Lu-Hf analyses of 685–730 Ma igneous zircons yield enriched initial εHf values in the range +2 to −17, indicating that they crystallized from magma that incorporated depleted Paleoproterozoic to Archean crustal components of the underlying Farmington Canyon Complex and Wyoming craton.

INTRODUCTION

Volcanic, siliciclastic, and minor carbonate strata of the Neoproterozoic Pocatello Formation in southeast Idaho record rifting of the supercontinent Rodinia during widespread regional glaciation (Fig. 1) (Ludlum, 1942; Crittenden et al., 1983; Link, 1983; Link et al., 1994; Smith et al., 1994; Lorentz et al., 2004; Corsetti et al., 2007). These strata bear on paleogeographic, Snowball Earth, and Rodinia rift models that seek to explain drastic tectonic and climatic fluctuations near the onset of complex life (Hoffman, 1991; Hoffman et al., 1998; Li et al., 2008; Hoffman and Li, 2009).

Neoproterozoic glacial successions are recognized worldwide, but the number, timing, and duration of glacial episodes remain controversial. Three generally agreed upon Neoproterozoic glacial phases are the Middle Cryogenian “Sturtian,” the Late Cryogenian “Marinoan,” and the Ediacaran “Gaskiers” (Hoffman and Li, 2009). The Sturtian, Middle Cryogenian glaciations derive their name from exposures of diamictite in the “gorge of the River Sturt,” south of Adelaide, South Australia (Howchin, 1901; Preiss, 2000; Preiss et al., 2011). This name has been applied worldwide, and current age constraints on the glaciation suggest it was diachronous between ca. 716 Ma and ca. 660 Ma. The older bound is the chemical abrasion–isotope dilution–thermal ionization mass spectrometry (CA-ID-TIMS) U-Pb zircon age of 716.47 ± 0.24 Ma from diamictites of the Mount Harper Group in the Yukon Territory representing the Rapitan glaciation (Macdonald et al., 2010). The Ghubrah Formation in Oman is slightly younger, at 711.5 ± 0.3 Ma (Browning et al., 2007). A considerably younger bound is the U-Pb sensitive high-resolution ion microprobe (SHRIMP) age of 660 ± 6 Ma from a tuff in a dropstone-bearing sandstone within the Sturtian glacial succession near Copley in South Australia (Fanning and Link, 2006, 2008). In South Australia, the postglacial Tapley Hill Formation is composed of thick transgressive shale that yielded a Re-Os age of 643 ± 2.4 Ma from the basal Tindalina Shale Member (Kendall et al., 2006).

The chronostratigraphic term “Marinoan” comes from the Marino Arkose Member of the Wilmington Formation, which underlies the glaciogenic Elatina Formation (Williams et al., 2008). The Elatina glaciation of Marinoan age in Australia has been interpreted to have occurred at 635 Ma on the basis of chemoostratigraphic correlation with dated glacials in South China and Namibia (Hoffman and Schrag, 2002). The
Marinoan, late Cryogenian glaciation is bracketed between 655 Ma (Zhang et al., 2008) and 632 Ma (Condon et al., 2005), with equivalent 635 Ma U-Pb zircon ages recorded from Namibia (Hoffmann et al., 2004), Oman (Bowring et al., 2007), and South China (Condon et al., 2005).

Whereas the Sturtian and Marinoan glaciations have been considered global glaciations, or “snowball Earth” periods (Kirschvink, 1992), the Gaskiers glaciation is generally limited to high latitudes (>45°) (Hoffman and Li, 2009) and is directly dated by the Gaskiers Formation on the Avalon Peninsula of Newfoundland, Canada (Bowring et al., 2003). It was suggested by Hoffman and Li (2009) that the time constraints for the Gaskiers glaciation are too brief for a global glaciation.

The diamictites that crop out throughout southeast Idaho and northern Utah (Fig. 1) have long been interpreted as having been deposited in a glaciomarine setting (Calkins and Butler, 1943; Crittenden et al., 1983; Link and Christie-Blick, 2011). Lines of evidence for regional glaciation are seen in the type area of the Scout Mountain Member, Pocatello Formation (Fig. 2), where massive and stratified diamictite facies associations have been interpreted as ice-proximal lodgment and ice-distal debris-flow tillite. Rare striated quartzite clasts are present.

Regional equivalents of the Pocatello Formation include the informal formation of Perry Canyon in the northern Wasatch Range, the Mineral Fork Formation in the central to southern Wasatch Range, and the Otts Canyon and overlying Dutch Peak Formations in the Sheeprock Mountains, Utah (Fig. 3; Crittenden et al., 1971, 1983; Link et al., 1993; Link and Christie-Blick, 2011). Glacial dropstones that pierce underlying lamination are common in the Mineral Fork Formation in northern Utah (Christie-Blick, 1983). Dropstones have also been found in the upper Otts Canyon Formation immediately underlying the ~1.5-km-thick stratified and massive diamictite of the Dutch Peak Formation (Christie-Blick, 1983; Link and Christie-Blick, 2011). The formation of Perry Canyon displays two horizons of thick diamictite separated by several hundred meters of glaciomarine strata. The lower of these diamictites has been interpreted to bear dropstones. The two diamictite units here have been interpreted to represent two stades in one glaciation (Crittenden et al., 1983). Overlying the Pocatello Formation, in the Brigham Group, incised valleys between the Caddy Canyon Quartzite and the Inkom Formation (Fig. 3) are interpreted as indirect evidence for eustatic changes due to late Cryogenian glaciation (Link and Christie-Blick, 2011).

The timing and duration of rifting of the Neoproterozoic to Paleozoic western North American passive margin also remain poorly constrained (Moore, 1991; Karlstrom et al., 1999). However, recent age constraints have significantly improved our understanding. Radiometric ages along the western Laurentian Cordillera begin with the ca. 780 Ma Gunbarrel mafic dike swarm, interpreted to suggest crustal extension induced by a mantle plume beneath Rodinia (Harlan et al., 2003). Mechanical rifting along the Cordillera is then recorded by thick volcanic- and diamictite-bearing successions from which volcanic rocks provide ages of 716–711 Ma (Macdonald et al., 2010), ca. 688 Ma (Ferri et al., 1999), and ca. 685 Ma (Lund et al., 2003). Thin volcanic rocks in the overlying Brigham Group and correlatives yield ages of 580–570 Ma (Fig. 3; Christie-Blick and Levy, 1989; Colpron et al., 2002). It is unknown whether the volcanics represent a protracted rift history or separate rift episodes.

The mafic volcanic rocks, diamictite, and abundant feldspathic sandstones that occur in the lower two members of the Pocatello Formation (and correlative strata in Utah) are interpreted to record the initial stages of rift basin development in response to first-order continental rifting during regional glaciation (Link,
Figure 2. Generalized stratigraphic sections. (A) Entire Neoproterozoic section in Pocatello area. (B) Pocatello Formation stratigraphy at Portneuf Narrows. *Black sandstone of Link (1987). Sensitive high-resolution ion microprobe (SHRIMP) maximum age constrains adjacent to column are U-Pb zircon concordia ages (Fig. DR2). (C) Revised stratigraphy of Pocatello Formation at Oxford Mountain. Chemical abrasion–isotope dilution–thermal ionization mass spectrometry (CA-ID-TIMS) age on Oxford Mountain is a U-Pb zircon weighted mean. Modified from Fanning and Link (2004) based on new mapping.

Figure 3. Composite correlation chart of Neoproterozoic and Lower Cambrian rocks of the Cordillera. The allochthonous and parautochthonous rocks in Utah-Idaho refer to rocks in the hanging wall and footwall, respectively, of the Paris-Willard thrust sheet in, respectively. BHF—Brown’s Hole Formation; CLG—Coates Lake Group; Cls—Cambrian limestone and shale undifferentiated; DGD—Daugherty Gulch diamictite; LMHG—Lower Mount Harper Group; MHVC—Mount Harper Volcanic Complex; MMS—Mackenzie Mountains Supergroup; qzt—quartzite undifferentiated. (1) Macdonald et al. (2010); (2) Macdonald et al. (2011); (3) Ross et al. (1995); (4) Atkin (1991); (5) Feni et al. (1999); (6) Colpron et al. (2002); (7) Kendal et al. (2004); (8) Lund et al. (2003); (9) Lund et al. (2010); (10) Dehler et al. (2010); (11) Mahon (2012); (12) Dehler et al. (2012); (13) Christie-Blick and Levy, (1989). *After Balgord (2011); †Proposed as Horse Thief Springs Formation by Mahon (2012). Adapted from Link and Christie-Blick (2011).
1987; Link et al., 1994; Link and Christie-Blick, 2011). The Paris-Willard thrust sheet of southeast Idaho contains the 1500-m-thick Pocatello Formation over lain by a 4000-m-thick package of westward-thickening quartzite strata, the Brigham Group (Crittenden et al., 1971; Link et al., 1987). Overlying the Brigham Group, there are ~2000 m of dominantly carbonate and subordinate siliciclastic Cambrian strata. This package thins to the east and gives way to Lower Paleozoic to Mesozoic rocks. This sedimentary package thickens west of and subparallel to ~112°W longitude, known as the Cordilleran hinge line, which is interpreted by many workers to be the W-facing, trailing edge of Laurentia after Rodinia rifting. Abundant W-directed paleocurrents (Link et al., 1987; Keeley et al., 2009) and numerous detrital zircon studies linking migogeosynclinal sedimentary provenance to basement provinces east of the hinge line (Rainbird et al., 1992; Stewart et al., 2001; Schoenborn et al., 2012; Rainbird et al., 2012) support this interpretation. West of the hinge line, near the Idaho–Oregon border, the 0.706 ⁰⁶Sr/⁰⁶Sr line is interpreted to mark the paleogeographic border of western Laurentia.

The purpose of this study is to better resolve age inconsistencies from the Pocatello Formation so as to better place it in a global framework for the Neoproterozoic. The work presented in this paper first describes new results of field mapping followed by a detrital zircon provenance analysis carried out by laser ablation–multicolonlector–inductively coupled plasma–mass spectrometry (LA–MC–ICP–MS). The latter was also used as a screening process to locate Neoproterozoic zircons. Maximum age constraints were then obtained by methods of increasing precision, first SHRIMP, followed by CA-ID–TIMS. The zircons from the former analyses were then analyzed for lutetium–hafnium isotopes in an effort to begin developing a Neoproterozoic Hf database and to make some inferences about the petrogenesis of the zircons. Finally, we used the new ages to place the Pocatello Formation in context with Neoproterozoic glaciations and Rodinia rifting.

**GEOLOGIC BACKGROUND**

**Geologic Setting and Regional Stratigraphy of the Pocatello Formation**

The Pocatello Formation is exposed in the Bannock and Pocatello Ranges (Ludlum, 1942; Link et al., 1993; Link and Christie-Blick, 2011). The outcrops in Idaho are aligned N-S and extend ~120 km from east of Pocatello in the north, to the southern end of the Bannock Range, 25 km south of Oxford Mountain (Fig. 1). The lower contact is not exposed (Fig. 2A). The correlative formation of Perry Canyon in northern Utah (Crittenden et al., 1983; Balgod et al., 2011) rests on Paleoproterozoic Facer Formation above the Willard thrust fault, in the footwall of the active Wasatch fault. Exposed below the Willard thrust fault, there is the Paleoproterozoic Farmington Canyon Complex, which is projected northward and thought to underlie the Bannock and Pocatello Ranges in SE Idaho (Foster et al., 2006).

Part of the type section of the Pocatello Formation is located in an overturned limb of an E-vergent fold, south of China Peak and north of the Portneuf Narrows near Pocatello (Ludlum, 1942; Trimble, 1976; Link, 1983). Here, the ~1500 m Pocatello Formation is divided into the lower basaltic Bannock Volcanic Member, the middle Scout Mountain Member, containing diamictite, siliciclastic rocks, and minor carbonate, and an informal upper member of phyllitic shale (Crittenden et al., 1971, 1983; Link, 1983) (Fig. 2B).

**Bannock Volcanic Member**

Harper and Link (1986) measured trace and rare earth elements of basalt samples from the Bannock Volcanic Member and determined a tholeiitic–alkaline to subalkaline composition, comparable to other within-plate basalts. Felsic porphyritic volcanic clasts within Scout Mountain Member diamictite have been interpreted to originate from felsic lava flows in the Bannock Volcanic Member (Link, 1983). However, the lack of rhyolite in the Bannock Volcanic Member led Harper and Link (1986) to conclude volcanism was not bimodal. Keeley (2011) reported major- and trace-element geochemistry that suggested a within-plate continental rift setting for both Bannock Volcanic Member basalts and the trachytic to rhyolitic suite of felsic volcanic clasts from the lower Scout Mountain Member. Assuming that the analyzed volcanic clasts (Keeley, 2011) are representative of the full compositional range, the clast chemistries indicate a span of intermediate to felsic volcanism interpreted to be associated with the basaltic Bannock Volcanic Member.

**Scout Mountain Member**

In more detail, the Scout Mountain Member at Portneuf Narrows (Fig. 2B) includes a lower diamictite interval, which contains locally stratified, green to brown, matrix-supported diamictite with mafic volcanic, argillite, and rare felsic porphyritic volcanic clasts. Above this lies locally ferruginous sandstone and conglomerate and an upper massive diamictite unit that contains quartzose, gneissic, granitic, and silicic volcanic clasts, including glacially striated quartzite clasts (Link, 1982). Overlying the upper diamictite is a bedded, locally cyclic, pink dolomite and reworked dolomite chip breccia with interstratified sandstone and argillite (Meyer et al., 2012). This “cap carbonate” (Dehler et al., 2011) exhibits a negative δ³⁴S excursion (~6 to ~3 ‰, Lorentz et al., 2004), similar to the Neo- proterozoic Noonday Dolomite of the Pahrump Group in Death Valley (Pettersen et al., 2011).

A porcellaneous 10-cm-thick reworked green tuff (667 ± 5 Ma, SHRIMP concordia age; Fanning and Link, 2004) overlies the pink dolomite–bearing unit and is overlain by limestone and argillite of the upper member of the Pocatello Formation (Link, 1983, 1987; Lorentz et al., 2004). Carbonates from this limestone interval at the top of the Scout Mountain Member record a transition to a positive δ³⁴S excursion (~5 to +5 ‰) and have been correlated with carbonates of the Johnnie Formation in Death Valley (Corsetti et al., 2007).

**Pocatello Formation on Oxford Ridge, Southern Bannock Range**

The east face of Oxford Ridge exposes the deepest structural and lowest stratigraphic levels in the Paris-Willard thrust sheet in southeast Idaho (Figs. 1 and 4), exposed in a modern Basin and Range extensional horst. Rocks of the Pocatello Formation on Oxford Ridge are metamorphosed to lower-middle greenschist facies, as evidenced by pervasive chlorite, and abundant albite and epidote in the mafic rocks. Outcrops of the Pocatello Formation contain at the base at least 200 m of pillow basalt (Fig. 5A) and hyaloclastite (Fig. 5B) of the Bannock Volcanic Member. This is overlain by up to 250 m of Scout Mountain Member (Figs. 5C, 5D, and 5E), with four stratigraphic units (Fig. 2C): a lower transitional unit of mainly diamictite with interbedded mafic volcanic rock (greenstone); a middle extrabasinal diamictite containing diverse quartzite, plutonic, and metamorphic clasts (Fig. 5E); a volcaniclastic interval (the Oxford Mountain tuffite) containing tuffaceous sandstone (Fig. 5D); and an upper sandstone. Since the base is not exposed, it is possible that the Bannock Volcanic Member is significantly thicker (Fig. 6).

The complex structure shown in Figure 6 is the result of two or more episodes of normal faulting involving the ca. 10–4 Ma Bannock detachment system (Janecke and Evans, 1999; Long et al., 2006) and modern Basin and Range extension. Several WSW-trending and gently W-dipping low-angle faults are involved in the large-offset and regionally extensive Bannock detachment system (Carney and Janecke, 2005).
Figure 4. (A) Simplified geologic map of the Cache Valley, Malad Valley, and Marsh Valley areas. New mapping of Oxford Ridge is bordered in black. Modified from Janecke et al. (2003). Inset is modified from Janecke and Evans (1999). (B) New geologic map of Oxford Ridge showing locations of geochronologic samples. CA-ID-TIMS—chemical abrasion–isotope dilution–thermal ionization mass spectrometry; LA-ICP-MS—laser ablation–inductively coupled plasma–mass spectrometry. Cross sections to section lines A–D are shown in Figure 6.
Pre-to synglacial rift-related volcanism in the Cryogenian Pocatello Formation

Figure 5. Lithologies of the Pocatello Formation on Oxford Mountain in ascending order. (A) Pillow basalt, Bannock Volcanic Member. Dashed white line marks contact between pillow basalt (below) and lobate flow (above). (B) Hyaloclastite and volcaniclastic rocks of the Bannock Volcanic Member. (C) Scoured unconformity (arrow) between extrabasinal conglomerate at base of diamicite on the Bannock Volcanic Member. (D) Oxford Mountain graded volcanic sandstone (sample 68JK09). (E) Stratified extrabasinal diamicite of the Scout Mountain Member (scale on left is 6.5 cm).
Figure 6. Cross sections A–D along Oxford Ridge from north to south. Locations are shown in Figure 4. (A) Near the northernmost extent of the Clifton fault with the Camelback Mountain Quartzite (C Zu) in its hanging wall and the underlying and folded Clifton Canyon fault. (B) Near the northernmost extent of the Oxford Ridge anticline and subsequently folded and back-tilted New Canyon, Clifton, and Clifton Canyon low-angle normal faults. (C) Back-tilted, east-dipping Clifton fault cut by the structurally higher New Canyon fault. (D) WSW-dipping and back-tilted Clifton Canyon fault intruded by a sheet-like Tertiary (?) intrusion.
They are, from bottom to top, the Clifton Canyon, Clifton, and New Canyon faults (Figs. 4 and 6). A sheet-like Tertiary mafic sill intrudes the Clifton Canyon fault, and its sheared lower contact suggests it was emplaced during faulting. Carney and Janecke (2005) interpreted that, during WSW-directed slip on the Bannock detachment system, removal of lithostatic overburden caused isostatic folding, leading to the present Oxford Ridge anticline (ORA; Figs. 4 and 6). High- to moderate-angle faults associated with Basin and Range extension include the E-dipping Oxford Basin fault and the range-bounding Dayton-Oxford fault (Figs. 4 and 6). An intervening extensional episode resulted in E-W striking cross faults such as the Gooseberry Creek fault (Fig. 4). More detailed descriptions of the regional geology and structural history of Oxford Ridge were presented by Carney and Janecke (2005), Keeley and Link (2011), and Keeley (2011).

**Bannock Volcanic Member on Oxford Ridge**

On the east face of Oxford Ridge, new mapping found two roughly 30-m-thick pillow basalt flows (Fig. 5A), separated by massive basalt flows and hyaloclastite (Fig. 5B) in the Bannock Volcanic Member. No porphyritic felsic volcanic flows (Link, 1983) or intrusions for direct radiometric dating were found. Locally, sedimentary textures on the margins of pillow basalt flows and hyaloclastite display soft-sediment deformation. Siltstone and sandstone injections, and basalt inclusions in soft sediment are evidence of peperite deposition and/or hypabyssal intrusion of basalt.

Just below the Clifton Canyon fault north of Davis Basin in section 18, T14S, R38E, extrabasinal conglomerate and diamictite of the Scout Mountain Member lie above a locally scoured unconformity on volcaniclastic rocks of the Bannock Volcanic Member (agglomerate facies of Link, 1982) (Fig. 5C). Further south, along Clifton Road (Fig. 7A), diamictite is interbedded with hyaloclastite and basalt flows in the transitional unit at the base of the Scout Mountain Member (Keeley and Link, 2011).

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**Figure 7.** Stratigraphic sections of the Pocatello Formation on Oxford Ridge. (A) Section exposed at Clifton Basin Road composed of the lower transitional unit with intercalated metabasalt and diamictite. (B) Section north of Clifton Basin showing diamictite and Oxford Mountain tuffite. (C) Section in the northern portion of the Clifton Quadrangle. (D) Section at the southern end of the cliffs above Oxford Basin. BVM—Bannock Volcanic Member; SMM—Scout Mountain Member.
Scout Mountain Member on Oxford Ridge

New mappable, intertonguing stratigraphic units have been identified in the Scout Mountain Member on Oxford Mountain (Fig. 7). The lower “transitional unit” of Link (1982) consists of 70 m of green-gray, massive to crudely-bedded quartzite diamictite intercalated with metabasalt flows and hyaloclastite units up to 10 m thick, with reworked basalt pillows (Fig. 7A). Vescular and nonvesicular basaltic clasts up to cobble size make up 80% of the clasts in the diamictite (Link, 1982). The remaining clasts are dominantly argillite with quartzite and sparse basement lithologies. Fine grained, light-colored, aphanitic clasts may represent a subordinate felsic volcanic or volcaniclastic component. The diversity of clasts from the Scout Mountain Member diamictite on Oxford Mountain suggests a mixture of intrabasinal and extrabasinal components. Locally, a gray quartzite granule conglomerate is discontinuously interbedded in this unit.

A 150–190-m-thick, brown-green to purple and locally sandy, massive to stratified, inferred glaciogenic diamictite, with a variety of extrabasinal clasts up to boulder size, overlies the transitional unit. Clast lithologies in the diamictite include porphyritic felsic volcanic rocks, basalt, chloritic argillite, plus granite and gneiss. This unit progressively to the east (Fig. 6). Facies and in gradational contact with the diamictite on Oxford Ridge is the Oxford Mountain tuffite (tuffite is a general term for a rock unit that contains a mixture of pyroclastic and epiclastic material; Schmid, 1981). The Oxford Mountain tuffite consists of 40–60 m of debris-flow diamictite and feldspathic volcaniclastic sandstone (Figs. 7B–7D). It is exposed for 5.5 km along strike parallel to and in the upper plate of the low-angle Clifton Canyon fault, which cuts out the top of the unit progressively to the east (Fig. 6). Facies within the Oxford Mountain tuffite include: (1) mafic volcaniclastic diamictite, with abundant pebble- to cobble-sized vesicular and nonvesicular basalt clasts, generally found at the base (Fig. 8A); (2) nonstratified volcaniclastic diamictite that includes rounded pebbles to rare boulders of gray to purple porphyritic trachyandesite, trachyte, dacite, and rhyolite (Fig. 8B, 15JK10); (3) one or more reworked plagioclase-phycric volcanic lithic wacke beds (up to 1 m thick) with soft-sediment deformation (Fig. 8C, 13JK10); (4) thin- to medium-bedded, graded volcanic sandstone (68JK09) intercalated with nonstratified volcanic diamictite (67JK09) (Fig. 5D); and (5) crudely stratified volcanic diamictite bearing trachyandesite and rhyolite clasts (Fig. 8D), including angular to subrounded, white to dark-gray aphanitic clasts. The individual tuffite beds contain angular to round felsic volcanic clasts, some of which have undergone negligible transport from the volcanic vent. The stratified volcaniclastic diamictite is interbedded with multiple beds having thin to thick, fine-grained green wavy laminations (04JK09, 23JK10, and 16JK10, Fig. 8E). Overlying the tuffite, there is normally graded breccia and conglomerate, containing argillite and sandstone clasts, and green feldspathic sandstone (Fig. 8F). Oxford Mountain does not expose Neoproterozoic carbonate or dark phyllicit strata as seen in the upper Scout Mountain Member at Portneuf Narrows.

Previous to the mapping in this study, the volcaniclastic unit (now the Oxford Mountain tuffite) examined by Fanning and Link (2004) to have a SHRIMP weighted mean 206Pb/238U zircon age of 708 ± 5 Ma. In total, 18 zircon grains were used in the calculation (n = 18) with a mean square weighted deviation (MSWD) of 1.7. Condon and Bowring (2011) suggested that given the sample size (average relative to SHRIMP standards) and the relative high MSWD of 1.7, the scatter in the data set was possibly due to real age variation in the sample rather than analytical error. Fanning and Link (2004) noted that there were both clear euhedral and possibly reworked zircons.

Further analysis on samples from the same stratigraphic level using CA-ID-TIMS was conducted by Condon and Bowring (2011). Condon and Bowring (2011) analyzed three zircon grains each from samples 06PL00 and 34PL05 (Fig. 4). Of the six total grains analyzed, four grains were equivalent at ca. 687 Ma, one was slightly younger at ca. 684 Ma, and one was older at ca. 705 Ma. Condon and Bowring (2011) reported a weighted mean ID-TIMS 206Pb/238U age of 687.4 ± 1.3 Ma (n = 4). They concluded that the lower precision of SHRIMP U-Pb dates does not offer the analytical precision needed to discern subpopulations at the <3% level, and confidently assign depositional ages. These results prompted a second SHRIMP analysis of the same sample (06PL00), which yielded a revised zircon age of 686 ± 4 Ma (Fanning and Link, 2008). At the time, structural complexities on Oxford Ridge made the exact stratigraphic position of the unit uncertain.

Several workers (Hoffman and Li, 2009; MacDonald et al., 2010; Petterson et al., 2011) have suggested that the volcaniclastic unit is not in contact with glacial strata and have questioned the reliability of the age constraints. The present paper presents the results of two seasons of geologic mapping aimed at resolving the age and stratigraphic relations between the Bannock Volcanic and Scout Mountain Members of the Pocatello Formation both on Oxford Ridge and at the type section south of Portneuf Narrows near Pocatello, Idaho. We address the age of the Neoproterozoic Pocatello Formation with new SHRIMP and CA-ID-TIMS U-Pb zircon ages.

Analytical Methods

Samples from the Pocatello Formation both at Oxford Mountain and at the type area south of Portneuf Narrows, were selected for U-Pb geochronology to (1) determine detrital zircon provenance ages (data presented in Keeley and Link, 2011; GSA Data Repository Table DR1) and (2) place constraints on the depositional ages of the succession. Zircon crystals were extracted from samples by traditional methods of crushing and grinding at Idaho State University, followed by separation with a Wilfley table, heavy fluids, and a Frantz magnetic separator at Boise State University. Samples were processed such that all zircons were retained in the final heavy mineral fraction.

Initial detrital zircon provenance analysis and screening were done at the Arizona Laser-Chron Center following Gehrels et al. (2006, 2008). Here, the grain separates were incorporated into a 1” (2.54 cm) epoxy mount together with fragments of the Sri Lanka standard zircon. The mounts were sanded down to a depth of ~20 µm, polished, imaged, and cleaned prior to isotopic analysis. U-Pb geochronologic analysis of zircons was conducted by LA-MC-ICP-MS. The analyses involved ablation of zircon with a New Wave UP193HE Excimer laser using a spot diameter of 30 µm. The ablated material was carried in helium into the plasma source of a Nu HR ICP-MS, which was equipped with a flight tube of sufficient width that U, Th, and Pb isotopes could be measured simultaneously. The ablation pit was ~15 µm in diameter.
Figure 8. Lithologies of the Oxford Mountain tuffite. (A) Mafic volcaniclastic diamictite. (B) Nonstratified volcaniclastic diamictite with trachyte epi-clast (15JK10). (C) Stratified plagioclase-phyric volcanic sandstone and diamictite exhibiting slump folds (13JK10). Dashed white lines mark folded planar beds. (D) Crudely stratified and tectonically stretched volcaniclastic diamictite bearing light-colored, aphanitic, volcanic clasts. (E) Volcaniclastic diamictite with fine-grained green wavy laminations (04JK09, 23JK10, and 16JK10). (F) Above the tuffite lies normally graded breccia, conglomerate, and feldspathic sandstone.
depth. Detrital zircon spectra were created with Isoplot (Ludwig, 2008).

Further analysis of tuffaceous units was conducted using SHRIMP. U-Pb zircon analyses were carried out at Research School of Earth Sciences, Australian National University, Canberra, Australia, using SHRIMP II following procedures described in Williams (1998) and references therein. Zircon grains from volcanoclastic diamictite samples were handpicked from heavy mineral concentrates and placed onto double-sided tape, together with Temora reference zircon grains. Epoxy disks were cast, and the zircon grains were sectioned approximately in half, and polished. Transmitted and reflected light photomicrographs and cathodoluminescence (CL) images were made for all grains (Fig. DR1 [see footnote 1]). Four samples (4JK09, 62JK09, 63JK09, and 64JK09) were analyzed in a single extended probe session. Uncertainty in Temora reference zircon U/Pb ratio calibration was 0.39%. Sample 15PL08 was run in a separate session, where uncertainty in Temora reference zircon U/Pb ratio calibration was 0.58%. Data were processed using the SQUID Excel macro of Ludwig (2000), and U-Pb ages of ~330 detrital zircon grains from seven samples of the Scout Mountain Member along Oxford Ridge were analyzed by LA-MC-ICP-MS. Data are listed in GSA Data Repository Table DR1 (see footnote 1). The detrital zircon spectra (Fig. 9) exhibit a dominant 1.6–1.7 Ga Paleoproterozoic-aged peak with locally important Archean, Mesoproterozoic, and Neoproterozoic peaks. Populations include a Paleoproterozoic to Neoproterozoic grouping at 2450–2700 Ma, Mesoproterozoic 1400–1500 Ma grains, and locally common Neoproterozoic plucked from the epoxy, and then chemically abraded following a modified version of Mattinson (2005). The residual crystals were washed, spiked with an EARTHTIME mixed 205Pb–205Pb–206Pb–207Pb–208Pb tracer solution (ET2535) and dissolved. U and Pb were separated from zircon solutions following Krogh (1973), loaded onto a Re filament after Gerstenberger and Haase (1997), and measured using a GVI (IsotopX) Isoprobe-T MC-TIMS.

GEochRONOLOGY RESULTS

Detrital Zircon Provenance on Oxford Mountain

U-Pb ages of ~330 detrital zircon grains from the four samples were analyzed for lutetium and hafnium isotopes using MC-ICP-MS (multicollector–inductively coupled plasma–mass spectrometer) at the Australian National University using procedures as described in Münziga et al. (2008). Two-stage crustal model ages were calculated using the $^{176}Lu$ decay constant of Söderlund et al. (2004), model chondritic values of Bouvier et al. (2008), present-day depleted mantle values of Vervoort and Blichert-Toft (1999), and average crustal values of Goode and Vervoort (2006).

Fifty zircons with determined ca. 700 Ma U-Pb ages from the four samples were analyzed for lutetium and hafnium isotopes using MC-ICP-MS (multicollector–inductively coupled plasma–mass spectrometer) at the Australian National University using procedures as described in Münziga et al. (2008). Two-stage crustal model ages were calculated using the $^{176}Lu$ decay constant of Söderlund et al. (2004), model chondritic values of Bouvier et al. (2008), present-day depleted mantle values of Vervoort and Blichert-Toft (1999), and average crustal values of Goode and Vervoort (2006).
ca. 650–720 Ma grains. Grains of Grenville age (950–1300 Ma) are rare and do not form a significant detrital zircon age grouping. Positive Kolmogorov-Smirnov (K-S) tests (Press et al., 1986) on two diamictite samples (75JK09 and 73JK09), one volcanic sandstone sample (68JK09), and one quartzite clast (74JK09) between the Fivemile Canyon area and Oxford Ridge suggest all four samples are not statistically different.

Maximum Age Constraints: Oxford Mountain

SHRIMP

Sample 04JK09 is from the Oxford Mountain tuffite unit resampled from the same stratigraphic unit as 06PL00 of Fanning and Link (2004) and 34PL05 of Condon and Bowring (2011). The rock is a medium-bedded plagioclase-rich volcaniclastic diamictite with gray and green, 2-cm-thick, thinly laminated siltstone beds (Fig. 8E). Subangular clasts include basalt cobbles, aphanitic gray, trachytic volcanics, and quartzite. Zircons in sample 04JK09 are generally small and fragmented. Twenty grains were analyzed using SHRIMP II (Table 1); one grain recorded a 207Pb/206Pb age of ca. 2445 Ma, while another high-U zircon is discordant with a Tertiary 207Pb/206U age. The remaining 18 areas analyzed are predominantly within uncertainty of the concordia curve. A probability density plot of 206Pb/238U ages shows a prominent peak at ca. 680–710 Ma with scattered older ages extending to 820 Ma (Fig. 10).

CA-ID-TIMS

The range of concordant SHRIMP 206Pb/238U ages for zircons from the Oxford Mountain tuffite suggests complex, multi-age components. Thus, CA-ID-TIMS analyses (approaching 0.1% precision on these samples) were made on five new samples along 5.5 km of strike (Fig. 4B) in order to isolate precise single grain ages and so determine the age of the youngest zircon age grouping. Five to 12 euhedral grains were selected on the basis of CL images from grain mounts of samples 04JK09, 23JK10, 13JK10, 15JK10, and 16JK10. Data are presented in Figures 11 and in Table 2.

The zircon grains from these samples are 50–200 µm in length, equant to subprismatic, and euhedral to subrounded. Cathodoluminescence (CL) images from the samples show sector zoning and rare narrow to broad oscillatory zoning patterns, suggesting that all grains are magmatic. Sector zoning is the dominant pattern for samples 04JK09 and 23JK10, whereas samples 13JK10, 15JK10, and 16JK10 exhibit more variation. Rare subrounded grains suggest minor transport. Despite our best efforts at picking first-cycle magmatic grains, one equant grain was Archean and one was Mesoproterozoic. Individual crystal age uncertainties are generally between 0.5 and 1 Ma.

A resample of 04JK09 (Fig. 8E; resample of 06PL00 of Fanning and Link, 2004) yielded three age groups at 703, 699, and 694 Ma, confirming the multimodal, detrital character of the Neoproterozoic crystal load of the volcaniclastic diamictite. A similar conclusion was made by Condon and Bowring (2011) and alluded to in Fanning and Link (2004). Sample 23JK10 of volcaniclastic diamictite, 20 m to the north along strike of sample 04JK09, has subrounded basalt and felsic volcanic clasts with several closely spaced fine-grained wavy laminations (Fig. 8E). Analysis of this sample yielded ages of 697 and 700 Ma, as well as a Mesoproterozoic grain.

Sample 13JK10, located ~3 km further north, is composed of plagioclase-phryic, laminated, volcanic sandstone with floating subangular to subrounded pebbles (Fig. 5C). The poorly sorted rock contains abundant subangular, light-colored, aphanitic clasts with carbonate-chlorite alteration on rims. Rounded epipelitic trachyte and basalt lithic clasts are also present. Laminations exhibit soft sediment deformation and are depressed and punctuated by basaltic and trachytic episclasts. This sample yielded crystal populations with diverse ages of 707, 703, 693, 696, and 688 Ma (Table 2; Fig. 11); and one Grenville-age grain.

Sample 15JK10, further north, is composed of volcaniclastic diamictite (Fig. 8B). Clasts within this diamictite are trachyte, basalt, feldspathic lithic wacke and quartz arenite. This sample yielded zircons ranging from 709 to 696 Ma (Fig. 11).

The northernmost sample, 16JK10, lies just below the Clifton fault and is lithographically higher than sample 62JK09. The probability density plot shows that zircon ages form a simple bell-shaped distribution (Fig. DR3 [see footnote 1]). However, in light of the multimodal age populations shown by CA-ID-TIMS data on other such zircon populations from Oxford Mountain, it is likely that any weighted mean calculation from SHRIMP data would encompass a range of actual zircon crystallization events.

Sample 63JK09 is a similarly intraclastic and mixed volcaniclastic diamictite 100 m stratigraphically higher than sample 62JK09. The probability density plot shows an even distribution of zircon ages between 728 Ma and 684 Ma, with one outlying grain dated at ca. 786 Ma (Fig. DR3 [see footnote 1]). The SHRIMP concordia age, and 206Pb/238U age, of the youngest grain is younger than, but within the SHRIMP analytical uncertainty, of the youngest grain in the underlying sample (Table 1; Fig. DR3 [see footnote 1]).

Immediately above the lower diamictite, a plagioclase-arkose (15PL08) yields Paleoproterozoic and Neoproterozoic detrital zircon populations. The Neoproterozoic population ranges in age from 700 to 662 Ma, excluding two discordant analyses as old as 774 Ma. The probability density plot shows that zircon ages form a simple bell-shaped distribution (Fig. DR3 [see footnote 1]). The SHRIMP concordia age, and 206Pb/238U age, of the youngest grain is younger than, but within the SHRIMP analytical uncertainty, of the youngest grain in the underlying sample (Table 1; Fig. DR3 [see footnote 1]).

Moving up section, a prominent cobble conglomerate lies below the upper diamictite south of Portneuf Narrows. Sample 64JK09 is a coarse of several felsic volcanic clasts from this conglomerate that exhibits flow-banding and possible fiamme. Most of the grains are euhedral and exhibit rare resorption in CL (Fig. DR1 [see footnote 1]). Several subrounded grains and one Paleoproterozoic grain suggest some reworking during eruption and/or deposition. The volcanic
| Grain | U | Pb | Th | Pb/Th | Th/U | U | 235U | 238U | 206Pb | 207Pb | 208Pb | 209Pb | 210Pb | 212Pb | 214Pb | 232Th | 238U | 235U |
|-------|---|----|----|-------|------|---|------|------|-------|-------|-------|-------|-------|-------|-------|-------|-------|
| 62JK09 | 13.1 | 61 | 36 | 0.60 | 6.0 0.000202 | 0.36 | 0.01 | 8.710 | 0.124 | 0.0617 | 0.0011 | 0.0058 | 0.00005 | 0.000005 | 0.000005 | 0.000005 |
| 63JK09 | 10.1 | 7.1 | 3.8 | 0.50 | 4.6 | 0.000221 | 0.39 | 0.13 | 8.678 | 0.132 | 0.0678 | 0.0013 | 0.0046 | 0.000046 | 0.0000046 | 0.0000046 | 0.0000046 |
| 11.1 | 16 | 22 | 13 | 0.60 | 2.0 | 0.000092 | 0.16 | 0.16 | 8.814 | 0.173 | 0.0629 | 0.0016 | 0.0050 | 0.000050 | 0.0000050 | 0.00000050 | 0.000000050 |
| 23.1 | 66 | 40 | 0.60 | 7 | 0.000236 | 0.41 | 0.19 | 8.628 | 0.119 | 0.0643 | 0.0010 | 0.0045 | 0.000045 | 0.0000045 | 0.00000045 | 0.000000045 |
| 5.1 | 46 | 25 | 0.55 | 4.6 | 0.000275 | 0.48 | 0.12 | 8.689 | 0.126 | 0.0624 | 0.0011 | 0.0046 | 0.000046 | 0.0000046 | 0.00000046 | 0.000000046 |
| 15.1 | 127 | 59 | 0.47 | 12.5 | 0.000152 | 0.27 | 0.06 | 8.708 | 0.104 | 0.0612 | 0.0007 | 0.0029 | 0.000029 | 0.0000029 | 0.00000029 | 0.000000029 |
| 6.1 | 31 | 17 | 10 | 0.40 | 3 | 0.000019 | 0.03 | 0.19 | 8.760 | 0.128 | 0.0624 | 0.0011 | 0.0046 | 0.000046 | 0.0000046 | 0.00000046 | 0.000000046 |
| 21.1 | 46 | 27 | 15 | 0.60 | 4.2 | 0.000242 | 0.42 | 0.11 | 8.515 | 0.123 | 0.0678 | 0.0012 | 0.0050 | 0.000050 | 0.0000050 | 0.00000050 | 0.000000050 |
| 11.1 | 162 | 88 | 0.54 | 16 | 0.000142 | 0.25 | 0.05 | 8.856 | 0.102 | 0.0639 | 0.0006 | 0.0024 | 0.000024 | 0.0000024 | 0.00000024 | 0.000000024 |
| 16.1 | 22 | 13 | 0.60 | 2 | 0.000092 | 0.16 | 0.16 | 8.814 | 0.173 | 0.0629 | 0.0016 | 0.0050 | 0.000050 | 0.0000050 | 0.00000050 | 0.000000050 |
| 7.1 | 38 | 22 | 0.58 | 4 | 0.000221 | 0.39 | 0.13 | 8.678 | 0.132 | 0.0678 | 0.0013 | 0.0046 | 0.000046 | 0.0000046 | 0.00000046 | 0.000000046 |
| 23.1 | 66 | 40 | 0.60 | 7 | 0.000236 | 0.41 | 0.19 | 8.628 | 0.119 | 0.0643 | 0.0010 | 0.0045 | 0.000045 | 0.0000045 | 0.00000045 | 0.000000045 |
| 5.1 | 46 | 25 | 0.55 | 4.6 | 0.000275 | 0.48 | 0.12 | 8.689 | 0.126 | 0.0624 | 0.0011 | 0.0046 | 0.000046 | 0.0000046 | 0.00000046 | 0.000000046 |
| 13.1 | 61 | 36 | 0.60 | 6.0 | 0.000202 | 0.36 | 0.01 | 8.710 | 0.124 | 0.0617 | 0.0011 | 0.0058 | 0.000058 | 0.0000058 | 0.00000058 | 0.000000058 |
| 5.1 | 46 | 25 | 0.55 | 4.6 | 0.000275 | 0.48 | 0.12 | 8.689 | 0.126 | 0.0624 | 0.0011 | 0.0046 | 0.000046 | 0.0000046 | 0.00000046 | 0.000000046 |
| 6.1 | 31 | 17 | 10 | 0.40 | 3 | 0.000019 | 0.03 | 0.19 | 8.760 | 0.128 | 0.0624 | 0.0011 | 0.0046 | 0.000046 | 0.0000046 | 0.00000046 | 0.000000046 |
| 21.1 | 46 | 27 | 15 | 0.60 | 4.2 | 0.000242 | 0.42 | 0.11 | 8.515 | 0.123 | 0.0678 | 0.0012 | 0.0050 | 0.000050 | 0.0000050 | 0.00000050 | 0.000000050 |
| 11.1 | 162 | 88 | 0.54 | 16 | 0.000142 | 0.25 | 0.05 | 8.856 | 0.102 | 0.0639 | 0.0006 | 0.0024 | 0.000024 | 0.0000024 | 0.00000024 | 0.000000024 |
TABLE 1. SUMMARY OF SHRIMP U-Pb ZIRCON RESULTS (Continued)

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Note: SHRIMP data is high resolution ion microprobe. Uncertainties are given at 1σ level. Error in Tera Wairunga common Pb calibration was 0.2% for all analytical sessions, except for sample 64JK09.

* f206 % denotes the percentage of 206Pb that is common Pb.
† For % disc, 0% denotes a discordant analysis.

Pre-to-synglacial rift-related volcanism in the Cryogenian Pocatello Formation
Figure 10. Normalized probability density plots for sensitive high-resolution ion microprobe (SHRIMP), chemical abrasion–isotope dilution–thermal ionization mass spectrometry (CA-ID-TIMS), and ID-TIMS analyses of the Oxford Mountain tuffite originally sampled by Fanning and Link (2004). Samples 06PL00 and 04JK09, ~200 m along strike south of 06PL00, were analyzed first by SHRIMP. SHRIMP histograms are plotted on the inverted y-axis. Sample 04JK09 was reanalyzed by CA-ID-TIMS (this study). *ID-TIMS age reported by Condon and Bowring (2011) from samples 34PL05 (25 m away from 04JK09) and 06PL00. †New CA-ID-TIMS age from sample 16JK10 (not plotted), this study. See sample locations in Figure 4 and Table DR2 (see text footnote 1).

Figure 11. Compilation diagram showing the stratigraphic location of all sensitive high-resolution ion microprobe (SHRIMP) and chemical abrasion–isotope dilution–thermal ionization mass spectrometry (CA-ID-TIMS) geochronologic samples and their single-grain $^{206}$Pb/$^{238}$U analyses. SHRIMP and CA-ID-TIMS error bars are 2σ. Grain analyses shown by black error bars were used in concordia age calculations (SHRIMP) and weighted mean age calculations (CA-ID-TIMS). Grain analyses shown by gray-filled bars were not used in age calculations. Data are listed in Tables 1 and 2. See sample locations in Figure 4 and Table DR2 (see text footnote 1).
Table 2. Summary of CA-ID-TIMS U-Th-Pb isotopic data

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Note: CA-TIMS—chemical abrasion-thermal ionization mass spectrometry.

1a—z2, etc., are single zircon grains or fragments extracted from single grain mounts, all annealed and chemically abraded after Mattinson (2005). b—Model Th/U ratio calculated from radiogenic 206Pb/238U ratio and 207Pb/235U age. c—Pb* and Pb c represent radiogenic and common Pb, respectively; mol % 206Pb* with respect to radiogenic, blank, and initial common Pb. d—Measured ratio corrected for spike and fractionation only. Pb and U fractionation were corrected internally using the double spike composition. e—Corrected for fractionation, spike, and common Pb. Pb and U fractionation were assumed to be procedural blank: 206Pb/204Pb = 18.35 ± 1.5%; 207Pb/204Pb = 11.97 ± 0.5%; 208Pb/204Pb = 38.08 ± 1.0% (all uncertainties 2 σ).

Pre-to synglacial rift-related volcanism in the Cryogenian Pocatello Formation

Research
texture and range of ages (704–673 Ma) support incorporation of xenocrystic zircons during eruption and deposition. All 18 Neoproterozoic grains were used in the age calculation to yield a weighted mean \(^{206}\text{Pb}/^{238}\text{U}\) age of 691 ± 4 Ma (n = 18, MSWD = 0.98), assuming a single eruptive age on the volcanic clast. Therefore, 691 ± 4 Ma is interpreted as a maximum depositional age for the cobble conglomerate.

**Lutetium-Hafnium Isotopic Analysis**

Lutetium-hafnium isotopic analyses for 50 ca. 700 Ma zircon grains from samples 4JK09, 62JK09, 63JK09, and 64JK09 yield \(\varepsilon_{\text{Hf}}\) values that range from +2 to −17 (Fig. 12; Table 3). The volcaniclastic diamictite sample (04JK09), the Oxford Mountain tuffite, and lower diamictite samples from Portneuf Narrows (62JK09 and 63JK09) show considerable variation due to xenocrystic and detrital mixing, whereas the 691 Ma volcanic clast, 64JK09, shows a narrow xenocrystic and detrital mixing, whereas the 63JK09 show considerable variation due to xenocrystic and detrital mixing, whereas the 691 Ma volcanic clast, 64JK09, shows a narrow range of ages (704–673 Ma) supporting the local incorporation of xenocrystic zircons during eruption and deposition. All 18 Neoproterozoic grains were used in the age calculation to yield a weighted mean \(^{206}\text{Pb}/^{238}\text{U}\) age of 691 ± 4 Ma (n = 18, MSWD = 0.98), assuming a single eruptive age on the volcanic clast. Therefore, 691 ± 4 Ma is interpreted as a maximum depositional age for the cobble conglomerate.

**DISCUSSION AND INTERPRETATION**

**Pocatello Formation at Oxford Mountain**

The gradational upper contact of the Bannock Volcanic Member and the intercalated metabasalt flows at the base of the Scout Mountain Member provide evidence that diamicite deposition was synchronous with terminal basaltic volcanism and suggest that basaltic volcanism was temporally related to silicic volcanism represented by felsic volcanic clasts and the Oxford Mountain tuffite.

The diamicite below the tuffite may have been deposited under glaciomarine conditions, but no direct evidence was noted in this study. Noting the variation in provenance between intrabasinal and extrabasinal diamictites may be the key in separating exclusively rift-related processes from glacial processes. The local increases of mafic clast components within the diamicite are interpreted to indicate uplift and reworking of the underlying Bannock Volcanic Member, whereas increases in basement quartzose clasts suggest glacial transport inland of the rifted margin. Facies associations between the two major facies types in this unit support a glacial origin. The stratigraphic columns (Figs. 7B–7D) generally begin with a thick section of unstratified diamictite topped by thinner, weakly...
stratified units, as shown in Figure 8E (04JK09, 16JK10, and 23JK10). We interpret this facies change to mark the transition from proximal-medial to medial-distal glacial deposition. We interpret the thin green wavy laminations (Fig. 8E) to represent tuffaceous horizons deposited by suspension settling through the water column during a time of relative quiescence and possible ice rafting. This period is followed by the deposition of the upper Oxford Mountain tuffite (sample 16JK10).

The detrital zircon provenance analysis on the diamictite and volcanic sandstone on Oxford Ridge shows a “mixed Laurentian” signature (i.e., a sedimentary provenance with ages matching known ages of composite basement provinces of Laurentia). The prominent 1.6–1.7 Ga Paleoproterozoic peak (Fig. 9) suggests that the dominant source of detritus was the composite 1.6–1.9 Ga Mojave-Yavapai-Mazatzal Province (Stewart et al., 2001). The granitic clast (24JK09) with a Paleoproterozoic to Neoarchean age peak (2450–2700 Ma) suggests that the granitic component in the basal conglomerate and diamictite came from the underlying Farmington Canyon Complex, the Wyoming Province, and/or the Grouse Creek block (Foster et al., 2006; Mueller et al., 2011). The 1.3–1.5 Ga peak is interpreted to be from the midcontinent Granite-Rhyolite Province (Goodge and Vervoort, 2006, and references therein). The small group of grains between 0.9 and 1.2 Ga may have come from the Grenville orogen from eastern or southeastern Laurentia (Dickinson et al., 2010), perhaps transported by a transcontinental river system (Rainbird et al., 1992, 2012). Successful K-S tests on diamictite, sandstone, and quartzite clast samples (73JK09, 74JK09, 75JK09, and 68JK09) suggest that the zircon spectra are statistically indistinguishable. We interpret the Scout Mountain Member units on Oxford Ridge to have a common provenance and to have been deposited in a single basin. The provenance of the quartzite cobbles (74JK09 and 03JK09) suggests that some of the zircon grains in the diamictite and sandstone may have been recycled out of the “mixed Laurentian” Uinta Mountain Group and Big Cottonwood Formation to the southeast (Dehler et al., 2010).

The span of concordant SHRIMP ages from the Oxford Mountain tuffite (04JK09) indicates that the zircons within the sample did not crystallize during a single volcanic eruption. One discordant, high-U zircon with a Tertiary age is interpreted as either contamination or one that was reset during Mesozoic thrusting. The data suggest recurrent Neoproterozoic felsic volcanism, with other extraneous detrital zircons also present. The ~1% (1σ) analytical uncertainty of the SHRIMP analyses is not precise enough to deconvolve the range of grain ages present in these samples. The SHRIMP results do, however, overlap with CA-ID-TIMS data. Both the 709 Ma and the 686 Ma age groupings reported by Fanning and Link (2004, 2008) and Condon and Bowring (2011), and the 700 ± 5 Ma grouping reported by Kool (2010) are within the bounds of the bell-shaped portion of lower 206Pb/238U ages in Figure 10.

The clast compositions within samples 13JK10 and 15JK10 demonstrate a mixed sedimentary and volcanic provenance, potentially reworking the Pocatello Formation. The sedimentary textures of these samples are interpreted to represent mixing of pyroclastic and epiclastic components in a subaqueous debris flow that was medial-proximal to the volcanic source. The Oxford Mountain tuffite itself is interpreted as a marine volcanoclastic deposit that was proximal to medial to its volcanic source, and redeposited by sediment gravity flows. This proximal to medial interpretation is based on its (1) abundant plagioclase crystals, (2) rare angular volcanic pyroclasts, (3) round and angular volcanic epiclasts that depress laminae in underlying volcanic siltstone, and (4) consistent, albeit mixed, zircon age populations from 705 to 685 Ma. It is unclear whether these basinal clasts are primary (e.g., volcanic bombs) or have been rounded during aqueous transport. Based on these relationships, we interpret the youngest zircon age grouping (685 Ma) from the Oxford Mountain tuffite to be close to its depositional age.

**Pocatello Formation at Portneuf Narrows**

SHRIMP concordia ages from two mixed volcaniclastic diamictites and one sandstone from the Portneuf Narrows area (Fig. 2), as well as the youngest single-grain SHRIMP analyses from each sample, young upward (Table 1), suggesting syntepositional volcanism. As the underlying assumption of a weighted mean calculation requires a single age population, the SHRIMP concordia ages from detrital samples (62JK09, 63JK09, and 15PL08) are only tentative. The youngest single-grain ages from samples 62JK09, 63JK09, and 15PL08 are 689 ± 8 Ma, 684 ± 8 Ma, and 662 ± 16 Ma (or 676 ± 8 Ma), respectively (Table 1). However, a minimum of three zircon grains with identical ages is required to be considered a single population, and only two overlapping populations of 699 Ma and 700 Ma occur in samples 62JK09 and 63JK09. Due to the insufficient analytical precision of SHRIMP data (Fig. 11), attempts to deconvolve the complex zircon populations are not considered to be viable. Therefore, the youngest single-grain SHRIMP analyses are also only tentatively mentioned to call attention to the apparent younging-upward analyses. However, this pattern may be coincidental as it may be affected by sample size, the proportions of different age-groupings in the samples, and selection-related bias of grains.

Three felsic porphyritic volcanic cobbles from the overlying cobble conglomerate member and upper diamictite yield ages of 691 ± 4 Ma, 701 ± 4 Ma, and 717 ± 4 Ma (Fanning and Link, 2004; 2008; this study), demonstrating that the period of silicic volcanic activity lasted at least 25 m.y. and up to 32 m.y., considering the 685 Ma age of the Oxford Mountain tuffite.

Maximum depositional ages of the upper diamictite and cap carbonate succession are permissively younger (youngest population 672 ± 6.3 Ma [FHDZ1] and 665 ± 5.7 Ma [FHDZ2]; LA-MC-ICP-MS ages; Dehler et al., 2009, 2011). A 667 Ma reworked tuff overlies the cap carbonate succession (Fanning and Link, 2004).
between the cobble conglomerate at Portneuf Narrows and the slightly younger 685 Ma Oxford Mountain tuffite (Fig. 11). Below this line of correlation, the age components in the lower diamictite of the Scout Mountain Member at Portneuf Narrows are comparable to the age range of 709–685 Ma from the volcanioclastic rocks of Oxford Mountain, suggesting a similar volcanic provenance. The lack of ≤667 Ma zircons in the lower diamictite at Portneuf Narrows supports a lithostratigraphic correlation to the diamictite on Oxford Mountain. Such grains have only been documented above the lower diamictite (three such grains in 15PL08) at Portneuf Narrows.

Regional Correlations

The CA-ID-TIMS maximum depositional age from the Oxford Mountain section overlaps with SHRIMP U-Pb ages of ca. 685 Ma obtained from the Edwardsburg Formation 415 km to the northwest in central Idaho (Lund et al., 2003). The 667 Ma maximum depositional age of the reworked tuff above the pink dolomite at Portneuf Narrows correlates well with the recent ca. 664 Ma U-Pb zircon ages from the Tuff of Daugherty Gulch and the syenite-diorite suite of Acorn Butte (Lund et al., 2010). The maximum depositional age from the Pocatello Formation is consistent with magmatic quiescence in central Idaho between ca. 650 Ma and ca. 500 Ma suggested by Lund et al. (2010). The new SHRIMP data from the Portneuf Narrows section also overlap with lower and upper SHRIMP ages of 708 ± 5 Ma and 671 ± 5 Ma from the formation of Perry Canyon in northern Utah (Balgord et al., 2010; Balgord, 2011). The new CA-ID-TIMS age of 685 Ma is also consistent with the 688 +9.5/-6.2 Ma U-Pb zircon age from felsic volcanioclastic rocks from Gataga Mountain, northern British Colombia (Ferri et al., 1999). However, the Idaho and Utah ages are ~25 m.y. younger than the older 716 Ma age constraints from the Mount Harper Group (Rapitan Formation) of the northern Canadian cordillera (Macdonald et al., 2010), implying diachronous Cordilleran rifting and diachronous middle Cryogenian (Sturtian) glacial episodes, as discussed next.

Correlation to the Sturtian Glaciation

The new ages from the Oxford Mountain tuffite place a maximum age constraint of 685 Ma above the diamictite on Oxford Ridge. The ages presented in this paper suggest that the lower diamictite at Portneuf Narrows and the diamictite on Oxford Ridge and the upper diamictite at Portneuf Narrows correlate with a late phase of the 715–660 Ma middle Cryogenian “Sturtian” glaciations described by Hoffman and Li (2009).

None of the tuffaceous strata sampled thus far from the Scout Mountain Member represents primary air-fall tuffs or volcanic flows. Furthermore, the “cap carbonate” above the upper diamictite has been described by Corsetti and Lorentz (2006) as having “Marinoan-style” characteristics (similar to those dated at 635 Ma in South China and Namibia). Therefore, several workers (Macdonald et al., 2010; Pettersen et al., 2011) have suggested the possibility that the 685 Ma (and 667 Ma) zircons represent detritus deposited into strata tens of millions of years later than eruption, during, for example, the 635 Ma Marinoan glaciations.

The tuffaceous rocks are, however, proximal and little-reworked, volcanioclastic deposits. Their deposition 50 m.y. after eruption is unlikely, based on the poor preservation potential of silicic volcanoes after kilometer-scale uplift and denudation of volcanic rifted margins (Bryan et al., 2002). The age range of the Neoproterozoic population and the ages of the subpopulations within it (709, 702, 700, 695, 690, and 685 Ma) are demonstrated in Table 2. Considering the oldest 717 Ma volcanic clast (Fanning and Link, 2004), volcanism in SE Idaho may have lasted 20–32 m.y. A long-lived volcanic center, or numerous episodic ones, would be expected to develop some topographic relief and be incised, leading to the delivery of synrift, juvenile, volcanic debris to the depositional system. This would support our preferred hypothesis involving syndepositional middle Cryogenian volcanism.

Alternatively, after uplift and erosion, the roots of the volcanic center may have been exposed (Bryan et al., 2002) and therefore been subject to reworking, possibly resulting in the multimodal xenocrystic zircon population documented in this paper. This scenario would be consistent with the Petterson et al. (2011) claim that the Scout Mountain diamictites are late Cryogenian (Marinoan). However, the diamictites are volcanioclastic rather than containing plutonic rocks, as would be expected from eroded roots of a volcanic center. Further high-precision CA-ID-TIMS work searching for younger zircon grains in the oldest glaciogenic strata is under way to test this alternative hypothesis.

Rodinian Rifting along the North American Cordillera

Basal Neoproterozoic volcanic- and diamictite-bearing strata underlying the Cordilleran miogeoclone have long been thought to either represent protracted rifting before the rift-drift transition or a separate rift altogether (Bond and Kominz, 1984). Stewart and Suczuk (1977) originally suggested rifting began around 650 Ma. Further subsidence analysis of Cambrian strata along the Cordillera, using the backstripping approach, suggests Rodinian breakup and initiation of a continental drift phase at 600–550 Ma (Bond et al., 1983, 1985; Bond and Kominz, 1984; Armin and Mayer, 1983; Levy and Christie-Blick, 1991; Christie-Blick and Levy, 1989). The Re-Os age of 607.8 ± 4.7 Ma from organic-rich transgressive shale of the Old Fort Point Formation (Kendall et al., 2004) in the Horsethief Creek Group in southeastern British Columbia supports this (Fig. 3).

Felsic volcanism in southern Idaho spanned 717 Ma to 667 Ma, perhaps 70 m.y. before initiation of thermal subsidence at 600–550 Ma and at least 25 m.y. after an earlier phase of amagmatic, intracrystalline extension at ≤766 Ma to 742 Ma, recorded by strata of the Uinta Mountain Group (Dehler et al., 2010). Our new data suggest that separation of Laurentia from its western counterpart in Idaho and Utah occurred no earlier than 685 Ma, with the rift-drift transition after 667 Ma and thermal subsidence initiating at 600–550 Ma.

The “Missing-Link” model for Rodinian rifting (Li et al., 1995, 2008) suggests a first stage of rifting between western Laurentia and South China as early as 750 Ma. This was based on ages of 780–750 Ma from felsic and mafic intrusions and volcanics in South China (Lee et al., 1998; Li et al., 2003; Lin et al., 2007) correlated with those of the Franklin large igneous province (Harlan et al., 2003; Ernst et al., 2008, and references therein). Recent revisions to zircon and baddeleyite ages from gabbroic sills and dikes of the Franklin large igneous province indicate an age of 716 Ma (Macdonald et al., 2010), bringing the correlation into question.

Whereas dike swarms may be representative of crustal extension, only the sedimentary record of rift basins gives a clear indication of mechanical rifting. Sedimentation and volcanism in rift basins in South China have been dated between 820 Ma and 750 Ma (U-Pb SHRIMP ages), with a final nonvolcanic phase between 750 Ma and 690 Ma (Wang and Li, 2003). Wang and Li (2003) interpreted the 750–690 Ma rift phase as the rift-drift transition, consistent with a 90° counterclockwise rotation of the South China block at 750 Ma (Li and Evans, 2011). A “rift-to-drift” transition and rotation of South China at 750 Ma is not compatible with ages from western Laurentia, casting doubt on the “Missing-Link” model of Li et al. (1995), unless an intervening rift block existed (Colpron et al., 2002).

Colpron et al. (2002) proposed a two-stage rift model for the Cordillera from southeastern...
British Colombia to Utah, with the first stage of rifting at 750 Ma (revised to 716 Ma by Macdonald et al., 2010) to 700 Ma. A second stage of rifting at 600–570 Ma is suggested by ca. 600 Ma xenocrystic and solitary detrital grains throughout the Cordillera, a U-Pb zircon age of ca. 570 Ma from a trachyandesite flow above a rift-related unconformity in the Hamill Group in southeastern British Columbia (Colpron et al., 2002, and references therein), and a K-Ar age of ca. 580 Ma from the trachytic Brown’s Hole Formation in northern Utah, which overlies incised valleys within Ediacaran quartzite (Christie-Blick and Levy, 1989). Our ages and available data require a revision of the first phase to 717–660 Ma; otherwise, our ages are consistent with this model.

The 717–660 Ma age of volcanism demonstrated in southeast Idaho correlates with similarly aged volcanics documented in East Antarctica (Googe et al., 2002; Cooper et al., 2011). Googe et al. (2002) reported a U-Pb zircon age of ca. 668 Ma from the Beardmore Group of East Antarctica. Cooper et al. (2011) reported two ca. 650 Ma ages from a suite of volcanics spanning mildly alkalic basalt to rhyolite, in the Transantarctic Mountains, southern Victoria Land, Antarctica. This geochemistry is comparable to that documented by Harper and Link (1986) and Keeley (2011). Both 668 Ma and 650 Ma ages overlap with the 665–650 Ma magmatic pulse of alkali magmatism preserved in several igneous suites in western Laurentia (Lund et al., 2010). These relationships support the SW Laurentia–East Antarctica (SWEAT) connection of Moores (1991), are consistent with an Australian connection to SWEAT (Wingate and Giddings, 2000), and are also consistent with a 600 Ma rift event along the southeast Australia passive margin of Gondwana (Dureen and Crawford, 2003).

CONCLUSIONS AND IMPLICATIONS

Detailed mapping along the east face of Oxford Ridge has revised the stratigraphy of Fanning and Link (2004) and resolved the stratigraphic context of volcanioclastic strata on Oxford Ridge (Fanning and Link, 2008). The dated Oxford Mountain tuffite lies above glaciogenic diamictite, itself above a gradational contact with the underlying Bannock Volcanic Member. These relationships refute that (1) the volcanioclastic rock is not in contact with glaciogenic diamictite (Hoffman and Li, 2009) and (2) the contact is tectonic (Macdonald et al., 2010).

SHRIMP and CA-ID-TIMS U-Pb zircon dating has revealed a complex set of zircon age populations in the Oxford Mountain tuffite, with components at ca. 709, 702, 700, 695, 690, and 685 Ma. This suggests protracted volcanism spanning 709–685 Ma. Inclusion of the ca. 717 Ma porphyritic silicic clast from Portneuf Narrows (Fanning and Link, 2004) increases the range to ~32 m.y., which is comparable to the duration of silicic volcanism in rift settings lasting ~40 m.y. (Bryan et al., 2002). The age components of the Oxford Mountain tuffite overlap with SHRIMP age components of ca. 705 Ma, 700 Ma, and 688 Ma, from the lower diamictite of Portneuf Narrows, providing a correlation to the diamictite on Oxford Ridge. The ages also overlap with the ca. 708 Ma age (SHRIMP) from the formation of Perry Canyon in northern Utah (Balgord et al., 2010), supporting regional correlations with the first of two diamictite intervals recognized by Crittenden et al. (1983). Indistinguishable detrital zircon U-Pb age spectra between Fivemile Canyon and Oxford Ridge suggest deposition within a single rift basin with a mixed Laurentian source. The 685.5 ± 0.4 Ma age correlates well with ages of 685.6 ± 7 Ma and 684.6 ± 4 Ma from the Edwardsburg Formation, in central Idaho and the 688 ± 9.5–6.2 Ma age from the Gataga volcanics in northern British Columbia, supporting regional magmatism at ca. 685 Ma. Further, the 685 Ma age provides a robust maximum depositional age constraint on the diamictite on Oxford Ridge and a maximum age for onset of mechanical rifting. Thermal subsidence is constrained to after 667 Ma.

Lu-Hf zircon analyses of ca. 705–685 Ma zircons from the Pocatello Formation yield initial εHf values in the range +2 to −17. The wide variation and weakly positive values suggest that the igneous zircons may have crystallized from initially juvenile magma that incorporated crustal components of the underlying Farmington Canyon Complex and the Archean Wyoming craton.

The 685 Ma and 667 Ma maximum age constraints on the two respective middle Cryogenian diamictic populations in the Pocatello Formation highlight that the strata are younger than the ca. 715 Ma age for Rapitan diamictite in Canada (Macdonald et al., 2010) and the ca. 711 Ma age for diamictite in Oman (Bowring et al., 2007). These precise ages may be consistent with a 50 m.y. duration for a single middle Cryogenian, “Sturtian,” glaciation. This is twice as long as the suggested duration for the late Cryogenian, “Marinoan,” glaciation (Hoffman and Li, 2009). Available data may also be consistent with two Sturtian glaciations separated by an interglacial period lasting millions of years. A seemingly problematic factor is the overlapping age constraints on the upper “Sturtian” bound at 660 Ma and lower “Marinoan” bound at 655 Ma. Revision of the current understanding of the “Sturtian” and “Marinoan” episodes to allow a third and intervening glaciation at ≥265–660 Ma may be required.

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