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Chapter 8

Aerial Field Guide

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There are two overflights planned for the field conference; one for the Cheney-Palouse tract of the eastern Channeled Scabland, the other covering the coulees and basins of the western region. The approximate flight lines are indicated on the accompanying LANDSAT images (Figures 8.1 and 8.2).

The first flight will leave Spokane International Airport; the approximate location of flood spillover from the Spokane River Valley into the Cheney-Palouse tract. The elevation here is about 175 meters above the floor of the present Spokane Valley. The flight will follow the eastern margin of this large scabland tract, passing a series of loess remnants, gravel bars and excavated rock basins. At the junction of the Palouse and Snake Rivers, one can observe the erosion of deep, joint-controlled canyons and the deposition of high-level gravel bars. The return flight will pass Devils Canyon, Washtucna and Staircase Rapids, a multi-level erosional cataract caused by a major flow constriction at the southern end of the Cheney-Palouse tract. Other excavated rock basins and streamlined loess remnants with associated gravel bars will be observed on the last part of the flight past Sprague Lake back to Spokane.

The western scablands overflight will provide a review of the structurally controlled complex pattern of large-scale erosion and deposition characteristic of the region between the upper Grand Coulee (Banks Lake) and the Pasco Basin. The flight will depart Grant County airport, cross the central Ephrata Fan and then head east along Crab Creek to Wilson Creek and Marlin. From here, it will turn west to Long Coulee and enter the Hartline Basin at the north flank of Pinto Ridge. The most dramatic erosional topography of the entire Channeled Scabland will be seen in this part of the flight across the Hartline Basin to Banks Lake and south across Dry Falls down the lower Grand Coulee to its efflux section at Soap Lake. The flight will then skirt the western margin of the Ephrata Fan and cut west, across the Beezly Hills Anticline, to the Three Devils Cataract in Moses Coule. The flight down Moses Coulee will show its well-developed inner channel segments and the complex bar at the junction with the Columbia River. Flying south along the Columbia River, one will see, on the left, the three major western spillways out of the Quincy Basin, Crater, Potholes and Frenchman Coulees. At Sentinel Gap, the river has developed a deep antecedent cut through the Saddle Mountains Anticline. South of this mountain ridge, one finds large flood gravel ripples and modern active sand dunes on the Wahluke Slope. From here, the flight will return to Grant Co. Airport via the large southeastern spillway out of the Quincy Basin (the Drumheller Channel system) and the active sand dune field of the region bordering the Potholes Reservoir.

Figure 8.1. LANDSAT image of the Cheney-Palouse scabland tract with the tentative flight line superimposed. NASA LANDSAT E-5 854-16504-5, 20 August 1977.

Figure 8.2. LANDSAT image of the western parts of the Channeled Scabland with the tentative flight line superimposed. NASA LANDSAT E-2 936-17451-5. 20 June 1976.

Chapter 9

The Touchet Beds of the Walla Walla Valley

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The Touchet beds, named by Flint (1938b) extend along the Columbia River and its tributaries in the region of the Pasco and Umatilla Basins in Washington and Oregon (Fig. 9.1). The type locality for the Touchet beds is the vicinity of Touchet, Washington in the Walla Walla Valley (Flint, 1938b). These deposits were also described by Bretz (1928c, 1929, 1930b), Allison (1933), Lupher (1944) and Baker (1973a). The Touchet beds were deposited in Lake Lewis; the history of the idea of a Pleistocene lake east of the Cascades was summarized by Bretz (1919) and Flint (1938a).

The Touchet beds are as much as 100 m thick (Jones and Deacon, 1966) in the centers of valleys and basins but taper to a thin veneer as they rise to a maximum elevation of 350 m (1150 feet) (Flint, 1938b). The Touchet deposits in the Walla Walla Valley generally grade from coarser to finer both in an upvalley (eastward) direction and from lower to higher elevation (Bretz, 1928c, 1929; Newcomb, 1965). The principal constituents of the Touchet beds are buff to cream-colored clay, slit and sand with gravel ranging in size from small pebbles to large boulders. The mineral composition of the fine-grained portion of the sediments was estimated by P. D. Krynine (in Flint, 1938b, p. 494) to be:

1. Fine-grained fresh rock material (rock flour?) 65-80%
2. Yellow colloidal aggregates and iron-stained mineral grains (reworked Palouse loess?) 20-35%
3. Volcanic ash 1-5%
4. Diatoms .... Trace

Much of the coarser sand and the gravel are derived from the Columbia Plateau basalts, whereas granitic and metamorphic erratics may have been ice-rafted from bedrock sources to the north and east (Bretz, 1928c).

The bedding of the Touchet sediments varies from massive to well-stratified. East of Reese (between Touchet and Wallula Gap on Fig. 9.1), the strata are graded and vary from 0.1 to 2 m thick, individual strata being thinner toward the tops of exposures. Individual rhythmites or turbidites (Baker, 1973a) are separated by disconformities and grade from a basal coarse sand upward to silt and clay (Fig. 9.2). Just above some disconformities are erratics (Fig. 9.3) as well as blocks of sediment (Fig. 9.4) which may have been frozen during transport.

The Touchet beds are slack-water sediments associated with the Missoula Flood(s). At least two major floods and perhaps several minor floods are recorded in eastern Washington (Brown, 1973; Baker and Patton, 1976; Baker, Ch. 2, this volume). Mullineaux and others (1977), working with Mount St. Helens tephra, demonstrated that the last major flood occurred about 13,000 years ago. In the Walla Walla Valley above the Touchet beds, there are patches of Mazama ash (Brown, 1971) and wind-blown silt ("post-scabland eolian deposits" of Flint, 1938a).

The origin of the rhythmites within the Touchet beds is problematic. Surely each of the more than 50 graded beds (Fig. 9.5) exposed near Lowden (between Touchet and Walla Walla on Fig. 9.1) cannot represent separate floods. Bretz (1929, p. 539) asked: "Is not some kind

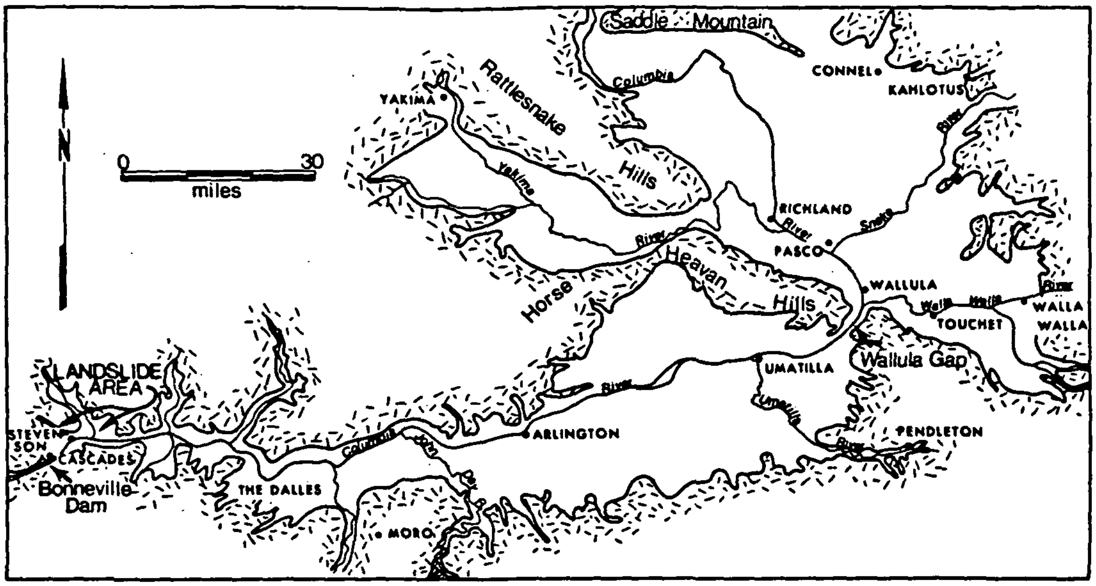

Figure 9.1. Approximate areal extent of Touchet beds and erratics. Areas of modern landslides in the Columbia Gorge are indicated. From Flint (1938b). Touchet-like valley deposits also extend up the Snake River and its tributaries to above Lewiston, Idaho. See Bretz (1929) and Flint (1938b, p. 507-510.)

Figure 9.2. Graded beds separated by disconformities (Burlingame Canyon, 4 km south of Lowden). Each rhythmite here grades from stratified sand at the base to massive silt and clay at the top. Note the thin sinuous clastic dike.

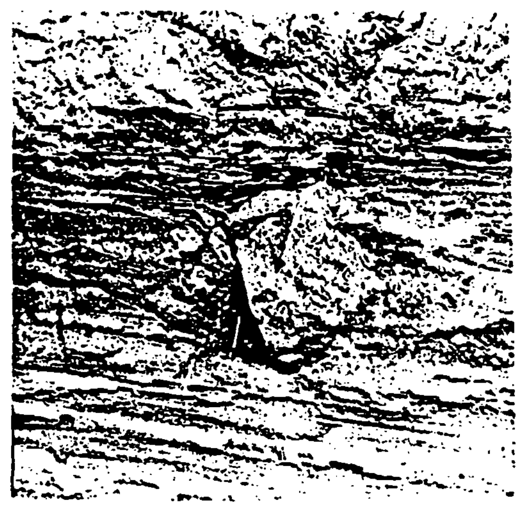

Figure 9.3. Granite erratic within the Touchet beds Burlingame Canyon). Stratified sand is at the base of the picture with massive silt and clay at the top.

of rhythmic pulsation in water supply and sediment supply recorded in the irregular and undulatory strata?" Lupher (1944, p. 1441) wrote, "The common alternation of stratified sand and massive silt layers indicates a periodic advance and retreat of the lake shore." The rhythmites probably record separate pulses of a flood or repeated surges or waves of water up the Walla

Figure 9.4. Sand block above silt block at same horizon as the granitic erratic in Fig. 9.3. The blocks may have beeen frozen during transport.

Walla Valley. Possible mechanisms to explain the graded beds include the following:

1. Variation in discharge at the ice dam as glacial Lake Missoula drained. This could be due to (a) gradual break-up of the ice dam, releasing water at a variable discharge, or (b) *periodic floating* of a diminishing ice dam by the glacial lake which drains beneath the ice (floating an ice dam is mentioned in Baker, 1973a).
2. Pulsations in the floodwater reaching Lake Lewis, due to (a) different lengths and sizes of the various scabland channels allowing portions of the flood to arrive at the lake at different times, (b) landslides, ice blocks, or gravel bars interrupting the waterflow in some of the scabland channels, or (c) surges of water from sudden channel deepening or the break-through of a cataract upstream from a scabland tributary (Baker, 1973a, p. 47).
3. Pulsations of floodwater between the major part of Lake Lewis in the Pasco Basin and the eastern arm of the lake extending up the Walla Walla Valley. At Reese (between Touchet and Wallula Gap, Fig. 9.1), there is a trench eroded into a basalt barrier (Bretz, 1929). The pulsations could reflect (a) seasonal precipitation which washed silt and sand into the lingering pond in the Walla Walla Valley (Bretz, 1929, p. 534-535), while the outlet of the pond was blocked by silt at Reese, or (b) ice jams, gravel bars, or possibly landslides which may have occurred near Reese could have varied the exchange of water through the pass.
4. Variations in flow rate at the dam sites (Wallula Gap and/or the Columbia River Gorge) of Lake Lewis. Such fluctuations could be envisioned as a result of (a) the break-up or floating of a Cascade glacier forming a dam at the Dalles (Russell, 1893; Flint, 1938b), (b) ice jams in the Columbia River Gorge (Allison, 1933, p. 719-721), or (c) landslides caused by flood erosion. Holocene slides in the Columbia Gorge are well known (Allison, 1933; Flint, 1938b), the evidence of the ones which temporarily dammed Lake Lewis could have been removed. In addition to ice jams, glaciers and landslides, damming of Lake Lewis could also have been controlled by lava flows (Allison, 1933), warping or faulting (Flint, 1938b) and hydraulic damming (Bretz, 1925; 1969; Baker, 1973a).

Another problematic feature of the Touchet beds are the clastic dikes (Figs. 9.2 and 9.5) which have been discussed by Jenkins (1925), Flint (1938b), Lupher (1944), and Newcomb (1962). The clastic dikes form a polygonal network according to Newcomb (1962) and have variable dips. The dikes vary in thickness from 1 mm to a few meters. Most of the large dikes tend toward the vertical (but are not planar). The dike sediment is as variable in grain size as the Touchet beds. In many cases, the dikes consist of vertical laminae of alternating silt and sand (Fig. 9.6). The clastic dikes may be related to the warping, folding and faulting of the Touchet beds. In many cases, the dikes consist of vertical laminae of alternating silt and sand (Fig. 9.6). The clastic dike may be related to the warping, folding and faulting of the Touchet beds (Flint, 1938b).

Below follows a summary of the many proposed origins for the clastic dikes.

1. The fissures which the dikes fill were produced by earthquakes (Jenkins, 1925; Jones and Deacon, 1966). Historic earth-

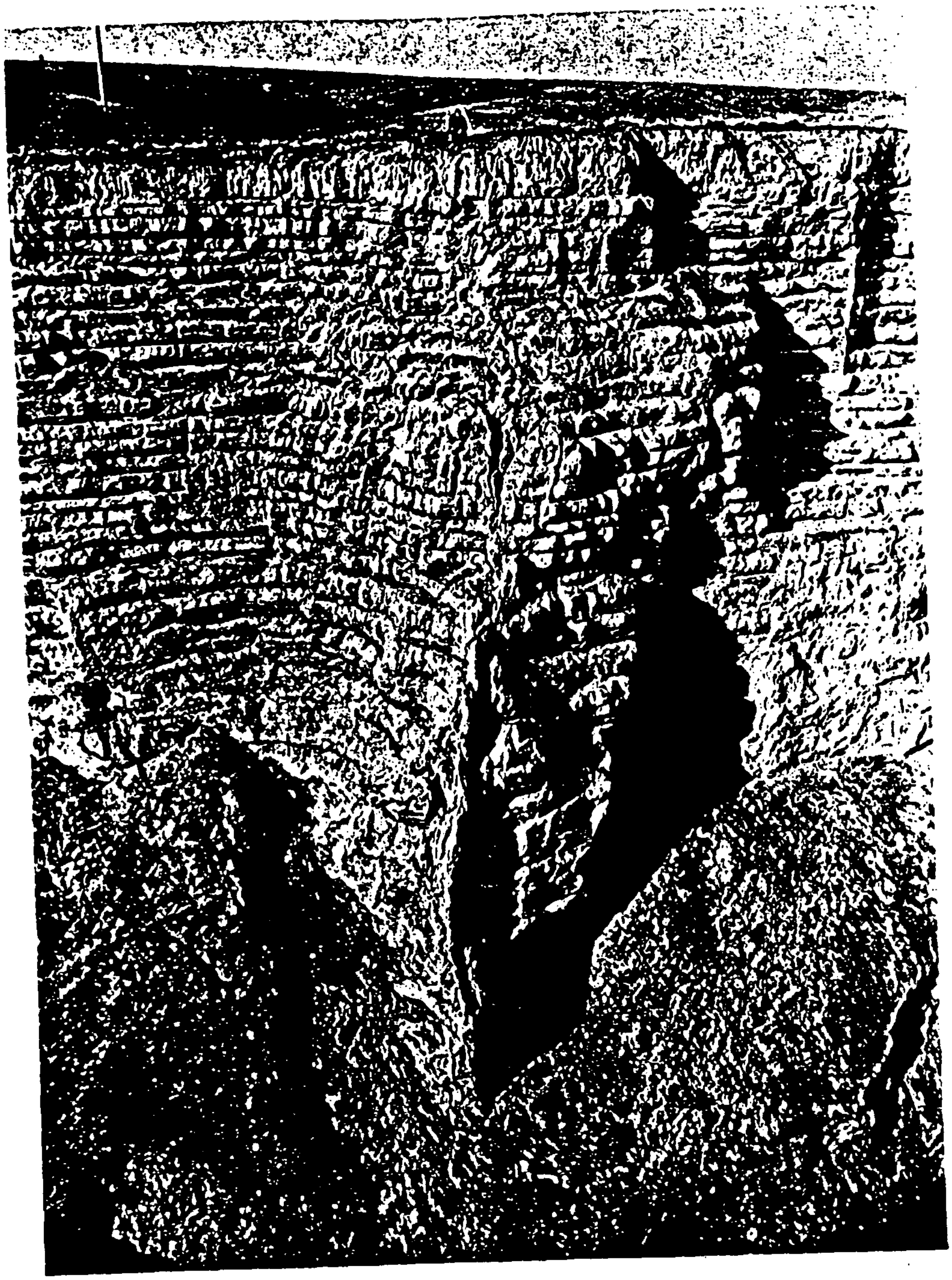

Figure 9.5. Rhythmites and clastic dikes in Burlingame Canyon. The clastic dike in the center of the photograph thins upward, whereas the dike in the upper right thins downward.

quakes in the Walla Walla Valley have not been uncommon (Brown, 1937; Jones and Deacon, 1966). The 1964 Alaskan earthquake produced clastic dikes with a polygonal pattern in saturated alluvium (Jones and Deacon, 1966).

2. The fissure development was caused by "(a) uneven settling and cracking through melting of buried ice, (b) gravity sliding and faulting on inclined zones of subsurface melting, (c) formation of cavities where ice blocks and layers melted" (Lupher, 1944, p. 1432).
3. A process of fissure development recognized in the Lewiston Basin was "erosion by underground streams" (Lupher, 1944, p. 1432).
4. Another dike-producing process noted in the Lewiston Basin, as well as eight kilometers northwest of Touchet, was "faulting and fissuring by landslides in the Columbia River basalts" (Lupher, 1944, p. 1432).
5. The dikes "represent fillings of permafrost-related crevices" (Alwin and Scott, 1970, p. 58).
6. During dessication of the Touchet beds, the contraction resulted in fractures.
7. The dikes resulted from repeated injections of groundwater, each hydraulic injection caused by a large lowering of Lake Lewis (Newcomb, 1962; Jones and Deacon, 1966).
8. The rapid deposition of a turbidite made the underlying sediments unstable. Slumping was accompanied by extension fractures (Baker, 1973a).

Jenkins (1925) preferred downward filling when the cracks opened beneath the mud of a lake bottom. Lupher (1944) also believed that the fissures filled from above, mostly by streams, lake currents, and waves. Baker (1973a) implies that the extension fractures were filled from above by later turbidity flows. Alwin and Scott (1970, p. 58) stated that "primary structures indicate a downward filling of the dikes by sand and silt." Newcomb (1962) believes in upward injections. In the Walla Walla Valley, individual dikes thin upward and/or downward (Fig. 9.5), suggesting that a combination of processes may be responsible for the clastic dikes.

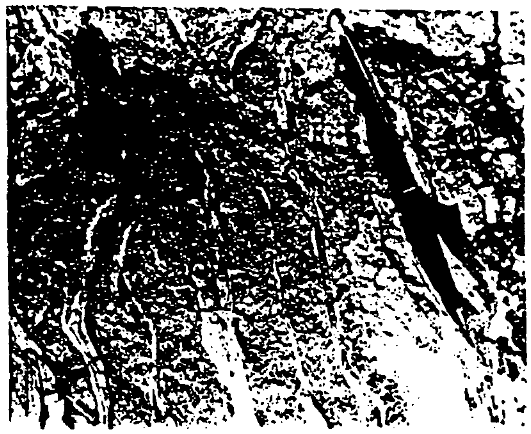

Figure 9.6. Portion of clastic dike (exposed in Burlingame Canyon) exhibiting vertical laminae of alternating silt and sand. Dike sediments are of the same general composition as the surrounding Touchet beds.