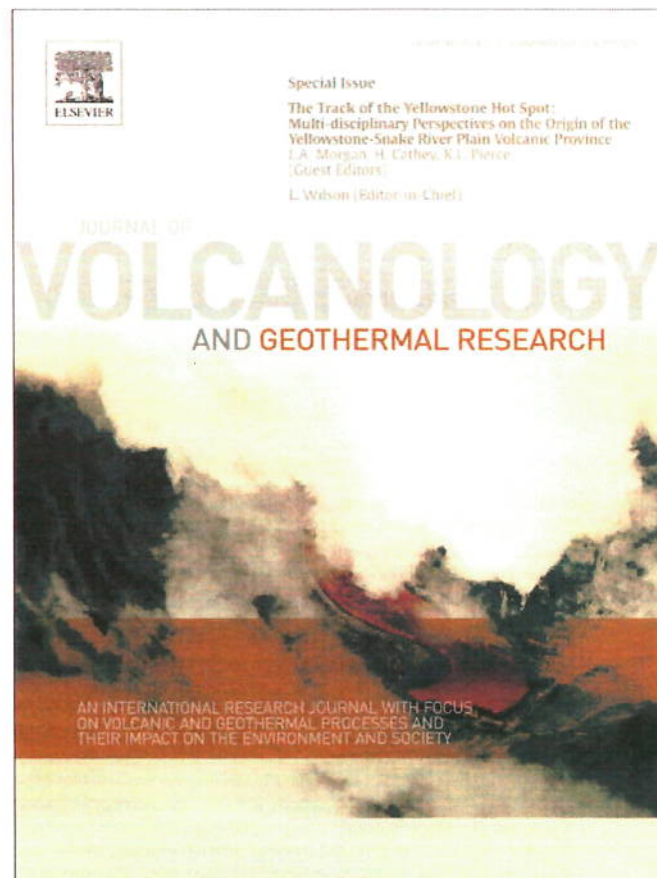


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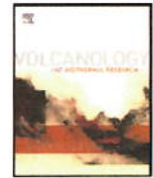


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The Pliocene Lost River found to west: Detrital zircon evidence of drainage disruption along a subsiding hotspot track

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ABSTRACT

SHRIMP analysis of U/Pb ages of detrital zircons in twelve late Miocene to Pleistocene sand samples from six drill cores on the Snake River Plain (SRP), Idaho, suggests that an ancestral Lost River system was drained westward along the northern side of the SRP. Neoproterozoic (650 to 740 Ma, Cryogenian) detrital zircon grains from the Wildhorse Creek drainage of the Pioneer Mountains core complex, with a source in 695 Ma orthogneiss, and which are characteristic of the Big Lost River system, are found in Pliocene sand from cores drilled in the central SRP (near Wendell) and western SRP (at Mountain Home). In addition to these Neoproterozoic grains, fluvial sands sourced from the northern margin of the SRP contain detrital zircons with the following ages: 42 to 52 Ma from the Challis magmatic belt, 80 to 100 Ma from the Atlanta lobe of the Idaho batholith, and mixed Paleozoic and Proterozoic ages (1400 to 2000 Ma). In contrast, sands in the Mountain Home Air Base well (MHAB) that contain 155-Ma Jurassic detrital grains with a source in northern Nevada are interpreted to represent an integrated Snake River, with provenance on the southern, eastern and northern sides of the SRP.

We propose that late Pliocene and early Pleistocene construction of basaltic volcanoes and rhyolitic domes of the Axial Volcanic Zone of the eastern SRP and the northwest-trending Arco Volcanic Rift Zone (including the Craters of the Moon volcanic center), disrupted the paleo-Lost River drainage, confining it to the Big Lost Trough, a volcanically dammed basin of internal drainage on the Idaho National Laboratory (INL). After the Axial Volcanic Zone and Arco Volcanic Rift Zone were constructed to form a volcanic eruptive and intrusive highland to the southwest, sediment from the Big Lost River was trapped in the Big Lost Trough instead of being delivered by surface streams to the western SRP. Today, water from drainages north of the SRP enters the Snake River Plain regional aquifer through sinks in the Big Lost Trough, and the water resurfaces at Thousand Springs, Idaho, about 195 km to the southwest.

Holocene to latest Pliocene samples from drill core in the Big Lost Trough reveal interplay between the glacio-fluvial outwash of the voluminous Big Lost River system and the relatively minor Little Lost River system. A mixed provenance signature is recognized in fine-grained sands deposited in a highstand of a Pleistocene pluvial-lake system.

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1. Introduction

Previous studies of age populations of detrital zircons from fluvial sands in the Snake River Plain (SRP) region demonstrate their utility as reliable indicators of the source rock geology of river drainages (Geslin et al., 1999; Link et al., 2002, 2005; Beranek et al., 2006). This paper applies detrital zircon population analysis using twelve samples from six boreholes at three widely-separated locations on the SRP (Fig. 1A and B) and refines models for drainage development as the surface of the SRP aggraded by basalt flows and sediment deposition

in the wake of the Yellowstone hotspot. We use ages of detrital zircon populations to trace drainage changes in the basaltic volcanic province that formed over the volcanic hotspot track.

Neogene river drainages developed as basins formed accompanying Basin-Range extension (Ore, 1999). These drainages carry distinctive age populations of zircons unique to the geology of each drainage (Link et al., 2005). Table 1 summarizes the detrital zircon populations of several drainages in the SRP area. Today, several tributaries to the Snake River have hook-shaped paths demonstrating capture of previous stream courses by the southwest-sloping SRP (for example South Fork Snake River and Portneuf River, Fig. 1A). These drainage diversions were controlled by: (1) post-hotspot subsidence, (2) external or internal changes in base level of the Snake River, and (3) for the Big Lost River, construction of basaltic edifices with lava

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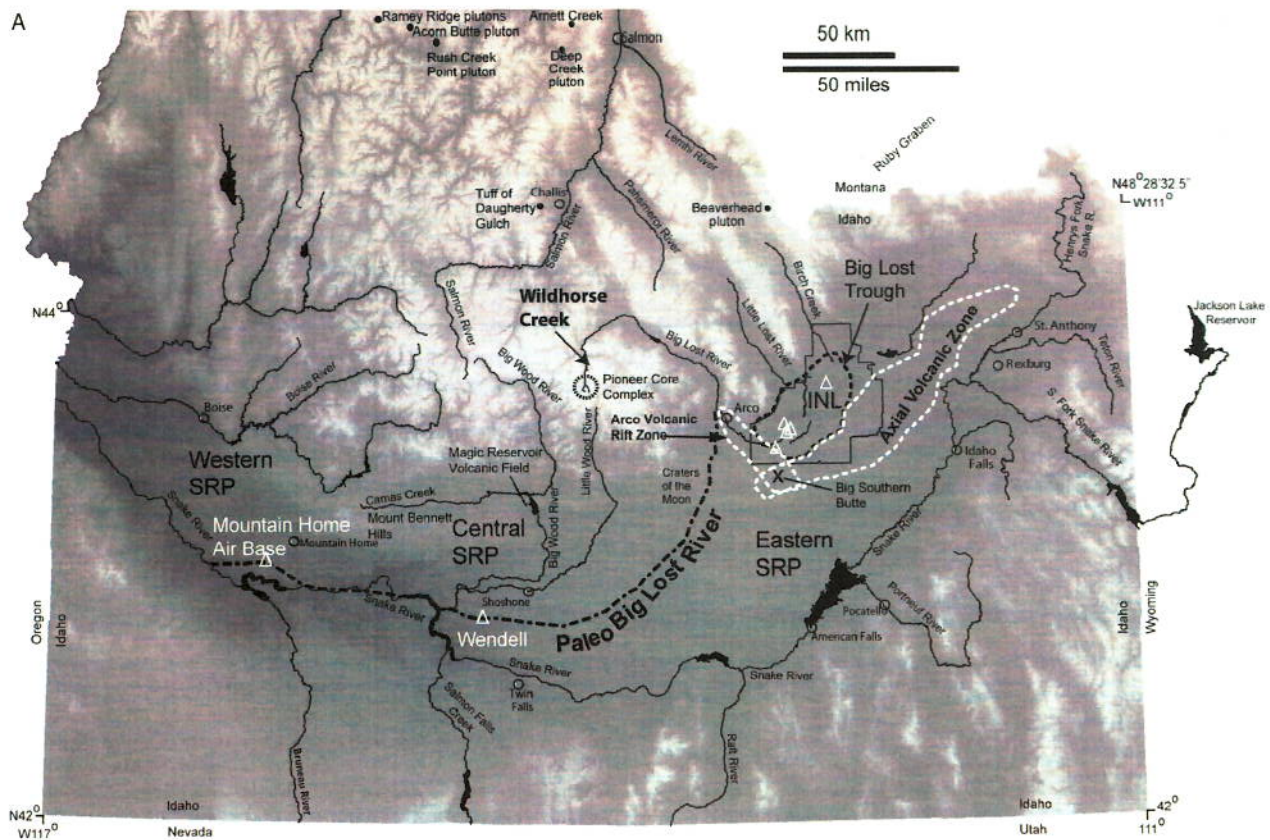


Fig. 1. Location maps. (A) map of southern Idaho area showing locations of wells and critical Neoproterozoic provenance areas. White triangles are drill holes sampled for detrital zircons. Dashed circle is Wildhorse Creek in the Pioneer Core Complex, the source for Cryogenian detrital grains; heavy dashed black line is inferred course of paleo-Big Lost River. Black X is Big Southern Butte. Dashed white line is Axial Volcanic Zone and Arco volcanic rift zone. Dashed black line is Big Lost Trough. Black dots are known locations of late Neoproterozoic and Cambrian albitic plutons of the Salmon River Mountains (Evans and Zartman, 1988; Lund et al., in press). (B) Idaho National Laboratory and vicinity showing locations of drill holes sampled for detrital zircons: USGS 134 (black triangle, point up), Middle 1823 (black triangle, point down), Middle 2050 (black dot), C1A (black square), and 2-2A (white open triangle). INL boundary is shown by thick solid black line. Other features include the Big Lost Trough (light gray shaded area), Axial Volcanic Zone, Arco-Big Southern Butte Volcanic Zone (medium gray shaded areas) and the Paleo-Big Lost River (dashed black line).

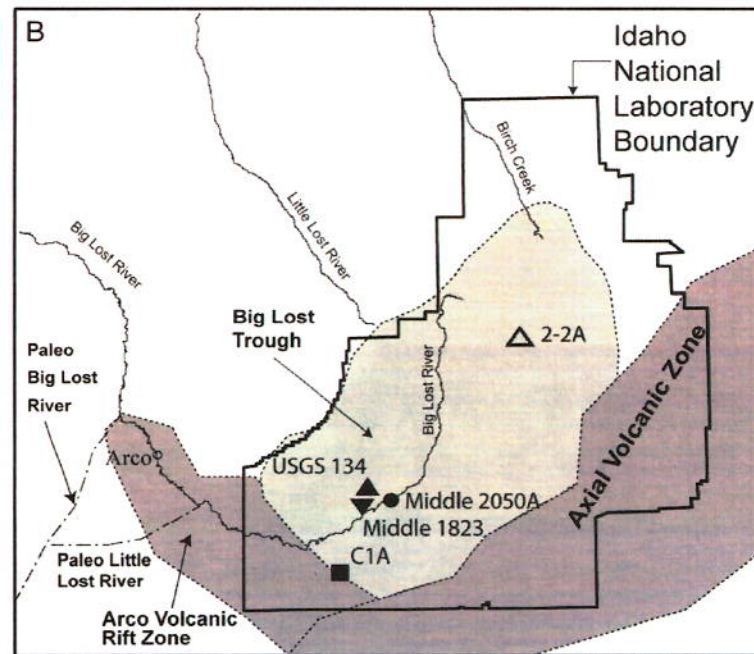


Fig. 1. (continued).

dams and associated lake systems (Pierce and Morgan, 1992; Scott, 1982; Kuntz et al., 1992; Geslin et al., 1999; Ore, 1999; Wood and Clemens, 2002; Link et al., 2002).

The Yellowstone–SRP hotspot track can be divided into three phases with respect to silicic volcanic activity. The oldest part, on the southwest, contains the 16.5– to 14-Ma McDermitt and Owyhee–Humboldt volcanic fields. The intermediate phase, located in the central SRP, contains the 13– to 5.5-Ma Bruneau–Jarbridge and Twin Falls volcanic fields (Bonnichsen et al., 2008). The youngest phase, in the area of the eastern SRP, includes the 10.2-Ma to present Picabo, Heise, and Yellowstone Plateau volcanic fields (Armstrong et al., 1975; Morgan, 1992; Pierce and Morgan, 1992; Smith and Braile, 1994; Kellogg et al., 1994; Christiansen, 2001; Perkins and Nash, 2002; Morgan and McIntosh, 2005; Bonnichsen et al., 2008). In the central SRP the young phase includes the 3– to 6-Ma Magic Reservoir volcanic field northeast of Wendell (Honjo et al., 1992). Fig. 2 (right column) shows ages of the youngest eastern SRP hotspot-related fields.

The western SRP is a northwest-trending, fault-bounded rift basin that began to develop northwest of the Bruneau–Jarbridge and Twin Falls volcanic fields about 11 Ma (Wood and Clemens, 2002; Bonnichsen and Godchaux, 2002; Shervais et al., 2002). Rifting was related to Basin and Range extension, loading by basaltic lava and sills,

and possibly formation of an extensional strike-slip duplex in part coeval with the 15.3 $\pm$ 0.5 Ma Oregon–Idaho graben (Cummings et al., 2000; Hooper et al., 2002). The western SRP contains late Miocene to Pleistocene arkosic and tuffaceous river and lake sediments of the Idaho Group, with the main sedimentary section sandwiched between underlying and overlying basalt (Lewis and Stone, 1988; Malde, 1991; Wood and Clemens, 2002; Tauxe et al., 2004).

2. Regional setting

Idaho and western North America were compressed during the Cretaceous to Paleogene Sevier and Laramide orogenies, resulting in the development of a western magmatic arc and eastern thrust belt (Burchfiel and Davis, 1975). Extension began in the Paleogene, accompanied by the rise of core complexes and development of a regional Paleogene rift system north of the SRP (Janecke, 1994). The voluminous Challis magmatic episode occurred in late Eocene time (52–42 Ma) mainly north of the SRP (Johnson et al., 1988; Sanford, 2005). Neogene extension produced Basin and Range block-fault mountain ranges both north and south of the SRP (Rodgers et al., 2002).

Table 1
Major rivers in the eastern SRP area and their detrital zircon barcodes.

Drainage basin	Defining detrital zircon populations
Big Lost River	Major population (>75%) 42– to 52-Ma Challis magmatic grains; minor population (~10%) 650– to 740-Ma grains from Cryogenian orthogneiss in Pioneer core complex
Little Lost River	Major population (>50%) 1650– to 1750-Ma grains recycled through Mesoproterozoic Lemhi Group; major population (20%) 42– to 52-Ma Challis magmatic grains
Big and Little Wood River	Major population (20%) 80– to 100-Ma Atlanta Lobe Idaho batholith; major population (40%) 42– to 52-Ma Challis magmatic field; minor populations of Proterozoic and Paleozoic grains, including 350– to 400-Ma grains from western assemblage of Antler thrust system
Salmon Falls Creek	Major population of 16– to 10-Ma (75%) grains from central SRP hotspot volcanic fields; medium population (20%) of 150– to 160-Ma zircons from Jurassic granitoids, 30– to 40-Ma grains (5%) from late Eocene to Oligocene magmatic rocks of northern Nevada scattered Paleozoic zircons
Eastern Snake River System	Medium population (15%) of 0– to 10-Ma eastern SRP hotspot grains; minor 42– to 52-Ma Challis (5%) and 80– to 100-Ma Idaho batholith (10%) populations; diverse populations of 1000– to 1800-Ma Proterozoic recycled zircons, 5 to 10% d Archean zircons

A major population is >25%; a medium population is >10%; a minor population is <10%.

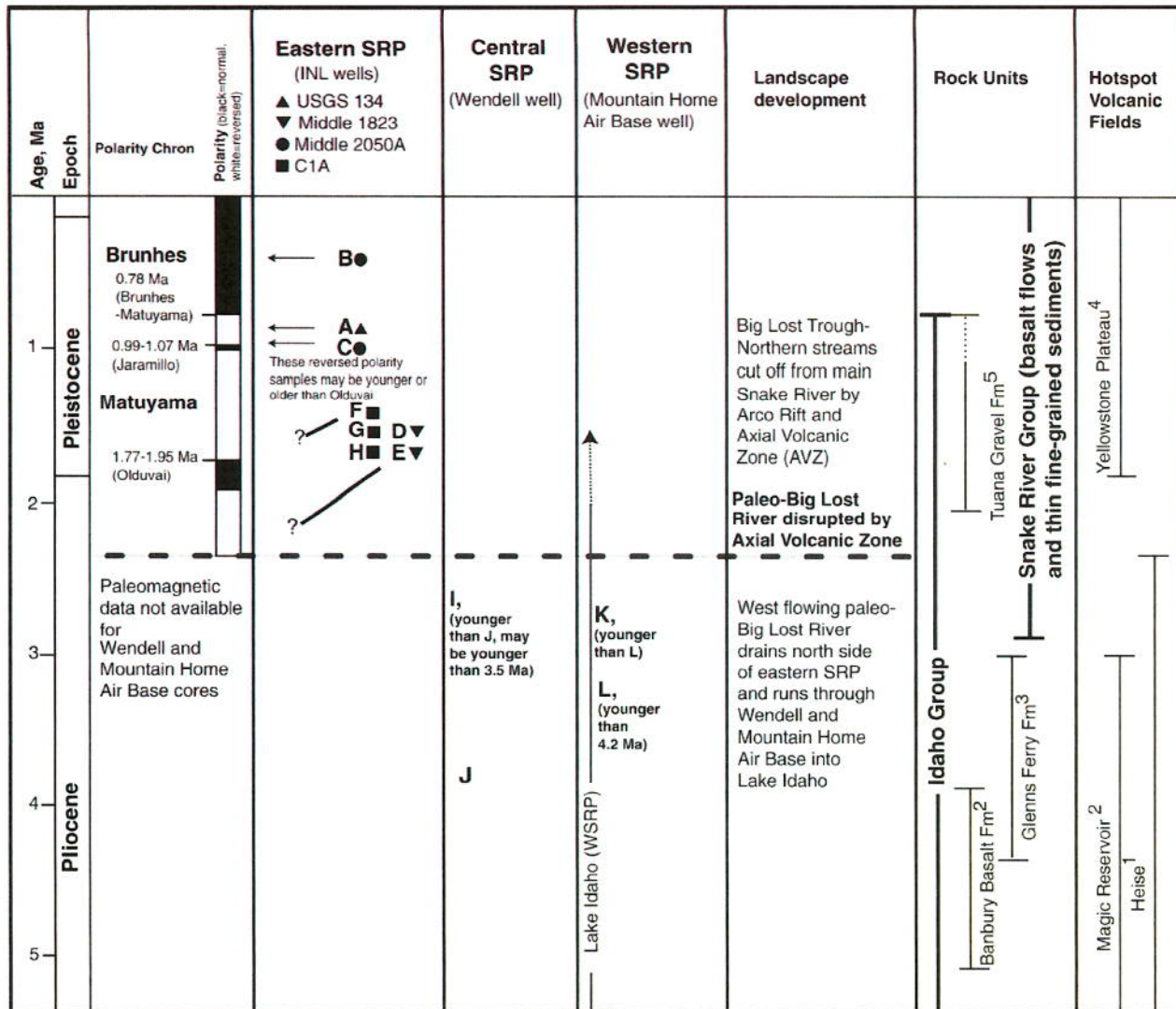


Fig. 2. Phases of drainage development on the SRP. The figure shows stratigraphic names, age constraints, magnetic polarity, and stratigraphic locations of detrital zircon samples. Transition from through-flowing paleo-Lost River to initiation of the Big Lost Trough is indicated by thick dashed line. General age constraints are after Pierce and Morgan (1992). Magnetic polarity data from Champion et al., (in press). Other references: ¹Morgan (1992) and Morgan and McIntosh (2005); ²Bonnichsen and Godchaux (2002); ³Hart and Brueske (1999); ⁴Christiansen (2001); ⁵Sadler and Link (1996); ⁶Tauxe et al. (2004, Table 4 and Fig. 12).

Large-scale silicic volcanism associated with the Yellowstone hotspot began about 17 million years ago in the McDermitt area on the Oregon–Nevada border (Pierce and Morgan, 1992). As the North American plate moved southwestward across the hotspot or deep melting anomaly a northeast-younging sequence of silicic volcanic fields was emplaced. On the eastern SRP, drainages were diverted radially away from each volcanic field as it developed, due to uplift associated with the hotspot (about 1 km, Pierce et al., 2002; Pierce and Morgan, 2009–this volume). After passage of the hotspot and subsidence these drainages were captured by the developing west-flowing Snake River system. As the North American plate continued to migrate over the hotspot, uplift and the location of the Continental Divide has apparently shifted eastward (Pierce and Morgan, 1992; Smith and Braile, 1994; Pierce et al., 2002). After passage over the hotspot, the uplifted SRP area subsided, due in part to emplacement of a mafic mid-crustal sill (Sparlin et al., 1982; Mabey, 1982; McQuarrie and Rodgers, 1998; Rodgers et al., 2002) and by thermal contraction (Pierce and Morgan, 1992; Smith and Braile, 1994).

Late Miocene to Holocene topographic development of the SRP, as predicted by Pierce and Morgan (1992) and Ore (1999) shows successive northeastward drainage capture by tributaries to the west-flowing Snake River system (Fritz and Sears, 1993). Proterozoic detrital zircons derived from the eastern Idaho thrust belt (Wood River or eastern Snake River drainage basins) only appear in sands of the western SRP after 7 Ma, suggesting the 10-Ma Continental Divide was in the Idaho batholith, west of the central Idaho thrust belt and over 200 km southwest of its present location (Beranek et al., 2006). Sandstones immediately below the 6-Ma Timber Hill Basalt, in the Sixmile Creek Formation of the Ruby Graben, southwest Montana, contain large populations of 9- to 12-Ma zircons deposited by a late-Miocene paleo-Missouri River and ultimately derived from the central SRP (Stroup et al., 2008a). Even if these grains were recycled from a proximal intermediate source area on the northern margin of the Heise volcanic field, the 6 Ma Continental Divide was at least 75 km southwest of its present location along the Idaho–Montana border (Fritz and Sears, 1993; Sears and Thomas, 2007; Stroup et al., 2008a).

In the central and eastern SRP, after eruption and subsequent subsidence of the Picabo (10.2–8.2 Ma; McCurry and Rodgers, 2009, Appendix B) and Heise (6.65–3.5 Ma; Morgan and McIntosh, 2005) volcanic fields, olivine tholeiite basaltic lavas filled the subsiding basins. These are intercalated with lacustrine, eolian, and fluvial sediments, which presently cover the SRP surface (LaPoint, 1977; Doherty et al., 1979; Scott, 1982; Kuntz et al., 1992; Geslin et al., 1999; Kuntz et al., 2003).

In late Pliocene time, eruption of basalt flows and intrusion of rhyolite domes along the axis of the eastern SRP formed the topographically elevated Axial Volcanic Zone (Hackett and Smith, 1992; Hackett et al., 2002). The oldest ages on Big Southern Butte (Fig. 1), the topographically highest point in the Axial Volcanic Zone, are 294 ± 15 ka and 309 ± 10 ka from aphanitic rhyolite and obsidian (Kuntz et al., 2007).

In general, Pleistocene basaltic eruptions along the length of the SRP (Bonnichsen and Godchaux, 2002) have ages with no systematic relation to the time-transgressive hotspot track. Ar/Ar ages of basalts exposed on the surface in the Craters of the Moon region, west of the Arco rift zone, range from less than 50 to 480 ka. USGS 135, a well drilled in the Arco rift zone in the southwestern corner of the INL site, penetrated to a depth of 365 m, and slightly more than 95% of the material penetrated is basalt (Brian Twining, USGS, pers. comm. 2009).

At about 2.5 Ma drainage of the Big Lost River, Little Lost River, Birch Creek and smaller streams to the east was ponded in an area known as the Big Lost Trough (Geslin et al., 1997; Gianinny et al., 1997; Geslin et al., 2002; Bestland et al., 2002; Blair, 2002). Deposition in the Big Lost Trough, which underlies much of the Idaho National Laboratory (INL), has been controlled by interplay between basalt and rhyolite eruptions and climate-related fluvial, lacustrine and eolian processes. The trough dates to at least 2.45 Ma based on Ar–Ar ages and paleomagnetic data from the 915-m deep core hole Well 2-2A (Fig. 1B), which terminates in the 4.45 Kilgore Tuff (Doherty, 1979; Bestland et al., 2002).

3. Detrital zircon “barcodes” from SRP tributary drainages

Previous work on detrital zircon grains in Holocene and Neogene drainages of the SRP indicates that provenance signatures of several modern drainages are distinctive. Multiple investigations show that detrital zircon provenance determinations are repeatable; these have been termed provenance “barcodes” (Link et al., 2005). The present study tests the premise that zircon-containing sands collected from drill core can be used to infer ancestral drainages.

Due to the complex history of compression, extension, and magmatism in the region, the geology north of the Snake River Plain is diverse. Drainages that developed in this Basin and Range system each contribute distinct age populations of zircon grains (see Table 1). Timing of drainage diversion and capture can be elucidated by determining the presence of distinctive detrital zircon age populations in sediments of known age recovered from boreholes or outcrops.

Previous detrital zircon studies, on the SRP and elsewhere (Gehrels and Dickinson, 1995; Geslin et al., 1999; Mahoney et al., 1999; Soreghan and Gehrels, 2000; Link et al., 2002; Ross and Villeneuve, 2003; Link et al., 2005, 2007; Dickinson and Gehrels, 2008; Stroup et al., 2008b) demonstrate that in fluvial systems the presence of a small defining population (5 to 10% of the total zircon grains) of a given age is a reliable indicator of provenance. In cases of small sample size (less than 60 zircon grains per sample) the absence of a particular population is a less-reliable provenance indicator.

There are several distinctive detrital zircon populations for river systems of the SRP region (Table 1). Some of these are defining populations that point to a specific source area. Other zircon populations are multiply recycled and only provide general provenance constraints. Of the recycled populations, Archean zircons are present in small amounts in all river drainages, likely recycled through Proterozoic or Phanerozoic sandstones. Recycled Laurentian Paleoproterozoic, Mesoproterozoic and Grenvillian (950–1300 Ma, making up 5% in the Wood River and >10% in the integrated Big and Little Lost River system) demonstrate provenance in the thrust belt of central and eastern Idaho. Paleozoic 350- to 400-Ma zircons in the Wood River and Salmon Falls Creek drainages are derived from Paleozoic western assemblages within the Antler allochthon.

A significant first-cycle defining detrital zircon population consists of Neoproterozoic 650- to 740-Ma zircons in the Big Lost River that are derived from a 695-Ma Neoproterozoic orthogneiss in the lower plate of the Pioneer core complex (Durk, 2007; Durk et al., 2007). Zircons from the southern, Atlanta lobe of the Idaho batholith (80–100 Ma) are present in the Wood River drainage, but not the Lost River drainages. Eocene grains from the Challis magmatic phase (42–52 Ma) are common in all drainages from the northern side of the SRP, and form over 80% of the grains in the Big Lost River. Jurassic (150–160 Ma) magmatic grains are derived from northern Nevada and are absent from drainages north of the SRP. Oligocene and late Eocene (30–40 Ma) grains from northern Nevada are distinctly younger than the Challis grains.

Yellowstone-SRP hotspot-aged grains are notably absent in all the modern drainages from the northern side of the eastern SRP, suggesting that while the adjacent Heise Volcanic Field was active, the area was generally a highland. This would be expected if the hotspot volcanic fields and their northern margin were tumescent, as proposed in a model for resurgent cauldrons (Smith and Bailey, 1968).

4. Materials and methods

Samples were collected from cores stored at the Lithologic Core Storage Library (CSL) of the Idaho National Laboratory, a Department of Energy facility administered by the U. S. Geological Survey. Drill core locations and depths are tabulated in Table 2. The stratigraphic positions of the samples and available paleomagnetic constraints are shown on the left side of Fig. 2. The zircon grain-ages are tabulated in Table 3 as well as in the digital database in the Appendix.

Table 2
Locations of core holes and sample depths.

Well name	Location (N latitude, W longitude)	Surface elevation (m)	Depth	Sample letters and depth in meters below land surface (BLS)	Depth of Brunhes–Matuyama boundary (m)	Depth of Jaramillo or Olduvai normal polarity interval (m)
USGS 134	43°36'11.15", 112°59'58.27"	1513	289	A-287–289 m	203	NA
Middle 2050A	43° 34'48", 112° 57'05.38"	1502	435	B-226 m; C-322–323 m	235	341
Middle 1823	43° 34'18.97", 112° 58'17.59"	1506	504	D-400 m; E-450 m	225	288
C1A	43° 30'23.63", 113° 02'03.72"	1532	550	F-402 m; G-443 m; H-465 m	213	343
Wendell core 7S-15E-12CBA5, originally 7S-15E-12CBA1	42° 49'55", 114° 39'03"	1097	342	I-137 m; J-329 m	NA	NA
Mountain Home Air Base—MHAB	43° 03'19", 115° 50'23"	920	1342	K-411 m; L-442 m	NA	NA

All core hole location information is north latitude, west longitude, NAD27, and altitude information is NGVD29. Altitudes were surveyed after completion of drilling, but depth information was recorded during drilling, which causes some inaccuracy in recorded altitudes.

Table 3

Sample designation, ages of zircon grains and one-sigma standard deviation for 12 samples from drill cores on the eastern and central SRP.

A. 59PL05 USGS 134 20 gr	B. 60PL05 Middle 2050A 23 gr	C. 61PL05 Middle 2050A 20 gr	D. 62PL05 Middle 1823 35 gr	E. 63PL05 Middle 1823 47 gr	F. 64PL05 CIA 39 gr	G. 65PL05 CIA 36 gr	H. 66PL05 CIA 38 gr	I. 67PL05 Wendell 101 gr	67PL05 (continued)	J. 68PL05 Wendell 60 gr	K. 51PL06 MHAB 60 gr	L. 52PL06 MHAB 25 gr														
Age (Ma)	±	Age (Ma)	=	Age (Ma)	±	Age (Ma)	=	Age (Ma)	±	Age (Ma)	=	Age (Ma)	±	Age (Ma)	=	Age (Ma)	±	Age (Ma)	±							
45.3	1.0	47.6	0.7	44.2	0.9	46.0	0.8	11.0	0.5	42.8	1.5	9.2	0.3	44.0	1.5	3.5	0.3	49.2	0.8	7.1	0.1	4.7	0.2	3.5	0.3	
46.5	0.9	47.8	1.0	47.0	0.6	46.2	0.8	26.5	1.4	46.0	1.7	10.5	0.3	44.5	0.8	7.2	0.2	49.6	0.9	7.2	0.6	7.2	0.4	4.2	0.2	
47.0	0.8	47.8	0.9	48.2	0.8	46.5	0.6	46.9	1.0	47.4	1.5	46.2	1.1	45.6	0.8	7.3	1.2	50.2	1.0	10.4	0.5	9.0	0.5	4.3	0.4	
47.5	0.6	48.9	1.1	48.4	0.8	47.5	0.7	47.6	9.4	47.7	1.0	46.5	0.9	45.7	0.7	7.7	0.9	50.4	1.3	25.1	0.7	9.2	0.6	4.7	0.2	
47.7	0.6	49.5	0.9	48.4	0.6	47.6	0.9	47.7	1.2	47.8	0.8	46.9	1.2	46.3	1.2	7.9	0.7	51.8	1.3	29.8	0.4	14.9	0.6	14.4	0.5	
48.3	1.3	49.8	0.7	48.5	0.9	47.6	0.9	48.4	1.3	48.1	0.7	47.2	0.9	46.6	1.4	8.4	0.6	52.0	0.9	45.6	1.1	15.0	0.8	14.9	0.3	
48.3	1.1	49.9	0.6	49.0	0.8	47.6	1.2	48.6	1.3	48.2	0.8	47.5	0.8	46.8	0.7	8.6	0.9	95.2	1.8	46.1	0.6	30.9	0.5	15.0	0.2	
48.8	1.0	50.4	0.8	49.4	1.1	48.7	0.8	49.1	1.1	48.7	0.8	48.0	0.9	47.1	0.8	8.7	1.6	98.5	1.3	46.5	2.1	34.6	0.6	15.3	0.4	
50.1	0.6	50.9	0.7	49.6	0.7	48.8	1.4	49.3	0.9	48.7	1.0	48.5	0.9	47.5	1.0	9.3	1.2	99.8	1.4	47.1	0.7	40.9	0.8	15.8	0.4	
50.2	1.4	51.0	1.1	49.8	0.6	48.9	0.7	49.8	1.0	48.8	1.0	48.5	1.2	48.0	0.8	9.4	1.1	204.7	2.3	47.2	0.8	49.0	0.6	16.5	0.5	
51.3	1.2	51.7	1.0	50.2	0.9	49.0	1.1	49.8	0.9	49.0	0.7	48.5	1.3	48.1	0.9	9.6	3.0	431.5	5.1	47.3	0.8	49.7	0.7	51.6	0.7	
52.8	1.2	53.2	0.6	50.7	1.0	49.1	1.2	49.9	0.9	49.0	0.6	48.8	0.8	48.1	0.7	9.7	1.7	435.9	5.0	47.7	0.8	49.7	0.8	52.9	0.6	
54.0	0.8	53.2	1.3	51.2	0.8	49.2	0.6	49.9	1.2	49.0	0.9	49.0	1.0	48.6	0.6	9.7	0.8	545.5	6.4	48.0	0.9	49.9	0.7	96.8	1.3	
123.9	1.9	54.1	0.6	456.9	5.5	49.9	1.2	50.6	0.8	49.3	1.0	49.2	0.8	48.9	0.7	9.9	0.5	558.7	6.3	48.2	0.9	50.0	0.7	99.9	1.5	
393.1	5.0	54.5	0.9	510.5	7.5	50.0	0.6	50.8	1.1	49.4	0.9	49.3	0.8	48.9	0.9	9.9	1.5	577.9	6.7	48.2	1.2	51.0	0.8	697.2	11.8	
443.5	4.9	130.2	2.1	701.1	11.4	50.0	0.9	50.8	0.9	49.4	0.7	49.4	1.2	48.9	1.0	9.9	0.5	649.1	8.0	48.2	1.3	51.0	0.6	827.5	11.5	
694.4	7.3	336.8	3.7	701.6	9.1	50.0	1.5	51.0	1.1	49.6	0.6	49.5	0.6	49.0	1.3	10.0	0.6	667.0	9.2	48.6	0.9	51.3	0.9	1042.1	11.2	
703.2	8.3	667.7	7.7	724.3	9.5	50.1	0.9	51.0	1.0	49.6	0.8	49.5	1.3	49.1	0.9	10.1	0.8	674.0	8.2	48.6	0.8	67.3	2.2	1255.0	14.5	
704.5	7.2	737.6	7.6	1772.4	16.2	50.6	1.6	51.8	1.2	49.6	0.9	49.5	1.2	49.1	0.8	10.2	1.2	1004.8	11.4	48.9	1.0	90.0	1.3	1436.8	17.8	
		1162.1	26.6	1830.1	30.5	50.9	0.6	52.3	0.9	49.9	0.8	49.6	0.7	49.2	1.5	10.3	0.5	1079.7	14.3	49.1	0.9	91.2	1.1	1480.1	12.9	
		1442.9	10.4			51.1	0.8	52.4	1.3	50.0	1.1	49.6	0.7	49.2	0.7	10.3	0.9	1131.9	18.4	49.2	1.0	94.7	1.9	1496.8	14.8	
		1946.7	22.2			51.1	4.6	52.5	1.0	50.6	0.6	49.7	0.6	49.3	0.8	10.5	0.3	1139.6	21.5	49.6	0.8	99.9	1.3	1628.2	14.4	
						51.7	0.6	53.5	1.9	50.6	0.8	49.8	0.9	49.9	0.9	10.7	0.6	1160.1	22.6	49.8	0.7	100.1	1.7	1728.5	7.1	
						52.4	1.6	53.8	1.3	51.0	1.0	49.9	0.8	50.0	0.9	11.0	0.3	1193.2	29.0	50.7	0.8	152.0	1.7	1779.6	8.1	
						62.4	0.8	56.7	7.8	52.7	0.6	50.5	0.6	51.4	1.0	11.1	0.6	1245.2	14.1	52.0	0.9	156.2	2.0			
						94.4	1.4	229.9	6.4	275.8	3.5	50.5	1.2	51.5	2.0	11.3	1.0	1299.3	21.0	53.5	0.8	159.5	1.9			
						110.0	1.5	701.5	11.3	548.4	8.5	51.9	0.8	52.6	0.7	11.5	1.1	1358.0	30.6	53.8	0.8	159.7	1.8			
						368.7	5.2	1458.9	23.5	602.7	6.1	52.9	0.6	52.7	1.0	11.5	0.6	1394.1	37.0	91.7	2.4	162.3	1.9			
						429.4	6.1	1471.6	30.3	758.7	8.4	53.0	0.8	471.6	8.1	11.7	0.6	1577.9	25.3	96.2	2.0	162.9	1.9			
						539.2	6.3	1686.7	13.7	1177.9	13.6	70.8	0.8	543.7	7.4	11.7	0.7	1687.9	29.4	99.1	1.6	342.6	4.0			
						634.3	7.2	1687.0	15.3	1190.7	13.5	404.9	5.1	624.2	10.7	11.8	1.5	1711.0	41.4	105.5	1.6	419.0	4.7			
						636.1	16.2	1697.5	15.8	1381.0	20.2	555.4	6.4	676.2	8.2	12.8	0.4	1745.5	14.3	168.4	1.9	601.3	9.9			
						662.1	10.2	1698.3	10.9	1434.3	5.7	576.4	7.5	679.9	9.7	13.0	0.8	1816.6	21.2	355.1	4.8	609.8	7.7			
						1462.5	35.7	1704.8	18.2	1654.1	10.2	678.7	7.9	684.3	11.1	13.6	0.3	1842.7	33.6	408.0	4.8	624.1	7.6			
						1957.1	54.2	1707.7	14.6	1710.7	30.5	1152.2	11.6	727.5	8.1	13.6	0.4	1846.6	35.5	501.4	7.0	983.4	10.9			
								1709.2	15.6	1718.2	14.0			1438.6	39.0	14.1	0.3	1936.8	26.0	574.2	8.5	1081.5	15.3			
								1731.7	16.1	1792.0	81.9					14.4	0.9	2512.0	25.3	578.7	8.5	1140.6	11.3			
								1742.4	13.1	1847.2	73.2					14.4	0.8	2548.0	31.1	668.0	13.1	1285.3	10.6			
								1745.5	12.0							14.4	2.4	2617.2	12.2	886.0	14.7	1326.0	11.4			
								1757.2	13.4							15.1	0.8	2737.1	20.3	995.1	13.1	1393.1	22.0			
								1762.6	17.8							31.4	0.6	2756.8	28.1	1003.9	11.2	1433.7	19.0			
								1780.7	40.7							46.2	0.8		1016.7	13.4	1453.8	14.1				
								1795.1	20.9							46.5	1.3		1030.4	12.1	1502.6	17.3				
								1834.2	27.7							46.6	1.1		1043.7	18.6	1593.8	12.4				
								1878.7	31.8							46.7	0.7		1056.2	15.2	1611.9	10.0				
								1879.6	33.2							47.1	0.8		1078.4	16.3	1632.8	11.8				
								1948.1	27.7							47.2	0.9		1166.0	20.9	1659.5	21.7				
																47.9	0.8		1171.3	17.0	1669.8	10.5				
																48.0	1.1		1449.3	49.2	1699.2	10.4				
																48.1	0.7		1450.7	32.6	1722.6	10.3				
																48.1	0.9		1528.0	18.5	1746.4	6.4				
																48.3	0.9		1630.4	27.3	1788.7	10.6				
																48.8	0.8		1655.9	16.9	1805.0	12.3				
																			1697.7	21.8	2261.3	8.7				
																			1778.2	18.6	2716.8	4.4				

Samples were processed, mineral separates prepared, cathodoluminescence images analyzed and grains were dated using Sensitive High Resolution Ion Microprobe (SHRIMP) facilities at The Australian National University following protocols described in Williams (1998) and Link et al. (2005, p.104). Spot sizes were generally 30 μm and were chosen either in the clearest part of magmatic grains, or in the youngest rim phase of polygenetic zircon grains. Uncertainty is reported at one sigma, and is about 1%. Data were reduced using the SQUID utility, and analyzed using the Microsoft Excel Isoplot Macro of Ludwig (2003). ^{207}Pb – ^{206}Pb ages more than 10% discordant (less than 5% of total) were removed from the sample and not shown in Fig. 3.

5. Results

Results are plotted in Fig. 3 (samples are capital letters A to L) with grains < 150 Ma shown in the upper part of the plots and older grains in the lower part of the plots. Grain ages for each sample are tabulated in Table 3. Ideally, 60 or 100 detrital zircon grains are needed to establish a provenance signature reliable to a population present at 5% (Dodson et al., 1988; Vermeesch, 2004). In the present study, some drill-hole samples were too small to yield sufficient zircon grains. In other cases, the analysis was stopped after as few as 20 gr because the provenance was clear. We used a 2-grain (preferably 3 grain) minimum grain population before recognizing with confidence the presence of that provenance. This applies in particular to samples from the Big Lost Trough, where a large Challis and small Neoproterozoic population identifies sediment from the Big Lost River.

5.1. USGS 134 core

Sample A is from core USGS 134 (287–289 m below land surface or BLS) and has a sample age greater than 780 ka (Fig. 2). This 20-grain sample has the Big Lost River signature, with a strong Challis age peak and two Neoproterozoic grains.

5.2. Middle 2050A core

Samples B and C are from Middle 2050A. The Brunhes–Matuyama boundary is at 235 m depth in Middle 2050A, therefore, sample B (226 m–23 gr analyzed) is a little younger than 780 ka and sample C (322–323 m–20 gr analyzed) is between 0.99 and 1.07 Ma (Champion et al., in press). Both samples have Big Lost River provenance with major Challis age peaks and small Neoproterozoic populations (5 to 12%).

5.3. Middle 1823 core

Samples D and E are from Middle 1823. The Brunhes–Matuyama boundary is at approximately 225 m BLS (Champion et al., in press). An interval of normal polarity basalt is found between 288 and 359 m BLS which may have erupted during either the Jaramillo (0.99 to 1.07 Ma) or Olduvai (1.77 to 1.95 Ma) normal polarity subchron. Sample D was from 400 m (35 gr analyzed) and sample E (47 gr analyzed) was from 450 m. Both samples lie beneath the undated deep normal polarity interval and are at least 0.99 Ma. If the undated deep normal polarity interval represents the Olduvai normal polarity subchron, they may be older than 1.95 Ma.

Sample D has a strong Challis peak and three Neoproterozoic grains (10% of total). Sample D is sand from an ancestral Big Lost River. Sample E has a strong Challis peak and a strong Paleoproterozoic peak. Neoproterozoic zircons are not found in sufficient quantity to interpret. Sample E is most probably from an ancestral Little Lost River.

5.4. C1A core

Samples F, G, and H are from C1A on the INL site (Fig. 1). In C1A, the Brunhes–Matuyama boundary is at 213 m BLS. An interval of normal

paleomagnetic polarity begins at about 343 m BLS and may represent either the Jaramillo Normal Polarity Subchron (0.99 to 1.07 Ma), or the Olduvai Subchron (1.77 to 1.95 Ma, Champion et al., in press). The samples are thus older than 0.99 Ma and may be older than 1.77 Ma.

Sample F (402 m BLS–39 gr) lies between normal polarity basalts (Fig. 2) considerably deeper than the Brunhes–Matuyama boundary. This sample has a strong Challis peak, one Neoproterozoic grain and several Mesoproterozoic and Paleoproterozoic grains. It is fine-grained lacustrine sand with interpreted mixed provenance from both the Big and Little Lost River.

Sample G (443 m BLS–36 gr) lies between a normal polarity basalt flow and a reversed polarity basalt flow. This sample has a strong Challis peak and two Neoproterozoic (Neoproterozoic) grains (5%). It is sand from an ancestral Big Lost River.

Sample H (465 m BLS–38 gr) lies between two reversed polarity basalts and it is older than sample G. This sample represents an ancestral Big Lost River, with a strong Challis peak and a moderate (10%) Neoproterozoic peak. Paleoproterozoic grains are absent.

In summary, these INL samples (all from the Snake River Group; Whitehead, 1992) demonstrate that the Big Lost Trough has existed since late Pliocene time. Provenance for the sampled sediments was mainly in the Big Lost River system, and secondarily in the Little Lost River system. The older Middle 1823 sample E was deposited by the Little Lost River. The fine-grained C1A sample F has zircon signatures of both the Big Lost River and Little Lost River.

5.5. Wendell core, Central SRP

Samples I and J are from the 342 m deep Wendell core (Fig. 1A), drilled on the central SRP, as part of the U. S. Geological Survey Regional Aquifer System Analysis. At the surface it penetrates the Pleistocene basalt of Notch Butte, which overlies the Pliocene Glens Ferry Formation (Fig. 2) (Hart and Brueseke, 1999; Kauffman and Othberg, 2004).

Sample I (137 m BLS–101 zircon grains) is from the Glens Ferry Formation (Fig. 2) (Whitehead, 1992, p. B9, Hart and Brueseke, 1999). It has a larger, more diverse population of zircon ages than samples from the INL cores, reflecting contributions from a more integrated second-order drainage basin (Link et al., 2005). This sample has a major Miocene rhyolite-derived zircon population (40%) with grains as young as 3.5 Ma, and most having ages of 7 to 15 Ma. There is a prominent population (20%) of zircons with ages from 42 to 52 Ma, representing the Challis magmatic episode. Dispersed Paleoproterozoic, Mesoproterozoic and Grenvillian (950 to 1300 Ma) grains suggest recycling through strata of the central Idaho thrust belt. Jurassic grains that would indicate provenance in northern Nevada are absent. There are three Cretaceous grains with ages consistent with the adjacent Atlanta Lobe, Idaho Batholith.

Six grains are Neoproterozoic (545 ± 7 Ma, 559 ± 6 Ma, 578 ± 7 Ma, 649 ± 8 Ma, 667 ± 9 Ma and 674 ± 8 Ma). The three older Cryogenian (>635 Ma) grains are interpreted to be derived from the Pioneer Mountains core complex via Wildhorse Creek and the Big Lost River. The three Ediacaran grains (540 to 600 Ma) may come ultimately from the belt of latest Neoproterozoic alkalic plutons found in central Idaho (Lund et al., in press, locations shown in Fig. 1A).

Thus, sample I represents Pliocene drainage from the northern side of the Snake River Plain, with sediment from both the Lost and Wood River systems. This suggests that a Pliocene paleo-Lost River drained westward joining the paleo-Wood River, forming a delta into Lake Idaho (the Glens Ferry Formation at Wendell).

Sample J (329 m BLS–60 gr) is from the Pliocene or late Miocene middle unit of the Banbury Basalt Formation (Whitehead, 1992, p. B9). The upper basalt in the Banbury Basalt is thought to be about 4 Ma, and the lower basalt that underlies this sand is thought to be about 10 Ma, (D. Champion, personal communication, 2009). The sample has a few Snake River Plain–Yellowstone grains, with a strong Challis peak (17 gr) and a small Idaho batholith peak (5 gr).

There are three Neoproterozoic grains (574 ± 9 Ma, 579 ± 9 Ma, and 668 ± 13 Ma). This sample probably also records drainage from the paleo-Wood and Lost River systems. It has few hotspot track grains, even though it was likely west of the Pliocene continental divide. The inference is that during late-Miocene and early-Pliocene formation of the Heise volcanic field, northern hotspot-related volcanics were drained by streams flowing northeast (Stroup et al., 2008a).

5.6. Mountain Home Air Base, western SRP

Two samples were taken from the 1342 m deep Mountain Home Air Base (MHAB) geothermal exploration borehole, which was drilled at Mountain Home Air Base in Elmore County, Idaho, on the western Snake River Plain. The upper 305 m were described from cuttings, as basalt with interbedded sediment that ranged from silt to coarse sand from the surface to 162 m depth, and as silty sand and silty clay from

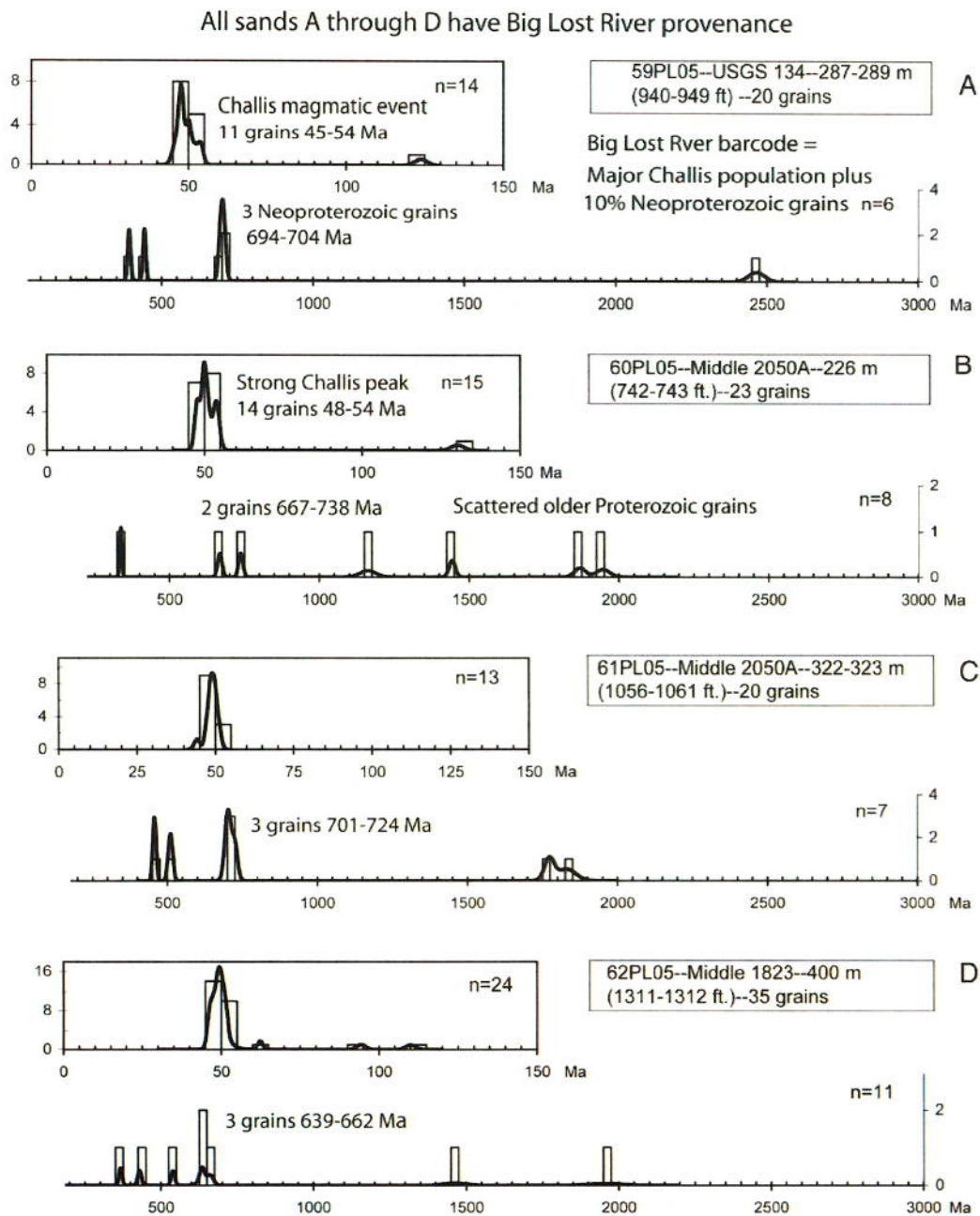


Fig. 3. Probability–frequency plots for SHRIMP U–Pb ages for detrital zircon grains in Neogene to Holocene Snake River Plain drill-hole sands. Grains younger than 150 Ma are plotted in the upper part of diagram and grains older than 150 Ma are shown in the lower half. Note that some samples have 60 to 100 gr, as is ideal, but many have less than 30 gr. The smaller sample sizes are due to either low grain-yield from drill-core sample, or the decision to stop analysis because a clear provenance signature had emerged. Definitive populations are labeled.

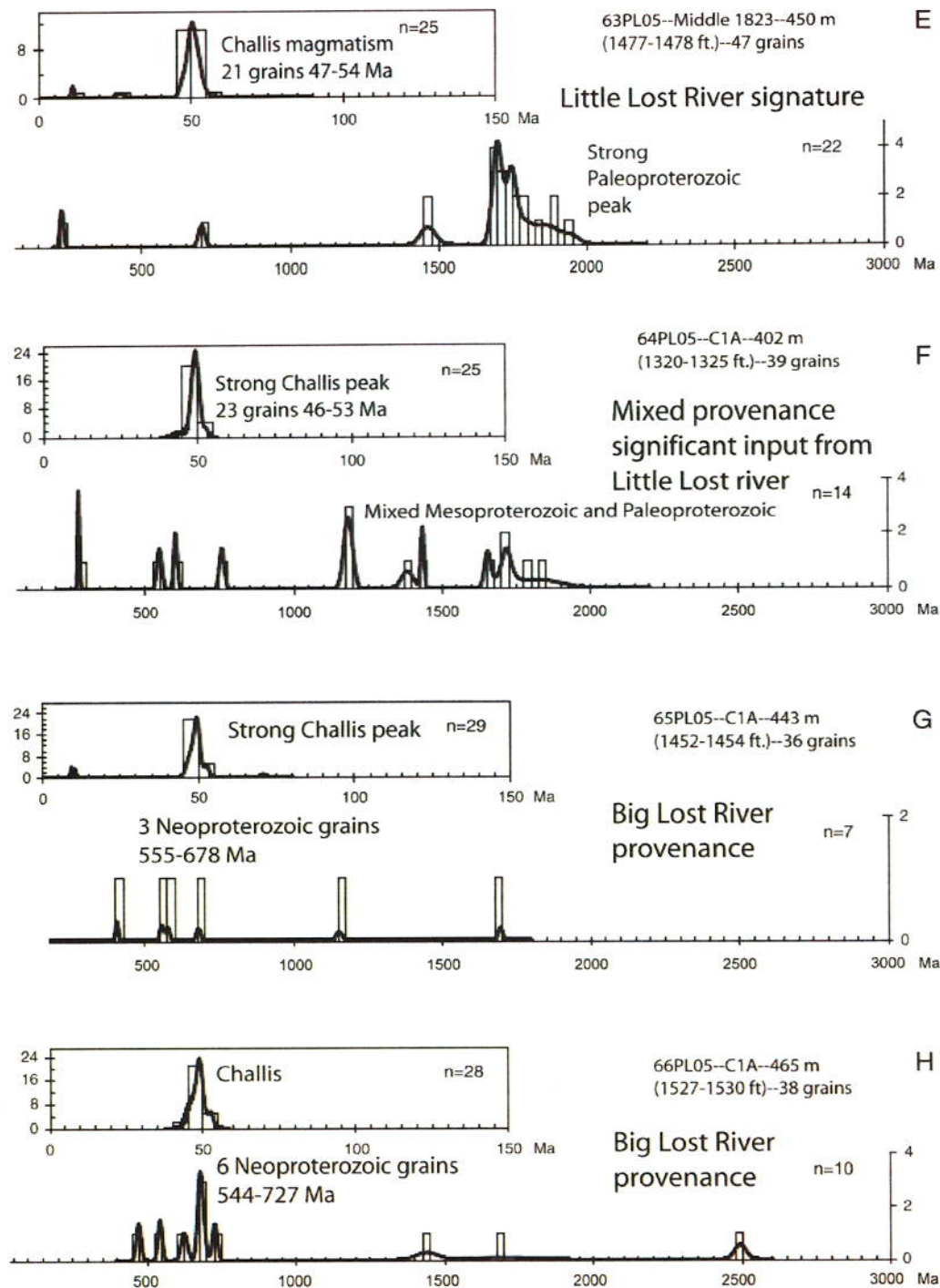


Fig. 3. (continued).

162 to 579 m BLS (Lewis and Stone, 1988). Both sample K (411 m BLS--60 gr) and sample L (442 m BLS--25 gr) are from the Idaho Group which underlies the plateau-forming Quaternary-Tertiary basalt of Rattlesnake Springs (Shervais et al., 2002, p. 349).

Sample K contains zircon populations consistent with the integrated, modern Snake River system, including Yellowstone hot-spot grains (5–15 Ma), 150- to 160-Ma Jurassic zircons and late

Eocene (35 to 40 Ma) zircons from northern Nevada, Challis (42 to 52 Ma) zircons, Idaho batholith zircons (80 to 100 Ma), and diverse Grenvillian (950–1300 Ma), Mesoproterozoic, Paleoproterozoic and Archean grains.

Sample L contains small populations of Challis age, Idaho batholith ages, and one Neoproterozoic zircon. It contains two distinct sets of Miocene zircons; 14 to 16 Ma and 3 to 5 Ma, but 8- to 12-Ma zircons

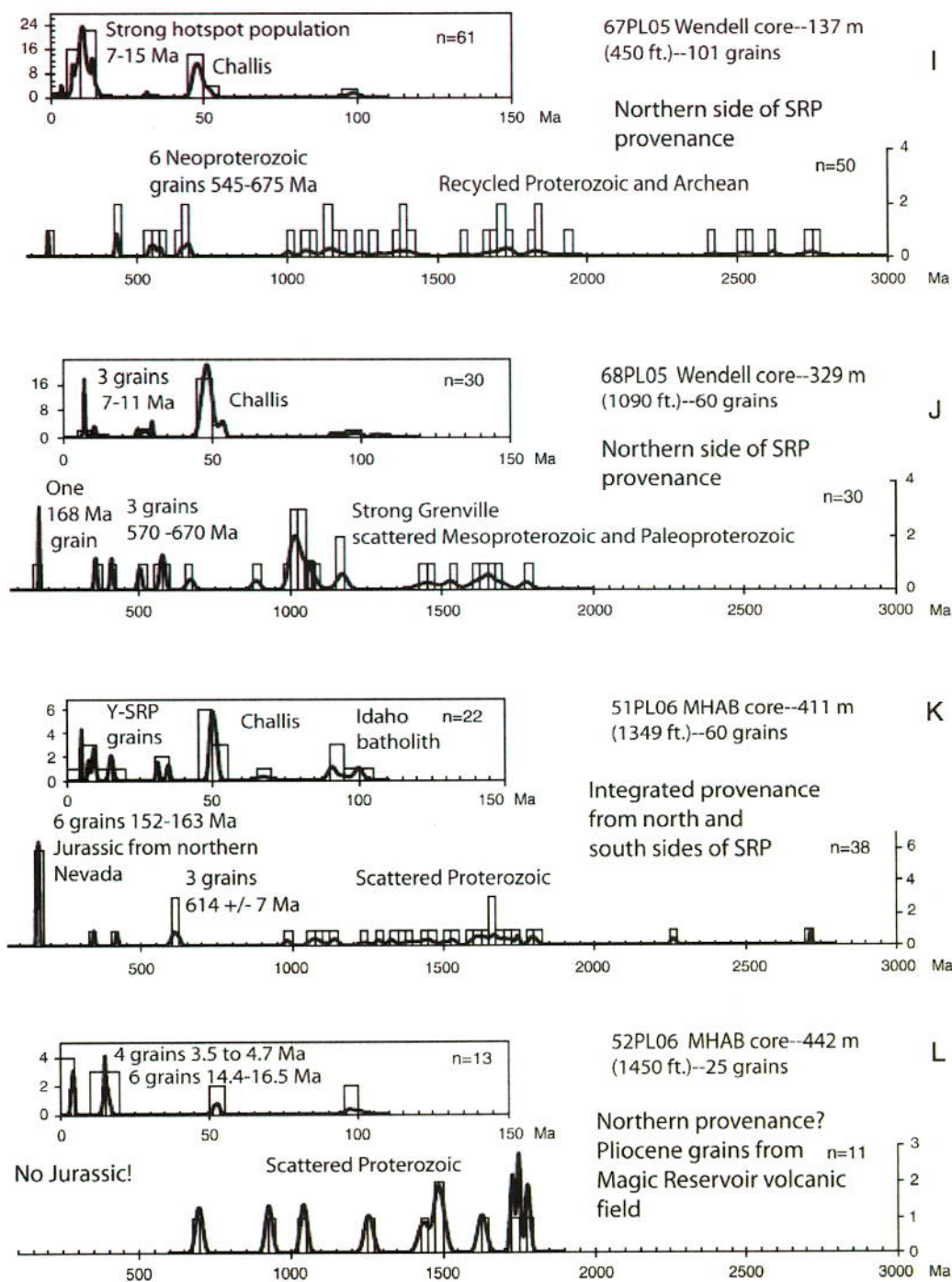


Fig. 3. (continued).

from the central SRP are conspicuously absent. The Pliocene zircons may come from the Magic Reservoir volcanic field, a late-stage rhyolitic field northeast of Wendell (Fig. 1A; Schmidt, 1962; Leeman, 1982). Other zircon ages in this sample are consistent with drainages from the north side of the eastern SRP, notably the Big Lost River. The lack of 8 to 12-Ma grains is surprising since the nearby Lake Hills and Mount Bennett Hills contain rhyolites of this age (Schmidt, 1962; Oakley and Link, 2006). This is not easily explained. More sampling

may reveal a more complex story. The most interesting zircon in sample L is the 697-Ma grain, which could only have come from the Pioneer Mountains in the headwaters of the Big Lost River.

6. Discussion

Neogene drainage development of the Snake River Plain region was controlled by several factors including subsidence of the western

Snake River Plain extensional rift, which was periodically filled by Lake Idaho after 9 Ma, subsidence of the central and eastern SRP hotspot track, and cutting of the drainage divide to Hell's Canyon and the Salmon River at 3 to 2 Ma (Malde and Powers, 1962; Swirydzuk et al., 1982; Malde, 1991; Sadler and Link, 1996; Wood and Clemens, 2002; Bonnicksen and Godchaux, 2002). Within this context, our new samples demonstrate the presence of paleo-Lost River sediment in the drillcores at Wendell and Mountain Home.

A late Miocene and Pliocene paleo-Lost and Wood River system flowed westward along the northern side of the SRP. The detrital zircon signature of this river, found in samples I, J and L, consists of major populations of Challis and Idaho batholith grains, a scattering of Paleozoic and older Proterozoic grains, and up to 10% Neoproterozoic grains. Jurassic zircons from northern Nevada are not found. Zircons of Yellowstone hotspot track age found in the Wendell and Mountain Home wells are interpreted to have been derived from the 3- to 6-Ma Magic Reservoir volcanic field northeast of Wendell (Honjo et al., 1992) and from 8- to 12-Ma central SRP rhyolites now exposed in the Mount Bennett Hills and Lake Hills north of Wendell (Oakley and Link, 2006).

At Mountain Home, by the time of late Pliocene deposition of sample K, an integrated west-flowing Snake River drainage system existed, carrying grain populations derived from both the northern and southern sides of the SRP. This is the Snake River system we know today.

A Neoproterozoic grain population (545 to 750 Ma) is present in several samples from the Wendell and Mountain Home cores. We interpret these grains to be recycled from the 650 to 700 Ma Cryogenian zircons that form about 10% of the zircon populations in sands of the Big Lost River system and in the Big Lost Trough. The ultimate source of the 650 to 700 Ma Cryogenian grains is a 695-Ma orthogneiss in the headwaters of Wildhorse Creek that intrudes Mesoproterozoic and Archean gneisses in the lower plate of the Pioneer Mountains metamorphic core complex (Dover, 1983; Rodgers et al., 1995; Durk, 2007; Durk et al., 2007). Some 550 to 650 Ma Ediacaran grains may come from alkalic plutons of central Idaho (Lund et al., in press).

The only other place in eastern Idaho that could be the source area for ~700-Ma zircons is the Pocatello Formation, east of Pocatello, which is now drained by the Portneuf River (Link et al., 2005). The Portneuf River enters the main drainage of the Snake River upstream from Salmon Falls Creek, and there is no discernable zircon signature in the Wendell well samples that indicates contributions from Salmon Falls Creek or other southern drainages. Since the detrital zircon signal from the ~700-Ma Bannock Volcanic Member of the Pocatello Formation is weak (less than 5% of Portneuf River sands), and no evidence of contributions from southern drainages is found in this sample, it is very unlikely that the Bannock Volcanics are the source of this age population of zircons.

After establishment of the Big Lost Trough about 2.5 Ma, in the area of the central INL where the several wells are drilled, coarse-grained glacio-fluvial sediment from the Big Lost River (samples A,B,C,D,G, and H) interfingered with finer sediment sourced in the Little Lost River (sample E). These two signatures are mixed in lacustrine sediment of sample F which represents an integrated Lost River system. Thus, during highstands of recurring Pleistocene lakes within the Big Lost Trough, detrital zircon populations were integrated. Sequence stratigraphy controlled provenance.

7. Conclusions

We infer that the Pliocene (~5 to 3 Ma) west-flowing paleo-Big and Little Lost River flowed westward through Wendell to Mountain Home where it provided sediment for Glens Ferry Formation lakebeds on the northern shores of Lake Idaho. We hypothesize that in late Pliocene or early Pleistocene time (3 to 2 Ma) this west-flowing drainage was dammed by lava from basaltic eruptive centers of the

Arco Volcanic Rift zone near Craters of the Moon, and uplift associated with rhyolite domes of the Axial Volcanic Zone. Our hypothesis predicts that 1- to 2-Ma olivine tholeiite basalt lava flows are present in the subsurface at Craters of the Moon, and that these flows, on the north flank of the Axial Volcanic Zone, stopped the integrated Big and Little Lost Rivers from flowing southwest after about 2 Ma. This northwest-striking volcanic upland, roughly parallel with the Basin and Range faults marginal to the SRP, dammed the paleo-Lost River and forced it to flow northeast into the Big Lost Trough (cf. Helm-Clark and Link, 2006; 2007). This hypothesis can be tested by drilling and geochronology in the area of the Craters of the Moon–Arco volcanic system (Fig. 1B). Detrital zircon provenance studies of Big Lost Trough subsurface sediment can thus detect changes from lowstand fluvial to highstand pluvial depositional systems.

The bedrock Holocene surface of the eastern SRP consists of late-stage basaltic volcanoes and rhyolite domes that overlie the subsided rhyolitic hotspot track. As these volcanic highlands were built, lava flows, vent and fissure systems, and domes dammed and disrupted streams and lakes. The internally-drained, pluvial-playa Big Lost Trough system is the result of largely basaltic volcanism since late Pliocene time.

The eastern SRP near the INL today contains internal drainage of the Big Lost Trough. If late-stage basaltic volcanoes are built in a variable manner on a subsided rhyolitic hotspot track, regional drainage will be disrupted and new patterns result. Detrital zircon studies can track such changes. Modeling of the Snake River Plain subsurface hydrology will be enhanced by inferences about first-order stratigraphic and volcanic architecture dictated by zircon-age data.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.jvolgeores.2009.08.019.

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