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Notes

Neogene tephra correlations in eastern Idaho and Wyoming: Implications for Yellowstone hotspot-related volcanism and tectonic activity

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ABSTRACT

The explosive rhyolitic eruptions that define the track of the Snake River Plain–Yellowstone volcanism have produced a large volume of tephra found in late Miocene and younger basin-fill sediments throughout the western United States. Here we use ⁴⁰Ar/³⁹Ar isotopic dating, paleomagnetic analysis, major- and trace-element geochemistry, and standard optical techniques to establish regional tephra correlations. We focus on tephra deposits in three Neogene basins in spatially separated areas—Grand Valley, in eastern Idaho; Jackson Hole, in northwestern Wyoming; and the Granite Mountains area, in central Wyoming. These basins have experienced relatively continuous deposition from the late Miocene to the Holocene. We found tephra layers that directly tie the stratigraphy between all three basins. Using these correlations we found that basins experienced discrete pulses of extension separated by long periods of relative quiescence, the dates of which are staggered between basins. An early pulse of extension occurred in Grand Valley and Jackson Hole between 10.3 Ma and 16.33 Ma with a second pulse initiating between 4.54 Ma and 2.09 Ma. The Granite Mountains basin experienced a single pulse of extension sometime between 11.14 Ma and 6.75 Ma. These pulses of accelerated extension, along with evidence of similar pulses in other basins, present a pattern of west-to-east migration that we suggest is related to the Yellowstone hotspot. The later pulse of activity in Grand Valley and Jackson Hole corresponds to the migration of the North American Plate over the tail of the Yellowstone hotspot. We speculate that the earliest pulse in each basin is related to the more rapid movement of the sublithospheric hotspot head as it spreads

out from its earliest known location, where the Columbia River Plateau Flood Basalt Province initiated in southeastern Oregon, to its outermost edge under central Wyoming. Our results are consistent with this model of a plume head, though not unique to it.

The results of our study also indicate that previous suggestions that the rate of Snake River Plain explosive volcanism has decreased by a factor of 2 or 3 since emplacement of the middle Miocene Trapper Creek tuffs likely underestimate post-Trapper Creek eruption rates. We have discovered a large number of previously unidentified post-middle Miocene major eruptive events, both as ash-flow tuffs and as vitric air-fall tuffs. Recalculation to include these newly discovered events results in a rate of major eruptions that is fairly uniform until ca. 4.54 Ma. However, there is a substantial gap in major silicic eruptions in the interval between 4.54 Ma and 2.09 Ma, which we call the “post-Heise eruptive gap.” With the exception of this gap, the rate of major eruptions on the Snake River Plain has been roughly constant since inception of the eastern Snake River Plain–Yellowstone volcanic track between 16 Ma and 17 Ma.

INTRODUCTION

A northeastward-propagating track of explosive silicic volcanism, with eruptions beginning on the Oregon–Nevada border and progressing across the Owyhee Plateau and the eastern Snake River Plain to the Yellowstone Plateau (Fig. 1), has long been thought to result from the motion of the North American Plate over a stationary plume or hotspot (Morgan, 1972; Armstrong et al., 1975; Suppe et al., 1975; Leeman, 1982; Anders et al., 1989; Rodgers et al., 1990; Anders and Sleep, 1992; Pierce and Morgan, 1992; Smith and Braile, 1994; Camp, 1995; Camp and Ross, 2004). The commonly accepted hotspot model is one in which a large plume head is fed by a narrow plume tail (e.g.,

Lecheminant and Heaman, 1989; Richards et al., 1989; White and McKenzie, 1989). In this model the plume head impinges on the base of the lithosphere resulting in production of a massive volume of basalt in a short interval (on the order of one million years). In the model the head spreads out along the base of the lithosphere for a distance of over 500 km. The narrow feeder tail (tens of kilometers in diameter; e.g., Sleep, 1990) remains fixed in the asthenosphere as the lithospheric plate moves, producing a volcanic track on the lithosphere. In this classic model as applied to the Yellowstone hotspot (see Pierce et al., 2002), the head is inferred to have produced the Columbia River Flood Basalt Province while the tail remained roughly fixed with respect to the asthenosphere and produced the volcanic track initiating at 16.6 Ma and culminating with the 649 ka Lava Creek eruption in the Yellowstone Plateau volcanic field. This view of the origin of the Columbia River Flood Basalt Province and Snake River Plain–Yellowstone volcanic system is not universally accepted (Hamilton and Myers, 1966; Christiansen and McKee, 1978; Hamilton, 1989; Humphreys, 1995; Humphreys et al., 2000; Christiansen et al., 2002; Tikoff et al., 2008). Although the track of volcanism clearly follows the eastern Snake River Plain, there is debate as to which eruptive events might or might not be related to hotspot activity prior to the first silicic eruptions on the eastern Snake River Plain (Duncan, 1982; Pollitz, 1988; Pierce and Morgan, 1992; Geist and Richards, 1993). A problem relating to the head model for the Columbia River Flood Basalt Province is the apparent mismatch of the presumed location of the track at ca. 16 Ma and the geographic center of the basalt province at ca. 17 Ma (see Christiansen et al., 2002). Moreover, there is the “Newberry trend” (MacLeod et al., 1976; Humphreys et al., 2000; Christiansen et al., 2002) of younger ages of silicic eruptions that trend in a northwest direction—a clear divergence from the northeast-trending eastern Snake River Plain–Yellowstone volcanic system. Several authors have addressed

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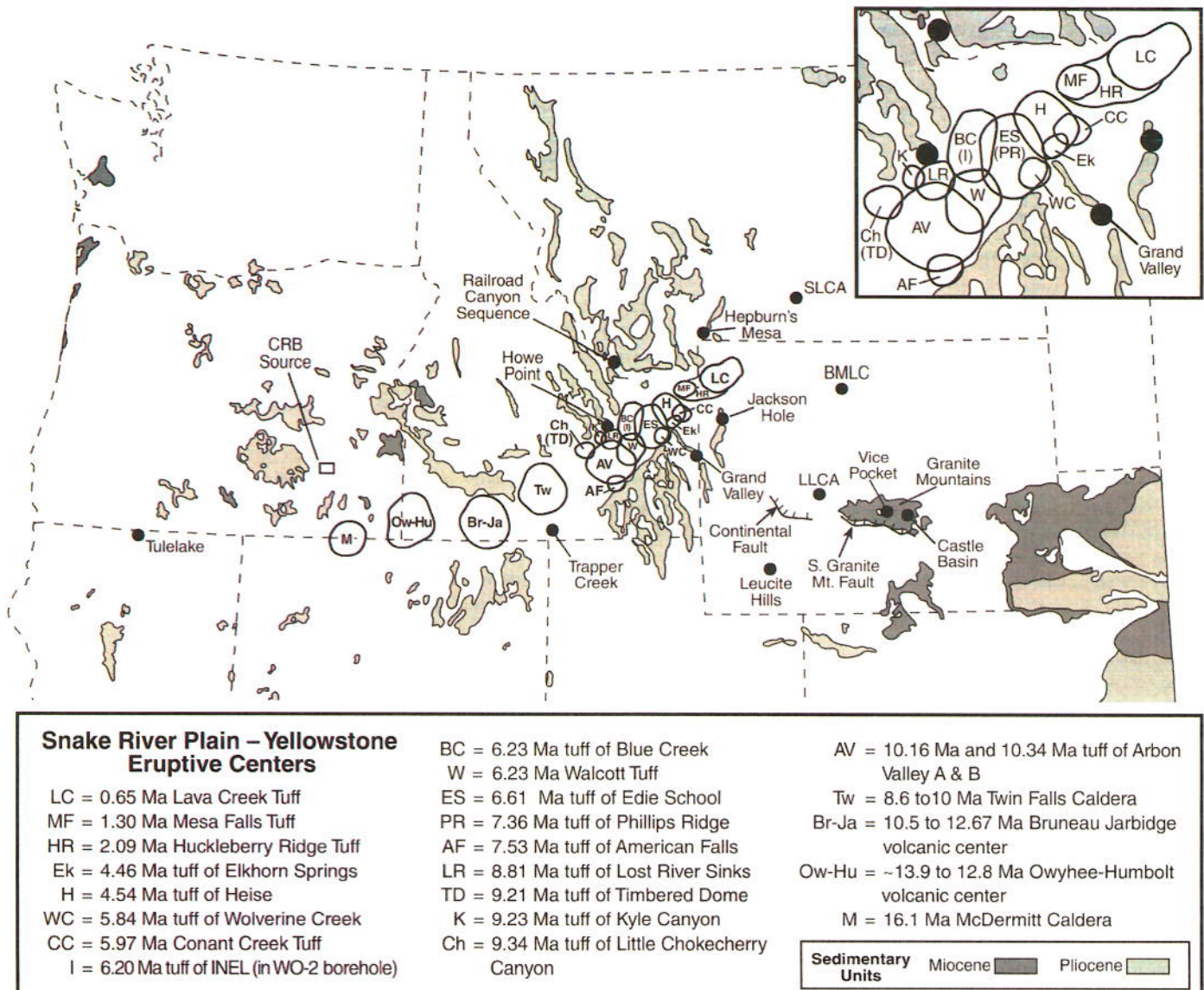


Figure 1. Map of Pliocene and Miocene sediments in the northwestern United States. Black dots represent the sampling areas discussed in the text. Caldera and volcanic center locations are modified from Christiansen (1982) and Perkins et al. (1995). The three youngest $^{40}\text{Ar}/^{39}\text{Ar}$ ages are from Lanphere et al. (2002; corrected based on Renne et al., 1998). Individual unit ages older than 2.09 Ma and younger than 10.35 Ma are $^{40}\text{Ar}/^{39}\text{Ar}$ age determinations from the Lamont-Doherty Earth Observatory argon laboratory.

the issue of the mismatched head location by suggesting that the location of the hotspot-head impingement on the lithosphere is actually near where the silicic track begins and not the geographic center of the Columbia River Province Flood Basalts (Camp and Ross, 2004). Another suggested solution is that the hotspot head and tail became separated by interaction with the subducted Vancouver slab (Geist and Richards, 1993; Pierce et al., 2002). Draper (1991) and Pierce et al. (2002) have suggested that misdirected Newberry trend is the result of the out-

ward spreading of a hotspot head. If so, this is currently the only evidence of a migratory pattern of hotspot-head-related volcanic activity. In the classic model the volcanic activity of the eastern Snake River Plain–Yellowstone system is commonly thought to be related to a hotspot tail (Anders et al., 1989; Anders and Sleep, 1992; Pierce and Morgan, 1992; Smith and Braile, 1994). Parsons et al. (1994) have suggested that initial tectonic thinning of the Basin and Range Province created space for the hotspot head to propagate in a generally southeastward direc-

tion providing buoyancy that could explain the anomalous kilometer of elevation found in the northern Basin and Range Province. Camp and Ross (2004) suggest that the head is more limited to the area west of the North American craton and east of the Cascades. Evidence suggesting an outward-spreading plume head under the thinned lithosphere of western North America includes the Newberry trend (MacLeod et al., 1976), the apparent spread of anomalous Basin and Range elevation (Parsons et al., 1994), and the migration of high-alumina olivine tholeiites

outward from the source area of the Columbia River flood basalts (Draper, 1991; Camp and Ross, 2004). However, there is currently no evidence of an outward-spreading hotspot head under the thicker lithosphere eastward under Idaho and Wyoming.

Although the pattern of eruptive ages of silicic calderas is the primary evidence for a hotspot model, the chemistry of the basalts (e.g., Duncan 1982; Draper, 1991) and rhyolites (e.g., Leeman, 1982; Perkins et al., 1995; Hughes and McCurry, 2002; Nash et al., 2006) as well as the high $^3\text{He}/^4\text{He}$ found at Yellowstone (Craig et al., 1978) are also consistent with the classic hotspot model (cf. Christiansen et al., 2002). Changes in the chemistry of the volcanic rocks have long been used to suggest changes in the composition of the lithosphere through which the magmas pass (e.g., Armstrong et al., 1977; Doe et al., 1982; Leeman, 1982; Farmer and DePaolo, 1983; Hart et al., 1984). Nash et al. (2006) found a stair-step drop in $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{176}\text{Hf}/^{177}\text{Hf}$ as well as a matching drop in Fe content of silicic rocks along the hotspot track. The first dropdown they report was at 15 Ma, which geographically corresponds roughly to the strontium 0.706 line (Armstrong et al., 1977) and the change from high-alumina olivine tholeiites to the Snake River Plain olivine tholeiites (Hart et al., 1984; Camp and Ross, 2004). Perkins and Nash (1995) and Nash et al. (2006) also report a less pronounced change in silicic volcanic rock chemistry at 7.5 Ma. The first change is generally thought to mark the western edge of the North American craton (Armstrong et al., 1977; Nash et al., 2006). Perkins and Nash (2002) and Nash et al. (2006) interpreted the second dropdown to correspond to a change in the rate of the North American Plate motion and concomitant change in the input of basalts into the lithosphere. Nash et al. report a 5 cm/yr rate prior to 7.5 Ma as opposed to the lesser rate after that. For the interval 10 Ma to present, Anders (1994) found a rate of 2.2 cm/yr and Pierce and Morgan (1992) a rate of 2.9 cm/yr. Also, from 2 to 3 m.y. before present, Gripp and Gordon (1990) reported a rate of 2.2 cm/yr, and Gripp and Gordon (2002) reported 2.69 cm/yr. Perkins et al. (1995) and Nash and Perkins (2002) suggest that there is a marked reduction in the eruption rate of silicic eruptions starting at ca. 8.5 Ma from 10 to 20 tuffs/m.y. down to ~2.5 tuffs/m.y.

Apart from the pattern of volcanic eruptions, the progress of the hotspot tail is also identified by the migratory deformation field resulting from the thermal effects of an outward-spreading tail plume (Anders et al., 1989; Rodgers et al., 1990; Anders and Sleep, 1992; Pierce and Morgan, 1992; Anders, 1994; Smith and Braile,

1994). Anders and Sleep (1992) suggested that the interaction of velocity fields of the radially outward-spreading tail plume and the North American Plate yields a parabolic shape that is represented by the parabolic distribution of elevated seismic activity centered on the axis of the eastern Snake River Plain–Yellowstone volcanic track. As the North American Plate moved, the seismic parabola moved in tandem resulting in discrete pulses of accelerated faulting at different locations along the margins of the eastern Snake River Plain at different times.

Here we develop a regional correlation of eastern Snake River Plain–Yellowstone Plateau volcanic field silicic units that is used to establish the extensional history of three major basins affected by the Yellowstone hotspot. These are the Grand Valley, Granite Mountains, and Jackson Hole basins. In them we identify several pulses of accelerated deformation that we suggest are related to the Yellowstone hotspot tail as well as its head. Moreover, we use the same basin-fill volcanic units to demonstrate that the rate of major silicic eruptions is roughly constant over the entire history of the hotspot track with the only exception being the “Heise volcanic gap” between 4.49 Ma and 2.06 Ma in which no major silicic eruptions are known to have occurred.

GEOLOGIC SETTING

Our study areas include three normal-fault-bounded basins that we believe contain the most complete late Miocene to Pliocene sedimentary sections in Wyoming and eastern Idaho.

Grand Valley

In Grand Valley (Fig. 2), a nearly continuous section of the Salt Lake Formation records the interval from before mid-Miocene to the latest Pliocene. This unit is capped by gravels of the latest Pliocene to Quaternary Long Spring Formation, which contains the 2.09 Ma Huckleberry Ridge Tuff (here we use the Lanphere et al. [2002] age of 2.059 ± 0.004 corrected for the new monitor standard age of 28.34 Ma for the Taylor Creek Rhyolite by Renne et al. [1998] and rounded to the nearest 10 ka). The contact between the Salt Lake and Long Spring Formations is at times difficult to distinguish but generally involves a slight angular unconformity beneath the roughly horizontal Long Spring Formation. These sediments fill the 4-km hanging-wall depression of the 140-km-long Grand Valley and Star Valley fault (see Dixon, 1982; Anders et al., 1989; and Anders, 1990). The hanging wall is separated into three interconnected subbasins that are from north-

west to southeast—Swan Valley, Grand Valley, and Star Valley. In Swan Valley, Anders et al. (1989) used tectonically tilted units within the Salt Lake Formation to demonstrate that there was an accelerated extension event between ca. 4 Ma and 2 Ma. They suggested this pulse of extension is associated with the migration of Snake River Plain–Yellowstone silicic volcanism of the Yellowstone hotspot. In the adjacent Grand Valley, Merritt (1958) described a detailed stratigraphy of more than 1.6 km of upper Salt Lake Formation (Fig. 3). He referred to this unit as the Teewinot Formation (Merritt, 1956, 1958), although subsequent workers used the older Salt Lake Formation name (Rubey, 1973; Oriel and Platt, 1980). Merritt's section was measured along the banks of the Snake River, an area that is now covered by the Palisades Reservoir and reservoir-deposited sediments. Here we describe a new section within Grand Valley exposed during low water along the southern side of Van Point (Fig. 2). Much of the section is covered by reservoir sediments; however, enough exposure exists to allow us to measure the section from its top to its bottom (see GSA Data Repository Appendix 1¹).

At Van Point, many of the units described by Merritt (1958), including several prominent vitric-ash and pumicite layers, are well exposed. However, there are some substantive differences between Merritt's classification of tephra layers and ours that make direct correlation difficult. Unfortunately, Merritt's original studies presented no data on the geochemistry or ages of the tephra units they described.

Granite Mountains

The Granite Mountains of central Wyoming (Fig. 1) constitute a series of Precambrian granites upon which a series of Cenozoic sediments is deposited. The youngest of these sediments are the Miocene Split Rock and Pliocene Moonstone Formations. These Tertiary units were deposited and subsequently tilted to the south-southwest by movement on the South Granite Mountains fault system. The northern margin of the basin is bounded by a smaller-displaced normal fault, thereby defining the basin as an asym-

¹GSA Data Repository item 2009021, including a stratigraphic column showing the thickness and general character of the volcanic units at Van Point; a discussion of age dating of the Conant Creek Tuff and Teewinot Formation and their role in assessing the uplift history of the Teton Range; a discussion of the microprobe analyses of tephra glasses; and two plots of the correlation coefficients between tephra units within the basins studied, is available at <http://www.geosociety.org/pubs/ft2009.htm> or by request to editing@geosociety.org.

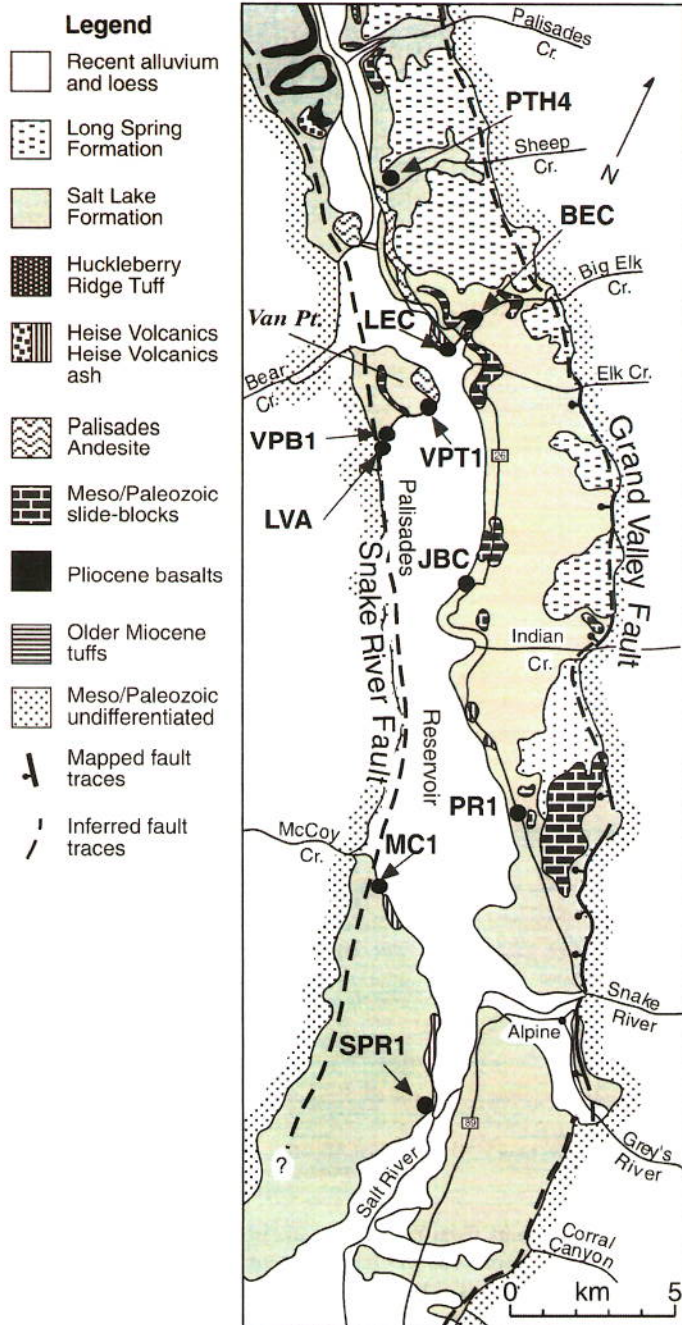


Figure 2. Map of Cenozoic rocks in Grand Valley. Black dots are sampling localities. Modified from Anders (1990).

metric graben. The Moonstone and Split Rock Formations measured by Love (1961) at Vice Pocket are 413 m thick. The Split Rock Formation, as measured near Vice Pocket, contains numerous tephra layers including many defined by Love (1961, 1970) as pumicites. Love's (1961) measured type section of the Moonstone Formation contains some 25 pumicite and other

tephra layers. As discussed by Love (1970), the Split Rock Formation is internally conformable. Moreover, Love (1961) reported a 2° southwest dip for the lowermost Moonstone Formation and 4° uniform dip for the Split Rock Formation. Love (1970) concluded the difference in dip indicated that tectonic activity began after the deposition of the Split Rock Formation.

The best exposures of the Moonstone and Split Rock Formations are not found at the same location. In the Vice Pocket area, the most complete section of the Moonstone Formation is found a few kilometers to the northeast of the most complete section of the Split Rock Formation. Another good section of Split Rock Formation, in the Castle Basin area, is found ~40 km east and 15 km south of the measured section near Vice Pocket (Fig. 1).

Jackson Hole

Love (1956) described the stratigraphy of the Pliocene to Miocene Teewinot Formation as a series of tuffaceous sandstones, fresh-water limestones, and tephra deposits. Love (1956) and Love et al. (1992) describe the Teewinot Formation as overlying the middle to lower Miocene Colter Formation. The Colter Formation comprises a series of middle to lower Miocene water-lain pyroclastic conglomerates, claystones, and sandstones (Love, 1956; Barnosky, 1984). Overlying the Teewinot Formation is the Conant Creek Tuff (Christiansen and Love, 1978; Gilbert et al., 1983; Love et al., 1992). The 2.09 Ma Huckleberry Ridge Tuff overlies the $5.97 \text{ Ma} \pm 0.01$ ($n = 8$) Conant Creek Tuff with a thin gravel separating the two units at Signal Mountain in northern Jackson Hole (Love et al., 1992; also see Appendix 2 [footnote 1] for further discussion of the age of the Conant Creek Tuff). As seen in Figure 4, there are a number of exposures of the Teewinot and Colter Formations (labeled Tpm) in the Jackson Hole area. Both these units are west tilted and thus younging upsection from east to west. Throughout Jackson Hole the Colter Formation has a greater westward tilting than the Teewinot Formation. Unfortunately, there is no place mapped as a continuous exposure of either unit (see Appendix 3 [footnote 1] for further discussion of these units). For the most part the Teewinot Formation is internally conformable and dips roughly $\sim 20^\circ$ to the west, although local folding and faulting as well as nonhorizontal deposition result in significant variation in tilt.

Gilbert et al. (1983) suggested the modern Teton Range initiated at ca. 5–6 Ma and continues today with latest Pleistocene displacement rates as high as 2.2 cm/yr. Pierce and Morgan (1992) suggested the modern Teton Range initiated at 5 ± 1 Ma. Fritz and Sears (1993) suggest there was a paleovalley system crossing the future Teton Range that was cut off sometime after development of the caldera that produced the tuff of Edie School, which is now dated at 6.61 ± 0.01 Ma. Byrd et al. (1994), based on a paleomagnetic study of the Huckleberry Ridge Tuff, concluded that the bulk of extension on

the Teton fault occurred after 2.09 Ma. Anders (1994) held that the Conant Creek Tuff and the Huckleberry Ridge Tuff were not significantly tectonically tilted with respect to one another and, like Byrd et al. (1994), assumed that movement initiated around the time of deposition of the 2.09 Ma Huckleberry Ridge Tuff. As we will present in the Discussion section, we believe that faulting initiated on the modern Teton fault at ca. 3 Ma.

EASTERN SNAKE RIVER PLAIN–YELLOWSTONE TEPHRA

Excellent data now exist enabling the correlation of tephra from the ca. 2 Ma and younger Yellowstone Plateau volcanic field (e.g., Reynolds, 1975; Izett, 1981) and for Snake River Plain silicic tephra older than 10 Ma (Perkins et al., 1995; Perkins and Nash, 2002; Nash et al., 2006). However, with the exception of some limited efforts by Anders (1990) and Morgan and McIntosh (2005), little work has been done to correlate the voluminous Snake River Plain silicic tephra for the interval between ca. 10 Ma and 2 Ma. This study uses geochronological, geochemical, paleomagnetic, and petrographic methods to correlate these tephra over a widespread area from eastern Idaho to central Wyoming (Fig. 1). The Miocene to Pliocene units studied are in three basins discussed above. These units contain a number of vitric ash, pumicite, and tuffaceous layers that we were able to correlate within and between basins and, in some cases, directly to individual ash-flow tuff units associated with major caldera eruptions on the eastern Snake River Plain.

Several early attempts were made to correlate the sections in the study areas (e.g., Love, 1956; Merritt, 1956, 1958; Love, 1970). These studies predominantly depended on paleontological control to establish temporal overlap with no attempt made to correlate individual tephra layers from one basin to another. Prior to this study, the only published dates for these units have been from the Jackson Hole area (Evernden et al., 1964; Burbank and Barnosky, 1990) and several from the Grand Valley area (Anders, 1990; Morgan and McIntosh, 2005). In a larger sense, the proximity of Grand Valley and Jackson Hole to the Snake River Plain (Fig. 1) logically suggests that the source for most of the tephra is Snake River Plain–Yellowstone silicic volcanism. However, one cannot assume a priori that any individual deposit in these basins is related to those volcanics. For example, in Jackson Hole there are Miocene volcanic deposits that by their proximal nature must have come from eruptive events located between Jackson Hole and Yellowstone Valley, north of Yellowstone Park

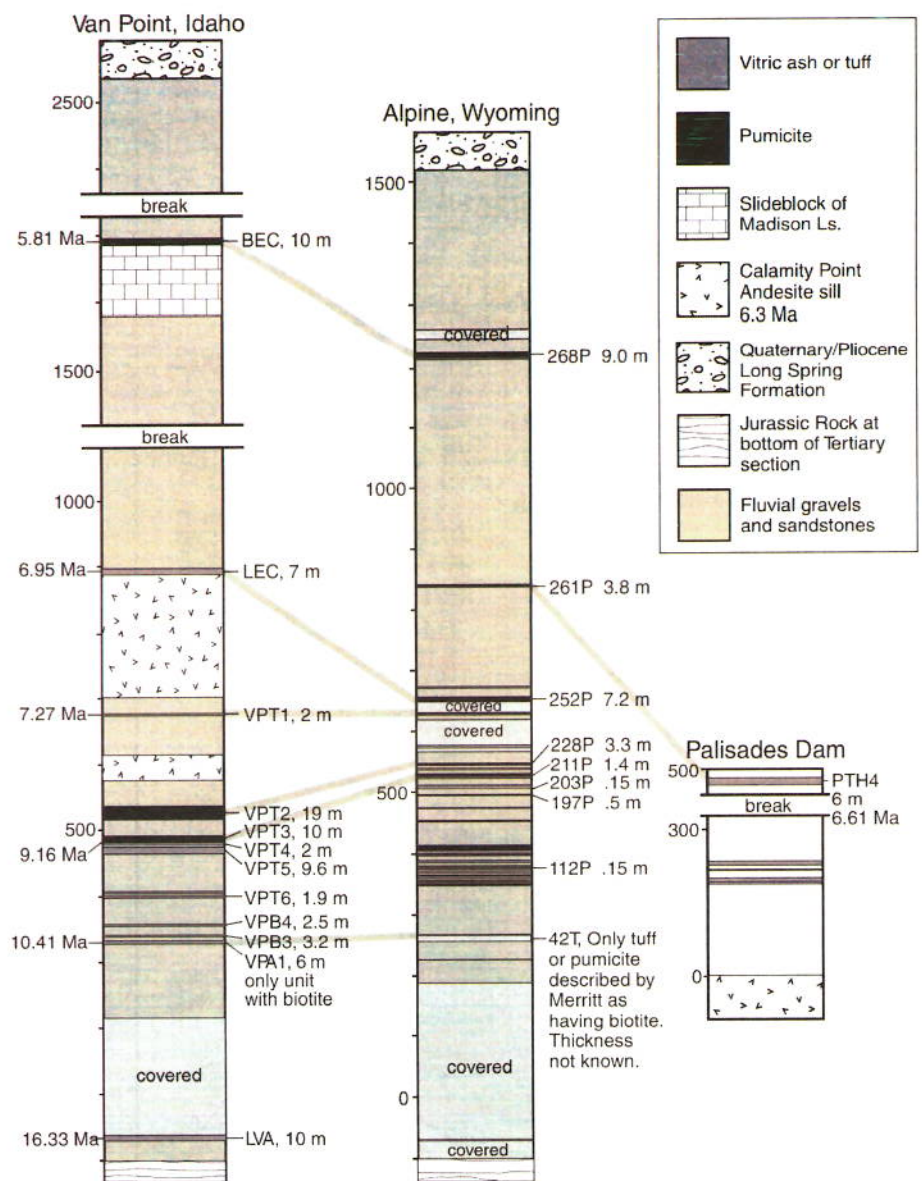


Figure 3. Stratigraphic columns from three locations in Grand Valley. The column on the left is based on our transect across the south side of Van Point (Fig. 2). The middle column is constructed from data in Merritt (1958) and represents a transect along the banks of the Snake River before the Palisades Reservoir was filled, extending from a few kilometers west of where the Snake River enters the valley to the area of McCoy Creek. The stratigraphic column on the right is from the base of the Palisades dam (Okeson, 1958). Gray lines suggest correlations between our stratigraphic column and the others. Numbers on the right of the Merritt column are his sample numbers with capital letters P for pumicite and T for tuff plus thicknesses measured in meters.

(Barnosky, 1984; Barnosky and Labar, 1989; Burbank and Barnosky, 1990). The age of these deposits, roughly 16 Ma and older, means they are too old and could not be exclusively related to the Snake River Plain–Yellowstone silicic volcanism. Also, in the southern Jackson Hole

area there are proximal volcanic deposits that may be as old as ca. 8 Ma and whose chemistry suggest they are also not related to the Snake River Plain–Yellowstone volcanism (Adams, 1997; Adams, 1999; Lageson et al., 1999). On the other hand, there are tephra deposits in the

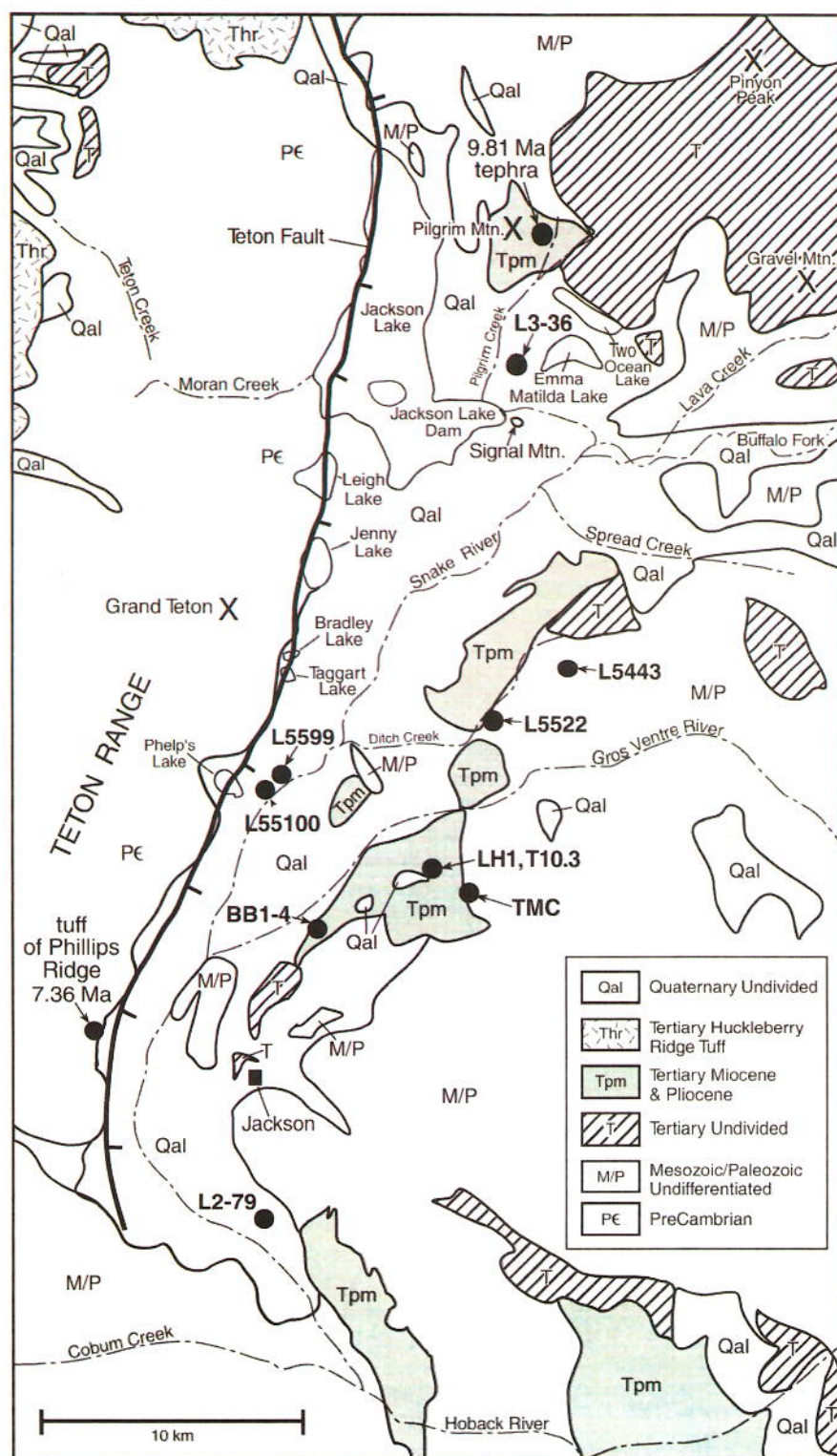


Figure 4. Geologic map of Cenozoic units in the Jackson Hole area. Black dots represent the sampling sites. Modified from Love (1956) and Love et al. (1992).

Granite Mountains area more than 300 km from the nearest eastern Snake River Plain source. Given the fine-grained nature of these deposits and assumed prevailing wind direction, they could have come from Snake River Plain silicic eruptions. However, it is just as possible that individual tephra layers may have been derived from other locations in the Basin and Range, or from as far as New Mexico, given the right wind conditions.

Our study of the tephra in the Grand Valley, Granite Mountains, and Jackson Hole basins, combined with an in-progress study of the ash-flow tuff deposits along the margin of the eastern Snake River Plain (Anders et al., 1997; some preliminary dates are shown in Fig. 1), provide an opportunity to test suppositions made about changes in the rate of volcanism as well as how the thermal structure associated with the Yellowstone hotspot affects regional patterns of extension.

CORRELATION TECHNIQUES

Tephra deposits are typically correlated based on a number of criteria (see Hahn et al., 1979; Izett, 1981; Sarna-Wojcicki et al., 1984; Perkins et al., 1995). These include (1) physical criteria such as color, shard shape, and the presence or absence of distinctive minerals or other identifiers such as obsidian balls; (2) relative criteria such as stratigraphic order; and (3) quantitative criteria such as shard geochemistry, paleomagnetic characterization, and $^{40}\text{Ar}/^{39}\text{Ar}$ isotopic analysis of feldspars. Some of the characteristics may vary with distance from the source. For example, obsidian balls or heavy minerals may be preferentially removed prior to deposition in more distal locations, and certain shard shapes may be able to remain aloft better than others. Nevertheless, these characteristics can and do serve as important criteria. However, we focused in this paper on paleomagnetic analysis, $^{40}\text{Ar}/^{39}\text{Ar}$ isotopic analysis, and the chemical analysis of glass shards including major- and trace-element chemistry.

Paleomagnetic Analyses

Paleomagnetic analysis of tephra also provides some constraints on correlations of deposits. Reynolds (1975) first used paleomagnetic analysis techniques on ash deposits from the Yellowstone Group silicic eruptions. Because each of the three major eruptions of the Yellowstone Group has a unique paleomagnetic signature, they can be correlated to individual air-fall tuffs when a primary magnetization signal can be identified. In our study of older air-fall tuffs exhibiting either viscous or chemical overprints,

it was extremely difficult to assess the primary magnetic direction. We used only alternating field (AF) demagnetization techniques to establish a site-mean direction. Unfortunately, it proved to be very difficult to remove overprints using even up to 900 Oe of AF demagnetization. Often there was little movement from the site natural remnant magnetization (NRM). Many sites were discarded because of this problem. We focused our efforts on tephra deposits in Grand Valley in an attempt to provide information about the potential correlations with ash-flow tuffs associated with Snake River Plain eruptions, whose magnetic signatures are well known (e.g., Anders et al., 1989; Morgan, 1992; Anders et al., 1993; Morgan and McIntosh, 2005).

$^{40}\text{Ar}/^{39}\text{Ar}$ Analyses

Age determinations on tephra from this study were done by laser fusing of single crystals of sanidine and plagioclase in our $^{40}\text{Ar}/^{39}\text{Ar}$ isotope laboratory at Lamont-Doherty Earth Observatory and at the Berkeley Geochronology Center. In most cases, sanidine was used; when no primary sanidine was present in the sample, plagioclase feldspar was used. Samples were irradiated at the Omega West research reactor of the Los Alamos National Laboratory and the Oregon State University reactor. The Fish Canyon Tuff sanidine with a reference age of 28.02 Ma (Renne et al., 1998) was used as the neutron flux monitor. A large number of grains were analyzed for each horizon studied. This was necessary because of the high probability of contamination of the feldspar population due to reworking. Statistical analysis of the results from multiple analyses of grains was performed using a weighted-average technique (Taylor, 1982; Turrin et al., 1998).

Geochemical Analyses

Microprobe analysis of shards for both trace- and major-element chemistry provides a distance-independent way of fingerprinting tephra deposits (e.g., Jack and Carmichael, 1969; Smith and Westgate, 1969). The assumption of chemical homogeneity of volcanic glass shards within one eruption is reasonable and is supported by previous observations. Sarna-Wojcicki et al. (1981) found that silicic volcanic glasses are relatively homogenous within an ash erupted from the initial explosive phase of volatile-rich eruptions. Many other workers have used glass chemistry for correlation of tephra (e.g., Hahn et al., 1979; Westgate et al., 1994).

Early workers (e.g., Smith and Westgate, 1969) found microprobe analysis of major elements in volcanic glass to be sufficient for fingerprinting tephra deposits. Recent attempts at tephra correlation have focused on trace- and major-element chemistry, isotopic dating, and petrographic analysis (e.g., Hahn et al., 1979; Izett, 1981; Sarna-Wojcicki et al., 1987; Perkins et al., 1995; Perkins and Nash, 2002). This study relies on trace- and major-element chemistry of glass shards and knowledge of relative stratigraphic position as major tools for correlation (see Appendix 4 [footnote 1] for details of our microprobe analysis).

Because such a large number of elemental analyses are available from microprobing individual shards, it is difficult to choose which elements provide the most information for correlation. Rather than pick one or two elements to compare abundance, we use a simple ratioing technique first developed by Borchardt et al. (1972) that allows us to compare the relative abundance of a number of elements at once. We simply compare the spectrum of analytical results from one suite of analyses from one sam-

pling site to another. The closer the match the higher the similarity coefficient and hence the greater likelihood that tephra deposits are from the same source.

To test the internal consistency of our geochemical techniques for identifying and correlating tephra over a wide geographic area, we conducted a control study of tephra from the 0.649 Ma Lava Creek eruption (corrected Lanphere et al. [2002] age of 0.639 Ma using the Renne et al. [1998] monitor standard age for the Taylor Creek Rhyolite), sampled within a similar geographic area as the late Miocene and Pliocene tephra deposits. The outcrops sampled, shown in Figure 1, are located at Silesia, Mountain (SLCA), the Bighorn Mountains, Wyoming (BMLC), and Lander, Wyoming (LLCA).

RESULTS

Paleomagnetic Analysis

Figure 5 shows the results of the paleomagnetic analysis of tephra from Grand Valley and, for comparison, the site-mean directions from Anders et al. (1993) for two areally extensive ash-flow tuffs exposed between Grand Valley and the eastern Snake River Plain. The right-hand panel in Figure 5 shows examples of the scatter exhibited at two of the individual sampling sites. Sample site BEC is the same unit described in Anders (1990) as "Big Elk Creek Ash" and in Morgan and McIntosh (2005) as Conant Creek Tuff sample sites #11 and #12. Not shown are several samples with normal polarity from unit BEC discussed in Anders (1990). The site-mean directions of normal polarity cores from BEC correspond to the present field direction and are thought to be an overprint unrelated to the field direction at the time of deposition.

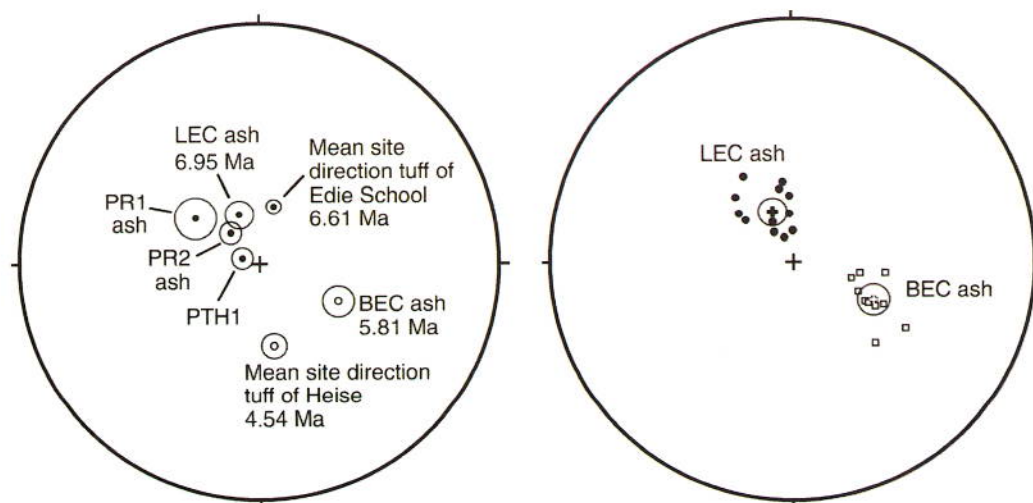


Figure 5. Results of paleomagnetic analysis of tephra layers in Grand Valley. Open symbols are upper hemisphere, and closed symbols are lower hemisphere. Circles represent 95% confidence level. Unit means for the tuff of Edie School and the tuff of Heise are from Anders et al. (1993). Panel on right shows the individual sample directions after alternating field demagnetization for two tephra layers shown on the left panel.

Below ash layer BEC is a stratigraphic horizon from which we sampled JBC, SPR1, PR1, and PR2 (Fig. 2). There is a close grouping of site-mean directions for these tephra layers, all of which show normal polarity. Correcting these ash units for tectonic tilting of the basin sediments results in an overlap of individual site directions with the mean site direction of the 6.61 Ma tuff of Edie School from Anders et al. (1993). Moreover, the ash at PTH1 is chemically similar to these four ash samples (see Fig. 6) and was determined to be the tuff of Edie School (Anders et al., 1989, therein called the tuff of Spring Creek).

Tephra horizon LEC is the same as Morgan and McIntosh's site #23. They dated the unit at 6.54 ± 0.06 ($n = 11$, corrected to 6.58 ± 0.06) and correlated it to the 6.61 ± 0.01 tuff of Edie School, which they called the Blacktail Creek Tuff. We date this unit at 6.94 ± 0.03 ($n = 3$). Although the Morgan and McIntosh ages seem a close match in time and paleomagnetic direction to the tuff of Edie School (Fig. 5), LEC is stratigraphically below the ash horizons JCB, SPR1, PR1, and PR2, and a tectonic tilt cor-

rection of LEC does not result in an overlap of respective α_{95} . Moreover, petrographic examination of shard morphology further suggests LEC is not correlative with the 6.61 Ma units.

In general the AF demagnetization of these ashes had little effect on their NRM. The danger in our analysis is that a resistant overprint, chemical or viscous, could go unnoticed. Since these ashes have clearly been tilted to the northeast by some 20° to 25° , NRM in any direction coincident with the present field direction must be viewed with suspicion.

$^{40}\text{Ar}/^{39}\text{Ar}$ Isotopic Dating

Table 1 shows the $^{40}\text{Ar}/^{39}\text{Ar}$ analyses done on tephra in this study. It is clear from Table 1 that a number of the tephra layers sampled were reworked. Several of the individual dates reported in Table 1 were not included in the weighted averages. The criteria used to exclude a particular sample included samples having significantly older age than the other samples or samples having low radiogenic argon, typically less than 50%. More distal

samples presented a greater problem, in that the feldspar grains were often at the lower limit of size for which good results could be expected given the young age of the tephra in this study. This is particularly true for samples LVA or MC1, McCoy Creek ash, from Grand Valley, and samples TMC from Jackson Hole and CB2 from the Granite Mountains. The oldest sample from Grand Valley (MC1) yielded an age of 16.33 ± 0.63 Ma. The oldest age we determined from the Granite Mountains area is 11.14 ± 0.23 Ma. This analysis was performed on feldspars from the tephra layer we labeled GSA, which is pumicite #20 in Love (1961), and is the highest pumicite layer in the Split Rock Formation. Each of these older dates has a relatively large error associated with it. Samples from VPT3 exhibited relatively low values of radiogenic argon that suggest some weathering, and therefore, some concern about the accuracy of its quoted age. Of the units analyzed, sample GVA, from the uppermost ash in the Moonstone Formation of the Granite Mountains, had the most surprising result. Although reworked, there is a significant population of grains with high radiogenic argon that yielded a 2.15 ± 0.01 Ma age (Table 1). Reported ages for the Huckleberry Ridge Tuff range from 2.003 ± 0.014 Ma (Ganseci et al., 1998) to 2.059 ± 0.004 Ma (Lanphere et al., 2002; again we use a corrected age 2.09 Ma rounded to the nearest 10 ka), inconsistent with our date. A possibility is that GVA could be related to a much smaller pre-Huckleberry Ridge Tuff silicic eruption at Yellowstone or a mixture of an older more distal ash with the ash from the Huckleberry Ridge Tuff ash or an eruption off the eastern Snake River Plain. Another possibility is that the ages represent variability in the excess ^{40}Ar found in the Huckleberry Ridge Tuff (Lanphere et al., 2002).

Major-Element and Trace-Element Geochemistry

We use a technique defined by Borchardt et al. (1972), and as modified for major-element chemistry (see Sama-Wojcicki et al., 1987), which matches the geochemical signature of one sampling site against all other sites in our study. The technique produces a "similarity coefficient" of one sampling site to each of the other sites. In order to test our technique we applied it to several known locations of the 649 ka Lava Creek ash (see Fig. 1 and Table 2). We compared sampling site BMLC2 against two other sites we sampled and against the results from Rieck et al. (1992). As hoped for, high similarity coefficients were found between our three Lava Creek ash deposits with similarity coefficients

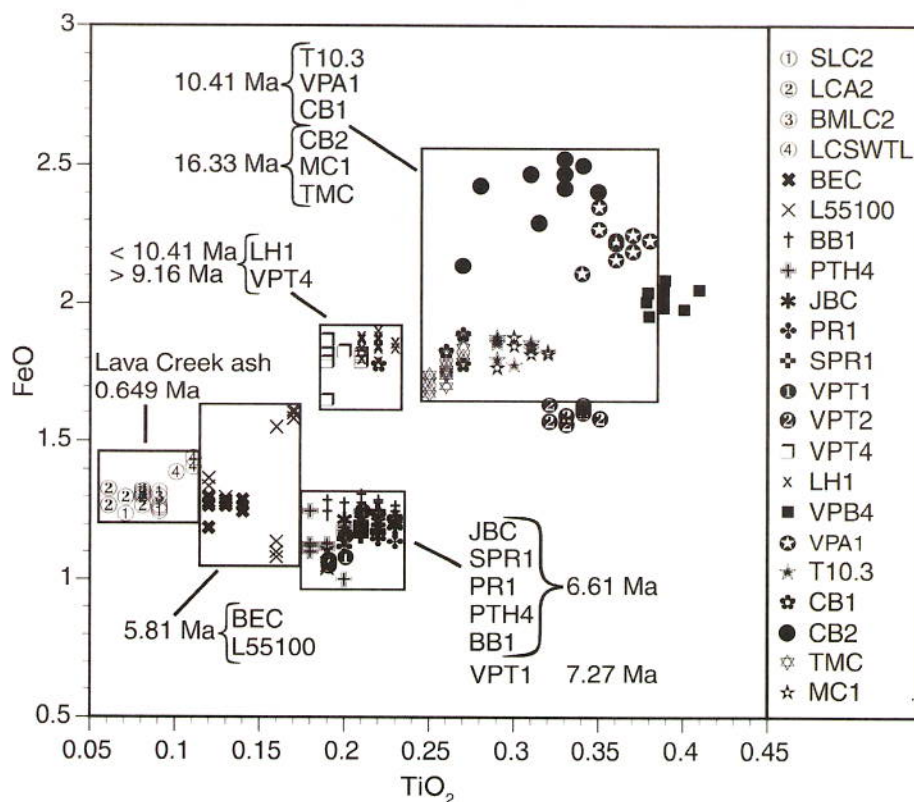


Figure 6. A plot of iron vs titanium from microprobe analyses of glasses. Tephra sample sites are discussed in the text. Note the general progression of increasing iron and titanium with age. Iron and titanium comprise only two of several elements used to establish similarity coefficients between all tephra studied.

TABLE 1. TEPHRA SINGLE-CRYSTAL, LASER-FUSION $^{39}\text{Ar}/^{40}\text{Ar}$ RESULTS (continued)

Unit	Age	Ca/K	% Rad. Ar	^{40}Ar moles	Mineral
GVA* (Granite Mountains)					
6187-02	2.24 ± 0.01	0.0092	99	1.00E-19	Sanidine
6187-08	2.19 ± 0.01	0.0100	99	9.01E-14	Sanidine
6186-03 ^c	2.07 ± 0.01	0.0089	99	1.16E-13	Sanidine
6186-06 ^c	2.08 ± 0.01	0.0094	99	7.83E-14	Sanidine
Average	2.15 ± 0.01				
6187-01 ¹	3.49 ± 0.01	0.0096	99	1.81E-13	Sanidine
6187-03 ¹	14.72 ± 0.35	0.0152	48	1.28E-15	Sanidine
6187-04 ¹	15.67 ± 0.50	0.0000	48	6.42E-16	Sanidine
6187-07 ¹	424.88 ± 4.09	0.1528	98	4.68E-14	Sanidine
6187-05 ¹	596.53 ± 8.12	0.0000	98	3.38E-14	Sanidine
GBA1¹⁰ (Granite Mountains)					
20163-01	6.75 ± 0.11	0.0002	99	5.00E-17	Sanidine
20163-02	102.96 ± 1.82	0.0081	81	9.30E-16	Sanidine
Average	6.75 ± 0.11				
GSA** (Granite Mountains)					
20078-01	10.91 ± 0.67	0.0305	98	1.30E-16	Sanidine
20078-02	11.16 ± 4.10	0.0893	78	3.00E-17	Plagioclase
20078-03	11.27 ± 0.34	0.0000	99	1.60E-16	Sanidine
20078-05	11.04 ± 0.47	0.1451	99	1.50E-16	Sanidine
20078-07	11.12 ± 0.57	0.2269	94	1.50E-16	Sanidine
Average	11.14 ± 0.23				
20078-04 ¹	14.43 ± 3.06	1.2397	89	4.00E-17	Plagioclase
20078-06 ¹	12.85 ± 1.27	0.3421	54	9.00E-17	Sanidine
MC1** (Grand Valley)					
20047-02	16.62 ± 0.13	0.0356	43	2.30E-16	Sanidine
20047-03	15.92 ± 0.87	0.0987	62	2.90E-16	Sanidine
Average	16.33 ± 0.63				
20047-04	9.7 ± 1.9	0.2734	8	8.20E-16	Sanidine
HPA¹⁵ (Howe Point)					
20041-01	16.61 ± 0.49	0.0542	84	7.79E-15	Sanidine
20041-02	15.49 ± 0.28	0.0903	78	4.40E-16	Sanidine
20041-03	16.38 ± 0.23	0.0679	90	4.00E-16	Sanidine
20041-04	16.19 ± 0.39	0.0458	89	3.40E-16	Sanidine
Average	16.12 ± 0.15				

Note: Sample numbers <30,000 at Lamont-Doherty Earth Observatory; all others at Berkeley Geochronology Center. Monitor standard for the Fish Canyon Tuff used 27.84 Ma converted to 28.02 Ma (Renne et al., 1998). Rad.—Radiogenic.

*J—4.08E-4 ± 1.0E-6.

**J—2.638E-6 ± 1.0E-6.

†J—2.664E-4 ± 4.3E-7.

‡J—8.881E-4 ± 6.280E-7.

^cSample run with label VPT 5 by mistake.

^dData not used; radiogenic ^{40}Ar percentage too low.

^eData not used; stratigraphically inconsistent.

TABLE 1. TEPHRA SINGLE-CRYSTAL, LASER-FUSION $^{39}\text{Ar}/^{40}\text{Ar}$ RESULTS

Unit	Age	Ca/K	% Rad. Ar	^{40}Ar moles	Mineral
BEC* (Grand Valley)					
20010-02	5.93 ± 0.07	1.1520	94	3.30E-16	Plagioclase
20007-02	5.67 ± 0.07	0.0000	77	3.00E-16	Sanidine
20085-01	5.97 ± 0.11	0.0141	76	5.50E-16	Sanidine
20085-02	5.71 ± 0.66	1.1101	68	1.40E-16	Plagioclase
20085-03	5.73 ± 0.07	0.7283	72	9.20E-16	Plagioclase
Average	5.81 ± 0.04				
20004-2 ¹	6.14 ± 1.0	1.1420	38	5.96E-15	Plagioclase
20001-3 ¹	5.88 ± 0.2	1.1750	26	9.20E-15	Plagioclase
LEC** (Grand Valley)					
20071-01	6.97 ± 0.05	0.0168	89	1.01E-15	Sanidine
20071-02	6.99 ± 0.09	0.0191	87	6.10E-16	Sanidine
20071-05	6.91 ± 0.05	0.0144	91	9.80E-16	Sanidine
Average	6.95 ± 0.09				
20071-04 ¹	15.74 ± 2.36	1.3028	24	2.90E-16	Plagioclase
20071-06 ¹	9.32 ± 0.42	1.4529	31	1.56E-15	Plagioclase
20071-03 ¹	8.39 ± 0.32	1.3212	91	2.00E-16	Plagioclase
VPT 1* (Grand Valley)					
6185-02	7.20 ± 0.03	0.0653	97	9.14E-15	Sanidine
6185-02	7.26 ± 0.03	0.0617	97	7.78E-15	Sanidine
6185-03	7.31 ± 0.04	0.0360	97	5.42E-15	Sanidine
6185-04	7.21 ± 0.03	0.0408	96	8.85E-15	Sanidine
6185-05	7.18 ± 0.05	0.0318	91	3.52E-15	Sanidine
6180-05	7.27 ± 0.03	0.0436	94	6.40E-15	Sanidine
6180-06	7.12 ± 0.04	0.0532	91	6.04E-15	Sanidine
6181-04	7.42 ± 0.04	0.0384	94	4.18E-15	Sanidine
6181-01	7.26 ± 0.03	0.0425	95	6.18E-15	Sanidine
6181-04	7.64 ± 0.15	0.0742	93	4.36E-15	Sanidine
Average	7.27 ± 0.04				
VPT 3* (Grand Valley)					
6183-02	8.94 ± 0.12	6.2281	84	1.86E-14	Plagioclase
6183-05	9.21 ± 0.12	5.1201	83	9.25E-16	Plagioclase
6183-07	9.04 ± 0.12	6.5611	79	9.88E-16	Plagioclase
6183-08	9.43 ± 0.13	5.5272	73	1.13E-15	Plagioclase
6183-09	9.21 ± 0.15	5.3278	83	1.12E-15	Plagioclase
Average	9.16 ± 0.06				
6183-06 ¹	11.09 ± 0.09	5.0028	88	2.31E-15	Plagioclase
6183-09 ¹	10.31 ± 1.19	5.5239	9	1.59E-15	Plagioclase
6183-08 ¹	9.63 ± 0.44	6.3713	15	7.81E-15	Plagioclase
VPA 1* (Grand Valley)					
6181-03	10.46 ± 0.04	0.0303	98	9.43E-15	Sanidine
6181-04	10.35 ± 0.04	0.0269	96	9.67E-15	Sanidine
6181-05	10.52 ± 0.05	0.0184	99	7.42E-15	Sanidine
6181-08	10.25 ± 0.04	0.0191	96	8.33E-16	Sanidine
6181-10	10.52 ± 0.04	0.5202	95	1.69E-14	Sanidine
Average	10.41 ± 0.02				
6181-02 ¹	8.68 ± 0.15	4.4018	68	7.86E-16	Plagioclase
6181-06 ¹	8.38 ± 0.16	4.0908	87	8.87E-16	Plagioclase
6181-09 ¹	9.07 ± 0.15	3.6854	65	1.29E-15	Plagioclase

(continued)

TABLE 2. MAJOR-ELEMENT GLASS GEOCHEMISTRY FROM THE 0.649 Ma LAVA CREEK ASH

	Na ₂ O	TiO ₂	SiO ₂	MgO	FeO	CaO	Al ₂ O ₃	MnO	K ₂ O
Av SD*	0.16	0.02	0.32	0.00	0.06	0.02	0.12	0.00	0.20
SLCA2	2.97	0.08	77.68	0.02	1.26	0.53	12.20	0.04	5.23
BMLC2	2.67	0.10	78.28	0.02	1.38	0.50	12.19	0.04	4.84
LLCA2	2.93	0.07	77.79	0.02	1.30	0.50	12.40	0.04	4.94
LC SW TL*	3.53	0.11	76.93	0.03	1.41	0.53	12.20	0.04	5.04

*Average standard deviation for first three in each column.

*Averaged data from Rieck et al. (1992).

of 94 and 93. A similarity coefficient of 89 was achieved with the analysis of the Lava Creek ash from Tule Lake, California (Fig. 1 and Appendix 5 [footnote 1]) by Rieck et al. (1992). Confident that this technique is useful, we applied it to all our samples. Again samples from the same stratigraphic horizon like JBC, PR1, and SPR1 from Grand Valley yield a 98 similarity coefficient (Appendix 6 [footnote 1]).

One of the ash horizons studied, the 6-m-thick VPA1 ash in Grand Valley (Fig. 3), exhibited a differing geochemical signature from top to bottom (Table 3). Similarly, one of the ash units

in Jackson Hole at sampling sites T10.3av and L5443 also exhibited variations from top to bottom. Moreover, ash horizon CB1 in the Granite Mountains basin also exhibited a change in chemistry from top to bottom. When comparing this Granite Mountains ash horizon (CB1) to the other units in Jackson Hole and Grand Valley, the similarity coefficients were, with one exception, all very high: VPA1B, 93; VPA1av, 89; VPA1D, 71; T10.3av, 92; and L5443, 95 (see Appendix 5 for further discussion).

Figure 6 shows a comparison of TiO₂ to FeO for samples from all three basins. There are

clear groupings of tephra within Ti/Fe space that in all cases corresponds to established stratigraphic order and ⁴⁰Ar/³⁹Ar isotopic dating. There is a clear trend in the Ti/Fe ratio with age, the younger units having lower values of both FeO and TiO₂ and older units having higher values. There appears to be a correlation among the lowest tephra sampled in each of the three basins. There is a high similarity coefficient of 91 between MC1 and TMC and not as high at 83 between TMC and CB2. These older deposits are all fine-grained ash from distant eruptive events. The ash deposits TMC and MC1 look very similar petrographically and have similar Fe and Ti (Fig. 6).

DISCUSSION

Based on chemical, paleomagnetic, and isotopic analyses of individual tephra deposits, combined with other factors such as stratigraphic order and petrographic similarities, we can identify several tephra layers within each of

TABLE 3. MAJOR- AND TRACE-ELEMENT GEOCHEMISTRY OF TEPHRA GLASSES

Av SD	Na ₂ O	TiO ₂	SiO ₂	MgO	FeO	CaO	Al ₂ O ₃	K ₂ O	Ti	Zr	Ba	Mg	Rb	Ce	Sr	Nb
	0.24	0.03	0.38	0.02	0.09	0.04	0.25	0.23	0.0125	0.0182	0.0225	0.0180	0.0794	0.1990	0.0181	0.0092
Salt Lake Formation, Grand Valley																
BEC	2.21	0.12	77.08	0.07	1.27	0.49	12.44	5.27	0.0640	0.0320	0.0575	0.0329	0.9404	0.2062	0.0127	0.0101
PTH4	2.61	0.19	77.24	0.08	1.15	0.49	11.95	6.27	0.1041	0.0249	0.0567	0.0451	0.8875	0.1667	0.0228	0.0051
JBC	2.33	0.22	77.23	0.10	1.21	0.50	12.50	5.88	0.1122	0.0230	0.0943	0.0513	0.7914	0.1776	0.0196	0.0077
PR1	2.11	0.22	77.34	0.10	1.18	0.51	12.56	5.60	0.1123	0.0243	0.0974	0.0513	0.7946	0.1318	0.0228	0.0062
SPR1	2.21	0.21	77.64	0.10	1.18	0.49	12.33	5.80	0.0939	0.0197	0.0377	0.0450	0.8474	0.1175	0.0215	0.0057
VPT1	1.73	0.19	77.90	0.09	1.12	0.48	12.29	6.16	0.1038	0.0251	0.0328	0.0453	0.8623	0.1506	0.0191	0.0078
VPT2	1.75	0.37	76.80	0.14	1.70	0.66	12.24	6.37								
VPT3	1.43	0.40	76.87	0.15	2.14	0.84	12.21	5.94								
VPT4	1.38	0.20	77.89	0.06	1.78	0.56	12.14	5.95	0.1043	0.0341	0.0535	0.0315	0.8215	0.0841	0.0143	0.0081
VPT5	1.29	0.30	78.61	0.10	1.47	0.50	12.00	5.69								
VPT6	0.98	0.37	77.14	0.14	1.69	0.67	12.20	6.76								
VPB4	1.34	0.39	76.20	0.17	2.02	0.85	12.52	6.32	0.2091	0.0507	0.1168	0.0944	0.8439	0.1526	0.0192	0.0058
VPA1A	1.39	0.34	76.43	0.12	2.09	0.81	12.46	6.31	0.1731	0.0487	0.1190	0.0932	0.8489	0.1770	0.0198	0.0070
VPA1B	1.51	0.32	76.18	0.11	2.06	0.77	12.69	6.32	0.1864	0.0201	0.0587	0.0549	0.8239	0.2677	0.0219	0.0079
VPA1D	1.73	0.12	77.13	0.09	1.13	0.63	12.60	6.54	0.0748	0.0428	0.1128	0.0669	0.8296	0.1527	0.0219	0.0062
MC1	1.61	0.31	77.28	0.11	1.80	0.66	12.55	5.71	0.1643	0.0380	0.0894	0.0657	0.8779	0.2254	0.0179	0.0037
Teewinot Formation, Jackson Hole																
L2-79	1.82	0.06	77.39	0.09	0.99	0.96	13.96	4.64								
L55100	2.54	0.16	77.35	0.08	1.24	0.51	12.54	5.53	0.0935	0.0240	0.0677	0.0386	0.9343	0.1058	0.0252	0.0061
BB1	2.13	0.22	78.41	0.08	1.26	0.46	12.22	5.20	0.1136	0.0243	0.0460	0.0453	0.8590	0.1077	0.0204	0.0063
L3-36	1.90	0.18	77.04	0.06	1.69	0.55	12.06	6.80	0.1867	0.0443	0.1031	0.0801	0.8368	0.1890	0.0185	0.0047
LH1	1.91	0.22	76.22	0.06	1.85	0.62	12.22	6.87	0.1221	0.0416	0.0632	0.0270	0.8822	0.1192	0.0181	0.0070
L5522	0.99	0.18	77.67	0.07	1.39	0.65	12.70	6.31								
T10.3 A	1.70	0.28	78.23	0.09	1.81	0.63	12.17	5.05	0.1577	0.0443	0.1160	0.0523	0.9096	0.8924	0.0168	0.0094
T10.3 B	1.79	0.36	77.52	0.12	2.37	0.86	12.28	4.65	0.1924	0.0569	0.1191	0.0713	0.8885	0.8179	0.0170	0.0087
L5443A	1.29	0.25	75.98	0.07	1.82	0.59	12.01	7.98	0.1400	0.0448	0.0688	0.0399	0.9292	0.1577	0.0121	0.0134
L5443B	1.16	0.38	75.13	0.15	2.16	0.81	12.23	7.95	0.2193	0.0549	0.1079	0.0934	0.9093	0.2226	0.0149	0.0077
TMC	1.60	0.26	77.15	0.08	1.76	0.79	12.19	6.15								
Split Rock Formation, Castle Basin																
CB1'A'	1.57	0.26	77.30	0.07	1.80	0.60	11.91	6.48	0.1281	0.0318	0.0708	0.0347	0.9145	0.1003	0.0214	0.0065
CB1'B'	1.63	0.35	76.08	0.12	2.35	0.85	12.43	6.16	0.1822	0.0508	0.1231	0.0677	0.8872	0.1500	0.0267	0.0051
CB2	1.30	0.32	77.12	0.08	2.43	0.75	12.18	5.79								
CB3	1.61	0.33	77.41	0.09	2.43	0.72	12.25	5.11								
CB4	1.52	0.24	78.01	0.10	1.65	0.64	12.69	5.12								
Split Rock Formation, Granite Mountains																
GSA2	1.58	0.30	76.60	0.13	1.81	0.69	12.59	6.27								
GSA	2.21	0.16	76.57	0.05	1.47	0.67	12.18	6.66								

Note: Av SD—Average standard deviation.

the basins studied that are derived from known eruptions on the eastern Snake River Plain. We can also correlate tephra deposits, whose source is not known, from one basin to another. Figure 7 summarizes our interpretation. The solid connecting lines indicate a high degree of confidence in our correlation; a dashed line indicates that we consider the tephra a good candidate for correlation. Below we will discuss some specifics of our correlations in detail.

Tephra Correlations

In Grand Valley, the stratigraphically highest tephra deposit that we identified is labeled BEC (Figs. 2 and 3) and yields a $^{40}\text{Ar}/^{39}\text{Ar}$ isotopic age of 5.81 ± 0.04 Ma (Table 1). Morgan and McIntosh (2005) dated the unit at 5.56 ± 0.08 ($n = 13$, corrected to 5.60 ± 0.08 , using the Renne et al. [1998] monitor age) and 5.43 ± 0.13 ($n = 12$, corrected to 5.47 ± 0.13). They suggest this unit correlates to the Conant Creek Tuff that we determined was 5.97 ± 0.01 Ma. Clearly, BEC does not correlate with the Conant Creek Tuff or with the 4.54 ± 0.01 Ma ($n = 23$) tuff of Heise as previously suggested by Anders (1990). Tephra layer BEC is an obsidian-rich unit close in physical character and in age to the tuff of Wolverine Creek. The tuff of Wolverine Creek was determined to be 5.84 ± 0.03 Ma ($n = 15$), all of which suggests to us that the BEC ash correlates with the tuff of Wolverine Creek.

In Grand Valley we determined, based on field relationships, that tephra outcrops JBC, PR1, PR2, and SPR1 in Grand Valley were from the same layer and that all four were stratigraphically below layer BEC and above sampling horizon LEC (Fig. 2). Site JBC has a fission-track age of 6.68 ± 0.40 Ma (Oriol and Moore, 1985). Dave Moore (1990, personal commun.) discounted an older age for this unit published in Oriol and Moore (1985) as possibly being from another unit. Morgan and McIntosh (2005) reported an age of 6.18 ± 0.22 Ma ($n = 3$, corrected to 6.22 ± 0.22 Ma) from this locally. They also reported an age of 6.35 ± 0.15 ($n = 7$, corrected to 6.39 ± 0.15) from this same horizon 2 km to the northwest. They suggested the ash at JBC (their sampling site #14) is correlative to the 6.23 ± 0.01 Ma ($n = 9$) Walcott Tuff. However, site PTH4 is unique with respect to the other tephra found in the area in that it grades upward from a welded ash-flow tuff to a nonwelded air-fall ash. The ash-flow tuff part of site PTH4 has been correlated with the 6.61 ± 0.01 Ma tuff of Edie School based on paleomagnetic characteristics (Anders et al., 1989). This unit is variously referred to in the literature as the tuff of Spring Creek, tuff of Blacktail, tuff of Blacktail Creek,

Blacktail Creek Tuff, and Blacktail Creek tuff (see Anders et al., 1989; Anders, 1990; Pierce and Morgan, 1992; Morgan, 1992; Anders et al., 1993; Morgan and McIntosh, 2005; Bindeman et al., 2007). The close age, physical characteristics, and similar magnetic signature suggest deposits JCB, PR1, PR2, SRP1, PHT4, and the tuff of Edie School are from the same eruption. This correlation is further supported by the closeness in their respective geochemical signatures, as discussed above (also see Appendix 6 [see footnote 1]).

We correlated tephra layer BEC to layer L55100, the highest tephra we sampled in the Teewinot Formation in Jackson Hole, based on their chemical and physical characteristics (see Appendices 5 and 6 [see footnote 1]). An isotopic age of 5.81 ± 0.04 Ma of BEC is significant because the 5.97 ± 0.07 Ma ($n = 8$) Conant Creek Tuff at Signal Mountain in Jackson Hole is thought to overlie Teewinot Formation (Love

et al., 1992). If our correlation is correct, the Conant Creek Tuff lies within the Teewinot Formation. This also suggests that the temporal match between the Salt Lake Formation in Grand Valley is extremely close and that Merritt's (1956) suggestion of calling the Salt Lake Formation in Grand Valley the Teewinot Formation is reasonable. Unfortunately, we could not make a direct geochemical correlation between the BEC/L55100/tuff of Wolverine Creek and any tephra in the Granite Mountains area.

In the Granite Mountains, tephra deposit GBA1 at the base of the Moonstone Formation (unit #5 in Love, 1961) yielded a date of 6.75 ± 0.11 Ma that is close to the age of the tuff of Edie School. Therefore, this unit could correlate with tephra layers JBC, PR1, and SPR1 in Grand Valley because of this similar age and also because of the similar shard morphology. The lack of geochemical correlation and the large error range of the age determination make

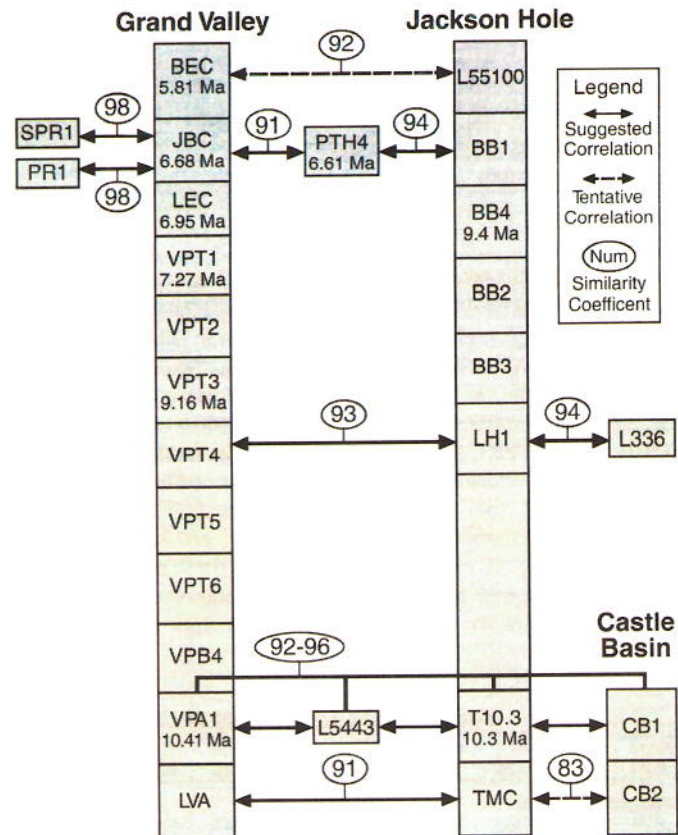


Figure 7. Columns showing schematic representation of stratigraphic order of units discussed in text. Bold arrows represent our best estimate of the correlations between tephra layers. Dashed arrows represent a lower level of confidence in the proposed correlation. Numbers in ellipses are the similarity coefficients between the tephra layers indicated.

for a weakly supported, but possible, correlation to the tuff of Edie School.

In Grand Valley, tephra VPT1 is chemically similar to those layers that we correlated with the tuff of Edie School. However, it is stratigraphically lower and slightly older, with a $^{40}\text{Ar}/^{39}\text{Ar}$ age of 7.27 ± 0.04 Ma (Table 1). No ash-flow tuff of this age has been identified on the eastern Snake River Plain. The chemical similarity and close age of this 2-m-thick tephra deposit suggest that it may represent a smaller precursor eruption to the larger eruption that produced the tuff of Edie School, an interpretation that would extend the maximum age of the Heise Volcanics. Moreover, $^{40}\text{Ar}/^{39}\text{Ar}$ age dating of the tuff of Phillips Ridge (sampled by Adams [1998]) at the southern end of the Tetons yielded an age of 7.36 ± 0.02 Ma. The chemistry of the tuff of Phillips Ridge is characteristic of silicic eruptions on the eastern Snake River Plain (D. Adams, 1998, personal commun.). Because of its geographical location, chemistry, and relative age, it is possible that the tuff of Phillips Ridge is also from a small early eruption from the same caldera as produced the tuff of Edie School, thus extending the age of the Heise Volcanics (*sensu* Morgan, 1992) even further to 7.36 Ma. There is a proximal ash-flow tuff near American Falls with an age of 7.53 ± 0.01 ($n = 4$) that is also a likely candidate for an eruption from the same source area as the tuff of Edie School (Fig. 1), thus possibly extending the Heise Volcanics even farther back in time.

Although we have no geochemical data from tephra deposit LEC, a similar paleomagnetic site-mean direction (Fig. 5) and its stratigraphic position between the 7.27 Ma VPT1 and the 6.61 Ma PR1, PR2, JCB, and SR1 tephra horizon, as well as its 6.95 ± 0.09 Ma age, all strongly suggest that it too may have the same caldera source as the 6.61 Ma tuff of Edie School but be earlier in time. This is consistent with our suggestion that an older Heise Volcanics eruption may have occurred as early as 7.53 Ma quickly followed by an eruption that produced an ash-flow tuff (tuff of Phillips Ridge) that made it into the Jackson Hole area well before the rise of the Teton Range.

In Grand Valley, the tephra layers below VPT1, layers VPT2 and VPT3 (see Appendix 1 [see footnote 1]), correlate chemically with tephra deposits in the Teewinot Formation. Three samples collected from the Teewinot Formation that lie stratigraphically below beds BB1, BB4, BB2, and BB3 all have unacceptably high standard deviations in their shard chemistry thus making their similarity coefficients suspect. However, layer BB4 was previously dated at 9.4 Ma using K/Ar (Evernden et al., 1964; corrected for 1979 decay constants).

We dated the Grand Valley tephra layer VPT3 at 9.16 ± 0.13 Ma using $^{40}\text{Ar}/^{39}\text{Ar}$ (Table 1); these dates are consistent with the assumption that the Grand Valley tephra VPT2 and VPT3 could be equivalent to tephra BB2, BB4, or BB3 from the Teewinot Formation in Jackson Hole. Without further analyses a direct correlation is speculation, but the close ages suggest that these parts of the two sections are roughly correlative.

Based on similarity coefficients (see Appendix 5 [see footnote 1]) and relative stratigraphic position, we suggest that tephra layer VPT4 in Grand Valley correlated with either tephra layer LH1 or layer L336 of the Teewinot Formation in Jackson Hole. The major-element similarity coefficient of VPT4 with LH1 is 92; and with L336 the coefficient is 93. For comparison, the similarity coefficient of LH1 with L336 is 94. The stratigraphic position of the tephra layers LH1 and L336 below the 9.4 Ma layer in Jackson Hole and the position of VPT4 below the 9.16 ± 0.06 Ma layer in Grand Valley further supports this correlation.

Near the bottom of the Grand Valley measured section at Van Point is the tephra layer VPA1, which we dated at 10.41 ± 0.02 Ma (Table 1). The Teewinot sample T10.3 was previously dated using K/Ar at 10.3 ± 0.6 Ma (D. Burbank, 1997 personal commun.). Tephra horizon VPA1 and T10.3 exhibit similar trends in changes in chemistry top to bottom. This is also a pattern observed in tephra horizon CB1 in the Castle Basin of the Granite Mountains basin. Based on the geochemical similarity coefficients and isotopic data, we consider tephra layer VPA1 to be equivalent with T10.3, L5443, and CB1. Moreover, the isotopic ages, 10.41 ± 0.02 Ma and 10.3 ± 0.6 Ma, are consistent with the age distribution determined for the tuff of Arbon Valley based on $^{40}\text{Ar}/^{39}\text{Ar}$ analyses from several locations on the margin of the eastern Snake River Plain. These $^{40}\text{Ar}/^{39}\text{Ar}$ ages are 10.16 ± 0.01 Ma ($n = 8$) and 10.34 ± 0.01 Ma ($n = 9$). This, plus the presence of biotite in these samples, which is uncommon for eastern Snake River Plain silicic air-fall or ash-flow tuffs, leads us to conclude that all these deposits are from at least two eruptions from the same caldera whose eruptive products are collectively called the tuff of Arbon Valley (source labeled AV in Fig. 1). Perkins and Nash (2002) reported that the Arbon Valley they sampled had Fe_2O_3 (1.1 wt%) and $\text{Al}_2\text{O}_3/\text{Fe}_2\text{O}_3$ (11.4) and $\text{MnO}/\text{Fe}_2\text{O}_3$ (0.08) ratios. We found that the upper layers of what we interpret as the tuff of Arbon Valley to be consistent with these results (e.g., Fe_2O_3 of 1.13 wt%, and $\text{Al}_2\text{O}_3/\text{Fe}_2\text{O}_3$ of 11.1). However, the stratigraphically lower parts of the layers we interpreted to be the earliest eruption of the tuff of Arbon Valley have chemistries similar to

the other metaluminous air-fall tuff we sampled (Table 3). Again, this supports our contention that the tuff of Arbon Valley is the result of two closely spaced eruptions events.

The lowest tephra in the Grand Valley sequence is air-fall tuff LVA. This layer crops out at Van Point and at the mouth of McCoy Creek (Fig. 2), where the sampling site is labeled MC1 and is dated at 16.33 ± 0.63 Ma (Table 1). The fluvial deposits directly below layers MC1 and LVA appear to be conformable to them; however, the bedding below is poorly exposed. Petrographically LVA looks similar to the Teewinot Formation's bottom air-fall tuff TMC, and has a similarity coefficient of 91 with it. The similarity coefficient of TMC with CB2, from Castle Basin, is only 83. Tephra layer LVA is relatively thick (~10 m), and like VPA1 above it, may contain more than one glass chemistry. Unfortunately, this layer has not yet been sampled bottom to top. We tentatively correlate TMC of the Teewinot Formation with CB2 of the Split Rock Formation (see Fig. 7). The correlation between air-fall tuff LVA in Grand Valley and air-fall tuff TMC in Jackson Hole also is robust. This correlation is somewhat surprising because the Teewinot Formation is underlain by the Colter Formation, whose upper age was thought to be ca. 13 Ma (Barnosky, 1984). However, the upper Colter Formation, called Pilgrim Conglomerate member at Pilgrim Creek, may be misidentified Teewinot Formation based on a $^{40}\text{Ar}/^{39}\text{Ar}$ age date we determined of 9.81 ± 0.14 Ma ($n = 3$).

The correlation of CB2 to either TMC or LVA is suggested. Although the chemistry of CB2, TMC, and LVA (Table 3) is consistent with the ca. 15 Ma to 16 Ma tephra described in Perkins and Nash (2002), the age (16.33 ± 0.63) is slightly older. One possibility is that our age determination is too old—note the large associated error. Another possibility is that the eruption(s) that produced these tephra layers is (are) not from the same geographic area as the older tephra discussed in Perkins and Nash (2002). A third possibility is that older tephra sampled by Perkins and Nash (2002) is not related to the hotspot track. We believe the latter possibility is the least likely.

No geochemical data were obtained for the Split Rock and Moonstone Formations from near the Vice Pocket area in the Granite Mountains due to technical difficulties. However, we obtained some geochemical data from the Split Rock Formation in the Castle Basin area. We determined $^{40}\text{Ar}/^{39}\text{Ar}$ isotopic ages from three tephra layers in the Vice Pocket area of the Granite Mountains basin. The stratigraphically highest tephra in the Moonstone Formation was dated isotopically using $^{40}\text{Ar}/^{39}\text{Ar}$ at 2.15 ± 0.01 Ma ($n = 4$, Table 1). If this date is

accurate, it is possible that much of the Moonstone Formation is too young to be directly correlative with the Teewinot Formation of Jackson Hole and Salt Formation of Grand Valley. This is especially true if the Teewinot Formation is defined (e.g., Love et al., 1992) as being stratigraphically below the 5.97 ± 0.07 Ma Conant Creek Tuff.

We obtained a $^{40}\text{Ar}/^{39}\text{Ar}$ age of 11.14 ± 0.23 Ma ($n = 5$, Table 1) for the upper tephra layer in the Split Rock Formation in the Vice Pocket area. The layer sampled corresponds to unit #20 in Love (1961). We also obtained an isotopic age for the lowest tephra layer in the Moonstone Formation at Vice Pocket of 6.75 ± 0.11 Ma ($n = 1$, Table 1). This unit is described as a 7-m-thick tuff and pumicite layer designated unit #5 in Love (1961). Therefore, we have $^{40}\text{Ar}/^{39}\text{Ar}$ dated both the highest and lowest tephra layers in the Moonstone Formation and the highest tephra layer in the Split Rock Formations within the Granite Mountains area. This is significant because we, as well as Love (1970), observed that there is an angular unconformity between the Moonstone and Split Rock Formations at Vice Pocket and that the Split Rock Formation is internally conformable.

EVOLUTION OF GRAND VALLEY, GRANITE MOUNTAINS, AND JACKSON HOLE BASINS

We have used the tephra units discussed above to evaluate the Miocene and younger structural evolution of the Grand Valley, Granite Mountains, and Jackson Hole basins. We use the tilting history of sedimentary units in the basins as a proxy to gauge the changes in extension rate within each basin. In two of the basins, Grand Valley and Jackson Hole, we have identified an early extension episode followed by quiescence again followed by a later episode of accelerated extension. In the Granite Mountains, only a single episode of accelerated extension is identified. As will be discussed later, we believe the timing of these extension episodes is related to the thermal effects of the Yellowstone hotspot.

Grand Valley Basin and the Grand Valley Fault

Grand Valley (Fig. 2) lies directly to the southeast of Swan Valley, and the two valleys form a continuous hanging-wall basin of the Grand Valley normal fault. In Swan Valley, 20 km northwest of Van Point, the relationship between accelerated extension and the position of the Yellowstone hotspot was first recognized (Anders et al., 1989). They showed that extension rates before 4.3 Ma (now corrected

to 4.54 Ma) were very low, but that the rate of extension increased an order of magnitude sometime after 4.3 (4.54) Ma and before 2.0 Ma (now corrected to 2.09 Ma). After 2.09 Ma, almost all activity on the fault stopped. Our results show a very similar pattern for Grand Valley. The 5.81 Ma ash layer BEC is tilted 23° to the northeast, and lower in the stratigraphic column the 10.41 Ma horizon VPA1 is tilted 26° to the northeast. Thus, there was only $\sim 3^\circ$ of tilting toward the Grand Valley normal fault in the 4.6 m.y. interval from 10.41 Ma to 5.81 Ma, for a tilting rate of $\sim 0.65^\circ/\text{m.y.}$ For the interval from 5.81 Ma to 2.09 Ma, the Huckleberry Ridge Tuff near the bottom of the 1° northeast-tilted Long Spring Formation; the calculated tilt rate is $\sim 5.9^\circ/\text{m.y.}$ for 22° of tectonic tilt. The tilting rate after 2.09 Ma is $\sim 0.5^\circ/\text{m.y.}$ This pattern of changes in rate of tilting is almost identical to that reported by Anders et al. (1989) and Anders (1990) for Swan Valley.

The older parts of the stratigraphic section in the Van Point area exhibit evidence of an earlier extensional event, discussed briefly in Anders (1990). The oldest tephra deposit exposed at Van Point is LVA, which is equivalent to the 16.33 ± 0.63 Ma MC1 farther to the southeast (Table 1 and Fig. 2). This unit is tilted 35° to the northeast; thus it experienced 9° of tilt between the time of its deposition and the deposition of the 10.41 Ma air-fall tuff VPA1. The strata directly below layer

LVA appears to be conformable. Therefore, sometime between 16.33 Ma and 10.41 Ma, Grand Valley began forming a basin associated with movement on the Grand Valley fault. The tilt rate for this earlier event is $\sim 1.5^\circ/\text{m.y.}$ The age of this older event suggests that the Grand Valley fault was active well in advance of any thermal and/or tectonic event caused by the tail of the Yellowstone hotspot. However, as we will discuss later, this earlier event may well be associated with the Yellowstone hotspot.

Jackson Hole Basin and the Teton Fault

Barnosky (1984) argued that there was an earlier advent of accelerated extension sometime between 18 Ma and 13 Ma in the northern region of the Jackson Hole area. Barnosky also indicated this extensional event occurred during deposition of the Colter Formation and was associated with a regional unconformity called the mid-Tertiary unconformity, which Barnosky et al. (2007) suggest initiated between 17.5 Ma and 16.73 Ma (timing based on paleomagnetic reversal patterns correlated to astronomical cycles). Where exposed on the east side of Jackson Hole, the older Colter Formation is tilted a few degrees more steeply than the overlying Teewinot Formation (see Love et al., 1992). Figure 8 is a plot of the tilt data from Love et al. (1992) for the Colter and Teewinot Forma-

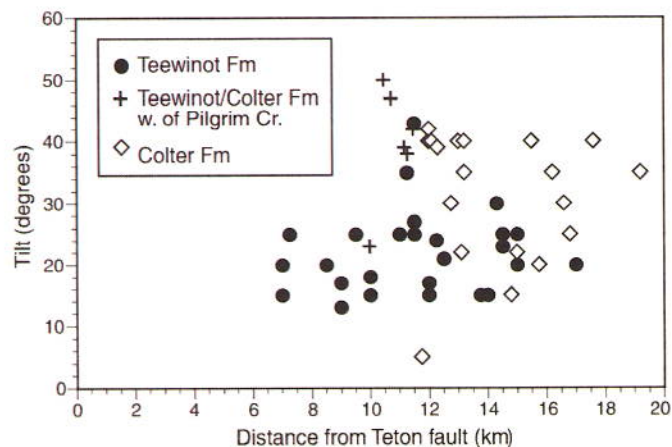


Figure 8. Fault dip vs distance from the Teton fault. Dip data are from Love et al. (1992). Solid circles are dip measurements from the Teewinot Formation; open diamonds are dip measurements from the Colter Formation; pluses are from the mapped Colter Formation in the area north of Signal Mountain and west of Pilgrim Creek (Fig. 4). Based on recent data discussed in the text, it is possible that dip locations marked with pluses could be either Teewinot Formation or Colter Formation. Moreover, the steepest dips, as discussed in the text, are likely due to local faulting. Abbreviations: Cr.—Creek; Fm.—Formation; w.—west.

tions. The tilt data from west of Pilgrim Creek are treated separately because, though mapped as Colter Formation (Barnosky, 1984; Love et al., 1992), sampling high in the section produced a Teewinot Formation $^{40}\text{Ar}/^{39}\text{Ar}$ age of 9.81 ± 0.14 Ma ($n = 3$), and, therefore, it is not clear where the transition between the two units is. Moreover, we interpret the rapid increase and wide range of measured tilt values in the Colter Formation (5° to 50° within a kilometer of each other) as representing localized faulting in the area west of Pilgrim Creek. In Figure 8 removing tilt data from west of Pilgrim Creek (Fig. 4) and plotting the measured tilt of the two formations from the hanging wall of the Teton fault demonstrates the significant variability of tilts in the formations. However, two important observations can be made about the tilt data. First, the range and average tilt of Teewinot Formation is less than that of the Colter Formation. Second, there is no demonstrable pattern of increased tilting of the Teewinot Formation as the Teton fault is approached. The lowest tephra layer sampled in the Teewinot Formation, air-fall tuff TMC (Fig. 4), we correlate with the 16.33 Ma MC1/LVA basal tephra layer in Grand Valley. If the tephra correlations are correct, the Teewinot Formation is older than Barnosky indicates by some 3 m.y., and the Colter Formation is younger by an equivalent amount. Moreover, as discussed previously, an age of 9.81 Ma of what is mapped as the upper Colter Formation (Love et al., 1992) and interpreted as Colter Formation by Barnosky (1984) suggests to us that some of what has been described as the upper Colter Formation is actually Teewinot Formation.

Because of intense folding near the base of the Teewinot Formation, especially in the area where we sampled air-fall tuff TMC, we cannot determine whether extension initiated just before or after deposition of tephra layer TMC. However, by the time of deposition of tephra layer T10.3 at ca. 10.3 Ma the earlier extension event was over. This is based on the conformability of layers stratigraphically above layer T10.3 in the Teewinot Formation. We prefer, based on the qualitative criterion that the Colter Formation is also folded and faulted on its easternmost exposure, that the extension event occurred after deposition of tephra layer TMC. Furthermore, we see no direct evidence supporting the Teewinot Formation as representing infill of an extensional basin while the character of the sediments in the Colter Formation is consistent with an evolving basin (Barnosky, 1984; Barnosky and Labar, 1989). Therefore, we conclude an extension event occurred sometime between 16.33 Ma and 10.3 Ma.

There is disagreement over when the accelerated extension that produced the mod-

ern Teton Range initiated; the disagreement evolves around the amount of tilt and the age of initiation of tilting of the Conant Creek Tuff at Signal Mountain (Fig. 4). Drilling of the Conant Creek Tuff in the area surrounding the Jackson Lake Dam site recorded a 20° to 22° tilt westward toward the Teton fault (Gilbert et al., 1983). In the same study, the Huckleberry Ridge Tuff was found to tilt westward toward the Teton fault 9° to 11° on the west side of Signal Mountain, and the Conant Creek Tuff has a 22° west tilt. Directly east of Signal Mountain there are outcrops of the Teewinot Formation that exhibit tilts of 17° and 20° , and about one-half kilometer north of Signal Mountain, there is one dipping 27° (Gilbert et al., 1983). These tilts of the Teewinot Formation are consistent with tilts throughout the Jackson Hole basin. Using the dip data on the two ash-flow tuffs of Gilbert et al. (1983), Pierce and Morgan (1992) concluded there was an angular unconformity between the two units that marked initiation of movement on the Teton fault. Pierce and Morgan (1992) concluded there was accelerated displacement on the modern Teton fault that began after 5 ± 1 Ma, assuming this as the approximate age of the Conant Creek Tuff, and before the 2.09 Ma Huckleberry Ridge Tuff was deposited. However, the age of the Conant Creek Tuff they used is somewhat problematic (see Appendix 2 [see footnote 1]).

Because the Teewinot Formation is tilted westward the same amount as the Conant Creek Tuff, significant extension on the Teton fault must have initiated after deposition of the Conant Creek Tuff. This is the same conclusion reached by Gilbert et al. (1983) and Pierce and Morgan (1992). Gilbert et al. (1983) estimated the initiation of the Teton fault, the main bounding fault of the Jackson Hole basin, to be between 5 Ma and 6 Ma based on their assumed age of the Conant Creek Tuff and on tilting patterns near the Jackson Lake Dam. Similarly, Pierce and Morgan (1992) held that the Teton fault experienced a rapid increase in displacement sometime after emplacement of the Conant Creek Tuff and before emplacement of the 2.09 Ma Huckleberry Ridge Tuff yielding an initiation of rapid extension at 5 ± 1 Ma (Fig. 4).

Anders (1994) interpreted the Conant Creek Tuff and the Huckleberry Ridge Tuff to have roughly the same tectonic tilt. This was based, in part, on the closeness between the presumed westward tectonic tilt in the Teton fault footwall. In the footwall the Conant Creek Tuff is westward tilted 6° to 9° and the Huckleberry Ridge Tuff is westward tilted 6° to 10° (as determined from contouring of outcrop data of Christiansen and Love [1978]) and by field observations as well. Moreover, Love et al. (1992) reported

similar dips of the two units at Signal Mountain which were different than those reported by Gilbert et al. (1983). Similarly, Byrd et al. (1994) found the tectonic footwall tilt of 10° for the Huckleberry Ridge Tuff was also the total tilt of the footwall. As discussed previously, Gilbert et al. (1983) found a hanging-wall tilt differential between the Huckleberry Ridge Tuff and the Conant Creek of 10° based mostly on drill core data from the area around the Jackson Lake Dam. In accepting the measures of Gilbert et al. (1983), there is clearly a significant difference between the tilting measurements made in the hanging wall and the footwall of the Teton fault. It follows that if a constant displacement rate for the Teton fault is assumed and if it is assumed at the tilting of the Huckleberry Ridge Tuff and Conant Creek Tuff are solely due to movement on the Teton fault, then the age of initiation of the fault is between 2 Ma and 4 Ma (i.e., a maximum of 10° of tilt from ca. 6 Ma to ca. 2 Ma and 10° of tilt from 2 Ma to the present, and a minimum of 0° tilt before 2 Ma). We therefore suggest, in the absence of more definitive information, that 3 ± 1 Ma is a reasonable estimate for the initiation time of the Teton fault.

Clearly there are two distinct extension episodes recorded in the sediments of the Jackson Hole basin. The earliest one initiates between 16.33 Ma and 10.3 Ma, and the latter starting at ca. 3 Ma.

Granite Mountains Basin and the South Granite Mountains Fault

The Granite Mountains area preserves evidence of an episode of accelerated extension after the youngest deposition of the Split Rock Formation and before the oldest deposition in the Moonstone Formation. We determined that the uppermost pumicite unit in the Moonstone Formation (unit 44 of the stratigraphic section of Love [1961]) has a 2.15 ± 0.01 Ma isotopic $^{40}\text{Ar}/^{39}\text{Ar}$ age (Table 1), an average slightly older than the Huckleberry Ridge Tuff, whose age has been variously reported as 2.003 ± 0.014 Ma, (Gansecki et al., 1998), 2.08 ± 0.02 Ma (Obadovich and Izett, 1991), 2.018 ± 0.016 Ma (Obadovich, 1992), and 2.059 ± 0.004 Ma (Lanphere et al., 2002), which we convert to 2.09 Ma using Renne et al. (1998). We could not get repeatable microprobe results from this unit, and thus could not rule out that the two older dates (2.19 ± 0.01 Ma and 2.24 ± 0.01 Ma) are from an earlier eruptive phase of Yellowstone Group volcanism or some non-Yellowstone Plateau volcanic field eruption. The two older $^{40}\text{Ar}/^{39}\text{Ar}$ ages combined with the two younger dates of 2.07 ± 0.01 Ma and 2.08 ± 0.01 Ma suggest the possibility of reworking of older ashes with those

from the Huckleberry Ridge Tuff. At near the base of the Moonstone Formation is a pumice unit (#5 in Love, 1961) that we dated at 6.75 ± 0.11 Ma (Table 1). Stratigraphically above the Moonstone Formation is the Bug Formation that is dated as Pleistocene in age (Love, 1970).

We correlated units CB1 and CB2, from the Split Rock Formation in the Castle Cliffs area, with the 10.41 Ma VPA1 ash horizon and the 16.33 ± 0.63 Ma LVA ash horizon, in Grand Valley, respectively. This age for the upper Split Rock Formation at Castle Cliffs is consistent with the date of 11.14 ± 0.23 Ma on the highest tephra layer in the Split Rock Formation at Vice Pocket (unit 1 in Love, 1961). Because the Split Rock Formation is internally conformable, no fault movement is believed to precede 11.14 Ma. The 2° angular unconformity between the Split Rock Formation and the younger Moonstone Formation suggests tectonic movement initiated sometime between the deposition of the top of the Split Rock Formation and the bottom of the Moonstone Formation (see Love, 1970) or between 11.14 Ma and 6.75 Ma. The stratigraphically higher Pleistocene Bug Formation is also tilted to the south (Love, 1970) suggesting tilting continued into the Quaternary and possibly to the present. Although 2° is a small amount of tilt, the larger basin width of almost 40 km, compared to 20 km for Jackson Hole and 10 km for Grand Valley, corresponds to a relatively greater fault offset on the South Granite Mountains fault for the same amount of tilt.

The Granite Mountain basin is clearly outside the region of accelerated faulting surrounding the eastern Snake River Plain–Yellowstone volcanic track (e.g., Anders et al., 1989; Pierce and Morgan, 1992; Smith and Braile, 1994) and therefore extension in this part of central Wyoming must have some other cause.

SPECULATION ON AN OUTWARD RADIATING HOTSPOT HEAD

The analysis presented above suggests an early pulse of extension that preceded the volcanism and accelerated extension associated with the track of the Yellowstone hotspot tail. This pulse migrated eastward across the northern Basin and Range in a direction away from the Columbia River Plateau and is characterized by an initial rapid eastward progression followed by a precipitous decay in the eastward migration commencing east of the Wyoming–Idaho border toward a cessation of activity in central Wyoming (Fig. 9). The timing of migration and its possible relationship to a plume head is discussed below.

At Howe Point, on the northern margin of the Snake River Plain (Fig. 1), the tuff of Arbon

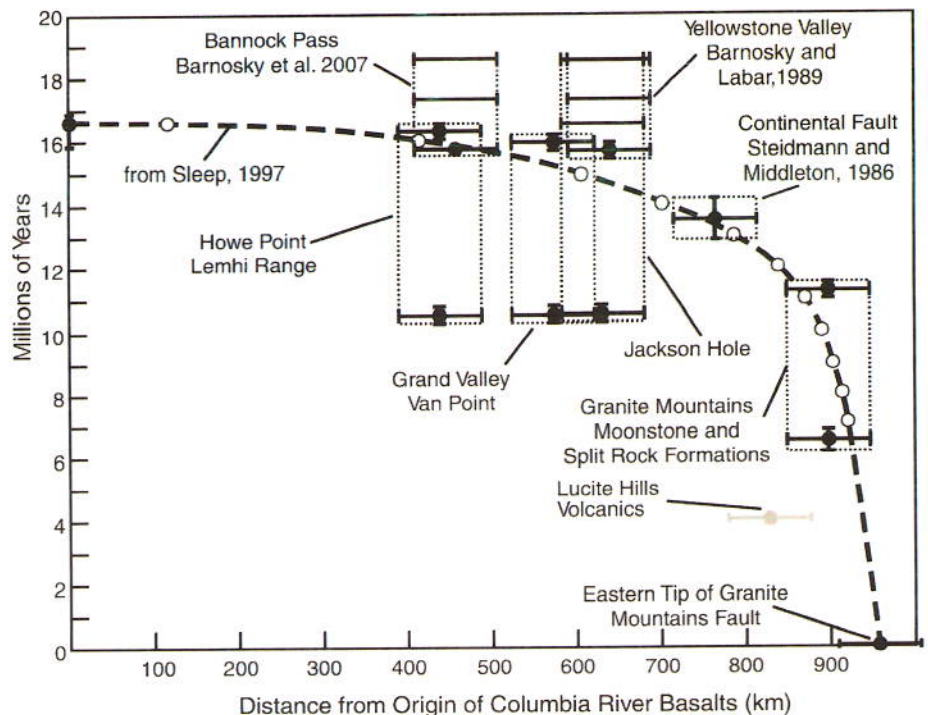


Figure 9. A plot of distance from initiation of Yellowstone hotspot (Camp and Ross, 2004) to sampling locations vs. age. Horizontal error bars represent estimated error in determining distance to the origination point of hotspot head. Solid dots represent radiometric age determinations. Dashed-line boxes represent the range of uncertainty in estimating the time of initiation of accelerated faulting at a given location. Horizontal error bars with no solid center circle are for estimates that do not directly involve radiometrically determined ages. Solid dashed curve and corresponding open circles are from the Sleep (1997) calculation of the outer limit of a hotspot head encountering the lithosphere. Gray dot and error bar is for the Leucite Hills volcanism.

Valley, the age of which is bimodal at 10.16 ± 0.01 Ma ($n = 8$) and 10.34 ± 0.01 Ma ($n = 9$), is underlain in angular discordance by a 16.12 ± 0.15 Ma air-fall tuff deposit (Table 1; Kuntz et al., 2003). There is some minor tilting of sediments directly below the lower ash suggesting some tilting might have occurred slightly prior to the ash deposition (D. Rodgers, 1999, personal commun.). Clearly there was active extension at Howe Point in this interval. This pulse of extension was followed by a relative tectonic quiescence again followed by accelerated tilting rates sometime between 9.9 Ma and 6.61 Ma (Rodgers and Anders, 1990; Anders, 1994). The later event is thought to be associated with the “tail” of the Yellowstone hotspot (e.g., Pierce and Morgan, 1992; Geist and Richards, 1993).

Farther to the east in Grand Valley, an early extensional event occurred between 10.41 Ma and 16.33 Ma. Prior to 16.33 Ma and in the interval between 10.41 Ma and 5.81 Ma the rates of extension were either very low or

zero as indicated by the tilting patterns. In the eastern Jackson Hole area, the Miocene Colter Formation dips more steeply to the west than the Teewinot Formation (see Love et al., 1992). The overlying Teewinot Formation is internally conformable, with the exception of some of the basal beds and tilts roughly 15° to 25° to the west, on average ~5° to 10° steeper than the Colter Formation. In our interpretation the interval of accelerated extension occurs between these units. Based on our age control this places the interval of accelerated extension in the Jackson Hole basin to initiate and end between 16.33 Ma and 10.3 Ma.

About halfway between Grand Valley and the Granite Mountains is the Continental fault (Fig. 1). Steidtmann and Middleton (1986, 1991) found that extension on the Continental fault began at ca. 13.5 Ma. It is possible that some extension precedes this time since the zircon dating is on ash deposits that came slightly after the first sediments associated with faulting.

In the Granite Mountains there is a single phase of extension, as defined by the 7° tilt of the conformable Split Rock Formation followed by a 2° change in dip between the 11.16 Ma top of the Split Rock Formation and the 6.75 Ma bottom of the conformable Moonstone Formation, thus restricting this interval to the initiation of accelerated faulting.

Early-phase tectonic activity has been suggested by the work of Barnosky and Labar (1989), Burbank and Barnosky (1990), and Barnosky et al. (2007). These authors suggest extensional tectonics, as represented by the mid-Tertiary unconformity (as discussed earlier), roughly between 17.5 Ma and 16.73 Ma at Hepburn's Mesa in the Yellowstone Valley and in the Railroad Canyon Sequence at Bannock Pass area (Fig. 1). This age range is based on paleomagnetic reversal stratigraphy, the ages of which are based on astronomical forcing. Within these basins only two tuffaceous units can be tied directly to isotopic dating. In the Yellowstone Valley, Barnosky et al. (2007) reported a unit CC-4 of the Hepburn's Mesa Formation dated at 15.82 ± 0.21 Ma that lies just above the mid-Tertiary unconformity and in the Bannock Pass area just below the mid-Tertiary unconformity of an ash horizon (unit +23 m of the Whisky Spring 3 section) that was dated 16.6 Ma to 15.8 Ma by correlation to tephra of known age.

A plot of the timing of this early extension event versus distance from the initiation point of the Columbia River basalts, as defined by Camp and Ross (2004), is shown in Figure 9. These earlier pulses of extension all precede the elevated thermal activity associated with the motion of the North American Plate over the proposed Yellowstone hotspot "tail" (Anders et al., 1989; Anders and Sleep, 1992; Pierce and Morgan, 1992; Smith and Braile, 1994; Rodgers et al., 2002). Parsons et al. (1994) and Pierce et al. (2002) have suggested that the head of a hotspot underlies much of the western Northern Basin and Range, but is absent under the thicker lithosphere to the east. A hotspot head according to White and McKenzie (1989) could extend outward on the order of 500–1500 km in diameter. As Anders and Sleep (1992) proposed for the plume "tail" of the hotspot, the outward spread of the head would progress toward some stagnation point. The spread initially would be rapid; thereafter the rate of outward spreading would drop precipitously as the plume loses the potential energy necessary to spread further. In our analogy to the "head" model, the outer limit of spreading would be somewhere near the Granite Mountains or the easternmost limit of extension. Based on modeling of Sleep (1997), the outward spread might proceed at a rate as high as 10 cm/yr for the first 13 m.y.,

dropping off to 3 cm/yr during the past 3 m.y. This outward spread is somewhat analogous to the standard "spreading drop" experiment of basic fluid mechanics (Koch and Koch, 1995). According to Camp and Ross (2004) the initial head arrived at an extended region of the lithosphere in southeastern Oregon at ca. 16.6 Ma. From its initial point of intersection, as determined by earliest tholeiitic basalts, Camp and Ross (2004) suggest the head spread beneath the previously thinned crust both northward and southward. As speculated by Parsons et al. (1994), the head filled the "low points" created by thinning lithosphere and migrated southward into the thinned lithosphere of the Basin and Range. Pierce et al. (2002) suggested that there is anomalous high elevation in the western Basin and Range caused by the buoyancy of the hotspot that could have affected climate as far east as central Wyoming. It is our contention that some of the buoyant plume head material first filled the thin lithosphere around the source area and then "spilled over" into regions of thicker lithosphere, migrating in a general eastward direction. The net effect of such a plume head would not only be increased elevation, but also accelerated extension due to heating of the lithosphere. This in turn should have resulted in a temporal and spatial outward-migrating pattern of accelerated extension.

On Figure 9 we have superimposed Sleep's (1997) migration rate for an assumed Yellowstone-sized hotspot head. The source is defined here as the centroid (see Fig. 1) of early Columbia River basalts emplaced between 16.6 Ma and 15.3 Ma (Camp and Ross, 2004). The farthest edge of the head plume is located on the easternmost edge of the Granite Mountains area, east of which there is only little or no evidence of significant post-Miocene extension in North America.

Given the timing of the earlier extension event discussed above, the distance-time relationships can be characterized by a rapid eastward spread of the extension pulse starting at ca. 16.6 Ma at the source of the Columbia River flood basalts in a pattern consistent with the "oil drop" modeling of a hotspot head by Sleep (1997). In just over one million years the head would have passed beneath Grand Valley and Jackson Hole. By 13 Ma its outer edge would be beneath the Continental fault, and sometime after 11.14 Ma it would be beneath the Granite Mountains area. We interpret the eastern extent of the South Granite Mountains fault to mark the head's most eastern extent.

Clearly, there are explanations possible for this eastward spread other than a hotspot head. It could be part of the observed center-to-margin spread of volcanism and extension discussed by

Armstrong et al. (1969), Scholz et al. (1971), and Allmendinger (1982) that is in some way related to the peeling away of a subducted plate (e.g., Snyder et al., 1976; Humphreys, 1995) or massive backarc upwelling (e.g., Scholz et al., 1971). Another equally plausible explanation is that there is no relationship or no single underlying physical mechanism relating one pulse of extension to another. Also of note is that there is no clear pattern of volcanism defining the progress of the head. The one clear example of volcanism that could fit the timing of a migrating head in interior Wyoming is the Lucite Hills volcanic intrusion, which does not fall on the curve of Sleep (1997). Since this volcanism occurs over our proposed sublithospheric plume limit, it could be a delayed thermal pulse from within the spreading plume. Also possibly, the Lucite Hills volcanism could be completely unrelated to a hotspot head but rather be a manifestation of eastward-migrating Basin and Range extension related to one of the other mechanisms mentioned above.

Assuming there is a plume head, a number of questions arise. For example, would the plume head have just enough volume to occupy the thinner lithosphere of Basin and Range (e.g., Parsons et al., 1994) or would it have enough remaining potential energy to permit spreading under the thicker lithosphere of Wyoming? Clearly, more data are needed to test this hypothesis, including fluid-dynamic modeling of the expected buoyancy driving forces and viscosities of the hypothesized plume head or as well as a more refined data set from other extensional basins within the Basin and Range.

CHANGES IN THE RATE OF SILICIC VOLCANISM ON THE EASTERN SNAKE RIVER PLAIN

Perkins et al. (1995) have suggested that the rate of large silicic eruptions spanning the ~16 m.y. of the Yellowstone hotspot track dropped off by a factor of 2 to 3 at between 8 Ma and 10 Ma. This calculation is based on defining large eruptions as vitric air-fall tuffs greater than 1.5 m in thickness or of reworked tephra deposits greater than 10 m in thickness in Miocene and/or Pliocene sections measured in the Trapper Creek area of Idaho (Fig. 1). Perkins et al. (1995) recorded 27 major events out of 51 tephra layers by this qualitative criterion. As they point out, this method is subject to error, and local geography may play an important role in the thickness and distribution of vitric tuffs. Merritt (1958) measured a section in Grand Valley that is now buried by Palisades Reservoir sediment (Fig. 1). He defined the tephra layers he measured as pumicite, tuff, or tuffaceous. He

identified nine pumicites, 36 tuffs, and 78 tuffaceous units in his study of the Tertiary section in Grand Valley (Fig. 3). Okeson (1958) also described several tuffaceous units (pumiceous sandstones) during excavation of the Palisades Dam (Fig. 3) but only delineated between pumicites and tuffaceous layers. Assuming that air-fall deposition of the tephra include only pumicites and tuffs but not tuffaceous units as defined by Merritt, we count 45 layers in the interval we define as between 10.41 Ma and 5.81 Ma (see Fig. 3). Of these, 24 qualify as "major" by the criteria established by Perkins et al. (1995). Again, we stress that there is not a direct comparison between the terminology used by Merritt (1958) and that of Perkins et al. (1995).

Between 10.41 Ma and the present we count 21 major eruptions as evidenced by the presence of significant ash-flow tuff deposits in, or on the margin of, the eastern Snake River Plain and Yellowstone Plateau. As shown in Figure 1, these include the 10.34 ± 0.01 Ma and 10.16 ± 0.01 Ma eruptions, which produced the tuff of Arbon Valley, the 9.40 ± 0.03 Ma ($n = 4$) tuff of Little Chokecherry Canyon, the 9.23 ± 0.01 Ma ($n = 6$) tuff of Kyle Canyon, the 8.81 ± 0.16 Ma ($n = 3$) tuff of Lost River Sinks, the 7.53 ± 0.01 Ma ($n = 4$) tuff of American Falls, the 7.36 ± 0.02 Ma ($n = 6$) tuff of Phillips Ridge, the 6.61 ± 0.01 Ma ($n = 17$) tuff of Edie School, the 6.23 ± 0.01 Ma ($n = 3$) Walcott Tuff, the 6.23 ± 0.05 Ma ($n = 15$) tuff of Blue Creek, the 6.20 ± 0.01 Ma ($n = 7$) tuff of INEL (only exposed in boreholes), the 5.97 ± 0.07 Ma ($n = 8$) Conant Creek Tuff, the 5.84 ± 0.03 Ma ($n = 15$) tuff of Wolverine Creek, the 5.46 ± 0.02 Ma ($n = 4$) tuff of Elkhorn Spring, the 4.54 ± 0.01 Ma ($n = 23$) tuff of Heise, the three eruptions of the 2.09 Ma Huckleberry Ridge Tuff, the 1.30 Ma Mesa Falls Tuff (Lanphere et al., 2002), and the two eruptions associated with the 0.649 ± 0.004 Ma Lava Creek Tuff (Lanphere et al., 2002). Again, all ages shown above were corrected for new monitor standards from Renne et al. (1998).

Not all of the major silicic eruptions on the eastern Snake River Plain are associated with ash-flow tuffs exposed along the margins. This is apparent from the Trapper Creek (Fig. 1) area (Perkins et al., 1995) and from Merritt (1958) as well as our study of tephra from the Grand Valley basin. The lack of these ash-flow tuff deposits is due to erosion and burial of the eastern Snake River Plain by younger sediments and basalts.

Using 10.34 Ma (oldest of the two tuffs of Arbon Valley ages) as a common dividing line, and subtracting the six post-10.34 Ma eruptions from the Perkins et al. (1995) compilation, yields 21 eruptions from 13.74 Ma to 10.34 Ma, for ~6.2 major eruptions per m.y. or one every

162 k.y. (compared with one every ~200 k.y. calculated by Perkins et al. [1995] for the interval ca. 13.9 Ma to 9.5 Ma; see Table 4). These rates can also be compared to changes in rate from before and after 8.5 Ma in Perkins and Nash (2002) of an eruption every 100–200 k.y. before and a ~400 k.y. interval after (see Table 4). Perkins et al. (1995) assumed a 1.5 m layer of air-fall tuff constituted a major eruption. Merritt's (1958) compilation, as interpreted by us, yields a record of 24 major eruptions between 10.34 Ma (assumed to be more accurate than the 10.41 Ma age for the tuff of Arbon Valley for site VPA 1) and 5.81 Ma assuming the 1.5 m criterion. Within this interval, we could not correlate five major Snake River Plain ash-flow tuff producing eruptions with equivalent deposits described in Merritt's compilation. These were all in the upper part of Merritt's measured section. These missing tephra deposits are likely from the eruptions that produced the Conant Creek Tuff, the tuff of INEL (found only in boreholes), the tuff of Phillips Ridge, the Walcott Tuff, and tuff of Blue Creek. All other eruptions on the eastern Snake River Plain that produced ash-flow tuffs could be linked to a "major" tephra layer, even if the correlation was only based on conjecture (i.e., an individual ash layer is undated but falls between two dated layers, allowing a possible correlation).

Within the interval 6.95 Ma and 5.81 Ma at Van Point there was an insufficient number of tephra layers to match equivalent ash-flow, tuff-producing eruptions on the Snake River Plain. For example, the tuff of American Falls and the tuff of Phillips Ridge could not be accounted for in the measured sections in Grand Valley. The corresponding tephra layers were not deposited at this location or were subsequently eroded. Taking this into account we calculated for Merritt's sections 29 major eruptions between 10.34 Ma and 5.81 Ma, assuming that the 5.81 Ma BEC tephra is not correlated to the 5.84 Ma tuff of Wolverine Creek. Assuming that these two units do correlate yields 28 major eruptions over 4.53 m.y., or on average 6.2 eruptions per million years, or a major eruption every 162 k.y. (see Table 4). This is the same rate Perkins et al. (1995) calculated for the interval from 13.9 Ma to 10.34 Ma.

We count eight major eruptions after 5.81 Ma; these eruptions correspond to the tuff of Elkhorn Spring, the tuff of Heise, and the six individual eruptions of the Yellowstone Group (Christiansen, 1982, 2001). Although there are only three cycles of caldera eruptions, we use the criteria that if individual eruption events are discernable in the outcrop, for comparison purposes they are counted as single events even though they are from the same caldera closely spaced in time. Therefore, over the

TABLE 4. RATE OF MAJOR SILICIC ERUPTIONS ON THE SNAKE RIVER PLAIN

Range	Interval between eruptions (k.y.)
Perkins et al. (1995)	
13.9 Ma to 9.5 Ma	ca. 200
<7.0 Ma	500–600
13.74 Ma to 10.34 Ma (21) ¹	162
Perkins and Nash (2002)	
15.2 Ma to 8.5 Ma	100–200
8.5 to present	ca. 400
This study	
10.34 Ma to 5.81 Ma (29)	162
10.34 Ma to 4.54 Ma (30)	193
10.34 Ma to present (33) ²	313
10.34 Ma to present (36) ³	287
8.5 Ma to present (32) ²	266
8.5 Ma to present (29) ³	293

(21) The number of eruptions during a particular interval.

¹Calculated from Perkins et al. (1995) data.

²Assumes six major Yellowstone Plateau volcanic field eruptions. Includes three eruptions of Huckleberry Ridge Tuff, one for the Mesa Fall Tuff, and two of the Lava Creek Tuff.

³Assumes only the three major Yellowstone Plateau volcanic field cycles.

interval 10.34 Ma to the present there are 36 major eruptions, which yields 3.5 eruptions per million years or an eruption every 287 k.y. on average (Table 4). Using only three major Yellowstone Group cycle eruptions yields an interval of 313 k.y. (Table 4). If we assume there is a major gap in volcanism between 4.54 Ma and 2.09 Ma, we can easily calculate the rate of eruption from 10.34 Ma to 4.54 Ma. With the post-Heise eruptive gap removed there are 30 eruptions over 5.8 m.y. or 5.2 eruptions per million years or one every 193 k.y. on average (Table 4). Beginning with the Huckleberry Ridge Tuff first eruption, there are six eruptions in 2 m.y. or 3 per million years or one every 333 k.y. There is risk in calculating eruption rates for small numbers of eruptions over geologically short intervals and it is not appropriate to project future eruptive events on this basis.

Based on our calculations, there is no significant dropoff of rate until after the eruption of the tuff of Heise at 4.54 Ma. The rate of eruption prior to 4.54 Ma is not significantly different than that calculated by Perkins et al. (1995). We calculated an eruption of every 193 k.y. on average for the 5.8 m.y. interval after 10.35 Ma. The Perkins et al. (1995) study site at Trapper Creek is directly adjacent to the Snake River Plain (Fig. 1) and Grand Valley is 70 km to the southeast of it. Therefore, the filter thickness of 1.5 m for defining a major eruption may not be equivalent to thicknesses Merritt (1958) recorded for

a similar sized eruption. In fact, if one were to use a minimum thickness of 0.75 m as representative of a major event, then Merritt (1958) found 35 tuff layers that would be considered major events. Calculating as above yields a rate of a major eruption every 141 k.y. on average over the interval 10.34 Ma to 4.54 Ma. This rate is a significantly higher rate than the Perkins et al. (1995) post-dropoff rate of a major eruption every 500–600 k.y. or the Perkins and Nash (2002) rate of 405 k.y. between eruptions and more in line with a constant rate from 13.9 Ma to 4.54 Ma.

In fairness, our criteria for these calculations, based as they are on Merritt's (1958) definition, may be less rigorous than those of Perkins et al. (1995). Nevertheless, our calculations do suggest a far less significant drop in the rate of explosive volcanism at ca. 10 Ma than that described by Perkins et al. (1995). Any perceived reduction in rate is strongly influenced by a "post-Heise eruptive gap" between 4.54 Ma and 2.09 Ma. Although there are no major ash-flow tuffs in this interval, there are a number of intercaldera lavas (Bindeman et al., 2007), which are a common feature following major silicic eruptions such as those following the Lava Creek eruption of the Yellowstone Plateau volcanic field (Christiansen, 2001).

Why the post-Heise eruptive gap occurred when it did during the roughly 16 m.y. history of the Snake River Plain–Yellowstone volcanism is not clear. We speculate that after the last Heise eruption the North America Plate would have placed the hotspot tail directly under the Eocene Absaroka Volcanics. Since the source of the rhyolitic magmas is the lower crust (see Leeman, 1982; Anders and Sleep, 1992), the lower crust may have already undergone significant fractionation in the Eocene resulting in a more refractory source region thus causing a delay in volcanism at the surface. As the plume moves farther under the Yellowstone Plateau, the effect of the previous Eocene volcanic activity results in the observed reduced eruption rate of the Yellowstone Group volcanism.

CONCLUSIONS

In a study of three Neogene fault-bounded basins in eastern Idaho and Wyoming, we have discovered that each experienced major pulses of extension. These basins are located in the Grand Valley, Jackson Hole, and the Granite Mountains areas, and all contain significant volumes of silicic tephra. Using geochemistry, argon geochronology, paleomagnetism, and petrographic techniques, we are able to correlate tephra from one basin to another and establish the timing of several pulses of extension. Using these results

we hypothesize that the earliest pulse in each basin is related to the outward migration of the head of the Yellowstone hotspot, whose eastern limit is presently beneath central Wyoming. The more recent pulses in extension rate observed in Grand Valley and Jackson Hole we believe are caused by the thermal-mechanical effects of the migration of the North American Plate over the fixed tail of the Yellowstone hotspot.

Using the stratigraphy we have established in these basins, we have been able to identify several new silicic eruptive events occurring during the past ca. 10 m.y., which we believe originated on the eastern Snake River Plain. When these new eruptions are added to previous compilations of the major eruptions, the results are interpreted to indicate that the rate of major silicic eruptions associated with the track of the Yellowstone hotspot was roughly constant from ca. 16 Ma to ca. 4.5 Ma with a gap of ~2.5 m.y. prior to initiation of the Yellowstone Plateau volcanic field at ca. 2 Ma.

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