500–490 Ma detrital zircons in Upper Cambrian Worm Creek and correlative sandstones, Idaho, Montana, and Wyoming: Magmatism and tectonism within the passive margin

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ABSTRACT

Upper Cambrian feldspathic sandstones deposited across southeast Idaho, Montana, and Wyoming (USA) during the Sauk II-Sauk III regression boundary contain distinctive 500–490 Ma detrital zircon grains, derived from Late Cambrian plutons in the Lemhi arch of east-central Idaho. The Worm Creek Quartzite Member of the St. Charles Formation in the Paris plate of the southeast Idaho thrust belt contains as much as 320 m of feldspathic fine-grained sandstone within a thick section of carbonate rocks. The near-unimodal age of hundreds of detrital zircons from 8 samples of the Worm Creek is 497 Ma. This age and the initial $\varepsilon_{\text{Hf}}$ values from these detrital zircons ($\varepsilon_{\text{Hf}}$ of $-8.0 \pm 1.9$ to $5.4 \pm 1.2$) overlap the age and isotopic composition of the Deep Creek and Beaverhead plutons intruded into the Lemhi arch ($\varepsilon_{\text{Hf}}$ of $-6.3 \pm 1.1$ to $2.7 \pm 1.4$). This suggests rapid unroofing of the hypabyssal alkalic plutons, which were the primary source for the sandstones. In the plutons, intermediate initial $\varepsilon_{\text{Hf}}$ values are neither juvenile nor evolved, suggesting mixing with a Mesoproterozoic component. A 493–488 Ma detrital zircon age peak is also found in Upper Cambrian sandstones (from the Sauk II-III boundary) in the Wind River Canyon on the Wyoming craton, the Melrose area of the southwest Montana thrust belt, and the Leaton Gulch area of the central Idaho thrust belt. The detrital zircon signatures of these Upper Cambrian rocks is markedly different from that of the Lower Cambrian upper Brigham Group in southeast Idaho and the Middle Cambrian Flathead Sandstone at Teton Pass, Wyoming (1790 Ma age peak). The overlying Middle Ordovician Swan Peak and Kinnikinic Quartzites from Idaho south to Nevada contain a different detrital zircon age population, with almost all grains older than 1800 Ma and a peak at 1860 Ma.

We suggest that the Lemhi arch is a relatively unextended crustal block coincident with the northwest-trending Mesoproterozoic Lemhi subbasin of the Belt Supergroup and with ca. 1.37 Ga mafic magmatism. This magmatism strengthened the lower crust and predisposed the Lemhi arch to remain intact during extension and Neoproterozoic rifting of western Laurentia. Oblique normal faulting and subsidence along the dextral normal Snake River transfer fault produced the Late Cambrian Worm Creek basin and juxtaposed active Cambrian magmatism and exhumation with passive-margin sedimentation to the south.

INTRODUCTION

In western North America, Neoproterozoic volcanism and siliciclastic sedimentation associated with protracted Rodinian rifting was followed by Cambrian passive margin sedimentation with deposition of thick carbonate platform strata (Bond et al., 1985; Christie-Blick and Levy, 1989; Link et al., 1993; Dickinson, 2004; Yonkee et al., 2014). South of the modern Snake River Plain, >6 km of Neoproterozoic and lower Paleozoic strata were deposited within the westward-thickening passive margin as part of the Sauk megasequence (Fig. 1; e.g., Sloss, 1963; Bond and Kominz, 1984; Bush et al., 2012). Cambrian strata are predominately carbonate rocks, but throughout the Paris plate of the southeast Idaho thrust belt, at the sequence boundary or regressive maximum of the Sauk II-Sauk III contact, feldspathic sandstones interrupt the carbonate-dominated sequence (Armstrong and Oriel, 1965).

In east-central Idaho, there are no Neoproterozoic and Cambrian passive margin strata across the Lemhi arch (Fig. 1; Sloss, 1954; Ruppel, 1986). In this area, Cambrian to Ordovician alkalic plutons of the Beaverhead magmatic belt were emplaced into Mesoproterozoic Belt Supergroup strata (Evans and Zartman, 1988; Lund et al., 2010). Because Middle Ordovician Kinnikinic Quartzite unconformably overlies the Mesoproterozoic Belt Supergroup (Fig. 1), there is no evidence of Neoproterozoic and Cambrian rift and post-rift subsidence and sedimentation. The Transcontinental Arch, a poorly defined uplift trending northeast-southwest across the western midcontinent of North America (Sloss, 1988; Amato and Mack, 2012), is interpreted to have been uplifted in Middle Cambrian time, such that it blocked transport of Grenville-aged detrital zircon grains from eastern Laurentia to the western passive margin (Yonkee et al., 2014; Linde et al., 2014). East of the arch, in southern New Mexico, 509–504 Ma Late Cambrian detrital zircons with an interpreted proximal magmatic source are found in the Bliss Sandstone (Amato and Mack, 2012). Early and Middle Cambrian magmatism (539–528 Ma) is recognized in the southern Oklahoma aulacogen (Hogan and Gilbert, 1998). West of the Transcontinental Arch, sparse 500 ± 15 Ma diabase dikes are present.
Figure 1. (A) Present-day outcrop map of the U.S. northern Rocky Mountains. Black square denotes the map extent of Figure 3. Modified from Muehlberger (1983), Vuke et al. (2007), Lund et al. (2010), and Lewis et al. (2012). Ck—Creek. U-Pb zircon sample localities (p = plutonic sample): BHp—Beaverhead pluton; DCp—Deep Creek pluton; LG—Leaton Gulch, MC—Midnight Creek; ML—Melrose; MP—McPherson Canyon; SC—Secret Canyon; TP—Teton Pass; WC—Weston Canyon; WR—Wind River Canyon. (B) Basement domains (modified from Gaschnig et al., 2013), approximate extent of the Lemhi arch (modified from Ruppel, 1986), and location of ca. 1.4 Ga plutons and isotopically delineated 1.4 Ga lithosphere (Doughty and Chamberlain, 1996; Elk City domain of Gaschnig et al., 2013). Line of cross section of Figure 11 is shown. SRTF—Snake River transfer fault (modified from Lund, 2008); GFTZ—Great Falls tectonic zone.
in west-central Colorado (Larsen et al., 1985), but such mafic rocks are zircon poor. Cambrian zircon grains have not been detected in regional studies of Paleozoic strata of the Cordillera (Gehrels and Pecha, 2014) or in sandstones from western and northwestern Canada (Hadlari et al., 2012).

Within the Upper Cambrian carbonate rocks1 of the passive margin in southeast Idaho on the Paris thrust plate (Fig. 1), the presence of 0–320 m of feldspathic sandstone in the uppermost Cambrian Worm Creek Quartzite Member of the St. Charles Formation has been known for more than 60 yr (Armstrong and Oriel, 1965, 1982). We demonstrate that the Beaverhead belt of plutons was the primary source for coeval Worm Creek Quartzite. In this paper we refer to the Worm Creek Quartzite Member of the St. Charles Formation and underlying sandstone in the top of the Nounan Dolomite as the “Worm Creek sandstone.” This link suggests that instead of a tectonically quiet subsiding margin during the latest Cambrian, intrusion and exhumation occurred within the Lemhi arch of central Idaho, and sand deposition occurred in what had been a carbonate platform in southeast Idaho. Furthermore, the coincidence of these events with the Sauk II–Sauk III boundary raises questions about linkage and causality, including the possibility that uplift of the Lemhi arch contributed to the Sauk II–Sauk III lowstand. In this paper we summarize Worm Creek sandstone cyclicity and present detrital zircon analyses of uppermost Cambrian sandstones from central Idaho east to central Wyoming. We compare U-Pb ages and initial εHf isotopic compositions of these zircons to zircons from the Beaverhead and Deep Creek plutons from the Beaverhead belt of east-central Idaho (Todt and Link, 2013; Todt, 2014).

Because our samples span from the Laurentian craton on the east to the central Idaho thrust belt, these data provide a test of the Rubia ribbon continent model of Johnston (2008) and Hildebrand (2009). This model holds that the western thrust belt contains a narrow continental block exotic to Laurentia.

**GEOLOGIC SETTING**

**South of the Snake River Plain**

The breakup of Rodinia and formation of the Laurentian passive margin in Utah and Idaho south of the Snake River Plain was protracted and occurred as multiple stages (Link et al., 1993; Yonkee et al., 2014). The final stage of rifting began near 570 Ma, with true development of the Cordilleran passive margin after 550 Ma, and thick carbonate deposition concomitant with eustatic sea-level rise through the Cambrian (Bond and Kominz, 1984; Thomas, 1993; Saltzman et al., 2004).

The sandstones studied in this paper belong to the siliciclastic inner detrital belt (as opposed to the shale-rich outer detrital belt; Palmer, 1971; Myrow et al., 2012), and were deposited during a Laurentian sea-level low across the Sauk II–Sauk III boundary (Sloss, 1988). This boundary, locally an unconformity or sequence boundary, is coincident with the Steptoean positive isotopic carbon excursion (SPICE), a strong positive δ13C swing in Fusacanian time (ca. 497–485 Ma; Fig. 2), correlated with a high in atmospheric oxygen and a punctuated increase in invertebrate complexity (Saltzman et al., 1998, 2004, 2011).

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1 Although “Lower, Early” and “Upper, Late Cambrian” are formal terms of the U.S. Geological Survey, the 1999 and subsequent versions of the GSA Geologic Time Scale and recommendations of the International Subcommission on Cambrian Stratigraphy (http://www.paleoontology.geo.uov.us/SGS/SCS/SCS_home.html) instead divide the Cambrian into 4 epochs, Terreneuvian, Epoch 2, Epoch 3, and Fusacanian. However, in keeping with common usage, we still use uppercapse when referring to Early, Middle and Late Cambrian time and Lower, Middle and Upper Cambrian rocks. We follow the 1999 time scale and use Laurentian stage terms (Steptoean and Sanawaptian) for Cambrian division I, which includes the Worm Creek Quartzite Member of the St. Charles Formation (see Fig. 2).

**Uppermost Cambrian Sandstone**

In southeast Idaho, south of the Snake River Plain, >1 km of Cambrian carbonate-rich strata of the Sauk sequence records overall sea-level rise (Fig. 2; Armstrong and Oriel, 1965; Trimble and Carr, 1976). The uppermost Nounan Dolomite and overlying feldspathic Worm Creek Quartzite Member of the St. Charles Formation were deposited during pulsed regressions across the Sauk II–Sauk III boundary ca. 495 Ma (Fusacanian Epoch, Steptoean stage [1999 GSA Geologic Time Scale]) (Fig. 3) (Sloss, 1988; Saltzman et al., 2004; Haq and Schutter, 2008). These are manifested as four siliciclastic to carbonate cycles (Fig. 3). The Sauk II–III contact was represented by Haq and Schutter (2008) as eustatic relative sea-level falls of 25–75 m at 499 Ma and >75 m at 495 Ma.

In southeast Idaho, the rocks of concern include the uppermost Nounan Dolomite (Crepicephalus, Aphelaspis, and Dunderbergia trilobite zones, 498.5–494 Ma), and the Worm Creek Quartzite Member of the St. Charles Formation (Dunderbergia and Elvinia trilobite zones; 495.2–493 Ma; time scale of Gradstein et al., 2012, here and henceforth) (Fig. 2; Howell et al., 1944; Haynie, 1957; Armstrong and Oriel, 1965; Lochman-Balk, 1974; Saltzman et al., 2004; Hintze and Kovallis, 2009, fig. 25 therein). The uppermost Nounan and overlying Worm Creek Quartzite Member form four upward-fining cycles, and are henceforth referred to as the Worm Creek sandstone. On the north at Midnight Creek (MC in Fig. 1), just south of the Snake River Plain, the Worm Creek totals 320 m in thickness (Fig. 3). Only 2 cycles totaling 26 m are evident toward the southeast at Weston Canyon (WC in Fig. 1) along the Idaho-Utah border. Paleocurrents derived from planar foresets in cross-bedded sandstone in northern Utah are bimodal, trending northwest (280°) and southeast (100°) in some localities and unimodal to the east-northeast (70°) in others (R.Q. Oaks, Utah State University, 2013, written commun.).

The Nounan Dolomite thickens northward from a feather edge at the Idaho-Utah border to >150 m thick just south of the Snake River Plain (Trimble and Carr, 1976). Its upper member contains the stratigraphically lowest sandstone of cycle 1 at Weston Canyon (Gardiner, 1974).

The Worm Creek Quartzite Member of the St. Charles Formation contains three siliciclastic to carbonate cycles of fine-grained subarkosic arenite (Fig. 3). Worm Creek sandstone contains detrital potassium feldspar (photomicrographs shown in Figs. 4A–4D). X-ray diffraction study of the feldspar (Todt, 2014) shows it is microcline. The sandstones locally contain plutonic myrmekite (Fig. 4C). Volcanic glass, or its altered manifestation, is lacking. Palmer (1971) hypothesized that the feldspar was derived from the Lemhi arch. The thickest section of the Worm Creek (320 m) is near Midnight Creek in the northern Bannock Range (MC in Fig. 1; A in Fig. 3), immediately south of the Snake River Plain. There, three cycles of potassium feldspar (5%–40%) arenite to limy and dolomitic mudstone, wackestone, and packstone are present (Wakeley, 1975; Trimble and Carr, 1976). The thickest package (cycle 3 of Fig. 3) is >150 m thick. It is made up of light pink to tan, poorly sorted, very fine to coarse, trough cross-bedded limy sandstone with 10% potassium feldspar (microcline) and 90% quartz (Todt, 2014). Shallow-marine trace fossils paired with the herringbone (Fig. 5A) and trough cross-stratification are indicative of terrigenous subtidal sand shoals that pass upward to tidal flats and sublittoral lagoons (Walcott, 1908; Gardiner, 1974; Wakeley, 1975). Saltzman et al. (2004) interpreted lower shoreface environments for cycles 2 and 3 at Smithfield, Utah (Fig. 1).

Upper Cambrian strata in the northern Lost River Range, east of Challis, Idaho (Fig. 1), broadly correlate with the Worm Creek sandstone, include >25 m of the upper part of the Wilbert Formation (Fig. 2; Hargraves et al., 2007). These strata consist of trace fossil-bearing (Skolithos) silstone and quartz arenite. They had previously been included in the informal formation of Leaton Gulch (McIntyre and Hobbs, 1987; Carr and Link, 1999).
This bed is 0.6 m below the informal Dry Creek shale (Thomas, 1993). It include the Du Noir Member of the Gallatin Limestone (Fig. 4). The Du Noir Member contains a 0.4-m-thick sandstone unit sampled for this paper. The sandstone just below the contact with the Upper Cambrian Red Lion Formation. It is a white, trough cross-bedded, mature, medium-grained quartz sandstone, containing scattered abraded Crepicephalus Zone trilobites, and is 0.3 m below the Marjumiid extinction boundary or the Marjumiid-Pterocephaliid extinction boundary. This is just below the contact with the Upper Cambrian Red Lion Formation.

North of the Snake River Plain: The Lemhi Arch

North of the Snake River Plain in east-central Idaho is a northwest-trending region of thinned or missing Neoproterozoic to lower Paleozoic strata, the Lemhi arch (Fig. 1; Sloss, 1954; Scholten, 1957; Armstrong, 1975; Ruppel, 1986). Across much of the Lemhi arch, Middle Ordovician Kinnikinic Quartzite (James and Oaks, 1977) unconformably overlies tilted Mesoproterozoic Belt Supergroup (Fig. 2; Ross, 1947; Sloss, 1950, 1954; Lochman-Balk, 1971; Ruppel, 1986; Bush et al., 2012). On the southeastern side of the Lemhi arch in the southern Lemhi and Beaverhead Ranges, latest Neoproterozoic and Cambrian Wilbert Formation overlaps the unconformity (Fig. 2; Skipper and Link, 1992). In central Idaho along the Salmon River southwest of Challis, Cambrian and Ordovician carbonate and siliciclastic rocks of the Bayhorse assemblage (Hobbs and Hays, 1990; Hobbs et al., 1991) were deposited on the downfaulted western side of the Lemhi arch (Pearson et al., 2013; Brigham Group correlative strata near Edwardsburg and Stibnite (Fig. 1) are west of the arch (Lund et al., 2003; Lewis et al., 2014).
Figure 3. Location map and measured stratigraphic columns for two of seven study locations; 500 point thin section point counts for each of the four siliciclastic layers (ternary diagrams: Qt—quartz, F—almost 100% K-feldspar, L—lithics); and isopach map of the third siliciclastic layer—the main siliciclastic wedge, contour interval 25 m, throughout the Nounan–Worm Creek depositional basin, southeast Idaho. Locations were chosen from regional map of Oriel and Platt (1980). Locations of samples are shown; letters are as follows: A—Midnight Creek (1MKT12, 2MKT12, 3MKT13); B—McPherson Canyon Road (7MKT12, 8MKT12, 9MKT12); C—Green Canyon Measured Section; D—Secret Canyon (7MKT13); E—Weston Canyon (1MKT13). The thickest column, located at Midnight Creek, shows all four cycles of siliciclastic sedimentation. Notice the change in scale in the stratigraphic column of the third cycle at the Midnight Creek location.
Figure 4. Thin section photomicrographs for select siliciclastic layers of the Worm Creek Quartzite, the Beaverhead pluton, and Deep Creek pluton, and cathodoluminescence (CL) images of the study detrital and igneous zircon. Mineral abbreviations: Q—quartz, F—feldspar, B—biotite. (A) Worm Creek Quartzite (01MKT12)—cycle 2 siliciclastic layer with >40% stained feldspar content (uncrossed polars). (B) Worm Creek Quartzite (02MKT12)—cycle 3 siliciclastic layer with <15% stained K-feldspar content (uncrossed polars). (C) Worm Creek Quartzite (05MKT13)—cycle 3 siliciclastic layer; large feldspar grain has perthitic texture (crossed polars). (D) Worm Creek Quartzite (07MKT12)—cycle 2 siliciclastic layer with subhedral and blobby feldspar grain shape (crossed polars). (E) Deep Creek pluton (01LKB12) (crossed polars). (F) Beaverhead pluton (05MKT12) (crossed polars). (G) Example CL image from 09MKT12 showing zircon morphologies for cycle four—Worm Creek sandstone. (H) CL image for 08MKT12, cycle three—Worm Creek sandstone. Ages of zircon grains shown in Ma. (I) CL image for Beaverhead pluton (05MKT12). (J) CL image for Deep Creek pluton (02LKB12).
Within the Lemhi arch, the Big Creek–Beaverhead magmatic belt contains the Big Creek (665–650 Ma) and Beaverhead (500–485 Ma) hypabyssal alkalic granitoid plutons. These intruded into Mesoproterozoic upper Belt Supergroup (Fig. 1 and 2; Evans and Zartman, 1988; Lund et al., 2010; Lewis et al., 2012). Within the Beaverhead suite of plutons, the Beaverhead and Deep Creek granitoids are specific geologic units that we sampled (Evans and Zartman, 1988; Lund et al., 2010).

The mildly bimodal, alkalic nature of the Big Creek–Beaverhead belt plutons suggests that they originated as lithospheric melts intruded during progressive crustal extension (Lund et al., 2010). The two discrete magmatic pulses, at 665–650 Ma and 500–485 Ma, may have reflected recurrent extension along an inherited structural weakness (Lund, 2008). They have been correlated with initial and final rifting of western Laurentia (Yonkee et al., 2014).

The contact between the Beaverhead pluton and the Kinnikinic Quartzite is planar, but irregular, with no contact metamorphism (photos in Figs. 5B, 5C). Scholten and Ramspott (1968) interpreted this contact as intrusive. This interpretation was supported by a K-Ar age on biotite of 441 ± 15 Ma. The age is 50 m.y. younger than the U-Pb zircon age (Evans and Zartman, 1988; Lund et al., 2010). The fact that the Kinnikinic Quartzite is Middle Ordovician (470–458 Ma; James and Oaks, 1977), and thus younger than the Late Cambrian and Early Ordovician Beaverhead pluton, indicates that an intrusive contact is not possible. We interpret the contact (Fig. 5B) as an unconformity. This requires post-intrusive uplift, even if the plutons were initially hypabyssal (within a few hundred meters of the surface). This interpretation is consistent with the map relations shown by Skipp (1984).

In contrast to the dominantly Archean lithospheric framework in southwestern Montana (e.g., Foster et al., 2012; Gaschnig et al., 2013), magmas that were intruded in the region of the Lemhi arch completely lack an Archean isotopic influence. Along the Salmon River (orange polygons north of DCp in Fig. 1) is a region of 1370 Ma magmatism, interpreted to represent melting of lower Belt Supergroup strata by juvenile magma (Doughty and Chamberlain, 1996). Doughty and Chamberlain (1996) interpreted that the region is underlain by Mesoproterozoic lithosphere generated after Belt basin extension.

In the Pioneer Mountains (Fig. 1) 2.7–2.6 Ga orthogneiss is intruded by rift-related 695 Ma granitic plutons (Durk et al., 2007; Gaschnig et al., 2013; Link et al., 2017). Yates (1968) hypothesized a 350 km trans-Idaho discontinuity, extending northwest across central Idaho and east of the Pioneer Mountains, between outcrops of diamicotics and rift-related volcanic rocks at Pocatello and Edwardsburg (Fig. 1). The U-Pb ages of the Wildhorse gneiss in the Pioneer Mountains obviate the concept, because Archean basement is present on both sides of the discontinuity.

Lund (2008) proposed that the Lemhi arch was part of the upper plate of an east-dipping low-angle rift fault (i.e., Lister et al., 1986), and was bounded on the south by the high-angle Snake River transfer fault (i.e., Thomas, 1977, 2014) (SRTF in Fig. 1 inset). In a lower plate setting south of the Snake River Plain, thousands of meters of rift-related and post-rift strata were deposited (Yonkee et al., 2014). A down-to-the-west pre-Ordovician normal fault was mapped along the west side of the Lemhi arch (Hansen, 2015; Hansen and Pearson, 2016). West of this fault, Neo-protoreozoic and Cambrian strata, not present on the Lemhi arch to the east, are recognized at Edwardsburg, Stibnite, and in the Sawtooth Mountains (Fig. 1) (Hobbs et al., 1991; Lund et al., 2003; Lewis et al., 2012, 2014; Ma et al., 2016).

**Upper Cambrian Sandstones on the Wyoming Craton**

To the east, on the craton (Fig. 2), Cambrian rocks of Wyoming compose a sandstone-shale-carbonate transgressive sequence punctuated by

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**Figure 5.** (A) Outcrop photograph of planar cross-bedded fine sandstone of the second siliciclastic layer—Worm Creek Quartzite, Midnight Creek location (MC in Fig. 1). Diameter of lens cap is 3 cm. (B) Outcrop photograph of the unconformable contact (dashed line) between the Late Cambrian Beaverhead pluton and the Ordovician Kinnikinic Quartzite; 18 Mile Creek, west side Beaverhead Mountains; east-central Idaho. Width of field of view is 2 m. (C) Bedded Kinnikinic Quartzite along ridge south of 18 Mile Creek, near Beaverhead pluton sample location. Photographs by M.K. Todt. Width of field of view is 100 m.
the Sauk II–Sauk III regression (Lochman-Balk, 1971; Middleton et al., 1980; Thomas, 1994). Middle Cambrian sandstone of the Flathead Sandstone overlies Archean Wyoming province basement (Middleton et al., 1980). Detrital zircons in the middle Flathead from the Bighorn Basin in northwest Wyoming and from southwest Montana contain a unimodal 1790 Ma peak (May et al., 2013; Mahoney et al., 2015), but this age peak is muted or missing in four samples of the Flathead from the southern Beartooth Mountains in northwest Wyoming (Malone et al., 2017). The bulk of the overlying Cambrian section is carbonate, but thin sandstone (that we sampled) is present at the top of the Upper Cambrian Du Noir Member of the Gallatin Formation, representing the Sauk II–Sauk III regression.

METHODS

Zircon U-Pb and εHf Isotopic Analysis

Seven stratigraphic sections of the Worm Creek sandstone within the Paris thrust plate were measured (Todt, 2014; Fig. 3). The rocks compose four siliciclastic to carbonate cycles. Thin sections were cut and 500 points per section were counted. Triangular sandstone diagrams are shown in Figure 3. U-Pb analysis was applied to randomly selected zircons from 11 detrital samples (Fig. 6 and 7). We analyzed two zircon separates from each of the four stratigraphic cycles within the Worm Creek sandstone, plus samples from Teton Pass and Wind River Canyon, Wyoming.
east of Melrose in southwest Montana, and Leaton Gulch near Challis, Idaho. Table 1 shows abbreviations for sample localities from Figures 1 and 3. Subsidence analysis (Fig. 8) was performed using BasinMod software (Platte River Associates, Inc.; www.platte.com/software/basinmod-2012.html).

We also analyzed magmatic zircons from four plutonic samples of the Deep Creek and Beaverhead granitoids, from the Beaverhead plutonic belt (Fig. 9). Lund et al. (2010) reported SHRIMP (sensitive high-resolution ion microprobe) U-Pb ages on these plutons. The reanalysis affords direct comparison of the ages and Lu-Hf isotope values of magmatic and detrital zircons using the same instrument and technique.

U-Pb zircon analyses from plutonic and detrital grains were acquired using laser ablation–multicollector–inductively coupled plasma–mass spectrometry (LA-MC-ICP-MS) at the Arizona LaserChron Center following procedures given by Gehrels et al. (2006). Plots and age calculations were carried out using Isoplot (Ludwig, 2003). For igneous samples, cathodoluminescence images were used to check for inherited cores within the zircon crystals.

Ages with >10% discordance or >5% precision error were rejected and are not included in the results presented here. Weighted mean averages of grain ages from each pluton were calculated using Isoplot (Ludwig, 2003; U-Pb age uncertainties henceforth include internal and external errors at the 2σ level). Probability density plots with superposed 20 Ma bin width histograms were generated for the detrital samples using Isoplot to determine age peaks of each detrital sample (Figs. 6 and 7; Ludwig, 2003). We used a laser spot size of 30 µm. For grains younger than 1.0 Ga, we used 206Pb/238U ages, whereas we used 206Pb/207Pb ages for grains older than 1.0 Ga. The Koolmogorov-Smirnoff (K-S) test was used to determine if there are statistical differences between two samples’ age distributions (DeGraaff-Surpless et al., 2003; Guynn and Gehrels, 2010).

LA-MC-ICP-MS was also used on selected 510–490 Ma zircon grains to measure the ratios of isotopes of Hf, using methods detailed in Cecil et al. (2011) and Gehrels and Pecha (2014). We conducted Hf analyses over the same pit as the U-Pb age spot. For large grains, we used a 50 µm spot for Hf analysis, whereas for smaller grains, we used a 40 µm spot size. Hf results are shown in Figure 10, where initial 176Hf/177Hf ratios are expressed in εHf notation, which represents the Hf isotopic composition relative to the chondritic uniform reservoir at the time of zircon crystallization (Bouvier et al., 2008). Initial εHf(t) values that are <5 units below the depleted mantle are considered juvenile in composition, while values >12 units below the depleted mantle are evolved in composition; initial εHf values between 5 and 12 units below the depleted mantle values are classified as intermediate (Bahlburg et al., 2011). For this study, detrital zircon grains were chosen for Hf analysis based on their measured U-Pb ages between 510 and 490 Ma.

Plutonic grains were chosen based on their low U-Pb uncertainty and lack of inherited cores seen in cathodoluminescence images (see Fig. 4). Initial εHf results for each zircon grain analyzed are shown in Figure...
Figure 8. Tectonic subsidence and total subsidence curves for Neoproterozoic (Cryogenian) to Ordovician strata in the northern Bannock Range, southeast Idaho. Steep parts of curve show initial and final rifting. Inflection of both subsidence curves at 500 Ma corresponds to subsidence of the Worm Creek basin. Curve flattens into the Ordovician with the deposition of the Garden City Formation and Swan Peak Quartzite. Thickness and depositional environment data start from the Caddy Canyon Quartzite (Fig. 2) and are from Trimble and Carr (1976) and Link et al. (1987). Subsidence was calculated using BasinMod software developed by Platte River Associates, Inc. (www.platte.com/software/basinmod-2012.html). The software takes input for geologic age (top of the unit), thickness, lithology and general depositional environment. This allows calculation of water depth and compaction. Compaction is calculated using the Statoil fluid flow porosity reduction method: \( \Phi = \Phi_i \times \exp(-C \times S_{eff}) \); \( \Phi_i \) = initial porosity; \( C \) = Statoil compaction exponent; and \( S_{eff} \) = effective stress.
Late Cambrian zircons, Worm Creek sandstone | RESEARCH

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Figure 9. (A–D) Concordia plots showing laser ablation–multicollector–inductively coupled plasma–mass spectrometry U-Pb isotopic data for pluton samples. Values of individual grains are shown with 1σ error ellipses; no discordant grains were analyzed. Included for each sample are weighted mean ages with 2σ uncertainty, number of grains plotted, and mean square of weighted deviates (MSWD).

Figure 10. Zircon εHf values versus age (Ma) for the two sampled plutons (white diamonds) and the 510–490 Ma analyzed detrital grains (black diamonds). The 176Hf is radiogenic, from decay of 176Lu; 177Hf is the stable isotope. The εHf is 176Hf/177Hf. DM is depleted mantle; CHUR is chondritic uniform reservoir, and represents the εHf of the average Earth; 2σ is 95% confidence level of the analysis.
10. They were plotted with their corresponding U-Pb ages and 1σ εHf uncertainties. The average internal precision for εHf(t) data presented here is 2 epsilon units (2σ).

RESULTS

Late Cambrian Subsidence and Upper Cambrian Stratigraphy

Ediacaran to Ordovician tectonic and total subsidence curves (Fig. 8) for the northern Bannock Range were calculated using BasinMod software. The subsidence curves show a decrease and then increase in subsidence rate, immediately before and during deposition of the Worm Creek sandstone. We suggest that this change in subsidence rate is linked with emplacement, cooling, and exhumation of Beaverhead plutons and punctuated subsidence of the Worm Creek sedimentary basin. The subsidence curves also show earlier periods of rapid subsidence corresponding with initial and final rifting of the passive margin (Yonkee et al., 2014).

Zircon U-Pb Ages and Hf Isotopes

Worm Creek Sandstone: Detrital Zircon Analyses

The extraordinary aspect of the U-Pb ages of detrital zircon grains in all of the sandstones is the large near-unimodal population of ages of 498–497 Ma (Fig. 6) that overlaps with the Crepicephalus trilobite zone (498.5–497 Ma). The presence of this detrital zircon population was suspected from analysis of zircons in modern streams draining areas where the Worm Creek sandstone crops out (Link et al., 2005). All 4 siliciclastic cycles have average ages of 498–497 Ma. These 498–497 Ma zircon grains are typically clear and subrounded, with simple oscillatory zoning (Figs. 4G, 4H). Subordinate Paleoproterozoic (mainly 1750–1700 Ma) and Archean zircon grains are darker colored, light pink to violet, rounded, and complexly zoned. These detrital grains are interpreted as recycled from uppermost Belt Supergroup (Fig. 7A; Burmester et al., 2016), the country rock for the Beaverhead granitoids (Skipp, 1984; Link et al., 2007, 2016).

The dominant age peak (54% of zircon grains) for the first cycle of siliciclastic sedimentation is 497 Ma, with a smaller age group (15%) of 1716 Ma grains (Fig. 6). The defining unimodal age peak (85%) for the second cycle is 498 Ma. Proterozoic grains are sparse. The third cycle has an age peak of similar magnitude to the first cycle peak at 498 Ma (44%); subordinate age groups have average peaks at 1719 and 1732 Ma (32%) and 3.0–2.5 Ga (8%). The fourth siliciclastic cycle has a signature similar to that of the third cycle. There is a slightly stronger Paleoproterozoic signature at 1740–1710 Ma and a dominant age peak at 498 Ma (41%).

The K-S two-sample test was used to assess the similarity of the 510–480 Ma siliciclasts from the 4 siliciclastic cycles (data are in Table DR2 in the GSA Data Repository Item2; see Todt, 2014). P values are consistently high (0.753–1.000), allowing that zircons in all of the four Upper Cambrian siliciclastic cycles could be derived from the same source population (e.g., Berry et al., 2001; DeGraaff-Surpless et al., 2003).

We utilized Hf isotopic analysis of detrital zircons with U-Pb ages of 510–490 Ma to compare Worm Creek sandstones with coeval plutons of the Beaverhead plutonic suite. The detrital zircons have a spread of intermediate to evolved initial εHf values from −8.0 ± 1.9 to 5.4 ± 1.2 (Fig. 9).

Other Cambrian Sandstones

Zircons from the Middle Cambrian Flathead Sandstone at Teton Pass, Wyoming, have a major age peak at 1787 Ma (Fig. 7D). In contrast, white sandstone beds within the Upper Cambrian Pilgrim Limestone, Melrose, Montana (Fig. 7H), and sand in the Upper Cambrian Du Noir Member of the Gallatin Limestone, Wind River Canyon, Wyoming (Fig. 7G), share a major 491 Ma age peak. Sandstone from the uppermost part of the Upper Cambrian Wilbert Formation at Leaton Gulch, south of Challis, Idaho (Carr and Link, 1999; Hargraves et al., 2007), has a 488 Ma zircon age peak (Fig. 7F).

Lu-Hf Isotopes from Plutonic Zircons

We analyzed U-Pb and Lu-Hf in zircon from four plutonic samples from the Deep Creek (two samples) and Beaverhead plutons (two samples) of the Beaverhead plutonic suite north of the Snake River Plain (Figs. 9 and 10). The Deep Creek pluton yielded U-Pb LA-ICP-MS ages of 492 ± 4 Ma and 491 ± 2 Ma, and the Beaverhead pluton produced ages of 496 ± 2 Ma and 500 ± 3 Ma (Fig. 9). These ages are close to, but not precisely the same as, the SHRIMP ages reported by Lund et al. (2010) of 497 ± 6 Ma (Deep Creek) and 488 ± 5 Ma (Beaverhead). Ages from the Deep Creek pluton overlap within error, while the SHRIMP age for the Beaverhead pluton is slightly younger than our ages. The 500–490 Ma zircons from the plutons have an intermediate initial εHf range of −5.4 ± 1.1 to −2.2 ± 0.9 for the Deep Creek pluton and −6.3 ± 1.1 to 2.7 ± 1.4 for the Beaverhead pluton (Fig. 10).

DISCUSSION

Provenance Relationships

Given the excellent match in U-Pb age and initial εHf values from both magmatic and detrital zircons, we conclude that the Beaverhead plutonic suite was the primary source for the flood of 500–490 Ma zircon grains and detrital potassium feldspar in the Upper Cambrian Worm Creek sandstone of the southeast Idaho thrust belt (Figs. 6 and 7E). By transport to more distal locations within the continental shelf, young 500–490 Ma zircon grains were also deposited in the Upper Cambrian Du Noir Limestone on the Wyoming craton and the Pilgrim Formation on the southwest Montana cratonal shelf (Figs. 7G, 7H). To the west, in the Cordilleran thrust belt, this young grain population is also present in the upper Wilbert Formation at Leaton Gulch in the northern Lost River Range (Fig. 7F). Within the Antler assemblage at Pete’s Summit in the Toquima Range of central Nevada (Fig. 11), the Ordovician lower Vinini Formation also contains this zircon population; 29 of 189 grains are between 503 and 480 Ma (Linde et al., 2016).

Transcontinental Arch

Detrital zircon evidence from Neoproterozoic and Cambrian sandstones in Idaho as well as in the Roberts Mountains allochthon of the Devonian Antler orogenic belt in central Nevada suggests that in Early Cambrian time, the Transcontinental Arch rose in the central United States (see Fig. 11). This cut off the supply of Mesoproterozoic, Grenville-age (1250–950 Ma) zircons to the western continental margin (Linde et al., 2014; Yonkee et al., 2014).

Rapid Pluton Exhumation or Coeval Vulcanism

The Worm Creek sandstones contain 500–490 Ma detrital zircons with ages nearly the same as biostratigraphic ages of the host strata, and

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2GSA Data Repository Item 2017339, four tables: DR1—Locations and description of samples, DR2—U-Pb zircon data, DR3—Lu-Hf isotope data on dated zircon grains, DR4—Results of Kolmogorov-Smirnov tests on detrital zircon samples, is available at http://www.geosociety.org/datarepository/2017, or on request from editing@geosociety.org.
Late Cambrian zircons, Worm Creek sandstone

A

B

Figure 11. (A) Map relations and location of cross section in B. Cyn—canyon. (B) Generalized cross section showing schematic interpretation for latest Cambrian time, with the Beaverhead pluton to the north of the Snake River Plain supplying arkosic sediment to the Worm Creek basin on the south side of the Snake River transfer fault. Key locations mentioned in text are also shown. Location of Worm Creek basin southern onlap is based on Coulter (1956). Hypothetical braided stream system that supplied plutonic debris to marine shoreline is shown. Yb—Mesoproterozoic Belt Supergroup; Czw—Neoproterozoic—Cambrian Wilbert Formation; C—Cambrian; Z—Neoproterozoic; A—away; T—toward. Trade winds (from Amato and Mack, 2012) are consistent with generally east-west shoreline inferred from paleocurrents in northern Utah. V.E.—vertical exaggeration.
Regional Implications

We are now able to reconstruct a predictable succession of Neoproterozoic to Paleozoic detrital zircon age populations in southern and central Idaho, shown in Figure 7. Sources of zircon grains of a given age were reviewed in Link et al. (2005), Balgord et al. (2013), and Yonkee et al. (2014). South of the Snake River Plain, the Ediacaran (635–540 Ma) lower Brigham Group (Fig. 7B) has a diverse provenance that includes 1250–950 Ma grains from the Grenville orogen, Mesoproterozoic grains (1.47–1.40 Ga), and Paleoproterozoic grains (1.8–1.6 Ga), likely derived or recycled from the Yavapai-Mazatzal provinces. Archean grains are also present. This Laurentian integrated provenance is common in Paleozoic passive margin sandstones of western North America (Gehrels and Pecha, 2014) and is labeled type II in Cambrian sandstones of the northern Canadian Cordillera (Hadrallari et al., 2012).

At the top of the Brigham Group (Fig. 7C, Lower Cambrian), and in the Middle Cambrian Flathead Sandstone on the Wyoming craton (Fig. 7D), the Grenville age peak disappears and the bulk of the zircons are 1.8–1.75 Ga (Figs. 7B, 7C). Strata mapped as Flathead Sandstone in southwest Montana are notable because they regionally overlie Archean basement but contain mainly 1790 Ma Paleoproterozoic grains (Mahoney et al., 2015). The upper Brigham Group lacks Grenville-age grains. We infer that these grains were cut off by rise of the Transcontinental Arch in Early Cambrian time (Yonkee et al., 2014; Linde et al., 2014).

The Worm Creek sandstones have very a different and much more specific detrital zircon distribution. The near unimodal, 500–490 Ma ages of detrital zircons in the Worm Creek sandstones (Fig. 7D) are interpreted to represent a local flood of grains from the Beaverhead plutons. This signals proximal siliciclastic sediment delivery to the passive margin. Smaller amounts of zircon of this age are found in more distal strata, from the Wilbert Formation at Leaton Gulch (Fig. 7F) in central Idaho, eastward to the Pilgrim Formation near Melrose, Montana (Fig. 7H), and sandstone in the Du Noir Member, Gallatin Limestone, in the Wind River Canyon on the Wyoming craton (Fig. 7G). Small groups of Paleoproterozoic (mainly 1750–1700 Ma) grains in the Worm Creek sandstones are interpreted to be recycled from the uppermost Belt Supergroup (Swauger Formation and overlying Jahnke Lake member of the Apple Creek Formation; Fig. 7A) (Link et al., 2007, 2016), which are country rock to the plutons. In the second siliciclastic cycle of the Worm Creek sandstone (Fig. 6B), Paleoproterozoic grains are sparse, their signature nearly flooded out by 500–490 Ma grains derived from the Beaverhead plutons.

Stratigraphically above the Worm Creek sandstone, the Middle Ordovician Kinnikinic Quartzite (and the correlative Swan Peak Quartzite south of the Snake River Plain; Wulf, 2011) have a mainly Paleoproterozoic provenance, with almost all grains older than 1.8 Ga, a peak at 1860 Ma, and no younger grains (Fig. 7I; Baar, 2009; Beranek et al., 2016). The provenance is interpreted to have been from the Peace River Arch in Alberta. This signature is prominent in Middle Ordovician sandstones of the Canadian Cordillera and Nevada (Gehrels and Pecha, 2014) as well as in central Idaho (Beranek et al., 2016). In the Sawtooth Mountains (Fig. 11) it is group B of Ma et al. (2016). In Cambrian strata of the northern Canadian Cordillera, it is provenance type I of Hadrallari et al. (2012).

HF Isotope Discussion: Mesoproterozoic Parent for Cambrian Plutons

Initial eHf isotope compositions of zircon help differentiate between a lithospheric (evolved) versus asthenospheric (juvenile) source, or the extent of crustal contamination during magmatism (Kinny and Maas, 2003; Goode and Vervoort, 2006; Cecil et al., 2011; Gehrels and Pecha, 2014). The 500–490 Ma Deep Creek and Beaverhead plutons have intermediate to evolved initial eHf of −6.3 ± 1.1 to 2.7 ± 1.4. This range overlaps that of Neoproterozoic plutons in the Pioneer Mountains, with initial eHf of −2.4–3.4 in 675 Ma detrital grains (Link et al., 2017).

These intermediate initial eHf values suggest limited reintegration of evolved Archean lithosphere. They are consistent with the Mesoproterozoic lithospheric source hypothesized to occur beneath the southwestern portion of the Belt basin (Doughty and Chamberlain, 1996; Elk City domain of Gaschnig et al., 2013; Fig. 1B). The Cambrian plutons occur in crust previously modified by a northwest-trending belt of Mesoproterozoic plutons and concomitant mafic sills (ca. 1370 Ma) intruded into the lower crust during the final stages of Belt basin subsidence (Fig. 1B; Doughty and Chamberlain, 1996). We suggest that a fundamental cause for the Lemhi arch, and its lack of overturning Neoproterozoic strata, as well as the Big Creek–Beaverhead belt of Neoproterozoic and Cambrian plutons, is this post-Belt Mesoproterozoic underpinning (Fig. 1 inset).

Tectonic Model

The flood of feldspathic sand in the Worm Creek sandstones was coupled with an increase in tectonic subsidence and accommodation (Fig. 8) within the thermally subsiding Ediacaran to Orдовician continental terrace in southeast Idaho. We interpret the Snake River dextral transfer fault (Fig. 1) of Lund et al. (2010) to have had dextral normal movement (Fig. 11), and to have accommodated the northward transition from the passive margin in southeast Idaho to the Lemhi arch. The amount of dextral strike slip on this fault must have been less than 100 km, since strata correlate with the passive margin at Pocatello are found to the northwest at Stibnite and Edwardsburg (Figs. 1, 11) (Lewis et al., 2014; Stewart et al., 2016). Sediment was transported southward from uplifted Beaverhead and Deep Creek plutons, across the Snake River transfer fault from the Lemhi arch and its upper Belt Supergroup country rock. The latest Cambrian Worm Creek basin formed on the hanging wall of the fault separating upper and lower plate margins (Fig. 11).

By east-west transport across the Laurentian continental shelf, 500 Ma Beaverhead magmatic zircons reached the Du Noir Limestone in Wind River Canyon in Wyoming (Fig. 7G), the Pilgrim Formation on Montana shelf to the north (Fig. 7H), and the thrust belt in central Idaho (Fig. 7F).

The presence of 500–490 Ma magmatic zircon grains in the Antler allochthon in central Nevada (Linde et al., 2016) demonstrates linkages from the Laurentian craton to the western thrust belt. This raises doubt about models of the exotic Rubia ribbon continent, the eastern margin of which is predicted to be within the Idaho thrust belt (Johnston, 2008; Hildebrand, 2009).
CONCLUSIONS

Similar U-Pb zircon ages, coupled with the overlap in εHf values between the Upper Cambrian Worm Creek sandstone and the Cambrian–Ordovician plutons of the alkaline Beaverhead plutonic suite, suggest that the plutons were the source for the sandstones. This requires rapid exhumation and erosion of the plutons into the actively subsiding Worm Creek basin. These 500–400 Ma zircons were transported across the continental shelf as far east as the Wind River Canyon on the Wyoming craton, and as far west as Leaton Gulch in the central Idaho thrust belt.

These results support the model of long-lived Rodinian rifting along the western margin of Laurentia; they also suggest the existence of the dextral-normal Snake River transfer fault, a major transverse structure separated a region of ongoing magmatism and exhumation from one of subsidence and passive margin sedimentation. The spatial overlap between the ca. 1.4 Ga Mesoproterozoic Belt basin and Lemhi subbasin, ca. 1.37 Ga mafic intrusions, and the partially exhumed Lemhi arch suggest that the shape of the Laurentian rifted margin was controlled by the preexisting lithospheric architecture, perhaps due to strengthening of the lower crust during Mesoproterozoic mafic magmatism.

Rather than being tectonically quiet in central Idaho, the Cambrian was a time of magmatism and uplift. The Beaverhead and Deep Creek alkaline intrusions were emplaced, uplifted, and eroded. The Worm Creek basin south of the Snake River Plain subsided and was filled close to sea level with feldspathic sand derived from those plutons.

The coincidence of the uplift of the Lemhi arch and erosion of the Beaverhead plutons with the Steptoean positive isotope carbon excursion begs consideration of causality. The fundamental issue is whether the volume of uplifted Lemhi arch (estimated as 2 x 10$^6$ km$^3$ from area shown in Fig. 1B, and 0.5 km uplift) could have caused one of the Steptoean eustatic drops documented by Haq and Schutter (2008). In Link and Janecke (2009), it was suggested that mantle drip may have been a cause of Lemhi arch exhumation. At this point we only note that latest Cambrian uplift in central Idaho was coeval with both Laurentian and worldwide sea-level drop.

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