Hydrogeology of the Upper Deschutes Basin, Central Oregon: A Young Basin Adjacent to the Cascade Volcanic Arc

David R. Sherrod, U.S. Geological Survey, Hawaii National Park, Hawaii 96718; dsherrod@usgs.gov

Marshall W. Gannett, U.S. Geological Survey, 10615 SE Cherry Blossom Drive, Portland, Oregon 97216; mgannett@usgs.gov

Kenneth E. Lite, Jr., Oregon Water Resources Department, 158 12th Street NE, Salem, Oregon 97310, kenneth.e.lite@wrd.state.or.us

INTRODUCTION

The upper Deschutes Basin encompasses about 11,700 km^2 of the Deschutes River drainage basin in central Oregon (Fig. 1). Chiefly draining the east flank of the Cascade Range, the upper Deschutes Basin extends northward from a drainage divide near Chemult that separates it from the Klamath Basin to the south. The eastern margin of the basin lies along the south part of the Ochoco Mountains and through the crest of Newberry Volcano. The northern boundary is near Warm Springs, northwest of Madras.

The upper Deschutes Basin is underlain by Quaternary and Tertiary volcanic and sedimentary rocks. The occurrence and movement of ground water and the interaction of ground water and streams are controlled by the distribution of permeability within the depositional sequence. The permeability distribution reflects the age, lithology, and depositional environment of the strata, along with the subsequently imposed geologic structure.

The crest of the Cascade Range, including a broad upland area east of the Three Sisters, is the principal source of recharge for the ground-water system. The average annual rate of recharge from precipitation in the upper Deschutes Basin is estimated to be roughly 108 m^3/s (3,800 ft^3/s) (Gannett and others, 2001).

Ground water moves eastward from the Cascade Range and then generally northward through permeable Quaternary and upper Tertiary deposits. North of Madras, the permeable deposits thin out against relatively impermeable lower Tertiary deposits of the John Day and Clarno Formations (Fig. 2), forcing nearly all the northward-flowing ground water to discharge into the Deschutes River and its tributaries. This massive amount of ground-water discharge, exceeding 60 m^3/s (2,000 ft^3/s) near the confluence of the Deschutes and Crooked Rivers, is the principal reason for the remarkably stable flow of the Deschutes River.

Participants on this trip will explore the visible and conceptual aspects of the regional ground-water hydrology of the upper Deschutes Basin, including the interaction between ground water and streams. The trip follows the general direction of ground-water flow northward from the headwaters of spring-fed streams at the margin of the Cascade Range to the principal regional discharge area near Lake Billy Chinook.

This guidebook describes a 2-day trip. Day 1 begins in the La Pine Subbasin (uppermost Deschutes Basin) and proceeds through Bend, concentrating chiefly on the hydrologic controls created by Quaternary stratigraphy and structure (Fig. 3). Day 2 examines strata of the Deschutes Formation from Bend to Madras and the geologic factors that influence regional ground-water discharge.

A road log for each day is at the end of the field-trip guide. The metric system is used for all scientific aspects of the guidebook except altitude and water discharge rate, which are given in both meters and feet (altitude) and cubic meters per second and cubic feet per second (rate) owing to the widespread familiarity with U.S. traditional units in these matters. The road log is reported in miles to match most car odometers.

Field Guide to Geologic Processes in Cascadia: Oregon Department of Geology and Mineral Industries Special Paper 36, 2002.

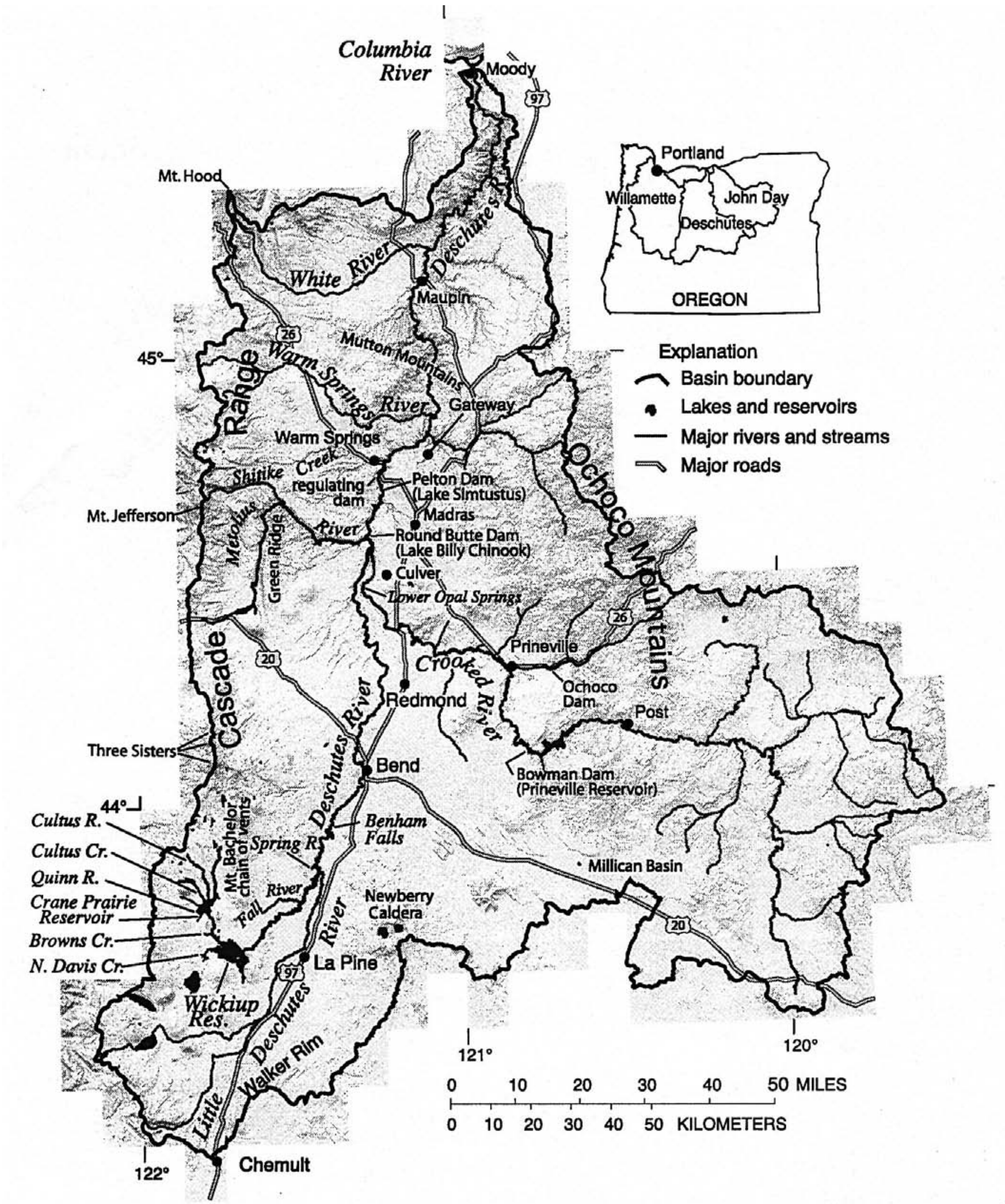


Figure 1. Extent of the Deschutes Basin (bold line), major rivers, and geographic features named in the text. Extent of the upper Deschutes Basin is shown on Figures 5 and 11.

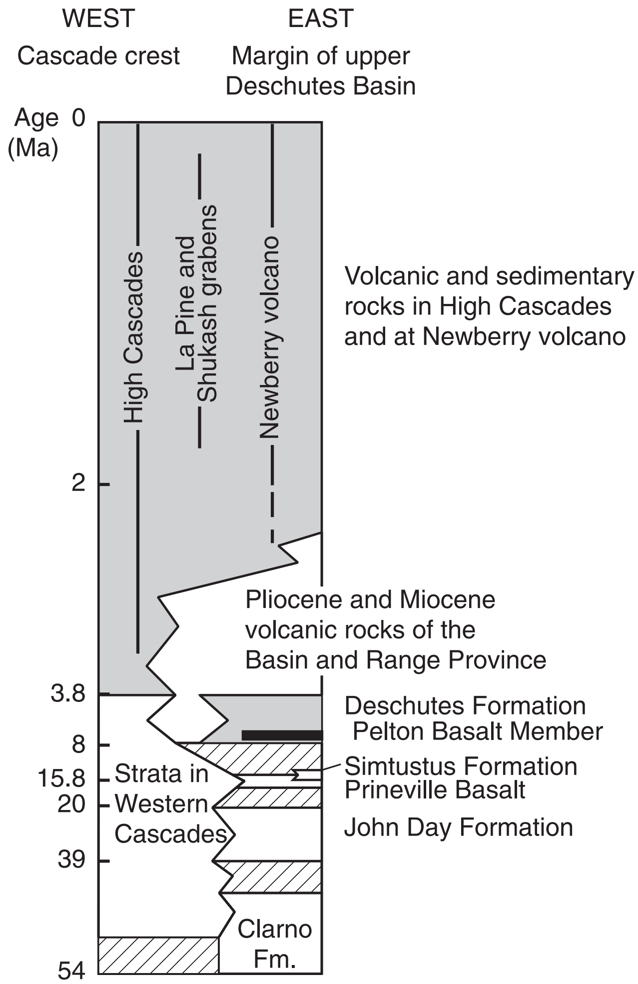


Figure 2. Generalized stratigraphic column for the upper Deschutes Basin. Shaded boxes show stratigraphic units visited on this trip. Open boxes, not visited. Diagonal pattern indicates nondeposition. Several names have descriptive geographic value but do not designate formally defined stratigraphic units.

Regional-scale geologic maps may aid travelers wishing a more thorough understanding of the geology along the trip route. Day 1 stops are within the area of the west half of the Crescent 1 by 2 Degree Quadrangle (MacLeod and Sherrod, 1992) and the Bend 30 by 60 Minute Quadrangle (Sherrod and others, in press). Maps of the Mount Bachelor volcanic chain (Scott and Gardner, 1992) and Newberry Volcano (MacLeod and others, 1995) provide more detail around the margins of Day 1 travel. A road map of the Deschutes National Forest is helpful for navigating the maze of Forest Service and county roadways during Day 1.

Most Day 2 stops are within the Bend Quadrangle (Sherrod and others, in press). The most northerly stops, however, are in an area that lacks regional-scale geologic mapping. Several 1:24,000-scale maps cover the northern part of the trip, including those by Smith (1987a, b), Smith and Hayman (1987), and Ferns and others (1996a). The Geologic Map of Oregon (scale 1:500,000, Walker and MacLeod, 1992) would also be helpful in the northern area.

ACKNOWLEDGMENTS

This acknowledgment has been 25 years in the making. Our deepest appreciation goes to Larry Chitwood, Rick Conrey, Mark Ferns, Kyle Gorman, Bob Jensen, Norm MacLeod, Bob Main, Andrei Sarna-Wojcicki, Willie Scott, Gary Smith, Ed Taylor, and George Walker for richly gestured discussions that have proved fundamental to our understanding of the geology and hydrology of the Deschutes Basin. Joy Gannett assisted in checking the driving instructions. This guidebook was reviewed for accuracy by Larry Chitwood and Bill McFarland and edited by George Moore for final publication. We thank them all.

DAY 1

Synopsis, Uppermost Deschutes Basin

Day 1 of this trip focuses on the uppermost Deschutes Basin, including the margin of the Cascade Range and the La Pine Subbasin (Fig. 3). This area is characterized at the surface by broad flatland areas and interspersed volcanic buttes. Relief increases toward the basin margins owing chiefly to volcanic buildup. The buttes, which consist of cinder cones, small shield volcanoes, and sparse silicic domes, are characteristic Cascade Range eruptive centers. From them have issued the numerous lava flows, predominantly basaltic andesite, seen in road cuts and river banks. A few tuff cones and maars resulted from eruptions during times when ground water saturated the upper strata or when shallow lakes occupied part of the basin. Silicic domes are scattered among the basalt and basaltic andesite vents; these range in composition from dacite to rhyolite. Pyroclastic-flow deposits are exposed sporadically, limited

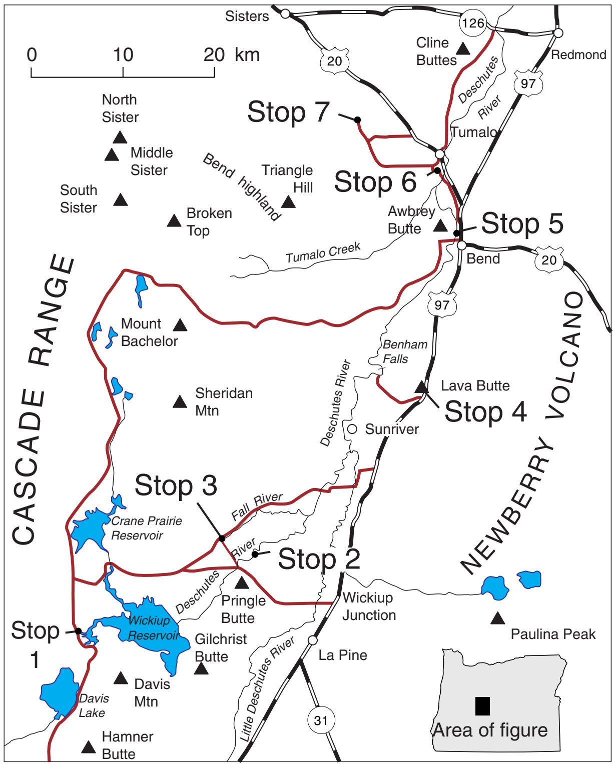


Figure 3. Route and stops for Day 1 of the field trip.

in extent, andesitic to rhyolitic in composition, and partially welded to nonwelded.

Sedimentary rocks, although rarely exposed, fill a substantial part of the La Pine Subbasin. The beds range from lacustrine to fluvial silt, sand, and gravel. We examine some of these beds at Pringle Falls (Stop 2). Sand- and gravel-rich strata are locally abundant and form important aquifers, especially near the town of La Pine.

Sedimentation occurred in the lowland created by the growth of adjacent volcanoes, as might be surmised by a casual glance at the landscape. Less obvious are within-basin horst-and-graben structures characteristic of the Basin and Range Province.

Precipitation and Orographic Effects

Much of the High Cascades in Oregon is a broad ridge whose topographic crest is defined by volcanoes along the axis of the Cascade Range. The range crest is a fundamental orographic barrier whose flanks receive precipitation from prevailing air flow across the North Pacific Ocean. The uppermost Deschutes Basin differs from other parts of the Cascade Range in Oregon by virtue of two additional substantial volcanic ridges east of the range crest—the Mount Bachelor chain of vents and Newberry Volcano.

Perhaps equally important hydrologically is a broad volcanic upland adjacent to the Cascade Range crest east of the South Sister. Known informally as the Bend Highland, this area encompasses 160 km^2 of terrain in excess of 1,830 m (6,000 ft) altitude, including Broken Top Volcano, Tam MacArthur Rim, and Triangle Hill. The infiltration of snowmelt and rainfall into the Three Sisters region and Bend Highland is the principal source of recharge to a ground-water system that dominates the hydrology of the Deschutes Basin. This extensive area of high altitude is unique in the Cascade Range in Oregon. In contrast, the upper Klamath River, which drains a comparably sized basin adjacent to the Cascade Range south of the Deschutes Basin has only about one-third of the mean discharge of the Deschutes River at Madras.

Ground-Water Flow in the Uppermost Basin

Much of the upper Deschutes Basin lacks a developed and integrated stream system. Permeability of the late Pleistocene and Holocene lava flows is high, so precipitation (including rapid snowmelt) tends to infiltrate the ground and not run off in established channels. Permeable lava of the Cascade Range encroaches eastward onto or interfingers with relatively impermeable fine-grained sedimentary strata in the La Pine Subbasin. Consequently, ground water flowing from the Cascade Range eastward through the permeable lava is likely to be diverted to the surface wherever it encounters fine-grained sedimentary strata. Ground water commonly issues abruptly from large springs in the lava along the west margin of the La Pine Subbasin, creating creeks and rivers flowing fully at 3 to 6 m^3/s (100-200 ft^3/s) within a few tens or hundreds of meters from their origin (Gannett and others, 2001).

Table 1. Altitude, Temperature, Discharge, and Isotopic Data for Selected Springs in the Upper Deschutes Basin (James and Others, 2000).

Spring name	Alt (m)	Alt (ft)	Temp (°C)	Disch (m^3)	$\delta^{18}O$ (‰)	$\delta^{18}O$ (‰)	^{14}C (% mod C)
Browns Creek	1332	4369	3.8	1.1	-13.9	-16.4	115.3
Cultus River	1356	4448	3.4	1.8	-14.1	-15.8	113.9
Fall River	1286	4219	6.1	4.2	-14.2	-11.8	110.9
Lower Opal Springs	597	1959	12.0	6.8	-15.3	-12.7	60.0
Metolius River	920	3018	8.2	3.1	-14.7	-11.5	61.3
North Davis Creek	1323	4340	3.4	—	-14.1	-15.8	114.0
Quinn River	1354	4442	3.4	0.7	-13.7	-16.5	112.3
Snow Creek	1378	4521	5.5	0.8	-14.1	-12.9	111.1
Spring River	1268	4160	8.0	3.5	-14.6	-11.5	67.5

Stop 1—Springs at North Davis Campground (Altitude 1,323 m; 4,340 ft)

Drive into North Davis Campground and try to find parking near Campsite 13. Walk north into the shallow drainage of North Davis Creek.

Numerous springs issue from lava flows exposed along North Davis Creek, a tributary to what is now the Davis Creek Arm of Wickiup Reservoir (Fig. 4). Streamflow measurements show that the Davis Creek Arm gains approximately 5.4 m^3/s (190 ft^3/s) from ground-water inflow over a distance of 3 km. Much of this flow comes from the spring complex of North Davis Creek seen at this stop. This spring complex is typical of several spring complexes along the margin of the Cascade Range in the southernmost Deschutes Basin. Similar full-fledged creeks take form over equally short distances along Browns Creek and at the heads of Quinn, Cultus, Fall, and Spring Rivers (Fig. 5). The total discharge from spring complexes in the area of Wickiup and Crane Prairie Reservoirs averages close to 17 m^3/s (600 ft^3/s).

The springs of North Davis Creek and similar springs in the area represent the westernmost (topographically highest) area of copious ground-water discharge in the Deschutes Basin. Temperature and isotopic data suggest that water emerging from these springs follows relatively short flow paths (James and others, 2000). Spring temperatures (about 3.5°C) essentially match the mean annual temperatures at the recharge altitudes as inferred from oxygen isotope ratios, indicating little or no geothermal warming (Table 1). Isotopes of carbon and noble gases show no evidence of magmatic volatile content (James and others, 2000).

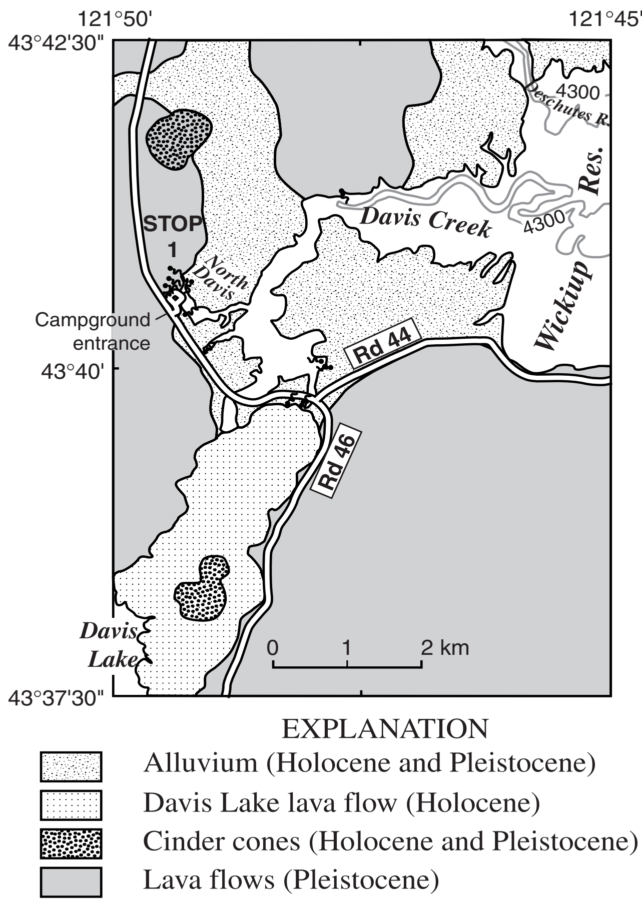


Figure 4. The area of North Davis Creek, Stop 1 (geology from MacLeod and Sherrod, 1992).

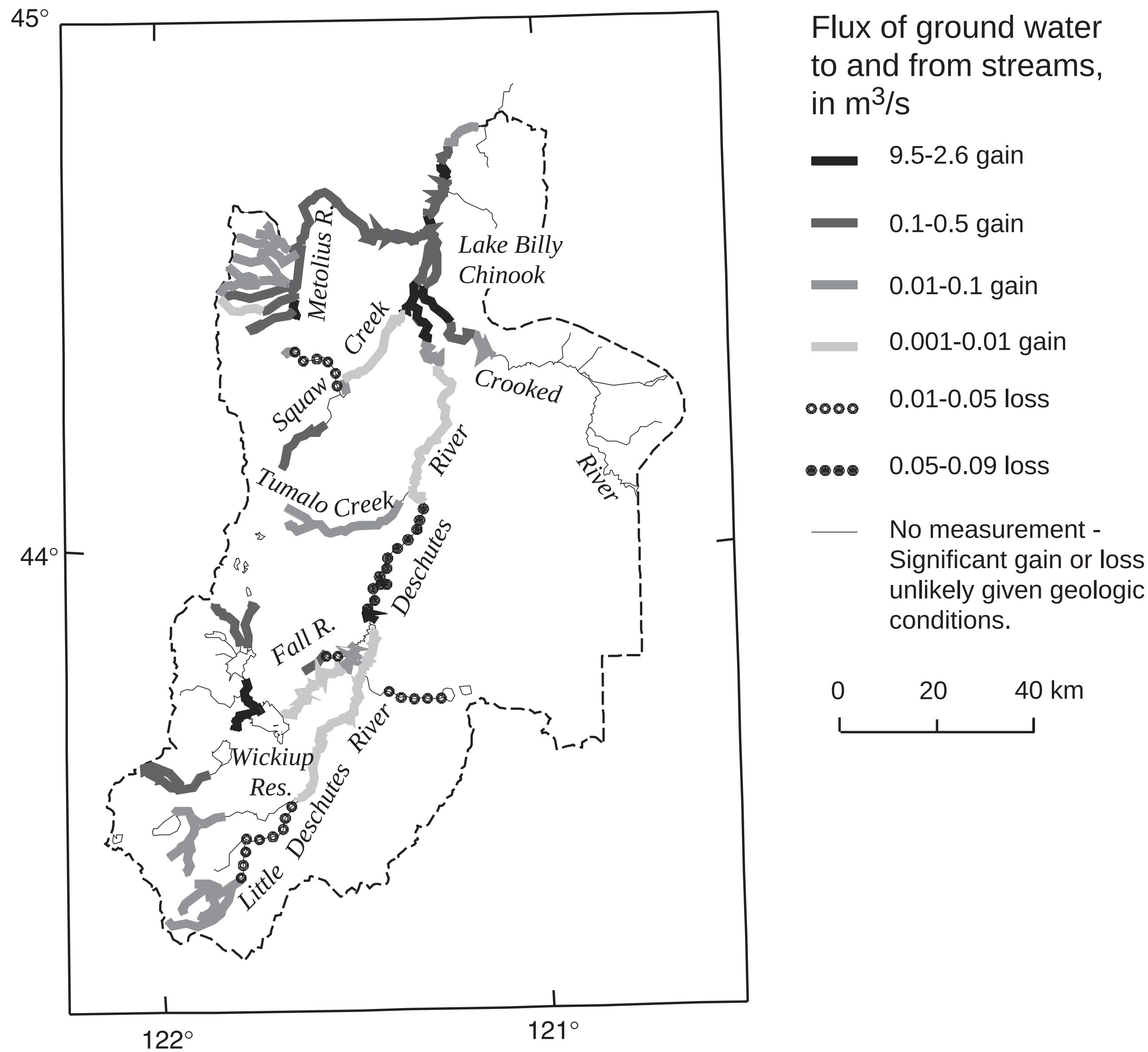


Figure 5. Estimated gain and loss rates for selected stream reaches in the upper Deschutes Basin, Oregon (from Gannett and others, 2001).

Stop 2—Pringle Falls Sedimentary Sequence (Altitude 1,295 m; 4,250 ft)

The Deschutes River at Pringle Falls has eroded through fine-grained sediment approximately 20 m thick. The falls themselves, more a dangerous rapid than a waterfall, are in an underlying basalt lava flow. The geologic exposure at Pringle Falls is unique because of the thickness of sedimentary strata exposed at the surface. Most of the upper basin is barely incised, which limits the extent of exposure for poorly indurated beds. The Mazama Ash, a Holo-cene tephra deposit erupted from Mount Mazama (Crater Lake), 100 km south-southwest, thickly blankets much of the area; it is about 70 cm thick in the Pringle Falls area. Even the few good sedimentary exposures are draped by colluvium from interbedded or capping lava flows. The floors of some reservoirs have been exposed during drought years, providing the only other sites where sedimentary strata may be viewed extensively.

Our stop is at the east end of the Pringle Falls area, one of two excellent exposures (Fig. 6). The other site is on private land to the west at the base of Pringle Falls. The deposits at Pringle Falls are thin-bedded, nearly

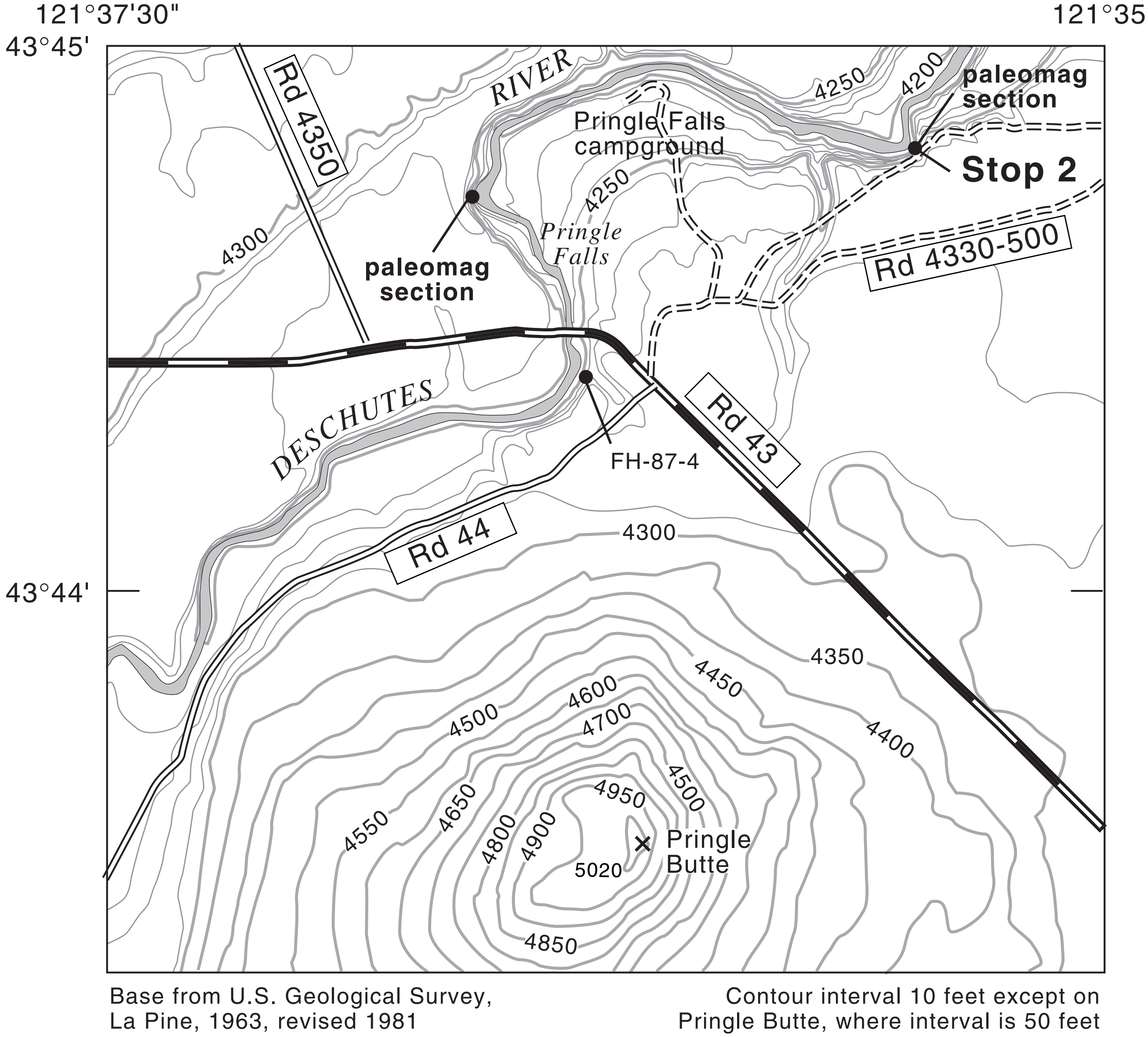


Figure 6. The Pringle Falls area, Stop 2.

undeformed lacustrine and fluvial mud, silt, and sand (Fig. 7). Gravel caps the sedimentary sequence. The beds are well exposed locally where the Deschutes River undercuts and maintains oversteepened banks.

The sedimentary sequence contains several fallout tephra layers. Pumice bed D, about 3.5 m above river level (Fig. 7), has an $^{40}Ar/^{39}Ar$ age of 218±10 ka (Herrero-Bervera and others, 1994; A.M. Sarna-Wojcicki, written commun., 2001). The lacustrine sequence also contains the Pringle Falls geomagnetic polarity episode, which occurred between about 218±10 and 169±17 ka (Herrero-Bervera and others, 1994). The latter age is obtained by regional correlation of lacustrine paleomagnetic data from Tule Lake, Calif. The polarity episode was determined from paleomagnetic directions of the lacustrine sediment at the west and east sites (Herrero-Bervera and others, 1989; Herrero-Bervera and Helsey, 1993).

The Pringle Falls sedimentary sequence is exposed on the north flank of a north-northeast-trending horst marked by Pringle and Gilchrist Buttes. This horst separates the La Pine Graben on the east from the Shukash Graben on the west (Fig. 8). The La Pine Graben is a lowland with numerous water wells that provide some understanding of the thickness and lithology of the strata there. Water wells in the La Pine

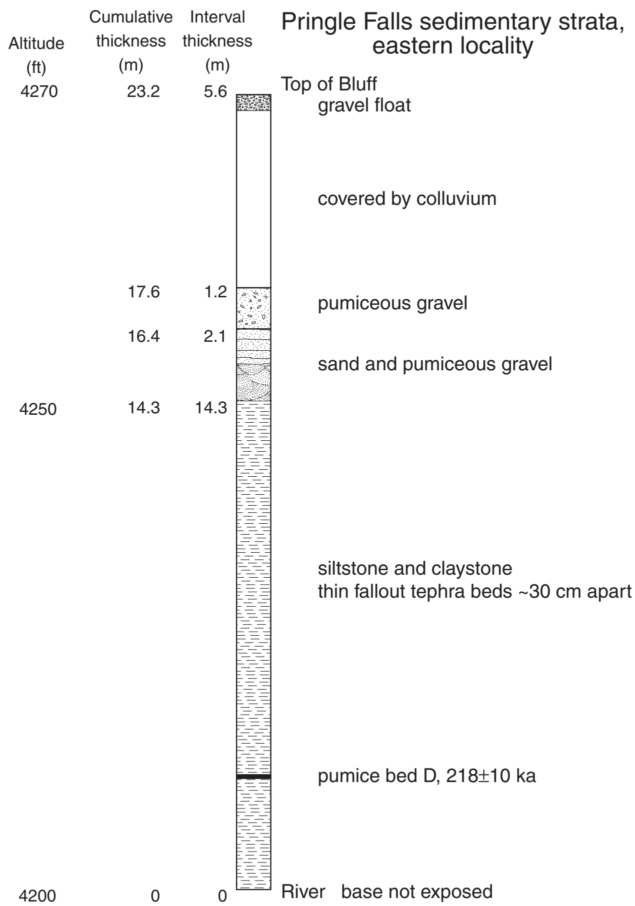


Figure 7. Stratigraphic section for the east exposure, Pringle Falls area, Stop 2.

Graben have penetrated as much as 395 m of sediment before encountering lava flows (Fig. 8; Couch and Foote, 1985). A well drilled at the Fall River Fish Hatchery, roughly 5.8 km north of Pringle Falls, penetrated fine-grained sediment to a depth of 150 m, and below that, interbedded basalt and cinders to a depth of 172 m. The thickest parts of the sedimentary fill, which may be as thick as 0.7-1.0 km (Gettings and Griscom, 1988), probably coincide with a -150 mGal gravity anomaly (Fig. 8). Deep holes drilled on the upper west flank of Newberry Volcano encountered no lacustrine deposits, and the alluvial strata on the volcano are complexly interbedded with primary volcaniclastic deposits. Therefore the volcano probably has formed an eastern buttress to basin-filling sediment for a substantial period of time (MacLeod and others, 1995). In contrast, the Shukash Graben has been obscured topographically by the growth of small shield volcanoes and cinder cones (Fig. 8). No deep water-well data are available from the Shukash Basin.

Although undated, the inception of the La Pine and Shukash Grabens likely began in late Pliocene or early Pleistocene time. The Chemult Graben, which may be of similar age, lies 30 km to the south (Fig. 8). An age from a lava flow capping the Chemult Graben's east rim (Walker Rim) is 2.33 Ma (Table 2), suggesting that the Chemult, and by analogy, the La Pine Grabens are younger. An age of 0.61 Ma was obtained from Gilchrist Butte, a normal-polarity shield at the south end of the horst between the La Pine and Shukash Grabens (Table 2). Gilchrist Butte lava has been faulted, so at least some of the faulting is younger than 0.6 Ma.

Other ages from the Pringle Butte–Gilchrist Butte Horst are thought meaningless, or at best, difficult to interpret. Ages ranging from 6.7 to 8.2 Ma were obtained from lava flows exposed on the horst (Hawkins and others, 1989). One of these samples was collected 150 m upstream from the Pringle Falls road bridge (FH-87-4, Fig. 6); it yielded an age of 6.9±0.8 Ma. These ages are much too old in view of the geomorphic landforms preserved by the volcanic vents along the horst. Gilchrist Butte buries most of the rhyolite dome of Eaton Butte, age 3.68±3.3 Ma (Table 2), which lies on its west flank. The large analytical error renders this age almost useless for determining the age of bedrock in the Pringle Butte–Gilchrist Butte Horst.

The geographic and geologic setting of the Pringle Falls sedimentary sequence suggests that a natural dam for the La Pine Subbasin may have once existed 23 km northeast, where lava flows from Newberry Volcano have ponded against Cascade Range lava flows north of Sunriver. Subsequent deformation along faults that bound the Pringle Butte–Gilchrist Butte Horst has probably elevated the sedimentary section slightly. This interpretation relies on a comparison of altitudes around the basin: 4,200 ft, where the dam might have existed north of Sunriver, and 4,250 ft at the top of the Pringle Falls lacustrine section (Fig. 6). The basin between Pringle Falls and Sunriver lacks remnant geomorphic surfaces that might correspond to an

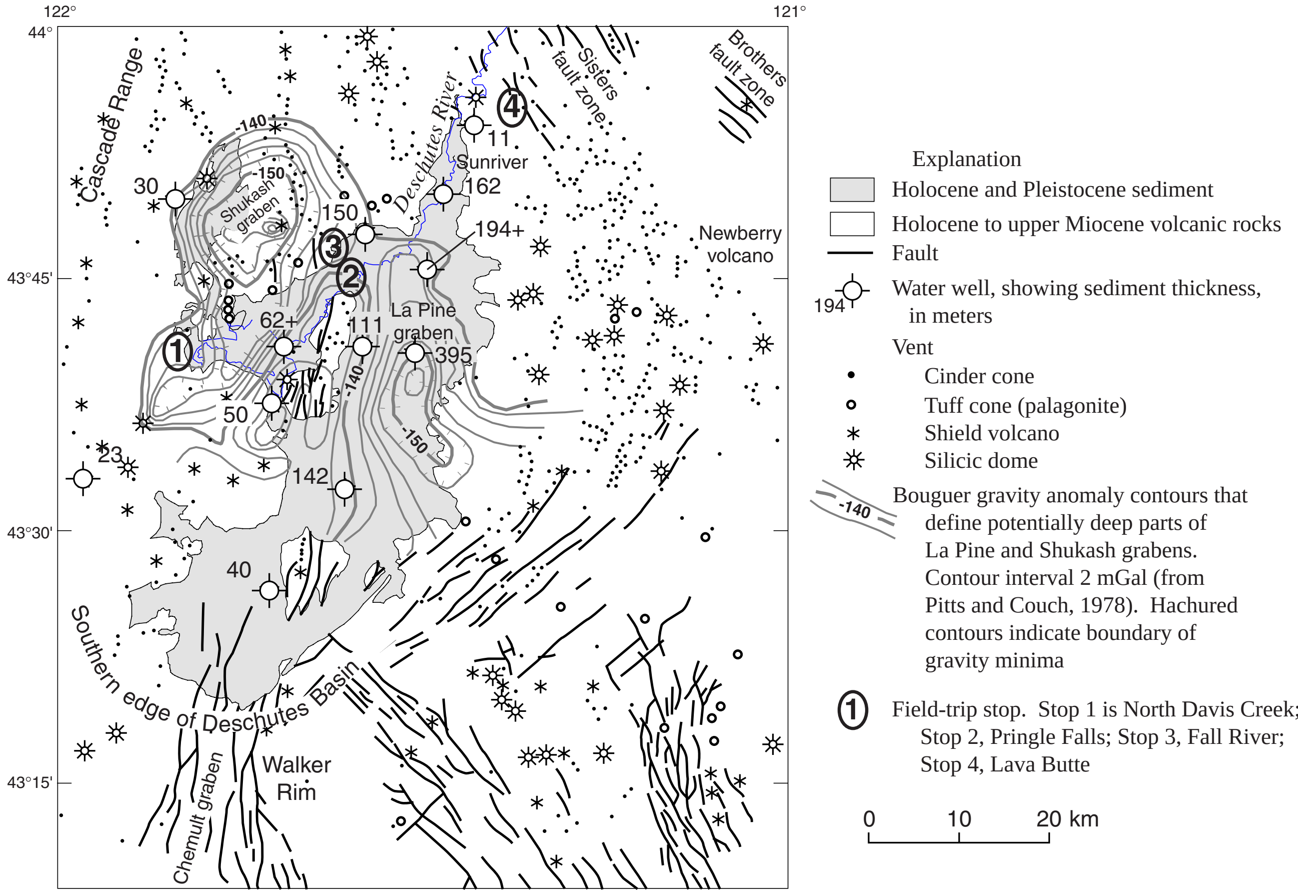


Figure 8. Map showing water wells, volcano-tectonic structure, and extent of grabens defined by gravity anomalies in the upper Deschutes Basin (from MacLeod and Sherrod, 1992).

extensive sediment sequence with an upper surface at 4,250-ft altitude; hence our reasoning that the Pringle Falls section has been elevated in response to uplift on the horst.

The fine-grained sediment exposed at Pringle Falls and inferred at depth from drilling data and geophysical studies is hydrologically significant. The low permeability of the sediment contrasts sharply with that of the lava flows and pyroclastic deposits to the east and west. As mentioned previously, the numerous springs on the west margin of the La Pine Subbasin are believed to result from eastward-flowing ground water diverted to the surface upon encountering these low-permeability deposits. The low permeability affects flow in the vertical direction as well. The area underlain by these fine-grained sedimentary deposits is characterized by a shallow water table, with the depth to water commonly less than 3 m.

Stop 3—Fall River Headwater Springs (Altitude 1,305 m; 4,280 ft)

Fall River is one of the larger spring-fed tributaries to the Deschutes River. Virtually the entire flow of Fall River originates from springs near its headwaters. Fall River springs emerge from Pleistocene basaltic andesite lava at the contact with Pleistocene alluvial and lacustrine sediment, presumably similar to deposits exposed at Pringle Falls (Stop 2). The Fall River headwaters are one of two spring complexes along the northwest margin of the La Pine Graben that discharge water that has followed intermediate-length flow paths. The other spring complex is that of Spring River, 16 km northeast of the springs at Fall River.

Temperatures of the Fall River and Spring River springs (6.1 and 8.0°C respectively, Table 1) are significantly warmer than mean annual temperatures at recharge altitudes inferred from oxygen isotope ratios, indicating the water

Table 2. Potassium-argon and $^{40}Ar/^{39}Ar$ isotopic ages of volcanic materials discussed in the text

Age (Ma)	Quality	Geologic unit or geographic location	Lat. (N)	Long. (W)	Rock type	Material dated	Reference
Volcanic rocks of High Cascades or Basin and Range provinces							
0.093±0.011	+	South Sister, early-erupted lava on northeast flank	44°08.29'	121°43.20'	Dacite	Whole rock	Hill and Duncan, 1990; Hill, 1992
0.61±0.05	+	Gilchrist Butte	43°38.75'	121°39.15'	Basalt	Whole rock	Sherrod and Pickthorn, 1989
*1.19±0.08	+	Bas. of The Island, intracanyon flows in Crooked River	44°25.96'	121°14.48'	Basalt	Whole rock	Smith, 1986a
1.43±0.33	+	Black Butte, 1 km northeast of summit	44°24.41'	121°37.53'	Basaltic andesite	Whole rock	Hill and Priest, 1992
2.33±0.09	+	Walker Rim (east side, Chemult graben)	43°16.20'	121°44.05'	Basalt	Whole rock	Sherrod and Pickthorn, 1989
3.68±3.3	-	Eaton Butte	43°38.6'	121°41.2'	Basalt	Whole rock	Fiebelkorn and others, 1983
Pliocene basalt and basaltic andesite							
2.9±0.2	+	Summit, Squaw Back Ridge shield volcano	44°28.70'	121°28.59'	Basaltic andesite	Whole rock	Armstrong and others, 1975
*3.56±0.30	+	Basalt of Redmond	44°23.20'	121°13.02'	Basalt	Whole rock	Smith, 1986a
Deschutes Formation							
3.97±0.05	+	Basalt of Round Butte	44°37.29'	121°11.80'	Basalt	Whole rock	Smith, 1986a
4.7±0.1	+	Deschutes Formation lava, near Bull Spring	44°06.63'	121°29.22'	Basaltic andesite	Whole rock	Hill, 1992
4.9±0.4	+	Lava flow at top of Deep Canyon grade (Oreg. Hwy 126)	44°17.38'	121°24.80'	Basaltic andesite	Whole rock	Armstrong and others, 1975
*5.06±0.03	+	Bas. andesite of Steamboat Rock, Deschutes Formation	44°21.45'	121°16.03'	Basaltic andesite	Whole rock	Smith, 1986a; Smith and others, 1987a
*5.31±0.05	o	Basalt of Tetherow Butte	44°32.70'	121°15.07'	Basaltic andesite	Whole rock	Smith, 1986a
*5.43±0.05	o	Basalt of Lower Desert	44°31.05'	121°18.57'	Basalt	Whole rock	Smith, 1986a
*5.77±0.07	+	Basalt of Opal Springs, Deschutes Formation	44°26.11'	121°14.50'	Basalt	Whole rock	Smith, 1986a; Smith and others, 1987a
*6.14±0.06	+	Rhyolite of Cline Buttes, Deschutes Formation	44°15.89'	121°17.50'	Rhyolite	Plagioclase	Sherrod and others, in press
*6.74±0.20	+	Rhyodacite southwest of Steelhead Falls, Deschutes Fm.	44°23.21'	121°22.44'	Rhyodacite	Plagioclase	Sherrod and others, in press
*7.42±0.22	+	Pelton Basalt	44°39.97'	121°12.10'	Basalt	Whole rock	Smith and others, 1987
John Day Formation							
*28.82±0.23	+	Gray Butte rhyolite, basal vitrophyre	44°24.30'	121°06.79'	Rhyolite	Anorthoclase	Smith and others, 1998
28.3±1.0	+	Powell Buttes dome	44°11.86'	120°58.03'	Rhyolite	Anorthoclase	Evans and Brown, 1981
*27.62±0.63	+	West end of Haystack Reservoir, quarry	44°29.87'	121°09.37'	Welded tuff	Sanidine	Smith and others, 1998
*29.57±0.17	+	West end of Haystack Reservoir, quarry	44°29.76'	121°09.36'	Welded tuff	Sanidine	Smith and others, 1998
*29.53±0.09	+	West end of Haystack Reservoir, quarry	44°29.85'	121°09.28'	Hydromagmatic tuff	Sanidine	Smith and others, 1998
30.8±0.5	+	Northeast of Gray Butte	44°26.10'	121°05.40'	Basalt (intrusion)	Whole rock	Fiebelkorn and others, 1983
*32.49±0.30	+	Tuff of Rodman Spring, at Rodman Spring	44°27.83'	121°06.91'	Fallout lapilli tuff	Sanidine	Smith and others, 1998

Grouped by stratigraphic unit and generally arranged from youngest to oldest. Ages are K-Ar unless preceded by *, which are 40Ar/39Ar. Symbols indicating "quality" show usefulness of age in stratigraphic interpretations: +, age thought meaningful; o, age probably meaningful but accuracy may be far poorer than indicated by the reported precision; -, age meaningless (owing to large analytical error) or incorrect (on basis of our knowledge obtained by all ages and regional stratigraphic relationships). Positional datum is NAD 27. Lithologically, material dated is lava except where noted

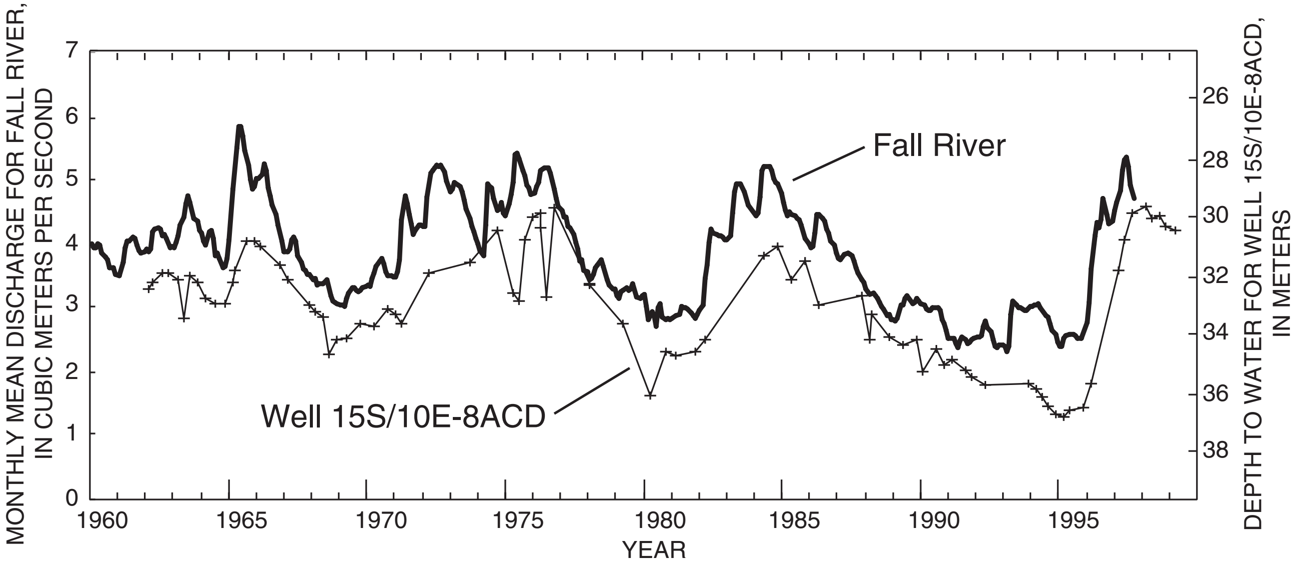


Figure 9. Hydrograph showing monthly mean streamflow of Fall River and the depth to water in a well near the margin of the Cascade Range at Sisters, about 60 km to the north.

has picked up geothermal heat along its flow path (Manga, 1998; James and others, 2000). Analysis of isotopes of carbon and noble gases indicates that the water of Spring River contains magmatically derived volatiles (James and others, 2000).

Stream-gage data show that Fall River has little seasonal variation; mean monthly flow varies only about 5 percent from the mean annual flow. On a decadal scale, however, the flow of the Fall River springs varies by a factor of more than two (Fig. 9). These decadal fluctuations correspond to climate cycles and snow-pack trends in the Cascade Range (Gannett and Lite, 2000; Gannett and others, 2001). Climate fluctuations represented by discharge variations in Fall River springs represent the largest transient signal in the hydrology of the upper Deschutes Basin. This pattern of variation occurs in spring discharge and water-table fluctuations throughout the basin (for example, Fig. 9).

Stop 4—Top of Lava Butte (Altitude 1,529 m; 5,016 ft)

A couple hundred large and small volcanoes and nearly two-thirds of the upper Deschutes Basin may be seen from the splendid vista atop Lava Butte. A simplified compass wheel shows the declination and distance to many features visible on a clear day from the lookout (Fig. 10).

Lava Butte itself is a Holocene cinder cone that erupted basaltic andesite lava flows. It lies at the downslope end of Newberry Volcano's northwest rift zone, which was active most recently between about 5,900 and 6,400 ^{14}C yr BP (MacLeod and others, 1995). A radiocarbon age of 6,160±70 ^{14}C yr BP was obtained from charcoal collected beneath Lava Butte's tephra plume where exposed in a highway road cut northeast of the butte (Chitwood and others, 1977).

Lava flows of the northwest rift zone merge with lava flows from the Cascade Range, forming the northern limit of the La Pine Subbasin today and probably for a substantial part of the subbasin's history. As mentioned at the Pringle Falls stop, lava flows probably formed natural dams in the past. Emplacement of the Lava Butte flow about 7,100 calendar years ago diverted the Deschutes River, forcing the river to establish a new channel along the west margin of the lava flow.

The topographic gradient increases abruptly north of this contact between the lava of Newberry Volcano and the Cascade Range, from relatively flat in the sediment-filled basin south of here to northward sloping to the north. The stream gradient reflects this topographic change. The Deschutes River drops a mere 0.48 m/km along the 71-km reach between Wickiup Reservoir and Benham Falls. North of Benham Falls the gradient steepens dramatically to 8.7

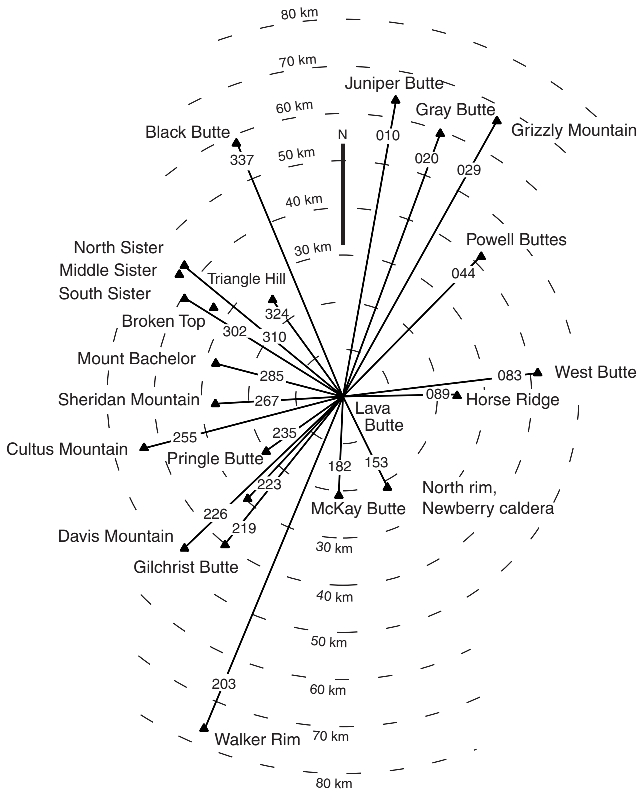


Figure 10. Compass wheel showing direction and distance to selected geographic features visible from the top of Lava Butte, Stop 4.

m/km along the 18.5-km reach to Bend.

The slope of the water table also increases north of Benham Falls, but to an even greater degree. The water table is roughly 3 m below the land surface in much of the La Pine subbasin. Starting at Sunriver, 8 km south of Lava Butte, the water table slopes northward so steeply that at Bend the depth to water is 170-200 m below the land surface. Near the river at Benham Falls, the water table is at an altitude of about 1,260 m (4,135 ft) and 1.5 to 5 meters below the land surface (Fig. 11). In downtown Bend, the water table is at an altitude of 940 m (3,080 ft) and the land surface is 1,110 m (3,630 ft)

The northward-increasing depth to water in this area has implications for the interaction of ground water and streams. South of Sunriver, the Deschutes River system commonly gains due to ground-water discharge, and major spring complexes are common. North of Sunriver, the streams lose water to the ground-water system as ground-water levels drop far below stream levels. For example, stream-gage data from the 1940s and 1950s showed that the Deschutes River lost an average of 0.68 m^3/s (24 ft^3/s) between Sunriver and Benham Falls. The average loss between Benham Falls and Lava Island, about 12 km downstream, was 2.3 m^3/s (83 ft^3/s). Most of the loss likely occurs where the channel crosses or is adjacent to the lava flows from Lava Butte. This lava is sufficiently young that fractures have not been sealed by sediment. Water easily leaks through the stream bed or channel walls into underlying lava flows.

In stops farther north we will see that conditions are reversed, and the regional water table is above river level, causing ground water once again to discharge to streams. The relation between topography and ground-water level in the upper Deschutes Basin is shown diagrammatically in Figure 12.

The rate of leakage from the Deschutes River in the Benham Falls area is proportional to the river stage and hence to streamflow (Gannett and others, 2001). The higher the stage, the greater the rate of loss. The ground-water level near the river varies in response to variations in the leakage rate. Continuous-recorder hydrographs show how the water table responds to changes in streamflow (Fig. 13). The stage and discharge in the Deschutes River in this reach are controlled by reservoir operations upstream. Streamflow is highest from April to October as water is routed down from the reservoirs to canal diversions near Bend. The water level in well 19S/11E-16ACC, about 150 m from the river near the Benham Falls parking area, rises and falls rapidly in response to river stage. Abrupt changes in streamflow usually become apparent in the well within a few to several days. These effects are much less pronounced, however, in wells farther from the river. The water level in well 18S/11E-21CDD, about 1.6 km from the river, also fluctuates in response to river stage, but the fluctuations are subdued, and the hydrograph is nearly sinusoidal, showing only the slightest inflections in response to abrupt changes in streamflow. In addition, the peaks and troughs in the hydrograph of well 18S/11E-21CDD lag those of well 19S/11E-16ACC by 1 to 2 months (Fig. 13).

Lithologic and geophysical logs from wells provide limited stratigraphic information for the Lava Butte area. The drillers' log from

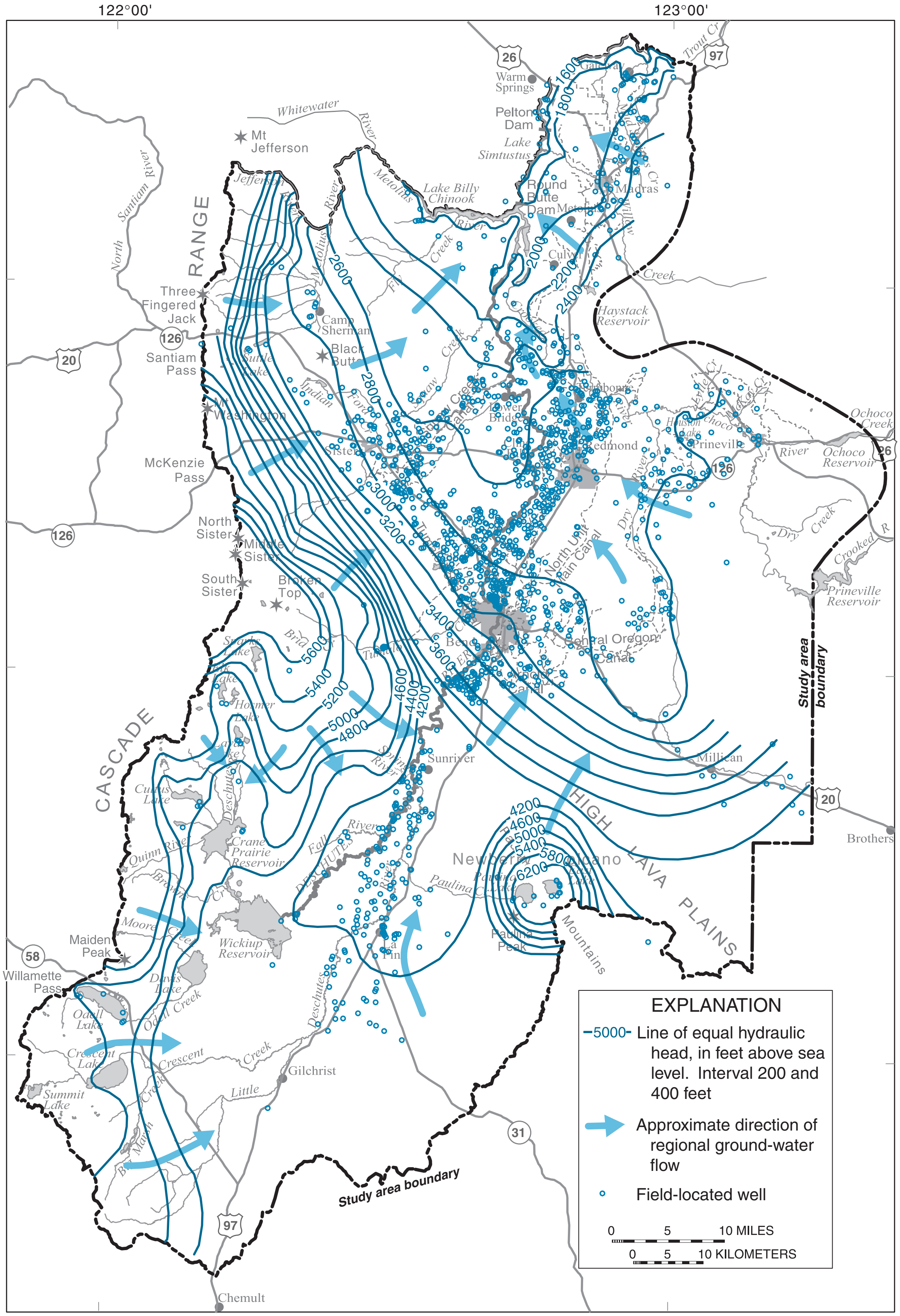


Figure 11. Map showing generalized lines of equal hydraulic head and ground-water flow directions in the upper Deschutes Basin (from Gannett and others, 2001).

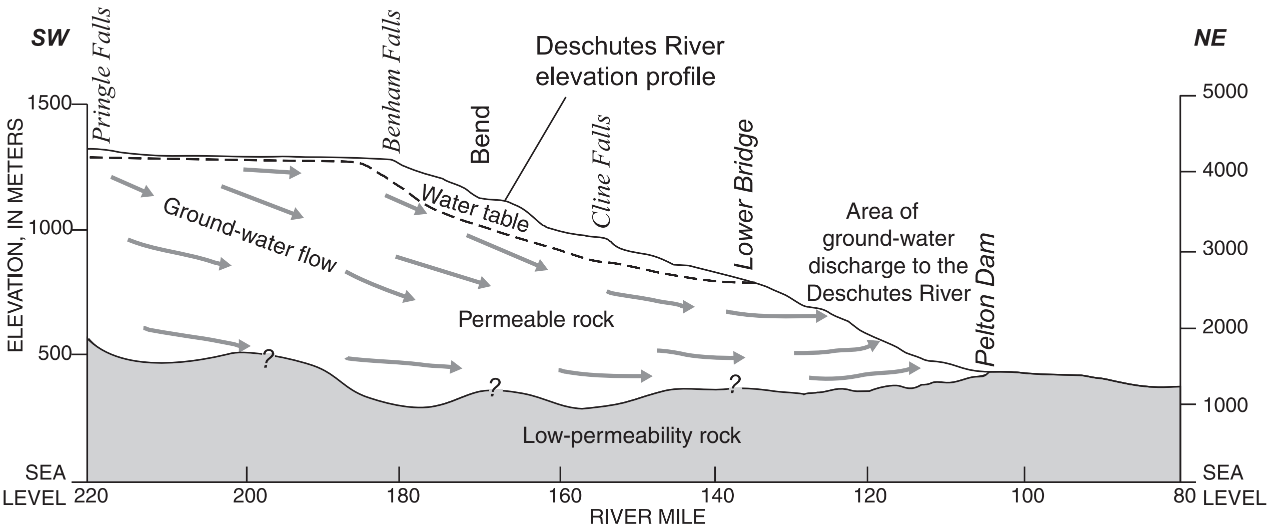


Figure 12. Diagrammatic section showing the effect of geology and topography on ground-water discharge along the Deschutes River from Benham Falls to Pelton Dam.

the well at the Lava Lands Visitor Center, just south of the butte, describes a more or less monotonous stack of interbedded lava and cinders to a depth of 155 m. The drillers' log for a test well east of US 97 shows predominantly lava and cinders to a depth of 91 m, underlain by 12 m of tan sandstone, in turn underlain by 12 m of white pumice. A gamma-ray log (Fig. 14; K.E. Lite, Jr., unpub data) indicates that these latter strata are silicic, so we interpret the "sandstone" and pumice to be the Tumalo Tuff and the Bend Pumice (described at Stop 6). Although ambiguous, descriptions for the 19.5-m section below the silicic deposits suggest interbedded lava and sediment. At the base of Lava Butte, the water table is at about 1,251 m (4,104 ft), and it lies about 122 m below the land surface.

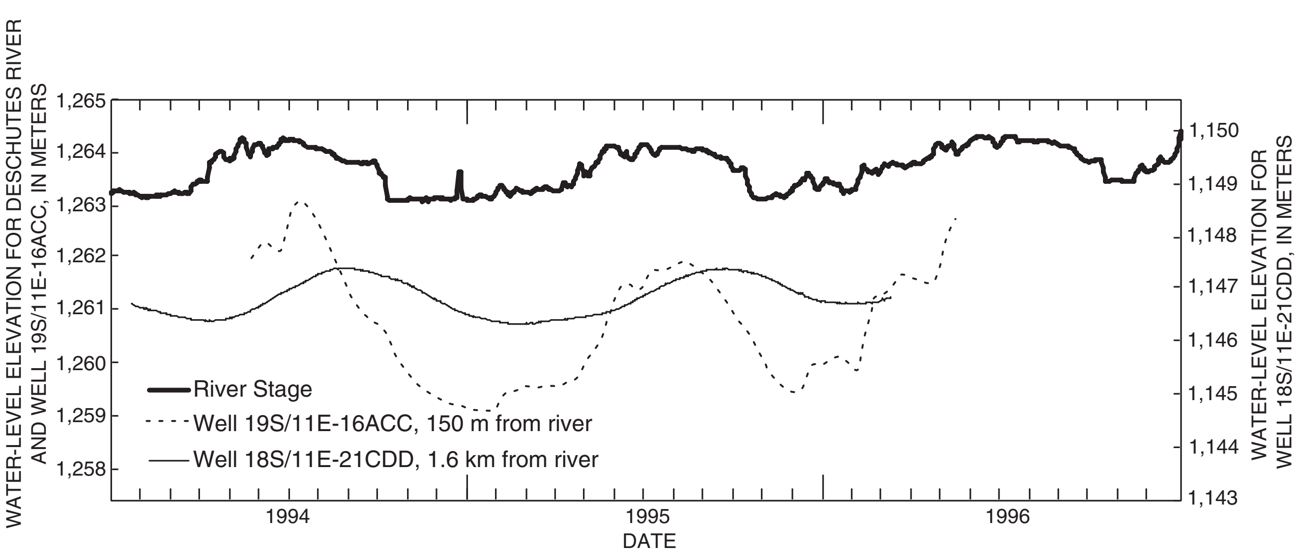


Figure 13. Hydrograph showing the relation between the stage of the Deschutes River at Benham Falls and water levels in wells 150 and 1,600 m from the river.

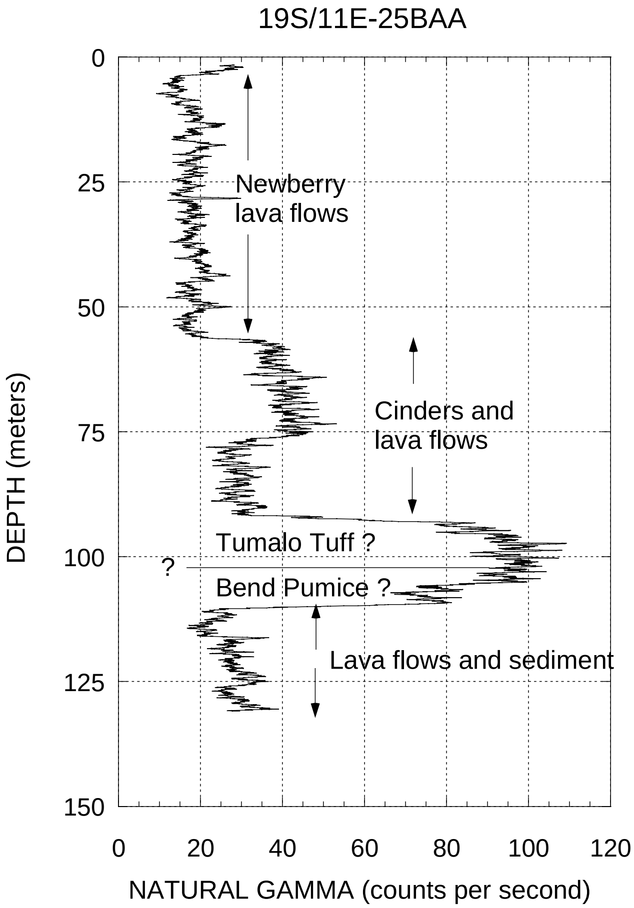


Figure 14. Natural gamma-ray log and inferred stratigraphy in a well east of US 97 near Lava Butte, Stop 4.

Stop 5—Surface-Water Diversions in Bend (Altitude 1,085 m; 3,560 ft)

At this stop we can see the dam and headgates for diversion into the North Unit and Pilot Butte Canals (Fig. 15). The dam, near River Mile 165, was constructed in the early 1900s as part of the diversions (K.G. Gorman, oral commun, 2001). It also once served as a hydropower impoundment. Some of the original power-generating equipment is visible downstream from the dam.

During the irrigation season, water is diverted from the Deschutes River between Lava Butte and Bend into irrigation canals at a rate of approximately 57 m^3/s (2,000 ft^3/s) (Gannett and others, 2001). The average annual rate is about 28 m^3/s (1,000 ft^3/s). The North Unit and Pilot Butte Canals account for roughly half the total diversion. Most canals are unlined and leak considerably. Overall transmission losses approach 50 percent and are even greater where the canals cross fractured Pleistocene lava. Water lost from canals infiltrates and recharges the ground-water system. Although the canals lose large amounts of water, the Deschutes River north of Bend shows little or no loss. The probable reason for the lack of stream leakage is that lava in the stream channel north of Bend is sufficiently old that fractures have been sealed by sediment.

Total canal losses north of Bend approach a mean annual rate of 14 m^3/s (500 ft^3/s), which is a substantial flux of water, more than 10 percent of the long-term average recharge from precipitation in the entire upper Deschutes Basin, about 108 m^3/s (3,800 ft^3/s) (Gannett and others, 2001). Ground-water flow directions

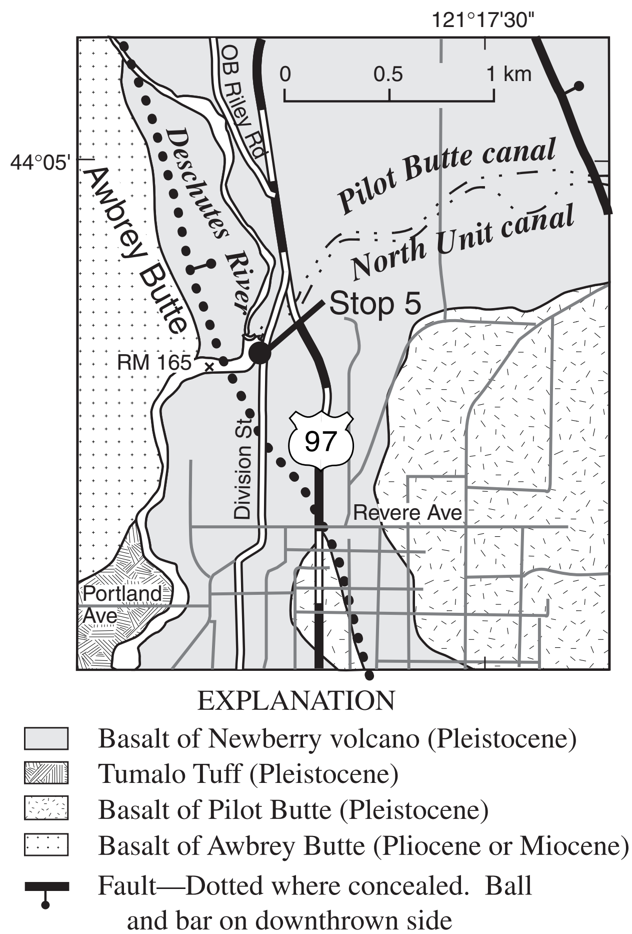


Figure 15. Map showing the Deschutes River diversions for the North Unit and Pilot Butte Canals near Bend, Stop 5 (geology from Sherrod and others, in press).

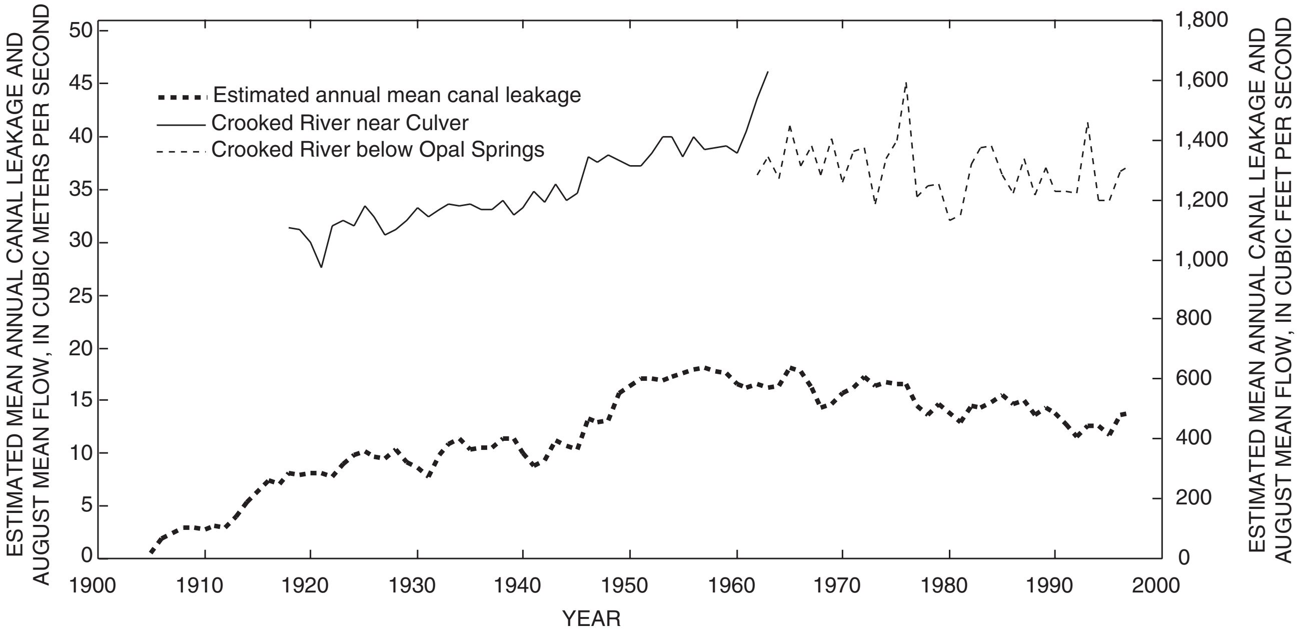


Figure 16. Hydrograph showing estimated canal losses north of Bend and August mean discharge of the Crooked River.

inferred from head maps (Fig. 11) suggest that the lost canal water flows toward the lower Crooked River. Comparing rates of estimated canal losses with long-term streamflow records confirms that this is the case. Figure 16 shows the estimated annual mean canal losses from 1905 to 1998. Also shown are the August mean flows of the lower Crooked River for the same period. During August, flows of the lower Crooked River are due almost entirely to spring discharge, and variation in August mean flow is a good proxy for variation in ground-water discharge (baseflow). As shown in Figure 16, baseflow to the lower Crooked River increased in a manner similar to the estimated canal losses throughout most of the past century. This relation has implications for water management. Efforts to conserve water by lining the canals may result in reduced streamflow in the lower Crooked River.

Many wells in the Bend area have static water levels greater than 200 m below the land surface. Some wells, however, have depths to water ranging from 30 to 60 m. Many of these shallower saturated zones are artificially recharged from canal leakage and deep percolation of irrigation water. Historic data are insufficient to determine precisely how much shallow ground water is canal derived and how much might result from natural stream leakage.

Ground-water levels show varying rates of response to changes in canal flow depending on the permeability of the bedrock, as described in detail by Gannett and others (2001). For example, the static water level in a well 5 km southeast of here, drilled in fractured lava 1 km from the Arnold Canal, responds within a matter of days to canal operation (Fig. 17A).

In contrast is a well on the north side of Redmond that spudded into the 3-4-Ma basalt of Dry River and yields water from underlying sedimentary strata of the Deschutes Formation. The water level in this well, 0.4 km west of the Pilot Butte Canal, has a greater lag time and a more subdued response (Fig. 17B). It begins to respond 2 months after canal operation begins and peaks 1 to 2 months after the canals are shut off for the year.

Stop 6—Sisters Fault Zone Outcrop, Tumalo Reservoir Road (Altitude 1,006 m; 3,300 ft)

This stop shows in cross section a minor fault of the Sisters Fault Zone. The quarried exposure shows Bend Pumice (fallout tephra), Tumalo Tuff (pyroclastic flow), and overlying gravel deposits cut by a steep northwest-striking fault with offset less than 4 m, down on the northeast side. The Tumalo Tuff is about 0.4 Ma in age, on the basis of ages by both conventional K-Ar (Sarna-Wojcicki and others, 1989) and step-heating $^{40}Ar/^{39}Ar$ methods (Lanphere and others, 1999). The capping gravel deposit is undated. Thus the latest fault motion here is known only to be younger than 0.4 Ma. Exposed in adjacent road cuts are older deposits, including the underlying Desert Spring Tuff, which is 0.6 Ma in age.

The north-northwest-trending Sisters Fault Zone comprises nearly 50 mapped faults across a breadth of 5-15 km. The total length of the zone is 60 km, and our stop here is about midway along it. The principal strand, the Tumalo Fault, extends nearly continuously for 47 km. Its trace passes 5.5 km southwest of this stop and will be visited at Tumalo Dam (Stop 7). The Tumalo Fault has displaced Pliocene lava flows of the Deschutes Formation by at least 60-70 m of normal separation at Tumalo Dam. Quaternary lava flows younger than 0.78 Ma in the same area have escarpments of only 6-10 m.

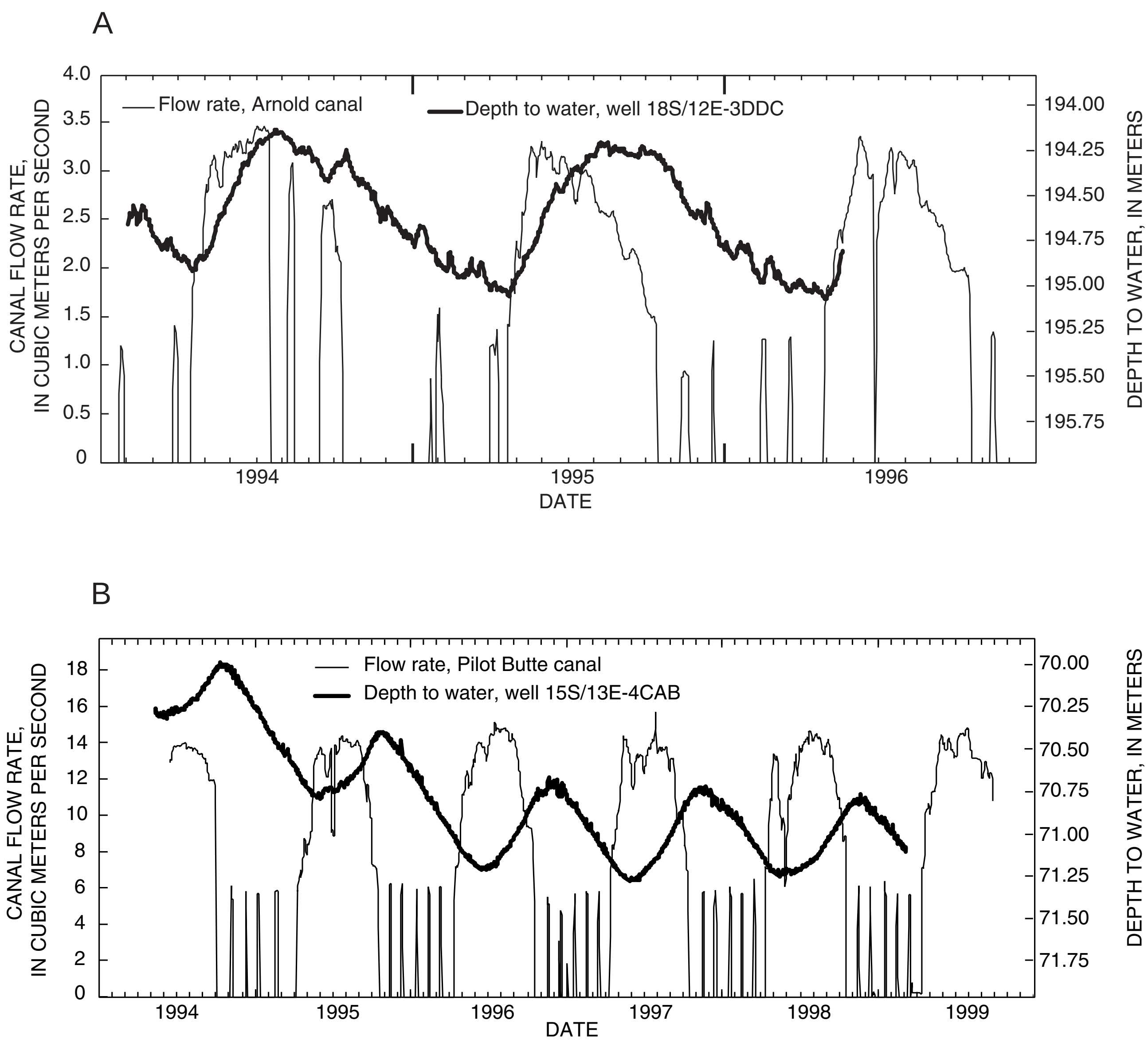
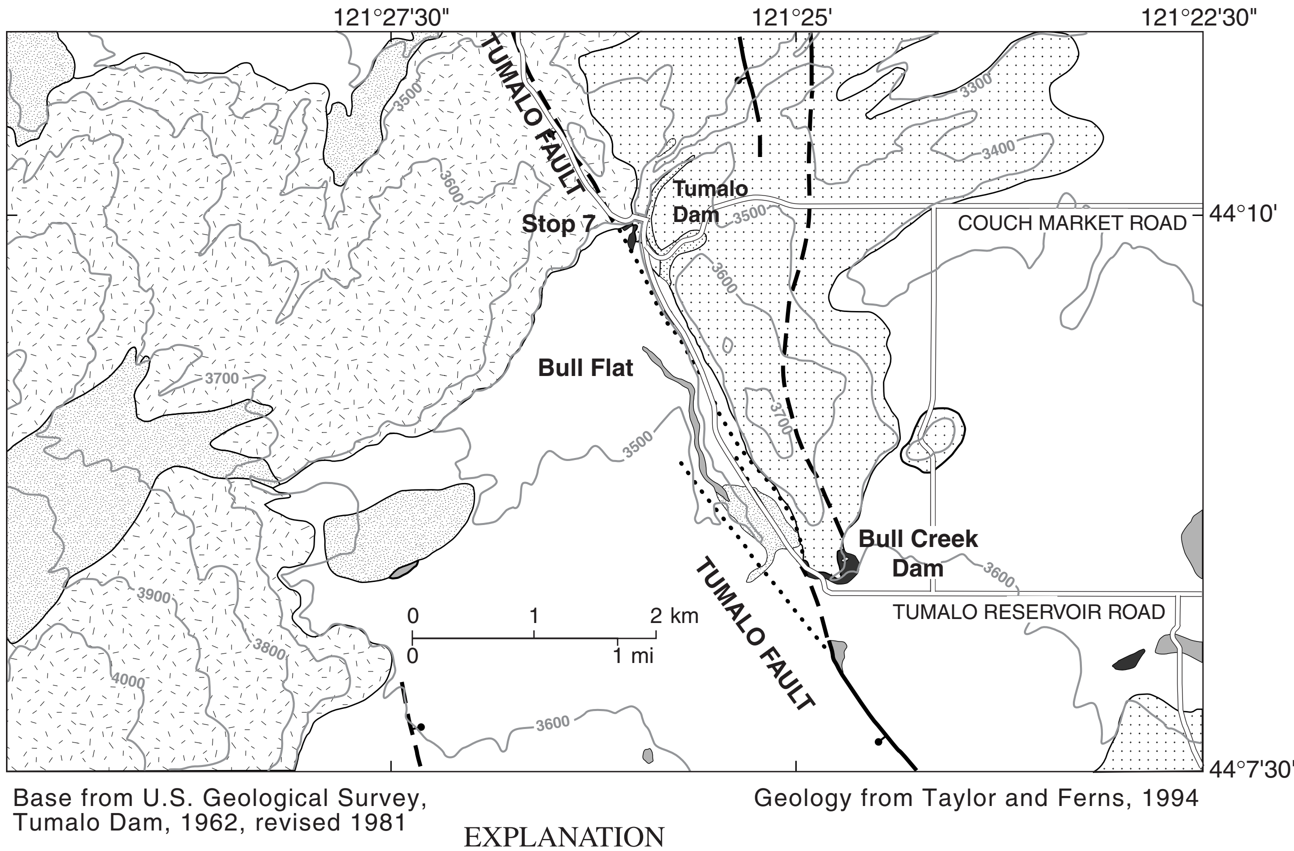


Figure 17. Hydrographs showing the relation between ground-water fluctuations and canal operation. *A.* Well and canal water fluctuations in highly permeable late Quaternary lava flows at Bend. *B.* Well and canal water fluctuations in early Pliocene and Miocene lava flows and sedimentary strata at Redmond.



EXPLANATION

Alluvium

Shevlin Park Tuff

Tumalo Tuff and Bend Pumice

Basaltic andesite

Desert Spring Tuff

Deschutes Formation

Lava flows

Pyroclastic flow deposits

Fault—Dashed where inferred, dotted where concealed. Ball and bar on downthrown side

Figure 18. Map showing geology of the Bull Flat–Tumalo Dam area, Stop 7.

The Sisters Fault Zone is roughly on trend with a pronounced ground-water gradient (high-head-gradient zone) that characterizes the Deschutes Basin northwest from Bend for 60 km to Suttle Lake in the Cascade Range (Fig. 11). The fault zone and ground-water gradient are displaced from each other by roughly 10 km, however, with the ground-water gradient lying closer to the topographic slope of the Cascade Range. There is no evidence that the Sisters Fault Zone has a measurable effect on the distribution of ground water.

This observation is unsurprising stratigraphically, because the upper Miocene and lower Pliocene strata of the Deschutes Formation on the upthrown side of faults along the zone are similar in permeability to upper Pliocene and Pleistocene deposits along the downthrown side. Thus, permeability contrasts across the faults likely are few. Ground-water damming would occur only if these faults had created substantial gouge, or if fault offsets occurred rapidly enough or were sufficiently great to create numerous or extensive closed basins.

Ground-water levels in this area are elevated slightly due to canal losses and irrigation (Gannett and others, 2001). The water table is at an altitude of about 930 m (3,050 ft), and depths to water range from 60 to 120 m de-

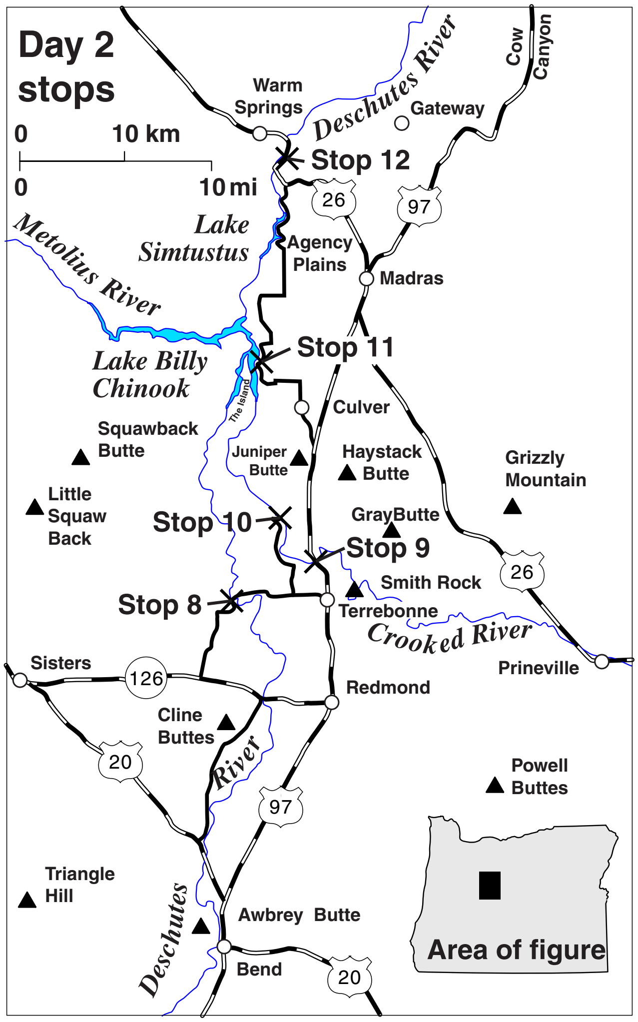


Figure 19. Route and stops for Day 2 of the field trip.

pending on land-surface altitude and proximity to the river.

Stop 7—Bull Flat and Tumalo Dam (Altitude 1,067 m; 3,500 ft)

Our travel northwest to here has been along the Tumalo Fault. En route we passed by the small Upper Tumalo Reservoir and its impoundment, Bull Creek Dam. Continuing 3.5 km farther, the road crosses Tumalo Dam, then resumes its northwest course along the fault. Tumalo Dam obstructs an unnamed drainage. Bull Flat is the treeless basin upstream from the dam. The water-table altitude is about 840 m (2,750 ft) in this area and depths to water in wells average about 150 m.

At Stop 7 we can see the geomorphic and stratigraphic contrast across the Tumalo Fault (Fig. 18). On the northeast (upthrown) side are Pliocene strata of the Deschutes Formation. They are chiefly basaltic andesite lava flows. An andesitic ignimbrite (62 percent SiO_2; Smith, 1986a) is interlayered in the Deschutes sequence and visible in the road cut at the east dam abutment. The Deschutes units exposed here are probably about 4-5 Ma in age, although no isotopic ages have been reported from this area. Offset of at least 70 m is indicated by the height of the Deschutes Formation ridge northeast of the fault and the absence of Deschutes Formation strata at the ground surface southwest of the fault. Cross sections by Taylor and Ferns (1994) suggest that the Deschutes Formation is buried by 50 m of volcaniclastic deposits in the downthrown block, in which case the dip separation is 120 m.

On the southwest (downthrown) side is alluvium that floors Bull Flat. Middle Pleistocene pyroclastic-flow deposits are exposed sporadically where they have banked against Deschutes Formation strata of the upthrown block or where alluvium is thin on the floor of Bull Flat. The ~0.6-Ma Desert Spring Tuff likely has been displaced several meters here, because no outcrops of the tuff are found on the downthrown side but they are present on the upthrown block of the Tumalo Fault near the dam. Displacement of younger pyroclastic-flow deposits is difficult to quantify, because the pyroclastic flows may have overtopped topographic barriers and been deposited across a wide altitude range.

Bull Flat is an alluvial basin that likely has received sedimentary and volcanic deposits during at least the past million years. Neither the depth nor lithology of the subsurface fill is known. The rate of stream incision across the Tumalo Fault in this area has kept pace with the rate of uplift, however, because the distribution of all pyroclastic flows of 0.6 Ma and younger suggests that they escaped through channels cut through the upthrown block.

The west dam abutment is a Cascade Range basaltic andesite lava flow. Topographic escarpments only 6–10 m high mark the trace of the Tumalo Fault across this flow. It possesses normal-polarity thermal-remanent magnetization and is thought to be younger

SIDEBAR 1. THREE BUTTES—70 YEARS OF UNCERTAINTY RESOLVED

Three volcanic buttes of previously uncertain age lie within a few kilometers of the Deschutes River between Redmond and Madras (Fig. 20). Our travel takes us by Cline Buttes, a prominent rhyolite dome complex west of Redmond. Less prominent is Forked Horn Butte, which lies 6 km east of Cline Buttes and may be visible along the drive. The third butte, which won't be visible, is an unnamed sprawling mass of rhyodacite lava southwest of Steelhead Falls, 15 km north of Cline Buttes. Each had been assigned to stratigraphic units several million years older than the Deschutes Formation (Stearns, 1931; Williams, 1957; Stensland, 1970; Robinson and Stensland, 1979; Smith, 1986a). By contrast, our mapping and dating have determined that these buttes are late Miocene in age and part of the Deschutes Formation.

The rhyolite of Cline Buttes has an isotopic age of 6.14±0.06 Ma (Table 2). Most rocks at Cline Buttes are hydrothermally altered, but the dated sample came from fresh obsidian in a quarry on the east flank. Forked Horn Butte is a knob within an extensive debris-avalanche deposit. It is undated isotopically, but it underlies the 3.68-Ma basalt of Redmond and overlies sedimentary strata in the Deschutes Formation. The rhyodacite southwest of Steelhead Falls is fairly fresh and homogeneous throughout. It has an age of 6.74±0.20 Ma (Table 2).

Depth to water in the Cline Buttes area averages 90 to 120 m on the south side and 60 to 100 m on the north side. The water-table altitude is 823 to 838 m (2,700 to 2,750 ft). Hydrologic evidence indicates that the rhyolite of Cline Buttes and the rhyodacite southwest of Steelhead Falls are fairly permeable relative to the surrounding terrain, a finding in keeping with their age (Ferns and others, 1996b; M.W. Gannett, unpub. data). Sparse head data from wells shows that the northward gradient steepens adjacent to Cline Buttes, suggesting higher relative permeability. In addition, substantial ground water discharges to the Deschutes River from springs issuing from the rhyodacite exposed in the canyon north of Steelhead Falls (Ferns and others, 1996b).

These late Miocene ages and stratigraphic assignments lead to a different view of the breadth and thickness of Deschutes Formation strata in the central part of the basin. Instead of thinning against steptoes, the Deschutes Formation extensively blankets the central basin with a thickness of 300-460 m. Indeed, the Deschutes Formation may be as thick as 500-600 m if measured from the summit of Cline Buttes.

than 0.78 Ma. The lava flow is overlain upslope by the 0.2-Ma Shevlin Park Tuff.

The story of the Tumalo Dam and its reservoir is one of broken promises and unfulfilled dreams, like many that surround the land-grab development of the arid west. An excellent history of the Tumalo Project, which we retell in the following, is provided by Winch (1984-85).

In the 1890s, homesteaders were lured to this area by promises of free land and abundant water for irrigation. By 1913, however, the private developers had gone bankrupt, and the promised water and delivery canals had not materialized. The State of Oregon stepped in to fulfill promises made to the homesteaders and to the federal government. Tumalo Dam was part of an attempt to provide the promised irrigation water.

Prior to construction, some engineers expressed concern about the permeability of the strata bounding the reservoir. The dam, constructed in 1914, is a 22-m-tall earth-fill structure with a steel-reinforced concrete core. The outlet was a 2.5 by 2.5-m concrete-lined tunnel 123 m long. Flow through the outlet was to be controlled from a small house at the end of the dam. The reservoir was to cover 447 ha and to impound 24.66 x 10^6 m^3 of water. The dam was completed and the sluice gates closed on December 5, 1914.

The concerns of the engineers were confirmed early in the winter of 1915, as the reservoir, at only a fraction of its capacity, saw its water level falling 0.2 m per day. When runoff increased in March and April, and the water level started to rise, large sinkholes developed (perhaps along the Tumalo Fault) at the eastern margin of the reservoir not far from the dam. The largest sinkhole was 9 by 15 m across and 3 to 8 m deep, swallowing water at an estimated rate of 5.7 m^3/s (200 ft^3/s). Attempts to seal the

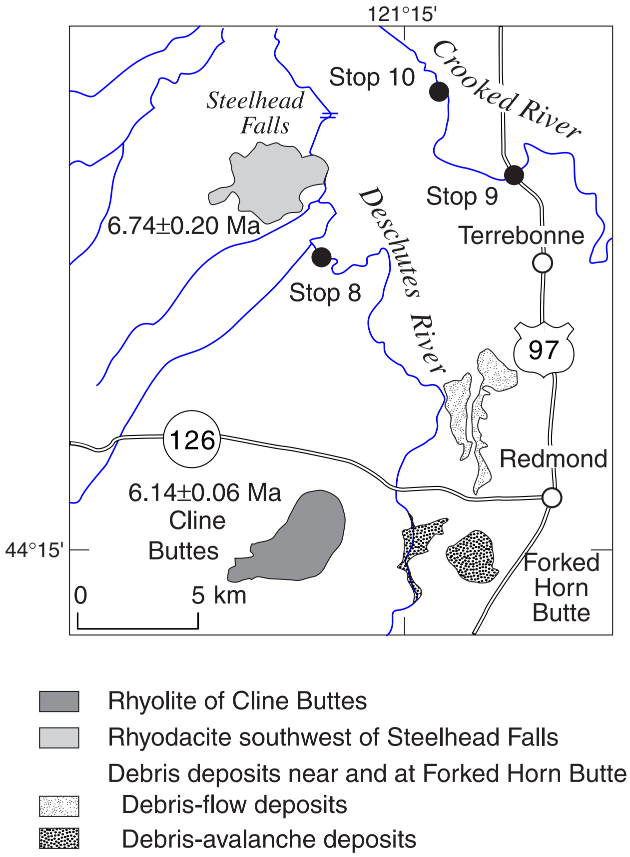


Figure 20. Cline Buttes, Forked Horn Butte, rhyodacite southwest of Steelhead Falls, and nearby debris deposits.

reservoir failed, and the Tumalo Dam has never been used.

DAY 2

Synopsis, Deschutes Basin from Bend to Madras

The traverse from Bend to Madras (Fig. 19) is within a broad valley dominated by a gently northward-sloping surface with scattered low hills. Many of the hills consist of cinder cones, small shield volcanoes, and silicic domes, mainly of the upper Miocene to Pliocene Deschutes Formation. Several small canyons form a northeast-trending topographic grain on the west side of the valley, exposing lava flows, ash-flow tuff, and sedimentary strata deposited as a result of volcanic eruptions during Deschutes time. We will drive down one of these canyons (Buckhorn) on our way to Lower Bridge (Stop 8).

The central and eastern surface in the southern part of the valley is flat and covered predominately with Pleistocene lava flows originating from Newberry Volcano, which forms part of the southeast basin margin. Some of the lava flows from Newberry traveled tens of kilometers and entered the ancestral Deschutes and Crooked Rivers as intracanyon flows. Remnants of these lava flows form conspicuous benches within the canyons of the Deschutes and Crooked Rivers in the northern part of the valley. The eastern boundary of the valley consists of hills of mainly silicic rocks of the Oligocene to Miocene John Day Formation.

The modern Deschutes and Crooked Rivers form shallow incised drainages in the southern part of the valley, but they have carved narrow deep canyons in the northern part. It is in these deep canyons where much of the section of the Deschutes Formation is exposed. Lava flows, ash-flow tuff, debris-flow deposits, and fluvial silt, sand, and gravel are exposed in cliffs along several kilometers of canyon walls. Our hike down into the canyon of the Crooked River (Stop 10) will be an opportunity to examine some of those deposits in detail. By about 7 km downstream from Round Butte Dam, the Deschutes River has cut through the entire thickness of the Deschutes Formation and exposed the underlying Simtustus Formation and Prineville Basalt. About 8 km downstream from that point, approximately at the regulator dam, the river has cut down to the John Day Formation.

The broad flat area between the canyon of the Deschutes River and Madras, known as the Agency Plains, is underlain by Deschutes Formation lava flows (the basalt of Tetherow Butte). These flows cover only the west side of the Agency Plains. The east side, including Madras, is underlain by fine-grained sedimentary deposits of the Deschutes Formation shed mainly from the uplands to the east.

Ground-Water Flow from Bend to Madras

Throughout the area between Bend and Madras, ground water flows generally northward through sediment and lava of the Deschutes Formation toward the confluence of the Deschutes and Crooked Rivers. The Pleistocene lava flows from Newberry Volcano and pyroclastic deposits from the Bend Highland

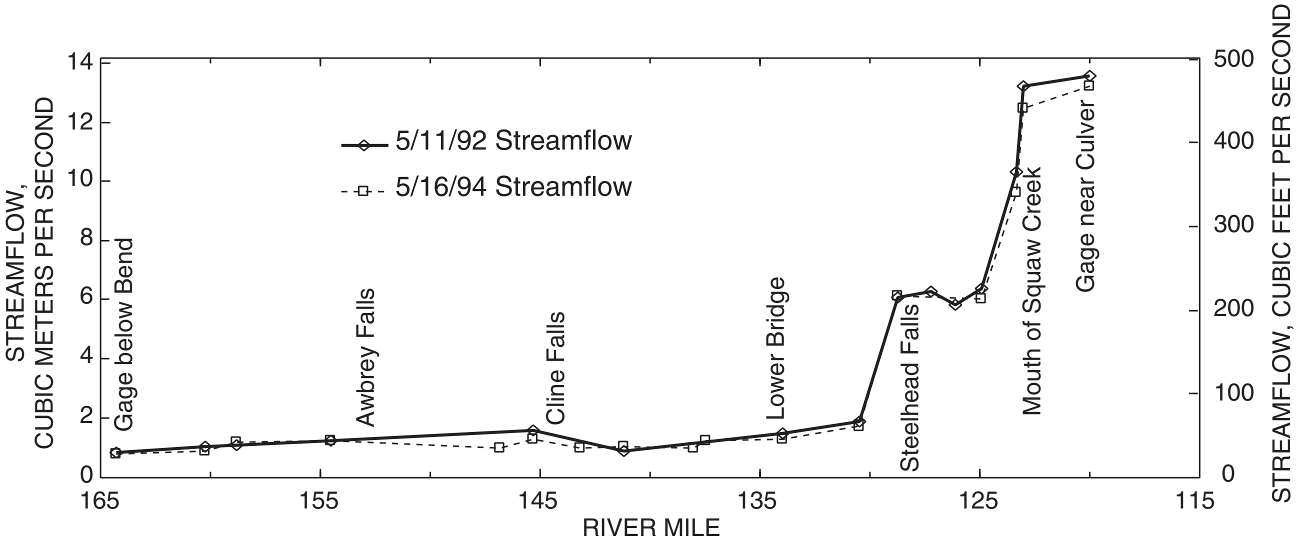


Figure 21. Graph showing gain in flow of the Deschutes River owing to ground-water discharge between River Miles 165 and 120, May 1992 and May 1994.

are largely unsaturated. The area around Bend extending north to Tumalo is characterized by a large vertical head gradient indicating strong downward flow. Shallow wells in the area commonly have static water levels less than 30 m below the land surface, whereas nearby deep wells may have depths to water exceeding 200 m. The presence of the shallow water table and the strong downward gradient is attributed to local and at least partly artificial recharge from canal and stream leakage. Such areas of artificial recharge and anomalously shallow saturated zones are far less common to the north around Redmond. The water table slopes gently northward from Bend toward the Deschutes–Crooked Rivers confluence area with a gradient of approximately 3.8 m/km (Fig. 11). The stream gradient, however, is somewhat steeper, averaging 6.2 m/km between Bend and Lower Bridge. As a consequence, the canyons of the Deschutes and Crooked Rivers intersect the regional water table several kilometers north of Redmond.

The Deschutes River intersects the regional water table at about Lower Bridge (Fig. 12; River Mile 134), and the Crooked River intersects the regional water table at about Trail Crossing (River Mile 20.6), about 2.1 km northeast of the US 97 bridge. Downstream from these sites, the altitude of the streams is below that of the regional water table, and ground water flows toward and discharges to the streams, resulting in dramatic increases in streamflow. Between the points where the Deschutes and Crooked Rivers intersect the regional water table and the point where the Deschutes River intersects the John Day Formation, the combined streams gain approximately 63 m^3/s (2,200 ft^3/s) from ground-water inflow (Gannett and others, 2001).

North of the regulator dam (at River Mile 100), the Deschutes Formation thins against the underlying units and pinches out by about River Mile 87. Below this point, the Deschutes River gains little if any flow from ground-water discharge, and the modest increases in flow that do occur are due to tributary inflow.

Stop 8—Lower Bridge, Where Ground Water Meets the Deschutes River (Altitude 770 m; 2,525 ft)

This stop is at the south side of the bridge on the left bank of the Deschutes River. It is about here where the Deschutes River has incised deeply enough to intercept the regional water table. Downstream from this point, the Deschutes River is below the regional water table, resulting in ground-water discharge to the river. Wells in the Lower Bridge area have static water levels coincident with the river level.

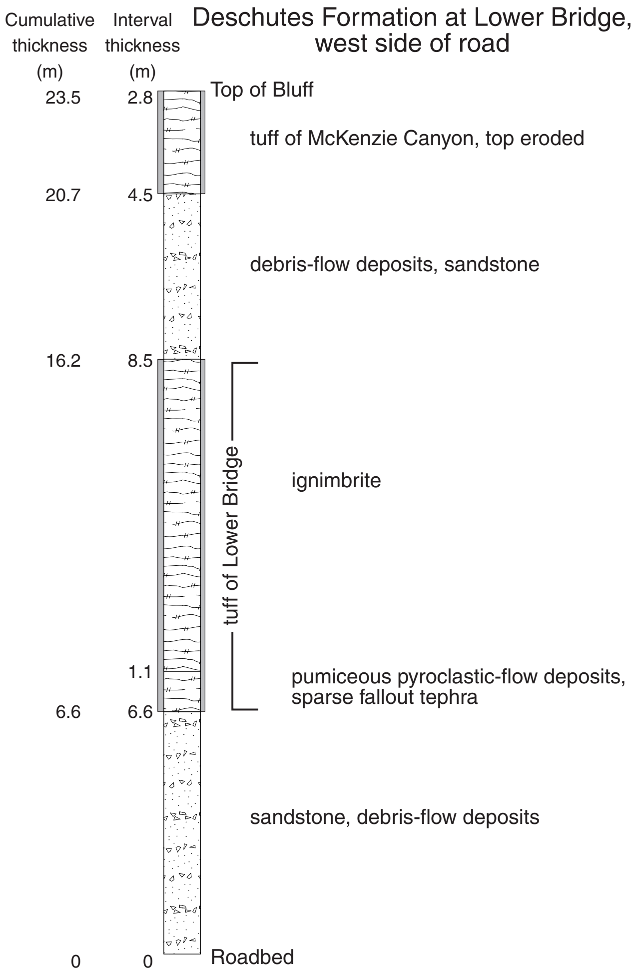


Figure 22. Stratigraphic section of the Lower Bridge area, Stop 8.

Streamflow measurements that are synoptic (gathered nearly simultaneously) show that the Deschutes River gains approximately 11.3 m^3/s (400 ft^3/s) from ground-water discharge between Lower Bridge and the stream gage just above Lake Billy Chinook near Culver (Fig. 21). About a quarter of this increase comes from ground-water discharge to the lower 3.2 km of Squaw Creek. Ground-water inflow is not uniform along the stream but emerges preferentially from permeable deposits such as coarse conglomerate and the fractured rhyodacite described in Sidebar 1 (Ferns and others, 1996b).

Exposures on the west side of the road provide easy access to about 24 m of the Deschutes sedimentary and volcanic sequence (Fig. 22). In the middle one-third is the tuff of Lower Bridge, about 8.5 m here, a typical thickness. At the top of the exposure is the tuff of McKenzie Canyon. It is only 2.8 m thick, but its top is eroded. The tuff of McKenzie Canyon commonly is 6-10 m thick in this part of the basin. Both units were erupted from vents in the Cascade Range. Neither tuff has been dated, but both were emplaced between about 5.77 and 5.06 Ma on the basis of their position above the basalt of Opal Springs and beneath the basaltic andesite of Steamboat Rock (Table 2). A gamma-ray log from a well about 1.3 km south of this exposure reflects the same stratigraphic section viewed at this stop (Fig. 23; K.E. Lite, Jr., unpub data). The principal water-bearing unit in the well is the sand and gravel deposit 12-25 m below the tuff of Lower Bridge.

Two Pleistocene stratigraphic units are widespread in the Lower Bridge area and visible from this stop. Poorly indurated earthy white diatomite caps the bluff west of our stop

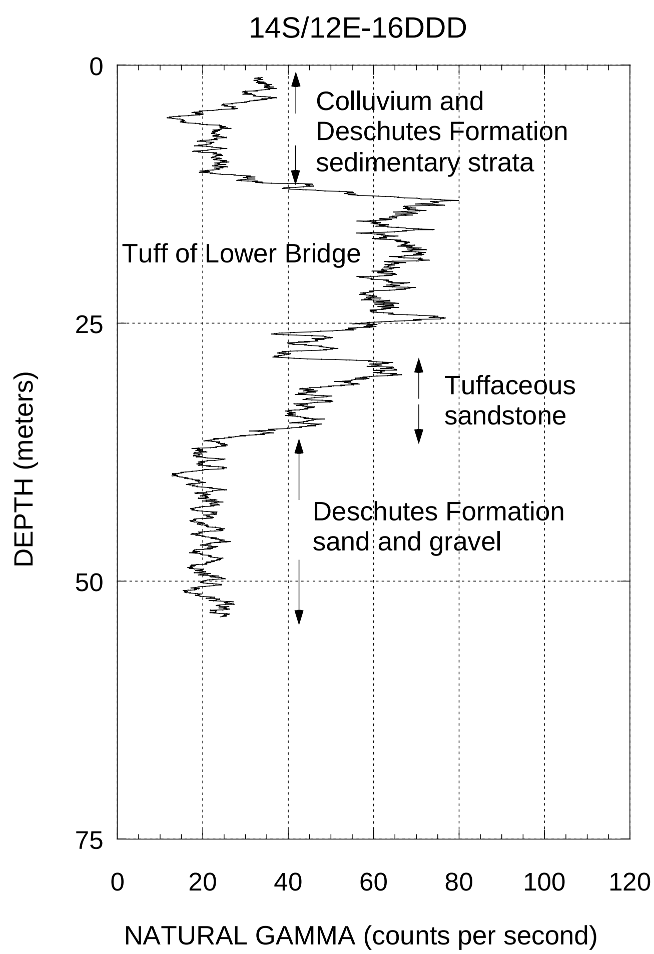


Figure 23. Natural gamma-ray log and inferred stratigraphy in a well near Lower Bridge, Stop 8.

and is also exposed in road cuts along the highway grade east from here. The diatomite has been mined extensively from the now-abandoned quarry above the bluff. Basalt from Newberry Volcano overlies the diatomite in several exposures.

The diatomite was once as thick as 20 m (Moore, 1937). Little of the original deposit remains, however, and the site is now mostly occupied by irregularly heaped overburden and waste from strip mining. Predominant diatom species are *Stephanodiscus niagarae* (K.E. Lohman in Moore, 1937) and *S. excentricus* (Smith and others, 1987), indicating a late Pliocene or Pleistocene age (Krebs and others, 1989). Volcanic ash bedded within the deposit has been tentatively correlated with the Loleta Ash (distal-fallout equivalent of the Bend Pumice) (Smith and others, 1987; A.M. Sarna-Wojcicki, oral commun, 1995), which is thought to be about 0.4 Ma in age. An alternative correlation with a 1.9-Ma ash bed found in drill core from a well near Tulelake, California (T-749, 191 m depth; Rieck and others, 1992), is nearly as satisfactory on the basis of statistical comparison coefficients (A.M. Sarna-Wojcicki, oral commun, 1995).

The age of the main diatomite body is probably middle Pleistocene, because the deposit fills a valley floor whose altitude is only slightly higher than the surface later mantled by basalt of Newberry Volcano. Presumably the prediatomite erosional stage is only slightly older than the basalt. Although overlain by the basalt of Newberry Volcano in road cuts northeast of Lower Bridge, earlier Newberry lava flows may have dammed the ancestral Deschutes River to create the lake in which the diatomite was deposited (Smith, 1986a).

Stop 9—Peter Skene Ogden Bridge (Older Bridge) and Adjacent New Bridge (Altitude 811 m; 2,660 ft)

US 97 crosses the Crooked River where middle Pleistocene basalt forms most of the canyon walls. These lava flows were erupted from vents on the north flank of Newberry Volcano and flowed north across the broad plain extending to Redmond. The lava poured into the canyon of the ancestral Crooked River 9 km southeast of the present Ogden Bridge and flowed downstream past the bridge site at least another 8 km—a distance of more than 55 km from the vent area. Lava flows also entered the Deschutes River and reached to Lake Billy Chinook, at least 65 km from the vent.

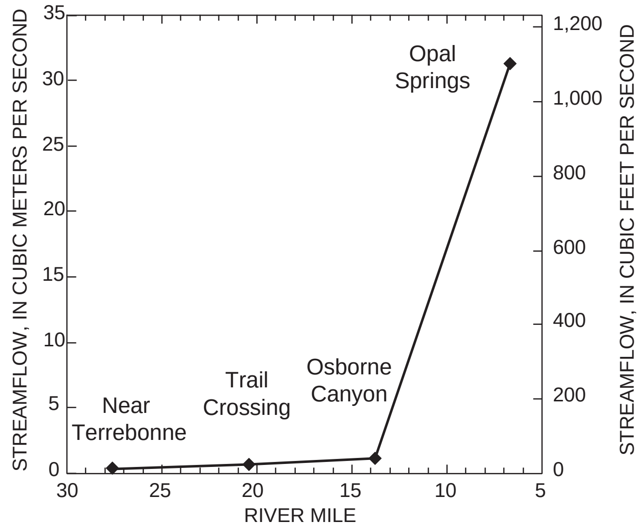


Figure 24. Graph showing gain in flow of the lower Crooked River owing to ground-water discharge between River Miles 28 and 7, July 1994.

The altitude of the Crooked River at the US 97 crossing is approximately 750 m (2,460 ft). At this level, the Crooked River has incised to the depth of the regional water table. Synoptic streamflow measurements gathered in 1994 (Fig. 24) show the Crooked River gaining about 2 m^3/s (70 ft^3/s) from ground-water discharge between Trail Crossing, about 3.2 km upstream from the US 97 crossing, to Osborne Canyon, about 7.2 km downstream. Along the 11.2-km downstream reach between Osborne Canyon and the gage above Lake Billy Chinook, the river gains an additional 28.3 m^3/s (1,000 ft^3/s), making this reach one of the principal ground-water discharge areas in the basin.

The hills to the east are underlain by southeast-dipping strata of the John Day Formation. They form the upthrown block of the Cyrus Springs Fault Zone (Smith and others, 1998). Rocks as young as the Prineville Basalt (15.8 Ma) were involved in the deformation, whereas the Deschutes Formation, the next youngest unit preserved, is undeformed. Thus the deformation occurred after early Oligocene and before late Miocene time. The deformation

SIDEBAR 2. WHAT WAS THE PATH OF THE ANCESTRAL DESCHUTES RIVER?

The Deschutes and Crooked Rivers of today were probably following similar paths before 1 Ma. The Crooked River is a long-lived system draining the southwest flank of the Ochoco Mountains. The geologic record for the Crooked River in the Deschutes Basin can be inferred back to about 6 Ma.

The Deschutes River history for the time back to 4 Ma is known only for its downstream reach north of Lower Bridge and Steelhead Falls. South (upstream) from there, the distribution of lava and pyroclastic flows indicate southwest-trending drainages, as if the Deschutes River may have had its source in the area of the Three Sisters. Evidence for an ancestral Deschutes path southward through the Bend area is buried, and there is no evidence that a river like the modern Deschutes flowed through the Bend area until sometime after 2 Ma.

As shown in Fig. 25A, the general path of the Deschutes and Crooked Rivers in late Miocene time is inferred from paleocurrent indicators, the orientation of intracanyon lava flows toward the basin's axis, and the position of older rocks forming a buttress to sediment transport and deposition along the east side of the basin (Stensland, 1970; Smith, 1986a). This alluvial region was choked by the basalt of Tetherow Butte about 5 Ma, forcing the ancestral Deschutes River westward to its present alignment (Fig. 25B).

The Crooked River subsequently established a course across the basalt of Tetherow Butte, incising a canyon from which it has never escaped. By about 3.5 Ma, the Crooked River upstream from the US 97 crossing was forced to follow its approximate current alignment near the foothills of the Ochoco Mountains owing to emplacement of the basalt of Dry River 3.5-4.0 Ma (Fig. 26).

Still unclear is when the Deschutes River expanded southeastward. The 3.56-Ma basalt of Redmond verged upon but did not enter the Deschutes River drainage northwest of Redmond. Instead it followed a path into the Crooked River drainage. Between 3 and 1 Ma, however, some Deschutes tributary reached southeast beyond Cline Falls and Redmond, judging from broad topographic incision northwest of Redmond. Also, the Redmond Channel, discussed later, probably was carved southeast through the Redmond area prior to about 1 Ma.

About 1 Ma, the basalt of The Island choked the lower Crooked River. This lava can be traced in patchy exposures along the walls of the Crooked River to within 600 m west of the US 97 bridge. The altitude along the lava's base shows that the Crooked River had cut down nearly to the present level by 1 Ma, but the basalt of The Island sheds little light on the paleoriver pattern, because nothing is known about its distribution beyond an already well-defined river valley.

The most recent intracanyon lava flows are part of the basalt of Newberry Volcano, which advanced across the Redmond Plain between 0.7 and 0.3 Ma. Some of these flows draped into and choked the canyon of the Crooked River, once again forcing the river northward to resistant bedrock. Today the Crooked River remains trapped in a deep gorge along the Smith Rock pathway. Other Newberry flows entered the Deschutes River system on each side of Forked Horn Butte (Fig. 27), the first rock-hard evidence we have that the Deschutes drainage was well developed south of Redmond.

One of the little-known features of the history of the Deschutes River is the 9-km-long Redmond Channel (Stensland, 1970). This gulch is only 6-12 m deep today, but its breadth is similar to the nearby modern gorge of the Deschutes River, which is 35 m deep. The Redmond Channel was carved after 3.5 Ma. It follows the contact between the basalt of Redmond and the basalt of Dry River, and in its lower reach, it cuts across the basalt of Redmond. The channel was then nearly filled with lava sometime after 0.7 Ma when the basalt of Newberry Volcano flowed into it. The Redmond Channel may mark the time when a river from the south (the head of the modern Deschutes River) first passed across the Bend area. We presume that the channel has an age of at least 1 Ma in order to account for the depth of erosion required by the breadth and length of its valley.

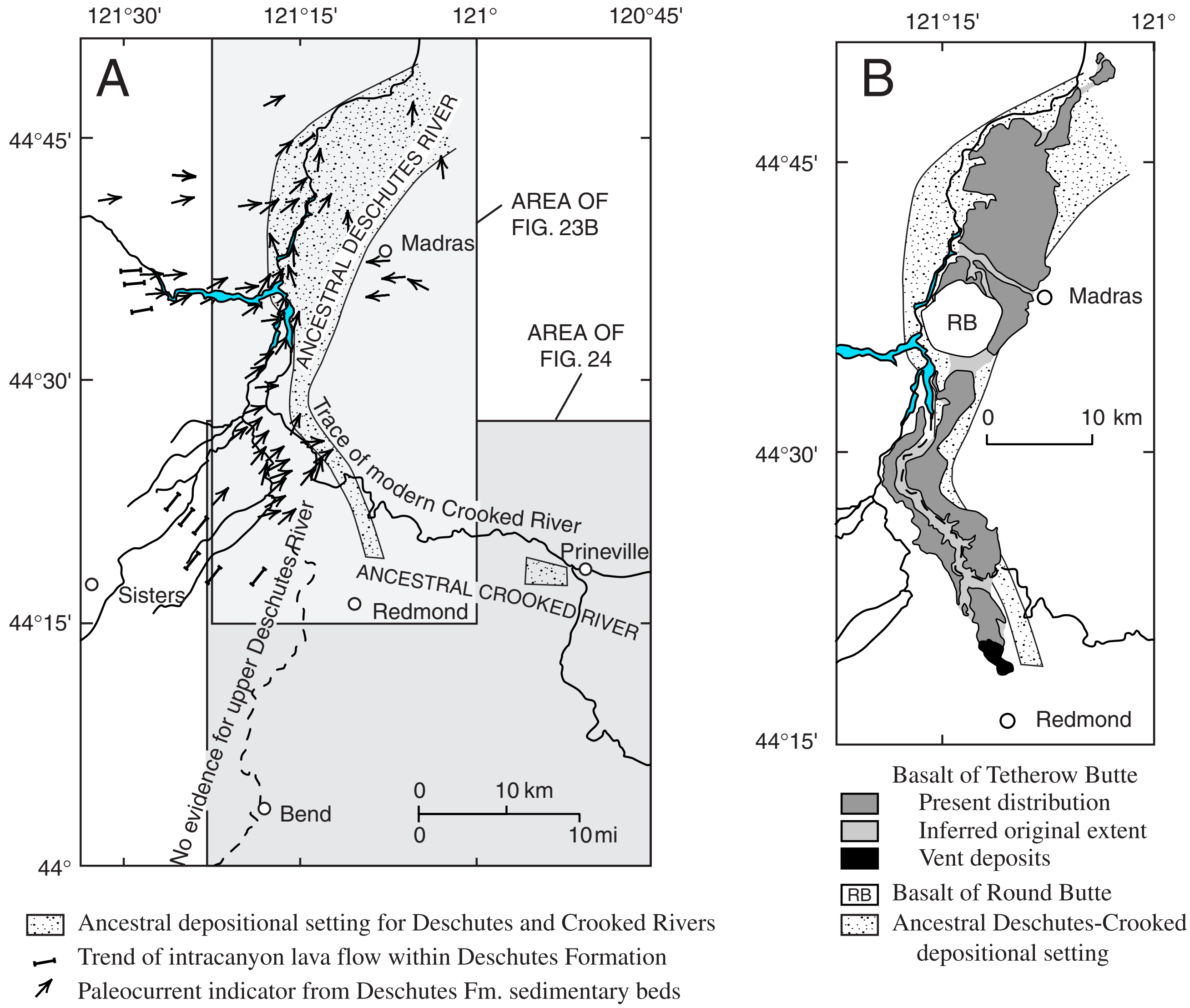


Figure 25. Maps showing path of the Deschutes and Crooked Rivers about 5 Ma, with modern rivers, tributaries, and Lake Billy Chinook shown for reference (data from Smith, 1986a). *A.* Extent of the Deschutes River depositional setting, and flow directions from paleocurrent and paleogeographic indicators. *B.* Extent of basalts of Tetherow Butte and Round Butte.

may be entirely middle or earliest late Miocene in age.

Rocks of the John Day Formation have low permeability because the tuffaceous material is mostly devitrified to clay and other minerals. Lava flows within the formation are weathered and contain abundant secondary minerals. The John Day Formation acts as a barrier to regional ground-water flow. It and age-equivalent Cascade Range strata are considered the lower boundary of the regional flow system throughout much of the Deschutes Basin. The overlying middle Miocene Prineville Basalt is locally fractured, contains permeable interflow zones, and is used as a source of water in some places.

Stop 10—Crooked River Gorge at Crooked River Ranch (River Altitude 670 m; 2,200 ft)

This stop includes a walk into the canyon of the Crooked River on a dirt access road known colloquially as the Hollywood Grade. Headworks for a flume system are preserved at the base of the grade. The term Hollywood stems from the use of this area for filming a major motion picture many years ago. Note: Access to the top of the road requires crossing private land, so check with the nearby motel

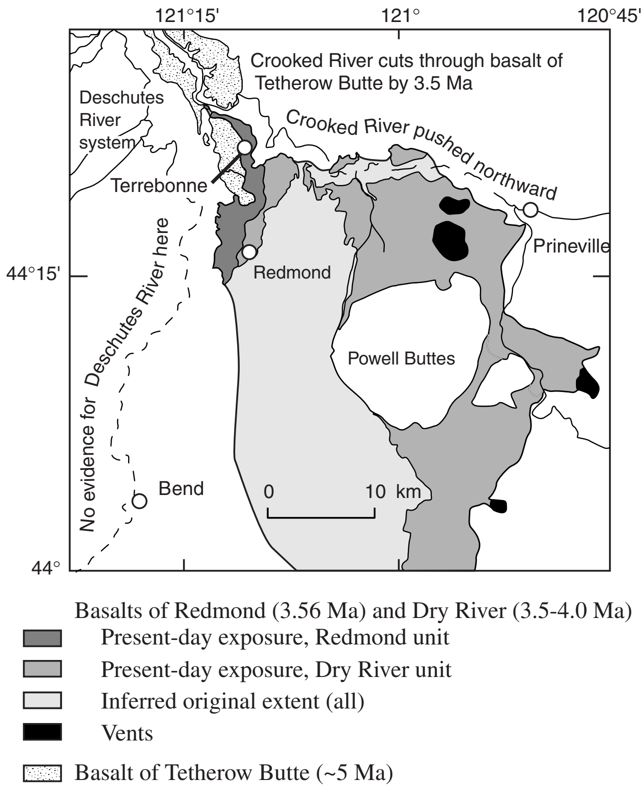


Figure 26. Extent of lava flows in the basalts of Redmond and Dry River across the Redmond Plain and into the ancestral canyon of the Crooked River.

staff or the office of Crooked River Ranch about permission.

The destination of this stop is about 2.4 km downstream from the top of the road, where we will encounter one of a large number of springs emerging from the canyon walls. Springs occur intermittently from about River Mile 13, just below Osborne Canyon, to Opal Springs, about River Mile 6.7. These springs issue from the 5.77-Ma basalt of Opal Springs. The Crooked River gains over 28 m^3/s (1,000 ft^3/s) from ground-water discharge through springs in the basalt of Opal Springs in this reach. Most prolific of these is Opal Springs itself, which discharges from the base of the basalt on the east bank of the river. Discharge from this single spring is estimated to be approximately 7 to 9 m^3/s (250 to 300 ft^3/s) (Stearns, 1931; Robert MacRostie, oral commun).

Isotope and temperature data (Table 1) indicate that the water from Opal Springs has followed a relatively long, deep flow path (Caldwell, 1997; James and others, 2000). The low tritium content indicates that the water was recharged prior to atmospheric testing of nuclear bombs in the early 1950s. Carbon and helium isotope data indicate that the water of Opal Springs contains a component of magmatic gas not present in many springs higher in the basin, such as those at North Davis Creek. Comparing the temperature of Opal Springs (12°C) with the mean annual temperature at the altitude of recharge inferred from oxygen isotope measurements indicates considerable geothermal warming (James and others, 2000).

About 170 m of Deschutes Formation strata are exposed across the valley in the area

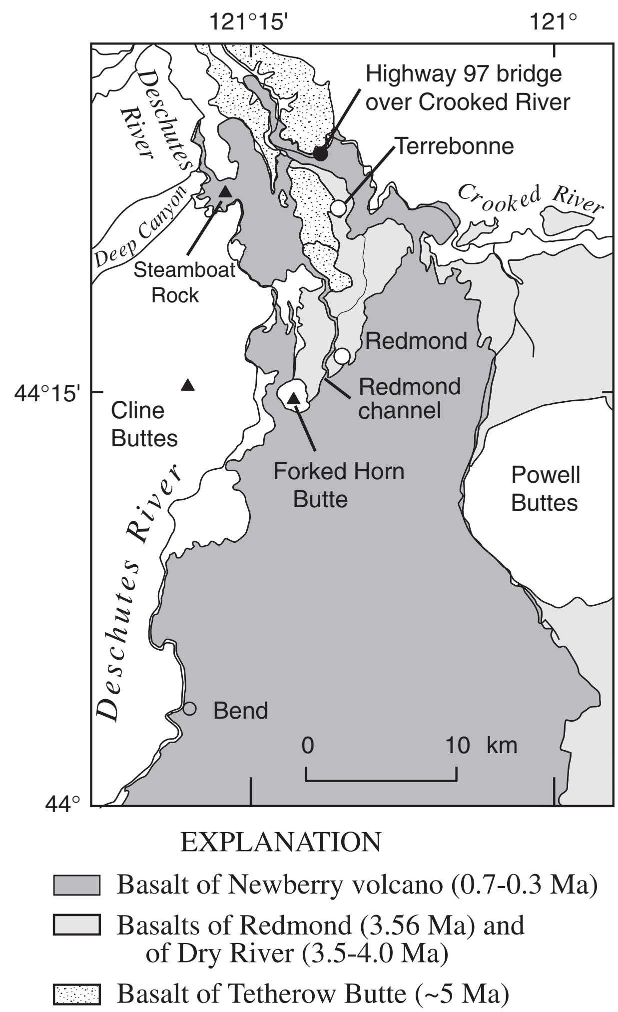


Figure 27. Extent of lava flows in the basalt of Newberry Volcano across the Redmond Plain and into the ancestral canyons of the Deschutes and Crooked Rivers.

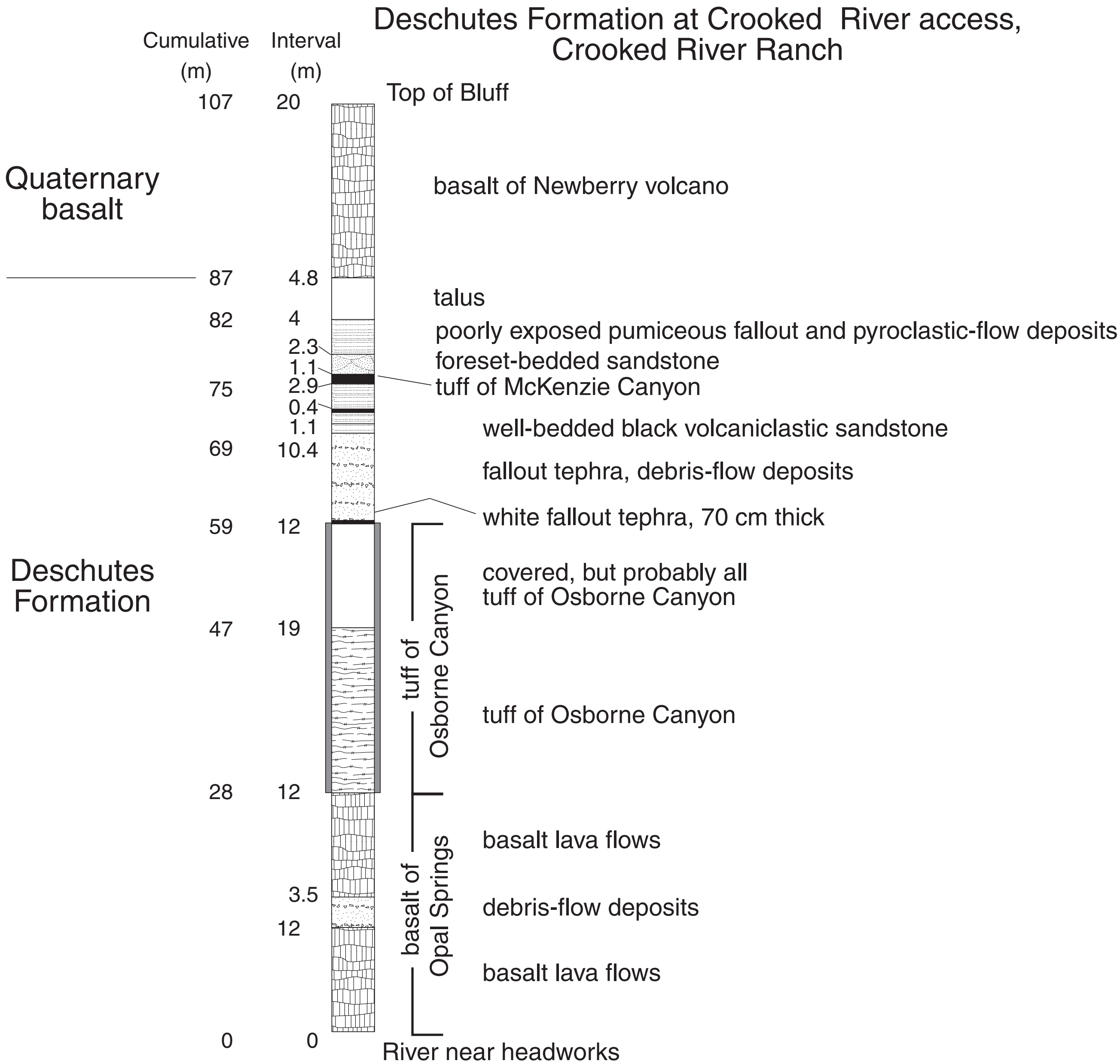


Figure 28. Stratigraphic section exposed along Hollywood Grade in the canyon of the Crooked River near Crooked River Ranch, Stop 10.

of this stop. At the top is the basalt of Tetherow Butte, a spectacularly jointed sequence of tholeiitic basalt 45-60 m thick. In the lower one-third of the valley wall is the orange-weathering tuff of Osborne Canyon, 20-30 m thick (Fig. 28). At the base of the sequence is the basalt of Opal Springs. The Opal Springs unit is exposed through 28 m, and it continues into the subsurface in this area. The basalt of Opal Springs is as thick as 40 m elsewhere in the basin (Smith, 1986a).

This part of the Deschutes Formation was deposited in less than 1 million years. The basalt of Opal Springs, at the canyon floor, has an isotopic age of 5.77±0.07 Ma (Table 2). The capping basalt of Tetherow Butte is probably younger than 5.04 Ma. It has an isotopic age of 5.31±0.05 Ma (Table 2), but it possesses normal-polarity thermal-remanent magnetization. An age younger than 5.04 Ma would agree with the currently accepted paleomagnetic time scale (Fig. 29).

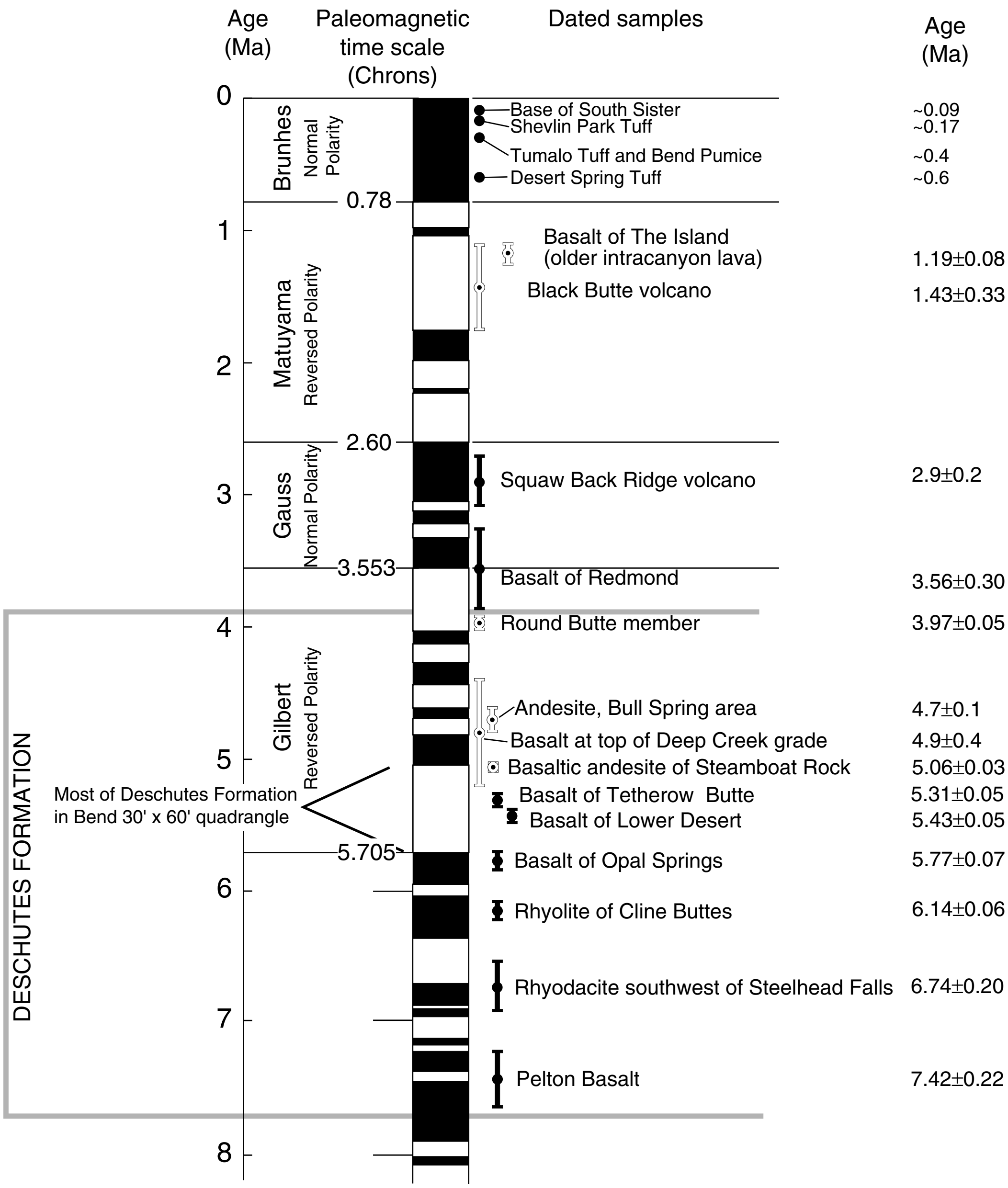


Figure 29. Correlation of selected dated samples with paleomagnetic time scale. Patterns show remanent magnetization: dark fill, normal polarity; white fill, reversed polarity. Bars showing age and standard deviation are similarly patterned. See Table 2 for references to age data. Remanent magnetization determined using portable fluxgate magnetometer (time scale from Cande and Kent, 1992).

Erosion along the ancestral canyon of the Crooked River has carved into and removed the basalt of Tetherow Butte and some underlying sedimentary beds from the area of our traverse. Quaternary intracanyon lava flows have invaded and partly refilled the inner canyon, creating the broad flat bench at Crooked River Ranch. The walk down the grade begins stratigraphically in the middle Pleistocene basalt of Newberry Volcano, then passes into sedimentary and volcanic strata of the Deschutes Formation (Fig. 28). The tuff of McKenzie Canyon, which we saw at Stop 7, lies partway down the grade. It is only 1–2 m thick and easily overlooked. Near the base of the grade is the conspicuous tuff of Osborne Canyon.

Directly beneath the tuff of Osborne Canyon is the basalt of Opal Springs. It comprises two lava-flow sequences, each containing several flows of open-textured olivine basalt. The two flow sequences are separated by 3–6 m of debris-flow deposits and other sedimentary beds. The isotopic age for the basalt of Opal Springs, 5.77±0.07 Ma, came from lava in the lower sequence (Smith, 1986a).

Both the capping basalt of Newberry Volcano and the stream-flooring basalt of Opal Springs possess normal-polarity thermal-remanent magnetization. A third basalt sequence, with reversed-polarity magnetization, underlies the basalt of Newberry Volcano at the base of Hollywood Grade. Named the basalt of The Island for a prominent mesa at Cove Palisade State Park, it is lithologically similar to the basalt of Newberry Volcano and the basalt of Opal Springs. It has an isotopic age of 1.19±0.08 Ma (whole-rock, $^{40}Ar/^{39}Ar$, Smith, 1986a).

The basalt of The Island reputedly was erupted from Newberry Volcano (Peterson and others, 1976; Smith, 1986a; Dill, 1992), but the unit cannot be traced farther south toward Newberry than the area near the Crooked River Ranch (Sidebar 2). Thus, the upstream pathway for the reversed-polarity basalt must now be overlain entirely by normal-polarity basalt of Newberry Volcano. An observation that may argue against a Newberry source is that the upper surface of these flows lies at an altitude of roughly 730 m (2,400 ft) in most locations (Dill, 1992; Ferns and others, 1996a), which would be unlikely if the lava were flowing downstream along its entire extent. An alternative explanation, therefore, is that the basalt of The Island was erupted from some yet-to-be-found fissure vent downstream from (north of) the Crooked River Ranch and backed up along the Crooked River (Sherrod and others, in press).

Stop 11—Lake Billy Chinook Overlook (Reservoir Altitude 593 m; 1,945 ft)

Stop 11 provides a view of the canyons of the Crooked and Deschutes Rivers just upstream from their confluence. Lake Billy Chinook is impounded behind Round Butte Dam, which was completed in 1964. Our east-rim overlook is on the ~5-Ma basalt of Tetherow Butte (Smith, 1986a). Sedimentary rocks of the Deschutes Formation form most of the canyon walls. The far rim (west rim of the canyon of the Deschutes River) is capped by the basalt of Lower Desert (Smith, 1986a). The basalt of Lower Desert has normal-polarity magnetization. Its isotopic age is 5.43±0.05 Ma (Table 2), but its magnetization suggests that it too is about 5 Ma in age (Fig. 29). The narrow flat-topped ridge separating the Crooked and Deschutes Rivers is The Island, an erosional remnant of a reversely polarized 1.19-Ma intracanyon lava flow, the basalt of The Island. Our viewpoint rim is at an altitude of 786 m (2,480 ft). The pool altitude is 593 m (1,945 ft), and the canyon floor, now flooded, is about 488 m (1,600 ft). A photograph taken from this spot in August 1925 by Harold T. Stearns (Fig. 30) provides a glimpse of the prereservoir exposures. In Figure 30B, the conspicuous lava flows at the base of the canyon belong to the 7.42-Ma Pelton Basalt Member, the lowest lava-flow member of the Deschutes Formation (Stearns, 1931; Smith, 1986a).

The Pelton Basalt Member is considered to be the base of the Deschutes Formation as defined by Smith (1986a). It comprises several lava flows of olivine tholeiite, some with thin sedimentary interbeds (Smith, 1986a). A dated flow in the Pelton has an age of 7.42±0.22 Ma (Table 2). The Pelton Basalt is exposed today as far south as Round Butte Dam and was mapped another 6.4 km southward from the mouth of the Crooked River prior to impoundment of Lake Billy Chinook (Stearns, 1931). It is as thick as 30 m near Pelton Dam (Smith, 1987b).

As is the case for the basalt of Opal Springs, considerable ground water discharges from springs in the Pelton Basalt Member. Stearns (1931) observed that all the springs in the lower Crooked River below about River Mile 4 and along the Deschutes from its confluence with the Crooked River to below the confluence with the Metolius River issue from the Pelton. He noted a large spring issuing from it in the forebay of the power plant that once existed about 1.6 km upstream from this point. He also documented a line of springs extending 1.2 km along the west bank of the Deschutes River, about 0.8 km upstream from the Metolius River, with an estimated discharge between 2.3 and 2.8 m^3/s (80 and 100 ft^3/s). Ground-water discharge to Lake Billy Chinook is estimated from stream-gage data to be approximately 12 m^3/s (420 ft^3/s). Most of this

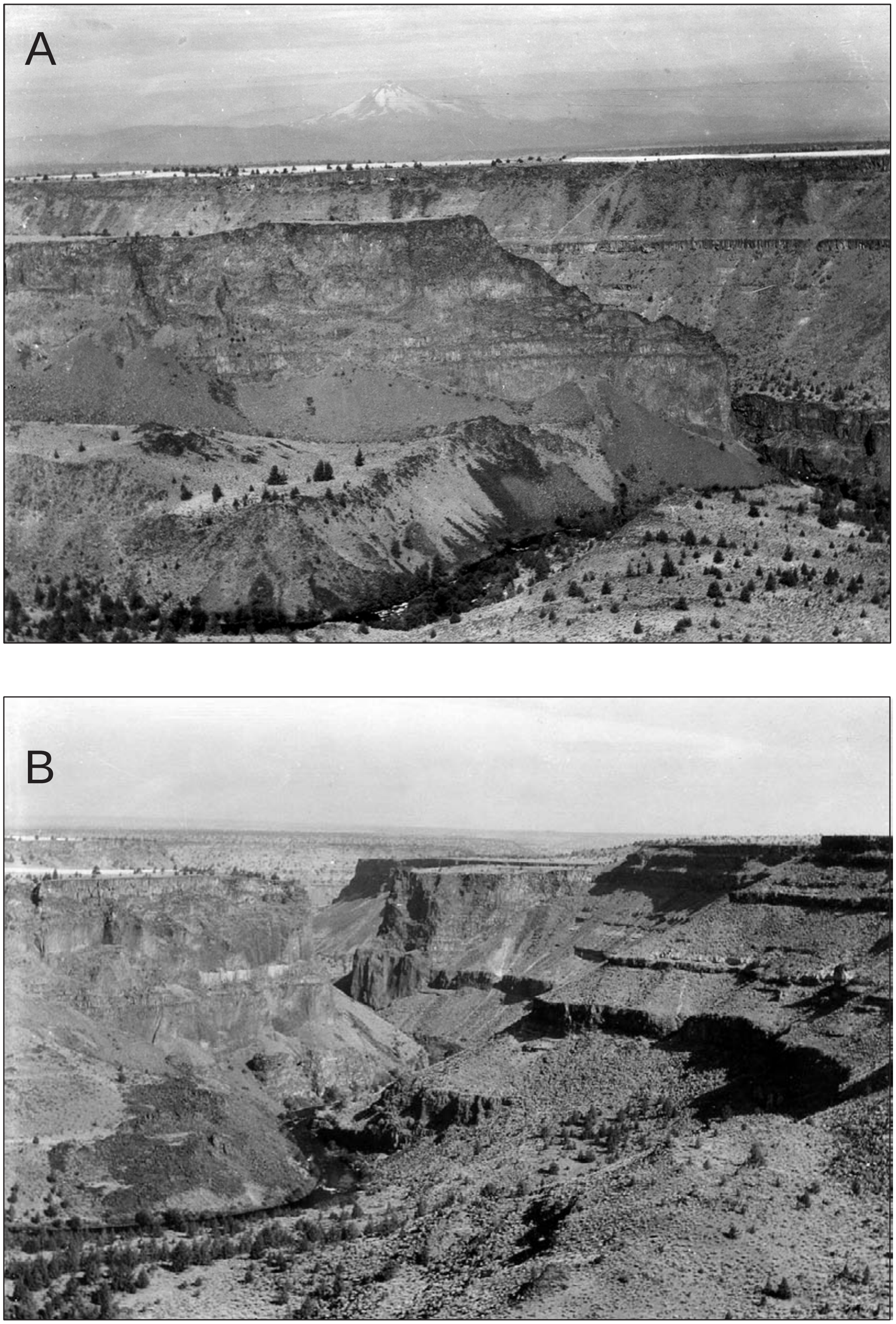


Figure 30. Photographs of the canyons of the Crooked and Deschutes Rivers from the SE 1/4 sec. 35, T. 11 S., R. 12 E. (Stop 11), taken by H.T. Stearns, August 1925, prior to inundation by Lake Billy Chinook. *A.* View west across the canyon of the Crooked River to The Island, and beyond it, to the west rim of the canyon of the Deschutes River. The Deschutes–Crooked confluence is at right side of photo in near ground, Mount Jefferson centered in background 45 km west-northwest. *B.* View north (downstream) into the canyon of the Deschutes River. The Deschutes–Crooked confluence is at lower left edge of photo.

discharge is likely from the Pelton (Gannett and others, 2001).

The Pelton Basalt Member extends in the subsurface south beyond its now-submerged exposure in the canyon. Wells drilled at river level near Opal Springs (River Mile 6.7) penetrated the Pelton at a depth of approximately 110 m. It proved to be a productive aquifer at this place. Wells encountered an artesian pressure in the Pelton Basalt Member of approximately 50 psi at the land surface and artesian flow rates of up to 0.32 m^3/s (5,000 gallons per minute).

The Pelton Basalt Member is the lowest sequence of lava in the Deschutes Formation, and although Smith (1987a,b) maps a thin section of Deschutes sediment underlying it, the Pelton is effectively the base of the permeable section in the Deschutes Formation. The Deschutes contact with underlying units is exposed about 8 km north of this stop. By that point, most of the ground water flowing from the upper basin in the Deschutes Formation has discharged to the river system. Depending on location, the Deschutes Formation may be seen in unconformable contact with the underlying middle Miocene Simtustus Formation or the upper Oligocene to lower Miocene John Day Formation (Fig. 2). Mapping by Smith (1987a, b) shows that the Simtustus Formation and Prineville Basalt typically separate the Deschutes and John Day Formations in the canyon of the Deschutes River. West of the canyon, the Deschutes Formation lies directly on the John Day Formation in many places.

The Simtustus Formation is a sequence of middle Miocene volcanogenic sandstone, mudstone, and tuff conformable on and interbedded with lava of the Prineville Basalt (Smith, 1986b). The Simtustus Formation, which is as thick as 65 m, was deposited across an area almost 20 km wide. Deposition occurred in response to drainage-system disruption by lava flows of the Columbia River Basalt Group and the Prineville Basalt (Smith, 1986b). The Simtustus Formation is exposed in areas ranging from the canyon of the Deschutes River to east of Gateway. Its hydrologic characteristics are unknown. Because of its location, limited areal extent, and relative thinness, it is a hydrologically insignificant stratigraphic unit.

The Prineville Basalt is a sequence of relatively evolved lava flows exposed sporadically across north-central Oregon. Of middle Miocene age, it includes flows with both normal- and reversed-polarity magnetization that are thought to have erupted during a short time period about 15.8 Ma when the Earth's magnetic field was changing polarity from reversed to normal (Hooper and others, 1993). In the Deschutes Basin, the Prineville Basalt is as thick as 200 m. It is exposed east of Powell Buttes, east of Smith Rock, and along the Deschutes River from Pelton Dam downstream toward Cow Canyon, 20 km northeast of Gateway (Smith, 1986a). The Prineville Basalt may underlie the Deschutes Formation in much of the east half of the basin. Stratigraphic separation between the two formations is relatively small, depending on the thickness of intervening middle Miocene strata of the Simtustus Formation.

The hydrologic characteristics of the Prineville Basalt are poorly known. It is generally less permeable than the Deschutes Formation, and we know of no published reports of major springs issuing from the unit. It is, however, used locally as a source of water from domestic wells and a few irrigation wells.

About 13 km north of this place, just below Pelton Dam, the John Day Formation is exposed in the canyon of the Deschutes River. The John Day Formation in this area consists primarily of light-colored tuff, lapillistone, fine-grained volcanic sandstone, and mudstone (Smith, 1987a, b). The John Day Formation has very low permeability and is considered the basement of the regional ground-water flow system (Sceva, 1968; Gannett and others, 2001). In contrast to the upper basin, little ground water discharges to the Deschutes River downstream from the point where it intersects the John Day Formation. The John Day Formation can be observed at optional Stop 12.

Stop 12—Canyon of the Deschutes River Near Warm Springs, Optional Stop (Altitude 421 m; 1,380 ft)

At this place the John Day Formation forms the lower canyon walls on either side of the Deschutes River. These strata have limited permeability. The permeable strata of the

Deschutes Formation are above the river at this point and thin abruptly northward. Virtually all of the regional ground-water flow has discharged to the Deschutes River and its tributaries upstream from this point. During the summer months, almost the entire flow of the Deschutes River at this place comes from ground-water discharge. Year-round, approximately 90 percent of the mean discharge of the Deschutes River here is attributable to ground water. In the 160 km between here and the mouth of the river, the Deschutes gains very little water from ground-water discharge, and its increases in flow are primarily from tributary streams.

ROAD GUIDE

Directions to Stop 1—Springs at North Davis Campground

From Bend: South on US 97 approximately 15 mi to Vandevert Road (6 mi south of entrance to Lava Lands Visitor Center). West on Vandevert Road 1.0 mi to South Century Drive. South on South Century Drive 1.1 mi to County Road 42 (which becomes Forest Road 42). West on Forest 42 for 23 mi to Forest 46. South on Forest 46 for 3.9 mi to North Davis Campground.

From Willamette Pass: Follow Oregon 58 southeast 5.5 mi beyond Odell Lake east access road to Crescent Cutoff Road. East on Crescent Cutoff Road (Forest 61) 3.2 mi to Forest 46. North 14.1 mi on Forest 46 to North Davis Campground.

From the south on US 97: In town of Crescent turn northwest onto Crescent Cutoff Road (County 61; may be marked as Ward Street in town). West on Crescent Cutoff Road 9 mi to Forest 46. North 14.1 mi on Forest 46 to North Davis Campground.

Directions from Stop 1 to Stop 2—Pringle Falls Sedimentary Section

Forest 46 north 3.9 mi to South Century Drive (Forest 42). East 9.3 mi to Burgess Road (Forest 43). East 3.6 mi to Forest 500 on left (north) side of the road after crossing Deschutes River. North on Forest 500 about 0.3 mi to fork. Veer left and follow road along power lines about 0.5 mi to where the road is adjacent to the Deschutes River.

Directions from Stop 2 to Stop 3—Fall River Headwater Springs

Return to Forest 43. Turn right, proceed west 0.6 mi to Forest 4350. North on Forest 4350 for 2.1 mi to Forest 42. East on Forest 42 about 0.1 mi to Fall River Guard Station. Turn into the guard station and park.

Directions from Stop 3 to Stop 4—Top of Lava Butte

Leave Fall River Guard Station and continue east 10.7 mi on Forest 42 to South Century Drive. North 1.1 mi to Vandevert Road. East 1.0 mi on Vandevert Road to US 97. North 6.0 mi on US 97 to Lava Lands Visitor Center. Depending on time of year, a fee may be charged to enter the center and its roads. Road leads approximately 1.7 mi to top of Lava Butte.

Directions from Stop 4 to Stop 5—Surface-Water Diversions in Bend

North 11.5 mi on US 97, which becomes the newly completed Bend Parkway through downtown Bend. Take Exit 137 (Revere Avenue—Downtown) straight through traffic signal onto Division Street. Proceed on Division Street 0.5 mi from traffic signal to Riverview Park's tiny parking area on west side of road. Diversion dam and headgates are seen just north of parking area.

Directions from Stop 5 to Stop 6—Sisters Fault Zone Outcrop, Tumalo Reservoir Road

North 0.2 mi on Division to intersection with US 97 business route. North 0.3 mi to O.B. Riley Road. West 4.1 mi on O.B. Riley Road to Tumalo Reservoir Road, which lies just beyond crossing of Deschutes River. West (left) 0.5 mi on Tumalo Reservoir Road, then park on road shoulder. Stop 6 is prominent exposure of pyroclastic deposits quarried north of road.

Directions from Stop 6 to Stop 7—Bull Flat and Tumalo Dam

West 0.6 mi on Tumalo Reservoir Road to T intersection. Left (west) 3.5 mi to where the road turns north and becomes Sisemore Road. North 2.3 mi on Sisemore Road (along the Tumalo Fault) to Tumalo Dam. Find place to park just west past the dam.

Directions from Stop 7 to Stop 8—Lower Bridge

Turn around and go east beyond Tumalo Dam 0.1 mi to Couch Market Road (Y intersection). East 3.5 mi to US 20. Southeast 2.4 mi to Cook Avenue in small town of Tumalo (see sign for Cline Falls State Park). North 10.3 mi on Cook Avenue (becomes Cline Falls Highway after a few blocks) to Oregon 126. West (toward Sisters) 3.7 mi on Oregon 126 to Buckhorn Road. North 4.2 mi on Buckhorn Road to Lower Bridge Road. East 1.6 mi to Deschutes River. Park just before crossing the bridge (south side).

Directions from Stop 8 to Stop 9—Peter Skene Ogden Bridge and Scenic Viewpoint

East 6.2 mi on Lower Bridge Road to US 97. North 2.4 mi on US 97 to wayside at Ogden Scenic Viewpoint. Pull into wayside, then walk to old highway bridge, which is now a pedestrian walkway.

Directions from Stop 9 to Stop 10—Crooked River Gorge at Crooked River Ranch

Return south 1.4 mi on US 97 to Wimp Way. West 0.3 mi on Wimp Way, which curves north to Ice Avenue. West 0.2 mi to 43rd Street. North 0.8 mi to a T intersection at Chinook Drive. West (then curves more or less north) 3.3 mi to Clubhouse Road. East and southeast 0.4 mi on Clubhouse Road past clubhouse and stores; park near chapel. Walk through parking lot for motel and hike into the canyon of the Crooked River on a road that starts beyond the gate south of the motel parking lot. The road crosses private property. Check at the office of the Crooked River Ranch for information on access to the road through the gate. Alternatively, check with motel for permission to follow the fence line north to the canyon rim, where the road is reached by an easy scramble. Watch for rattlesnakes.

Directions from Stop 10 to Stop 11—Lake Billy Chinook Overlook

Return to US 97 by retracing route from Stop 9. North 8.5 mi on US 97 to Culver Road. East, then north, 2.6 mi on Culver Road to C Street in town of Culver. Signs from here help by pointing route toward The Cove Palisades State Park. West 1.0 mi on C Street to Feather Drive. North 0.9 mi on Feather Drive to Fisch Lane. West 0.5 mi on Fisch to where it turns north and becomes Frazier. Follow Frazier 0.5 mi to intersection with Jordan Road. West on Jordan Road 0.3 mi to Mountainview (look for the sign to View Points). North on Mountainview, following it 2.1 mi along rim of the canyon of the Crooked River to third scenic viewpoint on left. Failure to turn onto Mountainview will result in a geologically rewarding drive into Cove Palisades State Park.

Directions from Stop 11 to Optional Stop 12—Canyon of the Deschutes River Near Warm Springs

Continue north on Mountainview about 2.2 mi to an intersection. Turn left and continue north on Mountainview 2.6 mi to a T intersection. Turn right on Belmont Lane and travel east 1.5 mi to Elk Drive. Left (north) on Elk Drive for 2.8 mi, where it turns left down into the canyon and becomes Pelton Dam Road. Follow Pelton Dam Road 6.8 mi north to its intersection with US 26. Turn left (north) on US 26 and go about 1.5 mi to where the canyon narrows. This is the stop. US 26 lacks pullouts in the area, and the road shoulder is narrow. Use extreme caution if disembarking to look at outcrops. A wayside and boat ramp about 0.7 mi north of this place provide an opportunity to examine the river.

REFERENCES CITED

Armstrong, R.L., Taylor, E.M., Hales, P.O., and Parker, D.J., 1975, K-Ar dates for volcanic rocks, central Cascade Range of Oregon: Isochron/West, no. 13, p. 5-10.

Caldwell, R.R., 1997, Chemical study of regional ground-water flow and ground-water/surface-water interaction in the upper Deschutes Basin, Oregon: U.S. Geological Survey Water-Resources Investigations Report 97-4233, 49 p.

Cande, S.C., and Kent, D.V., 1992, A new geomagnetic polarity time scale for the Late Cretaceous and Cenozoic: Journal of Geophysical Research, v. 97, p. 13,917-13,951.

Chitwood, L.A., Jensen, R.A., and Groh, E.A., 1977, The age of Lava Butte: Ore Bin, v. 39, p. 157-164.

Couch, R., and Foote, R., 1985, The Shukash and La Pine Basins: Pleistocene depressions in the Cascade Range of central Oregon: Eos, v. 66, no. 3, p. 24.

Dill, T.E., 1992, Stratigraphy of the Neogene volcanic rocks along the lower Metolius River, Jefferson County, central Oregon: Oregon State University MS Thesis, 343 p.

Evans, S.H., and Brown, F.H., 1981, Summary of Potassium/Argon dating—1981: U.S. Department of Energy, Division of Geothermal Energy DE-AC07-80-ID-12079-45, 29 p.

Ferns, M.L., Lite, K.E., Jr., and Clark, M.D., 1996b, Lithologic controls on groundwater discharge to the Deschutes River between Lower Bridge and Lake Billy Chinook, central Oregon: Geological Society of America Abstracts with Programs, v. 28, no. 5, p. 65.

Ferns, M.L., Stensland, D.E., and Smith, G.A., 1996a, Geologic map of the Steelhead Falls Quadrangle, Deschutes and Jefferson Counties, Oregon: Oregon Department and Geology and Mineral Industries Geological Map Series GMS-101, scale 1:24,000.

Fiebelkorn, R.B., Walker, G.W., MacLeod, N.S., McKee, E.H., and Smith, J.G., 1983, Index to K-Ar age determinations for the State of Oregon: Isochron/West, no. 37, p. 3-60.

Gannett, M.W., and Lite, K.E., Jr., 2000, Climate-driven fluctuations in hydraulic head and groundwater discharge to streams in the upper Deschutes Basin, Oregon: Eos, v. 81, no. 48, p. F531-F532.

Gannett, M.W., Lite, K.E., Jr., Morgan, D.S., and Collins, C.A., 2001, Ground-water hydrology of the upper Deschutes Basin, Oregon: U.S. Geological Survey Water Resources Investigation Report 00-4162, 77 p.

Gettings, M.E., and Griscom, A., 1988, Gravity model studies of Newberry Volcano, Oregon: Journal of Geophysical Research, v. 93, p. 10,109-10,118.

Hawkins, F.F., LaForge, R.C., and Gilbert, J.D., 1989, Seismotectonic study for Wickiup and Crane Prairie Dams, Deschutes Project, Oregon: U.S. Bureau of Reclamation Seismotectonic Report 89-2, 38 p.

Herrero-Bervera, E., and Helsey, C.E., 1993, Global paleomagnetic correlation of the Blake geomagnetic polarity episode, in Applications of paleomagnetism to sedimentary geology: SEPM Special Publication 49, p. 71-82.

Herrero-Bervera, E., Helsey, C.E., Hammond, S.R., and Chitwood, L.A., 1989, A possible lacustrine paleomagnetic record of the Blake episode from Pringle Falls, Oregon, U.S.A.: Physics of the Earth and Planetary Interiors, v. 56, p. 112-123.

Herrero-Bervera, E., Helsey, C.E., Sarna-Wojcicki, A.M., Lajoie, K.R., Meyer, C.E., McWilliams, M.O., Negrini, R.M., Turrin, B.D., Donnelly-Nolan, J.M., and Liddicoat, J.C., 1994, Age and correlation of a paleomagnetic episode in the western United States by $^{40}Ar/^{39}Ar$ dating and tephrochronology: The Jamaica, Blake, or a new polarity episode?: Journal of Geophysical Research, v. 99, p. 24,091-24,103.

Hill, B.E., 1992, Petrogenesis of compositionally distinct silicic volcanoes in the Three Sisters region of the Oregon Cascade Range: The effects of crustal extension on the development of continental arc silicic magmatism: Oregon State University PhD Dissertation, 235 p.

Hill, B.E., and Duncan, R.A., 1990, The timing and significance of silicic magmatism in the Three Sisters region of the Oregon High Cascades: Eos, v. 71, p. 1,614.

Hill, B.E., and Priest, G.R., 1992, Geologic setting of the Santiam Pass area, central Cascade Range, Oregon, *in* Hill, B.E., ed., Geology and geothermal resources of the Santiam Pass area of the Oregon Cascade Range, Deschutes, Jefferson, and Linn Counties, Oregon: Oregon Department of Geology and Mineral Industries Open-File Report O-92-3, p. 5-18.

Hooper, P.R., Steele, W.K., Conrey, R.M., Smith, G.A., Anderson, J.L., Bailey, D.G., Beeson, M.H., Tolan, T.L., and Urbanczyk, K.M., 1993, The Prineville Basalt, north-central Oregon: Oregon Geology, v. 55, p. 3-12.

James, E.R., Manga, M., Rose, T.P., and Hudson, G.B., 2000, The use of temperature and the isotopes of O, H, C, and noble gases to determine the pattern and spatial extent of groundwater flow: Journal of Hydrology, v. 237, p 100-112.

Krebs, W.N., Bradbury, J.P., and Theriot, E., 1989, Neogene and Quaternary lacustrine diatom biochronology, western U.S.A.: Palaios, v. 2, p. 505-513.

Lanphere, M.A., Champion, D.E., Christiansen, R.L., Donnelly-Nolan, J.M., Fleck, R.J., Sarna-Wojcicki, A.M., Obradovich, J.D., and Izett, G.A., 1999, Evolution of tephra dating in the western United States: Geological Society of America Abstracts with Programs, v. 31, no. 6, p. A73.

MacLeod, N.S., and Sherrod, D.R., 1992, Reconnaissance geologic map of the west half of the Crescent 1 by 2 Degree Quadrangle, central Oregon: U.S. Geological Survey Miscellaneous Investigations Series Map I-2215, scale 1:250,000.

MacLeod, N.S., Sherrod, D.R., Chitwood, L.A., and Jensen, R.A., 1995, Geologic map of Newberry Volcano, Deschutes, Klamath, and Lake Counties, Oregon: U.S. Geological Survey Miscellaneous Investigations Series Map I-2215, scales 1:62,500 and 1:24,000.

Manga, M., 1998, Advective heat transport by low-temperature discharge in the Oregon Cascades: Geology, v. 26, p. 799-802.

Moore, B.N., 1937, Nonmetallic mineral resources of eastern Oregon: U.S. Geological Survey Bulletin 875, 180 p.

Peterson, N.V., Groh, E.A., Taylor, E.M., and Stensland, D.E., 1976, Geology and mineral resources of Deschutes County, Oregon: Oregon Department of Geology and Mineral Industries Bulletin 89, 66 p.

Pitts, G.S., and Couch, R.W., 1978, Complete Bouguer gravity anomaly map, Cascade Mountain Range, central Oregon: Oregon Department of Geology and Mineral Industries Geological Map Series GMS-8, scale 1:125,000.

Rieck, H.J., Sarna-Wojcicki, A.M., Meyer, C.E., and Adam, D.P., 1992, Magnetostratigraphy and tephrochronology of an upper Pliocene to Holocene record in lake sediments at Tulelake, northern California: Geological Society of America Bulletin, v. 104, p. 409-428.

Robinson, P.T., and Stensland, D.E., 1979, Geologic map of the Smith Rock area, Jefferson, Deschutes, and Crook Counties, Oregon: U.S. Geological Survey Miscellaneous Investigations Map I-1142, scale 1:48,000.

Sarna-Wojcicki, A.M., Meyer, C.E., Nakata, J.K., Scott, W.E., Hill, B.E., Slate, J.L., and Russell, P.C., 1989, Age and correlation of mid-Quaternary ash beds and tuffs in the vicinity of Bend, Oregon, *in* Scott, W.E., Gardner, C.A., and Sarna-Wojcicki, A.M., eds., Guidebook for field trip to the Mount Bachelor–South Sister–Bend area, central Oregon High Cascades: U.S. Geological Survey Open-File Report 89-645, p. 55-62.

Sceva, J.E., 1968, Liquid waste disposal in the lava terrane of central Oregon: U.S. Department of the Interior, Federal Water Pollution Control Administration

Technical Projects Branch Report No. FR-4, 66 p., app. 96 p.

Scott, W.E., and Gardner, C.A., 1992, Geologic map of the Mount Bachelor volcanic chain and surrounding area, Cascade Range, Oregon: U.S. Geological Survey Miscellaneous Investigations Map I-1967, scale 1:50,000.

Sherrod, D.R., and Pickthorn, L.G., 1989, Some notes on the Neogene structural evolution of the Cascade Range in Oregon, *in* Muffler, L.J.P., Weaver, C.S., and Blackwell, D.D., eds., Geology, geophysics, and tectonic setting of the Cascade Range: U.S. Geological Survey Open-File Report 89-178, p. 351-368.

Sherrod, D.R., Taylor, E.M., Ferns, M.L., Scott, W.E., Conrey, R.M., and Smith, G.A., in press, Geologic map of the Bend 30 by 60 Minute Quadrangle, central Oregon: U.S. Geological Survey Miscellaneous Investigations Map I-2683, scale 1:100,000.

Smith, G.A., 1986a, Stratigraphy, sedimentology, and petrology of Neogene rocks in the Deschutes Basin, central Oregon: A record of continental-margin volcanism and its influence on fluvial sedimentation in an arc-adjacent basin: Oregon State University PhD Dissertation, 467 p.

Smith, G.A., 1986b, Simtustus Formation: Paleogeographic and stratigraphic significance of a newly defined Miocene unit in the Deschutes Basin, central Oregon: Oregon Geology, v. 48, p. 63-72.

Smith, G.A., 1987a, Geologic map of the Seekseequa Junction and a portion of the Metolius Bench Quadrangles, Jefferson County, Oregon: Oregon Department of Geology and Mineral Industries Geological Map Series GMS-44, scale 1:24,000.

Smith, G.A., 1987b, Geologic map of the Madras West and Madras East Quadrangles, Jefferson County, Oregon: Oregon Department of Geology and Mineral Industries Geological Map Series GMS-45, scale 1:24,000.

Smith, G.A., and Hayman, G.A., 1987, Geologic map of the Eagle Butte and Gateway Quadrangles, Jefferson and Wasco Counties, Oregon: Oregon Department of Geology and Mineral Industries Geological Map Series GMS-43, scale 1:24,000.

Smith, G.A., Manchester, S.R., Ashwill, M., McIntosh, W.C., and Conrey, R.M., 1998, Late Eocene–early Oligocene tectonism, volcanism, and floristic change near Gray Butte, central Oregon: Geological Society of American Bulletin, v. 110, p. 759-778.

Smith, G.A., Snee, L.W., and Taylor, E.M., 1987, Stratigraphic, sedimentologic, and petrologic record of late Miocene subsidence of the central Oregon High Cascades: Geology, v. 15, p. 389-392.

Stearns, H.T., 1931, Geology and water resources of the middle Deschutes River Basin, Oregon: U.S. Geological Survey Water-Supply Paper 637-D, p. 125-212.

Stensland, D.E., 1970, Geology of part of the northern half of the Bend Quadrangle, Jefferson and Deschutes Counties, Oregon: Oregon State University MS Thesis, 118 p.

Taylor, E.M., and Ferns, M.L., 1994, Geology and mineral resource map of the Tumalo Dam Quadrangle, Deschutes County, Oregon: Oregon Department of Geology and Mineral Industries Geological Map Series GMS-81, scale 1:24,000.

Walker, G.W., and MacLeod, N.S., 1992, Geologic map of Oregon: U.S. Geological Survey, scale 1:500,000.

Williams, H., 1957, A geologic map of the Bend Quadrangle, Oregon, and a reconnaissance geologic map of the central portion of the High Cascade Mountains: Oregon Department of Geology and Mineral Industries, scale 1:125,000 and 1:250,000.

Winch, M.T., 1984-85, Tumalo—Thirsty land: Oregon Historical Quarterly, v. 85, p. 341-374; v. 86, p. 47-79, 153-182, 269-297.