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From cirques to canyon cutting: New Quaternary research in the Uinta Mountains

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ABSTRACT

The Quaternary record of the Uinta Mountains of northeastern Utah has been studied extensively over the past decade, improving our understanding of the Pleistocene glacial record and fluvial system evolution in a previously understudied part of the Rocky Mountains. Glacial geomorphology throughout the Uintas has been mapped in detail and interpreted with reference to other well-studied localities in the region. In addition, studies in Browns Park and Little Hole in the northeastern part of the range have provided information about paleoflooding, canyon cutting, and integration of the Green River over the Uinta Mountain uplift. Notable contributions of these studies include (1) constraints on the timing of the local last glacial maximum in the southwestern Uintas based on cosmogenic surface exposure dating, (2) insight into the relationship between ice dynamics and bedrock structure on the northern side of the range, and (3) quantification of Quaternary incision rates along the Green River. This guide describes a circumnavigation of the Uintas, visiting particularly well-documented sites on the north and south flanks of the range and along the Green River at the eastern end.

Keywords: Uinta Mountains, Colorado Plateau, Last Glacial Maximum, Quaternary.

INTRODUCTION

For much of the past century, the Quaternary record of the Uinta Mountains escaped the attention focused on the neighboring Colorado Front Range, Wind River Range, and Yellowstone Plateau. That situation changed in the mid-1990s when researchers interested in glacial and fluvial geomorphology began working on aspects of these records in the range. Their work was unified by recognition of the characteristics that make the Uinta Mountains important to our overall understanding of landscape evolution in the interior western United States. First, the Uintas contain an unusually complete record of post-Laramide tectonics, erosion, and drainage integration, and they contained an extensive glacier complex during the Pleistocene glaciations. Second, the unique

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east-west orientation of the mountains allows study of paleoprecipitation gradients in a direction parallel to primary storm tracks and moisture transport. Third, the location and orientation of the range allow the Quaternary record in the Uintas to function as a link between the middle Rockies and Colorado Plateau to the Great Basin and the Sierra Nevada. Finally, numerous famous geologists worked in the Uintas in the late 1800s and early 1900s, including Hayden (1871), Powell (1876), King (1878), Atwood (1909), and Bradley (1936). Many of their seminal ideas regarding tectonics and landscape evolution evolved through consideration of the evidence observed in the Uinta Mountains.

Trip Overview

This field trip provides an overview of recent work illuminating aspects of the glacial and fluvial history of the Uinta Mountains from the Last Glacial Maximum (MIS-2, ca. 22–18 ka) to the present. Much of this work was conducted within the 456,704-acre High Uintas Wilderness Area, which is jointly administered by the Ashley and Wasatch-Cache National Forests. Depending upon the season, it is often not possible to visit sites in the interior of the Uintas. However, this trip takes full advantage of viewpoints and exposures around the perimeter of the range in order to summarize our work and to introduce lingering questions.

The route of this trip circumnavigates the Uinta Mountains in a counter-clockwise direction (Fig. 1). The first day focuses on glacial deposits on the south slope of the range, where county roads (CR) and Forest Service roads (FR) allow access to glacial deposits in the Lake Fork and Yellowstone River valleys. The second day of the trip explores the post-Laramide tectonic and drainage evolution of the eastern end of the range, including the Quaternary stratigraphy of the Green River and recognition of a significant paleoflood event. The final day continues westward along the north slope of the Uintas, visiting late Quaternary gla-

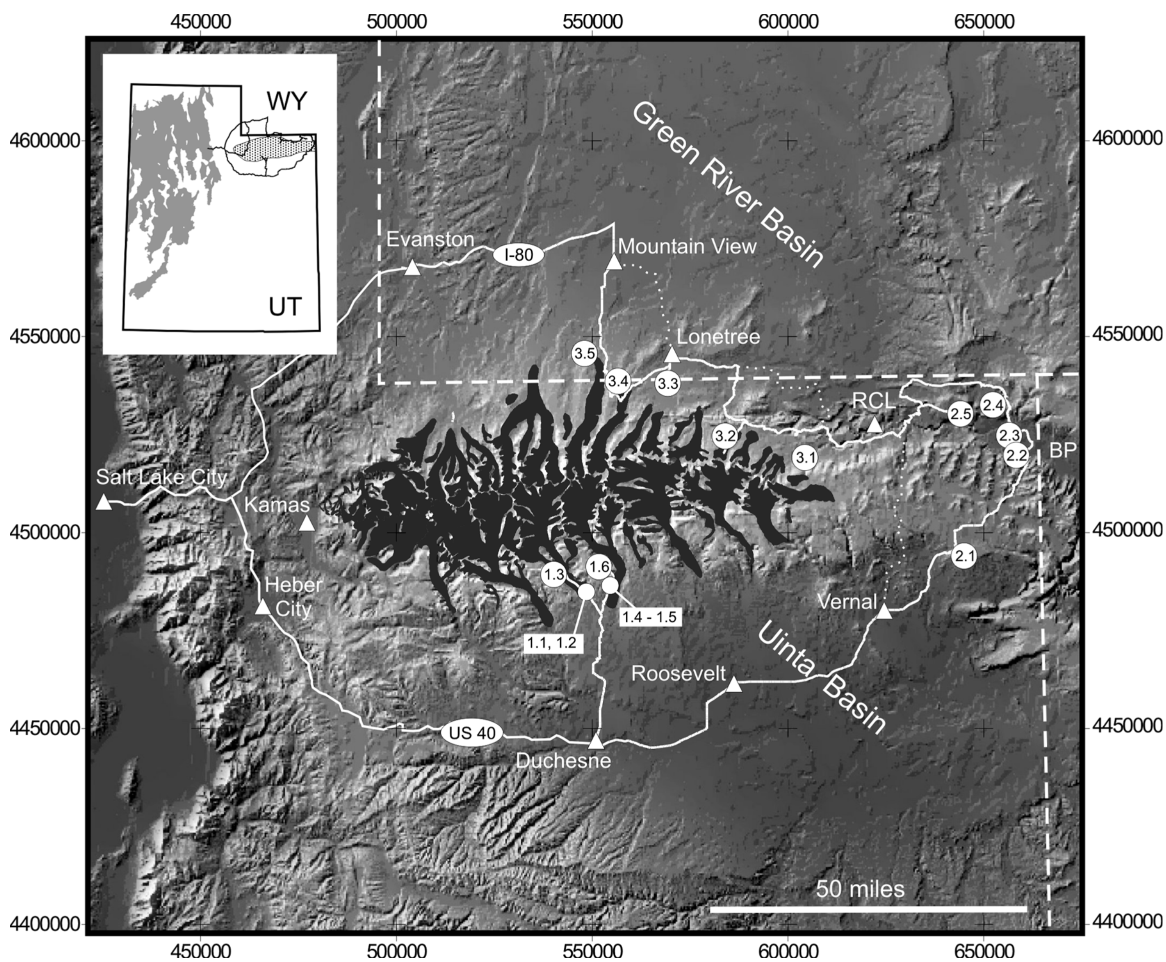

Figure 1. Route of the field trip through northeastern Utah and southwestern Wyoming. Inset shows the state of Utah, the extent of Lake Bonneville at the Last Glacial Maximum (LGM), the Uinta Mountains (stippled pattern on inset), and the field trip route. Larger map shows the reconstructed outlines of LGM glaciers (dark gray), the route of the trip (solid white line), alternate roads in case of early season snow (dotted white lines), and numbered stops (corresponding to text). BP—Browns Park; RCL—Red Canyon Lodge.

cial and fluvial deposits in several localities before returning to Salt Lake City.

As a note to guidebook users, mechanized equipment of any kind is prohibited within federally designated wilderness areas. Thus, access to field sites within the High Uintas Wilderness is by foot or by horseback only. Limitations also exist on the number of people and heads of stock allowed per party. Please check with either the Ashley National Forest (+1-435-789-1181, Vernal, Utah) or the Wasatch-Cache National Forest (+1-307-789-3194, Evanston, Wyoming) for details if you are interested in visiting the Uinta backcountry.

Physical Setting

The Uinta Mountains are the longest east-west–trending mountain range in the conterminous United States, extending ~200 km eastward from the Wasatch Mountains at Kamas, Utah, into northwestern Colorado. Physiographically, the range can be divided into two sections: the western glaciated Uintas and the eastern nonglaciated Uintas (Hansen, 1986). The boundary between the two subprovinces is located ~40 km west of the Utah-Colorado border, along a line extending north from Vernal (Fig. 1). The core of the western Uintas, centered on Kings Peak, the highest mountain in Utah at 4136 m, is referred to as the "High Uintas."

The drainage systems in the Uinta Mountains flow primarily north and south away from the crest of the range. On the north flank, short tributaries originating in alpine cirques drain into main streams in the glaciated valleys. Bear River, at the western end of the range, drains into Great Salt Lake; all other streams flow into the Green River upstream from Flaming Gorge dam. On the south flank, tributaries in large compound cirques feed into deep, narrow canyons that descend to the Uinta Basin. All streams on the south flank either flow into the Duchesne River (a tributary of the Green River) or directly into the Green River.

Bedrock Geology

The Uinta Mountains are cored by a series of Precambrian orthoquartzite, shale, and sandstone units known as the Uinta Mountain Group (UMG) (Bryant, 1992; Bradley, 1995; Hansen, 1969; Ritzma, 1959; Wallace and Crittenden, 1969; Dehler et al., 2006). These sediments, with a total thickness in excess of 7 km (Hansen, 1965; Stone, 1993), accumulated in an intracratonic rift basin near the suture of the Wyoming Archean craton (the Cheyenne Belt) and Paleoproterozoic terrane (the Colorado Province) ca. 800 Ma (Condie et al., 2001). The uppermost unit of the Uinta Mountain Group is the Red Pine Shale, which is unconformably overlain by thin (≤50-m-thick) sandstone of the Cambrian Tintic and Lodore Formations in parts of the western and eastern Uintas, but is more commonly overlain directly by Mississippian Madison Limestone. The Madison Limestone is locally >200 m thick (Bryant, 1992), forms hogbacks in some places where it is dissected by glacial valleys, and forms karst topography at ~3000 m in the headwaters of the Rock Creek and Duchesne River. Stratigraphically above this unit are the Mississippian Humbug (limestone and dolomite) and Doughnut (shale, limestone, and dolomite) Formations, which are 10–50 m thick. Large landslides are common near the contact of these two units, one of which can be observed at Stop 1.2. Upper Paleozoic strata include the Permian Weber Sandstone, which forms local hogbacks and which is overlain by upper Permian and Mesozoic strata of varying thickness at the mouths of glacial valleys.

Flat-lying, weakly cemented Tertiary gravel deposited during uplift of the Uinta Mountains is generally found above angular unconformities with tilted, pre-Cenozoic bedrock. On the north slope, these deposits are mapped as the extensive Wasatch Formation (Bryant, 1992). On the south flank, these gravels include fluvial sandstone, conglomerate, and colluvial or mudflow diamictite of the Duchesne River Formation, which are most common on unglaciated uplands between the Duchesne River and Whiterocks drainage basins (Bryant, 1992). Sandstone and conglomerate beds in this unit are poorly indurated and are a source for mass-wasting and large alluvial fan deposits in tributaries of glacial valleys. The age of this unit is considered to be early Tertiary; however, recent work by D. Sprinkel (2004, personal commun.) suggests that several gravels currently mapped as part of this unit may be significantly younger (late Tertiary or early Pleistocene).

Igneous rocks are almost entirely absent from the Uintas. Several small mafic dikes exist along the crest of the uplift (Ritzma, 1983). Dates from these dikes range from 552 ± 17 to 453 ± 29 Ma on the basis of Rb/Sr wholerock and K/Ar methods (Hansen et al., 1982; Ritzma, 1983; Rowley et al., 1985).

Structural Geology

Structurally, the Uintas are the surface expression of a broad doubly plunging anticlinal uplift, the axis of which is roughly concordant with the ridge crest. The axis of the uplift extends westward through the Wasatch Range and emerges as the Cottonwood Uplift at the Wasatch Front (Butler et al., 1920; Hansen, 1986; Bradley, 1995; Paulsen and Marshak, 1999). At the eastern end, the axis extends through northwestern Colorado and merges with the White River uplift (Tweto, 1976; Hansen, 1986). The Uinta anticline is asymmetrical, with its axis much closer to the north flank of the fold. Consistent with this geometry, the Precambrian and Paleozoic strata in the Uinta Mountains are subhorizontal at the crest of the range, gently dipping to the south on the southern flank and steeply dipping to the north on the northern flank.

Uplift of the Uinta Mountains occurred during the Laramide Orogeny (late Mesozoic to early Cenozoic time) and was characterized by folding and thrusting along a system of east-west–trending faults on the north and south flanks of the range. Folding of the Uinta arch accommodated deformation by crustal shortening, and thrust faulting accommodated brittle deformation on the north and south flanks (Bradley, 1995). North-dipping thrust faults on the south flank of the Uintas are now buried by sediments in the Uinta Basin; however, south-dipping thrust faults are well-exposed on the north flank (Bradley, 1995). Displacement also occurred to a lesser extent on normal faults during the later stages of the Laramide Orogeny; one prominent normal fault is the South Flank fault, which juxtaposed Madison Limestone and quartzite of the upper UMG (Bryant, 1992).

Once Laramide uplift ceased, there was an episode of tectonic stability throughout the Rocky Mountains during which significant pedimentation took place (Epis and Chapin, 1975; Mears, 1993). This includes the extensive Gilbert Peak erosion surface developed around the flanks of the Uintas (Sears, 1924; Bradley, 1936; Hansen 1986). Two Tertiary deposits record relatively subtle sedimentation and deformation in the Uintas: the Oligocene Bishop Conglomerate, which lies atop the Gilbert Peak erosion surface, and the Miocene Browns Park Formation. The regional tilting of the Bishop Conglomerate–Gilbert Peak surface and faulting associated with the Browns Park graben record the collapse of the eastern part of the Uinta uplift in late Oligocene–Miocene time (Bradley, 1936; Hansen, 1984; 1986).

Geomorphology

Although glaciers are absent from the modern Uinta Mountains, the spectacular alpine landscape of the High Uintas testifies to extensive Pleistocene glaciation. Cirque glaciers coalesced to form confined valley glaciers that apparently never surmounted the crest of the range or the major drainage divides, except in a few drainages at the west end of the range. As a result, the topography of the Uinta Mountains is dominated by deep, glacially scoured valleys separated by broad, unglaciated interfluves. Terminal moraines dating to the Last Glacial Maximum and older glaciations are present in most glaciated valleys, and outwash valley trains and meltwater channels are locally well developed.

At the highest elevations, abundant and diverse periglacial deposits cover many slopes and the tops of most ridges and peaks of the Uintas. These include talus, protalus ramparts, felsenmeer, solifluction features, patterned ground, nivation hollows, and rock glaciers. The ages of these deposits are controversial. Grogger (1974) proposed that current periglacial activity is only capable of locally modifying older deposits. However, apparently active nonsorted circles are present at higher elevations. Furthermore, many rock glaciers have features suggestive of motion, including fresh unstable frontal slopes steeper than the angle-of-repose, transverse and longitudinal surface furrows, springs discharging 0 °C water in late summer, and surface meltwater ponds (Wahrhaftig and Cox, 1959). On the basis of these characteristics, Osborn (1973) reported that several rock glaciers in the western Uintas appear to be active. It is also worth noting that Bauer (1985) reported the presence of perennial ice in talus at an elevation of 2500 m in the western Uintas, strongly suggesting that active rock glaciers and permafrost are a possibility at higher elevations.

At lower elevations, Holocene alluvial deposits mantle valley floors and merge with a series of outwash terraces and moraines where streams flow into the Green River and Uinta basins. The stream planforms are dictated by the glacial history. In headwater regions, streams originate from talus slopes and cirque lakes and feed directly into mainstem channels. The interfluves and steep valley walls are drained by distributed surface flow and short tributaries, often with summer discharges of no more than ~0.05 m^3/s. The mainstem channels are typically confined by the valley walls in areas where resistant Precambrian and Paleozoic bedrock outcrops. The stream channel gradients through these areas are steep (water surface slopes in excess of 0.002 m/m), with the bed comprised primarily of cobbles and boulders. In areas where less resistant bedrock is exposed, the valley floors widen appreciably into low-gradient meadows locally more than 300 m wide. Channel bed sediment in these meadows is dominated by sand and silt in point bars and sand and gravel on channel bottoms, and slopes are commonly <0.0005 m/m.

History of Research

Very little previous nonglacial geomorphic research has been conducted in the Uinta Mountains region. Other than hypotheses put forth by early workers on the integration of the Green River across the uplift (Powell, 1876; Sears, 1924; Bradley, 1936), the middle–late Cenozoic record has been explored mostly by W.R. Hansen (e.g., Hansen, 1965, 1984, 1986). He had several keen and interesting observations about regional landscape evolution, though none of his work focused on the Quaternary. Osborn (1973) and Nelson and Osborn (1991) mapped and employed relative dating of soil profiles on the extensive series of outwash terraces on the south flank of the Uintas. They distinguished at least twelve outwash surfaces in the Uinta Basin, some hypothetically pre-dating the Quaternary, and estimated relatively high incision rates of 170–320 m/m.y.

In terms of glacial geology, W.W. Atwood completed the first intensive study of the Uintas. In his seminal 1909 monograph, *Glaciation of the Wasatch and Uinta Mountains*, he described the glacial geomorphology of every major drainage in the range and identified evidence for two major glacial advances. Later, in a report on the geomorphology of the north slope, Bradley (1936) recognized three separate ice advances; from oldest to youngest, he named these the Little Dry, Blacks Fork, and Smiths Fork glaciations after the valleys where their deposits were particularly well preserved. In his summary report on glaciation of the Rocky Mountains, Richmond (1965) correlated the Little Dry deposits with the Illinoian and pre-Illinoian glacial stages of the U.S. Midwest, the Blacks Fork with the early Wisconsin, and the Smiths Fork with the late Wisconsin glacial stage. He also correlated the Blacks Fork advance with the Bull Lake glaciation in the Wind River Range and the Smiths Fork glaciation with the Pinedale. More recent mapping has also considered the Blacks Fork glaciation to be correlative with the Bull Lake glaciation (Richmond, 1986; Bryant, 1992), although the Bull Lake glaciation has been shown to pre-date the early Wisconsin (Chadwick et al., 1997; Sharp et al., 2003). Thus, the overall glacial framework in the Uintas mirrors that of numerous other western ranges with a well-preserved MIS-2 advance (Smiths Fork glaciation), and a slightly more extensive older advance, considered to have occurred during MIS-6 (Blacks Fork glaciation).

In the past few decades, several unpublished theses have focused on details of the glacial geology and geomorphology in specific areas of the Uintas (e.g., Schoenfeld, 1969; Barnhardt, 1973; Osborn, 1973; Grogger, 1974; Gilmer, 1986; Schlenker, 1988; Zimmer, 1996; Douglass, 2000). Most recently, Munroe (2001) mapped the glacial geomorphology of the northern Uintas at 1:24,000 scale and reconstructed the late Quaternary glacial history of this area on the basis of glacial and periglacial landforms. Similarly, Laabs (2004) mapped the south slope at 1:24,000 scale and applied cosmogenic surface exposure dating and numerical modeling of past glaciers in an attempt to better constrain the timing and paleoclimate of the local LGM. On the basis of thirteen cosmogenic ^{10}Be exposure ages of boulders on adjacent moraines, Laabs (2004) determined that the Smiths Fork glacial maximum ended in the south-central Uinta Mountains by 17.6 ± 1.1 ka.

Recent investigations into the fluvial record in the Uinta Mountains are fewer, and have largely concentrated on sediment transport and historic channel adjustments to dams, water diversions, and other human impacts. Ringen (1984) and Lenfest and Ringen (1985) quantified suspended sediment-discharge relationships for a number of stream gages on tributaries. Brink and Schmidt (1996) evaluated the historic record of geomorphic change on the Duchesne River, and a series of subsequent studies focused on that drainage. This includes studies of bedload transport, channel form, and channel migration (Smelser, 1997; Paepke, 2001), and the geomorphic effects of dams and water diversion in terms of sediment storage and channel response (Stamp, 2000; Gaeuman et al., 2005). The historic fluvial geomorphology of the Green River in the eastern Uinta Mountains has been investigated by Andrews (1986) and Grams and Schmidt (1999), especially in terms of sediment budget and the influence of Flaming Gorge dam. Finally, Martin (2000) and Larsen et al. (2004) studied debris-flow activity and the effects of low-magnitude floods on debris fans in the Green River canyons of Dinosaur National Monument.

DAY 1—SALT LAKE CITY TO VERNAL: THE SOUTHERN UINTA MOUNTAINS

Introduction

The first day of this field trip (total driving distance ≈247 mi [396 km]) focuses on the glacial geology of the southern Uinta Mountains. The trip crosses the Wasatch Mountains and proceeds eastward across the south flank of the Uintas, ending in Vernal, Utah. Stops in the Lake Fork and Yellowstone River drainages are to observe Pleistocene glacial deposits and landforms along with a few prominent mass-wasting features. Two stops include moderate hiking and climbing at elevations up to 3323 m above sea level (m asl; 10,900 ft asl). Please be prepared for hiking in rough and possibly wet terrain and for sudden changes in weather (temperatures at high elevations may be close to 0 °C). Access to the last stop (Stop 1.6, Lake Fork Mountain) may be prohibited by inclement weather or adverse road conditions.

Directions to Stop 1.1

From the Salt Palace Convention Center in downtown Salt Lake City, drive west on E 400 St./Utah Hwy 186 to South Main St. After 0.1 mi, turn left to follow Utah Hwy 186. After 0.1 mi, turn right onto South State St./U.S. Hwy 89. After 2.9 mi, merge onto I-80. Enter Parley's Canyon. After traveling 23.1 mi on I-80, merge onto U.S. Hwy 40 via Exit 148 toward Heber and Vernal and proceed 70 mi. In Duchesne, turn left onto Utah Hwy 87 East and *zero your trip odometer to begin the road log for Day 1*. At mile 15.5, turn left onto road 21000 W and follow signs for Mountain Home, Moon Lake, and Yellowstone Canyon. Just prior to mile 17.4, the road descends to an outwash surface mapped by Nelson and Osborn (1991) as pre–Bull Lake in age. Turn right at mile 19.1 onto an unnamed county road toward Moon Lake. This road enters Ute tribal land at mile 19.8; be advised that stopping vehicles or exiting the road are prohibited in this area. At mile 20.5, the road ascends to an Altonah-age outwash surface, descends through a landslide in this outwash at mile 21.5, and descends to a Smiths Fork–age outwash at mile 23. Here, the scarp on the left side is cut into pre–Smiths Fork outwash. After mile 24.8, the road ascends to a Smiths Fork–age outwash fan (which is pitted near its head) and crosses Smiths Fork–age moraines at miles 26.2 and 27. At mile 28.3, park along the right side of the road just north of the Ashley National Forest boundary. Hike 0.4 mi due east to the top of the small hill to observe moraines, bedrock structures, and mass wasting features near the mouth of Lake Fork Canyon.

Stop 1.1—Lake Fork Canyon (Mile 28.3)

This stop provides a spectacular view of the moraine sequence at the mouths of Lake Fork and Yellowstone River canyons. The small bedrock-cored hill we have climbed is composed of Triassic-Jurassic sedimentary rocks of the Ankareh Formation and Nugget Sandstone (Bryant, 1992). You likely noticed sandstone and orthoquartzite erratics of the Uinta Mountain Group during the ascent; these are derived from Blacks Fork–age till that caps most of the hill. Although some of the erratics may have been deposited during the last glaciation, this hill likely split the termini of the Lake Fork glacier. This has led to excellent preservation of moraines on its west side (where the terminal moraine has not been subject to erosion by the Lake Fork River). Looking to the west and south across the Twin Pots Reservoir affords views of the moraine sequence at the mouth of Lake Fork Canyon (Figs. 2 and 3). The low-relief (less than 10 m) moraine immediately south of Twin Pots Reservoir is a Smiths Fork recessional moraine; the higher relief (more than 70 m), Ponderosa pine–covered, compound ridge beyond it is the prominent, sharp-crested Smiths Fork terminal moraine. Nine cosmogenic ^{10}Be ages of this moraine from Laabs (2004) indicate this moraine was deposited prior to 16.9 ± 0.4 ka (based on a production rate of 5.1 atoms g SiO_2^{-1} yr^{-1} adjusted for altitude and latitude by using scaling factors from Stone [2000]), which approximates the termination of the Smiths Fork glacial maximum in this valley. This moraine is distinguished from the outer Blacks Fork moraine (which is not visible at this stop) by its sharp crests and steep distal slopes, higher frequency of moraine surface boulders, and the presence of an ice marginal drainage valley that exceeds 60 m depth in places (Fig. 3; see also Laabs and Carson, 2005). The Smiths Fork moraine grades to an outwash fan that is more than 20 m below a fan deposited during the Blacks Fork glaciation (Fig. 3). The outermost moraines visible to the south likely correlate to the Altonah moraine; these high-relief moraine ridges grade into outwash terraces that are more than 50 m above modern grade (Laabs and Carson, 2005). The age of this episode is poorly known, but may correlate to the Sacagawea Ridge glaciation believed to be correlative to marine oxygen-isotope stage 16 (Chadwick et al., 1997).

Outcrops to the east of the Lake Fork River expose an unconformity between south-dipping Mesozoic sedimentary rocks (namely the Triassic Woodside, Thaynes, and Ankareh Formations; the Triassic-Jurassic Nugget Sandstone; and the Jurassic Preuss Sandstone and Twin Creeks Limestone) and Oligocene sandstones and conglomerates of the Starr Flat Member of the Duchesne River Formation (Bryant, 1992). The Starr Flat Member was deposited by ancient tributaries of the Duchesne River while the Uinta Mountains were rising during late stages of the Laramide Orogeny. This unit may be correlative in places near Lake Fork canyon to the Bishop Conglomerate, a tuffaceous

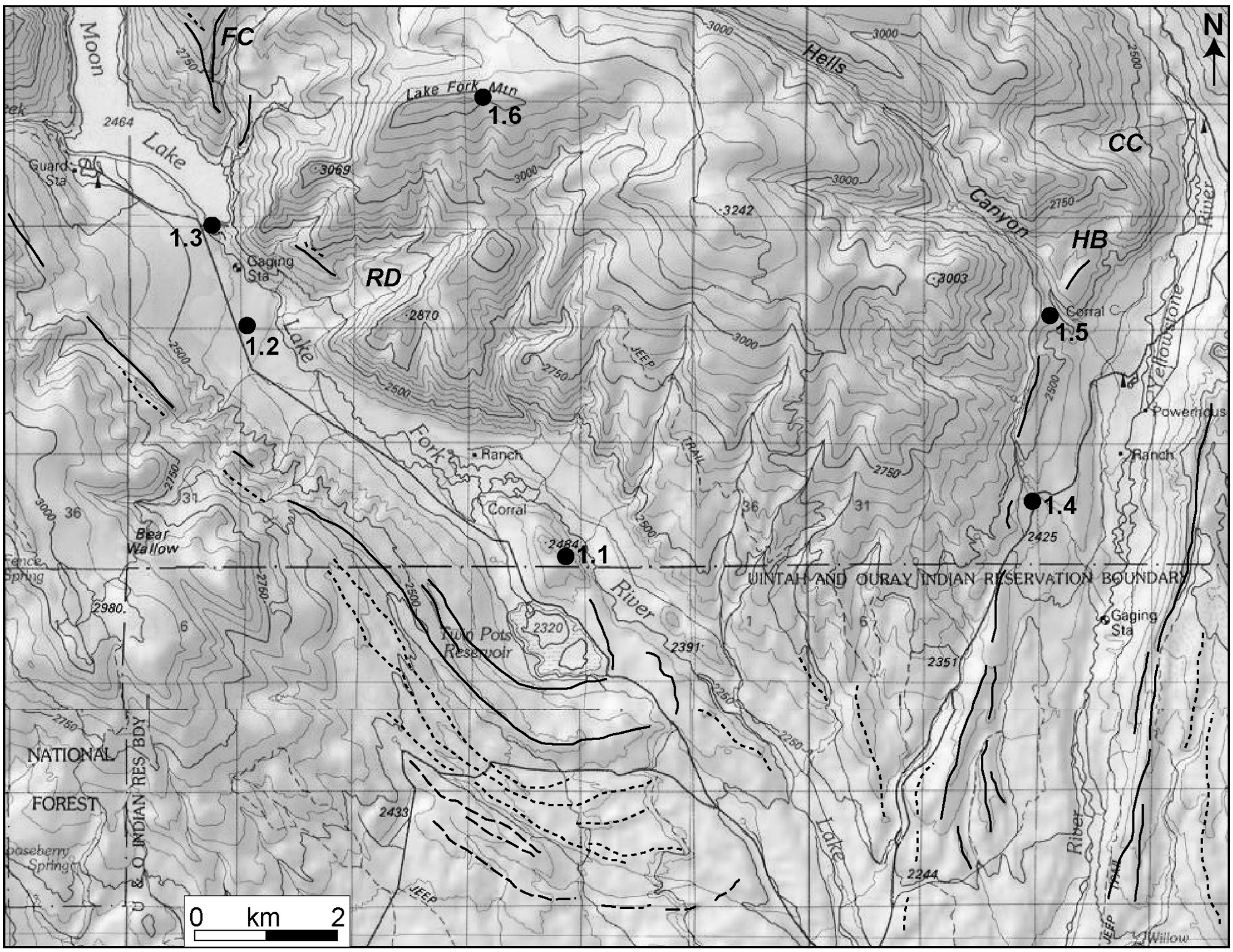

Figure 2. Map of field trip stops on Day 1 (black dots). Moraines deposited during the Smiths Fork (solid black lines), Blacks Fork (short-dashed lines), and Altonah glaciations (long-dashed lines) are shown (from Laabs, 2004). Notable locations are Fish Creek (FC), Raspberry Draw (RD), Harmston Basin (HB), and Cow Canyon (CC). Shaded topography is from portions of the U.S. Geological Survey 15′ Kings Peak and Duchesne quadrangles.

sandstone and conglomerate that overlies the Gilbert Peak erosion surface (which will be discussed at Stop 2.1).

Views to the north display the broad U-shape of the Lake Fork canyon. The western side of the canyon is covered mainly by till immediately north of this stop. Lateral moraines high above the valley bottom are easily visible and represent deposition during the Blacks Fork and Smiths Fork glaciations. The eastern side of the valley contains little till near the valley mouth, perhaps due to oversteepening of the valley side and subsequent mass wasting caused by toe-slope erosion by the Lake Fork River. Post-glacial mass wasting has taken place along both sides of the valley, as indicated by abundant rock falls, slumps, and landslides on the east side of the valley and dissected moraines on the west side.

Directions to Stop 1.2

Continue driving northwest toward Moon Lake. At the bridge crossing mile 28.5, notice at 3 o'clock the massive slope failure between two limestone ridges on the east wall of Lake Fork Canyon. Landslides are common on this side of the valley, which has been oversteepened near the mouth of the canyon due to toe-slope erosion by the Lake Fork River or by glacial erosion. After the bridge crossing, till covers most of the valley to the left of the road until mile 30, where the road ascends upon the broad alluvial fan discussed at Stop 1.1 (Fig. 2). At mile 31, a large landslide in Raspberry Draw is viewable to the right. Stop 1.2 is at the sign that indicates Raspberry Draw at mile 31.4; park along the right side of the road.

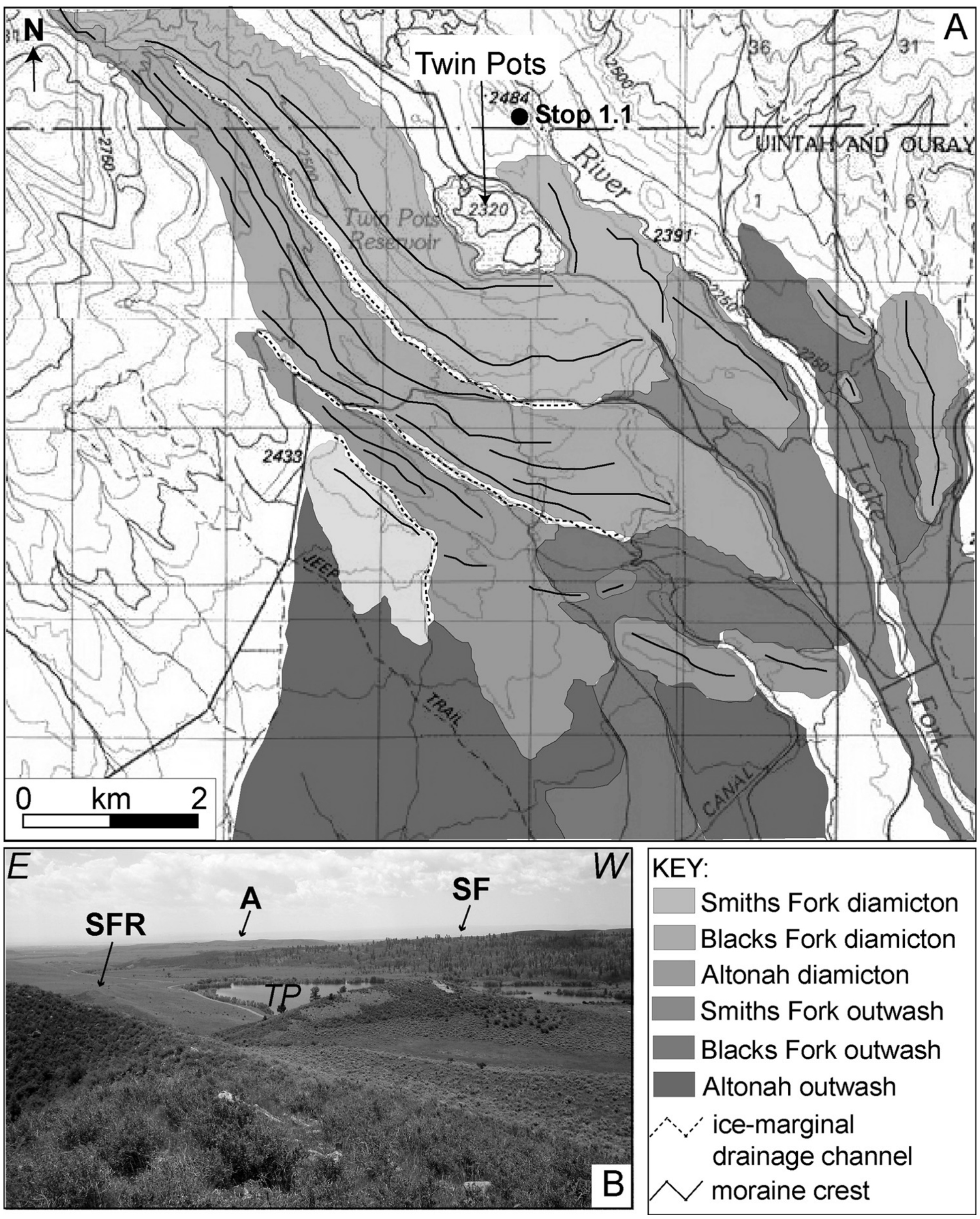

Figure 3. (A) Map of glacial deposits at the mouth of Lake Fork canyon (from Laabs and Carson, 2005). (B) View of moraine sequence to the south of Stop 1.1. A—Altonah moraine; SF—Smiths Fork moraine; SFR—Smiths Fork recessional moraine; TP—Twin Pots reservoir.

Stop 1.2—Raspberry Draw (Mile 31.4)

A large landslide in Raspberry Draw, an eastern tributary of the Lake Fork River (see Fig. 2 for location), cuts across two lateral moraines on the side of Lake Fork Canyon. The moraines are discontinuous and their ages are poorly known, but their crests are at elevations very close to Blacks Fork and Smiths Fork lateral moraines on the other side of the valley (Laabs, 2004); this indicates that the landslide occurred sometime after ca. 17 ka. The surface of the landslide is hummocky and several sag ponds are present, but it is heavily grazed by cattle that have likely disturbed pond sediments that might otherwise provide a limiting radiocarbon age of this feature.

The origin of the landslide is undoubtedly related to bedrock geology; the underlying strata are limestones and dolomites of the Mississippian Humbug Formation and shales of the Mississippian Doughnut Formation. As mentioned, landslides are common at the contact between these two units and are most prominent in Raspberry Draw and in Cow Canyon; the latter is a western tributary of the Yellowstone River (see Fig. 2 for locations; Laabs, 2004). One possible explanation for the origin of these landslides is saturation of weakly-cemented, fractured shales of the Doughnut Formation where they overlie low-permeability dolomite and/or limestone of the Humbug Formation. Increased pore-water pressure in saturated material weakens its shear strength and can cause it to move down slope, perhaps in response to oversteepening of the slope or seismic shaking (seismic shaking alone can also reduce the sheer strength of saturated material). Another possibility is that the lateral moraines north of the valley were once continuous across Raspberry Draw and dammed a small lake in the tributary valley. Failure of the moraine dam, the lake-bottom sediments, and underlying low-strength rocks may have triggered the landslide. Small, moraine-dammed lakes are common in tributaries to the Lake Fork and Yellowstone Valleys; however, distal slopes of the remaining lateral moraines on the north side of Raspberry Draw reveal no lake sediment.

Directions to Stop 1.3

Continue driving northwest across the alluvial fan toward Moon Lake. At mile 32.2, the head of the broad alluvial fan can be seen to the left. Flat-lying sandstone and conglomerates of the Duchesne River Formation seen above the head of the fan are the primary source of the fan sediment. Proceed northward to mile 32.2, turn right, and drive to the end of the driveway; this is Stop 1.3.

Stop 1.3—Lake Fork River Cutbank below Moon Lake Dam (Mile 32.3)

Moon Lake is a reservoir that enlarged a natural lake when an earthen dam was constructed just south of the original outflow in 1938. The lake maintains its original shape of a crescent moon (Fig. 2) but is now almost 6 km long (compared to an original length of ~5 km; Atwood, 1909) and has an average depth of ~14 m. Outflow from the dam generates hydroelectricity utilized locally, and discharge is used primarily for irrigation in the Uinta Basin.

The origin of Moon Lake is likely related to deposition of the alluvial fan on the western side of the valley (Fig. 2) and a back-rotational slump in bedrock on the east side, both of which are visible here. We are standing on the distal edge of the alluvial fan, which is composed primarily of debris-flow diamicton exposed in a cutbank just below this site (note: the trails that descend upon the cutbank are highly unstable; it is strongly recommended that you view it from above and stay off the trails). The source of this sediment is sandstone and conglomerate of the Duchesne River Formation exposed at the top of the high ridge to the west. Physical properties of the alluvial fan sediment strongly resemble those of its source rock, which contains medium- to coarse-grained sand and abundant rounded cobbles and boulders and is actively eroding today. Based on an observed breach in the Smiths Fork right-lateral moraine to the west and a lack of buried soils, erosional contacts, and glacial till in the cutbank, we believe that all of the fan sediment exposed here was deposited after the last glaciation. The back-rotational slump across the valley from this stop is in Madison Limestone, Tintic Formation, and Red Pine Shale; mass wasting in the latter unit is ubiquitous in the southern Uinta Mountains. Evidence for back rotation of the slump can be viewed to the east where beds of the Tintic Formation and Madison Limestone are dipping southward more steeply than in situ strata. The slump extends westward into Moon Lake, and part of it was probably excavated during construction of the dam. The dam is therefore built into alluvial fan sediment on the east side of the valley and a slump deposit on the west side of the valley. Atwood (1909) observed Moon Lake in its original form and did not describe evidence of a moraine dam or bedrock sill that could have formed a threshold for Moon Lake; therefore, it is not considered a paternoster lake.

Northeast of Moon Lake dam, prominent lateral moraines are visible on the sides of Fish Creek Valley, an eastern tributary of the Lake Fork River. These were deposited during the Smiths Fork glaciation when a small tributary glacier advanced southward from a broad high basin and intersected the main valley glacier. The headwaters of Fish Creek, a former glacier accumulation area, will be viewed from Stop 1.6.

In summary, the Quaternary history of lower Lake Fork Canyon includes repeated glaciations during the middle and late Pleistocene, the last of these terminated by ca. 17 ka (Laabs, 2004). Post-glacial events include mass wasting on steep valley slopes, erosion and mass wasting of lateral moraines, and deposition of the alluvial fan on the west side of the canyon. The Lake Fork River has incised its valley since the Smiths Fork glaciation, leaving an outwash terrace ~8 m above modern grade.

Directions to Stop 1.4

Return to the paved road, turn right, and enter the Moon Lake campground at mile 33.6. Restrooms and water are available at this lunch spot. When exiting the campground, *reset your odometer to zero at the cattle guard* for accurate mileage to afternoon stops. Drive southeast on the unnamed county road past all of the previous stops and leave Ashley National Forest. Turn left at mile 10.1 and drive east on the bridge over Lake Fork River. Smiths Fork–age outwash gravels appear in roadcuts at mile 10.4. Turn left at mile 10.5 and drive north toward the Yellowstone campgrounds in Ashley National Forest. Looking west at mile 11.0 provides a cross-sectional view of the moraine sequence at the mouth of Lake Fork Canyon (described at Stop 1.1). Immediately east of this road is a Blacks Fork–age latero-frontal moraine at the mouth of Yellowstone Canyon; parts of this moraine were deposited on Tertiary strata and sediment exposed in places immediately east of the road. At mile 12.5, the relatively deep gorge (up to 35 m) to the left was an ice-marginal drainage valley occupied during the last glaciation; this valley can be traced northward to the Ashley National Forest boundary. Enter Ashley National Forest at mile 14.4 and bear right at the fork at mile 14.6; here, the road descends onto Smiths Fork–age recessional moraines and kame terraces deposited during the last deglaciation. Stop 1.4 is the pullout area on the right at mile 15.3.

Stop 1.4—Moraine Sequence in Yellowstone Canyon (Mile 15.3)

Latero-frontal moraines that represent glacier termini during three Quaternary glaciations are well preserved and easily viewable just south of this stop. Lateral moraines on the east side of the valley are best preserved and are separated by broad ice-marginal drainage valleys. Smiths Fork terminal moraines are visible on both sides of the valley, where they locally have over 70 m of relief (Figs. 2 and 4). A well preserved sequence of Smiths Fork recessional moraines covers much of the valley center between terminal moraines; some of these features are clearly visible to the east. Part of a Blacks Fork moraine can be seen down valley to the south (Fig. 4), where a latero-frontal moraine curves toward the center of the valley. The outermost moraine is visible to the southeast and is the type section for the Altonah glaciation, where a prominent left-lateral moraine rises ~90 m above Mud Spring Draw (Fig. 4).

Laabs et al. (2004) attained seven cosmogenic ^{10}Be exposure ages of moraine boulders on the Smiths Fork left-lateral moraine, which is the highest visible ridge due east of this stop. The weighted-mean exposure age of four of these boulders (three samples were excluded based on statistics and boulder height) is 18.0 ± 0.7 ka, which they interpret to represent the end of the Smiths Fork maximum in this valley. This age estimate overlaps the weighted mean age of the Smiths Fork terminal moraine in the Lake Fork Canyon (16.9 ± 0.4 ka), and several boulder-exposure ages from these two moraines overlap within 2σ uncertainty. Laabs (2004) computed a weighted mean age of 13 samples from both moraines of 17.6 ± 1.1 ka, which represents the termination of the Smiths Fork maximum in the southwestern Uintas.

The Yellowstone River flows east of, and more than 50 m below, this stop. Glacial sediment has filled in much of the west side of Yellowstone Canyon; this stop provides a view of recessional moraines that comprise the surface deposits between Yellowstone River and the Smiths Fork latero-frontal moraine (Figs. 2, 4, and 5). Westlund (2005) mapped nine recessional moraines in this area, some of which are breached by Crystal Creek on the west side of the valley. This moraine sequence represents deposition during the last deglaciation; the Smiths Fork terminal moraine is visible upslope to the west (it is the ridge covered with prominent Ponderosa pines). Some recessional moraines are separated by kame terraces (Fig. 5), one of which is exposed in a road cut just east of the parking area. Here, glacial sediments are distinguished from tills elsewhere in the valley by the grain-size distribution of the matrix sediment (sandy silt versus silty sand) and by the presence of carbonate in the matrix and as coatings on cobbles and boulders. Clastic carbonate grains are rare in tills of the Uintas, and secondary carbonate is common only in pre–Smiths Fork tills. The source of carbonate in the kame terrace sediments is outcrops of Madison Limestone near the head of Hells Canyon (see Fig. 2 for location). Indeed, this implies that Hells Canyon was the source of ice-marginal drainage sediment on the west side of Yellowstone Canyon during the last glaciation; the evolution of fluvial drainage in this area will be discussed at the next stop.

Directions to Stop 1.5

Turn around and drive south-southeast to the fork encountered earlier. At mile 15.9, turn right (the turn is actually ~300° to the right) and follow FR 227 toward Hells Canyon. At mile 16.2, the Smiths Fork–age right-lateral moraine (covered by Ponderosa pines) can be viewed above the parking area to the west. This ridge, along with multiple recessional moraines below it, is cross-cut by Crystal Creek at mile 16.5; after crossing Crystal Creek, the Smiths Fork–age moraines are to the east of the road. At mile 16.6, outcrops of colluvium and landslide diamicton on top of a buried soil in Blacks Fork till can be viewed on the left side of the road. The road enters an ice-marginal drainage valley at mile 16.7 that separates the Smiths Fork moraines on the right from a Blacks Fork moraine on the left. At mile 18.0, park in the campsite on the right side of the road and climb on to the ridge on the east side of the campsite; this is Stop 1.5.

Stop 1.5—Ice-Marginal Drainage in Yellowstone Canyon (Mile 18.0)

The stream in Hells Canyon is visible directly north of this stop where it flows eastward to the Yellowstone River. The presence of an ice-marginal drainage valley to the south, however, suggests that drainage was to the south during the last glaciation and that the lateral moraine here once connected to the ridge on the north side of Hells Canyon (Fig. 6). A continuous lateral moraine would have blocked eastward drainage out of Hells Canyon and diverted flow southward along the distal side of the moraine, as evidenced by the presence of sand and rounded cobbles on the bottom of the valley through which we have been driving. Southward flow extended at least 3 mi south of this stop where the stream incised a relatively deep valley that was seen on the drive to Stop 1.4. Depending on the surface elevation of the

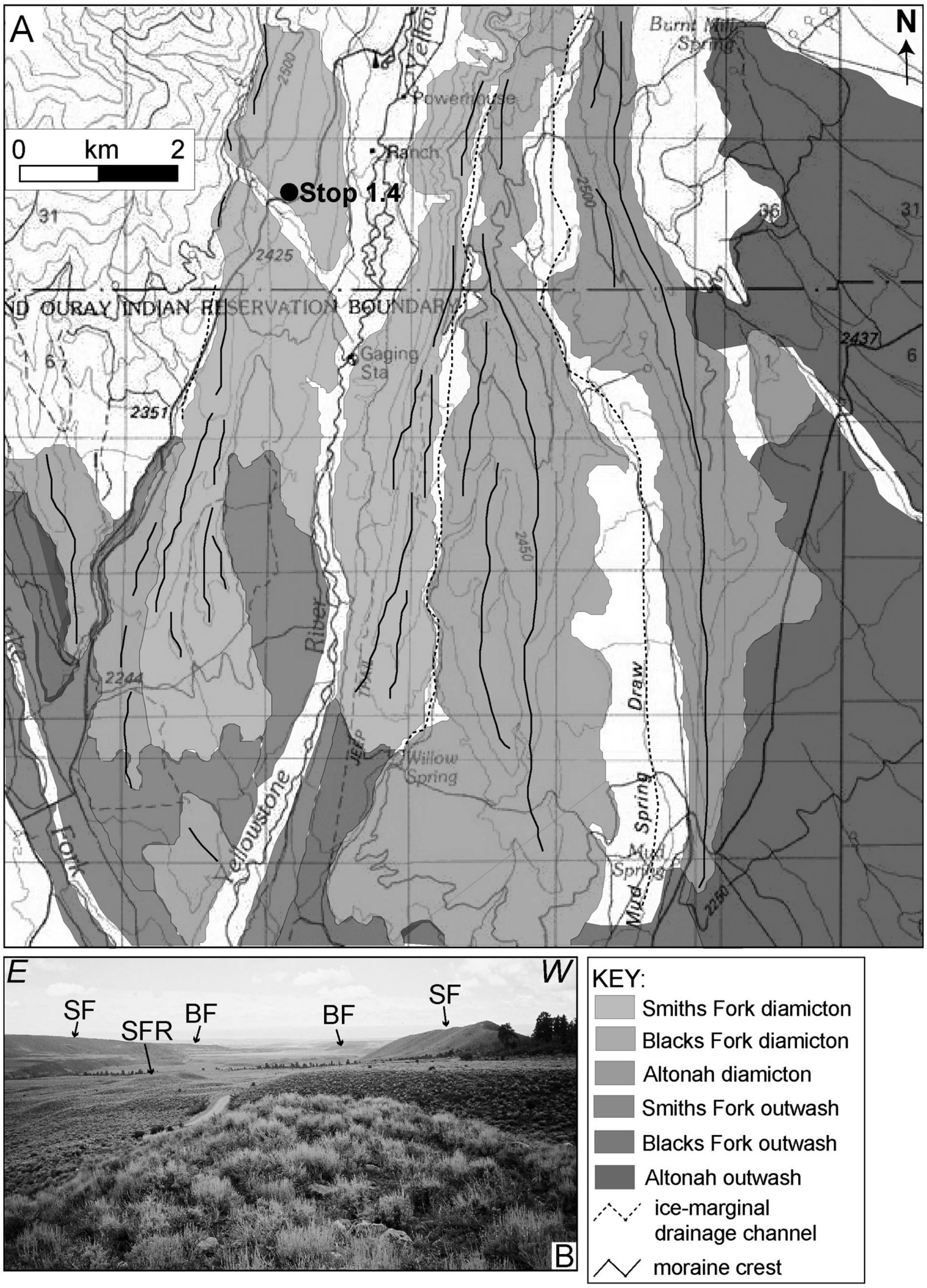

Figure 4. (A) Map of glacial deposits at the mouth of Yellowstone canyon. (B) View of moraine sequence to the south of Stop 1.4. This photo was taken at the top of a moraine just west of Stop 1.4). BF—Blacks Fork moraine; SF—Smiths Fork moraine; SFR—Smiths Fork recessional moraine. From Laabs and Carson (2005).

lateral moraine that has been removed from this area, drainage from Hells Canyon may have been ponded by the moraine and back-filled northward into Harmston Basin (see Fig. 2 for location; see also Fig. 6). Therefore, topographically closed depressions in Harmston Basin may contain a record of drainage from Hells Canyon during the last glaciation.

The stream in Hells Canyon breached the Smiths Fork lateral moraine here during the last deglaciation, flowed southward along the next successive lateral moraine to deposit kame terrace sediments, and then breached each successive lateral moraine and flowed southward before cutting through the youngest recessional moraine and adopting its current path eastward to Yellowstone River. As noted above, the presence of carbonate in kame terrace sediments indicates drainage from Hells Canyon along with the fact that no other source of ice-marginal drainage was available to deposit these sediments.

Directions to Stop 1.6

Continue driving along FR 227 and enter Hells Canyon at mile 18.6. The stream in Hells Canyon is ephemeral in the lower

Figure 5. Cross-sectional view of Smiths Fork lateral moraines and kame terraces west of Stop 1.4. This photo was taken at the top of a moraine just west of Stop 1.4. SF—Smiths Fork lateral moraine; SFR—Smiths Fork recessional moraine; SFK—Smiths Fork kame terrace.

part of the canyon, partly because some of the drainage is lost to the subsurface through sinkholes in the Madison Limestone. Note the abundance of sandstone and orthoquartzite cobbles and boulders in the talus along the right side of the road; these are derived from conglomerate and diamictite of the Duchesne River Formation. At mile 23.5, turn left onto FR 196 toward Mill Park. Weathered colluvial deposits derived from local outcrops of Tertiary gravels can be seen on either side of the road and at Mill Park. Lake Fork Mountain can be seen to the west across Mill Park at the junction at mile 24.4. Turn left at the junction, proceed ~50 ft, and turn right onto an unnamed forest road. The road intersects an ATV trail at mile 25.4; parking here is recommended, although the road continues toward the top of Lake Fork Mountain and can be driven in a high-clearance vehicle. If the road is wet, a four-wheel drive vehicle may be needed to continue driving past this point. Park near the intersection with the ATV trail and hike ~1 mi along the road to an air quality station at the top of Lake Fork Mountain (note: this hike ascends 400 vertical feet [122 m] at high elevations; although the climb is not steep, please use caution if you are not accustomed to hiking at elevations above 3000 m).

Stop 1.6—Lake Fork Mountain (Mile 25.4)

This site provides a broad view of the geomorphology of the southern Uintas and a rare opportunity to view terminal moraines and cirque headwalls from a single location that is (almost) accessible by car (Fig. 7). Lake Fork Mountain is an east-west–trending strike ridge of Madison Limestone bounded to the north by a normal fault. The Madison Limestone rarely forms continuous hogback ridges due to gently dipping bedrock in the southern Uintas, which contrasts with steeply dipping strata in the northern Uintas where hogbacks are common (see Stops 3.1 and 3.4).

Looking southward from Lake Fork Mountain affords views of the aforementioned moraine and outwash sequences on the piedmont below Lake Fork and Yellowstone River canyons. On a clear day, one can see across the Uinta Basin to the uplands of the San Rafael Swell. High peaks, cirques, and broad valleys in the headwaters of Lake Fork and Yellowstone River canyons can be viewed to the northwest and northeast, respectively (Fig. 7).

Broad, compound cirques that exceed 10 km in width (e.g., Brown Duck and Yellowstone basins; Fig. 7) and ultimately drain into either Lake Fork or Yellowstone River canyons are clearly visible to the northwest and northeast, respectively, and are representative of the distinct morphology of paleo-glacier accumulation zones in the southern Uintas. Such valleys must have contained many small glaciers at the onset of past glaciations that ultimately coalesced and joined the main valley glacier. Moraines in these valleys only exist in small cirques, and glacial deposits vary from several meters of till to a scattering of boulders over bedrock. The morphology of these high basins is distinct from those on the north slope (as viewed at Stop 3.1), where valleys are narrower and ice masses coalesced into smaller individual glaciers during the Smiths Fork glaciation. Although high basins lack topographic shading and receive higher amounts of solar

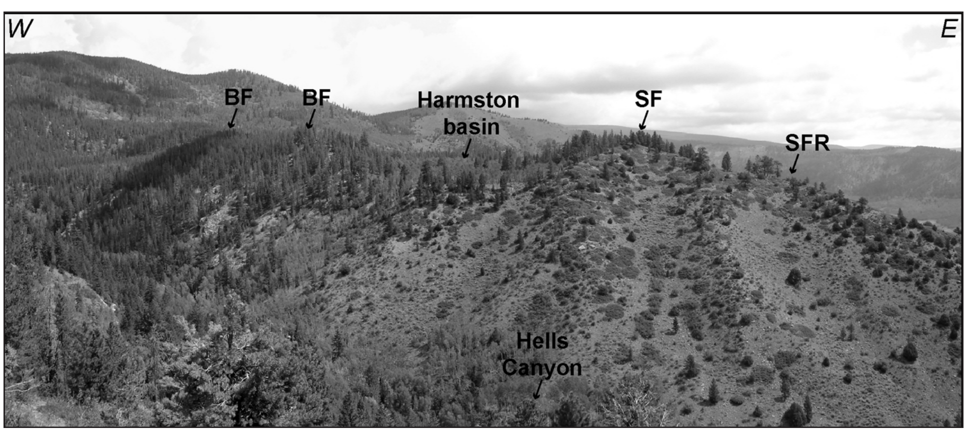

Figure 6. Photo of Harmston Basin as viewed to the north of Stop 1.6. BF—Blacks Fork–age moraine; SF—Smiths Fork–age moraine; SFR—Smiths Fork recessional moraine.

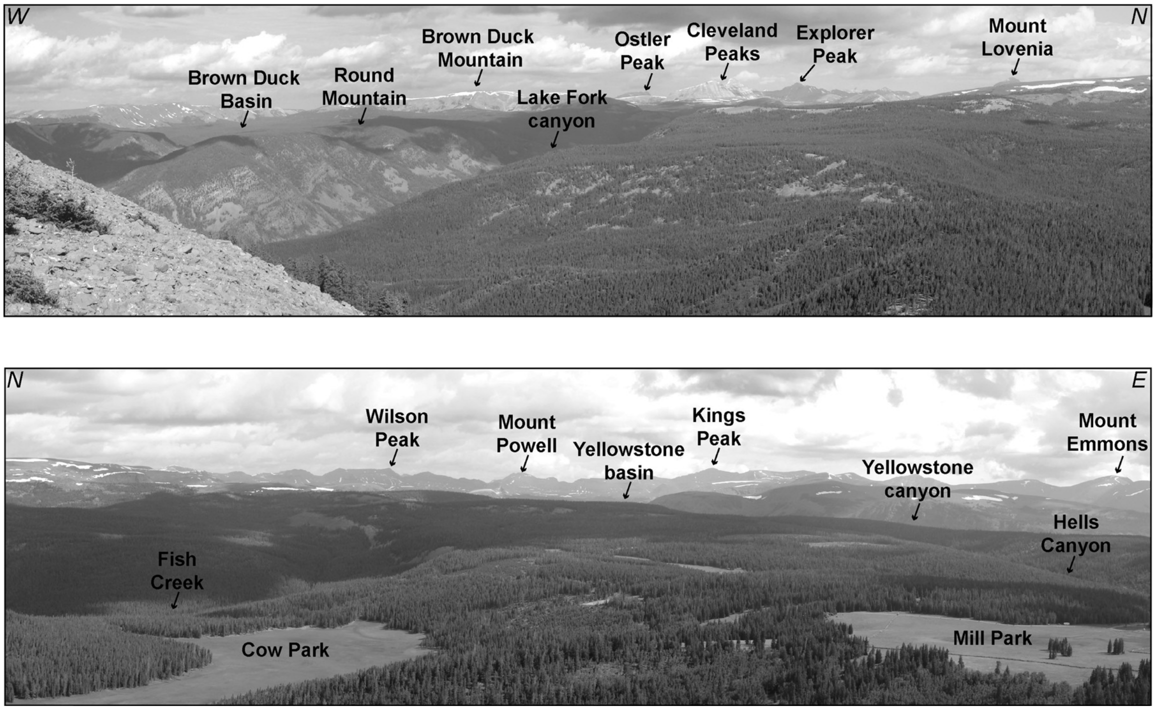

Figure 7. Panoramic photo of the high Uintas as seen from the top of Lake Fork Mountain. Top image shows the headwaters of Lake Fork basin as viewed to the northwest; lower image shows the headwaters of Yellowstone River basin as viewed to the northeast.

radiation than valleys on the north side, they are at elevations above the reconstructed equilibrium line altitude of the Smiths Fork maximum (~3000 m; Shakun, 2003; Laabs and Carson, 2005). The northeastern part of the Yellowstone drainage basin roughly coincides with the domal center of the Uinta Mountain anticline (see Stop 2.1, next section), which could explain why this is one of the highest areas in the Uinta Mountains and could therefore sustain ice masses in broad, poorly shaded basins under glacial climates.

Directions to Vernal

Descend Lake Fork Mountain and proceed southward. Do not turn right at mile 39.0; instead, proceed south toward Altonah. The Piñon-covered hills to the east at mile 39.9 are part of the Blacks Fork–age terminal moraine. The road climbs onto Smiths Fork–age outwash at mile 41.1 then ascends to Blacks Fork–age outwash at mile 41.5. At mile 42.2, the Blacks Fork–age terminal moraine is the tree-covered surface that rises above the outwash to the north. The Altonah left-lateral moraine comes into view at 10 o'clock just before mile 43.6. At mile 46.0, turn left at the stop sign onto 7000 North. Turn right onto 16000 West at mile 47.0 and proceed southward to a stop sign at mile 50.0. Turn left onto Utah Hwy 87 East, drive through Altamont, and bear right at mile 50.8 toward the towns of Mount Emmons and Upalco. Proceed on Hwy 87 and turn left onto U.S. Hwy 40 at mile 67.0. Bear right at mile 72.5 and follow signs to Vernal.

DAY 2—VERNAL TO RED CANYON LODGE—THE EASTERN UINTA MOUNTAINS

Introduction

Day 2 (total driving distance ≈106 mi [171 km]) focuses on the nonglacial Tertiary-Quaternary landscape evolution of the eastern Uinta Mountains. Ideas on the classic enigma of the integration of the Green River across the eastern Uinta uplift are reviewed, and new surficial mapping in western Browns Park is presented, which features deposits interpreted as a middle Pleistocene paleoflood and impoundment deposit associated with a landslide dam across the paleo–Green River.

Directions to Stop 2.1

Start in Vernal at the intersection of Main Street and 500 North (mile = 0.0). Turn east on 500 North, which becomes Jones Hole Road. Take the left fork after crossing Ashley Creek, continue over the south flank of the Buckskin Hills, and stay left at the intersection with Brush Creek Road at mile 5.7. At mile 6.5, note the unstudied sequence of Plio(?)-Pleistocene gravels and terraces along Brush Creek and at the top of the Buckskin Hills. Nelson and Osborn's (1991) work on the south flank to the west of here estimated relatively high Quaternary incision rates using soil-development indices on alluvial outwash surfaces. Stay right to cross Brush Creek, and climb to the Diamond Mountain

Plateau. At mile 13.8, the road has been damaged by landsliding of the Cretaceous Mancos shale. Visible at this point across the piedmont to the southeast, the Green River has incised the Split Mountain Anticline. The top of this Laramide structure is beveled by the Gilbert Peak erosion surface and capped by the Bishop Conglomerate, which is what the paleoriver probably flowed upon, enabling superimposition of the drainage.

Stop 2.1 (Mile 15.0)

This roadcut is an exposure of a tuffaceous facies in the middle of the Bishop Conglomerate atop the Diamond Mountain Plateau. Concordant plateaus and summits all around record a low relief post-Laramide landscape (Fig. 8). Two Tertiary deposits, the Bishop Conglomerate with its underlying Gilbert Peak erosion surface (seen here) and the Browns Park Formation (seen at next stop), provide a framework for understanding post-Laramide structural deformation, drainage development, and subsequent erosion of the eastern Uintas. The Gilbert Peak erosion surface is an extensive pediment that formed on the north, east, and south flanks of the range during a prolonged period of tectonic stability in Oligocene time. The Bishop Conglomerate was deposited basinward and atop the pediment as erosion continued at the retreating mountain front. An ash at this locality has produced a new Ar/Ar date of 30.54 ± 0.22 Ma (Kowallis et al., 2005).

Evidence for the regional tilting of the Bishop Conglomerate–Gilbert Peak surface lead Hansen (1986) to conclude that the northeastern part of the Uinta fault was reactivated with normal dip-slip motion in Neogene time. In Hansen's model, this and faulting associated with the formation of the Browns Park graben represent the gravitational collapse of the eastern part of the Uinta uplift (Bradley, 1936; Hansen, 1984, 1986). Note that the surface of the Bishop Conglomerate here on the Diamond Mountain Plateau has multiple terrace levels (Fig. 8). There has been significant later Tertiary and Quaternary reworking and beveling of this inherited planation surface.

Additional evidence for late Cenozoic deformation in the eastern Uinta Mountains includes the courses of numerous streams that have been altered and reversed (Hansen, 1984, 1986). Consequent streams flowed southward through the Diamond Mountain Plateau from the crest of the Laramide uplift in what is today the area of Browns Park. The courses of many of these drainages have now been reversed, as Hansen (1986) pointed out through evidence of an older, relict drainage divide and the barbed pattern of southerly flowing tributaries that make nearly 180° turns before joining northward-flowing trunk streams. The tilting and subsidence of Browns Park developed steep drainages that have eroded headward and captured south-flowing drainages, thus displacing the drainage divide southward (Fig. 9). This process continues today, as apparent along the upcoming field trip path, where Crouse Creek and Jackson Draw are very close to capturing the headwaters of Pot Creek.

Directions to Stop 2.2

Continue north and then east on paved Jones Hole Road across the Diamond Mountain Plateau. At mile 22.6, turn left onto gravel Crouse Canyon/Browns Park Road. Note the gently dipping hogback ridges held up by the Pennsylvanian Round Valley Limestone and the Mississippian Madison Limestone as we cross the beveled and buried southern flank of the Uinta uplift. Near the top of a Madison hogback, at mile 24.8, stay left at the fork (Crouse Canyon Road) and cross a drainage divide/windgap capped by Bishop Conglomerate(?). Ahead, to the north, broad valleys cut in UMG sandstones are filled with what has been mapped as Bishop Conglomerate. This is an example of the relict Eocene-Oligocene drainages Hansen (1986) interpreted as once flowing to the south (e.g., through this pass) from the high Laramide range crest, and which then foundered with the range, driving drainage change.

At mile 27.6, turn right on the intersecting gravel road and head east to mile 29.9, where we take a left turn and go north at

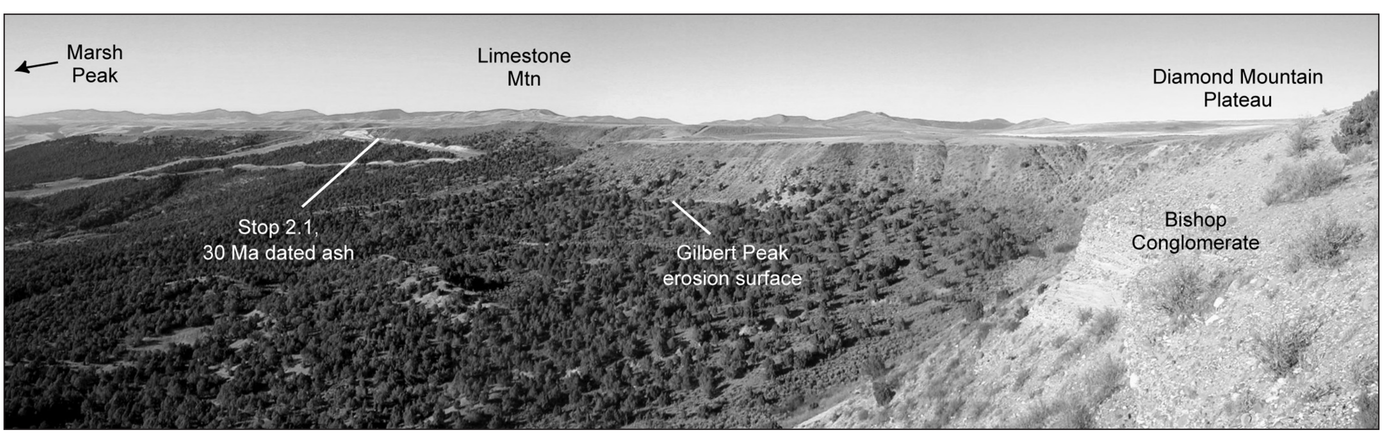

Figure 8. Photographic panel looking west from the south flank of Diamond Mountain Plateau toward the location of Stop 2.1. The broad topographic shoulder of the range is a bedrock pediment near the axis of the uplift, but it becomes progressively overlain by Oligocene gravel toward the basin margin, as epitomized by the Diamond Mountain Plateau.

the T. At mile 31.5, cross a subtle drainage divide and descend into the incised head of the Crouse Creek drainage, which is poised to capture the upper Pot Creek drainage system behind us. Note the last outcrops of mapped Bishop Conglomerate at mile 37.0, as we descend into the steepest reach of the youthful Crouse Canyon and out of the broad, relict Oligocene valley. Pit Draw, to the east, is an example of a barbed drainage meeting Crouse Creek. At mile 38.3, note significant talus with possible protalus ramparts (?) on the east side of the canyon. Turn left at mile 41.1 onto Taylors Flat Road after exiting the canyon mouth, cross Crouse Creek, and climb past roadcut exposures of tuffaceous Browns Park Formation. Continue to mile 42.7, a roadside stop overlooking Swallow Canyon.

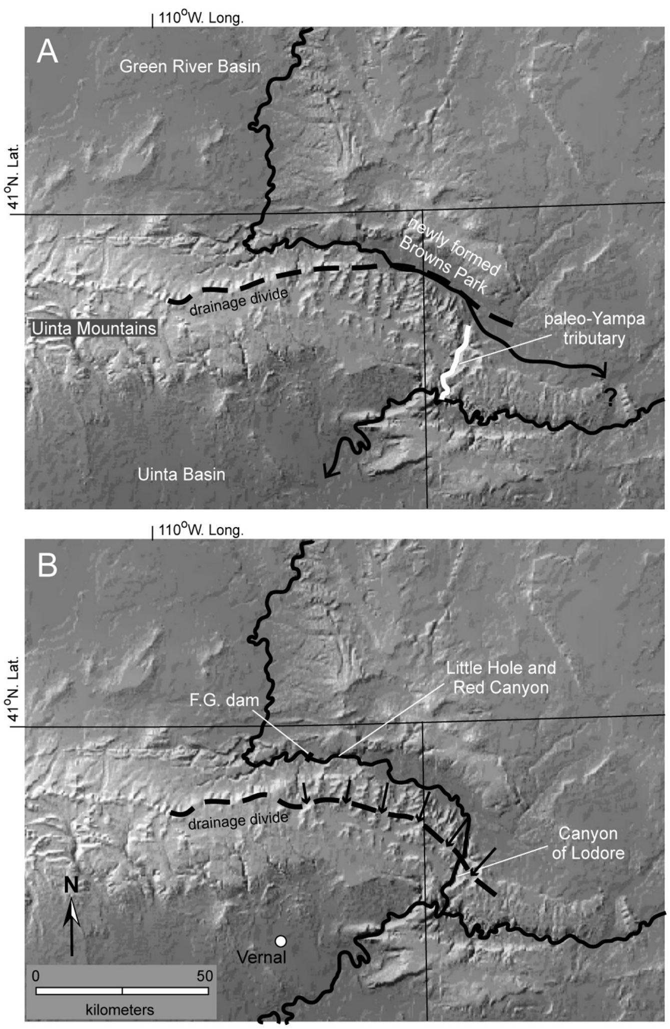

Figure 9. Map illustration of Green River integration history. (A) Oligocene-Miocene landscape with location of original paleodrainage divide indicated by dashed line. After Miocene subsidence of Browns Park and collapse of the eastern uplift, the upper Green River was diverted into the graben, and a tributary drainage of the paleo-Yampa underwent subsidence of its headwaters. (B) In post–Browns Park time (Pliocene?), the Green River was captured or diverted through the Canyon of Lodore, and the drainage divide shifted south to its present position in the eastern Uintas.

Stop 2.2 (Mile 42.7)

This stop provides an overview of where the Green River passes through a series of geologic transitions. Just west of Browns Park, the steep walls of Red Canyon temporarily open into a terraced, park-like area encompassing Little Hole and Devils Hole (Fig. 9). Farther downstream, the valley becomes narrow again as the Green River continues through the well-indurated UMG sandstones of lower Red Canyon. The canyon opens into western Browns Park, where bedrock changes to the weakly-consolidated tuffaceous sediment of the Tertiary Browns Park Formation. The Green River passes through the short bedrock gorge of Swallow Canyon at this stop and then continues through the heart of Browns Park. Quaternary landforms and deposits in Browns Park show evidence of minor displacement along the Home Mountain fault at the linear northwest basin edge, whereas the southern mountain front is generally highly sinuous and embayed. About 12 mi downstream at the center of Browns Park, the Green River turns south and punches through the Uinta uplift at the Gates of Lodore (Fig. 9). It then continues across the range past its confluence with the Yampa River, forming the deep canyons of Dinosaur Canyon National Monument.

There is a long history of debate on the timing and mechanisms of integration of the Green River across the Uintas and into the greater Colorado River. It was here that John Wesley Powell coined the term “antecedent” to relate his hypothesis about how the Green River came to cross the Uinta uplift and cut the Canyon of Lodore. This idea that the river path is older than a relatively young and active orogen was subsequently discarded in favor of some combination of superposition and stream capture. Swallow Canyon, seen here, is certainly an example of river superposition. Subsequent workers provided two hypotheses for the mechanisms of drainage integration across the Uinta uplift: (1) headward erosion of a small stream through the uplift and capture of the upper Green River drainage (Bradley, 1936; Hansen, 1965); and (2) superposition of the Green River over the eastern Uintas as it filled Browns Park with sediment and spilled over the range to the south (Sears, 1924; Hansen, 1986). These hypotheses each have problems. For example, headward erosion of a first-order stream through resistant sandstone is difficult, and there is no known evidence in the Browns Park Formation for sediment of a proto–Green River.

A review and new perspective on these options provides a preferred hypothesis for drainage integration (Pederson and Hadder, 2005). Hansen (1969) recognized that the subsidence associated with Miocene extensional collapse of the eastern Uintas must have played a key role in capturing or diverting an upper Green River south into Browns Park, as well as south from Browns Park across the eastern Uinta uplift. The paleo–Green River, once diverted into the subsiding Browns Park, would have found a gentle path continuing along the strike of the graben into northwestern Colorado (Fig. 9). But the path of the Canyon of Lodore follows the trend and spacing of the relict, subsided drainage headwaters of the Diamond Mountain Plateau, such as the one just followed along Crouse Canyon. The Laramide (pre–Browns Park) drainage divide would have extended into the heart of Browns Park in the area of the Gates of Lodore (Bradley, 1936; Hansen, 1986) (Fig. 9). That is, relict Oligocene topography suggests that the headwater drainage divides of a steeper, south-flowing tributary of the paleo-Yampa subsided into the Browns Park graben during Miocene faulting. This would have greatly eased the path for headward erosion of the tributary or for the spilling or diversion of the nascent upper Green River through the fallen divide atop the Browns Park Formation in late Miocene or Pliocene time (Fig. 9).

The surficial stratigraphy here in western Browns Park is being studied, in part, with the goal of quantifying incision rates. Given the identification of the Lava Creek B tephra (see Stop 2.4) and mapping correlations, Counts (2005) estimated a Quaternary incision rate of 80–115 m/m.y. in western Browns Park. This is only about half that calculated in most other places in the region, including many likewise based on the Lava Creek B tephra (Table 1). This relatively low incision rate estimate on the north flank of the Uintas also contrasts with estimates made just across on the southern flank of the Uintas (Nelson and Osborn, 1991). Similarly, though not necessarily related, GIS-based calculations indicate the total late Cenozoic exhumation of the Green River Basin to the north is approximately half that of the Uinta Basin to the south of the Uinta Range (Pederson, 2004) (Fig. 10).

An explanation is required for this lower total exhumation and, if initial results hold true with more geochronology, lower incision rates. There is no evidence in the Browns Park area for differential climate changes or significantly active subsidence or faulting in the Quaternary that could have reduced total exhumation and incision rates. To some degree, less exhumation to the north may be a function of the more recent integration of the Green River Basin into the overall system, as discussed above. But considering that incision is ultimately driven by baselevel fall, reduced exhumation to the north may also reflect buffering of the region from baselevel fall that has driven incision elsewhere. The upper Green River is isolated in the headwaters of the greater Colorado River system, far from its ultimate baselevel at the Gulf of California. In addition, the upper Green River is located above several steep canyon reaches (Fig. 10). Any baselevel fall downstream, whether caused by epeirogenesis or river integration, would diffuse and get absorbed or “hung up” in canyon reaches where stream power is concentrated to meet the resistance of harder bedrock or coarse sediment loading.

Directions to Stop 2.3

Continue west on the Taylor Flats Road. Cross Sears Creek at mile 48.7, and then cross the terrace associated with Quaternary alluvial gravel 5 (Qag5) for the next few miles to the scattered dwellings of Taylors Flats at mile 51.5. This landform dominates western Browns Park, and most of the dissected piedmont fans and slopes in the area grade to this level. Continue north across the Green River at Bridge Hollow at mile 52.8, turn left toward the Jarvie Ranch Historic Site, and pull into parking area.

Stop 2.3 (Mile 53.1)

Hike up a short but steep slope to south of Jarvie Ranch, over Browns Park Formation, to a very coarse basal Qag5 deposit that grades up to typical Green River gravel and is finally capped with distinct piedmont sand and gravel. This outcrop provides an overview of the Quaternary stratigraphy of the Green River and an introduction to the Qag5 flood deposit.

There are at least nine stratigraphically distinct and inset mainstem Green River deposits preserved in western Browns Park. These gravel and sand deposits generally range from 2 to 12 m thick and are associated with strath or thin fill terraces that have tread heights up to 227 m above modern grade (Fig. 11). Based on local piedmont equivalents as well as the landscape position of the latest Pleistocene (Smiths Fork) glacial outwash in the Henrys Fork field area (visited on Day 3), we interpret the

TABLE 1. CALCULATED INCISION RATES FOR REGIONAL RIVERS

Rivers	Location	Incision rate	Age control	Study
Yampa River	Northwestern Colorado	110–150 m/m.y.	Lava Creek B ash	Reheis et al., 1991
Bighorn River	Bighorn Basin, Wyoming	160 m/m.y.	Lava Creek B ash Field Camp ash	Palmquist, 1983
Wind River	Wind River Basin, Wyoming	150 m/m.y.	Lava Creek B ash	Chadwick et al, 1997
Lake Fork, Yellowstone, Uinta	NW Uinta Basin	170–320 m/m.y.*	Soil-development indices	Nelson and Osborn, 1991
Green River	NE Uinta Mountains, Browns Park	90–115 m/m.y.	Lava Creek B ash	Counts, 2005
Henrys Fork	NE Uinta Mountains, Manila, Utah	80–110 m/m.y.	Correlation to dated Wind River stratigraphy	Counts, 2005

*Reported Uinta Basin rates are for the mountain piedmont for comparison to same on the northern flank of range. These are higher than those reported (Nelson and Osborn, 1991) near the center of the Uinta basin.

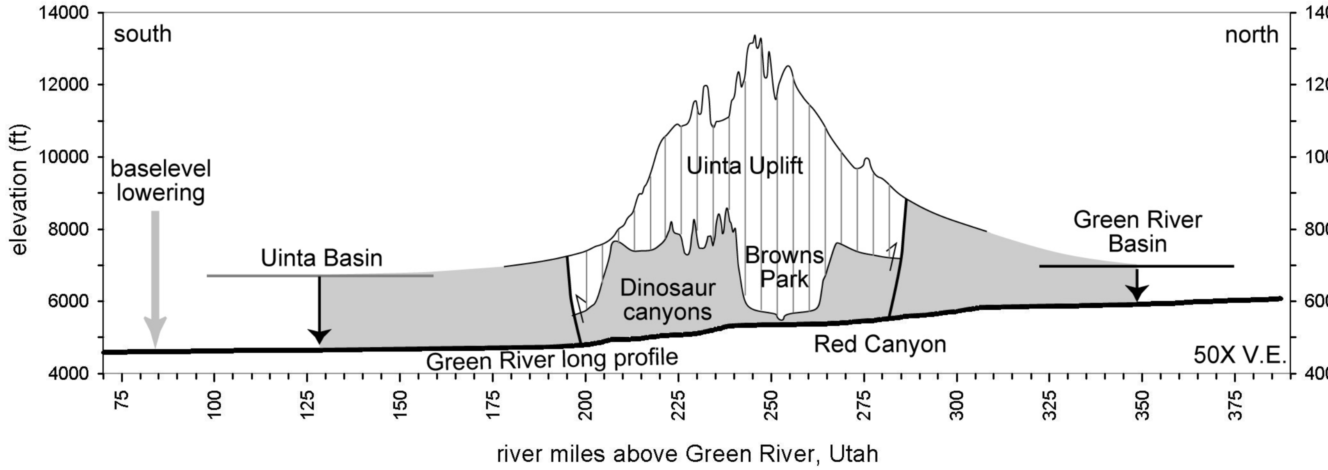

Figure 10. Schematic cross section along Green River, with river long-profile digitized from Trimble (1924). "Dinosaur canyons" are Lodore, Whirlpool, and Split Mountain canyons of Dinosaur National Monument. Uinta and Green River basins show basin-averaged late Cenozoic exhumation calculated following the methods of Pederson et al. (2002). The striped interval of Uinta uplift is a range of peak elevations between a higher profile across the peaks in the central-western part of range and a lower profile taken across the eastern Uintas. The signal of baselevel lowering that has been driving regional incision should diffuse and dampen as it propagates upstream and encounters steep canyon reaches.

presence of Qag1 Green River gravels to be mostly below grade, deposited during the LGM. The sandy upper alluvium and terraces of Qag1 visible in the floodplain are probably Holocene in age, and the youngest of these (historic) have been studied by Grams and Schmidt (1999). Qag1, Qag3, and Qag4 are distinct from this viewpoint, and higher gravels are visible, but the predominant surficial deposits in Browns Park and upstream in the lower reaches of Red Canyon are Qag5 sand and gravel (Fig. 11). These differ strongly in character between here and the next stop in lower Red Canyon at Little Hole, but they are correlative in terms of their position within the Quaternary stratigraphy and their interpreted genesis.

The Qag5 here is very distinct from typical Green River gravels. It is underlain by a planar strath cut in the Browns Park Formation 40–50 m above grade, and the deposit thins strongly downstream from over 25 m thick at the mouth of Red Canyon to 10 m thick near Stop 2.2. The exposure of Qag5 near Jarvie Ranch is 23 m thick (Fig. 12), and Qag5 has a fan-shaped geometry that spreads across much of the valley floor in this area. The basal 18 m at this locality is a clast-supported boulder gravel with boulders that are exclusively subangular UMG sandstone up to 3.3 m in intermediate-axis diameter, with a mean diameter of 0.84 m (Counts, 2005). The framework openings are filled with a matrix of sand and rounded pebbles and cobbles of polymictic Green River provenance. Boulders are abruptly absent above 18 m, grading sharply to better sorted and rounded pebble-cobble gravel of the Green River. The caliber and dominance of UMG clasts in the base of the Qag5 deposit decreases strongly downstream, starting with boulders up to 5 m in intermediate diameter at the mouth of Red Canyon. The placement of these largest boulders at the apex of a fan-shaped deposit indicates an origin

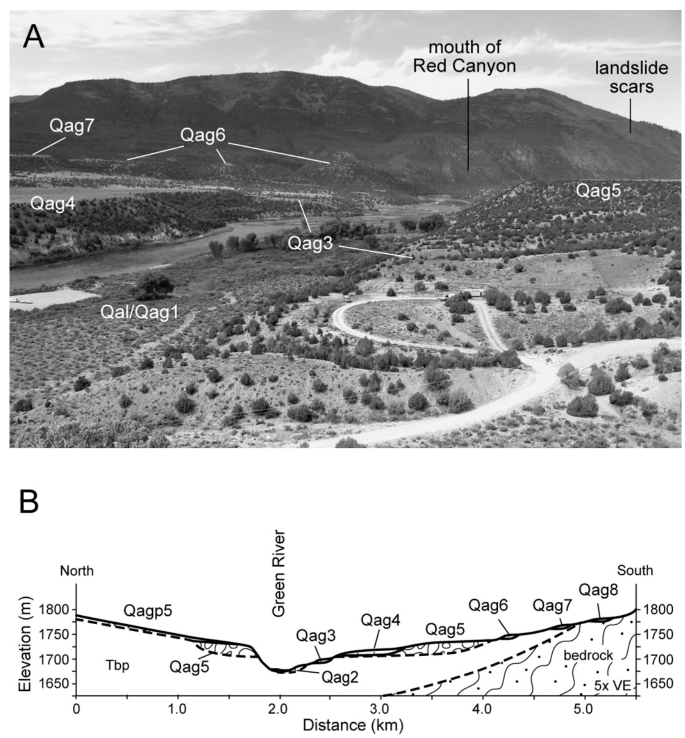

Figure 11. (A) Photograph looking upstream (west) from Jarvie Ranch at Green River terrace gravels and mouth of lower Red Canyon. A ~50-m-thick shale formation within the Uinta Mountain Group outcrops at the base of the south slope in lower Red Canyon on the outer bank of the undercutting river. (B) Composite surveyed cross section from the Jarvie Ranch area (modified from Counts, 2005). Qag5 deposits dominate the valley bottom and are anomalous in their sedimentology, thickness, and extent. Qag—Quaternary alluvial gravel of mainstem drainages; Qal—alluvium of active channels and floodplains.

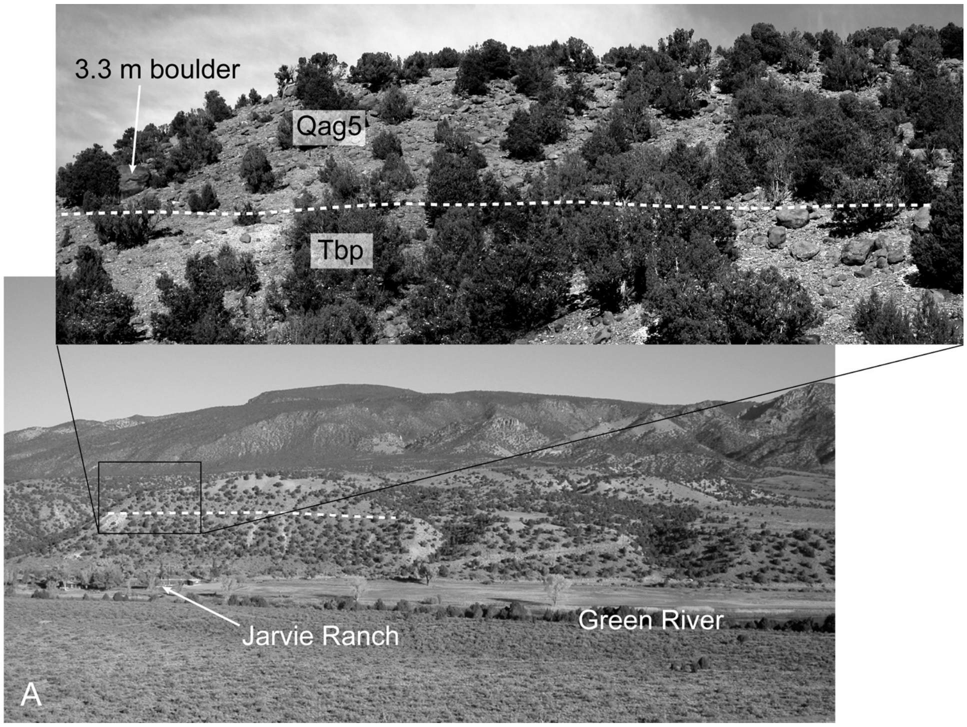

Figure 12. Overview of Qag5 deposit exposed immediately north of the historic Jarvie Ranch in western Browns Park; river flows from left to right. Inset is close up of the boulder gravel at the base of the deposit (from Counts, 2005). Qag—Quaternary alluvial gravel of mainstem drainages; Tbp—Tertiary Browns Park Formation.

from lower Red Canyon. Hansen (1965) interpreted these as a coarse facies of the underlying Browns Park Formation adjacent to steep paleocanyon walls, and there are, in fact, such deposits in the distinct Browns Park Formation nearby.

Landslide scars and deposits cover much of the south wall of lower Red Canyon between the confluence of Red Creek and the mouth of Red Canyon (Fig. 12). In this reach of the canyon, a ~50-m-thick formation of Neoproterozoic shale within the UMG is exposed at river level (Dehler et al., 2006). This creates a situation in which the outer bank of the river impinges upon shale, undercutting a steep slope of sandstone that rises nearly 800 m above the river. The surface of the massive landslide deposit, according to field relations, predates at least Qag3, and it has been modified by fluvial process into a sloping terrace that sits at an elevation ~40 m above the top of the Qag5 mainstem deposit here at Jarvie Ranch. However, its elevation is concordant with the top of the anomalous Qag5 deposit at Little Hole (next stop). We hypothesize this older landslide impounded the Green River and that the failure of this landslide dam is recorded in the basal Qag5 here in western Browns Park.

Directions to Stop 2.4

Drive out of Jarvie Ranch to the east, straight past Bridge Hollow, and turn left at the T-intersection toward the mouth of Jesse Ewing Canyon at mile 54.8. At mile 56.3, the Home Mountain fault bounding Browns Park is crossed and the road ascends the steepest part of Jesse Ewing Canyon. At mile 58.9, cross the drainage divide and enter the Clay Basin gas field after crossing the Uinta-Sparks fault zone at the northern structural edge of the Uinta uplift. Stay left at the intersections and cross Red Creek as the road changes to pavement at mile 64.1. The road ahead follows the Uinta fault, separating, at left, Precambrian metamorphic rocks of the Red Creek Quartzite, and to the right, tilted Cretaceous (Mesa Verde Group) to Eocene (Wasatch Formation on skyline) sedimentary rocks of the Green River Basin. Here is the greatest throw of any structure in the Uintas, and it presumably once hosted the highest part of the Laramide range to the south, before collapse.

Cross back onto the gravel road at the Wyoming state line at mile 72.1, and at mile 75.0, turn left on Hwy 191, crossing the Utah state line through a hogback of the Mesa Verde Group. Continue through the hogback of Jurassic Glen Canyon (Nugget) Sandstone to Dutch John at mile 83.6. Turn left and proceed to the Little Hole overlook turnout on the right side of the road at mile 89.0.

Stop 2.4 (Mile 89.0)

This stop at Little Hole, a small park-like area within lower Red Canyon 10 km downstream of Flaming Gorge Dam, focuses on the anomalous Qag5 fill and features a short optional hike to an outcrop containing Lava Creek B tephra.

Highly variable surficial deposits are preserved at Little Hole. The four lowest deposits are 5–10 m thick, typical Green River gravels that are each associated with a distinct terrace (Fig. 13). In contrast, the sand-dominated Qag5 deposit at Little Hole and just downstream at Devils Hole are at least 60 m thick (the base of the deposit is obscured here), and the regular Green River terrace gravels are inset into this older fill. Qag5 here is generally a medium-scale, cross-bedded, pebbly sand with clast-supported, relatively immature, pebble-boulder gravel lenses of locally derived UMG clasts (Fig. 13). The middle-upper exposed deposit generally coarsens upward, and pebbles of brown and tan quartzite, Paleozoic, and Mesozoic bedrock clasts typical of mainstem Green River deposits appear as a terrace-gravel cap near the top. Laterally equivalent, coarse, local piedmont gravels of UMG cobbles and boulders grade to the higher preserved terrace level near the top of the deposit. At both Little Hole and Devils Hole, thin- to medium-scale lenticular beds of fine grained, in most cases reworked, tephra lie near the preserved top of the deposit (Fig. 13). Geochemical analysis of these beds indicates the presence of Lava Creek B ash (Mike Perkins, 2004, personal commun.). Thus, the upper strata of this anomalous sandy deposit were emplaced soon after the Lava Creek B eruption at ca. 640 ka in the middle Pleistocene (Lanphere et al., 2002).

Hansen (1965) recognized that these deposits were in some ways more similar to the Browns Park Formation than to typical Pleistocene Green River deposits, and he understandably mapped them as the heterogeneous Browns Park basin fill. We have reinterpreted this middle Pleistocene deposit as sediment of the Green River and local tributaries deposited in the accommodation space upstream of the hypothesized landslide dam in lower Red Canyon. The Qag5 terrace here is at approximately the same elevation as the reworked top of the landslide toe. Results of sand petrography of the Qag5 deposit at Little Hole support this interpretation, with QFL point-count data of samples indicating a composition that matches Pleistocene Green River deposits rather than examples of basin fill in western Browns Park (Counts, 2005).

Directions to Red Canyon Lodge

Return to Dutch John (mile 94.4), turn left, and head south on Hwy 191, cross Flaming Gorge Dam, climb out of canyon, and turn right at the intersection with Rt. 44 at mile 102.5. Continue west to mile 106.3 then turn right on Red Canyon Road to Red Canyon Lodge.

DAY 3—RED CANYON LODGE TO SALT LAKE CITY—THE NORTHERN UINTA MOUNTAINS

Introduction

The final day of this trip (driving distance ≈225 mi [360 km]) focuses on the latest Pleistocene glacial geology and geomorphology of the north slope of the Uintas, returning to Salt Lake City on I-80 through Echo and Parley's canyons.

Directions to Stop 3.1

From Red Canyon Lodge (mile 0.0), backtrack to Rt. 44, and turn right. At 9.4 mi, the road descends into the deeply incised Carter Creek drainage, which carried meltwater away from the northeasternmost glaciers in the Uintas. A 12.1 mi, turn left on

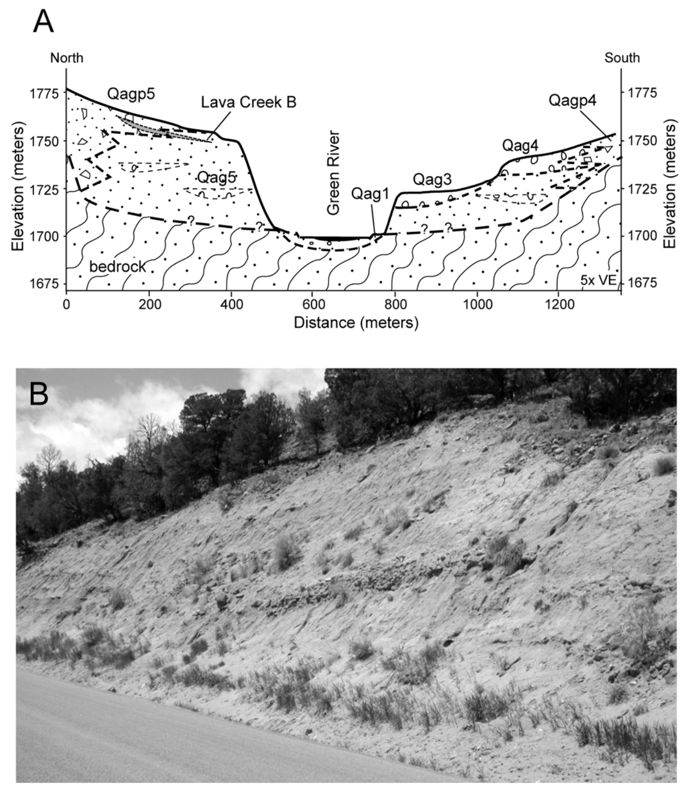

Figure 13. (A) Composite surveyed cross section from Little Hole (modified from Counts, 2005). The Qag5 deposit is anomalously sandy and thick (note that vertical exaggeration is the same as in Figure 11B). (B) Roadcut of Qag5 deposit at Little Hole is primarily fluvial sand containing lenses of UMG-dominated cobble gravel derived from local catchments. Typical mainstem and piedmont terrace gravels overlie and are inset into the sandy fill, and the Lava Creek B tephra lies near the top. Qag—Quaternary alluvial gravel of mainstem drainages; Qagp—Quaternary alluvial gravel of piedmont drainages.

the Sheep Creek Geologic Loop (FR 218). Views of Leidy Peak and the cirques at the head of Carter Creek appear at mile 15.2. Turn left at mile 15.3 on FR 221, and immediately bear right, following the signs for the Ute Tower. The fire tower is accessed via a short (1.3 mi) side road (FR 005) on the left at mile 16.6.

Stop 3.1—Ute Tower (Mile 16.6)

The Ute Tower provides a tremendous overview of the eastern High Uintas that is an illustrative way to start the third day of this field trip. Due south of the tower is the easternmost alpine summit of the High Uintas, Leidy Peak, named for nineteenth century paleontologist Joseph Leidy (Figs. 14 and 15). Leidy Peak rises above the extensive, gently sloping, summit flats known as "bollies" at this end of the range. In his monograph on the geomorphology of the northern Uinta, Bradley (1936) interpreted this summit surface as a remnant of a pediment formed prior to mountain uplift during a long period of erosion under a semiarid climate. However, recent research and modeling on

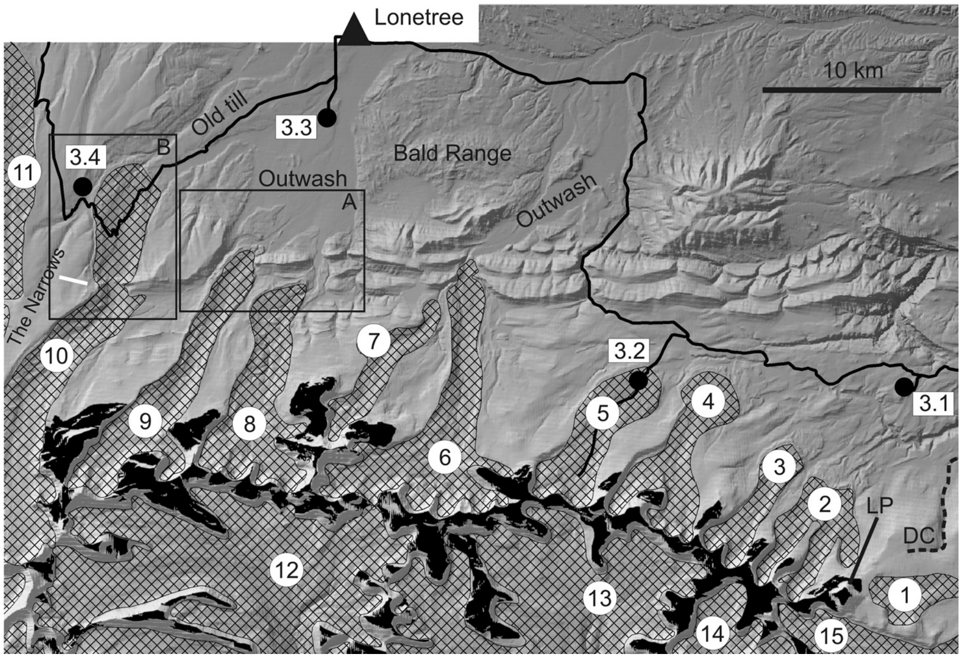

Figure 14. Map showing the locations of stops on the eastern north slope of the Uintas overlain on a 30-m digital elevation model. The dark line marks the field trip route (east to west); the stops are labeled as in Figure 1. Leidy Peak (LP), Deep Creek (DC), and The Narrows (see text) are highlighted. Box A marks the location of Figure 17; Box B marks Figure 20. Outlines of glaciers at the Smiths Fork maximum (Munroe, 2001; Laabs, 2004) are shown with the cross-hatched pattern. Black highlights the high elevation summit flats of the eastern Uintas (from Munroe, 2005). Glacier names: 1—North Fork Ashley Creek; 2—East Fork Carter Creek; 3—West Fork Carter Creek; 4—South Fork Sheep Creek; 5—Middle Fork Sheep Creek; 6—Burnt Fork; 7—Thompson Creek; 8—Middle Fork Beaver Creek; 9—West Fork Beaver Creek; 10—Henrys Fork; 11—East Fork Smiths Fork; 12—Uinta; 13—Whiterocks; 14—Dry Fork; 15—South Fork Ashley Creek.

summit flats in the Wind River Range, ~200 km to the north (Anderson, 2002; Small et al., 1997), has suggested that summit flats can form in place through periglacial hillslope processes and can result solely from the difference in rate between periglacial and glacial erosion. Given this theory, Munroe (2005) investigated the distribution of summit flats throughout the Uintas and found that these surfaces are considerably more extensive at the eastern end of the range (Fig. 14). He speculated that the paucity of summit flats at the western, upwind, end of the Uintas may reflect more effective glacial erosion there over the course of the Quaternary. This idea is corroborated by reconstructions of Smiths Fork glaciers by Oviatt (1994), Munroe (2001), and Laabs (2004), which indicate that the western Uintas were more extensively glaciated during at least the last few glacial cycles.

The contrasting morphologies of glaciated and nonglaciated valleys on the north side of Leidy Peak provide a good example of a "glaciation threshold" conditioned by local snow redistribution. Deep Creek (DC on Fig. 14) begins on the northeastern shoulder of Leidy Peak at an elevation of ~3120 m and flows eastward for 1.5 km before turning abruptly northward and descending almost 250 m in 3 km. Throughout this length, the Deep Creek valley exhibits the typical V-shaped profile of a fluvial system, and nowhere do deposits typical of glaciation interrupt the course of the river. In contrast, the East Fork Carter Creek, located immediately west of Deep Creek, originates in the easternmost cirque on the north slope. Red Lake (3015 m) in this cirque is overdeeped by at least 17 m (Pettengill, 1996) and is flanked along its western and southern sides by a steep headwall that rises 200 m to meet the bollies on the north side of Leidy Peak. The headwaters of both drainages lie at similar elevations and are only 4.5 km apart, yet Deep Creek was apparently never glaciated. This relationship underscores the sensitivity of small glaciers to local variations in snow accumulation. The East Fork Carter Creek lies immediately downwind of the summit flats flanking Leidy Peak (Fig. 15). Snow blown from this surface—the easternmost extensive alpine upland in the Uintas—nourished the accumulation area of the East Fork Carter Creek glacier, while Deep Creek, arising at a similar elevation, but slightly farther east of Leidy Peak, did not receive as large a volume of drifting snow.

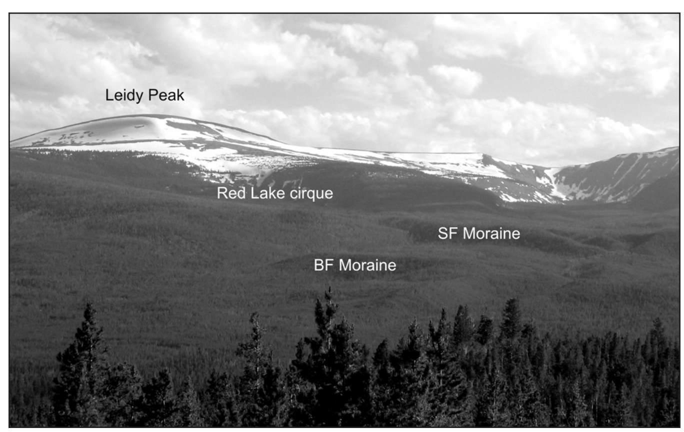

Figure 15. View southward from Ute Tower. Leidy Peak is the easternmost alpine summit in the Uinta Mountains; Red Lake cirque was the accumulation area for the easternmost glacier in the northern Uintas. The Smiths Fork (SF) and Blacks Fork (BF) age moraines are visible as prominent benches, highlighted by the evening light.

End moraines of the Smiths Fork and Blacks Fork glaciations are also visible to the south and southwest from Ute Tower, where Smiths Fork–age glaciers in the East and Middle Forks of Carter Creek coalesced to form a piedmont lobe (Fig. 15). The prominent terminal moraine runs ~E-W for several kilometers at an elevation of 2775 m, enclosing Tepee Lakes as well as other unnamed ponds (Leidy Peak quadrangle). The moraine is typically sharp-crested, with a steep frontal slope that rises up to 60 m above the older till deposits to the north. The analogous terminal moraine in the West Fork Carter Creek of Smiths Fork–age is over 35 m high and fronts a chaotic zone of hummocky topography including numerous water-filled closed depressions.

Older glacial deposits are also well preserved in the Carter Creek region. Immediately in front of the Smiths Fork terminal moraine is a lower relief, Blacks Fork–age, drift sheet that has a slightly hummocky surface and contains only isolated kettles (Fig. 15). Blacks Fork till is also present beyond the Smiths Fork–age moraine along the West Fork Carter Creek. This surface also displays subdued hummocks, partially filled kettles, and poorly drained areas. No evidence for pre–Blacks Fork till is found in the Carter Creek area. Instead, outcrops of bedrock ledge and flaggy surface clasts indicate that surficial deposits beyond the Blacks Fork moraines were derived from in situ weathering of bedrock.

Directions to Stop 3.2

Descend the road from Ute Tower and turn left on FR 221 (mile 19.2). End moraines of the West Fork Carter Creek are visible straight ahead at mile 22.0, after leaving Half Moon Park. Continue past the turn for Browne Lake at mile 22.7, passing several bedrock-cored hills similar to the one on which Ute Tower is located. At mile 28.8, turn left on FR 001 following the signs for Spirit Lake. The road passes around the outside of Hickerson Park before rising toward the terminal moraine complex in the South Fork Sheep Creek valley. At mile 29.4, the Smiths Fork–age terminal moraine forms the skyline ridge straight ahead, while the unglaciated North Fork Ashley Creek is visible on the right. The road climbs onto subdued Blacks Fork–age till at mile 30.2 and reaches the front of the Smiths Fork–age terminal moraine at mile 30.7. Park in the turnout on the left before the road curves to the right.

Stop 3.2—Smiths Fork–Age End Moraine, Middle Fork Sheep Creek (Mile 30.7)

During the Smiths Fork glaciation, glaciers in the Middle and South forks of Sheep Creek were larger than those in the Carter Creek drainages (~25 km^2 versus <20 km^2). However, glaciers in all of these valleys, more than those in other drainages of the Uintas, tended to broaden out in their terminal zones, forming piedmont lobes. The uniqueness of these glaciers is apparently due to the structural geology in this part of the eastern north slope. Because the glacial valleys dip less steeply than the Uinta Mountain Group in this area, the valleys become progressively shallower to the north, allowing the glaciers to broaden into piedmont lobes on the steeply dipping north flank of the Uinta arch. Furthermore, all four of these glaciers terminated upslope from the Paleozoic hogback, so they did not encounter this prominent topographic obstacle as glaciers farther west did. Divergent ice flow across the bedrock dipslope assisted ice stagnation within the terminal zones by increasing the rate of ice thinning toward the margin. As a result, Smiths Fork–age terminal moraines along the forks of Sheep Creek and Carter Creek incorporated large amounts of buried ice and consequently formed areas of hummocky topography.

Deposits of at least two glaciations are present along the forks of Sheep Creek, straddling the boundary between the Whiterocks Lake and Phil Pico Mountain 7.5′ quadrangles (Fig. 14). Immediately south of Hickerson Park is an unglaciated area of bedrock-controlled topography featuring linear strike ridges of quartzite, colluvial basins controlled by bedrock sills, and a conspicuous absence of subrounded quartzite erratics. Moving southward, this thin, locally derived regolith gives way to a heavily eroded till unit, identifiable by its abundance of subrounded erratic boulders. No moraine marks the outer limit of this drift sheet, but the appearance of the first erratics south of the Lodgepole Creek crossing along FR 001 is conspicuous, and the position and elevation (~2810 m) of this deposit suggest that it is analogous to the Blacks Fork–age deposits along Carter Creek.

The main focus of this stop is the Smiths Fork–age terminal moraine, found at an elevation of 2875 m along FR 001. Accessible, extensive exposures of Smiths Fork–age till are rare in the Uintas, and the prominent roadcut on the west side of the road is exceptional because of its scale; its freshness is ensured by road maintenance. The exposure reveals the diamicton typically found in Smiths Fork–age terminal moraines in the northern Uintas, which were apparently produced by a wide variety of sedimentary mechanisms. Large surface boulders (>50 cm in diameter), some partially buried by eolian silt, were likely deposited directly from a supraglacial position at the ice margin, either due to wasting of the underlying ice or by direct rolling and sliding to the glacier terminus. The sandy till present below the silt cap is unstratified and is interpreted as meltout till deposited at the ice margins. Fabric is difficult to evaluate in most Smiths Fork–age till due to the paucity of elongate clasts. Basal till is rarely exposed at the surface in the northern Uintas, although diamicton containing striated bullet clasts exposed by landsliding along the Little West Fork Blacks Fork may represent basal till. Overconsolidated till with well developed columnar jointing is exposed along the West Fork Beaver Creek upstream from the Smiths Fork–age terminal moraine complex. A unique feature of the exposure on Spirit Lake Road is the slabs of bedrock exposed in the sandy till matrix. Is this evidence for wholesale entrainment of bedrock ledges from points further south in the basin? Or is this moraine bedrock-cored? We will spend some time trying to answer this question before moving on to our next stop.

Directions to Stop 3.3

Retrace the route back to FR 221, and turn left at the intersection (mile 32.6). At mile 35.3, continue straight on FR

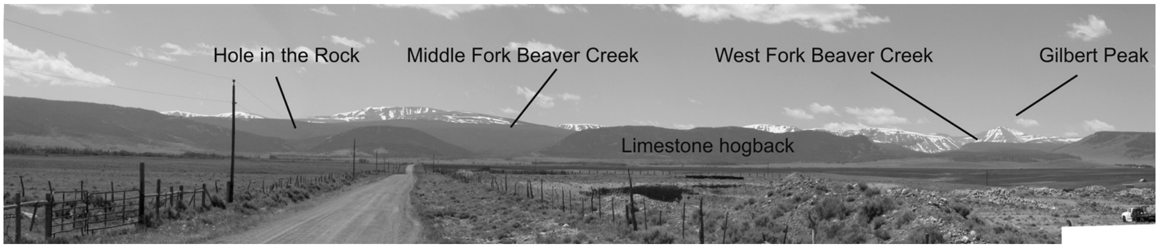

Figure 16. Hole in the Rock and the Beaver Creek area viewed from Stop 3.3 near Lonetree, Wyoming.

221 and descend Birch Canyon to the north. Red Pine Shale is exposed along the road near mile 35.7, and the valley narrows as the Madison Limestone is encountered near mile 37.4. Views of the Smiths Fork–age outwash surface graded to the terminal moraine in the Burnt Fork valley open up on the left at mile 42.4. Turn left on Hwy 414 toward Lonetree, Wyoming, at mile 45.9. At mile 46.6, the road curves left and crosses the Burnt Fork outwash surface, and at mile 47.8, passing the marker for the 1825 Rendezvous, the forested hills of the Bald Range are visible in the middle distance. Enter Uinta County at mile 50.8, with views of the Henrys Fork on the right at mile 52.1. Turn left at mile 56.4 on CR 290 toward the Henrys Fork, and then turn left on CR 291 at mile 57.4. The rounded sage-covered upland on the right after the turn is underlain by pre–Black Fork-age till. Continue straight through the intersection at mile 58.8, and park at the entrance to the gravel pit on the right at mile 59.9.

Stop 3.3—Gravel Pit South of Lonetree, Wyoming (Mile 59.9)

This stop provides a spectacular view of the north face of the High Uintas and allows us to review the alluvial stratigraphy below the glacial termini of the Henrys Fork drainage. Lonetree is situated at the northern end of an expansive compound outwash surface formed by the combined meltwater from the Henrys Fork, West Fork Beaver Creek, and Middle Fork Beaver Creek glaciers (Fig. 14). To the south, the prominent strike-ridge of the Mississippian Madison Limestone is visible (Fig. 16), forming "Hole in the Rock" at the left. The central gap in the limestone was formed by meltwater from the Middle Fork Beaver Creek; Smiths Fork–age moraines in this valley are hidden behind the limestone ridge (Fig. 17). The most westerly (and widest) gap in the limestone is partially blocked by end moraines of the West Fork Beaver Creek (Fig. 17). Bryant (1992) mapped deposits of the Smiths Fork, Blacks Fork, and pre–Blacks Fork glaciations in this moraine complex. More recently, Douglass (2000) examined the morphology of these moraines and their weathering characteristics. His results support the mapping of Bryant, and further suggest that the Smiths Fork–age moraine is a compound feature, possibly representing ice advances during MIS 2 and 4.

To the east-southeast of Lonetree lies the Bald Range, a complex upland landscape of random hills and local internal drainage (Fig. 14). Bradley (1936) interpreted these sediments as deposits of pre–Blacks Fork glaciers that advanced from the northern Uintas presumably before the modern Beaver Creek–Henrys Fork valley was developed. However, poorly consolidated sediment of the Tertiary Wasatch Formation is exposed along the west side of the Bald Range, and diagnostic evidence of a glacial origin of the capping sediments is lacking. Gibbons and Hansen (1980) argued for a reinterpretation of the Bald Range as landslide deposits, and Gibbons (1986) mapped the entire Bald Range as landslide deposits on the "Surficial materials map of the Evanston 30′ × 60′ quadrangle." The exact manner in which mass wasting produced the modern form of the Bald Range is unclear. However, fluvial erosion on three sides (Beaver Creek–Henrys Fork on the west, Henrys Fork on the north, and Burnt Fork on the west) clearly isolated this pile of weak sediments, which may have shifted and failed in response to a seismic trigger.

The gravel pit at this site also provides an opportunity to review the alluvial stratigraphy below the glacial termini of the Henrys Fork drainage, which has been mapped at 1:24,000 scale in the development of a nonglacial Quaternary stratigraphic

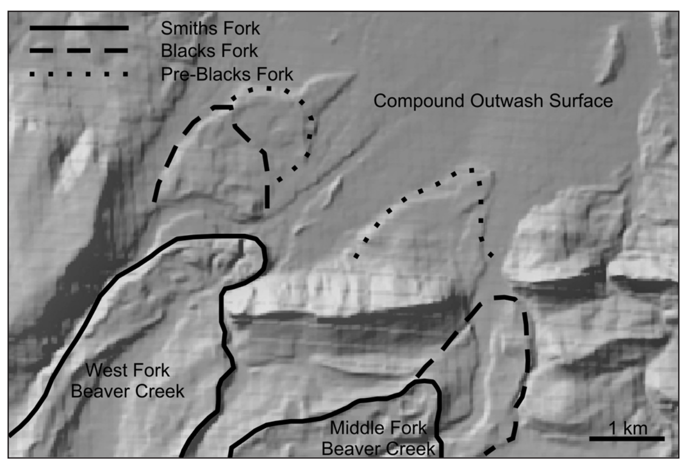

Figure 17. Moraines of the Middle Fork and West Fork Beaver Creek and their compound outwash surface south of Lonetree, Wyoming (see Fig. 14 for location). Prominent east-west ridge is a hogback of the Mississippian Madison Limestone. The Middle Fork Beaver Creek glacier terminated upslope from the limestone during the Smiths Fork Glaciation, while the West Fork glacier advanced slightly beyond the hogback. Moraines corresponding to the Blacks Fork Glaciation, and till of pre–Blacks Fork age, are present farther north. The compound outwash surface extends over 10 km northward to merge with outwash deposits in the Henrys Fork river valley.

framework for the northeastern Uinta Mountains (Counts, 2005). The Henrys Fork valley contains nine distinct mainstem gravels, six piedmont gravels, and landslide deposits (Fig. 18). Gravels on the Henrys Fork are grouped as older, higher remnant gravels that cannot be directly linked to glacial units and as younger gravels that can be traced from glacial moraines, through outwash plains, to stream-valley gravels with terraces formed upon them.

This example of Qag3 (Blacks Fork equivalent) gravels are representative of the clast-supported, cobble gravel of the Henrys Fork, derived mostly from the UMG and Paleozoic limestone units. Near moraines, gravels are thicker but they quickly thin downstream and lie on planar bedrock straths, and so form strath terraces that converge downstream. Besides Qag1, which is correlated to the dated Smiths Fork glaciation, Henrys Fork terraces Qag2 and Qag3 are tentatively correlated to relatively well dated Wind River terraces (Counts, 2005). Incision rate estimates for the Henrys Fork from this correlation are 80–110 m/my over the late Pleistocene (Table 1). Extrapolating a linear incision rate (maximum long-term estimate) suggests that the oldest gravels on the Henrys Fork were deposited in the early Pleistocene.

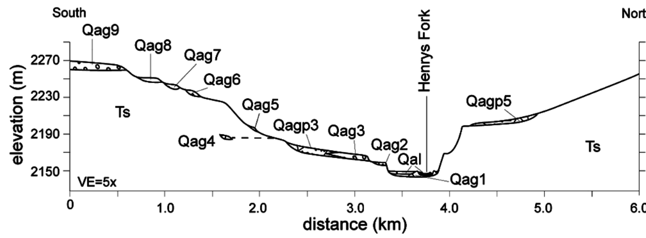

Figure 18. Composite cross-valley profile showing landscape position of major Henrys Fork terrace gravels (modified from Counts, 2005). Ts—Tertiary basin-fill sediment. Qag4 gravel was not identified along the Henrys Fork, but it exists along the tributary Beaver Creek and Burnt Fork. Qag—Quaternary alluvial gravel of mainstem drainages; Qagp—Quaternary alluvial gravel of piedmont drainages; Qal—alluvium of active channels and floodplains.

Directions to Stop 3.4

Retrace the route back to the last intersection (mile 61.1) and turn left on CR 263. Bear right at mile 61.5, following signs for the Henrys Fork. Cross back into Utah at mile 67.2, and pass the Forest Boundary at mile 67.9, with the Smiths Fork–age terminal moraine visible straight ahead. The road enters the moraine near mile 68.6 and passes through an area of high-relief hummocky topography before descending the proximal slope of the moraine at mile 69.3. Cross the Henrys Fork at mile 69.7, with a view of Gilbert Peak upstream, and turn right into the Quarter Corner Trailhead at mile 71.3 for lunch (bathroom, picnic tables). After lunch, return to the main road and continue to the right, taking a hard right at the intersection with FR 017 at mile 71.8. The road climbs up and over the west side of the terminal moraine and crosses Dahlgreen Meadow (mile 74.0) before climbing a series of switchbacks up the south face of Red Mountain. Turn right on FR 155 at mile 75.1 and park at the entrance to FR 132 on the right (mile 75.3). From the vans, we will hike for ~15 min up FR 132 to the summit of Red Mountain.

Stop 3.4—Red Mountain (Mile 75.3)

Red Mountain provides a spectacular view of the central High Uintas (Fig. 19). Several of the most prominent peaks visible from this point are named for famous geologists, including Kings Peak (4136 m—due south), Gilbert Peak (4110 m—SSE), and Mount Powell (4023 m—SSW). The broad alpine basin west of Gilbert Peak formed the accumulation area for a glacier in the Henrys Fork that covered 76 km^2 at the LGM. The glacier, which was over 24 km long, advanced beyond the hogback of Madison Limestone at The Narrows (visible 5 km upvalley) and formed a hummocky moraine east of Red Mountain (Figs. 14, 19, and 20). Because the cross-sectional area of the valley within The Narrows is 25% less than that immediately upstream, and because the Henrys Fork glacier overtopped The Narrows by only 20–50 m, relatively modest downwasting during deglaciation would have separated the ice in the terminal zone from the active ice in the valley to the south. This geometry likely caused the terminal zone of the Henrys Fork

Figure 19. View southward from Red Mountain. Several major peaks of the High Uintas are visible in the distance. The Narrows is a prominent valley constriction formed where the Henrys Fork passes through the Madison Limestone hogback. The Smiths Fork–age terminal moraine of the Henrys Fork glacier is visible as a forested ridge in the foreground. Dahlgreen Meadow marks the site of a former ice-marginal lake, formed when the Henrys Fork glacier impinged upon the base of Red Mountain, damming an ice-marginal drainage.

glacier to stagnate at the start of deglaciation, even when glaciers in neighboring valleys were able to keep pace with ameliorating conditions by actively retreating upvalley.

Interestingly, Atwood (1909) mapped this terminal moraine as a Blacks Fork equivalent and located the Smiths Fork–age terminal moraine near Alligator Lake, 6 km upvalley from The Narrows. However, the overall fresh form of this massive moraine loop, its steep frontal slope, and its dramatically kettled surface argue that it was formed during the Smiths Fork glaciation. Furthermore, outwash graded to this moraine merges with Smiths Fork–age outwash from the West and Middle forks of Beaver Creek. Thus, the moraine near Alligator Lake is most likely a rock-cored recessional feature formed when the retreating Smiths Fork–age glacier was temporarily pinned on a bedrock high.

The advance and retreat of the Henrys Fork glacier also impacted the fluvial system through formation of ice-marginal channels and ice-dammed lakes. Ice-marginal channels containing underfit streams are common components of the north slope landscape. Dahlgreen Creek, which runs alongside the western lateral moraine of the Henrys Fork below Red Mountain, is a good example (Fig. 19). In some cases, lateral moraines redirected ice-marginal drainages into other watersheds, for instance upstream from The Narrows on the east side of the Henrys Fork where the voluminous right lateral moraine contained a stagnant ice body that drained eastward through Bullocks Park into the Beaver Creek system along the route followed by FR 058. Meltwater derived from this stagnant ice mass appears to have helped carve a short canyon along Fallon Creek (Hole in the Rock quadrangle). Ice-marginal channels were also vulnerable to impoundment during periods of ice advance, leading to the formation of temporary ice-marginal lakes. The willow-filled meadow just south of Red Mountain was likely the bed of an ice-marginal lake during times when the Henrys Fork glacier blocked the northward drainage of Dahlgreen Creek (Fig. 19).

Finally, the summit of Red Mountain was mapped by Bryant (1992) as the Tertiary Bishop Conglomerate (see Stop 2.1). Red Mountain and the surface extending southward to increasingly higher elevations along the west side of the Henrys Fork are remnants of Bradley's (1936) Gilbert Peak erosion surface mantled by Bishop Conglomerate. Although the exact mechanism by which this extensive gently sloping landform was generated is unclear, Red Mountain appears to have been isolated by slumping along its southern and eastern faces, perhaps in response to slope steepening due to glacial erosion by the Henrys Fork glacier.

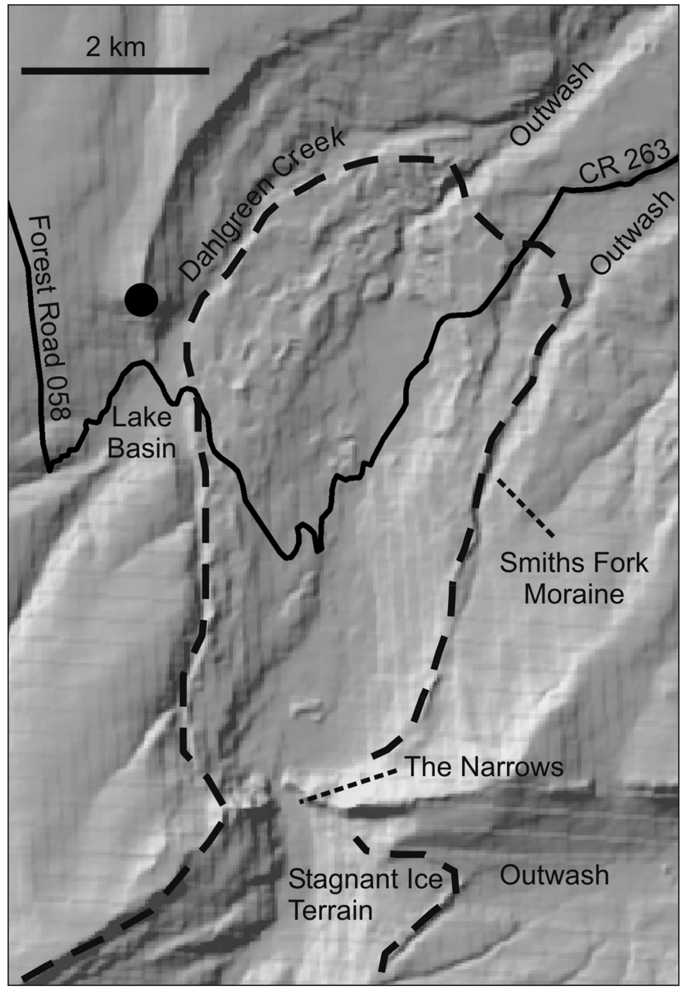

Figure 20. Location of the Smiths Fork–age moraine (dashed) in the Henrys Fork (see Fig. 14). The Smiths Fork–age glacier in the Henrys Fork advanced northward beyond the prominent valley constriction of The Narrows, formed by the Mississippian Madison limestone. The piedmont lobe formed by the glacier stagnated during deglaciation when the glacier surface lowered, dramatically reducing the cross-sectional area feeding the terminal zone. A smaller area of stagnant ice developed on the southeastern side of The Narrows; this zone drained eastward, depositing outwash on the south side of the limestone. Dahlgreen Creek (an ice-marginal drainage) was temporarily impounded by the glacier terminus, forming a lake in the basin south of Red Mountain (black dot). The route of the field trip southwest from Lonetree, Wyoming, on County Road 263, and westward on Forest Road 058, is also shown (black line).

Directions to Stop 3.5

After walking back to the vans, turn around and head south to the intersection with FR 017 (mile 75.6). Turn right to continue west into the East Fork Smiths Fork drainage. At the next intersection (mile 79.4), turn right toward Mountain View, Wyoming. The slope directly across the road from the stop sign is the distal side of the right lateral moraine in the East Fork Smiths Fork valley. Cross the Wasatch-Cache National Forest boundary at mile 83.1, and at mile 84.3, turn left on CR 285. Cross back into the forest at mile 86.0, with the terminal moraine of the East Fork Smiths Fork visible to the left of center. Outwash graded to this moraine is visible along the left side of the road at mile 86.8. Cross the river and park on the right at mile 86.9.

Stop 3.5—Smiths Fork Type Locality (Mile 86.9)

The terminal moraine considered by Bradley to be the Smiths Fork glaciation type locality is present just south of the road at the bridge, on either side of the stream. The western half of the terminal moraine is vegetated by lodgepole pine and stands in stark contrast to the sagebrush-covered outwash surface. The eastern half of the moraine (at least the part visible from this vantage) forms an unforested slope rising steeply above the river. The crests of both moraines are <1 km wide and marked by slightly sinuous subparallel ridges standing <10 m above their surroundings. Large erratic boulders of quartzite are rare, but several were located and sampled for cosmogenic surface exposure dating during the 2003 and 2004 field seasons. Results of these analyses, which should help constrain the timing of the local LGM in the northern Uintas, are pending.

Directions to Salt Lake City

Turn around and head back toward Mountain View, Wyoming, turning left at the stop sign (mile 89.5). Continue straight at the intersection with the paved road (mile 96.7) and turn left at the junction with Hwy 414 in Mountain View (mile 103). Pass through Mountain View and continue straight at the intersection in Urie (mile 106.8). Cross the Blacks Fork at mile 109.1 and enter I-80 westbound at mile 109.7. Follow I-80 west to Evanston (mile 144) and on to Salt Lake City (~mile 224).

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