INTRODUCTION

The Wyoming Province is an Archean craton that occupies most of Wyoming and portions of Montana, and adjacent states. The Archean rocks are exposed in the cores of basement-involved Laramide uplifts. The early mafic crust appears to have been Hadean (Frost et al., 2017), though most of the exposed area consists of Palearchean to Neoarchean quartzofeldspathic orthogneisses that retain an isotopic signature of that ancient crust (Frost, 1993). The Wyoming craton is subdivided into three main provinces (Fig. 1; Mueller and Frost, 2006). The northwestern province is the Montana metasedimentary province, which is an area composed of quartzo-feldspathic gneisses, all of which were accreted at ca. 2.55 Ga. The Beartooth-Bighorn magmatic zone, which occupies the core of the craton, is dominated by orthogneisses. Most of the Beartooth-Bighorn magmatic zone contains rocks that were last deformed between 2.86 Ga and 2.71 Ga. On the southern and western margins of the Beartooth-Bighorn magmatic zone, these older gneisses were overprinted by deformation that is as young as 2.63 Ga. The southern margin of the craton contains the southern accreted terranes, consisting of various fragments of arcs and continents that were accreted to the Wyoming Province at ca. 2.63 Ga.

The Teton Range, a small range of spectacular mountains in northwestern Wyoming, exposes some of the westernmost outcrops of the Archean Wyoming Province (Fig. 2). The northern portion of the range, described by B. Frost et al. (2006), Frost et al. (2016a), and Swapp et. al. (2018), contains some of the oldest high-pressure granulites in the world. In this paper, we summarize the past work on the northern Teton Range, identify a deformation zone that marks the contact between gneisses of the northern and southern Teton Ranges, and discuss how this structural belt relates to the final Neoarchean assembly of the Wyoming Province.

GEOLOGIC BACKGROUND

Preliminary geologic mapping of the range was conducted by John C. Reed Jr. from 1962 to 1970 (Fig. 2; Reed, 2014). Subsequent studies (Miller et al., 1986; B. Frost et al., 2006; Frost et al., 2016a; Swapp et al., 2018) concentrated...
on the Archean history of the northern part of the range, roughly north of Leigh Canyon, where evidence of high-pressure (high-P) granulite facies is preserved. These studies show that the high-P granulites were metamorphosed at 2695 Ma and were tectonically assembled with layered gneisses at 2685 Ga (Swapp et al., 2018). These rocks were intruded by trondhjemitic leucogranites between 2685 and 2675 Ma (Frost et al., 2016a).

The basement rocks of the southern Teton Range include quartzofeldspathic gneisses and hornblende gabbro (Reed, 1973; Love et al., 1992). They lack the 2675–2685 Ma leucogranites and all traces of the 2695 Ma granulite metamorphism. Intruding both domains is the Mount Owen batholith, an undeformed, peraluminous leucogranitic batholith that was emplaced at 2547 ± 3 Ma (Zartman and Reed, 1998). The Mount Owen batholith underlies the rugged high peaks in the central part of the range and forms granitic and pegmatitic dikes throughout the uplift.

Figure 1. Map of the Wyoming Province, showing the mountain ranges exposing Archean rocks and the location of the three major subprovinces: the Montana metasedimentary province, the Beartooth-Bighorn magmatic zone, and the southern accreted terranes. The study area in the Teton Range is indicated by the rectangle and shown in Figure 2. Colors identify time of final Archean deformation and magmatism; see text for discussion. TTG—tonalite-trondhjemite-granodiorite. Figure is adapted from Frost et al. (2016a).

Figure 2. Geologic map of the Archean rocks of the Teton Range, modified after Love et al. (1992), showing locations of samples included in this study.
Archean Geology of the Northern Teton Range

The Archean rocks of the northern Teton Range consist of three major rock types. Most distinctive of these is the Moose Basin gneiss, a suite of mafic and pelitic rocks that were metamorphosed into granulite facies (Swapp et al., 2018). The Layered Gneiss, a sequence of heterogeneous gneisses, consists dominantly of psammitic to pelitic paragneiss with minor amounts of quartzofeldspathic orthogneiss, and mafic and ultramafic rocks. The Layered Gneiss was mapped by Reed (1973) as a single unit that extended the whole length of the range. Frost et al. (2016a) and Swapp et al. (2018) separated the Moose Basin gneiss from the Layered Gneiss. In the northern Teton Range, the Layered Gneiss and the Moose Basin gneiss are intruded by two compositionally distinct leucogranitic gneisses, the Webb Canyon Gneiss and the Bitch Creek Gneiss (Frost et al., 2016a).

The Moose Basin gneiss extends from Moose Basin at the northern end of the Teton Range southwards along the crest of the range for ~10 km (Fig. 2). It is composed of kyanite-bearing metapelitic rocks with leucogranitic patches interpreted as leucosomes formed by partial melting (Swapp et al., 2018). Mafic rocks are also found within the Moose Basin gneiss and these likewise contain leucosomes formed by partial melting. The leucosomes within the mafic rocks contain assemblages that record high-pressure granulite-facies metamorphism. The Zr-in-rutile thermometer records temperatures around 900 °C. The assemblage garnet-kyanite-rutile-quartz (Grt-Ky-Rt-Qz; abbreviations after Whitney and Evans, 2010) allows application of the garnet-rutile-aluminosilicate-ilmenite-quartz (GRAIL) geobarometer (Bohlen et al., 1984). At 900 °C, the core garnet compositions yield a pressure of 12.0 kbar; rim garnet compositions yield a pressure of 12.7 kbar. Since ilmenite does not occur as inclusions in the garnets, these pressures are necessarily minimum estimates (Swapp et al., 2018). U-Pb sensitive high-resolution ion microprobe (SHRIMP) dating of zircon within the garnet-bearing leucosomes records an age of 2695 ± 7 Ma, which Swapp et al. (2018) interpreted as the age of the granulite metamorphism. Sparse zircon grains yielding older ages of 3.1–2.8 Ga, negative initial ɛNd values of metapelitic gneiss samples, and Nd model ages between 3.4 and 3.0 Ga suggest that the protoliths to the Moose Basin gneiss significantly predate granulite-facies metamorphism (Swapp et al., 2018).

In the northern Teton Range, the outcrops of the Layered Gneiss that we studied consist mostly of quartz-biotite-plagioclase paragneiss interlayered with minor amounts of quartzofeldspathic orthogneiss and low-Ca amphibolite, amphibolite, and metaperidotite. The gneiss has been metamorphosed in upper amphibolite facies and locally is migmatitic. Zircon from the leucosome in the migmatite yields a SHRIMP U-Pb date of 2685 ± 5 Ma, which Swapp et al. (2018) interpreted as the age of tectonic assembly of the Moose Basin gneiss and the Layered Gneiss. No older zircon areas were identified, consistent with mainly positive initial ɛNd values for the Layered Gneiss (Swapp et al., 2018).

Reed (1973) mapped large panels of leucogranitic gneisses that intruded the Moose Basin and the northern Layered Gneiss, which he named the Webb Canyon Gneiss. Based on the geochemistry of these leucogranitic gneisses, Frost et al. (2016a) recognized two compositionally distinct units, both of which are calcic and silica-rich. The dominant unit is the Webb Canyon Gneiss, which is ferroan and, based on the classification of Barker (1979), is low-Al. The less-voluminous unit, the Bitch Creek gneiss, forms small dikes and plutons within the Moose Basin and northern Layered gneisses. It is distinguished from the Webb Canyon Gneiss because it is magnesian and shows high-Al values, based on the classification of Barker (1979). Frost et al. (2016a) postulated that these compositional differences are produced by different modes of origin: water-excess melting for the Bitch Creek gneiss, and dehydration melting for the Webb Canyon Gneiss. Because the Webb Canyon Gneiss is by far the most voluminous, both types of leucogranitic gneiss are included in the Webb Canyon Gneiss unit on the map shown in Figure 2 and in the maps in Frost et al. (2016a) and Swapp et al. (2018). SHRIMP U-Pb zircon ages from the Webb Canyon and Bitch Creek gneisses range from 2675 to 2685 Ma (Frost et al., 2016a), suggesting that they were emplaced as multiple intrusions during or within a few million years following the tectonic assembly of the Moose Basin gneiss and the northern Layered Gneiss. Their apparent lack of inherited zircon and mainly positive initial ɛNd values are consistent with an origin by partial melting of juvenile sources with little to no involvement of significantly older crust (Frost et al., 2016a).

Tabular amphibolite bodies occur within the leucogranites of the Moose Basin gneiss, within the Layered Gneiss, and within the Webb Canyon and Bitch Creek gneisses. Locally, some amphibolites contain garnet, but most do not. Initial ɛNd values (τc = 2685 Ma) of the Moose Basin gneisses from the northern Teton Range are mainly negative, whereas the initial ɛNd values of the Layered Gneiss, Webb Canyon Gneiss, and Bitch Creek gneisses are mainly positive. The positive initial values of most of the leucogranitic rocks suggest that they could be derived by partial melting of the Layered Gneiss or other similarly juvenile sources (Frost et al., 2016a). Contributions from the Moose Basin gneiss, with its Nd isotopic signature indicating a more ancient continental provenance, appear to have been limited.

Archean Geology of the Southern Teton Range

The Archean rocks of the southern Teton Range consist of three units, the Layered Gneiss, the Rendezvous Gabbro, and the Mount Owen batholith (Fig. 2). In the southern Teton Range, the Layered Gneiss extends from Snowshoe Canyon south to Rendezvous Mountain. It is dominantly a quartzofeldspathic gneiss that contains minor amounts of amphibolite. Both paragneiss and orthogneiss have been identified, but much of it is a
quartzofeldspathic gneiss of uncertain parentage. The Layered Gneiss is variably foliated and folded. Grain size is variable on the outcrop scale (Fig. 3A). Biotite selvages on some leucocratic layers suggest that in some areas, the Layered Gneiss has undergone partial melting (Fig. 3B). Because both the northern and southern Layered Gneisses are highly heterogeneous, we have not been able to distinguish them in the field; the distinction is made based on their occurrence relative to the Moran deformation zone.

The “Augen” Gneiss (Reed, 1973) extends from Paintbrush Canyon north to Bivouac Peak. We put the word “augen” in quotation marks because by modern terminology, this is not an augen gneiss (i.e., a porphyroclastic rock with feldspar porphyroclasts). This rock is a gray, granodioritic orthogneiss with conspicuous, large (0.5–4 cm long) phenocrysts of white plagioclase and alkali feldspar. It contains biotite with accessory hornblende, and it is moderately well foliated and lineated, with little to no compositional layering. The foliation is concordant with the adjacent biotite quartzofeldspathic gneiss.

Another distinctive orthogneiss within the southern Layered Gneiss is the “bright-eyed gneiss” (Bradley, 1956) found in Death Canyon. This biotite-hornblende orthogneiss is distinguished by magnetite crystals surrounded by white haloes where biotite is absent, which have the appearance of hundreds of small eyes peering from the rock (Fig. 3C).

Biotite-rich paragneiss is found throughout the southern Layered Gneiss. We have examined it in three locations. At the east end of Moran Canyon, the paragneiss is interlayered with amphibolite, and leucosomes indicate that it has been partially melted (Fig. 3D). On the southeast face of the Grand Teton, paragneiss is preserved as enclaves within Mount Owen granite (Fig. 3E). Biotite paragneiss with the assemblage garnet-biotite-cordierite-staurolite (Grt-Bt-Crd-St) crops out in Paintbrush Canyon and at Paintbrush Divide (Fig. 3F). Concordant bodies of amphibolite with well-developed foliations, some garnet-bearing and some garnet-free, are found throughout the southern Layered Gneiss. The southern Layered Gneiss also includes minor occurrences of calc-silicate schist.

The Rendezvous Gabbro crops out over an area of ~25 km² in the southern-most Teton Range. It is a weakly metamorphosed rock that retains an igneous texture consisting of centimeter-sized hornblende and plagioclase. Several textural features indicate that the hornblende has replaced pyroxene: (1) The hornblende is green and not brown, indicating that it is poorer in TiO₂ than most igneous hornblendes; (2) many of the hornblende grains contain inclusions of quartz, which we interpret as having formed by the reaction: augite + orthopyroxene + plagioclase + H₂O = hornblende + quartz (see Tracy and Frost, 1991); and (3), the hornblende locally has zones of non-pleochroic tremolite, which we interpret as replacements of the remnants of the original clinopyroxene. Plagioclase for the most part preserves the tabular shape of igneous feldspar. Locally, it shows deformation twinning, and, in places, it is bent and contains abundant deformation twins. Some grains are rimmed by plagioclase neoblasts, indicating that it has undergone some high-temperature deformation involving fast grain-boundary migration. On the whole, the Rendezvous Gabbro has not undergone extensive deformation.

Initial ε⁹⁶Nd values (t₀ = 2685 Ma) of basement gneisses from the southern domain are variable. Initial ε⁹⁶Nd from the Layered Gneiss vary from -0.2 to +4.0, indicating relatively juvenile protoliths (B. Frost et al., 2006). The initial ε⁹⁶Nd values of porphyritic gneiss samples also are negative, and Nd model ages are 3.1–3.2 Ga, suggesting that the porphyritic gneiss magma sources included some older crustal materials (B. Frost et al., 2006). Initial ε⁹⁶Nd values for the Rendezvous Gabbro (2685 Ma) are intermediate, at +0.0 and +0.9 (B. Frost et al., 2006).

Mount Owen Batholith

The Mount Owen batholith is a peraluminous leucogranite that contains muscovite and locally garnet. The grain size of the rock is heterogeneous; in many outcrops, it grades from pegmatic to aplitic, suggesting that water activity was highly variable during emplacement. Pegmatic and granitic dikes from the Mount Owen batholith are present throughout the Teton Range. In some areas, the pegmatic dikes contain centimeter-sized grains of tourmaline. The 2547 ± 3 Ma Mount Owen batholith (Zartman and Reed, 1998) is undeformed, which indicates that deformation had ceased by that time. The initial Nd isotopic compositions of the Mount Owen batholith are consistent with an origin by partial melting of rocks with an isotopic composition similar to the older gneisses of the Teton Range, supporting a crustal origin for the peraluminous leucogranite batholith (B. Frost et al., 2006).

The Mount Owen batholith is composed of quartz, alkali feldspar, plagioclase, biotite, and muscovite. The alkali feldspar is perthitic. In some rocks, the alkali feldspar is orthoclase, in some, it has been partially inverted to microcline, and in others, it has been completely inverted. In many rocks, muscovite is coarse and tabular and is clearly magmatic. In others, it is a fine, fealty intergrowth that is likely to have formed deuterically. In some rocks, biotite is the dominant mica; in others, it is muscovite. Most rocks show small amounts of retrogression. Plagioclase in many rocks is riddled with sericite, and biotite and garnet are altered to chlorite. Biotite locally is altered to epidote; in places, this reaction also produced titanite. Garnet is present in some samples. In a few samples, quartz contains tiny needles of sillimanite (<50 mm in maximum dimension), suggesting that crystallization began in the sillimanite field and then moved into the muscovite field as temperature fell or as water fugacity increased. Zircon is abundant, and allanite and monazite were observed in some samples.

METHODS

We compiled structural data on the deformed Archean rocks of the Teton Range from two sources: measurements by the authors and measurements...
Figure 3. Photographs of the various types of Layered Gneiss in the southern Teton Range. (A) Southern Layered Gneiss, upper Garnet Canyon. (B) Southern Layered Gneiss, Avalanche Canyon. (C) Bright-Eyed gneiss, Death Canyon. (D) Biotite paragneiss, Moran Bay. (E) Biotite paragneiss, Upper Exum route near summit of the Grand Teton. (F) Strongly foliated biotite paragneiss at Paintbrush Divide within the Moran deformation zone. Field of view for A and B is about 1 meter. The horizontal pegmatite dike in Fig. 3E is about 0.7 m thick.
U-Pb geochronology was undertaken by two methods. U-Pb isotopic data on zircon from biotite paragneiss and porphyritic gneiss in the southern Teton Range were obtained using the sensitive high-resolution ion microprobe–reverse geometry (SHRIMP-RG) at Stanford University and at the Australian National University (ANU). Zircon grains were mounted in epoxy along with chips of reference zircon. At Stanford, the standards were Duluth Gabbro (1100 Ma; Paces and Miller, 1993) and R33 (Black et al., 2004), and at ANU, the standard was Temora (417 Ma; Black et al., 2003). After polishing, cathodoluminescence (CL) scanning electron microscope (SEM) images were taken for all zircon grains. Isotopic ratios and U, Th, and Pb concentrations were measured using procedures similar to those given in Williams (1998, and references therein). The data were reduced using the SQUID macro of Ludwig (2001). Uncertainties given for individual analyses (ratios and ages) are at the 1σ level, with correction for common Pb made using the measured 206Pb/235U ratios. Concordia plots and linear discordia regression fits were carried out using Isoplot (Ludwig, 2003), and uncertainties are reported at the 95% confidence level.

Zircon grains separated from two samples of the Rendezvous Gabbro and a boudinaged amphibolite dike in the Bitch Creek gneiss were dated by a procedure modified from the chemical abrasion–thermal ionization mass spectrometry (CA-TIMS) method of Mattinson (2005) at the University of Wyoming. Representative zircon grains from each morphological subpopulation were annealed at 850 °C for 50 h and then dissolved in two stages in HF and HNO3. The first step was for 12 h at 180 °C, which removed the most metamict domains. Single grains were then spiked with ET535, a mixed 205Pb-233U-235U tracer, completely dissolved at 240 °C for 30 h, converted to chlorides at 180 °C overnight, and loaded onto Re filaments with H3PO4 and silica gel. Pb and UO2 isotopic ratios were measured in single-collector mode using a Daly-photo-multiplier collector on a Micromass S54 TIMS. Pb fractionation was determined by multiple analyses of NIST 981, and U fractionation was determined internally. Pb procedural blanks averaged 2 pg, and U was consistently less than 0.1 pg. Isotopic compositions of additional common Pb from each analysis of the Rendezvous Gabbro were modeled using the model of Stacey and Kramers (1975); data from the red zircon grains in 04T16 were reduced with whole-rock Pb values, and data from colorless zircon grains in 04T16 were reduced with the common Pb solution to a three-dimensional (3-D) total Pb isochron. Raw data were reduced using PbMacDat and Isoplot, based on the algorithms of Ludwig (1988, 1991).

Titanite from three samples of Rendezvous Gabbro and from the boudinaged amphibolite dike in the Bitch Creek gneiss were selected based on size and dissolved in HCl/HF ratio (5:1) at 180 °C in Parrish-style dissolution bombs within a high-pressure Parr dissolution cell. Pb and U were purified on HBr-HCl and HNO3-HCl ion exchange columns, respectively. Isotope ratios were analyzed on the Micromass S54 at the University of Wyoming. Pb was run in static-multicollector mode with 204Pb in the Daly; U was loaded with graphite and run as a metal in static Faraday mode. Pb blanks for titanite averaged 8 pg, and U was less than 0.1 pg. Stacey and Kramers (1975) model values were used for initial Pb isotopic compositions for titanite from 07T23 and 98T12, and the Pb isotopic composition of coexisting feldspar was used for samples 07T22 and 04T16. U-Pb isotopic data for the samples analyzed by SHRIMP-RG are reported in the supplemental materials Table S6, and U-Pb isotopic data for samples analyzed by TIMS are reported in Table S7 (footnote 2).

Samples for whole-rock geochronology were prepared by crushing rock trimmed of weathered surfaces between tungsten carbide plates in a hydraulic press, followed by powdering in a ceramic or tungsten carbide ring mill. Major elements and a limited suite of trace elements were determined by X-ray fluorescence (XRF) at the University of Wyoming. Reproducibility determined from replicate analyses was better than 0.01% for all oxides, except SiO2, which was better than 0.1%. Trace-element concentrations, including rare earth element (REE) concentrations, were determined by inductively coupled plasma mass spectrometry (ICP-MS) at XRAL Laboratories, Don Mills, Ontario, Canada, or ALS Minerals, Reno, Nevada. Average reproducibility was <5% for all elements, except for Rb, Hf, Eu, Tb, Tm, and Yb, which were <15%. These data, together with previously published analyses, are presented in Table S8 (footnote 2). Sample locations are provided in Table S9 (footnote 2).

Table S10 (footnote 2) compiles Sm-Nd isotopic data for the southern Teton Range, most of which were previously published by B. Frost et al. (2006). Data on sample 07T3 was obtained following methods provided in Frost et al. (2016a).

Mineral analyses were obtained on polished thin sections for garnet, plagioclase, and hornblende using a JEOL 8900 Superprobe electron microprobe with a 15 kV accelerating voltage and sample current of 20–30 nA for all minerals. Both natural and synthetic silicates and oxides were used as standards. Analyses were corrected using the ZAF procedure. Errors were less than 1% of the measured values of major elements and increased greatly for elements with abundances below 1 wt% oxide. Representative analyses are presented in Table S11 (footnote 2).

### RESULTS

#### Structural Geology—Moran Deformation Zone

The major structural feature in the Tetons is a deformation zone that crops out between Leigh and Snowshoe Canyons (Fig. 2). This deformation zone, which has not been recognized previously, is over 2 km wide in Moran Canyon and extends south from Moran Canyon through the western portion of the Moran massif and over Thor Peak. Because of its exposure in the Moran Canyon and Moran massif, we call it the Moran deformation zone.
It includes high-strain zones, tens to hundreds of meters wide, that are marked by increasing intensity of foliation, grain-size reduction, and the development of porphyroclasts. Figure 4 shows one of these high-strain zones located near Triple Glacier on the north face of the Thor Peak–Mount Moran massif. The northern end of the Moran deformation zone has not been traced across Snowshoe Canyon, but based upon lithologic contrasts across the deformation zone, it probably projects to the east of the Archean outcrop north of Snowshoe Canyon. The southern end of the Moran deformation zone is exposed in the slopes south of Leigh Canyon but is obliterated by the Mount Owen batholith before reaching Paintbrush Divide. We project it to extend approximately to Hurricane Pass, west of the outcrops of the southern Layered Gneiss (Fig. 2).

Foliations in the Moran deformation zone in Moran Canyon give strikes that average 10° and dips of around 40° to the east. Considering that the sedimentary rocks on the west side of the Teton Range dip ~8° to the west, the dip of the zone may have been somewhat steeper before uplift on the Teton fault. The foliation in the high-strain zones folds around numerous phacoids of less-deformed gneisses, which account for the variations in strikes and dips we obtained from the deformation zone. Lineations generally are down dip, and the sense of shear as determined by asymmetrical porphyroblasts is top-to-the-west, which suggest that the deformation zone represents a broad area of strain marking a crustal-scale thrust fault.

**Structural Fabrics**

Apart from the Mount Owen batholith, which is undeformed, the Archean rocks of the Teton Range are variably foliated gneisses. Minor folds occur locally on all scales, as do higher-strain zones. Stereographic projections of foliations measured across the range suggest that the Archean rocks of Teton Range can be divided into three structural domains (Fig. 5). The northern area lies north and west of the Moran deformation zone, the central area extends from the vicinity of Mount Moran south to the east face of Teewinot, and the southern area extends from Avalanche Canyon to the southern limits of Precambrian outcrop.

In the northern area, the Moose Basin gneiss and Layered Gneiss show three generations of folds (Swapp et al., 2018). The earliest ($F_1$) folds are seen as isolated isoclinal fold hinges. The second generation ($F_2$) is seen as large iso-planar foliation related to the $F_3$ folds. Because the fabric on the east limb of the $F_1$ folds is parallel fabric to the Moran deformation zone (see Figs. 5A and 5B), we interpret the $F_2$ folds to have formed in response to the Moran deformation zone and that the major foliation formed before the Moran deformation zone.

The central area is characterized by consistent north-northeasterly striking foliations that dip moderately to the east (Fig 5B). This fabric is roughly parallel to the strike and dip of the shear zones within the Moran deformation zone, and we conclude that this fabric represents foliations that formed during the deformation, as well as any earlier foliations on either the footwall or headwall that were transposed into parallelism with the Moran deformation zone.

We have done limited field work in the southern Teton Range; the fabric data from this area are mainly those of Reed and his field assistants. Multiple fold generations exist (for example, see Fig. 3B), but the major foliation in the southern Teton Range displays broad open east-trending folds that plunge gently to the east (Fig. 5C). Considering that the trace of the Moran deformation zone likely lies ~5 km to the west of the westernmost outcrop of the Layered Gneiss in this area, we suggest that these folds and foliations represent the structure of the hanging wall of the Moran deformation zone and that they were not transposed during the thrusting event.

**Geochronology**

For this study, we obtained U-Pb isotopic analyses of zircon from six samples. The first, 07T1, is a sample of the "Augen" Gneiss in the southern Layered Gneiss collected in Paintbrush Canyon (Table S6 [footnote 2]). This sample was analyzed to constrain the intrusive age of the "Augen" Gneiss. Analyses of 13 areas on 10 grains yielded a wide range of U contents, from ~140 to ~2040 ppm. The four analyses with the U highest contents (>1450 ppm) were also the most discordant and were not interpreted. Another high-U grain area gave the oldest $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2870.7 ± 5.4 Ma. The remaining eight analyses yielded a concordia age of 2803 Ma, but there is some scatter in the $^{207}\text{Pb}/^{206}\text{Pb}$ ages, as indicated by the mean square of weighted deviates (MSWD) of 2.7. If the scatter is due to slight inheritance in two grain areas, a chord through the remaining six analyses yields a date of 2800 ± 8 Ma (MSWD = 1.4; Fig. 6A). We interpret this date as the best estimate for the magmatic age of the "Augen" Gneiss.

We also analyzed detrital zircon from two samples of biotite paragneiss from Paintbrush Canyon. One sample, 07T3, was composed mainly of low-U zircon (35–450 ppm), with only a few higher-U, more discordant zircon grains. Neglecting the four analyses that were more than 5% discordant, the remaining analyses define several age populations. Three grains defined an upper-intercept age of ca. 2840 ± 21 Ma, six grains yielded a date of 2663 ± 16 Ma, and six other grains gave 2711 ± 10 Ma. Two older grains yielded $^{207}\text{Pb}/^{206}\text{Pb}$ dates of 2.84 and 2.86 Ga (Fig. 6B).

Another biotite paragneiss sample, 10T4, contained a few zircon grains, many of which were relatively high in U and discordant. Of the 13 analyses, only six were less than 35% discordant. These analyses defined two age groups of 2630 ± 13 Ma and 2663 ± 9 Ma (Fig. 6C). We note that these two age groups are also present in sample 07T3.

The age of the Rendezvous Gabbro was constrained by U-Pb dates on single zircon grains from sample 98T12, a coarse-grained hornblende gabbro...
Figure 4. Field photographs of the Moran deformation zone. (A) Photograph of the north face of the Thor Peak–Mount Moran massif near the westernmost portion of T riple Glacier. Foliated gneisses within the Moran deformation zone are crosscut by several generations of undeformed dikes of Mount Owen granite. Red arrows indicate locations of structural measurements. Triangular peak in the middle of the photograph is about 100 m high. (B) Strongly foliated gneiss from a high-strain zone of the Moran deformation zone located to right of field of view shown in A.
excavated from the Jackson Hole Mountain Resort (Table S7 [footnote 2]). Four single-grain CA-TIMS analyses of red, anhedral grains gave an upper-intercept date of 2685.3 ± 2.2 Ma (MSWD = 1.3; Fig. 6D). Sample 07T23, collected from a garnet hornblende dike that cuts the Rendezvous Gabbro, contained three zircon morphologies, each of which yielded different dates. Three large anhedral grains yielded an upper-intercept date of 2691.7 ± 2.8 Ma, five medium euhedral grains produced a range of $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 2688 and 2683 Ma, and small euhedral grains gave the youngest weighted mean age of 2678 ± 3 Ma (Fig. 6D). These results suggest that the emplacement and crystallization of the Rendezvous Gabbro involved several episodes over a period of several million years.

The boudinaged mafic dike in the Bitch Creek gneiss (04T16) yielded two distinct euhedral zircon populations: red and colorless. Three single-grain CA-TIMS analyses of the red population gave a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ date of 2682.1 ± 1.6 Ma (Fig. 6E; Table S7 [footnote 2]). We interpret these grains to be xenocrysts from the host Bitch Creek gneiss (06T17), which yielded a U-Pb zircon date of 2685.7 ± 4.2 Ma (Frost et al., 2016a). Four single-grain CA-TIMS analyses of the colorless population had significantly lower U concentrations than those of the red zircons and relatively high amounts of common Pb, with blank-corrected $^{206}\text{Pb}/^{204}\text{Pb}$ from 1043 to 142 (Table S6 [footnote 2]). A total Pb 3-D isochron solution of these analyses indicated an isotopically evolved composition for this common Pb ($^{206}\text{Pb}/^{204}\text{Pb} = 21.4 ± 6.2$, $^{207}\text{Pb}/^{204}\text{Pb} = 16.4 ± 4.5$). Propagating errors with these values produced relatively large concordia ellipses (Fig. 6E) and a concordia age of 2667.6 ± 4.3 Ma.

Three sizes of titanite picks from Rendezvous Gabbro sample 98T12 yielded a range of concordant $^{207}\text{Pb}/^{206}\text{Pb}$ dates, from 2615.4 ± 1.7 Ma (large ≥300 μm diameter), to 2614.0 ± 1.7 Ma (medium ~200 μm), to 2611 ± 1.7 Ma (small
≤100 μm; Table S7 [footnote 2]). A large single titanite grain from 07T23 yielded a date of 2622 ± 5 Ma. Different dates for dark and pale titanite were obtained from sample 07T22, a sample of Rendezvous Gabbro cut by dike sample 07T23. All the titanite grains from 07T22 were mechanically abraded to try to isolate end-member growths and/or a diffusion gradient. The two most intensely abraded fractions of pale grains produced a weighted mean 207Pb/206Pb date of 2626.7 ± 2.5 Ma (MSWD = 0.0005). The four-point weighted mean, moderately abraded pale grains and a single extra-large dark grain gave 2622.4 ± 0.9 Ma (MSWD = 0.74), and a two-point weighted average of dark titanite picks gave 2616.5 ± 1.1 Ma (MSWD = 0.02). Five titanite fractions from 04T16, the boudinaged mafic dike intruding Bitch Creek gneiss in the northern domain, produced an upper-intercept age of 2622.8 ± 2.5 Ma (MSWD = 0.68; Fig. 6F; Table S7 [footnote 2]). These ages are consistent with titanite growth at ca. 2620 Ma, ca. 2622 Ma, and possibly as young as 2615 Ma.

**Geochemistry**

Thirty-two whole-rock geochemical analyses of samples from the Layered Gneiss, Moose Basin gneiss, Rendezvous Gabbro, and Mount Owen batholith...
are presented on Table S8 (footnote 2), along with data from these units previously published by Miller et al. (1986), B. Frost et al. (2006), and Wilks (1991). We subdivided the Layered Gneiss into northern and southern groups, based on location with respect to the Moran deformation zone. Together with data on the Webb Canyon and Bitch Creek leucogneisses published by Frost et al. (2016a), these data represent all the available geochemical analyses on Archean rocks from the Teton Range.

**Layered Gneiss**

The northern and southern Layered Gneiss groups are composed of quartz-rich rocks and amphibolites, a bimodal assemblage that is reflected in their silica contents (Table S8 [footnote 2]). The quartzofeldspathic rocks range from 68 to 78% SiO$_2$, whereas the mafic rocks contain less than 52% SiO$_2$. The only intermediate-composition sample is biotite paragneiss. Most are magnesian, although one sample of silica-rich northern Layered Gneiss is strongly ferroan (Fig. 7A). The silica-rich rocks are mainly calcic, but some are calc-alkalic (Fig. 7B). Almost all are peraluminous (Fig. 7C). The Na$_2$O contents of most felsic gneiss samples cluster between 2.1% and 4.1%, but K$_2$O varies widely from 0.1% to 4.5% (Figs. 8A and 8B). Almost all Layered Gneiss samples have Na$_2$O > K$_2$O. Rb contents are low (45–130 ppm). Sr varies from 65 to 450 ppm, suggesting a role for fractional crystallization or accumulation of plagioclase (Fig. 9). Other trace-element characteristics, including Y and Nb contents, are typical of magnesian, calc-alkalic to calcic granitoids generally (Table S8 [footnote 1]; cf. Frost et al., 2016a). There are two important differences between the geochemistry of the northern and southern Layered Gneisses. Because the K$_2$O contents of most northern Layered Gneisses are lower than those of the southern Layered Gneisses, the average northern Layered Gneiss group has lower K$_2$O/Na$_2$O than the average southern Layered Gneiss group (Fig 8C). In addition, Zr is high in northern Layered Gneiss (Zr = 436–692 ppm), whereas most southern Layered Gneiss samples have Zr between 100 and 300 ppm.

The two biotite paragneiss samples from the southern Layered Gneiss have REE abundances similar to those of average shales, yielding REE patterns with
light (L) REE enrichment and flat heavy (H) REE patterns (Fig. 10A). One sample has a slight positive Eu anomaly. The third southern Layered Gneiss sample, an orthogneiss from Death Canyon, is enriched in LREEs and has a negative Eu anomaly. REE patterns of northern Layered Gneisses are more variable and include a group with high REE abundances and deep negative Eu anomalies (Fig. 10B). These patterns are reminiscent of the “seagull”-shape patterns of the Webb Canyon Gneiss (shaded area on Fig. 10; Frost et al., 2016a). Others have steep LREE/HREE patterns with no Eu anomalies, and one sample has a steep pattern but low REE abundances and a strong positive Eu anomaly (Fig. 10B). Clearly, the Layered Gneisses in both domains include rocks with a variety of composition, origin, and petrogenetic evolution.

Amphibolite bodies and dikes from the northern Teton Range contain 45%–52% SiO₂, are metaluminous, and define an iron-enrichment trend (Fig. 7A). K₂O, Rb, Nb, Y, and Zr contents are low. REE contents are low, and REE patterns are flat. Modest negative or positive Eu anomalies suggest plagioclase fractionation or accumulation in some samples (Fig. 10C).

**Moose Basin Gneiss**

The Moose Basin gneiss samples range in silica content from 51% SiO₂ for biotite-rich metapelitic gneisses to 80% SiO₂ for felsic gneiss. The three most silica-rich samples are also strongly ferroan (Fig. 7A), have high REE contents with deep, negative Eu anomalies (Fig. 10D), and share other major- and trace-element characteristics in common with the Webb Canyon Gneiss (Table S8 [footnote 2]; cf. Frost et al., 2016a). We conclude from their geochemical composition that these are samples of the Webb Canyon Gneiss that intruded the Moose Basin gneiss.

The majority of the Moose Basin gneiss samples have the aluminous nature expected of metapelitic rocks (Fig. 7C) and LREE-enriched REE patterns (Fig. 10D). Three metapelitic rocks have silica contents much lower than average Archean shale (for example, average Canadian Archean shale; Cameron and Garrels, 1980; see Figs. 7 and 8) and may represent restites from which leucosome has been extracted. Other Moose Basin gneiss samples have SiO₂ contents higher than average Archean shale (Figs. 7 and 8). The protoliths of these more siliceous Moose Basin paragneiss may have been more quartzose than shale, or these samples may have incorporated partial melt.
Figure 7. Geochemical diagrams of (A) Fe-index, (B) modified alkali-lime index (MALI), and (C) aluminum saturation index (ASI) for Archean rocks from the Teton Range. Boundaries are from Frost et al. (2001a). Dark- and light-shaded fields represent compositions of the Webb Canyon and Bitch Creek trondhjemitic leucogranitic gneisses, respectively. Values shown for comparison are average tonalite-trondhjemite-granodiorite (TTG; Martin et al., 2005) and average Canadian Archean shale (Cameron and Garrels, 1980). MBG—Moose Basin gneiss; NLG—Northern Layered Gneiss.

Figure 8. (A) Na$_2$O, (B) K$_2$O, and (C) K$_2$O/Na$_2$O as a function of silica content for Archean rocks of the Teton Range. Most Layered Gneisses are less sodic than average tonalite-trondhjemite-granodiorite (TTG). Symbols as the same as in Figure 7.
Rendezvous Gabbro

Analyses of the Rendezvous Gabbro are limited to four samples: three coarse-grained hornblende-plagioclase gabbro samples and one fine-grained garnet-bearing gabbro dike (Table S8 [footnote 2]). The coarse gab- bros are magnesian, and the gabbro dike is weakly ferroan (Fig. 7A), and all samples are metaluminous (Fig. 7C). Trace-element contents are within the range defined by the amphibolites within the southern and northern Layered Gneiss and Moose Basin gneiss (Figs. 8 and 9).

Mount Owen Batholith

Analyses of 21 samples of the Mount Owen batholith are shown in Table S8 (footnote 2). Normative quartz-alkali feldspar-plagioclase (Q-A-P) compositions of the Mount Owen batholith calculated from these analyses plot within the stability field of granite, except for one that lies on the granodiorite-tonalite boundary (Fig. 11). To calculate these normative compositions, we subtracted the normative amount of alkali feldspar needed to combine with normative corundum to make muscovite. Because these rocks are dominated by quartz and feldspar, these calculated compositions should come close to true modal compositions. The modal abundance of plagioclase, however, will be overestimated because small amounts of albite will have dissolved into the alkali feldspar. We observe from Figure 11 that the rocks are true granites and that none of the samples is so poor in silica that it is a quartz monzonite. It is for this reason that we advocate for calling the unit the Mount Owen batholith, rather than the Mount Owen Quartz Monzonite, as named by Reed (1973) and Love et al. (1992).

Compositionally, the Mount Owen batholith is similar to other peraluminous leucogranites defined by Frost et al. (2001a) and Frost et al. (2016b) in that it spans the ferroan-magnesian boundary (Fig. 12A). It is both calc-alkalic and alkali-calcic, but the composition range covers a more narrow range of modified alkali-lime index (MALI) values than do peraluminous leucogranites as a group (Fig. 12B). The variation in Fe-index is likely to reflect degree of melting, with small-degree melts being more enriched in Fe than larger ones (Frost et al., 2001a; Frost et al., 2016b). The variation in MALI is ascribed to a different water fugacity during melting. For example, dehydration melting of micas will produce alkalic melts, whereas with increasing water activity, more plagioclase will be involved in the melting reactions, making the melt more calcic. The Mount Owen batholith shows a distinct increase in aluminum saturation index (ASI) with increasing silica. This probably reflects the fractionation of feldspars, which would increase both SiO₂ and ASI in the residual melt.

The Zr content of the Mount Owen batholith ranges from 40 to 185 ppm, which is similar to, but on the high side of, other peraluminous leucogranites (Frost et al., 2016b). Zr decreases with increasing silica, indicating zircon crystallization with differentiation; however, there is a broad spread in the pattern, suggesting that some zircon in the Mount Owen batholith may be xenocrystic or cumulate (Fig. 13). The zircon-saturation temperature of the rocks is high, ranging from below 750 °C to nearly 900 °C. These results indicate that in the rocks with apparently high zircon-saturation temperatures, some of the zircon is probably cumulate, but even the rocks with the lowest Zr abundances record temperatures that are high for melting of felsic crust.

The Mount Owen batholith is somewhat richer in REEs than the typical Himalayan leucogranite. The high Zr content of the Mount Owen batholith may reflect the presence of cumulate REE-bearing accessory minerals. The rocks have a distinct negative Eu anomaly, which is probably caused by melting in the presence of a plagioclase-bearing restite (Fig. 14).
Sm-Nd Isotopic Data

Sm-Nd isotopic data for the leucogranites of the northern domain have been published by Frost et al. (2016a), and Sm-Nd isotopic data for the Moose Basin gneiss and northern Layered Gneiss were reported by Swapp et al. (2018). Sm-Nd isotopic data for the southern Layered Gneiss, including the “Augen” Gneiss and the Rendezvous Gabbro, were reported by B. Frost et al. (2006). Table S10 (footnote 2) compiles these data along with one unpublished analysis for the southern domain, and \( \varepsilon_{\text{Nd}} \) values calculated for 2685 Ma are plotted on Figure 15.

The Nd contents of samples of southern Layered Gneiss vary widely from 26 to 228 ppm (Table S10 [footnote 2]). The analyzed samples are mainly quartzofeldspathic gneiss, but they also include an amphibolite (99T2) and a biotite paragneiss (07T3). Two samples with high Sm, Nd, and Y contents also have the most positive initial \( \varepsilon_{\text{Nd}} \) of 2.7 and 4.0. All other southern Layered Gneiss samples have initial \( \varepsilon_{\text{Nd}} \) between 0 and 1.0. Nd isotopic compositions of two samples of Rendezvous Gabbro, originally reported in B. Frost et al. (2006), also have \( \varepsilon_{\text{Nd}} \) at 2685 Ma of between 0 and 1.0. These values are within the range of initial \( \varepsilon_{\text{Nd}} \) documented for the northern Layered Gneiss.

Figure 10. Normalized rare earth element patterns for Archean rocks from the Teton Range. (A) Southern Layered Gneiss. (B) Northern Layered Gneiss. (C) Mafic rocks, with peridotite shown by dashed line. (D) Moose Basin gneiss. Also shown as shaded areas are the fields for Webb Canyon and Bitch Creek leucogranitic gneiss samples from the northern Teton Range from Frost et al. (2016a).
and are more radiogenic than \( \varepsilon_{Nd} \) for most samples of Moose Basin gneiss (Fig. 15).

The two analyses of the “Augen” Gneiss have similar Sm and Nd contents. At the estimated time of crystallization (2800 Ma), these gneisses had \( \varepsilon_{Nd} \) of ~0.7 to ~0.5, values that rule out derivation solely from depleted mantle sources. No older rocks have been identified within the Archean exposures of the Teton Range; however, the “Augen” Gneiss analyses lie within the range of values for 2.86–2.80 Ga gneisses from the eastern Beartooth Mountains, Washakie block of the Wind River Mountains, and western Owl Creek Mountains (Frost et al., 2006a).

Mineral Chemistry

Amphibolites with the assemblage Hbl-Pl-Ilm-Qz ± Grt ± Ttn ± Ep ± Bt ± Cum occur throughout the Teton Range. They can be divided into two groups. One has high Fe/(Fe + Mg) whole-rock and mineral compositions and comparatively sodic plagioclase; garnet is present in virtually all of these samples. The second set has lower Fe/(Fe + Mg) ratios and more calcic plagioclase; only a few samples in this category contain garnet (Fitz-Gerald, 2008). Both varieties occur in both the Moose Basin gneiss and in the Layered Gneiss. Fitz-Gerald evaluated pressures and temperatures for these rocks using the barometer of Kohn and Spear (1990) and the thermometer of Dale et al. (2000). Mineral chemistry appropriate for this barometer occurred only in the lower Fe/(Fe + Mg) amphibolites, represented by sample 03T5 and 07T9.

Garnet

Garnet from the garnet amphibolite of the Teton Range is dominantly an almandine-pyrope solid solution with minor amounts of spessartine and grossular end members (Alm\(_{70-77}\)Prp\(_{10-18}\)Sps\(_{3-4}\)Grs\(_{8-13}\); Table S11 [footnote 2]). Some garnet is moderately zoned with margins richer in almandine than the cores.

Plagioclase

Plagioclase from the garnet amphibolite is weakly to moderately normally zoned and ranges from An\(_{30}\) to An\(_{51}\).
Hornblende contains 1.49–1.83 atoms of AlIV pfu and has an $X_{Mg} = 0.444 \pm 0.096$. According to the classification of Leake et al. (1997), the hornblende is magnesiohastingsite to edenite, but the edenite lies very close to the edenite-magnesiohastingsite boundary.
Thermobarometry

Garnet Amphibolites—Moose Basin Gneiss and Layered Gneiss

Two garnet-bearing samples within the prescribed composition range for the barometer yielded 450–600 °C and 4.0 ± 1 kbar (sample 03T5, Moose Basin gneiss) and 550–600 °C and 5.0 ± 1 kbar (sample 07T9, Layered Gneiss; Fitz-Gerald, 2008). These results are consistent with the presence of titanite in the assemblage and the absence of garnet in all but the more iron-rich amphibolites in the region.

Sample 03T5 contained the assemblage Grt-Hbl-Pl-Cum-Qz-Ilm-Ttn (Fig. 16). A pseudosection calculated for this sample using Perple-X-6.7.7 of Connolly (2009) and the thermodynamic data set of Holland and Powell (1998, 2011) restricts the appearance of cummingtonite to pressure (P) <4.6 kbar and temperature (T) in the range 540–600 °C (shaded area in Fig. 17). These pressures and temperatures are consistent with the Kohn and Spear (1990) results above, and we conclude that the garnet amphibolites equilibrated at 540 °C < T < 600 °C and P < 5.0 kbar (Fig. 17). Considering that titanite was stable in this assemblage, we conclude that the titanite ages of ca. 2.62 Ga that we obtained from the northern and southern portions of the range formed during this event, which we call M3.

DISCUSSION

Archean Assembly of the Teton Range

The Archean rocks exposed in the northern Teton Range record geologic events that did not affect the southern part, nor are these events preserved elsewhere in the Wyoming Province. The ca. 2.7 Ga, >12 kbar high-pressure granulite metamorphism is the most striking of these features, particularly because granulite-facies metamorphism is rare throughout the Wyoming Province. Small exposures of granulite-facies rocks of approximately this age occur locally in the northwestern Wind River Range (Frost et al., 2000; Chamberlain and Frost, 2005), but they record lower pressures and temperatures (6–8 kbar, 800–900 °C) than the granulites in the Teton Range. The 2685 Ma Webb Canyon and Bitch Creek leucogranitic orthogneisses in the northern domain are other notable features that also have no counterparts, either in the southern domain or elsewhere in the Wyoming Province. These trondhjemitic orthogneisses are interpreted to have formed by water-excess melting and by dehydration melting of amphibolite (Frost et al., 2016a). Both melting mechanisms are expected in collisional environments, a process that appears not to have operated in the southern domain. Finally, the broad, open folding event (F3) recorded in the northern Teton Range (Fig. 4A) is not present to the south.

The southern Teton Range exposes rock units that are not present in the northern area but that have affinities to rocks elsewhere in the Wyoming Province.
Detrital zircon ages from biotite paragneiss of the southern Layered Gneiss also suggest that the northern and southern parts of the Teton Range had distinct crustal histories for most of the Archean. The dominant age populations of ca. 2.63 Ga and ca. 2.66 Ga and sparse grains with older ages in the range 2.86–2.84 Ga do not match the ages of any known geologic events recorded in the northern area. However, these are the ages of voluminous felsic magmatism elsewhere in the Wyoming Province, including the 2.63 Ga Louis Lake and 2.64–2.71 Ga Bridger batholiths (Frost et al., 1998; Aleinikoff et al., 1989; Table S12 [footnote 2]) and the 2.86–2.84 Ga Bighorn batholith (Frost and Fanning, 2008). The fact that the biotite paragneiss incorporates zircon that could have been derived from these sources, but none from units in the northern area (such as the zircon-bearing Webb Canyon and Bitch Creek leucogranitic gneisses), suggests either that the rocks of the northern Teton Range were not exposed to erosion, or that the biotite paragneiss depocenter was not located where it could receive detritus from the northern domain.

The northern and southern parts of the Teton Range both preserve ca. 2.62 Ga titanite associated with amphibolite-facies metamorphism. As noted above, we conclude that these grains grew during M3 (540 °C < T < 600 °C and P < 5.0 kbar), which is well below the closure temperature of titanite (Frost et al., 2001b). The boundary between the northern and southern domains of the Teton Range is marked by the Moran deformation zone. The biotite paragneiss of Paintbrush Canyon lies within the Moran deformation zone and exhibits a strong NNE-striking foliation that is characteristic of the deformation zone. The older age limit of deformation is well constrained by detrital zircon in both samples 07T3 and 10T4 to be no older than the ages of the youngest detrital zircon grains, which are ca. 2.63 Ga. The available ages constrain the juxtaposition of northern and southern domains to a short interval of ~10 m.y. between deposition of the biotite paragneiss and amphibolite metamorphism, which we interpret as coincident with collision.

The pressure, temperature, and age constraints for M3 allow us to put further constraints on the pressure-temperature-time (P-T-t) path followed by the Teton Range (Fig. 18). The M1 and M2 events occurred during the collisional orogeny in the northern Teton Range between 2695 and 2685 Ma. They are connected by the dashed arrow because they represent part of a single orogenic event (Swapp et al., 2018). The M3 event was associated with the accretion of the rocks of the northern Teton Range with those of the south. The M3 event is separate from the M1 and M2 events. It occurred as much as 60 million years later.

**Implications for the Accretion of the Wyoming Province**

Several crustal fragments of oceanic and continental affinity were accreted along the southern margin of the Wyoming craton (present-day coordinates) between 2.65 and 2.62 Ga, and comprise the southern accreted terranes (Chamberlain et al., 2003, C. Frost et al., 2006b). This accretion occurred prior to the intrusion of undeformed 2.63–2.62 Ga granitic batholiths (C. Frost et al., 2006).
Our data show that the Moran deformation zone occurred at roughly the same time as the later phase of accretion on the southern margin. During this deformation, the rocks in the northern Teton Range, which have a geologic record distinct from the rest of the Wyoming Province, were accreted to the Wyoming Province. This exotic terrane was accreted to the craton from the west at approximately the same time as accretion occurred along the Wyoming province’s southern boundary.

The discovery of the 2.62 Ga Moran deformation zone is a major addition to our understanding of the Neoarchean structural development of the Wyoming Province. From the precise dates and kinematics of individual accretion events that we have been developing from across the province, it is evident that the ancient core of the Wyoming Province was affected by multiple collisions during the span from 2.65 to 2.62 Ga.

Significance of the Mount Owen Batholith

The 2.55 Ga Mount Owen batholith is a typical peraluminous leucogranite: It is a muscovite-bearing, silica-rich granite that is not associated with more mafic rocks. The undeformed batholith intrudes the Moran deformation zone and, as such, may be considered a stitching pluton. Peraluminous granites are commonly interpreted to have formed by melting of a sedimentary source (Chappell and White, 1974), but they also may form by melting of biotite-bearing metaluminous felsic rocks (Miller, 1985; Frost and Frost, 2011). The Nd isotopic composition of the Mount Owen batholith is similar to that of the Layered Gneiss. From this, we conclude that the Mount Owen batholith was melted from a pelitic or intermediate to felsic crustal source with Nd isotopic compositions of the Layered Gneiss.

The Mount Owen batholith is the largest 2.55 Ga granitic intrusion in the Wyoming Province. Imprecise 2.55 Ga dates reported for granite plutons in the Wind River Range (Stuckless et al., 1985) and Granite Mountains (Ludwig and Stuckless, 1978) have not been confirmed by more recent U-Pb geochronology, which suggests older ages of 2.62 Ga (Wall, 2004; Bagdonašas et al., 2016). However, small volumes of 2.55 Ga granitic rocks have been identified in the northern Wyoming Province. Mogk et al. (1988) established an age of 2564 ± 13 Ma for an augen gneiss sill in the North Snowy block of the Beartooth Mountains, and Mueller et al. (2016) identified 2.55 Ga leucogranites emplaced along ductile shear zones in southwest Montana. Mueller et al. (1996) dated a 2.55 Ga hydrothermal zircon from a 3.5 Ga trondhjemitic mylonite from the boundary of the Beartooth-Bighorn magmatic zone and the Montana metasedimentary province.

Foster et al. (2006) also documented 2.55 Ga orthogneisses in the Grouse Creek block, an area underlain by Archean crust located in southern Idaho and northern Utah and Nevada that may have a shared history with the Wyoming Province (Gaschnig et al., 2013). Latest Archean granitic gneiss is exposed in the Albion Mountains, where Strickland et al. (2011) established a 2532 ± 33 Ma age for the Green Creek orthogneiss, a
two-mica granite with alkali-feldspar megacrysts. Circa 2.55 Ga zircon grains are also present in xenoliths in Snake River Plain basalts (Wolf et al., 2005), and as inherited zircon in the Southern Atlantic lobe of the Idaho batholith (Gaschnig et al., 2013).

It is evident that ca. 2.55 Ga deformation and magmatism are widespread in the western United States, but the lack of outcrop and the overprinting by younger deformation and metamorphism make it nearly impossible to characterize the tectonic environment of this event. The best indication for the tectonic origin of this event is the fact that the Mount Owen batholith is undeformed and that we have no obvious indication of metamorphism at 2.55 Ga. It is reasonable to assume that the Mount Owen batholith formed from magmas generated from intermediate to felsic crustal rocks that were thrust beneath the western Wyoming Province at 2.55 Ga.

### CONCLUSIONS

1. The Teton Range exposes two areas with contrasting Archean histories. The northern portion of the range preserves a high-pressure granulite event at ca. 2.70 Ga, followed by amphibolite metamorphism and intrusion of leucogranitic gneisses at 2.68 Ga. These features, which have not been documented elsewhere in the Wyoming Province, have been interpreted as recording a collisional orogeny (Frost et al., 2016a; Swapp et al., 2018). We interpret these rocks to represent a terrane exotic to the rest of the Wyoming Province.

2. The southern part of the Teton Range contains rock units that differ from those in the northern area but that share affinities with mafic and felsic rocks elsewhere in the Beartooth-Bighorn magmatic zone, suggesting that it should be considered a part of the Wyoming Province.

3. Both northern and southern domains experienced amphibolite-facies metamorphism at ca. 2.62 Ga. We interpret this to have occurred when the two domains were juxtaposed along the Moran deformation zone. This Archean assembly of the Teton Range was approximately coeval with the last phase of accretion of the southern accreted terranes along the southern margin of the Wyoming Province. The Moran deformation zone is the only exposed Neoarchean terrane boundary in the Wyoming Province.

4. The intrusion of the 2.55 Ga Mount Owen batholith was the latest igneous event in the Teton Range and the last major Archean event in the Wyoming Province. The next youngest event, the intrusion of the voluminous peraluminous event in the Teton Range, and the last major Archean event in the Wyoming Province, have been interpreted as recorders of a collisional orogeny (Frost et al., 2016a; Swapp et al., 2018). We interpret these rocks to represent a terrane exotic to the rest of the Wyoming Province.

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