Inaugural Bertz Club Field Trip, May 2010

We will be using a portion of Cashman et al. (2009)'s GSA field guide, "Fire and water: Volcanology, geomorphology, and hydrology of the Cascade Range, central Oregon" for this field trip. We are providing the relevant portion of the field guide (and encourage you to acquire the entire volume of field trips associated with the 2009 GSA Portland meeting, available from the GSA website). Below is a schedule for the field trip. The pertinent page numbers in the field guide are listed.

Stops 2 and 4 will involve hiking, so bring appropriate footwear and rain gear, and stop 5 may involve some rowing. We will have lunch at stop 3 – this will be a bag lunch which you will make yourself Saturday morning.

Sometime on Friday:

Lookout creek at HJ Andrews (p. 553)
Example of a western Cascades stream

Saturday:

8:30 am	Depart HJ Andrews
9:00 am	Stop 1: Lost Spring (p. 546 – 548) *Groundwater system*
9:45 am	Stop 2: Proxy Falls (p. 551 – 552) *Hike to Proxy falls, where a waterfall feeds the Collier Cone lava flow groundwater supply*
11:30 am	Stop 3: Olallie Creek (p. 554) *Example of a high Cascades stream (spring fed system)*
Bag lunch	
1:00 pm	Stop 4: Carmen Reservoir & upper McKenzie River (p. 555 – 557) *View the dry channel, hike from Carmen Reservoir to Sahalie falls*
3:00 pm	Stop 5: Clear Lake (p. 557 – 559) *View drowned trees and possible row boat excursion to Great Spring and lava flows.*
4:30 pm	Field trip ends at Clear Lake

The Geological Society of America
Field Guide 15
2009

Fire and water: Volcanology, geomorphology, and hydrogeology of the Cascade Range, central Oregon

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ABSTRACT

This field trip guide explores the interactions among the geologic evolution, hydrology, and fluvial geomorphology of the central Oregon Cascade Range. Key topics include the geologic control of hydrologic regimes on both the wet and dry sides of the Cascade Range crest, groundwater dynamics and interaction between surface and groundwater in young volcanic arcs, and interactions between rivers and lava flows. As we trace the Willamette and McKenzie Rivers back to source springs high in the young volcanic rocks of the Cascade Range, there is abundant evidence for the large permeability of young lava flows, as manifested in streams that dewater into lava flows, lava-dammed lakes in closed basins, and rivers that emerge from single springs. These dynamics contrast sharply with the older, lower permeability Western Cascades terrane and associated runoff-dominated fluvial systems. On the east side of the Cascades we encounter similar hydrologic characteristics resulting in complex interactions between surface water and groundwater as we follow the Deschutes River downstream to its confluence with the Crooked River. Here, deep canyons have cut through most of the permeable part of the geologic section, have been invaded by multiple large intracanyon lava flows, and are the locus of substantial regional groundwater discharge. The groundwater and surface-water interaction in the Deschutes Basin is further complicated by surface-water diversions and an extensive network of leaking irrigation canals. Our west-to-east transect offers an unparalleled opportunity to examine the co-evolution of the geology and hydrology of an active volcanic arc.

Cashman, K.V., Deligne, N.I., Gannett, M.W., Grant, G.E., and Jefferson, A., 2009, Fire and water: Volcanology, geomorphology, and hydrogeology of the Cascade Range, central Oregon, *in* O'Connor, J.E., Dorsey, R.J., and Madin, I.P., eds., Volcanoes to Vineyards: Geologic Field Trips through the Dynamic Landscape of the Pacific Northwest: Geological Society of America Field Guide 15, p. 539–582, doi: 10.1130/2009.fld015(26). For permission to copy, contact editing@geosociety.org.

INTRODUCTION AND TRIP OVERVIEW

The Cascade Range of Oregon is the result of ~40 million years of evolution of an active volcanic arc located on the leading edge of a continental plate downwind of an enormous source of moisture—the Pacific Ocean. The focus of our trip will be the interplay among these factors, particularly as expressed in the recent (Quaternary) history and styles of volcanism associated with the convergent plate margin just offshore, the flow paths and volumes of groundwater systems located within the range, and the geomorphic evolution of rivers that have developed on both sides of the Cascade Range crest.

This is very much a "West- and East-side story." The north-trending Cascade Range is perpendicular to the path of the prevailing westerly winds, creating a strong orographic lift and rain shadow effect. Precipitation on the windward western slopes of the Cascades ranges from 2000 to over 4000 mm per year. On the leeward east side, precipitation drops tenfold to around 300 mm per year, resulting in one of the steepest precipitation gradients in the United States. The topography, hydrogeology, vegetation, and landscape evolution reflect this sharp contrast in precipitation and provide a unique natural laboratory for examining both geologic and climatic controls on landscape development.

Our trip will begin in Portland and proceed south through the Willamette Valley toward the headwaters of the Willamette River (Fig. 1). As we travel south on Interstate 5, to the west (right) across the broad valley floor rises the Oregon Coast Range, an uplifted and dissected continental shelf of early Tertiary age. To the east (left), one can see the foothills of the Western Cascades, and, if the day is clear, an occasional glimpse of the large stratovolcanoes along the crest of the High Cascades. Along the way, we will cross several large rivers draining the Coast and Cascade Ranges.

At Eugene, we will turn east and follow the McKenzie River, a major tributary of the Willamette River. The McKenzie River and its tributaries, Lost Creek and White Branch, will lead us to the crest of the Cascade Range at McKenzie Pass. The stops on Day 1 are intended to display aspects of the regional geology, climatology, and landscape history, including the striking contrast between the Western and High Cascades geology, topography and dissection patterns, features and styles of Holocene lava flows, westside Cascade Range spring systems, and various erosional processes, including glacial erosion and mass movements. We will spend the night at the H.J. Andrews Experimental Forest, a National Science Foundation (NSF)–funded Long-Term Ecological Research site, which is surrounded by magnificent old-growth forest.

On the morning of the second day, we will continue along the McKenzie River to its source at Clear Lake, stopping to observe spectacular waterfalls, lava-filled canyons, and dry riverbeds. At Clear Lake, which owes its origin to a lava flow that dammed the river, we will board rowboats for the short crossing to the Great Spring, a high-volume cold spring emanating from the base of a late Holocene lava flow, and discuss different river and lava flow interactions. We will then drive over Santiam Pass, stopping to observe a range of young volcanic structures and styles and another very large spring and river system, the Metolius River.

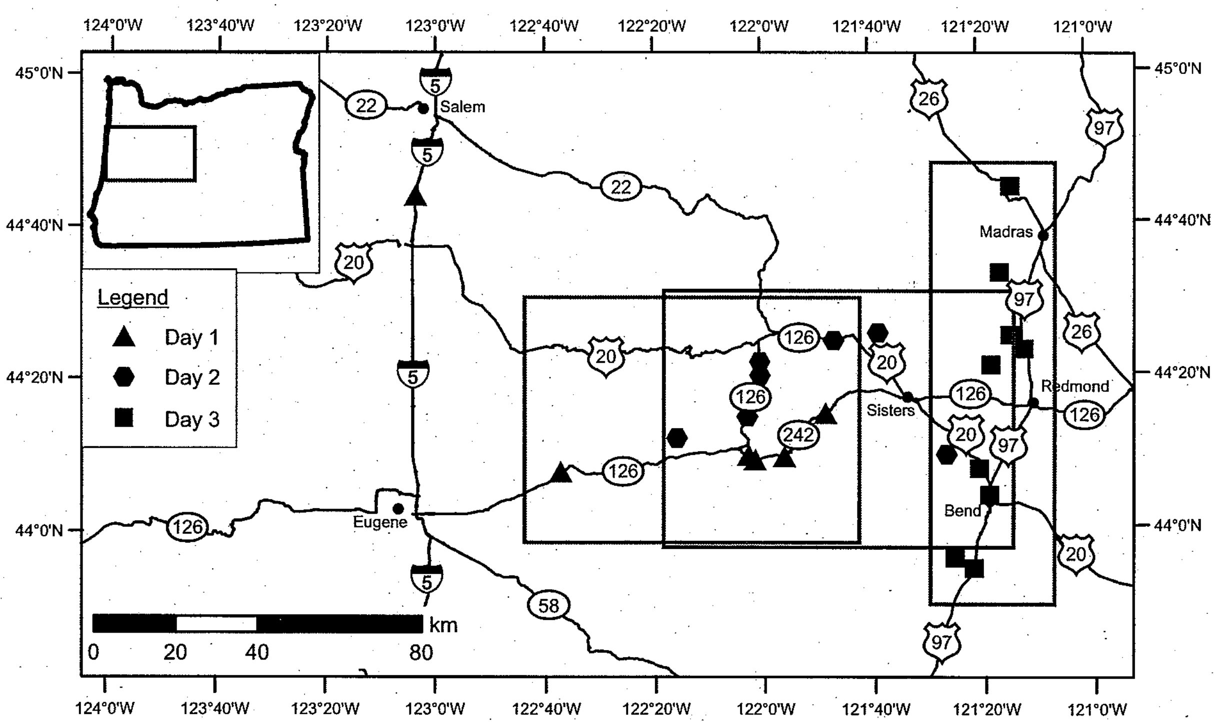

Figure 1. Map of northwestern Oregon showing field trip route and locations of individual stops for each day, plus areas covered by more detailed maps for each day of the field trip.

We will end the day at Bull Flat and Tumalo Dam, with an introduction to the east side story. Overnight will be in Bend, Oregon.

The third and final day of our trip will begin with a focus on lava and river interactions in the vicinity of Lava Butte, a Holocene cinder cone that was the source of lava flows that encroached on, and temporarily dammed, the Deschutes River. We will then travel north along the Deschutes River from Bend, where the river is hundreds of meters above the water table, to Lower Bridge, where the river is at the same elevation as the regional water table, and on to Pelton Dam, where the river has cut entirely through the permeable part of the geologic section and, hence, the regional groundwater flow system. This day will include a hike into the Crooked River canyon to observe the stratigraphy of diverse deposits of the Pliocene Cascade Range (the Deschutes Formation), Pleistocene intracanyon lava flows, and the springs through which the regional groundwater returns to the surface. Along the way we will consider the regional groundwater story in relation to the volcanic history and current issues and conflicts in water management associated with this groundwater system. Finally we will climb out of the Deschutes Basin, cross the Cascades on the lower slopes of Mount Hood, and return to Portland.

The discussion for last stop on the second day, Tumalo Dam, and discussions and figures for most stops on the third day are taken verbatim, or nearly so, from an earlier field trip guide by Dave Sherrod, Marshall Gannett, and Ken Lite (Sherrod et al., 2002).

A Brief Overview of the Geologic and Physiographic Setting of the Cascade Range

The Cascade Range extends from northern California to southern Canada. In central Oregon, the Cascade Range is 50–120 km wide, bounded on the west by the Willamette Valley and on the east by the Deschutes Basin. The Cascade volcanic arc has been active for ~40 million years due to the convergence of the Pacific and Juan de Fuca plates, although volcanism has not been continuous in either space or time throughout this period (Sherrod and Smith, 1990). In Oregon, the Cascade Range is commonly divided into two physiographic subprovinces—the Western Cascades and the High Cascades—which differ markedly in their degree of dissection, owing mainly to the near absence of Quaternary and Pliocene volcanoes in the Western Cascades. From the Three Sisters north to Mount Hood, the young (Quaternary to Pliocene) High Cascades occupy a structural graben formed by a northward-propagating rift (Sherrod et al., 2002), which has affected both the composition of the erupted magma and patterns of groundwater flow.

The Western Cascades are composed of a thick, mixed assemblage of mafic lava flows, mostly of andesitic composition, and ash-flow and ash-fall tuffs, with minor silicic intrusive bodies and stocks, which range in age from middle Eocene to early Pliocene (40–5 million years). Rocks along the western margin of the Western Cascades tend to be older and decrease in age toward the boundary with the High Cascades; some intracanyon lava flows from the High Cascades fill the upper valleys. Rocks have been locally altered by hydrothermal processes, particularly in the contact aureoles surrounding granitic stocks. The landscape has been covered repeatedly by montane glaciers, dissected by rivers, and is prone to frequent mass wasting by landslides, debris flows, and earthflows. Consequently, the topography is extremely rugged, ranging in elevation from 200 to 1800 m, with sharp dissected ridges, and steep slopes of 30° or more. Stream channels range from high-gradient bedrock channels to alluvial gravel to boulder-bed rivers.

In the Western Cascades, outcrops are commonly obscured by dense native coniferous forests of Douglas fir, western hemlock, and western red cedar. Trees in this region can grow to great height (>80 m) and age (>500 yr old) and are subject to episodic wildfires combined with, more recently, intensive logging on both public and private lands. At one point in the 1980s, timber harvested from the Willamette National Forest, which includes much of the Western Cascades, produced more than 20% of the nation's softwood timber. The legacy of this harvest remains in a distinct pattern of regenerating clearcuts of various sizes and shapes. Precipitation of up to 2500 mm/yr typically falls from November through April as both rain (below 400 m) and snow (above 1200 m), with the intervening elevations, which make up much of the landscape, constituting a "transitional snow zone."

Extending along the east margin of the Western Cascades is the modern volcanic arc of the High Cascades, a north-trending belt 30–50 km wide of upper Miocene to Holocene volcanic rocks. In central Oregon, the High Cascades form a broad ridge composed of a 2- to 3-km-thick sequence of lava flows that fill a graben formed in the older rocks (Sherrod and Smith, 1990; Sherrod et al., 2004). High Quaternary stratovolcanoes are constructed on top of the flows; they have rhyolitic to basaltic compositions and are composed of interlayered thin lava flows and pyroclastic deposits overlying cinder cones (Taylor, 1981). The location of the High Cascades at the western margin of the Basin and Range places it in a zone of crustal extension, which influences both its structural features and volcanic history. Most striking is the density of Quaternary volcanoes in the Oregon Cascade Range, with 1054 vents in 9500 km^2 (Hildreth, 2007). In the Sisters reach, which we will traverse, there are at least 466 Quaternary volcanoes, many with a pronounced N-S alignment of vents and most of which are mafic (basalt or basaltic andesite; Hildreth, 2007). Sherrod and Smith (1990) estimate an average mafic magma production rate in the central Oregon Cascades of 3–6 km^3 $m.y.^{-1}$ per linear km of arc during the Quaternary. Mafic activity has continued into post-glacial times, with 290 km^3 of magma erupted from the Cascade Range over the past 15 ka. Hildreth (2007) estimates that 21% of the erupted material forms mafic cones and shields, and that most of these edifices are within the Oregon Cascade Range.

The crest of the Oregon Cascade Range has an average altitude of 1500–2000 m, with several of the high volcanoes exceeding 3000 m. The conical morphology of the stratovolcanoes is best preserved on the younger edifices—Middle Sister, South Sister, and Mount Bachelor—as the older cones have been deeply eroded by Pleistocene glaciation. The High Cascades has also been extensively and repeatedly glaciated by thick montane ice sheets but is relatively undissected by streams (drainage density is ~1–2 km/km^2; Grant, 1997) and generally preserves many primary volcanic features. Most winter precipitation falls as snow in this zone, with occasional summer thunderstorms contributing to the water budget. Forests east of the crest are a mix of alpine and subalpine firs that transition abruptly into a more open forest of ponderosa and lodgepole pine in response to the abrupt rainfall gradient just east of the crest. Much of the land is in public ownership and managed by the Forest Service and Bureau of Land Management for timber, grazing, and recreation. Of the High Cascades subprovince in central Oregon, 25% is in wilderness areas managed by the U.S. Forest Service. On the east side, the Pleistocene glacial record is better preserved and mapped than the west side, due to the lower rainfall (~300 mm/yr), more subdued topography, and limited opportunity for fluvial erosion (Scott, 1977; Scott and Gardner, 1992; Sherrod et al., 2004).

On the east the Oregon Cascade Range is bounded by the upper Deschutes Basin, a volcanic landscape dominated by a thick (>700 m) sequence of lava flows, pyroclastic and volcaniclastic deposits of Cascade Range origin, and fluvial gravels deposited between ~7 and 4 million years ago in a broad depositional basin (Smith, 1986a). These deposits extend east to uplands consisting of early Tertiary volcanic rocks of the John Day and Clarno Formations. Interspersed throughout are local eruptive centers of a wide variety of sizes, compositions, and eruptive styles. The most prominent eruptive center off the Cascade Range axis is Newberry Volcano, which forms the southeastern boundary of the Deschutes Basin. Lava flows from Newberry blanket a large portion of the central Deschutes Basin, and partially fill canyons of the ancestral Deschutes and Crooked Rivers. The volcanic eruptions that generated the Deschutes Formation culminated with the formation of a downfaulted depression along the axis of the Cascade Range in central Oregon (Allen, 1966; Smith et al., 1987). The Pleistocene deposits in this part of the range are largely restricted to this axial graben.

The basic feature controlling regional groundwater flow on the east side is the pronounced permeability contrast between the early Tertiary units and the Pliocene and younger deposits (Lite and Gannett, 2002). Deposits of the Mio-Pliocene Deschutes Formation are highly permeable and, along with the younger volcanic deposits, host a continuous regional groundwater system that extends from the Cascade Range to the depositional contact with the early Tertiary deposits in the eastern part of the basin. The High Cascades subprovince is the principal source of recharge to this regional aquifer system.

DAY 1. PORTLAND, OREGON TO H.J. ANDREWS EXPERIMENTAL FOREST (BLUE RIVER, OREGON)

Starting in Portland, the trip takes us south through the Willamette Valley to Eugene, before turning eastward into the Cascade Range (Fig. 2). We stop to observe and discuss geologic controls on the hydrologic regimes of the westward flowing tributaries of the Willamette River (Stop 1). We then turn east at Eugene, and drive into the Western Cascades along the McKenzie River, stopping to consider the hydrogeology of the McKenzie (Stop 2). Our lunch stop is at Limberlost Campground (Stop 3), a nice example of a High Cascades stream. We then visit "Lost Spring" (Stop 4), where we discuss the relation between groundwater and lava flows and then continue up the White Branch glaciated valley to the crest of the Cascades at McKenzie Pass (Stop 5), to discuss the volcanic history and dynamics and emplacement features of young lava flows. After the stop at McKenzie Pass, we return to our evening's quarters at the H.J. Andrews Experimental Forest, near Blue River.

In Transit: I-5 Corridor and the Willamette Valley

Heading south from Portland, we traverse the Willamette Valley, part of a broad structural low that extends from the Puget Sound lowland to just south of Eugene and has existed for at least 15 million years (O'Connor et al., 2001b). The valley is a broad alluvial plain, 30–50 km wide, and is flanked by the sedimentary rocks of the Coast Range to the west and the volcanic rocks of the Cascade Range to the east. The valley slopes gently to the north, with elevations ranging from 120 m at Eugene to 20 m at Portland. The hills that the road traverses south of Portland and near Salem represent incursions and faulting of lava flows from the east; the Salem Hills in particular are underlain by Miocene Columbia River Basalt flows that were erupted in eastern Washington and followed the path of the paleo–Columbia River toward the ocean prior to the construction and uplift of the High Cascades.

The Quaternary history of the Willamette Valley has been the subject of geological investigation for over 100 yr. It reflects a dramatic interplay between erosion and deposition from the Willamette River and its tributaries, and backwater flooding and lacustrine deposition from the catastrophic flooding of the Columbia River. As summarized in O'Connor et al. (2001b), the valley is floored and filled with Quaternary gravels brought down by tributaries draining the Cascade Range and, to a lesser extent, the Coast Range. This thick (100+ m) alluvium is capped by a thinner (10+ m) but areally extensive sequence of sand, silt, and clay of late Pleistocene age that was deposited during backflooding of the Willamette Valley by the immense Missoula floods coming down the Columbia River. Multiple episodes of outbreak flooding occurred as glacial Lake Missoula repeatedly filled and failed catastrophically, resulting in rhythmically bedded, fine-grained units draped over the older alluvium, into which the modern Willamette River is now incised. The flat valley floor is thus a constructional surface ~15,000 yr old. The Missoula flood deposits thin to the south and can be traced as far south as just north of Eugene. The total volume of these deposits is ~50 km^3, and the volume of water inundating the Willamette Valley during the largest floods was ~250 km^3, equivalent to 10 yr of the annual flow of the Willamette River at Salem (Jim O'Connor, 2009, oral commun.).

Directions from Portland to Stop 1

Follow Interstate 5 south to Salem. After passing through the Salem Hills we will enter the wide southern portion of the Willamette Valley. The Santiam rest stop is near MP 241; pull in here for a restroom stop and introduction to the Willamette Valley.

Stop 1. Santiam Rest Stop—Willamette Valley Overview

The Santiam River is one of the major west-flowing tributaries of the Willamette River (Fig. 3). Drainage area at the U.S. Geological Survey (USGS) gage (Santiam River at Jefferson) 5.7 km upstream is 4580 km^2; the confluence of the Santiam and the Willamette Rivers is ~8 km downstream from our location.

The pattern of streamflow in the Santiam and the other rivers draining the western margin of the Cascades is one of the story lines of this field trip: the geological control of hydrologic regimes in volcanic landscapes. This characteristic of Cascade streams was generally described by Russell (1905), Meinzer (1927), and Stearns (1929, 1931) but has been the subject of extensive work since then (Ingebritsen et al., 1992, 1994; Grant, 1997; Gannett et al., 2001, 2003; Tague and Grant, 2004; Jefferson et al., 2006, 2007, 2008).

Understanding the role of geology in hydrologic regimes first requires an appreciation of how annual variation in precipitation controls runoff. The first rains of the hydrologic year typically begin in mid- to late October, following a prolonged summer drought of three to four months. Early fall storms must therefore satisfy a pronounced soil deficit before any significant runoff occurs. Once this deficit has been satisfied and soils are hydrologically "wetted up"—a condition that normally occurs by early to mid-November—streamflow becomes more synchronized with precipitation, rising and falling in response to passage of frontal storms from the Pacific. At higher elevations (above ~1200 m), however, precipitation typically falls as snow, building the winter snowpack, so the upper elevations of westward-draining rivers do not normally contribute much to streamflow until the spring melt.

A pattern of repeated rising and falling streamflow during the winter is clearly visible in the hydrograph for the Little North Santiam, one of the tributaries of the Santiam that exclusively drains the Western Cascades landscape (Fig. 4). Some of these rises may be augmented by melting snow during rain; these "rain-on-snow" events are generally responsible for the largest floods in the Willamette Valley, such as occurred in December

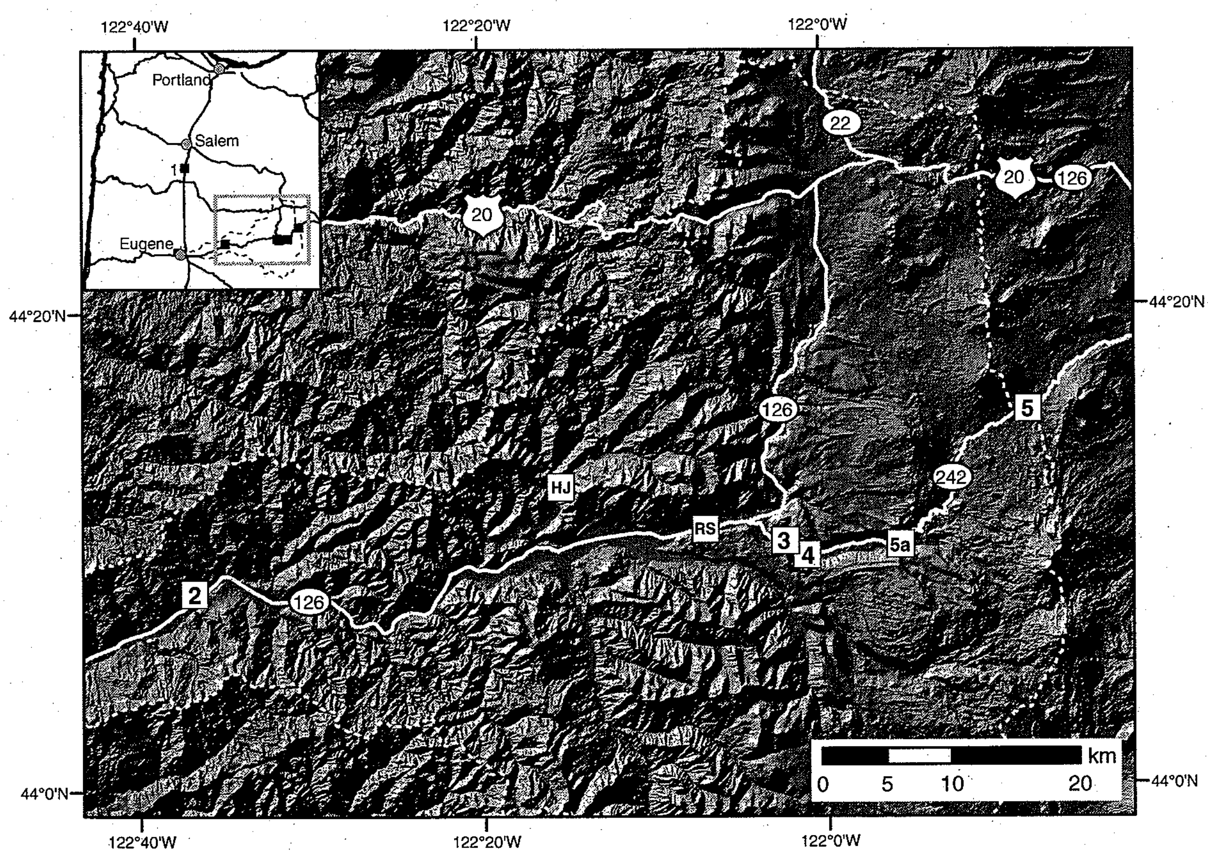

Figure 2. Overview map for Day 1, Stops 2 through 5 (squares). H.J. Andrews Experimental Forest (HJ) and McKenzie River Ranger Station (RS) are also indicated. Shaded relief map created from a 10 m digital elevation model (DEM) shows the striking contrast in morphology of the older, rugged Western Cascades and young High Cascades. The McKenzie River watershed is outlined with a dashed line.

1861, December 1964, and February 1996. As precipitation diminishes in the spring, there is a minor snowmelt rise in late April or May and then a distinct recession into very low flows that typically persist until the first fall rains begin the cycle anew. The steep, deeply dissected landscape and permeable soils of the Western Cascades respond quickly to precipitation recharge and have little storage (Tague and Grant, 2004). For this reason, the overarching pattern of flashy winter responses and very low summer flows is the characteristic signature of rivers draining Western Cascades landscapes.

Contrast this pattern with the hydrograph of the upper McKenzie River, which is located ~50 km southeast, and drains the High Cascades landscape (Fig. 4). This basin receives similar amounts of precipitation to the Little North Santiam, although more winter moisture falls as snow at its higher elevation. This snow-dominated system overlies the permeable volcanic rocks and relatively modest relief of the High Cascades subprovince, resulting in subsurface flow that buffers the hydrograph response. Major increases in streamflow are effectively limited to only the largest winter storms. A significant snowmelt peak occurs in the spring, the summer flow recession is much less pronounced, and high base flows are sustained throughout the summer, supported by discharge from volcanic aquifers (Jefferson et al., 2006).

Larger rivers, such as the Santiam and the Willamette, are hybrids of these two distinct flow regimes and demonstrate streamflow characteristics of the end members described above directly in proportion to the percentage of basin area that is in the High or Western Cascades (Tague and Grant, 2004). These trends are best illustrated by low flow regimes, including absolute flow volume (Fig. 5) and trajectories of longitudinal change in discharge with distance downstream (Fig. 6), which are both directly correlated with percent High Cascades rocks in the basin area. In essence, summer low flows come from young volcanic rocks in the High Cascades and winter peak flows are due

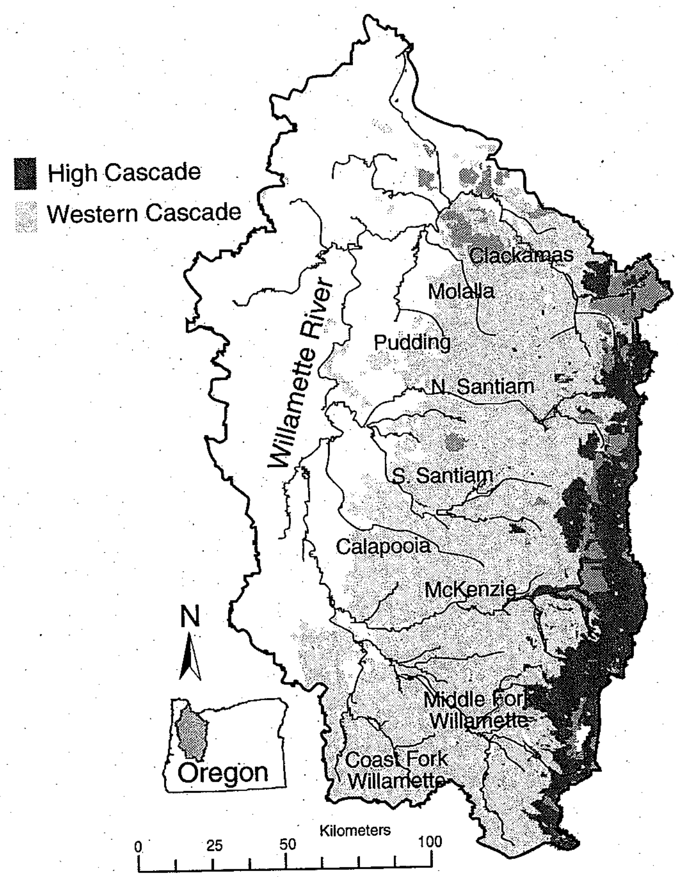

Figure 3. Willamette River Basin, Oregon, showing approximate location of High and Western Cascades subprovince boundary and major west-flowing tributaries. Geology from Walker and MacLeod (1991); figure from Tague and Grant (2004).

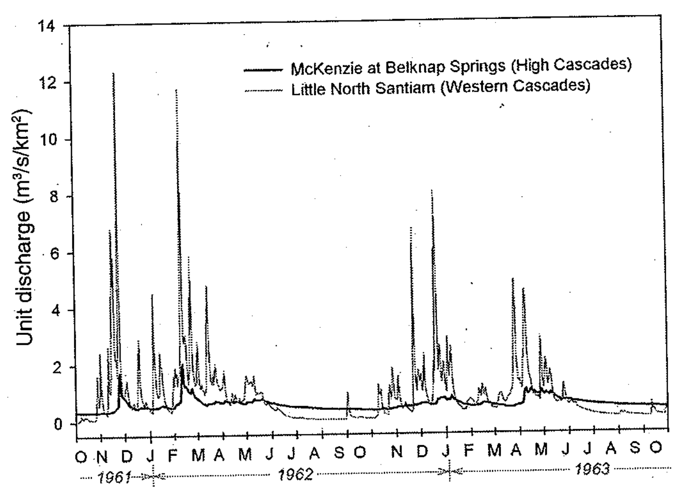

Figure 4. Daily streamflow hydrographs, normalized by drainage area, for a predominantly High Cascades (McKenzie River at Belknap Springs) and Western Cascades (Little North Santiam) river. Discharge data from U.S. Geological Survey streamflow data archive.

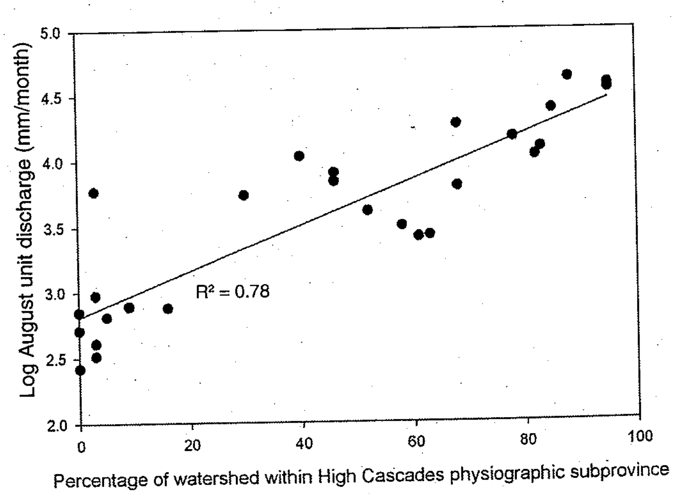

Figure 5. Mean August streamflow and percent High Cascades in contributing area for selected Willamette tributary basins (after Tague and Grant, 2004).

to runoff over older, less porous volcanic rocks in the Western Cascades. The Santiam River at Jefferson, just upstream of Stop 1, is ~15% High Cascades by area, with the North Fork Santiam (drainage area = 1990 km^2) at 28%, and the slightly larger South Fork (drainage area = 2590 km^2) at 5% (Fig. 6B). So the Santiam River is primarily a Western Cascades stream with some High Cascades influence. We'll contrast it to the McKenzie River at the next several stops.

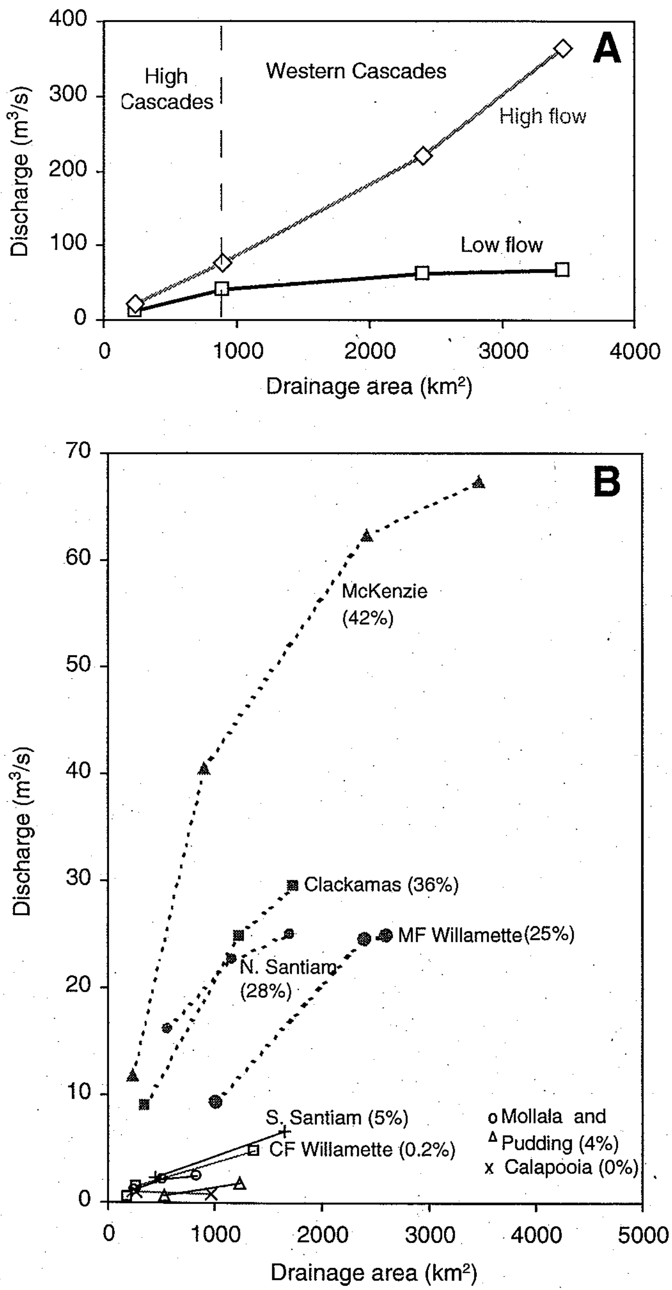

Figure 6. Discharge-drainage area relations showing impact of High Cascades contributions for (A) gages along the McKenzie River for high (1 March 1950) and low (1 September 1950) flow and (B) east-west–trending subbasins of the Willamette at low flow (1 September 1950). Percentages of High Cascades basin area shown in parentheses. Modified from Tague and Grant (2004).

Directions to Stop 2

Leave the rest stop and head south on Interstate 5. Follow I-5 south up the Willamette Valley, crossing the Santiam River immediately after leaving the rest stop. On the outskirts of Eugene, 44 mi to the south, we will cross the McKenzie River. Take Exit 194A (48 mi south of the rest stop) east onto I-105 and Oregon Highway 126 toward Springfield. Follow the highway for 6.4 mi to a stoplight at Main Street. Turn left and follow Highway 126 for 17.7 mi to Leaburg Dam. We will be traveling up the McKenzie River valley, and crossing the McKenzie River again at Hendricks Bridge State Wayside. After 13.8 mi we will cross a power canal, where water diverted from the McKenzie is transported almost 10 mi downstream to a hydroelectric generation facility. Leaburg Dam is just after mile marker 24; turn right into the parking area.

Stop 2. Leaburg Dam—Introduction to the West Side Story

The McKenzie River valley crosses the Western Cascades and extends to the High Cascades in its upper reaches (Fig. 2). At Leaburg Dam, the McKenzie River has a drainage area of 2637 km^2, of which 61% is classified as High Cascades (Tague and Grant, 2004). Contrast this with McKenzie River at Belknap Springs (40 km upstream), where the drainage area is only 374 km^2, of which 95% is in the High Cascades. The McKenzie at Leaburg Dam is a hybrid of flow regime types but more dominated by the High Cascades overall, particularly in summer. Comparison of the sources of water during low and high flow shows the relative contribution of water from the High and Western Cascades and highlights the importance of groundwater discharge during the summer (Fig. 7). As a result of the contrasting drainage mechanisms in the High and Western Cascades, the McKenzie River has a non-linear drainage-discharge relation, particularly during the summer dry season, when nearly two-thirds of the water in the McKenzie is derived from High Cascades aquifers (Fig. 6A). Over the next two days we will investigate the ways in which the geology of the High Cascades, which is dominated by Quaternary mafic lava flows, controls the hydrology of this region.

Directions to Stop 3

From the Leaburg Dam, follow Highway 126 past the Goodpasture covered bridge and through the town of Vida. Continue along Highway 126 past the towns of Finn Rock, Blue River, and McKenzie Bridge. In 28.6 mi you will see the McKenzie Bridge Ranger Station on the right. Turn into the parking lot; we'll stop briefly here to get an overview of the geography of the upper McKenzie. This is also a good place to purchase maps. Return to the vehicles and continue on Highway 126 for another 2.2 mi to the intersection with Oregon Highway 242, where we will turn right (east) off of Highway 126 to climb toward McKenzie Pass through the White Branch valley. Here outcrops of till mark the most recent termination of the glacial tongue that carved out the White Branch valley (Fig. 8A). Follow Highway 242 for 2.2 mi; turn left at the sign for Limberlost Campground (Forest Road 220).

Stop 3. Limberlost—A High Cascades Stream

This scenic Forest Service campground is bordered by Lost Creek, a channel fed by both large springs and glaciers located on the western flanks of the Three Sisters volcano complex. The geomorphic form of the channel is typical for spring-dominated streams: a rectangular channel cross section, absence of bedforms and exposed gravel bars, lack of well-developed floodplain, mature vegetation down to the water level, and stable woody debris accumulations. These attributes reflect the extremely stable flow regime and lack of floods that characterize High Cascades streams.

Directions to Stop 4

Follow Highway 242 for another 1.4 mi; there will be a small meadow on the left. Turn in on a dirt track; park under the big tree.

Field Trip Stop 1

Stop 4. Lost Spring—Groundwater System

"Lost Spring" (once called Lost Creek Spring) lies at the distal end of the White Branch valley, carved by glacial ice that accumulated and flowed from the vicinity of North and Middle Sister volcanoes. As noted by Lund (1977), the stream that enters the upper part of a valley usually has the same name as the stream that flows out the lower end. This is not the case here, where the stream name changes from the White Branch in the upper parts of the valley to Lost Creek in the lower. This change in name reflects the complex hydrology of the area and the flow of water largely through, rather than over, the young mafic lava flows that fill the valley. This valley, with its transient surface water flow, provides an elegant vignette of the hydrology of the High Cascades, where

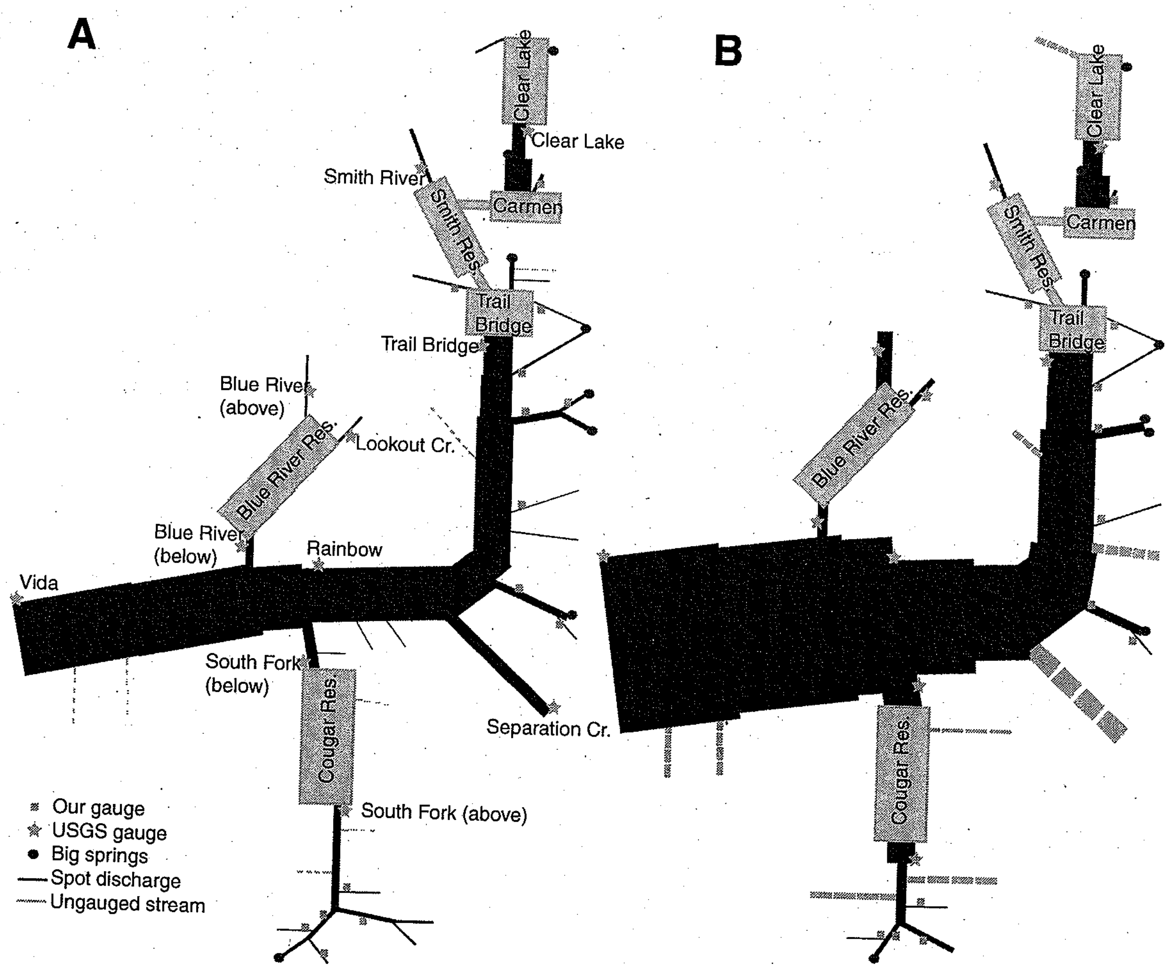

Figure 7. Sources of August and March flow in the McKenzie River. Discharges are schematically represented by thickness of each line. Distances are not to scale (after Jefferson et al., 2007).

young and highly permeable volcanic rocks form aquifers for subsurface transport of water from the high peaks to the valleys below.

Two young lava flows fill the White Branch valley. One lava flow originated at Sims Butte in the late Pleistocene (Fig. 9) and traveled nearly all the way down the White Branch valley. The younger lava flow originated at Collier Cone (ca. 1600 yr B.P.; Scott, 1990), just northwest of North Sister, and advanced west into the White Branch valley, partially covering the Sims Butte lava flow. The younger Collier Cone flow, in particular, developed a complex surface morphology because of multiple flow pulses (which created multiple levees) and interactions with the surrounding topography (Fig. 8B).

Lost Spring is actually a complex of small outlet springs emerging from the terminus of the Sims Butte lava flow (e.g., Stearns, 1929) and flows into a series of larger ponds. At a mean discharge of 6 m^3/s (212 ft^3/s), the spring complex has the largest discharge of any of the springs tributary to the McKenzie River, and also exhibits more seasonal fluctuations in discharge than other McKenzie springs. Like other springs that feed the upper McKenzie River, its estimated recharge area (114 km^2) is substantially

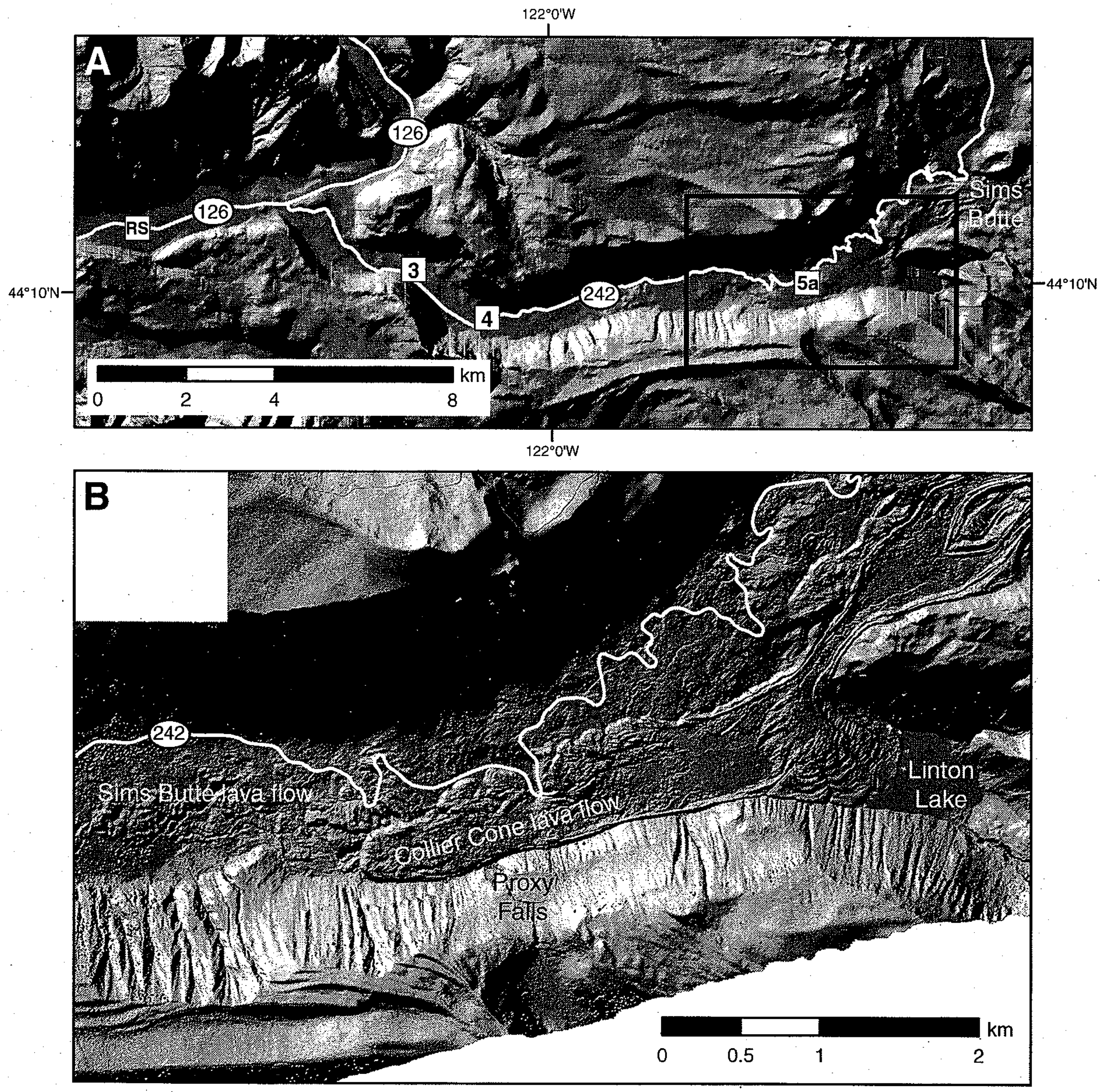

Figure 8. White Branch glacial trough. (A) Shaded relief map created from a 10 m digital elevation model of the entire White Branch valley shows stop locations and McKenzie River Ranger Station (squares), and location of (B), a light detection and ranging–generated shaded relief image showing the distal end of the Collier Cone flow (dashed line) and the lava dam that it formed to impound Linton Lake. Also shown are the locations of Proxy Falls and Highway 242.

different from its topographic watershed area (197 km²), indicating that modern topography is not the main constraint on groundwater catchments (Jefferson et al., 2006). It also has the highest temperature (6.3 °C) of the cold springs in the area, the largest component of mantle-derived volatile constituents (as estimated from $^3He/^4He$ ratios) and the longest estimated transit time (54.5 yr; Jefferson et al., 2006). These data combine to suggest that the water emerging from Lost Spring is a mixture of deep-derived (warm) water rising along a fault zone 5 km upstream (Ingebritsen et al., 1994) and shallow-derived (cold) groundwater.

To the south of Lost Spring, a channel of White Branch may also contain water. The White Branch originates at the toe of Collier Glacier, on the northwest flank of Middle Sister and just south of Collier Cone. As Collier Glacier retreated in the twentieth century, a lake formed behind the terminal moraine and then breached catastrophically in July 1942, and again in 1954–1956. Peak discharge from the 1942 breach is estimated at 140 m³/s (4944 ft³/s) and formed a debris flow that traveled 7.5 km downstream to where the preexisting White Branch channel disappears into the Collier Cone lava flow (O'Connor et al., 2001a). Under current conditions, White Branch is ephemeral from Collier Glacier downstream to its junction with discharge from Lost Spring. Surface flow is discontinuous in the stream, and in some places there is no identifiable channel (e.g., 1 km upstream from Lost Spring). In places, the channel is well defined on the Collier Lava flow (e.g., north of Linton Lake, shown on Fig. 8B), and we interpret that channel as resulting from vigorous glacial meltwater discharge during the advanced glaciers of the Little Ice Age.

About 750 m up valley from Lost Spring, White Branch emerges from the Sims Butte lava flow in a seasonally fluctuating spring. Discharge in this spring ranges from 0 to 2 m³/s (0–70 ft³/s), peaking in March and late June and drying up from late summer to mid-winter. When White Branch Spring is discharging water, the water is the same temperature and oxygen isotopic composition as water in Lost Spring. We infer that the two springs discharge from the same aquifer, but that seasonal fluctuations in the water table result in perennial flow to Lost Spring and intermittent flow to White Branch Spring (Jefferson et al., 2006). Just downstream and to the south of the springs, White Branch Creek plunges over a small 3 m waterfall into a short slot canyon, before joining Lost Creek and flowing into the McKenzie River below Belknap Hot Springs.

In-Transit: White Branch Glacial Trough to McKenzie Pass

From Lost Spring, we follow Highway 242 up the White Branch along the Collier Cone lava flow (Fig. 8B). The lava flow

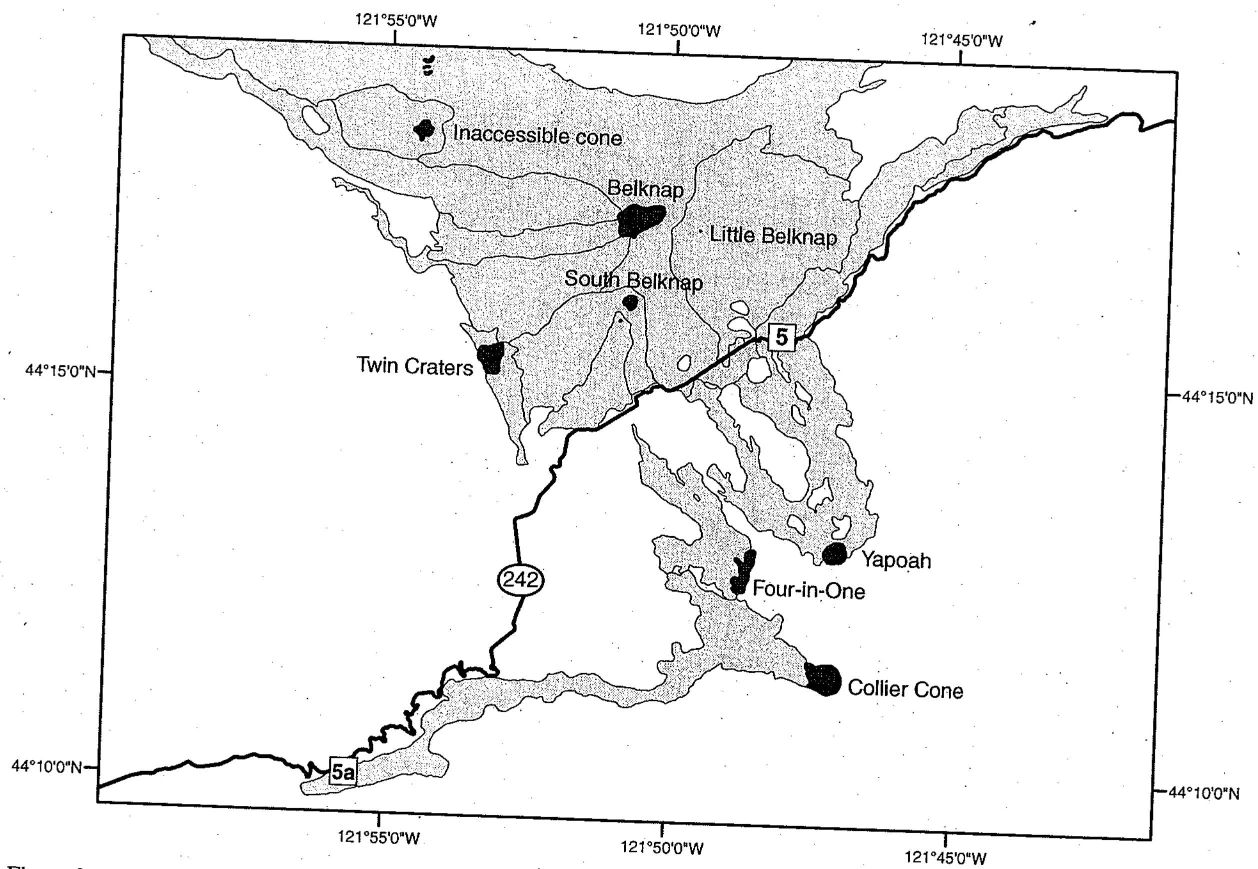

Figure 9. Post-glacial lava flows visible from McKenzie Pass. Stop numbers (squares), Young (<4000 ¹⁴C yr B.P.) lava flows (dark gray) and older flows (light gray) are indicated. After Sherrod et al. (2004).

extends more than 13 km from its source at Collier Cone down the White Branch valley to form one of the longest Holocene lava flows in the Cascade Range of central Oregon. The lava flow created two lakes as it flowed down the valley and blocked streams: Spring Lake, northeast of Sims Butte, and Linton Lake, which is fed by both Obsidian and Linton Creeks. Obsidian Creek seasonally carries snowmelt runoff but is dry in the summer; Linton Creek is perennial and fed from springs to the west of Middle Sister. Linton Lake has no surface outlet. In July and August 2004, Linton Creek had an average discharge of 1.46 m^3/s (51 ft^3/s), and over that same period, lake levels dropped ~3 m. This drop is unlikely to be solely the result of evaporation, based on a comparison with the evaporation rate of Crater Lake, located at 1883 m elevation, which is 1.2 m/yr (Redmond, 1990). Linton Lake is at 1067 m in elevation, so the evaporation rate is probably higher than at Crater Lake, but not likely to exceed 3 m/yr. Thus, all of the water discharge lost from Linton Creek in the summer, plus some water stored in the lake, is recharging groundwater. Several small sinkholes that visibly and audibly drain lake water are present along the Collier lava margin. The water that flows out of Linton Lake probably mingles with both upstream water from White Branch and downstream water introduced at Proxy Falls to emerge from the distal end of the Sims Butte flow at Lost Spring.

Directions to Stop 5

Depart the parking area at Lost Spring, turn left, and follow Highway 242 up the White Branch valley. After 3.4 mi we pass the parking lot for the Proxy Falls trailhead, which is our alternative Stop 5 in the event of inclement weather or road closures (see below). From Proxy Falls, follow Highway 242 up the headwall of the valley, around the cone of Sims Butte, past the trail to Hand Lake (dammed by a lava flow from Twin Craters) and along young lava flows from Belknap Crater. In 13.2 mi you will see Dee Wright Observatory, a small stone viewing shelter and associated restrooms, on your left. Park here.

Stop 5. Dee Wright Observatory—McKenzie Pass and the High Cascades

Dee Wright Observatory is located at an elevation of 1581 m at the summit of McKenzie Pass. This stone structure was constructed by the Civilian Conservation Corps in 1935 and named for the foreman who oversaw the project. From the top of the Observatory we will (if the day is clear) have a spectacular view of Pleistocene to Holocene volcanoes of the central Oregon Cascades. Table 1 lists features visible from this location. The most prominent peaks include North and Middle Sister to the south and Belknap Crater, Mount Washington, and Mount Jefferson to the north. Also evident are numerous older mafic composite volcanoes and cinder cones, as well as the young cinder cones to the south (Four-in-One Cone and Yapoah Crater) and Belknap Crater to the north, which are the source of the young lava flows that surround us (Fig. 9).

Dee Wright Observatory is built on a basaltic andesite lava flow from Yapoah Crater, one of the young cinder cones at the base of North Sister (the other is Collier Cone, the source of the lava flow that we traced up the White Branch valley). Although this eruption has not been dated directly, tephra deposits from Yapoah Crater immediately overlie silicic ash produced during the eruptions of Rock Mesa and the Devils Hill Chain, near South Sister (~2300–2000 ^{14}C yr B.P.; Scott, 1987). The Yapoah lava flow turns to the east at the pass and travels another 8 km toward the town of Sisters. Near the vent, lava flows from Yapoah are overlain by tephra from Four-in-One Cone, a NNE-trending chain of small cinder cones produced by an eruption ca. 2000 ^{14}C yr B.P. (Scott,

TABLE 1. VIEW OF CENTRAL OREGON VOLCANOES FROM DEE WRIGHT OBSERVATORY, MCKENZIE PASS

Azimuth (°)	Feature	Description
1	Mount Jefferson	Andesitic and dacitic stratovolcano
7	Cache Mountain	Glaciated basaltic andesite shield (0.88 Ma)
11	Bald Peter	Glaciated remnant of basaltic andesite shield (2.2 Ma)
20	Dugout Butte	Glaciated basaltic shield forested foreground
30	Green Ridge	East-bounding graben fault (5–7.5 Ma)
40	Black Butte	Basaltic andesite shield volcano (1.42 Ma)
82	Black Crater	Basaltic andesite shield volcano (ca. 50 ka?)
105–155	Matthieu Lakes Fissure	Basaltic andesite–andesite (11–75 ka)
167	Yapoah Crater	Basaltic andesite ca. 2000 yr B.P.
168	North Sister	Basaltic andesite shield volcano
171	Collier Cone	Basaltic andesite ca. 1600 yr B.P.
174	Middle Sister	Stratovolcano of variable age and composition
178	Little Brother	Basaltic shield volcano (coeval or older than North Sister)
188	Four-in-One cones	Basaltic andesite and andesite ca. 2000 yr B.P.
195	Huckleberry Butte	Glaciated basalt
197	The Husband	Eroded core of basaltic andesite shield volcano (<0.42 Ma)
218	Condon Butte	Basaltic andesite shield
235	Horsepasture Mountain	Western Cascades
256	Scott Mountain	Basaltic shield volcano
282	South Belknap cone	Basaltic andesite flank vent, 1800–1500 yr B.P.
309	Belknap Crater	Basaltic andesite and andesite shield volcano; 2635–1500 yr B.P.
321	Little Belknap	Basaltic andesite flank vent, 2900 yr B.P.
340	Mount Washington	Glaciated basaltic andesite shield volcano

Note: Descriptive data from U.S. Geological Survey (USGS) Cascade Volcano Observatory, Conrey et al. (2002), Sherrod et al. (2004), Schmidt and Grunder (2009)

1990). The tephra deposit has a maximum thickness of >2 m immediately east of the cones and thins rapidly to the east. Two small cones that lie along the same trend to the SSW of Four-in-One Cone are surrounded by basaltic andesite flows from Collier Cone (1600 ^{14}C yr B.P.). The eruption that produced Collier Cone also produced a thick (>2.5 m, 1 km east of the vent) tephra blanket, although its full extent is not known.

Collier Cone, Yapoah Crater, and Four-in-One Cone are monogenetic cones, that is, cones produced by a single eruption. A classic example of a monogenetic cone is Parícutin Volcano, Mexico (the volcano that erupted in a cornfield), which was active from 1943 to 1952 (e.g., Luhr and Simkin, 1993). Monogenetic eruptions are assumed to be fed by a single batch of magma (although see Strong and Wolff, 2003; Johnson et al., 2008) and typically persist for months to years. Strombolian explosions produce abundant bombs and scoria clasts, which accumulate close to the vent to form cinder (scoria) cones. At the same time, lava flows emerge from the base of the cone. Sufficiently high mass eruption rates may generate violent strombolian eruptions, which are characterized by high (6–8 km) ash columns that produce widespread tephra deposits (e.g., Walker, 1973; Pioli et al., 2008). The relative proportions of cone-forming scoria, lava, and tephra reflect the relative fluxes of gas and melt (and crystals). All of the young scoria cones in the McKenzie Pass region produced tephra blankets in addition to lava flows, indicating relatively high mass eruption rates.

Like Parícutin, Collier Cone, Yapoah Crater, and Four-in-One Cone erupted magma of basaltic andesite composition (Fig. 10), with genetic affinities to magma produced at North Sister (e.g., Conrey et al., 2002; Schmidt and Grunder, 2009). However, although basaltic andesite magmas from Collier Cone, Yapoah Crater, and Four-in-One Cone are similar, they are not identical in composition, suggesting that each batch of magma had a slightly different source. Additionally, all three show surprising variability in composition; Collier Cone lavas, for example, show extensive heterogeneities in both major elements and phenocryst content (Schick, 1994). Variations in SiO_2 are similar to those observed during the 1943–1952 eruption of Parícutin, Mexico, which have been interpreted to reflect assimilation of silicic upper crustal rocks in shallow dikes and sills (Wilcox, 1954). Silicic xenoliths present in tephra and lava deposits from both Collier Cone and Four-in-One Cone (Taylor, 1965; Schick, 1994) may explain much of the chemical variation observed in these two eruptive sequences (Fig. 10B).

To the west and northwest of Dee Wright Observatory are lava flows from Belknap Crater (Figs. 9 and 11) that cover 88 km^2 and are described by Williams (1976) as "one of the largest and most impressive sheets of recent lava anywhere in the United States." Unlike Collier Cone, Yapoah Crater, and Four-in-One Cone to the south, Belknap is composed of lavas that are primarily basaltic in composition and slightly enriched in K_2O (Fig. 10), continuing a trend noted by Schmidt and Grunder (2009) of a general decrease in SiO_2 in younger lavas from North Sister to McKenzie Pass (Fig. 12). Belknap also differs in being a mafic shield volcano formed by numerous eruptions from the same vent. The oldest exposed Belknap lavas lie to the northeast and have not been dated. Lava flows from South Belknap have an age of 2635 ± 50 ^{14}C yr B.P. (Licciardi et al., 1999) and form numerous interfingering small channels and flows with blocky surfaces. Overlying lava flows from Little Belknap have been dated at 2883 ^{14}C yr B.P. (Taylor, 1968, 1990), a date that seems too old for the young morphology and stratigraphic relationship with Belknap Crater (Sherrod et al., 2004). An eruption of Belknap ca. 1500 ^{14}C yr B.P. sent flows 15 km west into the McKenzie River

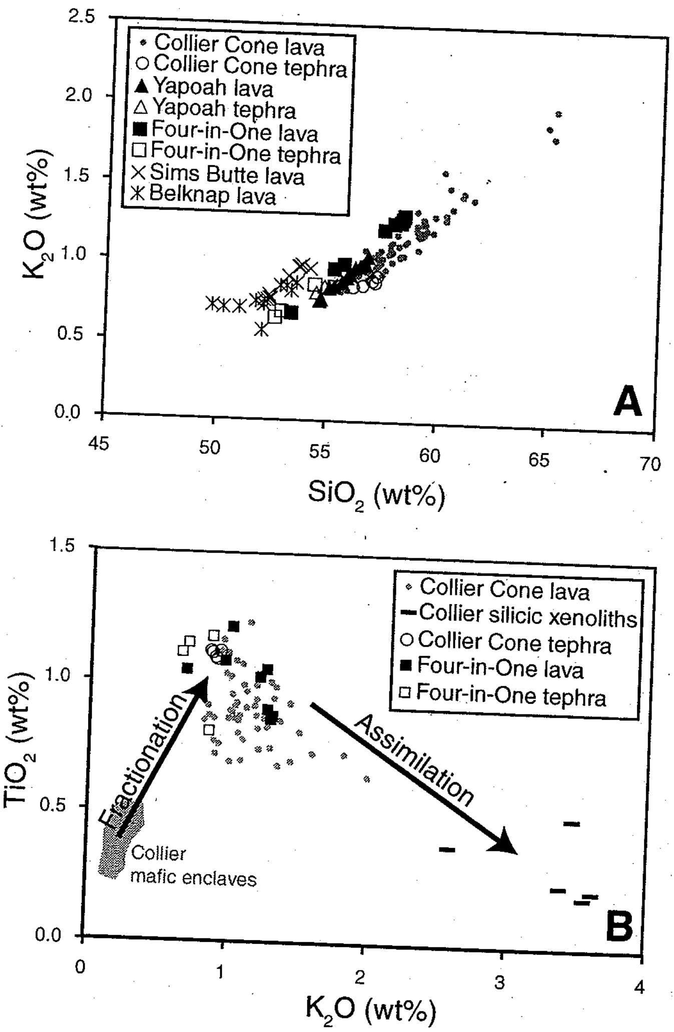

Figure 10. Composition of McKenzie Pass Holocene lavas. (A) A plot of wt% K_2O versus wt% SiO_2. Note the range in compositions of the young cinder cones north of North Sister (Collier Cone, Yapoah, Four-in-One) and the contrast with the basaltic lavas erupted from Belknap and Sims Butte. (B) A plot of wt% K_2O versus wt% TiO_2 for tephra, lava, xenoliths, and mafic enclaves from Collier Cone and Four-in-One Cone. Note that the tephra samples are less evolved than corresponding lava and that lava samples from both eruptions appear to have experienced varying degrees of assimilation of underlying silicic material. Data from Schick (1994), Conrey et al. (2002), and D. McKay (2009, personal commun.).

valley (see Day 2, Stop 8). Together these eruptions have created a shield-like edifice cored by a scoria cone. The lava flows have surface morphologies that range from blocky to pāhoehoe, with the latter flows from Little Belknap creating islands (kipukas) of older lavas within the younger (Fig. 11).

The permeable surfaces of the Holocene lava flows on the Cascade Range crest facilitate rapid infiltration and recharge to the groundwater flow system. Although average precipitation in this area exceeds 2500 mm annually (mostly as snow), there are no streams and no evidence of fluvial incision. Instead, this volume of water percolates into the permeable lava to re-emerge as spring flow, such as at Lost Spring, farther downstream. Jefferson et al. (2006) calculated the volume of mobile groundwater stored in the upper McKenzie River area as 4 km^3, based on a mean transit time of 7.2 yr and a total discharge of 17.1 m^3/s from seven major springs. Aquifer thickness may range from 30 to 120 m, derived from dividing aquifer volume by an effective porosity of 15% (following Ingebritsen et al., 1994). Thermal profiles in boreholes near Santiam Pass are isothermal for the uppermost several hundred meters, as a result of groundwater fluxes (Saar and Manga, 2004). Near the surface, hydraulic conductivity is ~10^{-3} m/s (Manga, 1996; Jefferson et al., 2006); hydraulic conductivity decreases rapidly with depth, such that at 500 m it is estimated to be ~10^{-6} m/s (Saar and Manga, 2004).

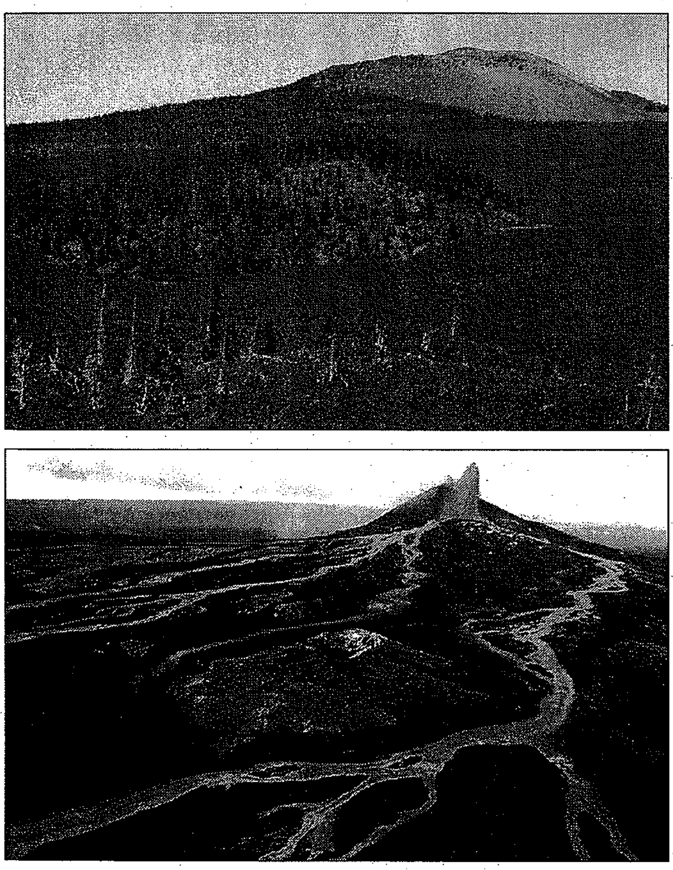

Figure 11. (A) Photograph of Belknap Crater (the cinder cone on the horizon) and pāhoehoe lava flows from Little Belknap (foreground), which surround an older cone to form a kipuka. (B) Photograph of Pu'u 'Ō'ō, Kilauea Volcano, Hawaii, showing the formation of a kipuka of older material surrounded by active lava channels (U.S. Geological Survey photo by J.D. Griggs, 1986).

Absence of fluvial incision in this young terrane raises the question of how long it takes for rivers to become established in a constructional volcanic landscape. To initiate fluvial incision, vertical infiltration through highly porous lava flows must evolve toward integrated development of surface flow paths. A fundamental question is whether surface flow develops from the top down (development of an impermeable surface layer through soil and clay formation) or bottom up (infilling of pores by translocation of fines and bulk weathering). An additional question includes the role of glaciers in jumpstarting drainage development both by incising canyons and by producing large quantities of rock flour that can reduce landscape permeability. Initial data suggest that timescales of 10^5 to 10^6 years are required for maximum soil development and measurable increase in surface drainage density.

Field trip Stop 2

Stop 5a. Proxy Falls—Traversing the Collier Cone Lava Flow

Note: this is an alternate to Stop 5, if McKenzie Pass is not open to traffic or if the weather is particularly inclement.

Starting from the Proxy Falls parking lot, 8.9 mi from the junction with Highway 126, we will hike 0.6 km to Proxy Falls. The trail crosses the ca. 1600 yr B.P. Collier Cone lava flow and the (typically dry) surface channel of the White Branch watercourse that has been carved into the flow (Fig. 8B). The light detection and ranging (LiDAR) image shows that here, at its distal end, the Collier Cone lava flow is confined on its south side by the steep walls of the glacial valley. In contrast, the north side has lobate structures that formed as the flow spread laterally.

The hydrology of the Proxy Falls area is described in detail by Lund (1977). He notes that Proxy Falls consists of several waterfalls with different sources. Upper Proxy Falls originates in

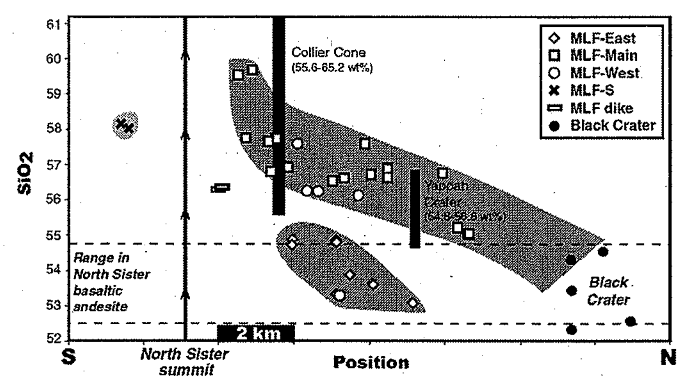

Figure 12. Illustration of the compositional variation of younger mafic lavas south of McKenzie Pass as a function of the distance from North Sister (from Schmidt and Grunder, 2009). This variation is interpreted to reflect the influence of a well-developed magma storage region beneath North Sister. MLF—Matthieu Lakes Fissure.

springs emerging from an older lava flow ~200 m above the valley floor. These springs feed streams that flow down two ravines and join at Upper Proxy Falls. Lower Proxy Falls is fed by Proxy Creek, which traverses a glacially carved hanging valley. Below the falls, Proxy Creek continues along the southern edge of the Collier Cone until the end, when it flows on top of the Sims Butte lava and loses much of its water to the underlying lava flow. All of the waters from this system, Linton Lake, and the White Branch ultimately emerge in the Lost Spring to form Lost Creek (Stop 4).

Directions to the H.J. Andrews Experimental Forest (End of Day 1)

Retrace the route west from Dee Wright Observatory, following Highway 242 for 22.1 mi to its intersection with Highway 126. Turn left (west) on Highway 126 and follow it 15.3 mi, passing though the town of McKenzie Bridge. Turn right (north) at the "Blue River Reservoir" sign onto Forest Road 15. Drive up the hill and away from the McKenzie River, following the edge of Blue River Reservoir. After 3.8 mi you will cross the Lookout Creek Bridge. Just past the bridge, turn right onto Forest Road 130 at the Andrews Forest Headquarters sign. Park in the parking lot.

DAY 2. H.J. ANDREWS EXPERIMENTAL FOREST (BLUE RIVER, OREGON) TO BEND, OREGON

Day 2 of the field trip will focus primarily on changes to patterns of water flow through the upper McKenzie River caused by the incursion of lava flows (Fig. 13). However, first we start the day by contrasting the morphology of a typical Western Cascades stream (Lookout Creek at the Andrews, Stop 6) with the morphology of rivers dominated by High Cascades streamflow regimes at Olallie Creek (Stop 7). We then travel up the McKenzie to Carmen Reservoir (Stops 8 and 9), where we examine the seasonally occupied channel of the McKenzie where the river traverses a lava flow from Belknap Crater. We then continue up the McKenzie to its headwaters in Clear Lake, a lava-dammed, spring-fed lake that we will explore using rowboats (Stop 10). After leaving Clear Lake we travel along the west side of the young basaltic Sand Mountain volcanic chain before crossing the chain to head east over Santiam Pass (Stop 11). Here we see tuyas formed during the last glaciation, have a good view of the glacially sculpted cone of Mount Washington, and look down on Blue Lake crater, the product of another recent eruption. From the Blue Lake–Suttle Lake valley, the trip descends the flank of the High Cascades, crossing glacial outwash deposits and Quaternary lava to the base of Black

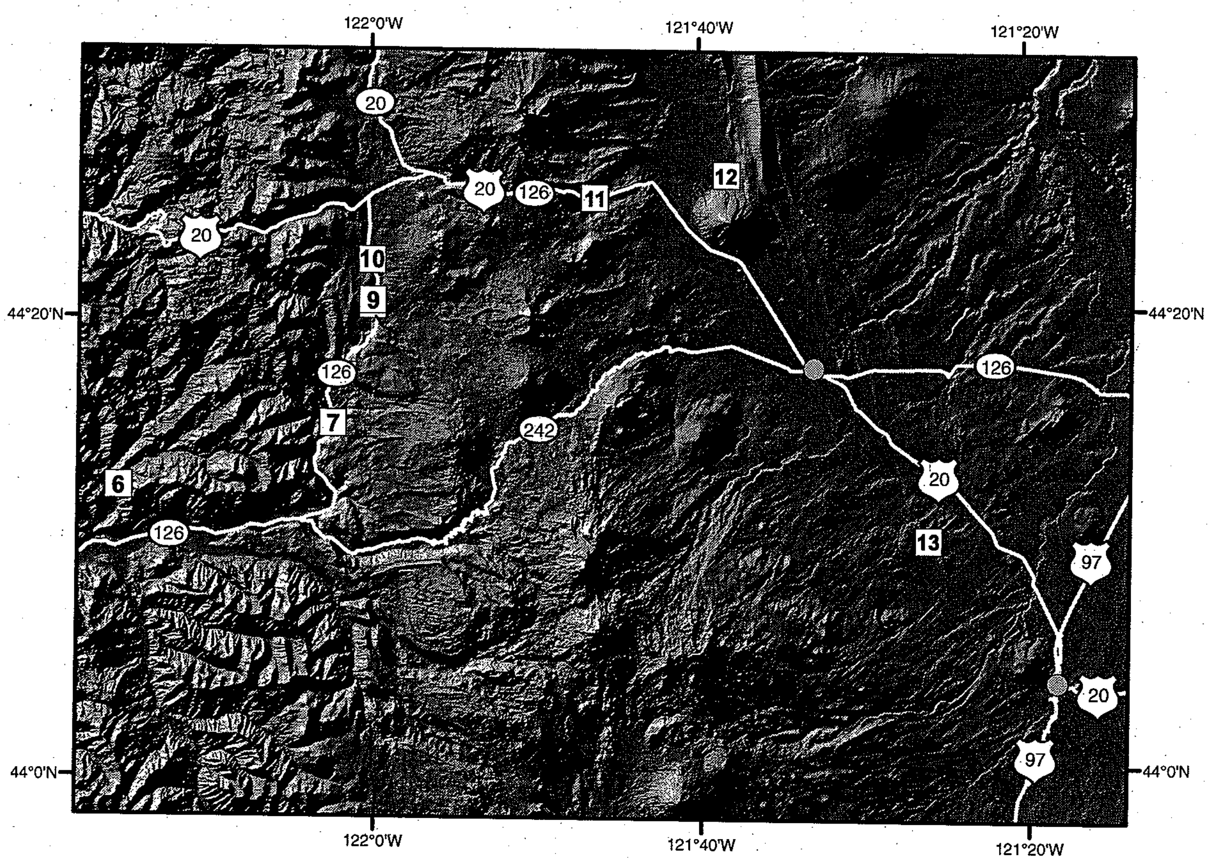

Figure 13. Overview map of Day 2, Stops 6 through 13 (squares). Shaded relief map created from a 10 m digital elevation model shows notable features including the Three Sisters volcanoes in the southern central part of the image, the pronounced Green Ridge fault, and symmetrical Black Butte cone to the north, and the broad lava plains to the east.

Butte, a large symmetrical cone prominent on the landscape. At Stop 12, on the north base of Black Butte, is the headwaters spring of the Metolius River, a major spring-fed stream tributary to the Deschutes River. Also prominent on the landscape at this location is Green Ridge, the escarpment of the eastern bounding fault of the graben in which this part of the High Cascades resides. From Black Butte the trip continues southeast along the Sisters Fault Zone, a 5–15 km zone of nearly 50 mapped faults (Sherrod et al., 2002, 2004) that extends from Green Ridge to the north side of Newberry Volcano. The final stop of Day 2 (Stop 13) examines Quaternary and Tertiary deposits on either side of the Tumalo fault, a principal strand of the fault zone, at the location of a dam and reservoir that never held water.

Field Trip - HJ Andrews

Stop 6. Lookout Creek—A Typical Western Cascades Stream

Lookout Creek near the main H.J. Andrews administration site is a classic Western Cascades stream, with a drainage area of 64 km^2. Characteristic features of this Western Cascades stream include: (1) a planform morphology dominated by coarse-grained lateral and marginal bars of flood origin, now colonized by broadleaf alders, cottonwoods, and willows; (2) a well-defined floodplain of mixed fluvial and debris-flow origin, now colonized by old-growth Douglas fir forest; (3) a well-defined channel morphology of step-pool sequences; and (4) marginal and occasionally channel-spanning, large, woody debris accumulations.

Channel and valley floor morphology, processes, and changes in this reach have been extensively studied and described (e.g., Grant et al., 1990; Grant and Swanson, 1995; Nakamura and Swanson, 1993; Swanson and Jones, 2002; Faustini, 2001; Dreher, 2004), and reveal interactions among fluvial processes, debris flows from upstream tributaries, growth and disturbance of riparian vegetation, and dynamics of large woody debris. In particular, this reach has been affected by repeated debris flows generated during major storms in 1964 and 1996. These debris flows entered the Lookout Creek channel ~2 km upstream, transitioned into bedload-laden floods that mobilized large woody debris accumulations, stripped mature and old-growth riparian forests, and deposited large coarse cobble bars that now support a young forest of alders and conifers. Stratigraphy of older deposits on which the current old-growth forest now grows reveals a similar origin. These reaches undergo a decades-long sequence of morphologic changes following large floods that is driven both by fluvial reworking of flood deposits and morphologic adjustments around large pieces of wood that fall in from the adjacent forest stand (Fig. 14).

In-Transit: Fluvial Incision and Hydrology along the Western Cascades-High Cascades Boundary

The Western Cascades-High Cascades boundary is traversed by several ridge-capping lavas resulting from topographic inversion and illustrating the persistent effects of erosion-resistant lava flows on landscape form. These lava flows can also be used to estimate incision rates across this boundary. Lookout Ridge, which separates Lookout Creek from the McKenzie River, is capped by 6–8 Ma lavas from the ancestral High Cascades (Conrey et al., 2002). These ridge-capping lavas probably originated as intracanyon lava flows and therefore mark the course of the ancestral McKenzie River. The elevation difference between the summit of Lookout Ridge (1341 m) and the modern McKenzie River at the town of McKenzie Bridge (396 m) yields an average incision rate of 0.12–0.16 mm/yr. Similar calculations for Foley Ridge (634 m), a 0.6–0.8 Ma intracanyon flow on the south side of the McKenzie River near the ranger station (454 m) indicate incision at 0.23–0.3 mm/yr. These rates bracket estimates for the Middle Santiam River of 0.14 mm/yr over the past 5 Ma (Conrey et al., 2002), and rates for the Western Cascades of 0.28–0.33 mm/yr from 3.3 to 2 Ma and 0.14–0.17 mm/yr since 2 Ma (Sherrod, 1986).

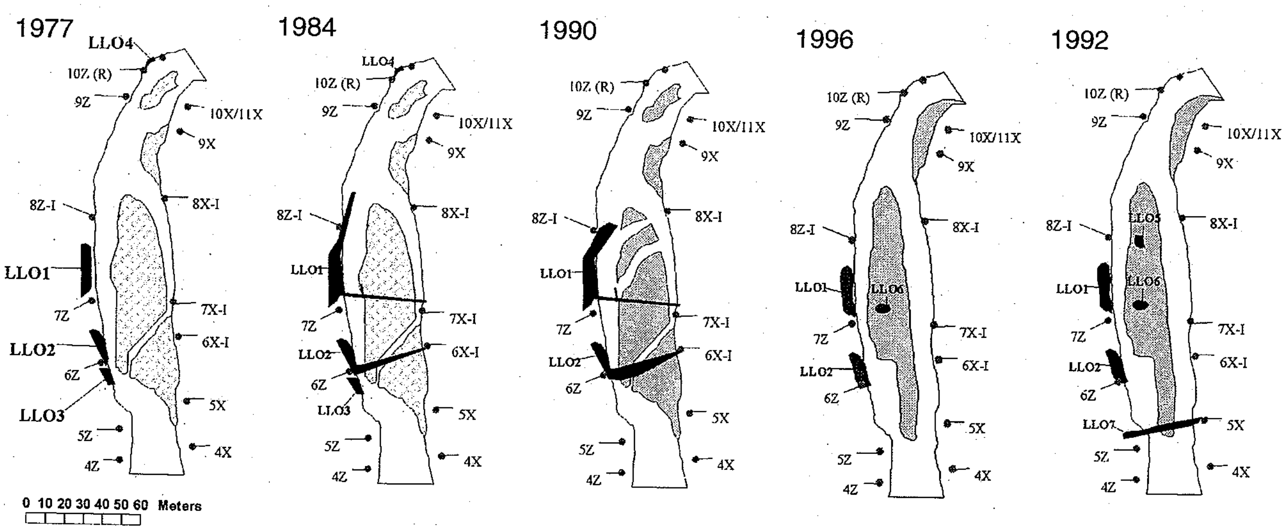

Figure 14. Changes in wood accumulation (black) and gravel bars (gray) in Lower Lookout Creek between 1977 and 2002. Numbers denote surveyed cross-section locations. Figure from Dreher (2004).

Just beyond the turnoff for Highway 242, site of yesterday's Stops 4 and 5, the McKenzie River (and Highway 126) makes an abrupt turn to the north. From here to its headwaters, the McKenzie River flows along the western graben boundary between the High Cascades and the Western Cascades. Within the graben, there has been nearly 3 km of subsidence over the past 5 million years (Conrey et al., 2002).

Along the graben-bounding fault zone, there are several hot springs discharging into the McKenzie River or its tributaries, including the privately-owned Belknap Hot Spring, and Bigelow Hot Spring on Deer Creek, which is submerged during high flows of the McKenzie River. The thermal waters are inferred to recharge near the Cascade crest, move in flow paths several km deep, and emerge along faults, which interrupt the downgradient flow of water (Ingebritsen et al., 1994). The waters discharge at temperatures of 46–79 °C, are enriched in Na, Ca, and Cl, and contain a magmatic signature in their helium isotopes. Discharge at individual hot springs ranges from 5 to 24 L/s. Total discharge of geothermal water in the Central Oregon Cascades is less than 0.2% of annual groundwater recharge but represents 148 MW of heat discharge (Ingebritsen et al., 1994). To date there has been no development for geothermal energy production, but numerous exploratory wells have been drilled.

The deep-flowing thermal water captures most of the geothermal heat and magmatic gases migrating upward through the west slope of the High Cascades (Fig. 15). As a result, water flowing from the large nonthermal springs is close to the temperature at the mean recharge elevation inferred from stable isotopes, has low total dissolved solids, and a helium isotope ratio close to that of the atmosphere.

Directions to Stop 7

Exit the Andrews Forest Headquarters and turn left onto Forest Road 15. Follow the Blue River Reservoir 3.8 mi to Highway 126. Turn left and follow the highway for ~18 mi to Olallie Campground. Park at the northern end of the campground.

Field Trip Stop 3

Stop 7. Olallie Creek—Geomorphology of Western and High Cascades Streams

This stop offers an excellent opportunity to contrast the morphology of Western and High Cascades streams. Channel-flanking unvegetated gravel bars, large boulders, well-developed step-pool sequences, and woody debris pieces and accumulations that indicate fluvial transport are all features characteristic of Western Cascades streams (Stop 6). Contrast this with the Olallie Creek channel here, a spring-fed High Cascades stream characterized by: (1) a planform morphology with few emergent gravel bars and dominated by stable wood accumulations, as indicated by moss growth and nurse logs; (2) mature conifers at the channel margin and absence of well-defined floodplain and broadleaf species; (3) a chaotic and poorly defined channel unit structure; and (4) stable, channel-spanning wood accumulations with little evidence of fluvial transport. High Cascades channels reflect the near-constant flow regimes, absence of flooding, and limited sediment and wood transport.

LiDAR imagery suggests that the path of Olallie Creek is controlled by the position of lava flow margins, and the creek is fed by springs <800 m apart at the heads of two tributary channels. The north spring discharges 1.7 m^3/s of constantly 4.5 °C water that cascades down a steep slope of lava from Scott Mountain, while the south spring emerges from under a talus pile, discharging 2.3 m^3/s at a constant 5.1 °C (Jefferson et al., 2006). Both springs discharge more water than falls as precipitation in their watersheds, showing that groundwater flowpaths cross modern topographic divides (Jefferson et al., 2006). Flowpaths are inferred to follow lava geometries that were influenced by now-obscured paleotopography. Using water temperature and hydrogen and oxygen stable isotopes, Jefferson et al. (2006) showed that the two springs feeding Olallie Creek receive groundwater from different source areas, with the south spring recharging at lower elevations than the north spring. Olallie Creek and its springs illustrate the dominant influence of constructional volcanic topography on pathways of both groundwater and surface water.

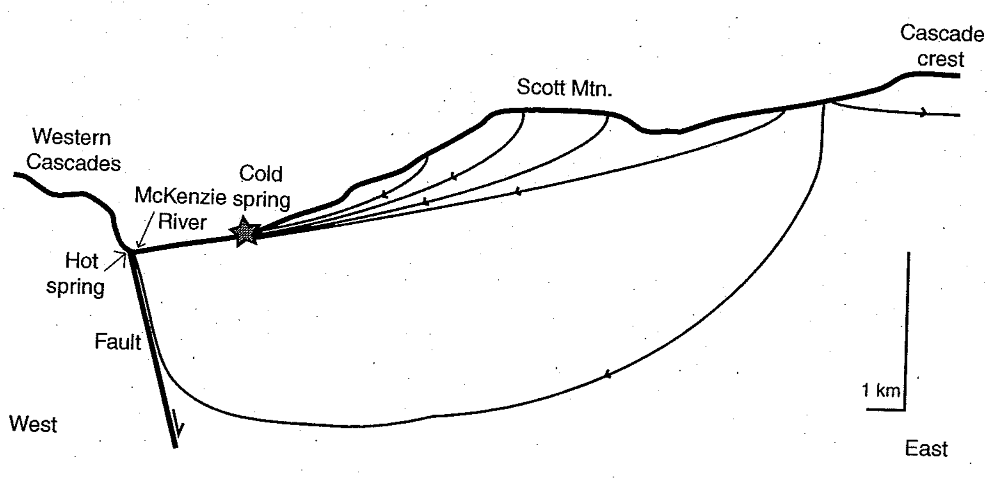

Figure 15. Cross section showing conceptual groundwater flow paths from the Cascade crest to the Western Cascades. Hot springs occur where deep groundwater flow paths are interrupted by faults (Ingebritsen et al., 1994), while cold springs are the result of shallower flow paths. From Jefferson et al. (2006).

Directions to Stop 8

Exit the campground and turn left (north) on Highway 126. Follow the highway for 7 mi to Forest Road 750, where you will turn left and follow the road around the northern end of Carmen Lake. Park at the trailhead at the end of the road.

Field Trip Stop 4

Stop 8. Carmen Reservoir—Where the McKenzie River Goes Underground

Carmen Reservoir is part of a water diversion and power generation complex operated by the Eugene Water and Electric Board (EWEB). In 1958, EWEB was issued a license by the Federal Energy Regulatory Commission to establish and operate the hydroelectric system (described below), which became operational in 1963. EWEB recently renewed the license for another 50 yr.

The McKenzie River begins at Clear Lake and flows for 2.5 km before entering Carmen Reservoir. Water is then diverted to Smith Reservoir via an underground, 3469-m-long, 3.9-m-diameter tunnel to join water from the Smith River. The combined water is then run through a second tunnel (2217 m long by 4.1 m in diameter) to Trail Bridge Reservoir. In this second tunnel, power is generated with a maximum power output of 118 MW. At Trail Bridge Reservoir, water is returned to the McKenzie River, with flow rates closely controlled to mimic natural conditions (Figs. 7 and 16).

Although there is no surface river flow directly south of Carmen Reservoir, this is not solely a function of the Carmen-Smith Diversion. Stearns (1929, p. 176) noted that at this location in September 1926 the McKenzie River "flows only in the spring ... during the rest of the year the water sinks into the permeable lava in the valley floor." Stearns noted that the McKenzie River flows underground between modern Carmen Reservoir and "Tamolitch Pool" (the "Lower Falls" in Stearns, 1929), where it re-emerges after flowing either beneath or through a young lava flow attributed to a ca. 1500 yr B.P. eruption from Belknap Crater (Taylor, 1965; Sherrod et al., 2004; Fig. 17). The diversion has decreased the amount of water traveling through this stretch by about one-half: the discharge rate at Tamolitch Pool in October 2003 was 4.1 m^3/s (Jefferson et al., 2006), which is roughly half the 9.9 m^3/s discharge estimated by Stearns (1929).

At this stop, we hike down the McKenzie River trail, beginning at the southwestern margin of Carmen Reservoir and following the trail for 0.8 km to a footbridge over the dry McKenzie River channel. Note that despite the lack of surface flow and the young age of the lava, the channel is well defined and composed of large angular boulders. Farther downstream, boulders within the channel are imbricated. Both channel creation and boulder imbrication require substantially larger flows than observed historically in this stretch of the McKenzie River. The valley itself was glacially carved, and several of the unusual morphologic features in this area, including the steep cliffs and the amphitheater immediately north of Tamolitch spring, reflect this origin.

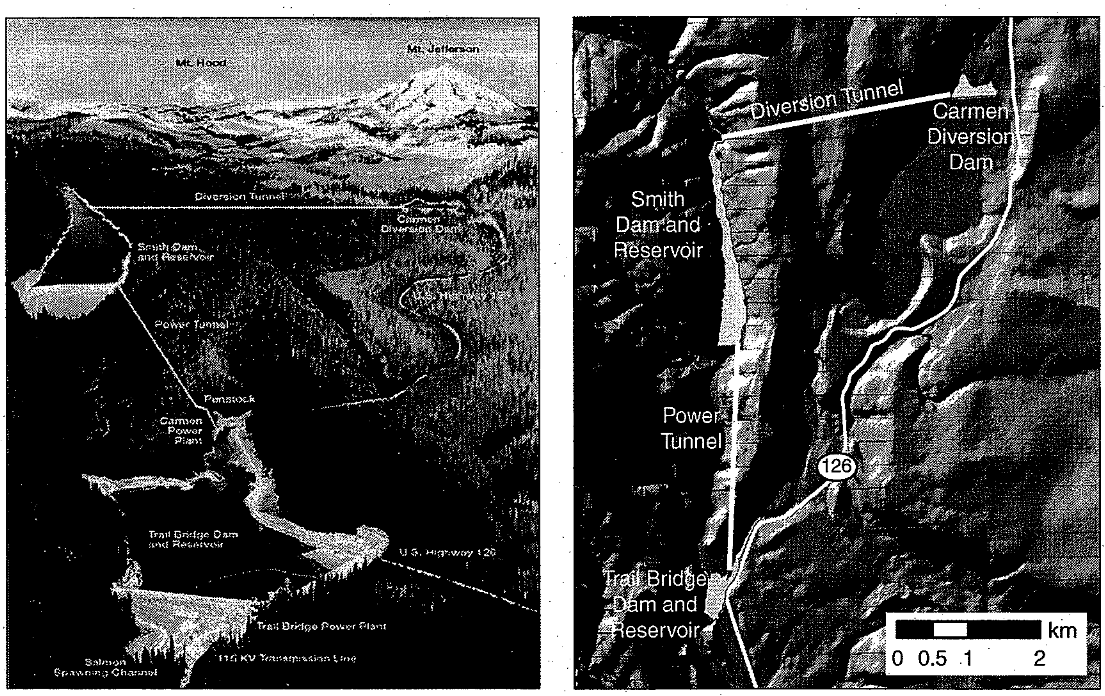

Figure 16. The Carmen Smith Project on the Upper McKenzie River in (A) illustration from Eugene Water Electric Board and (B) map view in shaded relief. Water is collected at Carmen Reservoir and transferred to the Smith River drainage and then back to the McKenzie River drainage at Trailbridge Reservoir for the purpose of power generation.

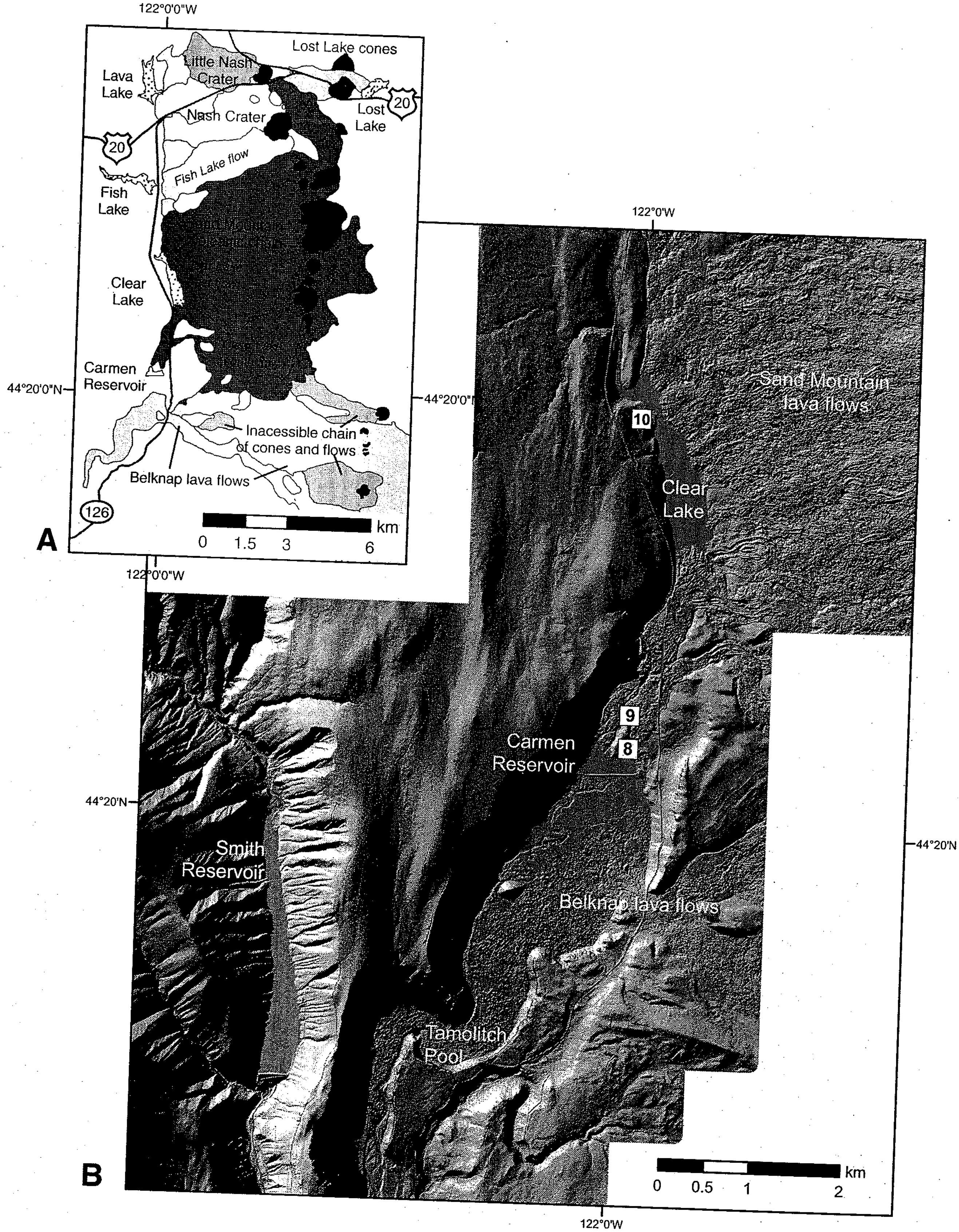

Figure 17. (A) Young lava flows (<4000 ^{14}C yr B.P.) that entered the McKenzie River Valley. Modern lakes and reservoirs are shown in light gray. Geology from Sherrod et al. (2004). (B) Light detection and ranging (LiDAR) of upper McKenzie River, with stop locations (squares) and Tamolitch pool (black dot) shown.

Directions to Stop 9

Retrace your route around the northern end of Carmen Lake and follow Forest Road 750 to Highway 126. Turn left and in less than 0.1 mi, turn left into Koosah Falls and Ice Cap Campground. Follow the road to the day use area in the campground, where we will park and walk to Koosah Falls.

Field Trip Stop 4 continued

Stop 9. Koosah Falls—Water Flow over and through Young Lava

Koosah Falls is the lower of two waterfalls along the McKenzie River between Clear Lake and Carmen Reservoir; the upper is Sahalie Falls. The river along this stretch flows from Clear Lake, along the graben structure that defines the uppermost McKenzie River valley, and over two lava flow lobes to form Sahalie and Koosah Falls (Fig. 18A). These flow lobes are part of the lava flow complex responsible for damming Clear Lake (e.g., Taylor, 1981; see Stop 10).

Koosah Falls and nearby Ice Cap Spring illustrate the multiple surface and subsurface flow paths of water through the young volcanic terrains. Although water flows over a young lava flow at this location, there is also evidence for additional flow of water through the lava flows in this part of the McKenzie River. Between Clear Lake and Carmen Reservoir, a distance of only ~3 km, ~6 m^3/s of flow is added to the river. Much of this water is discharged directly into the channel, as illustrated just below Koosah Falls, where springs discharge along flow boundaries within the valley walls. Ice Cap Spring, on the other side of the Koosah Falls parking area, demonstrates that flow paths and discharge areas can be strongly controlled by local lava topography. Ice Cap Spring, appropriately named for its 4.7 °C temperature, feeds a 0.35 m^3/s creek that parallels the McKenzie River for 400 m before entering Carmen Reservoir.

Directions to Stop 10

Leave the Koosah Falls–Ice Cap Campground parking lot and drive 1.9 mi to the entrance to the Clear Lake Resort on the right. Turn right into the resort and follow the road down the hill to the parking lot by the lake. In front of you is the restaurant providing rowboats required to access Stop 10.

Field Trip Stop 5

Stop 10. Clear Lake—The Source of the McKenzie River

Clear Lake, the highest permanent source of the McKenzie River, is bordered by lava flows on the northern, eastern, and southern margins, and bounded by a fault along the western margin of the lake (Fig. 17). The lava flows on the northern and southern margins have considerable soil development and are covered with old-growth forest, while the lava flow on the eastern margin has relatively little soil development and sparse vegetation cover. Clear Lake's name stems from its remarkable clarity, stemming from the inability of most organisms to grow in the 4.0 °C spring water feeding the lake (Jefferson et al., 2006). Roughly 10% of

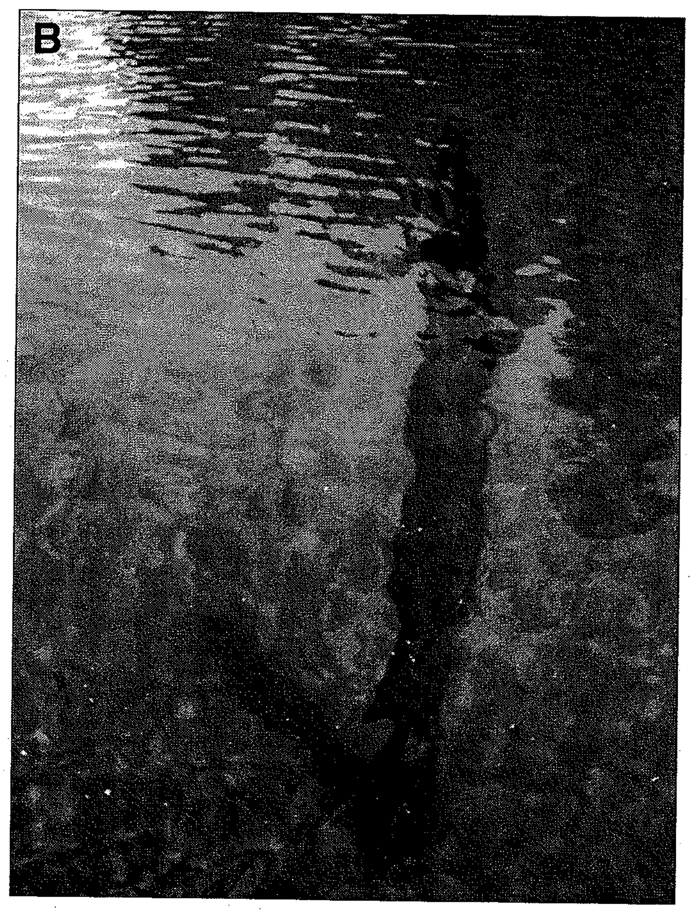

Figure 18. (A) Photograph of Koosah Falls, formed where the McKenzie River plunges over the lowermost of the lava flow lobes associated with the flow that dammed Clear Lake. Note the well-developed columnar joints within the lava flow and the obvious water seep from the base of the flow on the far canyon wall. (B) Photograph of a "ghost tree" that was drowned when Clear Lake was dammed. A similar tree has an age of 2700–3000 ^{14}C yr B.P. (Sherrod et al., 2004).

the water comes from the Great Spring situated along the northeastern margin of the lake (Fig. 19); the rest comes from small springs along the edges and bottom of the lake. There is seasonal runoff into Clear Lake from Ikenick and Fish Lake Creeks from the north and a small, unnamed creek from the west.

Clear Lake formed when lava flows entered and dammed the ancestral McKenzie River. Trees drowned by creation of the lake can still be seen in the shallower waters at the north end, preserved by the cold spring-fed water (Fig. 18B). The first scientific investigation of Clear Lake was by Stearns (1929), who evaluated the suitability of this and other areas in the McKenzie watershed as dam and reservoir sites. Stearns (1929) noted the remarkable clarity of the water, the drowned trees, and the striking difference between the lava flows on the eastern and the southern (damming) margins. He concluded that the trees were drowned when the southern flow was emplaced, and that the eastern flow was emplaced more recently. Stearns's observation that the lake is leaky, with clear evidence of water flow through the damming flow, led him to conclude that Clear Lake was an unsuitable dam or reservoir site.

Subsequently Taylor (1965) linked the Clear Lake lava flows to the Sand Mountain volcanic chain vents. He traced the flows exposed on the north shore of Clear Lake to the base of the north Sand Mountain cone and the basalt at Koosah Falls to the southern end of the Sand Mountain chain. Taylor (1965) grouped the flows that bound the eastern and southern shores of the lake as the "Clear Lake flow." However, subsequent geochemical analyses show that two flows differ substantially in composition (Conrey et al., 2002; Fig. 20). The sparsely vegetated flows on the

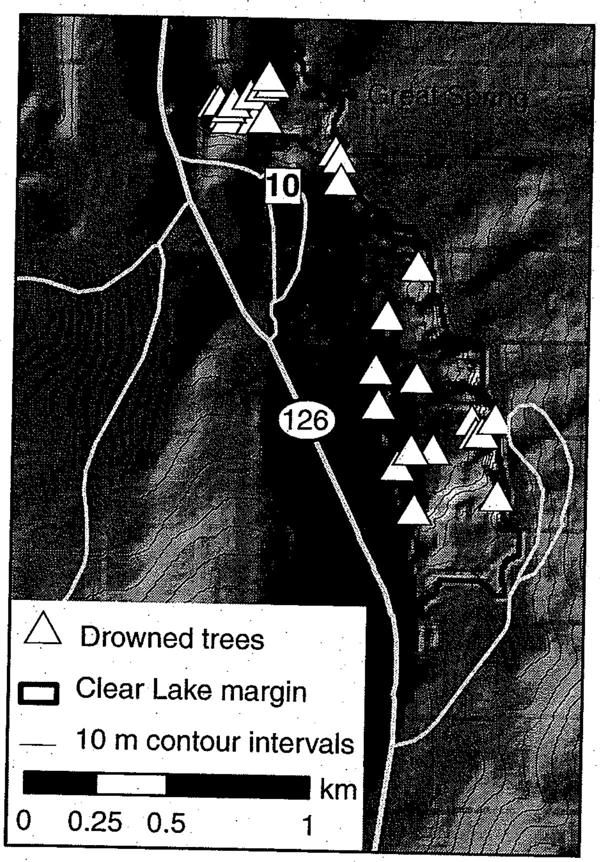

Figure 19. Topographic map of Clear Lake area including bathymetry mapped in the summer of 2008. Stop number indicates parking area; drowned tree locations (open triangles) and Great Spring (black dot) are also shown.

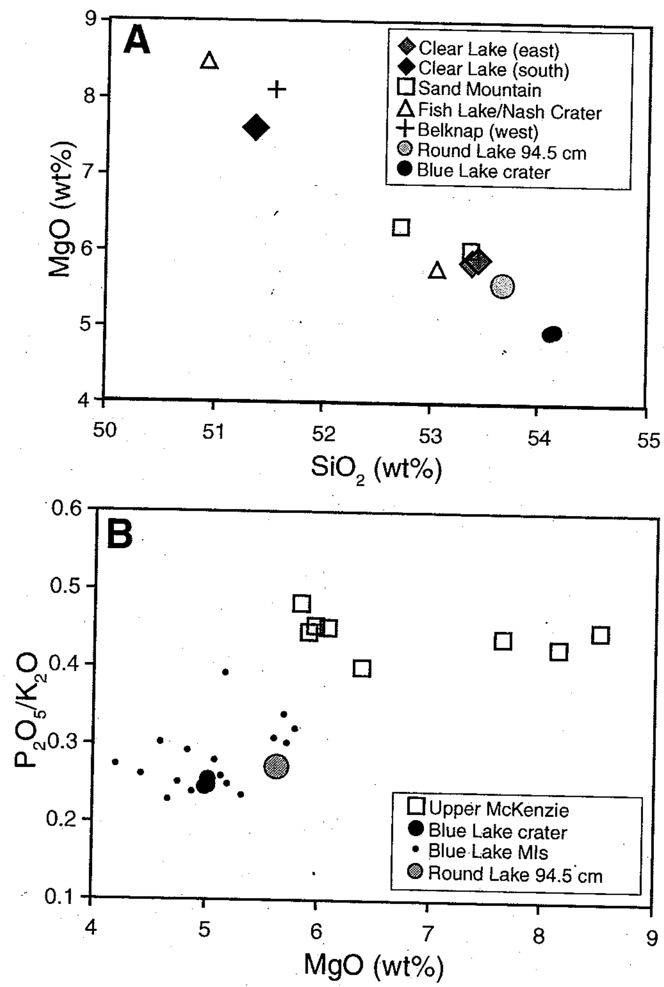

Figure 20. Geochemistry of Holocene flows from the upper McKenzie River and Santiam Pass areas. (A) A plot of wt% MgO versus wt% SiO_2 for lavas from Holocene vents in the upper McKenzie River (Sand Mountain chain, Fish Lake–Nash Crater, and the distal end of a flow from Belknap Crater). For comparison are shown lava samples from the southern and eastern side of Clear Lake, and tephra from Blue Lake crater. Note that young flows from the eastern margin of Clear Lake have compositions that are similar to those of the Sand Mountain chain, consistent with these being the sources of the lava flows (e.g., Taylor, 1990). (B) Comparison of bulk rock P_2O_5/K_2O compositions from upper McKenzie lavas (Fish Lake, Nash Crater, Sand Mountain chain, and Clear Lake lava flows), tephra and melt inclusion (MI) analysis from Blue Lake crater, and an ash layer from a Round Lake core. Because both P_2O_5 and K_2O are expected to act as incompatible elements except for late crystallization (as seen in the melt inclusion compositions), this ratio provides a useful fingerprint to distinguish the different magma types. Data from Conrey et al. (2002) and E. Johnson and D. McKay (2009, personal commun.).

east side of the lake are similar in composition to Sand Mountain (~53.4 wt% SiO_2, 5.9 wt% MgO) while the southern (damming) flow is substantially more mafic (51.3 wt% SiO_2; 7.6 wt% MgO). Placed in a larger context, the mafic composition of the damming flow is similar (although not identical) to early lavas from the northern end of the Sand Mountain chain (e.g., early Fish Lake lava), while the eastern flows are similar in composition to other lava samples from the southern Sand Mountain cone (Conrey et al., 2002).

Radiocarbon ages of ~2700–3000 yr B.P. obtained from a piece of a submerged tree at the southeast end of the lake (R. Conrey, 2009, oral commun.) correspond to dates obtained from charcoal fragments beneath the young lava flows on the lake's eastern shore, and have thus been used to date both the lake formation and part of the eruptive activity of the Sand Mountain chain (Champion, 1980; Taylor, 1965, 1990). Re-analysis of the submerged tree by anisotropy of magnetic susceptibility (AMS) yielded a radiocarbon age of 2750 ± 45 ^{14}C yr B.P. (2848 ± 69 cal. yr B.P.; Licciardi et al., 1999). Although considered by Taylor (1990) to be part of the same eruptive episode, no dates have been obtained for the damming flow.

Bathymetry measurements of Clear Lake made in September 2008 (Fig. 19) show that the lake is separated into two regions: a smaller northern region with a maximum depth of 18.5 m, and a large southern region with a maximum depth of 54 m. The two parts of the lake are separated by a bottleneck that has a maximum depth of 3 m. Drowned trees located during the same survey show that prior to inundation, the western slope was tree-covered to the valley floor (Fig. 19). The southern (damming) lava flow appears to correlate topographically with a submerged plateau at the southern end of the lake. However, there are standing trees on this platform, which indicate that this platform was not submerged prior to the last damming event. This observation, together with the more extensive vegetation cover and distinct composition of the damming flow, suggests that this flow may be substantially older than the accepted ca. 3000 yr B.P. age.

Clear Lake also provides a good opportunity to consider how the hydrology of this volcanic landscape is likely to change under conditions of climate warming (Fig. 21). Recent work suggests that Western and High Cascades streams are likely to respond quite differently (Tague and Grant, 2009). Both downscaled energy-balance models and empirical observations document that Cascade Range snow packs are diminishing and melting on the order of one to three weeks earlier, and will continue to do so in the future as the climate warms (Mote, 2003; Mote et al., 2005; Nolin and Daly, 2006; Jefferson et al., 2008; Tague et al., 2008). The consequences of reduced snow packs and earlier melting on streamflow will likely have different consequences in the Western and High Cascades. Simulations of effects of future climate on streamflow using RHESSys (Regional Hydro-Ecologic Simulation System), a distributed-parameter, spatially explicit, hydrologic model, reveal that Western Cascades streams will recede earlier in the year to minimum streamflow levels similar to those currently observed in summer months (Fig. 21B; from Tague and Grant, 2009). These low flows are remarkably similar from year to year under all climatic conditions, because they are constrained by an absence of significant groundwater storage. High Cascades streams, in contrast, will lose snowmelt peaks and also recede earlier, but because of large groundwater storage flux, will continue to recede throughout the summer (Fig. 21A). These results suggest that High Cascades streams will have higher winter flows than in the past but will also lose a larger proportion of their summer discharge than Western Cascades streams under conditions of climate warming.

In-Transit: Clear Lake to Santiam Pass

From Clear Lake we will drive north along the west side of the Sand Mountain chain of cinder cones to the extension of the chain at Nash and Little Nash cones and associated flows (Fig. 17). The volcanic history of the Sand Mountain region is described by Taylor (1965, 1990), who summarized the eruption history as including at least three distinct episodes: (1) early eruptions (ca. 3850 ^{14}C yr B.P.) from Nash Crater produced lava flows that traveled to the north and west and blocked drainages to create Fish Lake and Lava Lake; (2) eruptions that produced Sand Mountain and an extensive ash fall deposit (dated at ca. 3440 ^{14}C yr B.P.); and (3) eruptions along the southern segment that included the younger flows into Clear Lake (dated at ~2700–3000 ^{14}C yr B.P.). A lake core from Round Lake, north of Highway 22 and ~12 km east of the Sand Mountain chain, preserves ~0.75 m of tephra with geochemical similarities to Sand Mountain and is probably

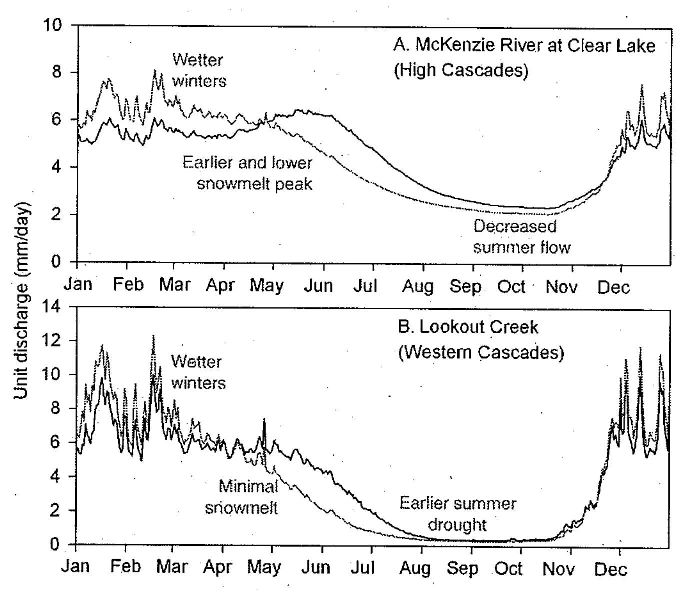

Figure 21. Modeled estimates of annual hydrograph and sensitivity to climate warming for (A) McKenzie River at Clear Lake and (B) Lookout Creek. Mean unit discharge is computed by averaging RHESSys (Regional Hydro-Ecologic Simulation System) estimates of daily streamflow (normalized by drainage area) for each day of year for 30 yr climate record. Black lines show estimated streamflow using baseline meteorologic data, and gray lines show estimated streamflow given a 1.5 °C warming. Modified from Tague and Grant (2009).