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Geology and geomorphology of Bear Lake Valley and upper Bear River, Utah and Idaho

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ABSTRACT

Bear Lake, on the Idaho-Utah border, lies in a fault-bounded valley through which the Bear River flows en route to the Great Salt Lake. Surficial deposits in the Bear Lake drainage basin provide a geologic context for interpretation of cores from Bear Lake deposits. In addition to groundwater discharge, Bear Lake received water and sediment from its own small drainage basin and sometimes from the Bear River and its glaciated headwaters. The lake basin interacts with the river in complex ways that are modulated by climatically induced lake-level changes, by the distribution of active Quaternary faults, and by the migration of the river across its fluvial fan north of the present lake.

The upper Bear River flows northward for ~150 km from its headwaters in the northwestern Uinta Mountains, generally following the strike of regional Laramide and late Cenozoic structures. These structures likely also control the flow paths of groundwater that feeds Bear Lake, and groundwater-fed streams are the largest source of water when the lake is isolated from the Bear River. The present configuration of the Bear River with respect to Bear Lake Valley may not have been established until the late Pliocene. The absence of Uinta Range–derived quartzites in fluvial gravel on the crest of the Bear Lake Plateau east of Bear Lake suggests that the present headwaters were not part of the drainage basin in the late Tertiary. Newly mapped glacial deposits in the Bear River Range west of Bear Lake indicate several advances of valley glaciers that were probably coeval with glaciations in the Uinta Mountains. Much of the meltwater from these glaciers may have reached Bear Lake via groundwater pathways through infiltration in the karst terrain of the Bear River Range.

At times during the Pleistocene, the Bear River flowed into Bear Lake and water level rose to the valley threshold at Nounan narrows. This threshold has been modified by aggradation, downcutting, and tectonics. Maximum lake levels have decreased

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from as high as 1830 m to 1806 m above sea level since the early Pleistocene due to episodic downcutting by the Bear River. The oldest exposed lacustrine sediments in Bear Lake Valley are probably of Pliocene age. Several high-lake phases during the early and middle Pleistocene were separated by episodes of fluvial incision. Threshold incision was not constant, however, because lake highstands of as much as 8 m above bedrock threshold level resulted from aggradation and possibly landsliding at least twice during the late-middle and late Pleistocene. Abandoned stream channels within the low-lying, fault-bounded region between Bear Lake and the modern Bear River show that Bear River progressively shifted northward during the Holocene. Several factors including faulting, location of the fluvial fan, and channel migration across the fluvial fan probably interacted to produce these changes in channel position.

Late Quaternary slip rates on the east Bear Lake fault zone are estimated by using the water-level history of Bear Lake, assuming little or no displacement on dated deposits on the west side of the valley. Uplifted lacustrine deposits representing Pliocene to middle Pleistocene highstands of Bear Lake on the footwall block of the east Bear Lake fault zone provide dramatic evidence of long-term slip. Slip rates during the late Pleistocene increased from north to south along the east Bear Lake fault zone, consistent with the tectonic geomorphology. In addition, slip rates on the southern section of the fault zone have apparently decreased over the past 50 k.y.

INTRODUCTION

Sediments deposited in Bear Lake, on the Idaho-Utah border (Figs. 1 and 2), provide a record of changing inputs of water and sediment derived from voluminous calcium-bicarbonate-charged groundwater discharge and from the Bear River. In addition to these variable water and sediment sources, the river has interacted with the lake basin in complex ways that are modulated by lake-level fluctuations, by the distribution of active Quaternary faults, and by migration of the river across its fluvial fan north of the present lake. An understanding of the timing and magnitude of all these changes is required to unravel the environmental record preserved in the lake sediment.

This chapter uses published literature to review the bedrock and pre-Quaternary tectonic setting of Bear Lake Valley and the upper Bear River in the context of their influence on sediment supply and groundwater discharge to the lake. We then discuss and summarize recently published and new information on the distribution and relative age of Quaternary glacial and fluvial deposits along the Bear River upstream of Bear Lake and the implications for the timing of sediment transport by Bear River. New mapping of moraines near the crest of the Bear River Range (west of Bear Lake; Fig. 2) improves the understanding of the former extent of glaciers and their potential influence on the lake.

At intervals when the Bear River flowed directly into Bear Lake, the lake would tend to overtop its sill to the north en route to Lake Bonneville (Fig. 1). Without the input of the Bear River, the lake would fluctuate within a topographically (but not hydrologically) closed basin. This relationship is controlled by several factors, including (1) climatic changes that cause the lake to transgress and intersect the Bear River, or to retreat into a closed basin; (2) active tectonics, especially along the east Bear Lake fault zone (Fig. 2); (3) migration of Bear River across its fluvial fan where it debouches into the valley; and (4) aggradation and incision that affect the altitude of the basin's threshold. The relationship between the Bear River and Bear Lake during the late Quaternary is documented by preserved shoreline deposits at various altitudes (dated by radiocarbon, amino acid racemization, and tephrochronologic techniques) and by river terraces. We present new information on the progressive migration of Bear River northward away from Bear Lake in the past 8000 years based on studies of sediments preserved in abandoned channels. Finally, we discuss new information on displacements and slip rates on valley-bounding normal faults based on the dated lacustrine and marsh deposits, tectonic geomorphology, and seismic data of Colman (2006).

TECTONIC SETTING AND BEDROCK GEOLOGY

The upper Bear River flows northward for ~150 km from its headwaters in glaciated valleys on the north flank of the Uinta Mountains (Fig. 1), generally following the strike of regional structures in the Laramide overthrust belt and on the northeastern margin of the Basin and Range province. Along this margin is a northeasterly trending zone of right-stepping normal faults controlling fault-block mountains and basins that developed in response to east-west Neogene extension superimposed on Cretaceous to early Tertiary folds and thrust faults (Armstrong, 1968; McCalpin, 1993) south of the Snake River Plain. The normal faults developed along the same strikes as the older thrust faults, and in some cases coincident with them. The northeasterly extensional pattern has been related to the probable influence of the Yellowstone hotspot (Anders et al., 1989; Pierce and Morgan, 1992; Parsons et al., 1994). Pierce and Morgan (1992) subdivided a parabola-shaped region of extension around Yellowstone and the Snake River Plain into "seismic belts" by using the frequency

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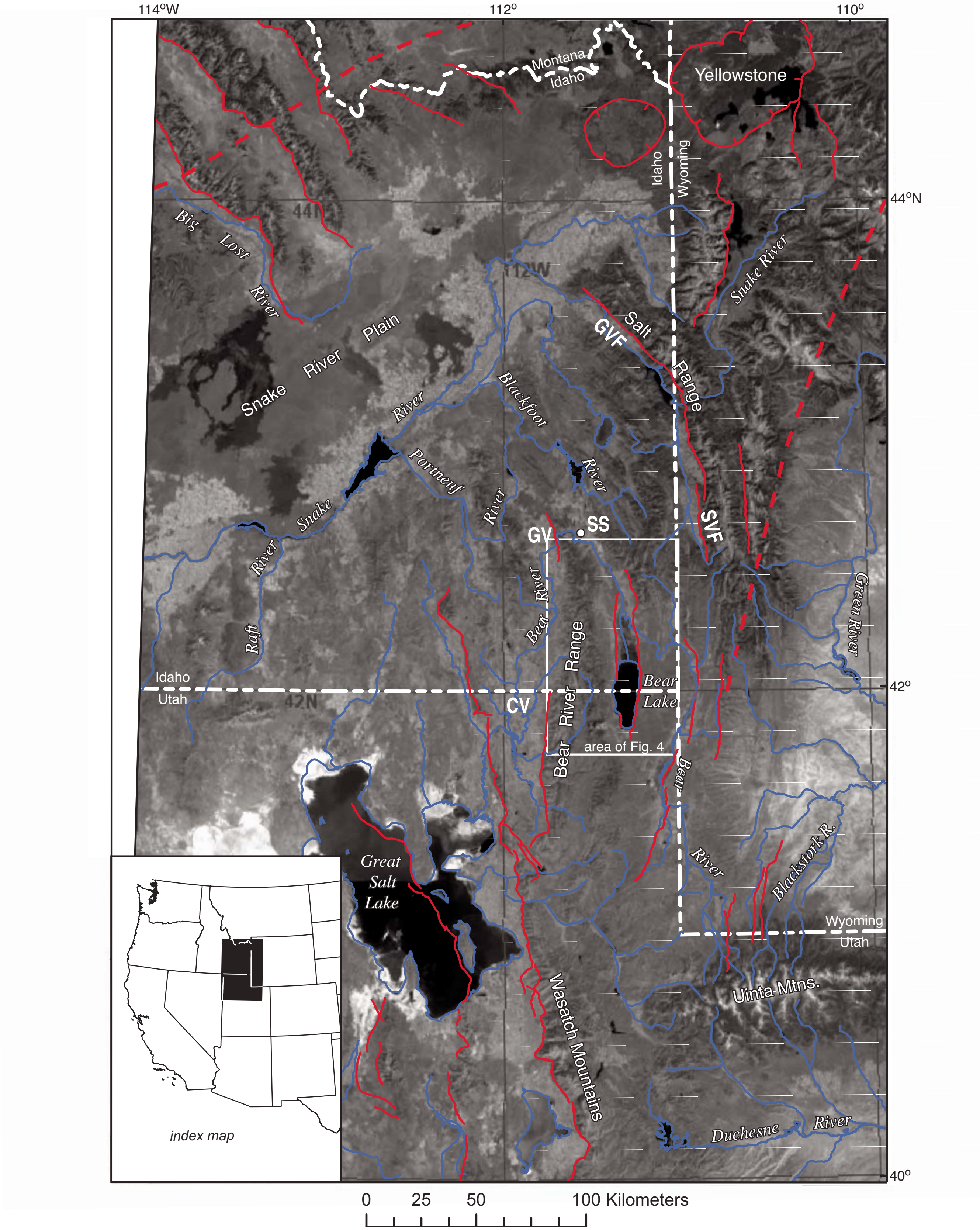

Figure 1. Location index map (inset) and regional setting, including drainages (blue) and principal Quaternary faults (red; modified from U.S. Geological Survey, 2004). Red hachured lines are boundaries of Quaternary calderas. Red dashed line is axis of high elevations within tectonic "parabola" of Pierce and Morgan (1992). CV—Cache Valley; GV—Gem Valley; GVF—Grand Valley fault; SS—Soda Springs; SVF—Star Valley fault. Box shows area of Figure 4.

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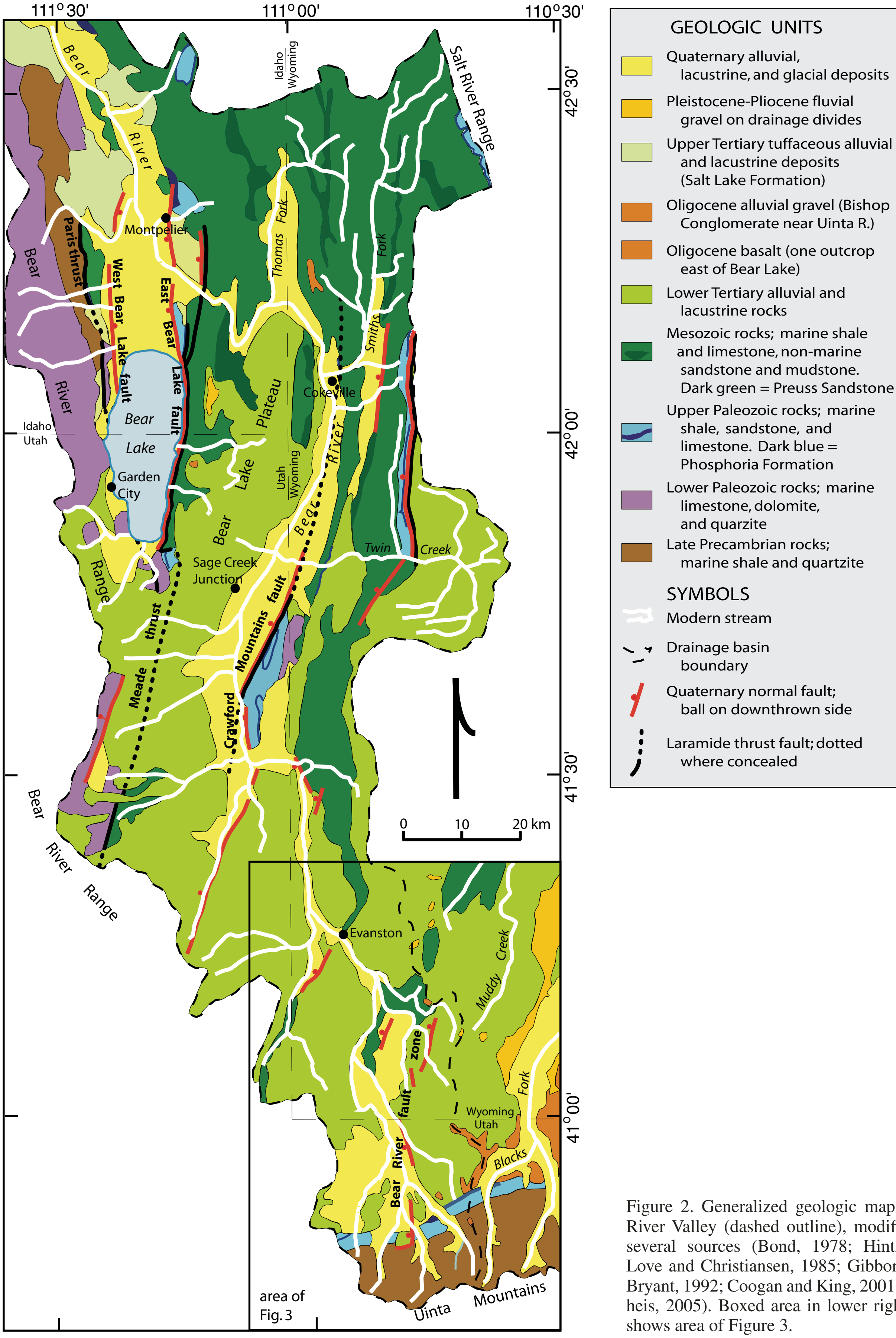

Figure 2. Generalized geologic map of Bear River Valley (dashed outline), modified from several sources (Bond, 1978; Hintz, 1980; Love and Christiansen, 1985; Gibbons, 1986; Bryant, 1992; Coogan and King, 2001; and Reheis, 2005). Boxed area in lower right corner shows area of Figure 3.

of recent seismic activity and the height of normal-fault scarps in each valley.

Bear Lake Valley lies within the most active seismic zone on the northeast-southwest–trending southern arm of the parabola (Fig. 1). The valley forms a complex graben between two normal faults, informally named the east and west Bear Lake fault zones (Fig. 2; McCalpin, 1993, 2003). Several reaches of the Bear River and its tributaries coincide with Pleistocene normal faults (Fig. 2). Northeast of Bear Lake, the river bends sharply west and crosses folds and faults of the Bear Lake Plateau and the Preuss Range at a nearly right angle to emerge into Bear Lake Valley, where it resumes a northward course parallel to structural trends. South of latitude 42°N, the upper Bear River lies within a less conspicuous zone of faulting that includes faults with Holocene and late Pleistocene displacement but mostly lacking high, steep range fronts (Pierce and Morgan, 1992).

The Bear River drains a wide variety of rock types and ages (Fig. 2). Its headwaters lie in Precambrian orthoquartzite, siliceous sandstone, siltstone, and shale in the core of the Uinta Range (Bryant, 1992); detritus derived from these rocks by glacial erosion and rock flour production contributed much of the sediment load of the river during glacial periods (Rosenbaum et al., this volume; Rosenbaum and Heil, this volume). Downstream, the majority of exposed bedrock consists of Paleocene–Eocene conglomerate, sandstone, and shale (Bond, 1978; Hintz, 1980; Love and Christiansen, 1985). These rocks are poorly indurated and are probably easily erodible, especially during times when vegetation cover is reduced. The Tertiary sediments overlie complexly faulted and folded Paleozoic and Mesozoic sedimentary rocks, mainly exposed on the east and north sides of the river upstream of Bear Lake. The lower Paleozoic sequence is dominated by limestone and dolomite, and the upper Paleozoic rocks include marine sandstone and phosphate-bearing rocks. In contrast, the Mesozoic rocks in the drainage basin mainly consist of continental sandstones and shales.

The local basin of Bear Lake drains a much more restricted suite of rock types and ages (Fig. 2). Although upper Paleozoic and Mesozoic rocks crop out along the east side of the lake, there is little surface runoff from the small creeks that drain the flank of the Bear Lake Plateau. The majority of streams entering the basin drain extensive areas of lower Paleozoic limestone, dolomite, and quartzite west of the lake. Bear Lake Valley also contains locally thick deposits of upper Tertiary sediment (Salt Lake Formation) deposited during the early phases of extensional faulting, suggesting persistence of the valley as a fault-controlled basin probably since the Miocene. The most recent mapping restricts the Salt Lake Formation mainly to the area north of the lake (Dover, 1995), in contrast to earlier maps that showed this unit throughout the valley (Oriel and Platt, 1980). In general, the pre-Tertiary rocks are much more resistant to erosion than the Tertiary rocks and may contribute less detritus to the Bear River. However, the older rocks also contain readily soluble carbonate and evaporite units that likely contribute to the solute load of surface streams and groundwater. Closed depressions, sinkholes, and collapse basins in the highlands of the Bear River Range west of Bear Lake (Dover, 1995) attest to wholesale subsurface solution of Paleozoic carbonate rocks, probably facilitated by the dense network of generally north-striking thrust and normal faults.

Structural Influences on Groundwater Flow

During much of the Holocene, Bear Lake received no surface water from Bear River (Dean, this volume; Rosenbaum et al., this volume), and the input was dominated by spring-fed streams originating in the Bear River Range west of the lake (Bright, this volume). The structural setting of Bear Lake and the surrounding highlands likely controls the flow paths of groundwater to these and other springs, but not in simple or obvious ways. North-striking, west-dipping splays of the Paris and Meade thrust faults crop out on the west and east sides of Bear Lake Valley, respectively (Fig. 2), and basin-and-range-style normal faults are thought to sole into the thrust faults at depth (Coogan, 1992; Dover, 1995; Evans et al., 2003). Farther east, the west-dipping Crawford fault, a Mesozoic thrust fault reactivated as a normal fault, similarly bounds the east side of the Bear River and normal faults complementary to the Crawford fault bound the west side of the river (Bear River graben of Dover, 1995). Thus, east of Bear Lake the structural configuration should inhibit Bear River–derived groundwater from reaching the lake, except by upward infiltration along normal faults that intersect the thrust faults at depth. Discharge and solute data indicate that springs on the east side of the lake contribute only minor amounts of water (Bright, this volume; Dean et al., 2007). Flows from springs on the west side are much larger but extrabasinal sources may also contribute to the lake (Bright, this volume).

One likely source of extrabasinal water is groundwater derived from recharge from the topographically high, southern part of the Bear River Range (Fig. 2) along the western boundary of the Bear River drainage basin. Paleozoic carbonate rocks in this divide area are overlain to the north by the Wasatch Formation and are displaced by normal faults with Tertiary to Quaternary displacement that sole into the Meade thrust fault. These normal faults are part of a continuous zone of anastomosing faults that extends directly to the south end of Bear Lake Valley and merges with faults west of Bear Lake. Groundwater that originates by infiltration through the carbonate rocks could be confined by the overlying Wasatch beds, permitting the normal faults to act as conduits for groundwater that originates outside the Bear Lake drainage basin. Because the rocks in this recharge area are carbonates, isotopic compositions of groundwater from this source would probably resemble those of groundwater from the Bear River Range within the Bear Lake drainage basin.

Faults clearly play important roles in controlling groundwater discharge on a local scale. Along the west side of Bear Lake, large and small springs are located on north-striking Tertiary normal faults (Oriel and Platt, 1980; Dover, 1995); in some cases these faults also offset Quaternary deposits (Reheis, 2005). Higher in the Bear River Range, parallel north-striking faults cut gently west-dipping, lower Paleozoic carbonate rocks (Oriel and Platt, 1980; Dover, 1995), and these faults likely provide conduits for snowpack recharge into carbonate-rock aquifers, as suggested by numerous closed depressions and sinkholes (Reheis, 2005). Dye tests indicate that groundwater flow paths in the Bear River Range cross topographic divides (Spangler, 2001). Stable isotope data on spring waters (Bright, this volume) support such a groundwater source for west-side springs. Along the east side of Bear Lake, isotopic data (Bright, this volume) indicate a sharp change in the relatively small-volume groundwater sources across the east Bear Lake fault zone, which soles into the Meade thrust (Dover, 1995; Evans et al., 2003). West of the Meade thrust, near-vertically dipping upper Paleozoic and lower Mesozoic sedimentary rocks, including the phosphate-rich Phosphoria Formation, are thought to underlie the east side of Bear Lake. Solutes derived from these formations may cause the sharp change in groundwater composition. A similar abrupt change in isotopic composition of surface water (Bright, this volume) occurs along the Bear River just upstream of Cokeville (Fig. 2). This change approximately coincides with the point where the river crosses the subsurface trace of the Crawford thrust (Rubey et al., 1980; M'Gonigle and Dover, 1992).

THE UPPER BEAR RIVER DRAINAGE BASIN

Tertiary and Quaternary deposits in the upper Bear River reveal a long history of drainage-basin evolution, glaciation, and landsliding in high basin-bounding ranges, and incision and deposition of fluvial and alluvial-fan deposits. Profound changes in drainage boundaries and directions occurred in this region around the Utah-Idaho-Wyoming borders during the late Cenozoic. These changes were driven by base-level fall due to the encroachment of basin-and-range extensional faulting from the west, and by the topographic changes caused by the progressive northeastward motion of the Yellowstone hotspot and the development of the eastern Snake River Plain (Fig. 1; Pierce and Morgan, 1992).

The oldest fluvial geomorphic features preserved in the study area are the remains of coalescing piedmont gravels composed of the Oligocene Bishop Conglomerate (Bradley, 1936) that extend northward from the Uinta Mountains into the Green River Basin. These gravels presently crop out on drainage divides in the headwaters of Bear River (Fig. 2) and are much more extensively preserved to the east (Bryant, 1992). In Oligocene time, much of the study area to the north of these outcrops was probably a gentle alluvial plain that capped the sedimentary fill of the older Wasatch and Green River Formations.

With the onset of extensional faulting and the development of large calderas in the central Snake River Plain (Fig. 1) during the Miocene, deposition of the Salt Lake Formation began in valleys along the southern margin of the Snake River Plain in southeastern Idaho, including what is now northern Bear Lake Valley (Fig. 2). On the basis of fossil snail faunas, Taylor and Bright (1987) suggested that the western part of the area had affinities to western drainages, whereas the eastern part of the area had affinities to the south; if so, the northern Bear Lake Valley was still a drainage divide in the early(?) Miocene. Previous workers (Mansfield, 1927; Williams et al., 1962; Taylor and Bright, 1987) suggested that the Bear River attained its present course, particularly its westward jog across structure upstream of Bear Lake, in Pliocene time when the valleys were presumably filled with sediments of the Salt Lake Formation. More recent geologic mapping (Dover, 1995) has reinterpreted the sediments capping the Bear Lake Plateau as the Eocene Wasatch Formation rather than the younger Salt Lake Formation as mapped by Oriel and Platt (1980), and suggested that Salt Lake Formation sediments are thin or absent within the Bear Lake Valley south of the present Bear River. If this interpretation is correct, the Bear River could not have attained its present course into Bear Lake Valley by superposition across structures buried by Pliocene sediment.

The oldest deposits of what may have been a through-flowing Bear River are preserved on the very crest of the Bear Lake Plateau (Fig. 2), as much as 360 m above the present Bear River. These remnants, discovered during the present study (Reheis, 2005), are inset below the level of an Oligocene basalt and consist of cross-bedded fluvial gravel and sand as much as 5 m thick. The dominance of well-rounded pebble- to small-cobble-sized clasts and the lateral extent of outcrop indicate that the deposits represent a river similar to Bear River in size. The deposits slope gently northward along the drainage divide for at least 8 km and also slope to the east, toward the present course of Bear River. In one locality, nested channel fills in a roadcut show that the river migrated eastward and incised successively lower channels (Reheis, 2005). The most likely candidate for this river is the ancestral Bear River; however, the distinctive pink and purple quartzites derived from the Precambrian rocks of the Uinta Range, characteristic of the Pleistocene gravels of Bear River, are absent from these gravels.

There are three possibilities to explain the absence of the Precambrian quartzite clasts. First, the present-day northward slope of the gravel remnants may not represent the gradient of the ancient river, but may be an artifact of northward tilting of the footwall block along the east Bear Lake fault zone. However, the base of the deposits descends nearly 60 m over 8 km. Tilting along the fault would have had to exceed that amount to reverse the apparent flow direction of the paleo-river. The overall topography of the Bear Lake Plateau does not support northward tilting; the highest summits along the plateau are ~2225 m in the south, 2350 m atop the Oligocene basalt remnant (Fig. 2), and 2350 m northeast of the plateau gravels. Previous workers have inferred that *southward* tilting may have occurred along the hanging-wall block (Laabs and Kaufman, 2003; McCalpin, 2003). Second, the ancestral Bear River may not have originated in its present-day headwaters in the Uinta Range. Hansen (1985) suggested that at some time in the past, these headwaters flowed into the Green River via Muddy Creek (Fig. 2), and that the Bear River subsequently captured its present headwaters. Hansen (1985) observed drainages east of Hilliard Flat (Fig. 3) that presently drain to Bear

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River, but that in their upper reaches point toward upper Muddy Creek in the area of a topographic gap in the divide. In addition, later surficial geologic maps (Gibbons, 1986; Dover and M'Gonigle, 1993) show old terrace gravels (younger and lower than the Bishop Conglomerate) that lie on and near the present drainage divide (Figs. 2 and 3). The proposed capture area lies within the Bear River fault zone (West, 1993); displacement on these down-to-the-west faults and on other Pleistocene normal faults to the east along Muddy Creek could have been the proximate cause of the capture of the Uinta headwaters by an ancestral Bear River (West, 1993) that did not extend south of about Evanston prior to the capture. Third, the gravels predate the unroofing of the distinctive Precambrian quartzites in the Uinta Range. We reject this hypothesis because the gravels are younger than the Oligocene basalt, and the Oligocene Bishop Conglomerate contains such quartzites (Dover and M'Gonigle, 1993); thus, younger gravels from the same source area should also contain them.

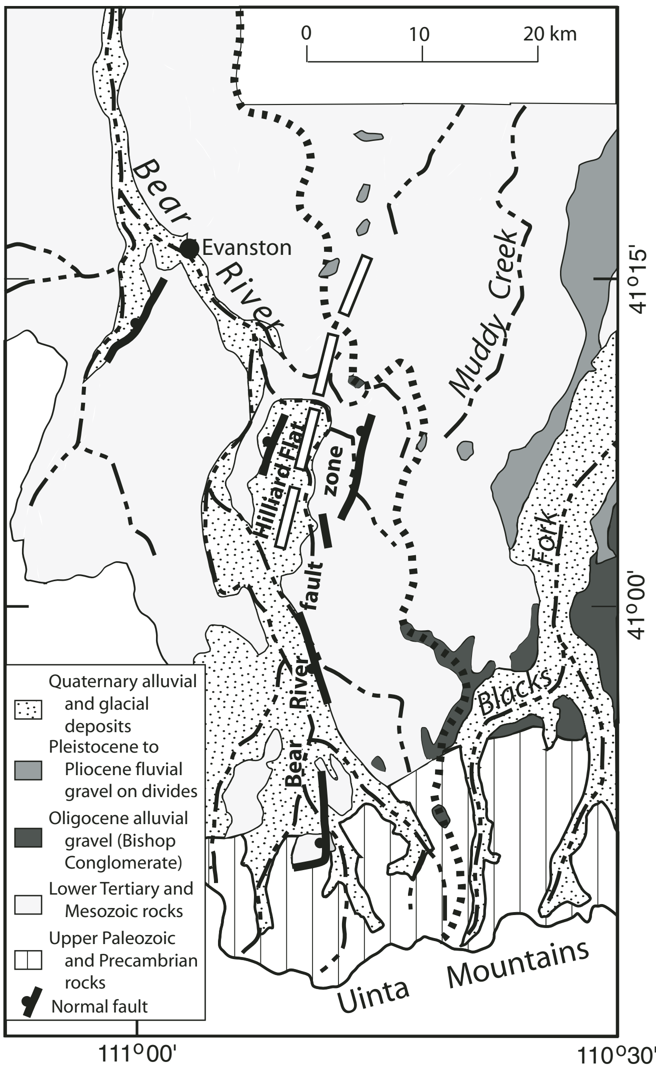

Figure 3. Generalized geologic map of capture area near Hilliard Flat showing postulated former course (white blocks) of headwaters of Bear River into Muddy Creek drainage, tributary to Green River (Hansen, 1985). Note black areas showing remnants of Bishop Conglomerate (Oligocene) and dark gray areas of younger Pliocene-Pleistocene terrace gravels capping divides, including Bear River–Muddy Creek divide (heavy dotted line) northeast of Hilliard Flat.

Addition of water from the Uinta Range would have added considerable volume to the ancestral Bear River. Further, the eastward migration and downcutting of the ancestral river atop the Bear Lake Plateau suggest that the east Bear Lake fault zone had become active, creating the depression to the west that is now Bear Lake Valley and lifting the area of the plateau along the footwall block. Thus, the Bear River may have gained a significant amount of discharge at about the same time that accommodation space was being created in Bear Lake Valley along with the potential for river diversion. We hypothesize that the subsequent diversion of the Bear River into Bear Lake Valley and the beginning of lacustrine and fine-grained basinal deposition may have been concurrent events.

The late Pliocene to early Pleistocene course of the Bear River downstream of Bear Lake Valley is speculative. Terrace deposits downstream of Bear Lake so far have been found no higher than ~40 m above river level (several roadcut exposures south of Soda Springs along U.S. Highway 30 in the northwestern part of the Fossil Valley, Idaho 7.5′quadrangle); thus, they are likely much younger than the gravels atop the Bear Lake Plateau. Previous workers have speculated that Bear River once flowed northwest to the Snake River Plain via the present-day Portneuf River Gorge (Fig. 1; Bright, 1963; Mabey, 1971) and was diverted southward by eruption of lava flows in Gem Valley west of Soda Springs, Idaho (Bouchard et al., 1998). However, given the altitude of a few Bear River terrace remnants south of Soda Springs and the estimated thickness of basalt in the Blackfoot River Canyon downstream of Blackfoot Reservoir (Mabey, 1971), it seems equally likely that the Bear River could have originally continued directly north to the Snake River via the Blackfoot River and was diverted westward by voluminous Pleistocene eruptions of the Blackfoot volcanic field north and east of Soda Springs. A few radiogenic ages on volcanic rocks in these two fields range from ca. 1.0 to 0.05 Ma (Armstrong et al., 1975; Heumann, 2004; Pickett, 2004; Scott et al., 1982). Comprehensive dating and mapping of terrace gravels and basalt flows would be required to investigate these alternative ancestral courses.

The north flank of the Uinta Mountains was extensively and repeatedly glaciated during the Pleistocene (Bryant, 1992; Munroe, 2001) and minor glacial advances may have also occurred during the Holocene (Munroe, 2000). Deposits of at least two major glaciations have been mapped in the Bear River headwaters. Outlet glaciers of a broad ice field in the western Uinta Mountains (termed the Western Uinta Ice Field by Refsnider et al., 2007) occupied valleys of the East, Hayden, and West Forks of Bear River at the maximum of the Smiths Fork glaciation (equivalent to the late Pleistocene Pinedale glaciation of the Rocky Mountains) ca. 19–18 ka. (Laabs et al., 2007). Glaciers of the Hayden and East Fork valleys coalesced on the piedmont beyond the mouths of tributary canyons to deposit a broad area of hummocky topography. Cosmogenic ^{10}Be surface-exposure dating indicates that this area was abandoned at the start of ice retreat ca. 18 ka (Laabs et al., 2007), although ice retreat may have started as much as 2 k.y. later in other parts of the Uinta Range (Munroe et al., 2006). These ages compare favorably with the age for the maximum extent of glaciation, 19.7–18.9 ka as inferred from rock flour abundance in sediment cores from Bear Lake (Rosenbaum and Heil, this volume). Deposits of older Blacks Fork or pre–Blacks Fork glaciations are mainly preserved on interfluves between tributary canyons of Bear River, but one small remnant thought to be older till was mapped on the valley floor ~4 km north of the Smiths Fork till limit (Dover and M'Gonigle, 1993).

Gravelly outwash terraces are preserved downvalley of the moraine sequence in the Bear River valley (Bryant, 1992; Dover and M'Gonigle, 1993; Munroe, 2001; Reheis, 2005). The lower two gravel terraces have been traced to terminal and recessional moraines of the Smiths Fork glaciation. A suite of at least four progressively higher and older terraces is preserved in places along the river between the glacial limit and Evanston (Fig. 2; Reheis, 2005), and these terraces probably represent outwash of pre–Smiths Fork glacial advances. Downstream of Evanston, the high terraces are discontinuous and correlations among remnants have not been attempted. Relatively extensive terrace deposits only a few meters above the floor of the valley are assumed to be of late Pleistocene age (marine oxygen isotope stages 2–4) based on limited soil data and absence of meander scars; most of the valley floor is covered with Holocene alluvium (Reheis, 2005). Remnants of pre–late Pleistocene river terraces are apparently absent along the Bear River between Sage Creek Junction and Bear Lake Valley, possibly due to subsidence. The well-preserved, incisional terrace sequence along the upper Bear River and along the upper part of unglaciated Yellow Creek (Coogan and King, 2001; Reheis, 2005), which joins Bear River at Evanston, may record downcutting as the headwater adjusted to the stream capture discussed above (Hansen, 1985).

A notable feature of the surficial geology of the Bear River headwaters is the abundance and lateral extent of landslides developed in the Wasatch Formation (Bryant, 1992; Dover and M'Gonigle, 1993). These landslides have not been dated, but their appearance (undrained depressions and obvious hummocky surfaces; Reheis, 2005) suggests that most were active during the late Quaternary, and they may have contributed sediment to tributaries of Bear River.

In the northeastern Bear River drainage basin, the Smiths Fork drains the Salt River Range (Fig. 2). Cirque and small valley glaciers occupied a few north-facing positions but little outwash is preserved (Reheis, 2005). A laterally extensive, thick deposit of fluvial gravel and sand crops out near the confluence of Smiths Fork and the Bear River northeast of Cokeville (Rubey et al., 1980), with a surface elevation ~60 m above present river level. The deposits rise in altitude up Smiths Fork, indicating deposition by that stream, and also grade laterally upslope to the east and south into alluvial fan deposits (locally faulted against bedrock along a west-down normal fault; Fig. 2) and pediment gravel. Outcrops in two gravel pits show that these deposits, locally at least 25 m thick and possibly as much as 45 m thick (Rubey et al., 1980), consist of very well rounded, well-washed, cross-bedded gravel with a few sand lenses (Reheis, 2005). This thickness and the apparent absence of buried soils suggest rapid aggradation by Smiths Fork and the adjacent Bear River. Such aggradation must have occurred in response either to local subsidence or to a significant rise in base level downstream, perhaps due to an expansion of Bear Lake in the early to middle Pleistocene (discussed below; Laabs and Kaufman, 2003).

SURFICIAL GEOLOGY OF BEAR LAKE VALLEY

Sedimentation in Bear Lake responds to a complex set of geomorphic and hydrologic influences. Previous work has focused on the shoreline record of Bear Lake, and other chapters in this volume discuss interpretations of lake fluctuations and interactions with the Bear River largely based on data from lake and marsh cores. In this section, we present new information on glacial deposits in the Bear River Range west of Bear Lake and on the fluvial terraces of the Bear River near and downstream of its entrance into Bear Lake Valley. We synthesize previously published information on the exposed shoreline deposits (Laabs, 2001; Laabs and Kaufman, 2003) with recent observations (Reheis et al., 2005; Reheis, 2005). We also present new data on Holocene marsh deposits and associated Bear River alluvium, with implications for the northward migration of Bear River since the last highstand of Bear Lake.

Glacial Deposits in the Bear River Range

Two studies documented glaciation in the Bear River Range south of the study area, west of the Bear River drainage basin (DeGraff, 1979; Williams, 1964). To our knowledge, however, no previous studies indicated glaciation in the northern Bear River Range, which is lower in altitude than the range to the south. The drainage divide of the Bear River Range west of Bear Lake (Fig. 4) lies at ~2750 m above sea level (asl) and isolated peaks east of the divide are as high as 2900 m; there is no summit plateau that would provide a source for wind-driven snow to accumulate on the lee (eastern) side. Precipitation in the Bear River Range is high, averaging 125 cm yr^{-1}, three-fourths as snowfall, at Tony Grove Lake (2415 m asl; http://www.wcc.nrcs.usda.gov/snow/) to the southwest of Bear Lake. Effective moisture was likely much greater during the maximum expansions of Lake Bonneville to the west due to cooler temperatures and perhaps to increased lake-effect precipitation (Munroe et al., 2006).

Three major streams head in shallow cirques and drain the east slope of the Bear River Range. Paris, Bloomington, and St. Charles Creeks contained valley glaciers as much as 10 km long (Fig. 5; Reheis, 2005). Nearly undissected, bouldery moraines that are very fresh in appearance and which have surface soils

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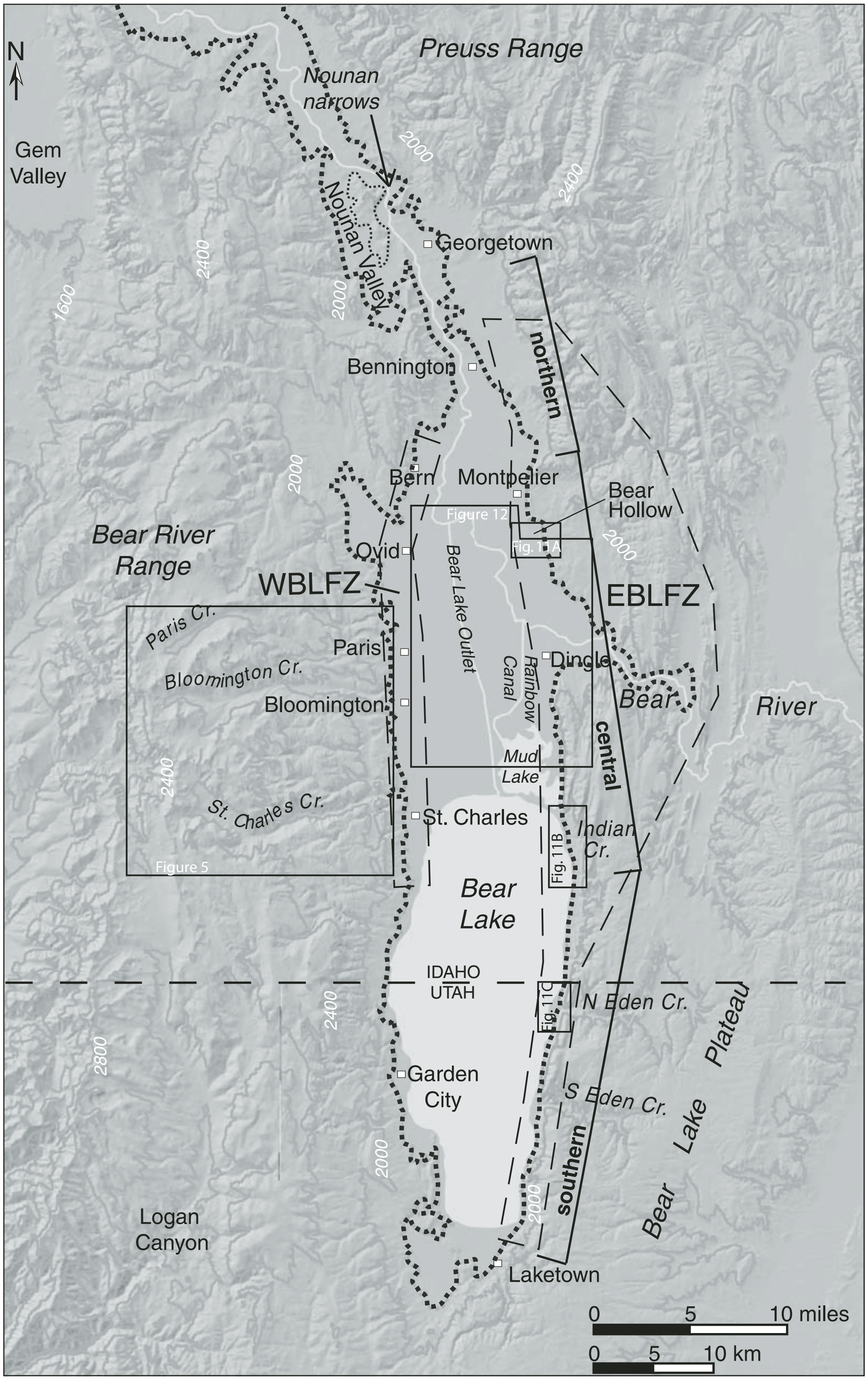

Figure 4. Shaded relief and topography of Bear Lake Valley. White rectangles show locations of towns; boxes show location of Figures 5, 11, and 12. Dashed lines enclose fault zones: west Bear Lake (WBLFZ, left) and east Bear Lake (EBLFZ, right). Brackets show north, central, and south sections of east Bear Lake fault zone. Dark gray lines are topographic contours (m asl) labeled with white text (contour interval = 400 m). Dashed heavy black contour marks elevation 1830 m asl in Bear Lake Valley, the approximate highest altitude of Bear Lake during the Pleistocene.

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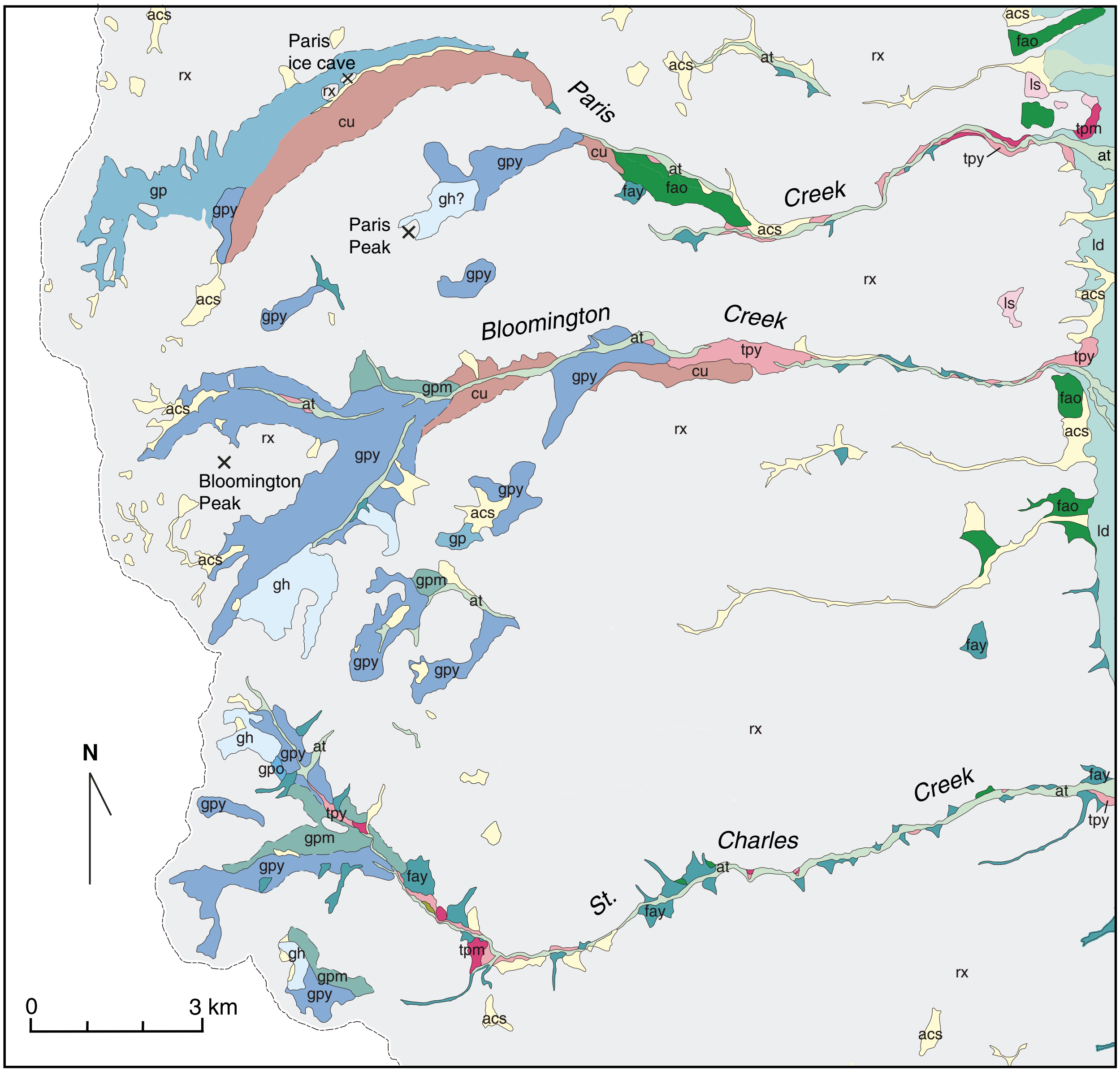

EXPLANATION

Alluvial deposits

- at — Fluvial channel and flood plain deposits (late Holocene)
- acs — Alluvium and colluvium (Holocene and late Pleistocene)--Includes fills in closed depressions (sinkholes)
- fay — Alluvial fans (Holocene and late Pleistocene)--Mostly undissected, smooth surfaces
- tpy — Fluvial terraces (late Pleistocene)--Includes outwash of Smiths Fork glaciation
- tpm — Fluvial terraces (middle Pleistocene)--Includes outwash of Blacks Fork glaciation
- fao — Alluvial fans (middle and early Pleistocene)--Incised irregular surfaces

Lake and marsh deposits

- ld — Lacustrine and associated marsh deposits (Holocene and Pleistocene)

Glacial deposits

- gh — Till of young cirque glaciers (early Holocene to latest Pleistocene)--Includes rock glacier deposits
- gp — Older till, undifferentiated (Pleistocene)--Subdivided into:
  - gpy — Till of Smiths Fork glaciation (late Pleistocene)--Moraine surfaces irregular, with undrained depressions
  - gpm — Till of Blacks Fork glaciation (middle Pleistocene)--Moraine surfaces smooth and dissected
  - gpo — Till of pre-Blacks Fork glaciations (middle and early? Pleistocene)--Moraine forms not preserved

Mass-wasting deposits

- ls — Landslide deposits (Holocene and Pleistocene)
- cu — Undifferentiated colluvium (Pleistocene)--Mainly glacial deposits modified by mass wasting

Bedrock

- rx — Bedrock, undifferentiated (Tertiary through Precambrian)

Figure 5. Surficial geologic map emphasizing glacial deposits and outwash in Bear River Range (Reheis, 2005). Queried map symbol (for example, gh?) indicates uncertain identification of map unit. Note that highest points (Bloomington and Paris Peaks) do not lie on Bear Lake drainage divide, and that outwash (tpy and tpm) is of limited extent.

containing very little infiltrated fine-grained eolian sediment lie as much as 1 km from cirque headwalls. The landform properties suggest a small Holocene or latest Pleistocene advance (Figs. 6B and 6C). Well-preserved, sharp-crested, bouldery moraines with undrained depressions, and with oxidized surface soils containing eolian sediment to a depth of ~20 cm, extend 5–6 km from the cirque headwalls (Fig. 6A) and are correlated here with the late Pleistocene Smiths Fork glaciation (Munroe, 2001). Subdued, broad-crested moraines with boulders set in a finer-grained matrix than the younger moraines are locally preserved higher on valley walls and as much as 1 km downvalley from the Smiths Fork-equivalent moraines; these are correlated to the Blacks Fork glaciation, probably of middle Pleistocene age (Munroe, 2001; marine oxygen isotope stage 6). Diamictons mapped as much as 1–2 km downvalley of these older moraines may represent one or more older glacial advances. Moraine complexes of short valley glaciers are also preserved on the northeast flanks and downvalley of high peaks and ridges that lie east of the drainage divide, such as Paris Peak (Fig. 5; Reheis, 2005).

Within the glaciated terrain and especially in the cirques are numerous sinkholes and collapse features developed in the mostly carbonate bedrock (Fig. 2; Oriel and Platt, 1980). Large portions of some valleys below the cirques, such as the upper several kilometers of Paris Creek valley, have no integrated drainages or surface streams. Closed basins surround many of the sinkholes, indicating that all runoff must exit as groundwater. Sinkholes in valleys are commonly floored by fine-grained deposits (unit acs, Fig. 5) that accumulated after the glaciers withdrew; locally, small collapse features have formed in the basin floors. Other collapse features in bedrock, such as Paris Ice Cave (Fig. 6D) and a pit near the Bloomington Lake trailhead (Fig. 5), permit free drainage into open cracks. This karst topography, similar to that reported in the glaciated areas farther south in the Bear River Range (Wilson, 1979), may have permitted basal glacier meltwater and runoff to enter the groundwater directly. Such direct recharge to groundwater would efficiently deliver glacial meltwater through a carbonate aquifer into Bear Lake. Limited glacial surface runoff is consistent with the paucity of outwash terraces downvalley of moraines (Reheis, 2005). For example, terraces that probably represent late Pleistocene outwash were mapped only ~2 km downstream of terminal moraines (Fig. 5). If outwash streams were not effective at transporting sediment, this implies that much of the glacially eroded sediment is still stored within the valleys upstream of the lake. Much further work would be required to test these hypotheses and the glacial correlations.

Fluvial Terraces of Bear River

Bear River and its terraces in the vicinity of Bear Lake can be divided into three reaches based on location relative to tectonic displacements (Figs. 4 and 7): (1) the footwall block upstream of the east Bear Lake fault zone, (2) the graben area between the east and west Bear Lake fault zones, and (3) downstream of the graben area, beginning at the confluence of Ovid Creek and Bear River. The terrace deposits (Reheis, 2005) have not been previously studied, probably because they are commonly buried or obscured by thick deposits of loess on the footwall block

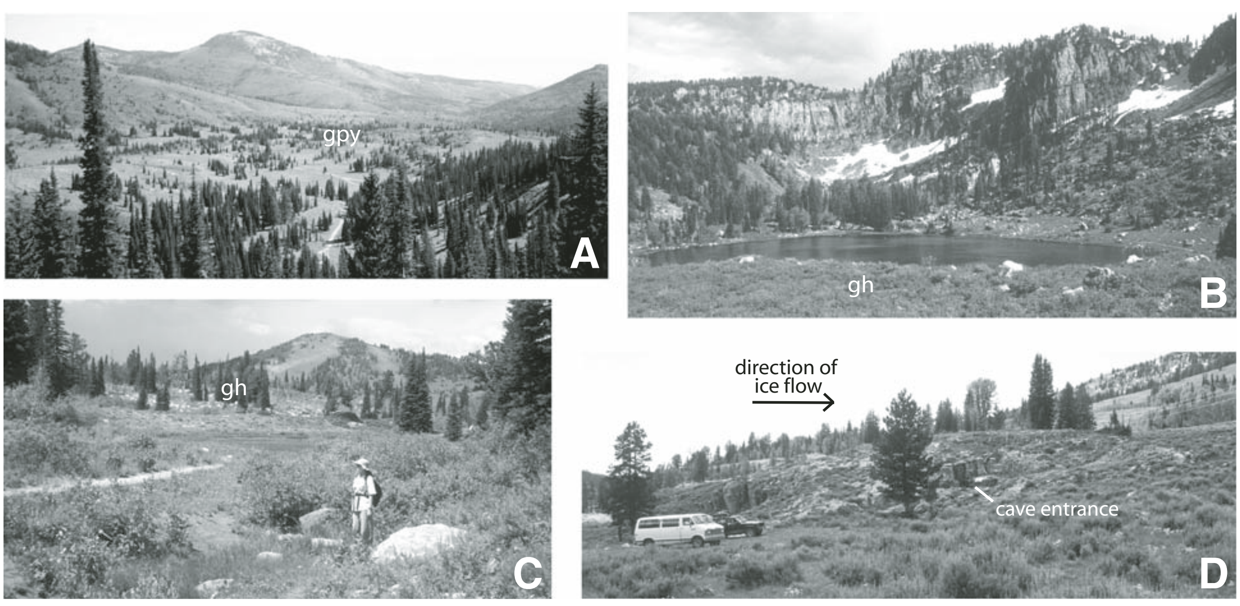

Figure 6. Photographs of glacial deposits in Bear River Range (see Fig. 5 for locations). (A) View downvalley of recessional moraine of Smiths Fork glaciation (gpy) in valley of Bloomington Creek. (B) Holocene moraine (gh) and part of compound cirque at head of Bloomington Creek. (C) View of outer part of Holocene moraine (gh) in Bloomington Creek. (D) View to southwest of roche moutoneé at Paris ice cave in valley of Paris Creek.

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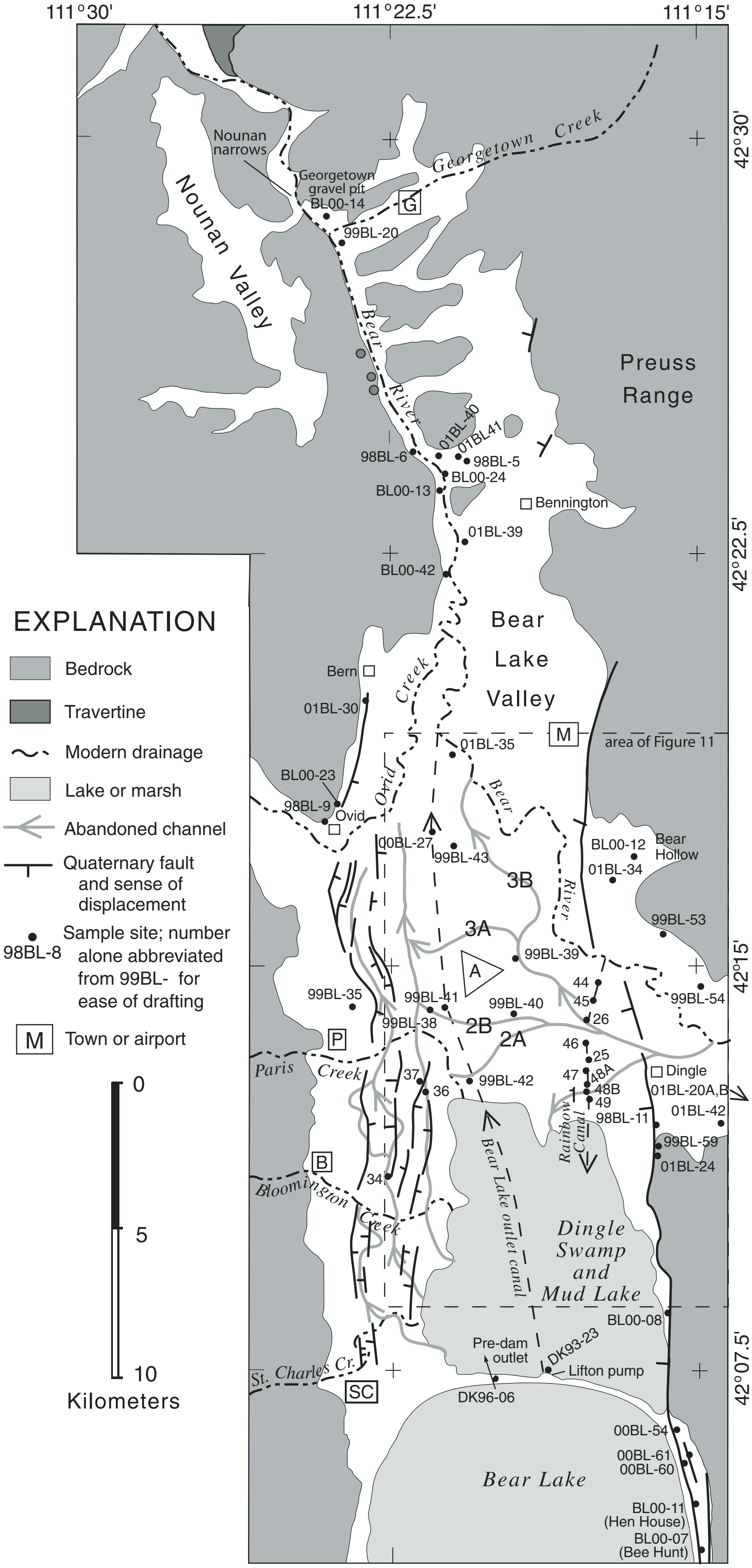

Figure 7. Sketch map showing Bear River and inferred former courses (2B, etc.), abandoned channels on the west side of the valley, faults with Holocene displacement, and selected terrace and lake-deposit study sites. Stratigraphy of some sites is shown in Figures 9 and 13. Dashed box shows area of composite aerial photograph (Fig. 12). A—airport; B—Bloomington; G—Georgetown; M—Montpelier; P—Paris; SC—St. Charles.

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and by lake and marsh deposits on the hanging-wall block of the east Bear Lake fault zone. The fluvial terraces, combined with evidence of lake-level fluctuations, record a complex history of uplift, downcutting, and aggradation in response to faulting, threshold incision, lake-level fluctuations, and glacial sediment loading. We use a variety of sources to draw tentative correlations among terrace remnants and lake highstands. These include surficial mapping; stratigraphic relations among alluvium, loess, and lake sediments; and chronologic data from radiocarbon, amino-acid racemization, and tephrochronologic techniques. In this section we address the terrace mapping, dating, and correlation; in the two following sections we discuss the exposed record of lake-level fluctuations and the synthesis of these two data sets to interpret the record of downcutting and aggradation caused by tectonics, lake-level fluctuations, and outwash deposition.

Terraces along the Footwall Block

East of the east Bear Lake fault zone, terrace gravels lie on the footwall block (Fig. 7). Within this reach, a group of four terraces lie between 40 and 80 m above the modern river (Fig. 8, sites 99BL-53 and -54, 01BL-20 and -42) and locally cap fine-grained sediment interpreted as lacustrine deposits of Bear Lake (Table 1, Figs. 8 and 9). Because lacustrine deposits have not been found at equivalent altitudes west of this fault zone, these very high terrace gravels and lake deposits probably record uplift of the footwall block. They are nested within the valley cut by the Bear River during uplift of the Bear Lake Plateau, and lie more than 200 m lower than the ancient river gravels preserved on top of the plateau (Fig. 2). A group of three younger terraces less than 15 m above the river are preserved where the Bear River enters the Bear Lake Valley.

Limited age control suggests that the older group of terraces ranges from Pliocene(?) to middle Pleistocene in age (Table 1). At site 99BL-53 north of the river (Fig. 7), a deep roadcut exposes ~5 m of terrace gravel 80 m above the modern river. The gravel caps 20 m of fine-grained lacustrine deposits that in turn overlie bedrock (Fig. 9). Reworked tephra within the fine-grained beds yielded no definitive correlations with dated rocks (A. Sarna-Wojcicki, U.S. Geological Survey, 2000, personal commun.). These deposits form a deeply dissected belt parallel to the modern river and are thickly blanketed with loess. A loess-buried terrace gravel farther upstream at site 01BL-20B may be correlative (Figs. 7 and 8). A lower terrace gravel south of the river at site 01BL-42, also buried by loess, overlies several meters of deposits interpreted as lacustrine fan-delta deposits (Figs. 8 and 9). These lake deposits conformably overlie locally

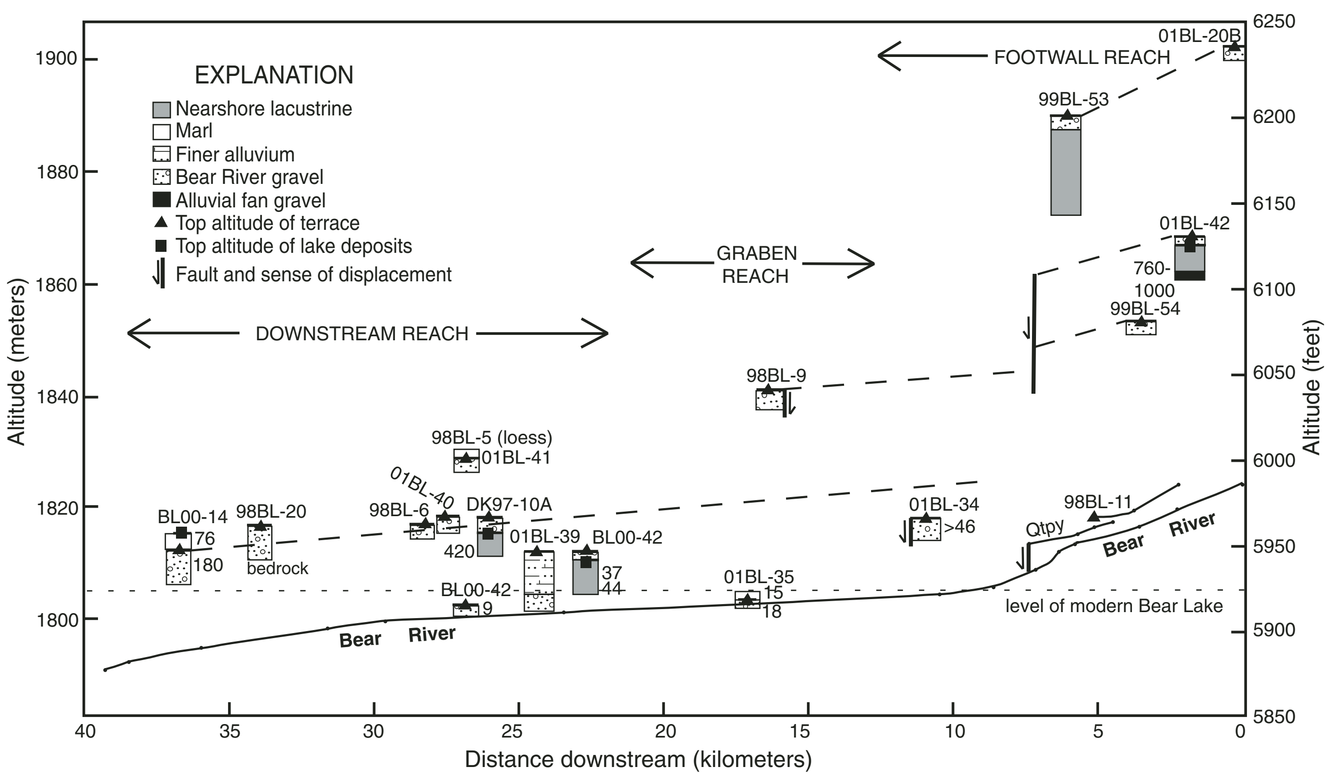

Figure 8. Bear River terrace sites and modern gradient of river showing study sites and simplified stratigraphic columns. Dashed lines show inferred terrace correlations based on height above river and limited age control; ages in ka are listed next to columns. Heavy gray arrows separate reaches of Bear River and terraces defined by major faults: footwall reach includes east Bear Lake fault zone and sites upstream, graben reach includes one site on footwall block of west Bear Lake fault zone and one site between fault zones, and downstream reach includes sites downstream of active faulting.

TABLE 1. BEAR RIVER TERRACE SITES GROUPED BY RELATIVE AGES, WITH ALTITUDES, UNDERLYING STRATIGRAPHY, AND CORRELATIONS WITH BEAR LAKE PHASES (LOCATIONS ON FIGS. 4, 7, AND 8)

Site number	Location	Position relative to fault*	Gravel altitude (m)†	Height above river (m)	Age control§	Stratigraphy and key soil horizons
BL00-24	Bennington Bridge	Downstream reach	1804	3	8.2 ± 1.6 ka (AAR)	1.5 m loess/t.g.
Willis Ranch Phase, ca. 9 cal ka, ~1814 m (many preserved barriers)						
Raspberry Square Phase, 16–15 cal ka, ~1814 m (Garden City pit)						
01BL-35	Bear River floodplain cut	Graben reach	1803	1	16.3 ka (^{14}C)	2 m marl/t.g.
BL00-42	Bear River cut (Red Pine Hollow)	Downstream reach	1812	9	<44.3 ka, <36.8 ka (^{14}C)	t.g./rippled sand
01BL-39#	Pescadero	Downstream reach	1794–1807	4	Altitude, stratigraphy	8 m sandy silt/t.g.
Jensen Spring Phase, 46–39 ka, ~1817 m (Ovid spit); 98BL-11 on FW block may be correlative fan-delta unit of Bear River						
01BL-34	Gravel pit W of US-30	Footwall block of EBLFZ	1818	14	>46 ka (^{14}C)	Bw/Bk (stage II $CaCO_3$)
99BL-56-57	Gravel pits E of Dingle cemetery	Footwall block of EBLFZ	1838	15	Soils, stratigraphy	2 m loess (Bw/Bk)/2 m loess (Bk)/t.g.
Late Bear Hollow Phase, 170–90 ka, ≥1818 m (Georgetown pit), <1830 m (FW; upper Bear Hollow pit), <<1866 m (FW; powerline)						
01BL-20A	E of old fan-delta site	Footwall block of EBLFZ	1850	26	Altitude	>4 m loess/t.g.
BL00-13	S of Bennington Bridge	Downstream reach	1818	15	>43.3 ka (^{14}C); <<419 ± 81 ka (AAR)	t.g./2.5 m marl, sand, and silt
01BL-40	N of Bennington Bridge	Downstream reach	1818	17	Soils, stratigraphy	25 cm loess/Bw/Bk
98BL-6	W side river	Downstream reach	1817	17	Altitude	Not described
99BL-20	Gravel pit S of old railroad	Downstream reach	1811–1817	21	Soils, stratigraphy	2 m loess/t.g. (Btb/Bkb, stage III $CaCO_3$)
BL00-14	Georgetown gravel pit, lower	Downstream reach	1810–1816	20	180 ± 60 ka; >76 ± 35 ka (AAR)	t.g. (Bt)/3 m marl and mud/t.g.
Middle Bear Hollow Phase, 450–380 ka, ≥1817 m (S of Bennington Bridge), ≤1824 m (FW?; culvert cut)						
99BL-54	E of Dingle siding	Footwall block of EBLFZ	1853	35	Soils, stratigraphy	Bk (stage IV $CaCO_3$)
98BL-5, 01BL-41	E of Bennington Bridge	Downstream reach	1829	27	Soils, stratigraphy	1.3 m loess (Bt/Bk)/>1 m loess (Bt/Bk)/t.g.
Early Bear Hollow Phase, 1100–820 ka, <1829 m (FW, lower Bear Hollow pit), <<1863 m (FW; old fan-delta site)						
01BL-42, 99BL-58	Old fan-delta site	Footwall block of EBLFZ	1868	46	<760-1200 ka (tephra)	>3 m loess/t.g./fan-delta/alluvial fan with tephra
98BL-9**	Ovid cemetery gravel pit	Footwall block of WBLFZ	1841	38	Soils, altitude	Eroded Bkm (stage V $CaCO_3$)
Other phases, late Pliocene–early Pleistocene (ostracodes): <1846 m (FW, 00BL-60), <<1870 m (FW, 00BL-61; <<1876 (99BL-53)						
01BL-20B	E of fan-delta site 01BL-42	Footwall block of EBLFZ	1905	81	Soils, stratigraphy	>4 m loess/alluvial fan/t.g.
99BL-53	US-30 cut, prodelta site	Footwall block of EBLFZ	1890	79	Reworked Tertiary tephra	t.g./20 m bedded sand and silt, clay at base

Note: AAR—amino acid racemization; FW—footwall; t.g.—terrace gravel.
*EBLFZ—East Bear Lake fault zone; WBLFZ—West Bear Lake fault zone; river reaches shown on Figure 8.
†Elevation in meters above sea level. Range of altitudes shows significant aggradation of fluvial deposits.
§ ^{14}C data given in Table 2; AAR ages from Laabs and Kaufman (2003).
#May postdate Raspberry Square phase and predate Willis Ranch phase.
**Assumed to correlate to 99BL-54; is definitely too high to correlate to sites 99BL-5 or 01BL-41 (see Fig. 8).

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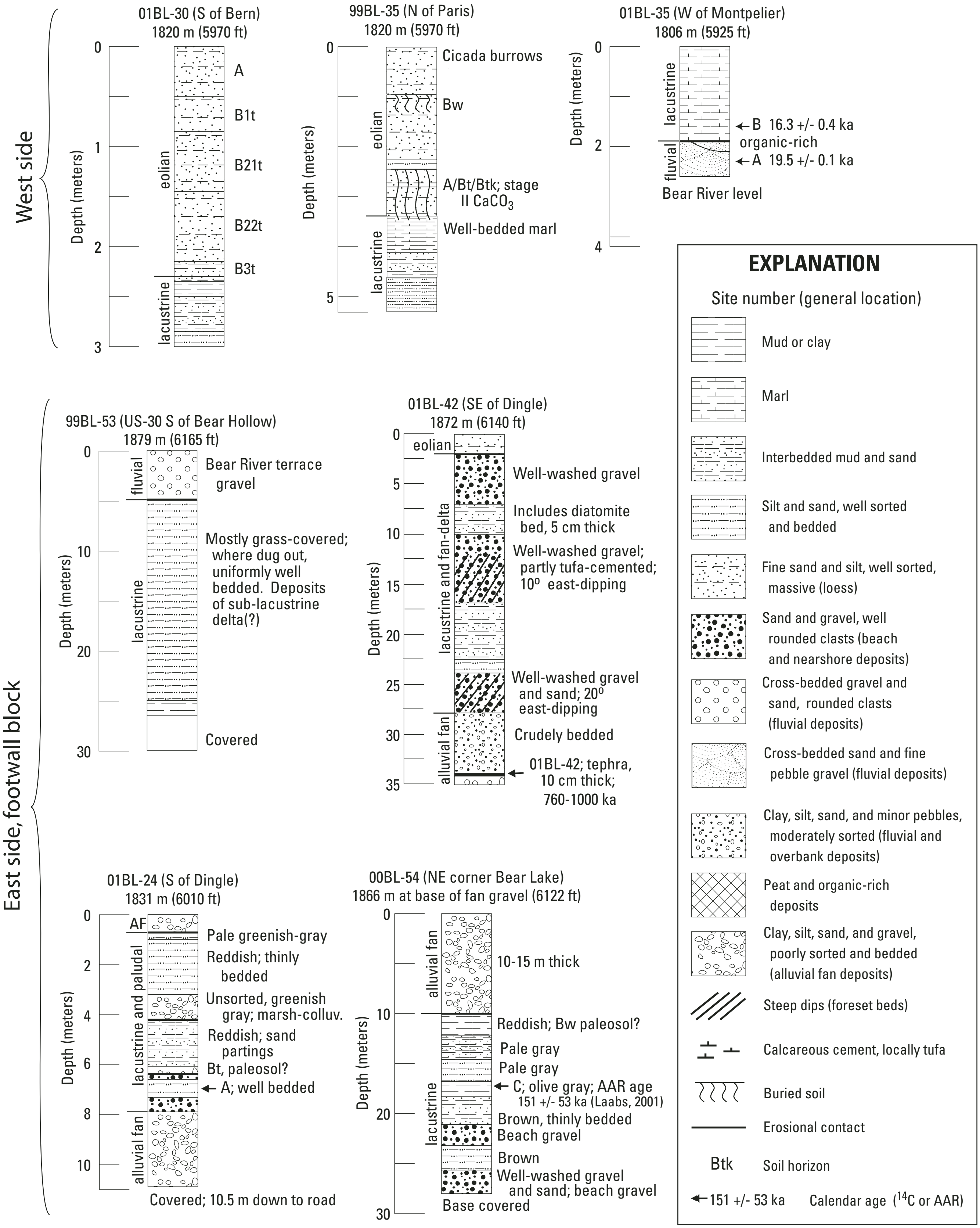

Figure 9. Selected stratigraphic sections related to Bear River terraces and lake deposits measured from outcrops and auger holes in Bear Lake Valley (data from Reheis et al., 2005). Site locations are on Figure 7. Column to right of lithology gives descriptive information such as color, sample data, and soil horizons, and column to left gives interpreted environment of deposition (AF—alluvial fan). See Table 2 for radiocarbon ages and Laabs and Kaufman (2003) for amino acid racemization (AAR) ages.

derived alluvial-fan deposits containing a 10-cm-thick, reworked, rhyolitic tephra layer that is chemically correlative with either the Bishop ash bed (760 ka) or Glass Mountain tephra (ca. 1 Ma) (A. Sarna-Wojcicki, U.S. Geological Survey, 2000, personal commun.; Sarna-Wojcicki et al., 2005). The lowest terrace in the older group (site 99BL-54) has a soil characterized by stage IV pedogenic $CaCO_3$, suggesting an early middle Pleistocene age by comparison with the carbonate morphology of soils on younger deposits (e.g., 99BL-20, Table 1). On the other side of the Bear River Valley, a terrace gravel preserved on the footwall block of the west Bear Lake fault zone (98BL-9) bears a truncated, strongly developed soil with stage V pedogenic $CaCO_3$. On the basis of this morphology and elevation above the river, it may be early Pleistocene in age.

Younger terraces in the footwall-block reach are late Pleistocene in age on the basis of soil development, one ^{14}C age, and relations to lacustrine deposits. These terraces are abruptly truncated by left-stepping strands of the east Bear Lake fault zone (Figs. 7 and 8), and fine-grained lake deposits on the hanging-wall block are in fault contact with terrace deposits. Terrace gravel ~15 m above Bear River is well exposed in 5-m-deep gravel pits at site 01BL-34, where gastropod shells yielded an age of 49,950 ± 1020 ^{14}C yr B.P. (Tables 1 and 2); we interpret this to be a minimum age for the deposit. Lower, younger terrace deposits have not been directly dated. Terrace height and soil development suggest that a terrace or fan-delta deposit at site 98BL-11 (Reheis, 2005) may correlate with or slightly postdate lacustrine deposits with ages of 46–39 ka (Laabs and Kaufman, 2003; discussed below).

Terraces and Deposits within the Graben

Within the graben reach between the east and west Bear Lake fault zones (Figs. 7 and 8), subsidence has resulted in burial of terraces by younger sediments. One cutbank of the Bear River (01BL-35, Fig. 9) exposes marl overlying cross-bedded fluvial sand; gastropod shells in these deposits yielded ages of 16,320 ± 380 (in marl) and 19,480 ± 100 (in sand) cal yr B.P. (Tables 1 and 2). No data are available for the possible reservoir effect on gastropods in fluvial and marsh deposits near Bear Lake; the reservoir effect for ostracode shells in Bear Lake cores is 370 ± 105 yr (Colman et al., this volume). Outcrops along the Rainbow Canal, transverse to the modern course of Bear River, show that river gravel and sand representing the aggradational surface of a fluvial fan extend at least 5 km south of the river (see Fig. 13A discussed below). One bivalve shell from this alluvium (site 99BL-45) yielded an age of 12,980 ± 130 cal yr B.P., and gastropod shells from sand interpreted as fluvial more than 2 m below the surface in auger hole 99BL-42 gave an age of 14,210 ± 390 cal yr B.P. These deposits descend ~3 m in altitude southward along the canal exposure and are overlain by fine-grained fluvial and marsh deposits of Holocene age.

Terraces Downstream of Active Faults

Terraces are common downstream of the northern limits of active Quaternary faults (as mapped by Reheis [2005]) in Bear Lake Valley (Table 1, Fig. 7). Because altitudes of these terraces appear to be little influenced by local fault displacement during the late Quaternary, we assume they more accurately reflect base-level and climatic changes and effects of upstream faulting than terraces to the south and east. The highest river gravel (site 01BL-41) northeast of Bennington is poorly exposed on the flank of a low plateau buried by loess. A backhoe trench at nearby site 98BL-5 (Fig. 7) exposed more than 2.2 m of loess including a surface soil and a buried soil, each with 60- to 80-cm-thick argillic horizons and stage II-III calcic horizons, suggesting an age of middle Pleistocene or older for the buried gravel.

The next lower set of terrace deposits includes five localities close to the river northward from Bennington (Table 1, Figs. 7 and 8). Their elevations fall within a range of 15–20 m above the modern river, but they are not all the same age because their stratigraphic sequences and surface soils differ. From south to north, at site BL00-13, fluvial sand and gravel unconformably overlie nearshore lacustrine deposits. A nonfinite age of >43,260 ^{14}C yr B.P. was obtained from gastropod shells in the alluvium (sample DK97-10A, Table 2) and an age of ca. 420 ka was obtained on shells in the lacustrine deposits by using amino acid racemization (AAR; Table 1; Laabs and Kaufman, 2003). At sites 01BL-40 and 98BL-6, thin loess (~25 cm) overlies terrace gravel and the surface soils are weakly developed (A/Bw/Bk, stage II $CaCO_3$), suggesting a late Pleistocene age. At site 99BL-20, 2 m of loess with a weak surface soil (A/Bw) and a stronger buried soil (Bt/Bk, stage III $CaCO_3$) overlies 6–7 m of river gravel, in turn overlying bedrock (Salt Lake Formation). At site BL00-14, in the Georgetown gravel pit, calcareous marsh or shallow lacustrine deposits 3 m thick conformably overlie nearly 6 m of fluvial sand and gravel. Gastropod shells gave AAR ages of ca. 76 ka (marsh) and 180 ka (gravel) (Table 1; Laabs and Kaufman, 2003). Such ages are compatible with the soil development observed at nearby site 99BL-20 and perhaps with the nonfinite radiocarbon age at BL00-13, but the latter two fluvial units are not buried by marsh or lacustrine deposits. Soils at sites 98BL-6 and 01BL-40 seem too weakly developed to assign a middle Pleistocene age to these sites, unless the soils were eroded.

Two adjacent terrace localities ~8 m above the modern river show evidence for incision followed by rapid aggradation. At site BL00-42 (Figs. 7 and 8; Bear River cutbank of Laabs and Kaufman, 2003), 1.4 m of alluvial sand and gravel overlies lacustrine sand containing gastropods with an age of 47,000 ± 1000 cal yr B.P. (Tables 1 and 2). At a slightly lower surface altitude 2 km downstream (site 01BL-39), terrace gravel at river level is 4 m thick and grades up into ~8 m of interbedded silt and fine sand. No datable materials were found in this outcrop, but the weak surface soil suggests an age no older than late Pleistocene. Low terraces ~3–4 m above the modern river are locally preserved. In one of these low terraces (site BL00-24), gastropod shells yielded an AAR age estimate of ca. 9 ka. These variable terrace altitudes and stratigraphic relations are discussed below in the context of shoreline fluctuations.

TABLE 2. RADIOCARBON AGES FROM OUTCROPS AND AUGER HOLES IN THE BEAR LAKE DRAINAGE BASIN, 1998–2001

Sample number*	General location	Latitude (N)	Longitude (W)	Material dated	Sample depth (cm)	^{14}C lab number[†]	Radiocarbon age (yr B.P.)	Calibrated age[§] (yr B.P.)
99BL-26B	Rainbow Canal N of airport road bridge	42°13.992’	111°49.166’	Gastropod shells	80–100	WW-3049	1400 ± 40	1325 ± 55
99BL-34	Auger hole E of Bloomington	42°11.064’	111°22.625’	Gastropod shells	0–50	WW-2583	7760 ± 70	8530 ± 130
99BL-37	Auger hole E of Bloomington, footwall of fault cutting paleochannel	42°12.775’	111°21.814’	Shell fragments	70–80	WW-2584	1955 ± 70	1820 ± 95
				Gastropod shells	80–100	WW-2585	10,420 ± 80	12,360 ± 310
				Gastropod shells	200–230	WW-2586	9820 ± 75	11,250 ± 160
99BL-38	Auger hole, Paris-Dingle road, paleo-course BR2	42°14.140’	111°21.611’	Gastropod shells	0–20	WW-2587	2645 ± 55	2530 ± 180
				Gastropod shells	230–250	WW-2588	7985 ± 70	8830 ± 200
99BL-39	Auger hole E of airport, paleo-course BR3	42°15.119’	111°19.498’	Gastropod shells	60–70	WW-2589	2445 ± 55	2530 ± 180
99BL-42	Auger hole W of outlet canal; Bear River or Bear Lake channel?	42°12.860’	111°20.583’	Gastropod shells	0–25	WW-590	1720 ± 55	1630 ± 110
				Gastropod shells	200–220	WW-2591	12,220 ± 100	14,210 ± 390
99BL-45	Cut, Rainbow Canal; paleo-course BR3	42°14.270’	111°17.639’	Gastropod shells	60–65	WW-2592	5130 ± 65	5280 ± 150
				Gastropod shells	120–130	WW-2593	8520 ± 70	9520 ± 120
				Bivalve shells	200–220	WW-2594	11,015 ± 85	12,980 ± 130
99BL-47	Cut, Rainbow Canal	42°13.001’	111°17.762’	Gastropod shells	0–80	WW-2595	7150 ± 70	8000 ± 160
99BL-48A	Cut, Rainbow Canal	42°12.658’	111°17.048’	Gastropod shells	165–185	WW-3048	8460 ± 40	9480 ± 50
99BL-48B	Cut, Rainbow Canal	42°12.654’	111°17.706’	Gastropod shells	110–140	WW-3050	7870 ± 40	8670 ± 120
99BL-49	Auger hole, Rainbow Canal; paleo-course BR1	42°12.472’	111°17.721’	Gastropod shells	20–50	WW-2596	4000 ± 60	4460 ± 180
					270–300	WW-2597	6925 ± 70	7770 ± 120
99BL-59	Culvert cut S of Dingle	42°11.640’	111°16.082’	Shell fragments	B, 180	WW-2599	>40,800 ± 1600	N.D.
					A, 200–250	WW-2598	>39,100 ± 1100	N.D.
00BL-27	Cut, outlet canal	42°30.0’	111°21.5’	Gastropod shells	10–20	WW-3047	4880 ± 40	5600 ± 110
00BL-63A	Waterline trench S of Indian Creek	42°05.181’	111°15.309’	Gastropod and bivalve shells	Several meters	WW-3369	39,870 ± 490	N.D.
01BL-34	Borrow pit W of US-30, Bear Hollow	42°16.5’	111°17’	Gastropod shells	350	WW-3721	>45,950 ± 1020	N.D.
01BL-35	Bear River cutbank N of US 30	42°19.81’	111°23.26’	Gastropod shells	B, 150–190	WW-3723	13,675 ± 50	16,320 ± 380
					A, 190–250	WW-3722	16,350 ± 50	19,480 ± 100
DK96-01 (BL00-10)	Cisco Beach, edge of North Eden fan	41°58.45’	111°16.17’	Mollusk shell	150	WW-1557	10,420 ± 50	12,510 ± 170
DK96-06D	Fish ladder, W of Lifton pumping station	42°07.35’	111°20.15’	*Stagnicola* shell	70	NSRL-1566	7210 ± 40	8160 ± 100
DK96-06B	Fish ladder, W of Lifton pumping station	42°07.35’	111°20.15’	*Stagnicola* shell	150–160	WW-1561	5650 ± 40	6410 ± 100
DK96-06B	Fish ladder, W of Lifton pumping station	42°07.35’	111°20.15’	Charcoal	150–160	WW-1566	5530 ± 50	6410 ± 70
DK96-06A	Fish ladder, W of Lifton pumping station	42°07.35’	111°20.15’	*Valvata* shell	210–220	NSRL-10940	8520 ± 65	9500 ± 420
DK97-10A (BL00-14)	Bennington, N of Pescadero	42°23.80’	111°21.20’	*Sphaerium* shell	Several meters	WW-1559	>43260	N.D.
DK98-02A (99BL-25)	Rainbow Canal at airport road bridge	42° 14.27’	111°17.63’	*Stagnicola* shell	200–250	NSRL-10569	8350 ± 70	9290 ± 160
DK98-03A (BL00-02C)	Road cut north of North Eden canyon	41°59.53’	111°15.88’	*Stagnicola;* 2 shells	Several meters	NSRL-10570	37,900 ± 460	41,000 ± 1000
DK98-03B (BL00-02C)	Road cut north of North Eden canyon	41°59.53’	111°15.88’	*Stagnicola* shell	Several meters	NSRL-10571	38,700 ± 790	41,500 ± 1000
DK99-11 (BL00-23)	Ovid spit, N of Ovid	42°18.00’	111°23.62’	Shell fragments	180–210	NSRL-11353	41,240 ± 640	43,700 ± 1000

(*Continued*)

TABLE 2. RADIOCARBON AGES FROM OUTCROPS AND AUGER HOLES IN THE BEAR LAKE DRAINAGE BASIN, 1998–2001 (*Continued*)

Sample number*	General location	Latitude (N)	Longitude (W)	Material dated	Sample depth (cm)	^{14}C lab number†	Radiocarbon age (yr B.P.)	Calibrated age§ (yr B.P.)
DK99-13 (BL00-42)	Bear River cutbank, W of Bear River	42°22.17'	111°21.33'	*Stagnicola* shell	170–180	NSRL-11354	36,800 ± 790	30,500 ± 1000
DK99-18A (BL00-02)	North Eden canyon, W of highway	41°59.22'	111°15. 56'	Charcoal	50–150	NSRL-11355	8780 ± 90	9840 ± 130
DK99-18B (BL00-02)	North Eden canyon, W of highway	41°59.38'	111°15.95'	*Discus* shell	50–150	NSRL-11356	10,490 ± 100	12,830 ± 190
DK99-28C (BL00-14)	Georgetown gravel pit	42°28.83'	111°24.00'	*Discus* shell	100–140	NSRL-11359	>45,200	N.D.
BL00-11	Hen House, E of highway	42°05.02'	111°15.28'	*Sphaerium* shell	100–150	NSRL-12061	36,000 ± 320	39,000 ± 1000
BL00-41A	Garden City borrow pit	41°57.13'	111°23.82'	Mollusk shell	125–160	NSRL-12062	13,540 ± 70	16,310 ± 240
BL00-41B	Garden City borrow pit	41°57.13'	111°23.82'	Mollusk shell	125–160	NSRL-12063	13,280 ± 70	16,010 ± 270
BL00-42	Bear River cutbank, W of Bear River	42°22.17'	111°21.33'	*Stagnicola* shells	175–185	NSRL-12064	44,300 ± 920	47,000 ± 1000

*Site number given in parentheses if different from sample number. DK- and BL- data taken from Laabs and Kaufman (2003); other samples reported in Reheis et al. (2005).
†Letter prefixes indicate laboratory: WW—U.S. Geological Survey, Reston, Virginia; NSRL—Institute of Arctic and Alpine Research, Boulder, Colorado.
§For ages <21,000 radiocarbon years, calibrations performed using CALIB program, version 4.4 (Stuiver et al., 2005). Some ages >21,000 radiocarbon years corrected using data in Kitagawa and van der Plicht (1998). N.D.—not determined. Calibrated age values are midpoints plus or minus one-half the 2σ range.

Fluctuations of Bear Lake

Bear Lake fluctuates between a topographically closed basin when lake level is largely controlled by groundwater discharge and a topographically open basin when Bear River flows into and out of the basin (Laabs and Kaufman, 2003; Bright et al., 2006; Kaufman et al., this volume). Although Bear Lake experienced regional climate changes similar to nearby Lake Bonneville, climate change alone cannot account for Bear Lake highstands because they did not always coincide with pluvial maximum conditions (Kaufman et al., this volume). Interpreting the history of water-level changes in Bear Lake Valley is complicated by active tectonism and geomorphic processes that control basin geometry and the altitude of the basin's threshold. Because the present threshold of Bear Lake is the northern shoreline (Fig. 7), only ~2 m above modern lake level (nominally 1805.5 m asl; at times lower prior to human modifications), the altitude and location of the threshold must have been different in the past to allow higher lake levels to occur. At such times, the lake expanded to a northern, formerly higher threshold at or south of Nounan narrows (Fig. 7; Robertson, 1978; Laabs and Kaufman, 2003). Thus, maximum lake level is ultimately controlled by the altitude of this bedrock threshold, which can be changed by downcutting, fault displacement, aggradation, or landsliding (Laabs and Kaufman, 2003). An additional factor in threshold rise, which we cannot evaluate with available data, is the possible effect of travertine dams in the area of Nounan narrows (Fig. 7). Thick, extensive sheets of travertine lie on the east side of Bear River downstream of the narrows (mapped by Oriel and Platt, 1980). These sheets range from 1805 m asl to as high as 1830 m asl along side tributaries and on the west side of the river south of the Georgetown gravel pit (BL00-14, Fig. 7). We do not know the age of the travertine or whether such outcrops may once have been extensive enough to fill the valley, thus temporarily raising the threshold. If this occurred, it was likely to have been at a time of greatly reduced flow in Bear River.

In the following discussion, results from Laabs and Kaufman (2003) are augmented with stratigraphic and chronologic data from other outcrops (Reheis, 2005; Reheis et al., 2005). The paucity of exposures, uncertainty of water depth for some sediment, and post-depositional deformation hampers an accurate reconstruction of water level in Bear Lake Valley. This problem is particularly acute for the Bear Hollow phases because of tectonic subsidence and footwall uplift. We assume that sites within a given age range belong to the same lake phase, which may include more than one lake-level fluctuation. A second major assumption is that sites containing lake sediments along the west side of Bear River Valley and north of most mapped Pleistocene faults (Fig. 7) are at or near their original depositional altitudes; tectonic displacement is greatest on the east side of the valley and increases southward (discussed below).

Pliocene or Early Pleistocene Lake

Lacustrine deposits as high as 1876 m asl (99BL-53, Figs. 7 and 9) on the footwall block of the east Bear Lake fault zone

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represent the oldest exposed evidence of a perennial lake in Bear Lake Valley during the Pliocene or early Pleistocene (Reheis et al., 2005). Well-sorted, well-bedded, fine sand and silt with minor green clay constitute 20 m of deposits that are interpreted as lacustrine sediment deposited near the mouth of the ancestral Bear River. Although undated, these deposits are nearly 70 m above modern lake level, suggesting much more fault displacement and hence a much older age than lower deposits. Two sites near the mouth of Indian Creek, 00BL-61 and 00BL-60, are ~65 and 41 m, respectively, above modern lake level (Figs. 4 and 10). These sites expose very similar sequences 10–12 m thick including, from base to top, green clay, gray, bedded fine sand, reddish mud, and beach gravel (sections not measured in detail). Reddish mud implies a contribution from the ancestral Bear River, if the sediment characteristics are similar to those in cores of the late Pleistocene age Bear Lake units (Rosenbaum et al., this volume). The green clay at 00BL-60 and the gray sand at 00BL-61 contain previously unknown ostracode species of the genera *Candona* and *Limnocythere* that resemble Pliocene and early Pleistocene ostracodes known from large lakes elsewhere (Table 2 *in* Reheis

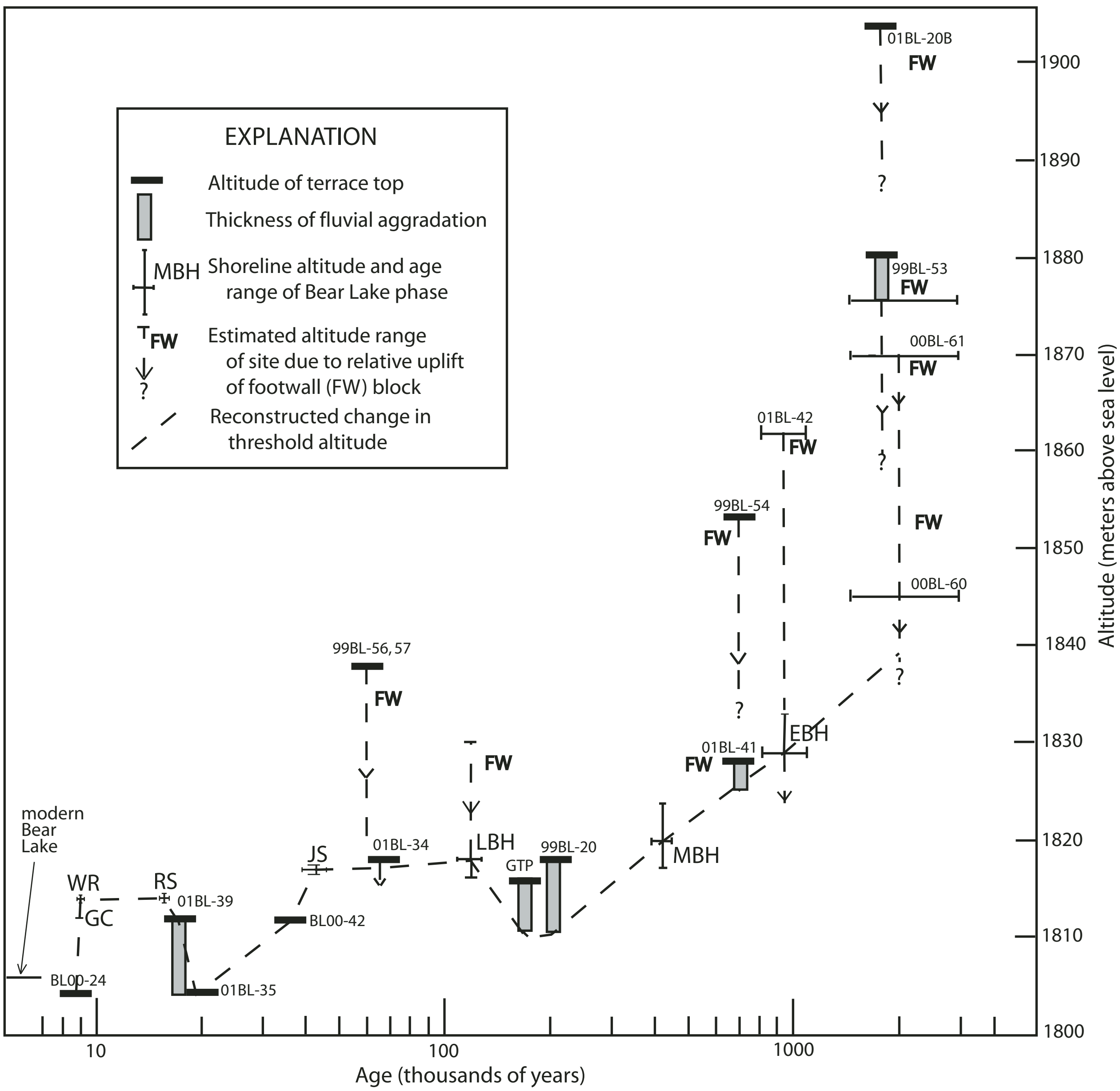

Figure 10. Synthesis of inferred Quaternary incision and aggradation history from lake levels and river terraces; key sites denoted by site name (see Figs. 4 and 7). Large capital letters indicate altitudinal phases of Bear Lake (Laabs, 2001): EBH—early Bear Hollow; GC—Garden City; GTP—Georgetown gravel pit site of Laabs and Kaufman (2003); JS—Jensen Spring; LBH—late Bear Hollow; MBH—middle Bear Hollow; RS—Raspberry Square; WR—Willis Ranch.

et al., 2005).The two sites are separated by a strand of the east Bear Lake fault zone (Reheis, 2005; see Fig. 11B), which may account for their altitude difference.

Early Bear Hollow Phase (Early Pleistocene to Early Middle Pleistocene)

Lacustrine deposits at two sites suggest a highstand of Bear Lake ca. 1200–700 ka. At Bear Hollow, deposits interpreted as fan-delta sand and gravel at 1829 m asl contain aquatic snails with AAR age estimates ranging from 1100 ± 160 ka to 830 ± 180 ka (Figs. 7 and 11A; site BL00-12 of Laabs and Kaufman, 2003). At site 01BL-42 (Figs. 7 and 9), steeply dipping gravel and sand interpreted as delta foreset beds are interbedded with horizontally stratified silt and sand as well as thin marl and diatomite beds at and below 1863 m asl (Reheis et al., 2005). These deposits conformably overlie fan gravel containing tephra correlated with either a Glass Mountain tephra or the Bishop ash (discussed earlier); thus, the lacustrine deposits must be somewhat younger than 1000 or 760 ka and hence are broadly similar in age to those at Bear Hollow. Despite being >30 m different in altitude, both sites are located on the footwall block of the east Bear Lake fault zone. A conspicuous depositional break in slope along the western (relatively unfaulted) margin of Bear Lake Valley lies at ~1830 m asl (Fig. 4); this altitude also coincides with the vertical extent of "lake" deposits mapped in this area by Mansfield (1927) and Robertson (1978). We tentatively interpret the coincidence in altitudes to indicate that the Bear Hollow deposits lie at or near their original depositional altitude; if so, they have experienced much less footwall uplift than sites to the south (discussed in tectonics section).

Middle Bear Hollow Phase (Middle Pleistocene)

Water level probably was as high as 1820 m asl (15 m above modern lake level) at least once ca. 500–300 ka, as shown by lacustrine or wetland marl and beach gravel south of Dingle exposed at Culvert cut (site 99BL-59, Fig. 7) and an adjacent site (01BL-24), and by fine-grained lacustrine deposits west of Bennington (site BL00-13; Laabs and Kaufman, 2003; Reheis et al., 2005). Gastropods at the Culvert cut and Bennington sites yielded AAR age estimates between 445 ± 105 and 385 ± 85 ka. Lacustrine deposits at the Bennington site consist mainly of marl, silt, and mud, indicating deposition below wave base. This site lies north of mapped Quaternary displacement on the east and west Bear Lake fault zones and we infer that it has been relatively unaffected by faulting. Thus the altitude of ~1817 m asl at the top of lacustrine sediments (modified slightly from the altitude given by Laabs and Kaufman, 2003) is a minimum for lake level at this time. South of Dingle, sediments exposed in the Culvert cut represent nearshore deposition. The outcrop near the fault is poor; although the deposits may lie on the hanging-wall block of the east Bear Lake fault zone, they more likely lie between two strands of the fault or are plastered onto the footwall block. Their altitude of ~1821 m asl is a few meters higher than that at the Bennington site. Nearby deposits exposed in a roadcut at site 01BL-24 (Figs. 7 and 9) extend to an altitude of ~1831 m and clearly lie on the footwall block, east of the active fault strands. These deposits include interbedded beach gravel, thinly bedded reddish sand and silt, and unsorted greenish-gray mixed mud and gravel, interpreted to represent nearshore lacustrine and marsh deposition with incursions of colluvium in this fault-marginal setting. Sand above the basal beach gravel contains an assemblage of ostracodes interpreted to represent a lacustrine setting with a nearby stream or wetland (Table 3 *in* Reheis et al., 2005). The reddish sand and silt imply that Bear River was feeding Bear Lake. The deposits in this outcrop can be traced northward to within ~200 m of those exposed in the Culvert cut (Reheis, 2005), suggesting they represent the same lacustrine phase. If so, their separation in altitude indicates at least 10 m of displacement on one fault strand.

Late Bear Hollow Phase (Late Middle to Late Pleistocene)

Lacustrine and marsh deposits of middle to late Pleistocene age crop out in three places in Bear Lake Valley: Bear Hollow, Georgetown gravel pit, and site 00BL-54 (Figs. 7, 8, and 9; latter same as Power line site of Laabs and Kaufman, 2003). AAR estimates of these deposits range from ca. 200 to 100 ka (Laabs and Kaufman, 2003). Lacustrine deposits at site 00BL-54 (Fig. 11B) extend as high as 1864 m asl. Prominent fault scarps to the west, with a combined height of nearly 35 m above the modern lake, indicate that footwall-block uplift has affected this site. Lacustrine silt and sand in this outcrop contain ostracodes endemic to Bear Lake and indicative of nearshore deposition, as well as other ostracodes indicating adjacent streams and wetlands (Reheis et al., 2005). Gastropods from this deposit yielded an AAR age of 151 ± 53 ka (Laabs and Kaufman, 2003). A second set of lake deposits at Bear Hollow (Fig. 11A), much younger than those identified as the early Bear Hollow phase, have AAR age estimates ranging from 180 ± 58 ka to 88 ± 33 ka (Laabs and Kaufman, 2003). As noted above, the Bear Hollow site is on the footwall block. At the Georgetown gravel pit (site BL00-14, Figs. 7, 8), 3 m of marl, bedded sand, and mud overlie 4 m of Bear River gravel and sand; the highest fine-grained deposits lie at ~1817 m asl. AAR age estimates for gastropods are 75 ± 35 and 90 ± 38 ka in the marl and 180 ± 58 ka in the gravel. The fine-grained deposits may be either lacustrine or spring-discharge deposits, but in either case they are interpreted to reflect a lake level at or near 1817 m (Laabs and Kaufman, 2003).

At least one very low lake level, to as low as 1781–1783 m asl (22–24 m below present lake level) may have occurred during the late Bear Hollow phase. Colman (2006) interpreted acoustic stratigraphy of Bear Lake sediments to indicate a wave-cut bench along the western side of the lake, at about the same altitude as the top of a submerged paleodelta in the northwestern part of the lake. Correlations of seismic reflectors to core data (Colman, 2006) suggest an age of somewhat younger than 97 ka for this delta and perhaps also the wave-cut bench. If this delta represents incursion of the Bear River into the lake, it implies that streamflow must have been so small that it did not cause a rise in lake

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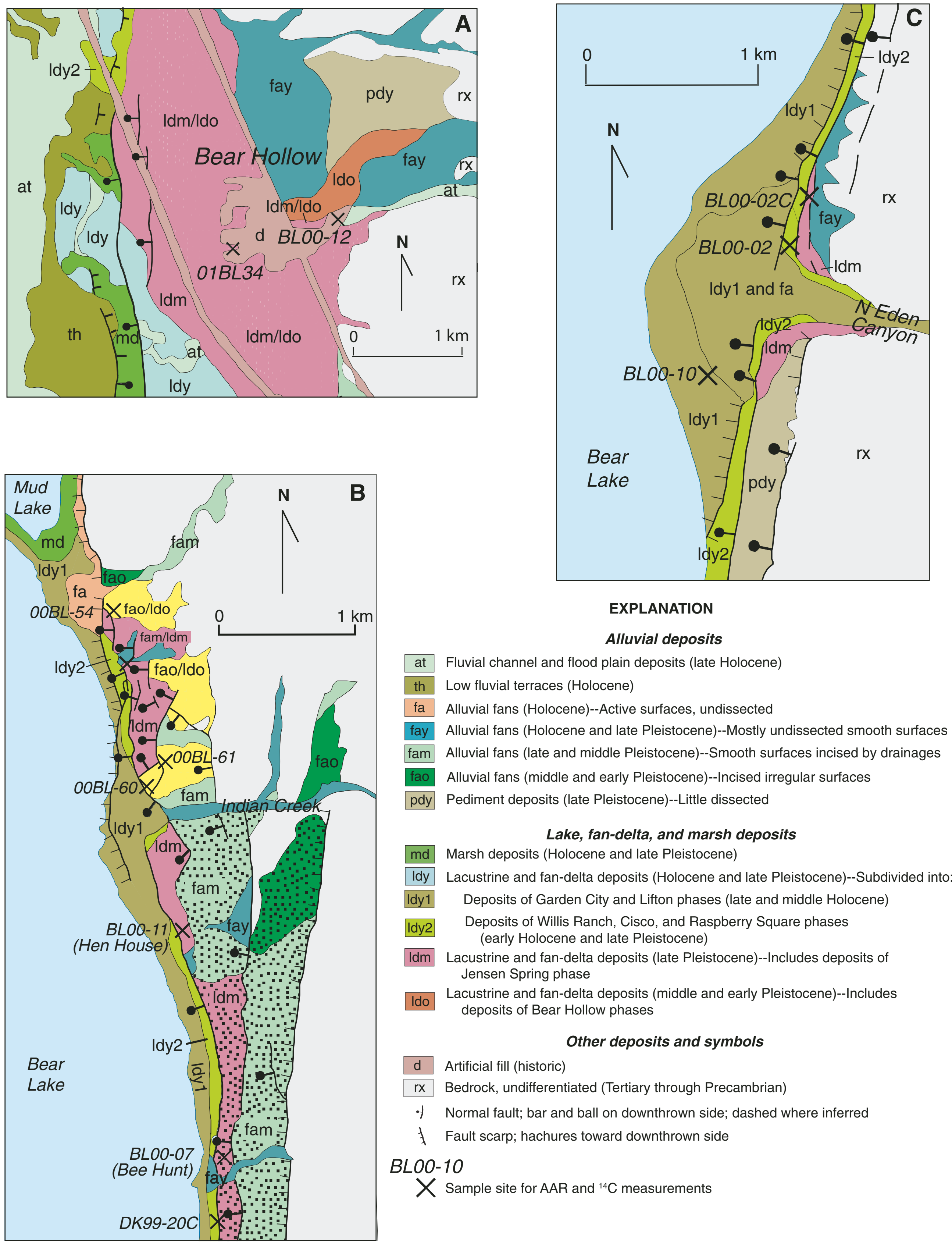

Figure 11. Surficial geologic maps of selected sites along east Bear Lake fault zone (locations on Fig. 4). Stipple indicates alluvial fan or lacustrine deposits partly buried by loess. (A) Surficial geologic map of Bear Hollow, south of Montpelier. (B) Geologic map around Indian Creek, northeastern margin of Bear Lake. (C) Geologic map of North Eden Creek.

level such that Bear Lake expanded to the northern threshold. This period approximately coincides with major soil-forming periods noted by Oviatt et al. (1999) in the Burmester core from Lake Bonneville.

Jensen Spring Phase (Late Pleistocene)

During the Jensen Spring phase, Bear Lake extended up to ~1817 m asl (11 m above modern lake level) between ca. 47 and 39 cal ka (Table 2). Lake deposits of this age crop out at several sites on the footwall blocks of both the east and west Bear Lake fault zones, and also north of active faulting at the Bear River cutbank site (site BL00-42, Fig. 7; Laabs and Kaufman, 2003). At Ovid spit (BL00-23) near Jensen Spring, at ~1817 m asl (11 m above modern lake level), well-sorted, ripple-laminated sand forms a 2-m-high, >400-m-long spit. The eastern edge of Ovid spit is cut by a ~5-m-high scarp of the west Bear Lake fault zone. Because Quaternary slip rates on this fault are low (McCalpin, 2003), the Ovid spit probably has not experienced much footwall uplift. However, its elevation could be as much as 5 m too high, as suggested by deposits at Bear River cutbank site (BL00-42) north of mapped fault scarps (Fig. 7; Reheis, 2005), where lake sands of the same age crop out at ~1811 m asl (Laabs and Kaufman, 2003).

Lacustrine and lake-marginal deposits containing gastropods with similar ^{14}C ages crop out along the footwall block of the east Bear Lake fault zone above the east shore of Bear Lake between 1826 and 1828 m asl (Fig. 11B) (21–23 m above modern lake level; Laabs and Kaufman, 2003). Their higher elevation relative to Ovid spit indicates footwall uplift. One of the deposits (Bee Hunt, Fig. 11B) was analyzed for ostracodes and yielded a rich assemblage indicative of a marginal wetland (J. Bright, 2002, personal commun.). Gypsum-permeated spring deposits as well as beach gravel are associated with the marl at Hen House. Beach gravel also underlies the marl at North Eden Canyon (Fig. 11C).

Raspberry Square Phase (Latest Pleistocene)

No outcrop evidence indicates a lake highstand during the last glacial maximum. A terrestrial megafauna site beneath lake level at the south end of the modern lake suggests that lake level was probably lower than present ca. 22 cal ka (D. Madsen, 2001, personal commun.). Sedimentological evidence from lake cores indicates that the lake level was near the modern level during the last glacial period, ca. 26–18 cal ka (Smoot and Rosenbaum, this volume), when abundant glacial flour was deposited in the lake (Rosenbaum and Heil, this volume). The sedimentological evidence from the cores further indicates that lake level lowered to ~40 m below modern ca. 17.5–15.5 cal ka.

Nearshore sediment exposed in a gravel pit at Garden City (Fig. 4; site location in Laabs and Kaufman, 2003) indicates the lake rose to ~1814 m asl (9 m above present lake level) during the Raspberry Square phase late during the last glacial period, ca. 16–15 cal ka. This altitude is consistent with faulted fan-delta gravels of similar age at North Eden canyon (Figs. 4 and 11C); these deposits lie at ~1824 m asl on the footwall block and 1814 m asl on the hanging-wall block (McCalpin, 1993, 2003). The lake-level curve of Smoot and Rosenbaum (this volume) also shows a highstand centered ca. 15 cal ka. On the basis of their reconstruction, lake level regressed again ca. 14.8–11.8 cal ka. This regression culminated in a drop to ~30 m below modern before a rapid rise to a level above modern ca. 11.5 cal ka. At 11.0 cal ka, salinity in the lake had reached a threshold and aragonite began to precipitate (Dean, this volume).

Holocene Phases

During the Holocene, sedimentological evidence from lake cores indicates that the level of Bear Lake was typically below the modern, but with high-frequency fluctuations of 10–20 m (Smoot and Rosenbaum, this volume). On shore, the Willis Ranch phase (Williams et al., 1962) occurred ca. 9 cal ka, when Bear Lake filled the valley south of Bennington, up to ~8 m above modern lake level (1814 m asl). This timing is consistent with isotopic and other evidence in sediment cores (Liu et al., 1999; Dean et al., 2006; Dean, this volume; Smoot and Rosenbaum, this volume). Two lower and younger shorelines are not well dated, because their deposits largely consist of materials reworked from those of Willis Ranch age (Laabs and Kaufman, 2003). The Garden City shoreline formed at 1811 m asl (Williams et al., 1962) and the Lifton shoreline formed at 1808 m asl (Williams et al., 1962; Robertson, 1978) sometime later.

Synthesis of Downcutting, Aggradation, and Lake Levels

Interpreting the record of aggradation and incision by the river and highstands of Bear Lake (Table 1) is very difficult because of the effects of tectonics on terrace and shoreline altitudes and the paucity of ages on deposits older than ca. 100 ka. Although much further research could be done, a broad outline can be made. Figure 10 presents a synthesis of our current understanding of the interplay between the lake and river during the past million years. We emphasize that there probably were many more fluctuations in lake level, threshold incision, and river aggradation in addition to those we have documented.

The oldest exposed lacustrine sediments in Bear Lake Valley are probably of Pliocene or early Pleistocene age, but their depositional altitude cannot be constrained because they have been identified only on the footwall block of the east Bear Lake fault zone. Only one fluvial terrace (01BL-20B, Figs. 7 and 10) has been identified that is higher in altitude than the possible Pliocene deposits, and this may be due to displacement on presently inactive fault strands between the sites (Reheis, 2005). Lake level must then have fallen and the river established a course separated from the lake, judging from the presence of Bear River gravel overlying lacustrine deposits at site 99BL-53.

During the early Bear Hollow phase, ancestral Bear Lake reached a maximum altitude of 1829 m asl at least once between ca. 1200 and 700 ka. Cross-valley faults near Nounan narrows (Figs. 4 and 7) displace Pliocene deposits (Salt Lake Formation; Oriel and Platt, 1980). If the faults were active during the early

Pleistocene, they may have caused these high lake levels by raising the northern threshold, as suggested by Laabs and Kaufman (2003). Alternatively, if the Bear River bypassed Bear Lake prior to ca. 1.2 Ma, then southward subsidence of the valley floor along the graben-bounding faults followed by capture of the Bear River by Bear Lake due to minor water-level rise or to an avulsion of the Bear River could have caused water level to rise and expand to the valley threshold.

A prolonged period of episodic incision of the threshold to as low as ~1810 m followed the early Bear Hollow phase, with incision possibly persisting as late as 200 ka (Fig. 10). During this interval of ~600 k.y., Bear Lake filled the valley to the threshold at ~1820 m asl at least once during the middle Bear Hollow phase, at 500–300 ka. This altitude is consistent with progressive lowering of the threshold and only requires that the river rejoined the lake to contribute to lake-level rise. Several terrace remnants in the reach downstream of the graben record incision of the threshold to ~1810 m after this lake-filling episode, followed by at least 6 m of aggradation indicated by fluvial sediments at site 99BL-20 (Reheis et al., 2005) and at the Georgetown gravel pit (BL00-14; Laabs and Kaufman, 2003), just upstream of Nounan narrows (Fig. 7). This aggradation occurred ca. 200–180 ka as suggested by an AAR age estimate from shells at the Georgetown site. During this time the Bear River was probably separated from Bear Lake, because gravel containing Uinta Mountain clasts would not have aggraded in northern Bear Lake Valley if Bear River had debouched into Bear Lake.

During the late Bear Hollow phase, water level probably rose to at least 1818 m shortly after ca. 180 ka as recorded by sediments at the Georgetown pit and at Bear Hollow (Figs. 7 and 8). Such an increase in threshold altitude could be explained by one or more of the following: (1) aggradation due to a very large sediment load associated with extensive glaciation during oxygen isotope stage 6 (Blacks Fork glaciation), (2) normal faulting south of Nounan narrows, or (3) landsliding downstream of the Georgetown pit. We have not found extensive terraces of Blacks Fork-equivalent age as would be expected if aggradation were due to glacial sediment loading. Nor have we observed Quaternary deposits displaced by the cross-valley faults near Nounan narrows, or landslide deposits in the reach downstream of the Georgetown pit. The threshold appears to have remained at nearly the same altitude (at or slightly above 1817 m asl) until after the Jensen Spring phase, ca. 47–39 cal ka, suggesting that flow in Bear River was so low that it could not effectively incise a bedrock threshold. Such low flow and possibly a large reduction in ground water input is also suggested by the submerged Bear Lake shoreline at ~1782 m asl (Colman, 2006).

After the Jensen Spring phase, the Bear River began incising its threshold relatively rapidly to essentially its present gradient by ca. 18 cal ka (Fig. 10). Shortly after this time, by ca. 16 cal ka, the Bear River aggraded once again, piling up nearly 8 m of fine sand and silt at site 01BL-39 just south of Bennington (Figs. 8, 9, and 10). The rapid aggradation argues for an abrupt threshold change, possibly due to a landslide in the narrow part of the valley west of Bennington (Laabs and Kaufman, 2003). The lake rose in concert with this aggradation to a highstand at ~1814 m asl by 16–15 cal ka during the Raspberry Square phase. In addition to the temporarily higher sill, the climate was favorable for a highstand during this time of generally cooler and moister conditions in the northern Great Basin and Snake River Plain regions (Thompson et al., 1993). Renewed threshold incision or a marked decrease in river flow caused the north shore of the lake to regress southward and Bear River to separate from Bear Lake by ca. 12 cal ka on the basis of oxygen isotope data in core BL96–2 (Liu et al., 1999; Dean et al., 2006; Dean, this volume) and ca. 12.5 cal ka on the basis of sedimentary indicators of lake level in several cores (Smoot and Rosenbaum, this volume). Then, ca. 11.5 ka, the lake rose once again (Smoot and Rosenbaum, this volume). During the early Holocene, channel migration in the delta area (discussed below) diverted all or part of the Bear River toward the lake, forming the Willis Ranch and Garden City shorelines at 1814 and 1811 m asl, respectively, by sometime after ca. 9 cal ka. Since then, the river has incised to ~1800 m asl west of Bennington (Williams et al., 1962) and to 1790 m asl at Nounan narrows (Fig. 7).

Holocene Marsh Deposits and Migration of the Bear River

Between Bear Lake and the present course of the Bear River is a low-lying region bounded by the east and west Bear Lake fault zones. This region is filled with marshes (including present-day Mud Lake), meandering channels, and scattered surfaces slightly elevated above water level (e.g., the airport area, Fig. 7). Outcrops are rare and generally restricted to artificial cuts along canals. Our observations in this region are limited to these few outcrops, sediments in hand-augered holes, and interpretation of aerial photographs (Fig. 12).

Most of the marsh is at or below the altitude of modern Bear Lake (1805.5 m) and separated from the lake only by the 2-m-high Lifton barrier beach (Figs. 4 and 7). Except for the fan-delta area where the Bear River enters Bear Lake Valley, the marsh must have been mostly submerged during the early Holocene when Bear Lake formed shorelines 6–9 m higher than present. Thus, surficial sediments in this low-lying area mostly record Holocene events, including shifting channels of the Bear River, tributary streams entering the valley from the Bear River Range on the west (St. Charles, Paris, and Ovid Creeks, Fig. 7), and probably the natural outlet of Bear Lake.

Northward Migration of the Bear River

The Bear River presently flows near the northern limit of its low valley floor, beginning to the east of the east Bear Lake fault zone (Fig. 12) and entering Bear Lake Valley where the young fault scarps display an en echelon left step. Analysis of aerial photographs shows that several abandoned channels similar in width to the modern channel lie to the south, and they become increasingly indistinct southward toward Mud Lake. These abandoned channels all appear to emanate from an abandoned course

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Figure 12. Composite aerial photograph showing Bear River (solid white line), inferred paleochannels (BR1, etc., white dashed lines), faults (black lines), and locations of study sites. Although BR1 cannot be traced south of its entry into Dingle Swamp, it likely extended south to the present area of Mud Lake and thereby fed Bear Lake. See Figure 13 for stratigraphy at sites.

of the river that crosses the active fault scarp just north of Dingle. The Rainbow Canal provides a continuous, 5-km-long exposure transverse to the axis of the Bear River and to the older channels, forming a cross-section view of stratigraphy and facies changes descending from the higher part of the fluvial fan toward the marsh (Fig. 13A).

The oldest sediment exposed along the canal and encountered in auger holes elsewhere in northern Bear Lake Valley consists of fluvial sand and pebble-cobble gravel of Bear River. On the basis of topographic contours and outcrops, these deposits form a large braided fluvial fan emanating from the Bear River entrance into Bear Lake Valley. Two ^{14}C ages on shells collected

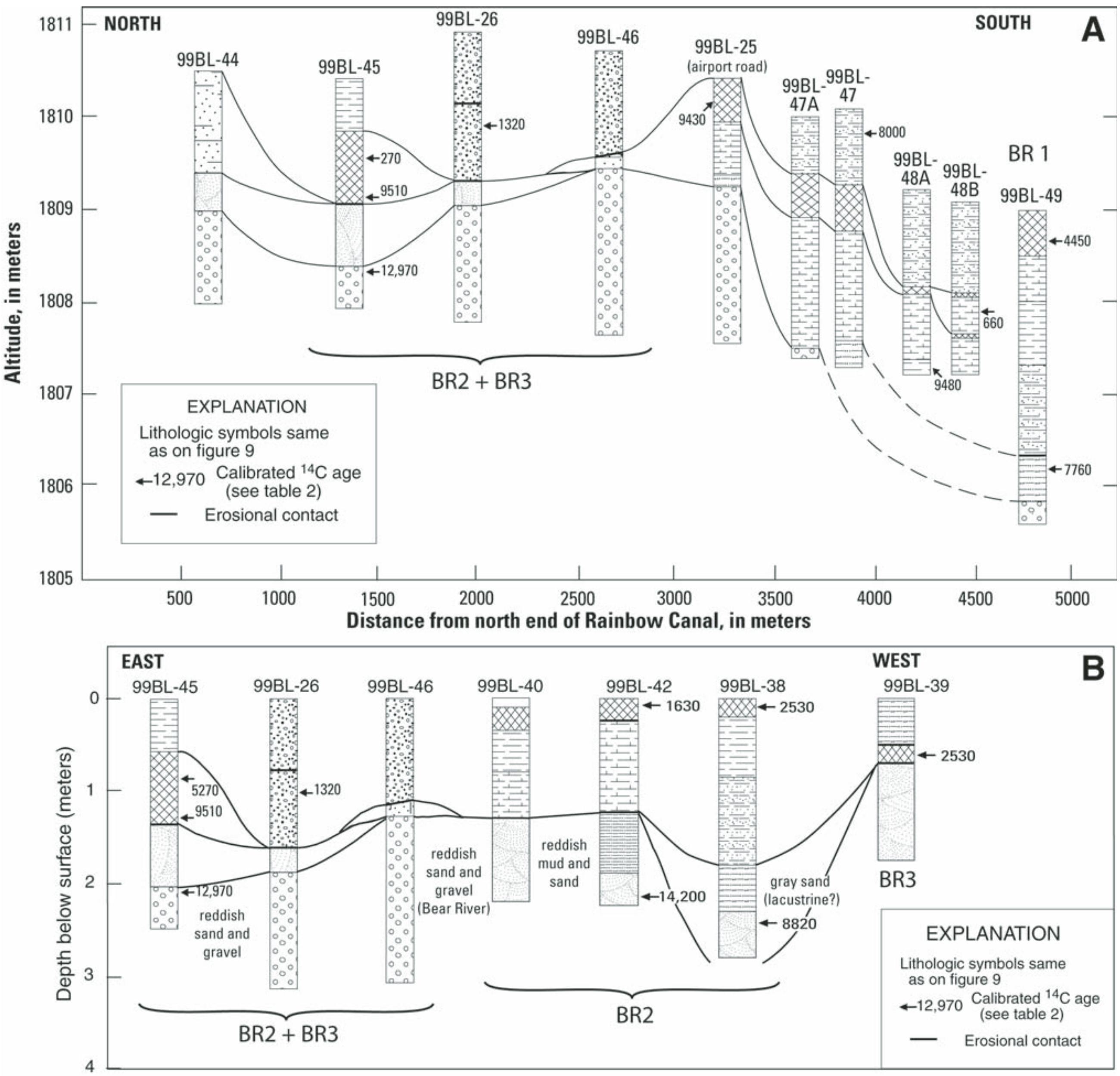

Figure 13. Stratigraphic sections of exposures and auger holes north of Bear Lake (location shown on Figs. 7 and 12). (A) Rainbow Canal. Most sections are outcrops and correlation lines were traced physically, except that 99BL-49 is an auger hole within a channel fill cut into older sediment and correlations are uncertain. Surface altitudes were plotted using differentially corrected GPS measurements. The base of the outcrop sections is the water level; thus the section bases essentially reflect the water gradient in the canal over a two-day period when flow rate in the canal remained relatively constant. Measured sections were then plotted and their altitudes slightly adjusted to yield a smoothly sloping water level at the base of the outcrop sections. BR1, 2, and 3 indicate sections relevant to reconstructing age and location of former courses of Bear River. (B) Selected stratigraphic sections from auger holes west of Rainbow Canal providing age control for BR2 and BR3.

within the gravel and sand at two different sites (99BL-45 and 99BL-42, Figs. 12 and 13 and Table 2) indicate an age of between 14 and 13 cal ka for this part of the fan, so it accumulated after the Raspberry Square highstand of Bear Lake (16–15 cal ka), presumably coincident with a lake-level decrease to below the present lake surface between ca. 15 and 12 cal ka (Smoot and Rosenbaum, this volume). Specific courses of the Bear River have not been identified during this fan-building period.

The river course farthest south, termed BR1, is the most indistinct on the aerial photographs and is crossed diagonally by the Rainbow Canal (Fig. 12). It consists of a meander belt of subdued channels partly or completely filled by younger marsh deposits. The surface trace of this channel cannot be distinguished more than ~0.5 km southwest of the canal, but at its southernmost visible extent, it appears to be directed toward modern Bear Lake. Fluvial deposits, including lenses of pebble gravel, sand, and mud, crop out at the surface along the southern part of the canal and were encountered 1.5 m below the surface in an auger hole within an infilled channel at the south end of the transect. The fluvial deposits overlie a peat bed that thins and fines southward, and the peat overlies and is intercalated with marl formed in a wetland environment (interpreted from microfossils; Reheis et al., 2005); these in turn rest on the 14 to 13 ka Bear River fan gravel. Shells from the peat and marl yielded ages of 9480–8670 cal yr B.P. (Table 2, Fig. 13A) and from the stratigraphically younger fluvial deposits, a single age of 8000 ± 160 cal yr B.P. Thus, the oldest river course is roughly 8.6–8.0 cal ka and overlaps in time with the Willis Ranch highstand of Bear Lake, and with a relatively high lake level as interpreted from sediment cores from Bear Lake (8.5–8.0 cal ka, Smoot and Rosenbaum, this volume). The correspondence in age and the orientation of the meander belt of BR1 suggest that this river course conveyed the discharge that caused Bear Lake to rise to this highstand. However, the Willis Ranch shoreline reaches 1814 m asl on the mostly unfaulted west margin of Bear Lake, an altitude well above the entire length of the Rainbow Canal (Fig. 13A). Because the fluvial deposits of BR1 either lie at the surface or are shallowly buried, these relations suggest that the deposits have been displaced downward since deposition by motion on the east Bear Lake fault zone.

Another meander belt, more sharply defined on aerial photographs, lies between the southern course and the present course of the river (Fig. 12). This middle course is directed due west across the Rainbow Canal and then splits into two main courses, termed BR2 and BR3. BR2 continues straight west, dividing into two courses, both of which appear to merge with north-trending drainages that carried the flow of the west-side creeks (St. Charles and Paris) and probably also the discharge from Bear Lake itself (Williams et al., 1962). BR3 trends northwest from the canal and also bifurcates. Fluvial deposits exposed along the northern part of the canal and encountered in auger holes to the west do not precisely limit the timing of the shifts in river courses (Figs. 12 and 13B), but do suggest that the initial jump north from BR1 occurred no later than 5.3 cal ka (99BL-45, Table 2). The shift from BR2 to BR3 occurred by ca. 2.5 cal ka (99BL-39), and the shift to the modern course of Bear River probably occurred after ca. 1.3 cal ka (sample 99BL-26B).

North-Flowing Axial Channels

The history of the north-flowing channels carrying water from the west-side tributary streams and from Bear Lake is not well defined. Williams et al. (1962) first identified the existence of north-flowing channels or sloughs, which they interpreted as former outlets of Bear Lake, and inferred that discharge had shifted eastward through time. However, Williams et al. (1962) did not recognize the fault-bounded nature of these channels. Several ages on shells from the auger holes of this study and from McCalpin's (1993, 2003) trenching studies on the west Bear Lake fault zone near Bloomington provide rough constraints on the establishment of these drainages following the retreat of Bear Lake in latest Pleistocene time. Due east of Bloomington, ^{14}C ages indicate that a low-energy fluvial channel occupied a north-flowing course sometime after 13 cal ka and was abandoned by ca. 7.4–6.7 cal ka (McCalpin, 2003), a time that McCalpin interpreted as closely dating the most recent earthquake on this part of the west Bear Lake fault zone. On aerial photographs (WAC on Fig. 12, and shown as western channel on Fig. 7), this channel can be traced southward through a complex set of grabens at least as far as St. Charles Creek, and one indistinct remnant south of this creek also suggests that this drainage conducted discharge from the area of present-day Mud Lake and Bear Lake itself (Williams et al., 1962; Reheis, 2005). North of Bloomington Creek at site 99BL-34, this channel is incised into lacustrine deposits thinly overlain by ~50 cm of marsh or possibly overbank fluvial deposits that yielded an age of ca. 8.5 cal ka (Table 2), consistent with McCalpin's (2003) age constraints for this channel. To the north, the same channel (Figs. 7 and 12) continues northward transverse to the modern course of Paris Creek and can be traced nearly to Ovid Creek.

Another north-trending channel parallel to and just east of the channel described above is less confined to the graben structures (EAC on Fig. 12). Auger holes 99BL-37 and -36 within this channel yielded stratigraphic relations and ^{14}C ages that indicated this channel was cut after ca. 12 cal ka and was filled and abandoned by ca. 2 cal ka (Table 2). Because the BR2 and BR3 river courses merge with this eastern north-trending channel, we infer that the channel formed after the western north-trending channel identified near the Bloomington scarp and is coeval with BR2 and BR3, that is, older than ca. 5.3 cal ka to as young as 1.3 cal ka.

Processes Affecting Channel Shifts

Several factors, including climate change, faulting, location of the fluvial gravel fan, and fluvial geomorphic processes may have interacted to produce the Holocene changes in channel position and lake level. BR1 may simply represent an avulsion of Bear River to the south across its post-glacial fluvial fan, and this diversion resulted in the river's merging with the lake, causing the lake to rise to a higher level. Displacement to the south along the east and west Bear Lake fault zones may have encouraged the diversion. On the west Bear Lake fault, two or more events resulting in ~6 m of displacement probably occurred between ca. 12 and 7 ka (McCalpin, 1993, 2003).

The location of the north-flowing, west-side axial channel north of Bear Lake was probably controlled by two factors. First, the Bear River fluvial fan formed a natural dam that was lowest on the west side of the valley (Williams et al., 1962), and the fan was probably higher in elevation before Holocene displacement on the east Bear Lake fault zone. We interpret the stratigraphic and chronologic data and map relations to indicate that northward drainage from the tributary creeks and Bear Lake was established after deposition of the fluvial fan of Bear River and probably just after the land was exposed following the ca. 9 ka Willis Ranch highstand of Bear Lake (as suggested by Williams et al., 1962). This northward drainage followed active graben structures on the west side of the valley.

The former river courses show that the Bear River has migrated northward in the past 8000 years and that it eventually abandoned the west-side axial drainage system. The reasons for this northward shift include (1) gradual channel migration, (2) earthquake events on the east Bear Lake fault zone, and (3) inactivity of the west Bear Lake fault zone since the early Holocene such that graben formation ceased to aid north-flowing drainage. The second possibility is supported by the coincidence of the shift from BR1 to BR2–3 near Dingle with a left step in the fault (Fig. 11). The third possibility is supported by McCalpin's (2003) work showing that the last earthquake event on the Bloomington scarp predates ca. 7 ka.

NEOTECTONICS

Bear Lake Valley resides in the zone of active extension affected by the Yellowstone hotspot (Fig. 1; Pierce and Morgan, 1992), and forms a relatively simple half-graben between the lystric eastern Bear Lake fault zone and the steeply dipping, antithetic western Bear Lake fault zone (Fig. 2; Evans et al., 2003; Colman, 2006). We examine the tectonic geomorphology of range fronts bounding the valley to characterize the pattern of normal faulting in Bear Lake Valley, and use ages of faulted lacustrine and wetland deposits to estimate minimum average slip rates during the past ca. 200,000 years on the east Bear Lake fault zone.

Studies of surficial deposits surrounding Bear Lake have yielded evidence for late Quaternary activity on both the east and west Bear Lake fault zones. McCalpin (1993, 2003) trenched and dated faulted colluvial and lacustrine sediment at the mouth of North Eden canyon where it intersects the east Bear Lake fault zone (Fig. 14A) and calculated a late Quaternary slip rate of 1.1 m k.y.$^{-1}$. McCalpin (1993, 2003) also studied faulted lacustrine, fluvial, and marsh deposits on the west Bear Lake fault zone (Fig. 14D) and estimated a slip rate of ca. 0.5 m k.y.$^{-1}$ since ca. 13 ka; he found that the last earthquake event on this fault zone occurred at or just before ca. 7 ka. Surficial mapping has expanded knowledge of the distribution of fault scarps and youngest displaced deposits along both faults (Reheis, 2005). Seismic profiling along the lake bottom has also revealed evidence of recent faulting (Colman, 2001, 2006; Denny and Colman, 2003).

Tectonic Geomorphology

Several morphological properties of East Bear Lake fault scarps (Figs. 4, 7, and 14) were studied to assess the pattern and relative activity of Quaternary normal faulting in Bear Lake Valley. Fault scarps were examined in the field and on aerial photographs, 1:24,000 U.S. Geological Survey topographic maps, and digital elevation models (DEMs). Mountain front–piedmont slope angle, length, and range-front sinuosity were measured using geographic information software (Laabs, 2001).

Along-strike changes in mountain-front morphology can be used both to identify fault-zone discontinuities (dePolo et al., 1990) and to indicate different ages of faulting (Crone and Haller, 1991). Distinct, along-strike changes in scarp height, slope angle, and orientation were used to divide the east Bear Lake fault zone into fault sections (Fig. 4; modified from segments as defined in Laabs, 2001). The range-front length and strike and the length of the mountain-piedmont junction of each segment were measured where applicable. The sinuosity was calculated as the ratio of the length of the mountain-piedmont junction to the length of the range front (Burbank and Anderson, 2001). We avoided river-eroded reaches while measuring sinuosity.

The morphology of fault scarps in Bear Lake Valley suggests that the east Bear Lake fault zone consists of three sections (Fig. 4) similar to those defined by McCalpin (1993), bounded by changes in fault geometry and recency of motion. The northern section of the east Bear Lake fault zone extends north of Montpelier, Idaho, where scarp strike changes from north-northwest to north-northeast (Fig. 7) and late Quaternary displacement has been minimal or zero. Only two possible short faults cutting late Pleistocene or Holocene deposits have been observed north of Bennington (Fig. 7), and unlike the fault zone to the south, these strike northeast. The central section lies between Montpelier and Indian Creek (Fig. 4). Along this section, several strands of the fault splay off to the north-northeast into bedrock where they show little or no evidence of late Quaternary displacement (Reheis, 2005), whereas the active strands strike north-northwest and bound steep, straight bedrock range fronts or exhibit youthful scarps in late Pleistocene and Holocene deposits (Figs. 14B and 14C). South of Indian Creek, the southern section of the east Bear Lake fault zone strikes north and is essentially confined to a single active trace or to two parallel, closely spaced faults between the range front and Bear Lake (Fig. 14A). Seismic data show additional faults offshore, some forming (paired) grabens (Colman, 2006).

The southern section of the east Bear Lake fault zone and that part of the central section south of Dingle (Figs. 4 and 7) reveal morphological evidence for "maximal" activity based on parameters outlined in McCalpin (1996) for mountain range fronts in semiarid or arid regions. Piedmont landforms include undissected alluvial and debris fans, talus cones, and triangular-faceted ridges (Figs. 14B and 14C). Alluvial fans at the mouths of North and South Eden Canyons are dissected by modern streams and cut by individual fault scarps as much as 14 m high (Fig. 14A; McCalpin, 1993, 2003). Drainage development on the mountain front is incipient, forming shallow V-shaped valleys in bedrock (Fig. 4). The low sinuosity (1.23) and steepness of the Bear Lake Plateau front (30–45°) and fresh scarps that mark surface ruptures suggest that faulting is active (McCalpin, 1996). Finally, a bathymetric low in Bear Lake west of South Eden Canyon suggests that normal faulting on the southern east Bear Lake fault zone has been recently active (Denny and Colman, 2003; Colman, 2006).

The northern section of the east Bear Lake fault zone (Fig. 4) reveals morphological evidence for "minimal" activity according to classifications of McCalpin (1996). Piedmont landforms include deeply dissected alluvial fans near the range fronts and deeply dissected pediments on Tertiary bedrock in the distal zones of alluvial fans (Oriel and Platt, 1980). Exposures of unfaulted alluvial fan gravel between Bennington and Georgetown reveal soils with Bk horizons as much as 2.5 m thick (K and Km horizons). Normal-fault scarps in alluvial fan deposits and pediments are scarce to absent. Alluvium-filled valleys are present in drainages of the Preuss Range, and the mountain front is generally draped by coalesced alluvial fans, suggesting that erosion and deposition have obliterated evidence of fault displacement. The

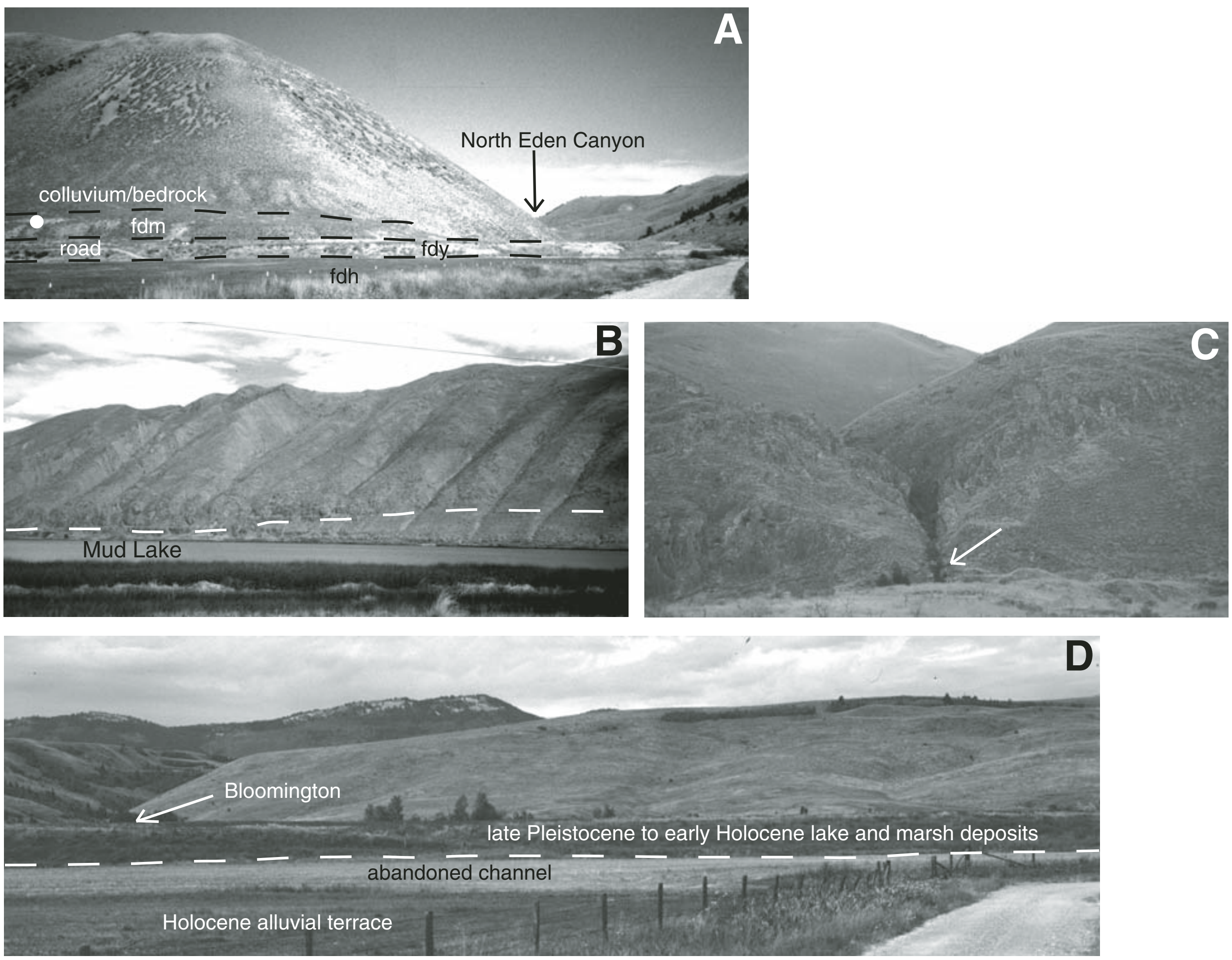

Figure 14. Photographs of fault scarps (see Figs. 4 and 7 for locations). (A) Faults separate fan-delta units at North Eden Creek on southern section of east Bear Lake fault zone; compare with map in Figure 11C. Black dashed lines show approximate locations of fault splays separating units fdm (fan-delta deposit of Jensen Spring phase, 49–37 cal ka), fdy (fan-delta deposit of Raspberry Square phase, 16–15 cal ka), and fdh (fan-delta deposit of Willis Ranch and younger phases, <8.5 cal ka). White dot is location of gastropod samples dated ca. 41 cal ka (Table 2). Fault trenches of McCalpin (2003) located above and below gravel road in lower left. (B) Central section of east Bear Lake fault zone looking east across Mud Lake; dashed white line shows active fault trace. (C) Wineglass canyon and fault scarp on east Bear Lake fault zone adjacent to Mud Lake. Arrow indicates outcrop exposing Holocene alluvial-fan and shoreface deposits faulted against bedrock at site 00BL-37 (Reheis, 2005). (D) View west of fault scarp near Bloomington on west Bear Lake fault zone. Scarp is 6–8 m high; white dashed line marks approximate fault trace. Fault trench of McCalpin (2003) located just off photo to left.

relatively high sinuosity (2.12) and lower slope angles (15–30°) of the Preuss Range front indicate that it has undergone slow or minimal tectonic activity in recent times (McCalpin, 1996).

On the west Bear Lake fault zone, fault scarps in Quaternary deposits are abundant between Ovid and St. Charles (Figs. 4 and 14D). Such scarps have not been identified north of Bern and are rare along the west side of Bear Lake, although fault-line scarps along steep bedrock fronts are present and seismic profiles within the lake (Colman, 2001, 2006; Denny and Colman, 2003) indicate that minor faults displace lake-bottom sediments along the western margin. In the absence of more definite evidence for changes in fault behavior along strike, we do not divide the west Bear Lake fault zone into sections. Previous studies have suggested that the west Bear Lake fault zone extends north of Bern either as several Neogene fault strands that displace Tertiary deposits (Oriel and Platt, 1980) or as a fault with morphologic evidence for recent activity (Laabs, 2001). However, the mountain-piedmont junction in this area is difficult to define and most of it has been eroded by the Bear River. In addition, the gentle eastward dip of deposits of the Miocene-Pliocene Salt Lake Formation beneath pediments east of the river (Oriel and Platt, 1980; Laabs, 2001) implies that displacement along this possible northward extension of the west Bear Lake fault zone must have been very limited (as it is within the lake basin; Colman, 2006); otherwise, the dips would be horizontal or even westward toward the footwall block of the west Bear Lake fault zone.

Slip Rates on the East Bear Lake Fault Zone

We use the water-level history developed above, founded on AAR and ^{14}C geochronology of Laabs and Kaufman (2003) and Reheis et al. (2005), to estimate late Quaternary slip rates on the east Bear Lake fault zone (Table 3). Uplifted fluvial and lacustrine deposits representing Pliocene(?) to middle Pleistocene highstands of Bear Lake on the footwall block of the east Bear Lake fault zone, including terrace gravel on the crest of the Bear Lake Plateau (Fig. 2), provide dramatic evidence of long-term slip, but their counterparts have not been identified in undeformed locations or on the hanging-wall block and thus cannot be used to calculate slip rates. Estimates of depth beneath the lake to certain prominent seismic reflectors, along with the estimated ages of the reflectors and their overlying sediment thicknesses taken from Colman (2006), allow constraints to be placed on the total slip rate and component of footwall uplift between North Eden Creek and Indian Creek (Figs. 4 and 14) during different time intervals since ca. 200 ka.

During the late Bear Hollow phase, water level rose to at least 1818 m asl sometime between ca. 200 and 100 ka. Marl of this age at the Georgetown pit (BL00-14, Fig. 7) lies in a little-deformed area of the valley and is assumed to represent a reference datum. Two other sites of the same lake phase lie on the footwall block of the east Bear Lake fault zone and thus constrain footwall uplift. Fan-delta deposits at Bear Hollow (BL00-12, Figs. 7 and 11A) have a similar age range and lie on the footwall block of the north-central section of the east Bear Lake fault zone at ~1830 m asl. Shorezone deposits at site 00BL-54 (Figs. 7 and 11B) in the south-central section of the fault zone extend as high as 1864 m asl. Fault scarps west of the Bear Hollow and 00BL-54 sites show that displacement is a minimum of 6 m and 35 m, respectively (Table 3, Figs. 11A and 11B). We estimate the component of footwall uplift as 7–17 m to the north at Bear Hollow and 41–51 m at site 00BL-54 in the past 100–200 k.y., yielding uplift rates of 0.04–0.15 and 0.23–0.51 m k.y.$^{-1}$, respectively. Total slip rates can be estimated assuming that footwall uplift has been either 20% of total fault slip (Stein and Barrientos, 1985; Demsey, 1987; Morley, 1995), yielding minimum slip rates, or 10% of total slip, yielding maximum slip rates (Jackson and McKenzie, 1983). These calculations yield minimum slip

TABLE 3. COMPARISON OF SLIP RATES ALONG THE EAST BEAR LAKE FAULT ZONE (SEE FIGS. 4, 7, AND 11 FOR LOCATIONS)

Site number or location	Age range (ka)	Footwall uplift (m)	Footwall uplift rate (m/k.y.)	Max. total slip (m)	Max. slip rate (m/k.y.)	Min. total slip (m)	Min. slip rate (m/k.y.)
Central section of fault zone							
BL00-12 (Bear Hollow)	200–100	7–17	0.04–0.15	70–170*	0.4–1.5*	35–85†	0.2–0.8†
98BL-11 (Dingle)	47–39	N.D.	N.D.	≥6	≥0.12–0.15	≥6	≥0.12–0.15
North of 98BL-11	16–15	N.D.	N.D.	≥4	≥0.25	≥4	≥0.25
Central-southern border area of fault zone							
00BL-54 (North of Indian Creek)	200–100	41–51	0.2–0.5	410–510*	2.0–5.1*	205–255†	1.0–2.6†
00BL-54 and seismic reflector R7	200–100	41–51	0.2–0.5	110‡	1.1‡	60‡	0.3‡
Southern section of fault zone							
BL00-07, DK99-20C, BL00-02C (Hen House, Bee Hunt, North Eden)	47–39	9–12	0.2–0.3	90–120*	1.9–3.1*	45–60†	1.0–1.5†
BL00-07 (Hen House) and seismic reflector R3	47–35	9	0.2–0.3	65‡	1.9‡	45‡	1.0‡
BL00-02C (North Eden) and seismic reflector R3	47–35	12	0.3	91‡	2.6‡	31‡	0.8‡
North Eden (McCalpin, 2003)	47–39	N.D.	N.D.	>22.9	>1.4–1.5	>22.9	>1.4–1.5
North Eden (McCalpin, 2003)	16–15	N.D.	N.D.	10.5	0.7	10.5	0.7

*Assumes footwall uplift is ~10% of magnitude of hanging-wall subsidence (Jackson and McKenzie, 1983).

†Assumes footwall uplift is ~20% of magnitude of hanging-wall subsidence (Stein and Barrientos, 1985; Demsey, 1987; Morley, 1995).

‡Maximum slip and slip rates calculated using elevation of site above lake level plus estimated modern water depth offshore plus estimated thickness of sediment above marker seismic horizon of similar age (R3, about 35 ka; R7, about 97 ka; seismic data and horizon age from Colman et al., this volume). Minimum slip and slip rates exclude modern water depth. See discussion in text. N.D.—no data.

rates of 0.2–0.8 m k.y.$^{-1}$ and maximum rates of 0.4–1.5 m k.y.$^{-1}$ for the north-central section of the east Bear Lake fault zone, and minimum rates of 1.0–2.6 m k.y.$^{-1}$ and maximum rates of 2.0–5.1 m k.y.$^{-1}$ for the south-central section of the fault zone.

During the Jensen Spring phase, ca. 47–39 ka based on ^{14}C ages, Bear Lake attained an altitude of ~1817 m (Table 1). At site 98BL-11 south of Dingle (Fig. 7), fan-delta deposits correlated with the Jensen Spring phase on the basis of soil morphology lie at ~1817 m asl and are cut by a 6-m-high fault scarp; assuming a similar age range for these deposits yields a minimum slip rate of 0.12–0.15 m k.y.$^{-1}$ for one strand of the central section of the fault zone. Deposits of the Jensen Spring phase at three localities on the footwall block of the southern section of the east Bear Lake fault zone—Bee Hunt, Hen House, and North Eden Canyon (Figs. 11B and 11C)—lie above the east shore of Bear Lake between 1826 and 1829 m asl (Laabs and Kaufman, 2003; Reheis, 2005). Their higher elevations represent footwall uplift of ~9–12 m (increasing to the south and highest at North Eden Canyon) since ca. 47–39 ka, and yield uplift rates of 0.19–0.31 m k.y.$^{-1}$ (Table 3). Again assuming this is either 20% (Stein and Barrientos, 1985; Demsey, 1987; Morley, 1995) or 10% (Jackson and McKenzie, 1983) of the total displacement, estimated total slip rates are 1.9–3.1 m k.y.$^{-1}$ (maximum) and 1.0–1.5 m k.y.$^{-1}$ (minimum) along the southern section of the fault zone, increasing to the south. These slip rates are somewhat higher than, but entirely consistent with, a minimum slip rate of ~1.4–1.5 m k.y.$^{-1}$ calculated from the minimum displacement of beach gravel of the Jensen Spring phase exposed in fault trenches at North Eden Canyon (McCalpin, 1993, 2003). Slip rates estimated from fault trenches at North Eden Canyon are minimum values because additional faults lie offshore, and because the Jensen Spring-equivalent beach deposits were not exposed in the fault trench west of the western fault trace.

During the Raspberry Square phase, ca. 16–15 cal ka, interpreted nearshore deposits at Garden City on the west shore (Fig. 4) indicate that Bear Lake rose to ~1814 m asl (Table 1). Faulted fan-delta gravels of similar age at North Eden Canyon on the east shore (Fig. 11C) at an altitude of ~1815 m are inset below a fault scarp below the older Jensen Spring-equivalent deposits; these inset deposits have been displaced a cumulative 10.5 m across two fault strands (McCalpin, 1993, 2003). Fan-delta deposits to the north at the mouth of Bear River between Dingle and site 98BL-11 are mapped as correlative with the Raspberry Square phase on the basis of soil morphology (Reheis, 2005) and are cut by a 4-m-high fault scarp. These displacements yield slip rates of 0.7 m k.y.$^{-1}$ and 0.25 m k.y.$^{-1}$, respectively, for the southern and central sections of the east Bear Lake fault zone (Table 3).

To estimate total slip rates and footwall-uplift rates for the east Bear Lake fault zone near Indian Creek and at North Eden Creek, we can compare slip-rate estimates based on an assumed ratio of footwall uplift to total displacement to those made by combining altitudes of footwall shoreline sites with seismic data of Colman (2006) (Table 3, Fig. 15). Colman (2006) identified and mapped prominent seismic reflectors and their overlying sediment thicknesses beneath Bear Lake, and correlated these markers with sediments in core BL00-1E. The estimated ages of two of these reflectors are 35 ka for R3 and 97 ka for R7. Assuming that R3 represents a time near the end of the Jensen Spring phase (47–39 ka) and that R7 represents a time near the end of the late Bear Hollow phase (200–100 ka) allows us to calculate maximum total slip by summing (1) the altitude of similar-aged deposits above modern lake level, (2) modern water depth offshore, and (3) depth to the seismic reflector. The calculation yields maximum total slip because the seismic-reflector strata must have been deposited at some water depth; hence the use of modern water depth is an overestimate for that part of the hanging-wall slip not accounted for by sediment thickness. Minimum slip estimates exclude modern water depth. For site BL00-7 (Hen House) just south of Indian Creek (Fig. 11B), these calculations yield minimum and maximum slip rates of 1.0 and 1.9 m k.y.$^{-1}$, and for site BL00-02C at North Eden Canyon, 0.8 and 2.6 m k.y.$^{-1}$. Comparing maximum total displacement with footwall uplift at these sites (Table 3) indicates that footwall uplift has been a minimum of 13% of total displacement since ca. 50 ka, a result that supports the assumed 10%–20% range for the footwall-uplift component.

Applying the same method to reflector R7 and site 00BL-54 (200–100 ka) north of Indian Creek yields minimum and maximum slip rates of 0.3 and 1.1 m k.y.$^{-1}$ and a value of 35%–45% footwall uplift relative to maximum total displacement (Table 3). Although the slip rates seem reasonable, the proportion of footwall uplift is much larger than the 13% estimated for younger deposits. The discrepancy could be caused by three factors: (1) the age of deposits at 00BL-54 is not the same as the age estimated for R7; (2) the depth to R7, here estimated by extrapolating incomplete isopachs of sediment thickness northward to Indian Creek from Figure 6 of Colman (2006), is incorrect; or (3) the estimated proportion of footwall uplift to total slip is incorrect over the longer time scale. For example, in a study of the Borah Peak, Idaho, earthquake, Stein and Barrientos (1985) pointed out that measured footwall-to-hanging-wall displacement (1:5) caused by the seismic event was similar to that indicated by geologic data during the past ca. 15,000 years but the cumulative Pliocene displacement across the fault suggested that interseismic deformation increased uplift to approximately equal subsidence.

Discussion of Neotectonic Effects

The presence of steeper and straighter mountain fronts to the south along the east Bear Lake fault zone (Fig. 4), along with greater slip rates on the central and southern sections of the fault relative to the northern section (Table 3), means either that the focus of normal faulting has shifted southward with time, or that slip rates have always been higher to the south. The northern section of the east Bear Lake fault zone was probably more active during the Pliocene than later. In this area, Quaternary pediment gravel lying on Tertiary mudstone and tuff of the Salt Lake Formation has not been faulted and dips gently westward toward the valley floor (Oriel and Platt, 1980; Laabs, 2001). In addition, the

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subdued, sinuous morphology of the Preuss Range front suggests that the north segment of the east Bear Lake fault zone has had minimal tectonic activity during the Quaternary. In contrast, the east-dipping Salt Lake Formation strata suggest that this part of the fault was active in the Tertiary. The distribution of fault scarps (Reheis, 2005) and the morphology of range fronts suggest that the southward increase in slip rate coincides with confinement of the fault zone to fewer, more closely spaced faults. The morphology of the southern section of the east Bear Lake fault zone, along with seismic data of Colman (2006) and results of McCalpin (1993, 2003), indicates that it is the focus of most recent normal faulting in Bear Lake Valley. Our interpretation of a southward shift in fault activity since Pliocene time is similar to observations of Anders et al. (1989) on the Grand Valley and Star Valley faults to the northeast of Bear Lake (Fig. 1) and to general conclusions about patterns of fault activity associated with the track of the Yellowstone hotspot (Pierce and Morgan, 1992).

Slip rates on the southern section of the east Bear Lake fault zone have apparently decreased over the past 50 k.y. (Table 3), from at least 1.0 m k.y.$^{-1}$ since 50 ka to ~0.7 m k.y.$^{-1}$ since 15 ka (data of McCalpin, 2003). McCalpin (1993) suggested several hypotheses to account for apparently lower slip rates during the past 15 k.y. than the long-term, late Cenozoic rates on the east Bear Lake fault zone. These include (1) a seismic cycle greater than 15 k.y., (2) an earlier time of fault initiation, and (3) a possible "slip deficit" caused by some factor that has restrained slip since 15 ka. We cannot assess these hypotheses independently, but suggest that slip could have been restrained during the past 15 k.y. by relatively low lake levels that reduced water loading in the valley.

CONCLUSIONS

Our studies of the surficial geology, geomorphology, and neotectonics shed light on the complex interactions that drive the relationship between the Bear River and Bear Lake. The chemistry of water and the lithology of sediments in Bear Lake largely reflect whether Bear River was merged with Bear Lake to form a flow-through, river-dominated lake, or was separated from Bear Lake, resulting in a topographically closed lake fed largely by groundwater flow. Whether the river and lake are merged depends on faulting, fluvial geomorphic processes, changes in threshold altitude, and changes in paleoclimate.

An ancestral Bear River course, probably of Pliocene age, that lacks Uinta Range–derived clasts is preserved atop the Bear River Plateau east of Bear Lake. This ancestral river likely began

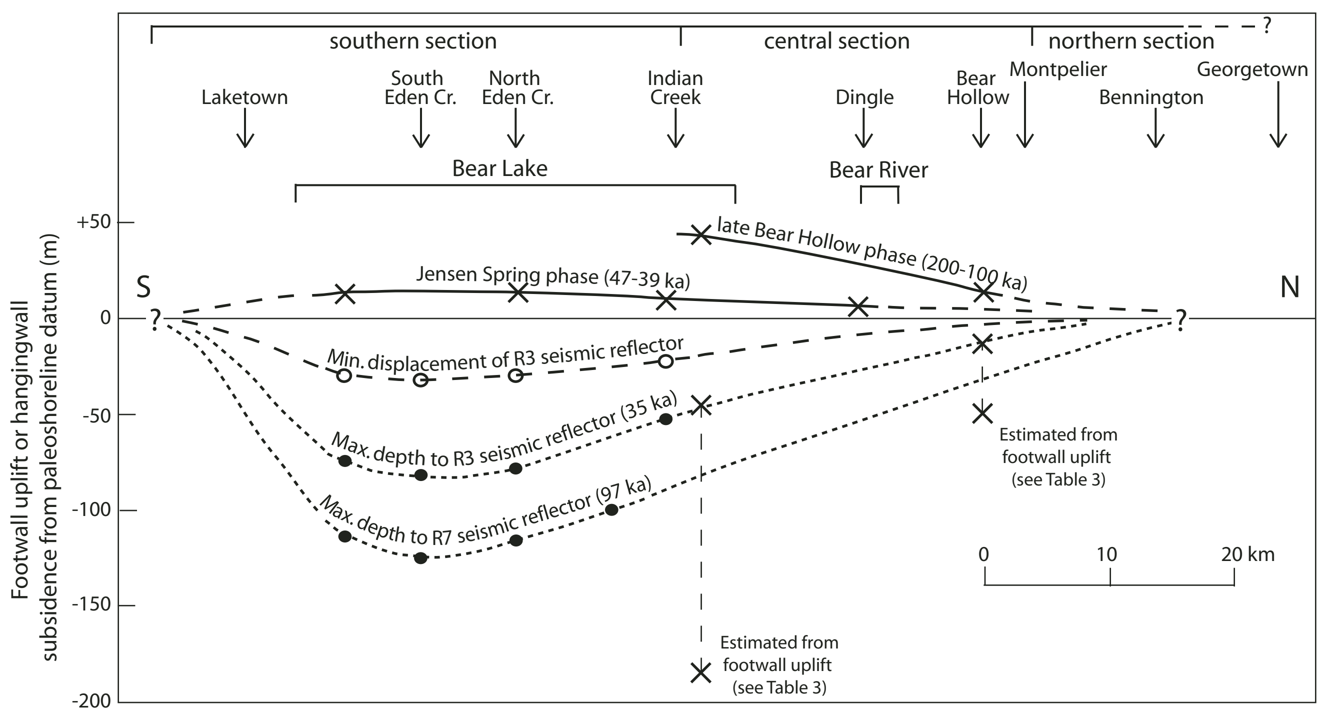

Figure 15. Diagram showing relative displacement of footwall (positive) and hanging-wall (negative) blocks along east Bear Lake fault zone. Footwall sites (X's) based on sites discussed in text (Figs. 3 and 7) and in Reheis (2005). Hanging-wall sites of late Bear Hollow phase (X's; 200–100 ka) north of Indian Creek and at Bear Hollow estimated from footwall uplift:subsidence ratios of 1:5 (maximum) and 1:1 (minimum) (Stein and Barrientos, 1985). Other hanging-wall sites derived from seismic data of Colman (2006). Maximum estimates of hanging-wall subsidence (filled circles) obtained by summing (1) water depth, (2) thickness of sediment overlying seismic reflector, and (3) difference between modern lake level (1805 m asl) and reconstructed lake highstand during Jensen Spring and late Bear Hollow phases (Fig. 10). Minimum estimates (open circles) exclude water depth. Minimum estimates not shown for late Bear Hollow phase (R7 reflector) for clarity, because those points generally fall on the same line as that of maximum depth to the R3 reflector.

to incise and migrate eastward in response to uplift along the east Bear Lake fault zone at the same time that subsidence of the hanging-wall block began to create an accommodation space that became Bear Lake. Capture of the modern headwaters in the Uinta Range, which were formerly tributary to the Green River, added considerable volume to this ancestral Bear River. The subsequent diversion of the Bear River into Bear Lake Valley and the beginning of lacustrine deposition may have been concurrent.

Our studies have established the following broad framework for the record of aggradation and incision by the river and highstands of Bear Lake. Reconstruction of this record is complicated by the differential effects of tectonics on terrace and shoreline altitudes and the paucity of ages on deposits older than ca. 100 ka. The oldest exposed lacustrine sediments in Bear Lake Valley are probably of Pliocene or early Pleistocene age, but their depositional altitude cannot be constrained. Lake level must then have fallen and the river established a course separated from the lake, as indicated by Bear River gravel overlying lacustrine deposits.

During the early Bear Hollow phase, ancestral Bear Lake reached a maximum altitude of 1829 m asl at least once between ca. 1200 and 700 ka, constrained by a threshold at that altitude near Nounan narrows. A prolonged, ~600 k.y. period of episodic incision of the threshold to as low as ~1810 m asl followed this phase, with incision possibly persisting as late as 200 ka. Bear Lake filled the valley to the threshold at ~1820 m asl at least once during the middle Bear Hollow phase, 500–300 ka. This altitude is consistent with progressive lowering of the threshold and only requires that the river merged with the lake to cause the lake to rise. Several terrace remnants in the reach downstream of the graben record incision of the threshold to ~1810 m asl after this lake-filling episode, followed by at least 6 m of aggradation just upstream of Nounan narrows by ca. 180 ka. During this time the Bear River was probably separated from Bear Lake.

During the late Bear Hollow phase, water level rose to at least 1818 m asl shortly after ca. 180 ka. Such an increase in threshold altitude could be explained by one or more of the following: (1) aggradation due to a very large sediment load associated with extensive glaciation during oxygen isotope stage 6, (2) normal faulting to the south, or (3) landsliding upstream of Nounan narrows. This lake phase probably coincided with the Blacks Fork glaciation in the Uinta Range; correlative moraines in the Bear River Range west of the lake indicate that valley glaciers there were as much as 6–7 km long. The lake threshold appears to have remained at nearly the same altitude (at or slightly above 1817 m asl) until the Jensen Spring phase, ca. 47–39 cal ka, suggesting that flow in Bear River was so low that it could not effectively incise a bedrock threshold.

After the Jensen Spring phase, the Bear River incised its threshold relatively rapidly to essentially its present gradient by ca. 18 cal ka, but by ca. 16 cal ka had aggraded 8 m of sediment again. The rapid aggradation may have been caused by a landslide in a narrow part of the valley west of Bennington (Laabs and Kaufman, 2003). The lake simultaneously rose to a highstand at ~1814 m asl by 16–15 cal ka during the Raspberry Square phase. This period coincided with the start of the last deglaciation in the Uinta Range and likely the Bear River Range. Renewed threshold incision or a marked decrease in river flow caused the north shore of the lake to regress southward and Bear River to separate from Bear Lake by ca. 12 cal ka (Dean et al., 2006; Dean, this volume), but the lake once again rose during the early Holocene when channel migration in the delta area diverted all or part of the Bear River toward the lake. Since then, the river has incised to ~1790 m asl at Nounan narrows.

Dating of sediments within the former river courses visible in aerial photography of the extensive modern marsh between Bear Lake and the Bear River shows that the Bear River has migrated northward during the past 8000 years. In addition, two west-side axial drainages, partly controlled by grabens, were eventually abandoned; these drainages had previously carried streamflow northward from tributaries and from Bear Lake. The causes of the northward shift and abandonment of the west-side drainages include gradual channel migration, earthquake events on the east Bear Lake fault zone, and inactivity of the west Bear Lake fault zone since the early Holocene such that graben formation ceased to aid north-flowing drainage.

We interpret the presence of steeper and straighter mountain fronts to the south along the east Bear Lake fault zone, along with greater slip rates on the central and southern sections of the fault relative to the northern section, to indicate either that the focus of normal faulting has shifted southward with time, or less likely, that slip rates have always been higher to the south. The distribution of fault scarps and the morphology of range fronts suggest that the southward increase in slip rate coincides with confinement of the fault zone to fewer, more closely spaced faults. Slip rates on the southern section of the east Bear Lake fault zone have apparently decreased over the past 50 k.y., from at least 1.0 m k.y.$^{-1}$ since 50 ka to ~0.7 m k.y.$^{-1}$ since 15 ka (McCalpin, 2003).

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