Evolution of Eureka Flat: A dust-producing engine of the Palouse loess, USA

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Abstract

Sedimentologic, stratigraphic, and pedologic evidence demonstrates that Eureka Flat (EF), a narrow, 80-km-long deflation plain in south-central Washington, USA, is the source for the thick and widespread loess of the Palouse region. Located in the southern Columbia Plateau, EF has been a depocenter for abundant clay-to-sand-sized sediment derived from repeated glacial outburst floods that swept through the area during the Pleistocene. Prevailing SW winds that were funneled across EF remobilized flood sediment into sand dunes, sand sheets, and loess. An eolian sand sheet buried beneath 1.8 m of loess at the downwind end of EF attests to bioclimatically driven changes in the style of eolian deposition. The depositional transition from sand sheet to loess is characterized by a change from wind-rippled strata to structureless, bimodal sand and silt with insect burrows and rhizoliths, to unimodal sandy loess. Soil moisture and density of dust-trapping vegetation control where sand dunes and loess accumulate. Up to 4.5 m of post-LGM loess mantles the landscape adjacent to and downwind from EF. Some of the thickest loess deposits owe their origins to the sequestration of eolian sand particles by topographic traps such as the Touchet River and its tributary valleys. The removal of saltating sand from the eolian system promotes thick accumulation of loess downwind of topographic traps. Patterns of loess accumulation across the Columbia Plateau are identified on isopach maps of major pre- and post-LGM loess units and show that EF has been the major engine for the production of atmospheric dust and loess during both times.

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1. Introduction

Generation of extensive loess accumulations over long time scales requires an abundant, periodically regenerated source of silt and conditions suitable for its preservation (Pye, 1995). The long record of dust production and accumulation preserved on the Chinese Loess Plateau has continued in some places for nearly 22 m yr, and may have had the same source areas for just as long (Guo et al., 2002; Sun, 2002). In contrast, the extant loess in the North American Great Plains has been fed by multiple sources (Muhs and Bettis, 2003). Other major regions of loess, such as in Alaska, have long records of eolian activity, but the deposits are discontinuous and found in small, isolated areas, each with their own provenance and history (Busacca et al., 2003). Whereas thick loess derived from a vast and stationary source has been accumulating on the Chinese Loess Plateau throughout the Quaternary, shorter discontinuous records from numerous shifting dust sources are found in other parts of the world.

The Palouse loess of the Pacific Northwest, USA (Fig. 1) is perhaps as old as 2 Ma (Busacca, 1991). Cataclysmic glacial outburst floods periodically replenished the source of eolian sediment (McDonald and Busacca, 1998). Although the area of the Palouse loess is smaller than that of China or the North American Great Plains, the combination of strong prevailing winds and a long-term sediment supply has produced a long Quaternary record that rivals other great loess regions.

Eureka Flat (EF) in south-central Washington is a deflational plain aligned with prevailing dust-transporting winds (Fig. 2). The flat occupies a central location upwind of the deepest accumulations of Palouse loess and is within a major depocenter of glacial outburst flood sediments. We used field stratigraphic relationships, grain size analysis, primary sedimentary structures, and pedogenic features to define seven distinct stratigraphic units that range from...
pre-oxygen isotope stage (OIS) 2 to the Holocene. Prevailing southwesterly winds have driven this *engine of the Palouse* by reworking glacial outburst flood sediments into dunes on the upwind portion of EF and loess on the downwind portion and uplands surrounding EF. To reconstruct the history of EF as the dominant source basin for dust that supplied the Palouse loess, we use a descriptivem model that accounts for the movement and trapping of eolian sand and deposition of loess, namely topographic interactions and variations in bioclimate, defined by soil moisture and the geographic distribution of vegetative cover (Fig. 3).

Accumulations of sand dunes and loess on the Columbia Plateau (CP) are genetically linked to their shared source areas, and it is important to understand the interactions of saltation-dominated versus suspension-dominated eolian processes in the evolution of the eolian system. Loess accumulated contemporaneously with upwind eolian sand dune activity in many areas (Mason et al., 1999, 2003; Busacca et al., 2003; Muhs and Bettis, 2003). The role of sand transport is key to the evolution of the loess system. Saltating sand grains that bombard a source bed containing other sand and silt grains initiate saltation of other sand grains and eject silt- and clay-sized particles into suspension (Bagnold, 1941; Shao et al., 1993). A source bed composed predominantly of silt- and clay-sized particles would require high wind velocities to entrain particles owing to electrostatic forces that hold particles together and the low profiles of small grains to the wind (Bagnold, 1941; Pye, 1995). On the other hand, source sediments

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**Fig. 1.** Location map of eastern Washington depicting the Palouse loess, sand dunes, the Channeled Scabland, and the extent of the Cordilleran Ice Sheet. CL = Carp Lake, EF = Eureka Flat dunes, H = Hanford dunes, J = Juniper/Smith Canyon dunes, Q = Quincy dunes, U = Umatilla Basin dunes; PB = Pasco Basin, UB = Umatilla Basin, WWV = Walla Walla Valley, YV = Yakima Valley. DU-1 and WA-3 are site locations.

**Fig. 2.** Shaded relief map of Eureka Flat in relation to Walulla Gap and the prevailing wind direction. Dashed line surrounds areas >365 masl (O’Connor and Baker, 1992) that were not modified by OIS 2 glacial outburst floods. Major site locations are depicted.
composed primarily of sand would yield very little suspended eolian material. Sedimentary sources containing a range of grain sizes have the potential to produce an abundant supply of dust that will eventually settle to form loess.

Mason et al. (1999) proposed that topographic traps (Fig. 3A) influenced the thickness and distribution of thick loess in the Upper Mississippi valley. Topographic traps function in the following way: As sand dunes migrate downwind they may encounter a stream valley or canyon and the sand becomes trapped, inhibiting further migration of the dunes. This trap for saltating sand protects sediments deposited by suspension fallout in front of advancing dunes that would otherwise be re-entrained by saltation. Incised streams on EF influenced the distribution and thickness of loess downwind from eolian sand.

In the bioclimatic model of loess distribution (Fig. 3B), the alternating deposition of eolian sand and loess through time is explained as a function of climate change (Pye, 1995). Bioclimates refer to soil moisture and vegetation cover and density. Both soil moisture and vegetation are controlled by precipitation, counterbalanced by evapotranspiration. During dry climate episodes, sand dunes and saltating sand grains migrate downwind along a climatic gradient of increasing precipitation until inhibited by plants or increased soil moisture. Loess accumulates on vegetated surfaces downwind of the dunes. If precipitation increases over time, the area of dune activity is restricted upwind by a larger area of increased soil moisture and vegetation cover, allowing dust deposition on stabilized eolian sand accumulations. The boundary between dune and loess areas may shift through time, in a pattern similar to the deposition of alternating eolian sand and loess during the Pleistocene at the desert–loess boundary in China (Ding et al., 1999; Sun et al., 1999; Rokosh et al., 2003). Bioclimatic interactions with eolian sedimentation on EF dominate where topographic barriers to saltating sand are absent.

2. Background

2.1. Setting, modern climate and soils

Loess in the Pacific Northwest discontinuously mantles ~50,000 km² in southeastern Washington, northeastern Oregon, and the panhandle of Idaho (Fig. 1). Up to 76 m of Quaternary loess occur in the eastern part of this area known as the Palouse (Ringe, 1970). Strong, prevailing winds from the southwest occur during Fall and Spring (Phillips, 1970). The modern climate for eastern Washington is characterized by cold, wet winters and warm to hot, dry summers (Phillips, 1970). Precipitation increases from west to east across the CP due to the rain shadow caused by the Cascade Mountains to the west. Mean annual precipitation (MAP) increases from 150 mm in the EF area to greater than 800 mm at the eastern margin of the Palouse. Natural vegetation prior to agricultural development changed with increasing MAP from sagebrush-steppe in the driest areas, to pure bunchgrass steppe, to meadow-steppe containing mesophytic shrubs, and finally, to several community types of coniferous forest (Daubenmire, 1970). The gradients of MAP, vegetation and differing parent materials across the CP have dictated the types of soils that have formed in eolian sediments, from Entisols in dune
lands, to Aridisols in loess under <230 mm MAP and shrub steppe, to Mollisols from 230 to 600 mm MAP in steppe or meadow steppe, to Alfisols and Andisols in higher-elevation forest settings (Boling et al., 1998).

2.2. Glacial outburst floods and eolian sources

The geomorphology and sedimentology of the CP were strongly influenced by glacial outburst flooding during OIS 2 (Waitt, 1985; McDonald and Busacca, 1988; Busacca and McDonald, 1994) but floods occurred episodically for the last 2 Ma (Bjornstad et al., 2001; Pluhar et al., 2002). Catastrophic flooding during OIS 2 occurred from \( \sim 15.3–12.7 \) \(^{14}\)C yr BP (Waitt, 1985). It is widely accepted that an ice dam impounding glacial Lake Missoula released large volumes of water through a part of southeastern Washington known as the Channeled Scabland (Bretz, 1923; Baker and Bunker, 1985; Waitt, 1985; Clague et al., 2003). Floods scoured and eroded the Miocene-age Columbia River Basalt (Hooper, 1982), forming steep-sided channels called coulees, partially removing pre-existing loess, and forming streamlined remnants of the thick loess cover called loess islands. Outburst floods were hydrologically dammed at Wallula Gap, a narrow canyon where the Columbia River cuts through the Horse Heaven Hills (Figs. 1 and 2). Backflooding behind this hydraulic dam created temporary lakes that occupied the Yakima Valley, Pasco Basin and Walla Walla Valley. Maximum elevations of slackwater sediment (fine-grained outburst flood sediment deposited in backflooded valleys) and ice-rafted debris are estimated to have reached about 365 m above sea level (asl) in south-central Washington (Fig. 2) (Baker, et al., 1991; O’Connor and Baker, 1992). Other constrictions along the Columbia River resulted in slackwater deposition in the Umatilla Basin in Oregon (O’Connor and Waitt, 1995).

Source sediments capable of providing the volume and textural range of Quaternary eolian sediments on the CP include Pleistocene glacial outburst flood sediment and loosely consolidated sediment of the Miocene–Pliocene Ringold Formation (Lindsey and Gaylord, 1990). Both sources lie immediately upwind of sand dunes and loess, but the glacial outburst flood sediments cover a larger spatial extent (Busacca and McDonald, 1994). Many sand dunes on the CP are mineralogically similar to sands from glacial outburst floods (Gaylord et al., 1991; 2001). Major and trace element geochemistry suggests that the primary source of the loess is indeed outburst flood deposits, with minor contributions from the Ringold Formation (Sweeney et al., 2002).

2.3. Sand dunes

Sand dunes on the CP are confined to areas with <250 mm MAP at present, and are found surrounding the upwind perimeter of the Palouse loess (Fig. 1). Most modern sand dunes lie within basins inundated by glacial outburst floods and rest on fine- to coarse-grained outburst flood sediments (Gaylord and Stetler, 1994; Gaylord et al., 1999, 2001; Sweeney et al., 2001). All sand dunes are aligned with the prevailing south-southwesterly winds. Cropping and irrigation have artificially stabilized most sand dunes today.

The largest areas of active dunes on the plateau occur between Quincy and Moses Lake, on the Hanford site, and near Smith Canyon (Fig. 1). Parabolic dunes dominate within these dune fields but some contain barchan or barchanoid ridge dunes (Petrone, 1970; Gaylord and Stetler, 1994; Gaylord et al., 2001). Sand sheets are found on the perimeter of dune fields, but their extent and character is poorly understood because of agricultural modification. Major phases of dune activity apparently occurred just before and after the Mazama climactic eruption (ca. 7600 cal yr BP; Zdanowicz et al., 1999) in the mid-Holocene (Gaylord et al., 2001, 2003).

2.4. Loess stratigraphy

Palouse loess accumulations are characterized by 19 or more pedogenically modified beds (Busacca, 1989). The maximum age of the loess is as old as 2 Ma (Busacca, 1989).
The latest Pleistocene loess is divided into two informally named chronostratigraphic units: L1 and L2 (Fig. 4). Major episodes of loess accumulation occurred primarily during interglacials or interstadials, while major episodes of soil formation occurred primarily during glacial periods (Berger and Busacca, 1995; Richardson et al., 1997, 1999). In most other loess regions, soil formation is most pronounced during interglacials and interstadials, while loess accumulation is associated with dry and windy glacial episodes (Pye, 1995). In the Palouse, major paleosols that formed during glacial stages define boundaries between loess units, and weakly formed paleosols that are found within loess units represent minor changes in accumulation rates and/or climatic conditions (Busacca and McDonald, 1994).

The most recent loess unit is L1 (ca. 0–15 ka). The L1 contains the modern surface soil and the Sand Hills Coulee Soil, a weakly developed paleosol defined by a zone of filamentous soil carbonates (McDonald and Busacca, 1992) (Fig. 4). The age of the Sand Hills Coulee Soil is poorly constrained at present. It typically lies decameters to a meter or more above the Mt. St. Helens set ‘S’ tephra (15,400 cal yr BP; Mullineaux, 1986; calibration of Stuiver and Reimer, 1993), and therefore is likely to have formed in the latest Pleistocene or early Holocene. The base of the L1 loess is defined by the Mt. St. Helens set ‘S’ tephra. Where this tephra is missing, the base of the L1 is delimited by: (1) the base of the Sand Hills Coulee Soil, (2) the top of the Washtucna Soil, a prominent paleosol developed in the upper L2 loess, or (3) the contact with OIS 2 outburst flood deposits, where present (Busacca and McDonald, 1994).

The next older loess unit is L2 (ca. 15–70 ka). It contains the Washtucna Soil, a well-developed paleosol in the upper L2 that formed from about 20–40 ka (Richardson et al., 1997) during arid climate conditions under shrub steppe vegetation (O’Geen and Busacca, 2001; Blinnikov et al., 2001) (Figs. 4 and 5). The Washtucna soil has abundant pedogenic carbonates that locally form a laminar cap (McDonald and Busacca, 1990) reaching Stage IV development (Gile et al., 1966). The soil is almost completely bioturbated with 1–2 cm in diameter cylindrical insect burrows (McDonald and Busacca, 1990; O’Geen and Busacca, 2001). The plantopal phytolith assemblage of this soil, a proxy for vegetation type, indicates that sagebrush steppe was the dominant plant community on the CP, whereas bunchgrass steppe has dominated most parts of the CP since the beginning of the Holocene (Blinnikov et al., 2001). Nymphs of cicadidae (cicadas) live in close association with woody shrubs, principally Artemisia, and are responsible for the cylindrical, back-filled burrows (O’Geen and Busacca, 2001).

The Old Maid Coulee Soil is a weakly developed L2 paleosol below the Washtucna Soil that is recognized by an increase in filamentous soil carbonates (McDonald and Busacca, 1992). The Mt. St. Helens Set ‘C’ tephra, with a TL age on volcanic glass of 46,300±4800 yr (Berger and Busacca, 1995), lies immediately below the Old Maid Coulee Soil in localities where the two occur together. Luminescence ages near the base of the L2 loess are about 70 ka (Richardson et al., 1997). The L2 loess rests upon the Devils Canyon Soil, formed in the upper part of the L3 loess unit. The Devils Canyon Soil (Fig. 4) is another well-developed paleosol that contains pervasive cicada burrows and is cemented by calcium carbonate (McDonald and Busacca, 1992).

### 2.5. Eureka Flat

EF is immediately downwind of Walulla Gap (Fig. 2) and is bounded by the Snake River on its northern and western margins and by the Touchet River on its southeastern margin. The Snake and Columbia Rivers meet at its southwestern corner. EF rises gently in elevation from ~100 masl at the confluence of the Columbia and Snake rivers to over 500 masl at its northeastern, or downwind, margin. Precipitation increases from about 180 mm MAP at the upwind margin of EF to about 350 mm at the downwind margin. Airflow–terrain interactions prompted by prevailing southwestern winds and prominent terrain features focus winds through the Columbia River Gorge (Gregg, 1964) and through Walulla Gap, resulting in airflow acceleration and downwind dune formation similar to the Windy Gap area of Wyoming (Gaylord and Dawson, 1987). The focusing effect of Walulla Gap across EF is the likely driving force behind the development of an elongated deflational zone as well as the EF dust engine. EF is underlain by the Columbia River Basalt, which slopes gently to the southwest (Ringe, 1970). To the west of EF, anticlines in the basalt generated by north–south compression were formed (Reidel et al., 1989) that are roughly perpendicular to the elongate trough in which EF resides, suggesting EF is an erosional rather than structural feature.
3. Methods

Eolian sediments and paleosols were characterized using outcrop exposures, hand-auger cuttings, and hydraulic soil-probe cores. The location of each sample site was recorded using a global positioning system (GPS) unit. Surface samples were collected from a depth of 30 cm in order to avoid sediment mixing by agricultural tillage. Soil descriptions followed standard methods outlined by the National Soil Survey Center (Schoeneberger et al., 1998). Samples were collected at 30 cm intervals and analyzed for grain size, tephras, sediment chemistry, and weight % soil carbonate. Grain size distributions were determined using a Malvern Mastersizer S laser diffractionometer that measures volume percent of particles in 64 size classes from 0.5 to 850 μm. Samples were pretreated prior to analysis using sodium acetate to dissolve carbonates and hydrogen peroxide to oxidize organic matter. Samples were rinsed with deionized water, centrifuged, and excess supernatant was decanted. Each sample was dispersed with sodium hexametaphosphate and analyzed in a de-ionized water suspension with no sonication (Supplementary Data, Appendix A). Tephras were analyzed with a Cameca Camebax electron microprobe at Washington State University, using an acceleration voltage of 15 kV, a beam current of 10 nA, and a beam diameter of 6 μm. Approximately 20 glass shards from each sample were chemically analyzed and compared to a database of Pacific Northwest tephras and known tephra standards using a similarity coefficient (SC) calculated for each sample (Borchardt et al., 1972). X-ray fluorescence (XRF) elemental composition was used to determine provenance of eolian units. Samples for XRF were pretreated with sodium acetate to remove carbonates and sieved to between 62.5 and 500 μm (very fine to medium sand range). Analytical techniques for XRF are those of Johnson et al. (1999). Calcium carbonate content of loess-derived paleosols was determined using a titration method (Loeppert and Suarez, 1996). The grain size and thickness of key stratigraphic units were compiled and mapped using ArcGIS software. Grain size and unit thickness data were interpolated using an ordinary kriging method within the ArcGIS Geostatistical Analyst.

4. Results

Seven distinct sedimentary units were identified in EF (Fig. 6). Unit 1 includes the oldest exposed sediments of EF, composed of eolian and fluvial sediments that are strongly modified by pedogenesis. Unit 2 is composed of pre-OIS 2 outburst flood deposits. Units 3 and 4 include older eolian sand and contemporaneous L2 loess, respectively. Unit 5 is composed of OIS 2 glacial outburst flood deposits. Units 6 and 7 include recent eolian sand and contemporaneous L1 loess, respectively. Fig. 7 demonstrates the spatial extent and depositional relations between the units. All named field sites discussed below are shown in Fig. 2.
by sand-rich flood sediment, and Gravel 2b, composed of reworked pebble-sized nodules of soil carbonate.

Gravel 2a is composed predominantly of cross-stratified basalt pebbles and cobbles. Exposures of Gravel 2a on EF are found close to the Snake and Columbia Rivers, which were major flood channels. The DODD-1 site at the southwestern margin of EF is a 10 m exposure that includes 5 m of fluviolacustrine sands and gravels (Gravel 2a) overlain by 6 m of Unit 5 sediments. A well-developed petrocalcic horizon formed in the top of Unit 2. Gravel 2a is composed of basalt-rich cross-stratified and imbricated pebbles suggesting east or northeast paleocurrent directions. Site LM-1 on the northwestern margin of EF in the deep Palouse loess exposes L1 and L2 loess resting upon 2 m thick foresets that document a northeast-directed paleocurrent (Fig. 8). Calcium carbonate accumulated at the top of the foresets with some seams parallel to cross-strata.

Gravel 2b is composed of pebble-sized nodules of carbonate-cemented loess that are rounded to subangular. A small percentage (<10%) of the clasts are carbonate-coated basalt pebbles. Gravel 2b is generally <1 m thick and is grain- or matrix-supported (Fig. 9C). The matrix is silt and sand-rich. Gravel 2b is locally imbricated; however, the data are inadequate to determine a paleocurrent direction. Gravel 2b is located within EF and exposed

Table 1
Glass chemistry of tephras from Eureka Flat

<table>
<thead>
<tr>
<th>Oxide</th>
<th>BAB-6</th>
<th>TOU-2 upper</th>
<th>TOU-2 lower</th>
<th>SHEF-1</th>
<th>ANW1</th>
<th>ANW2</th>
<th>MSH Sog</th>
<th>MSH Sg</th>
<th>Mazamaa std</th>
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<tr>
<td>SiO₂</td>
<td>74.93 (0.43)c</td>
<td>76.79 (0.79)</td>
<td>76.31 (0.18)</td>
<td>76.56 (0.24)</td>
<td>73.25 (0.31)</td>
<td>76.74 (0.25)</td>
<td>77.08 (0.28)</td>
<td>76.50 (0.14)</td>
<td>73.26 (0.19)</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>14.15 (0.20)</td>
<td>13.38 (0.48)</td>
<td>13.70 (0.08)</td>
<td>13.70 (0.14)</td>
<td>14.40 (0.20)</td>
<td>13.44 (0.15)</td>
<td>13.39 (0.18)</td>
<td>13.80 (0.09)</td>
<td>14.34 (0.11)</td>
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<tr>
<td>Fe₂O₃</td>
<td>1.52 (0.10)</td>
<td>1.27 (0.16)</td>
<td>1.30 (0.06)</td>
<td>1.35 (0.08)</td>
<td>2.14 (0.06)</td>
<td>1.27 (0.07)</td>
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<td>1.29 (0.03)</td>
<td>2.26 (0.10)</td>
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<td>0.18 (0.02)</td>
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<td>CaO</td>
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<td>MSH Sg</td>
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<td>MSH Sog</td>
<td>MSH Sg</td>
<td>Mazama</td>
<td>MSH Sog</td>
</tr>
</tbody>
</table>

a MSH standards, Clague et al. (2003).

c Standard deviations of the analyses are given in parentheses.
d Analyses normalized to 100 wt%.
below Units 3 or 4. At BAB-5 in north-central EF, Gravel 2b is deformed into 0.25 m high antiforms and synforms (Fig. 9C). Site PV-1 on the northwestern margin of EF exposes 5 m of L1 and L2 loess, overlying Gravel 2b. At PV-1, Gravel 2b rests on top of an older paleosol (possibly the Devils Canyon Soil) dominated by cicada burrow fabric.

4.1.3. Unit 3: sand sheet

Sand sheet deposits of Unit 3 were mapped in topographic lows and on ridgetops covering an area of about 38 km² (Figs. 9 and 10). The sand sheet underlies the L1 loess (Unit 7) at the downwind, northeastern margin of EF. Thickness of Unit 3 ranges from a few dm to 1 m where exposed on ridges, and > 1 m in topographic lows. The sand appears to blanket preexisting topography, as does the overlying Unit 7 loess. In most exposures, the sand is structureless and several exposures contain burrows and calcified rhizoliths. The WEL-1 exposure in the northern part of EF exposes weakly expressed cross-strata within Unit 3 that dip between 20° and 27° NE. The BAB-5 site exposes thin, inversely graded laminations (Fig. 9B). The fine to medium sand of the sand sheet is poorly to moderately well sorted, and well- to sub-rounded. The sand is rich in quartz and feldspar and contains a small percentage (~10%) of basaltic lithic grains and other heavy minerals. A plot of major elements comparing outburst flood sands with sand sheet sands shows that the sand sheet is depleted in MnO, P₂O₅, TiO₂, MgO, CaO, and FeO, and enriched in SiO₂ relative to modern dune sand (Table 2; Fig. 11).

Stacked sets of inversely graded laminations are exposed in the lower meter of the sand sheet at BAB-5. The upper part of the sand sheet is structureless and contains calcified rhizoliths and silt-filled burrows. The contact with the overlying loess is gradational. Silt content increases upward with a corresponding decrease in sand content. The structureless sands have a bimodal grain size distribution with sand and silt modes. In some locations, the sand is interbedded with silt-dominated strata up to 30 cm thick. The silt strata have a mean grain size in the coarse silt range, are unimodal and finely skewed. Sites BAB-5 and EF-4 both have two silt
carbonate content in this paleosol falls within the range reported from the Sand Hills Coulee Soil (1.6–13.5 wt% CaCO₃) and at the low end of the range reported for the Washtucna Soil (6–24 wt% CaCO₃) (McDonald and Busacca, 1992). This paleosol lacks any of the cicada burrow fabric or Stage IV carbonate morphology seen in the Washtucna Soil.

4.1.5. Unit 4: L2 loess

Though no exposures of Unit 4 L2 loess were found within EF, L2 loess is found in the uplands surrounding EF, including at CLY-1, where a 7.5 m thick fining-up succession of loess is preserved (Fig. 4). At the CLY-1 site and other loess sites close to EF, the Washtucna Soil is bifurcated (McDonald and Busacca, 1990). This bifurcated soil has two densely burrowed and calcified zones separated by a thin, less-resistant bed of loess containing a lower density of burrows. A single Washtucna Soil formed in loess north of the Walla Walla Valley and at other sites far from loess sources such as EF.

4.1.6. Unit 5: OIS 2 glacial outburst flood sediment

Glacial outburst flood sediments that comprise Unit 5 are mapped at elevations below 365 m asl and are characterized by thick structureless beds (about 0.5 m thick) or thinly bedded (cm scale) rhythmites (Fig. 12). Structureless beds of sand and silt containing granule lenses are the most common flood deposit, whereas rhythmites are preserved at elevations below 250 m asl. Samples of tephra that occur at the interface between Unit 5 flood sediment and Unit 7 L1 loess were compared to standards (Clague et al., 2003). Rhythmites at the TOU-2 site are separated from overlying L1 loess by the MSH set So tephra (TOU2-upper; SC = 0.98) (Fig. 12; Table 1). The MSH set Sg tephra (TOU2-lower; SC = 0.98) (Table 1) was identified below several meters of flood sediment at TOU-2.

At some sites, floods scoured L2 loess and older sediments, rather than depositing sediment. At SHEF-1, floods scoured the top of the L2 loess and removed the Washtucna Soil. The presence of the MSH set Sg (SC = 0.98) (Table 1) allowed the distinction between the L1 and L2 in the absence of other pedostratigraphic indicators. An unconformity located at ANW-1 in the central part of EF is interpreted as outburst flood-generated because it is draped by MSH set S tephra (ANW2; SC = 0.96; Table 1). The unconformity separates older Unit 1 sediments from Holocene-aged fluvial and eolian sediments containing rip-up clasts and burrow-fill of Mazama tephra (ANW1; SC = 0.99; Table 1).

Flood sediments have a wide range of sizes and distributions. Much of the sediment that accumulated during backflooding of EF has a mean grain size of 30–50 µm and has a unimodal distribution with a peak in the coarse silt range (Fig. 13). Some silt-rich flood sediments have a bimodal or polymodal distribution and all are poorly sorted. Most of the fine-grained sediment

interbeds that contain rhizoliths, burrows, and Fe and Mn mottles.

The sand sheet at the southwestern margin of EF was modified by flooding and is found in isolated, discontinuous exposures. At EF-18, below the maximum flood elevation, the sand sheet is preserved on an erosional remnant of Unit 1. Weakly defined cross-strata within the sand sheet dip to the north or northeast. Along the southeastern margin of EF (TOU-1), 1.5 m of sand underlies flood sediment of Unit 5 and has a higher percentage of basalt lithic grains (~15%) and mica than the sand sheet to the north. Unit 3 sands at TOU-1 are predominantly composed of mm-scale laminated, horizontal to low-angle strata (<20°) with rare interspersed gravel lenses 0.5 m thick and 1.0 m wide, and scour and fill structures.

4.1.4. Paleosol development within Unit 3

A paleosol is exposed at BAB-5 at the interface between loess (Unit 7) and sand sheet strata (Unit 3; Fig. 9A). This paleosol contains ~6–9% total calcium carbonate (Table 3) creating a horizon about 40 cm thick. The paleosol is characterized by filamentous and pore-filling carbonate, weak vertical seams of carbonate, and moderately developed very coarse prismatic and blocky structure. The

![Fig. 10. Contour map depicting occurrence and thickness of the buried sand sheet of Eureka Flat. Contours indicate thickness of the sand sheet in m. White dots depict sample locations. A continuous sheet was mapped in the northern part of Eureka Flat. Remnants spared from flood erosion were found in the southern part of Eureka Flat.](image-url)
Table 2
Major and trace element geochemistry of eolian and flood sands

<table>
<thead>
<tr>
<th>Sand dunes</th>
<th>Sand sheets</th>
<th>Outburst flood</th>
</tr>
</thead>
<tbody>
<tr>
<td>L5 SE2 S8 S45 S52</td>
<td>EF18-250 TOU1U TOU1B W300 B1-240 EF1-180</td>
<td>S10 DG8a DG8b FL1 FL2</td>
</tr>
<tr>
<td>Unnormalized Major Elements (Weight%):</td>
<td></td>
<td></td>
</tr>
<tr>
<td>SiO₂</td>
<td>72.13</td>
<td>72.04</td>
</tr>
<tr>
<td>Al₂O₃</td>
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<td>12.60</td>
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<tr>
<td>TiO₂</td>
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<td>0.855</td>
</tr>
<tr>
<td>FeO*</td>
<td>2.78</td>
<td>1.75</td>
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<tr>
<td>MnO</td>
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<td>0.079</td>
</tr>
<tr>
<td>CaO</td>
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<td>2.99</td>
</tr>
<tr>
<td>MgO</td>
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<td>Na₂O</td>
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<td>2.87</td>
</tr>
<tr>
<td>P₂O₅</td>
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<td>0.182</td>
</tr>
<tr>
<td>Total</td>
<td>100.29</td>
<td>100.09</td>
</tr>
</tbody>
</table>

Normalized major elements (ppm): |
Ni | 18 | 19 | 19 | 12 | 14 | 14 | 14 | 13 | 12 | 10 | 9 |
Cr | 35 | 42 | 47 | 26 | 32 | 22 | 24 | 21 | 14 | 16 | 9 |
Sc | 14 | 12 | 14 | 7 | 16 | 6 | 6 | 5 | 4 | 4 | 5 |
V | 18 | 12 | 19 | 13 | 103 | 135 | 48 | 45 | 41 | 24 | 36 | 29 |
Ba | 127 | 119 | 133 | 103 | 135 | 48 | 45 | 41 | 24 | 36 | 29 |
Sr | 326 | 390 | 380 | 386 | 337 | 347 | 321 | 347 | 317 | 345 | 317 |
Zr | 168 | 162 | 174 | 136 | 181 | 121 | 177 | 128 | 103 | 130 | 112 |
Y | 26 | 23 | 24 | 24 | 26 | 13 | 15 | 13 | 9 | 12 | 10 |
Nb | 14.4 | 13.5 | 13.9 | 10.5 | 11.9 | 11.3 | 13.9 | 10.7 | 8.7 | 9.7 | 6.6 |
Ga | 15 | 15 | 14 | 14 | 15 | 14 | 12 | 13 | 15 | 13 | 11 |
Cu | 12 | 12 | 13 | 7 | 12 | 8 | 9 | 9 | 8 | 6 | 5 |
Zn | 59 | 54 | 56 | 49 | 62 | 27 | 27 | 27 | 20 | 27 | 16 |
Pb | 14 | 15 | 14 | 12 | 9 | 14 | 19 | 14 | 16 | 15 | 12 |
La | 27 | 27 | 31 | 19 | 17 | 24 | 24 | 24 | 19 | 17 | 19 |
Ce | 61 | 55 | 61 | 45 | 54 | 39 | 45 | 33 | 30 | 24 | 26 |
Th | 8 | 8 | 7 | 4 | 10 | 7 | 5 | 2 | 6 | 6 | 5 |

<table>
<thead>
<tr>
<th>Elements</th>
<th>EF1-250 TOU1U TOU1B W300 B1-240 EF1-180</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>72.13</td>
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<td>2.53</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.194</td>
</tr>
</tbody>
</table>

from outburst floods has a similar texture to loess, making differentiation between these two deposits difficult at times. In some places, bioturbation obscures the boundary between flood sediment and overlying loess. The tops of some flood sediment contained abundant root pores. Fine-grained flood sediment becomes coarser and dominated by sand towards the upwind margin of EF near Wallula Gap, demonstrating that this was one of the main points of entry for floodwaters entering into EF from the Columbia River.

4.1.7. Unit 6: eolian sand

Sand dunes were generated at the upwind margin of EF in an active deflationary zone floored by gravel and granule lags, which armor sand and silty-rich glacial outburst-flood sediments (Unit 5). Sand dunes include stabilized to active parabolic, blowout, and coppice dunes, as well as sand stringers and sand sheets (Fig. 14). Sand stringers are thin, low-standing remnant limbs of parabolic dunes. The orientation of the dunes and sand stringers is about N50E, reflecting the prevailing wind direction. Many dune limbs and sand stringers extend at least 1 km parallel to the wind direction, and several exceed 2 km in length.

Eolian sand grains of Unit 6 are composed of quartz, feldspar, mica, basalt, and a heavy mineral suite that is compositionally similar to that of the outburst flood sands. Recent dunes and sand-rich flood deposits contain ~20–30% basaltic lithic grains. A plot of major element chemistry from Unit 5 outburst flood sand and Unit 6...
eolian sand exhibits a 1:1 relationship (Fig. 11; Table 2), which is presumptive evidence that dune sands were derived from the flood sediments.

The bulk of the eolian sand deposits of Unit 6 were mapped in the southwestern margin of EF, with eolian sediments fining progressively downwind (Fig. 15). Eolian sands are poorly to well sorted and have mean grain sizes of 200–300 µm (fine to medium sand) (Fig. 13). Active dune sands are typically well sorted, while vegetated dunes are poorly sorted. Sand sheets are usually <1 m thick and are composed of poorly sorted sand and silt. The sand sheets have some horizontal laminations and crude stratification but most are structureless.

4.1.8. Unit 7: L1 loess

L1 loess (Unit 7) rests upon flood deposits (Unit 5) and sand sheets (Unit 3) and is downwind of eolian sand on the northeastern part of EF (Fig. 9). The thickness of L1 loess...
on EF is 1.2–1.8 m. The loess is texturally unimodal, with a mean grain size ranging from 40 to 60 μm that includes 20–30% sand-sized particles (>63 μm) and ~10% clay-sized particles (<4 μm). In many localities, the loess fines upward from the base, with the mean grain size decreasing by about 20 μm and sand percent decreasing by about 10% (Figs. 9 and 13). The modern surface soils formed in the loess have coarse prismatic structure, a weakly developed A horizon, and are classified as Mollisols. Calcium carbonate was leached to a depth of ~1.2 m, forming a Bk horizon.

The uplands surrounding EF contain thick L1. At the CLY-1 site east of EF (Fig. 2), the L1 is 4.5 m thick (Fig. 5). Particle size within the L1 at CLY-1 fines upward from its base from a mean of 76–44 μm, accompanied by a 25% decrease in sand content (Figs. 4 and 13).
4.2. Interpretation of units and the evolution of EF

4.2.1. Unit 1: older sedimentary units

The oldest sediment exposed in EF was likely emplaced by fluvial and eolian processes, then stabilized, vegetated, and modified during long intervals of pedogenesis. The cm-scale rhizoliths are most likely roots of woody shrubs (Retallack, 1990). Conditions for the formation of well-developed calcium carbonate horizons similar to these can take thousands to tens of thousands of years to form in the Palouse (McDonald and Busacca, 1990). The soil carbonates in Unit 1 are a likely source of the carbonate within Gravel 2b of Unit 2. Higher chroma, pervasive soil features, and presence of a possible Mt. Rainier tephra in the upper part of the unit suggests they were deposited before OIS 2.

4.2.2. Unit 2: pre-OIS 2 flood gravels

Paleomagnetically constrained sediments and stratigraphic evidence indicate that glacial outburst flooding occurred several times prior to the LGM (Patton and Baker, 1978; McDonald and Busacca, 1988; Bjornstad et al., 2001; Pluhar et al., 2002). The stratigraphic, pedologic, and geographic characteristics of Gravel 2a of Unit 2 are consistent with a pre-OIS 2 flood origin. Gravel 2a at DODD-1 is separated from Unit 5 flood sediments by a petrocalcic paleosol that required tens of thousands of years to form. Stratigraphic evidence suggests both Gravels 2a and 2b predate the OIS 2 episode of floods.

Clast imbrication within Gravel 2b suggests that the nodules of pedogenic carbonates formed elsewhere and were eroded and transported as tractional, subaqueous transport. The convolute nature of Gravel 2b suggests soft-sediment deformation. Concentrations of matrix-poor gravel beds lacking imbrication may have formed as deflationary lags during eolian activity that generated the overlying sand sheet (Unit 3). The interleaving of Gravel 2b between the L2 and L3 loess units at PV-1 is compelling evidence for emplacement during an outburst flood event prior to L2 loess deposition (~70 ka; OIS 4). Sites at the northern margin of EF, including PV-1 that contain Gravel 2b at elevations of up to 500 m are well above OIS 2 flood stages, suggesting that pre-OIS 2 flooding may have been even more voluminous. We propose that the pre-OIS 2
flood waters overtopped the Snake River-Palouse divide, sending waters southward across EF, scouring older sedimentary units, leaving only erosional remnants of Unit 1, and depositing reworked nodules derived from older soil petrocalcic horizons.

4.2.3. Unit 3: sand sheet

The sand sheet exposed at the northern margin of EF is interpreted as an eolian sand sheet. We attribute the inversely graded, subcritically climbing translatent strata exposed at BAB-5 to migrating wind ripples (Hunter, 1977). Structureless sand units in the same stratigraphic position as the inversely graded strata suggest they have a similar eolian origin. The rare gravel lenses within the sand sheet suggest either deflation or local reworking of the eolian sand by streams. The above suite of features is consistent with those found in a vegetated eolian sand sheet.

Sand sheets commonly have bimodal grain size distributions due to poor sorting of the sand (Fryberger et al., 1979). When vegetation is present, bioturbation commonly produces structureless deposits while, in areas where vegetation is absent, wind ripples may be generated and preserved (Kocurek and Nielson, 1986). Ephemeral streams also commonly rework sand sheets (Fryberger et al., 1979; Schwan, 1987). Sand sheet development is favored if there is an adequate source of sand, sparse vegetation, low relief, and formation of dunes is inhibited (Kocurek and Nielson, 1986; Kasse, 1997). Kasse (1997) argued that low sand availability favored the generation of sand sheets instead of dunes. It is unclear, however, what factors ultimately control whether sand sheets or sand dunes form (Kocurek and Nielson, 1986).

Lower concentrations of major elemental oxides in the sand sheet (Table 2) probably reflect lower percentages of basaltic lithic grains within the sand compared to modern dune sands. Sand sheet sediments are compositionally equivalent to modern dune sands in all respects except for basalt content. Assuming a flood sediment source, it appears that the sand sheet formed from basalt-depleted sediments or that compositional sorting (due to the higher density of basaltic lithic grains) during eolian transport produced a basalt-depleted sand sheet.

Alternating beds of sand and silt are usually preserved at the base of the Unit 3 sand sheet (e.g. Fig. 9) where it is thickest; these beds may reflect intermittent episodes of loess accumulation onto the sand sheet during times of increased vegetation density or soil moisture. The silt strata have similar grain size distributions to modern loess, suggesting they are eolian. Rhizoliths and Fe and Mn mottles within the lower sand and silt strata of Unit 4 record the presence of plants early in the evolution of the sand sheet.

4.2.4. Paleosol development and sand sheet age

The paleosol located at the interface between Units 3 and 7 probably formed during the stabilization of the sand sheet and the onset of loess accumulation. Soil morphology and carbonate content of the paleosol suggest that it is a variant of the Sand Hills Coulee Soil.

The L2 loess and associated Washtucna Soil were not found on EF. There is no evidence that OIS 2 outburst floods inundated the north end of EF that could have eroded the L2 and the Washtucna Soil. The absence of the L2 and the Washtucna Soil is most plausibly explained if deflation on EF during L2 accumulation predominated and the surface was never stabilized by vegetation long enough to result in loess accumulation and soil formation. Intervals of incipient soil formation apparently were never preserved in the sand sheet on EF because episodic eolian activity cannibalized soils by deflation and sand blowing.

The Unit 3 sand sheet is equivalent in age to the L2 loess because the sand sheet: (1) underlies the L1 loess and the Sand Hills Coulee Soil, (2) underlies OIS 2 slackwater sediments of Unit 5 at lower elevations on EF, and (3) overlies pre-OIS 2 outburst flood deposits (Gravel 2b, Unit 2). Additionally, the texture and mineralogy of the sand in the sand sheet is more mature than dune sand derived from OIS 2-age outburst flood sediment. Sands in the sand sheet have a lower percentage of basaltic grains and are better rounded, reflecting a longer duration of eolian activity. Dunes of EF contain more basaltic grains and the sand grains are subangular.

4.2.5. Unit 4: L2 loess

The L2 loess is preserved in the upland areas surrounding EF and documents a significant episode of landscape stability that followed its accumulation. The L2 was derived from the eolian reworking of fine-grained sediments of pre-OIS 2 floods (McDonald and Busacca, 1988, 1998). Luminescence ages suggest that most of the L2 accumulated between 60 and 45 ka and that lower rates of accumulation occurred between 40 and 18 ka (Richardson et al., 1999). The timing of slower accumulation rates corresponds to the formation of the Washtucna Soil at the top of the L2. At many sites in the Palouse, the Washtucna Soil is a single paleosol, but at sites such as CLY-1, KP-1, and CON-1, the Washtucna Soil occurs as two distinct paleosols (McDonald and Busacca, 1990). The geographic distribution of the bifurcated Washtucna Soil at sites KP-1 and CLY-1 is relatively close to EF, suggesting that EF contributed to a short-lived pulse of dust that interrupted soil formation. Trends in loess thickness between 40 and 15 ka demonstrate a region-wide decrease in dust production that has been linked to the presence of a glacial anticyclone (Sweeney et al., 2004).

4.2.6. Unit 5: OIS 2 glacial outburst flood sediment

OIS 2 glacial outburst floods significantly modified the upwind margin of EF, and soil features and tephrachronology have given insights into timing of floods and the onset of L1 loess accumulation. The flood sediments are geochemically similar to dune sand (Fig. 11; Table 2)
and loess (Sweeney et al., 2002). Flood sediment was distinguished from loess by the presence of granule-rich zones and ice-rafted debris (pebbles and cobbles). Root pores at the top of flood strata represent weak soil development under vegetated conditions between the deposition of flood sediment and the beginning of loess accumulation. The occurrence of Mt. St. Helens set So tephra on top of Unit 5 flood deposits suggests that few floods, if any, entered EF after 15 ka. The So–Sg tephra couplet bracketing flood deposits at TOU-2 indicates that several decades passed between eruptions (Clague et al., 2003).

4.2.7. Units 6 and 7: development of the post-OIS 2 eolian system

Surface samples of sediments on EF fine downwind, corresponding to the progression from sand dunes to loess (Fig. 15). Several factors have influenced the distribution of eolian sand (Unit 2) and loess (Unit 1) on EF including physical barriers such as incised river valleys and bioclimatic influences such as soil moisture and vegetation density, which are driven by climate (Fig. 3). The downwind migration of sand on the southeastern margin of EF was interrupted at least partially by the Touchet River and its tributaries (Fig. 2). Migrating dunes intersected these incised drainages resulting in the trapping of saltating sand grains that restricted farther downwind migration of sand (Sweeney et al., 2001). Tributary drainages entering the southwestern margin of EF are aligned parallel or oblique to prevailing winds and did not impede downwind dune migration.

Deposition of sand-rich loess in the lower 1.5 m of the L1 loess at CLY-1 (Fig. 4) and at other sites in the upland hills surrounding EF suggests that conditions were either windier and/or CLY-1 was more proximal to the migrating dune front immediately after the LGM. As effective moisture increased after the LGM, saltation may have been restricted to positions further upwind on the flat, an explanation supported by a decrease in sand content as the loess accumulated.

4.3. Loess thickness and morphology

L1 is at most 1.8 m thick on EF, whereas it is 4.5 m thick at CLY-1. The distinct change in the thickness of L1 loess over less than 1 km from EF to CLY-1 (Figs. 2 and 16) may have been caused by several factors, including vegetation density and topography. Relatively thin loess on EF may be a function of low trapping efficiency. The density of vegetation may never have been sufficiently high to inhibit saltating sand grains, resulting in thin, sand-rich loess accumulation. During times of heightened eolian activity, saltating sand on EF may have re-entrained silt particles resulting in only thin loess accumulation on EF itself. In contrast, the thick loess around CLY-1 is out of the main zone of eolian sand transport (Figs. 2 and 15). With finer soil textures affording a higher water holding capacity compared to the sediments at BAB-5, a higher density of vegetation probably existed at CLY-1 consisting of shrub steppe (O’Geen and Busacca, 2001; Blinnikov et al., 2002) making a better trap for dust.

Topographic influences, such as incised streams or the interaction of winds with escarpments, might also have influenced loess thickness. The Touchet River and its tributaries, which lie between EF and CLY-1, trapped migrating sand dunes and sand sheets (Fig. 3A), reducing the occurrence of saltating sand at CLY-1, promoting thick loess accumulation there (Sweeney et al., 2001). The very fine sand that dominates the basal part of the L1 could have reached CLY-1 through short-term suspension transport during a prolonged time of high-winds at the end of the LGM.

Loess accumulations such as those at CLY-1 and EF-12 are part of the Skyrocket Hills, an area downwind of a 30-m escarpment bordering EF and the Touchet River. The coarse texture and thickness of the loess in the Skyrocket Hills could also be explained by cliff-top deceleration and flow separation of winds downwind of the escarpment (Mason et al., 2003; Rawling et al., 2003).

The relatively thin, flat blanket of loess on EF itself contrasts with the thick accumulations of loess in hills with linear topography that surround the flat (Fig. 2). The origin of the linear loess ridges may be related to loess accumulation in the wind shadow of topographic obstacles, such as basalt knobs (Lewis, 1960). Compared to the linear loess ridges in the Central Lowlands region (known as paha) that are cored by till or linear sand dunes (Flemal et al., 1972; Ruhe, 1983), loess ridges in the Palouse are cored by older loess (Lewis, 1960). Their accumulation and growth during the Pleistocene and Holocene resulted in the formation of pseudoanticlinoles, illustrated at the CLY-1 site (Fig. 5). As envisioned here, pseudoanticlinoles develop as successive loess units mantle mounds of loess deposited in the wind shadow of knobs of underlying basalt bedrock. The thickness of loess units is relatively uniform from the crest of the ridges to the swales. The orientation of loess ridges is N25E to N30E, which is offset 20°–25° to the trends of modern sand dunes on EF (average N50E). Lewis (1960) and Wells (1983) speculated that the two orientations may reflect a shift in direction of the prevailing southwesterly winds from the onset of loess accumulation to modern dune formation. We hypothesize that the orientation of the ridges would not change once the core of the ridge was in place, since loess accumulates as a blanket of sediment and would conform to pre-existing topography. Higher soil moisture on the north-facing slopes of the basalt knobs may have allowed denser vegetation growth, promoting higher dust trapping efficiency and enhancing dust accumulation in the wind shadow. The linear ridges are limited to that part of the southwest Palouse where loess contains greater than 20% sand.
4.4. Bioclimatic model control on eolian sediment distribution

The bioclimatic model explains the modern distribution of eolian sediments on EF (Fig. 3B) and the observed sequence at BAB-5 (Fig. 9A). We interpret these deposits as reflecting environmental changes that began with arid conditions that promoted sand sheet development. The sand sheet stabilized as the density of vegetation increased. Denser vegetation and higher soil moisture prevented saltating sand from reaching BAB-5 and instead promoted loess accumulation there.

Fig. 16. Isopach maps depicting the thickness of L1 and L2 loess in southeast Washington, from Busacca and McDonald (1994). Contours of approximate loess thickness overlay maps showing present occurrence of loess and the Channeled Scabland. (A) L1 loess thickness, illustrating the thickest accumulations downwind of Eureka Flat and the Pasco Basin. (B) L2 loess thickness, illustrating a similar pattern in thickness to the L1. Note that for both loess units, a thick finger of loess extents downwind of Eureka Flat for nearly 150 km. P = Pullman, S = Spokane, star symbol = location of Eureka Flat.
Similar observations of a shifting boundary between saltation-dominated and suspension-dominated eolian processes have been described at the northern margin of the Chinese Loess Plateau (Ding et al., 1999; Sun et al., 1999). There, bodies of sand are interstratified with bodies of loess. Shifting climatic factors, principally the strength of the winter monsoon and accompanying changes in soil moisture and plant density, caused changes in the geographic boundary separating the active dunes of the desert basins and the vegetated landscapes of the Chinese Loess Plateau (Rokosch et al., 2003; Sun et al., 1999). Grain size distributions within the loess also reflect the proximity of loess to upwind sand dunes (Vandenberghhe and Nugteren, 2001; Ding et al., 1999).

4.5. Regional paleoclimate and Eureka Flat

Loess deposits are important geologic repositories of paleoclimate information, some of which is archived in the characteristics of pedogenic horizons that formed during the accumulation of loess (Kemp, 2001). The continuing dominance of eolian transport on EF and regional paleoclimate records from fossil pollen, opal phytoliths, and faunal burrows (Whitlock and Bartlein, 1997; Blinnikov et al., 2001; O’Geeen and Busacca, 2001) suggest that the climate has been arid to semi-arid on the CP since the LGM. The dominance of sagebrush (Artemisia) during the LGM over large areas of the CP versus pure bunchgrass steppe through the Holocene until the onset of farming suggests that cold and dry conditions prevailed during the LGM at many loess study sites (Blinnikov et al., 2002). As conditions became warmer and wetter at the start of the Holocene, shrubs were replaced by grassland at low elevations and by forests at high elevations (Whitlock and Bartlein, 1997; Blinnikov et al., 2001; O’Geeen and Busacca, 2001). Since the end of the LGM when the dominance of sagebrush on the CP waned, climate has fluctuated moderately, from wetter conditions just after the LGM, to a warm and dry interval in the mid Holocene, and back to more equitable conditions for the last 4 kyr (Whitlock et al., 2000; Blinnikov et al., 2002). Fluvial sediments, including those of the Columbia River, record similar climate fluctuations, as documented by alternating episodes of aggradation and incision (Chatters and Hoover, 1992). During arid times, flood plains tended not to aggrade but were sites of eolian deflation and sand dune formation (Chatters and Hoover, 1992).

Prior to the LGM, a relatively dry climate was required for the formation of the eolian sand sheet (Unit 3) that was responsible for the generation of the L2 loess (Unit 4). Since ~70 ka and throughout the LGM (during the onset of L2 loess deposition), opal phytoliths record drier conditions than at present across loess sites on the plateau (Blinnikov et al., 2002). The pollen record at Carp Lake (Fig. 1) suggests a similar climate, with dry conditions from 72 to 58 ka (Whitlock et al., 2000).

4.6. Eureka Flat: the engine of the Palouse

Regional thickness trends for the L1 and L2 loess units show that the thickest accumulations are immediately downwind of major flood slackwater sediment depocenters, principally EF (Busacca and McDonald, 1994) (Fig. 16). Loess isopach maps depict very similar distribution patterns for the L1 and L2, in particular, a finger of relatively thick loess extends in the downwind direction from EF to just south of Spokane, WA. Distribution of thick loess in this narrow geographic area can be explained by small variability in the direction of dust-transporting winds. Winds blowing nearly parallel to the source area should result in a narrow zone of loess deposition adjacent to and downwind of the source (Handy, 1976).

Loess thickness patterns suggest that EF has been a persistent source of dust for at least 70 ka. Although the total thickness of the loess cover is not known with great accuracy across the expanse of the CP, the thickest measured accumulation is in Whitman Co., WA (46°N50°, 117°W30°) at 76 m (Ringe, 1970). The thickest total accumulation of Palouse loess is aligned on the northeastward trajectory of dust derived from EF, suggesting that it has served as the major source of dust for the Palouse.

Magnetostatigraphy of the Palouse loess suggests a long history of loess activity. The Bruhnes–Matuyama magnetic reversal (~790 ka) occurs at 11 m in the 25 m thick section of loess at the WA-3 site (Fig. 1) ~25 km downwind of EF (Kukla and Opdyke, 1980; Busacca, 1991), suggesting that the lowest exposed loess at this site is at least as old as 1 Ma. Site DU-1 (Fig. 1) ~60 km downwind of EF records a normal-reverse-normal polarity zonation. Due to potential erosion by outburst floods and other events, it is unclear if the lower normal interval represents the Jamarillo or Olduvai magnetic polarity subchrons (0.90–0.97 Ma and 1.67–1.87 Ma, respectively). This led Busacca (1991) to speculate that the Palouse loess is as old as 2 Ma. The geographic position of WA-3 and DU-1 downwind of EF suggests that the source of dust for the older units may be from EF, assuming that the wind directions have been relatively unchanged over the past 2 Ma. Paleoind directions inferred from older loess units are difficult to demonstrate because of the lack of well-exposed or continuous outcrops.

The development of EF is complex, modified by several phases of catastrophic outburst flooding followed by eolian deflation and deposition. Deflation and flood erosion has prevented a long and complete record of eolian accumulation within EF itself. Fig. 17 summarizes the evolution of EF starting at 70 ka with the modification of floods during OIS 4 followed by eolian activity, OIS 2 floods, and recent eolian activity. This cycle has been driven by glacial–interglacial climate changes that controlled the timing of outburst floods, loess accumulation, and soil formation.
5. Conclusions

(1) EF is a deflational zone aligned with prevailing winds and has been a major source of dust for the Palouse loess. Sand dunes and sand sheets that have played key roles in the ejection of dust from a sand- and silt-rich sediment source document modern and ancient eolian activity.

(2) EF was a major depocenter for glacial outburst flood sediments for approximately 2 Ma. This periodic replenishment of sediment has been key in driving the eolian system throughout the Quaternary.

(3) Bioclimatic and topographic effects have controlled the distribution and thickness of loess in the EF area, and have provided evidence for paleoclimate fluctuations.
throughout the Quaternary (especially since the LGM) that can be paired with other paleoclimate proxies. The role of bioclimatic in eolian sediment distribution involves the oscillation of the dune-loess boundary through time as a function of climate change. Topographic trapping involves the physical segregation of eolian sand from transport, resulting in thick loess accumulations downwind of the “trap”.

(4) Future work and implications of this study include obtaining luminescence ages that span the L1 loess, which will provide insights into ages of soil formation, timing of loess accumulation, and calculation of mass accumulation rates that can be used in the generation of a regional atmospheric dust model.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found in the online version at doi:10.1016/j.quaint.2006.10.034.

References


