

Track of the Yellowstone Hot Spot: Young and On-Going Geologic Processes from the Snake River Plain to Yellowstone Plateau

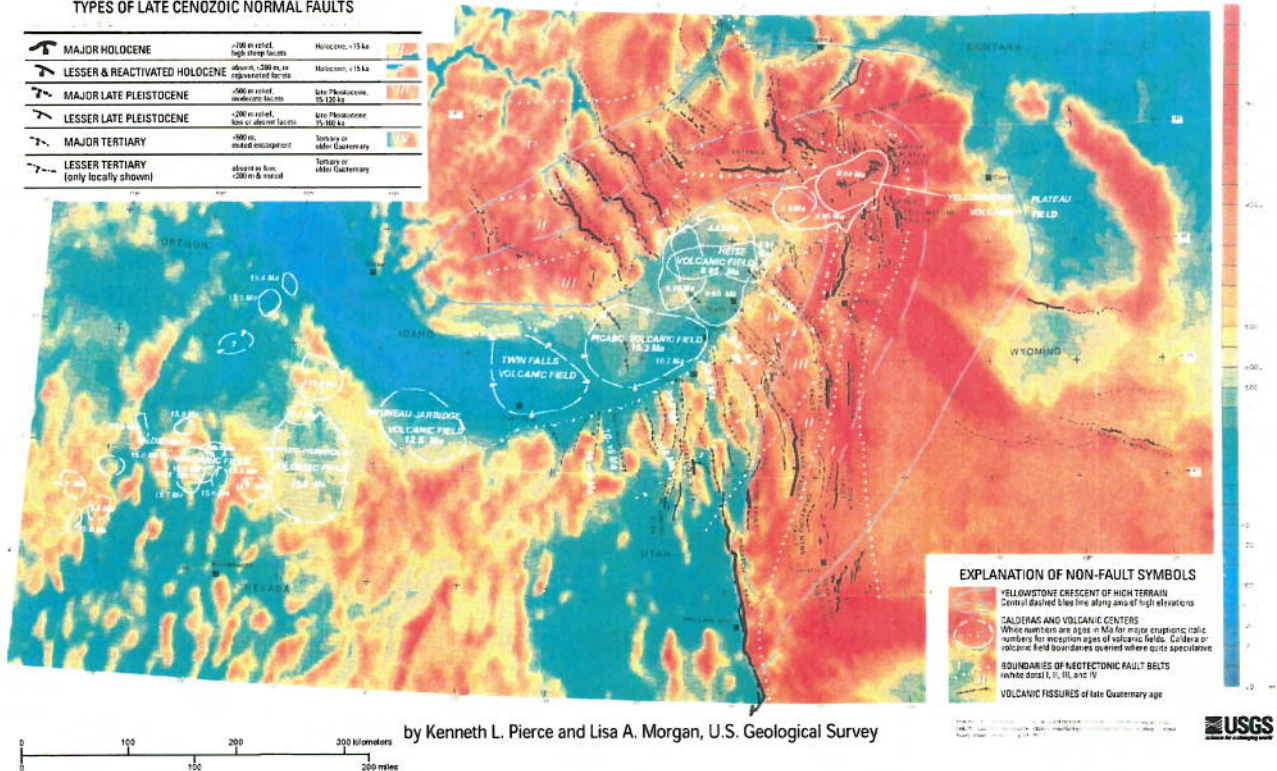
Geological Society of America Field Guide, October 23-27, 2007

L.A. Morgan¹, K.L. Pierce², and W.C. Pat Shanks¹

THE TRACK OF THE YELLOWSTONE HOTSPOT: VOLCANISM, FAULTING, AND UPLIFT

TYPES OF LATE CENOZOIC NORMAL FAULTS

MAJOR HOLOCENE	>100 m relief, high steep faults	Holocene, <15 ka
LESSER & REACTIVATED HOLOCENE	<100 m relief, moderate to steep	Holocene, <15 ka
MAJOR LATE PLEISTOCENE	>100 m relief, moderate to steep	late Pleistocene, 50-100 ka
LESSER LATE PLEISTOCENE	<100 m relief, low to moderate slopes	late Pleistocene, 50-100 ka
MAJOR TERTIARY	>100 m relief, moderate to steep	Tertiary or older Quaternary
LESSER TERTIARY (only locally shown)	<100 m relief, low to moderate slopes	Tertiary or older Quaternary



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The 16-Ma Yellowstone hot spot track is one of the few places on Earth where a time-transgressive suite of processes on continental crust can be seen in the volcanic and tectonic (faulting and uplift) record at the rate and direction as predicted by plate motion. Recent interest in young and possible renewed volcanism at Yellowstone along with new discoveries and synthesis of previous studies (i.e., tomographic, deformation, bathymetric, and seismic surveys) are fueling discussion as to Yellowstone's possible plume origin. This field trip will highlight various stages in the evolution of the Snake River Plain-Yellowstone Plateau bimodal volcanic province, also known as the track of the Yellowstone hot spot.

Field trip stops will include the young basaltic Craters of the Moon, exposures of 12-4-Ma rhyolites and edges of their collapsed calderas on the Snake River Plain, and faulting progressing with the volcanic fields. An essential stop is Yellowstone National Park, where the last major caldera-forming event occurred 640,000 years ago and now is host to the world's largest hydrothermal field. This 3-day trip will present a quick, intensive overview into volcanism and tectonism in this dynamically active region. We will discuss origin models and examine features that may be reflective of a tilted thermal mantle plume as indicated in recent tomographic studies. Our drive home will pass through Grand Teton National Park, where the Teton Range is currently rising in response to the passage of the North America plate over the Yellowstone hot spot.

Overview

The Snake River Plain-Yellowstone Plateau (SRP-YP) volcanic province is one of the best examples of bimodal (basalt and rhyolite) volcanism on Earth. For decades, until the 1970's most people thought of the Snake River Plain as a basaltic province due to its extensive cover of basaltic lava flows, shield volcanoes, fissure vents, rift systems, and cinder cones (eg., Greeley, 1982; King, 1982). While early workers in this area such as Mansfield and Ross (1935) recognized the large sheet exposures of very densely welded ignimbrites, occasional rhyolitic lavas and domes, and fall deposits, few recognized that the basalts were related to the rhyolites. On the Yellowstone Plateau, early explorers and scientists recognized the volcanic terrain on the plateau as dominantly rhyolitic with subordinate amounts of basalt (refer to quote from Albert Peale, U.S. Geological Survey, Hayden survey, August 6, 1871, Merrill, 1999). No one linked the volcanism, especially the rhyolitic, on the Yellowstone Plateau with earlier volcanism (rhyolitic and basaltic) on the Snake River Plain.

In the early 1970's, Jason Morgan (1972) first postulated that the SRP-YP volcanic province represented a hotspot track and later Armstrong et al. (1975) showed that the initiation of rhyolitic and later basaltic volcanism along this track becomes systematically younger northeastward towards the Yellowstone Plateau. A vigorous debate has ensued as to whether or not this volcanic progression results from a mantle plume beneath the southwest-moving North American plate. Arguments in favor of a mantle plume origin can be found in Suppe et al., 1975; Anders et al., 1989; Westaway, 1989; Draper, 1991; Pierce and Morgan, 1992; Smith and Braile, 1993; Camp, 1995; Takahashi et al., 1998; Pierce et al., 2002; Camp and Ross, 2004. Arguments not supportive of a mantle origin can be found in Hamilton, 2003; Humphreys et al., 2000; and Christiansen et al., 2002.

The track of the Yellowstone hotspot began ~16 Ma with eruption of Columbia River Plateau and associated flood basalts in Oregon and Washington and extensive rhyolite volcanism and mafic dikes in Nevada and Oregon over an N-S distance of ~1,000 km (Figure 1) (Pierce and Morgan, 1992). A large-diameter plume head (Draper, 1991; Camp, 1995; Pierce et al., 2002) at the base of a lithosphere with a variety of compositions depending on location may explain the widespread volcanism and intrusions. Basaltic magma and associated heat were produced by decompression melting. The Columbia River and related basalts were emplaced through thin Mesozoic oceanic crust, whereas the 14-17-Ma rhyolites were generated above thicker and more silicic crust (Draper, 1991; Pierce et al., 2002; Camp and Ross, 2004). The basalt versus rhyolite dichotomy can be attributed to these contrasting crustal compositions that are delineated by the $^{87}\text{Sr}/^{86}\text{Sr}$ 0.706 and 0.704 contours (Figure 1).

Volcanism. Caldera-forming events on the SRP-YP volcanic province produce some of the largest eruptions known on Earth and range in volume from several hundred cubic kilometers to greater than 2500 km³. The regional extent and character of volcanism changes from the initial widespread extent at 16 Ma to a more focused area of volcanism about 10 Ma and may represent a radical change in plume dimension and development as well as changes in crustal composition. Prior to 10 Ma, volcanism was widespread, had high eruption temperatures, and produced more frequent eruptions. Around 10 Ma, the track narrows to an area <100 km wide.

This plume “head” to plume “tail” transition may be explained as a more-evolved, steady state phase of the plume which we refer to as a plume “tail” with a diameter of ~100 km. The track is represented by a series of volcanic fields, each spanning a couple million years and characterized by multiple caldera-forming, large ignimbrite eruptions. A 1-2-million-

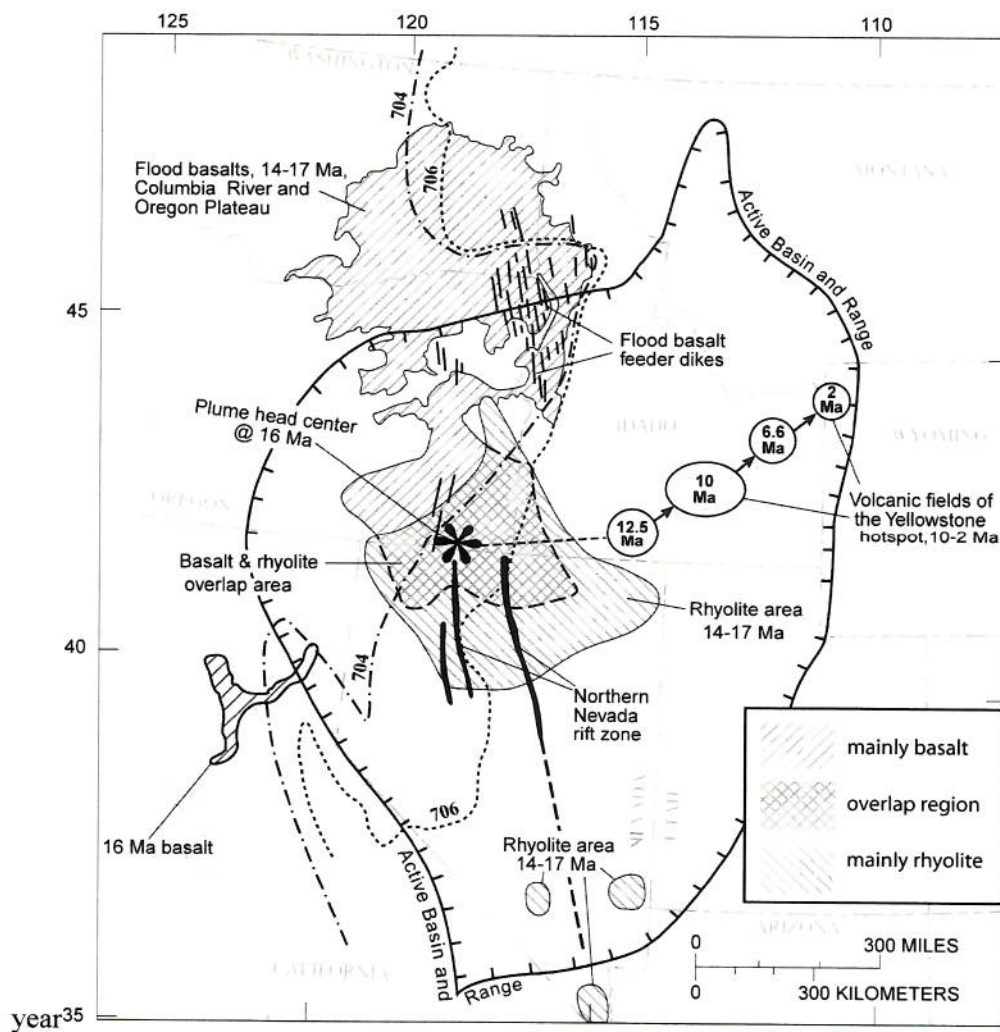


FIGURE 1.

hiatus in caldera-forming events and up to 50-100 km separate adjacent volcanic fields (Plate 1). The initial caldera-forming ignimbrite eruption within each field defines a systematic progression that has advanced north-northeast to Yellowstone at the same rate (25 km/m.y) and in the opposite direction as motion of the North American plate. Within a particular volcanic field, no obvious age progression is apparent for location of a caldera.

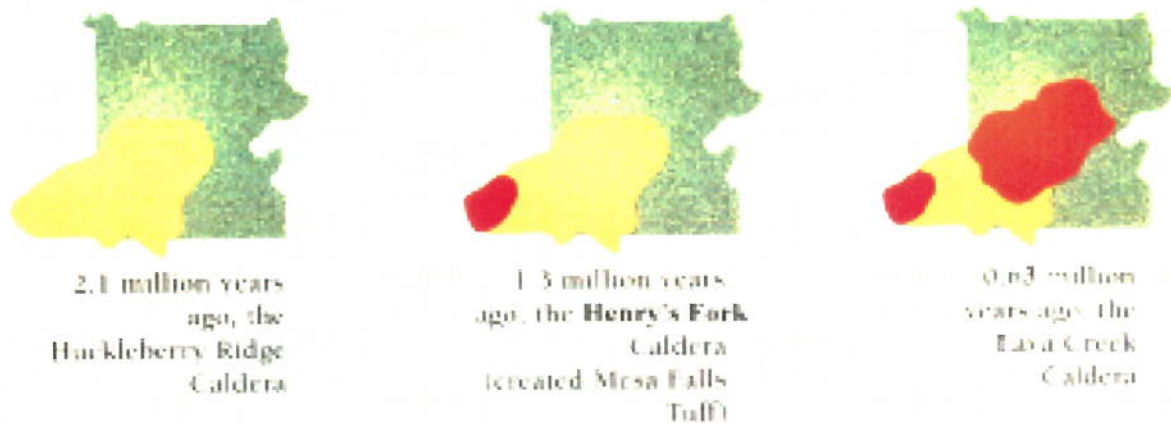


FIG. 2. Development of the Yellowstone Plateau volcanic field (after Christiansen, 2001).

A model for development of a particular volcanic field in the <10 Ma part of the hot spot track is illustrated in examining development of the Yellowstone Plateau (Figure 2) and Heise volcanic fields (Figure 3) (Morgan and McIntosh, 2005). At Heise, the initial caldera-forming eruption started with emplacement of the Blacktail Creek Tuff at 6.65 Ma probably from several smaller volcanic vents followed by collapse and formation of the Blacktail Creek caldera. This eruption signaled the start of the Heise volcanic field. The four large ($>200 \text{ km}^3$ - 1800 km^3) caldera-forming events spanned a 2+ million year interval, each being separated by 300,000 to 1,000,000 years. Rhyolitic lava flows fill and surround the caldera (Figure 2). Basaltic volcanism erupted on the sides of the rhyolitic magma chamber but while this chamber is ductile or partially molten cannot erupt through and onto the caldera floor. Basaltic volcanism continues several hundred to million years after the major caldera-forming events. Multiple caldera-forming events within a restricted area of 100 km^2 produces nested calderas filled with rhyolitic ignimbrites and lava flows intercalated with sediments, many lacustrine, and minor volume basalt flows. After cessation of caldera-forming events, basaltic magma is able to ascend along fractures or faults in the now-brittle crystallizing rhyolitic magma chamber body and erupt onto the floor of the caldera; basaltic lavas eventually cover the entire volcanic field and conceal it.

Reminders of this volcanic past are exposed along the margins of the Snake River Plain. On the Plain, hints of this geologic past are few.

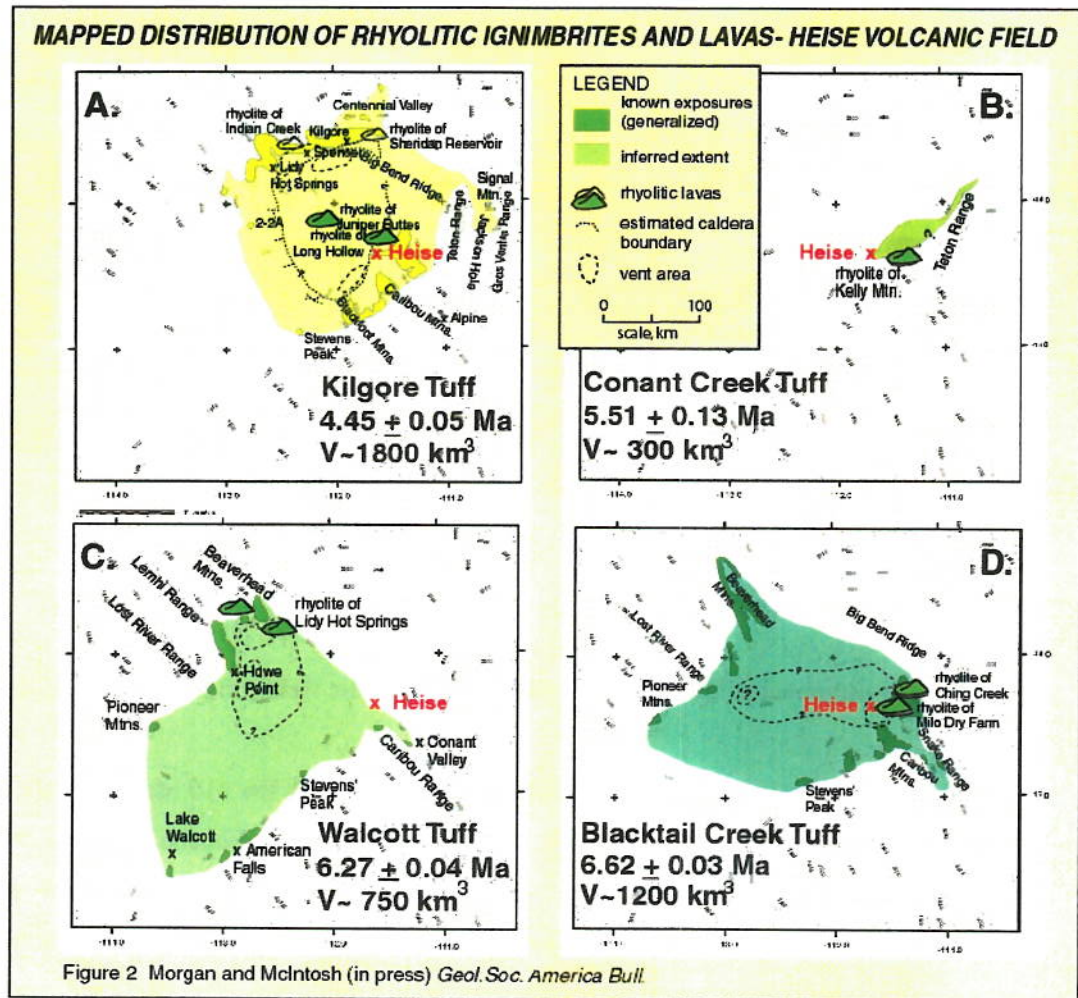


FIGURE 3. Evolution of the Heise volcanic field (from Morgan and McIntosh, 2005)

Figure 4 shows the evolution of a single caldera within a volcanic field, the 0.64-Ma Yellowstone caldera, the youngest caldera and the third cycle in Yellowstone Plateau volcanic field.

Uplift. The Yellowstone Crescent of High Terrain (YCHT) has a “bow-wave” configuration that is characterized by uplift in advance of volcanism. Several observations are well explained by hotspot uplift. The Yellowstone area stands high on the western North American plate and supports the Continental Divide. The axis of the Yellowstone Crescent of High Terrain, outlined by the wide grey line in Figure 5, flares out in a “bow wave”-like pattern outward from the northeast progression of hotspot volcanism. This Crescent forms an arc that crests ahead (northeast) of the youngest volcanism of the hotspot track. High mountain areas in this Crescent near the east boundary of Yellowstone National Park are formed of weakly consolidated, erodible

bedrock of the Eocene Absaroka Volcanic Supergroup. There, a gently rolling upland above tree line is sharply incised by steep valleys a kilometer deep. Prior to the cutting of the modern canyons, 3.5-Ma basalt erupted onto this surface. This incision can be

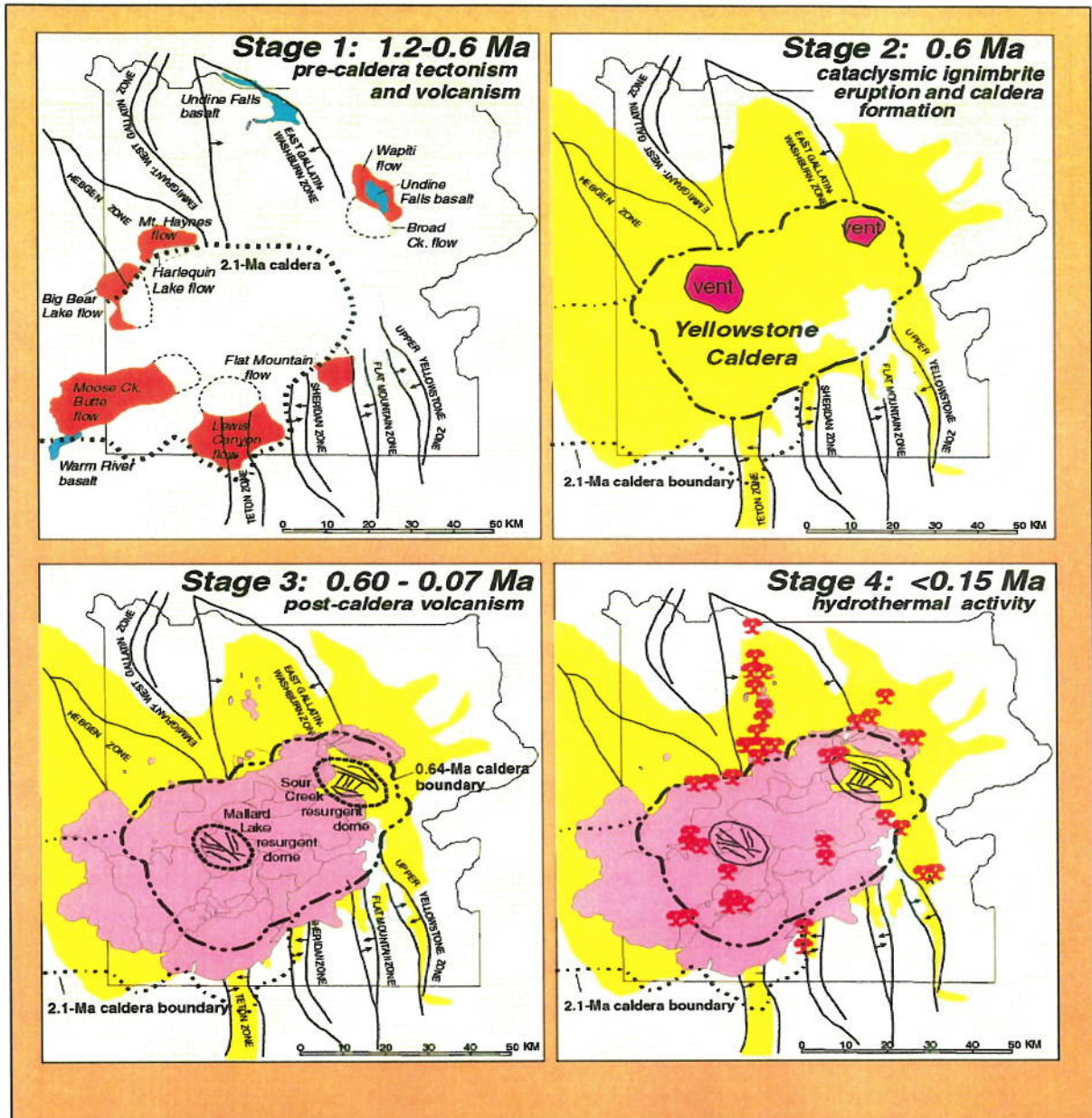
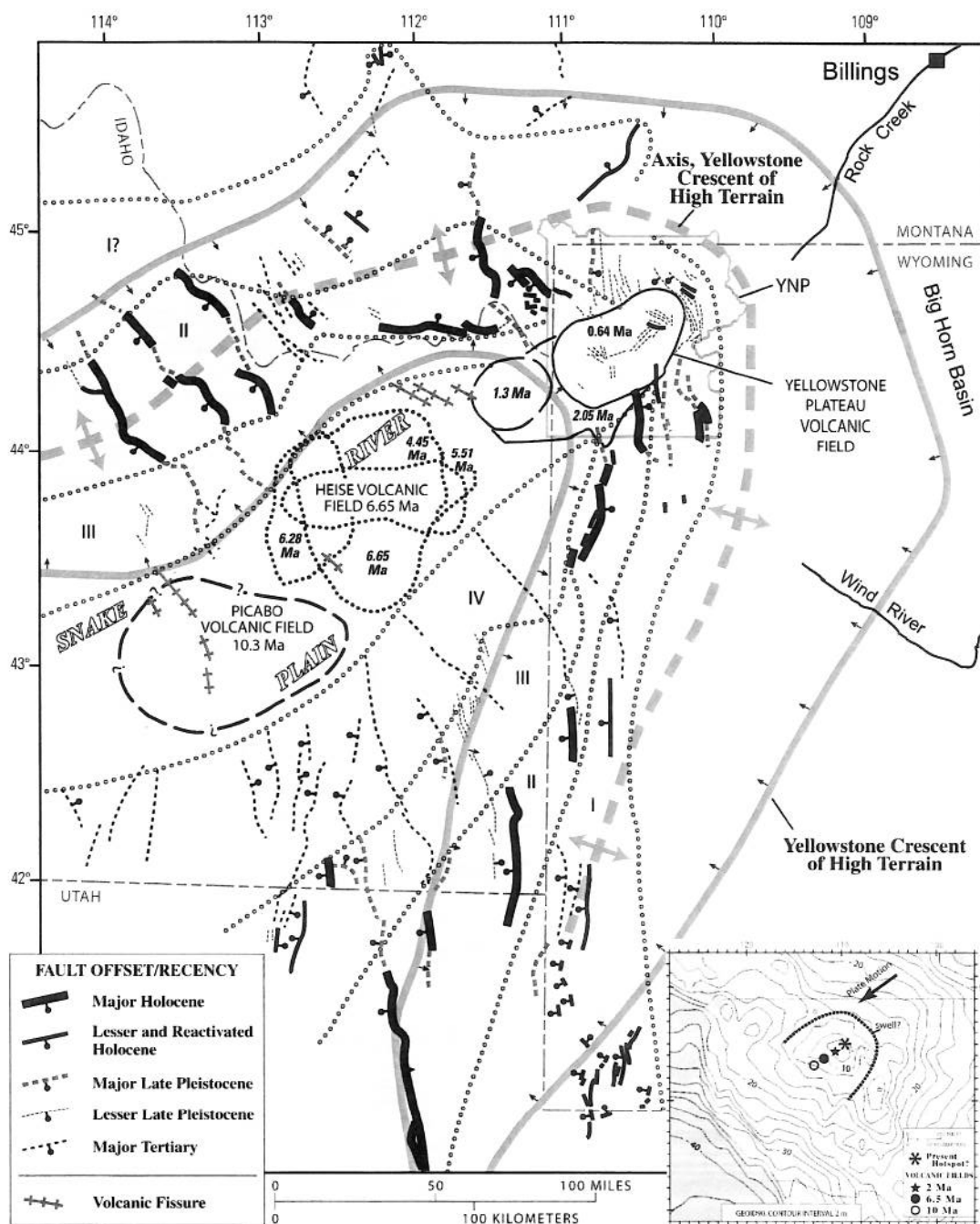


FIGURE 4. Evolution of the Yellowstone caldera, youngest and third volcanic cycle in the Yellowstone Plateau volcanic field (from Finn and Morgan; compiled from Christiansen, 2001 and Hildreth and others, 1984).

explained by young ongoing uplift (Pierce and Morgan, 1992). Additional evidence for young uplift and incision can be seen along the southeast and northwest part of the Crescent near the borders of Yellowstone National Park, where readily erodible bedrock of Mesozoic shales and sandstones form high mountains (Pierce and Morgan, 1992). Smith and Braile (1993) conclude that a 650 m topographic high centers on Yellowstone. Uplift associated with older position of the hotspot, Paul Link and students conclude from zircon provenance data that the Yellowstone hotspot controlled the east-migrating continental divide (see Beranek and others, 2006)

Faulting. Zones of late Cenozoic faulting extend south and west from Yellowstone. Within these zones are 4 belts: Belt I-faults with post-glacial (<12 ka) offset but little accumulated offset; Belt II-faults with post-glacial offset and >700 m range front relief; Belt III- faults with late Pleistocene offset and > 500 m relief, and Belt IV- faults with Tertiary or older Quaternary offset and >500 m relief (Pierce and Morgan, 1992). These belts may be explained by activity arranged about the volcanic track in a “bow wave” fashion with the following progression: Belt I- initiation of Quaternary faulting, Belt II-culmination of activity, Belt III- waning activity and Belt IV- cessation or great diminution of activity (on south side of eastern Snake River Plain only). These belts of faulting are roughly on the crest and inner slope of the Yellowstone Crescent of High Terrain and thus the faulting may be gravity driven by this topographic gradient down the inner slope of this “bow-wave” form. Both the YCHT and belts of faulting flare out more (1.6 x farther) on the south side of the track, possibly associated with the SE rise of the plume. This larger flaring on the south largely results from fault Belt IV that occurs only on the south side of the plain.

The nature of large-scale processes driving the Yellowstone hotspot is reflected in its surface geologic record, geophysical expression, and similarity to other hotspots. Recent tomographic images reveal a plume-like feature extending from the Yellowstone caldera to ~500-600 km depth and inclined 70° NW (Yuan and Dueker, 2005; Waite and Smith, 2005). This “plume” is an inclined 100-km-wide thermal anomaly whose temperature is about 180°C above ambient temperatures. The Yellowstone geoid anomaly, an 800-km-wide topographic swell centered on Yellowstone, is the largest geoid anomaly in the conterminous US. This U.S. geoid culmination on Yellowstone contrasts with both the U.S. Bouguer gravity and topography that culminate in western Colorado, with Yellowstone having a lesser anomaly. “Geoid anomalies reveal deep isostatic compensation. The geoid height of an isostatically compensated region is proportional

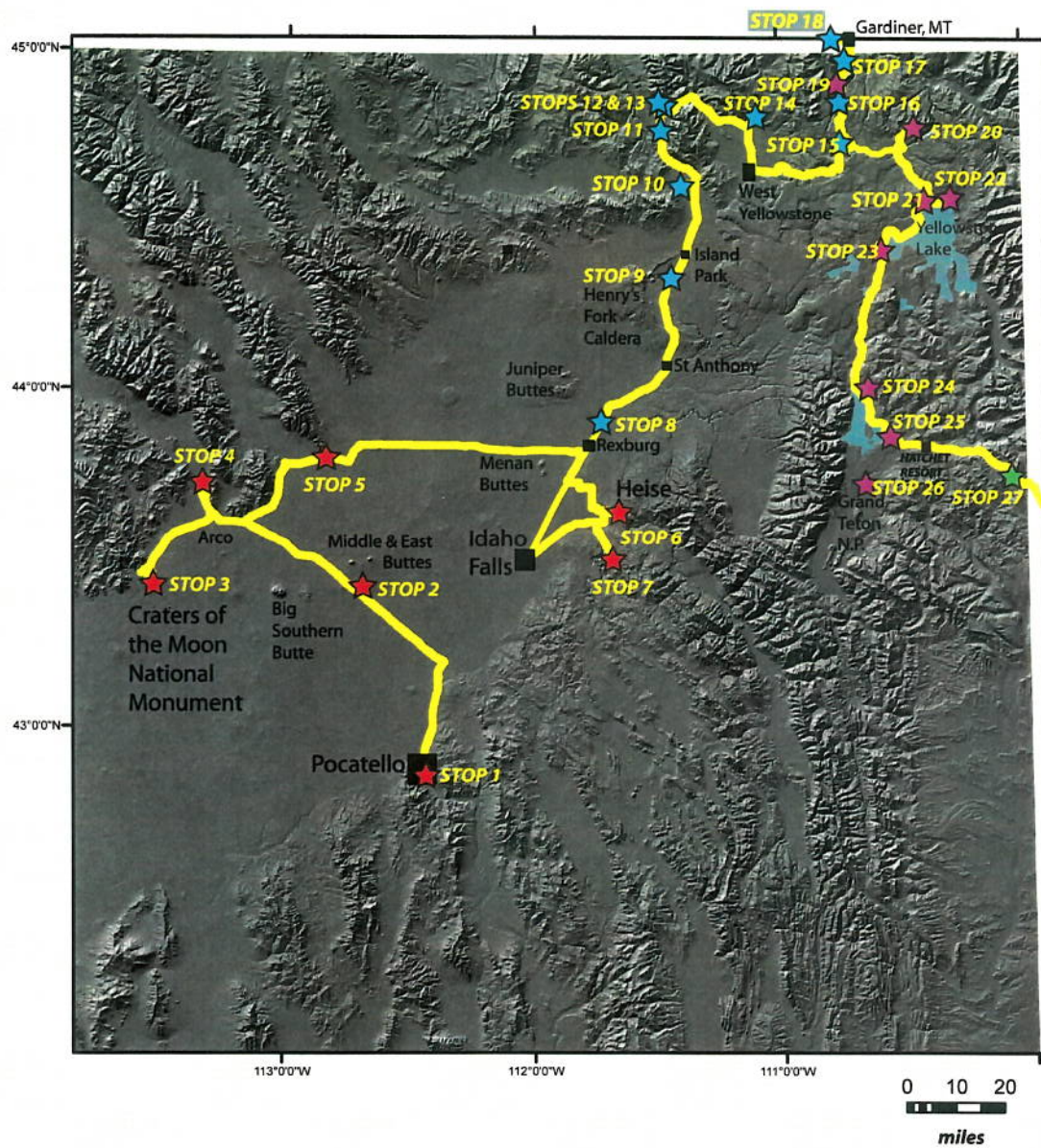


to the elevation times the depth of compensation. The Bouguer gravity anomaly of a broad feature is proportional to elevation.” (Norm Sleep, written commun., 2006).

The continuity of the Yellowstone hotspot track from its present position back in time to an apparent start with extensive volcanism in Nevada, Oregon, and Washington conforms to the plume-head/plume-tail model proposed by Richards et al. (1989). A voluminous plume head that pancaked outward to a radius of 500-1,000 km at the base of the North American plate has been invoked to explain the initial large area of volcanism (Draper, 1991; Pierce and Morgan, 1990, 1992; Camp, 1995; Takahashi and others, 1998; Pierce et al., 2002; Camp and Ross, 2004; Jordan et al., 2004). Later volcanism was much more restricted and aligned along the hotspot track (Figure 9), possibly because the magmatism was associated with a narrow plume ‘tail’. Seismic tomography images the plume into the mantle transition zone at 500-600 km depth but “... lower mantle plumes may be so altered by the transition zone that they appear to originate from within it.” (Schubert and others, 2001). The ~1,000-km-wide N-S geologic manifestations of the starting plume suggest inflation of the inferred plume-head was sourced from deeper than 500 km, as may the present 800-km-wide geoid anomaly. The plume head is centered in the Basin and Range province; its 16 Ma arrival correlates with changes in tectonics and volcanism (Pierce et al., 2002).

FIGURE 6. INDEX MAP TO FIELD TRIP

SNAKE RIVER PLAIN- YELLOWSTONE NATIONAL PARK FIELD TRIP



DAY 1 (October 23, 2007) Drive from Denver, Colorado across Wyoming to Pocatello, Idaho

Participants for this field trip will meet at the Denver Convention Center (address) between 7:00-7:30 am early Tuesday morning. We will depart promptly at 7:30 am and drive from Denver north on I-25 and west on I-80 across Wyoming. until we arrive at Pocatello, Idaho. We anticipate arriving in Pocatello around 6:00 pm. Our first night we will stay at the Rodeway Inn. Dinner is on your own; several restaurants are within walking distance of the motel. Breakfast Wednesday morning also is on your own; a restaurant is next door to our motel.

INFORMATION: Participants can reach Lisa Morgan on her cell phone at 720-530-4080

DAY 2 (October 24, 2007) Eastern Snake River Plain (Official start of field trip.)

We will depart from our motel in Pocatello at Lawton and 5th. The 5-minute drive to our first stop will proceed N on 5th, turn right on Humboldt and head toward the columns on hill overlooking Idaho State University. Continue and curve left on Cesar Chavez, turn right (east) on Martin Luther King Way, just past light at Memorial Dr. Turn right at the Performing Arts Center (Bartz Way), continue past tennis courts, pass Red Hill Rd. and turn right on Bartz Way. Curve left and pass McIntosh Manor. Stop at Schubert Heights sign. Walk up hill.

STOP 1- GATEWAY TO EASTERN SNAKE RIVER PLAIN: : ~1.3 miles, Mark 230, coordinates: 42.85960, -112.42770

Here we briefly summarize the volcanism, faulting, and uplift that define the track of the Yellowstone hotspot. Plate 1 portrays this pattern of volcanism, faulting, and uplift. Over the last 10 Ma, the onset of caldera-forming volcanism has migrated NE at a rate of about 25 km/my from near here to the Yellowstone Plateau. Faulting has accompanied this volcanic track and, over time, has flared out in a “bow wave” fashion. The Yellowstone Crescent of High Terrain (YCHT) is on the leading margin of the hotspot track, also flares out in a “bow-wave” fashion, and is attributed to uplift associated with the hotspot.

Basin-Range filling of the Portneuf valley occurred mostly between deposition of two ashes dated 8.3 and 7.4 Ma (Rodgers and others, 2002). About 15 miles east of here is a major fault on the east side of the Portneuf Range. At the northern end of the range adjacent to the eastern Snake River Plain, major Basin and Range normal faulting occurred mostly between 7.0-6.5 Ma

(Kellogg and Marvin, 1988, p 15), which is near the time of inception of caldera-forming activity in the Heise volcanic field on the adjacent eastern Snake River Plain associated with the Yellowstone hotspot. Our stop and these nearby sites are in Belt IV of faulting that is defined by late Tertiary or early Quaternary activity but later Quaternary quiescence (Pierce and Morgan, 1992).

The Bonneville flood roared down Portneuf valley ~18 ka reaching almost as high as the knob we are standing on. It eroded back the valley sides creating the steep slopes that truncate the perched alluvial fans. Up valley from here, the 0.5 Ma basalt of Portneuf valley was close to base level at the time of the flood (Rodgers and others, 2002). The Bonneville flood eroded scablands on this basalt and deposited a huge bouldery fan at the margin of the Snake River Plain.

To the west in the divide at the head of Trail Creek, a layer of Lava Creek ash occurs well down in the loess sequence (Izett and Wilcox, 1982), and records some of the last caldera-forming activity associated with the Yellowstone hotspot.

From this vantage point, we can look across the Snake River Plain to the northern margin of the plain where we see the Lemhi and Lost River Ranges. The 90-km width of the Snake River Plain reflects the width of the various volcanic fields each containing several nested calderas. In this part of the Plain, the Heise volcanic field stretches across the plain and contains several nested calderas. For the most part, deposits erupted from the four major caldera-forming events in the Heise field are exposed along the margins in the foothills of the Basin and Range terrain. Figure x shows the evolution of calderas and the distribution of associated deposits with each caldera event in the Heise volcanic field.

The Snake River Plain has a veneer of younger (<2.0 Ma) basalt concealing the calderas. Several deep boreholes have been drilled on the Snake River Plain and penetrate the basaltic cover. In general, the thickness of the basaltic cover with intercalated sediments is less than 800 m; the basalts overlie a thin sequence of tuffaceous lacustrine sediments (<100 m) which in turn overlie a thick sequence of rhyolitic ignimbrites, lava flows, and fall deposits. In the deepest well drilled to date on the eastern Snake River Plain, INEL-1 at 3159 m, over 2500 meters of rhyolite are present below the basaltic cover (Doherty and others, 1979). In the distance to the north, several large Quaternary rhyolitic buttes protrude the basaltic cover on the Snake River Plain. These buttes include Big Southern (300 ka, Hughes 1999), East (600 ka), and Middle (<1000 ka) Buttes. While Big Southern and East Buttes are clearly rhyolitic, Middle Butte is covered with Snake River Group basalts whose slopes are covered with talus from its basaltic cap. Magnetic and gravity data, however, suggest its core is rhyolite and thus the age of the rhyolite here is < 2.0 Ma.

Return to bus and go north on I-15 toward Blackfoot and Idaho Falls. To get to I-15, drive north on Cesar Chavez Road, go to right on Center, head east and get on I-15N.

As we drive north on I-15, we drive along the southern edge of the 10.3-Ma Arbon Valley caldera, identified in Kellogg and others, 1992. The tuff of Arbon Valley is a relatively crystal-rich welded to non-welded ignimbrite containing bipyramidal quartz and biotite. Hydrous minerals are rare in rhyolites from the SRP-YP volcanic province and thus, the biotite-bearing tuff of Arbon Valley is unusual and can be used as a regional stratigraphic marker. The tuff of Arbon Valley is the oldest identified ignimbrite in the Picabo volcanic field (Plate 1). Several rhyolitic domes have been mapped and have been used to infer the approximate location of the Arbon Valley caldera. Much of the landscape is mantled with thick loess, which in places is several to many meters thick. As we pass the Fort Hall Reservation, a basaltic shield volcano can be observed off to the left. As we continue north on I-15, we pass Buckskin Basin (10s of feet in

As we pass the Fort Hall exit, a large shield volcano can be seen on the left

As we continue north, we pass Buckskin Basin where several of the ignimbrites from the Heise volcanic field are exposed and in places are above exposures of the 10.3-Ma tuff of Arbon Valley and other minor units

As we approach Blackfoot, we traverse Snake River gravels of Pinedale age that are less than 25 ka. At exit 93, we drop down onto the floodplain of the Snake River. **Note about the Snake River:** the headwaters of the Snake River are in the southern part of Yellowstone National Park. From there, the Snake River flows west to near the South Entrance of Yellowstone National Park where it turns south and flows along the eastern front of the Teton Range continuing south through Palisades Reservoir and the Swan Valley of Idaho where it turns northwestward to the south side of the eastern Snake River Plain. The river flows along the south side of the plain cutting numerous water falls through basaltic and rhyolitic flows. Many towns along the Snake River have “falls” in their name and include Idaho Falls, American Falls, Twin Falls. It should also be noted that no river crosses the Snake River Plain—more later in the trip.)

At Blackfoot, take exit 93 off of I-15N and head northwest on US 26. Drive toward Arco, ID (57 miles). We cross the Snake River in Blackfoot and at exit 93, we climb from the floodplain of the Snake River to a higher terrace of late Pleistocene to Holocene age cut into Pinedale gravels (Scott, 1982). In about 5 miles as we continue to drive northwest on US 26, (Mark 231) we climb onto a remnant of fresh Holocene basalt with tumuli and pressure ridges. After a mile, Highway 26 traverses agricultural land on older, loess-mantled basalts.

In the distance, a basaltic shield volcano is located at about the 2:00 position. We have now driven onto loess-mantled basalt. The gently rolling agricultural land may represent a series of coalescing shield volcanoes.

As we continue on US 26, we see around us several buttes, most of which represent Quaternary rhyolitic domes or smaller vents that erupted after the main phase of hot spot activity had passed this part of the track. Big Southern Butte is in the distance located at about the 10:00, Table Legs Butte is located at about 11:30, Middle Butte is at 2:00, and East Butte with the antennas on top is at 3:00. Middle Butte is an endogenous dome with a basaltic cap of Snake River Group basalt; East and Big Southern Buttes are exogenous and represent excellent examples of these forms. The low elongate landform between Middle and East Buttes is Twin Buttes.

As we continue to drive northwest on US 26, the Lost River Range can be seen straight ahead in the distance on the northern margin of the SRP. Cedar Butte now comes into view just south of Big Southern Butte at the Atomic City turnoff.

STOP 2: OVEVIEW OF SRP AND RHYOLITE DOMES (55 miles, mark #232, 43.45040, -112.78968): This stop is in the middle of Snake River Plain and is due north of Atomic City. Here we have an excellent view of all the buttes. This stop is in the center of the Heise volcanic field and represents an evolved stage of hot spot volcanism. At least three large Yellowstone-style calderas are present below this location. Over time, these large collapse calderas have been filled in by rhyolitic lavas and later basaltic lavas. In the Quaternary and most recent Holocene, this area has evolved into a much younger volcanic landscape covered with basaltic fissure vents, shield volcanoes, and rhyolitic buttes.

Return to bus and continue northwest on US 26. Here we see the Lemhi Range on the northern margin of the plain. Howe Point, where we will stop later in the day, is on the southernmost tip of the Lemhi Range. Turn left (west) onto US 20 & 26.

INL is a government reservation located in the southeastern Idaho desert. At 890 square miles (569,135 acres), INL is roughly 85 percent the size of Rhode Island. It was established in 1949 as the National Reactor Testing Station, and for many years was the site of the largest concentration of nuclear reactors in the world. Fifty-two nuclear reactors were built, including the U.S. Navy's first prototype nuclear propulsion plant. During the 1970s, the laboratory's mission broadened into other areas, such as biotechnology, energy and materials research, and

conservation and renewable energy. At the end of the Cold War, waste treatment and cleanup of previously contaminated sites became a priority.

As we drive west, we cross the 890-square-mile Idaho National Laboratory (INL) where at one point more nuclear reactors were concentrated at this site than anywhere else in the world. Established in 1949, INL current focus is nuclear waste treatment and cleanup of previously contaminated sites. The charge of the Idaho Cleanup Project is to safely and cost-effectively complete the majority of cleanup work from past laboratory missions by 2012. On the left, we pass EBR-1 (Experimental Breeder Reactor) site is located on the left and is a National Historic site. INL has significantly influenced the economy of the whole region.

The Snake River aquifer underlies the Snake River Plain. No river crosses the SRP. The Big Lost River and the Little Lost River both flow from basins north of the plain out onto the plain along the northern margin and then flow into sink holes at Howe Point, the southern tip of the Lemhi Range. The Snake River aquifer is about 400ft deep here. The Water Resources Division of the USGS has been at INL for many years and monitors wells to track the movement of a radioactive plume which flows from INL to the southwest toward Twin Falls. On the southern margin of the plain, the Snake River aquifer surfaces at Thousand Springs where fish hatcheries use the emergent waters of the aquifer before it joins the Snake River.

The basin ranges north of the plain are east-tilted ranges with listric faults on their west sides adjacent to basins forming half graben also tilted east.

As we continue driving on US 20 and 26, we begin to see off in the distance to the west the Craters of the Moon lava field as the low dark terrain with nonvegetated knobs. On the left, we pass the small enclave of Butte City which was heated by geothermal energy during the 1970's and 1980's. Harmon Craig obtained high $^3\text{He}/^4\text{He}$ values from thermal waters collected here in the early 1980's.

As we approach the town of Arco, note the Arco Hills to right and Arco Peak ahead, composed of deformed Paleozoic limestones. Note the long history of Arco High School marking outcrops on the hillside with graduation dates. Arco's name came from its original settlers who came in the 1860s from Arco, Georgia.

Optional Rest room stop at Texaco—^{Dave's} Carol's travel plaza. (84 miles, MARK 233)

Return to bus, continuing west on US 20-26. At lighted intersection, turn left or south onto US 93. Here we cross low terraces and the floodplain of the Big Lost River which forms the drainage between the Lost River Range and the Pioneer Mountains. Note the sign on the Arco City Hall as

we pass: Arco was the first city in the world to be entirely powered by atomic power. As we continue through town on US 93, note Atoms for Peace Park on our left.

Leaving Arco on US 93, we continue to head southwest towards Craters of the Moon. Note the beautiful shield volcanoes located at about 11 o'clock.

102 miles- As we approach Craters of the Moon (18 miles southwest of Arco), we see 12,000-year-old Sunset Crater located about 1 o'clock on the right side of the road. Sunset Crater with its vegetative cover is one of the older eruptive sites at Craters of the Moon. On our left to the south is the 2,000-year-old Blue Dragon flow which erupted from Inferno Cone. Note the large blocks of welded cinder from crater wall which were rafted on the lava flow. We will continue past the entrance to the Craters of the Moon National Monument. Note the kapuka protruding through a a'a lava on our left at the overview.

Rafted on
Dev. O. & Serrate
Flows

primarily from fissures S of Big Craters

STOP 4: ENTRANCE TO CRATERS OF THE MOON NATIONAL MONUMENT (104 miles, Mark 235, 43.46188, -113.56308)

We will briefly visit the visitor center where they have a 3-dimensional solid terrain model of the Snake River Plain and Yellowstone showing the track of the Yellowstone hot spot over time. We will return to the bus and proceed into the Monument. We will park at the base of Inferno Cone and take a 10-minute hike to the top where, weather permitting, we will have lunch and have an incredible view of the Craters of the Moon (COM) lava field, the largest Holocene basaltic field in the conterminous United States. At >1600 km², the COM lava field has >30 km³ of lava flows and associated vent and pyroclastic deposits (Kuntz and others, 1992). The COM lava field formed during the past 15,000 years in 8 eruptive periods. The last eruptive period was 2100 years ago. Each eruptive period lasted several hundred years and each period was separated by an interval of several hundred to 3000 years. Given that the last eruptive period was 2100 years ago, Kuntz and others (1992) inferred that another eruptive period is likely within the next 1000 years. Craters of the Moon lava field is distinct from most of the other lava fields on the eastern Snake River Plain in that it is a polygenetic field with over 60 lava flows which erupted from 25 tephra cones and at least 8 eruptive fissures in the northern Great Rift Zone (Kuntz and others, 1982). Other Holocene basaltic lava fields on the eastern Snake River Plain have monogenetic eruptions from vents scattered in space and time. Additionally, the chemical composition of its lavas show that they are more evolved; their SiO₂ content ranges from 44% up to 64%. Some of these lavas clearly show crustal contamination whereas others clearly have mineralogical and chemical variations due to crystal fractionation (Kuntz and others, 1992). In

contrast, most of the other basaltic lava fields on the eastern Snake River Plain have a much narrower range of SiO₂ (45-48%) and have not been fractionated.

Stop at Inferno Cone (43.44503, -113.55555): Short hike to top of Inferno Cone is one of the youngest cinder cones in the COM lava field. Prevailing winds from the west and southwest distributed tephra from the erupting Inferno Cone vent to the east and northeast resulting in an asymmetric tephra cone. From here to the south, we have a great views of Big Cinder Butte, the ~~Blue Dragon~~ flow with rafted blocks of its cratered wall, a tan kapuka to the northeast, and small bombs and agglutinated spatter at summit.

Dev Arch. • 2:52p, 3932 mi. Return to bus and leave Craters of the Moon National Monument. Head to the northeast on US 20-26 back to Arco. At Arco, turn left west at intersection and drive west on US 93 toward Mackey. We are in the Big Lost River valley, which is floored by Holocene and late Pleistocene gravels. Alluvial fans are built out from the Lost River Range to the east and from the Pioneer and other mountains to the west. About 4 miles from Arco, we turn right onto a gravel road. Shortly after we turn onto the gravel road we will turn left onto King Mountain Road. All along this road to the east, the Arco fault forms high scarps in old alluvial fan gravels. Drive along here for ~3.6 miles.

STOP 4: FAULTING ASSOCIATED WITH THE YELLOWSTONE HOT SPOT: the A-2 trench of Hal Malde. (139 miles, Mark 240, 43.7209, -113.32768)

Park where a two-track road goes up valley to the right. Walk up small drainage and climb up to trench and then climb to top of fault scarp.

The Arco segment of the Lost River fault is in fault Belt III that is characterized by Quaternary faults on range fronts >500 m high with late Pleistocene (<130 ka) offset but that have not been active since the last glaciation (Plate 1) (Pierce and Morgan, 1992). Forty miles north, the Thousand Springs segment is in Belt II faulting, and includes scarps of the 1983 Borah Peak earthquake in addition of a previous Holocene offset. At the start of the day in Pocatello, we saw inactive faulting from Belt IV in the Portneuf range at Pocatello. This is all part of a pattern of faulting flairs out from the track of the Yellowstone hotspot. The Yellowstone Crescent of High terrain also flares out, and includes Borah Peak, the highest peak in Idaho near the northern end of the Lost River Range.

At this stop, the foot wall of the fault scarp is in ~160,000 yr old alluvial fan gravels (Pierce, 1985). Pierce estimated 19 m total offset, resulting in an overall slip rate of 0.12 m/ka. In the hanging wall, the bottom of the trench just above the inferred same gravel at the top of the

footwall are two volcanic ashes in loess just above gravel inferred to be that at the top of the footwall. The upper ash is probably from Yellowstone and is about 76 ± 5 ka which would correspond to some of the youngest rhyolitic volcanic activity on the Yellowstone Plateau. The other ash may be from the Cascade Range in the Pacific northwest (see Pierce, 1985 for description). The inferred late Pleistocene gravels forming the floor of the valley just north of the trench are not offset, indicating the age of the youngest faulting is older than 12-15 ka.

From the Arco Peak trench site farther south, Olig and others (1995; 1997) concluded the last two offsets occurred between 20 ± 2 ka and 23 ± 2 ka. They also conclude that offset events clustered in time.

Carbonate coats form on the undersides of stones in the soil profile. They are useful as a relative age indicator and can be dated by U/Th analysis. Carbonate coats in the soil on the footwall fan gravels are 10.1 ± 5.4 mm thick. John Rosholt (in Pierce, 1985) determined that three subdivisions of the carbonate coat became successively older towards the rock surface and that the start of coat deposition was about 160 ± 35 ka. More recently, Jennifer Pierce and others (2007) dated this surface as 178 ± 8 ka. Dating coats on different aged alluvial fans yielded an overall rate of buildup of ~ 0.6 mm/10 ka.

Depart trench site. Drive ~ 0.8 mile and turn right on gravel road (3000N) (different than the way we came). After ~ 1 mile, turn left on highway 93 and head back toward Arco. Proceed east on US 20-26 out of Arco and from Arco, continue ~ 7 miles until the T-shaped intersection with Idaho Highway 22 and 33 heading toward Howe. As we turn onto highway 33, we can see in the distance the Lost River sinks on the right at the southern tip of the Lemhi Range. As you look to the east, you can see the axial high of the eastern Snake River Plain.

Our 16-mile drive toward Howe takes us along the east side of the Lost River Range which is mostly Paleozoic limestones. As we proceed north we can start seeing dark beds of Miocene volcanic rocks from the Heise volcanic field plastered against at the base of the mountain front. The main ignimbrite exposed here is the densely welded 6.27 ± 0.04 -Ma Walcott Tuff (Morgan and McIntosh, 2005). The 4.45-Ma Kilgore Tuff is exposed farther north in the Lost River Range. Out on the Plain, all the large-volume ignimbrites or fall deposits from the Heise volcanic field have been identified in various drill holes (well 2-2A, well 1, INEL-1, WO-2) (Doherty, 1979; Doherty and others, 1979; Morgan and others, 1992).

As we continue to drive north, two high-angle subsurface faults are present at ~ 1.8 km and 6.0 km east from the range front as inferred from seismic reflection (Pankratz and Ackerman, 1982). Displacement of 1000 m and 500 m, respectively, along these faults brings the volcanic sequence into contact with Paleozoic carbonate basement. Morgan and others (1984) noted that

these subsurface structures strike parallel to a steep north-south gravity gradient (Mabey, 1978) and are very different in their orientation from local basin-range faults. They suggested these faults may be caldera related (Morgan and others, 1984). We suggest these structures are associated with the structural margin of the Blue Creek caldera, source of the Walcott Tuff. As we approach Howe, the Lemhi Range is to the east and at 2:00 Howe Point marks a topographic edge to the Blue Creek caldera. The high peak to the north is M Mt. Tyler. The low flat bench extending into the plain is comprised of Miocene volcanic rocks. Here the 10.3-Ma tuff of Arbon Valley is exposed and is below a package of volcanic rocks from the Heise volcanic field, including a basalt, the 6.65-Ma Blacktail Creek, 6.27-Ma Walcott, and 4.45-Ma Kilgore Tuffs (Morgan and McIntosh, 2005).

Sixteen miles from where we came onto highway 33, we enter Howe, formerly known as Sweet Sage. In the early 20th century, Howe had over 400 residents, 4 barbershops, and several bars and restaurants due to the lead and zinc mining operations up the Little Lost River Valley. At Howe, turn right, and continue east on highway 33. From Howe to the next stop, Scott (1982) shows that the highway first traverses a Holocene floodplain of the Little Lost River for about a mile, and then traverses the flat Late Pleistocene lake sediments of Pleistocene Lake Terreton for most of the way east to Howe Point. North of the highway beneath the wide flat Little Lost River Valley are Pinedale gravels representing a major depositional event from the the Little Lost River before it flowed into Lake Terreton.

On our left at 10:00 are a series of alluvial fans, the older of which are faulted. You may see a dark trench across a scarp in an older fan. The younger fans do not have fault scarps. In his landmark Nature paper, Anders (1996) noted that an interval of high rates of faulting followed passage of the tectonic parabola.

As we approach Howe Point at the southern tip of the Lemhi Range, we see a thick section (~30 m) of the densely welded, crystal-poor, Walcott Tuff exposed as the cap rock in this sequence of bedded volcanic rocks. This ignimbrite has a 2- to 50-cm-thick basal vitrophyre that is of artifact quality and was highly valued and traded among various Native American people.

STOP 5: HOWE POINT – (179 miles, Mark 241, 43.80388, -112.84840).

Pull off on the south side of the highway. The road marks near where we cross a caldera-bounding fault related to the Blue Creek caldera. North of the highway, the Walcott Tuff is exposed as near-source out-flow sheets; in contrast, to the south we cross a faulted topographic boundary of the caldera, where the Walcott Tuff is faulted down into the plain with minimum displacements of ~50 m. The thickness of the Walcott Tuff increases dramatically from ~30 m

outside of the topographic margin of the caldera to >100 m within the caldera. The intracaldera unit shows well-developed rheomorphic folds. Thermal demagnetization studies of various oriented samples of tightly folded Walcott Tuff show that deformation of the ignimbrite occurred above its Curie temperature; each sample regardless of its orientation has the same remanent magnetization.

On the north side of the highway, an older densely welded lithophysal ignimbrite, the tuff of Howe Point, is exposed. Fission-track analyses on zircon microphenocrysts from this unit give a radiometric age of ~12 Ma (McBroome, 1981). The eruptive source for this unit may be to the southwest of Twin Falls where other 12-Ma ignimbrites are abundant, thick, and well-exposed. Figure 4 is a composite section of the volcanic units exposed here. The 10.3-Ma tuff of Arbon Valley is exposed near here to the north and in turn is covered by a sequence of lacustrine sediments intercalated with basaltic lava flows. Above these units the Blacktail Creek Tuff overlies a series of ashes that grade upward into the densely welded Walcott Tuff. We will look at exposures on both sides of the caldera-related fault or road and spend ~45-55 minutes here.

Howe Point and east. On the northern margin of the eastern Snake River Plain, range-front faults are on the western sides of each range, including the Lost River and Lemhi Ranges and Beaverhead Mountains. At Howe Point, the Lemhi fault flanks the west side of the Lemhi Range; to the east the Beaverhead fault is on the west side of the Beaverhead Mountains. For the Lemhi fault, Mark Anders (1994) concluded the highest activity was between ~10 and 6.5 Ma. To the northeast, fault activity along the Beaverhead fault was around 6-7 Ma. These are two of ten localities marginal to the Snake River Plain that indicated a northeast progression of faulting in a parabolic pattern has been migrating for at least the last 10-12 m.y. at a rate of 2.2 cm/yr and oriented north-northeast.

As we continue east on Highway 33, we cross the Test Area North of INL, where much of the early nuclear submarine reactor development was done.

Turn left at stop sign, continue to follow highway 33. Continue 4.7 miles before turning right toward Rexburg and continuing on highway 33. Off to the right is Circular Butte, a basaltic shield volcano. The highway traverses largely stabilized sand dunes (Scott, 1982).

Pass through town of Mud Lake and two miles farther east Terreton. Several miles north of Terreton is the Mud Lake National Wildlife Refuge, whose heart is Mud Lake, a remnant of late Pleistocene Lake Terreton.

About 14 miles east of Terreton, we are headed straight for Menan Buttes, basaltic maar volcanoes (Hackett and Morgan, 1988). Winds from the southwest to the NE during eruption resulted in the asymmetric form of these cinder cones.

As we continue north, on our left is Juniper Buttes, the resurgent dome for Kilgore caldera (Morgan, 1988; Morgan and McIntosh, 2005). Juniper Buttes are closely flanked by active dunes with inactive dunes farther in the distance.

Cross and continue east on I-15. Two miles east of I-15, one starts getting great views of the backside of the Teton Range. Turn south on US20.

Over time, the Snake River has entered the Snake River Plain and has deposited a huge alluvial fan of outwash and glacial-climate gravel. Glacial outwash here is in large part from the southern margin of the greater Yellowstone glacial system in Jackson Hole which we will visit on this field trip in the day-after tomorrow. The alluvial fan forms a right angle cone with a radius of 30 km and a periphery of more than 40 km. This large, low gradient fan made an ideal landform for extensive early irrigation farming using radial canals and lateral ditches. It is interesting that there is no fan recognized of Bull Lake age, although there are Bull Lake terraces in Jackson Hole that are more than 100 m above the present Snake River. Perhaps the Bull Lake fan is buried. Continue following road. At the turn-off east for US20 at Archer, turn east (left). In the distance located at 2:00 PM is a great view of the Heise cliffs, type section for the Heise Volcanic Group. About 3 miles later, turn right (S) on 2000W, which curves left and turns into 7800S, curve right into 600E, Heise Cliffs to left. Top down: Kilgore, Conant Cr., Walcott, Blacktail, Hawley Spring at 7.2 Ma (lava flow). Further on is the 5.7-Ma rhyolite of Kelly Mountain, a massive rhyolitic lava flow.

Continue another 6 miles from where you cross the Snake River. Continue another ~1.5 miles, turn left on 200 N to Heise Hot Springs.

STOP 6: HEISE HOT SPRINGS/HEISE CLIFFS. (243 miles, mark 243, 43.64102, -111.68385)

The Heise cliffs contain many of the type or reference sections of major units of the Heise Group, a series of 4-7.5-Ma volcanic rocks including basaltic tephra, rhyolitic lava flows and large-volume densely welded ignimbrites, and subordinate non-welded ignimbrites and fall deposits. At the base of the cliffs is the 2.0-5.0-Ma Grand Valley fault (Anders et al., 1989).

Head out south on 160W to east on US 26, turn right (south) on 145W- Meadow Cr Rd. Pass Frisbee golf course at Ririe reservoir. Continue on road as it changes to gravel base.

STOP 7: MEADOW CREEK DUGWAY SITE. (280 miles, mark 245, 43.53328, -111.70110). Here we have an incredible exposure of several ignimbrites from the Yellowstone Plateau and Heise volcanic fields exposed along the Meadow Creek Dugway. Draping the landscape and capping the exposure is the 2.05-Ma Huckleberry Ridge Tuff from the Yellowstone Plateau volcanic field whose source is over 50 km away to the northeast. The Huckleberry Ridge Tuff lies above volcanic units from the Heise volcanic field and is immediately above the 4.45-Ma Kilgore Tuff which has filled an undulating topography; its thickness changes rapidly from <2 m to >15 m in places nearby. The Kilgore Tuff overlies the 5.57-Ma tuff of Wolverine Creek, exposed here as a series of pyroclastic fall, surge, and flow deposits. The source for the tuff of Wolverine Creek is close; this unit typically is non-welded and contains large amounts of obsidian. The tuff of Wolverine Creek overlies the 6.65-Ma Blacktail Creek Tuff exposed at the base of the Dugout in the drainage (Fig. xx-strat column).

Return to Idaho Falls where we will spend the night at Guesthouse Suites. Dinner will be on your own. A good beer and pizza (with other Italian food) is at the Brownstone, just a couple blocks south down the road.

DAY 2 (October 24, 2007)

LEAVE IDAHO FALLS AT 7:30 AM AND GO NORTH ON US 20.

As we approach Rexburg, we again traverse a huge Snake River alluvial fan of Pinedale age. From here, we also have great views of the backside of the Teton Range peaking to the east. The Rexburg bench forms the broad high flat-topped ridge to the east and has excellent exposures of the 2.05-Ma Huckleberry Ridge Tuff as well as many major and minor units associated with the Heise volcanic field (fig. Heise). The northwest edge of the bench is caldera-related; near here

the topographic margins of the 4-45-Ma Kilgore and 6.65-Ma Blacktail Creek calderas are located. Out onto the plain to the west-northwest, the low hills are the Juniper Buttes, a series of rectilinearly faulted lavas related to resurgence of the Kilgore caldera (Morgan, 1988; Morgan and McIntosh, 2005).

Between Rexburg to St. Anthony, US 20 follows Pinedale outwash terrace gravels of the Henrys Fork and tributary streams (Scott, 1982).

STOP 8: OVERLOOK OF JUNIPER BUTTES RESURGENT DOME. (40 miles, mark 251, 43.88375) -111.75237. As shown on the geologic map of Juniper Buttes (Kuntz, 1979), the buttes represent a series of rectilinear fault blocks cut into rhyolitic lava flows, similar to those mapped at many resurgent domes elsewhere (eg., Yellowstone, Timber Mountain, Valles calderas). Several post-Kilgore Tuff-rhyolitic lavas with K-Ar ages of 3.31 ± 0.06 Ma, 3.47 ± 0.06 Ma, and 3.72 ± 0.08 Ma are exposed here. Quartz phenocrysts from these lavas have similar dO18 values as the Kilgore Tuff and are distinct from the older ignimbrites of the Heise volcanic field. The age of the lavas at Juniper Buttes, the central location of Juniper Buttes dome complex in relation to the buried vents for the Kilgore Tuff suggest that Juniper Buttes is a relatively young rhyolitic feature on the eastern Snake River Plain and is a resurgent dome or up-faulted block associated with the Kilgore caldera (Morgan, 1988; Morgan and McIntosh, 2005).

Quartz phenocrysts from the rhyolitic lavas at Juniper Buttes have similar dO18 values as the Kilgore Tuff and are distinct and have significantly lower values than the older ignimbrites of the Heise volcanic field (Morgan, 1988). Low dO18 values of the Kilgore Tuff and lavas from its resurgent dome may indicate that the source of these magmas comes from hydrothermally altered terrains of previous caldera landscapes within the Heise volcanic field. Hydrothermal sinter deposits in older parts of the Heise volcanic field are part of the stratigraphic record in exposures including those at Heise, Green Mountain, Stinking Springs, Lidy, Spencer Hot Springs, and elsewhere and may contribute significantly to the overall low dO18 value of crust that is later melted and incorporated into younger magmas (Morgan and others, 2005; Morgan and others, 2006).

Return to bus and continue north on US 20 toward Ashton, Idaho.

Ashton is just inside the outer margin of the older Bull Lake-age moraines of the greater Yellowstone glacial system. Obsidian hydration rims and basalt weathering rinds (Coleman and Pierce, 1981) collected here are similar to those for the 40-ka Bull Lake moraines at West Yellowstone. On the west and south sides of the present position of the Yellowstone hotspot, the

Bull Lake glaciation was much more extensive than the more recent (16,000-24,000 years ago) Pinedale glaciation. Subsidence on the trailing edge (ie the southwest edge) of the Yellowstone Crescent of High Terrain resulted in the extent of the Bull Lake glaciation being much more extensive here than was the younger Pinedale glaciation.

As we drive north of Ashton, we leave the Heise volcanic field and enter the Yellowstone Plateau volcanic field as we drive up a large rampart known as Big Bend Ridge. This prominent, semi-circular high ridge represents segments of the topographic margins of the 2.05-Ma Huckleberry Ridge caldera and the younger 1.3-Ma Henrys Fork caldera. As we drive up Big Bend Ridge, we will first drive through member B of the Huckleberry Ridge Tuff, the same member we saw exposed at Meadow Creek Dugway on Day 1. Member B here is capped by the 1.3-Ma Mesa Falls Tuff. Thick fall deposits that come coarser as we go up in the section are present at the base of the Mesa Falls Tuff and are exposed here. Near the top of these fall deposits are a series of bedded, cross-stratified, and channeled pyroclastic deposits interpreted as emplaced as ground-surge units associated with the eruption of the Mesa Falls Tuff (Doherty, 1976; Neace, 1986; Neace and others, 1986).

As we come over and into Big Bend Ridge at 6124 feet, we come into the 1.3-Ma Henrys Fork caldera—the caldera which formed in response to the second cycle eruption from the Yellowstone Plateau volcanic field. The name “Henrys Fork caldera” (Christiansen, 2001) replaces the previous term Island Park caldera, first described by Hamilton (1965). The structure described by Hamilton (1965) is not a single caldera but rather forms a basin related to all three volcanic cycles in the Yellowstone Plateau volcanic field. The Island Park basin is flanked on its southern edge by Big Bend Ridge (representing segments of both the 2.05-Ma and the 1.3-Ma calderas), on its west by lavas from the first 2.05-Ma cycle and Thurmon Ridge from the 1.3-Ma caldera-forming event, and on its eastern flank by post-caldera lavas from the 3rd cycle, 0.64-Ma Yellowstone caldera event.

As we come into the basin floor, we cross the Henrys Fork caldera continuing north on U.S. 20. The 0.64-Ma Lava Creek Tuff for the Yellowstone caldera is exposed on the floor of the second-cycle caldera and to contribute to filling this caldera. Younger basalts, such as the 200-ka Gerrit Basalt, can be seen also filling the floor of the Henrys Lake caldera; eventually this caldera, like its buried equivalents on the Snake River Plain, will be mostly concealed by younger volcanic deposits. Based on the age of the 1.3-Ma Mesa Falls Tuff and the age and distribution of the intracaldera 200-ka Gerrit Basalt, it took approximately 1.0 million years for the rhyolitic magma chamber which erupted the Mesa Falls Tuff to become brittle enough to fracture and allow for the ascension of the lower basaltic magmas.

STOP 9: INSIDE THE HENRY'S FORK CALDERA, SECOND-CYCLE CALDERA OF THE YELLOWSTONE PLATEAU VOLCANIC FIELD (83 miles, Mark 255, 44.3311, - 111.44117)

This is a brief stop to look at the 200-ka Gannet Basalt covering the floor of the Henrys Fork caldera and features surrounding the Island Park basin. The Mesa Falls Tuffs erupted 1.3 Ma resulting in the collapse and formation of the Henrys Fork caldera.

Return to bus and continue driving north on US 20.

At the entrance to Henrys Lake State Park, turn left and drive a short way before road drops down to entrance station. Disembark bus and walk to fault scarp to the north.

STOP 10: CENTENNIAL FAULT IN HENRY'S LAKE STATE PARK (106 miles)

Disembark on surface of old alluvial fan surface just before road descends towards Park entrance station. Walk north across old alluvial fan to viewpoint at the top of strand of Centennial fault scarp about 9 m high (Falene Petrick, written., commun., 2007, site AR). We are standing on the southern of two east-west scarps with an intervening tilted "ramp relay". Carbonate coats from a badger hole on the older fan here have an average coat of 5.25 mm thick, favoring a Bull Lake age. For calcic soils in the region, gravels of the Pinedale glaciation have carbonate coats that average a little more than 1 mm. If the gravel is Bull Lake as suggested by Witkind (1972), this surface is about 140 ka old based on the age of Bull Lake moraines nearby in the West Yellowstone Basin (Pierce and others, 1976). Alternatively, assuming a constant rate of coat buildup, the age would be 5 times the Pinedale (15 ka) or about 75 ka.

The surficial geology is complex. East of here, Pinedale fan gravels may have been deposited on the downdropped side of the fault. An inferred northern fault strand offsets the northern margin of the ramp relay with surface offset of 10.8 m (Falene Petrick, written commun., 2007, site AP). The eroding north margin of the ramp relay has carbonate coats about 1.4 mm thick, compatible with a Pinedale age. Recent events here include: 1) faulting of the Bull Lake? fan, 2) deposition of the present ramp relay surface against the fault in Pinedale? time, and 3) probable faulting at the north margin of the ramp relay and tilting of the ramp.

The east-west-trending Centennial normal fault is about 67 km long. Along the Centennial Range front westward from here, Pinedale moraines and alluvial fan gravels of inferred late Pleistocene age are typically offset 8-11 m (Falene Petrick, written commun., 2007). The 2.05-

Ma Huckleberry Ridge Tuff on the north side of the Centennial Valley is tilted about 5 degrees southward into the fault whereas south of the fault this ignimbrite is tilted at the southern base of the Centennial Range (Christiansen, 2001, Plate 3) and may be the ignimbrite on the crest of the Centennial Range (O'Neil and Christiansen, 2001). In either case, this indicates about 1.8 km (6,000 ft) of offset over the last 2 Ma.

Return to bus and get back on U.S. 20 going north. Continue until turnoff for the town of Ennis, Montana. Turn left onto US 87.

After ~2 miles on US 87, at about 2 o'clock, you can see a fault scarp at the base of the Madison Range.

STOP 11: MADISON VALLEY OVERVIEW (125 miles, mark 256, 44.79203, -111.47407)

Structural deformation of the Huckleberry Ridge Tuff. Here the Huckleberry Ridge Tuff dips into the Madison Range fault and forms the grassy ridge to the west as well as the forested ridges farther in the distance (Christiansen, 2001). The forested bench to the northeast is also the Huckleberry Ridge Tuff that is dipping into the Madison Valley. In the distance to the north just east of the Madison River, the Huckleberry Ridge Tuff forms a high point in the Madison Valley with ~150 m of structural relief along longitudinally oriented dome in this Basin and Range Valley (O'Neill and Christiansen, 2004). The Madison Range front fault is clearly visible on the lower slopes of the Madison Range to the north. To east is the Madison landslide that was triggered and formed during the 1959 Hebgen Lake earthquake. The open bench lands are Missouri Flats, which is a terrace sequence of the Madison River (Lundstrom, 1986). The Madison River is one of four major rivers draining Yellowstone National Park.

About 3 miles farther on US 87, we cross the Madison River where the lowest terrace complex is 3-15 m above the Madison River and considered Pinedale in age based on weakly developed soil and carbonate coats <1mm (Lundstrom, 1986). This surface has abundant evidence of large floods most likely from glacially dammed lakes upstream, especially by the Beaver Creek glacier. Flood indicators include giant ripple-like features, huge boulders, and closed depressions and eddy scour marks (Lundstrom, 1986).

Continue on US 87, turning right on US 287 toward West Yellowstone. A terrace 25 m above the Madison River is also considered to be Pinedale based on soils and 1-2 mm carbonate coats

(Lundstrom, 1986). Such incision may represent ongoing uplift even on the downstream side of the Madison Range front fault. The Madison Valley fault scarp is near top of fan on left.

Continue ~6-7 more miles on U.S. 287 until we have a view of the headwall of the Madison Rockslide at 6500 feet elevation.

STOP 12-- EARTHQUAKE LAKE VISITOR CENTER. (133 miles, mark 257, 44.8311, - 111.42465). Fee of \$15 for bus and use of rest rooms. **We will stop here for lunch.**

The Madison slide occurred on the south side of the canyon in Precambrian rocks of gneiss, schist, and dolomite dipping into the canyon. The slide accompanied the 1959 Hebgen Lake earthquake. The slide crossed the canyon with enough momentum to climb the south wall to a height of 130 m above the riverbed. Air blast related to the slide blew people and cars away. Nineteen people were buried by the slide and a total of 28 perished from the earthquake. The slide dammed the Madison River to create Quake Lake. Heavy equipment was rapidly brought in to create a stable overflow channel and lower the level of Quake Lake.

The Madison mylonite zone occurs in Precambrian rocks 2 km north of here and across the valley from the slide and can be seen as dark grey, shiny mylonite fragments in the landslide. The Madison mylonite forms a major zone of crustal weakness that trends northeast and parallel to the track of the hotspot, but was not used or taken advantage of by the plume which formed the track. In fact, major crustal boundaries defined by both geologic and geophysical studies are identified here (30 km north of the track), farther north at the Great Falls tectonic zone (O'Neill and Lopez, 1985) 200 miles north of the plain, and similar zones of weakness south of the track 200-300 km south, but NOT on the track. In fact, no geologic or geophysical evidence supports any pre-existing zone of crustal weakness along the track of the Yellowstone hot spot and supports a deep origin for the hotspot track. Crustal flaws and crustal tectonics do not control the location of the hotspot track (see Pierce and Morgan 1992 for a full discussion of this and associated references).

Return to bus. We continue to head east on US 287. As we drive into the valley, the hilly topography on our right are moraines of Beaver Creek. This may be the glacial dam whose failure created the flood deposits noted earlier. About 1 mile farther east, we can look to the left of the Beaver Creek valley in the Madison Range. Beaver Creek drains some of the higher areas of the range. A couple miles later, we cross Cabin Creek and turn left to park near the Hebgen Lake fault scarp on Cabin Creek.

STOP 13: HEBGEN LAKE FAULT SCARP. (141 miles, mark 259, 44.87168, -111.34217)

Cabin Creek is at the northwest end of the August 17, 1959 surface rupture of the Hebgen fault. This fault lies in Belt II of major Holocene faults; the Madison fault is the most active fault nearby (Pierce and Morgan, 1992). The scarp near the creek offsets a 1-3-ka terrace and represents the actual displacement due to the 1959 quake. The 1959 earthquake created a waterfall on Cabin Creek. Some back-rotation below the fault also flattened the stream gradient there. No evidence of the 1959 earthquake is present in tree rings examined from trees at this location

Surveys were made of the scarps and a trench was dug across the fault scarp (Pierce and others, 2001; Pierce and others, 2003). There is a small graben at the foot of the scarp. The scarp offsets two different-aged terraces. The lower terrace is about 1-3 ka and has 3.1 m of surface offset from the 1959 event. The upper terrace is about 11-17 ka and is offset 5.3 m. In addition to the 1959 offset, the scarp offsetting the upper terrace includes a pre-1959 event with surface offset of about 2.2 m that occurred about 3 ka (cosmogenic ages on terraces from Jerome Van Der Woerd, written commun., 2001). Since the 1959 scarp was formed, Cabin Creek has built an alluvial fan on the hanging wall and incised the footwall.

Return to bus and drive by Hebgen Lake dam.

To the south across the lake is the Bull Lake moraine. Ash possibly correlative with the <200-ka tuff of Bluff Point from the Yellowstone Plateau (Christiansen, 2001) is found associated with moraines of Bull Lake age at "The Narrows" of Hebgen Lake.

About 10 miles farther on US 287, we cross the Red Canyon fan. The Red Canyon fault is to the northwest and behind the range that is fronted by the Hebgen fault. Offset of both faults occurred during the 1959 earthquake.

4300.9 As we continue east due west of Grayling Creek, landslide deposits are exposed. In about a half mile, we cross the Grayling Creek alluvial fan of Pinedale age. The Red Canyon fault offsets this fan and was recently trenched and studied by Kathy Haller and Mike Machette of the US Geological Survey.

Come to a stop sign at the intersection of US 287 and US 191. Turn left onto US 191 and head toward Bozeman. Road climbs through a succession of subdued moraines of Bull Lake age. Turn right into cemetery.

STOP 14: BULL LAKE MORAINES AT GNEISS CREEK TRAILHEAD (153 miles, mark 260, 44.79532, -111.10307)

Walk respectfully southward through the graveyard built into a meadow developed on Bull Lake moraine with classic muted forms. During the Bull Lake glaciation, the West Yellowstone Basin was nearly filled with ice flowing west off the Yellowstone Plateau. Note the basalt boulders nearly flush with the ground, and pebbles of obsidian. The ~110 ka-West Yellowstone rhyolite flow (Obradovich, 1992) forms the skyline south across the basin, and that flow advanced across the Bull Lake moraines. Meadows on the gently sloping Bull Lake moraine surface are outlined against the forest of Horse Butte. Based on combined K-Ar and obsidian hydration dating, these Bull Lake moraines are about 140 ka and correlate with Marine Isotope Stage 6 (Pierce and others, 1976).

The moraines have a loess mantle and support lush herbaceous meadowlands. Here on the west side of the greater Yellowstone glacial system, the older Bull Lake moraines extend 22 km to the west beyond the younger Pinedale moraines. In contrast on the north and east sides of the Yellowstone Plateau, Pinedale glaciers have generally overridden Bull Lake moraines. This is consistent with subsidence on the trailing older margin of the Yellowstone Crescent of High Terrain and uplift on the leading younger margin.

Excellent exposures of the Huckleberry Ridge Tuff are present farther east on the Gneiss Creek Trail as well as farther north U.S. 191. Head south on U.S. 191 toward West Yellowstone.

The forested flat terrain from here through West Yellowstone and beyond to the west entrance into Yellowstone Park is all on an obsidian sand plain. The sand plain heads in Pinedale moraines farther to the east and was probably deposited by floods, probably from recessional ice-dammed lakes in the geyser basins on the Firehole River. Obsidian hydration on granules from the sand plain is 7 μ m thick, the same as for Pinedale moraines (15-20,000 yrs). The bedding of the obsidian sand plain is sub-horizontal, with primary beds about 15 cm thick, capped by a thin (a few mm) silty zone. The gradient of the sand plain is very low, about 1.5 m/km. These and other features suggest that the obsidian sand plain was built by large sheet flow discharges carrying sand and granules in traction across a surface of wide extent. Soils derived from these obsidian-rich sands are well drained and poor in nutrients and support a monoculture of stunted

lodgepole pine trees. Antithetic normal fault offsets the obsidian sand plane and is manifest just west of airport runway by an alignment of a conical depression.

Passing through the town of West Yellowstone, we continue across the obsidian sand plain.

WEST ENTRANCE TO YELLOWSTONE NATIONAL PARK. (165 miles,)

Three miles east of the West Entrance booth, we see the margin of the 110-ka West Yellowstone rhyolite flow, a post-Yellowstone caldera rhyolitic lava flow. A terminal moraine of Pinedale age crosses the road and has new growth of lodgepole pine since the 1988 fires. More extensive end moraines are across the Madison River. To the right about 2 o'clock are low rounded ridges with lodgepoles on top, which are the outflow of the 0.64-Ma Lava Creek Tuff.

171.5 miles, MARK 262 As we drive along the Madison River, we come to a clearing straight ahead where the ridge on the north side is the pre-caldera Mt. Haynes rhyolite flow (part of the Mt. Jackson Rhyolite) and the ridge on the south side is the post-caldera West Yellowstone flow (110,000 yrs). This clearing is a segment of the topographic margin of the Yellowstone caldera. Another pre-caldera rhyolite flow (the Harlequin rhyolite flow) is at the base of the Mt. Haynes flow and is reversely magnetized, indicating its age is greater than 780 ka (Finn and Morgan, 2002). The 0.64-Ma Lava Creek Tuff and all the post-caldera lavas are normally magnetized (Reynolds, 1977; Harlan and Morgan, 2006, 2007). Cliffs to the north across the Madison River are exposures of near-source facies of the Lava Creek Tuff, member A.

The Lava Creek Tuff erupted from two smaller (<10 km in diameter) vents inside but near the edge of the later collapsed structures (figure 4b) associated with the caldera and has two members, distinguished from each other based on slight differences in mineralogy and trace element geochemistry (Christiansen, 2001). Based on the distribution of each member as reflected in isopach (Christiansen, 2001) and isopleth (Morgan and Christiansen, 19xx) maps, member A of the Lava Creek Tuff came from a vent near Purple Mountain near Madison Junction whereas member B erupted shortly thereafter from a vent east of Fern Lake in the northeast part of the caldera. The Lava Creek Tuff came from one of these two vents; no evidence exists for other vent sources for the Lava Creek Tuff suggesting that collapse of the caldera may have been piecemeal and a structural, non-volcanic, response to the massive evacuation of high-level magma chambers (Morgan and Christiansen, 1998). Fractures along the margin of the Yellowstone caldera were zones of weakness for the ascension of later rhyolitic lava flows. Over 30 volcanic events, mostly rhyolitic lava flow eruptions, comprise the last 200,000 years of volcanic activity from ~200,000 years to 70,000 years.

Very little time elapsed between successive eruptions from vent A and vent B. At exposures where member A is in contact with and beneath member B, no soils and little to no erosional surfaces are developed. Fall and surge deposits can be identified at various localities which grade upward into Member B of the Lava Creek Tuff. Based on stratigraphic position, eruption of member A was complete prior to eruption of member B.

As we continue driving east, we pass on the left the –reversely magnetized Harlequin rhyolitic lava flow.

At Madison Junction, we will turn right and take a quick break to visit the newly constructed facilities. Madison Junction represents the confluence of the Firehole River from the southeast and the Gibbon River from the northeast. From here, the Madison River starts at the confluence and flows west. After visiting the facilities at Madison Junction, we will turn left and head northeast toward Norris Geyser Basin, driving along the Gibbon River and the topographic margin of the Yellowstone caldera. We have arrived!

As we drive along the topographic margin of the Yellowstone caldera, on our left is Terrace Hot Springs, a CO₂-rich hot spring. Here, the carbonate-rich hot springs are due to subsurface water flowing through and reacting with with slide blocks of Paleozoic limestones. As we continue, we pass Tuff Cliffs, a massive cliff of Lava Creek Tuff member A. Much of the lower part of the cliffs consist of non-welded to incipiently welded ignimbrite. Higher in the section, the degree of welding increases. Columnar joints can be seen high up on the cliffs in the densely welded upper part of the Lava Creek Tuff.

We follow the road toward Norris much of which coincides with the topographic margin of the Yellowstone caldera. Outcrops of the Lava Creek Tuff dominate the landscape outside and north (on our left) of the caldera margin while post-caldera lava flows dominate the landscape inside the caldera. To our right, we start seeing the Nez Perce rhyolite; this lava erupted about 154 ka (Morgan and McIntosh, unpublished data, 2007). At pull offs on either side of the road, one can look back at Tuff Cliff and see the topographic margin of the Yellowstone caldera margin defined by the contact between pre-caldera Mt. Haynes rhyolite, the post-caldera West Yellowstone rhyolite, and the outflow facies of the Lava Creek Tuff.

We pass Gibbon Falls on our right (south). After Gibbon Falls, Lava Creek Tuff is exposed on both sides of the road so we are now outside but near the caldera margin. Hills around us are covered with the Lava Creek Tuff with thicknesses outside the caldera exceeding 300 m. As we head east, member A is now mantled with member B flowing in from the east.

Continuing toward Norris, we pass Beryl Springs on the left. Beryl Springs is a high-temperature, neutral Cl hot spring with high $^3\text{He}/^4\text{He}$ values (Fournier and Kennedy, xxxx). It is also one of the few hot springs in the park with significant gold in the sinter (1-1.4 ppm; Fournier, Kennedy, xxxx). Beryl Springs, located near the saturated groundwater table, is at the base of a northwest-trending fissure system that starts here outside the caldera and continues northwest. Monument Geyser Basin, 250 m above at the top of the ridge to the northwest, is on this same northwest-trending fracture but rather is located above a perched water table and is an acid steam system with unusual siliceous spires. Morgan and Shanks (2005) suggested these spires formed underwater in a glacially dammed lake which has since been breached leaving the spires rising above the extensive sinter terrain around Monument Geyser Basin. Similar spires are present today in Yellowstone Lake (Morgan and others, 2003). Across the road from Beryl Hot Springs and Monument Geyser Basin (here coinciding with the margin of the caldera) are exposures of the 154-ka Nez Perce rhyolite flow, the 90-ka Gibbon River flow (Obradovich, 1992) and the 116-ka Gibbon River dome. On our left at the top of the Gibbon River flow are well-bedded glacial sediments.

The peak straight ahead or slightly to the northwest is Electric Peak, the highest peak in the Gallatin Range. As we continue driving north to our left is the Gibbon Geyser Basin. Continue north and in the distance we get our first glimpse of the south end of Norris Geyser Basin, located just to the north of the topographic margin of the Yellowstone caldera and along the tectonically active north-south-trending Norris-Mammoth corridor. As we approach the junction for Norris Geyser Basin, exposures of Lava Creek Tuff member B are exposed on the east side of the road. Drive to stop sign and turn left into Norris Geyser Basin.

STOP 15: NORRIS GEYSER BASIN 194 miles:

We plan to stay at Norris for about 1 hour and will walk around the geyser basin. See Morgan and Shanks, 2005 for a discussion of location of geyser basins inside and outside the Yellowstone caldera.

Depart Norris Geyser Basin—Turn left at stop sign and head north toward Mammoth Hot Springs. As we drive north, we are now outside of the Yellowstone caldera and are traveling along the Norris-Mammoth tectonic corridor.

As we drive north, exposures of the Lava Creek Tuff are on our right and show extensive hydrothermal alteration. Samples analyzed for magnetic susceptibility show the Lava Creek Tuff to have extremely low values (10^{-6}). In contrast, fresh unaltered samples of Lava Creek Tuff have susceptibility value of 10^{-3} (Finn and Morgan, 2002). To our left are the new hydrothermal vents north of Nymph Lake. Initial activity of this vent system started in late summer of 2003. Structurally controlled and parallel to faults on the Norris-Mammoth corridor, the new vents can be easily identified along a north-south-trending fracture. Initially two vents were active on the slope. A vent located higher on the slope began as an acidic steam vent while a lower vent <15 m away was a neutral Cl hot water vent. Prior to 2003, this area was just another tree-covered slope.

Roaring Mountain is the white, steaming mountain on our right as we continue north (E). This is a significant acid steam system cut into the Lava Creek Tuff and aligned along the active Norris-Mammoth corridor. H_2S in steam is oxidized to H_2SO_4 by reaction with atmospheric O_2 or O_2 -saturated groundwaters. The high-silica rhyolitic composition of the Lava Creek Tuff is intensely altered to an argillic assemblage of kaolinite, quartz, alunite, and smectites. On top of cliffs are a series of large hydrothermal explosion craters- age of explosion is unknown.

Less than 1 mile north of the turn off for Roaring Mountain, we pass on the left (west) Semi-centennial pool, a hydrothermal explosion crater that formed in 1926.

STOP 16: OBSIDIAN CLIFF (203 miles, MARK 267, 44.82427, -110.72920) The $183\text{-ka} \pm 3\text{ ka}$ Obsidian Cliff flow (from Obradovich, 1992 as revised by Christiansen, 2001) is a glassy rhyolitic flow-banded obsidian. Major archeological excavations for obsidian are on top of the flow; obsidian from this flow is widely distributed in archeological sites. Obsidian knives as long as 25 cm from this flow are found in the Mound-Builder sites in Ohio. The flow is one of many extracaldera post-Yellowstone lavas in the Norris-Mammoth Corridor. The Norris-Mammoth Corridor is a north-south-trending set of faults along which are located vents from 14 young flows and many geothermal springs (White and others, 1988). The near-vertical volcanic vents would intercept the Gallatin Range-front fault at 8-12 km depth. Extension that once would have reached the surface by way of the Gallatin Range front may subsequently have occurred in the Norris-Mammoth Corridor (Pierce and others, 1991, Figure C-7). The origin of this corridor or trench now occupied by Obsidian Creek is difficult to explain by either glacial or stream action. This trench is old, for the Obsidian Cliff flow spills into this trench. . One hypothesis is formation above a blind dike.

As we drive north, we see to the west the peaks of the Gallatin Range including Mt. Holmes (a laccolith), Trilobite Peak, Quadrant Mountain, and Electric Peak at the northern end. The range dips north with Precambrian rocks at the south end and Cretaceous rocks with intrusives at Electric Peak. On the right (east), we pass the entrance to Sheepeater Cliffs, where we have physically mixed magmas with a bulk composition of andesite- actually mixed basalt and rhyolite. Continue north and come onto Swan Lake Flats (elevation 7317 feet). Here post-caldera basalts erupted along this zone of weakness outside of the Yellowstone caldera. Its magnetic remanence is normal (Finn and Morgan, 2002; Harlan and Morgan, 2006). Swan Lake, on our left, has been cored and analyzed for climate history studies. Farther north at the northern end of the Norris-Mammoth corridor, Bunsen Peak is on our right. Bunsen Peak is a Tertiary andesitic intrusive; the peak has glacial striations.

Ahead is Terrace Mountain mostly made up of a thick sequence of Huckleberry Ridge Tuff, from the first-cycle caldera in the Yellowstone Plateau volcanic field, and capped with travertine. The travertine at Terrace Mountain is nearly horizontal as it was deposited on a flat surface of Huckleberry Ridge Tuff. The travertine here is the oldest part of the sequence of travertines related to Mammoth Hot Springs, located at the northern tip of the tectonic corridor. Neill Sturchio obtained a U/Th age of 406 ± 30 ka.

As we descend toward Mammoth Hot Springs, we enter Golden Gate as the road cuts through this 2.05-Ma densely welded ignimbrite. Continuing down the road, we pass the "Hoodooos", a landslide with back-rotated blocks of travertine from Terrace Mountain. In the distance, we see Mt. Everts across the valley. This mountain of Cretaceous shale and sandstone is capped by the 2.05-Ma Huckleberry Ridge Tuff filling paleovalleys which in turn is capped by the younger 0.64-Ma Lava Creek Tuff filling paleovalleys cut into the older unit. Excellent exposures of the ignimbrites and underlying fall deposits can be seen as the capping red oxidized ash beds overlain by black vitric and densely welded ignimbrite. Continuing north we drive past the travertine terraces and hot springs at Mammoth. Hot water reacts with Mesozoic carbonates at depth, becomes charged with CO₂, and precipitates travertine upon cooling.

Optional STOP 17: OVERLOOK OF MT. EVERTS AND NORTH INTO PARADISE VALLEY

Park in lot across from Mammoth Post office behind the Mammoth Hotel. Behind the Hotel is the old unpaved north entrance road which climbs up to a great viewpoint.

Leave Mammoth Hot Springs and continue to drive north to the North Entrance and town of Gardiner, where we will stay for the night. As we drive toward the North Entrance, we pass

Boiling River on our right, a hot spring site on the Gardner River. Tracer tests indicate that flow from the sinkhole near Liberty Cap at Mammoth Hot Springs carries water to the site at Boiling River in 18 hours, a very rapid subsurface flow rate (David Janecky, written communication, 1998). Along the Gardner River canyon, mountain goats frequent the steep cliffs of Cretaceous shales. Note large boulder Holocene conglomerate on the left exposed in the river bank on the west side of the Gardner River buried by postglacial landslides.

Exit Yellowstone National Park through the North Entrance, drive past the Heritage Center and continue driving west out onto rolling terrain northwest of Gardiner, Montana.

STOP 18: GIANT FLOOD DEPOSITS: Flood bar near Stevens Creek (elevation at 5299 feet) MARK 270 45.0548, -110.76782

This mid-channel flood bar has “giant ripples” spaced about 15 m apart and up to 2 m high, with boulders up to 2 m in diameter. Cosmogenic (^{10}Be) ages on the boulders average 13.7 ± 0.5 ka. The flood-rippled fronts of alluvial fans of Reese Creek 0.5 km to the northwest provides evidence of three floods 45-60 m deep (Pierce, 1979). These floods were probably from release of glacially dammed lakes upstream, most likely form a lake dammed in the Lamar Valley by the Slough Creek recessional glacier.

Incision of the Yellowstone River valley is recorded by: 1) incision of ~2-Ma basalt (probably equivalent in age to the basalt of Junction Butte, a pre-Huckleberry Ridge Tuff basalt) by ~400 m, and 2) incision of ~0.650-Ma Undine Falls basalt by 200 m. During the last glaciation, the valley may have been over-deepened by ~100 m below its present level.

This site is where glaciers from the east, south, and west converged to form the northern Yellowstone outlet glacier more than 1 km thick, and terminated about 50 km downvalley. About 30 km down valley, the ~2-Ma Emigrant basalt is exposed and is notable in that olivine phenocrysts from the basalt yield magmatic $^3\text{He}/^4\text{He}$ values ranging up to 21 times that of atmospheric (Mark Kurtz and Joe Licciardi, written commun., 2007), representing undegassed and possibly a deep mantle source.

Return to Gardiner and stay at Wilson's Yellowstone River Motel. All are on their own for dinner; several restaurants are within walking distance.

DAY 3 (October 25, 2007)

LEAVE GARDINER PROMPTLY AT 7:30 AM. Drive south back into Yellowstone National Park through the North Entrance. This morning we will backtrack some of our route from yesterday given that our alternative route over Dunraven Pass is closed due to snow.

STOP 19: SWAN LAKE FLATS (mark 268, 44.9157, -110.73055).

Pull in at Swan Lake parking area.

The Gallatin Range is 7 km to the west from here. Between the highway and the Gallatin Range is Swan Lake Flats and farther west Gardners Hole. Both Swan Lake Flats and Gardners Hole are underlain by glacial deposits covered with a meadowland of steppe vegetation. Gardners Hole is a dramatic landscape of glacial drumlins formed by glacial flow northeast towards Gardiner (Pierce, 1979). Glacial ice was about 2,200 ft (670 m) thick here. Farther south from here are landscapes mostly underlain by rhyolite and vegetated by lodgepole pine. Glacial deposits support meadowland vegetation because they have adequate nutrients, are fine grained and hold the snowmelt water at shallow depths. Rhyolite lacks many nutrients, forms sandy soil that permits deeper percolation of water, and is vegetated by lodgepole pine which can survive under these nutrient and permeability conditions. Continue south toward intersection with Norris Geyser Basin. Turn left heading east toward Canyon. Here we are driving near the northern margin of the 0.64-Ma Yellowstone caldera. From Norris, the road passes through outflow facies of the Lava Creek Tuff. As we continue east, we drive through pass Virginia Meadows we come into the caldera driving over the 110-ka Solfatara Plateau rhyolite flow (Obradovich, 1992). Note the undulating surface with obsidian sands as we drive on top of this flow; at several sites along this way, hydrothermally cemented sediments are exposed associated with steam vents. Forests here in the rhyolites are dominated by lodge pole pines. Continue east descending off the steep flow front of the Solfatara Plateau flow. North of the road are outflow sheets of the Lava Creek Tuff whereas to our right (south) we are looking into the caldera. Come to stop sign and turn left driving several miles north toward Mount Washburn and Dunraven Pass. For the first couple miles, to the immediate east are post-caldera lavas that erupted shortly after the cataclysmic eruption of the Yellowstone caldera. These rhyolitic lavas (the Dunraven and Canyon flows of the Upper Basin Member of the Plateau Rhyolite) are 0.486 ± 0.042 Ma and 0.484 ± 0.010 Ma, respectively (Gansek and others, 1996). As we drive north and out of the Yellowstone caldera, Tertiary volcanic rock of andesitic compositions are to our north and west. Note the diversity of

vegetation developed on these soils in contrast to the monotonous lodge pole pine forests which dominate the rhyolitic terrain. Lodgepole pine forests flourish on rhyolitic terrain surviving well on sandy soils with poor nutrients. Pull over at parking area with overview of Yellowstone caldera.

STOP 20: OVERVIEW OF YELLOWSTONE CALDERA FROM MT. WASHBURN

(mark 277, 44.7677, -110.45525)

We are standing on the south-facing slope of Mount Washburn, a Tertiary stratovolcano composed of andesitic lavas, debris flows, and subordinate ignimbrites. From here, we can look from the north edge of the caldera about 45 km across to its southern rim, at the base of Flat Mountain. The Yellowstone caldera is one of the largest in the world at ~75 km (northeast-southwest dimension) by ~45 km. Looking south into the Yellowstone caldera, we can look southeast into the Canyon of the Yellowstone River. The river cuts through the 0.479±0.010-Ma tuff of Sulfur Creek (Gansek and others, 1996), and the Canyon flow, both of which are extensively hydrothermally altered. These flows have very low $\delta^{18}\text{O}$ values associated low $\delta^{18}\text{O}$ magma formed by melting of hydrothermally altered rhyolite (Hildreth and others, 1984, Bindeman and Valley, 2001).

As we look farther south into the caldera, we are looking at the high Yellowstone Plateau, a landscape filled with much younger (<200 ka) post-caldera rhyolitic lava flows. The morphology of these units is distinctive in their steep flow fronts, hummocky tops, and extensive nature. These rhyolitic lava flows are some of the largest on Earth having volumes in excess of several tens of cubic kilometers, extending up to 40 km from their source vents, and having thicknesses in excess of 100 m. These lava flows fill the caldera. In the distance, you can see Yellowstone Lake, the largest high-altitude (ie., >7000 feet elevation) lake in North America.

During the last glaciation (Pinedale), ice from the north advanced as far south as Canyon Village. An icecap built up on the Yellowstone Plateau to a height where it began to flow back toward the north across the top of Mount Washburn. Late in that glaciation, the ice-cap on the Yellowstone Plateau down-wasted and ice from the north now reversed its direction once again and flowed to the south; however this glacial advance did not reach as far south as Canyon Village.

Return to bus and backtrack driving south to Canyon Village. We will stop here for a brief break before continuing south toward Lake Village.

Continue driving south toward Lake Village along the Yellowstone River. Depending on time and accessibility, we may pull into Artists' Point to have a glimpse of the Lower or Upper Falls of the Yellowstone River. Continuing south into the Yellowstone caldera, the river cuts between the Solfatara Plateau rhyolite on the west and the Hayden Valley rhyolite on the east. The Solfatara Plateau rhyolite is a subaerial lava flow forming steep slopes and consists of a series of thick and dense individual flow units; in contrast, the Hayden Valley flow is exposed along the east bank of the Yellowstone River as gentle receding slopes composed of very fragmented perlitic rhyolite. We infer that the Hayden Valley rhyolite may have been emplaced when the valley floor was glacially occupied.

As we continue south along the Yellowstone River, we come into Hayden Valley, the serengeti of Yellowstone, known for its vast meadows developed on glacial lake sediments. Ancestral Lake Yellowstone (Love and others, 2007) occupied this valley during glacial times. The soils developed on these glacial lake sediments are rich in nutrients, have a much better water holding capacity, and support steppe meadowlands and a diverse fauna.

To the southeast, we see the Sour Creek dome coming into view. Note the fault-bounded blocks present on the dome which form a rectilinear pattern, similar to that present at Juniper Buttes on the eastern Snake River Plain. Sour Creek dome is one of two resurgent domes associated with development of the Yellowstone caldera. The Sour Creek dome is capped by both an intracaldera facies of the Lava Creek Tuff and the older 2.05-Ma Huckleberry Ridge Tuff. These units were in the subsurface prior to being domed up shortly ($<100,000$ years) after the cataclysmic eruption of the Yellowstone caldera and represent parts of the caldera floor which remained intact following the eruption. Both members of the Lava Creek Tuff (A overlain by B) are exposed here. The other resurgent dome is the Mallard Lake dome to the southwest and also has rectilinear faulting characteristic of resurgence. At Mallard Lake dome, a series of rhyolitic lava flows have been emplaced beginning with eruption of the 0.516 ± 0.007 -Ma Biscuit Basin flow (Gansek and others, 1996), followed by eruption of the 0.198 ± 0.008 -Ma Scaup Lake flow, and later by eruption of the 0.151 ± 0.004 -Ma Mallard Lake rhyolite flow (Obradovich, 1992). The Yellowstone caldera is unusual in that it has 2 resurgent domes, each whose history and composition differ somewhat. Elephant Back Mountain with its predominant northeast-trending fractures is located between Sour Creek dome to the northeast and Mallard Lake dome to the southwest and shows signs of active deformation of the Yellowstone caldera (Dzurisin and others, 1991).

We will slow down the bus as we drive past LeHardy Rapids on our right. Note the rapid change in stream gradient of the Yellowstone River. Located on the Yellowstone River between Sour Creek dome and Elephant Back Mountain, LeHardy Rapids acts a northeast-trending hinge fault associated with the active deformation of the caldera. Bedrock (densely welded Lava Creek Tuff) and serves as a threshold that controls the level of the Yellowstone River and Yellowstone Lake upstream.

STOP 21: OVERVIEW OF FISHING BRIDGE AND POST-GLACIAL CALDERA

UNREST. (mark 283, 44.5676, -110.38415)

Park in bus-parking area on west side of Fishing Bridge, cross road and climb up stairs to viewpoint.

Between 1923 and 1985, the center of the Yellowstone caldera rose ~1 m (Dzurisin, 1991) and from 1985 to 1995-6, it subsided at about 2 cm/yr (for here and other places, please see references in Pierce, 2002). More recent radar interferometry studies show renewed inflation (Wicks and others, 1998). Doming along the northeast-trending caldera axis reduces the gradient of the Yellowstone River from Le Hardys Rapids to the Yellowstone Lake outlet and ultimately causes an increase in lake level. Figure 7 shows the history of oscillating lake level changes and other events (Pierce and others, 2002). This “heavy breathing” of the central part of the Yellowstone caldera may reflect a combination of several processes: 1) Magmatic inflation, 2) tectonic stretching and deflation, and 3) inflation from hydrothermal fluid sealing, followed by cracking of the seal, pressure release, and deflation. Paleoshorelines descend towards the caldera axis indicating that although inflation has periodically occurred, no net buildup of the center of the caldera has occurred. Thus, the mechanism of hydrothermal inflation and deflation is preferred.

Stop briefly at rest stop at Fishing Bridge Visitor Center. If open, we will look at the 3-d, solid terrain model of Yellowstone Lake based on recent high-resolution bathymetric mapping (Morgan and others, 2003).

Ma Yellowstone caldera where heat flow values commonly exceed $1,500 \text{ mW/m}^2$ (Morgan and others, 1977). A few are present outside the caldera along the tectonically controlled, north-trending Norris-Mammoth corridor (such as Semi-centennial crater) and along the East Mount Sheridan fault in the south. Large explosion craters in Yellowstone occur on land or on the floor of Yellowstone Lake, and some of the older subaerial craters may have formed beneath glacially dammed lakes (Muffler and others, 1971; Morgan and others, 2003).

Hydrothermal explosions are violent local events which may occur with little or no warning, and thus are potential natural hazards. In many cases, mineral deposits containing gold, uranium, mercury, lithium, and other strategic minerals are known to be associated with hydrothermal explosion deposits (Sillitoe and others, 1984; Sillitoe, 1985; Silberman and Berger, 1985; Nelson and Giles, 1985; Vikre, 1985). Compared to other hydrothermal features such as geysers and hot springs in Yellowstone National Park, large hydrothermal explosions are relatively uncommon events occurring on the order of about every 700-1000 years over the past 14,000 years (Morgan and others, in press; fig. xxx). Small historic hydrothermal explosion events sometimes have been documented in the Park (Fournier and others, 1991; Marler and White, 1975), whereas others are noted but not described in sufficient detail to interpret the processes and triggering mechanisms. However, none of the very large explosion events resulting in features such as Indian Pond have been witnessed and recorded; thus much uncertainty remains about precursor signals and processes leading up to an explosion.

Large explosions, which are certainly less destructive than catastrophic events such as caldera-forming eruptions and large magnitude earthquakes, are still a potentially significant hazard. The relative frequency of large hydrothermal explosions warrants monitoring of potentially explosive systems (Lowenstern and others, 2005). Ejected large blocks of rock and fine material, boiling water, steam, and mud, and the associated flow of muddy material several meters in thickness hundreds of meters from source constitutes a significant hazard that could affect a local area of many square kilometers. Understanding precursors and potential triggers to explosions and the processes involved are factors necessary in anticipating their formation.

Most of the large subaerial hydrothermal-explosion craters, such as Indian Pond, are now occupied by lakes (Christiansen, 1974; Richmond, 1973; U.S. Geological Survey, 1972). All show topographic or bathymetric evidence of nested craters within the larger parent crater indicative of multiple cratering events over a period of time. Many of the smaller craters within the parent craters may have formed from a variety of processes including multiple steam explosions and collapse due to dissolution of the underlying material beneath the siliceous

hydrothermal cap rock (Shanks and others, 2005). Some of these lakes are perched above the normal groundwater level reflecting a lake floor of low permeability clays and silica.

Several large (>500 m) craters have been mapped in Yellowstone Lake, and including: the (6 ka?) Evil Twin explosion crater in the western West Thumb basin, the (>10 ka) Frank Island crater in the south central basin, the >8-ka Elliott's crater in the northern basin (Johnson and others, 2003; Morgan and others, 2003) and the 13-ka Mary Bay explosion crater (fig. 5; Pierce and others, 2002; Wold and others, 1977; Morgan and others, 2003). Additional large hydrothermal craters also occur east and southeast of Frank Island.

Storm Point hydrothermal dome

Located along the northern shore of Yellowstone Lake, Storm Point stands above the surrounding terrain and at least 20 meters above lake level. Storm Point is a structural dome related to a large hydrothermal center. Doming is evident in high-resolution LIDAR data (Pierce and others, 2002) and in the diversion of stream channels away from these uplifted areas. The structure measures approximately 840 meters by 795 meters and has multiple craters exposed on its top (fig. 23). Dips measured on bedded, cemented beach sands and gravels and altered laminated lake sediments on the eastern, southern, and western edges of the dome range from 8-15° East, 15-22° South, and 5-8° West, respectively. West of Storm Point, the 8-ka S2 shoreline and younger shorelines are tilted away from Storm Point about 6 m over a distance of one kilometer (Pierce and others, 2002). Deformation associated with doming of Storm Point area is estimated to have occurred 4 to 6 ka (Pierce and others, 2002).

At least thirty hydrothermal craters occupy the top of the Storm Point dome; individual craters range in diameter from approximately 5 meters to as large as 80 meters. Many are compound craters with smaller craters nested in a larger parent crater. The top surface of the eastern half of the Storm Point dome is mostly bare, has little vegetation except for some grasses and *Dichanthelium lanuginosum* (Stout and Al-Niemi, 2002), a broadleaf flowering species known as hot spring panic grass that is common in many thermal areas in Yellowstone. Most of the craters at Storm Point cannot be unambiguously interpreted as subaerial hydrothermal explosion craters because very little hydrothermal explosion breccia is present and most craters do not have a rim of ejecta around their edges. It is possible that if the ejected material were sinter, as is typical around smaller craters, it would disintegrate rapidly. Marler and White (1975) described large sinter blocks ejected from Link Geyser in Upper Geyser Basin, disintegrating within a decade. However, the substrata at Storm Point are mostly resistant silicified gravels and lake sediments and thus, alternatively, most craters may have formed either by collapse and/or

dissolution as supported by the absence of ejected blocks. Some craters are rimmed by isolated blocks of slightly tilted hydrothermally cemented beach sands and sediments and contain no breccia fragments. Fine, well-sorted eolian sands cover the floors of the craters.

Some thermal activity continues at present. Active hydrothermal vents occur at the edges of the Storm Point dome in the lake and the floor of many of the craters on Storm Point are warm to the touch and retain little snow in winter despite a heavy snow cover nearby. Temperatures as high as 68°C have been measured with shallow probes in the sandy, eolian crater fill; typical temperatures range between 18°C and 56°C.

Return to bus and backtrack driving now west on the East Entrance road until the stop sign to turn toward Lake Village. Turn left and drive south along the western side of Yellowstone Lake. To the west is Elephant Back Mountain. Continue driving past Lake Village and past Bridge Bay. In the recent high resolution bathymetric mapping of Yellowstone Lake, 12-18 siliceous spires were identified in Bridge Bay. These unusual conical-shaped structures are composed of amorphous silica and range in height from >1 m to >8 m and are up to 10 m at their base and are hydrothermal in origin. Scanning electron microscopic images of the spires show them to be composed of filamentous bacteria and diatoms (Shanks and others, 2005).

As we pass Pumice Point going west, we look south into the West Thumb basin of Yellowstone Lake. West Thumb is a smaller caldera, about the diameter of Crater Lake (Oregon), within the Yellowstone caldera that erupted about 173- to 191-ka (Morgan and McIntosh, unpublished data, 2007; Morgan and Shanks, 2005; Calvert, unpublished data, 2007). A series of post-caldera lavas, discovered during the recent mapping of Yellowstone Lake, flow into the Lake and form the main features on the floor of the northern two-thirds of the lake. Prior to this research, the morphology on the floor of Yellowstone Lake was attributed to glacial features. In fact, glaciolacustrine sediments merely mantle the rhyolitic lava flows on the lake floor. Prior to eruption of the West Thumb caldera, the Dry Creek rhyolitic lava flow exposed to the west, represented the extent of younger post-caldera volcanism. The tuff of Bluff Point whose source is attributed to be the West Thumb caldera, is one of the few ignimbrites erupted during post-caldera volcanism which was dominated primarily by rhyolitic lava flow eruptions. As we continue south along the west side of the West Thumb basin, excellent exposures of the tuff of Bluff Point are present on our right. Most of the tuff of Bluff Point is distributed to the west of West Thumb; additional exposures are along the eastern side of Shoshone Lake and the northern shores of Lewis Lake.

On our left is Potts Springs Geyser Basin on the west side of West Thumb Geyser Basin. Immediately ahead and slightly to the southwest (about 1:00) is the slope of the Duck Lake hydrothermal explosion crater whose dimensions are approximately 700 meters by 500 meters. The crater is rimmed with an apron of hydrothermal explosion breccia that built the slope in front of us. Duck Lake is located along the edge of the Dry Creek rhyolite lava flow about 0.6 km northwest of the West Thumb Geyser Basin. Even though Duck Lake is at the edge of the West Thumb Basin of Yellowstone Lake, water in the lake is perched at a level that is approximately 15 meters above Yellowstone Lake level. The pH measured at thermal springs along the south edge of the crater is 4.9. The lake has several active hydrothermal vents on its bottom as evidenced by bubbles rising to the lake surface. The explosion breccia deposit from Duck Lake crater is composed of subangular, silicified clasts containing cemented beach sands, sinter, lake sediments, obsidian fragments, and hydrothermally altered pumice. Many of the clasts look like they may be derived from the tuff of Bluff Point, exposed nearby.

A newly discovered 500-m-diameter sublacustrine explosion crater is 300 m southwest of Duck Lake in West Thumb basin and is referred to informally as the Evil Twin explosion crater. In the western part of West Thumb basin, heat-flow values have been measured as high as 1,500 mW/m² (Morgan and others, 1977) reflecting the hydrothermal activity that contributed to the formation of the offshore explosion crater. The 500-m-wide Evil Twin explosion crater has 12- to 20-m-high, nearly vertical walls and has several smaller nested craters along its eastern edge. These nested craters are as deep as 40 m and are younger than the main crater. Temperatures of hydrothermal fluids emanating from the smaller northeast nested crater have been measured at 72°C by the submersible ROV.

Stop 23: WEST THUMB GEYSER BASIN (Mark 291, 44.41573, -110.57453)

Turn left into the parking area for West Thumb Geyser Basin. Restroom facilities are available as is a nice bookstore. We will spend about 45 minutes here walking around the geyser basin.

A rise of Yellowstone Lake had drowned sinter aprons whereas the vents are still active.

Return to bus. Turn left onto road and proceed to stop sign. Turn left heading south toward Jackson. As we drive south to the west under the bridge are hydrothermally cemented beach sediments. We continue south driving on top of the mostly subdued topography of the Aster Creek rhyolite flow, emplaced shortly after eruption of the tuff of Bluff Point. Driving along the east side of Lewis Lake, look west to the thermal areas on the west shore of Lewis Lake as well

as those higher up at Shoshone Lake. Near the southern end of Lewis Lake, the older Aster Creek rhyolite flow comes into contact with the younger 70-ka Pitchstone Plateau rhyolite flow, representing one of the last eruptions from the Yellowstone Plateau. The Pitchstone Plateau flow is an excellent example of large rhyolite lava morphology with steep flow fronts, talused slopes, and hummocky tops. In places, the Pitchstone Plateau flow is ~100 m thick. As we continue south to the east we look outside the Yellowstone caldera toward the Red Mountains caldera segment of the 2.05-Ma Huckleberry Ridge caldera from where member A of the Huckleberry Ridge Tuff erupted. We continue driving south cutting through outflow facies of the densely welded Lava Creek Tuff member B. As we drive along the Lewis River, look down into the steep-walled canyon where exposures of the pre-caldera, reversely magnetized Lewis Canyon rhyolite flow are exposed below the Lava Creek Tuff. As we continue south, we pass another pre-caldera rhyolite, the Moose Creek Butte rhyolite flow.

Leave Yellowstone National Park at the South Entrance where the Lewis River flows into the Snake River. Enter the John D. Rockefeller Memorial highway. On the sides of the road in this stretch, outflow sheets of the Lava Creek Tuff member B. As we cross the Snake River, we see on the south bank an excellent exposure of Lava Creek Tuff overlying a non-welded ash-fall deposit that grades upward into a welded ash grading upward into the welded body of the ignimbrite.

ENTER GRAND TETON NATIONAL PARK

STOP 24 JACKSON LAKE PICNIC AREA (mark 298, 43.98183, -110.66248, also toilet and place to get out of bus after long ride from West Thumb) Pull over on right just beyond the long strip parking area along Jackson Lake. REST STOP.

The reservoir here covers the Snake River delta natural Jackson Lake where the dam can raise water level by as much as 11 m. When the lake was low during rebuilding of the Jackson Lake extensive archeological and geoarcheological studies were made. Glacial scour excavated Jackson Lake which is 120 m deep and has as much as 120 m additional sedimentary fill (Smith and others, 1993. On the bench 500 m above the reservoir from here are Pinedale moraines of the outlet glacier that flow south from the Yellowstone Plateau.

Post-glacial scarps on the Teton fault are locally apparent near the base of the Teton Range. Events in the natural delta indicate the last two offsets on the Teton fault: 1) a submergence event

back flooding the delta about 2 ka, and 2) formation in deltaic sediment in ridge and basin terrain about 4 ka apparently by jolting during fault offset (Pierce and others, 1998).

STOP 25. WILLOW FLATS VIEWPOINT. (mark 299, 43.87265, -110.57362

To the south on Signal Mountain two ignimbrites are tilted into the Teton fault. These units combined with the tilt of ~10 Ma Teewinot Formation define movement on the Teton fault (Figure ____). The ~10 Ma Teewinot is tilted about 22° similar to the tilt in the 4.45-Ma Kilgore Tuff, which indicates little or no deformation occurred between 10 and 4.45 Ma. The Huckleberry Ridge Tuff is tilted 11° indicating the tilting between 4.5 and 2 Ma was similar to that over a similar time between 2 Ma and present Figure ____). The post-glacial rate of tilting based on the offset of Pinedale moraines and a tilting model indicates a similar rate of faulting in post-glacial time and in post-Huckleberry Ridge time.

STOP 26. CATHEDRAL GROUP OVERLOOK.

The Teton fault scarp offsets moraine of Pinedale “slab” glaciers heading high on the Teton front as well as steep slope colluvium. The fault scarp is as steep as 39° in cohesive glacial till and colluvium. The scarp is 34 m high and surface offset is about 19 m. (Gilbert and others, 1983).

The Teton fault is in fault Belt II that is defined by major Quaternary faults. The Teton and associated en echelon faults are adjacent to the 2 Ma-to-present location of the Yellowstone volcanic hotspot. The high, fresh, abrupt range front reflects the high rates of young faulting. Range fronts elsewhere with such morphology probably indicate similarly large, young fault offset.

Return to bus and continue east. To the north is Pilgrim Peak where the 4.45-Ma Kilgore Tuff, whose eruptive source is on the eastern Snake River Plain, is exposed. First identified as the Conant Creek Tuff, whose source is also on the eastern Snake River Plain in the Heise volcanic field, by Love and Christiansen, 1988, it has been later correlated with and is distinct from the Kilgore Tuff (Morgan, 1992). Farther south, the Kilgore Tuff is exposed along the western front of the Gros Ventres Range and at Signal Mountain. Implications derived from the eastern extent of the distribution of the Kilgore Tuff suggest that the northern Teton Range could not have been a significant topographic feature at the time of emplacement of the Kilgore Tuff (Morgan and McIntosh, 2005).

Leave Grand Teton National Park. Turn left at stop sign and head toward Dubois, Wyoming. We will drive to the **Hatchett Resort** where we will stay for the night. A special dinner will be prepared for us for ~\$30-35 each.

LAST DAY (October 27, 2007)

RETURN TO DENVER.

Leave Hatchet Motel at 7:30 am, drive east on US 287.

STOP 27: TOGWOTEE MOUNTAIN LODGE: NEW FAULT, 7.8 miles east from Hatchet Motel.

“New” basin and Range fault 0.7 miles west of Togwotee Mountain Lodge. A post-glacial fault scarp faces east and has a sag-pond at its foot. The scarp is about 10 m high, has a maximum scarp angle of about 40°, and is part of a set of scarps that are recognized for at least 3 km to the south. A related fault 1.5 km south-west of here offsets Bull Lake moraines 20 m. This fault is in Belt I of “Lesser Quaternary faults” characterized by post-glacial offset but total offset only minor as indicated by small escarpments. It is on the leading edge of faulting associated with the Yellowstone hotspot and considered basin and Range faults early in their offset history. (or reactivated on the west side of the hotspot). In a few million years perhaps an escarpment like the Teton front will develop in this area.

At Scenic off-road parking area 0.4 miles beyond Togowotee Mountain Lodge, drive up to view point east to Teton Range. Note bulky moraine where we turned off the highway. This is a large moraine of the second phase of the Pinedale glaciation (Pinedale-2 or Hedrick Pond).

Continuing east another mile and continuing for the next 5, we come into Black Rock Meadows. In the early phase of the Pinedale glaciation, an icecap-like glacier 150 m thick filled Black Rock Meadows and poured westward into Jackson Hole where its margin is at the Snake River Overlook.

About 90 miles from Hatchet Motel and 40 miles past Dubois.

The Bull Lake Reservoir is to the southwest of the highway. Centering in this area over a distance of 55 miles (90 km), terraces of the Wind River are consistent with tilting away from

Yellowstone (Figure 8). A 640 ka terrace rises towards Yellowstone 100 ft (30 m) more than

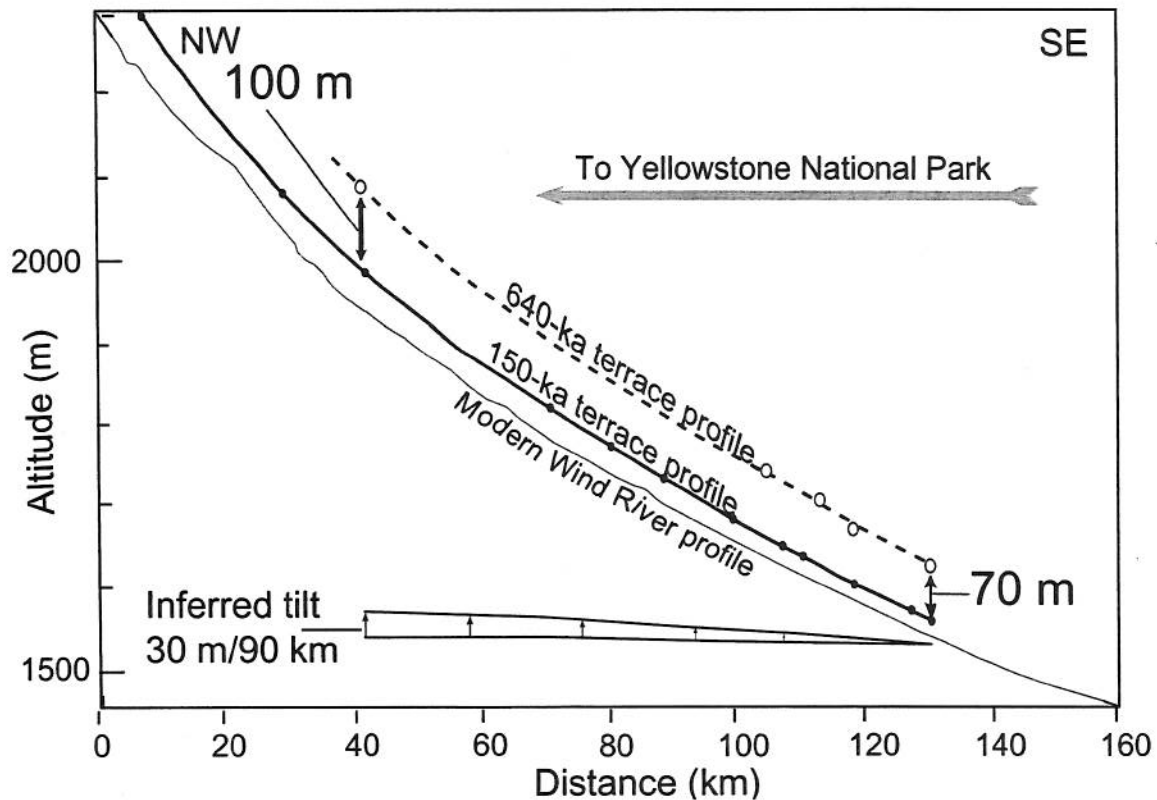


Figure 8. from Pierce and Morgan, 1992

does a lower terrace dated 150 ka. Such tilting over the full age and distance associated with the hotspot uplift computes to about 1,000 m total hotspot uplift (Pierce and Morgan *in* Saunders and others, 2006). Similarly, for terraces in the Bighorn Basin, older terraces are more tilted away from Yellowstone than younger ones for streams flowing both away and towards Yellowstone. Uplift on the on the leading margin of the Yellowstone hotspot is consistent with this tilting of these terraces.

On to Denver!!

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