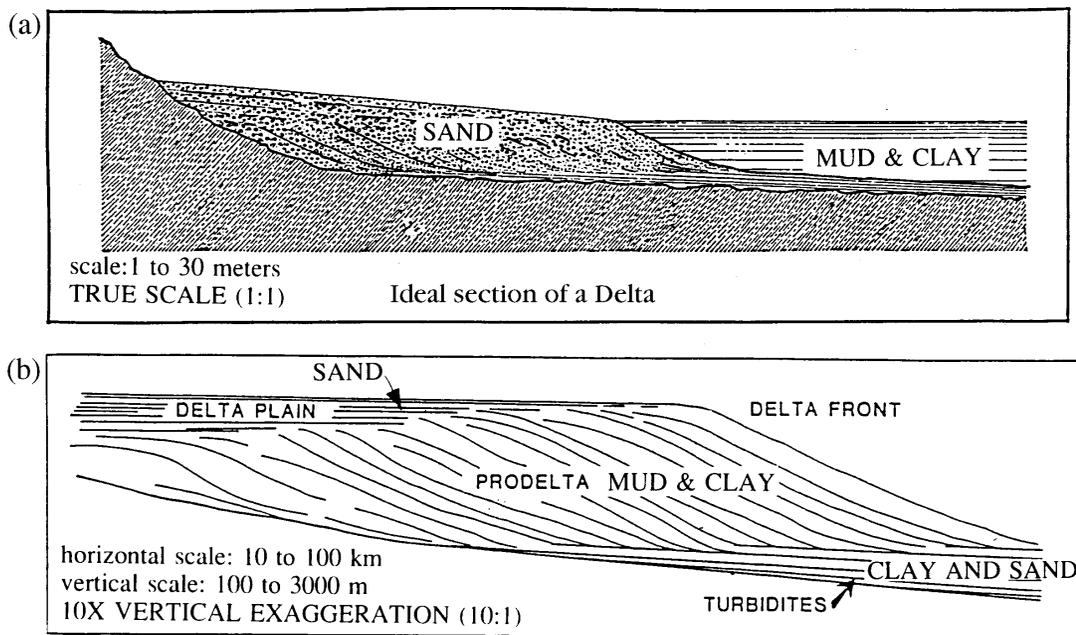
Beyond the delta front is the prodelta slope, formed mostly by the deposition of the suspended load of silt and clay, as illustrated in Figure 11. Slopes of the prodelta beds are generally a few degrees (Figure 9). The prodelta deposits, however, make up most of the volume of deposits of the fine-grained 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 16×10^9 m³/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 km² (5000 mi²) evaporation could have exceeded 13×10^9 m³/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 Glens 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 Glens 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 km² (1500 mi²). 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

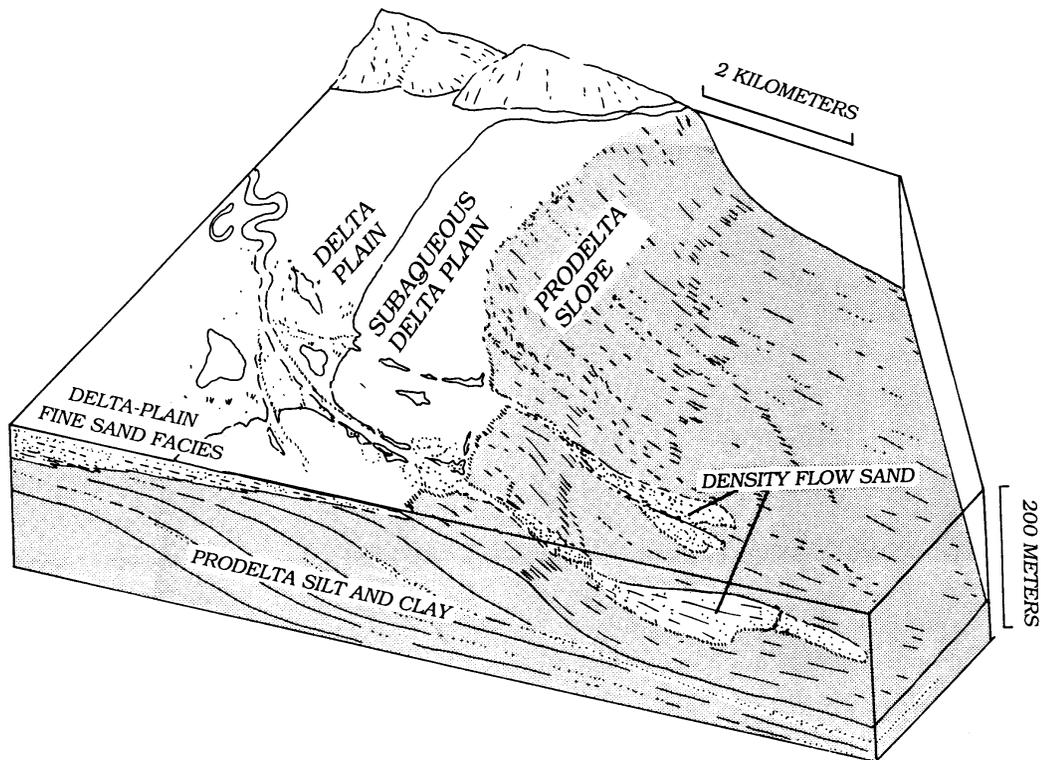


Figure 11—Drawing of a lacustrine delta environment, with a 10× exaggerated vertical scale, showing distribution of sand units within the dominantly mud and clay sequence.

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 down-

warping 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 km² (5000 mi²) had filled by the time the delta sequence had reached Caldwell, Idaho, so that the existing lake area was about 4000 km² (1500 mi²). 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.

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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.

