The Palouse loess and the Channeled Scabland:
A paired Ice–Age geologic system

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INTRODUCTION

The Palouse loess and Channeled Scabland of the Columbia Plateau are genetically–paired geomorphic wonders whose evolution is tied to global–scale, Quaternary climate shifts. Cataclysmic glacial–outburst floods that created the scabland supplied the bulk of loose sediment from which a *second* 'paired' system of eolian deposits–sand dunes and loess–arose. In order to appreciate these pairings, we have designed a three–day field trip that ties together the various parts of these sedimentary–geomorphic associations.

The starting point, Spokane, Washington (Fig. 1), lies on the southwestern edge of the Rathdrum Prairie, a gravel plain whose thick sedimentary cover accumulated during multiple, glacial, outburst floods. The rolling hills immediately south and east of Spokane are, by contrast, mantled with variably thick loess, much of it derived from eolian reworking of fine–grained Missoula flood sediment that bypassed the Spokane area. Over the next three days, we will trace a circuitous path that mimics the transport route of Missoula flood sediment that passed through the Spokane area, was transported to the south–central reaches of the Columbia Basin, and then carried back to Spokane by the prevailing wind.

FIELD TRIP ROUTE: THEMES AND TOPICS

The field trip route and stops highlight two themes: (1) The role that Quaternary events had on Columbia Plateau landscapes, and (2) the timing and development of the genetically–paired, dune–loess system that dominates the central and eastern plateau. Other topics include the: wind energy potential of the region, role of Pleistocene and Holocene landforms and soils in modern agriculture (including a world–class wine industry), and contemporary environmental problems associated with dust aerosols.

HISTORY OF THE CHANNELED SCABLAND

The Channeled Scabland in eastern Washington (Fig. 2) resulted primarily from cataclysmic outburst floods from glacially dammed Lake Missoula (Bretz, 1923; Bretz et al., 1956) both during and prior to the last glaciation. Outburst floods resulting from failures of ice dams on the Columbia and Spokane Rivers (Atwater, 1984, 1986, 1987) also may have helped sculpt the scabland topography. These outburst floods constituted some of the largest documented floods on earth, overwhelming the

Columbia River drainage with up to 2500 km^3 (500 mi^3) of water per flood event (Baker, 1973, 2002; Baker and Nummedal, 1978; Waitt, 1985). Mean peak flood discharges estimated at 7 x 10^6 ft^3/sec (2 x 10^7 m^3/sec) and mean flow velocities as high as 68 miles/hr (110 km/hr) (Baker, 1973, 1978) attest to the enormity of these flood events.

The Channeled Scabland is characterized by an anastomosing complex of now–largely–abandoned, steep–walled channels or coulees that were incised into bedrock and loess. This scabland terrain includes km–long cataracts with >100 m cliffs (now recognized as dry falls), loess islands (streamlined erosional remnants of thick loess separated by former flood channels), ice–rafted boulders, and immense gravel bars. In south–central Washington, the many paths taken by the onrushing floods converged on Wallula Gap (Fig. 2), a narrow water gap along the Columbia River. Temporarily impounded behind this constriction, sediment–laden waters backflooded tributaries of the lower Columbia River
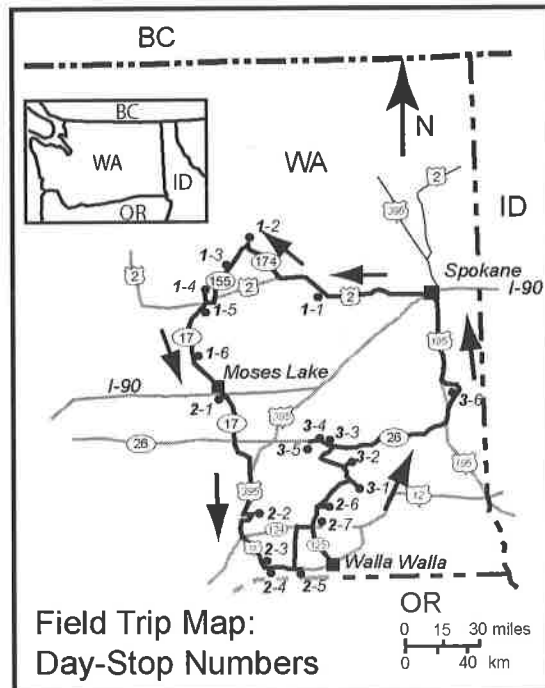

Figure 1. Field trip road guide and map. Stops are identified by day numbers (bold) and stop numbers.
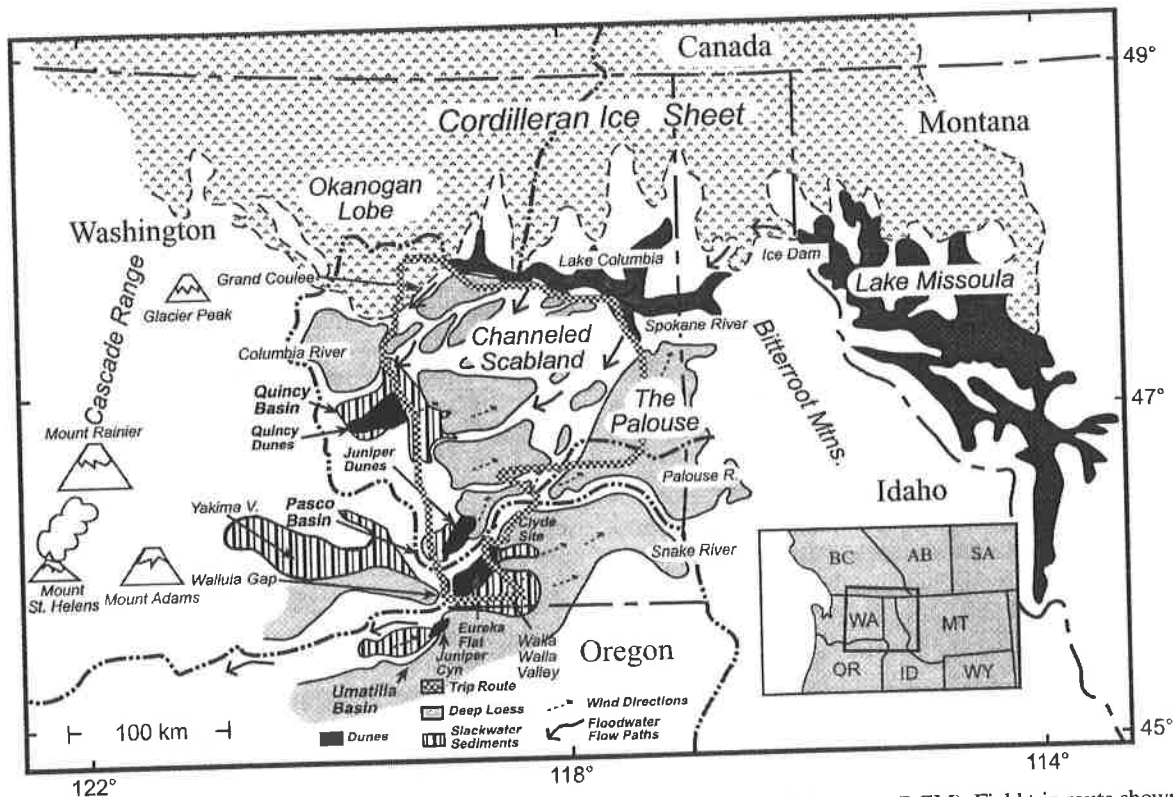
Figure 2. Schematic map of the Pacific Northwest at the height of the Last Glacial Maximum (LGM). Field trip route shown with broad, cross–hatched line.

including the Snake, Yakima, and Walla Walla Rivers. Backflooding resulted in accumulations of normally graded, fine sand, silt, and clay, slackwater rhythmites, 10's of m thick.

The interpretation of a giant–flood origin for the features of the Channeled Scabland made by J Harlen Bretz in journal articles beginning in 1923, (Bretz, 1923, 1925, 1928, 1932, 1969; Bretz *et al.*, 1956), precipitated one of the most celebrated scientific debates in American geology (Allen *et al.*, 1991). Bretz's ideas of giant floods were in direct conflict with the uniformitarian principles that had brought geology to its modern era. Even after additional evidence was found for a floodwater source (Pardee, 1942), not until 1956 (Bretz *et al.*, 1956) did his ideas become universally accepted. In recent decades, researchers have estimated the hydraulics of giant flood flows and explained the origin of numerous scabland features (Baker, 1973; Baker and Nummedal, 1978; Baker and Bunker, 1985), and identified multiple last–glacial floods (Atwater, 1984, 1986, 1987; Waitt, 1980, 1984, 1985), a number that may be as high as 89 individual events (Atwater, 1986, 1987). While the number and origins of outburst floods associated with late Wisconsin glaciation have been the focal point for continued debate (Shaw *et al.*, 1999, 2000; Atwater *et al.*, 2000), late Wisconsin outburst floods were not unique within the Pleistocene. In fact, a number of episodes of pre–late Wisconsin, glacial–outburst floods on the Columbia Plateau have been recognized (Bretz *et al.*, 1956; Patton and Baker, 1978; McDonald and Busacca, 1988; Bjornstad *et al.*, 2001).

DAY 1

The first day is largely dedicated to examining the glacial and outburst flood record of the northern Columbia Plateau.

Depart Spokane, and drive west on U.S. Highway 2 to Airway Heights. Set odometers to zero at Airway Heights and drive 39 miles (63 km) to the rest area on the south side of the road. The highway climbs in and out of a number of steep–sided, outburst flood coulees that incise the Columbia River Basalt Group (CRBG). Readers interested in learning more about the CRBG are directed to Hooper (1984; 2002). The smooth, cultivated hills and plateaus are generally loess–capped islands that escaped erosion by the floods. The Cordilleran Ice Sheet advanced within about 12 miles (19 km) (north) of U.S. 2.

STOP 1–1. CHANNELED SCABLAND

The first stop provides an excellent opportunity to view scabland topography and contemplate the broader history of the outburst flooding. This site is located within the Creston–Davenport scabland (Kiver and Stradling, 1989) where a relatively low–relief, butte–and–basin topography is developed on basalt. Note the typical scabland angular basalt gravel and scoured basalt hollows in which ponds now reside.

Because this setting is primarily erosional, the flood record here is sparse. However, ~25 miles (40 km) to the northwest, varves, out-

burst flood and glacial deposits exposed along Manila Creek, a tributary of the Sanpoil and Columbia Rivers, has preserved one of the longest and most detailed chronologies of flooding and glaciation anywhere on the Columbia Plateau. Glacial Lake Columbia, generated when the Okanogan lobe of the Cordilleran Ice Sheet blocked the Columbia River, influenced the sedimentation along Manila Creek from at least ca. 15.5 to 13 ^{14}C yrs. B.P (Atwater, 1984, 1986. 1987). This ~2,500–year interval essentially covers the late Wisconsin history of Missoula outburst flooding (Waitt, 1980, 1985; Baker *et al.*, 1991). Varve counts at Manila Creek (Atwater, 1986, 1987) indicate that floods triggered by ice–dam collapses were smaller and relatively more frequent when glacial Lakes Columbia and Missoula first began to form. Until the last glacial maximum (LGM), the volumes of these floods increased while their frequency decreased, with 50 to >65 yrs passing between flood events (Atwater, 1986). After the LGM, their frequency increased and volume decreased. The 89 individual floods that Atwater (1986) identified at Manila Creek more than double the approximately 40 outburst floods identified by Chambers (1971) in Montana and by Waitt (1980; 1985) in southern Washington. Atwater (1986, 1987) attributed the greater number of floods to the relative nearness of the Manila Creek site to local base level, a factor that enhanced the preservation of even small volume flood deposits. By contrast, sedimentary deposits from small–volume floods in the southern plateau tended not to be preserved (Baker and Bunker, 1985; Waitt, 1985; Atwater, 1986). Ice–rafted erratics at elevations of up to 750 m (Atwater, 1986), indicate that STOP 1 (695 m) was inundated frequently during outburst flooding.

Exit the rest stop and continue west on U.S. Highway 2 for 17.1 miles (27.5 km) to its junction with State Highway 174. Turn right onto Highway 174 and drive toward Grand Coulee. The road passes through channeled scabland and loess islands for approximately 30 miles (48 km) before descending into the steep, basalt–walled Columbia River valley. On the outskirts of the town of Grand Coulee, watch for the 'Business District sign'; turn right at this sign onto Spokane Way, driving 0.5 mile (0.9 km) to its junction with State Highway 155. Turn right onto Highway 155 and proceed 1.7 miles (2.7 km) to the Grand Coulee Dam Visitors Center parking area.

STOP 1–2 GRAND COULEE DAM VISITORS CENTER

Construction of Grand Coulee dam began in 1933 and was completed in 1942. It produces about 6,800 megawatts annually (Pitzer, 1994). The dam is operated by the U.S. Bureau of Reclamation, and backs up the Columbia River (Lake Franklin D. Roosevelt) nearly to the Canadian border, a distance of ~150 river miles (241 km). Rising ~550 feet (182 m) above its granite bedrock base, the ~4200–foot–(1380 m) –wide dam is a key element of the Columbia Basin Irrigation Project. A portion of the electricity produced by the dam is used to pump water from Lake Roosevelt ~95 m *up* to Banks Lake, a ~27–mile–(43.5 km) –long and 1–4.5–mi. (1.6–7.2 km)–wide reservoir in the upper Grand Coulee, ~2 mi. (3.2 km) to the southwest. Water from Banks

Lake feeds a canal system that supplies more than 250,000 acres of irrigated cropland (Pitzer, 1994).

Turn left onto State Highway 155 and drive 13 miles (21 km) south toward Grand Coulee and Banks Lake. As the road climbs out of the Columbia River Valley it traverses numerous granite and gneiss exposures that are capped in the near distance by Miocene Columbia River basalt flows.

STOP 1–3. GRAND COULEE AND STEAMBOAT ROCK

Grand Coulee is the deepest and arguably the most spectacular of the scabland coulees, ~1.6–7.2 km wide, 300 m deep, 70 km long. It is subdivided into incised upper and lower Grand Coulee segments separated by an unconfined, kms–long scabland (Salisbury, 1901; Bretz, 1932, 1969).

The history of the Grand Coulee was first described in detail by Bretz (1932). After the Okanogan lobe dammed the Columbia River in the late Wisconsin, creating Lake Columbia, Grand Coulee provided the principal drainage outlet for the lake (Bretz, 1932; Richmond *et al.*, 1965; Atwater, 1986; Waitt and Atwater, 1989). Glacial ice blocked the coulee at the present Grand Coulee dam when the Okanogan lobe was at its maximum (Bretz, 1932; Atwater, 1986, 1987; Waitt and Atwater, 1989). Glacial drift that caps Steamboat Rock probably was deposited during a pre–late Wisconsin glaciation (Bretz, 1932; Atwater, 1986, 1987; Waitt and Atwater, 1989). Lake Columbia continued to occupy Grand Coulee for several centuries following the last Missoula flood (Atwater, 1986). Varved, silt and clay lake sediments, some containing dropstones, are intercalated with thin flood deposits (Atwater, 1987). Following the last flood, several decades of ponding led to continued silt–clay deposition here (Atwater, 1987; Waitt and Atwater, 1989).

Turn right (south) at Steamboat Rock State Park and drive 15.6 miles (25 km) toward Coulee City State Highway 155. At the junction of State Highway 155 and U.S. Highway 2 follow U.S. 2 west 5.9 miles (9.5 km) to its junction with State Highway 17. Turn right onto State Highway 17. Two miles (3.2 km) north of the junction, note the hummocky, rubble–strewn ridge to the northwest (on your left). This ridge marks the southern terminus of the Withrow end moraine (Easterbrook, 1976, 1979). Stop 5.5 miles (8.8 km) from the junction of Highways 17 and U.S. 2, where a basalt 'haystack' rock rises from the prairie on the right (east) side of the road. This enormous glacial erratic was transported by the Okanogan lobe. (Optional) Drive 8.5 miles (13.7 km) north from this stop to Sims Corner (Junction of Highways 17 and 172). Driving west on Highway 172 takes one past an array of recessional glacial topography, including kames, eskers, and kettles (Easterbrook, 1979).

STOP 1–4. WITHROW END MORAINE AND THE WATERVILLE PLATEAU

This stop provides an opportunity to examine glacial drift left by the Okanogan lobe after it crossed the Columbia River

~30 miles (48 km) to the north. As the Okanogan lobe emerged from the Okanogan valley, it radiated outward across the Waterville Plateau. The terminus of this glacial advance is marked by a distinctive end moraine, several kilometers wide and 30–70 m high. The hummocky moraine extends approximately 35 miles (56 km) from near Lake Chelan on the west to upper Grand Coulee (Bretz, 1932; Flint, 1935). Termed the Withrow moraine (Easterbrook, 1976, 1979) for the settlement where it achieves its maximum height (Flint, 1935), this prominent feature records a prolonged ice front presence (Easterbrook, 1979; Waitt and Thorson, 1983).

The glacial drift of the Withrow moraine overlies glacially polished, striated, and scoured basalt (Bretz, 1932; Flint, 1935, 1936; Easterbrook, 1976, 1979). Behind the terminal moraine, the low, rolling topography bears the imprint of glacial recession, with numerous stagnation–related kettles, kames, and eskers (Bretz, 1932; Flint, 1935, 1936; Easterbrook, 1976, 1979). The timing of Waterville Plateau glaciation has been obscured by a lack of in–place, datable materials (Easterbrook, 1976, 1979). Atwater's (1986, 1987) reconstructions of outburst flood activity at Manila Creek, suggest that the Okanogan lobe had retreated north of the Columbia River by a few centuries after the final Missoula outburst flood, ca. 12.5 ka (Waitt, 1985).

Drive 5.5 miles (8.8 km) south on Highway 17 to the junction with U.S. Highway 2. Turn left (east) on Highway 2 and drive 1.7 miles to the junction with Highway 17; turn right and drive 1.9 miles (3 km) to the Dry Falls overlook.

STOP 1–5. DRY FALLS OVERLOOK

This spectacular, dry waterfall marks the northern margin of the lower Grand Coulee. From this overlook, only the western edge of a continuous 5.5–km–wide, 120–m–high cataract, created by the progressive recession of the cliff face, is visible (Bretz, 1932; Baker, 1973, 1989). Bretz (1969) attributed the Dry Falls cataract to highly focused flood flows caused by downstream incision of weaker basalt in the Coulee monocline. Deep, plunge–pool lakes, like Dry Falls Lake, attest to the erosive power of the flood flows. Falls–related, plunge–pool scouring and subaqueous plucking of the coulee bottom generated many of the cataracts in the lower Grand Coulee (Baker, 1989). Subaqueous plucking was facilitated by vortices and near–bed fluid turbulence that excavated the less–resistant, columnar–jointed portions of the basalt flows (Baker, 1987).

Turn left (south) on State Highway 17 and drive 23.2 miles (37.3 km) south past Soap Lake. Turn left on Hatchery Road; at 0.7 miles (1.2 km) note the giant basalt megaclast amidst the boulder field on the left. Continue another 1.9 miles (3 km) to the stratified and cross–stratified gravel exposure along the road leading out of the Rocky Ford Creek valley.

STOP 1–6. EPHRATA FAN

The concentrated outflow of sediment–laden water from lower Grand Coulee near Soap Lake into the unconfined Quincy Basin generated a broad, 200 mi^2 (520 km^2), fan–shaped accumulation of coarse–to–fine gravel. Reaching a maximum thickness of 56 m (Bretz, 1932), Ephrata fan deposits generally fine upward from thick, high–angle–cross–bedded, cobble–boulder gravel to horizontally stratified, coarse sand and granule–pebble gravel. The giant basalt megaclast along the road has a maximum dimension of 18 m (Baker, 1987) and is part of an armored surface generated by waning–stage, outburst flood incision (Baker, 1973). The northwest–oriented, crescent–shaped depression surrounding the megaclast records turbulent scour (Baker, 1973, 1987). Rocky Ford Creek has incised the fan surface to an average depth of 15 m revealing the coarse–grained, fines–depleted nature of the fan interior. Sand, silt, and clay transported here by outburst floods largely bypassed this region, ultimately accumulating in slackwater deposits or leaving the plateau via the Columbia River.

Drive 2.6 miles (4.2 km) to the junction with State Highway 17. Turn left (south) and drive 3.8 miles (6 km) to the junction with State Highway 282; bear left and continue south on Highway 17 for 9.4 miles (15 km) to Moses Lake. End of Day 1.

DAY 2

Much of the second day is dedicated to examining the paired, eolian, sand–dune–and–loess systems and their relation to paleoclimatic fluctuations on the Columbia Plateau. Special tours of a wind energy facility and slackwater flood stratigraphic succession have been arranged.

SAND DUNES ON THE COLUMBIA PLATEAU

Numerous stabilized–to–active dune fields, sand sheets, deflation flats, and areas of deflationary gravel lags mantle the plateau. Dune fields cover from 20 km^2 at the Sand Hills Coulee (Schatz, 1996; Gaylord et al., 1999) to >1000 km^2 in the Quincy dunes (Petrone, 1970). Relatively large dune fields exist in the southern portion of the scabland on the Hanford site (Gaylord et al., 1991; Smith, 1992; Gaylord and Stetler, 1994; Stetler and Gaylord, 1996), near Smith Canyon (Schatz, 1996) and within the Juniper dunes Wilderness area (Gaylord et al., 1997; Gaylord et al., 1998; Gaylord et al., 2001). Extensive sand sheets mantle many areas of the plateau including the important loess generating region of Eureka Flat (Sweeney et al., 2001). All of these dune fields have SW–NE orientations and are located upwind from loess. Sediments from the dunes commonly are linked compositionally to outburst–flood deposits and Palouse loess (Busacca and McDonald, 1994; Gaylord et al., 1991; Gaylord et al., 2001); but at Sand Hills Coulee, dune sand compositions reflect a Miocene–Pliocene, Ringold Formation

(paleo–Columbia River ancestry) provenance (Schatz, 1996; Gaylord *et al.*, 1999).

Tephra chronology and limited geoarcheology from the Hanford and Smith Canyon dunes suggests that sand dune activity on the plateau peaked during the early and middle Holocene (Gaylord and Stetler, 1994; Gaylord *et al.*, 2001). The prevalence of vegetation since the start of the Holocene has favored the development of parabolic and blowout dune types. The 20% of dunes that are actively migrating today also include barchan and barchanoid ridge types. These active dunes tend to form in elevated locations subject to higher wind speeds, in zones of low subsurface moisture, or within areas subject to persistent anthropogenic disturbance.

LOESS ON THE COLUMBIA PLATEAU

The loess that blankets the Columbia Plateau is one of the more prominent and widespread geologic features produced during the Quaternary in the Pacific Northwest. The core area of the deepest and most continuous loess in eastern Washington is termed the 'Palouse' (Fig. 2), a region that covers >50,000 km^2 with deposits that are from a few meters to 75 m thick. Palouse loess accumulated at elevations ranging from about 200 m along the eastern margin of the Pasco Basin to approximately 800 m near Moscow, Idaho. The loess on the plateau largely accumulated on the Miocene Columbia River Basalt. Incised perennial streams within the Palouse reveal basalt–floored valleys. First–and–second–order intermittent streams tend to bottom in loess rather than basalt (Fryxell, 1968). The drainage pattern of major streams on parts of the Palouse not affected by floods probably was established before outburst flooding began (Ringe, 1970).

Forty six natural and artificial loess exposures on the Columbia Plateau have been described, mapped, and correlated during the past 20 years (McDonald and Busacca, 1992; Busacca *et al.*, 1992). Region–wide correlations using paleosol and tephra markers have revealed two loess units that span approximately the last 70,000 yrs. These are informally named L1 and L2 (McDonald and Busacca, 1992) (Fig. 3). Two review articles (Busacca, 1989; 1991) summarize much of what had been written about the Palouse between the 1880's and 1980's. Three articles since 1991, McDonald and Busacca, 1992; Busacca *et al.*, 1992; Busacca and McDonald, 1994, add considerable detail concerning late Quaternary loess and paleosols.

The Palouse loess has attributes that have made it attractive for paleopedologic and paleoclimatic research, including: (1) a depositional record that spans much of the Quaternary , (2) its relatively great thickness per unit time, especially in areas proximal to outburst flood (source) sediments, (3) numerous distal tephra layers that provide chronostratigraphic control, and (4) paleoecologic features preserved in surface and buried soils. Research on the significance of cicada–burrow fabrics (Tate, 1998; King, 2000; O'Geen and Busacca, 2001) and on plant opal phytoliths (Blinnikov *et al.*, 2001, 2002) indicates that an *Artemisia*–dominated, periglacial steppe that supported huge populations of sagebrush–dependent,

soil–dwelling, cicada nymphs and the above–ground, adult forms was widespread on the Columbia Plateau during development of the Washtucna and other cold–phase soils. The striking visual and morphologic characteristics of the majority of buried soils in loess across the plateau largely reflect the abundance of cicada burrows that have been cemented by carbonates. The widespread periglacial, shrub–steppe was replaced by a pure, perennial, bunchgrass–steppe at the opening of the Holocene (except that sage persisted in the driest core area of the plateau). Lacking a shrub component, the bunchgrass soils that developed during the Holocene did not support cicada nymphs. Consequently, the L1 loess is all but devoid of large burrows that typify the Washtucna soil in L2 (O'Geen and Busacca, 2001). In summary, significant soil development on the Palouse seems to have occurred during cold–climate phases, while warm, interglacial phases were characterized by rapid loess deposition and relatively little soil development, except in distal areas along the loess field margin that have received higher rainfall.
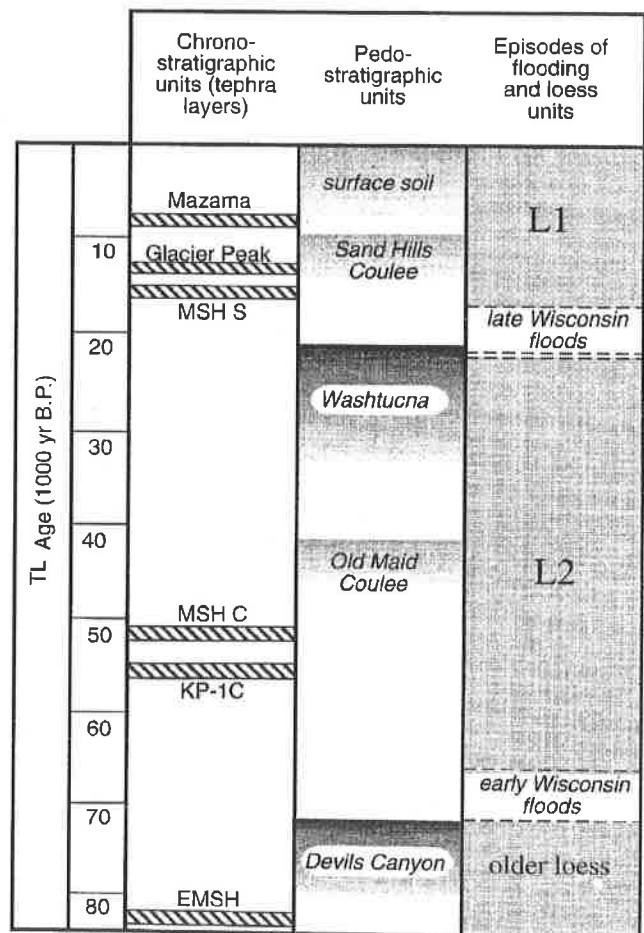

Figure 3. Composite schematic stratigraphic diagram of late Quaternary chronostratigraphic, pedostratigraphic, and loess units in Palouse loess (revised from McDonald and Busacca, 1992) and episodes of cataclysmic flooding. Correlated tephras from Busacca *et al.*, (1992); TL age scale after Berger and Busacca (1995).

EOLIAN TRANSPORT AND DEPOSITION: CONTROLS ON THE THICKNESS AND DISTRIBUTION OF LOESS ON THE PLATEAU

Field and wind tunnel evidence demonstrate that saltating sand is critical to entrainment of silt–and–clay–sized particles (Bagnold, 1941; Gillette *et al.*, 1974; Shao et al., 1993; Saxton et al., 2000). Saltating sand grains impact the surface and eject the smaller grains that otherwise would be difficult to entrain because of their cohesive nature and low profiles to the wind. Once ejected, particles of silt and clay may remain in suspension for prolonged periods of time, riding turbulent eddies well into the atmosphere. Ejected particles of silt and clay remain in suspension until: (1) the wind velocity falls below the settling velocity of the particles, (2) electrostatic bonding of aggregates produces particles with sufficient settling velocities to fall, or (3) individual dust particles or aggregates become incorporated in rain or snowfall (Pye, 1995). Typically, loess deposits are thickest and coarsest near their sources and become progressively thinner and finer–grained downwind.

Preliminary data collected by the authors from four test sites on the plateau indicate that the thickness and distribution of loess have been controlled by three primary factors, topography, bioclimate (vegetation and moisture), and source sediment supply (Fig. 4; Sweeney et al., 2000, 2001; Gaylord et al., 2002).

Topographic Controls

Distribution and thickness of loess is a function of topography, such as incised stream valleys or other topographic impediments that can trap saltating particles (Mason et al., 1999).
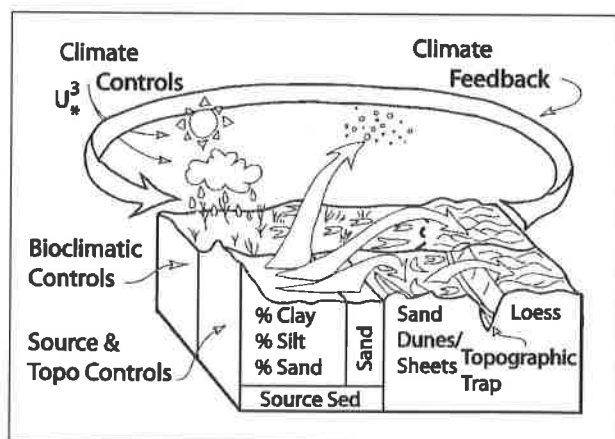

Figure 4. Conceptual diagram of pathways and fates of near–source eolian sediment and atmospheric dust (adapted from C. Zender, personal communication). Source sediment and topographic trap controls on dust and loess production (via saltation–induced ejection of silt and clay) are depicted in the front half of diagram. Bioclimatic (moisture and vegetation) controls on loess production (in the absence of a topographic trap) are depicted in the rear half of the diagram.

Saltating sand activity on a mixed–texture, source sediment inhibits the accumulation of thick loess upwind from incised valleys by persistently re–entraining particles of silt and clay or by temporarily promoting the storage of sand in dunes or sand sheets. However, thick loess can accumulate immediately downwind from incised valley 'topographic traps' because the saltating sand grains are removed from the system as they migrate into traps from which they do not escape, a situation well described for late Pleistocene loess deposits from the U.S. upper Midwest by Mason et al. (1999).

Bioclimatic Controls

Distribution and thickness of loess is a function of precipitation and vegetation density, or bioclimate. If vegetative cover or soil moisture increase, then the potential exists to maximize loess deposition. At the same time, as the loess area expands, saltating activity shifts upwind; if the bioclimatic factors expand to the point where saltation is not a major eolian process then loess generation and deposition ceases. Conversely, sand dunes and sheets will expand during times of increased aridity, effectively pushing loess deposition farther downwind.

Source Sediment Controls

Distribution and thickness of loess is also a function of the source sediment textural character. Bioclimatic and eolian conditions permitting, source sediments cannot produce thick loess deposits downwind if the source is so depleted in silt and clay (e.g., the Ephrata Fan; Stop 1–6) that even if saltating sand particles are present insufficient fine–grained particles are ejected to produce thick loess. Sand–and–gravel–rich source sediment instead tends to be reworked into dunes or sand sheets.

EOLIAN SOILS AND CONTEMPORARY WIND EROSION

The Columbia Plateau of eastern Washington, northern Oregon, and the Idaho Panhandle have experienced numerous exceedences of the mean 24–hr PM_{10} health standard (PM_{10}= airborne particles < 10–mm in diameter) since measurements were started in 1985. Evidence from recent research, largely concerning occupational health risks, has strongly implicated airborne PM_{10} as a cause of respiratory problems such as pneumoconiosis. A broad–scale research project, called the "Columbia Plateau Wind Erosion/Air Quality Project" ("CP3" or PM_{10} project for short) has provided a significantly enhanced understanding of the wind erosion and potential dust emission control strategies for the region. For more information on outcomes from this research the reader is directed to Saxton *et al.* (2000).

Starting at the I–90 underpass, drive 3.6 miles (5.8 km) south on State Highway 17 to Road 2 SE. Turn right on Road 2SE and drive 1.2 miles (1.9 km) west to a ridge crest with a view of the Quincy dunes on the horizon.

STOP 2–1. QUINCY DUNES

The Quincy dune field encompasses the largest concentration of dune sand (>1000 km^2) on the Columbia Plateau (Petrone, 1970). Although formerly more active, these dunes now are relatively dormant because extensive irrigation in the area has promoted high surface moisture and widespread growth of stabilizing vegetation. The nearby Potholes Reservoir (completed in 1952) (Pitzer, 1994) drowns a formerly active dune field. Present–day islands in this reservoir are vegetated and many retain original barchan and barchanoid dune shapes.

Drive 1.2 miles (1.9 km) to the junction with State Highway 17; turn right and drive 47.6 miles (76.7 km) south to the junction with U.S. Highway 395, passing through scabland, loess, and cultivated fine–grained outburst flood deposits. Near the town of Mesa, the road briefly descends into the steep–sided Esquatzel Coulee. At the junction with U.S. Highway 395, turn right (south) and drive 15.5 miles (25 km), to Phend Road. Turn left (east) on Phend Rd until it ends (4.2 miles; 6.9 km). Turn right onto Frontier Road and drive 1 mile (1.6 km) south to its junction with Eltopia Canal and Falls Roads. Turn left onto the south–side Eltopia irrigation canal road. Drive 1.6 miles east.

STOP 2–2. ELTOPIA CANAL

This site reveals an unusually well–exposed stratigraphic section of late Pleistocene, paleosol–capped, slackwater deposits, overlain by Holocene fluvial and eolian strata. The ~14–m–thick, Eltopia stratigraphic section (Fig. 5) preserves the most complete record of late Pleistocene and Holocene sand–dune and fluvial sedimentation yet described from the Columbia Plateau (Gaylord *et al.*, 2001). Notable features in the Eltopia Canal exposure include Mazama tephra clasts that accumulated on the slip face of an ancient sand dune (during the mid–Holocene), cicada–burrowed, dune sediments, and mixed eolian–fluvial deposits. The Mazama tephra clasts (ca. 6845 ^{14}C yrs. B.P.) (Bacon, 1983) provide the most definitive temporal control. The sedimentary deposits exposed at the eastern end of Eltopia Canal are subdivided into five informal units that span deposition from the time of outburst flooding to the Holocene (Gaylord *et al.*, 2001).

Drive Back to the junction of the Eltopia Canal Road and Falls Road and proceed along Falls Road for 5.1 (8.2 km) to the junction with U.S. Highway 395. Turn left (south) on U.S. 395 and drive 6.5 miles (10.5 km) south to junction with U.S. Highway 12. Turn southeast on U.S. 12 toward Walla Walla, driving along the Columbia River. Continue on U.S. 12 for 14.7 miles (23.6 km) to the town of Wallula. Turn left into town to the parking area across from the post office.

STOP 2–3. WALLULA GAP

Wallula Gap is a narrow canyon in the Horse Heaven Hills through which the Columbia River now passes. Outburst flood waters ponded behind this flow constriction caused backflooding of tributary valleys and widespread deposition of fine–grained, slackwater sediment. High–divide channels eroded into basalt above Wallula Gap indicate that the maximum flood stage was at least 350 m (1150 ft), or ~ 240 m above present day Wallula. Loess scarps at the entrance to Wallula Gap suggest that the maximum flood stage could have reached as high as 365 m (1200 ft) (O'Connor and Baker, 1992), or close to the highest elevations at which ice–rafted erratics are found around the Pasco Basin. Calculations of maximum flood discharges based on the high–water evidence suggest that about 10 million m^{3}s^{-1} passed through Wallula Gap (O'Connor and Baker, 1992), an amount that is 300 times the maximum flows of the 1993 Mississippi River flood (O'Connor and Waitt, 1994)!

*Drive 1.8 miles (2.9 km) south on U.S. Highway 12 to Wallula Junction, and turn east toward Walla Walla, driving < 0.1 miles to the stop sign. Cross the road to reach the parking area. A special arrangement has been made for the INQUA group to receive a guided tour of the FPL Stateline wind farm. **Note: No permission has been given to others who might be following this field***
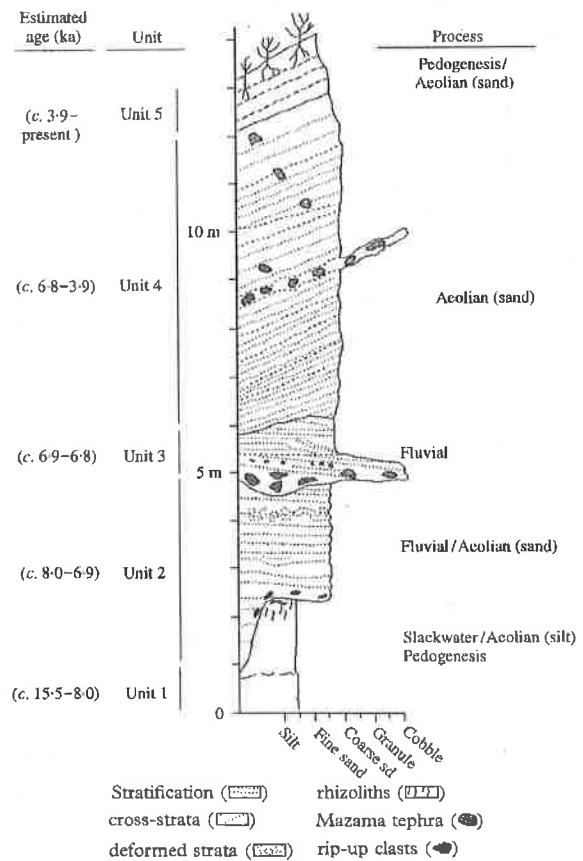

Figure 5. Eltopia Canal composite stratigraphic section (after Gaylord *et al.*, 2001). Ages loosely constrained by slackwater deposits (minimum age ca. 12.5 ka) and Mazama tephra (ca. 6.8 ka).

guide after the time of the initial trip; others who wish to visit the Stateline property must first contact FPL.

STOP 2–4. HORSE HEAVEN HILLS WIND FARM

Developed, owned, and operated by FPL Energy, the 263–megawatt, Stateline Energy Center provides clean, renewable energy to PacificCorp Power Marketing (PPM) for its Pacific Northwest customers. With more than 200 wind turbines, this site became fully operational in December, 2001.

Drive east on U.S. Highway 12 to the east side of the town of Touchet (13.2 miles; 21.3 km). A representative from the Gardena Farms Irrigation District will escort the INQUA group onto their property. Note: The Burlingame Canyon exposure is on private property of the Gardena Farms Irrigation District and persons are allowed to visit during this field trip only by special permission. Trespassing will result in access being denied to all.

STOP 2–5. BURLINGAME CANYON: THE "LITTLE GRAND CANYON": AN EOLIAN SEDIMENT SOURCE

The thick exposures of stacked slackwater rhythmites at sites such as Burlingame Canyon formed the basis for Richard Waitt's hypothesis for multiple, Missoula outburst floods during the last glaciation (Waitt, 1980, 1984, 1985). Others have argued that these stacked rhythmites may instead reflect multiple flood surges during a few large floods (Baker, 1973; Bjornstad, 1982; Bunker, 1982; Baker and Bunker, 1985). This interesting debate is central to discussions of outburst flood dynamics but less important to the paired eolian system that we are examining because regardless of their origin, these fine–grained sedimentary deposits were an ideal (and abundant) source for the eolian deposits downwind.

Turn north from U.S. Highway 12 onto Touchet Road at the east end of Touchet and drive north 21 miles (33.8 km) to the junction with State Highway 124. Turn right (east) onto Highway 124 and continue 5.3 miles (8.5 km) to the junction with Lyons Ferry Road. Turn left (north) on Lyons Ferry Road and drive 9.5 miles (15.3 km) to Smith Springs Road. Turn right (east), drive 2 miles (3.2 km) and pull off the road.

STOP 2–6. BABCOCK ROAD SECTION

The exposures on Babcock Road are located in the north–central part of Eureka Flat, a deflationary plain that has been a persistent source of loess–producing, wind–blown silt throughout the Pleistocene. These roadcuts reveal an overlying blanket of L1 loess (~1m) capping eolian–sand–sheet, outburst–flood Pleistocene sediments and older paleosols. Eureka Flat was inundated during the late Wisconsin by backflooding from the lower, southern end of the flat and possibly by flood surges that breached the Snake River divide to the northeast. Some of the subdued hills nestled within Eureka Flat, including this exposure, appear to be deeply eroded,

loess–island remnants mantled by outburst flood and sand sheet sediment and loess. Flood sediments in this part of Eureka Flat are characterized by poorly sorted silt and sand, with occasional weakly imbricated, pebble–to–cobble soil carbonate clasts. These deposits record a transition from semi–arid and arid conditions during sand sheet deposition to progressively less arid conditions during loess accumulation.

Return to the junction with Lyons Ferry Road and turn left (south), driving 7.7 miles (12.4 km) to the Clyde Site. The Skyrocket Hills, a repository for some of the thickest Palouse loess, are visible on the horizon to the left. The substantial accumulation of loess on these hills attests to the persistent loess–generating "engine" that Eureka Flat represents. Note: This exposure is on a blind curve of a road traveled by large grain trucks. Caution is warranted.

STOP 2–7. CLYDE SITE–1 (CLY–1)

This roadcut exposes the thickest known Pleistocene loess in the Palouse. The exposure includes loess units L1 and L2 as well as the Devils Canyon buried soil complex in the core of the exposure and several meters of loess beneath that. Tephras identified here include the Mt. St. Helens sets S, C, and a pre–set–C Mt. St. Helens tephra. The loess at road level has a probable age of >100 ka (Busacca and McDonald, 1994) based on stratigraphic position. The site lies immediately downwind from Eureka Flat where accumulation of loess since the LGM is almost 6 m. CLY–1 is representative of late Quaternary loess stratigraphy for sites that are proximal to major sources of eolian sediment. This section has a much thicker sequence of the same buried soils and loess layers than at sites farther downwind from their source areas, such as site WA–5 (DAY 3, STOP 3). The area that includes the CLY–1 site has the thickest known occurrences of L1 and L2 loess on the plateau (Fig. 3). Regional thickness trends for L1 loess strongly imply that Eureka Flat and the Walla Walla valley were the source of voluminous latest Pleistocene and Holocene loess.

Continue south on Lyons Ferry Road to the junction with State Highway 124. Drive 7 miles (12.3 km) east on Highway 124 to the junction with State Highway 125. Turn right (south) on Highway 125 and drive 16 miles (25.7 km) to Walla Walla. End of Day 2.

THE WASHINGTON WINE INDUSTRY

Washington State is an agricultural region with world class wines whose character is a function of the complex interplay between climate, soil, and geology, a concept known as *terroir*. In the Walla Walla area the terroir is strongly influenced by the soils that have been derived from weathering of the basalt and from slackwater and eolian deposits that have been central to this field trip. To learn more about the terroir of Washington state wines, readers are directed to Meinert and Busacca (2002).

DAY 3

The third day further explores the record of outburst flood and eolian activity as the route continues towards Spokane.

From the junction of U.S. Highway 12 and State Highway 125, north side of Walla Walla, drive 16 miles (25.7 km) north on Highway 125 to the junction with Highway 124. Turn left on Highway 124 and drive 7 miles (12.3 km) west to the junction with the Lyons Ferry Rd. Turn right (north) onto Lyons Ferry Road and drive 16.2 miles (26 km) to the junction with State Highway 261. Turn right on Highway 261 and drive 4.8 miles (7.7 km) to a turnout on the left side of the road, just before the Little Goose Dam turnoff. **Note: Be extremely cautious at this road cut because of blind corners and heavy traffic.**

STOP 3–1. STARBUCK SITE: PROTOLOESS AND PROTODUNES

This stop provides an opportunity to examine sedimentary structures and texturally diverse deposits generated during multiple, late Wisconsin, outburst floods whose waters flowed upstream into the Tucannon River valley. Although flood deposits along the Tucannon River have been known for many years, only Smith (1993) has conducted a detailed sedimentologic examination of them. His strategy was to describe and interpret sedimentary facies in measured stratigraphic sections. The deposits here are dominated by granule–to–cobble, flood–surge sediments intercalated with very–fine–sand–to–clay slackwater deposits. Six distinct flooding events are preserved at this site (Smith, 1993). The abundance of sand, silt, and clay in these flood deposits again attests to their importance as sand dune, sand sheet, and loess sources.

Drive back down State Highway 261 for 13.4 miles (21.4 km) past Lyons Ferry and the confluence of the Palouse and Snake Rivers to the turnoff for Palouse Falls State Park. On the right side of the road, 2.5 miles (4.0 km) from the last stop numerous, elongated, gravel ridges are oriented perpendicular to the road. These ridges are giant ripples or 'megaripples' (Bretz *et al.*, 1956, Bretz 1969; Baker, 1973) that record relatively late–stage, outburst–flood, traction–transport, rounded pebble–cobble–and– boulder clasts, most of which are basalt. Closer inspection will reveal a few rounded quartzite and granitic cobbles derived from plutonic and metamorphic sources in Canada, Montana, and Idaho.

STOP 3–2. PALOUSE FALLS

The road into Palouse Falls provides classic views of butte–and–basin scabland topography. At the falls, the Palouse River flows over an ~60 m high cataract cut into multiple basalt flows during the last episode of outburst flooding. The floods that sculpted the surrounding terrain flowed down the Cheney–Palouse scabland tract, overtopped the Snake–Palouse–River divide south of Washtucna Coulee and excavated a new route for the Palouse River to the Snake. Now simply an underfit stream

dwarfed by its valley, the perennial Palouse River is nevertheless a major resource for this area.

Turn right (northeast) at the Park Road entrance onto State Highway 261 and drive 8.2 miles (13.1 km) to the junction with Highway 260. Turn right and drive northeast 6.5 miles (10.5 km) to Washtucna and the junction with State Highway 26. Turn left (west) on Highway 26, drive 4 miles (6.4 km) and pull off the highway adjacent to the staired–stepped exposure.

STOP 3–3. WASHTUCNA SITE–5 (WA–5)

This roadcut contains a distinctive sequence of loess layers, buried calcic soils, sand–sheet deposits, and marker tephras that define the late Quaternary loess stratigraphy of the Channeled Scabland. It is the same sequence, although much thinner, seen at CLY–1 (STOP 2–7). Buried soils and tephras at this site were first examined by Lucy Foley and Henry Smith, who established correlations to eruptions of Mount St. Helens for several of the tephras (Foley, 1982). This site was central in developing the regional framework of late Quaternary soil–stratigraphic units (McDonald and Busacca, 1992; Busacca *et al.*, 1992). These workers defined four soil–stratigraphic units within the upper 7 m of the WA–5 section that are recognized at most of the more than 100 sites they described across the Channeled Scabland and Palouse: the Sand Hills Coulee soil, Washtucna soil, Old Maid Coulee soil, and the Devils Canyon soil.

At this stop the L1 loess overlies the Mt. St. Helens set S tephra and contains the Sand Hills Coulee soil. The L2 loess layer as noted earlier consists of eolian sediment that was deposited following the penultimate episode of flooding. The L2 layer here contains the Mt. St. Helens set C tephra and the Old Maid Coulee and Washtucna soils.

Continue west on State Highway 26 for 1.5 miles (2.4 km) and pull off the road.

STOP 3–4. WASHTUCNA SITE– 9 (WA–9)

This site exposes evidence for multiple episodes of outburst flooding prior to the late Wisconsin (McDonald and Busacca, 1988). Evidence gathered at sites such as this one indicate that loess hills adjacent to scabland coulees can contain one–to– several, large, angular unconformities (Fig. 6) that interrupt the layer–cake stratigraphy typical of sites away from scabland channels. These unconformities, along with the sedimentology of material overlying the unconformities, led McDonald and Busacca (1988) to conclude that at least six episodes of Missoula floods occurred in the Channeled Scabland during the Quaternary. All of the loess units at this site are normally magnetized (McDonald and Busacca, 1988), suggesting up to five episodes of flooding during the Brunhes Normal Magnetic Polarity Chron, or more recently than 780,000 yrs. B.P. A flood–cut unconformity at the nearby Benge site lies stratigraphically within reversely magnetized strata,

suggesting at least one episode of outburst flooding prior to 780,000 yrs. B.P. (McDonald and Busacca, 1988).

Drive 6.5 miles (10.4 km) west on State Highway 26 to the east edge of Sand Hills Coulee dunes. **Note: These dunes are on private property; permission must be granted to gain access.**

STOP 3–5. SAND HILLS COULEE DUNES

The Sand Hills Coulee dune field is a relatively small concentration of semi–active–to–stabilized, Holocene, sand dunes that partially fill a former outburst flood channel incised into Palouse loess (Schatz, 1996; Gaylord *et al.*, 1999). Blowout exposures reveal interstratified sand dune and interdune deposits, paleosols, and Mt. Mazama (ca. 6845 ^{14}C yrs. B.P.) and Mt. St. Helens set D (1980) tephras. These dunes cover ~20 km^2, have SW–NE alignments, and consist of iron–stained, quartzo–feldspathic sediment that is compositionally distinct from virtually all other Quaternary dune sands on the plateau. Unlike the Quincy, Hanford, Smith Canyon, and Juniper dunes that were derived from Missoula flood sediment, the Sand Hills Coulee dunes originated from reworking of the Miocene(?)/ Pliocene Ringold Formation that underlies flood deposits <30 km to the west (Schatz, 1996; Gaylord *et al.*, 1999). Except for occasional sand–sheet deposits intercalated with loess, such as at WA–9 (Stop 3–4), eolian sand deposits on the plateau largely post–date the LGM. Dune histories at the Hanford site (Gaylord and Stetler, 1994; Stetler and Gaylord, 1996), Juniper dunes (Gaylord *et al.*, 2001), and here (Gaylord *et al.*, 1999) all indicate that the main phase of sand dune mobility on the plateau occurred during the mid–Holocene. The absence of older, Pleistocene, sand dune deposits attests to their relative impermanence and fragility as landforms.

Drive 61.1 miles east on State Highway 26 to Colfax. Turn left (north) on U.S. Highway 195 and drive 6.6 miles (10.6 km) to the McCroskey Road turn off (on the right). This turn–off is 0.4 miles past Dry Creek Road. Follow McCroskey and Hume Roads for 4 miles (6.4 km) to the turn–off for Steptoe Butte State Park. The top is 2.5 miles (4 km) from the park entrance.

STOP 3–6. STEPTOE BUTTE

The final stop at the top of Steptoe Butte offers panoramic views of eastern Washington and northern Idaho. Rising ~300 m above the surrounding terrain, Steptoe Butte is cored by late Proterozoic metaquartzite from the Belt Supergroup. During the Miocene, Columbia River Basalt fissure flows engulfed the lower flanks of the butte, leaving it isolated from the more contiguous metamorphic and crystalline terrains to the east. During the Pleistocene, loess, locally >30 m thick, accumulated on the basalt but did not bury Steptoe Butte, likely because the steep slopes prompted relatively high rates of loess erosion.

The origin of the distinctive, steeply–rolling, complex, hill patterns of the Palouse (Kaiser *et al.*, 1951) has been the subject of
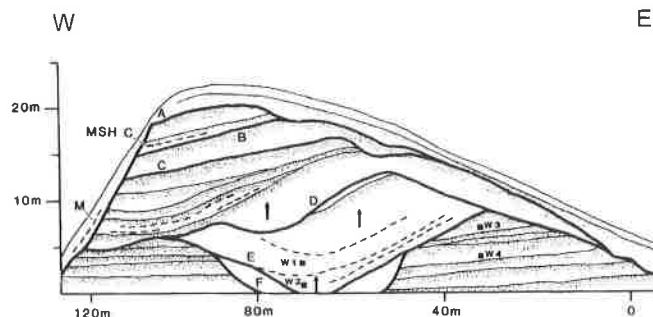

Figure 6. Low–angle–oblique photo and interpretive map of the Washtucna–9 (WA–9) roadcut. View is to the north. Heavy lines on the diagram highlight interpreted outburst–flood cut unconformities (lettered A through F); line and stipple patterns represent calcic paleosols; M is ca. 6,845 ^{14}C yrs. B.P., Mazama tephra (Bacon, 1983); C is the 36,000 ^{14}C yrs B.P., Mt. St. Helens set C tephra. Square boxes indicate samples for magnetostratigraphy; all samples were normally magnetized. Arrows represent fining–upward from sand to silt. See also Foley (1982) and McDonald and Busacca (1988).

considerable debate among geomorphologsits and pedologists. The drainage pattern is dendritic in the eastern Palouse but distinctly trellis–like with strict alignment of linear loess ridges in the southwestern Palouse (Lewis, 1960). The Palouse landscape here lies in a transition zone between the linear loess hill landscapes of the southwestern Palouse and the rounded loess hills of the eastern Palouse. Many visitors to this area misinterpret the asymmetric hill shapes of the rounded loess hills immediately below the butte as those formed by migrating sand dunes. The fine–grain sizes of the sediment making up the hills (Busacca and McDonald, 1994) and the absence of sand–dune–specific sedimentary structures precludes this possibility. We regard the asymmetric loess hill morphologies to have resulted from water erosion and seasonally controlled mass wasting. Accumulation of snow on north–facing slopes, followed by slow thawing and water saturation contributes to over–steepening of these slopes via rotational slumps and liquefied flows during the late spring. The contact between the L1 loess and underlying paleo–argillic horizons often is the plane of failure for these slumps and flows.

At the entrance to the State Park, turn left on Hume Rd and drive 1 mile (1.6 km). Turn left again onto Ragon Road and fol-

low it 6.5 miles (10.5 km) to the junction with U.S. Highway 195. Drive north 48 miles (76.8 km) to reach Spokane.

ACKNOWLEDGMENTS

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