Proterozoic evolution of the western margin of the Wyoming craton: implications for the tectonic and magmatic evolution of the northern Rocky Mountains[1]

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Abstract: Defining the extent and age of basement provinces west of the exposed western margin of the Archean Wyoming craton has been elusive because of thick sedimentary cover and voluminous Cretaceous–Tertiary magmatism. U–Pb zircon geochronological data from small exposures of pre-Belt supergroup basement along the western side of the Wyoming craton, in southwestern Montana, reveal crystallization ages ranging from ~2.4 to ~1.8 Ga. Rock-forming events in the area as young as ~1.6 Ga are also indicated by isotopic (Nd, Pb, Sr) signatures and xenocrystic zircon populations in Cretaceous–Eocene granitoids. Most of this lithosphere is primitive, gives ages ~1.7–1.86 Ga, and occurs in a zone that extends west to the Neoproterozoic rifted margin of Laurentia. These data suggest that the basement west of the exposed Archean Wyoming craton contains accreted juvenile Paleoproterozoic arc-like terranes, along with a possible mafic underplate of similar age. This area is largely under the Mesoproterozoic Belt basin and intruded by the Idaho batholith. We refer to this Paleoproterozoic crust herein as the Selway terrane. The Selway terrane has been more easily reactivated and much more fertile for magma production and mineralization than the thick lithosphere of the Wyoming craton, and is of prime importance for evaluating Neoproterozoic continental reconstructions.

Résumé : Il a toujours été difficile de définir l'étendue et l'âge des provinces du socle à l'ouest de l'affleurement de la bordure ouest du craton de Wyoming en raison de l'épaisse couverture de sédiments et du volumineux magmatisme au Crétacé–Tertiaire. Des données géochronologiques U–Pb sur des zircons de petits affleurements du socle, pré-ceinture supergroupe, le long du coté ouest du craton du Wyoming, dans le sud-ouest du Montana, ont donné des âges de cristallisation de 2,4 à 1,8 Ga. Des événements aussi jeunes que ~1,6 Ga formant des roches dans le secteur sont aussi indiqués par des signatures isotopiques (Nd, Pb, Sr) et des populations de xénocristaux de zircon dans des granitoïdes du Crétacé–Éocène. La plus grande partie de cette lithosphère est primitive, avec des âges d'environ ~1,7 – 1,86 Ga et se retrouve dans une zone qui s'étend vers l'ouest, à la bordure fragmentée de Laurentia (Néoprotérozoïque). Selon ces données, le socle à l'ouest de l'affleurement du craton de Wyoming (Archéen) contient des terranes juvéniles accrétés de type arc (Paléoprotérozoïque) ainsi qu'une possible sous-plaque mafique d'âge semblable. Ce secteur se retrouve principalement sous le bassin de ceinture mésoprotérozoïque et il est pénétré par le batholite d'Idaho. Dans le présent document nous donnons le nom de terrane de Selway à cette croûte paléoprotérozoïque. Le terrane de Selway a été réactivé beaucoup plus facilement et il est beaucoup plus fertile pour la production de magma et de minéralisation que l'épaisse lithosphère du craton de Wyoming; il est, de plus, d'une importance primaire pour l'évaluation des reconstructions continentales au Néoprotérozoïque.

[Traduit par la Rédaction]

Introduction

A large section of southwest Laurentia consists of a poorly constrained mosaic of Archean and Proterozoic crust that accreted to the Wyoming craton subsequent to its incorporation into Laurentia at ~1.86 Ga. In many earlier portrayals of

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the Laurentian basement, the Archean crust of the Wyoming craton was assumed to extended westward from the exposures in southwestern Montana and Wyoming to the Neoproterozoic passive margin (Hoffman 1988). More recent data suggest, however, that much of the crust that underlies the Belt basin, Cordilleran fold–thrust belt, and Idaho batholith is Paleoproterozoic and not Archean (Mueller et al. 1995; Foster and Fanning 1997; O'Neill 1998; Foster et al. 2002). This crust, referred to here as the Selway terrane, appears to be a continuation of the Paleoproterozoic crust in the Great Falls tectonic zone (O'Neill 1998; Mueller et al. 2002) and may extend south to the Mojave block. Many features of the post-Paleoproterozoic geology of the northern Rocky Mountains show some basement control, in part owing to the nature of the Selway terrane basement.

The Selway terrane abuts the contiguous, thick Archean lithosphere of the Wyoming craton along its western margin. The Archean crust acted as a rigid and impenetrable lithospheric buttress to the repeated juxtaposition of younger Archean and Proterozoic crust, which now extends westward to the Neoproterozoic rifted margin and comprises the Selway terrane (Fig. 1). Geochronologic and thermochronologic data indicate significant, widespread events in the Selway terrane involving magmatic additions to the crust, metamorphism, and crustal melting at 2.2–2.5 Ga. Evidence for these events is found within the western Great Falls tectonic zone (Montana and Idaho) and as far south as the Farmington Canyon complex (Utah) (Mueller et al. 2004*a*). Younger Proterozoic events are not as widespread, based on present data, but include burial and crustal melting of the northwest margin of the Wyoming craton at ~1.77 Ga and a widespread tectonothermal event at ~1.65–1.63 Ga that is recorded from Utah to southwestern Montana and eastern Idaho (Mueller et al. 2004*a*, 2005*b*; Chamberlain et al. 2005). Crust of intermediate ages (1.8–1.9 Ga) and indeterminate origins has also been documented in this terrane (Foster et al. 2002; Mueller et al. 2002; Roberts et al. 2002). Unravelling the spatial and temporal distribution of this crust is critical to understanding the overall plate tectonic regime that characterized Proterozoic accretion along the southern margin of Laurentia and for identifying the specific terranes and continents that may have been joined to Laurentia prior to Neoproterozoic rifting. As we will show herein, this lithosphere was relatively enriched in incompatible elements and has been more fertile for partial melting than the areas underlain by Archean lithosphere.

Nature of the basement

Basement west of the exposed Wyoming craton in southwest Montana and central Idaho and extending south to Nevada and Utah is mostly covered and contains the western continuation of the Great Falls tectonic zone along with the Archean rocks of the Wyoming Province (Fig. 1). The distribution of crustal age provinces shown on Fig. 1 is derived from exposures of mainly Archean rocks from the Wyoming craton (Mogk et al. 1992; Chamberlain et al. 2003), Paleoproterozoic and less abundant Archean rocks in the Great Falls tectonic zone (Mueller et al. 2002, 2004*a*; Kellogg et al. 2003; Vogl et al. 2003, 2004*a*, 2004*b*), along with evidence for 2.4–1.6 Ga basement beneath the Belt basin, and the Idaho batholith (Mueller et al. 1995, 2004*a*; Foster et al. 2002; Sims et al. 2004). The following section summaries the basement terranes west and northwest of the Wyoming craton. The Paleoproterozoic and Mesoproterozoic basement south of the Wyoming craton (Fig. 1) is summarized in Condie (1992) and Karlstrom et al. (2002).

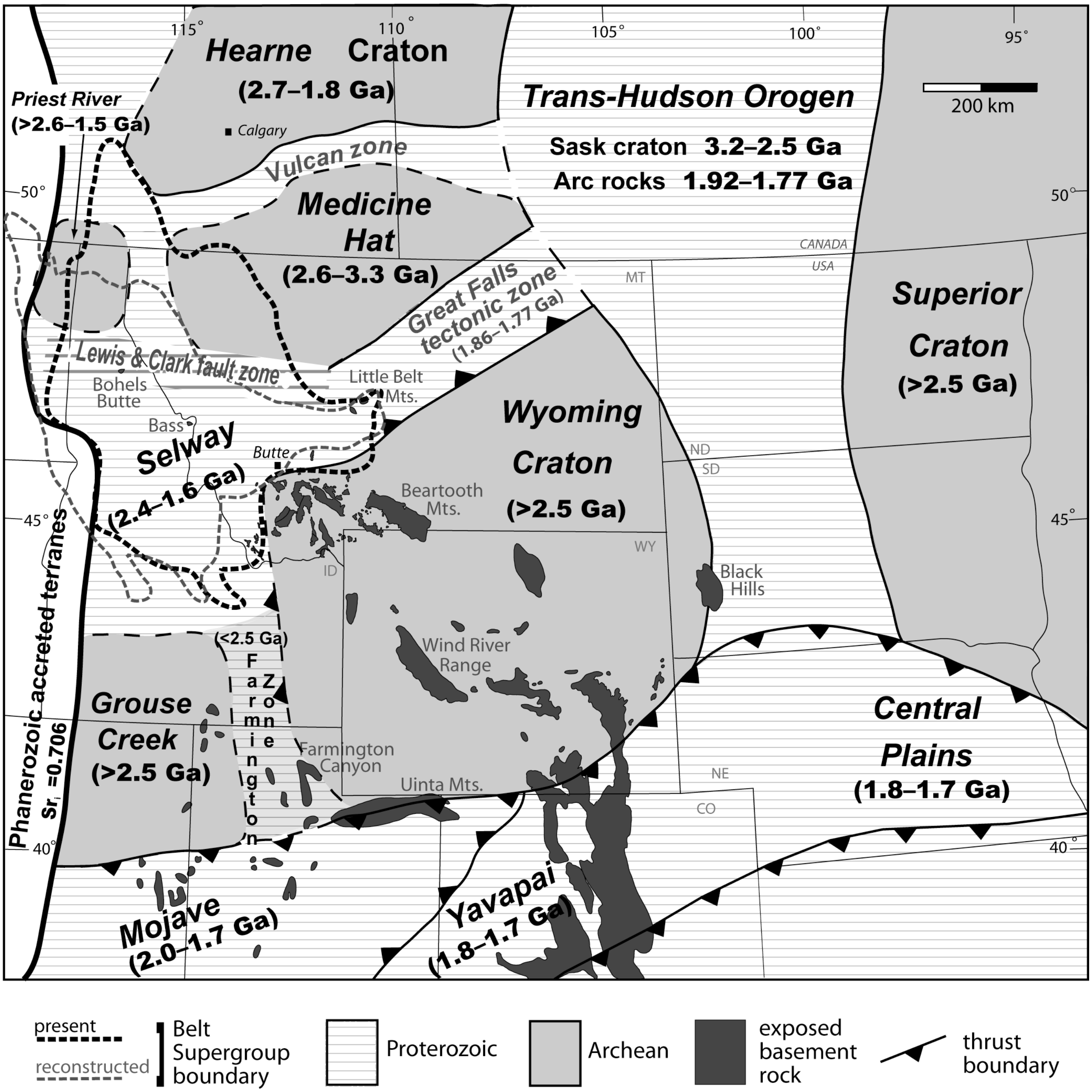
Wyoming Province

The Wyoming Province is a geophysically and geochemically distinct Archean craton that contains a unique assemblage of primarily Late Archean rocks (Wooden and Mueller 1988; Wooden et al. 1988; Mogk et al. 1992; Frost et al. 1998; Henstock et al. 1998; Chamberlain et al. 2003). The lithosphere of the Wyoming Province is unusually thick and strong, and has a distinctive velocity structure that readily distinguishes it from adjacent younger lithosphere (Thomas et al. 1987; Henstock et al. 1998; Dueker et al. 2001). The northwestern most part of the Wyoming Province consists of a distinct subdivision known as the Montana metasedimentary terrane. This terrane is dominated by middle Archean gneisses (3.2–3.5 Ga, Mueller et al. 1993; Kroh 1997; Mueller et al. 2004*b*) intercalated with metasedimentary rocks (Peale 1896). Crust of the Montana metasedimentary terrane is geochemically distinct from the crust of other parts of the Wyoming Province, as well as that of the Superior Province and other Archean terranes within Laurentia and globally (Wooden and Mueller 1988; Mueller and Wooden 1988; Zartman 1992; Frost et al. 1998). Rock ages from 2.5 to 3.5 Ga, detrital zircon and Sm–Nd model ages to 4.0 Ga, and uniquely enriched Pb isotopic signatures provide significant age and isotopic contrasts compared with adjacent, younger terranes (Mueller and Wooden 1988; Frost 1993). To the south, tonalite–trondhjemite–granodiorite suite rocks of the Beartooth–Bighorn magmatic zone (~2.9 Ga) and the Wyoming greenstone terrane (2.5–2.8 Ga) constitute the western boundary of the Wyoming craton (Mogk et al. 1992; Chamberlain et al. 2003).

Medicine Hat block

The Medicine Hat block of northwestern Montana and southern Alberta (Ross et al. 1991; Villeneuve et al. 1993; Lemieux et al. 2000; Fig. 1) is another Archean terrane known only from borehole intersections and geophysical data. This Archean block may be the southernmost part of the Hearne Province (Hoffman 1989), but based on seismic studies it has also been proposed to be part of the Wyoming Province (Henstock et al. 1998). The western extent of the Medicine Hat block is poorly defined. It may continue west into northwestern Washington and northern Idaho where Archean rocks are exposed in the Priest River complex (Evans and Fischer 1986; Doughty et al. 1997) and inferred by isotopic tracer data and xenocrystic zircons from Eocene plutons (Whitehouse et al. 1992). It may, however, extend only as far west as the eastern part of the Mesoproterozoic Belt basin (Fig. 1). Sims et al. (2004) separate the Medicine Hat block from the basement in the Priest River complex, based on magnetic anomalies. The southeastern border of the Medicine Hat block in central Montana is the Great Falls tectonic zone (O'Neill and Lopez 1985; O'Neill 1998) and it is bounded on the east by the Trans-Hudson orogen. The southwestern border may lie along the trace of the Lewis and Clark fault zone (Foster et al. 2003, 2006).

Fig. 1. Map of basement provinces of southwest Laurentia (after Ross et al. 1991; Condie 1992; Karlstrom et al. 2002; Vogl et al. 2004*b*; this study). Exposures of basement in Laramide-style uplifts are shown in the dark gray shaded areas. The outline of the Belt supergroup is shown in the present location and the reconstructed location based on Price and Sears (2000).



The Lewis and Clark fault zone is a complex, long-lived series of strike–slip and oblique slip faults that predate the Mesoproterozoic Belt basin and extend from northeastern Washington through central Montana (Reynolds 1979). The fault zone formed a major boundary between depocenters in the central Belt basin and the northern margin of the eastern Belt basin (Harrison et al. 1974; Winston 1986; Reynolds 1979). The fault zone was reactivated during Cretaceous to Paleogene thrusting (Sears et al. 2000) and served as a major transfer structure for Eocene extension (Reynolds 1979; Doughty and Sheriff 1992; Foster et al. 2003, 2006). Proterozoic anorthosite exposed along the southern side of the Lewis and Clark zone in the Bohels Butte block in the Clearwater metamorphic core complex (Figs. 1, 3) gives a U–Pb zircon crystallization age of ~1.79 Ga (Doughty and Chamberlain 2004).

Grouse Creek block

We propose the name Grouse Creek block to include Archean rocks (ca. 2.5–2.6 Ga) that occur west of the present western limit of semi-continuous exposures of the Wyoming craton. Archean rocks crop out in the Grouse Creek Range, Albion Range, and the East Humbolt Range, where they are strongly reworked by Paleoproterozoic and Phanerozoic events (Armstrong and Hills 1967; Wright and Snoke 1993; Egger et al. 2003; Premo et al. 2005). The deep crust beneath the central Snake River Plain is Archean based on Pb and Nd isotope data of Cenozoic lavas and xenoliths, and zircons in metaigneous xenoliths, with ages centered on 2.6–3.2 Ga (Leeman et al. 1985; Wolf et al. 2005). Isotopic signatures from Cretaceous igneous rocks also suggest that >2.5 Ga crust exists south of the Snake River Plain (Fleck and Wooden 1997). Exposures of Archean rocks in the northern Great basin and inferred Archean crust beneath the Snake River Plain are apparently separated from the contiguous Archean of the Wyoming craton by Paleoproterozoic rocks (2.45–1.6 Ga) exposed in the Wasatch Range (e.g., Farmington Canyon complex, Bryant 1988; Nelson et al. 2002; Mueller et al. 2004*a*), which is why we define it as a separate block. The southern margin of the Grouse Creek block may be located between the East Humbolt Range and the Ruby Mountains in Nevada (Wright and Snoke 1993; Premo et al. 2005), which puts it near the westward trend of the Cheyenne Belt (Karlstrom and Houston 1984). The northern margin is probably near the northern edge of the Snake River Plain. This block may or may not extend west to the Neoproterozoic rifted margin.

It is currently unclear if the Grouse Creek block is completely separated from the Wyoming craton in the deep crust. It is possible that the metasedimenary rocks in the Farmington Canyon complex, which comprise Archean and earliest Paleoproterozoic detritus, were deposited within a late Archean-earliest Paleoproterozoic rift within Wyoming, or along the late Archean passive margin of Wyoming. Orogeny and magmatism in the Farmington complex at ~2.45 and ~1.8 Ga (Barnett et al. 1993; Nelson et al. 2002; Mueller et al. 2004*a*) suggests accretion of Grouse Creek occurred in Paleoproterozoic time. The Grouse Creek block, therefore, could be an accreted Archean block, a rifted fragment of Wyoming that re-accreted, or part of Wyoming separated by a younger intracratonic rift and mobile belt. The area between the known Archean in the Wyoming craton and the Grouse Creek block is referred to here as the Farmington zone.

Great Falls tectonic zone

Proterozoic reactivation of Archean rocks of the northwestern Wyoming craton was first recognized by Hayden and Wehernberg (1959), who reported Paleoproterozoic K–Ar mineral ages from Archean gneisses. O'Neill and Lopez (1985) used these mineral ages and other thermochronologic data (Giletti 1966), along with geophysical and structural trends in younger rocks, to propose the existence of the Great Falls tectonic zone. They speculated that the Great Falls tectonic zone marks the Paleoproterozoic collision between the Archean Wyoming and Medicine Hat – Hearne provinces.

Geophysical studies have suggested the presence of a north-dipping paleosubducted slab beneath the Medicine Hat – Hearne block that may be related to the collision that led to the development of the Great Falls tectonic zone (Gorman et al. 2002; Ross 2002). Although the age of this slab cannot be specified, Mueller et al. (2002) have shown that Precambrian rocks in the Little Belt Mountains are dominated by 1.86 Ga calc–alkaline metaigneous rocks that exhibit trace element and Nd isotopic signatures suggestive of petrogenesis in a convergent environment in which juvenile lithosphere was consumed. These data suggest that the slab is likely to be Proterozoic and that oceanic lithosphere was being subducted beneath the Medicine Hat block at ca. 1.86 Ga. We refer to this calc-alkaline assemblage as the Little Belt arc.

The western Great Falls tectonic zone appears to have had a distinctly different tectonothermal history than the part exposed in the Little Belt Mountains (Giletti 1966; Harlan 1996; Roberts et al. 2002; Brady et al. 2004; Mueller et al. 2004*b*, 2005). U–Pb data from zircon and monazite demonstrate that the Archean gneisses in the Tobacco Root and Highland mountains experienced granulite facies metamorphism and partial melting ca. 1.77 Ga ago (Cheney et al. 2004; Mueller et al. 2004*b*, 2005). This is nearly 100 Ma later than the peak of magmatism and metamorphism in the exposed rocks of the Little Belt arc.

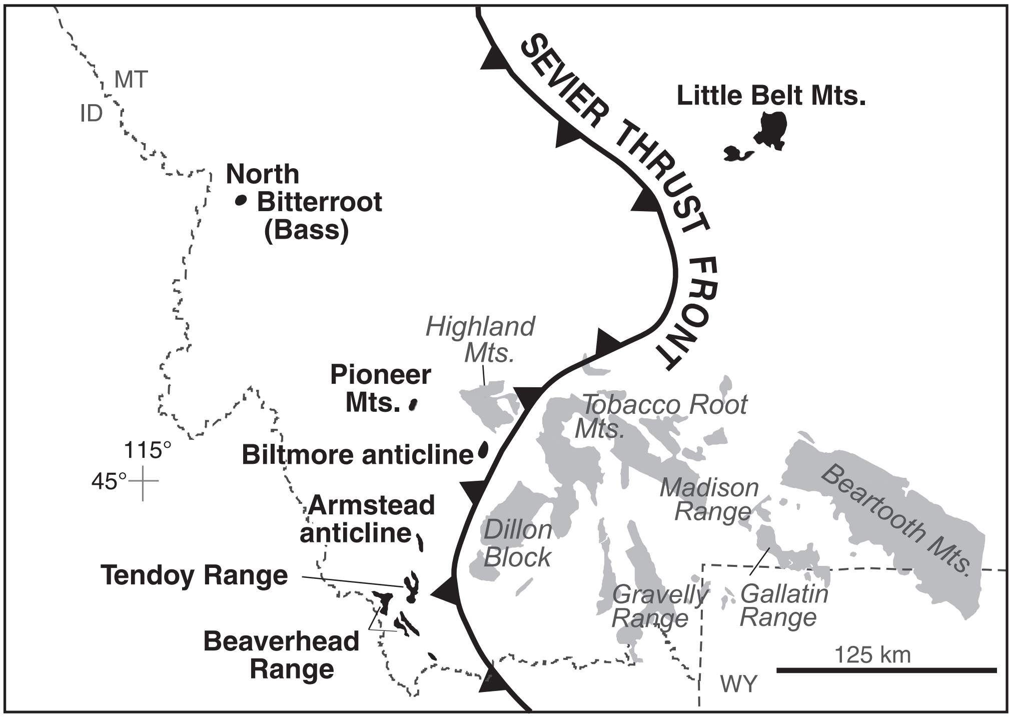
Holm and Schneider (2002) and Vogl et al. (2004*a*) reported $^{40}Ar/^{39}Ar$ ages for biotite and amphibole from the Little Belt Mountains that suggest rapid cooling through the biotite closure temperature (~350–300 °C) by ~1.75–1.71 Ga. K–Ar and $^{40}Ar/^{39}Ar$ data from the western Great Falls tectonic zone, where Archean rocks are overprinted, scatter considerably (~1.6–1.9 Ga, summary in Brady et al. 2004), and in some cases exceed U–Pb zircon and monazite ages. These apparent ages may reflect a single event with variable cooling histories, more than one event (± variable cooling histories), or a complex combination of cooling ages and variable excess argon contamination. It is unclear, therefore, whether the deformation and heating recorded in the central and western Great Falls tectonic zone reflect a single collisional event (i.e., Wyoming – Medicine Hat) or two distinct events.

The northwestern border of the Wyoming craton, therefore, lies along the 1.86–1.77 Ga Great Falls tectonic zone. The western border must be between the exposed Archean in Southwest Montana and western Wyoming and the Idaho batholith, which is underlain by mostly Paleoproterozoic crust (Mueller et al. 1995; Foster and Fanning 1997). Boundaries between these basement provinces are obscured by the Belt supergroup, Phanerozoic strata, Idaho batholith, and other Cretaceous–Eocene igneous rocks.

Selway terrane

We propose the term Selway terrane for the region underlain by Paleoproterozoic (1.6–2.4 Ga) metamorphic and igneous rocks west of the largely Archean rocks in the basement-cored, Laramide uplifts of Southwest Montana (Fig. 1). These Paleoproterozoic rocks crop out in the Pioneer Range, Biltmore anticline, Tendoy Range, Highland Range, Beaverhead Range, Bitterroot Range, and at Bohels Butte (Fig. 2), as suggested in previous studies or as documented in this study (O'Neill et al. 1988; Zen 1988; Toth and Stacey 1992; Ruppel et al. 1993; Foster et al. 2002; Kellogg et al. 2003; Mueller et al. 2004*a*, 2005; Doughty and Chamberlain 2004). Archean rocks have not been documented in any of these areas except the Highland Mountains (O'Neill et al. 1988). The outcrops are all within the Sevier fold and thrust belt, and most occur at the tops of major footwall ramps near the eastern part of the thrust belt (Sears et al. 1988; Kalakay et al. 2001). This is consistent with the interpretation that the western edge of the

Fig. 2. Sketch map of mountain ranges in Southwest Montana that expose Archean (gray shade) or Paleoproterozoic crystalline basement (black fill) (modified from Sims et al. 2004).



thick Wyoming craton is within this area. Magnetic anomaly patterns also suggest the presence of a major basement boundary along the edge of the thrust belt (Sims et al. 2004).

Results and methods

In the next section we present new U–Pb zircon data from basement rocks exposed along the boundary between the northwestern Wyoming craton and the Selway basement terrane in Southwest Montana. The sample locations are shown on Fig. 3 and the data are summarized in Table 1. Samples were crushed and zircon was extracted using conventional density and magnetic methods. U–Pb measurements were made using the sensitive high-resolution ion microprobe (SHRIMP) – reverse geometry (RG) at the US Geological Survey - Stanford Microanalytical Center, Standford University, using standard methods similar to those summarized by (Williams 1998). All zircon grains were imaged by cathodoluminescence, reflected light, and transmitted light before analysis. Data were reduced using SQUID and Isoplot-3.0 (Ludwig 2003) and errors are quoted at 2σ.

Pioneer Mountains

Sample P982 was collected from dioritic orthogneiss in the northern Pioneer Mountains. This exposure of basement is a km-scale pendant within the Cretaceous Pioneer batholith that Zen (1988) suggested was Proterozoic in age, based on highly discordant U–Pb zircon data. The basement block is within the eastern edge of the Sevier thrust belt, and, therefore, is probably detached, and thrust eastward up to 10–20 km (Kalakay et al. 2001). Magmatically zoned euhedral zircon from the dominant metaigneous phase gave a concordant ^{207}Pb/^{206}Pb age of 1.86 ± 0.01 Ga (Fig. 4) that we interpret to be the crystallization age of the protolith of the gneiss.

Biltmore anticline

Sample DF99-102 is a biotite–quartzofeldspathic orthogneiss from an exposure in the core of the Biltmore anticline, which is a small basement cored uplift immediately east of the McCartney Mountain thrust salient of the Sevier thrust belt (Sears et al. 1988). The McCartney Mountain salient is the easternmost Cretaceous thin-skin style thrust structure in this part of Southwest Montana; this means that this basement block has not been thrust east or west any significant distance. The majority of the spot analyses from euhedral-zoned zircon lie along a discordia line with an upper intercept of 1.89 ± 0.02 Ga (Fig. 5) that is taken to be the best estimate for the crystallization age of the protolith of the Biltmore anticline gneiss. Several analyses yield older concordant ages >2.0 Ga and these grains are considered to be xenocrysts.

Little Belt amphibolite

Sample LB01-168 is an amphibolite from the western Little Belt Mountains within the Sheep Creek intrusive complex (Vogl et al. 2004*b*). Based on field relationships, the Sheep Creek intrusive complex is the youngest major plutonic assemblage in the Little Belt arc. Numerous attempts to precisely date the leucogranites in this complex have failed because of high common Pb in the zircons. The granitic rocks are associated with sheets of amphibolite that appear to be metamorphosed, synplutonic dykes, based on some mutually crosscutting relationships. Sample LB01-168 gives a concordia age of 1.82 ± 0.01 Ga for euhedral magmatically zoned zircon (Fig. 6). The magmatic age of this

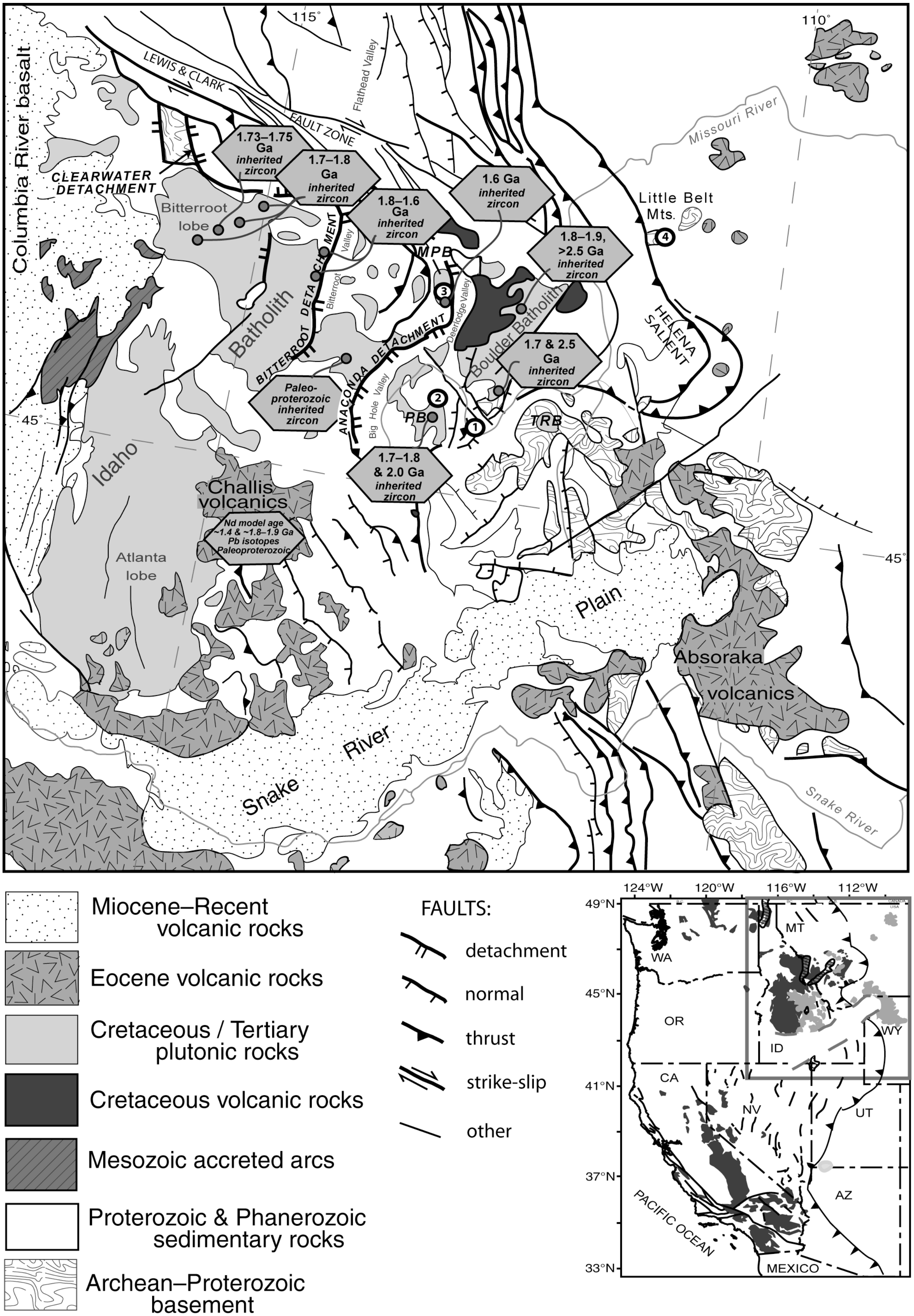
Table 1. Ion probe U–Pb data.

Grain number	U (ppm)	Th (ppm)	Rad ^{206}Pb (ppm)	$\frac{^{206}Pb}{^{238}U}$ age	Err. 1σ	$\frac{^{207}Pb}{^{206}Pb}$ age	Err. 1σ	Disc. (%)	$\frac{^{238}U}{^{206}Pb}$ *	Err (%)	$\frac{^{207}Pb}{^{206}Pb}$ *	Err (%)	$\frac{^{207}Pb}{^{235}U}$ *	Err (%)	$\frac{^{206}Pb}{^{238}U}$ *	Err (%)
Biltmore anticline (45°28.735′N, 112°28.449′W)																
DF99–102–1	105	55	31	1887.5	15.8	1904.1	25.0	1	2.94	1.0	0.1166	1.4	5.47	1.7	0.3402	1.0
DF99–102–1.2	86	53	25	1867.2	18.2	1903.1	31.1	2	2.98	1.1	0.1165	1.7	5.40	2.1	0.3360	1.1
DF99–102–2	378	146	109	1868.8	8.5	1855.4	14.1	–1	2.97	0.5	0.1134	0.8	5.26	0.9	0.3363	0.5
DF99–102–3	1192	792	218	1239.0	4.6	1915.3	17.8	55	4.72	0.4	0.1173	1.0	3.43	1.1	0.2119	0.4
DF99–102–4	1037	196	58	406.7	2.0	1652.1	28.1	306	15.36	0.5	0.1015	1.5	0.91	1.6	0.0651	0.5
DF99–102–5	151	218	42	1818.3	13.4	1884.9	23.5	4	3.07	0.8	0.1153	1.3	5.18	1.6	0.3259	0.8
DF99–102–6	216	147	62	1861.7	13.1	1889.7	19.1	2	2.99	0.8	0.1156	1.1	5.34	1.3	0.3348	0.8
DF99–102–7	290	71	85	1889.7	9.7	1882.6	15.6	0	2.94	0.6	0.1152	0.9	5.41	1.1	0.3406	0.6
DF99–102–8	397	151	106	1752.0	8.1	1874.3	14.2	7	3.20	0.5	0.1146	0.8	4.94	0.9	0.3123	0.5
DF99–102–9	971	469	150	1065.2	3.8	1976.5	12.1	86	5.57	0.4	0.1214	0.7	3.01	0.8	0.1797	0.4
DF99–102–10	239	105	69	1856.7	10.5	1868.5	17.7	1	3.00	0.7	0.1143	1.0	5.26	1.2	0.3338	0.7
DF99–102–11	986	679	293	1915.5	7.6	2491.6	6.8	30	2.89	0.5	0.1634	0.4	7.80	0.6	0.3460	0.5
DF99–102–12	198	56	60	1955.6	12.1	2016.8	17.9	3	2.82	0.7	0.1242	1.0	6.07	1.2	0.3544	0.7
DF99–102–13	971	902	132	930.8	3.8	1955.8	28.7	110	6.44	0.4	0.1200	1.6	2.57	1.7	0.1553	0.4
DF99–102–14	1343	1100	143	740.0	3.1	1936.7	29.9	162	8.22	0.4	0.1187	1.7	1.99	1.7	0.1216	0.4
DF99–102–14.2	3308	154	79	170.7	1.0	941.5	111.3	452	37.26	0.6	0.0705	5.4	0.26	5.5	0.0268	0.6
DF99–102–15	183	89	51	1825.6	17.9	1887.3	21.6	3	3.05	1.1	0.1155	1.2	5.21	1.6	0.3274	1.1
DF99–102–16	165	57	52	1999.3	13.5	1984.6	19.7	–1	2.75	0.8	0.1219	1.1	6.11	1.4	0.3636	0.8
DF99–102–17	277	189	65	1552.1	9.4	1892.9	19.4	22	3.67	0.7	0.1158	1.1	4.35	1.3	0.2722	0.7
DF99–102–18	146	44	45	1966.3	14.4	1959.2	23.1	0	2.80	0.9	0.1202	1.3	5.91	1.5	0.3567	0.9
Pioneer Mountains, P982 (45°40.217′N, 112°57.153′W)																
P982–1.1	159	70	56	1836.5	19.6	1901.8	33.0	3	3.03	0.0	0.1164	1.8	5.29	2.3	0.3296	1.2
P982–1.2	220	132	80	1824.7	24.5	1830.3	35.0	0	3.06	0.0	0.1119	1.9	5.05	2.6	0.3272	1.5
P982–2.1	243	148	91	1866.3	25.5	1856.0	24.7	0	2.98	0.0	0.1135	1.3	5.25	2.2	0.3358	1.6
P982–3.1	227	130	82	1826.3	21.0	1847.3	21.9	1	3.05	0.0	0.1129	1.1	5.10	1.9	0.3275	1.3
P982–4.1	193	78	68	1857.9	19.6	1869.7	23.2	0	2.99	0.0	0.1144	1.3	5.27	1.9	0.3340	1.2
P982–5.1	276	90	93	1795.6	23.4	1889.9	24.9	5	3.11	0.0	0.1156	1.4	5.12	2.1	0.3212	1.5
P982–6.1	132	50	47	1876.2	30.0	1839.0	39.6	–2	2.96	0.1	0.1124	2.2	5.24	3.0	0.3378	1.8
P982–7.1	170	84	62	1879.1	20.9	1875.5	37.1	0	2.95	0.0	0.1147	2	5.35	2.5	0.3384	1.3
P982–8.1	229	90	83	1891.8	14.9	1874.6	16.5	–1	2.93	0.0	0.1147	0.9	5.39	1.4	0.3411	0.9
P982–9.1	466	475	108	1162.4	22.1	1780.2	24.1	35	5.06	0.1	0.1089	1.3	2.97	2.6	0.1976	2.1
P982–10.1	218	121	80	1850.6	35.0	1870.3	19.5	1	3.01	0.1	0.1144	1.1	5.24	2.5	0.3325	2.2
P982–11.1	197	77	72	1910.7	31.9	1863.1	22.4	–3	2.90	0.1	0.1139	1.2	5.42	2.4	0.3450	1.9
P982–12.1	209	111	77	1884.3	18.8	1865.3	22.9	–1	2.95	0.0	0.1141	1.3	5.34	1.8	0.3395	1.1
P982–13.1	218	119	80	1862.9	33.6	1837.4	21.7	–1	2.98	0.1	0.1123	1.2	5.19	2.5	0.3351	2.1
P982–14.1	246	161	91	1832.8	97.7	1881.1	23.4	3	3.04	0.2	0.1151	1.3	5.22	6.4	0.3289	6.1
P982–15.1	258	106	89	1823.3	24.7	1815.1	16.1	0	3.06	0.0	0.1110	0.9	5.00	1.9	0.3269	1.6
P982–16.1	120	57	44	1878.2	20.5	1887.8	19.5	0	2.96	0.0	0.1155	1.1	5.39	1.8	0.3383	1.3
P982–17.1	279	154	101	1837.0	17.0	1849.5	18.8	1	3.03	0.0	0.1131	1	5.14	1.6	0.3297	1.1

Table 1 (*concluded*).

Grain number	U (ppm)	Th (ppm)	Rad ^{206}Pb (ppm)	$\frac{^{206}Pb}{^{238}U}$ age	Err. 1σ	$\frac{^{207}Pb}{^{206}Pb}$ age	Err. 1σ	Disc. (%)	$\frac{^{238}U}{^{206}Pb}$*	Err (%)	$\frac{^{207}Pb}{^{206}Pb}$*	Err (%)	$\frac{^{207}Pb}{^{235}U}$*	Err (%)	$\frac{^{206}Pb}{^{238}U}$*	Err (%)
Mount Powell batholith (46°16′N, 112°58′W)																
DF02–119A-1	1102	134	10	68.4	0.5	26.1	85.3	–62	93.70	0.8	0.0465	3.6	0.07	3.6	0.0107	0.8
DF02–119A-2C	231	56	22	688.1	4.9	1455.7	24.2	112	8.88	0.8	0.0914	1.3	1.42	1.5	0.1127	0.8
DF02–119A-3C	147	64	38	1679.9	11.9	1708.1	17.4	2	3.36	0.8	0.1046	0.9	4.30	1.2	0.2977	0.8
DF02–119A-4C	77	95	18	1588.7	15.9	1610.3	26.6	1	3.58	1.1	0.0993	1.4	3.82	1.8	0.2795	1.1
DF02–119A-5C	254	128	83	2087.9	10.7	2472.5	12.7	18	2.61	0.6	0.1616	0.8	8.52	1.0	0.3825	0.6
DF02–119A-6C	1577	291	111	506.8	1.6	1588.7	16.7	214	12.23	0.3	0.0981	0.9	1.11	0.9	0.0818	0.3
DF02–119A-7C	188	160	36	1308.1	9.0	1625.0	18.8	24	4.45	0.8	0.1001	1.0	3.10	1.3	0.2250	0.8
DF02–119A-8C	2294	104	552	1592.2	2.9	1764.6	4.5	11	3.57	0.2	0.1079	0.2	4.17	0.3	0.2802	0.2
DF02–119A-9C	112	71	26	1530.9	13.0	1619.4	28.1	6	3.73	1.0	0.0997	1.5	3.69	1.8	0.2681	1.0
DF02–119A-10C	97	81	18	1283.5	12.2	1589.7	33.5	24	4.54	1.0	0.0982	1.8	2.98	2.1	0.2203	1.0
DF02–119A-11C	189	65	50	1711.8	10.8	1811.1	15.2	6	3.29	0.7	0.1107	0.8	4.64	1.1	0.3041	0.7
DF02–119A-12C	195	188	48	1633.9	10.4	1673.4	16.0	2	3.47	0.7	0.1027	0.9	4.08	1.1	0.2885	0.7
DF02–119A-13C	88	40	13	1051.6	11.1	1535.7	38.2	46	5.64	1.1	0.0954	2.0	2.33	2.3	0.1772	1.1
DF02–119A-14C	572	719	129	1502.1	5.6	1557.2	10.3	4	3.81	0.4	0.0965	0.5	3.49	0.7	0.2624	0.4
Sheep Creek amphibolite (46°52.310′N, 11°58.668′W)																
LB0168–1	150	33	41	1795.2	26.6	1800.9	17.1	0	3.11	1.7	0.1101	0.9	4.87	1.9	0.3211	1.7
LB0168–2	141	36	40	1830.4	27.4	1818.8	16.6	–1	3.05	1.7	0.1112	0.9	5.03	1.9	0.3283	1.7
LB0168–3	74	18	20	1796.9	28.6	1748.9	35.7	–3	3.11	1.8	0.1070	2.0	4.74	2.7	0.3215	1.8
LB0168–4	77	17	21	1807.9	28.8	1793.6	23.3	–1	3.09	1.8	0.1096	1.3	4.89	2.2	0.3237	1.8
LB0168–5	82	24	22	1787.7	29.5	1831.8	22.8	2	3.13	1.9	0.1120	1.3	4.93	2.3	0.3196	1.9
LB0168–6	286	95	73	1682.8	24.6	1807.4	13.4	7	3.35	1.7	0.1105	0.7	4.54	1.8	0.2983	1.7
LB0168–7	188	49	53	1821.3	26.8	1794.4	14.7	–2	3.06	1.7	0.1097	0.8	4.94	1.9	0.3265	1.7
LB0168–8	310	91	86	1794.8	25.8	1830.6	11.3	2	3.11	1.6	0.1119	0.6	4.95	1.8	0.3210	1.6
LB0168–9	107	24	30	1812.4	27.8	1769.2	20.8	–2	3.08	1.8	0.1082	1.1	4.84	2.1	0.3246	1.8
LB0168–10	170	36	47	1789.7	26.4	1810.7	15.2	1	3.13	1.7	0.1107	0.8	4.88	1.9	0.3200	1.7
LB0168–11	108	39	28	1719.0	26.4	1778.0	25.5	3	3.27	1.8	0.1087	1.4	4.58	2.2	0.3056	1.8
LB0168–12	315	106	90	1848.6	26.4	1840.6	11.1	0	3.01	1.6	0.1125	0.6	5.15	1.8	0.3321	1.6
LB0168–13	106	27	30	1837.7	28.2	1815.6	18.9	–1	3.03	1.8	0.1110	1.0	5.05	2.0	0.3299	1.8
LB0168–14	2159	568	192	599.3	9.5	1584.8	67.2	62	10.26	1.7	0.0979	3.6	1.32	4.0	0.0974	1.7
LB0168–15	178	36	50	1829.3	26.9	1828.3	15.2	0	3.05	1.7	0.1118	0.8	5.06	1.9	0.3281	1.7
LB0168–16	77	20	20	1730.1	27.3	1790.2	32.7	3	3.25	1.8	0.1094	1.8	4.65	2.5	0.3078	1.8
LB0168–18	232	33	64	1780.2	25.9	1833.3	19.0	3	3.14	1.7	0.1121	1.0	4.91	2.0	0.3180	1.7
LB0168–19	314	190	80	1656.2	24.1	1829.5	19.6	9	3.41	1.6	0.1118	1.1	4.52	2.0	0.2929	1.6
LB0168–20	117	20	32	1777.7	27.1	1762.6	19.4	–1	3.15	1.7	0.1078	1.1	4.72	2.0	0.3175	1.7
LB0168–21	366	124	63	1168.0	18.2	1865.5	19.9	37	5.03	1.7	0.1141	1.1	3.12	2.0	0.1986	1.7
LB0168–22	116	41	32	1805.3	27.4	1827.4	18.0	1	3.09	1.7	0.1117	1.0	4.98	2.0	0.3232	1.7

Note: Common Pb correction using measured ^{204}Pb. Pb*, radiogenic Pb. Disc., discordance; Err, error.



115°
110°
LEWIS & CLARK
FAULT ZONE
Flathead Valley
Missouri River
Columbia River basalt
CLEARWATER DETACHMENT
1.73–1.75 Ga inherited zircon
1.7–1.8 Ga inherited zircon
1.8–1.6 Ga inherited zircon
1.6 Ga inherited zircon
1.8–1.9, >2.5 Ga inherited zircon
1.7 & 2.5 Ga inherited zircon
Paleo-proterozoic inherited zircon
1.7–1.8 & 2.0 Ga inherited zircon
Nd model age ~1.4 & ~1.8–1.9 Ga Pb isotopes Paleoproterozoic
Bitterroot lobe
Batholith
BITTERROOT DETACHMENT
Bitterroot Valley
MPB
Deerlodge Valley
Boulder Batholith
ANACONDA DETACHMENT
Big Hole Valley
PB
TRB
Little Belt Mts.
HELENA SALIENT
45°
Idaho
Challis volcanics
Atlanta lobe
Plain
Absoraka volcanics
Snake River Plain
Snake River
1
2
3
4
Miocene–Recent volcanic rocks
Eocene volcanic rocks
Cretaceous / Tertiary plutonic rocks
Cretaceous volcanic rocks
Mesozoic accreted arcs
Proterozoic & Phanerozoic sedimentary rocks
Archean–Proterozoic basement
FAULTS:
detachment
normal
thrust
strike-slip
other
124°W
120°W
116°W
112°W
49°N
45°N
41°N
37°N
33°N
WA
OR
CA
NV
ID
MT
WY
UT
AZ
PACIFIC OCEAN
MEXICO

Fig. 3. Geologic and sample location map of western Montana and Idaho showing Cretaceous and Tertiary igneous rocks and their relationships to outcropping basement in Southwest Montana, the Helena embayment of the Belt basin, and major structures. Labels indicate the age of xenocrystic zircons and isotopic model ages reflecting the sources of the igneous rocks. Sample locations are numbered (1) Biltmore anticline, (2) Pioneer basement gneiss, (3) Mount Powell batholith granite, and (4) Sheep Creek amphibolite. MPB, Mount Powell batholith; TRB, Tobacco Root batholith; PB, Pioneer batholith. Index map in lower right corner shows the map in the context of southwestern North America and major Cretaceous and Eocene igneous provinces.

Fig. 4. U–Pb concordia plot of zircon analyses from basement gneiss (P982) in the Pioneer Mountains, Montana.

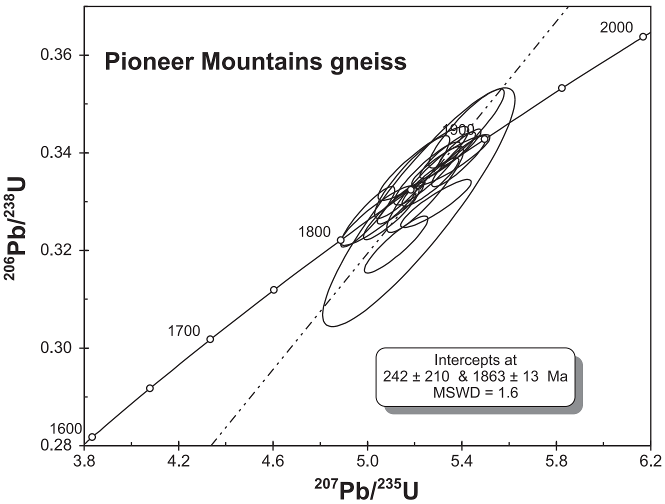
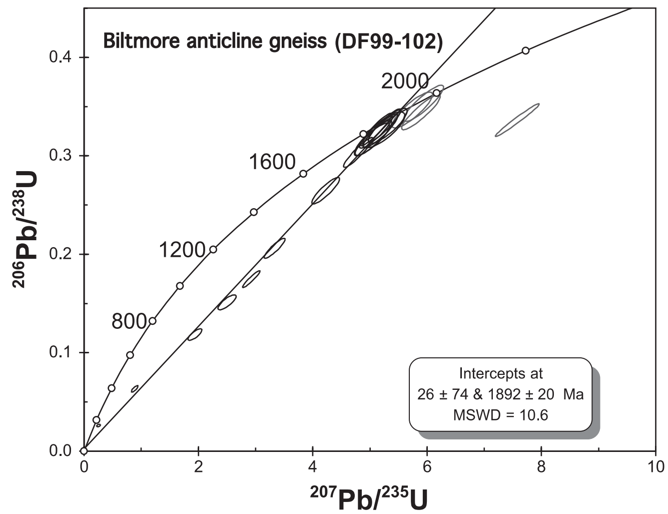


Fig. 5. U–Pb concordia plot for zircons from gneiss (DF99-102) from the Biltmore anticline.



amphibolite is younger than metadioritic rocks from the northeastern Little Belt Mountains (Mueller et al. 2002) and extends the temporal range of magmatism in the Little Belt arc of the central Great Falls tectonic zone.

Mount Powell batholith inherited zircon

The Mount Powell batholith is a Cretaceous–Paleocene granitic to monzodioritic suite of plutons in the Flint Creek Range, west of the Boulder batholith. Xenocrystic zircons from granitic plutons typically provide insight into the age(s) of lower crustal rocks involved in their petrogenesis (Mueller et al. 1995; Foster and Fanning 1997). Sample DF119A is a monzogranite of the Mount Powell batholith from the Race Track Creek drainage. Most of the inherited zircons in this sample show magmatic zoning patterns and lie along a discordia line with a relatively poorly constrained upper intercept age of 1.63 ± 0.03 Ga (Fig. 7). The large error reflects the scatter in individual grain ages, but suggests a population of grains ~1.63 Ga. Older grains with <20% discordance give ^{207}Pb/^{206}Pb ages of ~1.76, ~1.81, and ~2.47 Ga, commensurate with ages of rocks known from the western Great Falls tectonic zone.

Interpretation of the U–Pb data

The 1.86 ± 0.01 Ga crystallization age of the dioritic gneiss in the Pioneer mountains is concordant with several dioritic gneisses within the Great Falls tectonic zone exposed in the Little Belt Mountains (Mueller et al. 2002; Vogl et al. 2003, 2004*b*), and extends the range of metaigneous rocks of this age further southwest in the Great Falls tectonic zone. The protolith of the amphibolite from the Little Belt Mountains, which was intruded at ~1.82 Ga, demonstrates that magmatism within the Paleoproterozoic Little Belt arc was either continuous for ~40 Ma or that renewed magmatism occurred ~40 Ma after crystallization of the dioritic suite of plutons. The Biltmore anticline gneiss, which gives an age of ~1.89 Ga, is most likely also indicative of the crust that was accreted to the Archean Wyoming craton along with the Little Belt arc.

Inherited zircons in the Mount Powell batholith indicate that magmatic events in the lower crust of the western Great Falls tectonic zone include material as young as ~1.63 Ga. This overlaps the age of magmatic rocks along the western margin of the Wyoming craton in the Farmington Canyon complex (Mueller et al. 2004*a*) and events to the south of the Wyoming craton in the Mazatzal and Yavapai provinces (Karlstrom et al. 2002; Chamberlain et al. 2005). The three older inherited grains reflect ages from the Great Falls tectonic zone and Selway terrane and provide no evidence for Archean crust beneath the Flint Creek Range.

Discussion

The results presented here, along with previously published data, document Paleoproterozoic magmatism and metamorphism west of the Wyoming Province and have implications for the Mesozoproterozoic and younger evolution of the northern Rocky Mountains and the assembly of Laurentia.

Evidence for the Selway terrane

U–Pb zircon crystallization ages of the basement rocks west of the large, Archean-cored, Laramide basement uplifts in Southwest Montana — along with secondary isotope systematics, isotopic evidence, and xenocrystic zircons from Cretaceous and Tertiary granitoids, including the northern Idaho batholith (Bitterroot lobe), Pioneer batholith, Boulder

Fig. 6. U–Pb concordia plot for zircons from amphibolite (LB0168) from the Little Belt Mountains.

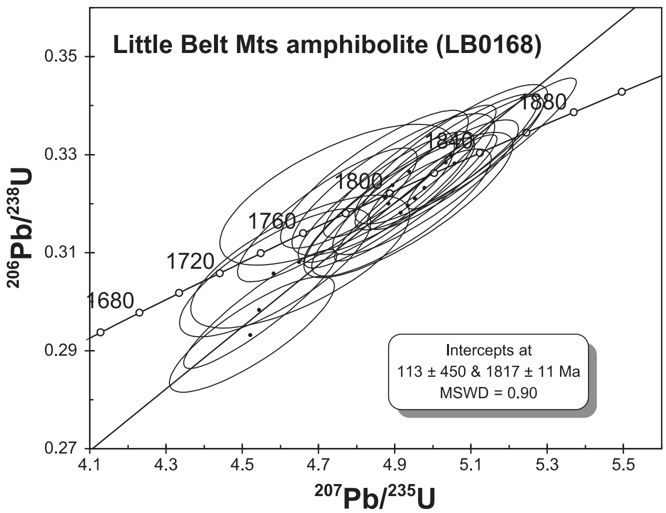


Fig. 7. U–Pb concordia plot for xenocrystic zircon from Cretaceous granite (DF02-119A) from the Mount Powell batholith. MSWD, mean square of weighted deviation.

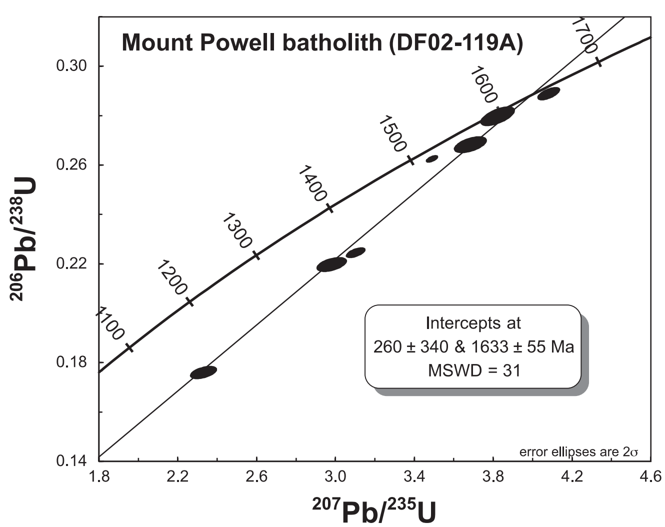
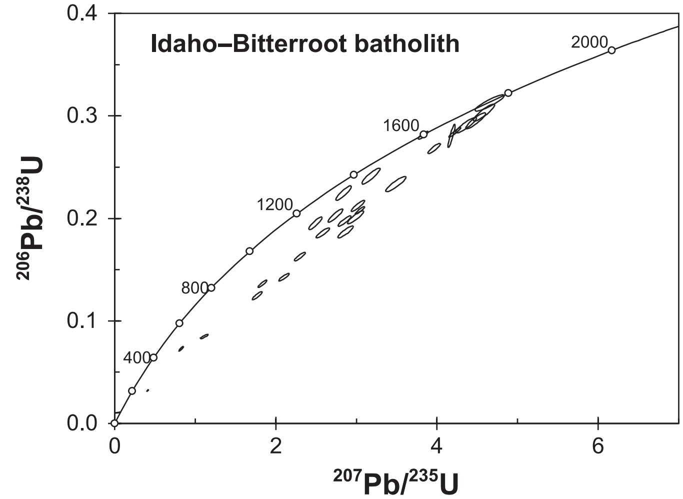


Fig. 8. U–Pb concordia plot showing a composite of ion probe data of xenocrystic zircons from the Bitterroot lobe of the Idaho batholith (data from Mueller et al. 1995, Foster and Fanning 1997; Foster et al. 2001).



batholith, Tobacco Root batholith, and Mount Powell batholith — suggest that the crust west of exposed Archean crust in southwest Montana is largely Paleoproterozoic in age (Fig. 3) (Doe et al. 1968; Bickford et al. 1981; Schuster and Bickford 1985; Toth and Stacey 1992; Fleck 1990; Mueller et al. 1995, 1996; Foster and Fanning 1997; O'Neill 1998; Foster et al. 2001; Lund et al. 2002; Murphy et al. 2002).

Most of the Idaho batholith intruded Precambrian continental crust, with only the western margin developed over Phanerozoic accreted terranes (Armstrong et al. 1977; Hamilton 1978; Hyndman 1983; Fleck and Chris 1985; Lund and Snee 1988). Initial $^{87}Sr/^{86}Sr$ ratios of northern Idaho batholith plutons generally increase from values of ~0.704 in the western border quartz diorites and tonalites to >0.706 to 0.712 in the "main-phase" (Armstrong et al. 1977; Fleck and Chris 1985). Epsilon Nd values from the western border zone plutons range from +2 to +6; those from the "main phase" plutons, range from –17.7 to –21.2 (Fleck 1990; Mueller et al. 1995). The Sm–Nd data from the main phase plutons give depleted mantle model ages of 1.72–1.93 Ga (Mueller et al. 1995). U–Pb ages of inherited zircons in the main phase granitoids dominantly range from 1.5–2.3 Ga (Bickford et al. 1981; Shuster and Bickford 1985; Toth and Stacey 1992; Mueller et al. 1995; Foster and Fanning 1997; Foster et al. 2001). SHRIMP results from single grains of inherited zircon in the northern Idaho batholith (Idaho–Bitterroot batholith) give ages ranging from 0.7 to 2.5 Ga, but several samples give single-age populations and concordant grains of 1.74–1.75 Ga (Mueller et al. 1995; Foster and Fanning 1997) (Fig. 8). These data, combined with the similar Nd-depleted mantle model age for these individual samples, suggest that a significant portion of the crust beneath the Idaho–Bitterroot batholith is made up of relatively juvenile ~1.75 Ga material (Mueller et al. 1995; Foster and Fanning 1997). Pb-isotopic data from the central Idaho–Bitterroot batholith also indicate a primitive, arc-like source with an age of 1.6–1.8 Ga (Toth and Stacey 1992).

Initial Sr isotopic values for the Late Cretaceous Pioneer batholith range from 0.7110 to 0.7160 (Arth et al. 1986; Mueller et al. 1997), and Pb isotopes give a model age of 1.9 ± 0.2 Ga (Doe et al. 1968; Zen 1992). These results suggest involvement of mainly Paleoproterozoic crust, as do inherited zircon ages of ~1.4 and 1.7–1.8 Ga (SHRIMP-II) for the Grayling Lake granite and Uphill Creek granodiorite reported by Murphy et al. (2002).

Initial Sr isotopic ratios for the Late Cretaceous Boulder batholith, including the Butte quartz monzonite, Marysville stock, and Hell Creek pluton range from 0.7055 to 0.7092, and are significantly lower than values for the Pioneer batholith (Doe et al. 1968; Mueller et al. 1997). A Proterozoic Pb isotopic model age for the Boulder batholith (2.1 ± 0.3 Ga) is similar to that of the Pioneer plutons (Zen 1992). Inherited zircons from several granitic intrusions in the Boulder batholith give abundant ca. 1.86–2.4 Ga ages, with the exception of the Radar Creek pluton, which intrudes Archean rocks and contains Archean inherited zircon (Mueller et al. 1997; Lund et al. 2002).

The ~75 Ma Tobacco Root batholith also intrudes exposed Archean rocks (Mueller et al. 1996; Vitaliano et al. 1980). The Tobacco Root batholith has epsilon Nd values ranging from –18.9 to –19.1 and Sm–Nd-depleted mantle model ages of 1.63–1.90 Ga (Mueller et al. 1996). Inherited zircons in the Tobacco Root batholith range in age from 2.2 to 3.0 Ga, generally consistent with the Archean country rocks. These results suggest derivation from Archean and Proterozoic crustal sources along with a volumetrically significant mantle-derived component that resulted in the relatively low initial Sr value of 0.7057–0.7067 for this pluton. These data from the Tobacco Root batholith suggest that some tectonic or magmatic layering of Archean and Proterozoic crust is present along the edge of the Wyoming craton.

In summary, the inherited zircon data from the Boulder batholith and Pioneer batholith suggest the involvement of crust of similar age to that exposed in the central Great Falls tectonic zone in the Little Belt (Mueller et al. 2002) and Pioneer Mountains (this study). The sources of the Idaho batholith are somewhat more varied, but show a major component of the crust beneath it formed at ~1.73–1.75, or ~100 Ma younger than most of the crust in the central Great Falls tectonic zone.

The variation in U–Pb zircon ages for the basement exposures in the Selway terrane indicate Proterozoic rock-forming events ranging in age from ~2.45 Ga (Kellogg et al. 2003; Mueller et al. 2004*a*) to ~1.8 Ga (this study). Isotopic tracer data and inherited zircons from the Cretaceous plutons indicate that Proterozoic lower crust west of this area includes rock-forming events of ca. 1.8–1.6 Ga. High-grade metamorphism and partial melting associated with orogenic events that overprint Archean rocks in the northwestern Wyoming craton (e.g., Tobacco Root and Highland Mountains) indicate orogenic events between ~2.4 and ~1.77 Ga (Mueller et al. 1994, 2004*a*, 2004*b*, 2005; Roberts et al. 2002; Sims et al. 2004). Because of the limited and discontinuous exposure of the Paleoproterozoic rocks and the currently limited geochronologic data, it is not possible to decipher the structural relationships between different segments of the material accreted to Wyoming or the number of orogenic events involved. What is apparent is that successive tracts of relatively primitive rocks were progressively accreted to the western margin of Laurentia during and following the 1.9–1.8 Ga growth of the continent and inclusion of Wyoming into Laurentia (Mueller et al. 2005).

The collage of mainly metamorphosed arc-like igneous rocks and associated metasedimentary assembledges of the Selway terrane ranges in age from ~2.45 to 1.6 Ga in western Montana and central Idaho. Crust of this age extends southward to the southern part of the Idaho–Atlanta batholith in southwestern Idaho (Norman and Mertzman 1991) and into the Farmington Canyon complex in northern Utah. Farmington complex rocks include Precambrian metasedimentary rocks, quartzofeldspathic gneisses, and amphibolites (Hedge et al. 1983; Bryant 1988). Our recent U–Pb geochronology of zircons (via ion probe) extracted from lithologies within the layered metasedimentary succession indicates that, although much of the detritus is likely to be Archean, younger (~2.4 Ga) zircons are also present; this demonstrates a Proterozoic (not Archean) depositional age for the protoliths (Mueller et al. 2004*a*). In addition, U–Pb ages of zircons extracted from orthogneisses in the northern part of the complex give a distinct crystallization age of ~2.45 Ga as opposed to the age of ~1.8 Ga reported from discordant multi-grain zircon data by Hedge et al. (1983). U–Pb monazite and $^{40}Ar/^{39}Ar$ studies (Barnett et al. 1993; Nelson et al. 2002) indicate that these rocks experienced amphibolite facies metamorphism at ~1.8 Ga.

Archean rocks exposed in the Grouse Creek block and under the western Snake River Plain (Leeman et al. 1985; Wolf et al. 2005) disrupt the belt of continuous Paleoproterozoic crust west of the Wyoming craton and partially separate the Selway terrane from the Mojave block, which contains crust of similar age. It is not yet clear if the Grouse Creek block is part of the Wyoming craton, separated by the Farmington complex, or a unique Archean block within the Paleoproterozoic accreted material of the Selway terrane and possibly the Mojave block.

Paleoproterozoic underplating and metamorphism

Lithoprobe Deep Probe seismic data reveal that the Great Falls tectonic zone and adjacent northern Wyoming Province are underlain by thick lower crustal layers with P-wave velocities >7 km/s that probably represent a very thick mafic underplate (Gorman et al. 2002). High P-wave velocity layers exist beneath both Archean provinces and the Paleoproterozoic terranes, and they have been variably interpreted to stem from Archean (Chamberlain et al. 2003), Paleoproterozoic (Gorman et al. 2002), and Mesoproterozoic (Chamberlain et al. 2003) processes. Correspondence among late Archean magmatism, the Wyoming Province >7 km/s layer (Chamberlain et al. 2003), and velocity gradients near the Great Falls tectonic zone (Gorman et al. 2002; Chamberlain et al. 2003) argue for separate, potentially Archean origins, whereas Proterozoic lower crustal xenoliths and secondary isotopic compositions of Tertiary volcanic rocks from the Great Falls tectonic zone derived from depths of the underplate (Rudnick et al. 1993; Carlson and Irving 1994; Gorman et al. 2002) and the occurrence of the layers on both sides of Proterozoic–Archean boundary argue for Proterozoic origins. Mafic dykes in the Wyoming Province with ages of 2.1–2.0 and 1.4–1.5 Ga (Chamberlain et al. 2003; Mueller et al. 2004), which could be related to post-Archean rifting of Wyoming and the development of the Belt basin respectively, may also hint at the age(s) of the mafic lower crust.

The presence of late Paleoproterozoic anorthosite in the lower middle crust exposed in the Clearwater metamorphic core complex, Idaho (Fig. 3) (Doughty and Chamberlain 2004) and anorthosite of perhaps similar age in the Bass Canyon area of the northern Bitterroot Mountains, Montana (Fig. 2) could be the middle crustal expression of a Paleoproterozoic (ca. 1.8–1.77 Ga) mafic lower crust. This inferred age for mafic lower crust in the Great Falls tectonic zone and Selway terrane is consistent with the age of the crustal xenoliths and the Nd-depleted mantle model ages of Cretaceous tonalitic plutons, like the Tobacco Root batholith, that are the result of partial melting of mafic lower crust (Mueller et al. 1996).

Most of the exposed rocks with the Great Falls tectonic zone give crystallization ages between ~1.86 and ~1.82 Ga (Mueller et al. 2002, this study). These rocks are all older than the granulite facies metamorphism and partial melting that overprinted Archean rocks in the Tobacco Root and Highland mountains at ~1.77 Ga (Mueller et al. 2005). $^{40}Ar/^{39}Ar$ thermochronology in the Tobacco Root Mountains (summary by Brady et al. 2004) and Little Belt Mountains (Holm and Schneider 2002; Vogl et al. 2003) indicates that relatively rapid cooling and exhumation of the high-grade metamorphic rocks occurred after ~1.77 Ga. This sequence of events is consistent with a collision between a ca. 1.86 Ga Little Belt magmatic arc, constructed on the edge of the Medicine Hat block, with the northwestern Wyoming Province at ~1.81 Ga. Crustal thickening and thrusting associated with this collision caused the leading edge of the Wyoming Province crust to be tectonically buried beneath the Little Belt arc (Mueller et al. 2005; O'Neill, 1998; Sims et al. 2004; Vogl et al. 2004*a*). By ~1.77 Ga, postorogenic extension and rapid exhumation of this orogenic belt resulted in granulite facies metamorphism and partial melting (Mueller et al. 2005) and perhaps metamorphic core complex formation in the Highland Mountains (O'Neill et al. 1988). Postorogenic extension is commonly associated with lithospheric delamination and mafic magmatism (Collins 2002). Such a process could have produced some of the Paleoproterozioc mafic lower crustal underplate that also appears to have been emplaced at ~1.80–1.75 Ga (Davis et al. 1995).

Basement influence on post-Paleoproteozoic evolution

The Wyoming Archean lithospheric mantle and lower crust were highly depleted in heat-producing elements by an extensive crust-forming event in late Archean time, and were thus depleted and refractory compared with the accreted Proterozoic arcs and mobile belts. The resultant differences in the overall composition and strength of the lithosphere (and tectosphere) between the provinces were apparently well established prior to Mesoproterozoic time. The distribution of this Paleoproterozoic lithosphere west of the Wyoming craton appears to have had a significant influence on the development of Mesoproterozoic, Neoproterozoic, and Phanerozoic sedimentary basins, magmatic provinces, structural provinces, and metallogenesis in the northern Rocky Mountains (O'Neill and Lopez 1985; O'Neill 1998; Foster et al. 2002; Sims et al. 2005).

Belt basin

The large-scale continental rifting that led to the development of the Mesoproterozoic Belt basin largely took place within Proterozoic crust of the Selway terrane along the edges of the Archean blocks (Fig. 9). The coincidence of the basement provinces and the margins of the Belt basin is clearer when the Belt rocks are palinspastically restored to the prethrust position following the reconstruction of Price and Sears (2000). The southeastern margin of thick Belt supergroup (and presumably a major rift boundary fault) is located just to the west of the outcropping Archean rocks in southwestern Montana. The Helena embayment is a significant eastward arm of the Belt rift, which developed where the Great Falls tectonic zone intersects the basin (O'Neill and Lopez 1985; Winston 1986). The southern margin of the Helena embayment lies along the northern edge of Wyoming craton, whereas the northern margin is along the Lewis and Clark fault zone and the southern margin of the Medicine Hat block. The Helena embayment is, therefore, largely coincident with the Great Falls tectonic zone. Major intrabasin, down-to-the-south faults lie along the Lewis and Clark zone as well, suggesting that the Medicine Hat block may be present beneath the northeastern part of the Belt basin. It appears that Mesoproterozoic Belt rifting largely avoided the areas of thick, relatively stronger lithosphere of the Wyoming and Medicine Hat blocks and was concentrated in the Paleoproterozoic lithosphere or in zones of mixed Archean and Proterozoic crust.

Phanerozoic magmatism

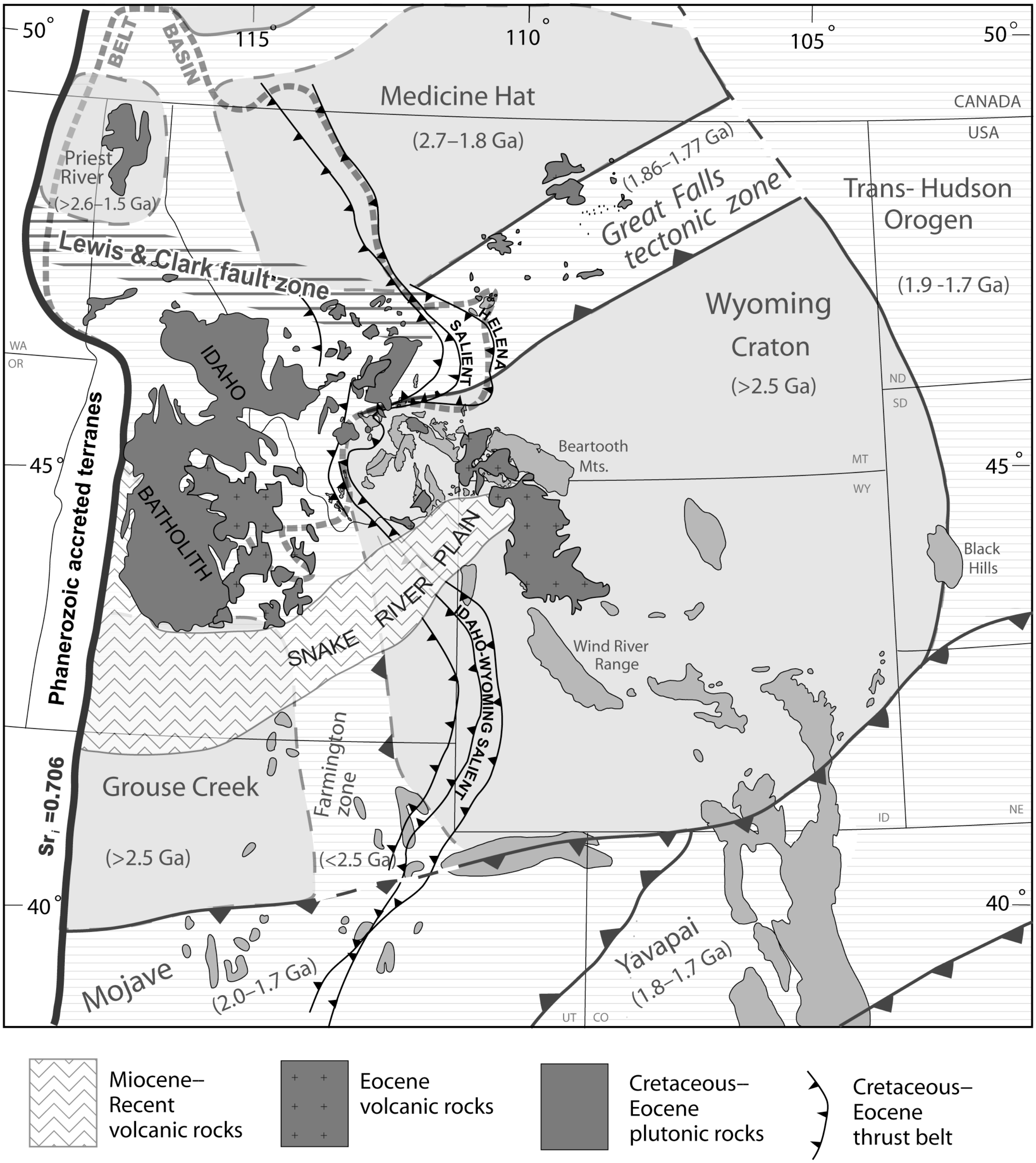
The Bitterroot lobe of the northern Idaho batholith is unique when compared with the other Mesozoic–Tertiary continental margin batholiths in the North American Cordillera orogen because it intruded thick Precambrian continental crust (Hamilton 1978; Hyndman 1983). Elsewhere in the Cordilleran orogen, Cretaceous plutons that intruded east of the Sr_i 0.706 isotopic line (Armstrong et al. 1977; Kistler and Peterman 1978) possess highly evolved isotopic signatures and are far less voluminous (Armstrong and Ward 1991; Elison et al. 1990; Miller and Barton 1990). In southwestern Montana, Cretaceous–Tertiary magmatic provinces extend eastward from the Idaho batholith through and beyond the eastern limit of the foreland fold–thrust belt. Large tracts of foreland magmatism in the Cordillera are only found south of the Wyoming craton in the Colorado Province (Karlstrom et al. 2002). The margin-orthogonal Cretaceous Southwest Montana granitic province, as well as the Eocene Challis volcanic province, Absaroka volcanic province, and Montana alkaline province are mainly restricted to the Great Falls tectonic zone or parts of the Wyoming craton thought to contain Paleoproterozoic mafic lower crust (Fig. 9). This implies that the relative fertility of the Paleoproterozoic lithosphere controlled the nature and distribution of Phanerozoic magmatic activity.

Paleoproterozoic lithosphere in the northern Rocky Mountains, therefore, appears to have been far more fertile for partial melting and more easily reactivated than the lithospheres of the Archean Wyoming and Medicine Hat – Hearne cratons. We hypothesize that the anomalous characteristics and fertility of this region are because of an integrated Paleoproterozoic–Mesoproterozoic history of enrichment in incompatible and heat-producing elements by primitive, asthenospheric-derived material (cf. Hyndman et al. (2005) for the modern Cascade subduction system). Partial melting of the accreted crust and possibly subduction-metasomatized mantle could have produced much more tonalitic to granitic magma than areas underlain by the Archean Wyoming craton to the south and Medicine Hat – Hearne province to the north. The area underlain by the ~2.4–1.7 Ga accreted crust of the Selway Province could, therefore, partly explain the shape and distribution of the Idaho batholith and southwest Montana granitic province (Fig. 9).

Phanerozoic structures

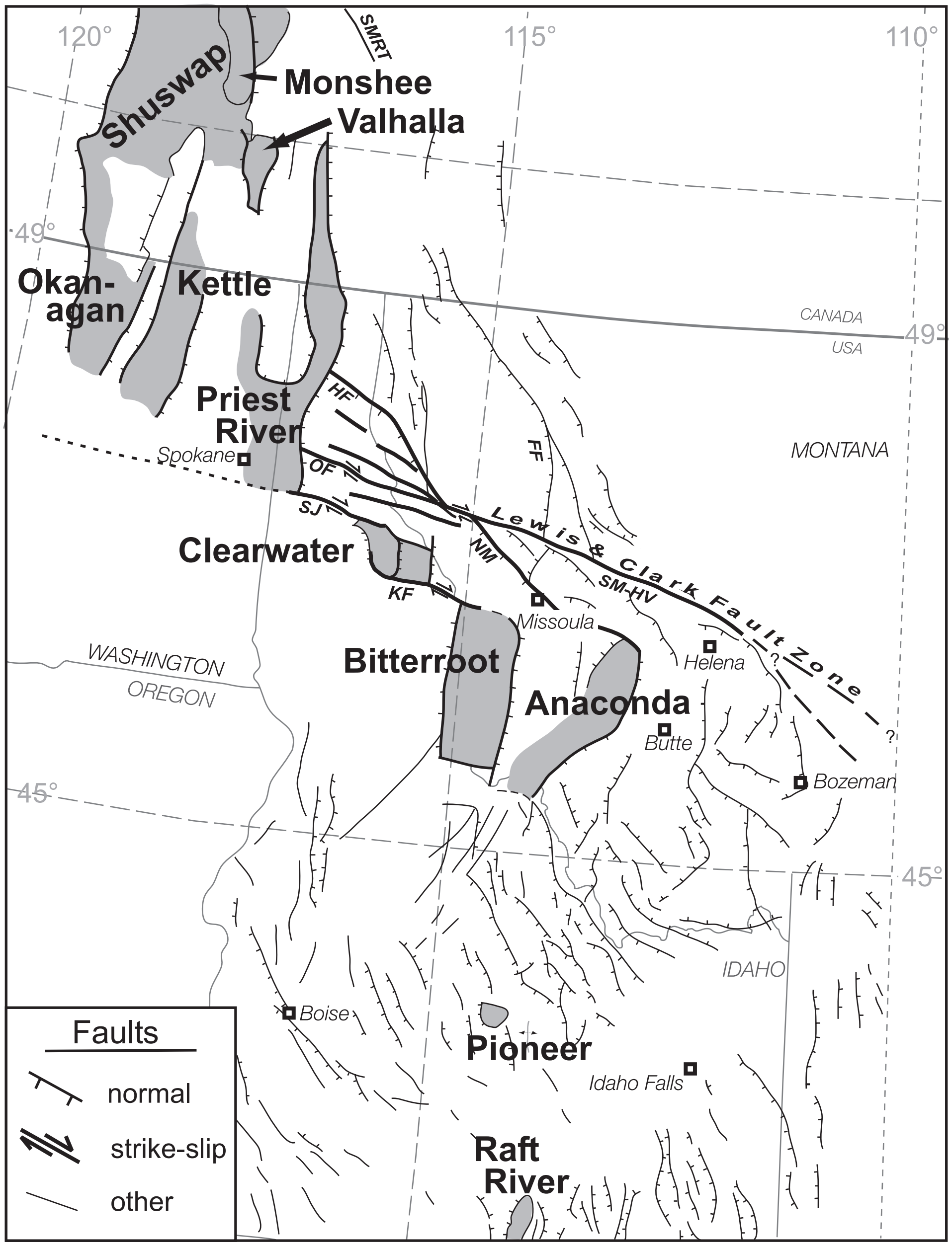
Large Laramide-style uplifts cored by basement in Southwest Montana do not extend west or northwest of the boundaries among the Wyoming Province, Great Falls tectonic zone, and Selway terrane (Fig. 9). The formation of these thick-

Fig. 9. Map showing the distribution of Cretaceous and Tertiary igneous rocks (medium gray, plutons; medium gray with crosses, volcanic rocks), fold–thrust belt structures, and basement exposures in Laramide-style uplifts (dark gray) in the northern Rocky Mountains overlain on the basement provinces from Fig. 1.



skinned, basement-involved structures appears to be restricted to the Archean domains in the northern Rocky Mountains. If these foreland structures were formed by Laramide flat-slab subduction (Bird 1998), then subduction was either steeper in central Montana or the Paleoproterozoic lithosphere responded differently than the Archean lithosphere to shallow subduction. We suggest that the tectonic environment was similar, but that the relatively stronger Archean crust de-

Fig. 10. Map of Tertiary faults and metamorphic core complexes along with associated transfer faults active during the Tertiary (modified from Foster et al. 2006).



formed mainly by reactivation of older structures with limited shortening compared with the Proterozoic crust in the northern Rocky Mountains.

The location of the east-vergent, thin-skinned Cordilleran fold–thrust belt in southwestern Montana – central Idaho also appears to be defined by the geometry of the eastern margin of the Belt basin, Great Falls tectonic zone, Northwest Wyoming craton, and the Lewis and Clark strike–slip fault system. Thin-skinned thrusting east of the Belt basin in southwestern Montana is limited to relatively thin Paleozoic platform sequences (Sears et al. 1988; Kalakay et al. 2001). The Helena salient in the fold–thrust belt extends just east of the Helena embayment of the Belt basin and is truncated to the north by the Lewis and Clark fault system, which probably developed during Belt rifting along the southern boundary of the Medicine Hat block. To the south, the thrust belt swings west around the promontory of the Dillon block, where there is some interaction between Sevier-style thrusts and Laramide-style basement uplifts.

The locus of major Eocene extension and metamorphic core complex formation in the northern Rocky Mountains, like the Bitterroot, Anaconda, Clearwater, and Priest River complexes, is restricted to the area underlain largely by Paleoproterozoic basement or mixed Paleoproterozoic–Archean basement (Foster et al. 2006) (Fig. 10). Furthermore, kinematic linkages between some of the large core complexes follow reactivated splays of the Precambrian structures of the Lewis and Clark zone (Doughty and Sheriff 1992; Foster et al. 2003, 2006). This may partly be because of the locus of voluminous Eocene magamatism and maximum crustal thickening in the Cordilleran orogen or that Laramide subduction produced areas of more intense hydration (Humphreys et al. 2003), but these features also seem to be dominantly within the Paleoproterzoic areas. Metamorphic core complexes south of the Snake River Plain, like the Ruby – East Humblolt and Raft River – Grouse Creek complexes, expose some Archean rocks, but these are highly overprinted and intercalated with younger rocks during multiple metamorphic events in the Proterozoic. The Great Basin core complexes also occur within a part of the Cordillera that has a much thicker Neoproterozoic–Phanerozoic sedimentary section because it lies west of the shelf hinge line (Bond et al. 1984) and, therefore, probably has a thinner, more fertile lithosphere than the basalt-depleted Archean lithosphere characteristic of the contiguous Wyoming craton (Humphreys et al. 2003).

Conclusions

U–Pb zircon data — along with Sr, Nd, and Pb isotopic data — demonstrate that Paleoproterozoic crust, which is at least in part juvenile in composition, occurs immediately west of the exposed Archean rocks of the Northwest Wyoming craton. Crystallization ages of basement slivers exposed in the cores of Phanerozoic thrust faults and ages of xenocrystic zircons from Phanerozoic plutonic rocks confirm a wider distribution of this Paleoproterozoic crust than indicated by the surface exposures. We interpret the evidence for abundant Paleoproterozoic crust and limited evidence for Archean crust to define the western limit of the Archean Wyoming craton. This crust, here named the Selway terrane, bounds the western Wyoming Province south of the Great Falls tectonic zone, perhaps as far as the Mojave Province. The Grouse Creek block, which lies generally west of the Selway terrane, contains heavily reworked Archean rocks that may or may not be part of the Wyoming craton. Rocks of the Selway and Grouse Creek blocks, therefore, represent important targets for developing piercing points for Neoproterozoic continental reconstructions, more so than the Archean rocks of the Wyoming craton. In this regard, it is important to note that the Selway terrane itself contains a complex and not yet deciphered record of Paleoproterozoic crustal evolution that spans almost 1 Ga.

The distribution of Precambrian structures and lithosphere strongly influenced the Phanerozoic evolution of the crust–mantle system of western North America. Structural and lithologic boundaries among the Archean Wyoming craton, Paleoproterozic Great Falls tectonic zone, Archean Medicine Hat block, and Proterozoic accreted terranes (Selway terrane) along the southwestern margin of Laurentia separate lithospheres of contrasting strength and fertility, such that these features controlled development of Phanerozoic structures and the overall tectonothermal histories of these blocks.

The correlation between basement provinces and Phanerozoic tectonic and magmatic features implies that the processes that formed and stabilized the Wyoming craton were different from those that formed the Proterozoic provinces that surround it. Variations in lithospheric strength and composition inherited from episodes of crustal growth and accretion explain many of the apparently anomalous features of the northern Rocky Mountains. In terms of fertility for later magmatism, mineralization, and crustal reactivation, the Great Falls tectonic zone and contiguous Paleoproterozoic crust (accreted arcs?) west of the Wyoming Province in the Selway terrane appear to share many similarities to the Proterozoic accretionary belts south of the Cheyenne Belt (Karlstrom et al. 2002).

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