Quaternary geology and ecology of the greater Yellowstone area

Authors: **Kenneth L. Pierce,** *U.S. Geological Survey, Bozeman, MT 5971*
Don G. Despain, *U.S. Geological Survey, Bozeman, MT 5971*
Cathy Whitlock, *Univ. of Oregon, Corvallis, OR, 97403–1251*
Kenneth P. Cannon, *National Park Service, Lincoln, NE, 68508–3873*
Grant Meyer, *Univ. of New Mexico, Albuquerque, NM, 87131–1116*
Lisa Morgan, *U.S. Geological Survey, Denver, CO 80225*

INTRODUCTION

This field guide focuses on the glacial geology, ecology, paleoecology, caldera unrest, and archeology in Yellowstone and Grand Teton national parks and vicinity (Fig. 1). Some previous field guides of Yellowstone are Locke et al. (1995) for the Yellowstone valley, Fournier et al. (1994) for hydrothermal and volcanic geology of Yellowstone, and Pierce and Good (1992) for the Quaternary of Jackson Hole. Non–technical overviews of Yellowstone and Grand Teton National Parks are Good and Pierce (1996) and Smith and Siegel (2000). Geologic maps are: Grand Teton (Love et al., 1992), and Yellowstone (bedrock and surficial geology (USGS, 1972a; 1972b). Christiansen (2001) extensively describes Yellowstone's volcanic geology, and Pierce (1979) describes the glacial geology of the northern Yellowstone region. We suggest that you obtain detailed maps.

GLACIAL AND ARCHEOLOGY OVERVIEW

In the Rocky Mts., Bull Lake terminal moraines are typically farther downvalley than Pinedale moraines, which are generally correlated with marine oxygen isotope stages (OIS) 6 and 2 respectively. However, for the Greater Yellowstone system, Bull Lake terminal moraines to the west, southwest, and south of Yellowstone are much farther downvalley than Pinedale moraines, whereas to the north in the Yellowstone valley and to the east, Pinedale glaciers have overridden Bull Lake terminal moraines.

Pleistocene glacial flow patterns for Yellowstone were complex. Glaciation was initiated in the mountains that surround the Yellowstone Plateau. Glaciers then built up on the plateau to about 3,000 ft (~1 km) thickness and flowed outward, down the major valleys that drain Yellowstone. During deglaciation, the plateau icecap stagnated, and glaciers from the adjacent mountains flowed into terrain previously occupied by the plateau icecap (Pierce, 1979). This pattern of buildup, full glacial conditions, and recessional changes resulted in shifts by up to 180º in direction of glacial flow and the transport of glacial erratics. Changing flow patterns sometimes dammed glacial lakes, including lakes on the Yellowstone Plateau. Outlet glaciers, augmented by ice from the Yellowstone Plateau, probably culminated later than typical mountain–valley glaciers, apparently because of both the interval needed to build up the plateau icecap and its self–amplifying nature once established.

Under full glacial conditions, the northern Yellowstone outlet glacier was formed by convergence from multiple sources in the northern Yellowstone area. Sources, from east to west, were (1) glaciers from the Beartooth uplift, and the Absaroka Range, (2) ice flowing over the Washburn Range from the eastern Yellowstone Plateau, (3) ice flowing from the Yellowstone Plateau between the Washburn and Gallatin Range, and (4) ice from the Gallatin Range (Fig. 4A). From the northern park boundary, this outlet glacier flowed 40 miles (65 km) downvalley to the Eightmile terminal moraines.

Interest in the archeological record of Paradise Valley extends back to the 19th century with the description of rock alignments, bison traps, and quarry sites, which probably represent the earliest archeological work in Montana (Norris, 1880; Brackett, 1893). Barnum Brown (1932) conducted the first investigations at the Emigrant bison drives in 1932. At Rigler Bluffs (24PA401; Haines 1966) fired–rock hearths eroding from a forty–foot clay bluff dated 5.6 cal ka B.P. (4900 ± 300 ^{14}C yrs. B.P. and 5.8 cal ka B.P. (5040 ± 150 ^{14}C yrs. B.P.). Good (1964) interpreted the overlying fine silts as having been deposited in a lake dammed by the Yankee Jim landslide.

George Arthur (1966a) recorded 47 sites and classified them as occupations, bison drives (jumps and pounds), wickiups, stone alignments, stone circles, pictographs, burials, and stone piles. The most impressive of these sites is the late Holocene complex of drive lines and bison kills associated with the glaciated basalt terraces near Emigrant. The Carbella site (24PA302) is located at the mouth of Yankee Jim Canyon. Arthur (1966a) noted that the artifacts were recovered from underneath and within large glacial boulders that may be the result of periodic or seasonal flooding or related to the rupture of the Yankee Jim dam. A middle Holocene age is suggested by the projectile point styles. The Eagle Creek site (24PA301) is located above the town of Gardiner within alluvial deposits. A ^{14}C date of 1200 cal yrs. B.P. (1230 ± 160 ^{14}C yrs. B.P.) was obtained from a fired rock hearth in tan, sandy loam (Arthur 1966b). Work by Arthur and others indicates a rather extensive archeological record in Paradise Valley that probably includes Paleo–indian occupation in latest Pleistocene time

Pierce, K.L., Despain, D.,Whitlock, C., Cannon, K.P., Meyer, G., and Morgan, L., 2003, Quaternary geology and ecology of the greater Yellowstone area: in Easterbrook, D.J., ed., Quaternary Geology of the United States, INQUA 2003 Field Guide Volume, [Desert Research Institute], p. XXX–XXX

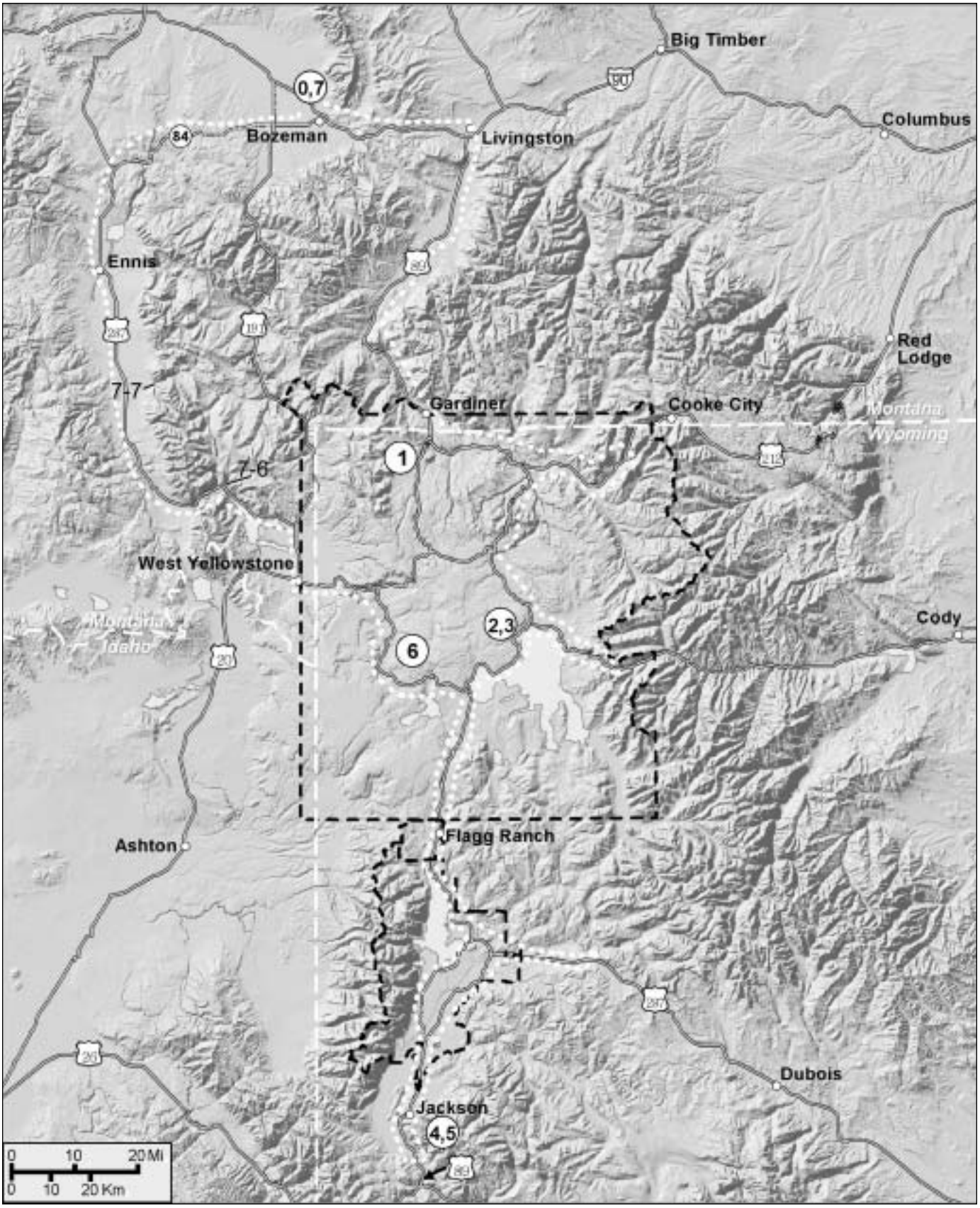


Figure 1. Shaded relief map of the Greater Yellowstone area showing the route of the Greater Yellowstone area field trip (white dots) and the place of at end of each day (number in white circle).

Figure 2. Map of the northern Yellowstone area showing location of stops on Day 1 and 2. C– Chico moraines, E–Eightmile moraines, D–Deckard Flats moraines, JB–Junction Butte moraines.

DAY 1. BOZEMAN TO MAMMOTH

STOP 1–1. MALLARDS REST FISHING ACCESS

West of the Yellowstone River, the outwash fan of the Northern Yellowstone outlet glacier emanates from the Eightmile moraines (Figs. 3, 4). This outwash fan is 200 ft (60 m) above the Yellowstone River at the Eightmile terminal moraines, 60 feet (18 m) above the river at this stop, and only 10–20 feet (3–6 m) above the river this side of the Livingston canyon. On the east side of the valley is the Beartooth uplift that supported local valley glaciers (Montagne and Locke, 1989), including Pine Creek with

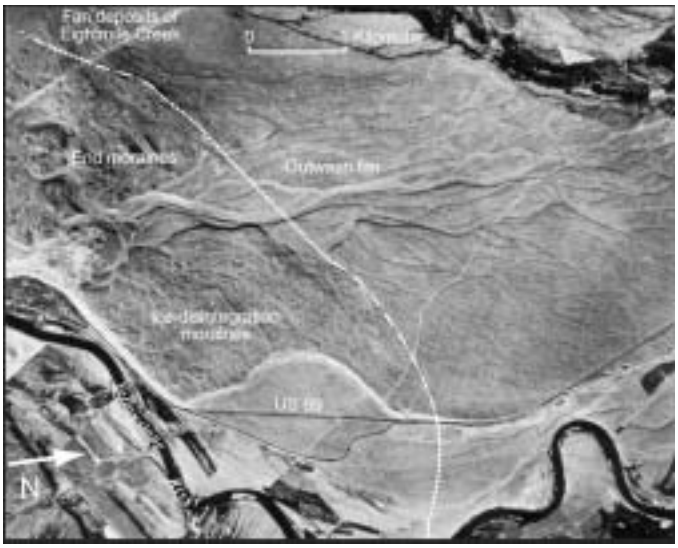


Figure 3. Pinedale end moraines and outwash fan of the northern Yellowstone outlet glacier, showing preservation of detailed braided channel pattern (from Pierce, 1979, Fig.16). The head of the fan is 200 ft above the Yellowstone River whereas it converges with the river level about 15 miles to the north (right).

Pinedale (~20 ka) and Bull Lake (~140 ka) moraines. The scarp of an active fault with about 15 ft (5 m) of post–glacial offset occurs near the base of the range at sites such as Barney Creek.

The modern Yellowstone River is eroding its banks downstream in the Livingston area, but upstream from here, it is remarkably stable. The history of glaciation and glacial floods exert important geomorphic controls on channel pattern and type. These, in turn, largely control the distribution of the three main alluvial types from here upstream to Gardiner, MT: (1) moderately to very stable, (2) incised to entrenched, and (3) single pool–riffle. Plane–bed channels form most of the channel. Several disturbance/sedimentation zones occur with multiple channels that range from anastomosing to anabranching. Downstream from Mallards Rest, multiple channels predominate and the Yellowstone River is a classic high–gradient, wandering, gravel–bed

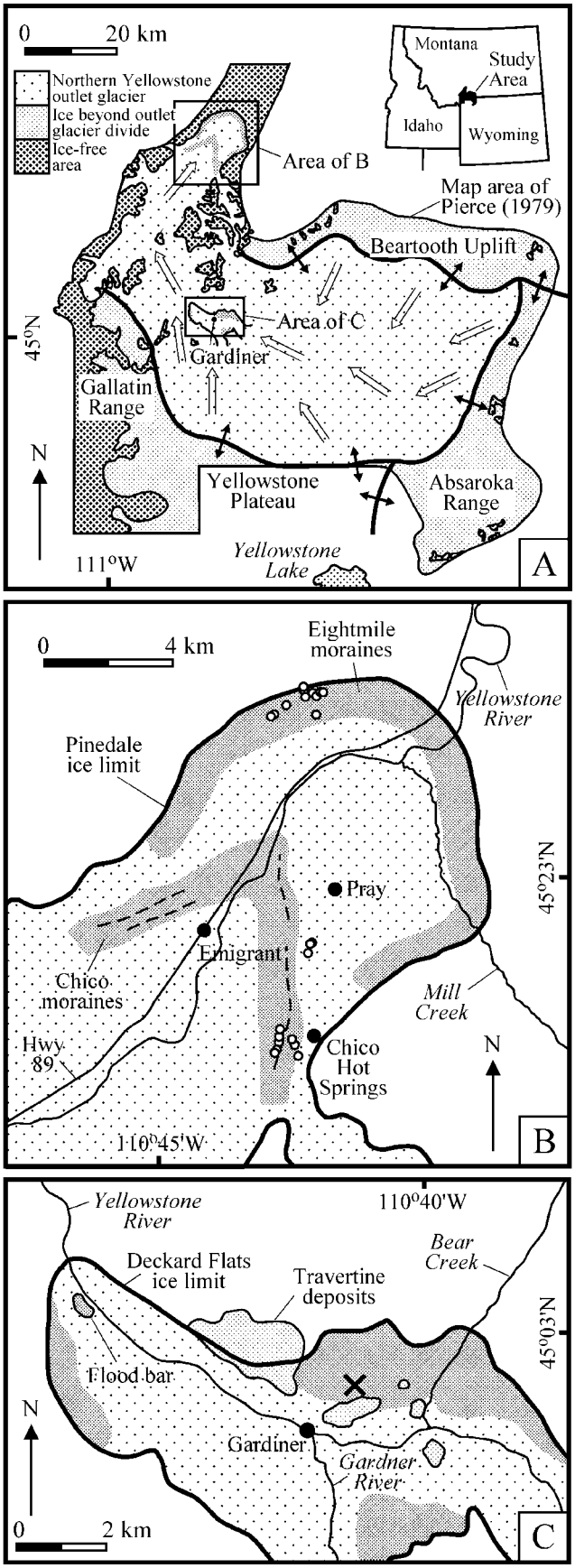


Figure 4. Pinedale glaciation of the in northern Yellowstone area (from Licciardi et al., 2001; adapted from Pierce, 1979). **A**: Northern Yellowstone ice cap. Thick black lines with double–pointed arrows indicate main ice divides of various ice masses that fed northern Yellowstone outlet glacier. Open arrows indicate flow directions of major ice drainageways. **B**: Sample sites (open circles) on Eightmile and Chico moraines. Dashed lines show prominent crests in Chico moraine complex. Three sampled boulders located 2 km south–southwest of the town of Pray are considered to belong to proximal part of Eightmile moraines. **C**: Sample sites on Deckard Flats moraines and late–glacial flood deposit. Individual sample sites on Deckard Flats moraines (dark stipple) are clustered in location marked ×, and sites on flood bar are clustered within outline of deposit. Travertine deposits are those used by Sturchio et al. (1994) to construct their U–series chronology.

river. In response to the floods of 1996 and 1997, the channel changed downstream from Mallards Rest, but changed little upstream. These contrasts have important management implications (River activity from Chuck Dalby, written communication 2002). Newly created gravel bars provide fresh substrates for primary plant succession, particularly by willow (*Salix exigua*) and cottonwood (*Populus angustifolia* and *P. trichocarpa*).

STOP 1–2. CHICO MORAINES

Just before Chico Hot Springs, turn right and climb up onto Chico outwash terrace and then Chico moraines. The Chico moraines (Fig. 4) of the Northern Yellowstone outlet glacier are bounded on the north by the Chico meltwater channel. Note the succession of outwash terraces that line the meltwater channel now occupied by the Chico road. Cosmogenic ages on boulders on the Chico moraines average 15.7 ± 0.5 ^{10}Be ka. Eightmile moraines average 16.5 ± 0.4 ^{3}He ka and 16.2 ± 0.3 ^{10}Be ka (Fig. 5) (Licciardi et al., 2001). These cosmogenic ages are younger than those based on ^{14}C dating (Licciardi et al., 2001: Pierce, 2003).

The Eightmile moraines (Figs. 3, 4) are well expressed just to the north on the other side of the Chico meltwater channel, where they arc around to their terminus 4.5 miles (7 km) beyond the Chico moraines. Across the valley, ice–marginal channels 150 ft (50 m) deep were eroded into bedrock at the Eightmile and Chico ice margins. To fit the Bull Lake–Pinedale paradigm, Horberg (1940) thought irregular stony deposits in the inner valley of the Yellowstone demonstrated a Pinedale age for the Chico moraines in contrast to the Eightmile moraines on the terrace seen at STOP 1–1, which he considered Bull Lake. But this inner valley deposit is more likely flood gravel rather than a morainal deposit because of the pattern of highs and lows and the gravelly, but not boulder studded, excessively drained highs.

STOP 1–3. HIGHWAY REST AREA

The low "terraces" of the Yellowstone River are actually flood deposits. In contrast to normal, flat–topped, terraces, the surface of these flood deposits has spoon–shaped forms: the lows are concave–up spoon–forms and highs are convex up spoon–forms.

Across the Yellowstone River is the basalt of Hepburns Mesa, whose age of 2.2 Ma (Smith et al., 1995) is consistent with the start of the Yellowstone volcanic field. The surface of this basalt is striated and polished. Locally, it forms cliffs that were used by Native Americans as buffalo jumps to procure bison.

STOP 1–4. FLOOD BAR, STEVENS CREEK

Cross Yellowstone River at Corwin Springs and continue just into Yellowstone Park. This mid–channel flood bar (Fig. 4C) has "giant ripples" spaced about 15 m apart and up to 2 m high, with boulders up to 2 m in diameter. Cosmogenic ages on the boulders average 13.7 ± 0.5 ^{10}Be ka (Fig. 5). The alluvial fan of Reese Creek 0.5 km to the northwest provides evidence of at least two

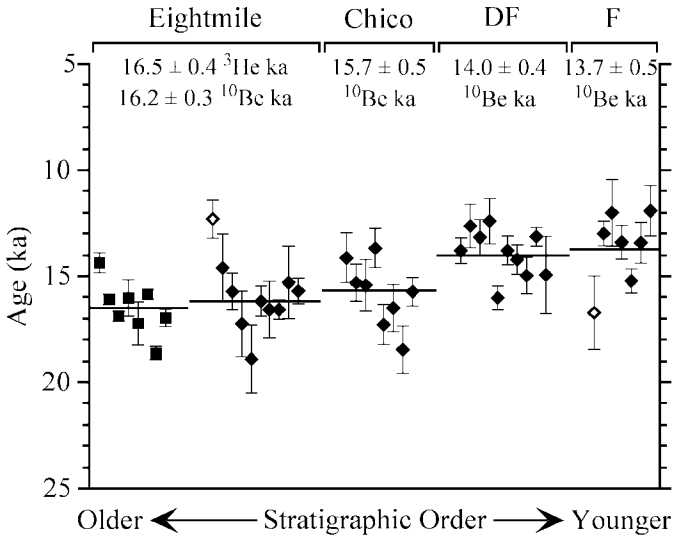


Figure 5. Cosmogenic ages of Yellowstone moraines and flood deposits, plotted in stratigraphic order from oldest (left side of graph) to youngest (right side) (from Licciardi et al., 2001). Solid squares are ^{3}He ages, and solid diamonds are ^{10}Be ages. Error bars on each age represent 1σ analytical uncertainty only, and do not include errors due to production rate, scaling, and other uncertainties. Horizontal lines and quoted ages indicate weighted means of each landform. Open diamonds indicate two outliers not included in weighted means. DF – Deckard Flats moraines; F – late–glacial flood deposits.

floods 150–200 ft (45–60 m) deep. First, an alluvial fan was deposited and then a flood down the Yellowstone valley eroded the fan front and left a "flood–ripped" fan front. The younger fan was then built and, later, another flood undercut and eroded the fan front (Pierce, 1979, Fig. 45). These floods were probably from release of glacially dammed lakes upstream, most likely from a lake dammed in the Lamar Valley by the Slough Creek glacier (STOP 2–3).

STOP 1–5. DECKARD FLATS MORAINES; TRAVERTINE BENCH; COSMOGENIC AGES

The 0.6 Ma Undine Falls basalt underlies the travertine bench and Deckard Flats bench (Christiansen, 2001) and this and other volcanic units date the incision of the Yellowstone valley (Fig. 6) The type area of the Deckard Flats moraines is south across Bear Creek. Large boulders in the moraines of the Deckard Flats 'readjustment' date 14.0 ± 0.4 ^{10}Be ka (Fig. 5). The Deckard Flats position is termed a 'readjustment' rather than an advance because it represents the time when the icecap on the Yellowstone Plateau ceased contributing to the northern Yellowstone outlet glacier, and the glaciers from adjacent mountains (Beartooth uplift, Absaroka Range, and Gallatin Range) readjusted to this loss and established a stable ice margin that can be traced over much of the northern Yellowstone area. In full–glacial time, the outlet glacier was about 3,500 ft (1 km) thick at Gardiner and terminated about 40 miles (65 km) downvalley,

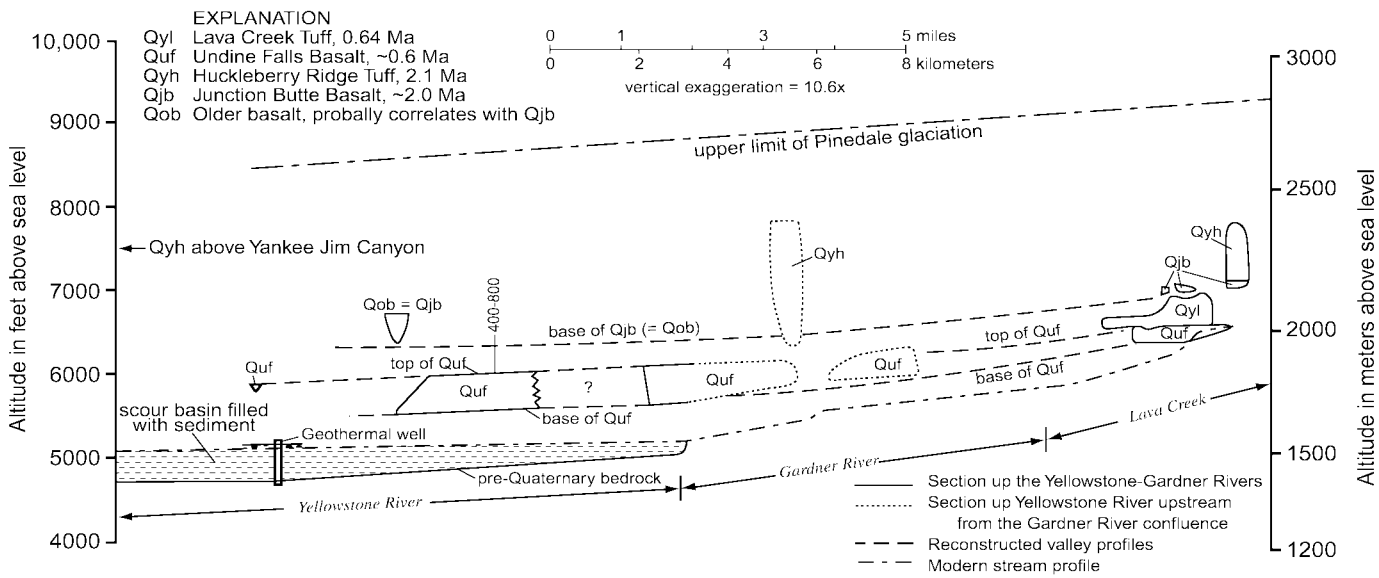


Figure 6. Valley deepening in Gardiner–Mammoth area based on incision of dated volcanic units (from Pierce et al., 1991). K–Ar ages from Obradovich (1992).

whereas during the Deckard Flats, it was about 900 ft (275 m) thick and terminated 4 miles (7 km) downvalley.

STOP 1–6. MAMMOTH OVERLOOK

The bench north of the Mammoth Hotel marks the Deckard Flats margin of glaciers originating to the southwest in the Gallatin Range. The bench is probably kame gravel and includes boulders of travertine. The conical hills to the south and east are thought to be thermal kames formed where hot springs melted cavities in the ice that filled with kame gravel. To the north above the Deckard Flats bench, a grassy, rounded hill of sub–glacial till also bears erratics of travertine. North in the Yellowstone valley near Gardiner, the full–glacial northern Yellowstone outlet glacier was about 3,500 ft (1 km) thick and created glacial streaming from northward flow. Under full–glacial conditions, ice from three points of the compass converged in this area to form the northern Yellowstone outlet glacier. Mt. Everts to the east exposes readily erodible, Cretaceous shale and sandstone on its west face. The southern part of Mt Everts is capped by 2.1–Ma Huckleberry Ridge tuff, with local, small, valley fills of 0.64–Ma Lava Creek tuff. The upland on top of Mt. Everts is glacially scoured with ridges and lakes forming a diverse wildlife habitat of lush meadows bordered by bands of forest.

Hydrothermal activity at the Mammoth terraces changes frequently. The following list shows the sequence of travertine deposits in the Mammoth area: (1) Terrace Mountain, the highest travertine near Mammoth Hot Springs, rests on 2.1–Ma Huckleberry Ridge tuff and has a U–Th age of 406 ± 30 ka (Pierce et al., 1991). Pinedale glacial erratics rest on this travertine, which was scoured by northward–flowing ice that reached an altitude of almost 9,000 feet (2750 m) on Sepulcher Mountain, (2) Above the active terraces, the inactive and forested Pinyon terrace is post–glacial with U–Th ages near 10 ka (Pierce et al., 1991). (3) The lowest travertine terraces are post–glacial. A deep research hole was drilled in the 1960's in the main terrace and a U–Th age of 7.72 ± 0.88 ka was obtained on travertine from a depth of 239 ft (72.8 m). We encourage you to tour the Mammoth Terraces in evening.

DAY 2. MAMMOTH TO LAKE VILLAGE

Vegetation and 1988 Fires

From the Mammoth area eastward to the Lamar Valley, we will look at recessional deposits of the last glaciation. The major mountain vegetation zones in this part of the Rocky Mts. are diagrammed in Fig. 7. Lower timberline is at about 6400 ft (1950 m) and upper timberline very close to 10,000 ft (3050 m). Below the forested zone, vegetation is dominated by semiarid sagebrush shrublands with basin big sage (*Artemisia tridentata* spp *tridentata)* and bluebunch wheatgrass (*Agropyron spicatum*). Alpine tundra above upper timberline has a large number of plant communities populated by species from the arctic tundra and species that have evolved from subalpine floras. Between these extremes are two forested zones: (1) the lower, montane zone, ca. 6400–7600 ft (1950–2300 m), dominated by Douglas–fir (*Pseudotsuga menziesii*) with occasional stands of limber pine (*Pinus flexilis*) and aspen (*Populus tremuloides*), and (2) the upper, subalpine zone, 7600–10,000 ft (2300–3050 m). The climax forest is dominated by Engelmann spruce (*Picea engelmannii*) and subalpine fir (*Abies lasiocarpa*) with lodgepole pine

(*Pinus contorta*) as the pioneer species in forest successions. Above 8600 ft. (2600 m), whitebark pine (*Pinus albicaulis*) is the dominant pioneer species but also forms climax stands on sites that are too cold and dry for spruce and fir. Areas underlain by fine soils are generally covered with shrublands or grasslands that provide most of the forage to support populations of large mammals. Those in the montane zone provide winter range and those in the subalpine and alpine zones provide summer range.

In 1988, extreme drought and high winds from mid–June to mid–September resulted in forest fires that eventually burned approximately 1.1 million acres (450,000 ha) in and around the park (Fig. 8). The total area encompassed by the fire was 1.7 million acres (690,000 ha) (Fig. 8). About 50% of the burned area was burned by crown fire, 40% by mixed–severity fire, and 10% by grassland/shrubland fires. Most of the sagebrush areas visible to the west of this stop were burned in the fires and burned forest is visible on the mountains forming the horizon to the south. Areas that were previously covered by lodgepole pine forests are now covered by lodgepole pine seedlings 4–8 ft (1.5–2.5 m) tall. Where Douglas–fir or spruce/fir dominated the forest stands the seedlings are sparser. Geological influences are also apparent in the vegetation response. The good soils developed from andesitic and sedimentary substrates have produced herbaceous growth and sparse tree seedlings while the poor soils derived from rhyolitic substrates are covered by dense lodgepole pine seedlings

The burned areas seen here are part of the North Fork fire, which started July 22, 1988 just west of the park boundary, due west of Old Faithful. It was immediately reported and crews were dispatched to it. High winds made it impossible to contain. It burned 111 acres (45 ha) the first day, 922 ac (373 ha) the second day, and 1317 ac (533 ha) the third day. The overhead team in charge of the fire requested bulldozers and the park told them that they could use them if they would be effective. The overhead team decided that the long distance spotting (fire brands carried by winds) and high spread rate at the fire head would not be contained by bulldozers, and that also made having men on the fire lines at night too dangerous. For the next 57 days, for only five days it did not increase in size, for only 15 days it gained less than 1000 ac (400 ha), for 28 days between 1000 and 5000 acres (400–2000 ha) burned, and for 9 days more than 5000 acres (2000 ha) burned. On 10 September, fires burned 60,000 acres (24,300 ha). The fires overran the developments at Old Faithful, Madison Junction, and Norris Junction, and threatened Mammoth Hot Springs, West Yellowstone, and Canyon Village. Fires eventually burned over 215,000 ac (87,000 ha).

Pull over to view: (1) High Bridge moraines deposited during the recession from the Deckard Flats position, (2) Lava Creek, which is the type area of the 0.64 Ma Lava Creek Tuff, and (3) recessional meltwater channel with Blacktail Ponds at the east end.

STOP 2–1. BLACKTAIL DEER PLATEAU

The Blacktail Deer Plateau is formed by moraines of Deckard Flats age that extend about 3 mi (5 km) south and west

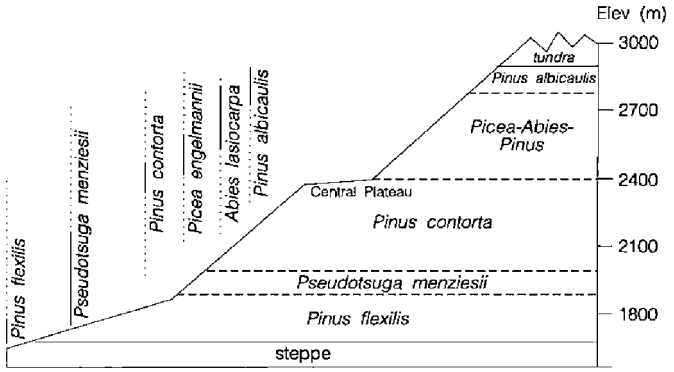


Figure 7. Vegetation zones of Yellowstone and Grand Teton national parks. Solid vertical lines show the altitudinal range of important species. Dotted vertical lines indicate where species is present (modified from Whitlock, 1993).

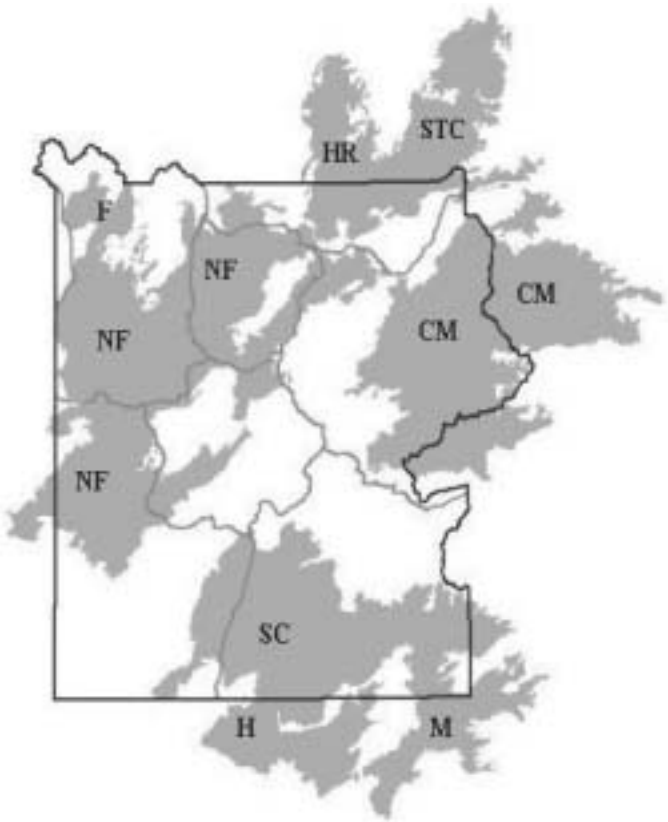


Figure 8. Areas (shaded) burned during 1988 Yellowstone fires. **CM**= Clover Mist complex, **F**= Fan fire, **H=** Huck fire, **HR**= Hellroaring fire, **M**= Mink fire, **NF**= North Fork fire, **SC**= Snake complex, **STC=** Storm Creek fire. The Clover Mist, Fan, Mink, Snake Complex and Storm Creek fires were all lightning caused fires that were allowed to burn early in the season. The Huck, Hellroaring, and North Fork fires were human caused and suppression action applied from first detection and throughout the season.

from this viewpoint. The successively lower morainal benches, locally including kame gravel, define about 10 recessional ice margins. Wind–drift of soil and snow makes for a loamier and moister soil on the northeast sides of ridges, whereas the south sides are drier due to greater solar insolation, wind exposure, and erosion of fine soil.

The Deckard Flats moraines of the Blacktail Deer Plateau are an important part of Yellowstone's northern winter range. Grasslands at the site are dominated by Idaho fescue (*Festuca Idahoensis*), western wheatgrass (*Agropyron smithii*), Junegrass (*Koeleria cristata*), bluebunch wheatgrass horsebush (*Tetradymia canescens*), green rabbit brush (*Chrysothamnus viscidiflorus*), and mountain big sage (*Artemisia tridentata spp menziesii*). Local vegetation patterns are strongly controlled by the surficial geology (Despain, 1990). The sagebrush–grasslands are underlain by glacial deposits and contrast with the adjacent bedrock areas. Limestone supports limber pine, Douglas–fir, and aspen, depending on moisture availability. Andesite of the Washburn Range and northern Grand Teton region supports spruce/fir forest at the higher elevations and Douglas–fir and aspen at lower elevations, interspersed with meadows, while rhyolite plateaus beyond the Deckard Flats ice margin to the southwest are dominated by lodgepole pine. One way to appreciate the influence of geologic processes on the vegetation is to note that the Deckard Flats moraines form extensive grasslands. If glaciation had not occurred, rhyolite would have been at the surface and the region would have probably been covered with lodgepole pine forest. If rhyolite had not been emplaced, then andesite and sedimentary rocks would have been at the surface and the northern range would probably have been covered by Douglas–fir forest. These substrate differences have exerted long–term influences on the vegetation that are evident in the paleo–ecological record as well (Whitlock, 1993).

Blacktail Creek descends into the Black Canyon of the Yellowstone River and joins the river near Crevice Lake, a deep lake with a varved–sediment record. In February of 2001, scientists from the USGS and Universities of Oregon and Nebraska recovered an 8–m–long core that spans the last 11,000 ^{14}C years. The exceptional water depth (30 m), compared with its relatively small surface area prevents yearly mixing of the water column. The availability of a record with this high temporal resolution creates opportunities to examine the environmental history of Yellowstone with a precision not possible in most paleo–ecological studies. In addition, tree–ring studies in and around Crevice Lake offer an opportunity to compare hydrological and ecological variables with climate–related changes in tree growth. Limber pine (*Pinus flexilis*), Douglas–fir, and other species from climatically sensitive sites in the Greater Yellowstone ecosystem (GYE) have been used to develop a tree–ring–based hydrologic record that spans several centuries (Graumlich and Littell, 2001; Graumlich et al., in press). Analysis of fire–scars on Douglas–fir around Crevice Lake discloses past fire events in the watershed (Littell, 2002), which can be compared with events found in previous studies (Houston, 1973) and charcoal–based studies (Millspaugh and Whitlock, 2002).

Pull over to view Phantom–Lake meltwater channel eroded more than 200 feet into bedrock by the ice–marginal "Yellowstone River." A lake dammed by a small alluvial fan was commonly present here in the early part of the summer.

STOP 2–2. TOWER FALLS VIEWPOINT

The Narrows Canyon of the Yellowstone River was eroded in postglacial time. A complex assemblage of Pleistocene basalt and glacial, fluvial, and lacustrine sediment known as the "sediments and basalts of The Narrows" are exposed in the steep, narrow canyon. Pinedale till caps the section, locally resting on Lava Creek tuff. These units rest on the eroded top of the upper basalt, which in turn is underlain by the till of Bumpus Butte, containing erratics of Precambrian rocks from sources to the north. Conformably beneath the till is lacustrine sediment of an ice–dammed lake, containing ice–rafted stones of Precambrian crystalline rock and coarse, silicic volcanic ash from a nearby source. The argon–argon age of the tephra is 1.3 ± 0.1 Ma and the age and chemistry suggest correlation with the Mesa Falls Tuff (Obradovich and Izett, 1991). The close tie between the ice–dammed lake sediments containing the ash and the overlying till also date the glaciation as ~1.3 ± 0.1 Ma, an age two orders of magnitude older than the Pinedale.

STOP 2–3. JUNCTION BUTTE MORAINES

In The Junction Butte moraines of the post–Deckard Flats recession are the most prominent and occupy a position typically referred to as late Pinedale in the Rocky Mts. Large boulders of Precambrian crystalline rocks stud their surface and permanent and seasonal ponds occupy common depressions. A large glacier flowing down Slough Creek deposited these moraines and blocked the Lamar valley upstream from Slough Creek. The till has enough fines for good water–holding capacity and adequate nutrients. The mineral soil is well covered by the vegetation. The vegetation at this site is mountain big sage, Idaho fescue, Junegrass, bluebunch wheatgrass, silvery lupine (*Lupinus argenteus*), Indian paintbrush (*Castilleja* sp.), and green rabbit brush. The morainal topography with ridges, slopes and swales produces a mosaic of vegetation assemblages. Near the pond, note willow (well browsed), sedge (*Carex* spp.), tufted hairgrass (*Deschampsia caespitosa*), spikerush (*Eleocharis* sp), and bulrush (*Scirpus* sp.). The water level in the ponds has diminished since the 1970's as a result of prolonged winter drought.

STOP 2–4. SLOUGH CREEK PITTED OUTWASH, ECOLOGY AND PALEOECOLOGY

Walk up onto kame–gravel bench south of road before the Lamar River bridge. Here, near the confluence of Slough Creek and the Lamar River, six, nested, pitted, outwash terraces formed

during recession of the Slough Creek glacier. In contrast to the boulder–studded Junction Butte moraines, the surfaces of the kame gravel terraces are boulder–free. Lakes dammed by the Slough Creek glacier existed in the Lamar valley at the time of the nested kame terraces as well as when the Slough Creek glacier was even larger. Draining of such lakes is the prime candidate for the flood deposits near Gardiner. The kame gravel with its well–washed sand matrix forms well–drained soils with low moisture–holding capacity and sparse vegetation.

Walk southwest to northeast from a dry SW–facing slope to a moist, NE–facing slope, noting plant changes in this dry–to–moist transect. The dry, SW–facing slope has high solar insolation, wind deflation of snow, and a sandy soil at the surface, with a high proportion of mineral soil not covered by vegetation. The dry–slope vegetation is composed of needle–and–thread grass, bluebunch wheatgrass, Junegrass, Hood's phlox (*Phlox hoodii*), fringed sage (*Artemisia frigida*) but no Idaho fescue. The NW–facing slope accumulates snowdrifts and soil blown from the dry site and has lower solar insolation and a silty soil with a complete vegetation cover of the soil. The vegetation is silky lupine (*Lupinus sericeus*), Idaho fescue, bluebunch wheatgrass, (tall) prairie smoke (*Geum triflorum*), a non–native smooth brome (*Bromus inermis*), sticky geranium (*Geranium viscosissimum*), and wild buckwheat (*Eriogonum umbellatum*).

In this pitted outwash complex, Slough Creek Lake provides a vegetation and fire history for the last 16,000 cal yrs, B.P. (Fig. 9). Following deglaciation, the area supported tundra vegetation and conditions were cooler and drier than at present. A period of pine–juniper (*Pinus–Juniperus*) forest and low fire frequency was present between ca. 11,000 and 7000 cal. yrs. B.P. (10,000 and 6000 ^{14}C yrs. B.P.), when the climate was warmer and wetter than at present. Fire frequency increased from four fires/1000 yrs. in the early Holocene to >10 fires/1000 yrs. in the last 7000 cal. years. The establishment of Douglas–fir parkland occurred as a result of cooler drier conditions and increased fire activity in the late Holocene.

Slough Creek Lake and other northern sites record a different climate history than sites in the southern region (Fig. 10). Today, the northern part of the Park experiences higher summer and annual precipitation as a result of convectional storms associated with summer monsoonal circulation, while the rest of the Park is relatively dry and under the influence of the northeastern Pacific subtropical high–pressure system. In the early Holocene, greater–than–present summer insolation amplified the contrast between wet–summer and dry–summer regions (Fig. 11). Central and southern Yellowstone (the dry–summer area) became drier in the early Holocene as a result of a stronger–than–present northeast Pacific subtropical high–pressure system, while wet–

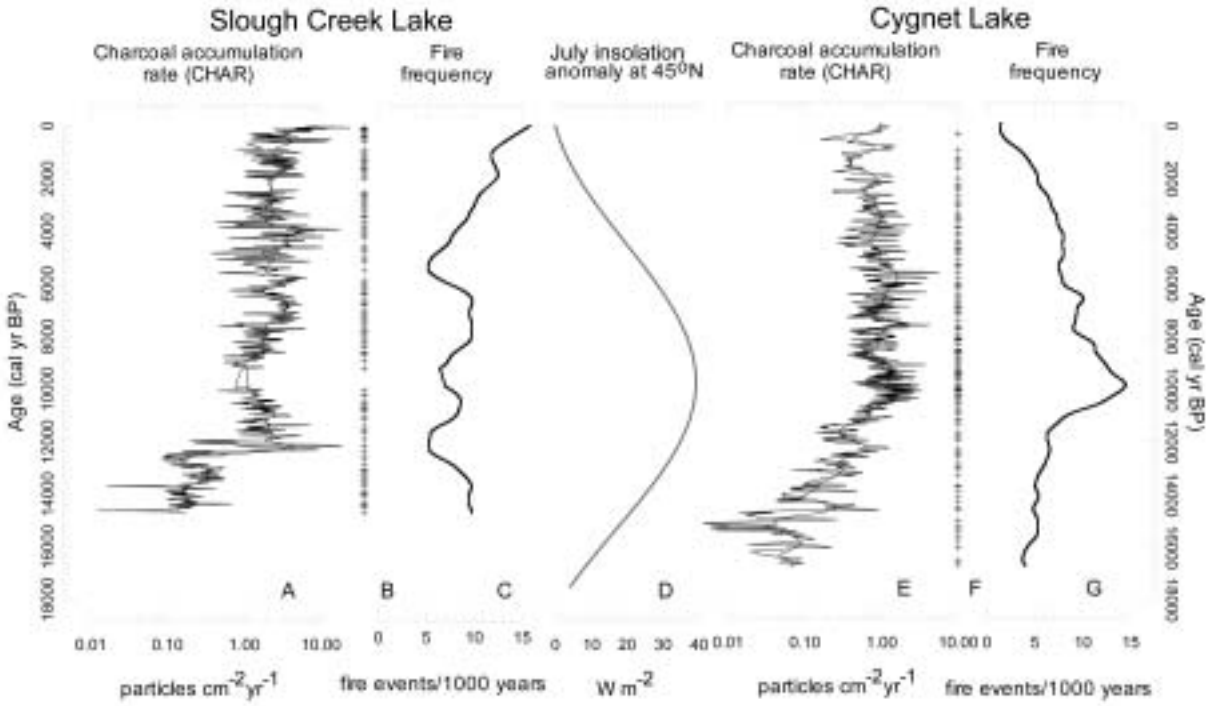
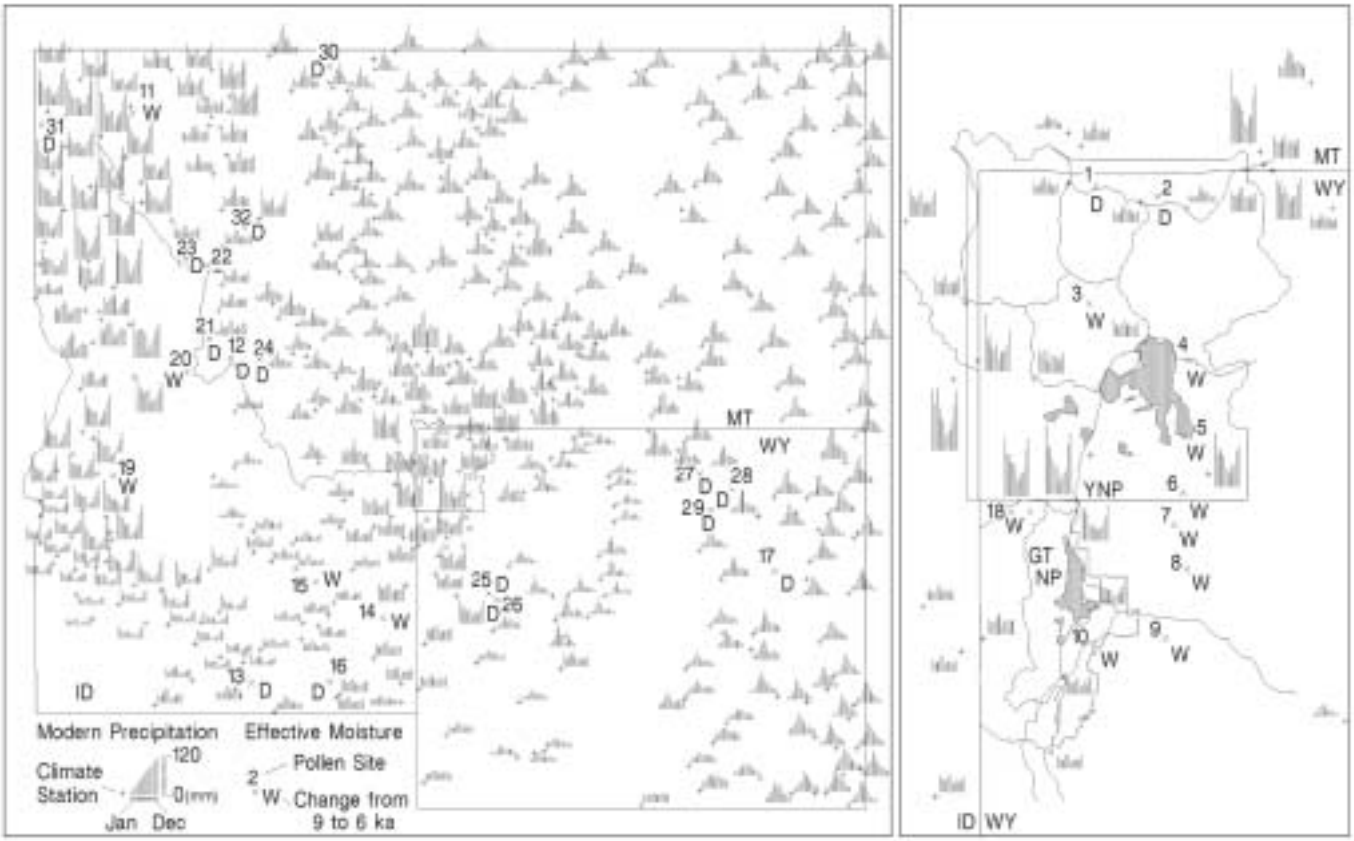


Figure 9. Comparison of charcoal data from Slough Creek Pond in the Yellowstone–Lamar region (summer–wet) and Cygnet Lake in the Central Plateau (summer–dry). The reconstructions are derived from accumulation rates of macroscopic charcoal particles (CHAR) in contiguous 1–cm–thick core samples. Results are plotted as log–transformed CHAR values (A, E). Charcoal peaks (+) above background levels define fire events (B, F), which are summaries as number of fires/1000 years (C, G). Long–term trend in fire frequency compares well with changes in precipitation regime associated with the July insolation anomaly (D, Berger, 1978).



1. Gennett & Baker (1986) Palynology 10:61-71.
2. Whitlock & Bartlein (1993) Quaternary Research 39:231-238; Millspaugh & Whitlock (in press) In: After the Fires: The Ecology of Change in Yellowstone National Park, L. Wallace (Ed.).
3. Millspaugh *et al.* (2000) Geology 28:211-214.
4. Waddington & Wright (1974) Quaternary Research 4:175-184.
5. Baker (1976) USGS Professional Paper 729-E:E1-E48.
6, 7, 8, 9, 10. Whitlock (1993) Ecol. Monogr. 63:173-198
11. Mack *et al.* (1983) Quaternary Research 20:177-193.
12. Mehringer *et al.* (1977) Arctic and Alpine Research 9:345-368.
13. Davis *et al.* (1986) Quaternary Research 26:321-339.
14. Beiswenger (1991) Ecol. Monogr. 61:165-182.
15. Davis *et al.* (1986) Quaternary Research 26:321-339.
16. Bright (1966) Tebiwa 9:1-47.
17. Markgraf & Lennon (1986) Plains Anthropologist 31:1-12.
18. Whitlock *et al.* (1995) Quaternary Research 43:433-436.
19. Doerner & Carrara (1999) Arctic, Antarctic, and Alpine Research 31:303-311.
20, 21, 22. Brunelle-Daines (2002) Dissertation, Univ. of Oregon.
23. Karsian (1995) Master's Thesis, Univ. of Montana.
24. Brunelle-Daines (2002) Dissertation, Univ. of Oregon.
25. Fall *et al.* (1995) Quaternary Research 43:393-404.
26. Lynch (1998) Ecology 79:1320-1338.
27, 28, 29. Burkart (1976) Dissertation, Univ. of Iowa.
30. Barnosky (1989) Quaternary Research 31:57-73.
31. Mack *et al.* (1995) Quaternary Research 10:241-255.
32. Mehringer (1985) In: Pollen Records of Late-Quaternary North American Sediments, V. M. Bryant and R. G. Holloway (Eds.).

Figure 10. Annual distribution of precipitation at climate stations in Idaho, Montana, and Wyoming (World WeatherDisc Associates, n.d.), and inferred effective moisture changes from 9000 to 6000 cal yrs. B.P. at paleoecological sites. "W" indicates an inferred effective moisture change to wetter conditions; "D" indicates an inferred effective moisture change to drier conditions; and "–" indicates no inferred direction of change. (Right) The same information for the region of Yellowstone National Park (YNP) and Grand Teton National Park (GTNP) (adapted after Whitlock and Bartlein, 1993; Shafer et al., in press).

summer northern Yellowstone became wetter in response to enhanced monsoonal circulation (Fig. 11). The boundary between wet–summer and dry–summer regimes is controlled by topography and has not changed significantly during the Holocene (Whitlock and Bartlein, 1993).

STOP 2–5., FIRE–RELATED EROSION AND SEDIMENTATION

Glacial ice completely filled the Lamar and Soda Butte valleys. During deglaciation, glaciers in the upper Lamar drainage

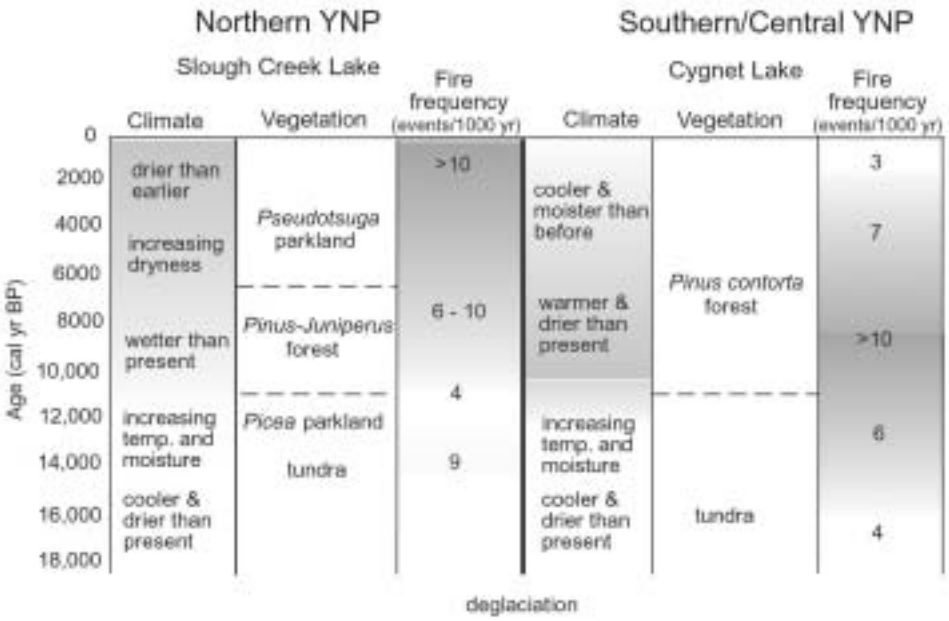


Figure 11. Summary of late–glacial and Holocene changes in climate, vegetation, and fire frequency for the Yellowstone–Lamar and Central Plateau provinces, based on pollen and charcoal data. Shading indicates intensity of aridity (darker shading=greater aridity) (from Millspaugh and Whitlock, in press).

downwasted and extensive kame gravel and lake sediments accumulated. Later, during deglaciation, the upper Lamar River built a sandy delta into glacial–lake Lamar that was dammed in the Lamar Canyon area by the Slough Creek glacier.

The dense, lodgepole–pine and mixed–conifer forests of northeastern Yellowstone tend to burn infrequently in large, stand–replacing fires, as in 1988. These severe burns promote debris flows and floods because of greatly enhanced runoff, especially in summer thunderstorms. Alluvial fans along the sides of glacial trough valleys, like those of Soda Butte Creek, were built in significant part by post–fire sedimentation. Approximately 30% of the stratigraphic thickness of these fans consists of fire–related deposits. Radiocarbon dating shows that fan aggradation was greatly accelerated by post–fire sedimentation during warm, drought–prone periods (e.g., Medieval warm period ca. 900–1200 AD). In contrast, fire–induced sedimentation is minimal at times that coincide with the ~1400–year Holocene cycle of cold episodes in the North Atlantic (Bond et al., 1997) (Fig. 12). During these cool, wet intervals, active lateral migration of lower Soda Butte Creek created broad floodplains cut on underlying glacial sediments, as shown by ^{14}C ages on overbank deposits. These surfaces were later preserved as fill–cut terrace treads. Millennial–scale climate change has had a major influence on both fires and geomorphic processes during the Holocene (Fig. 12).

STOP 2–6. ROUND PRAIRIE HISTORIC FLOODS

Infrequent, major floods are part of a suite of catastrophic geomorphic processes that exert a strong influence on valley–floor ecosystems in northeastern Yellowstone. Along Soda Butte Creek, these floods carried bedload high enough to deposit extensive overbank gravel that allowed conifers to replace floodplain meadows, as in upper Round Prairie. Tree–ring methods date these floods to 1918, ~ 873, and possibly ~1790. Short–term gage records indicate that snowmelt runoff approached the 100–year flood in 1996 and 1997, but indirect estimates show that peak discharge was 2–3 times greater in 1918 than in 1996. Gravelly overbank deposits create well–drained, more xeric surfaces and also reduces bank stability.

Anthropogenic impacts to Soda Butte Creek include a mine–tailings dam failure at Cooke City in 1950. The resulting flood produced extreme peak discharges, but was too brief to cause much erosion. It did emplace a large volume of acidic, metal–contaminated sediment over high floodplain surfaces. Because these tailings deposits lie above the reach of most natural floods, they may persist over long periods, with documented impacts to vegetation and aquatic organisms.

Pull over at Lamar Ranger Station to view Lamar Valley. This area was once the Buffalo Ranch where a few bison were sheparded until they formed large herds. These in turn became the stock for much of the bison now surviving in the U.S. Downvalley across the Lamar River, large erratics of the Slough Creek glacier provide evidence for damming of a deep lake in the Lamar Valley. In the roadcut going into the Lamar Ranger Station, foreset bedding suggests a delta of Rose Creek build into glacial lake Lamar. The arcuate perimeter also suggests a delta. Absaroka volcanic rocks of Eocene age form both sides of the valley, and those across the valley are noted for multiple beds containing petrified trees. Continue back to Tower Junction and head south.

STOP 2–7. WASHBURN RANGE VIEWPOINT

During the last glaciation, the Washburn Range (Absaroka andesite) was overridden, scoured, and striated by the north–flowing icecap from the Yellowstone Plateau (Pierce, 1979). The crest of Specimen Ridge was also striated by glacial

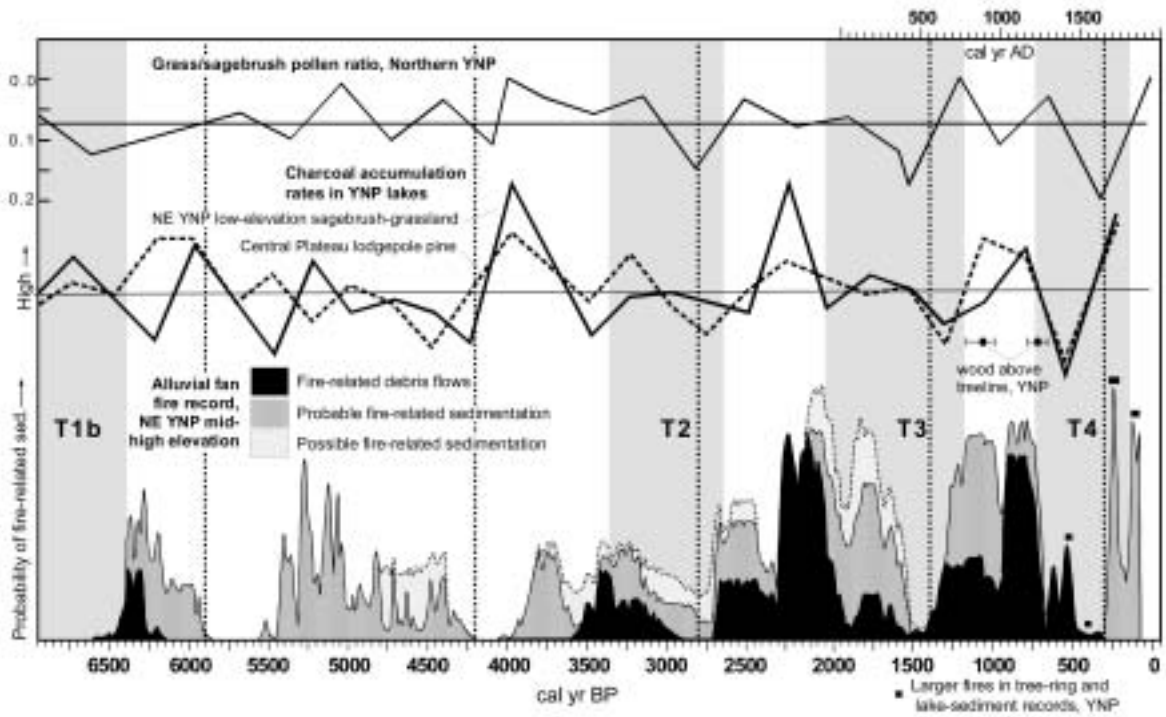


Figure 12. Comparison of records of middle to late Holocene fire–related alluvial–fan sedimentation (Meyer et al., 1995) with other fire and climate proxy records. Scale in cal yrs. B.P. indicates ^{14}C–calibrated calendar year ages before present, where present is defined by convention as 1950 (Stuiver and Reimer, 1993). The spectrum of probability of fire–related sedimentation is constructed by summation of calibrated probability distributions for radiocarbon dates on individual events. Note correspondence of fire–related sedimentation events with large fires identified in high–resolution lake sediment study of the last ~750 yr (black squares; Millspaugh and Whitlock, 1995). 500–yr average charcoal accumulation rates in Yellowstone lakes (Millspaugh, 1997; Millspaugh and Whitlock, this volume) show correspondence with major fire–related sedimentation peaks in the last 3000 years. Wood above present treeline (K.L. Pierce, written comm. 1993) indicates warmer temperatures ca. AD 800–1200 corresponding to increased fire sedimentation in the "Medieval Warm Period". Periods of overbank sedimentation on mainstem Soda Butte Creek terraces (gray shading, T1b–T4) correspond with generally wetter conditions indicated by high grass/sagebrush pollen ratios (note inverted scale; from Gennett and Baker, 1986). Vertical dotted lines indicate maximum North Atlantic cooling events identified by Bond et al. (1997), all of which correspond to a distinct minima in fire–related sedimentation, strongly suggesting control of large fires by hemispheric– and millennial–scale climatic variations.

ice flowing toward Gardiner. During glacial recession, as the plateau icecap stagnated, glaciers flowed south from the Beartooth uplift, left southerly striations on Specimen Ridge, and reached as high as 8800 ft (2680 m) on the north side of Mt Washburn. This Deckard Flats "readjustment" carried boulders of granite and limestone and locally left substantial deposits of till and kame gravel (Pierce, 1979).

This upland areas on Absaroka volcanic bedrock has soils of variable thickness. Most glaciated ridges have very thin soils with bedrock just below the surface. Such ridges form the dry meadows on the north side of Mt Washburn. Where the soil is thicker, particularly in concavities, are forests of spruce and subalpine fir. At this stop, till several meters thick supports lusher meadows, with mixed grass and forbs including Idaho fescue, bearded wheatgrass (*Agropyron caninum*), mountain brome (*Bromus ciliatus*), bluegrasses (*Poa* spp.), sticky geranium, cinquefoils (*Potentilla* spp.), little sunflower (*Helianthella uniflora*), balsamroot (*Balsamorhiza sagittata*), and mountain big sagebrush. Across Yellowstone Canyon, benches of Lava Creek Tuff support lodgepole pine. Above this bench on Specimen Ridge the Absaroka volcanic terrain is a mosaic of spruce/fir forest interspersed with meadows. Above the modern treeline on Mt Washburn, old stumps date from the Medieval warm period. Continue driving south to Lake Village.

DAY 3. CANYON–LAKE VILLAGE AREA

STOP 3–1. DUNRAVEN PASS, CALDERA, AND PLATEAU OVERVIEW

Park in Dunraven Pass and hike a few hundred meters up the old road. This south–facing escarpment is formed in Absaroka volcanic rocks of an Eocene volcano whose southern half was down–faulted into the 640 ka caldera. The caldera extends from

here to the Red Mts. about 30 mi. to the south. To the southwest, rhyolite flows form the nearly flat surface on the distant horizon. Most flows in and near the caldera are 70,000 to 160,000 years old. The vegetation on rhyolite is lodgepole pine whereas that on andesite is spruce/fir with extensive meadows. Well above this site, the top of Mt Washburn was striated by north–flowing ice from the Yellowstone Plateau. The Washburn Range was rounded and sculpted by overriding ice during the last and earlier glaciations.

STOP 3–2 (OPTIONAL). DUNRAVEN PASS PICNIC AREA. GEO–ECOLOGY AND FIRE HISTORY

Park in the Dunraven Pass picnic area. We are near the contact between Absaroka volcanic rocks that form the Washburn Range to the north and rhyolite lava flows that form the Yellowstone Plateau to the south. A small drainage occupied by the paved road separates the two. This north–to–south traverse illustrates the strong geologic control of forest types. We will walk from Absaroka volcanic terrain with a spruce/fir forest southward up onto rhyolite terrain (Dunraven Road flow) with a lodgepole forest.

STOP 3–3. GRAND CANYON AT ARTIST POINT AND REST STOP

Take the bridge across the Yellowstone River and park at the Artist Point parking area. The hydrothermally altered rhyolite in the deeper part of the canyon near Artist Point experienced pyrite mineralization before the canyon was cut, which was then oxidized to yield the yellow and red colors. Similar extensive alteration probably also exists elsewhere at depth within the caldera. Alteration decreases upstream from Artist Point to firmer rock at Lower Falls. Emplacement of rhyolite built up the Yellowstone Plateau and created steep gradients at the margin of this plateau. Stream erosion has incised the margins of this plateau creating canyons. The largest river, the Yellowstone, has eroded farthest into the plateau and created the deepest canyon. Across the canyon and extending about halfway down into the canyon is a paleo–valley of the Yellowstone River filled with sediment. Glacial advances from the Beartooth uplift both early and late in a glacial cycle have dammed the Yellowstone River.

Pull over into the parking area for Canyon horseback stables. The smooth grassy slopes are part of an extensive area of thick, glacial till containing Precambrian rocks carried southward from the Beartooth uplift, probably early in the last glacial cycle before an icecap became established on the Yellowstone Plateau. This south–flowing glacier also dammed the Yellowstone River creating a lake in the Hayden Valley–Yellowstone Lake area.

STOP 3–4. HAYDEN VALLEY

Hayden valley is a productive grassland on glacial lake sediments surrounded by lodgepole pine on rhyolite. These meadows can exist here because of the accumulation of lake and other sediments that have both better water holding capacity and more nutrients than the surrounding rhyolite terrain. Hayden Valley was the site of glacially dammed lakes: (1) before the last glacial maximum when an advance from the north reached the Canyon Village area, and (2) during recession from the last glacial maximum when the icecap on the Yellowstone Plateau had melted but glaciers from the Beartooth uplift blocked the Yellowstone River east of Mount Washburn (Pierce, 1979; Richmond, 1976, 1977).

At this stop, a thin veneer of alluvium mantles lake sediments resulting in a high water table that favors silver sage (*Artemisia cana*) and Richardson's needlegrass (*Stipa richardsonii*). Varved lake sediments deposited during Pinedale recession are exposed in the Yellowstone River bank. Where the overlying sediments are thick enough to allow deeper water drainage, mountain big sage replaces the silver sage and Idaho fescue that is the dominant understory species. Other flora includes sticky geranium, cinquefoils, bluegrasses, and bearded wheatgrass. Lodgepole pine forests surrounding Hayden valley grown on glaciated rhyolite that has sandy soils with low nutrients and water holding capacity.

A pollen study from Cygnet Lake (Fig. 9) in the Central plateau indicates that high percentages of sagebrush and grass (Poaceae) were present between 17,000 and 12,800 cal yrs. B.P. (14,800–10,900 ^{14}C yrs. B.P.) when the region was colonized by tundra. An increase in pine (*Pinus*) between 12,800 and 11,300 cal yrs. B.P. (10,900 and 10,100 ^{14}C yrs. B.P.) marks the establishment of lodgepole pine forest at the site. Unlike sites in non–rhyolitic regions, the pine forest at Cygnet Lake has persisted with little change for the last 11,300 cal yrs., despite changes in temperature and effective moisture (Whitlock, 1993). In contrast, fire frequency variations inferred from the charcoal record are well correlated with the July insolation anomaly and reach highest levels ca. 9900 cal yrs. B.P., when summer drought was most intense (Millspaugh et al. 2000).

Pull over at the Le Hardys Rapids parking area. Le Hardys Rapids serves as the bedrock threshold of Yellowstone Lake. It is also on the crest of the historic axis of uplift and subsidence of the Yellowstone caldera. Upstream from here, note that the Yellowstone River is quite wide, tranquil, and quite low gradient.

OVERVIEW OF CALDERA UNREST

Between 1923 and 1985, the center of the Yellowstone caldera rose ~1 m (Pelton and Smith, 1979, Dzurisin and Yamashita, 1987). From 1985 until 1995–6, it subsided at about two cm/yr (Dzurisin et al., 1994). More recent radar interferometry studies show renewed inflation (Wicks et al., 1998). Doming along the caldera axis reduces the gradient of the Yellowstone River from Le Hardys Rapids to the Yellowstone Lake outlet and ultimately causes an increase in lake level. Meyer and Locke (1986) and Locke and Meyer (1994) suggested that the shorelines were cut during intra– caldera uplift episodes that produced rising water levels. Figure 14 shows the history of lake level changes and other events (Pierce et al., 2002). The lake in the outlet region has been below or near its present level for about half the time

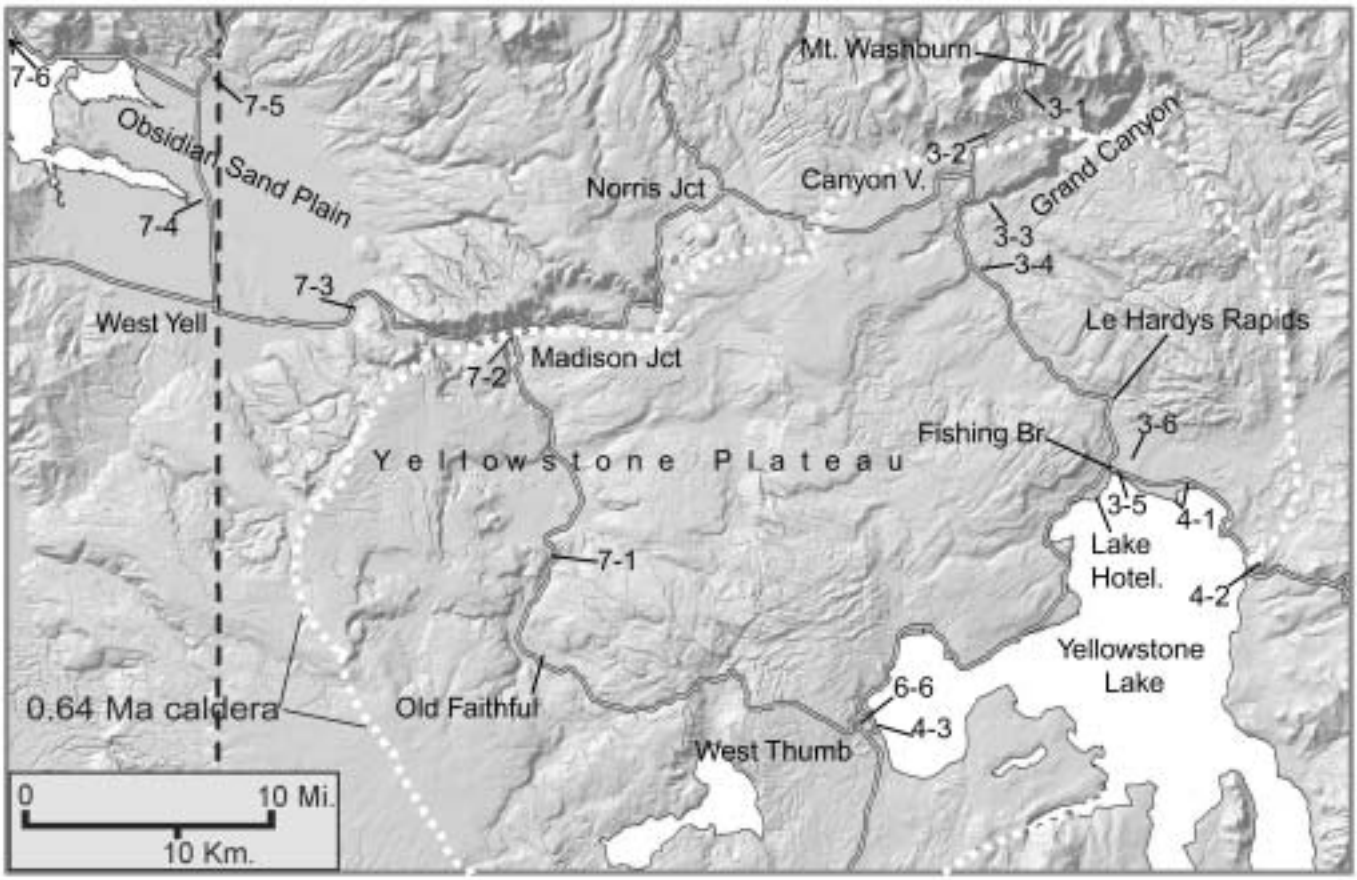


Figure 13. Map of central Yellowstone showing location of many stops on Day 3, 4, and 6. White dashed line is the approximate outline of 0.64 caldera.

since a 3,300 ft (1 km)–thick icecap melted from the Yellowstone Lake basin about 16 cal ka B.P..

Based on reconstructions of the river gradient from the lake outlet to Le Hardys rapids and indicators of former lake and river levels, Le Hardys rapids was uplifted about 8 meters relative to the lake outlet at least twice, once ~9.5–8.5 cal ka B.P. and again after ~3 cal ka B.P. (Pierce et al., 2002). Following these uplifts, it appears to have subsided a similar amount. Older possible rises in lake level are suggested by *S4* truncating older shorelines and *S5* truncating valleys. A plot of lake level through time shows 5–7 millennial–scale oscillations after 14.5 cal ka B.P. (Fig. 14). Le Hardys Rapids to Fishing bridge spans only the central 25% of the historic caldera doming, so if historic doming is used as a model, the total projected uplift is four–fold greater. This "heavy breathing" of the central part of the Yellowstone caldera may reflect a combination of several processes, (1) magmatic inflation, (2) tectonic stretching and deflation, and (3) inflation from hydrothermal fluid sealing, followed by cracking of the seal, pressure release, and deflation. Over the entire postglacial period, subsidence has balanced or slightly exceeded uplift, as shown by older shorelines that *descend* towards the caldera axis. We favor a hydrothermal mechanism for inflation and deflation because it provides for both inflation and deflation with little overall change. Other mechanisms such as inflation by magma intrusion and deflation by extensional stretching require two separate processes to alternate and yet result in no net elevation change.

STOP 3–5. WALK, FISHING BRIDGE AREA

Depart busses on the east side of Fishing bridge and drive over to Fishing Bridge Museum and walk to S4 barrier beach on the east side of the outlet. Shoreline designations are after Meyer and Locke (1986), starting with the modern shoreline as *S1* that is 1.9 m above datum (the Bridge Bay gauge). Walk east across Fishing Bridge fault with 1 m offset into the S4 lagoon and discuss archeological findings. Then hike south past Hamilton Store on S4 shoreline and across highway down to S2 shoreline near the Fishing Bridge Museum. Restrooms and busses will be near the museum.

The level of Yellowstone Lake has decreased and increased through time (Fig. 14; Pierce et al., 2002). Lidar images (Fig. 15) show the shorelines and other geomorphology important to deciphering this lake level history as well as tilting and faulting. The *S4* shoreline forms a barrier beach and lagoon and is about 10.7 cal ka B.P., based on buried charcoal. Cody Complex archeological material of late Paleoindian age also supports an age for *S4* of

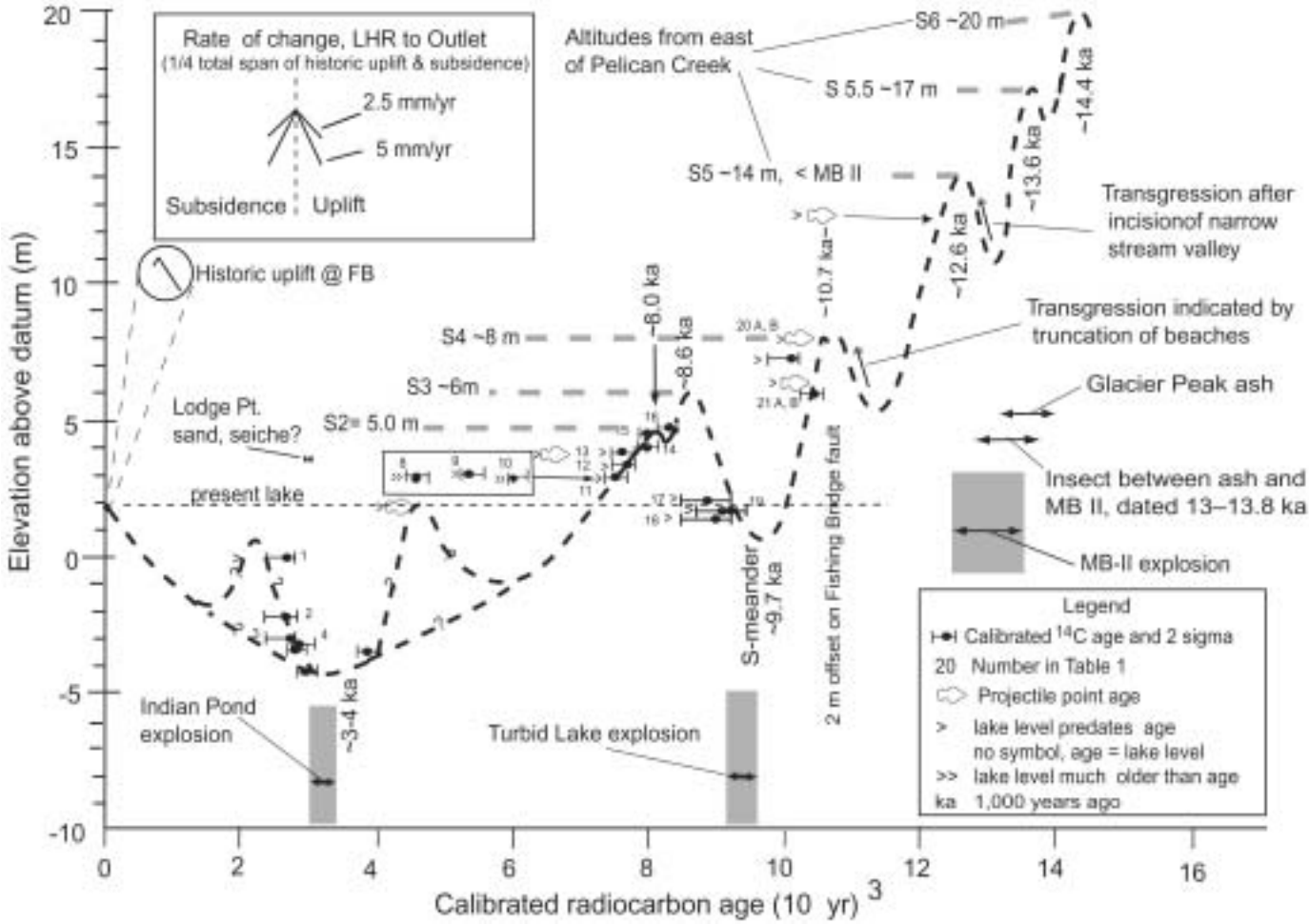


Figure 14. Holocene history of changes in Yellowstone Lake level suggesting cycles of inflation and deflation of the Yellowstone caldera (from Pierce et al., 2002). S2–S6, numbered shorelines following Meyer and Locke (1986).

10–11 ka (Cannon et al., 1994). The *S2* shoreline here is a wave–cut shoreline and is about 8 cal ka B.P. as dated by archeological material, and by charcoal ages on the *S2* shoreline and on the Pelican Creek terrace that truncates the *S2* shoreline (Cannon et al., 1994). When viewed from the *S2* shoreline, all post–8 ka history of the lake is either at present lake level or is now underwater (Pierce et al., 2002). To the southwest in Bridge Bay, a 3.8 cal ka B.P. beach is submerged in 5 m of water. To the east, Pelican Creek lies in a drowned valley with gravel down to 5 m below the present lake.

STOP 3–6. DROWNED RIVER CHANNELS

Drive north from the Hamilton store across the *S4* barrier beach and lagoon and 0.4 mi (0.7 km) beyond the locked gate. Park in the S–meander and walk over to the riverbank. The steep river bank here was undercut by the Yellowstone River at a time when the river was much more vigorous than it is now (Pierce et al. 2002). Cores in the abandoned channel here went through water and lake mud into gravel dated at 2.9 cal ka B.P. at a depth of ~5 m. The outlet reach of the Yellowstone River is now very low gradient, but about 3 ka B.P., it was a gravel–transporting river undercutting its banks, and at this site depositing eolian sand at the top of the eroding river bank. Inflation of Le Hardys rapids diminished the river gradient such that the gravel at 5 m below the surface here is now well below the level of the bedrock threshold 4 km downstream at Le Hardys Rapids. Accounting for this and a 1m/km original gradient of the gravel–transporting river, Le Hardys rapids has come up 7 m relative to this site (Pierce et al., 2002).

Across the road in the S–meander, we dug four backhoe pits across the old channel of the Yellowstone River (Pierce et al., 2002). The outside of this meander is a high, steep bank formed at a time when the Yellowstone River was undercutting its banks and carrying gravel up to 3 cm. Then, as the river lost its competence, charcoal accumulated at about 9 cal ka B.P., and was buried by shoreface gravel. Water level rose until this reach of the Yellowstone River became a northern arm of Yellowstone Lake and the *S3* and *S2* shorelines were cut. Return to Lake Village for second night and possible evening talks.

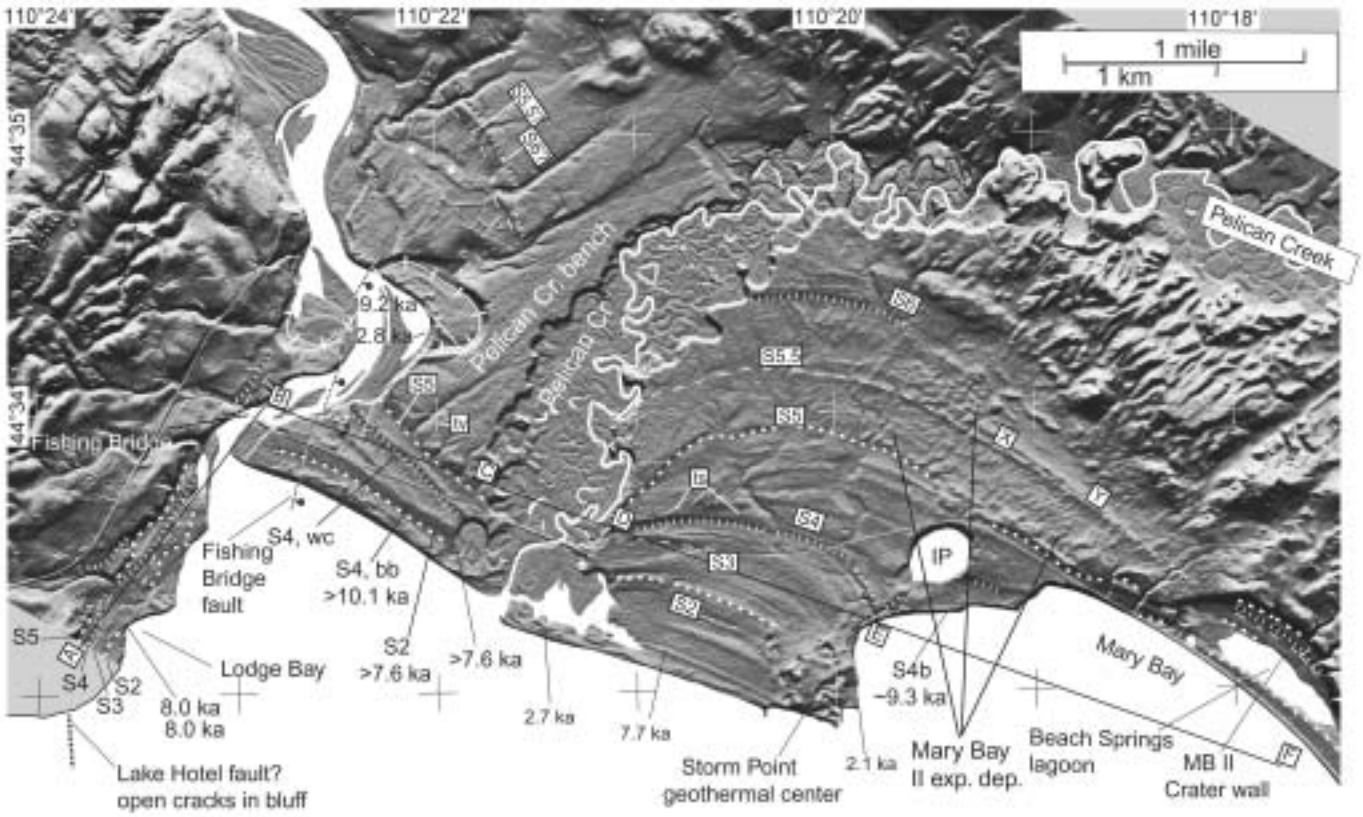


Figure 15. Lidar image of north shore area of Yellowstone Lake (from Pierce et al., 2002). Annotations show numbered shorelines, location of radiocarbon ages, Holocene faulting, and Mary Bay II explosion deposits. The wide channel of Yellowstone River is depositing sand bars and has been drowned by uplift downstream. The S–meander is shown by white line with crossing bar. IP– Indian Pond explosion crater, tv– truncated valley, ts– truncated shorelines.

DAY 4. LAKE VILLAGE TO JACKSON, WY

STOP 4–1. HIKE TO STORM POINT

About 3 cal ka B.P., a hydrothermal explosion ejected low–grade hydrothermal materials (weakly cemented lake sediments, sand, and gravel) and left the crater occupied by Indian Pond. The creek west of Indian Pond exposes platform sand, ~13–ka Mary Bay II hydrothermal explosion deposits, lake gravel, lake sediments with bryophytes, and varved lake sediments with Glacier Peak ash. Storm Point is a complex of hydrothermal craters as well as exposures of steeply dipping Pleistocene gravel. Uplift of Storm Point center has tilted shorelines westward at 6 m/km (Pierce et al., 2002).

YELLOWSTONE LAKE BOTTOM

Discoveries from multi–beam sonar mapping and seismic–reflection surveys of the northern, central, and West Thumb basins of Yellowstone Lake provide new insight into the extent of post–collapse volcanism and active hydrothermal processes occurring in a large lake above a large magma chamber (Morgan et al., 2003). Yellowstone Lake has an irregular bottom covered with dozens of features directly related to hydrothermal, tectonic, volcanic, and sedimentary processes (Fig. 16). Detailed bathymetric, seismic reflection, and magnetic evidence reveals that rhyolitic lava flows underlie much of Yellowstone Lake and exert fundamental control on lake bathymetry and localization of hydrothermal activity. Many previously unknown features have been identified and include over 250 hydrothermal vents, several >1,500 ft (500 m) diameter, hydrothermal explosion craters, many ~300–500 ft (100–200 m) diameter, hydrothermal vent craters, domed lacustrine sediments related to hydrothermal activity, elongate fissures cutting post–glacial sediments, hydrothermal siliceous spires, sub–lacustrine landslide deposits, submerged former shorelines, and a recently active graben. Sampling and observations with a submersible, remotely operated vehicle (ROV) confirm and extend our understanding of the identified features. Faults, fissures, hydrothermally inflated, domal structures, hydrothermal explosion craters, and sublacustrine landslides constitute potentially significant geologic hazards. Toxic elements derived from

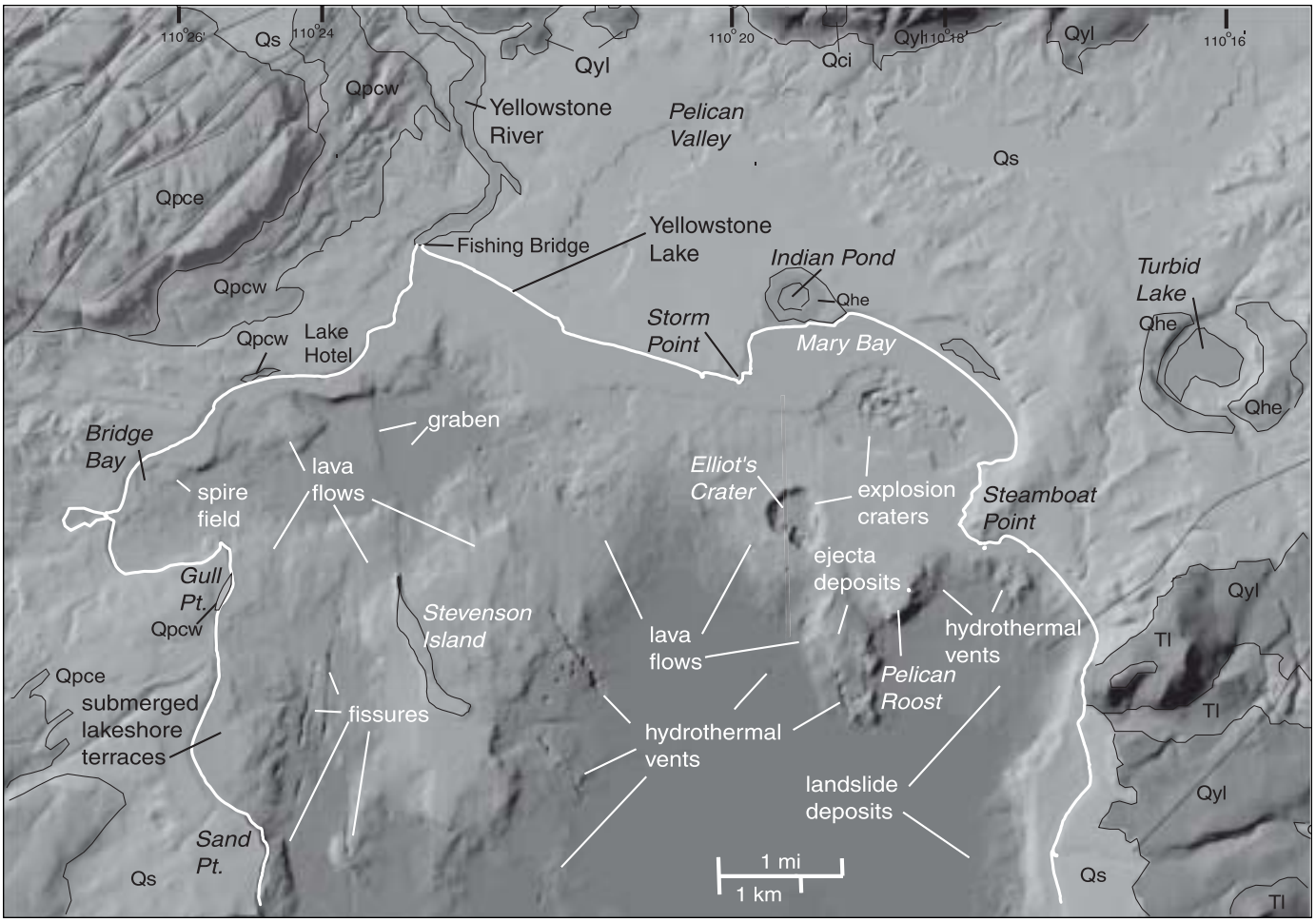


Figure 16. Shaded relief map showing underwater features on the floor of northern Yellowstone Lake (from Lisa Morgan, 2003). Onshore symbols: Qyl– Lava Creek Tuff, Qpcw– West Thumb flow, Qpce– Elephant Back flow, Qs– Quaternary sediments, and Qhe– hydrothermal explosion deposits.

hydrothermal processes may also significantly affect the Yellowstone ecosystem.

STOP 4–2. LAKE BUTTE VIEWPOINT

Absaroka volcanic and intrusive rocks here form the rim of the 640–ka Yellowstone caldera. During full–glacial time, the Yellowstone Plateau icecap formed a dome over Yellowstone Lake with ice flowing south from this icecap into Jackson Hole, west toward West Yellowstone, north over the top of the Washburn Range, and east through Sylvan Pass. The glacier flowing eastward through Sylvan Pass was so high that it flowed *up* the southern tributary valleys of Middle Creek and deposited lateral moraines 1200 ft (400 m) above the stream that slope *up* the tributary valley. During deglaciation, the Plateau icecap downwasted, and a succession of kame deposits was built from Sylvan Pass westward to Yellowstone Lake as well as around most of the lake basin. In a valley tributary to the southeast arm of Yellowstone Lake, peat dated at 15.5–16 cal ka B.P. overlies kame deposits and underlies gravel considered late Pinedale outwash from small glaciers upstream.

About halfway between here and Cody, Wyoming is Mummy Cave, which has an unusually complete archeologically record and serves as a major reference for archeological changes through time. (Husted and Edgar, 2002, Hughes, 1988).

The origin of Yellowstone Lake is complex. Excepting south arm, southeast arm, and the ~160–ka West Thumb caldera, the lake is in the southeast sector of the moat of the 0.64–Ma caldera. As concluded by Morgan et al. (2003), rhyolite flows are present on the west and south part of the lake, and a flow or intrusion is completely buried near the northeastern margin of the lake, but these flows are not as high as the flows that completely fill the “moat” in the north, west, and south parts of the caldera. The 300 ft (100 m) deep east side of the lake and its continuation up the southeast arm are a fiord–like trough apparently excavated by a glacier flowing northward from the high upper Yellowstone terrain during the early part of a glacial cycle.

STOP 4.3. WEST THUMB THERMAL AREA

West Thumb is located where a NNW–trending band of young intra–caldera vents intersect the ring–fracture zone of the Yellowstone caldera (Christiansen, 2001). Eruption of the tuff of Bluff Point 162 ± 2 ka formed the West Thumb caldera. Some sinter cones protrude above the lake and similar sinter cones are entirely underwater and were formed when the lake was at least 14–18 ft (4–5 m) lower. One of these is the "cutthroat Jacuzzi", often populated by circling trout. Along the south shore or Yellowstone Lake shorelines S2 and S4 are dated and of similar height as along the north shore. Although inflation and deflation of the caldera has occurred, the horizontality of shorelines suggests no net buildup of magmatic material at depth associated with the inflation–deflation cycles. This is compatible with inflation deflation cycles driven by geothermal fluid buildup and release.

Pull over to view the valleys of Thumb Creek and Little Thumb Creeks drowned by rising lake levels. The area burned here was part of the 1988 Shoshone fire. Further ahead in the Lewis River Canyon, the steep margin of the Yellowstone Plateau is deeply incising.

STOP 4–4. JACKSON LAKE PICNIC AREA

This stop (Fig. 17) is on the margin of the Snake River delta with its abundant archeology. Across the reservoir from here, the bench about 1,500 ft (500 m) above the lake is mantled by hummocky moraines of the last glaciation, deposited by the Snake River lobe from Yellowstone. The terminus of this glacial lobe was 13 mi (20 km) to the south beyond the south shore of Jackson Lake, and much of the terrain in between has drumlin–like glacial morphology. Glacial scour excavated Jackson Lake, which is 400 ft (120 m) deep with as much as 400 additional feet of sedimentary fill (Smith et al., 1993).

About 100 years ago, a dam was built to raise the lake 35 ft (10m). Before damming, the Snake River had built a delta southward past this stop. Post–glacial scarps on the Teton fault are locally apparent near the base of the Teton Range. Because the outlet of Jackson Lake is farther away (east) of the Teton fault than any other part of the lake, fault offsets would result in submergence of shorelines. Events here in the natural delta area are considered to represent the last two offsets of the Teton fault: (1) transgression of the lake up onto the riverine delta and construction of beaches on top of the scroll–work pattern of the riverine delta; this can be explained by ~1 m downfaulting of the delta relative to the lake outlet dated by habitation of the beaches ~1.6 cal ka B.P., and (2) formation of ridge–and–basin terrain apparently by shaking during fault offset, followed by Native American occupation of these ridges about 4 ka B.P..

Figure 17. Shaded relief map of Jackson Hole area showing location of stops on Day 4, 5, and 6. Locations of boundaries of Bull Lake and Pinedale glaciations transferred by eye and not accurate; see Pierce and Good (1992) for better rendition. Ice margins labeled as follows: BL– Bull lake, BR– Burned Ridge, HP– Hedrick Pond, JL– Jackson Lake.

Overview of Jackson Hole archeology

Jackson Lake has been the focus of intense archeological investigations (Reeve et al., 1979), including work associated with reconstruction of the Jackson Lake dam (Connor 1998). Although archeological sites experienced up to 1 m reservoir wave erosion, Boeka (2002) found only minimal horizontal movement of artifacts, thereby allowing human behavior to be confidently interpreted from the spatial patterning of artifacts

The interdisciplinary nature of the Jackson Lake investigations provided for the development of a model of landform evolution and subsequent human occupation (Pierce et al., 1998). Earliest occupation dates from the early Paleoindian Period (14.0 to 10.3 cal ka B.P.). Later Paleoindian complexes (10.3–7.6 cal ka B.P.), which are characterized by various lanceolate projectile point styles, are common in both Jackson Hole (Connor 1998) and the Yellowstone Plateau (Cannon and Hughes 1997), but few stratified sites have been found. A broad–based subsistence economy is suggested for this time period.

The earliest radiocarbon date (6.7 cal ka B.P., 5850 ± 50 14C yrs. B.P.) from the shores of Jackson Lake coincides with Archaic cultures of the mid–Holocene (7.6 to 1.9 cal ka B.P.) character-

ized by large side–notched and corner–notched projectile points, fired–rock earth ovens, and possible increased reliance on plant resources. Two subsistence–settlement patterns have been suggested for this period—one focused on bison which is common in the intermountain basins and on the plains, the other a mountain–oriented pattern for which mule deer and mountain sheep were the primary prey (Frison 1992). Connor (1998) suggests a pattern of change in subsistence–settlement systems occurred around 6 cal ka B.P.

The late Holocene of Jackson Hole (2.8 to 1.9 cal ka B.P.) mirrors a trend of increased site quantity and density seen throughout the Central Rocky Mts. (Connor 1998) and Northwestern Plains (Frison, 1992). Corner–notched points, such as Pelican Lake and Elko, are common during this period, as are smaller side– and corner–notched points representing bow–and–arrow technology. Subsistence patterns in the mountains appear to be a continuation from earlier times with mule deer and mountain sheep the primary prey. Fire pits, stone circles, and grinding stones are also common. However, Cannon (2001) has suggested that subsistence patterns in Jackson Hole may be different with bison being a more common prey. During the last 200 years, population seems to have decreased (Connor 1998). While Wright (1984) suggests collapse of the system around AD 1600, Conner (1998) suggests decreased temperatures associated with the Little Ice Age may have reduced seasonal use.

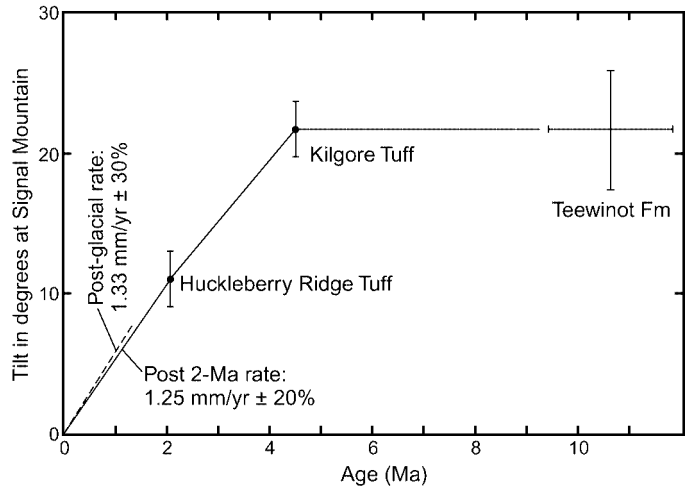


Figure 18. Tilting of late Cenozoic units in the Signal Mountain area westward into the Teton fault showing uniform high rate of tilting after about 5Ma. Dashed line is tilt rate reconstructed from post–glacial offset.

Pinedale Glacial Phase	NW Snake River glacial lobe	Pacific Creek glacial lobe	SE Buffalo Fork glacial lobe
Burned Ridge (oldest)	not even conflent	smaller	largest
Hedrick Pond	smaller	largest	not even conflent
Jackson Lake (youngest)	large	large	not even conflent

Figure 19. Changes in the relative size of the three glacial lobes flowing into northern Jackson Hole during the three phases of the Pinedale glaciation. Arrows indicate migration of dominant lobes from southeast to northwest towards direction of moisture supply.

STOP 4–5. WILLOW FLATS VIEWPOINT

Willow Flats is part of a huge, post–glacial fan of Pilgrim Creek where natural sub–irrigation of the lower part of the fan fosters a willow habitat ideal for moose. On and near Signal Mt., the dip of dated volcanic units suggests that tilting into the Teton fault started about 5 Ma and has been relatively constant since then (Fig. 18). The 4.45–Ma Kilgore tuff flowed into this area from its source caldera on the west side of the Teton Range (Lisa Morgan, oral communication, 2002) indicating the Teton Range was no barrier then. The high, steep, imposing front of the Teton Range is another reflection of this young, ongoing formation of the range.

This stop is within the Jackson Lake phase of the Pinedale glaciation (Fig. 19). East of here is a "two–sided" kame terrace between Christian Pond on the west and Emma Matilda Lake on the east that formed between the Snake River glacial lobe on the west and the Pacific Creek glacial lobe on the east and defines the exact the relation between the two lobes at a given time.

Earlier in the Pinedale history during Burned Ridge time (Fig. 19), the Pacific Creek lobe flowed westward through this area and excavated a deep basin that now extends from Pacific Creek westward up the Snake River, beneath the present dam, and under the lake north of Signal Mountain. Drill holes beneath the dike section of the dam encountered 600 ft of silty lake sediments now filling this glacial trough. During an earthquake on the Teton fault, the fill of this trough, particularly the unconsolidated sand beneath the dam and dike, might liquefy and result in failure of the dam. The dam and dike were recently rebuilt and liquefiable material firmed up by both dynamic compaction and in–situ manufacture of concrete columns by injection of cement into long, rotating augers.

STOP 4–6. SNAKE RIVER OXBOWS

This is a popular viewing area of waterfowl, moose, otter, and raptors. At the edge of the meadows east of here is a barrier beach of a late glacial lake at an altitude of 6800 ft (2072 m), 65 ft (20 m) above the natural level of Jackson Lake. The lush, flat meadows below this barrier beach are on the bottom of the associated lake. The water in the Oxbows reaches 15 ft (5 m) in depth.

This deep, tranquil stretch of the Snake River may result from back tilting on the Teton fault. We are above a deep bedrock trough excavated in Burned Ridge time. The Snake River established its present course by following the lowest post–glacial terrain from this Pacific scour basin into the Triangle X scour basin. Continue south to Jackson.

DAY 5 JACKSON HOLE

STOP 5–1. BULL LAKE GLACIAL TERRAIN

Drive east from Jackson through Elk Refuge to the viewpoint at the Curtis Canyon Campground. This viewpoint is in glacial terrain of Bull Lake age and overlooks Jackson Hole, which was filled by glacial ice during the Bull Lake glaciation. The strong, southwesterly grain of the local topography (Fig. 20) is the result of both the southwest flow of Bull Lake ice, whose upper limit is shown by till and erratics 1,200 ft (360 m) vertically upslope from this stop, and multiple ice–marginal, meltwater channels 30–300 ft (10–100 m) deep. The upper limit of Bull Lake deposits descends southward and at Jackson is at the top of the Snow King chair lift. Bull Lake terminal moraines and erratics drape the middle and lower slopes of the mountains at the southern end of Jackson Hole, 30 miles (50 km) south of the Pinedale terminal moraines.

The view from here includes: (1) Burned Ridge moraines and outwash, overlapped downvalley by successively younger Hedrick Pond outwash on antelope Flats and Jackson Lake(?) outwash at the airport, (2) Blacktail Butte and the hills of the National Elk Refuge overridden by Bull Lake ice, with the loess–mantled Bull Lake fan of the Gros Ventre River southwest of the Elk Refuge hills, (3) below that, the large, Pinedale, outwash fan of Flat Creek, (4) Pinedale moraines at the mouths of seven canyons along the front of the Teton Range, (5) the Teton fault, forming a scarp on the lower part of the range and offsetting the lateral moraines, (6) Gros Ventre Buttes streamlined by overriding Bull Lake glacier, and (7) Phillips Ridge, above these buttes on the far side of Jackson Hole, where the Bull Lake glacial limit is at 8,100 ft (2,470 m).

If the large glaciation filling all of Jackson Hole was pre–Bull Lake, the local valley glaciers would be expected to display Bull Lake and Pinedale moraines. Only Pinedale moraines are recognized for the seven Teton valley glaciers or for the several valley glaciers from the Gros Ventre Ranges, such as Sheep Creek just east of this stop (Fig. 20). As we will see later in the trip, the Bull Lake glaciation near West Yellowstone is dated 140 ka by combined obsidian–hydration and K–Ar dating techniques. For the greater Yellowstone glacial system, Bull Lake moraines are far beyond Pinedale ones to the west (West Yellowstone), southwest (Ashton), and south (Jackson Hole), but not recognized to the north and east.

Where slopes are gentle, Bull Lake deposits are mantled by several meters of loess with a well–developed, buried soil near its base. For a section 10 miles (15 km) south of Jackson, Figure 21 shows loess, buried soils, and ages on Bull Lake outwash terrace 400 ft (120 m) above the Snake River and about 2.5 mi (4 km) inside the Bull Lake terminus. Return to Jackson and head north.

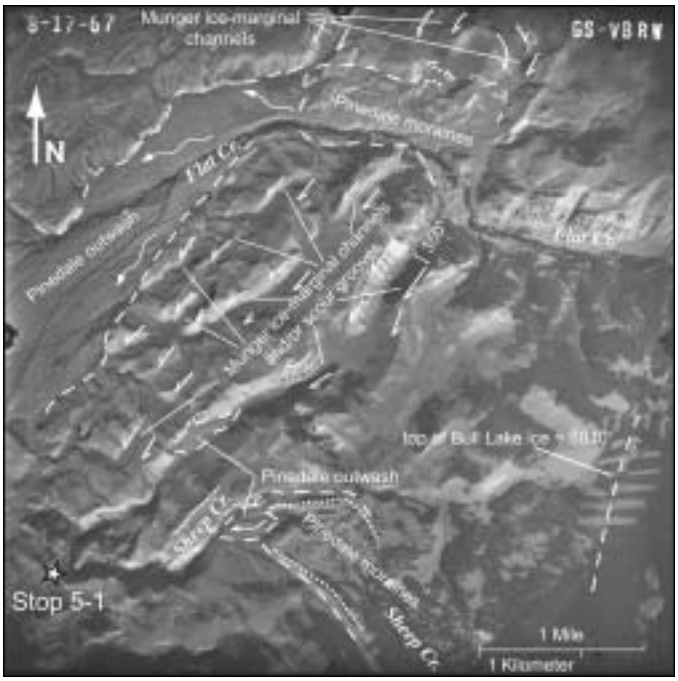


Figure 20. Moraines and meltwater channels on the eastern side of Jackson Hole from the Bull Lake glaciation that filled Jackson Hole. Also shown are Pinedale moraines and outwash from local mountain–valley glaciers.

STOP 5–2. TERRACES NEAR THE AIRPORT

Park on the road into an old gravel pit (now a NPS firing range) and climb up onto loess–mantled terrace (Fig. 22. This terrace is a Gros Ventre River deposit and a candidate for an early Wisconsin age. A soil pit 11 ft (3.3 m) deep encountered only a reddened zone near the base of the loess and thin carbonate coats in the underlying gravel, indicating only a weakly developed, buried soil.

A graben offsets this terrace 1 mi. (2 km) northeast of here. The scarp at this stop has also been mapped as a postglacial fault (Love et al., 1992), but Pierce considers it a fluvial scarp because (1) alluvial fans deposited across the scarp show no evidence of offsets, whereas multiple offset events less than 10 ft (3 m) would be expected to create this 60 ft (20m) high scarp, (2) the scarp is scalloped by channels at the margin of the lower terrace, perhaps by floods, and (3) the scarp is more linear, gently arcuate, and constant than typically seen in normal fault scarps and truncates abruptly to the south.

Beneath the terrace of Jackson Lake (?) age near the airport, well logs note clays near 100 ft (30 m) and resistivity surveys confirm a highly conductive material. These may be lake sediments deposited in a lake scoured out during the Bull Lake glaciation.

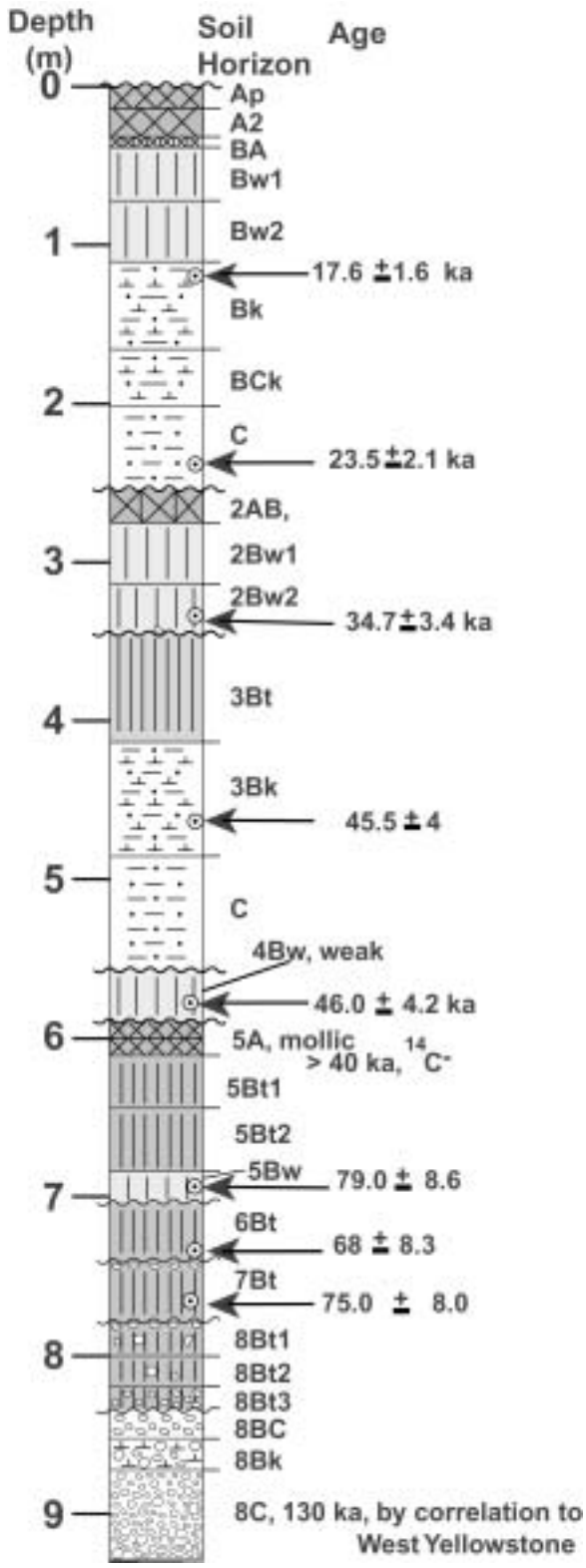


Figure 21. Loess, soil, and provisional TL ages in Porcupine Creek section, southern Jackson Hole. Symbols as follows: stronger development of B–horizons– darker shading and closer spaced vertical lines; A–horizons– cross hatch and shaded, carbonate– horizontal dash with vertical tic; and unweathered loess– dash dot pattern and no shading

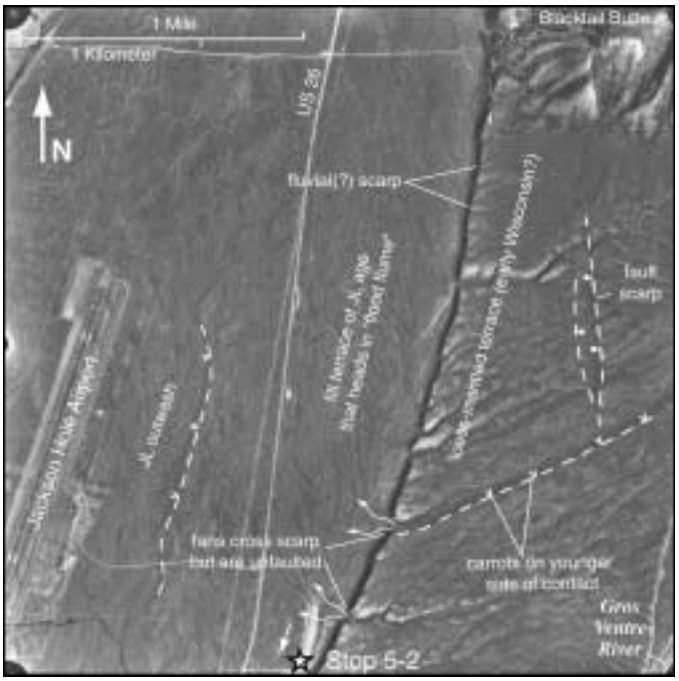


Figure 22. Snake River outwash terrace of Jackson Lake age around the Jackson Hole Airport undercutting loess–mantled Gros Ventre River terrace of early Wisconsin? age.

STOP 5–3. TETON POINT TURNOUT

The highest peak is Grand Teton, altitude 13,771 ft (4197 m), south of which are Middle Teton, South Teton, and Buck Mountain. North of Grand Teton are Mt Owen and Teewinot. The Snake River floodplain is down in the cottonwoods and is characterized by branching and continuously changing channels.

For locations that follow, the clock position of 12:00 is the upvalley trend of the bluff on which we are standing. This stop is near the onlap edge of the Spaulding Bay outwash fan (11:00) of Hedrick Pond age onto the Antelope flats outwash fan (1:00) of Burned Ridge age (Fig. 23). The Snake River follows the approximate boundary between these two large outwash fans of different ages and discharge points. The inset terrace is of Jackson Lake age. We call it a flood flume, for it displays large–scale bedforms produced by flood waters 30–45 ft. (10–15 m) deep and ~1 mi. (1.8 km) wide, and flood bars occur up to the top of the fill terrace. Looking downvalley along the length of the Snake River, this Jackson Lake terrace converges with the Hedrick Pond terrace.

Burned Ridge moraines occur at the head of outwash between 12:00 and 1:00 and represent the youngest (BR–3) of three sets of Burned Ridge moraines. The BR–2 moraines protrude through the outwash at 1:30. Kettles where the grassy ridge at 2:00 comes down to the outwash level represent the BR–1 limit. Along the Teton front at 8:00, moraines of Pinedale valley glaciers form large mounds that encompass Bradley and Taggart Lakes. The Teton fault scarp is along the front above these moraines.

Along the southern margin of the greater Yellowstone glacial system in Pinedale time, three glacial lobes poured into northern Jackson Hole (Fig. 19). The dominance of these lobes shifted through time as follows: Buffalo Fork lobe (1:00–2:00) from the Absaroka Range, Pacific Creek lobe (12:30) from the Two Ocean Plateau, and Snake River lobe (12:30), from the Yellowstone Plateau. This northwestward migration is towards the storm conduit of the Snake River Plain.

STOP 5–4A. BURNED RIDGE MORAINES

Turn right on Lost Creek road and depart busses at next sharp right turn. Hike north past a kettle and up onto the youngest of three Burned Ridge moraines (BR3; Fig. 24). Erratics are mostly Pinyon–type quartzite, with minor Tensleep sandstone, Paleozoic limestone, Absaroka volcanic rocks, Mesozoic sandstone, rhyolite tuff, and Precambrian crystalline rock.

To the north is a drop–off 100–200 ft (30–60 m) high that represents the cast of Burned Ridge ice, the front of which was largely buried by outwash. The middle Burned Ridge stand (BR2) is represented by moraines mostly buried by outwash 1.5 mi (2.5 km) south of here (Fig. 24), and the oldest stand (BR–1) is represented by kettles in outwash 2.5 mi (3.8 km) south of here. The Buffalo Fork lobe deposited these moraines and outwash. The outwash terrace rises to the northeast (Fig. 24) and heads at the ice–marginal position now occupied by Spread Creek. Outwash channel patterns (Fig. 24) show stream flow was to the west. The original depositional slope is an order of magnitude greater than the estimated tectonic tilting into the Teton fault.

The glacial deposits along the southern margin of the greater Yellowstone glacial system are predominantly outwash gravel of Pinyon–type quartzite. The great thickness of outwash gravel results from glacial erosion of weakly indurated quartzite conglomerate of upper Cretaceous to Paleocene age. Great quantities of these quartzite clasts were glacially transported to the head of outwash where they overloaded the outwash system and deposited outwash 200 ft (60 m) or more thick. These thick outwash accumulations bury much of the associated glacial moraines. Walk from moraine to the highway and carefully cross to Snake River overlook (Fig. 24) to where vehicles are parked.

STOP 5–4B. SNAKE RIVER OVERLOOK

This stop is on west–sloping, Burned Ridge outwash (Fig. 24). Across the Snake River, the forested landform is Burned Ridge, a moraine mantled or modified by Hedrick Pond deposits. South of Burned Ridge is outwash of Hedrick Pond age of the Spaulding Bay outwash fan. It has distinctly fresher channel morphology on aerial photographs than Burned Ridge outwash.

About 100 ft (30 m) lower than the Hedrick Pond outwash and to the north of Burned Ridge is outwash of Jackson Lake age of the Potholes channelway, where ice blocks of Hedrick Pond age were buried by Jackson Lake outwash. Jackson Lake out-

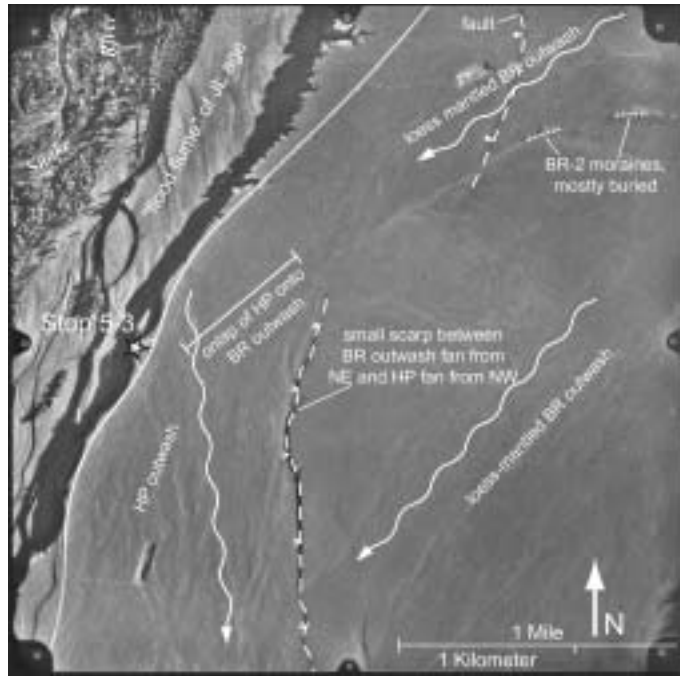


Figure 23. Onlap of outwash fan of Hedrick Pond age from north onto outwash fan of Burned Ridge age from northeast near Teton Point turnout.

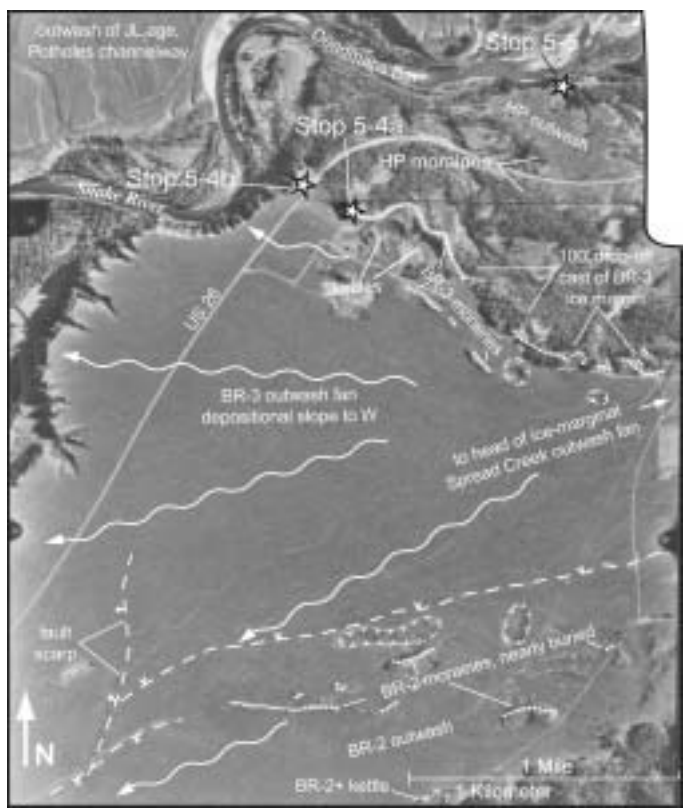


Figure 24. Outwash and moraines and of Burned Ridge and Hedrick Pond age near Snake River Overlook.

wash prograded southeast into the Triangle X2 lake. The outlet to this lake was here at the head of the "flood flume." Repeated flooding lowered the outlet from the ~6840 ft (2085 m) Hedrick Pond spill point to 6720 ft (2048 m), the present inset terrace level. The floods may be from release of glacially dammed lakes along the near STOP 5–6. The eroding bluffs from the Snake River up expose: gray gravel mantled by loess (?), yellow gravel (from the northeast?), gray gravel of Jackson Lake age.

STOP 5–5. HEDRICK POND SECTION AND PALEOECOLOGY

Turn left on the road to Deadman's Bar and park in the first parking area. Hike north across Hedrick Pond moraines, intersect the Snake River Bluffs on Hedrick Pond outwash, and walk up the Snake River to where it impinges on the bluff, creating a fresh, high exposure (Fig. 24). The following section is exposed (Pierce and Good, 1992):

Ft	(m)	Description
16	(5)	Gravel with scattered lenses of flow till. Bedded, but beds discontinuous; surface has kettles
6	(1.8)	Diamicton, flow till
11	(3.4)	Gravel
11	(3.4)	Sand, cross – bedded
2.5	(0.8)	Gravel
80	(24.1)	Diamicton, muddy sand matrix with scattered pebbles, locally wispy silt beds. The diamicton is rich in fines and poor in gravel and was probably eroded from lake sediments that accumulated in the Triangle X–1 lake that extended from the Snake River overlook to Moran Junction and up the Buffalo Fork valley from there. Diamicton deposited near ice as a lacustral till or flow till.

Kettles occur at this stop, as well as across the highway at the Hedrick–Pond core site (Fig. 25). An open forest of lodgepole pine and Douglas–fir and sagebrush surrounds Hedrick Pond. A 22.0–ft–(6.71 m)–long core yielded a basal date of ~20.3 cal ka B.P. (17,160 ± 210 ^{14}C yrs. B.P.). The pollen record indicates a period of alpine meadow and shrub communities prior to 13.5 cal ka B.P. (11,500 ^{14}C yrs. B.P.). Juniper (probably *Juniperus communis*) was the first conifer to become widespread in the deglaciated region. Spruce parkland developed at ca. 17.3 cal ka B.P. (14,750 ^{14}C yrs. B.P.), suggesting that spruce was growing close to the ice margin. Fir and pine (probably *whitebark pine*) were present with spruce in a mixed conifer forest between 15.3 and 10.5 cal ka B.P. (12,900 and 9500 ^{14}C yrs. B.P.). Lodgepole pine and Douglas–fir were abundant in the early Holocene ca. 10.5–6.3 cal ka B.P. (9500–5500 ^{14}C yrs. B.P.), implying that conditions were warmer than at present. In the last 6,000 years, spruce and fir reestablished near the site, compatible with the present mesophytic communities and a concomitant decrease in fire frequency (Whitlock, 1993). Figure 26 shows the five stages of the postglacial vegetation history in southern Yellowstone and Grand Teton National Parks.

Pull over before reaching the Triangle X Ranch turnoff. The white sediments are recycled from the Teewinot Formation and deposited as deltas in the Triangle X–2 lake. Deltas from 6,800 to 6720 ft (2,073–2,048 m) altitude represent lowering of lake outlet. Farther north on highway, Spread Creek is actively depositing gravel resulting from erosion along this ice marginal drainage. At Moran Junction, bear right on US 285 and where the highway the highway leaves the flood plain and climbs up onto a moraine, turn right and take the two–track road to fence.

STOP 5–6. ADVANCES UP BUFFALO FORK

During Burned Ridge time, the Buffalo Fork lobe covered nearly all the terrain visible up Buffalo Fork and was more than 1,500 ft (500 m) thick here. It excavated a scour basin probably at least as deep as 100 ft (30 m) drill holes in unconsolidated material. This stop is on a moraine of the Pacific Creek lobe that advanced *up* the valley of Buffalo Fork in Jackson Lake time. Ridges and depressions in the moraine arc up the valley of Buffalo Fork and lake sediments occur on the upvalley side from these moraines (Fig. 27). Across the valley of Buffalo Fork and about 500 feet (150 m) higher, Hedrick Pond moraines also descend upvalley, revealing that the Pacific Creek lobe pushed into the deglaciated Buffalo Fork valley. On this side of the valley, the Pacific Creek lobe scoured south into the ice–free lower Buffalo Fork (Fig. 27). Similarly in the next drainage to the south on the north slope of "Spread Creek hill," Hedrick Pond moraines slope eastward towards the source of the Buffalo Fork lobe. Sudden releases of glacially dammed lakes from Buffalo Fork valley are the favored source for the floods down the "flood flume". Return to highway and continue east.

Pull Over as you descend from these moraines to view lake sediments overlying kame gravels in roadcut on north side of road.

STOP 5–7. THE BUFFALO FORK LOBE

Park in the large area north of the road, carefully cross the road, and climb up onto the moraines. We are on kettled moraine built across the valley of Blackrock Creek and that rises to the north across the highway. The age of this backfill moraine is Hedrick Pond, forming after the Buffalo Fork lobe had receded from its maximal size in Burned Ridge time. About 0.3 miles (1/2 km) south of here, Blackrock Creek exposes lake sediments, sand, and gravel more than 100 ft (30 m) thick.

During full–glacial, Burned Ridge time, the Buffalo Fork lobe headed both north of here in three large valleys draining the Absaroka Range as well as in the Blackrock Meadows just east of here. Evidence for a Pinedale glacier filling Blackrock Meadows consists of sharp–crested Pinedale moraines 500 ft (150 m) above the Meadows and fresh glacial striations with rat tails demonstrating flow westward from of Blackrock Meadows. This ice margin connects along the trend of the ice marginal Spread Creek with the Burned Ridge moraines on the floor of Jackson Hole. Soil profiles are weakly developed, morainal topography is rela-

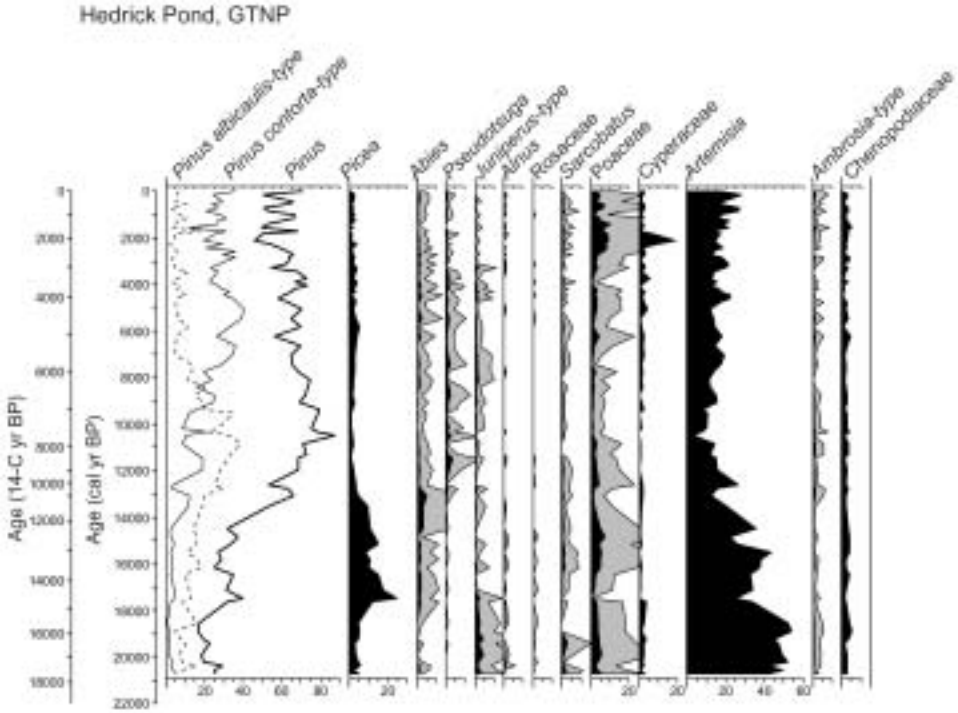


Figure 25. Pollen percentage diagram for Hedrick Pond. Shaded curves are 5X exaggeration (from Whitlock, 1993).

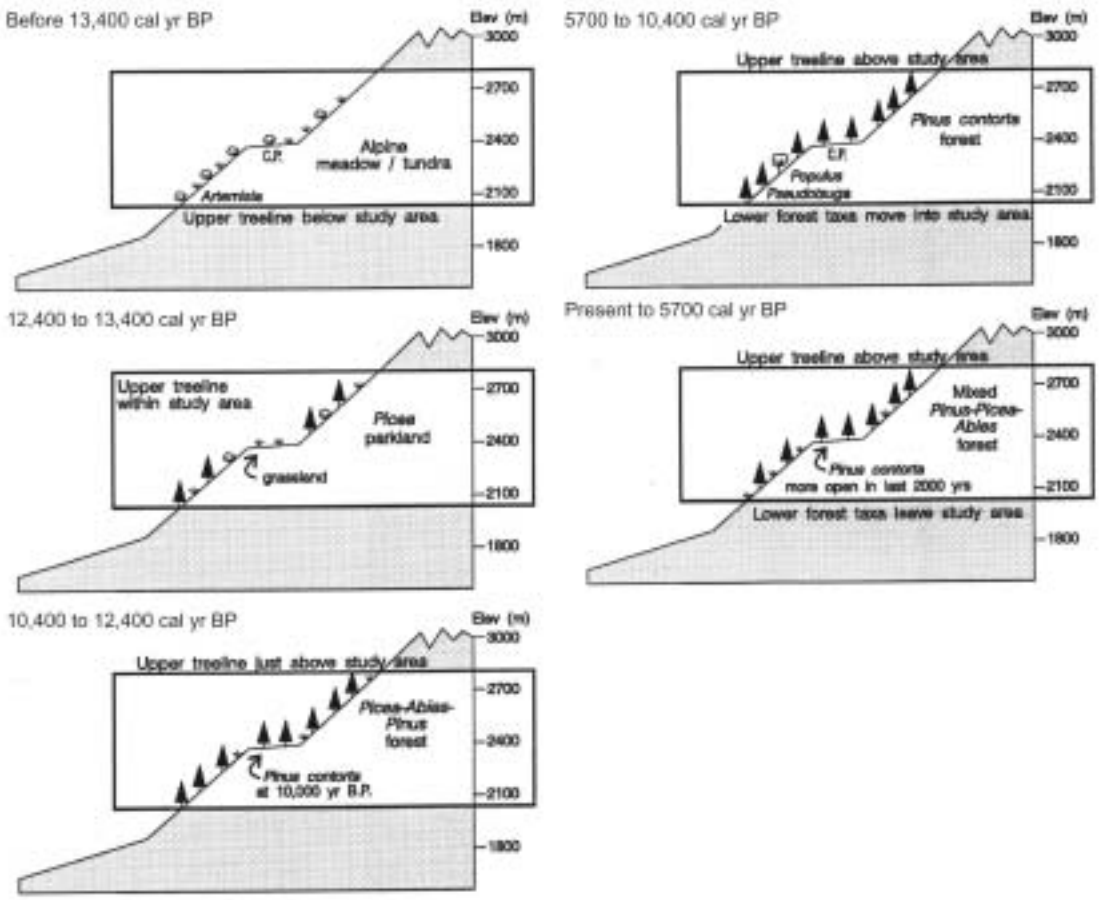


Figure 26. Vegetation history of the montane forest zones in the Yellowstone–Grand Teton region (based on Whitlock, 1993).

tively fresh, and weathering rinds are thin (0.1 mm), consistent with a Pinedale age.

Buffalo Fork flows in a deep valley 2 miles (3 km) north and 1,500 ft (460 m) below here. Blackwelder (1915) attributed this 1,500 ft high bench to the Blackrock erosion cycle and the deposits on it to the "Buffalo" Glaciation, inferring the height above drainage indicated a pre–Bull Lake age. But this height above the drainage here simply reflects the thickness of the Buffalo Fork lobe.

The Teton Range is about 30 miles (45 km) to the west and part of the reason for this stop is to show that the major glaciers did not come from the Teton Range. One mile (1.7 km) west of here, the Togowotee Lodge fault dams a sag pond on its east side (Love and Love, 1997, p. 15). The post–glacial scarp is 30 ft (10 m) high, 40º steep, and part of a set of fault scarps that are recognized for 2 miles (3 km) to the south, including a related fault that offsets Bull Lake moraines by 70 ft (20 m). This new faulting is part of an overall pattern of active faulting associated with northeast migration of the Yellowstone hotspot (Pierce and Morgan, 1992).

About 70 miles (110 km) farther east on the highway is Blackwelder's (1915) type Bull Lake Glaciation. Cosmogenic ages by Phillips et al. (1997, primarily ^{36}Cl) are >130 ka for older Bull Lake moraines, and 95–120 ka for younger Bull Lake moraines. But new Uranium series ages for carbonate coats on pebbles are 167 ± 6.4 ka and 150 ± 8.6 ka for older (WR4) and younger (WR3) Bull Lake terraces (Sharp et al., in press). In the same area, a high terrace dated by the 0.64 Ma Lava Creek ash increases in altitude 100 feet (30 m) more toward Yellowstone than does the 150 ka Bull Lake "Circle terrace." Pierce and Morgan (1992) interpret this 100 ft increase to represent uplift and tilting away from the Yellowstone hotspot. Continue east but turn right on Flagstaff Road.

Pull over and view Blackrock Meadows where glacier 500 ft (152 m) feet thick formed in Burned Ridge time. This area was snowier in Buffalo Fork time than in Hedrick Pond time, apparently due to the progressive northwestward buildup of ice towards the Yellowstone Plateau. Continue down gravel road to Lily Lake along the general trend of the ice–marginal Spread Creek

STOP 5–8. LILY LAKE PALEOECOLOGY, BURNED RIDGE ICE MARGIN, FAULTING

Lily Lake is one of several kettles in the sediment–filled valley of Spread Creek. Spread Creek has incised a canyon 400 feet (120 m) deep to the south of here. South of Spread Creek, north–flowing drainages were blocked along the southern margin of Buffalo Fork lobe and inwash kame gravel hundreds of feet thick accumulated. To the north, Burned Ridge moraines drape around Baldy Mountain. A north–south fault offsets these moraines 15 ft (5 m), which is at least half the total offset on this fault. This "new" fault is on the leading margin of the northeastward progression of faulting associated with the Yellowstone hotspot.

Figure 27. Lower Buffalo Fork area showing backfill and damming by Pacific Creek lobe in Jackson Lake (JL) and Hedrick Pond time.

A sediment core from the fen adjacent to Lily Lake has a basal date of ~18.9 cal ka B.P. (16,040 ±220 ^{14}C yrs. B.P.). The pollen record resembles that of Hedrick Pond, with an early period of alpine meadow, juniper, and birch (*Betula*), followed by the development of spruce (*Picea*) parkland and then subalpine forest with *Picea*, *Abies*, and *Pinus albicaulis* in the late–glacial period. The early Holocene featured an expansion of lodgepole pine, Douglas–fir and quaking aspen during the summer insolation maximum (Whitlock, 1993). Return to Jackson and possible talk in evening on Yellowstone hotspot.

DAY 6. JACKSON TO OLD FAITHFUL

STOP 6–1. TETON FAULT SCARP

Drive through Moose to the String Lake area and park near the rest room facility. Walk to the trailhead at footbridge across String Lake outlet (Fig. 28. Note the bouldery, outer Jenny Lake moraine that descends to the outlet and is buried east side of outlet by outwash from the Spaulding Bay terminus. From the footbridge, hike 0.3 miles (0.5 km) along the Jenny Lake moraine to where a trail comes in from the right, turn right (north) on the trail toward Indian Paintbrush Canyon walking over forested Jenny Lake moraines. Enter an open, avalanche area where in 1986, snow and trees from a large avalanche littered this area and extended out into String Lake. The steep slope above the trail was formed by a cascade of morainal debris coming to rest at its angle of repose after being delivered

to the snout of a Pinedale glacier in the Laurel Lake basin. The trail enters the forest and then comes into a meadow formed on a steep alluvial fan. Near the far side of the fan, look upslope to see the oversteepened fault scarp to the right of the gully at the fan head and leave trail and hike up the fan to the fault scarp. *(If you hike on the trail past the large, isolated Douglas–firs, you have gone too far).*

At the base of the fault scarp is a partially filled graben. The fault scarp is 113 ft (34 m) high, as steep as 39º, and has surface offset of 63 ft (19 m) (Gilbert et al., 1983). Pinedale moraines of steep slab–glaciers occur just above the fault scarp and the scarp offsets debris from this slab glacier estimated to have receded about 14 ka. This age and offset yield a high rate of 1.4 mm/yr.

The lakes along the Teton front are, from south to north, Jenny, String, Leigh, and Jackson. String Lake is a shallow, pan–like lake only about 5 ft (2 m) deep, which on sunny summer days is the warmest lake around for swimming. In Jackson Lake time, a large outwash stream headed at the upper end of String Lake. The outlet is 0.2 mi (0.6 km) farther away from the Teton fault than is the west lake shore. Tilting of Jackson Hole into the Teton fault combined with additional strands of faulting noted east of the main fault would submerge the western part of the lake area more than its outlet, thus backflooding this glacial stream course and converting it into a lake.

Given the clock direction to Mt. Leidy at 12:00, some key features are: 1:30–the southern margin of Snake River–Teton lobe at the narrows of String Lake; 11:00–the head of Spaulding Bay outwash fan of Hedrick Pond age with incised channels of Jackson Lake age, and on the far side of this fan, forested landform of Burned Ridge age draped with moraines of Hedrick Pond age; 12:00–across the Snake River at the far side of Antelope Flats meadows, the Burned Ridge outwash fan heading on the ice–marginal Spread Creek. The source areas for three glacial lobes and interlobe areas are: 10:00–12:30–Buffalo Fork lobe heading in the Absaroka Range; 10:00–11:00–highlands between Buffalo Fork and Pacific Creek lobes including Gravel Mountain and Mt Randolph; 10:00–Pacific Creek lobe; 9:00–10:00–Pilgrim Creek highlands that were invaded and surrounded by ice, but not traversed by ice, and 8:00–the Snake River lobe. The Gros Ventre slide of 1925 is at 1:30, and Sheep Mt., with the upper limit of Bull Lake ice near the upper treeline, is at 2:00. Return to busses parked near toilet facility at String Lake.

OPTIONAL STOP 6–1.5 CATHEDRAL GROUP OVERLOOK

The Teton fault scarp offsets moraines and debris of Pinedale slab–glaciers. Tilting or this surface into the Teton fault is much less than the original depositional slope. To the north are forested moraines of Jackson Lake (?) age containing large boulders from the Teton Range (Fig. 28). Quartzite cobbles are common in the outwash in front of the moraine, but are uncommon in the moraine itself, indicating glacial flow from the Tetons was fronted by outwash from the Spaulding Bay terminus.

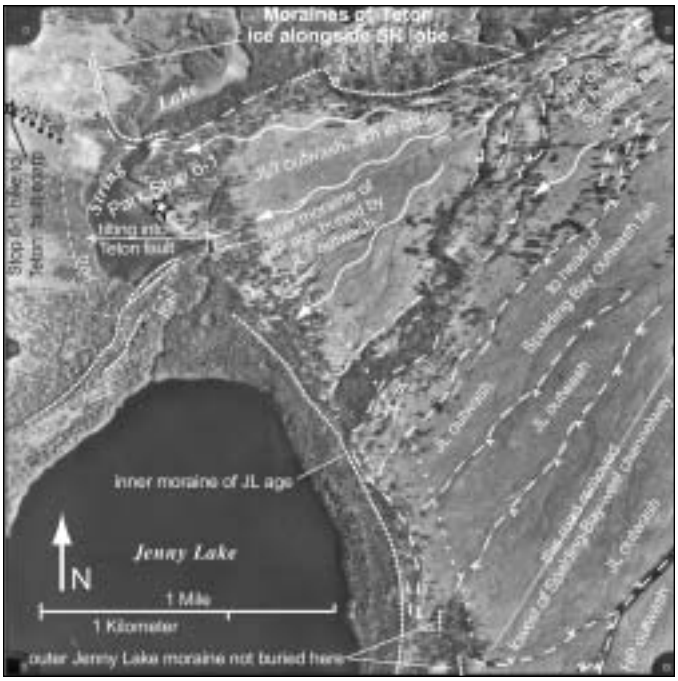


Figure 28. String Lake area showing outwash from greater Yellowstone glaciation and local moraines enclosing Jenny Lake.

STOP 6–2. HEAD OF SPAULDING BAY OUTWASH FAN

Turn into off–road parking area. We are on the Hedrick Pond level that heads at the Spaulding Bay terminus. Note the same level to the east across the channelway incised in Jackson Lake time. The outwash fan extends northward into “thin air” and to understand its origin, one needs to visualize the snout of a large glacier in the vacant space extending hundreds of feet above Spaulding Bay. This outwash fan extends eastward almost to the Snake River Overlook and southward to Moose. Much of the gravel that forms this fan was excavated from the deep, glacial trough of Jackson Lake. The head of this fan was incised in Jackson Lake time and both the original fan and its incised channels became relict features when the Snake River lobe pulled back from the Spaulding Bay terminus and outwash drained farther east. Across the channelway at the head of outwash are forested moraines of Hedrick Pond age that trend easterly. These moraines have a tilted–slab–like form and may be a glacially thrust mass.

STOP 6–3. POTHOLES CHANNELWAY VIEWPOINT

Park in Mount Moran scenic turnout (or pullout 0.4 miles to the east) and hike north across outwash flats to the top of the bare slope (Fig. 29). After the Burned Ridge recession and prior to Hedrick Pond time, the Triangle X–1 lake occupied the glacial scour basin that extended 20 miles (32 km) from Deadman’s bar up the Snake–Buffalo Fork drainage system. We are looking down the Potholes channelway, named for the abundant glacial

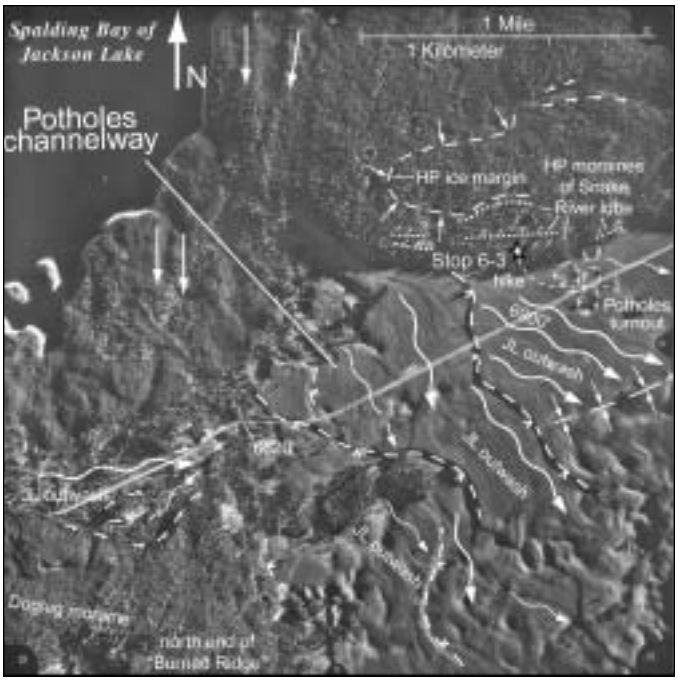


Figure Figure 29. Head of Potholes channelway where ice–blocks of Hedrick Pond age were buried by outwash of Jackson Lake age.

kettles that represent ice blocks of Hedrick Pond age buried by outwash of Jackson Lake age (Fig. 29). The kettles form alignments that indicate several ice–margin positions of the Snake River lobe of Hedrick Pond age. The highest outwash terrace along the left side of the Potholes channel is of Jackson Lake age and has linear trends of kettles that suggest the west margin of buried blocks the Pacific Creek lobe of Hedrick Pond age.

Looking down The Potholes channelway and across the Snake River, the higher alluvial bench is Burned Ridge (BR–3) outwash and moraines, which climb to the left where the ice–marginal Spread Creek followed the southern margin of the Buffalo Fork lobe. The Hedrick Pond bench is 100–200 feet (30–60 m) below the Burned Ridge bench and climbs more gently to the left. This gentler, ice–margin gradient may result from the Hedrick Pond advance moving across Triangle X–1 lake sediments with basal sliding aided by shearing in the lake sediments. The biggest difference in the extent of ice between Hedrick Pond and Jackson Lake time is that the Triangle X lake basin was ice filled in Hedrick Pond time, but in Jackson Lake time only its northeast margin was occupied by the Pacific Creek lobe and only its north margin was occupied by the Snake River lobe. The escarpment at this stop is one of a set of kame moraines that descend to the east. The distal slope of the moraine is a gently sloping (5–8º) bench rather than a ridge.

The glacial geology of the Burned Ridge landform is puzzling. The northern half of Burned Ridge has Hedrick Pond moraines of the Snake River lobe draped on its east side and part of its west side. Hedrick Pond ice advanced through a breach in Burned Ridge through a gap from 0.9 and 1.5 mi (1.4–2.3 km) this side of the Snake River. The margins of this breach are marked by a loop of aligned kettles on the outside of this breach and by west–sloping small moraines in the breach through Burned Ridge.

Pull over into parking area on north side of South Landing Channelway. This outwash fan of Jackson Lake age has remarkably well–preserved channel morphology with downstream branching channels. It has much fewer kettles than the Potholes Channelway.

OPTIONAL STOP 6–4A. TOP OF SIGNAL MT.

Signal Mt. was ice covered in Burned Ridge time. Pinedale moraines drape around it with the highest moraines draped as consistent with draping from a northeastern source and the lower moraines consistent for a northern source, compatible with the change through time in dominance from the Buffalo Fork lobe to the Snake River lobe (Fig. 19). The view from the top of Signal Mt. shows many aspects of the glacial geology of the floor of Jackson Hole. Also visible are the sources of the three glacial lobes: Buffalo Fork lobe and its scoured basin to the east, the Pacific Creek lobe and the scour basin to the northeast, Snake River lobe, and the Jackson Lake scour basin to the north.

STOP 6–4B. SIGNAL MT. LODGE

Between Donoho Point (northwest across the lake) and us is a deep scour basin excavated by the Pacific Creek lobe in Burned Ridge time. If offset were to occur today on the Teton fault, the western shoreline of Jackson Lake would move down more than the natural outlet at Jackson Lake dam, and when the water in the lake rose to spill at the outlet, shorelines on the west side of the lake would be submerged. During rebuilding of the Jackson Lake dam in the late 1980's, the lake was at times lowered to its natural level. From a canoe in calm water, we could see deeply enough to distinguish two submerged shorelines at depths of 3.5–4 feet (1–1.2 m) and 8–9 feet (2.4–2.7 m). Combined side–scan sonar and fathometer records gathered in 1986 by Steve Colman and Ken Pierce show shorelines east of Bearpaw Bay about 2 km east of the Teton fault at depths below the natural shoreline of about 3, 8, 15, 20, 24, 27, 31, and 36 feet (1, 2.4, 4.6, 6.1, 7.3, 8.2, 9.4, and 11 m). These shorelines are inferred to be submergence events related to downfaulting of Jackson Hole, and if so, suggest at least 8 faulting events in post–glacial time.

STOP 6–5. FLAGG RANCH

The ridge to the east is Huckleberry Mt., which is the type section of the 2.1–Ma Huckleberry Ridge Tuff (Christiansen, 2001). The tuff forms a dip slope inclined about 9º to the west. On the western skyline, the same tuff caps ridges at the northern end of the Tetons and also dips west. East dipping normal faults are responsible for these tilts and offsets (Christiansen, 2001). Thirteen miles (20 km) northeast of here on the east side of Mt

Sheridan, post–glacial scarps as high as 70 ft (20 m) represent the main postglacial faulting near the southern Yellowstone National Park boundary.

On Huckleberry Ridge, a well–defined Pinedale lateral moraine occurs at an altitude of 9,100 ft (2774 m) and defines the maximum of the Snake River lobe in Hedrick Pond time. Southward 9 miles (14 km), the ice surface descends about 600 feet (180 m) to the high bench across the lake from Stop 4–6. Continue driving to just past West Thumb.

STOP 6–6. DUCK LAKE AND CENTRAL PLATEAU

Duck Lake is a hydrothermal explosion crater of Holocene age. Another explosion crater beneath Yellowstone Lake northeast of here has lurked unnoticed on the floor of Yellowstone Lake until recently labeled Duck Lake's "evil" twin by Lisa Morgan. Duck Lake has been the site of core studies related to recent fire history and patterns of charcoal accumulation following the 1988 fires in Yellowstone (Millspaugh and Whitlock, 1995; Whitlock and Millspaugh, 1996).

Between here and Old Faithful, the road traverses rhyolite flows mostly in the 150 ka range. Evidence of glaciation, such as erratics, striated stones, and glacial till, is not easy to recognize in this rhyolite terrain. The most recognizable deposits of glacial origin are recessional kame and outwash gravel generally in low areas between rhyolite flows. A sector of the plateau icecap flowed southwest from this area to beyond the southwest corner of Yellowstone National Park. Richmond mapped about eight places on the Central Plateau as nunataks (U.S. Geological Survey, 1972a). Pierce found Richmond's upper limit of Pinedale ice resulted in basal shear stress values in the range of 0.1–0.01 bars, unreasonably low values above permeable bedrock. Pierce also found obsidian pebbles with hydration the same as Pinedale end moraines (7 μm) on the one nunatak examined.

Drive to lodging at Old Faithful and in the evening watch Old Faithful erupt and stroll through Upper Geyser Basin.

DAY 7. OLD FAITHFUL TO BOZEMAN

STOP 7–1. MIDWAY GEYSER BASIN

Cross the Firehole River on the footbridge and view the steamy Excelsior Geyser Crater and colorful Grand Prismatic Spring, which are two of the largest geothermal pools in Yellowstone. Southwest along the skyline is the faulted profile of the Mallard Lake resurgent dome, the southwest of the two resurgent domes of the 0.64 Ma caldera. These faults strike NW, trend towards the Midway Geyser Basin, and are likely to serve as conduits for thermal fluids.

STOP 7–2. MADISON JUNCTION (Rest stop)

Madison Junction is along 0.64–Ma caldera margin. National Park Mt. to the southwest is the northeast margin of the large West Yellowstone rhyolite flow that was emplaced between the Bull Lake and Pinedale glaciations. Pinedale ice almost reached the top of the mountain to the north. The road through the Madison Canyon to West Yellowstone obliquely crosses the caldera margin.

STOP 7–3. PINEDALE TERMINAL MORAINE

Park in off–road parking area on north (right) side of the road and walk up onto the Pinedale moraine of an outlet glacier that flowed down the Madison Canyon from the Yellowstone Plateau icecap. This moraine consists mostly of rhyolite from the Madison Canyon. The soils are weakly developed, except that the porous till has gray, clay lamellae translocated as deep as 7 ft. These moraines probably correlate with the Eightmile moraines north of Yellowstone and the Hedrick Pond moraines in Jackson Hole because the Yellowstone Plateau icecap was a key source in the glacial system that deposited these three end moraines. Across the Madison River from here, the end moraine complex is well expressed with more than 5 morainal positions. Outwash channels from the inner moraines trend westward through the outer moraines. The Madison Canyon moraines are at the head of the West Yellowstone obsidian sand–plain, a low–gradient, outwash fan rich in granule–sized obsidian that floors most of the West Yellowstone basin. The obsidian hydration thickness on glacial cracks is about 7 μm, yielding an age of 25 ka (Pierce et al., 1976). The hydration on Deckard Flats deposits is about 5 μm. The thickness of weathering rinds on basalt is about 0.4 mm, whereas that on Bull Lake deposits is 0.8 mm (Colman and Pierce, 1981). Continue into West Yellowstone and drive north on US 191.

STOP 7–3. OBSIDIAN SAND–PLAIN AND ITS ECOLOGY

Cross the Madison River, climb to the top of the obsidian sand–plain, and turn west onto a two–track road leading to a bluff above Madison River. The obsidian sand plain heads in the Pinedale moraines seen at the last stop. Glacial–pressure cracks on obsidian pebbles from here have hydration rinds 7 μm thick, the same as that on the end moraines. The soil with only a color B–horizon is consistent with a Pinedale age. The bedding of the obsidian sand deposit is sub–horizontal, with primary beds about 0.5 feet (15 cm) thick, capped by a silty zone a few millimeters thick. The sand plain has a remarkably low gradient of about 8 ft/mile (1.5 m/km). These and other features suggest that the obsidian sand plain was built by large sheet flow discharges carrying sand and granules in traction across a wide extend of the surface. The western part of the obsidian sand plain may be deltaic and deposited in a lake dammed by the Beaver Creek moraines just beyond the west margin of the West Yellowstone Basin. Such large discharge might result from release of water in the geyser basins, held either as glacially dammed lakes or as subglacial water. Coring for bridge abutments in the Lower Geyser Basin revealed at least three cycles of lake sediments capped with gravel, suggesting filling and sudden draining of glacially dammed lakes. The upper surface of the sand plain has

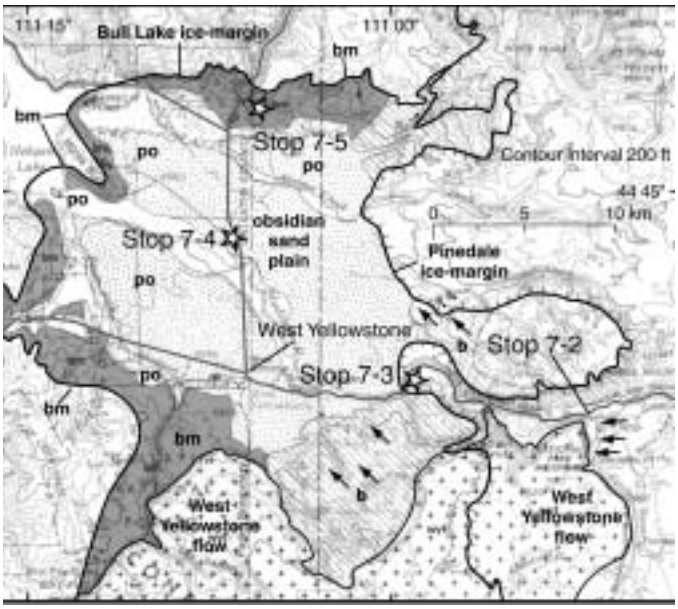


Figure 30. Map of Pinedale and Bull Lake glacial extents in West Yellowstone Basin (after Pierce, 1979, Fig. 8). bm– Bull Lake moraines, b–area glaciated by Bull Lake ice, po– Pinedale outwash forming obsidian sand plain. The West Yellowstone flow is ~120 ka and post–dates the Bull Lake moraines.

a thin topsoil of silty sand, and is excessively well drained, with an even lower water–holding capacity than rhyolite flows. The obsidian sand has the chemistry of rhyolite and thus is also low in nutrients. Such conditions allow lodgepole pine to form a climax community with Idaho fescue and antelope bitter brush (Persia tridentata) on the forest floor. The trees don't get very large here, even though most of them are more than 300 yrs. old.

STOP 7–4. BULL LAKE MORAINES

Continue north on US 191 past the Duck Creek Y and turn right into the old cemetery. Park and walk south onto a broad moraine crest in the open meadow. These Bull Lake moraines consist of glacially transported debris from the southern Gallatin Range and the Yellowstone Plateau. The moraines are broad, subdued, and capped with loess, commonly about 1.5–3 ft. (1/2–1 m) thick. A moderately developed, clayey–to–loamy soil has a high water–holding capacity. The roadcut near the turnoff shows a complex colluvial history with oxidized soil material filling concavities in the moraines. Although boulders protruding from the moraines are uncommon, boulders nearly flush with the surface are readily found. In soil pits, contortions and obliteration of soil horizons suggest solifluction has locally obliterated textural B–horizons and submerged boulders into the moraines.

The West Yellowstone Basin was nearly filled by Bull Lake ice from the Yellowstone Plateau. Bull Lake moraines from west–drape around Horse Butte in the western part of the basin. Two grassy moraines descend westward to The Narrows of Hebgen Lake and connect with the moraines near this stop. The Bull Lake terminal moraines are 14 miles (22 km) beyond the Pinedale terminal moraines of STOP 7–3 but this large separation is much less than the 30 miles (50 km) in Jackson Hole. The Bull Lake moraines are not significantly tilted northward into the Hebgen–Red Canyon fault, suggesting that few offsets of 15 ft (5 m) have occurred in the last 140 ka.

On the southern skyline is the West Yellowstone rhyolite flow that overlaps and post–dates the Bull Lake moraines. When the Bull Lake Glaciation occurred, the West Yellowstone and several more flows south of it were not yet emplaced. Hydration rinds on obsidian pebbles from moraines in the West Yellowstone Basin averaged 14.7 μm thick. Hydration was measured in thin–sections across pebble corners where pressure cracks are produced by glacial abrasion. Based on 12.3 μm hydration on the 123 ± 3 ka West Yellowstone flow and 16.2 μm on the 183 ± 3 ka Obsidian Cliff flow, the 14.7 hydration yields an age of about 140 ka for glacial abrasion prior to deposition of obsidian clasts in the Bull Lake moraines (K–Ar ages from Obradovich, 1992, processed by Christiansen, 2001 and age of West Yellowstone flow is from Obradovich, oral communication, 2002) with temperature and rate adjustments of hydration in Pierce et al. (1976.).

The Bull Lake moraines support a productive rangeland for bison and cattle, including mountain big sage, Idaho fescue, sticky geranium, cinquefoils, stickseeds (*Lappula redowskii*), and little sunflower. After absorbing water from spring snowmelt, these moraines are covered with lush, knee–high grasses and forbs with a carpet of blooming wildflowers.

STOP 7–5. HEBGEN FAULT AT CABIN CREEK

Return to the Duck Creek Y and turn right on U.S. 287 to Cabin Creek. The magnitude 7.5 Hebgen Lake earthquake (Doser, 1985) occurred on August 18, 1959. The Cabin Creek stop is near the west end of the 1959 fault scarp and offsets two terrace levels. The lower terrace is about 1–3 ka and is offset 10.2 ft (3.1 m); the upper terrace is about 11–17 ka and is offset 17.4 ft (5.3 m) (cosmogenic ages from Jerome Van Der Woerd, written communication, 2001). A trench across fault on the upper terrace exposed: (1) the 1959 colluvial wedge with a height of 8.9 ft (2.7 m), equivalent to ~17 ft (5.2 m) vertical throw on the main fault (Tm) and (2) a penultimate colluvial wedge with a height 3.9 ft (1.2 m) equivalent to 7.9 ft (2.4 m) Tm, totaling 25 ft (7.6 m) Tm. At Cabin Creek, only one pre–1959 Holocene event is documented, which occurred about 3 ka, based on radiocarbon ages and cosmogenic ages.

Continue west driving past the Beaver Creek moraines of Pinedale age that may have dammed a lake into which the obsidian sand plain was deposited. Farther west is the Madison Canyon slide triggered by the 1959 Hebgen Lake earthquake. Note the trees blown down and pointing away from the slide, indicating an air blast accompanied the slide.

Pull over to view Missouri Flats terrace sequence near highway 87 turnoff. Scott Lundstrom (1986) studied sediments,

soils, obsidian hydration, and range front faulting. He concluded this terrace 30 m above the Madison River was of Pinedale age, whereas the terrace 60–70 m on the far side of the river was of Bull Lake age.

STOP 7–6. INDIAN CREEK MORAINES AND FAULTING

A high lateral moraine is built of both Pinedale and Bull Lake deposits. The Pinedale moraines with lumpy morphology spill out beyond the high lateral. Smooth morainal ridges on the outside of the lateral appear to represent the Bull Lake. Across Indian Creek, Pinedale fan deposits are offset 13–20 ft (4–6 m) (Ruleman, 2002).

Continue north on US 287 through Ennis, and at Norris Hot Springs, take 289 to Bozeman. Arrive in Bozeman and end of field trip by 5:30 PM at City Center Motel. People are responsible for their own lodging.

REFERENCES

Arthur, G.W., 1962, The Emigrant Bison drives of Paradise Valley, Montana: *in*, Malouf, C., and Connor, S. eds., Symposium on Buffalo Jumps: Montana Archaeological Society Memoir No. 1, p. 16–27.

Arthur, G.W., 1966a, An Archaeological Survey of the Upper Yellowstone River Drainage, Montana, Montana Agricultural Experiment Station, Agricultural Economics Research Report No. 26, 111 p.

Arthur, G.W., 1966b, The Eagle Creek Site, 24PA301: Archaeology in Montana, v. 7(2), p. 11.

Berger, A.L., 1978, Long–term variations of caloric insolation resulting from the Earth's orbital elements: Quaternary Research, v.9, p. 139–167.

Blackwelder, Eliot, 1915, Post–Cretaceous history of the mountains of central western Wyoming: Journal of Geology, v. 23, p. 97–117, 193–217, 307–340.

Boeka, M.S., 2002, Spatial Analysis of Cultural and Natural Distributions of Artifacts at the Lawrence Site, Jackson Lake, Wyoming: Investigating Inundated Deposits: MA thesis, Department of Anthropology, University of Wyoming, Laramie, 59 p.

Bond, G., Showers, W., Cheseby, M., Lotti, R., Almasi, P., deMenocal, P., Priore, P., Cullen, H., Hajdas, I., and Bonani, G., 1997, A pervasive millennial–scale cycle in North Atlantic Holocene and glacial climates: Science, v. 278, p. 1257–1266.

Brackett, W.S., 1893, Indian Remains on the Upper Yellowstone: Annual Report of the Board of Regents of the Smithsonian Institution to July, 1892: Government Printing Office, Washington, D.C., p. 677–580.

Brown, B., 1932, The Buffalo Drive: Natural History, v 32(1), p. 75–82.

Christiansen, R.L., 2001, The Quaternary and Pliocene Yellowstone Plateau Volcanic Field of Wyoming, Idaho, and Montana: U.S. Geological Survey Professional Paper 729–G, 145 p.

Cannon, K.P., 2001, What the past can provide: contribution of prehistoric bison studies to modern bison management: Great Plains Research, v. 11, p. 145–174.

Cannon, G.M., Crothers, G.M., and Pierce, K.L., 1994, Archeological investigations along the Fishing Bridge Peninsula, Yellowstone National Park, Wyoming—the archeology, geology, and paleoenvironment: Manuscript on file at the Midwest Archeological Center, Lincoln, Nebraska, 200 p.

Cannon, K.P., and Hughes, R.E., 1993, Obsidian source characterization of Paleoindian projectile points from Yellowstone National Park: Current Research in the Pleistocene, v 10, p. 54–56.

Cannon, K.P., and Hughes, R.E., 1997, Provenance analysis of obsidian Paleoindian projectile points from Yellowstone National Park, Wyoming: Current Research in the Pleistocene, v 14, p 101–104.

Connor, M.A., 1998, Final Report on the Jackson Lake Archeological Project, Grand Teton National Park, Wyoming: Midwest Archeological Center, Technical Report No. 46, Lincoln, Nebraska, 278 p.

Colman, S.M., and Pierce, K.L., 1981, Weathering rinds on andesitic and basaltic stones as a Quaternary age indicator, western United States: U.S. Geological Survey Professional Paper 1210, 56 p.

Despain DG. 1990. Yellowstone Vegetation: Consequences of environment and history in a natural setting: Boulder, Colorado, Roberts Rinehart, Inc., 239 p.

Doser, D.I., 1985, Source parameters and faulting processes of the 1959 Hebgen Lake, Montana, earthquake sequence: Journal of Geophysical Research, v. 90, p. 4537–4555.

Dzurisin, Daniel and Yamashita, K.M., 1987, Vertical surface displacements at Yellowstone Caldera, Wyoming, 1976–1986: Journal of Geophysical Research, v. 92, no. B13, p. 13,753–13,766.

Dzurisin, Daniel, Yamashita, K.M., and Kleiman, J.W., 1994, Mechanisms of crustal uplift and subsidence at the Yellowstone caldera, Wyoming: Bulletin of Volcanology, v. 56, p. 261–270.

Fournier, R.O., Christiansen, R.L., Hutchinson, R.A., and Pierce, K.L., 1994, A field–trip guide to Yellowstone National Park, Wyoming, Montana, and Idaho—Volcanic, hydrothermal, and glacial activity in the region: U.S. Geological Survey Bulletin 2099, 46 p.

Frison, G.C., 1992, The foothills–mountains and the open plains: The dichotomy in Paleoindian subsistence strategies between two ecosystems: *in* Stanford, D.J. and Day, J.S., eds., Ice Age Hunters of the Rockies: University Press of Colorado, Niwot, CO, p. 323–342.

Gennett, J.A., and Baker, R.G., 1986, A late Quaternary pollen sequence from Blacktail Pond, Yellowstone National Park, Wyoming, U.S.A.: Palynology, v. 10, p. 61–71.

Gilbert, J.D., Ostenaa, Dean, and Wood, Christopher, 1983, Seismotectonic study, Jackson Lake Dam and Reservoir, Minidoka Project, Idaho–Wyoming: U.S. Bureau of Reclamation Seismotectonic Report 83–8, 122 p.

Good, J.D., 1964, Prehistoric landslide in Yankee Jim Canyon, Park County, Montana: Geological Society of America Special Paper 82, p. 327–328.

Good, J.D., and Pierce, K.L., 1996, Interpreting the landscapes of Grand Teton and Yellowstone national parks, recent and ongoing geology: Grand Teton National History Association, Moose, Wyoming, 58 p.

Graumlich, L.J., Pisaric, M.F.J, Waggoner, L.A., Littell, J.S., and King, J.C., in press, Upper Yellowstone River flow and teleconnections with Pacific basin climate variability during the past three centuries: Climatic Change.

Graumlich, L.J., and Littell, J., 2001, Multi–decadal variability of precipitation in the Greater Yellowstone region as inferred from tree rings: Proceedings, 18th Annual Pacific Climate Workshop on Climate Variability of the Eastern North Pacific and Western North America.

Haines, A.L., 1966, The Rigler Bluff hearth site, 24PA401: Archaeology in Montana, v. 7(2), p. 5.

Horberg, L., 1940, Geomorphic problems and glacial geology of the Yellowstone Valley, Park County, Montana: Journal of Geology, v. 48, p. 275–303.

Houston, D.B., 1973, Wild fires in Yellowstone National Park: Ecology, v. 54, p. 1111–1117.

Hughes, S.S., 1998, Getting to the point: Evolutionary change in Prehistoric weaponry: Journal of Archaeological Method and Theory, v. 5(4), p. 345–408.

Husted, W.M., and Edgar, R., 2002, The archeology of Mummy Cave, Wyoming: An introduction to Shoshonean prehistory: Midwest Archeological Center Special Report No. 4, Lincoln, Nebraska, 232 p.

Licciardi, J.M., Clark, P.U., Brook, E.J., Pierce, K.L., Kurz, M.D., Elmore, D., and Sharma, P., 2001, Cosmogenic ^{3}He and ^{10}Be chronologies of the northern outlet glacier of the Yellowstone ice cap, Montana, USA: Quaternary Research, v. 29, no 12, p. 1095–1098.

Littell, J. S. 2002. Determinants of fire regime variability in lower elevation forests of the northern Greater Yellowstone Ecosystem: MS thesis, Montana State University, Bozeman, MT, 122 p.

Locke, W.W., Clarke, W.D., Elliott, J.E., Lageson, D.R., Mokt, D.W., Montagne, J., Schmidt, J.G., and Smith, Michael, 1995, The middle Yellowstone valley

from Livingston to Gardiner, Montana: A microcosm of northern Rocky Mountain geology: Northwest Geology, v. 24, p. 1–65.

Locke, W.W., and Meyer, G.A., 1994, A 12,000–year record of vertical deformation across the Yellowstone caldera margin: The shorelines of Yellowstone Lake: Journal of Geophysical Research, v. 99, No. B10, p. 20,079–20,094.

Love, J.D., and Love, J.M., 1997, Road log, Jackson to Dinwoody and return: Geological Survey of Wyoming Public Information Circular 20, 37 p.

Love, J.D., Reed, J.C. Jr., and Christiansen, A.C., 1992, Geologic map of Grand Teton National Park: U.S. Geological Survey Miscellaneous Investigations Series Map I–2031.

Lundstrom, S.C., 1986, Soil stratigraphy and scarp morphology studies applied to the Quaternary geology of the southern Madison Valley, Montana: MS thesis, Humboldt State University, Arcata, CA, 53 p.

Meyer, G.A., and Locke, W.W., 1986, Origin and deformation of Holocene shoreline terraces, Yellowstone Lake, Wyoming: Geology, v. 14, p. 699–702.

Meyer, G.A., Wells, S.G., and Jull, A.J.T., 1995, Fire and alluvial chronology in Yellowstone National Park: Climatic and intrinsic controls on Holocene geomorphic processes: Geological Society of America Bulletin, v. 107, p. 1211–1230.

Millspaugh, S.H., and Whitlock, C., 1995, A 750–year fire history based on lake sediment records in central Yellowstone National Park, USA: The Holocene, v. 5, p. 283–292.

Millspaugh, S.H. and Whitlock C., in press, Postglacial fire, vegetation, and climate history of the Yellowstone–Lamar and Central Plateau provinces, Yellowstone National Park: *in* After the Fires, Wallace, L., ed., The ecology of change in Yellowstone National Park: Yale University Press, New Haven, Conn.

Millspaugh, S. H., Whitlock, C., and Bartlein, P. J., 2000, A 17,000–year history of fire for the Central Plateau of Yellowstone National Park: Geology, v. 28, p. 211–214.

Montagne, John, and Locke, W.W., 1989, Trip 7 road log, Cenozoic history of Yellowstone Valley between Livingston and Gardiner, Montana: *in* French, D.E., and Grabb, R.F., eds., Geologic resources of Montana, Guidebook 1989 Field Conference: Montana Geological Society, v. II, p. 502–521

Morgan, L.A., Shanks, W.C. III, Lovalvo, D.A., Johnson, S.Y., Stephenson, W.J., Pierce, K.L., Harlan, S.S., Finn, C.A., Lee, G., Webring, M., Schulze, B., Dühn, J., Sweeney, R., and Balistrieri, L., 2003, Exploration and discovery in Yellowstone Lake: Results from high–resolution sonar imaging, seismic reflection profiling, and submersible studies: Journal of Volcanology and Geothermal Research, p. 221–242.

Norris, P.W., 1880, Prehistoric remains in Montana between Fort Ellis and the Yellowstone River: Annual report of the Board of Regents of the Smithsonian Institution for the year 1879: Government Printing Office, Washington, D.C., p. 327–328.

Obradovich, J.D., 1992, Geochronology of the late Cenozoic volcanism of Yellowstone National Park and adjoining areas, Wyoming and Idaho: U.S. Geological Survey Open–File Report 92–408, 45 p.

Obradovich, J.D., and Izett, G.A., 1991, $^{40}Ar/^{39}Ar$ ages of upper Cenozoic Yellowstone Group tuffs: Geological Society of America, Abstracts with Programs, v. 23, No. 2, p. 84.

Pelton, J.R., and Smith, R.B., 1979, Recent crustal uplift in Yellowstone National Park: Science, v. 206, p. 1179–1182.

Phillips, F.M., Zreda, M.G., Gosse, J.C., Klein, Jeffrey, Evenson, E.B., Hall, R.D., Chadwick, O.A., and Sharma, Pankaj, 1997, Cosmogenic ^{36}Cl and ^{10}Be ages of Quaternary glacial and fluvial deposits of the Wind River Range, Wyoming: Geological Society of America Bulletin, v. 109, p. 1453–1463.

Pierce, K.L., 2003, Pleistocene glaciations of the Rocky Mountains: *in* Gillespie, A..R., Porter, S.C., and Atwater, B.F., eds., Quaternary glacial ages in the United States–Advances since 1965 in the study of the Quaternary Period, Elsevier Science B.V., in press.

Pierce, K.L., 1979, History and dynamics of glaciation in the northern Yellowstone National Park area: U.S. Geological Survey Professional Paper 729 F, 91 p.

Pierce, K.L., Adams, K.D., and Sturchio, N.C., 1991, Geologic setting of the Corwin Springs Known Geothermal Resource area–Mammoth Hot Springs area in and adjacent to Yellowstone National Park: *in* Sorey, M.L., ed., Effects of potential geothermal development in the Corwin Springs known geothermal resource area, Montana, on the thermal features of Yellowstone National Park: U.S. Geological Survey Water Resources Investigations Report 91–4052, p. C1–C37.

Pierce, K.L., Cannon, K.P., Meyer, G.A., Trebesch, M.J., and Watts, Raymond, 2002, Post–glacial inflation–deflation cycles, tilting, and faulting in the Yellowstone caldera based on Yellowstone Lake shorelines: U.S. Geological Survey Open–File Report 02–0142, 30 p.

Pierce, K.L., and Good, J.D., 1992, Field guide to the Quaternary geology of Jackson Hole, Wyoming: U.S. Geological Survey Open–File Report 92–504, 49 p.

Pierce, K.L., Lundstrom, Scott, and Good, John, 1998, Geologic setting of archeological sites in the Jackson Lake area, Wyoming: *in* Connor, Melissa, Final Report on the Jackson Lake Archeological Project, Grand Teton National Park, Wyoming, Midwest Archeological Center, National Park Service, Lincoln, Neb., Technical Report No. 46, p. 29–48, p. 222–242.

Pierce, K.L., and Morgan, L.A., 1992, The track of the Yellowstone hot spot: volcanism, faulting and uplift: *in* Link, P.K., Kuntz, M.A., and Platt, L.W., eds., Regional geology of eastern Idaho and western Wyoming: Geological Society of America Memoir 179, p. 1–53.

Pierce, K.L., Obradovich, J.D., and Friedman, Irving, 1976, Obsidian hydration dating and correlation of Bull Lake and Pinedale glaciations near West Yellowstone, Montana: Geological Society of America Bulletin, v. 87, p. 703–710.

Richmond, G.M., 1977, Surficial geologic map of the Canyon Village quadrangle, Yellowstone National Park, Wyoming: U.S. Geological Survey Miscellaneous Geological Investigations Map I–652, scale 1:62,500.

Richmond, G.M., 1976, Surficial geologic history of the Canyon Village quadrangle, Yellowstone National Park, Wyoming, for use with Map I–652: U.S. Geological Survey Bulletin 1427, 35 p.

Reeve, S.A.,. Wright, G.A., Marceau, T.E., and Meecham, P., 1979, Archeological Investigations at the Lawrence Site (48TE509), Grand Teton National Park, Wyoming: Manuscript on file at the Midwest Archeological Center, Lincoln, Nebraska. 111 p.

Ruleman, C.A. III, 2002, Quaternary tectonic activity within the northern arm of the Yellowstone tectonic parabola and associated seismic hazards, southwest Montana: MS thesis, Montana State University, Bozeman, MT, 157 p.

Shafer, S.L., Bartlein, P.J., and Whitlock, C. in press, Understanding the spatial heterogeneity of global environmental change in mountain regions: *in* Huber, U., Reasoner, M., and Bugmann, H., eds., Global Change and Mountain Regions, Kluwer, Dordrecht, The Netherlands.

Sharp, Warren, Ludwig, K.R., Chadwick, O.A., Amundson, Ronald, and Glaser, L.L., 2003, Dating fluvial terraces by $^{230}Th/U$ on pedogenic carbonate, Wind River Basin, Wyoming: Quaternary Research, in press.

Smith, M., Lageson, D., Heatherington, A., and Harlan, S., 1995, Geochronology, geochemistry, and isotopic systematics of the basalt of Hepburns Mesa, Yellowstone River Valley, Montana: Geological Society of America Abstracts with Programs, v. 27, no. 4, p. 56.

Smith, R.B., Pierce, K.L., and Wold, R.J., 1993, Seismic surveys and Quaternary history of Jackson Lake, Wyoming: *in* Snoke, A.W., Steidtmann, J.R., and Roberts, S.M., eds., Geology of Wyoming: Wyoming Geological Survey Memoir No. 5, p. 668–693.

Smith, R.B., and Siegel, L.J., 2000, Windows into the Earth, the Geologic Story of Yellowstone and Grand Teton National Parks: Oxford University Press, 242 p.

Stuiver, Minze, and Reimer, P.J. 1993, Extended ^{14}C data base and revised CALIB 3.0 ^{14}C age calibration program: Radiocarbon, v. 35, p. 215–230.

Sturchio, N.C., Pierce, K.L., Morrell, M.T., and Sorey, M.L., 1994, Uranium–series ages of travertines and timing of the last glaciation in the northern Yellowstone area, Wyoming–Montana: Quaternary Research, v. 41, p. 265–277.

U.S. Geological Survey, 1972a, Surficial geologic map of Yellowstone National Park: U.S. Geological Survey Miscellaneous Investigations Map 710.

U.S. Geological Survey, 1972b, Bedrock geologic map of Yellowstone National Park: U.S. Geological Survey Miscellaneous Geologic Investigations Map I–711.

Whitlock, C., 1993, Postglacial vegetation and climate of Grand Teton and southern Yellowstone National Parks: Ecological Monographs, v. 63, p. 173–198.

Whitlock, C., and Bartlein, P.J., 1993, Spatial variations of Holocene climatic change in the Yellowstone region: Quaternary Research, v. 39, p. 231–238.

Whitlock, C., and Millspaugh, S.H., 1996, Testing the assumptions of fire–history studies: An examination of modern charcoal accumulation in Yellowstone National Park, USA: The Holocene, v. 6, p. 7–15.

Wicks, Charles Jr., Thatcher, Wayne, and Dzurisin, Daniel, 1998, Migration of fluids beneath Yellowstone caldera inferred from satellite radar interferometry: Science, v. 282, p. 458–462.

Wright, G.A., 1984, People of the high country: Jackson Hole before the settlers: Peter Lang, NY, 197 p.