ABSTRACT

The gneiss complex of Wildhorse Creek (Wildhorse gneiss) forms the central component of the lowest structural plate in the Pioneer metamorphic core complex of south-central Idaho. The oldest rock in the complex is a felsic orthogneiss, with Neoarchean U-Pb magmatic zircon ages of 2.60–2.67 Ga. The orthogneiss overlaps in age and is interpreted to be part of the Grouse Creek block of the Albion Mountains to the south. This Archaean metagranitoid is structurally interleaved with paragneiss containing quartzite and calc-silicate rock. Structurally below the orthogneiss, some quartzites have multiple concordant populations of detrital-zircon grains as young as ca. 1700 Ma, while others have no zircon grains younger than ca. 2500 Ma.

Structurally above the Archaean gneiss is a heterogeneous paragneiss that contains calc-silicate and quartzitic rocks with detrital zircons as young as ca. 1460 Ma. Amphibolite in this unit contains zircons dated at ca. 1850 Ma, indicating that this rock can be no older than that and implying considerable structural complexity. The upper part of the Wildhorse gneiss contains metaquartzites bearing zircons as young as ca. 1400 Ma. The protolith of this paragneiss is interpreted as the southernmost exposures of the Lemhi subbasin of the Mesoproterozoic Belt Supergroup.

The upper Wildhorse gneiss includes ca. 695 Ma intrusive orthogneiss that is coeval with Neoproterozoic rift-related volcanic or intrusive rocks near Pocatello, House Mountain, and Edwardsburg, Idaho. This Cryogenian meta-intrusive rock is the likely source of the 650–710 Ma detrital-zircon population in the Big Lost River that drains the core complex. Initial εH values from 675 Ma zircons are between 3.4 and –2.4, suggesting the granitoids had a mixed source in both continental crust and juvenile mantle.

INTRODUCTION

The Pioneer Mountains core complex (PMCC) contains the Wildhorse gneiss, the largest exposure of Precambrian basement in central Idaho. The protolith ages of these metamorphic rocks are important because they place constraints on the boundary between the Archaean Grouse Creek block and Paleoproterozoic metamorphic basement to the north within the Great Falls tectonic zone and Selway terrane (Fig. 1). Further, some gneisses in the Pioneer Mountains may be metamorphosed Belt Supergroup and can thus define the south end of the Belt Basin. Finally, detrital-zircon studies of Big Lost River alluvium demonstrate the presence of a ca. 650–710 Ma population, which may be derived from Cryogenian intrusive rocks in the Wildhorse complex. The Wildhorse gneiss is therefore an important and unique window into the age and tectonic affinity of the Precambrian crust of central Idaho.

Our goals were to investigate the age and correlation of protoliths for the gneiss complex of Wildhorse Creek, the structurally lowest lithodeme in the Pioneer Mountains, south-central Idaho (Dover, 1969, 1981, 1983; O’Neill and Pavlis, 1988). Prior to this study, the only available radiometric date from the Wildhorse complex was a ca. 2.0 Ga Rb-Sr whole-rock age reported by Dover (1983). The exposures were recognized as part of the Pioneer Mountains metamorphic core complex by Wust (1986). Much of the deformation and magmatism within the PMCC has been interpreted as Eocene (Vogl et al., 2012).

In this paper, we present U-Pb zircon data for 15 rock samples, representing a complete structural section of the Wildhorse gneiss, and one sample of Wildhorse Creek alluvium. These data help to address three primary issues: (1) whether the Wildhorse gneiss in the PMCC is part of the Archaean Grouse Creek basement block (Fig. 1) (Gaschnig et al., 2013); (2) the location of the southern extent of the Mesoproterozoic Belt Supergroup, which is extensively exposed to the north and east (Fig. 1; Link et al., 2007); and (3) the origin of the population of anomalous 650–710 Ma zircon grains found in Big Lost River alluvium (Link et al., 2005).

Precambrian Basement in the Northern Rockies

North of the Snake River Plain, Precambrian metamorphic basement is exposed in the Pioneer Mountains and in isolated areas within the Atlanta lobe of the Idaho batholith to the southwest (Fig. 1 inset) (O’Neill and Pavlis, 1988; Mueller et al., 2002; Foster et al., 2006; Gaschnig et al., 2013; Ma et al., 2016). Proterozoic metamorphic rocks also occur in the Beaverhead Mountains along the Idaho-Montana border where several exposures have U-Pb zircon crystallization ages ca. 2.45 Ga (Kellogg et al., 2003). Immediately to the south of the Coyote Creek quadrangle (Fig. 1 inset), the gneiss of Bloody Dick Creek contains a significant population (60%) of ca. 1.88 Ga zircons (Sherwin et al., 2016).
Figure 1. General tectonic map of western Laurentia showing ages of terranes and geographic areas mentioned in text (after Foster et al., 2006, and Mueller et al., 2011). General trend of belt of metamorphic core complexes is shown. Dark areas are Paleoproterozoic and older metamorphic rock exposures. Belt Basin contains Mesoproterozoic metasedimentary rock. The Pioneer Mountain core complex (PMCC, shown in red) and House Mountain exposures are labeled. Big Creek (650 Ma) and Beaverhead (500 Ma) plutons in east-central Idaho (Lund et al., 2010) are shown in red. Clump of exposures in northwestern Utah and adjacent Idaho is the Albion–Raft River core complex. Inset map: Geographic map of Idaho showing features and geologic units mentioned in text. The Big Lost River drains from the Pioneer Mountains to the Snake River Plain and transports distinctive 675 Ma zircon grains. Location of U.S. Geological Survey (USGS) Drill Hole 142 is shown.
Mueller et al. (2016) report on two samples from the gneiss of Bloody Dick Creek. Sample BDC-2 has a primary age of 1904 ± 13 Ma and sample BDC-6 has a primary age of 1799 ± 7 Ma. To the north in Montana, the Selway terrane and Great Falls teutonic zone (Fig. 1; O’Neill and Lopez, 1985) contain Paleoproterozoic juvenile crust (1.7–1.86 Ga) with older fragments at 2.2–2.6 Ga (Foster et al., 2006, 2012; Gifford et al., 2014). Alcock and Muller (2012) and Alcock et al. (2013) identify Paleoproterozoic strata (Montana metasedimentary terrane) in what had been considered Archean rocks of the Wyoming Province in the Ruby Range of southwest Montana (Fig. 1). The northern Bitterroot lobe of the Idaho batholith intrudes Paleoproterozoic 1.6–1.8 Ga primitive arc-like rocks (Toth and Stacey, 1992; Mueller et al., 1996; Foster and Fanning, 1997).

In the Albion Mountains to the south of the Neogene Snake River Plain (Fig. 1), the basement comprises the Neoproterozoic Grouse Creek complex (2.5–2.6 Ga; see Egger et al., 2003; Strickland et al., 2011). Xenoliths of felsic gneiss in Snake River Plain basalt lavas contain zircons with ages of 2.5–3.2 Ga (Lee et al., 1985; Wolf et al., 2005). East of the exposed Grouse Creek block, the Farmington complex contains gneisses with earliest Paleoproterozoic ages of ca. 2.45 Ga (Foster et al., 2006; Shervais, 2006; Mueller et al., 2011).

The intracratonic Belt basin north and east of the Pioneer Mountains contains the Belt Supergroup, principally quartzose and locally feldspathic strata that were deposited between ca. 1470 and ca. 1390 Ma (Harrison et al., 1974; Ruppel, 1975; Winston and Link, 1993) (Fig. 1). Comparable sequences in Idaho were deposited in the Lemhi subbasin (Fig. 1 inset). The rocks of the Belt basin and the Lemhi subbasin share broad stratigraphic relations, detrital-zircon age populations, have overlapping depositional age constraints, and cannot be structurally separated (Winston et al., 1999; Link et al., 2007, 2016; Stewart et al., 2010; Burmester et al., 2016). The southern extent of the Belt Supergroup is currently unknown.

**PIONEER MOUNTAINS**

The Pioneer metamorphic core complex (Wust, 1986; O’Neill and Pavlis, 1988; Silverberg, 1990; Wori et al., 1995; Vogl et al., 2012, 2014; McFadden et al., 2015) is part of the belt of Cordilleran metamorphic core complexes (Fig. 1) that formed during Cenozoic postorogenic extension in the hinterland of the Sevier orogenic belt. Precambrian and lower Paleozoic rocks are exposed in the footwall of the Wildhorse detachment fault (Dover, 1981, 1983; Wust, 1988). Within that footwall, Umpleby et al. (1930) recognized the Wildhorse gneiss in the structurally lowest exposures. Dover (1983) divided the gneiss complex of Wildhorse Creek into a lower (quartzitic) gneiss, a middle (felsic) orthogneiss, a mafic gneiss, and an upper (quartzitic) gneiss. The ages for all those units were designated as Paleoproterozoic on the basis of preliminary Rb-Sr whole-rock dating. A revised geologic map of the Wildhorse gneiss that incorporates our recent U-Pb zircon age data is given as Figure 2.

The footwall of the PMCC contains a NNW-trending, doubly plunging antiform (Wildhorse dome) that is cored by the Wildhorse complex (Fig. 2). Above the gneiss, on the southwest flank of the Wildhorse dome, is a broadly concordant granodiorite sheet with a U-Pb zircon crystallization age of 48.6 ± 0.4 Ma (Vogl et al., 2012). On the east side of the dome is a large expanse of this Eocene granodiorite, locally termed the Pioneer Intrusive Suite (Fig. 2).

The PMCC developed in the Eocene, with extension beginning before ca. 49 Ma and continuing (perhaps episodically) through the Late Eocene. The Wildhorse gneiss underwent high-grade metamorphism and partial melting during the early stages of extension, during and/or before emplacement of voluminous granitoids at ca. 47–50 Ma (Vogl et al., 2012). The base of the middle structural plate of the Wildhorse gneiss (Fig. 2, cross section) underwent metamorphism at depths of ~11–15 km as a result of emplacement of the 48–49 Ma granodiorite sheet. Penetrative strain accompanied amphibolite-facies metamorphism throughout the Wildhorse gneiss and the basal middle plate (Vogl et al., 2012). The pervasive strain, metamorphism, melting, and widespread small-scale intrusions obscure the primary contact relationships between individual units within the basement exposures.

Structurally above the Eocene granodiorite is quartzite, marble, and schist of the middle structural plate of the PMCC. Along this boundary, Vogl et al. (2012) have identified significant structural omissions that predate intrusion of the granodiorite sheet. They suggested that the boundary was an extensional fault along which the granodiorite intruded (Fig. 2, cross section).

Ma et al. (2016) examined zircons in Paleozoic wall rocks of the Idaho batholith in the Sawtooth Mountains (location shown in Fig. 1 inset). Their samples all contain Mesoproterozoic (Grenville-age) zircons 1200–1000 Ma and are therefore younger than any of the Wildhorse gneiss complex. These rocks likely are of the same age as metaquartzites and calc-silicate schists in the middle structural plate within the Pioneer Mountains, structurally above the Wildhorse gneiss (unit OCZs on Fig. 2).

The upper plate of the core complex, above the Wildhorse detachment fault, contains Paleozoic sedimentary rocks of the central Idaho thrust belt including the Devonian Milligen Formation and Pennsylvanian and Permian Sun Valley Group (Mahoney et al., 1991; Link et al., 1995, 2014). To the north and east of the PMCC and the Pioneer thrust fault (Fig. 1 inset), upper-plate strata belong to the Lower Mississippian Copper Basin Group (Link et al., 1996; Beranek et al., 2016).

**METHODS**

Durk (2007) conducted field mapping and sampling of the headwaters of Wildhorse Creek. Cameron (2010) mapped and sampled a section of the Wildhorse gneiss on the southwest flank of the dome. To complement this Idaho State University senior thesis work, Vogl and Foster conducted reconnaissance sampling in 2013.

Zircon grains from Wildhorse Creek alluvium and Wildhorse complex orthogneiss and paragneiss were handpicked from heavy-mineral concentrates, placed onto double-sided tape, together with either Duluth Gabbro (FC1) or...
Figure 2. Geologic map and cross section of the gneiss complex of Wildhorse Creek, showing locations of samples discussed in this paper. Inset map shows regional context of Pioneer Mountain core complex (PMCC). Cross section shows structural arrangement of map units, and serves also as a stratigraphic diagram. WGC—Wildhorse Gneiss Complex.
Temora reference zircons, cast into epoxy disks and sectioned approximately in half, and polished. Transmitted- and reflected-light photomicrographs and cathodoluminescence (CL) images were made for all grains. Because the zircon grains from these gneissic samples have thin metamorphic rims, likely Eocene in age, for the purposes of this study, the areas analyzed were centered on the central areas that were considered to record the primary protolith history. The U-Pb analyses for 14 samples were carried out using high-resolution ion microprobe-reverse geometry (SHRIMP-RG and SHRIMP II) at the Research School of Earth Sciences, The Australian National University, Canberra, Australia, following procedures described in Williams (1998) and references therein. The data have been processed using the SQUID Excel macro of Ludwig (2000). Plots and age calculations were carried out using Isoplot/Ex, version 3.00 (Ludwig, 2003). A concordance filter of 20% was used for the paragenesis samples. GPS locations and sample descriptions are in Table S1. Data for samples run on the SHRIMP are shown in Table S2. Two other samples, DF-13-01 and DF-13-02, were dated by laser ablation–multicollector–inductively coupled mass spectrometry (LA-MC-ICPMS) at the University of Florida using methods similar to those described in Foster et al. (2012); data are given in Table S3.

We ran Lu-Hf isotopes on selected 675 Ma detrital zircons from a subsurface sample of Big Lost River alluvium, using LA-MC-ICPMS at the Arizona Laserchron Laboratory, following the methods outlined in Cecil et al. (2011) and Gehrels and Pecha (2014). Data are in Table S4. Hf isotopic values bear on the nature of the crust that was partially melted to produce the plutons in which the zircons crystallized (Kinny and Maas, 2003; Goode and Vervoort, 2006; Bahlburg et al., 2011).

### LITHOLOGIES AND U-Pb GEOCHRONOLOGY

Here, we report the first U-Pb zircon geochronology of the Wildhorse gneiss complex. We have dated all the map units designated by Dover (1983), spanning the complete exposed structural section around the Wildhorse dome. The new geologic map, incorporating these age assignments and revised informal unit names, is shown in Figure 2. The cross section on this figure shows the interpreted structural positions of those mapped units. Field photographs of representative lithologies are shown in Figure 3. Representative cathodoluminescence (CL) images are shown in Figure 4 and Figure S1.

#### Lower Paragneiss

The lower paragneiss (map unit ZYwg; lower gneiss of Dover, 1983) crops out in the canyon and lower slopes of middle Wildhorse Creek, as well as in three cirques to the southwest in the flat-lying core of the dome (Fig. 2). Paragneiss is recognized by the presence of lenses of gneissose quartzite (metamorphosed sandstone), marble, calc-silicate schist, and biotite schist. The unit contains mainly quartzite-bearing gneiss at the base with calcareous schist in the upper part. Foliation is defined by biotite. The units are intruded by meter-scale dikes, variably migmatisic, with in situ melt in the quartzofeldspathic parts. The detrital quartz grains in the quartzite pods and lenses have been recrystallized and display sutured contacts. The minimum unit thickness is ~200 m.

Sample 28PL09 from the lower paragneiss is an orthoquartzite from the Wildhorse mine (Fig. 2); this sample appears to be cross laminated (Fig. 3A). Zircon plots for the lower paragneiss are shown in Figure 5. The analyses scatter widely on a Wetherill concordia plot (Fig. 5A). The probability density plot (Fig. 5B) of radiogenic 238U/206Pb and dates for 40 of the 50 grains analyzed shows a considerable range from ca. 2500 Ma to ca. 3250 Ma, with few clear age groupings (Table S2C [see footnote 2])

Sample DF-13-02 is an orthoquartzite interlayered with marble and calc-silicate rocks. The zircon grains all yield discordant analyses and scatter widely on a Wetherill concordia plot (Fig. 5C; Table S3B [see footnote 3]). Nevertheless, all grains record Neoarchean 238U/206Pb dates.

Sample 54PL09 (Figs. 5D and 5E; Table S2D [see footnote 2]) is an orthoquartzite adjacent to partially melted zones and multiple crosscutting granitic dikes near the confluence of the east and west forks of Wildhorse Creek. Thin light-colored rims are present on some grains, but overall, the analyses scatter widely on a Wetherill concordia plot (Fig. 5D). The 14 least discordant grains in 54PL09 have 206Pb/238Pb dates ranging from ca. 3.2 to ca. 2.4 Ga (Fig. 5E), suggesting a diverse Neoarchean provenance. The youngest discordant zircon is 687 Ma.

Sample 1AC09 is from a 1 m lens of quartzite within impure marble from the Left Fork of Wildhorse Creek. The zircon grains from this sample are characterized by mottled surfaces under transmitted light, which indicates that they
Felsic Orthogneiss

Felsic orthogneiss ~700 m thick overlies the lower paragneiss around the entire Wildhorse dome (map unit Wwfg, Middle gneiss of Dover, 1983). The rock is light-gray to white, equigranular, fine to medium grained, and varies from biotite quartz monzonite to trondhjemite gneiss. The lower contact of this paragneiss is sharp but not well exposed. We follow Cameron (2010) and interpret the contact as a thrust fault. The felsic orthogneiss forms the high walls of many of the cirques in the area.
Zircons from four samples were dated from the north and southwest flanks of the Wildhorse dome (Fig. 6; samples 6PL05, 1KD06, 14AC09, and 17AC09; Tables S2E–S2H [see footnote 2]); a field photograph is given in Figure 3B with the Wetherill concordia plots given in Figure 6. Magmatic zircon grains in the felsic orthogneiss are clear, elongate prismatic, euhedral to subhedral, with a maximum size of ~275 microns (Fig. 4B). The larger zircons have well-defined oscillatory CL zoning. Most of the grains have diffuse central areas with a thin rim, and some have metamict cores.

The U-Pb data from all four samples define simple discordia regression arrays with upper intercept ages ranging from ca. 2595 Ma to ca. 2670 Ma (Fig. 6). We interpret the crystallization of this orthogneiss to be between 2.60 Ga to 2.65 Ga. The lower concordia intercepts for the regression lines range from ca. 100 to ca. 55 Ma, consistent with variable radiogenic lead loss during Eocene magmatism and metamorphism. The elevated U and lower Th/U ratios in some rims suggest that they are due to Eocene partial melting.
Figure 5. U-Pb detrital-zircon data from quartzitic lower paragneiss. (A) Wetherill concordia plot showing U-Pb analyses of dated detrital-zircon from sample 26PL09 (<20% discordance, n = 40/50; Table S2C [see footnote 2]). Analyses ≤20% discordant plotted as 1σ error ellipses. The detrital grains have dominantly Archean 207Pb/206Pb ages (see Fig. 4B). (B) Probability density plot of 207Pb/206Pb ages for detrital-zircon grains analyzed from sample 26PL09. (C) U-Pb concordia plot from sample DF-13-02. Analyzed at University of Florida (Table S3B [see footnote 3]). (D) Concordia plot of sample 54PL09, lower gneiss at the East Fork Wildhorse Creek crossing, showing U-Pb analyses of detrital zircons. Twenty-six of 41 grains were <20% discordant; Table S2D (see footnote 2). Analyses plotted as 1σ error ellipses. (E) Probability density plot of detrital-zircon ages, sample 54PL09. Fifteen of 41 grains were ≤20% discordant. (F) Lower end of concordia plot (Paleozoic and younger grains) from sample 1AC09 (Table S2B [see footnote 2]). Analyses plotted as 1σ error ellipses. (G) Upper end of concordia plot (Proterozoic grains) from sample 1AC09. Analyses plotted as 1σ error ellipses. MSWD—mean square of weighted deviates; conf—confidence.
Middle Paragneiss

The felsic orthogneiss is overlain across a sheared zone or a sharp contact by a heterogeneous package of metasedimentary rocks, which we refer to as the middle paragneiss (map unit XYwmg; Mafic gneiss of Dover, 1983). This unit is dominated by fine- to medium-grained, equigranular, thinly layered metapsammitic paragneiss that is interlayered with quartzite gneiss and contains several intervals of diopside-bearing calc-silicate rock (Fig. 3C). The unit contains locally abundant amphibolite that cuts the calc-silicate. The amphibolite is commonly isoclinally folded and boudinaged. Overall, this unit is highly strained as indicated by the folds, boudins of Eocene leucogranite, and porphyroclastic quartz-feldspar aggregates. The structural thickness of this unit is ~500–700 m.

Zircon age data from the middle paragneiss are shown in Figure 7. Detrital zircons were analyzed from a calc-silicate (4PL05) from the lower part of the middle paragneiss at the north end of the Wildhorse dome. The zircons are predominantly round and subround grains with a few that are euhedral in shape; overall they are considered to be detrital. Twenty-one of 29 grains analyzed record U-Pb ages that are less than 20% discordant (Fig. 7A and Table S2I [see footnote 2]). The 207Pb/206Pb ages of these low-discordance grains range from ca. 1800 Ma to ca. 1400 Ma, with one grain at ca. 2600 Ma (Fig. 7B). The morphology and age data suggest that the sandstone protolith was...
not deposited earlier than 1400 Ma and clearly could be much younger given the limited data set on this sample.

Sample DF-13-01 comes from a medium-grained lens of amphibolite within the middle part of the middle paragneiss unit where mafic boudins are abundant. The analyses scatter about a discordia trend with an upper intercept of 1860 ± 20 Ma (Fig. 7C; Table S3A [see footnote 3]). There are three possibilities for this age, all of which require the middle part of the middle paragneiss to be Paleoproterozoic or older. (1) If these boudins were basalt flows or synsedimentary sills, then this part of the section was deposited at ca. 1860 Ma. (2) If the zircon is metamorphic in origin, then the 1860 Ma zircon ages represent a minimum age of deposition of this part of the section. (3) If the boudins were originally mafic dikes, then the section must be older than 1860 Ma. Regardless of the interpretation, these data indicate that the entire middle paragneiss is not a Mesoproterozoic clastic package as previously suggested (Link et al., 2010) and that the protoliths of the lower exposures are younger than at least some of the structurally higher units.
Upper Paragneiss

Above the biotite-rich middle paragneiss is the upper paragneiss (map unit ZYwug; Upper gneiss of Dover, 1983), comprising gneissose quartzite, quartzofeldspathic gneiss, minor calc-silicate, amphibolite, and fine- to coarse-grained augen gneiss. The unit is medium-gray to white, sometimes with a green hue, and medium fine to medium grained. The unit contains gneissose quartzite in the lower 10–30 m. Grains to granule size (>2 mm) are present east of Wildhorse Creek. The unit is intruded by Eocene and Neoproterozoic plutonic rocks in the upper part; the basal contact is irregular. On the southwest side of the Wildhorse dome, this unit is ~300 m thick (Cameron, 2010).

Data from detrital-zircon grains found in the gneissose quartzite are shown in Figure 8 (samples 14KD06 [Figs. 8A and 8B] and 24KD06 [Figs. 8C and 8D]; Tables S2J and S2K [see footnote 2]). The zircons are clear, some with a slight red hue, round to subrounded, ranging to elongate prismatic, subhedral grains that are up to 200 µm length, with an overall average size of around 100 µm. The CL images (see Fig. 4C) show little-zoned to unzoned central areas that dominate each sectioned grain, with a very narrow bright CL rim to most.

The $^{207}\text{Pb}/^{206}\text{Pb}$ ages for the detrital zircons from both samples are generally between ca. 1450 and ca. 1800 Ma with prominent peaks ca. 1650–1700 and ca. 1680 Ma, respectively (Fig. 8). Two older grains are recorded in sample 24KD06 (ca. 2080 Ma and 2850 Ma) and also two Neoproterozoic grains. These age dis-
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The weighted mean 207Pb/206Pb age therefore has some scatter (MSWD = 1.3) ca. 693 ± 7 Ma (Fig. 9F). This correspondence in the weighted mean of upper-intercept ages from all three samples, within two sigma error, leads to the conclusion that the crystallization age of the orthogneiss is ca. 695 Ma.

DISCUSSION

Timing of Igneous Activity

Neoproterozoic

The felsic orthogneiss in the Pioneer Mountains was intruded in the Neoproterozoic. In all four samples, there is variable and at times significant radiogenic Pb loss, with the more concordant ages for two samples at ca. 2.66–2.67 Ga and the other two at ca. 2.60–2.61 Ga. The older 207Pb/206Pb age groupings are identical to intrusive ages from the Clearwater and Priest River complexes to the north as reported by Vervoort et al. (2015). The younger set of ages (2.60–2.61 Ga) from the Pioneer Mountains approach the ca. 2.57 Ga ages from xenoliths in Miocene volcanics from ~50 km to the southwest and ca. 2.50–2.60 Ga from Craters of the Moon National Monument to the south (Wolf et al., 2005). Furthermore, basement orthogneiss (Green Creek gneiss) from the Albion and Raft River Mountains immediately south of the Snake River Plain (Fig. 1) has yielded ages of ca. 2.57 Ga (Strickland et al., 2011).

Paleoproterozoic

Our new age of ca. 1.85 Ga for amphibolite in the middle paragneiss indicates that this unit, which was previously thought to have been deposited after ca. 1.45 Ga, locally contains metasedimentary rocks that are Paleoproterozoic or older. These ages also represent the southernmost exposures of rocks of this age that are prevalent to the north within both the Clearwater–Priest River block (Vervoort et al., 2015) and as inherited zircons within the Bitterroot lobe of the Idaho batholith (Gaschnig et al., 2013), as well as to the northeast along the Great Falls tectonic zone (e.g., Mueller et al., 2002, 2016; Vogl et al., 2004; Foster et al., 2006).
Figure 9. U-Pb zircon data from the upper orthogneiss (samples 12KD06, 23KD06, and 28KD06). (A) Sample 12KD06 concordia plot showing U-Pb analyses of dated zircon (Table S2L [see footnote 2]). Analyses plotted as 1σ error ellipses. (B) Sample 23KD06, concordia plot showing U-Pb analyses of dated zircon (Table S2M [see footnote 2]). Analyses plotted as 1σ error ellipses. (C) Sample 28KD06 concordia plot showing U-Pb analyses of dated zircon (Table S2N [see footnote 2]). Analyses plotted as 1σ error ellipses. (D) Sample 12KD06 weighted average 207Pb/206Pb age. (E) Sample 23KD06 weighted average 207Pb/206Pb age. (F) Sample 28KD06 weighted average 207Pb/206Pb age. MSWD—mean square of weighted deviates; conf—confidence.
The association of ca. 1.85 Ga gneisses with 2.6–2.7 Ga gneisses appears to be common throughout limited exposures and xenoliths of the Medicine Hat block (Foster et al., 2006, 2012; Gifford et al., 2012; Vervoort et al., 2015) suggesting a relationship between the basement of the Grouse Creek and Medicine Hat blocks. Although we have only dated one sample of this age thus far, we note that there is less than 1000 m of pre-Mesoproterozoic structural section (Archean felsic orthogneiss and part of the middle para gneiss) exposed within the PMCC, providing a very limited view of the lowest basement.

Neoproterozoic

The 695 Ma orthogneiss is of the same age as Cryogenian volcanic rocks at Edwardsburg 170 km to the northwest (Lund et al., 2003, 2010). The ages also overlap with zircon populations in epiclastic volcanic sandstones of the Cryogenian Pocatello Formation south of the Snake River Plain (Keeley et al., 2013). Gaschnig et al. (2013) reported a significant component of inherited zircons with an age peak of ca. 670 Ma from the southern lobe of the Idaho Batholith. Yonkee et al. (2014) proposed that these magmatic rocks are related to rifting of Laurentia and the opening of the paleo–Pacific Ocean.

To the west in the Atlanta lobe of the Cretaceous Idaho batholith, Gaschnig et al. (2013) found a population of inherited zircons with an age peak of ca. 2.55 Ga, as well as a smaller population at ca. 671 Ma. They interpreted these inherited zircons as being derived from melting of an igneous protolith, rather than as detrital zircons from metasedimentary rocks. At House Mountain (Fig. 1), along the South Fork Boise River 100 km southwest of the PMCC, Cryogenian (725 Ma) orthogneiss intrudes Archean orthogneiss with an age of 2.55 Ga (Alexander et al., 2006; Alexander, 2007; Schmitz, 2011). This growing database of basement ages suggests that Neoarchean crust intruded by Neoproterozoic granitoids is present along a north-south distance of >500 km along the southwest edge of Laurentia (Fig. 1). The ages of these Neoarchean orthogneisses span a period of ~100 m.y.

Depositional History and Sediment Sources

Lower and Middle Paragneiss Units

The data from the lower paragneiss samples (Fig. 5) suggest deposition after ca. 2.5 Ga, and for sample 1AC09, the data suggest deposition after 1.7 Ga. Because of the differing detrital-zircon spectra, we suggest that several stratigraphic units are present. The age of ca. 1850 Ma from amphibolite within the middle paragneiss suggests that part of this package is Paleoproterozoic or older and must be more structurally complex than previously thought (Link et al., 2010).
We compared the detrital signatures of several of the PMCC units and Proterozoic quartzites regionally (Fig. 12). Three samples of the lower paragneiss of the Wildhorse complex contain Archean detrital zircons with a range in ages from 2.5 to 3.3 Ga (two samples shown in Fig. 12E). This distribution does not resemble any part of the Lemhi Group or Belt Supergroup (Figs. 12A and 12B) but is similar to detrital-zircon patterns reported from the Cryogenian schist of the Upper Narrows (Fig. 12F) of the Raft River Range (Yonce et al., 2014). One sample of quartzite in the lower paragneiss (1AC09) contains a small population of concordant zircons with ages near 1710 Ma (see Table S2B [see footnote 2]). This age is similar to the largest age peak in the Lawson Creek, Coyote Creek quadrangle (Sherwin et al., 2016). Paleoproterozoic rocks are present in the Wildhorse gneiss. The age of ca. 1850 Ma from amphibolite within the middle paragneiss suggests that part of this package is Paleoproterozoic or older and more structurally complex than previously thought (Link et al., 2010). This age overlaps 2.45 and 1.8–1.9 Ga Paleoproterozoic (Farmington zone) metavolcanic rocks dated in the Tendoy Mountains, Montana (Fig. 1; Kellogg et al., 2003; Mueller et al., 2016) and the Coyote Creek quadrangle (Sherwin et al., 2016).

**Regional Basement Relations**

Wildhorse complex orthogneiss has a similar 2.6 Ga Neoarchean age as granites of the Grouse Creek block. The northwestern Wyoming province has metasedimentary rocks older than 3 Ga (Foster et al., 2006). We do not find evidence in the Wildhorse gneiss for pre–2.6 Ga magmatic rocks and southwest structural trends predicted by Sims et al. (2004, 2005). These authors recognize northeast-striking aeromagnetic discontinuities in central Idaho that are interpreted to continue along strike into Archean terranes of southwest Montana. Extension of the Wyoming province southwestward would suggest that the Archean rocks under central Idaho should contain Mesoproterozoic amphibolite or granulite-grade metavolcanic and metasedimentary rocks. We have found no such ancient strata in the Pioneer Mountains.

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**CONCLUSIONS**

This paper establishes several geochronologic and tectonic geologic relations in the Pioneer Mountains.

1. We elucidate age relations between lithologic units in the Wildhorse complex. These include 2.6 Ga Neoarchean orthogneiss of the Grouse Creek block, 1.8 Ga Paleoproterozoic amphibolite, 1.45 Ga Mesoproterozoic paragneiss, and 695 Ma Neoproterozoic orthogneiss.
2. We clarify tectonic relationships between the Grouse Creek block and neighboring Archean Wyoming and Medicine Hat blocks and Paleoproterozoic Great Falls tectonic zone. The Pioneer Mountains represent a northern extension of the Grouse Creek block north of the Snake River Plain against the southern margin of the Great Falls tectonic zone on the north and the Farmington zone on the east.
3. We identify Mesoproterozoic depositional ages and provenance similarities that support a southward extension of the Mesoproterozoic Belt Basin to the Pioneer Mountains.
4. We identify zircon-rich Neoproterozoic intrusive rocks (metamorphosed to orthogneiss) in the Wildhorse complex. These rocks likely record rifting along the western margin of Laurentia and are the source for 710–650 Ma detrital-zircon populations from the Lost River drainage.
5. Initial εHf values from 675 Ma detrital zircons of the lower Big Lost River reflect a mixture between older continental material and more juvenile mantle-derived magma as the source for the Neoproterozoic orthogneiss. Thus if there was an Archean Grouse Creek block component, there was also a juvenile mantle component.

![Figure 11. εHf plot for 675 Ma zircons in Big Lost River alluvium. These grains are interpreted to come from the Neo-proterozoic orthogneiss in the Pioneer Mountains, and the Hf isotopic values thus reflect the evolved Precambrian crust from which these plutons were partially melted. CHUR—chondritic uniform reservoir; DM—depleted mantle.](image)
Figure 12. Probability density plots of selected metasandstones from the Pioneer Mountains, the Belt Supergroup and one sample from the Albion Range, Idaho. (A) Lawson Creek and overlying strata, Lemhi and Beaverhead Ranges (Link et al., 2016). (B) Apple Creek Formation, Blackbird district, Salmon River Mountains (sample SS95-19), Link et al. (2007). (C) Upper paragneiss quartzite (samples 14KD and 24KD06). (D) Middle paragneiss calc-silicate (sample 4PL05). (E) Lower paragneiss quartzite (samples 26PL09 and 54PL09). (F) Neoproterozoic Upper Narrows schist, Raft River Range (Yonkee et al., 2014; sample 14PL07).
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