



Leucogranites of the Teton Range, Wyoming: A record of Archean collisional orogeny

Carol D. Frost^{a,*}, Susan M. Swapp^a, B. Ronald Frost^a, Lee Finley-Blasi^{a,b},
D. Braden Fitz-Gerald^{a,c}

^a Department of Geology and Geophysics, University of Wyoming, Laramie, WY 82071, USA

^b 151 Kings Road, Evergreen, CO 80439, USA

^c ExxonMobil Corporation, Houston, TX, USA

Abstract

Leucogranitic rocks formed by crustal melting are a prominent feature of collisional orogens of all ages. This study describes leucogranitic gneisses associated with an Archean collisional orogeny preserved in the Teton Range of northwestern Wyoming, USA. These leucogneisses formed at 2.68 Ga, and initial Nd isotopic compositions suggest they are derived from relatively juvenile sources. Two distinct groups of leucogneisses, both trondhjemitic, are identified on the basis of field relations, petrology, and geochemistry. The Webb Canyon gneiss forms large, sheet-like bodies of hornblende biotite trondhjemite and granodiorite. This gneiss is silica-rich ($\text{SiO}_2 = 70\text{--}80\%$), strongly ferroan, comparatively low in alumina, and is characterized by high Zr and Y, low Sr, and high REE contents that define “seagull”-shaped REE patterns. The Bitch Creek gneiss forms small sills, dikes, and plutons of biotite trondhjemite. Silica, Zr, Y, and REE are lower and alumina and Sr are higher than in the Webb Canyon gneiss. These differences reflect different melting conditions: the Webb Canyon gneiss formed by dehydration melting in which amphibole and quartz breaks down, accounting for the low alumina, high FeO, high silica content and observed trace element characteristics. The Bitch Creek gneiss formed by H_2O -excess melting in which plagioclase breaks down leaving an amphibole-rich restite, producing magmas higher in alumina and Sr and lower in FeO and HREE. Both melt mechanisms are expected in collisional environments: dehydration melting accompanies gravitational collapse and tectonic extension of dramatically thickened crust, and water-excess melting may occur when collision places a relatively cool, hydrous lower plate beneath a hotter upper plate. The Archean leucogranitic gneisses of the Teton Range are calcic trondhjemites and granodiorites whereas younger collisional leucogranites typically are true granites. The difference in leucogranite composition reflects the geochemically immature Archean crust that partially melted in the Teton Range compared to the more geochemically evolved rocks typically involved in younger collisional orogens.

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1. INTRODUCTION

Leucogranites and migmatites formed by crustal melting are a hallmark of collisional orogens of all ages, including the Himalayas (e.g., Searle et al., 2010; Reichardt and Weinberg, 2012a), Caledonides (e.g. Roberts, 2003;

Streule et al., 2010), and Proterozoic collision of the Wyoming and Superior provinces (Nabelek et al., 2006). The presence of partial melt in the middle to lower crust contributes to crustal flow during the construction and collapse of orogenic mountain systems (Royden, 1996; Beaumont et al., 2001), and crustal anatexis and magma migration are an effective means of transporting heat from depth to shallower levels in the crust (De Yoreo et al., 1989).

Most crustal melt leucogranites are strongly peraluminous, alkali-calcic granites. For example, the Miocene

* Corresponding author.

E-mail address: frost@uwyo.edu (C.D. Frost).

Makalu granite from the Himalayas (Visona and Lombardo, 2002) is a tourmaline, two-mica alkali-calcic granite proposed to form by partial melting of metapelite. The peraluminous alkali-calcic migmatites and granites of the East Greenland Caledonian orogeny are also interpreted to form by anatexis of paragneisses (Kalsbeek et al., 2001). The Proterozoic Harney Peak biotite- and tourmaline-bearing leucogranites that formed following collision of the Wyoming and Superior provinces are likewise strongly peraluminous. Their more variable major and minor element composition reflects derivation from multiple batches of magmas that were generated from two distinct metasedimentary sources (Nabelek and Glascock, 1995).

A number of different petrogenetic processes have been proposed for collisional leucogranites, despite their petrologic similarities. For example, workers studying the Himalayan leucogranites have attributed their origin to a variety of mechanisms. LeFort et al. (1987) proposed that thrusting and devolatilization of the underthrust Indian plate would cause fluid-present melting in the overriding Asian crust. Harrison et al. (1998) called on shear heating associated with thrusting, and Searle et al. (2010) recognized a possible role for local radiogenic heat production. Multiple studies call on decompression melting following orogenic collapse, a suggestion supported by experimental work by Patiño Douce and Harris (1998), who showed that dehydration melting of metapelitic rocks at 6–8 kb can produce melts that are virtually identical to Himalayan peraluminous leucogranites.

An exhumed Archean collisional orogen underlies the Teton Range of northwestern Wyoming, U.S.A. (Frost et al., 2006). Two contrasting rock assemblages are exposed in the northern part of the range: granulite facies metapelitic rocks that experienced pressures in excess of 10 kb (Fitz-Gerald, 2008) are juxtaposed with amphibolite facies, immature, migmatitic paragneisses. Occupying the boundary between these units are concordant sheets of leucogranite (*sensu lato*) up to 2 km wide in outcrop. These sheets, along with migmatite and small bodies of leucogranite within the paragneisses, record the presence of partial melt in the middle crust during construction and collapse of an ancient mountain-building system. The purpose of this study is to describe the field relations, petrology, and geochemistry of these leucogranites, and determine their age and petrogenesis. The relatively easy access and near complete mid-crustal exposure of this Archean terrain renders it an ideal location to compare and contrast processes and products of melting in ancient collisional orogens with more modern examples.

2. GEOLOGIC SETTING AND SAMPLES

The Precambrian rocks of the Teton Range represent some of the westernmost outcrops of the Archean Wyoming Province (Mueller and Frost, 2006). The Neoproterozoic basement rocks were uplifted and exposed by Laramide tectonism and subsequently by Basin and Range-style extension that continues to the present. The area of exposed Precambrian rocks is relatively small, approximately 50 km by 16 km. Together with the undeformed 2.55 Ga

Mount Owen Quartz Monzonite that intrudes the older gneisses, these rocks compose the scenic peaks of Grand Teton National Park.

The leucogranitic gneisses are located within the northern portion of the Tetons (Love et al., 1992) (Fig. 1). Two gneiss units that predate the leucogranitic gneisses were recognized in this portion of the range by Reed (1973): Layered Gneiss and Augen Gneiss. A portion of the Layered Gneiss preserves evidence of high-pressure metamorphism not present in the rest of the Layered Gneiss. We refer to the gneisses that experienced this high-pressure metamorphism as the Moose Basin gneiss (Frost et al., 2006).

The Moose Basin gneiss consists of kyanite-bearing pelitic gneisses and mafic gneiss that locally preserve relict granulite-facies metamorphism, including garnet- and pyroxene-bearing leucosomes. It is most extensive in the vicinity of Moose Basin in the northern part of the range (Frost et al., 2006), and extends at least as far south as the Window Peak area (Fig. 1). Zr in rutile thermometry on pelitic rocks suggests that the maximum temperatures were approximately 900 °C. The GRAIL (Garnet-Rutile-Aluminosilicate-Ilmenite) assemblage yields pressures in excess of 10 kb (Fitz-Gerald, 2008). Initial ϵ_{Nd} isotopic compositions of the Moose Basin pelitic gneisses are negative, indicating that their protoliths are composed of older continental crust. The age of the Moose Basin gneisses is constrained by dates of two populations of zircon from a leucosome in partially melted mafic granulite: a group of ~3.1 Ga ages from zircon cores interpreted as reflecting the age of the protolith, and a younger group that yields a U-Pb age of 2695 ± 7 Ma, interpreted as the age of peak metamorphism and the partial melting of the mafic rocks (Fitz-Gerald, 2008).

The remainder of the Layered Gneiss, which is exposed in the central and eastern portions of the uplift, consists of interlayered quartzofeldspathic paragneiss, orthogneiss, amphibolite, metagabbro, and metaperidotite (Reed, 1963; Hildebrandt, 1989). The Layered Gneiss is locally migmatitic (Fig. 2). The metamorphic conditions of the Layered Gneiss are poorly constrained because it contains very few assemblages appropriate for thermobarometry. Some samples contain garnet and a few also contain staurolite. We interpret fine-grained mats of chlorite and white mica in two garnet-bearing samples as altered cordierite. The presence of cordierite + garnet in the presence of melt suggests pressures of 5 kb or below (Spear, 1993). This estimate is consistent with the absence of garnet in all but extremely iron-rich associated amphibolites, which indicates that the peak pressure was probably below 6 kb. Leucocratic orthogneiss layers interpreted as leucosomes suggest that peak metamorphic temperatures lay above the reaction $\text{bi} + \text{pl} + \text{q} + \text{vapor} = \text{melt}$, which at less than 6 kb suggests temperatures in the range of 700–750 °C (Fitz-Gerald, 2008). Zircon grains from leucosome in the layered gneiss yield U-Pb ages of 2685 ± 5 Ma (Fitz-Gerald, 2008), providing a younger limit to the age of the Layered Gneiss. The protoliths for the Layered Gneiss are relatively juvenile, with positive initial ϵ_{Nd} isotopic compositions (Frost et al., 2006).

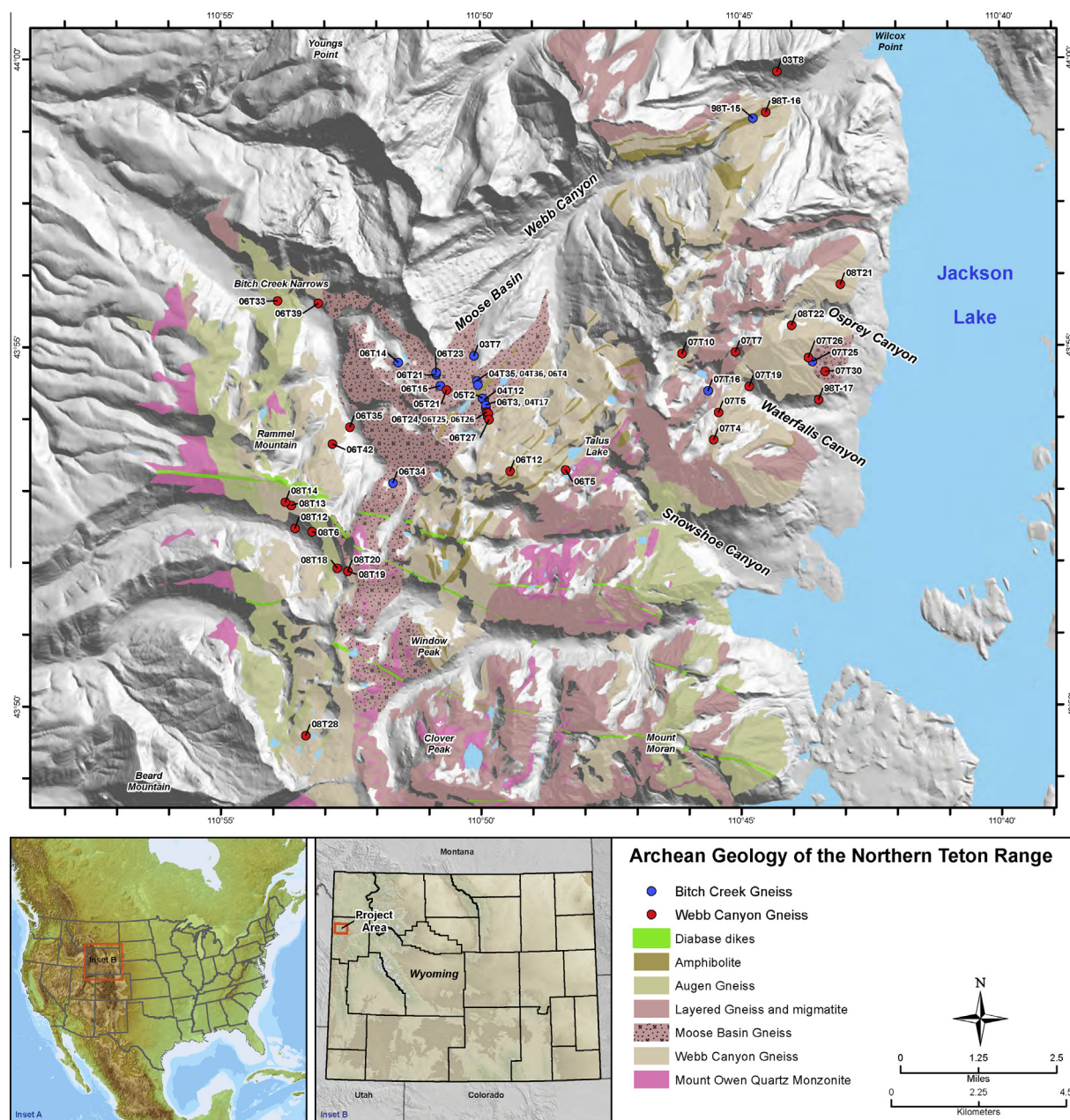


Fig. 1. Geologic map of the northern Teton Range, showing sample locations of Webb Canyon Gneiss and Birch Creek Gneiss samples analyzed in this study. Geology is from Love et al. (1992), with the extent of the Moose Basin Gneiss as identified by the authors shown in the stippled pattern. Note the elongate exposures of Webb Canyon Gneiss in the western, central, and eastern portions of the uplift.

Reed (1973) mapped a third unit he called Augen Gneiss, which occurs in two belts. One extends north from Mount Moran to Snowshoe Canyon and the other lies near the western limits of Precambrian outcrop (Fig. 1). It is a gray-colored, granite to granodiorite gneiss with conspicuous phenocrysts of white alkali feldspar (not an augen gneiss by modern terminology). The major ferromagnesian mineral is biotite with accessory hornblende. It is moderately well-foliated and lineated, with little or no compositional layering. Foliation in the Augen Gneiss is concordant with that of the adjacent layered gneisses

(Miller et al., 1986). A separate area mapped as Augen Gneiss on the northeastern edge of the Precambrian outcrop (Fig. 1) has been reinterpreted as Layered Gneiss and porphyritic Webb Canyon Gneiss (Miller et al., 1986).

Two groups of leucogranitic gneisses are intrusive into these older units. The most extensive is the Webb Canyon Gneiss, a weakly-layered to non-layered leucogranitic orthogneiss that is exposed as a series of elongate, sheet-like bodies approximately 2 km wide and up to 10 km in length (Fig. 1) (Fig. 3a). The Webb Canyon Gneiss generally shows a strong foliation defined by biotite (Fig. 3b).



Fig. 2. Boulder of migmatitic layered gneiss near Talus Lake. Leucosome from layered gneiss in this area was dated by U-Pb zircon at 2685 ± 5 Ma (Fitz-Gerald, 2008), and is interpreted as the time of partial melting of the layered gneiss. Boulder is approximately 1 m across.

Foliation is particularly intense adjacent to contacts, which are everywhere concordant to foliation. Some contacts are sheared, and in others the Webb Canyon Gneiss is complexly interleaved within the Layered Gneiss (Miller et al., 1986). In some localities, particularly in the western portion of the uplift, the Webb Canyon Gneiss contains megacrysts of alkali feldspar; near contacts where the rocks become more strongly foliated these are porphyroclasts. Most Webb Canyon Gneiss contains hornblende and biotite although the easternmost sheet lacks hornblende. Initial ϵ_{Nd} isotopic compositions of three samples of Webb Canyon Gneiss are slightly positive, similar to Layered Gneiss (Frost et al., 2006). U-Pb zircon ages have been determined by Zartman and Reed (1998) and Frost et al. (2006) at 2680 ± 12 Ma and 2673 ± 2 Ma, respectively. The Webb Canyon Gneiss contains numerous concordant bodies of amphibolite up to 3 km long by 200 m wide (Fig. 1; Miller et al., 1986). Because both are strongly foliated, the age relationship of the leucogranitic gneiss and amphibolites cannot be determined from field observations.

A second group of leucogranitic gneiss is recognized in this study. These occur as small sill- or dike-like bodies and plutons within the Moose Basin gneiss or Layered gneiss. We refer to these leucogranitic gneisses as the Bitch Creek gneiss because they are most abundant at the headwaters of Bitch Creek and in nearby Moose Basin¹. They are dominated by quartz and feldspar, and contain little to no alkali feldspar. Bitch Creek gneisses contain biotite

but no hornblende. Although superficially similar in appearance to Webb Canyon Gneiss, Bitch Creek gneisses are in general less intensely foliated (Fig. 3c). In many places they form conformable leucocratic layers within host gneisses (Fig. 3d), and in places preserve evidence of extension (Fig. 3e). Some smaller dike-like amphibolite bodies within Bitch Creek gneiss locally show extensional sense-of-shear (Fig. 3f).

Three episodes of folding are recognized in the northern Teton Range (Frost et al., 2006). The first, F_1 , involves rarely preserved isoclinal fold hinges within paragneiss. The second, F_2 , involves isoclinal folds the limbs of which define the major foliation in the region. The F_1 and F_2 structural elements are present in both the Layered and Moose Basin gneisses. The Webb Canyon and Bitch Creek gneisses do not contain F_1 elements, and F_2 fold hinges were observed only in the eastern exposures of the Webb Canyon gneiss. The F_3 event is manifest as broad, open folds (Fig. 1) that contain no axial planar schistosity and involves all metamorphic rocks in the area. The Moose Basin gneiss occurs in the core of a major F_3 synform and hence lies structurally above the Layered gneiss, although the direct contact is nowhere preserved (Fig. 1).

This study presents data for 31 samples of Webb Canyon Gneiss and 16 samples of Bitch Creek gneiss from the northern Teton Range, including analyses of three Webb Canyon Gneiss samples previously reported by Frost et al. (2006). Locations are shown on Fig. 1, and UTM coordinates are provided in Supplementary Table 1.

3. PETROGRAPHY

3.1. Webb Canyon gneiss

The Webb Canyon Gneiss is fine-grained, with crystals typically less than 2 mm in their longest dimension. The strong foliation of the gneiss is expressed primarily by preferred orientation of biotite, but also by minor compositional layering. Few igneous textures are preserved and 120° grain boundaries between quartz and feldspar suggest that the rock has annealed. Undulose extinction in quartz was likely developed by later minor deformation (Fig. 4a). The Webb Canyon Gneiss consists mainly of quartz and albitic plagioclase, along with 5–20% alkali feldspar. Hornblende and biotite are present in most samples, but the relative abundance varies and some samples lack one or the other mineral. Garnet is present in some samples. Garnet crystals range from subhedral to anhedral, and from more than 10 mm to less than 2 mm in diameter. In some samples garnets are surrounded by halos depleted in ferromagnesian minerals. Allanite, epidote, zircon, titanite, and apatite are common accessory phases. Intergrown biotite and oxide is interpreted to suggest that pyroxene was initially present in some samples.

3.2. Bitch Creek gneiss

Bitch Creek gneiss samples display little to no foliation and are equigranular. Grain-size varies from <1 to 5 mm, although one sample (06T4) is distinctly coarser-grained,

¹ The name Bitch Creek probably originated from a corruption of “biche,” the name for the waterway given by French trappers. (“Biche” is French for doe.) In Owen Wister’s 1902 classic, *The Virginian*, the eponymous hero turns his horse towards the northern Tetons. “Next day we swung southward from what is known to many as the Conant Trail, and headed for that short cut through the Tetons which is known to but a few. Bitch Creek was the name of the stream we now followed.” We name the Bitch Creek gneiss in honor of the Native Americans, trappers, outlaws, and settlers who first explored the area.

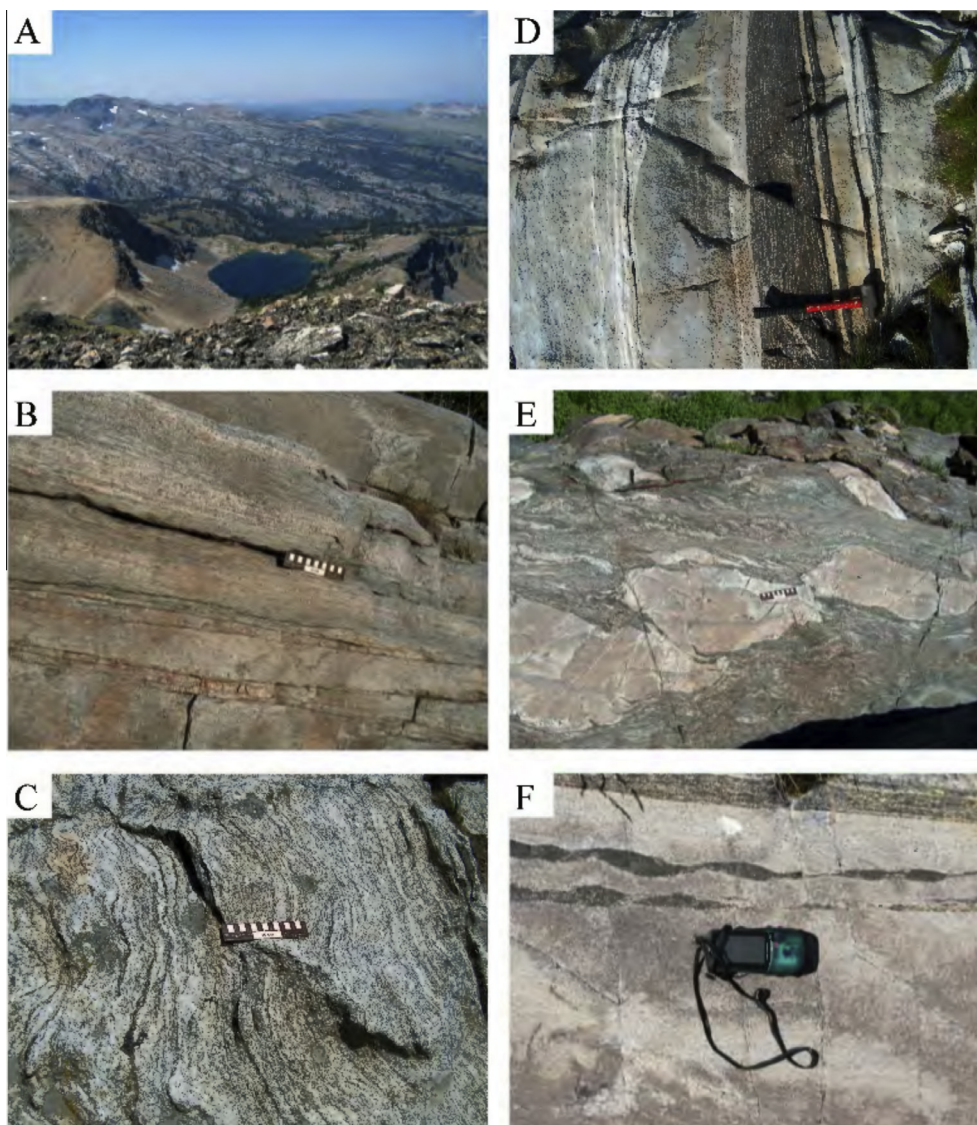


Fig. 3. (A) View to the west from Doane Peak (located west of sample 07T10 on Fig. 1). A 2 km wide tabular body of Webb Canyon Gneiss containing multiple concordant bands of amphibolite gneiss is oriented southwest–northeast and occupies the valley between the photographer and the ridge to the west. (B) Outcrop photograph of strongly foliated Webb Canyon Gneiss sample 07T16, upper Waterfalls Canyon. Scale is 15 cm long. (C) Outcrop photograph of Bitch Creek Gneiss sample 07T25 from above Jackson Lake, showing the less intensely foliation typical of many Bitch Creek Gneiss samples. Scale is 15 cm long. (D) Moose Basin Gneiss showing sills of leucocratic Bitch Creek Gneiss cross-cutting more mafic layers. Hammer is 45 cm long. (E) Boudined Bitch Creek Gneiss layer within the Moose Basin Gneiss showing the more competent behavior of the leucogranite compared to pelitic gneiss. Scale is 15 cm long. (F) Outcrop photograph of Bitch Creek Gneiss sample 04T17 with boudined amphibolite. GPS is 10 cm long.

with all crystals greater than 5 mm. These rocks are generally lighter in color than Webb Canyon gneisses and are distinguished by a lack of hornblende. Bitch Creek gneisses are dominated by quartz and plagioclase; alkali feldspar occurs only locally (Fig. 4b). Biotite is the primary ferromagnesian mineral. Some samples are garnet-bearing. Garnet is poikilitic, strongly resorbed, and surrounded by a light-colored halo in which biotite is absent (Fig. 4c). Accessory minerals include zircon and allanite, but these are less abundant than in Webb Canyon gneisses.

3.3. Summary

Both the Webb Canyon and Bitch Creek gneisses are dominated by plagioclase and quartz, and mafic minerals make up a small proportion of the rock. Because of the potassium contained in biotite, IUGS classification based on normative compositions places them as granite and granodiorite (Miller et al., 1986; Finley-Blasi, 2009). However, based on modal mineralogy Bitch Creek gneisses are

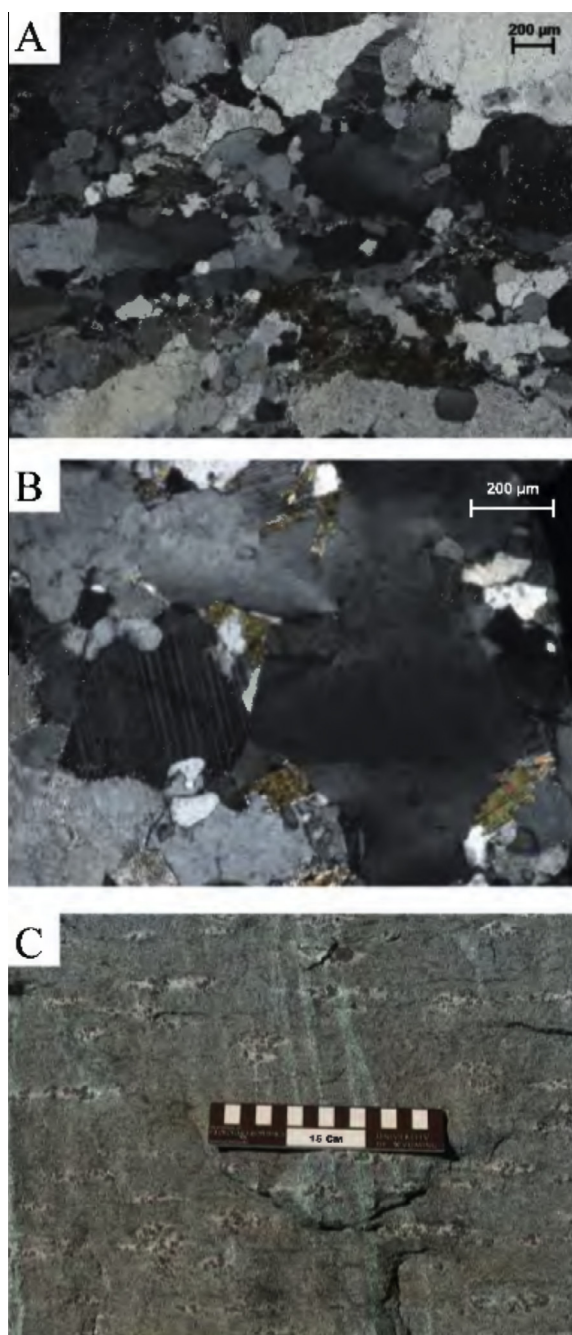


Fig. 4. (A) Photomicrograph of Webb Canyon Gneiss sample 07T7 showing foliation defined by biotite and lack of igneous textures. Serrated grain boundaries and core-and-mantle structure (especially in quartz) are evidence of dynamic recrystallization followed by incomplete recovery. Crossed-polarized light. (B) Photomicrograph of Bitch Creek Gneiss sample 07T25 showing relatively large grain size of quartz and plagioclase. Biotite defines a weak foliation in this rock. (C) Outcrop photograph of garnet-bearing Bitch Creek Gneiss showing resorbed garnet surrounded by haloes of quartz and feldspar, a texture interpreted as indicative of decompression. Upper Moose Basin area.

trondhjemites and the Webb Canyon gneisses, with 5–20% alkali feldspar, are trondhjemites and granodiorites. For this reason, our description of the gneisses as “leucogran-

ites” uses the term in the broad sense to denote a light-colored quartz- and feldspar-rich rock. The recognition that these gneisses are trondhjemitic identifies a fundamental difference between these leucogneisses and the true leucogranites identified within younger collisional orogens.

4. GEOCHEMISTRY

Although Webb Canyon and Bitch Creek gneisses are only subtly distinguishable in the field and in thin section, certain aspects of their major and trace element characteristics are quite different (Table 1). Webb Canyon gneisses are silica-rich, with SiO_2 varying from 70% to 80%. Bitch Creek gneisses have overlapping but on average lower SiO_2 , varying from 64% to 75% (Table 1; Fig. 5a). Both groups of leucogranitic gneisses are calcic, and most samples are peraluminous (Fig. 5b and c). Webb Canyon gneisses are distinguished by low Al_2O_3 compared to Bitch Creek gneisses (Fig. 5d). Webb Canyon gneisses have higher K_2O and lower Na_2O (Fig. 5e and f). They also are mostly ferroan because of their lower MgO compared to Bitch Creek gneisses (Fig. 5a). Six Webb Canyon gneisses samples have higher MgO than other Webb Canyon samples and plot in the magnesian field (Fig. 5a, see also Fig. 6b). These were collected from the easternmost sheet-like body of Webb Canyon gneiss accessible from Waterfalls and Osprey canyons. This geochemical distinction suggests that different sheets of Webb Canyon gneiss may have somewhat different geochemical compositions, a proposal that is consistent with the observation of Love et al. (1992), who noted that lack of hornblende in this portion of the Webb Canyon means it “probably should have been included in a separate map unit”.

Some trace element characteristics also distinguish Webb Canyon from Bitch Creek gneisses. Sr contents of Webb Canyon gneiss are relatively low, from 20 to 200 ppm, whereas Bitch Creek gneisses have higher Sr contents averaging 300 ppm. Similar Rb contents in both sets of gneisses yield higher Rb/Sr in the Webb Canyon gneisses (Fig. 6a). Zr contents of Webb Canyon gneisses are high (240–590 ppm), whereas Zr contents of Bitch Creek gneisses are lower (Fig. 6b). Zircon saturation temperatures (Watson and Harrison, 1983) are correspondingly higher for the Webb Canyon gneiss than for the Bitch Creek gneiss (875–1000 °C versus 750–900 °C; Fig. 6c). Yttrium is also higher in Webb Canyon gneiss compared to Bitch Creek gneiss (Fig. 6d).

The geochemical distinction between Webb Canyon and Bitch Creek gneisses also extends to their rare earth element characteristics (Fig. 7). REE contents of Webb Canyon gneisses are REE-enriched and relatively flat, and most samples have a deep negative Eu anomaly, producing a pattern that has been described as “seagull” shaped by Glazner et al. (2008) (Fig. 7a). Five Webb Canyon Gneiss samples have slightly lower REE contents and lack pronounced Eu anomalies (Fig. 7b); these samples contain abundant plagioclase, most are garnet-bearing, and all are higher in CaO and lower in K_2O and Rb/Sr than other Webb Canyon Gneiss samples. A sixth sample plotted on Fig. 7b has a pattern similar in shape to the typical Webb Canyon Gneiss

Table 1
Geochemical data for leucogranitic gneisses of the Teton Range.

Sample Analysis method	Bitch Creek Gneiss															
	98T-15 UW	03T7 ALS	04T12 ALS	04T17 UW	04T35 ALS	04T36 ALS	05T2 ALS	06T3 ALS	06T4 UW	06T14 UW	06T15 UW	06T21 UW	06T23 UW	06T34 UW	07T16 UW	07T25 XRAL
SiO ₂	71.58	68.3	74.5	72.32	72.4	70.3	64.2	72.8	70.54	71.86	69.00	72.13	68.73	72.96	72.47	72.86
TiO ₂	0.22	0.23	0.1	0.25	0.28	0.25	0.7	0.15	0.47	0.35	0.29	0.29	0.51	0.37	0.35	0.20
Al ₂ O ₃	16.26	17.15	13.3	15.97	14.85	14.85	16.2	15.2	17.60	15.72	18.08	16.12	16.35	15.46	15.15	14.84
FeO ^{tot}	1.64	1.06	1.17	1.86	2.52	3.47	5.38	1.48	1.60	2.16	1.85	1.85	4.87	2.01	3.42	1.65
MnO	0.02	0.01	0.02	0.03	0.03	0.06	0.04	0.02	0.05	0.02	0.02	0.08	0.07	0.03	0.04	0.03
MgO	0.50	0.91	0.19	0.56	0.71	0.69	1.61	0.36	0.64	0.57	0.65	0.65	1.13	0.51	0.79	0.50
CaO	2.65	4.18	1.81	2.85	3.49	3.51	5.72	3.05	3.64	3.09	3.90	2.39	4.49	2.59	3.84	1.72
Na ₂ O	6.02	4.54	4.32	5.27	4.05	3.95	3.7	5.12	5.57	5.19	5.34	5.59	3.45	5.25	3.79	4.81
K ₂ O	0.97	0.87	2.24	1.20	0.59	0.44	0.63	0.85	0.69	0.68	0.86	0.84	0.20	0.86	0.62	2.62
P ₂ O ₅	0.06	<0.01	0.01	0.07	0.08	0.07	0.36	0.05	0.07	0.09	0.05	0.09	0.16	0.07	0.11	0.07
LOI	0.12	1.04	0.36	0.08	0.49	0.31	0.6	0.51	0.11	0.10	0.13	0.09	0.09	0.04	0.10	0.69
Total	100.04	98.29	98.02	100.47	99.49	97.90	99.14	99.59	100.97	99.82	100.18	100.11	100.03	100.15	100.66	100.00
Fe-index	0.765	0.539	0.860	0.768	0.780	0.834	0.770	0.804	0.714	0.790	0.741	0.739	0.812	0.798	0.813	0.766
MAFI	4.336	1.230	4.750	3.625	1.150	0.880	-1.390	2.920	2.623	2.782	2.296	4.045	-0.839	3.528	0.567	5.707
ASI	1.031	1.071	1.037	1.054	1.088	1.112	0.944	1.021	1.065	1.056	1.076	1.116	1.164	1.083	1.092	1.069
Rb	25.4	14.1	69.3	50.3	13	10.4	15.1	36.2	17.7	17.1	22.5	110	4.7	19.4	19.9	51.9
Sr	176.2	891	167	217.5	221	210	181.5	322	472.2	349.3	454.9	241.5	225.8	325.3	171.6	231
Ba	231.1	438	879	490.7	262	244	159.5	385	463.2	246.5	258.8	257.4	45.3	384.5	547.4	945
Y	12.9	<0.5	3.8	5.5	4.4	10.6	25.2	3.2	3.2	2.8	3.2	6.6	5.3	3.2	19.5	9.8
Nb	3.2	0.4	5	3.5	8.8	8	16.6	3.6	1.6	1.2	1.2	4.9	6.2	3.1	6.4	2
Zr	113.9	168	71	170	214	197	397	118	90.5	194.2	122.9	165	295.1	155.9	290.8	116
Zircon saturation T	800	835	762	839	864	855	908	803	781	852	807	839	898	833	895	805
La	17.9	18.3	8.2	26.7	25.5	25.3	34.4	13.5	12.2	19.1	6.4	22.9	32.7	23	53.9	25.6
Ce	31.2	23	16.2	44.8	46.4	44.8	72.5	22.3	23.1	32	10.9	38.2	59.4	40.3	89	42.9
Pr	3.51	1.76	1.54	4.54	4.73	4.57	8.15	2.11	2.52	3.17	1.16	3.83	6.24	3.98	9.47	4.66
Nd	13.5	4.6	5.5	15.9	16.2	15.2	30	6.9	9.5	10.5	4.2	13.1	21.7	13.8	32.6	16
Sm	2.9	0.27	1.03	2.6	2.61	2.9	6.04	1.07	1.5	1.4	0.9	2	3.2	2	5.7	3.3
Eu	1.01	1.08	0.54	0.86	0.97	0.86	1.69	0.41	0.56	0.51	0.44	0.63	1.15	0.55	1.35	0.82
Gd	3.18	0.06	0.85	2.2	1.89	2.65	5.9	0.73	1.34	1.26	1.05	2.09	2.81	1.56	5.69	2.89
Tb	0.56	0.01	0.12	0.33	0.22	0.38	0.86	0.09	0.18	0.18	0.16	0.37	0.39	0.22	0.82	0.42
Dy	2.74	<0.05	0.58	1.26	1.02	2.15	4.79	0.49	0.75	0.64	0.76	1.68	1.39	0.76	4.01	2.06
Ho	0.56	<0.01	0.12	0.24	0.17	0.38	0.92	0.1	0.11	0.12	0.17	0.36	0.27	0.14	0.75	0.33
Er	1.59	0.04	0.35	0.64	0.48	0.99	2.5	0.31	0.25	0.34	0.42	0.96	0.72	0.41	2.23	1.03
Tm	0.21	0.01	0.05	0.08	0.06	0.13	0.38	0.06	<0.05	<0.05	0.06	0.12	0.08	<0.05	0.31	0.13
Yb	1.5	0.07	0.39	0.6	0.37	0.8	2.15	0.41	0.3	0.3	0.4	0.8	0.7	0.4	2.1	1
Lu	0.3	0.01	0.07	0.11	0.06	0.12	0.33	0.06	<0.05	<0.05	0.06	0.11	0.1	<0.05	0.27	0.12

(continued on next page)

Table 1 (continued)

Sample Analysis method	Webb Canyon gneiss														
	98T-16 CJES	98T-17 CJES	03T8 UW	05T21 ALS	06T5 UW	06T12 UW	06T24 UW	06T25 UW	06T26 UW	06T27 UW	06T33 XRAL	06T35 UW	06T39 UW	06T42 UW	07T4 XRAL
SiO ₂	76.20	79.40	77.52	77	77.00	78.12	70.50	70.20	73.69	77.69	78.76	72.06	72.89	74.44	76.47
TiO ₂	0.19	0.11	0.16	0.09	0.10	0.14	0.53	0.46	0.35	0.14	0.21	0.57	0.50	0.33	0.11
Al ₂ O ₃	10.40	10.80	11.43	11.35	11.62	11.42	12.25	12.65	11.87	11.37	10.31	12.11	12.36	12.10	11.11
FeO ^{tot}	3.67	2.46	2.38	1.71	2.75	1.86	7.06	7.39	5.96	2.69	2.68	5.72	4.80	3.43	3.37
MnO	0.05	0.02	0.03	0.01	0.05	0.04	0.14	0.14	0.13	0.06	0.00	0.10	0.09	0.07	0.01
MgO	0.06	0.10	0.01	0.1	0.09	0.01	0.13	0.16	0.11	0.07	0.21	0.29	0.30	0.13	1.16
CaO	1.46	0.76	1.03	0.69	0.83	1.61	3.44	3.77	2.70	1.04	0.93	2.59	2.09	1.64	0.25
Na ₂ O	3.36	3.98	4.01	2.78	3.95	4.65	4.28	4.12	4.32	4.10	3.40	2.66	2.58	3.29	2.88
K ₂ O	2.79	2.37	3.06	4.2	3.20	1.76	0.73	0.68	0.77	2.43	2.68	2.91	3.45	3.72	3.52
P ₂ O ₅	0.01	0.00	0.01	0.01	0.03	0.01	0.05	0.09	0.04	0.01	0.10	0.12	0.09	0.04	0.00
LOI	0.15	0.10	0.03	0.26	0.09	0.06	0.05	0.06	0.04	0.12	0.72	0.80	0.10	0.04	1.11
Total	98.34	100.10	99.65	98.20	99.70	99.67	99.15	99.70	99.98	99.71	100.00	99.91	99.23	99.22	100.00
Fe-index	0.98	0.96	0.997	0.945	0.969	0.995	0.982	0.978	0.982	0.975	0.929	0.951	0.942	0.963	0.744
MALI	4.69	5.59	6.041	6.290	6.325	4.801	1.568	1.033	2.394	5.492	5.155	2.972	3.947	5.369	6.154
ASI	0.93	1.03	0.971	1.094	1.013	0.916	0.870	0.881	0.924	1.009	1.012	0.991	1.050	0.974	1.235
Rb	41	37	64.5	104.5	82.5	49.5	13.1	8.8	22.5	41.9	86.1	71.9	81.1	111.1	88.6
Sr	96	35	41.9	87.5	33.2	59.4	156.2	177.8	117.5	50.4	103.6	134.8	125.8	98.3	29.6
Ba	1220	350	825.7	1400	654.9	687.8	246.2	285.5	398.9	814.6	821.5	3466.2	3152.7	2665.2	801
Y	275	156	188.4	184	181.3	213	59.1	57.5	96.7	221.5	169.5	161.8	154.2	189.8	157
Nb	13	38	50.4	22.7	41.1	47.9	8.3	8.1	8.6	26.7	55.5	31.9	29.4	28.2	49
Zr	508	402	266.3	439	245.3	286.3	781.3	827	587	305.8	487	530.3	464.9	340.4	388
Zircon saturation T	949	933	883	945	877	887	987	995	962	900	953	954	945	906	939
La	132	–	102	60.7	85.1	80.2	16.8	19	25.9	84.8	129	205	223	158	97.6
Ce	313	–	220	129.5	184	173	44.1	48	71.7	225	272	432	445	327	203
Pr	42.2	–	26.5	15.7	23	22.1	6.32	6.91	10.8	28.9	35.6	51.1	54.5	39.9	26.8
Nd	190	–	111	63.3	95.9	94.8	32.1	34.1	57	131	153	207	216	165	111
Sm	47.5	–	26.8	17.65	23	23.8	9.3	9.8	16.4	33.7	35.5	42.2	41.6	36.6	28
Eu	6.75	–	4.09	2.69	3.26	3.52	7.85	7.72	7.79	4.89	2.96	9.99	9.23	8.03	2.62
Gd	58.2	–	31.9	23.5	27.7	28.6	11.8	12	21.3	40.3	37.7	46.2	43.9	40.6	30.4
Tb	11	–	6.35	4.36	5.63	5.9	2.25	2.24	4	8.01	6.36	7.94	7.56	7.7	5.25
Dy	58.1	–	34.5	30.6	31.5	34	12.1	11.9	21.2	44.3	29.5	38.3	36	39.2	32.5
Ho	12.4	–	7.9	6.53	7.12	8.1	2.73	2.75	4.73	10.1	6.16	8.09	7.6	8.63	6.27
Er	33.6	–	23.8	17.7	21.6	25.5	7.95	8.49	13.8	30.1	16.4	22	20.6	24.8	20.1
Tm	3.92	–	3.29	2.28	2.97	3.54	1.1	1.14	1.8	3.97	1.94	2.65	2.53	3	2.91
Yb	22.7	–	21.8	13.05	20	24.1	7.7	8.2	12.6	25.6	11.7	16.4	15.9	18.6	19.4
Lu	4.2	–	3.8	1.86	3.47	4.16	1.54	1.7	2.44	4.32	1.98	3.05	2.76	3.21	2.38

Sample	07T5	07T7	07T10	07T19	07T26	07T30	08T6	08T12	08T13	08T14	08T18	08T19	08T20	08T21	08T22	08T28
Analysis method	XRAL	UW	XRAL	UW	XRAL	XRAL	UW	UW	UW	UW	UW	UW	UW	UW	UW	UW
SiO ₂	78.11	79.54	77.53	78.44	79.56	72.76	74.02	77.13	76.06	74.21	77.86	76.81	75.09	78.36	78.20	78.48
TiO ₂	0.13	0.12	0.11	0.10	0.16	0.44	0.34	0.21	0.21	0.23	0.18	0.21	0.35	0.17	0.18	0.15
Al ₂ O ₃	11.14	10.84	11.40	11.45	10.17	12.84	11.59	12.40	11.83	12.58	12.15	12.19	12.41	10.68	11.18	11.17
FeO ^{tot}	2.16	2.55	2.52	1.94	3.66	5.25	5.73	2.40	4.58	4.80	1.89	2.42	3.55	2.51	2.69	2.52
MnO	0.03	0.04	0.04	0.02	0.10	0.12	0.13	0.03	0.08	0.08	0.04	0.02	0.05	0.04	0.03	0.03
MgO	0.56	0.49	0.04	0.56	0.05	0.24	0.06	0.08	0.06	0.04	0.10	0.09	0.17	0.58	0.67	0.04
CaO	0.21	0.23	0.58	0.36	2.04	4.22	2.33	0.80	1.39	1.72	1.10	0.84	1.66	0.38	0.37	0.75
Na ₂ O	2.86	3.41	3.90	3.51	3.43	3.36	3.62	4.02	4.15	3.95	3.88	3.04	2.73	3.01	3.33	3.23
K ₂ O	3.97	2.43	3.30	3.36	0.45	0.24	1.35	2.98	1.70	2.44	2.56	3.84	3.19	3.71	2.82	3.13
P ₂ O ₅	0.00	0.01	0.00	0.02	0.03	0.05	0.04	0.03	0.03	0.04	0.01	0.02	0.07	0.01	0.01	0.01
LOI	0.82	0.27	0.58	0.08	0.35	0.47	0.05	0.07	0.05	0.07	0.09	0.09	0.07	0.08	0.09	0.05
Total	100.00	99.93	100.00	99.84	100.00	100.00	99.25	100.15	100.14	100.15	99.86	99.57	99.33	99.52	99.55	99.55
Fe-index	0.794	0.839	0.984	0.778	0.986	0.955	0.990	0.967	0.987	0.991	0.948	0.965	0.954	0.812	0.801	0.985
MALI	6.615	5.606	6.624	6.500	1.841	−0.611	2.639	6.200	4.455	4.674	5.339	6.036	4.261	6.329	5.775	5.611
ASI	1.186	1.252	1.033	1.138	1.034	0.953	0.996	1.099	1.057	1.026	1.089	1.141	1.133	1.106	1.215	1.111
Rb	79.6	34.2	103	84.3	10.9	2.8	45.5	73.4	67	65.7	48.9	164.1	113.8	77.6	41.8	92.6
Sr	38.6	20.9	31.9	28.3	96.8	140	93.2	75.1	62.2	74.1	59.1	57.8	99.4	21	25.9	40.2
Ba	921	748.2	711	718	258	101	579.4	1368.2	669.5	839.6	722.9	1787.6	2667.7	816.6	852.7	730.2
Y	188	177.1	190	183.1	95.1	44	187.8	41.7	287.6	303.9	174.9	166.3	165.9	172.3	131.3	119.2
Nb	35	46.2	43	42.9	14	7	29.2	26.2	29.4	27.6	33.9	46.5	30.6	40	53.3	39.4
Zr	393	321.1	351	306.9	403	572	557.7	329.1	463.8	492.9	242.1	327.9	366.1	281.2	339.8	318.2
Zircon saturation T	939	920	917	909	934	961	963	913	948	951	881	915	925	898	924	911
La	85	125	90.1	86.6	28.9	18.2	64	80.7	124	90	82.9	123	126	87.2	142	113
Ce	189	284	192	185	68.1	37.2	161	161	302	252	172	312	267	190	304	244
Pr	25.9	38.1	26.6	25.4	9.85	4.83	22.3	17.8	45.6	38.3	23.7	31.4	32.4	25.3	35.8	32.6
Nd	115	155	114	108	44.9	20	108	70.5	216	188	100	122	132	109	146	140
Sm	31	36	31.1	28.4	12.2	5	29.8	15.1	55.7	51.5	24.2	28.1	29.5	27.3	33.4	33.7
Eu	3.47	3.7	3.54	2.82	3.38	2.21	7.54	2.78	8.18	7.81	3.42	5.05	7.27	3.89	4.16	4.01
Gd	34.5	37.4	34.8	30.4	14.4	6.24	36.9	14.8	66.5	62.1	28.1	30.9	33.4	32	36.3	36.4
Tb	6.13	6.23	6.13	5.35	2.43	1.08	7.11	2.48	12.8	11.9	5.65	6.2	6.44	6.11	6.53	6.24
Dy	39.1	36.5	39.3	33	17.5	7.5	37.7	10.8	64.6	64	31.6	33.5	34.1	32.6	30.8	28.1
Ho	7.75	6.64	7.61	6.43	3.64	1.53	8.57	2.06	13.8	14.2	7.29	7.52	7.59	7.3	6.24	5.5
Er	25.4	19.7	24.4	20.8	12.5	5	24.9	5.53	38.1	39.9	22.2	22.4	21.9	21.9	17	14.4
Tm	3.7	2.76	3.67	3.04	1.91	0.81	3.25	0.67	4.73	4.87	2.97	2.91	2.74	2.99	2.14	1.61
Yb	24.5	17.6	24.8	20.8	13.1	5.5	21.7	4.7	30.2	29.8	19.2	18.6	17.1	19.1	13.9	9.9
Lu	2.89	2	2.99	2.53	1.71	0.68	3.92	0.83	5.06	5	3.38	3.12	3	3.41	2.44	1.57

Oxides in weight percent; trace elements in parts per million. Zircon saturation temperatures are calculated from [Watson and Harrison \(1983\)](#).

Major elements were analyzed by several different methods. UW = analysis by X-ray fluorescence at the University of Wyoming.

ALS = analysis by inductively coupled plasma-optical emission spectroscopy at ALS Minerals Division, Reno, Nevada.

XRAL = analysis by inductively coupled mass spectrometry (ICP-MS) at XRAL Laboratories, Don Mills, Canada, and results normalized to 100%. CJES = data previously published in [Frost et al. \(2006\)](#). All trace elements were analyzed by ICP-MS.

UW and XRAL samples were analyzed at XRAL Laboratories; ALS samples were analyzed at ALS Minerals.

LOI = Loss on ignition; Fe-index = $\text{FeO}^{\text{tot}}/(\text{FeO}^{\text{tot}} + \text{MgO})$; MALI = Modified Alkali Lime Index ([Frost et al., 2001](#)); ASI = Aluminum saturation index.

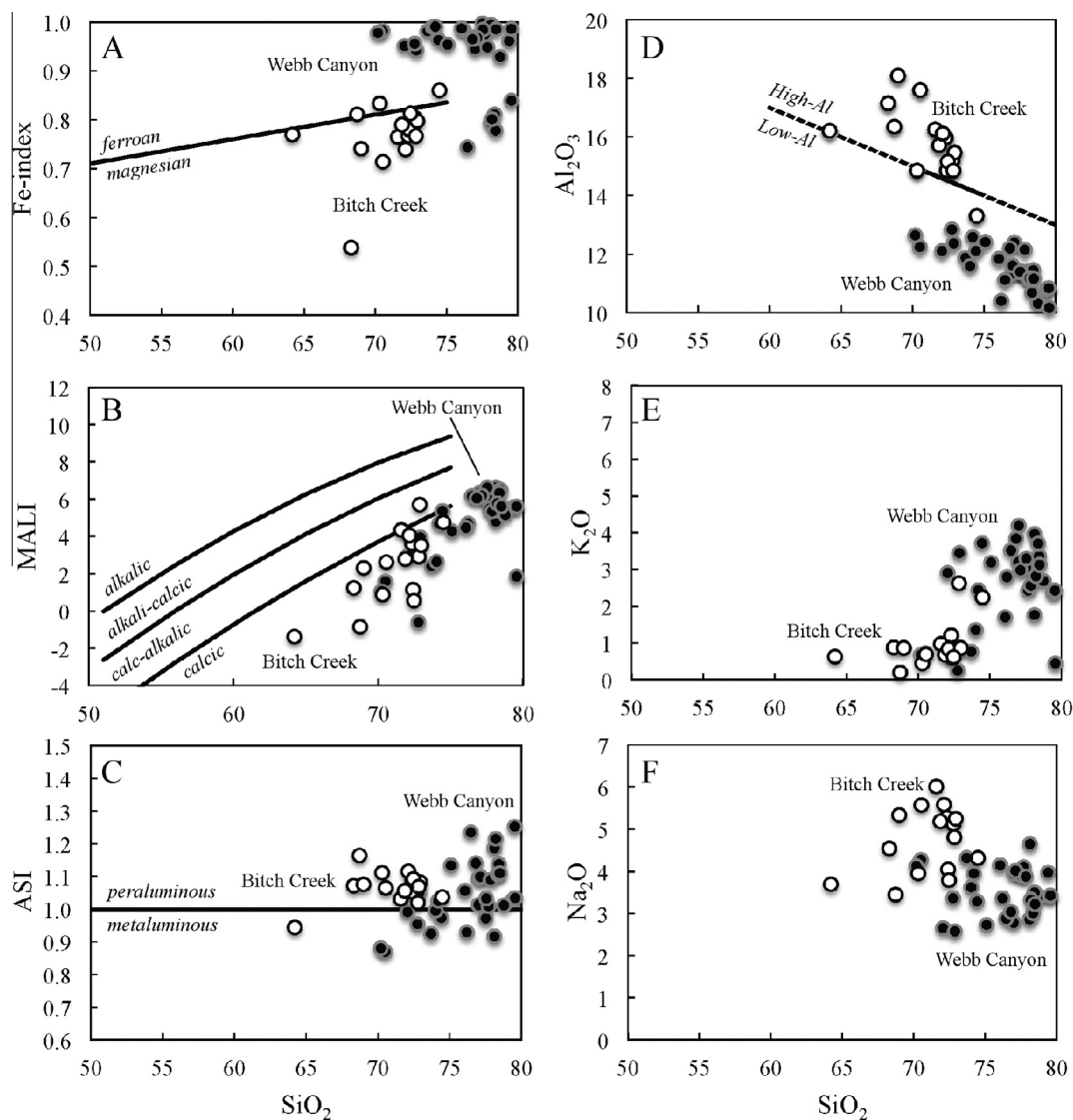


Fig. 5. (A) Fe index ($\text{FeO}^{\text{tot}}/(\text{FeO}^{\text{tot}} + \text{MgO})$) of Webb Canyon and Bitch Creek gneisses. Most Bitch Creek gneisses are magnesian, whereas most Webb Canyon gneisses are strongly ferroan. The six magnesian Webb Canyon gneisses are collected from the easternmost sheet of Webb Canyon gneiss exposed in the Tetons, suggesting that not all sheets have the same geochemical composition. Boundary between fields from Frost and Frost (2008). (B) MALI (Modified Alkali-Lime Index; $\text{Na}_2\text{O} + \text{K}_2\text{O} - \text{CaO}$; Frost et al., 2001) showing calcic nature of both Webb Canyon and Bitch Creek gneisses. (C) ASI (Aluminum-saturation index) for Webb Canyon and Bitch Creek gneisses, showing that most but not all are peraluminous. (D) Alumina content of Webb Canyon and Bitch Creek gneisses showing that Bitch Creek gneisses fall into Barker (1979) high-Al field whereas Webb Canyon gneisses are in the low-Al field. Bitch Creek gneisses are lower in K_2O (E) and higher in Na_2O (F) than Webb Canyon gneisses, reflecting the paucity of K-feldspar in Bitch Creek gneisses. Data from Table 1.

but with lower total REE abundances. Bitch Creek gneisses have distinctly different REE characteristics: REE abundances are lower, Eu anomalies are modest and include both small negative and small positive anomalies, and patterns are LREE-enriched and HREE-depleted. One sample (indicated by a dashed line on Fig. 7c) with a strong positive Eu anomaly and very low HREE abundances also has an extreme Sr content (890 ppm) and textural evidence of plagioclase accumulation.

5. GEOCHRONOLOGY

Two U-Pb zircon age determinations of Webb Canyon Gneiss of 2680 ± 12 Ma and 2673 ± 2 Ma were produced by Zartman and Reed (1998) and Frost et al. (2006), respectively. Both of these dates are based upon TIMS multi-grain analyses. This study reports a further 7 U-Pb zircon age determinations of Webb Canyon Gneiss and 3 from Bitch Creek gneiss. One analysis (06T12) was obtained by LA-ICP-MS and the others by SHRIMP. Results are sum-

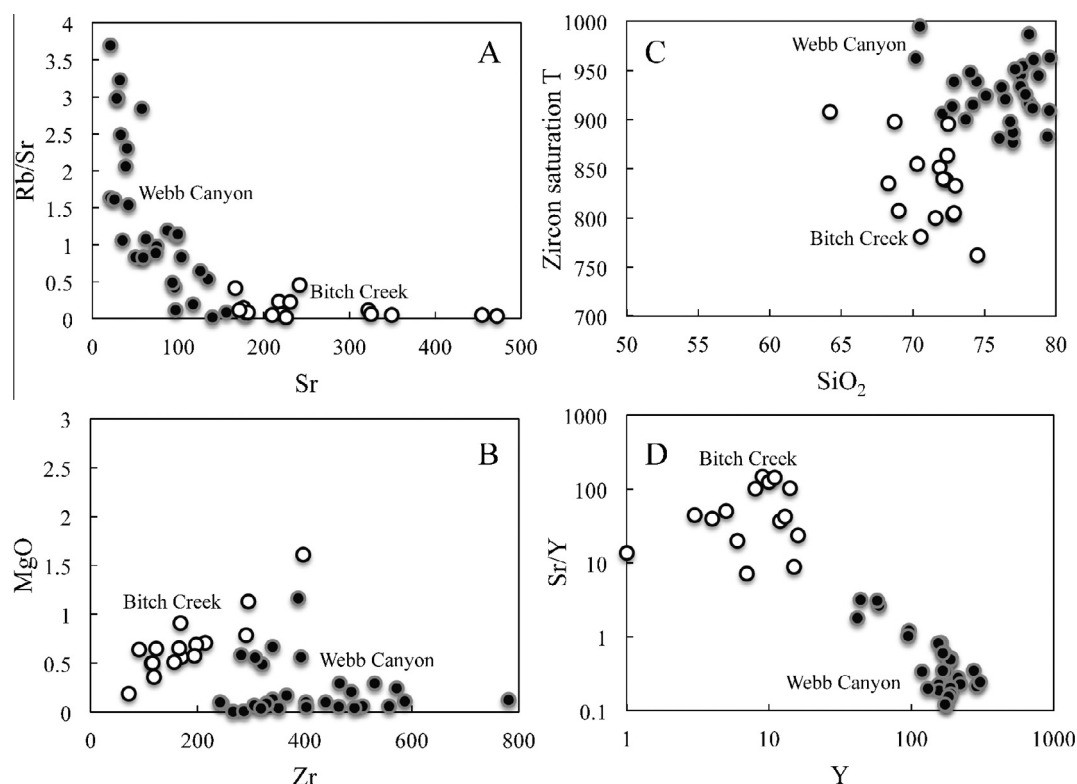


Fig. 6. Trace element and other characteristics of Webb Canyon and Bitch Creek gneisses. (A) Rb/Sr plotted against Sr (ppm). (B) Weight% MgO plotted against Zr (ppm) (C) Zircon saturation temperatures calculated from [Watson and Harrison \(1983\)](#). (D) Sr/Y plotted against Y (ppm). Data from [Table 1](#).

marized on [Table 3](#); data for individual samples are provided in [Supplementary Tables S2a–j](#). Both concordia and weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ ages are reported. Because lower intercepts on concordia diagrams are within error of zero age, Pb loss appears to have been recent and therefore weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ ages are likely to be geologically meaningful.

Evidence for inherited zircon was sparse in these samples. In only two samples was one older grain identified among those analyzed, and both grains were only slightly older, ~ 2.7 Ga. Uranium contents of the zircon extracted from these samples was variable. The highest U contents were measured in samples of the Bitch Creek gneiss. Because alpha-recoil damage in Archean grains with U contents greater than 500 ppm can render the grains metamict and susceptible to lead loss, in some samples only results from grains with lower U contents were interpreted. For one sample in which all zircon had $\text{U} > 500$ ppm (06T34), no age information was obtained.

5.1. Webb Canyon gneiss

Sample 03T8 is a sample of moderately deformed, well-foliated, fine-grained orthogneiss of type Webb Canyon Gneiss from Webb Canyon at the northeast end of the study area. Fourteen analyses on twelve large grains yielded individual $^{207}\text{Pb}/^{206}\text{Pb}$ ages that range from 2666 to

2694 Ma. Thirteen of the fourteen individual measurements are less than 3% discordant. A weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2683 ± 3 Ma (MSWD = 1.6) and a upper intercept concordia age of 2685 ± 4 Ma (MSWD = 1.0) are determined from all fourteen analyses ([Fig. 8a](#)).

Sample 06T12 is from the center of the middle exposure of Webb Canyon Gneiss ([Fig. 1](#)), and is a moderately deformed sample of typical biotite-bearing Webb Canyon Gneiss. Two zircon morphologies are observed, one small (50–150 μm) and rounded and the other large (200–450 μm), euhedral, and prismatic. Thirty individual grains were analyzed by LA-ICP-MS. No correlation exists between grain morphology and $^{207}\text{Pb}/^{206}\text{Pb}$ age. A weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2674 ± 5 Ma (MSWD = 0.3) and concordia age of 2675 ± 8 (MSWD = 0.1) were obtained ([Fig. 8b](#)).

Sample 06T33 is an orthogneiss located at Bitch Creek Narrows at the northwestern edge of the study area. Both euhedral and round zircons were identified in this sample; they are indistinguishable in age. 11 analyses yielded a weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2684 ± 4 Ma (MSWD = 0.1) and a concordia age of 2684 ± 4 (MSWD = 0.8) ([Fig. 8c](#)).

Sample 06T35 is a porphyritic Webb Canyon Gneiss from the westernmost body Webb Canyon gneiss. All eight analyses are concordant. One grain gave a slightly older age

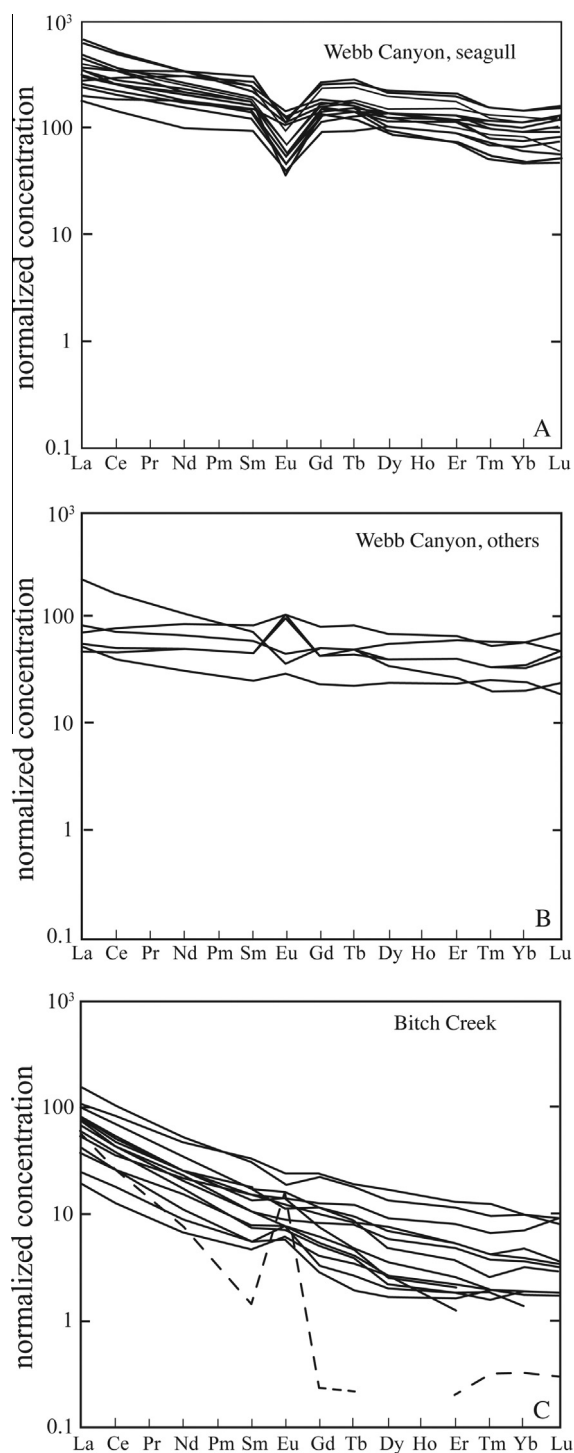


Fig. 7. Rare earth patterns normalized to chondrites. (A) The main group of Webb Canyon gneisses is high in total REE and exhibits a “seagull” pattern. (B) Some Webb Canyon gneisses have less pronounced Eu anomalies and slightly lower overall REE contents. (C) Bitch Creek gneisses have much lower REE contents, lack Eu anomalies, and are LREE-enriched and HREE-depleted. Data from Table 1.

of 2694 Ma. Excluding the slightly older grain, this sample gave a weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2679 ± 4

(MSWD = 1.6), indistinguishable from a concordia age of 2681 ± 6 (MSWD = 0.2) (Fig. 8d).

Sample 06T39 is a typical fine-grained porphyritic Webb Canyon Gneiss collected near Bitch Creek Narrows in the western part of the study area (Fig. 1). This sample has a population of small and large zircon grains. Fourteen individual large zircon grains were analyzed. Thirteen of the fourteen analyses are less than 7% discordant (Fig. 8e). One grain with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2691 Ma may contain an inherited component. A weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2677 ± 3 Ma (MSWD = 1.2) is determined from all fourteen analyses, indistinguishable from a concordia age of 2680 ± 5 Ma (MSWD = 0.9), which excludes the one older, discordant grain.

Sample 07T4 of Webb Canyon Gneiss was collected from upper Waterfalls Canyon (Fig. 1). Zircon grains, which range in size from $\sim 60 \mu\text{m}$ to $300 \mu\text{m}$ are typically elongate, euhedral, and strongly zoned. Uranium contents are less than 440 ppm, and analyses are less than 3% discordant. A weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2686 ± 5 Ma (MSWD = 0.2) and a concordia age of 2686 ± 5 Ma (MSWD = 0.2) were determined from seven analyses (Fig. 8f).

Sample 07T30 was collected west of Jackson Lake from a gully north of Waterfalls canyon. Zircon is low U (<370 ppm), and all analyses are concordant except for one reversely discordant grain. All 14 analyses yield a weighted $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2677 ± 6 Ma (MSWD = 2.0) and a concordia age of 2679 ± 7 Ma (MSWD = 2.0) (Fig. 8g).

5.2. Bitch Creek gneiss

Sample 04T17 of Bitch Creek gneiss collected south of Moose Basin was intruded by boudinaged mafic dikes. Fourteen zircon grains were analyzed. U varies from 158 to 970 ppm but discordance does not correlate with U concentration. The fourteen analyses define a single array with a mean weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2685 ± 3 Ma (MSWD = 0.7) and a concordia age of 2686 ± 4 (MSWD = 0.9) (Fig. 8h).

Sample 06T34 was collected from the upper reaches of Bitch Creek, where the Bitch Creek gneiss intrudes Moose Basin gneiss. Uranium contents of zircon from this sample ranged from 535 to 2919 ppm. U–Pb analyses are extremely discordant as is expected of metamict grains, and no age information was interpreted (Fig. 8i).

Sample 07T16 is a garnet-bearing Bitch Creek gneiss sample collected from Waterfall Canyon (Fig. 1). Zircon was less abundant in this sample than others analyzed in this study, possibly because garnet has incorporated Zr into its crystal structure. Individual grains range from $100 \mu\text{m}$ to $300 \mu\text{m}$ in length, and are needle shaped euhedral prisms. Four analyses have high uranium contents (1386–3162 ppm) and correspondingly high discordance. Four other grains yield $^{207}\text{Pb}/^{206}\text{Pb}$ ages from 2668 to 2678 Ma. Three of the four are less than 5% discordant. A weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2675 ± 6 Ma (MSWD = 0.34) and a concordia age of 2675 ± 7 (MSWD = 0.5) are determined from these four collinear analyses (Fig. 8j).

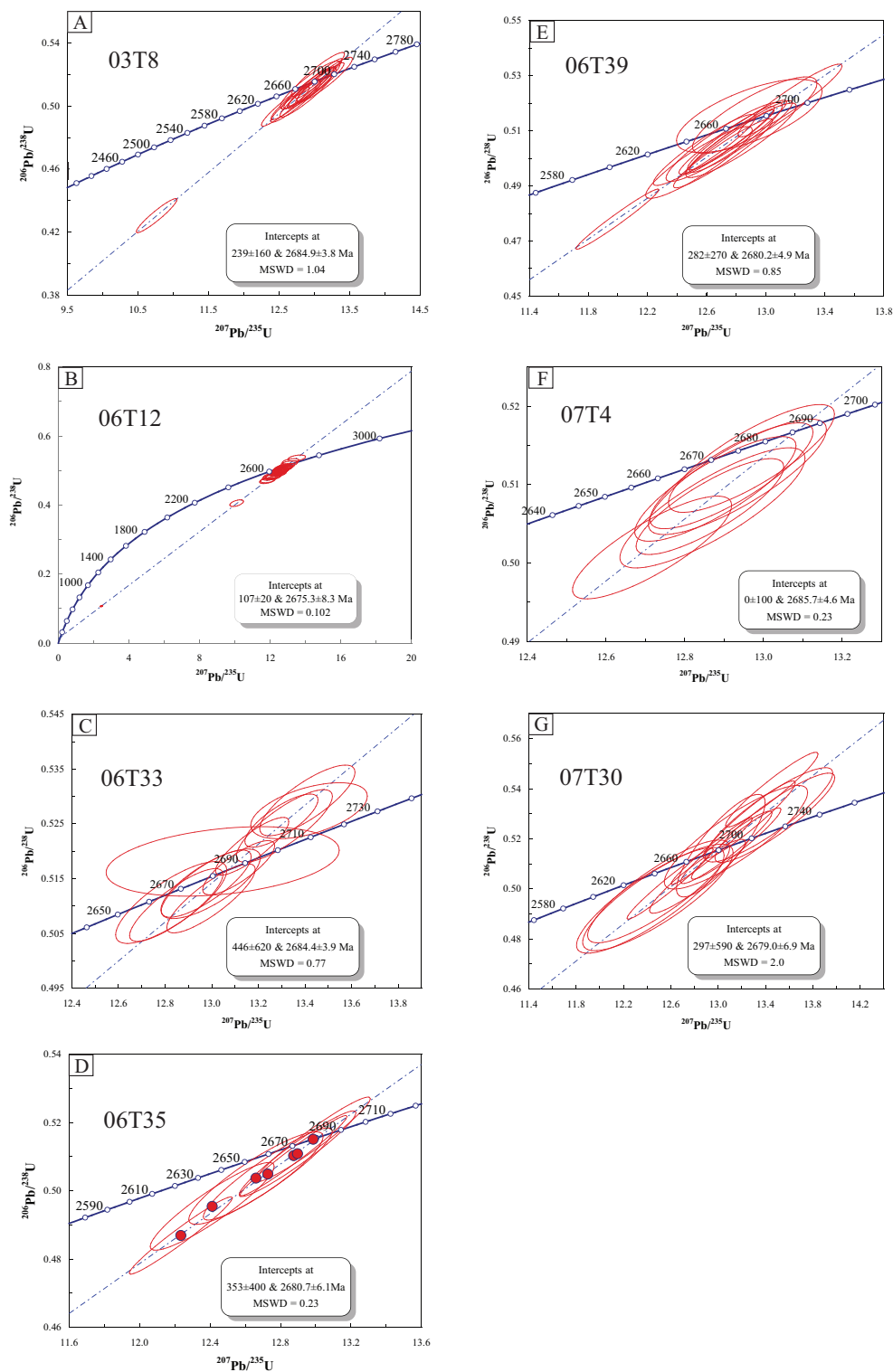


Fig. 8. Part 1. U-Pb zircon concordia diagrams for samples of Webb Canyon Gneiss. (A) 03T8, (B) 06T12, (C) 06T33, (D) 06T35, (E) 06T39, (F) 07T4, (G) 07T30. Part 2. U-Pb zircon concordia diagrams for samples of Bitch Creek Gneiss. (H) 04T17, (I) 06T34, (J) 07T16.

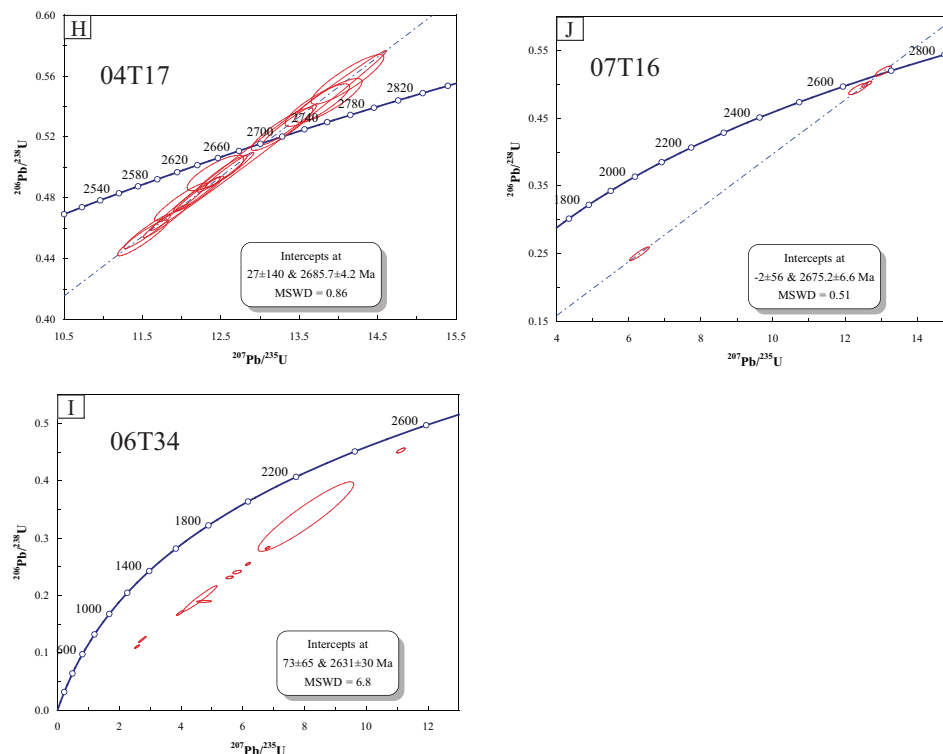


Fig 8. (continued)

5.3. Summary of geochronology

U-Pb dates for the Webb Canyon and Bitch Creek gneisses range from 2674 ± 5 Ma to 2686 ± 5 Ma (Table 2), and the dates for Webb Canyon and Bitch Creek gneisses are indistinguishable. Moreover there appears to be no spatial correlation with age, and adjacent samples yield different ages outside of uncertainty. Because the closure temperature for U-Pb in zircon is above 900 °C (Lee et al., 1997) and many of the calculated Zr saturation temperatures for the Webb Canyon gneiss are well above this, some dates may record the time when the rock cooled past this closure temperature after crystallization. Accordingly, it is not possible to determine unequivocally whether leuco-

granitic magmas were repeatedly intruded and crystallized over a period of time or as a single magmatic event. Nevertheless, two lines of evidence suggest the former. First, calculated zircon saturation temperatures suggest that Bitch Creek magmas crystallized below the zircon closure temperature, yet the weighted mean average $^{207}\text{Pb}/^{206}\text{Pb}$ ages for these gneisses are slightly different, outside uncertainty. Second, the observation that some areas of the Webb Canyon gneiss show F_2 folds, whereas others do not is consistent with crystallization at different times in the history of this mountain system.

The older U-Pb dates of the gneisses correspond to the time of amphibolite grade metamorphism and partial melting of the layered gneiss at 2685 Ma (Fitz-Gerald, 2008)

Table 2
Summary of U-Pb zircon ages for Webb Canyon and Bitch Creek gneisses.

Unit	Sample	Concordia age	Weighted mean average age
Webb Canyon	03T8	2685 ± 4 Ma (MSWD = 1.0)	2683 ± 3 Ma (MSWD = 1.6)
	06T12	2675 ± 8 (MSWD = 0.1)	2674 ± 5 Ma (MSWD = 0.3)
	06T33	2684 ± 4 (MSWD = 0.8)	2684 ± 4 Ma (MSWD = 0.1)
	06T35	2681 ± 6 (MSWD = 0.2)	2679 ± 4 (MSWD = 1.6)
	06T39	2680 ± 5 Ma (MSWD = 0.9)	2677 ± 3 Ma (MSWD = 1.2)
	07T4	2686 ± 5 Ma (MSWD = 0.2)	2686 ± 5 Ma (MSWD = 0.2)
	07T30	2679 ± 7 Ma (MSWD = 2.0)	2677 ± 6 Ma (MSWD = 2.0)
Bitch Creek	04T17	2686 ± 4 (MSWD = 0.9)	2685 ± 3 Ma (MSWD = 0.7)
	06T34	No age information	No age information
	07T16	2675 ± 7 (MSWD = 0.5)	2675 ± 6 Ma (MSWD = 0.34)

Table 3

Rb–Sr and Sm–Nd isotopic data for leucogranitic gneisses of the Teton Range.

Sample	Description	Rb (ppm)	Sr (ppm)	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$ (2685 Ma)	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	Initial $^{143}\text{Nd}/^{144}\text{Nd}$ (2685 Ma)	Initial ε_{Nd} (2685 Ma)
<i>Bitch Creek Gneiss</i>												
04T17		47.72	223.53	0.6185	0.72612	0.70209	2.80	16.69	0.1015	0.511049	0.509252	1.9
06T4		15.09	488.89	0.0894	0.70512	0.70165	2.24	13.55	0.1001	0.511093	0.509321	3.3
06T14		15.82	360.78	0.1269	0.70650	0.70157	1.64	11.26	0.0880	0.510840	0.509281	2.5
06T15		20.39	100.62	0.5863	0.70535	0.68257	0.85	4.05	0.1275	0.511547	0.509288	2.6
06T21		111.16	244.99	1.3146	0.74633	0.69524	1.67	9.73	0.1041	0.511088	0.509245	1.8
06T23		3.61	227.43	0.0459	0.70304	0.70126	3.40	22.10	0.0931	0.510869	0.509220	1.3
06T34		24.17	336.10	0.2083	0.70772	0.69962	2.18	14.30	0.0922	0.510898	0.509264	2.2
07T16		21.28	180.05	0.3424	0.71635	0.70304	5.37	30.42	0.1066	0.511154	0.509265	2.2
07T25		73.85	357.73	0.5986	0.72796	0.70469	4.90	24.39	0.1214	0.511178	0.509027	–2.5
<i>Webb Canyon Gneiss</i>												
98T-16		34.44	84.62	1.1833	0.75596	0.70997	44.07	166.50	0.1601	0.512013	0.509177	0.5
98T-17		33.71	31.25	3.1653	0.85023	0.72722	27.78	104.50	0.1600	0.512057	0.509223	1.4
03T8		45.00	44.50	2.9304	0.86074	0.74685	26.82	104.10	7.8769	0.512093	0.372554	3.5
06T5		75.50	35.48	6.3242	0.98286	0.73708	25.72	101.96	0.1525	0.511984	0.509282	2.5
06T12		51.34	44.78	3.3474	0.79947	0.66938	29.32	108.25	0.1637	0.511969	0.509068	–1.7
06T24		10.17	128.90	0.2285	0.71091	0.70203	7.63	25.99	0.1775	0.512446	0.509301	2.9
06T25		7.38	180.12	0.1185	0.70609	0.70148	9.36	32.11	0.1762	0.512341	0.509220	1.3
06T26		21.95	121.89	0.5219	0.72174	0.70146	17.51	57.89	0.1829	0.512433	0.509192	0.8
06T27		44.51	52.49	2.4733	0.78808	0.69197	31.64	116.01	0.1649	0.512126	0.509205	1.0
06T33		89.92	103.84	2.5275	0.79527	0.69704	39.79	174.26	0.1381	0.511733	0.509287	2.6
06T35		71.09	139.28	1.4848	0.76091	0.70321	48.60	202.90	0.1448	0.511385	0.508820	–6.5
06T39		67.83	481.29	0.4104	0.77113	0.75518	44.23	224.61	0.1190	0.511279	0.509170	0.3
06T42		106.29	108.62	2.8655	0.82944	0.71808	38.49	169.35	0.1374	0.511591	0.509156	0.1
07T4		76.37	25.58	8.9187	1.03805	0.69144	24.71	97.09	0.1539	0.511960	0.509234	1.6
07T5		65.95	28.43	6.8712	0.94971	0.68267	27.34	102.54	0.1612	0.512100	0.509244	1.8
07T7		35.85	21.47	4.8375	0.88246	0.69446	30.02	128.04	0.1418	0.511765	0.509254	2.0
07T10		84.89	34.04	7.2261	0.98977	0.70894	29.62	117.28	0.1527	0.512051	0.509345	3.8
07T19		76.05	28.46	7.9628	1.01036	0.70090	25.63	99.59	0.1556	0.511990	0.509233	1.6
07T26		9.08	96.24	0.2733	0.71202	0.70140	11.67	41.73	0.1691	0.512248	0.509252	1.9
07T30		2.06	85.93	0.0694	0.70457	0.70187	4.85	18.84	0.1558	0.511945	0.509185	0.6
08T6		42.05	94.47	1.2938	0.75338	0.70310	36.90	132.50	0.1683	0.512272	0.509290	2.7
08T12		61.67	75.61	2.3833	0.80749	0.71487	13.10	59.19	0.1340	0.511627	0.509254	2.0
08T13		61.30	64.12	2.6988	0.82066	0.71578	60.40	231.60	0.1578	0.512149	0.509354	3.9
08T18		49.89	59.94	2.4284	0.79054	0.69617	25.77	101.26	0.1539	0.511909	0.509183	0.6
08T19		128.69	59.12	6.4791	1.00057	0.74877	30.90	127.50	0.1467	0.511682	0.509082	–1.4
08T20		76.53	103.79	2.1585	0.82650	0.74261	31.40	138.75	0.1370	0.511631	0.509204	1.0
08T21		64.28	21.09	9.1499	1.08838	0.73279	26.40	104.40	0.1532	0.511987	0.509274	2.4
08T28		75.94	42.15	5.3423	0.95999	0.75237	34.38	145.02	0.1433	0.511903	0.509364	4.1

Analytical details: ~80–100 mg of sample were dissolved in HF–HNO₃. After conversion to chlorides, one-third of the sample was spiked with ^{87}Rb , ^{84}Sr , ^{149}Sm and ^{146}Nd .

Rb, Sr, and REE were separated by conventional cation-exchange procedures. Sm and Nd were further separated in di-ethyl-hexyl orthophosphoric acid columns.

All isotopic measurements were made on a VG Sector multi-collector mass spectrometer at the University of Wyoming.

An average $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratio of 0.710251 ± 20 (2s) was measured for NBS 987 Sr, and an average $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.511846 ± 11 (2s) was measured for the LaJolla Nd standard. Uncertainties in Sr isotopic ratio measurements are ± 0.00002 and uncertainties in Nd isotopic ratio measurements are ± 0.00001 (2s). Blanks are <50 pg for Rb, Sr, Nd, Sr and no blank correction was made. Uncertainties in Rb, Sr, Nd, and Sm concentrations are $\pm 2\%$ of the measured value; uncertainties on initial ε_{Nd} are ± 0.5 epsilon units.

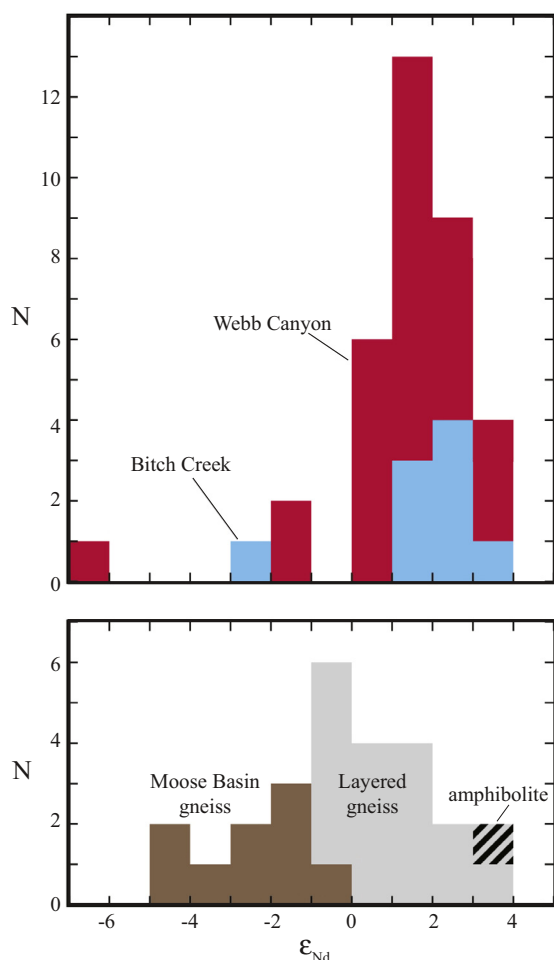


Fig. 9. Initial ϵ_{Nd} values (calculated at 2685 Ma) for Webb Canyon and Bitch Creek gneisses (A) compared to ϵ_{Nd} values at 2685 Ma for Moose Basin and Layered gneisses (B). The positive initial ϵ_{Nd} values of most of the leucogranitic rocks suggest that they were derived from a juvenile source such as the Layered gneisses.

suggesting that leucogranitic melts were first formed during that event and continued to intrude and crystallize for some millions of years thereafter. The leucogranites appear to postdate granulite-facies metamorphism of the Moose Basin gneiss (2695 ± 7 Ma; [Fitz-Gerald, 2008](#)) by between 10 to 20 million years.

6. ND AND SR ISOTOPIC DATA

6.1. Sm–Nd isotopic data

As their different REE patterns indicate, Sm and Nd concentrations and Sm/Nd ratios are higher for Webb Canyon gneisses than for Bitch Creek gneisses ([Table 3](#)). Nevertheless, both the Webb Canyon Gneiss and Bitch Creek gneiss are dominated by samples with positive initial ϵ_{Nd} values ([Fig. 9a](#)). The few samples yielding negative initial ϵ_{Nd} values are from locations scattered across the study area and their isotopic compositions do not appear to cor-

relate with other petrographic or geochemical characteristics.

The similar, positive initial ϵ_{Nd} values of both Webb Canyon and Bitch Creek gneiss samples suggests they may share a common provenance from relatively juvenile sources. Potential local sources include the Layered Gneiss and the Moose Basin gneiss. At 2685 Ma only the Layered Gneiss has positive initial ϵ_{Nd} values ([Fig. 9b](#)). The similar range of ϵ_{Nd} values for the Layered Gneiss and the Webb Canyon and Bitch Creek gneisses suggests that leucogranitic magma could be derived from partial melting of the Layered quartzofeldspathic and amphibolite gneisses, or from other similarly juvenile sources. Contributions from the Moose Basin gneiss, with its Nd isotopic signature indicating a more ancient continental provenance, appears to have been limited to a few disparate locations.

6.2. Rb–Sr isotopic data

As noted previously, Webb Canyon Gneiss samples are characterized by variable but generally high Rb/Sr. As a result, the correction for radiogenic growth of ^{87}Sr since 2680 Ma is large and calculated initial ratios vary greatly ([Table 3](#)). By comparison, the Bitch Creek gneisses have much lower Rb/Sr. The Rb–Sr system in both sets of gneisses almost certainly has been affected by open system behavior, either during deformation of the gneisses or during the intrusion of the 2.55 Ga Mount Owen Quartz Monzonite ([Zartman and Reed, 1998](#)). Dikes of the latter pervade the Archean exposures of the Teton Range, including throughout the study area. Fluids associated with these dikes, many of which are very coarse-grained pegmatites, may be responsible for retrogression of the high-grade metamorphic assemblages in the northern Teton Range. For this reason we do not interpret the Rb–Sr isotopic results in detail, except to note that the samples with the lowest $^{87}\text{Rb}/^{86}\text{Sr}$, and hence the samples least likely to have been affected by later interaction with fluids from the Mount Owen Quartz Monzonite, have initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.701 to 0.702. These relatively unradiogenic Sr isotopic compositions are consistent with the Nd isotopic results suggesting derivation by partial melting of a relatively young or unradiogenic source such as basalts or graywackes that compose the Layered Gneiss.

7. THERMOMETRY

Sample 07T30 is a garnet-bearing Webb Canyon Gneiss sample from the eastern side of the Teton Range north of Waterfalls Canyon. Garnet in this sample is not poikilitic, and is not surrounded by haloes depleted in ferromagnesian minerals ([Fig. 10a](#) and [b](#)). Magnetite, biotite, and hornblende intergrowths are interpreted as pseudomorphs after orthopyroxene. ([Fig. 10c](#)). We obtained compositional data on garnet, hornblende, and plagioclase from this sample in order to apply the thermobarometry calculations of [Kohn and Spear \(1990\)](#) ([Table S3](#)). The garnet has compositionally homogeneous cores with weak iron enrichment near edges and inclusions, and only minor textural and compositional evidence of post-crystallization resorption

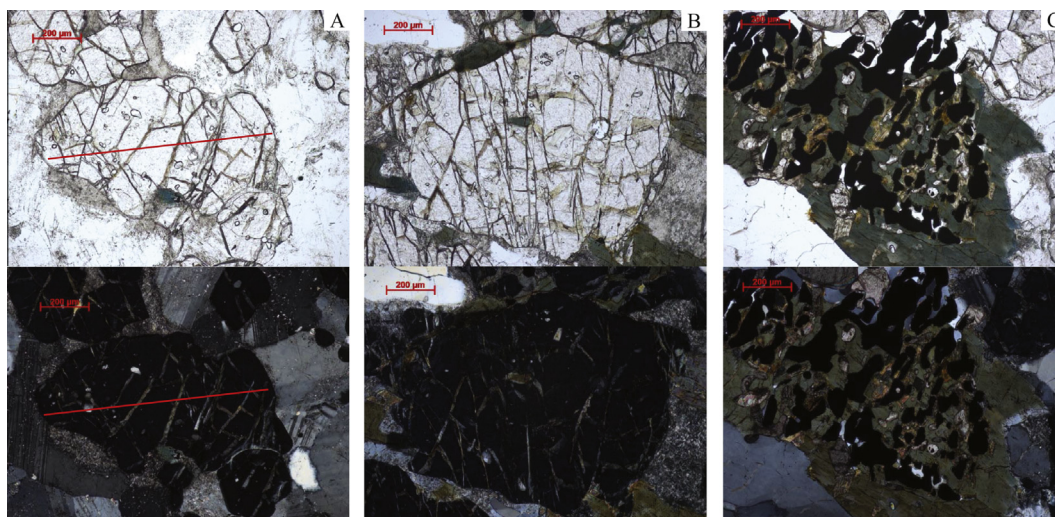


Fig. 10. Textures in Webb Canyon Gneiss (sample 07T30). (A) Anhedral, non-poikilitic garnet showing line of traverse for quantitative analyses shown in Fig. 11. (B) A subhedral, non-poikilitic garnet with abundant adjacent hornblende (no depletion halo). (C) Presumed pseudomorph of magnetite, biotite, and hornblende after orthopyroxene. All images are at the same scale.

(Fig. 11). Plagioclase rims are slightly enriched in anorthite component relative to cores (An_{30} vs. An_{25}), and hornblende rims are slightly more tschermakitic than cores (Table S3). Data from cores of all minerals yield temperature of 680 °C and pressure of 11 kb. The textural evidence that the rock originally contained orthopyroxene and the higher zircon saturation temperature of this sample (960 °C) both suggest that the maximum temperature achieved by this rock was considerably higher than is preserved by the garnet-hornblende-plagioclase compositions. Variation in edge compositions of garnet, plagioclase, and hornblende all are consistent with retrograde re-equilibration of mineral compositions.

The 11 kb pressure recorded in the Webb Canyon gneiss sample 07T30 is almost certainly higher than the pressures experienced by the Layered Gneiss and amphibolites along the eastern side of the Teton Range. As described above, the metamorphic conditions of the Layered Gneiss are poorly constrained, but the rare occurrence of cordierite + garnet in samples that experienced partial melting suggests pressures of 5 kb or below (Spear, 1993). Thermobarometry from low-iron amphibolites suggests that these rocks did not experience pressures above 6 kb (FitzGerald, 2008). Unfortunately the composition of the high-iron, garnet-bearing amphibolites lies outside the calibration of the barometer of Kohn and Spear (1990),

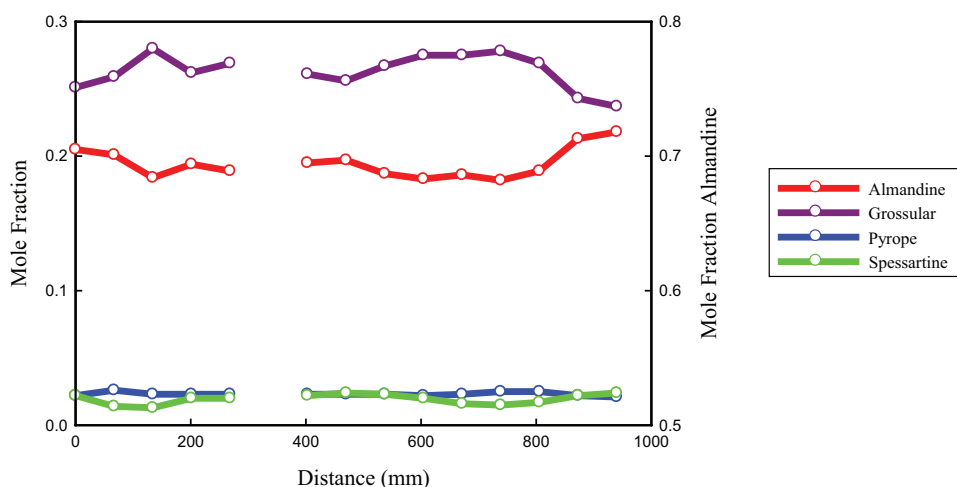


Fig. 11. Compositional zoning in garnet from sample 07T30. Profile consists of 15 microprobe analyses along red line shown in Fig. 10. One analysis was rejected because it fell on a crack. Note that the almandine composition (red line) is given by the right-hand axis; all other components are given on the left-hand axis. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

so no quantitative estimates of pressure could be obtained from these amphibolites.

8. DISCUSSION AND CONCLUSIONS

8.1. Petrogenesis of Webb Canyon gneiss and Bitch Creek gneiss

Both Webb Canyon and Bitch Creek gneisses are dominated by quartz and plagioclase, which accounts for their calcic, silica-rich compositions. However, the Webb Canyon and Bitch Creek gneisses are not petrologically and chemically identical, and any petrogenetic model for their origin must take into account the following differences, summarized in Table 4, below.

An important distinction between the two groups of gneisses is their difference in Zr contents and correspondingly different zircon saturation temperatures. Zr contents of Webb Canyon gneisses are high (240–590 ppm, plus two samples with ~800 ppm), whereas Zr contents of Bitch Creek gneisses are lower (70–295 ppm, plus one sample with 397 ppm) (Fig. 6b). The higher Zr contents in the Webb Canyon gneisses could indicate the presence of entrained and inherited zircon grains. However, the nearly total absence of older zircon in these rocks renders this explanation unlikely. We conclude that the different zircon-saturation temperatures indicate that Webb Canyon magmas were hotter than Bitch Creek gneiss magmas (Fig. 6b and c).

Because Sr is compatible in feldspars, the high Sr content and the high-Al nature of the Bitch Creek gneiss most likely reflect that melting of plagioclase played a role in the formation of Bitch Creek gneiss magmas. By contrast, Sr and Al are lower in the Webb Canyon gneiss samples, and their negative Eu anomalies indicate plagioclase fractionation or retention in the source. Yttrium, with behavior similar to heavy rare earth elements, is lower in the Bitch Creek gneiss than in the Webb Canyon gneiss. Because Y is strongly partitioned into garnet and hornblende, the higher Y of the Webb Canyon gneiss may indicate the melting of garnet or hornblende in the production of Webb Canyon magmas.

The contrasting REE patterns of the two groups of leucogranitic gneisses support the different roles of plagioclase and hornblende and/or garnet in the formation of the magmas. The “seagull” shaped patterns of the Webb Canyon gneiss (Fig. 7a) are common in rocks that form from hot, dry magmas (Bachmann and Bergantz, 2008). The high REE contents could reflect participation of hornblende and/or garnet in the melting reactions. The negative Eu anomalies may be due to retention of plagioclase in the residue, although variable Rb/Sr in these rocks suggests that fractionation of plagioclase may also be important. By contrast, the HREE-depleted patterns and lack of negative Eu anomalies of the Bitch Creek gneisses suggest that during production of those magmas HREE-enriched minerals were retained in the source rather than plagioclase. Two arguments suggest that hornblende is the likely phase. First, the Bitch Creek gneisses do not exhibit the extreme HREE-depleted, “J-shaped” REE patterns of many Archean gneisses for which a garnet-bearing residua is hypothesized (i.e., Moyen and Martin, 2012). Second, Reichardt and Weinberg (2012b) showed that water-fluxed partial melting in the mid-crust in which hornblende is retained in the source produces leucogranites with the HREE-depleted patterns (and high Sr/Y values) like those of the Bitch Creek gneisses.

Experimental studies help identify the petrogenetic processes that may be responsible for two types of trondhjemites in the Teton Range. Beard and Lofgren (1991) conducted partial melting experiments on greenstones and amphibolites from the Smartville complex at 800–1000 °C and pressures up to 6.9 kb. The partial melts obtained in water-excess and dehydration melting experiments at 3 and 6.9 kb were quite different. In the water-excess experiments, plagioclase broke down, yielding an amphibole-rich restite and a strongly peraluminous melt. Alumina contents of the melts all lie within Barker (1979) high-Al field (Fig. 12a) and are more Fe-poor and Ca-rich (Fig. 12b and c). Dehydration melting experiments contained only the water bound in hydrous phases, mainly amphiboles. In these experiments, amphibole and quartz broke down to produce pyroxenes in the restite and mildly peraluminous to metaluminous granodioritic and

Table 4
Summary characteristics of Webb Canyon and Bitch Creek gneisses.

	Webb Canyon gneiss	Bitch Creek gneiss
Outcrop character	Large, tabular exposures	Small sills, dikes, and plutons
Mineralogy	Hornblende biotite trondhjemite and granodiorite (except easternmost sheet, which lacks hornblende)	Biotite trondhjemite
Major element characteristics	Very high silica	High silica
	Low Al	High Al
	Strongly ferroan (except easternmost sheet)	Magnesian
Trace element characteristics	High Zr	Low Zr
	High Y	Low Y
	Low Sr	High Sr
	High REE, seagull shape pattern	LREE-enriched patterns, no pronounced Eu anomaly

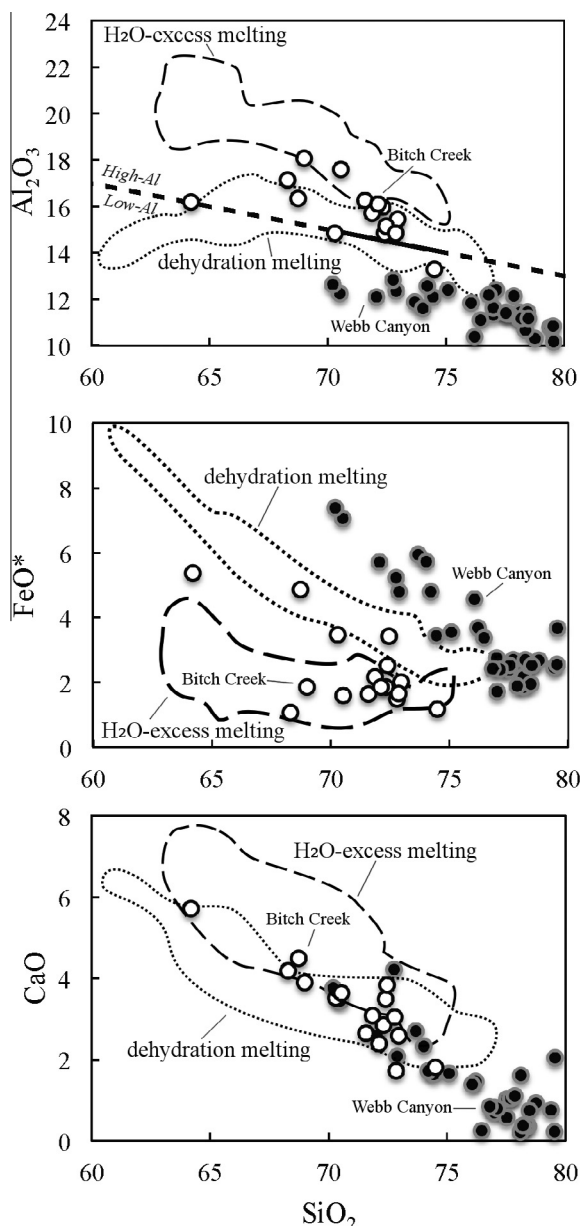


Fig. 12. Alumina (A), FeO^{tot} (B) and CaO (C) contents of Webb Canyon and Bitch Creek gneisses, compared to the fields of melt compositions produced by experimental melting of mafic rocks at moderate to low pressures (Beard and Lofgren, 1991).

trondhjemitic melts. The experimental melts have compositions that plot along Barker's high-Al/low-Al boundary (Fig. 12a), and are high in iron and poor in CaO (Fig. 12b and c).

Although experimental systems are simplified and the precise results depend upon starting compositions, these experiments nevertheless provide insights into role of water in producing the Webb Canyon and Bitch Creek magmas. Specifically, H_2O -excess melting by the reaction $\text{hbl} + \text{pl} + \text{qz} + \text{vapor} = \text{melt}$, in which plagioclase breaks down leaving an amphibole-rich restite, should produce magmas higher in alumina and lower in FeO along with high Sr

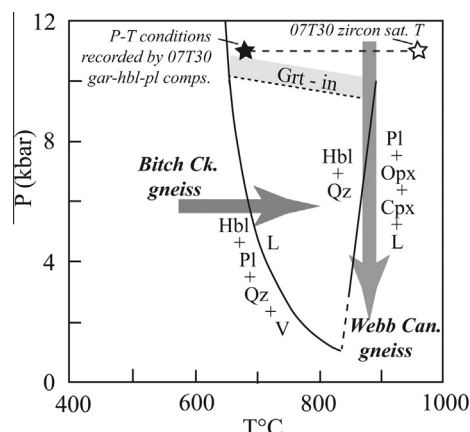


Fig. 13. Experimental and thermodynamically predicted positions of water-excess melting and dehydration melting of hornblende-plagioclase-quartz rocks from Pattison et al. (2003). $\text{Hbl} + \text{Pl} + \text{Qz} + \text{V} = \text{L}$ reaction from Piwinski (1968). $\text{Hbl} + \text{Qz} = \text{Pl} + \text{Opx} + \text{Cpx} + \text{L}$ from Patiño-Douce and Beard (1995). Garnet-in reaction from Vielzeuf and Schmidt (2001); its exact position is composition-dependent. The exact temperatures and pressures of the water-excess and dehydration melting reactions producing Bitch Creek and Webb Canyon gneisses, respectively, are poorly constrained; arrows are illustrative only. Filled star indicates temperature and pressure calculated from garnet-hornblende-plagioclase compositions of sample Webb Canyon Gneiss sample 07T30; open star indicates zircon saturation temperature calculated for the same sample.

and HREE depletion, as is observed in Bitch Creek gneisses. By contrast, dehydration melting involving breakdown of amphibole and quartz by the reaction $\text{hbl} + \text{qz} = \text{pl} + \text{opx} + \text{cpx} + \text{melt}$ can account for the low alumina, high FeO, high silica content and trace element characteristics of the Webb Canyon gneiss.

The temperatures and pressures for water-excess and dehydration melting of hornblende-plagioclase-quartz rocks can be determined from experimental constraints compiled by Pattison et al. (2003) (Fig. 13). The water-excess melting reaction that produces the Bitch Creek gneisses takes place at lower temperature than the dehydration reaction that generates the Webb Canyon Gneiss, as is also suggested by the lower zircon-saturation temperatures of the Bitch Creek gneisses. If typical of Webb Canyon gneisses generally, the mineral chemistry of garnet-bearing Webb Canyon Gneiss sample 07T30 suggests that dehydration partial melting took place at pressures of 11 kb or greater. The pressure of the water-excess melting reaction is poorly constrained. If the source of Bitch Creek magmas is the Layered Gneiss, then because garnet is absent in all but the most Fe-rich amphibolites within the Layered Gneiss, partial melting probably took place at or below 6 kb (Fig. 13) (Fitz-Gerald, 2008).

8.2. Tectonic implications

Magmas in collisional environments may form both by water-excess melting and dehydration melting. Water-excess melting may occur in a collision-related overthrust, where a relatively cool, hydrous lower plate releases water

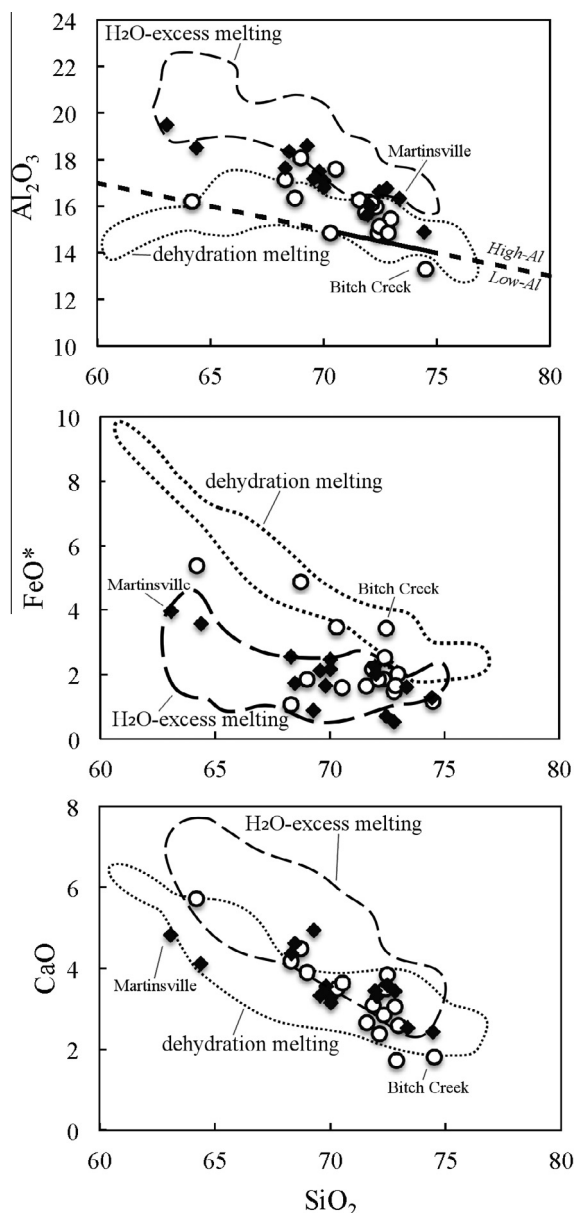


Fig. 14. Alumina (A), FeO^{tot} (B) and CaO (C) contents of Bitch Creek gneisses, compared to trondhjemites of the Martinsville igneous complex of Virginia (Beard, 1997). Also shown are the fields of melt compositions produced by experimental melting of mafic rocks at moderate to low pressures (Beard and Lofgren, 1991).

into a hotter upper plate. This environment was invoked for tonalites of the Paleozoic Martinsville igneous complex of Virginia (Beard, 1997). The apparent source rocks for the Martinsville tonalites were amphibolitic, and the major element composition of the partial melts produced are remarkably similar to the Bitch Creek gneisses (Fig. 14). Water-excess melting has also been called on to explain the origin of some of the Early Miocene leucogranites of the Himalaya (LeFort et al., 1987; Reichardt and Weinberg, 2012a). Many of the Bitch Creek gneisses are found within the Moose Basin gneiss, but they could not have formed

there: their Nd isotopic compositions preclude a Moose Basin gneiss source and the presence of garnets formed during decompression both indicate that Bitch Creek (Fig. 4c) magmas have migrated to higher structural levels. Water-saturated magmas normally will not ascend significant distances within the crust because the slope of the wet solidus is negative (Fig. 13). However, Reichardt and Weinberg (2012a) note that if external water fluxes the system, then partial melting could take place above the wet solidus temperature but below temperatures required for dehydration melting.

Dehydration melts are also likely in collisional environments, where dramatically thickened crust undergoes gravitational collapse and tectonic extension. Melt production may accompany the resulting decompression. The Webb Canyon magmas hypothesized to form by this mechanism migrated to form large layer-parallel sills or sheets. As shown on Fig. 13, the slope of the dehydration melting reaction is positive, such that melts can ascend to shallower levels without intersecting their solidus. Under such conditions Webb Canyon magmas generated by dehydration melting at depths in excess of 11 kb could ascend to solidify within Layered Gneiss that never experienced such high pressures. In part because of the ability of dehydration partial melts to migrate, Searle et al. (2010), Visona et al. (2012), and others have called on this mechanism to produce the majority of the two-mica leucogranites of the Himalayas.

Development of a detailed model for emplacement of leucogranites during continent–continent collision in the Teton Range is challenging for a number of reasons. First and foremost, the paleohorizontal and paleovertical directions are unknown, making it impossible to distinguish lateral or vertical flow from transpressional strike-slip motion. Furthermore, the rocks have undergone subsequent deformation, beginning with the open folding of the Archean F_3 event. Intrusion of the voluminous and fluid-rich 2.55 Ga Mount Owen Quartz Monzonite has further obscured relationships among the older rocks. Nevertheless, a continental collisional setting is supported by the similarity of a number of features observed in the Teton Range to those in the Himalayan orogen. Like in the Himalayas, leucogranitic magmas were generated by multiple melting mechanisms. In both orogens, the leucogranites are 10 to 20 million years younger than peak metamorphism, which is approximately the incubation period needed for crustal temperatures to rise, partial melt to form, and migration to commence (Godin et al., 2006). Like in the Himalayas, the host gneisses in the Tetons experienced shortening deformation that is not present in the leucogranites: Webb Canyon and Bitch Creek gneisses lack the F_1 isoclinal folds present in the older gneisses and only locally display the F_2 folds. The Webb Canyon leucogranite sheets are concordant with foliation in the host rocks, and the intensity of foliation within the sheets increases towards the margins. Finally, in both orogens there is a discontinuity of protoliths on either side of the leucogranite sheets reflecting the juxtaposition of two distinct continental blocks. In the Tetons the Layered and Moose Basin gneisses have strongly contrasting compositions, protoliths,

and metamorphic histories as would be expected in a collisional setting.

However, even if leucogranite magma generation and migration is similar in Teton and Himalayan orogens, there is a significant disparity: the rock types are different. The leucogranites of the Tetons are calcic trondhjemites and granodiorites whereas the Himalayan leucogranites are true granites. The difference in leucogranite composition reflects fundamental differences between the crust in the Tetons and the Himalayas. Although granitic crust was relatively widespread in Archean cratons by Neoproterozoic time, and even within the Wyoming province as early as 3.3 Ga (Frost et al., 2011), the crust that partially melted in the Tetons was composed of quartz-, plagioclase-, and hornblende-bearing rocks, such as greywacke, tonalite, and amphibolite. These rocks, which are less potassic than the pelitic sources of the Himalayan leucogranites, also required higher temperatures (>750 °C) to partially melt than were required to partially melt the pelitic rocks in the Himalayas (640–750 °C; Searle et al., 1997). As further examples of Archean collisional orogens are described, we may expect to document a variety of leucocratic crustal melt rocks depending on composition of the crust that melted. Study of the leucogranitic gneisses of the Teton Range suggests that compositional differences aside, the same general processes of melt formation, migration, and solidification in collisional orogens have taken place for over 2.6 billion years of Earth history.

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APPENDIX A. SUPPLEMENTARY DATA

Supplementary data associated with this article can be found, in the online version, at <http://dx.doi.org/10.1016/j.gca.2015.12.015>.

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