

SHRIMP U-Pb dating of recurrent Cryogenian and Late Cambrian–Early Ordovician alkalic magmatism in central Idaho: Implications for Rodinian rift tectonics

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ABSTRACT

Composite alkalic plutonic suites and tuffaceous diamictite, although discontinuously exposed across central Idaho in roof pendants and inliers within the Idaho batholith and Challis volcanic-plutonic complex, define the >200-km-long northwest-aligned Big Creek–Beaverhead belt. Sensitive high-resolution ion microprobe (SHRIMP) U-Pb zircon dates on these igneous rocks provide direct evidence for the orientation and location of the Neoproterozoic–Paleozoic western Laurentian rift margin in the northern U.S. Cordillera. Dating delimits two discrete magmatic pulses at ca. 665–650 Ma and 500–485 Ma at the western and eastern ends, respectively, of this belt. Together with the nearby 685 Ma volcanic rocks of the Edwardsburg Formation, there is a 200 Ma history of recurrent extensional magmatic pulses along the belt. A similar history of recurrent uplift is reflected in the stratigraphic record of the associated miogeoclinal and cratonal platform basins, suggesting that the Big Creek–Beaverhead belt originated as a border fault during continental rift events. The magmatic belt is paired with the recurrently emergent Lemhi Arch and narrow miogeoclinal facies belts and it lies inboard of a northwest-striking narrow zone of thinned continental crust. These features define a northeast-extending upper-plate extensional system between southeast Washington and southeast Idaho that formed a segment of the Neoproterozoic–Paleozoic miogeocline. This segment was flanked on the north by the St. Mary–Moyie transform zone (south of a narrow southern Canadian upper-plate margin) and on the south by the Snake River transfer zone (north of a broad Great Basin lower-plate margin).

These are the central segments of a zigzag-shaped Cordilleran rift system of alternating northwest-striking extensional zones offset by northeast-striking transfers and transforms. The data substantiate polyphase rift and continental separation events that included (1) pre- and syn-Windermere rifting, (2) Windermere margin subsidence, (3) late Ediacaran–Cambrian rifting, and (4) well-developed late Ediacaran–Devonian passive margin subsidence and deposition. Timing and geometries support synchronous but opposing divergence along Cordilleran and Atlantic rifts with a junction in Southern California–Sonora.

INTRODUCTION

Neoproterozoic to early Paleozoic continental rifting of Rodinia formed the western margin (present coordinates) of Laurentia. Along much of its length, the margin geometry is well documented by lithofacies or fossil-community data that define the resultant passive-margin sedimentary (miogeoclinal) belts (Stewart and Suczek, 1977; Stevens, 1991; Cecile et al., 1997) or by isotopic data that define the edge of unexposed Precambrian basement (Armstrong et al., 1977; Armstrong, 1988; Elison et al., 1990; Wright and Wooden, 1991; Tosdal et al., 2000). However, between the southern Canadian Cordillera and the northern Great Basin, there is a significant gap in rocks identified as part of the Cordilleran miogeocline (Fig. 1). Most depictions of the miogeocline connect exposures in northeastern Washington (south end of Canadian Cordillera) to those in southeastern Idaho (northern Great Basin), assuming a lack of preservation across the Idaho batholith (Stewart, 1991; Poole et al., 1992; Dickinson, 2006). Other depictions assume that most of this Laurentian-margin segment, including the miogeocline, was removed by a possible Mesozoic rift event (Davis et al., 1978).

As it is known from Canadian and Great Basin segments, and as it is projected through the northern U.S. Cordillera, the miogeocline forms a sinuous, zigzag shape (Stewart, 1972; Aitken, 1993; Speed, 1994; Sears and Price, 2003). Additionally, palinspastic reconstructions (Stewart and Poole, 1974; Stevens, 1981; Abbott et al., 1986; Levy and Christie-Blick, 1989) indicate that the present sinuous trace of the Cordilleran miogeocline represents the shape of the rift margin as it persisted through geologic time despite subsequent deformation (Aitken and Long, 1978; Beutner, 1977; Stevens, 1981; Oldow, 1984; Thompson et al., 1987).

In eastern Laurentia, embayments and promontories along that rift margin (terminology as defined by Rankin, 1976; Marshak, 2004) formed as extensional and transform (offsetting extensional segments of like asymmetry) or transfer (offsetting segments of different asymmetry) segments of the Neoproterozoic Appalachian rift margin (Thomas, 1977, 2006). Furthermore, differences in the width and thickness of associated sedimentary rocks and the relative thinning of basement in the different segments were the result of asymmetric extension (Thomas, 1977).

Similar arguments for the origin and character of rift-margin segments have been made for the Cordillera in Canada (Hansen et al., 1993; Cecile et al., 1997) and for north-central Nevada (Tosdal et al., 2000). In those region-specific models, northeast-southwest extension is proposed for specific areas (Canadian Cordillera and Great Basin). However, in more generalized portrayals, northwest-southeast extension is depicted (Speed, 1994; Sears and Price, 2003; Poole et al., 2005; Dickinson, 2006). The lack of information about miogeoclinal rocks in the northern U.S. Cordillera has made it difficult to connect margin segments and to resolve the different kinematic models. Recent studies in central Idaho, including identification of metamorphosed Cryogenian (following usage of

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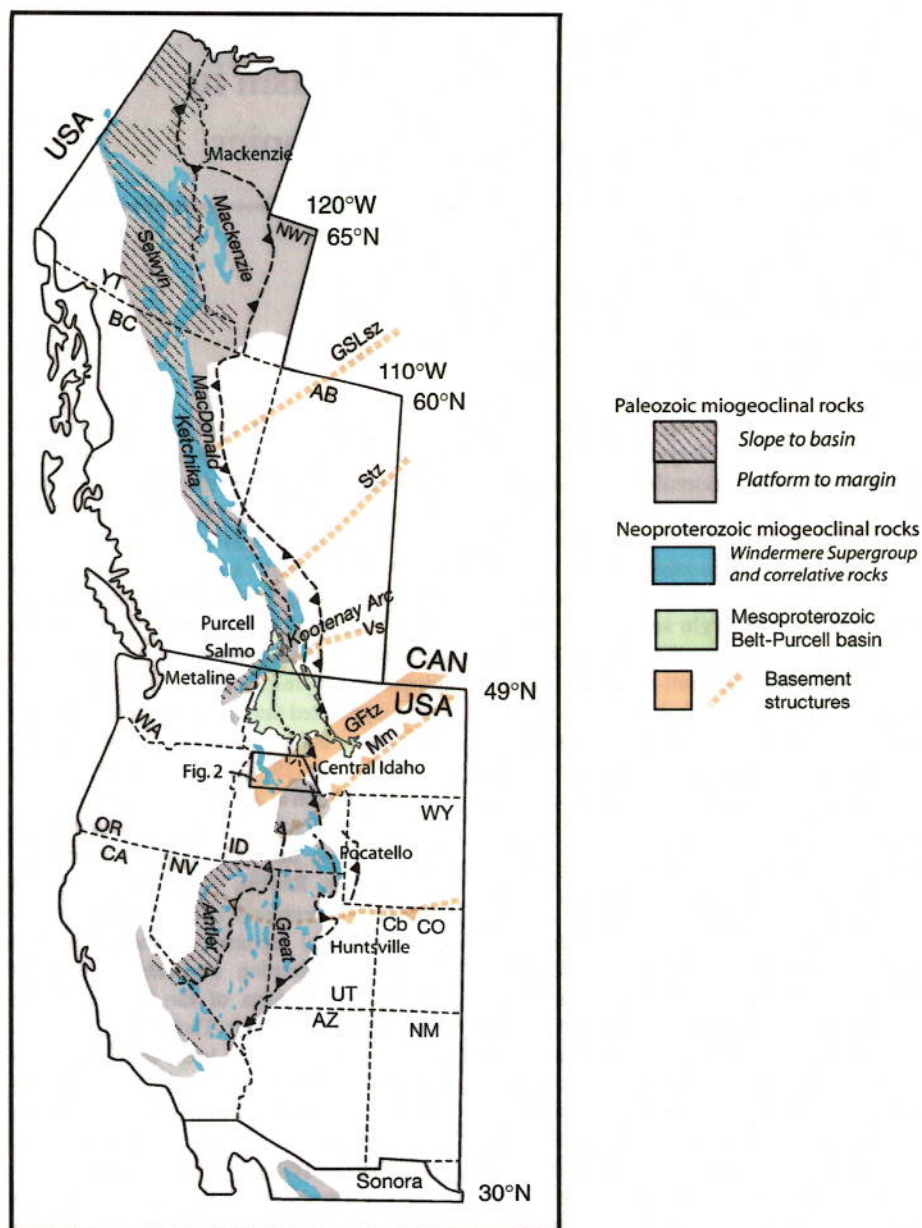


Figure 1. Index map of U.S. and Canadian Cordillera showing location of Neoproterozoic and Paleozoic miogeoclinal strata in relation to basement structures and younger thrust belts. Data were compiled from Ross et al. (2001), Lund et al. (2003a), Sims et al. (2005), and Lund (2008). YT—Yukon Territory; BC—British Columbia; AB—Alberta; GSLsz—Great Slave Lake shear zone; Vs—Vulcan Low; Stz—Snowbird tectonic zone; GFTz—Great Falls tectonic zone; Mm—Madison mylonite zone.

Plumb, 1991; Knoll et al., 2004) to Cambrian(?) rocks of the Windermere Supergroup, have advanced the understanding of depositional environment, continuity, and geometry of the late Neoproterozoic rocks through the northern U.S. Cordillera (Lund et al., 2003a).

In the absence of a well-preserved miogeoclinal section in central Idaho, the age and

genesis of any potential Neoproterozoic and Paleozoic rocks are important for determining the geometry and evolution of the western edge of Laurentia in the northern U.S. Cordillera. In east-central Idaho, southeast of the known extent of west-central Idaho Windermere rocks, an undated tuffaceous diamictite that underlies poorly dated Neoproterozoic–Cambrian(?) mio-

geoclinal rocks (Hobbs and Hayes, 1990) has been suggested to be Neoproterozoic (Jacob, 1990; Oliver and Blackwell, 2002) and may extend the Windermere rocks southeast of the known exposures. A series of seven syenitic plutons is exposed east of the Idaho batholith and in roof pendants within it. Several of the plutons are known to be early Paleozoic in age, whereas others are possibly older (Cater et al., 1973; Evans and Zartman, 1988). They form a northwest-striking belt across central Idaho, herein named the Big Creek–Beaverhead belt, which is parallel to, but east of, the belt of Windermere Supergroup rocks. Similar alkalic igneous rocks exposed along the miogeoclinal belt of the Canadian Cordillera are dated as Neoproterozoic and early Paleozoic and are interpreted to be manifestations of continental rifting (Goodfellow et al., 1995). Resolution of the ages of the syenitic plutons in the Big Creek–Beaverhead belt and of other possible Neoproterozoic stratified rocks is important to clarify the regional tectonic history.

The purposes of this paper are to: (1) describe the setting and character of syenitic plutons that cross central Idaho, (2) present U-Pb data for the plutonic suites and for the diamictite, and (3) discuss the origins of these rocks and their setting with respect to the Neoproterozoic to early Paleozoic rift tectonics of the Cordillera. Our new age data help to more fully describe the timing, geometry, and kinematics of the breakup of the Rodinian supercontinent as manifested along the western margin of Laurentia.

GEOLOGIC SETTING

Central Idaho is a complex region of metamorphosed sedimentary rocks intruded by voluminous plutons of several ages, including Mesoproterozoic and Proterozoic–early Paleozoic plutons, the Late Cretaceous Idaho batholith, and the Eocene Challis plutons. Because the pre-Cretaceous rocks are preserved as roof pendants and inliers in the Idaho batholith and Challis volcanic-plutonic complex and because of inaccessibility, many questions remain about the tectonic history of these complex rocks. The age and origin of the metasedimentary country rocks are poorly known but were long considered to be either Mesoproterozoic (Bond, 1978) or Archean (Harrison, 1972). However, recent mapping and dating studies document Neoproterozoic volcanic, glaciogenic, and basinal strata in a fragmented northwest-trending facies belt across west-central Idaho (Lund et al., 2003a, 2008). The Neoproterozoic rocks, considered to be part of the Windermere Supergroup (as used by Link et al., 1993), include the volcanic Edwardsburg Formation of probable continental-rift

origin (Lund et al., 2003a), which is similar in composition, age, and origin to Windermere volcanic units in northeastern Washington and southeastern Idaho (Miller et al., 1973; Harper and Link, 1986; Devlin et al., 1988; Colpron et al., 2002; Fanning and Link, 2004). Although the Neoproterozoic rocks of west-central Idaho do not completely link Neoproterozoic belts in northeastern Washington to those in southeastern Idaho-Utah, their northwest trend implies the orientation and location of Neoproterozoic rift structures and of the resultant passive margin through the region (Lund et al., 2003a).

The stratigraphy of the Windermere Supergroup in west-central Idaho (west of about longitude 115°W) is based on composite sections of upper-greenschist- to lower-amphibolite-facies rocks in roof pendants (Gospel Peaks and Big Creek) and isolated exposures (Fig. 2). In these, the basal Windermere unit is the Edwardsburg Formation, which is composed of two diamictites interlayered with mafic to felsic volcanic

rocks that are dated at 685 ± 7 Ma and 684 ± 4 Ma. These are succeeded by two thick carbonate, phyllite, and siltite successions that are separated by another level of diamictite. The uppermost Windermere unit is a thick conglomerate and dominantly quartzite unit. In west-central Idaho, Windermere rocks above the Edwardsburg Formation are undated. Tuffaceous and volcanoclastic diamictite and associated carbonate and phyllitic rocks are described from a drill core in east-central Idaho (Jacob, 1990; Oliver and Blackwell, 2002), the tuff matrix from diamictite is dated in the present study.

Within this region of igneous and metasedimentary rocks, seven syenitic plutonic suites (composed of interdigitated mafic and felsic syenitic rocks) are exposed in a northwest-striking belt across central Idaho, herein referred to as the Big Creek–Beaverhead belt (Fig. 2). Three eastern suites (Deep Creek, Arnett Creek, and Beaverhead) are constrained to be between Late Cambrian and Early Ordovician age, based

on several dating techniques (Ruppel, 1968; Scholten and Ramspott, 1968; Skipp, 1984; Evans and Zartman, 1988). The three western suites (Ramey Ridge, Acorn Butte, and Rush Creek Point) are only constrained by cross-cutting relations to be between Cretaceous and Mesoproterozoic but were inferred to be Neoproterozoic (Leonard, 1963; Cater et al., 1973) based on U-Pb data from the Ramey Ridge suite, which suggest an approximate age of 725 Ma (Leonard and Stern, 1966). The age of the central plutonic suite (Yellowjacket) has been speculated to range from Tertiary to Proterozoic age (Ross, 1934; Anderson, 1963; Ekren, 1988). The easternmost suite, the Beaverhead pluton, is reported to have intruded Middle Ordovician miogeoclinal strata (Scholten and Ramspott, 1968; Skipp, 1984) and to contain Mesoproterozoic metasedimentary xenoliths (Scholten and Ramspott, 1968; Evans and Zartman, 1988). The six western suites intrude Mesoproterozoic metasedimentary strata of varying metamorphic

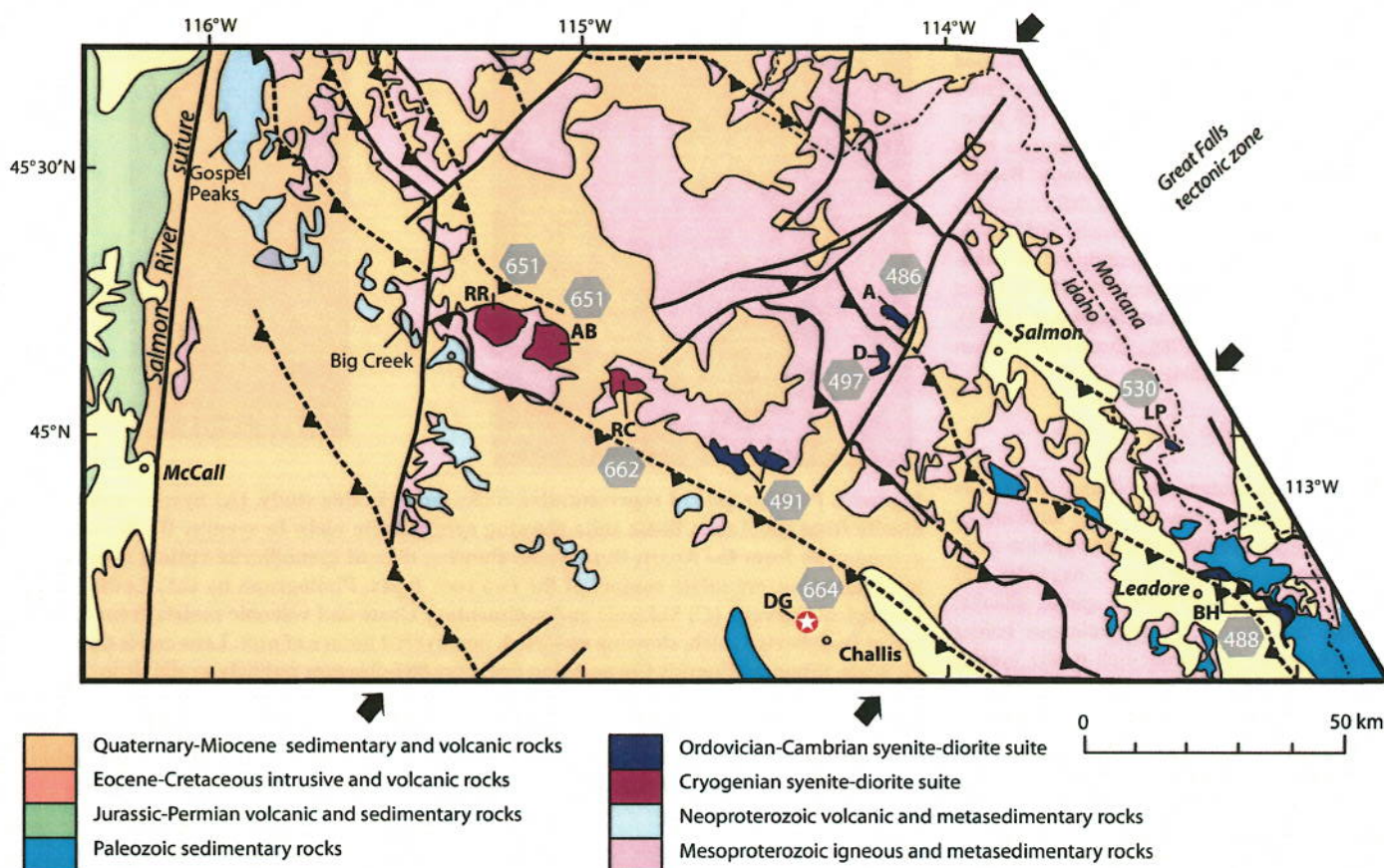


Figure 2. Generalized geologic map of central Idaho showing location of Cryogenian and Late Cambrian–Early Ordovician syenite-diorite suites and their ages. Figure also shows location of Gospel Peaks and Big Creek roof pendants containing Neoproterozoic Edwardsburg Formation. Map was compiled from Evans and Green (2003) and Lund (2004). RR—Ramey Ridge suite; AB—Acorn Butte suite; RC—Rush Creek Point suite; Y—Yellowjacket suite; D—Deep Creek suite; A—Arnett Creek suite; DG—Daugherty Gulch drill site; BH—Beaverhead pluton.

grade, and the suites are preserved in roof pendants and inliers in the Late Cretaceous Idaho batholith and Eocene Challis-age volcanic rocks. Those western suites lie parallel to, but a few kilometers north of, Neoproterozoic volcanic rocks of the Edwardsburg Formation (Lund et al., 2003a; Lund, 2004). Despite their close geographic relations, the plutonic suites are not known to have intruded Neoproterozoic strata.

SYENITE-DIORITE SUITES

Compositional, textural, and internal intrusive relationships of the seven plutonic suites are remarkably similar. They are each composite intrusive bodies composed of felsic (syenite, quartz syenite, and syenogranite) and mafic hornblende-rich (syenodiorite, amphibolite, and minor pyroxenite and gabbro) compositions. Within these composite plutonic bodies, the felsic and mafic phases have mutual crosscutting relations, indicating commingling of two related but chemically distinct magmas (Fig. 3A). The western plutonic suites have undergone greenschist-facies regional metamorphism and penetrative ductile deformation as well as localized deformation along faults (Lund, 2004). The eastern suites are less penetratively deformed and have undergone only minor metamorphism; the easternmost, Beaverhead, is extremely altered, and deformation is localized along structures (Evans and Green, 2003). The following descriptions are summarized primarily from Leonard (1963), Cater et al. (1973), Scholten and Ramspott (1968), Skipp (1984), Ekren (1988), Evans and Zartman (1988), and Lund (2004).

Ramey Ridge Suite

The Ramey Ridge suite (Lund, 2004) has an inner phase of primarily gray, medium- to coarse-grained syenite and quartz syenite composed of alkali feldspar, quartz, magnetite, and biotite, with accessory zircon, apatite, allanite, titanite, and fluorite. A discontinuous border phase is mafic hornblende-rich quartz syenite and syenodiorite. Leonard (1963) described all units as both grading into and locally intruding each other, but he interpreted the mafic border phase as mafic metavolcanic xenoliths. We interpret the mafic inclusions to be comagmatic mafic enclaves, as indicated by the common occurrence of crenulate margins on the enclaves (as pictured from the Acorn Butte plutonic suite; Fig. 3A). Leonard and Stern (1966) reported a U-Th-Pb multigrain zircon age of ca. 725 Ma for the Ramey Ridge suite.

The Ramey Ridge suite intruded Mesoproterozoic metasedimentary strata interpreted

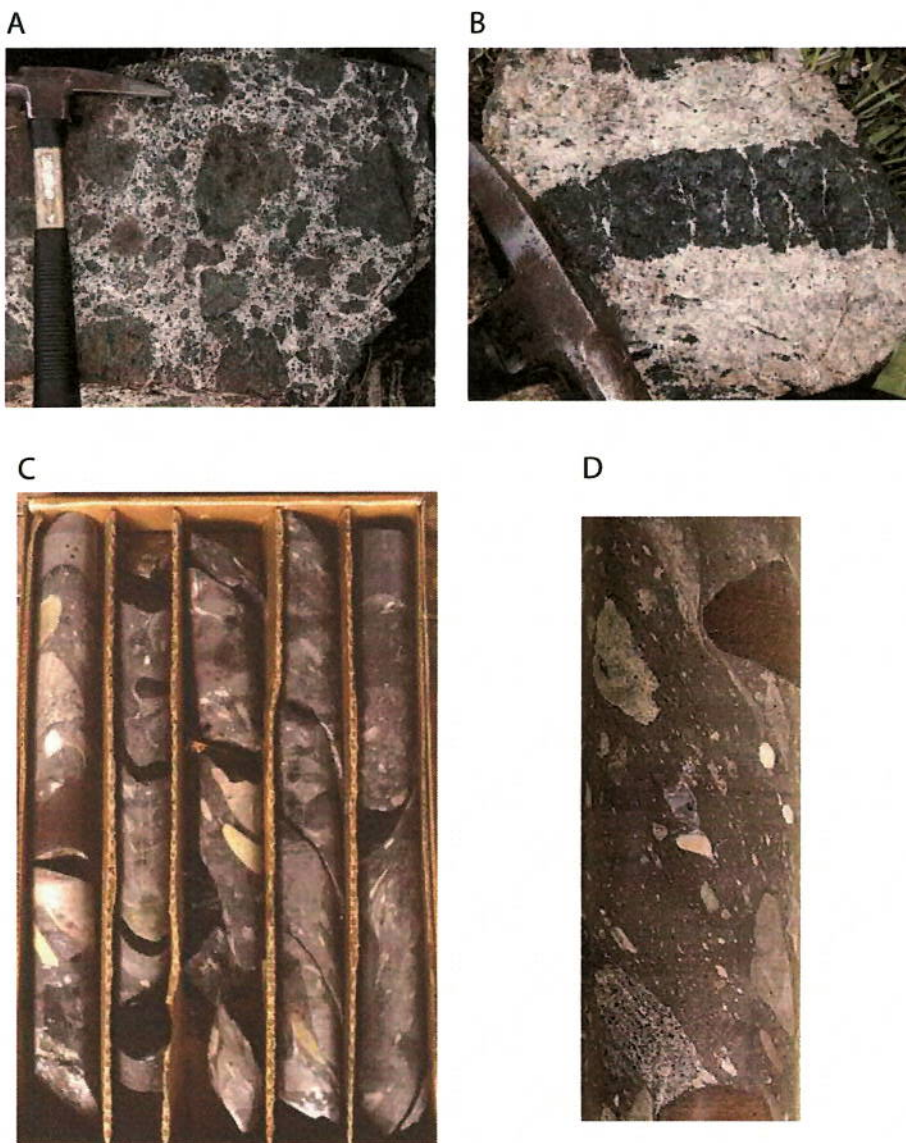


Figure 3. Photographs of representative rocks dated in this study. (A) Syenite and syenodiorite from the Acorn Butte suite showing syenodiorite blebs in syenite. (B) Syenite and syenodiorite from the Acorn Butte suite, showing dike of syenodiorite cutting syenite and interdigitating, crenulate contact of the two rock types. Photograph by R.S. Lewis (Idaho Geological Survey). (C) Volcanic and sedimentary clasts and volcanic matrix from diamictite of Daugherty Gulch, showing unsorted, nonlayered nature of unit. Lens cap is 6 cm wide for scale. (Shama Minerals Corporation drill core #87-3 is now publicly available in the U.S. Geological Survey Core Library, Denver, Colorado.) Core is ~6 cm wide. (D) Diamictite of Daugherty Gulch from core, showing detail of unsorted matrix in dated portion of core and deformation of volcanic clasts (light colored) against sandstone clasts (dark reddish) in upper right and bottom of view. View is ~6 cm wide.

by Lund (2004) to correlate with part of the Lemhi Group and Swauger Quartzite in east-central Idaho (Ruppel, 1975; Tysdal, 2000). The Ramey Ridge suite hosts gold-silver-copper deposits in sheared quartz veins and pods (Lund et al., 1990; Cater et al., 1973).

Acorn Butte Suite

The Acorn Butte suite consists of medium- to coarse-grained syenite and quartz syenite, coarse-grained amphibolite (meta-syenodiorite), and medium-grained alkali gabbro. Mineralogically,

the syenite and quartz syenite are similar to rocks of the Ramey Ridge Complex, but, generally, biotite is more abundant than hornblende. Mafic phases, interpreted as comagmatic enclaves (Fig. 3A), range from hybrid mixtures to nearly pure amphibolite. Alkali gabbro consists of labradorite, clinopyroxene, and uraltic hornblende with accessory magnetite, biotite, and titanite. Rocks intruded by the Acorn Butte suite are interpreted to be Mesoproterozoic Lemhi Group and Swauger Formation (Lund, 2004). Cater et al. (1973) inferred a Proterozoic age for the Acorn Butte suite. Gold-silver-copper deposits are hosted in sheared quartz veins and lenses (Cater et al., 1973).

Rush Creek Point Suite

The Rush Creek Point suite consists of medium- to coarse-grained syenite and quartz syenite, coarse-grained amphibolite (meta-syenodiorite) and pyroxenite, and medium-grained gabbro. Biotite is more abundant than hornblende except for hornblende-rich comagmatic enclaves. Gabbro and pyroxenite parts of the suite are similar to those of Acorn Butte suite. The Rush Creek Point suite intruded the Mesoproterozoic Yellowjacket Formation and Hoodoo Quartzite (Lund, 2004). A Proterozoic age is inferred for the Rush Creek Point suite (Cater et al., 1973).

Yellowjacket Suite

The Yellowjacket suite includes gabbro, syenodiorite, quartz diorite, syenite, quartz syenite, and granite. Felsic rocks consist of alkali feldspar, albite, quartz, hornblende, biotite, magnetite, and accessory zircon. Mafic rocks consist of plagioclase, hornblende (after pyroxene), biotite, actinolite-tremolite, chlorite, magnetite, and accessory apatite. This suite intruded the Mesoproterozoic Yellowjacket Formation and Hoodoo Quartzite (Ekren, 1988; Tysdal et al., 2000). The suite was originally mapped as Tertiary (Ross, 1934; Anderson, 1953, 1963), later inferred to be Proterozoic (Cater et al., 1973), and reinterpreted as possibly Ordovician (Peale, 1982; Ekren, 1988) on the basis of a preliminary U-Pb zircon date of ca. 490 Ma (Evans, 1984).

Deep Creek Suite

The Deep Creek suite is dominantly medium-grained equigranular alkali syenite to alkali quartz syenite and minor syenodiorite. It is composed of perthitic alkali feldspar, biotite, and plagioclase, minor pyroxene and hornblende, and accessory zircon, rutile, and opaque minerals. Calcite and epidote may be alteration phases. The Deep

Creek suite intruded the Apple Creek Formation (Evans and Green, 2003). However, faulted contacts (Evans and Zartman, 1988; Evans and Green, 2003) resulted from Cretaceous-age tight anticlinal folding of the plutonic suite and its country rocks (Lund and Tysdal, 2007). An early Paleozoic age for the Deep Creek suite was inferred from preliminary U-Pb and Rb-Sr dating (Evans and Zartman, 1988).

Arnett Creek Suite

The composition of the Arnett Creek suite ranges from alkali syenite to alkali granite, with textures ranging from medium to coarse grained and equigranular to porphyritic. These rocks contain perthitic alkali-feldspar and two generations of biotite, as well as quartz, opaque minerals, and accessory zircon and rutile (Evans and Zartman, 1988). Calcite in the rocks may be an alteration product. The reported presence of miarolitic cavities (E.H. Bennett, *in* Evans and Zartman, 1988) suggests shallow (epizonal) emplacement. The Arnett Creek suite intruded the Mesoproterozoic Apple Creek Formation (Evans and Green, 2003) and hosts Late Cretaceous gold deposits. A U-Pb zircon date of 492 ± 39 Ma was previously reported for the Arnett Creek suite (Evans and Zartman, 1988).

Beaverhead Pluton

The Beaverhead pluton is predominantly a pink to red syenogranite with lesser amounts of leucosyenite and syenodiorite. Syenogranite is medium to coarse grained, equigranular to subporphyritic, and contains miarolitic cavities in some localities. It is composed of quartz, alkali feldspar, biotite, and opaque minerals, with accessory zircon and apatite. Alteration is widespread and produced sericite, hematite, leucosyenite, limonite, magnetite, chlorite, and quartz. Leucosyenite is a medium- to coarse-grained, one-feldspar (hypersolvus) rock that grades from alkali-feldspar syenite to alkali-feldspar quartz syenite. It consists of microcline-albite micropertite and cryptoperthite, quartz, and lesser biotite, opaque minerals, zircon, and apatite. Intrusive contacts with the Ordovician Kinnikinnick Quartzite are reported (Scholten and Ramspott, 1968; Skipp, 1984). The Beaverhead suite is preserved on the upper plate of the Hawley Creek thrust fault (Lucchitta, 1966; Evans and Green, 2003).

There have been several attempts to date the Beaverhead pluton, but a definitive age has remained elusive. Scholten and Ramspott (1968) cited a K-Ar age of 441 ± 15 Ma for biotite from a microsyenite dike, whereas

Ruppel (1968) reported Rb-Sr model ages of 410 ± 60 Ma for potassium feldspar and 405 ± 40 Ma for a whole-rock sample of granite. Skipp (1984) subsequently reported K-Ar dates from a meladiorite dike of 482 ± 16 (biotite) and 467 ± 16 (hornblende) Ma. Most recently, Evans and Zartman (1988) interpreted an age of ca. 483 Ma for the syenogranite based on $^{207}\text{Pb}/^{206}\text{Pb}$ ages from two discordant, nonmagnetic, multigrain zircon fractions. However, two additional magnetic fractions were discordant but not collinear with the first two, thereby preventing any straightforward interpretation based on concordia intercepts. An indirect age of ca. 500 Ma for this plutonic suite is provided by sensitive high-resolution ion microprobe (SHRIMP) analyses on detrital zircons from Pliocene to Holocene fluvial sediments deposited by streams that drained the Beaverhead pluton (Geslin et al., 2002).

Geochemical Data

Limited trace- and major-element data for the syenite-diorite suites and for the nearby Cryogenian volcanic rocks of the Edwardsburg Formation provide preliminary geochemical characterizations (see GSA Data Repository Tables DR1 and DR2 and Document 1¹). The geochemical data for the seven composite plutonic suites reflect similar characteristics among them and, although broadly similar, reflect some differences between the plutonic and older volcanic rocks (Fig. DR1 [see footnote 1]). The suites have dominantly ferroan tholeiitic composition (Figs. DR1A–DR1C [see footnote 1]) and are alkaline and metaluminous (Figs. DR1D–DR1F [see footnote 1]). Tectonic discrimination diagrams indicate that these formed as within-plate magmas (Figs. DR1G–DR1K [see footnote 1]). There is no systematic variation in geochemistry through time along the suites or with geography from west to east. A comparison with geochemical data for the Cryogenian metavolcanic Edwardsburg Formation indicates that the volcanic rocks are also ferroan tholeiitic-trend magmas but are alkali-calcic and peraluminous (Figs. DR1A–DR1F [see footnote 1]). In terms of origin, the trace-element data show that both the plutonic suites and the precursor Edwardsburg volcanic rocks formed from within-plate tholeiitic magmas.

¹GSA Data Repository item 2009240, includes selected trace-element and major-element geochemical data with plots and preliminary interpretation and TIMS and SHRIMP U-Pb data with TIMS plots and SHRIMP methods, is available at <http://www.geosociety.org/pubs/ft2009.htm> or by request to editing@geosociety.org.

SEDIMENTARY UNITS OF DAUGHERTY GULCH

A drill hole located 10 km west of Challis, east-central Idaho (Fig. 2), intersected rocks that are probably Cambrian to Neoproterozoic, but some or all of which are not known to crop out (Jacob, 1990). The drill hole is set in intensely folded and faulted, structurally juxtaposed, poorly dated, Neoproterozoic(?) to Middle Ordovician(?) miogeoclinal sections, in fragmentary exposures that are isolated within younger plutonic and volcanic rocks (Hobbs and Hays, 1990; Hobbs et al., 1991). The 1142-m-deep drill hole is collared in the Eocene Challis Volcanic Group (11 m) and, below that, intersects several sedimentary units interpreted as an uninterrupted, straight stratigraphic section. From top to bottom in the core, the pre-Eocene units are described as (1) gray, sandy, and phyllitic dolostone correlated with the Upper Cambrian–Lower Ordovician Bayhorse Dolomite; (2) dark gray, pyrite-bearing, carbonaceous phyllite correlated with the undated Middle Cambrian(?) Garden Creek Phyllite; (3) muscovite-bearing, medium-grained, white, dolomitic marble correlated with the undated dolomite of Bayhorse Creek; and (4) Neoproterozoic(?) lithic tuff (Jacob, 1990; Oliver and Blackwell, 2002).

At the bottom of the Daugherty Gulch drill hole, there are two types of diamictite showing lower-greenschist-facies metamorphism and prominent cleavage. The lower type is more than 55.2 m (181 ft as marked on the core, from upper intersection of this unit to the bottom of the hole) of cleaved, heterogeneous clast diamictite (Fig. 3C). The diamictite matrix is unsorted clay- to sand-size fractions containing a variety of minerals and lithic fragments. Quartz is the most abundant mineral (20%–30%) in the matrix and occurs as fractured grains, some strained grains, and some embayed or dipyratidial quartz grains. Another 25% of the matrix is lithic fragments, including quartz siltites, volcanic chips, and carbonate grains. The rest of the matrix is a mixture of fine-grained sericite, quartz, opaque grains, and much intermixed sericitic and carbonate-altered material that originated as feldspar or mafic minerals of a tuffaceous matrix. Tuffaceous matrix is dominant in the lowest parts of the core. Clasts make up ~50% of the diamictite and are as large as 25 cm in diameter. These are dominantly rhyolite, lesser quartzite and metasandstone (but these form the larger clasts), minor carbonaceous phyllite and dolostone, and a few plutonic clasts. The proportion of volcanic to sedimentary clasts decreases upward in the core. Volcanic clasts are light-yellow, pink, green, and gray. Relict volcanic textures such as spherulites

are present in clasts (Oliver and Blackwell, 2002). The source of the volcanic clasts is undetermined. Metasandstone and quartzite clasts are dominantly pink to reddish or buff-colored and are well rounded and undeformed. They are similar to the Mesoproterozoic Gunsight and Swauger Formations exposed in the nearby region. The less abundant phyllite and dolostone pebbles did not originate from Mesoproterozoic formations of the region but, rather, are similar to Neoproterozoic units such as those exposed in west-central Idaho (near Big Creek; Fig. 2). Phyllitic cleavage is moderately steep and defined by preferred orientation of metamorphic sericite, lamination of opaque grains that may be remnants of volcanic biotite (now destroyed by alteration), and elongation of lithic fragments in the matrix. In the matrix, slightly coarser sericite formed in pressure shadows where the fine-grained foliation is deflected around lithic fragments or quartz grains. Similarly, cleavage is deflected around clasts, particularly around the metasandstone to quartzite clasts. Volcanic clasts are elongated and aligned parallel to sericitic cleavage and help define the phyllitic fabric in hand specimen. The volcanic clasts are commonly plastically deformed where in contact with rounded but not stretched quartzitic clasts (Fig. 3D). Welded textures were inferred in an earlier study (B.F. Leonard *in* Jacob, 1990), but volcanic glass grains, feldspars, and mafic minerals are not preserved due to metamorphism and deformation. The metamorphic grade and fabric, reflected by sericitic recrystallization and phyllitic cleavage, are responsible for the wavy lamination observed in thin section.

The top 2.4 m (8 ft) of the diamictite is light-greenish, chloritic, unsorted siltite to metasandstone matrix. Grit- to pebble-sized clasts are dominantly quartzite and metasandstone; volcanic clasts that dominate the lower diamictite are uncommon. This upper clast-bearing zone does not have a tuffaceous matrix and is probably of different origin from the underlying rocks.

Above the diamictites, there is a 1082-m-thick section of lower-greenschist-facies, interlayered carbonate and carbonaceous phyllite. The lowest part of this section is 59.7 m of white, muscovite-bearing, medium-grained, dolomitic marble. Above that, 873.7 m of cleaved, dark-gray, pyrite-bearing, carbonaceous phyllite is gradationally interlayered on the centimeter scale with lighter-gray, calcareous and dolomitic phyllitic marble. In this thick unit, soft-sediment bedding features are preserved locally. The features include load structures that indicate the section is upright. These phyllites are folded, as evidenced by common variations in layer orientation, and some small folds are preserved in the core. Spaced cleavage is defined by sericite

foliation and is generally ~50° steeper than bedding. An upper 148.3-m-thick unit of sandy, phyllitic dolostone (described by Jacob, 1990) was not observed for this study (because this part of the core was not archived).

Published correlations for the rocks in the upper part of the core (Jacobs, 1990) were based partly on a succession of compositions that grossly matched nearby units (i.e., carbonate, phyllite, carbonate). However, in detail, units in the core are dissimilar to those in the local vicinity. The white marble in the core is unlike the small exposure of dolomite of Bayhorse Creek (fine-grained, laminated, tan dolostone; Hobbs and Hays, 1990). Likewise, the carbonaceous phyllite from the core is significantly less phyllitic and friable and contains more carbonate than the exposed Garden Creek Phyllite. The previous correlations were also based partly on interpretations of no deformation between the drill site and the nearest exposures of pre-Mesozoic rocks. This is unlikely considering the complex folding and thrust and normal faulting within and between the fragmentary exposures. Significantly, rocks along structural trend east and west of the drill site are both intensely folded into southeast-plunging folds but belong to different “terrane” (see Hobbs and Hays, 1990), indicating faults between the exposures. The details of structural setting, together with more recent documentation of Neoproterozoic rocks in central Idaho, suggest that rocks in the drill core are part of a section not exposed in east-central Idaho and are probably older than rocks exposed in nearby outcrops, either because they are a deeper part of the same section or they are part of a lower structural plate.

For our study, we dated zircons from the tuffaceous matrix of the diamictite at 1140.3 m (3741 ft) depth in the core, ~1.5 m (5 ft) above the bottom of the hole and 55.2 m (181 ft) below the top of the diamictite unit.

U-Pb GEOCHRONOLOGY

Zircon U-Pb geochronology was previously attempted on rocks of the Deep Creek, Arnett Creek, Yellowjacket, Beaverhead, and Ramey Ridge suites (Leonard and Stern, 1966; Evans, 1984; Evans and Zartman, 1988) with mixed results. No age data have been previously published for the Acorn Butte and Rush Creek Point suites, nor for the diamictite of Daugherty Gulch. Our study was initiated in 1997, using multigrain fractions and thermal ionization mass spectrometry (TIMS). These results yielded somewhat scattered isotopic data suggesting the presence of inherited components and Pb-loss due to subsequent metamorphic events (GSA Data Repository Tables DR2 and

DR3 and Fig. DR2 [see footnote 1]). In 2003, all eight samples were dated by in situ microanalysis using an ion microprobe (SHRIMP, GSA Data Repository, Document 2 and Table DR4 [see footnote 1]). These analyses (using a 20–30 μm analytical spot) of pristine areas free of cracks and inclusions enabled determination of age components within individual grains. Thus, although precision of each age was at best about $\pm 1\%$ (2 σ , GSA Data Repository Table DR4 [see footnote 1]), the calculated ages are thought to be quite accurate. In the following discussion, we refer briefly to TIMS data (both from our study and previous studies) for historical perspective, but we used the SHRIMP ages to devise our geologic model. Data are discussed in order of decreasing age.

Results

Zircons from the Acorn Butte pluton (sample 3KE043) are very pale yellow to colorless, euhedral, and prismatic (length-to-width ratio [l/w] = 4–6). In cathodoluminescence (CL) imaging, nearly all grains are composed of partially resorbed, dark, oscillatory-zoned cores and light-colored, faintly zoned rims. Volumetrically, some rims make up as much as 80%–90% of the grains, but in most cases, they are ~10%–40% (Fig. 4A). TIMS analyses of multigrain fractions suggest an age of ca. 653 Ma (Table DR3 [see footnote 1]). Complex zoning of cores and rims revealed by CL imaging suggests that the TIMS data are mixtures of ages. SHRIMP analyses indicate that dark cores have moderate U contents (mostly 100–320 ppm) and Th/U (0.45–0.91), whereas light rims have much lower U (~10–55 ppm) and lower Th/U (0.28–0.51; see GSA Data Repository Table DR4 [see footnote 1]). Ten (of eleven) cores yield an age of 665 ± 6 Ma (Fig. 4B); one analysis is slightly older ($^{206}\text{Pb}/^{238}\text{U}$ age of 709 Ma; see GSA Data Repository Table DR4 [see footnote 1]). Analyses of five rims yield less precise isotopic data (note size of error ellipses in Fig. 4B) and a relatively imprecise weighted average age of 638 ± 13 Ma. We conclude that the Acorn Butte pluton was emplaced at ca. 665 Ma and metamorphosed ~15–40 Ma later, at which time the igneous zircon cores were resorbed and new rims formed.

Zircons from the tuff of Daugherty Gulch (sample DG1) are pale to medium pink, euhedral, and blocky (l/w of 2–4). In CL, most grains show fine, concentric oscillatory zoning (Fig. 4C). SHRIMP isotopic data from 15 grains form a coherent grouping with an age of 664 ± 6 Ma, interpreted as the time of extrusion of the tuff (Fig. 4D).

Zircons from the Rush Creek Point pluton (sample 3KE044) are light to medium pink, and

subhedral to anhedral (l/w = 1–2). Most grains have fine concentric oscillatory zoning. There is no evidence in CL images of metamorphic overgrowths (Fig. 4E). TIMS analyses suggest an age of 661 ± 3 Ma (Table DR3 [see footnote 1]). Eleven SHRIMP analyses from oscillatory-zoned internal areas form a coherent age group with a weighted average age of 651 ± 5 Ma (Fig. 4F), which is interpreted as the time of emplacement of the Rush Creek Point pluton.

Zircons from the Ramey Ridge pluton (sample 2KE085) are very pale pink to colorless, subhedral to euhedral, and stubby to prismatic (l/w = 1–4). Most zircons have fine oscillatory zoning, either concentric or parallel to the long axis of the grains (Fig. 4G). TIMS analyses suggest an age of 651 ± 5 Ma. Fifteen (of 17) SHRIMP analyses obtained from oscillatory-zoned internal areas yield a coherent age group with a weighted average age of 651 ± 5 Ma (Fig. 4H), which is in agreement with the TIMS data (Table DR3 [see footnote 1]). This age is interpreted as the time of emplacement of the Ramey Ridge suite. One SHRIMP analysis is slightly older, presumably due to minor inheritance, and one analysis is slightly younger, possibly due to minor Pb loss.

Zircons from the Deep Creek pluton (sample SRM-238) are colorless, euhedral, and prismatic (l/w = 3–5). Most grains show a distinctive zoning in CL, characterized by fine oscillatory zoning within light or dark patchy areas (appearance similar to sector zoning) of the grains (Fig. 4I). No new TIMS data were collected for this sample, although multigrain TIMS analyses by Evans and Zartman (1988) show a scattered data array indicative of presence of both inheritance and Pb loss, with an apparent maximum age of ca. 520 Ma. SHRIMP analyses indicate that these zircons have unusually low U concentrations (mostly ~20–70 ppm), as previously noted by Evans and Zartman (1988). Seventeen (of 18) SHRIMP analyses form a coherent group with a weighted average age of 497 ± 6 Ma (Fig. 4J). One analysis (SRM238–8.1; GSA Data Repository Table DR4 [see footnote 1]) of an unusual unzoned grain is 450 ± 5 Ma. We do not know if this significantly younger age is due to Pb loss or represents a new growth event.

Zircons from the Yellowjacket suite (sample 3KE051) are very pale yellow or pale pink to colorless, subhedral to euhedral, and stubby (l/w = 1–3). Most grains have fine concentric oscillatory zoning (Fig. 4K). Previous TIMS data (Evans, 1984) yield an age of ca. 490 Ma (no error given). Our TIMS analyses suggest an age of 495 ± 2 Ma. Fifteen SHRIMP analyses from internal zones yield a coherent group with a weighted average age of 491 ± 4 Ma (Fig. 4L),

within uncertainty of the TIMS data. This suite was emplaced at 491 ± 4 Ma.

Zircons from the Arnett Creek pluton (sample 03KL070) are pale yellow to light tan, anhedral to euhedral, and blocky (l/w = 1–3). In CL, they are mostly dark and unzoned, or they have limited, relatively broad oscillatory zoning (Fig. 4M). These zircons contain fairly high U concentrations of ~480–870 ppm (GSA Data Repository Table DR4 [see footnote 1]). No new TIMS data were collected for this sample; Evans and Zartman (1988) obtained somewhat scattered isotopic data from zircons from a different sample of the pluton, suggesting a TIMS age of 492 ± 39 Ma. Fifteen (of 16) SHRIMP analyses yield a coherent group with an age of 486 ± 6 Ma (Fig. 4N), which is interpreted as the time of emplacement of the Arnett Creek pluton. One analysis is slightly younger, probably due to minor Pb loss.

Zircons from the Beaverhead pluton (99KL074 and 99KL075) are pale to medium pink, and anhedral to euhedral. Many grains are fragments with rounded edges. In CL, all grains have fine concentric oscillatory zoning that is truncated at the grain edges (Fig. 4O). The external morphology and CL zoning patterns suggest that most of the zircons have been broken and/or partially resorbed. Previous multigrain TIMS analyses (Evans and Zartman, 1988) yielded scattered, discordant age data, with an apparent minimum age of ca. 483 Ma. Our TIMS analyses suggest a minimum age of ca. 484 Ma. Thirteen (of 15) SHRIMP analyses of zircons from sample 99KL075 are slightly scattered, but they form a coherent group with a weighted average age of 488 ± 5 Ma (Fig. 4P), which is interpreted as the age of emplacement of the Beaverhead pluton. The rather high mean square of weighted deviates ($\text{MSWD} = 1.85$) is indicative of complications to the isotopic systematics. Two analyses are slightly older (ca. 522 Ma). The older ages are consistent with detrital zircon ages obtained by Geslin et al. (2002) from modern surficial deposits of streams draining the Beaverhead pluton, and they interpreted that date as the probable age of the pluton.

In summary, these aligned igneous rocks formed during two magmatic episodes (Table 1). The Acorn Butte, Rush Creek Point, and Ramey Ridge suites, plus the diamicite of Daugherty Gulch, crystallized during the interval 665–650 Ma, whereas the Deep Creek, Yellowjacket, Arnett Creek, and Beaverhead suites were emplaced during the interval 500–485 Ma. The spread of ages for this Cambrian–Early Ordovician set is extended to 530 Ma by a date for syenite in a mine exposure (Gillerman et al., 2008) ~30 km north of the Beaverhead suite.

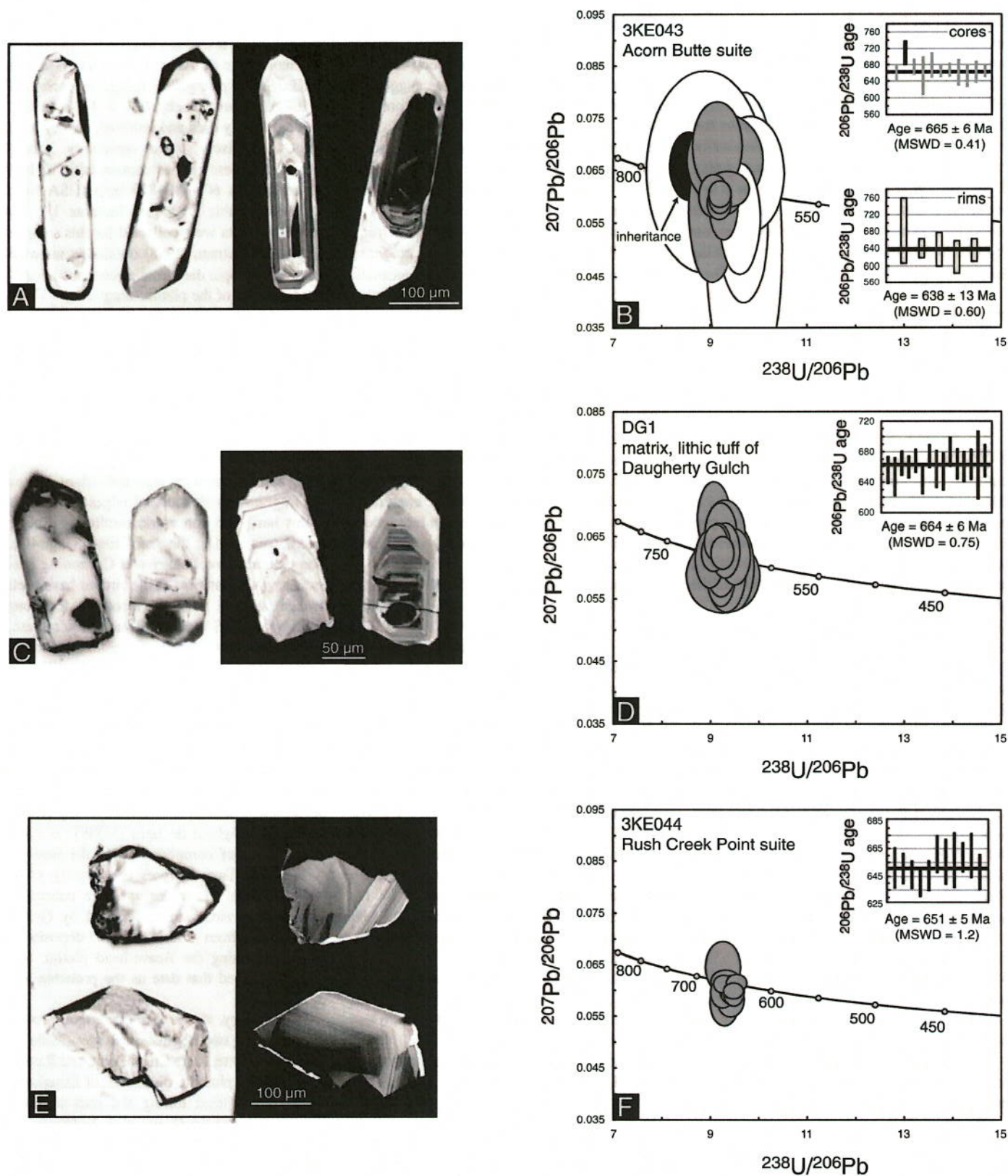


Figure 4 (on this and following two pages). Photographs of representative zircons and Terra-Wasserburg plots of sensitive high-resolution ion microprobe (SHRIMP) U-Pb dates from each sample dated in this study. Transmitted-light images are on left, cathodoluminescence images are in center, and Terra-Wasserburg plots of SHRIMP data are on right for each sample. MSWD—mean square of weighted deviates.

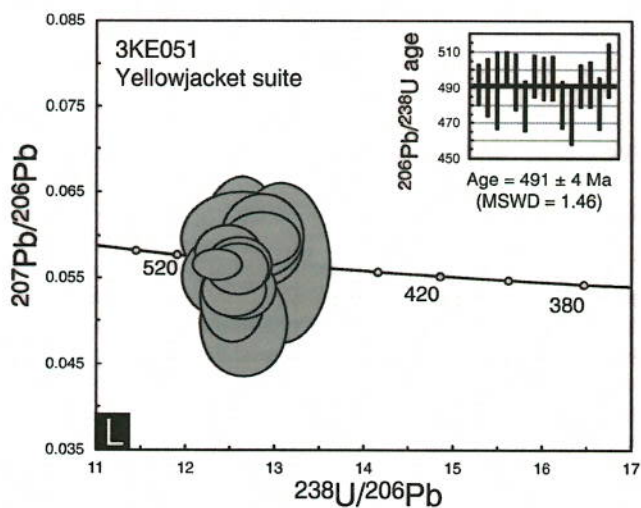
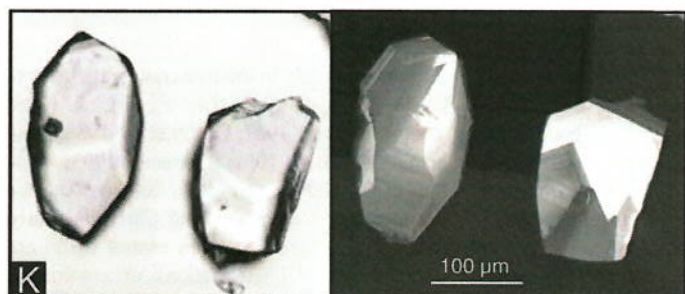
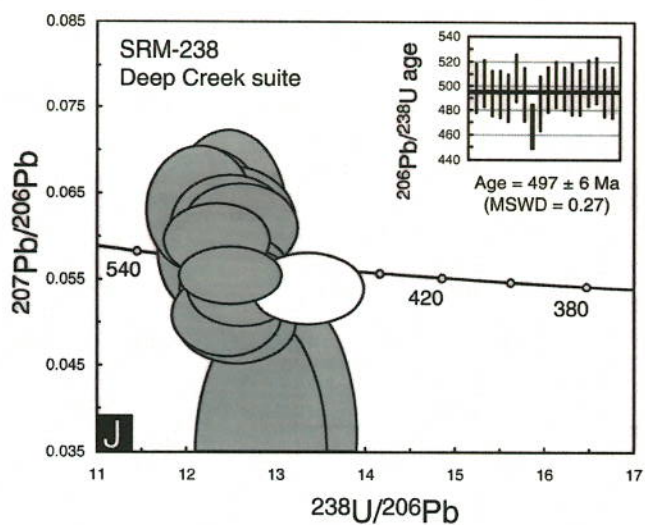
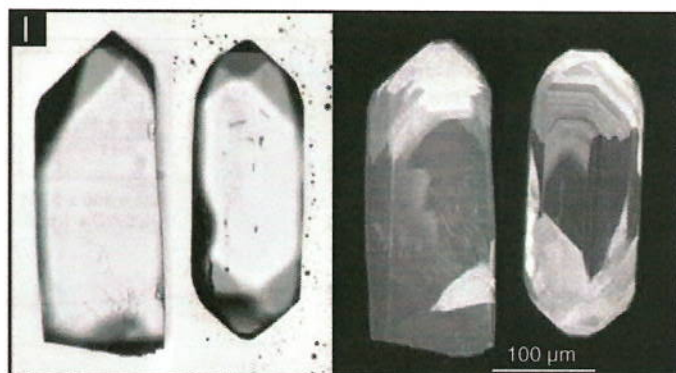
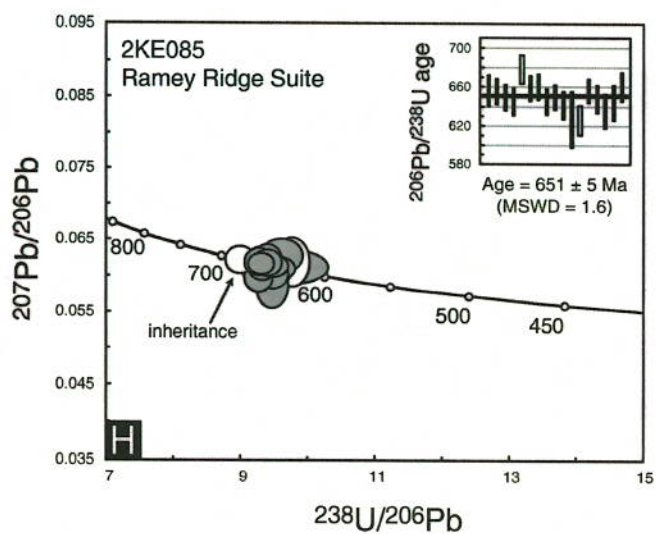
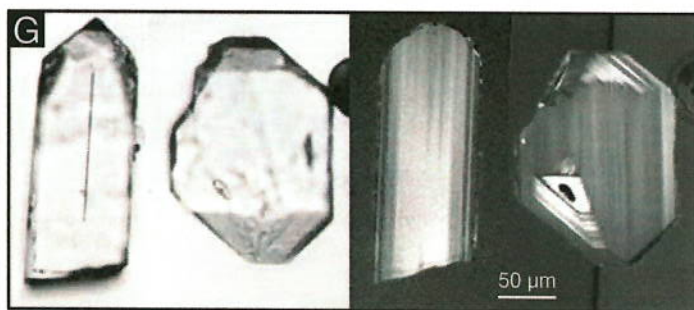


Figure 4 (continued).

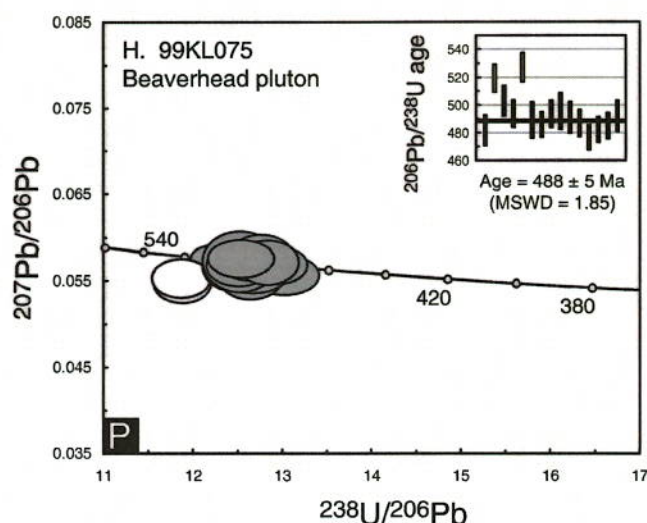
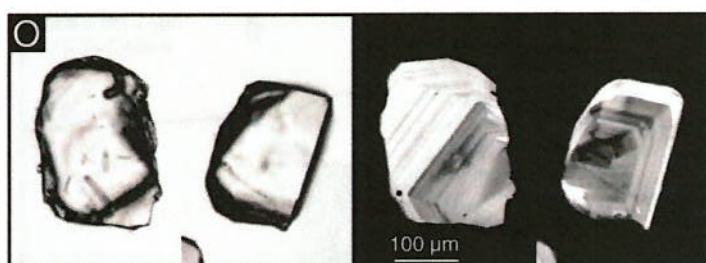
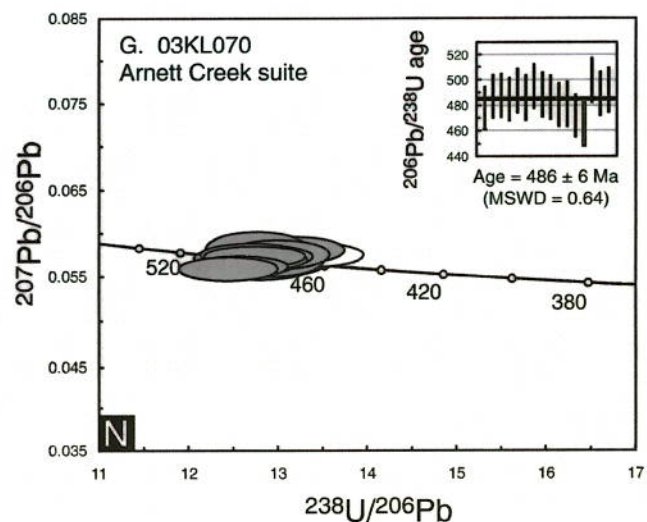
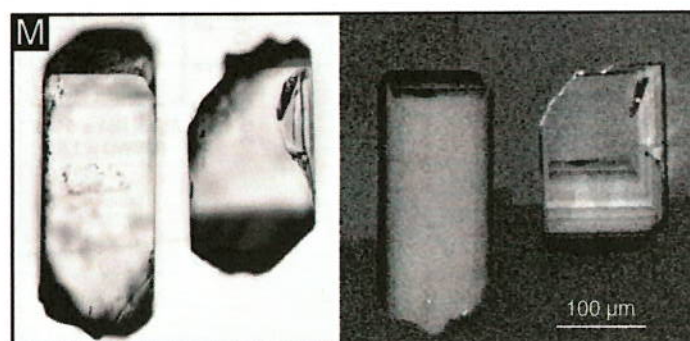


Figure 4 (continued).

TECTONIC SETTING

Genesis of Igneous Rocks

The preliminary geochemistry for the central Idaho Cryogenian and Late Cambrian–Early Ordovician plutonic suites and for Cryogenian volcanic rocks of the Edwardsburg Formation presents a relatively homogeneous data set. The mildly bimodal, alkalic, tholeiitic trend in Cryogenian and Cambrian–Early Ordovician plutonic rocks shows they originated as within-plate magmas. Igneous rocks with similar characteristics are interpreted to have resulted from continental-rift magmatism (Barberi et al., 1982; Suneson

and Lucchitta, 1983; Ruppel, 1995; Betts et al., 1998); significantly, these include Cryogenian and early Paleozoic plutons that occur along the Canadian Cordillera (Goodfellow et al., 1995; Pigage and Mortensen, 2004). The geochemical data indicate that the older Cryogenian, alkali-calcic tholeiitic-trend, bimodal volcanic rocks of the nearby Edwardsburg Formation also formed as within-plate magmas. Furthermore, based on geochemistry and geologic setting, Cryogenian mafic volcanic rocks of the Windermere Supergroup, including those of the Pocatello and Huckleberry Formations (southeastern Idaho and British Columbia, respectively; Fig. 5) and those of sections elsewhere in the Cordillera, are

interpreted widely to indicate continental-rift extension (Stewart, 1972; Harper and Link, 1986; Thompson et al., 1987; Devlin et al., 1988; Ross et al., 1989; Ross, 1991; Link et al., 1993; Rainbird et al., 1996; Lund et al., 2003a). Data for the recurrent Cryogenian and Cambrian–Early Ordovician igneous rocks in central Idaho are consistent with interpretations of continental rift–related magmatism.

The only significant geochemical difference between the 685 Ma volcanic rocks of the Edwardsburg Formation and the 665–650 and 500–485 Ma plutonic suites are that the older volcanic rocks are more distinctly bimodal and are alkali-calcic, whereas the plutonic suites are

TABLE 1. SUMMARY OF U-Pb GEOCHRONOLOGY FOR ZIRCON FROM CRYOGENIAN AND CAMBRIAN SUITES, CENTRAL IDAHO

Sample number	Suite name	SHRIMP	New TIMS	Comments
		age (Ma)	age (Ma)	
3KE043A	Acorn Butte	665 ± 6	653 ± 2	Firms = 638 ± 13 Ma (SHRIMP)
DG1	Tuff of Daugherty Gulch	664 ± 6		
3KE044A	Rush Creek Point	651 ± 5	661 ± 3	6 (of 7) TIMS analyses
2KE085A	Ramey Ridge	651 ± 5	651 ± 5	3 (of 7) TIMS analyses
SRM-238	Deep Creek	497 ± 6		<521 Ma (Evans and Zartman, 1988)
3KE051A	Yellowjacket	491 ± 4	495 ± 2	Ca. 490 Ma (Evans, 1984)
03KL070	Arnett Creek	486 ± 6		492 ± 39 Ma (Evans and Zartman, 1988)
99KL075	Beaverhead	488 ± 6	Ca. 484*	>483 Ma (Evans and Zartman, 1988)

Note: SHRIMP—sensitive high-resolution ion microprobe; TIMS—thermal ionization mass spectrometry.

*Concordant analysis of one grain.

only mildly bimodal and alkalic. Such chemical differences (Ruppel, 1995; Betts et al., 1998) suggest possible increased crustal extension and thinning during the ~35 Ma gap in time from the 685 Ma volcanic episode to the 665–650 Ma plutonic events. However, the available geochemical data do not indicate differences between the Cryogenian (665–650 Ma) and Cambrian–Early Ordovician (500–485 Ma) plutonic suites. Furthermore, there is remarkable similarity in the chemistry of all the plutonic suites from west to east along the Big Creek–Beaverhead belt. Thus, the crust underwent active thinning during the Cryogenian igneous activities, but the crustal setting may not have changed significantly from the Cryogenian plutonic activity to the Cambrian–Early Ordovician activity. The data suggest that for the duration of 200 Ma, recurrent extension-related igneous activity occurred under similar lithospheric, tectonic, and magma genesis conditions along an inherited structural zone.

Concurrent Regional Depositional History

Although Neoproterozoic to early Paleozoic strata are poorly preserved or fragmentary in central Idaho, and particularly in the Idaho batholith province in west-central Idaho, the age and stratigraphic information for early Paleozoic facies trends and for newly identified Neoproterozoic layered rocks in the region can be compared with the age and geographic distribution of the syenite-diorite suites for interpretation of tectonic setting. Classic stratigraphic studies in east-central and southern Idaho and in southwestern Montana delimit the location and character of early Paleozoic miogeoclinal versus cratonic basins and of the facies belts of the miogeocline. Although the miogeoclinal strata were disrupted and transported by numerous thrust faults that were incompletely mapped at the time of the most comprehensive stratigraphic studies, isopach maps and stratigraphic sections have resulted in interpretation of a northwest-striking recurrent Proterozoic to early Paleozoic uplift that has been named,

variously, (1) the Devonian and possibly pre-Devonian Lemhi Arch (Sloss, 1950, 1954), (2) Proterozoic to Devonian Salmon River Arch (Armstrong, 1975), or (3) more inclusive Neoproterozoic to Late Devonian Lemhi Arch that spatially overlapped (and was a southeastern temporal successor to) Mesoproterozoic Belt Island (Harrison, 1972; Ruppel, 1986). Additionally, a set of northeast-striking uplifts have been interpreted, including (1) the early Paleozoic Arco Arch, which separated a central Idaho depocenter from north-striking facies belts in southeastern Idaho (James and Oaks, 1977) and (2) the Cambrian or Early Ordovician Skull Canyon Disturbance (Scholten, 1957), a refinement of Sloss's (1954) mid-Paleozoic uplift.

The thick metamorphosed Cryogenian to Cambrian(?) rocks, which cross central Idaho, lie between the better-known and traditionally correlated sections in northeastern Washington and southeastern Idaho (Fig. 5). The succession of lithologic units in west-central Idaho is similar to the successions in the Windermere Supergroup in northwestern Washington and southeastern Idaho, and dating studies are providing limited chronostratigraphic ties among the sections (Colpron et al., 2002; Lund et al., 2003a; Fanning and Link, 2004, 2008). In these northern U.S. Cordilleran sections, 685 to 650 Ma rift-related volcanic and plutonic activity was ongoing during deposition of diamictites. Magmatic quiescence after ca. 650 Ma is matched in the stratigraphic section by deposition of thick interlayered carbonate and carbonaceous phyllite units after ca. 665 Ma (maximum dates from tuffaceous diamictite of Daugherty Gulch and Scout Mountain Member, east-central Idaho and Pocatello sections; Fig. 5). Deposition of these deeper-water strata, although interrupted by a glaciation (possibly equivalent to the Icebrook Formation, at or before ca. 607 Ma; Kendall et al., 2004), continued through most of the Ediacaran. Thick sandstone units were deposited from before 570 Ma (Crittenden and Wallace, 1973; Colpron et al., 2002) until the Early Cambrian (Hamill Group and up-

per Brigham Group; Fig. 5), after which no deposits are preserved in west-central Idaho until deposition of Eocene volcanic rocks.

In east-central Idaho, partially preserved stratigraphic sections of Ediacaran to Cambrian strata suggest intermittent Ediacaran deposition there, starting in the late Ediacaran (probably after 570 Ma) and continuing through Middle Cambrian time (Figs. 5 and 6). Based on mapping of early Paleozoic units, widespread unconformities have been interpreted after Middle Cambrian and before late Early to Middle Ordovician time (Scholten and Ramspott, 1968; Ruppel, 1986). However, our 488 Ma (latest Cambrian–earliest Ordovician) date for the Beaverhead plutonic suite demonstrates either that the quartzite intruded by this pluton is older than the Kinnikinnick Quartzite (which is Middle Ordovician; Hobbs et al., 1968; Hobbs and Hays, 1990; Ruppel et al., 1975), or that the field relations between the quartzite and pluton need to be reevaluated. The most reasonable alternative correlative unit for this quartzite is the late Ediacaran to Early Cambrian Wilbert Formation. Farther west in central Idaho, stratigraphic information for the outer shelf is incomplete because preserved Neoproterozoic(?) to Ordovician(?) strata are poorly dated and imbricate thrust-stacked (see Hobbs and Hays, 1990). Available data for central to east-central Idaho indicate that deposition occurred from late Ediacaran to the beginning of the Cambrian during the interval in which there was no known magmatic activity in central Idaho. Across east-central Idaho, renewed deposition occurred in a westward deepening subsiding basin from ca. 530 Ma (the onset of the second episode of alkalic plutonism) until the Silurian, after which, Silurian to Devonian unconformities are recorded widely in the rock record (Fig. 6).

In southwestern Montana, preserved early Paleozoic strata (Fig. 6) were deposited in a trough on the continental platform rather than in the miogeocline (Scholten, 1957; Armstrong, 1975; Ruppel, 1986). On the continental platform, major unconformities in the stratigraphic section span from the Mesoproterozoic (ca. 1400 Ma) to Middle Cambrian and from Upper Cambrian to Devonian (Scholten, 1957; Ruppel, 1986). Thus, despite the different settings, the cratonic stratigraphy of southwestern Montana records periods of nondeposition or erosion generally contemporaneous with those in the miogeoclinal stratigraphy of central Idaho.

Structural Synthesis

Recent geologic mapping indicates that the western four plutonic suites are on a single major Cretaceous thrust sheet, whereas the eastern

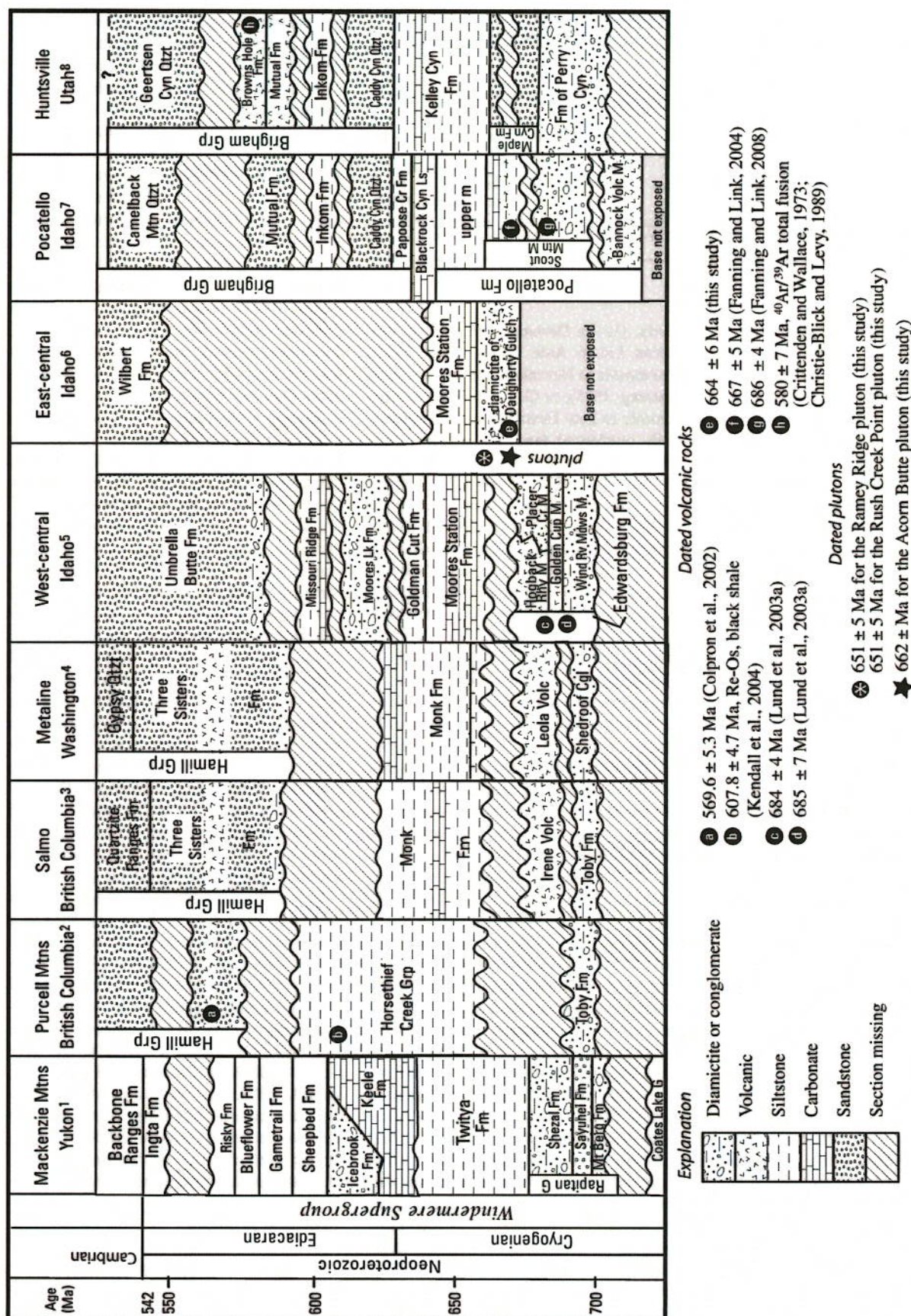


Figure 5. Correlation chart for Cryogenian to basal Cambrian strata from Yukon Territory to northern Utah, showing age and composition of strata, age of related igneous rocks, and general position of unconformities. Locations of sections shown on Figure 1. Cgl—conglomerate; Cny—canyon; Cr—creek; Fm—formation; Grp—group; Lk—lake; Ls—limestone; M—member; Mdws—meadows; Mtn—mountain; Qtz—quartzite; Rhy—rhyolite; Rv—river; Volc—volcanics. 1—Rainbird et al. (1996), Jefferson and Ruelle (1986), Roots and Parrish (1988); 2—Hein and McMechan (1994), Devlin and Bond (1988); 3—Hein and McMechan (1994), Devlin and Bond (1988), Devlin (1989), Colpron et al. (2002); 4—Miller (1994), Evans (1987), Miller and Whipple (1989), Lindsey et al. (1990); 5—this study, Lund et al. (2003a); 6—this study, McCandless (1982), Evans and Green (2003); 7—Crittenden et al. (1971, 1983), Link et al. (1993), Christie-Blick (1997); 8—Crittenden et al. (1971, 1983), Link et al. (1993).

Age (Ma)		Idaho			SW Montana ⁴
		Challis ¹ <i>outer shelf</i>	Big Creek ² <i>shelf</i>	Leadore ³ <i>inner shelf</i>	cratonic
416	Devonian	Grand View Jefferson	<i>not preserved</i>	Jefferson	Jefferson
445	Silurian			Laketown Saturday Mountain Kinnikinic Summerhouse	
488	Ordovician	Kinnikinic Ella			
542	Cambrian	Bayhorse Dol Garden Creek Ph Dol Bayhorse Cr		Tyler Peak Wilbert	Pilgrim Woolsey Meagher Flathead
630	Ediacaran		Umbrella Butte		
			Missouri Ridge Moore's Lake		
	glacial	Moore's Station	Goldman Cut Moore's Station		
	Cryogenian	Daugherty Gulch	Edwardsburg		
800					

1. Hobbs and Hays (1990), Hobbs et al. (1991), this study
2. Lund et al. (2003a)
3. Scholten (1957), Ruppel et al. (1975), McCandless (1982), Ruppel (1986)
4. Scholten (1957), Ruppel (1986)

Figure 6. Generalized stratigraphic chart for Cryogenian to Devonian strata and unconformities from west-central Idaho to southwestern Montana. Italicized units in Challis, Idaho, column are units that are poorly dated in the region (Hobbs and Hays, 1990; Hobbs et al., 1991).

three suites (plus a fourth at Lemhi Pass; Gillerman et al., 2008) are on several lower thrust sheets (Fig. 2; Evans and Green, 2003; Lund, 2004). The distribution of five of the suites (the three Cryogenian, Yellowjacket, Beaverhead) is essentially northwest-trending and linear. In detail, the Late Cambrian–Early Ordovician Deep Creek and Arnett Creek suites and the syenite of Lemhi Pass are on a northeast trend that is marginally offset from the overall northwest regional trend. The limited data for displacement geometries of the thrust faults indicate east- or northeast-directed motion, but the amount of displacement is unknown (Tysdal, 2002; Lund

et al., 2003b). From those movement directions, the suites (and associated sedimentary rocks) formed in a northwest-trending belt that was an unknown amount wider than its present exposure. The northeast offset of three suites may reflect control by inherited structures of the underlying northeast-striking Paleoproterozoic Great Falls tectonic zone, a northwest-southeast component in early Paleozoic extension, or Cretaceous tear faults. Whatever the local influence, the Cryogenian and early Paleozoic suites and the diamictite of Daugherty Gulch originally formed a 200-km-long, northwest-striking belt.

The Big Creek–Beaverhead belt is coincident with a major aeromagnetic anomaly that is interpreted to be one of a series of northwest-trending basement structures that formed at ca. 1500 Ma and controlled Mesoproterozoic sedimentation (McCafferty et al., 1999; Sims et al., 2005). Localization of Cryogenian and Late Cambrian–Early Ordovician alkalic igneous activity along the structure suggests an additional period of prolonged extensional deformation spanning ~200 Ma (recurrent episodes at 685, 665–650, and 530–485 Ma). This recurrent extension is reflected also in the stratigraphic record (Fig. 6). There are progressively larger gaps in the Cryogenian–Middle Cambrian stratigraphic record eastward across central Idaho (area shown on Fig. 2), such that Neoproterozoic strata of these ages are not preserved along or northeast of the Big Creek–Beaverhead belt (Fig. 7). Early Paleozoic rocks are localized to the elongated depositional centers on the flanks of the Big Creek–Beaverhead belt. These patterns resulted from recurrent uplift of the northwest-trending high between the miogeoclinal and cratonic basins.

The name Lemhi Arch is traditionally applied to the uplifted area between the miogeoclinal and cratonic sedimentary basins (Sloss, 1954; Scholten, 1957; Ruppel, 1986). A synthesis of the data indicates that the Lemhi Arch and Big Creek–Beaverhead belt are collocated and that flanking, northwest-trending depocenters located were actively receiving sediments during magmatic quiescence. The Lemhi Arch and Big Creek–Beaverhead belt were fundamental extensional rift-border fault and rotated block structures resulting in a recurrently emergent Cryogenian and early Paleozoic, northwest-striking continental drainage divide and paired basins reflecting uplift-subsidence geometries in response to crustal extension (Fig. 7C). Recurrent extension-related magmatism is commonly sited along preexisting zones of lithospheric weakness (Sykes, 1980; Storti et al., 2003; Vauchez and Tommasi, 2003; Vaughan and Scarrow, 2003), specifically, along rift-border faults (Hayward and Ebinger, 1996). Concurrent magmatism may significantly influence evolution of rift systems by accommodating large amounts of the total strain (Ebinger and Casey, 2001; Rosenbaum et al., 2008). These models apply to the other magmatically active Cordilleran rift segments where Neoproterozoic to early Paleozoic extension-related magmatic systems and rotated fault-block highs (arches) are commonly associated (Fig. 8). The arches bound local depositional centers in the miogeocline (i.e., Dogtooth, Windermere, and Purcell Highs of southern British Columbia; Wheeler, 1966; Reesor, 1973; Kubli and Simony, 1992; Colpron and Price, 1995) or characteristically

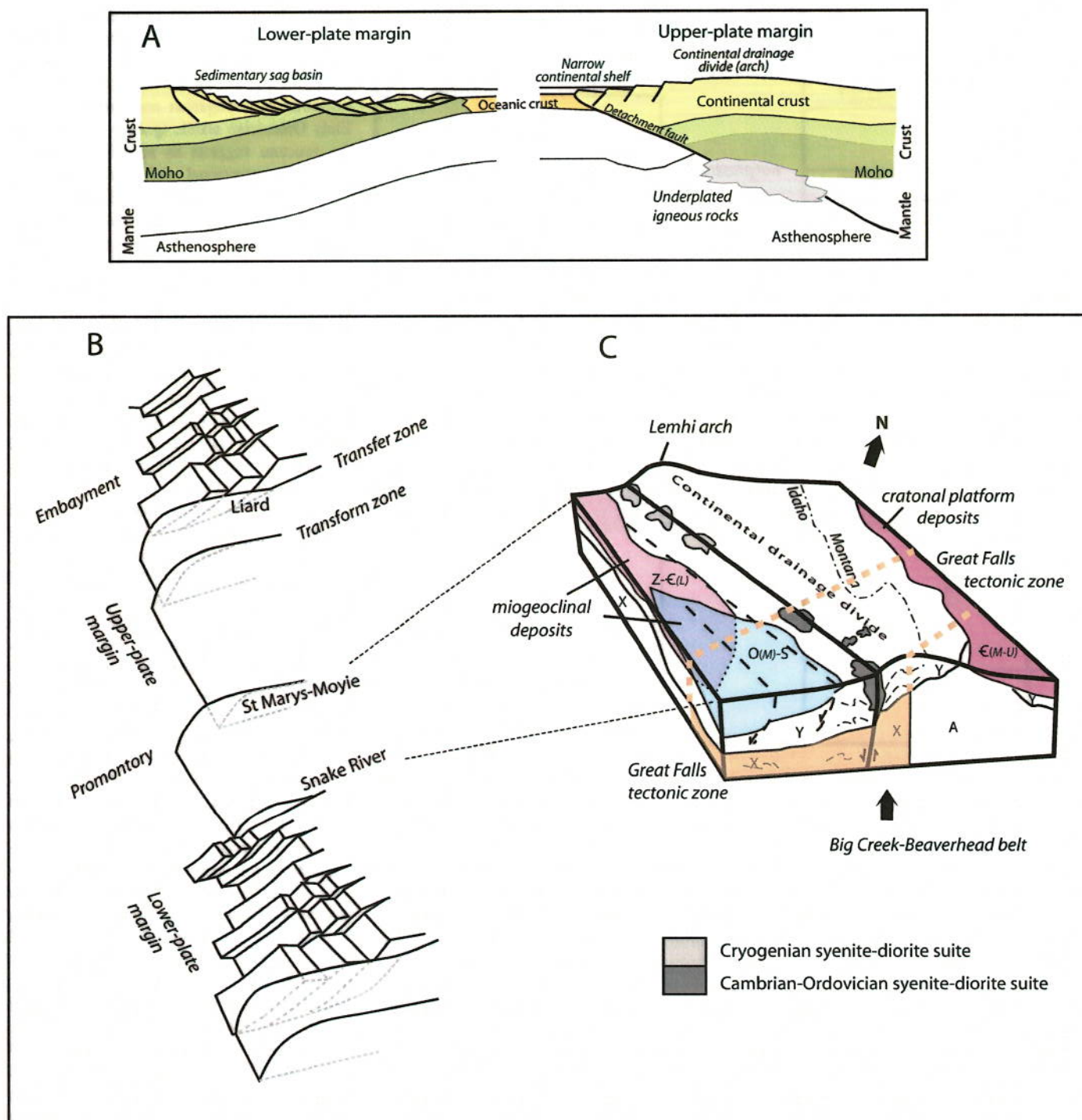


Figure 7. Models for asymmetric rifting of continental crust and for the Big Creek-Beaverhead belt. (A) Schematic cross-section view of structures in asymmetric extension (modified from Lister et al., 1986). (B) Schematic block diagram (after Thomas, 1993) illustrating asymmetric extension for segments along the Cordilleran margin of Laurentia. (C) Block diagram showing generalized relationships along the Big Creek-Beaverhead belt. A—Archean; X—Paleoproterozoic; Y—Mesoproterozoic; Z-E(L)—Lower Cambrian to Neoproterozoic; E(M-U)—Middle to Upper Cambrian; O(M)-S—Middle Ordovician-Silurian.

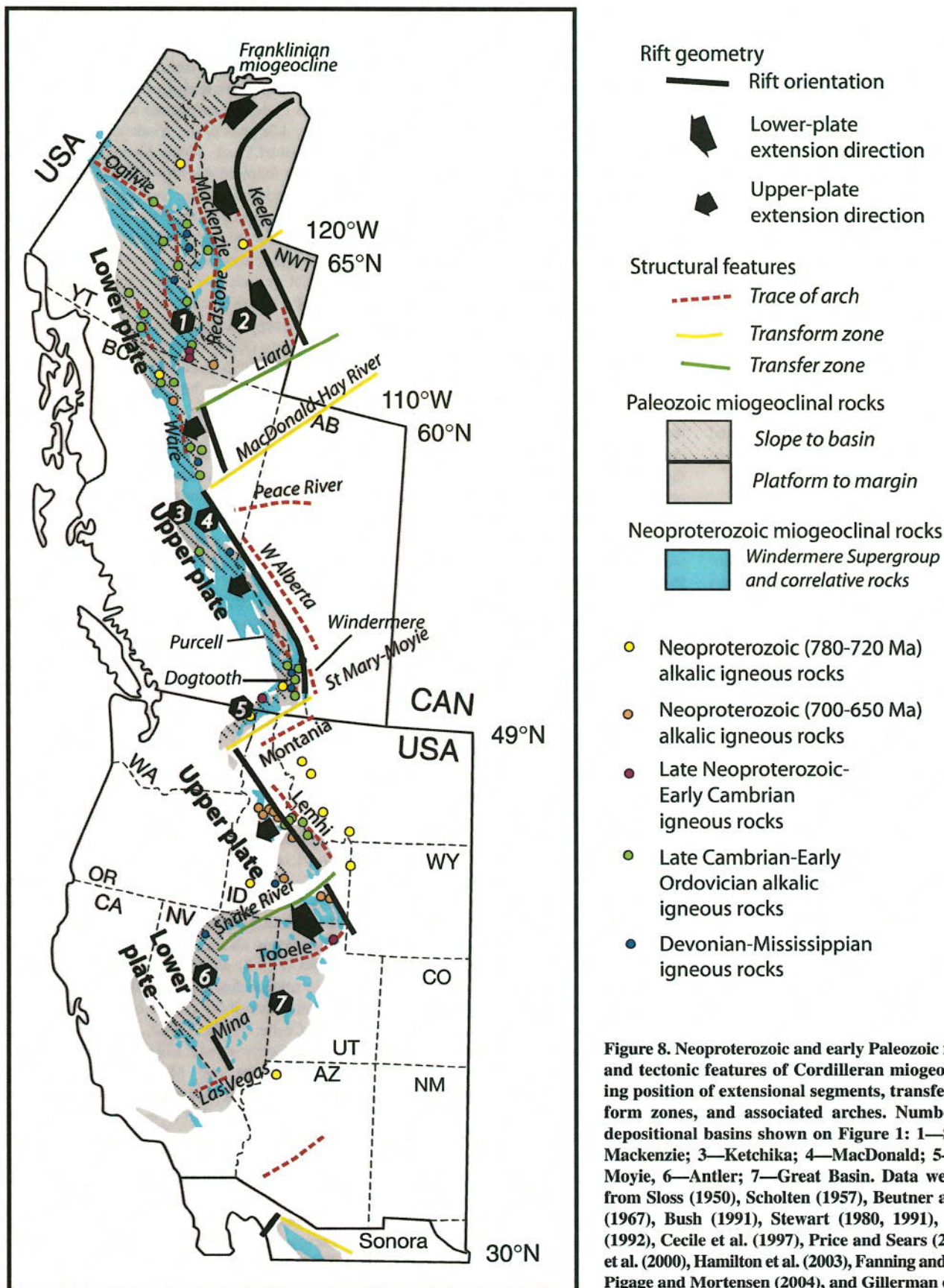


Figure 8. Neoproterozoic and early Paleozoic facies trends and tectonic features of Cordilleran miogeocline, showing position of extensional segments, transfer and transform zones, and associated arches. Numbers refer to depositional basins shown on Figure 1: 1—Selwyn; 2—Mackenzie; 3—Ketchika; 4—MacDonald; 5—St. Mary–Moyie; 6—Antler; 7—Great Basin. Data were compiled from Sloss (1950), Scholten (1957), Beutner and Scholten (1967), Bush (1991), Stewart (1980, 1991), Poole et al. (1992), Cecile et al. (1997), Price and Sears (2000), Tosdal et al. (2000), Hamilton et al. (2003), Fanning and Link (2004), Pigage and Mortensen (2004), and Gillerman et al. (2008).

separate miogeoclinal strata on the outboard side of the arch from cratonic platform strata on the inboard side (i.e., West Alberta Ridge of southern British Columbia; Aitken, 1993).

IMPLICATIONS FOR CONTINENTAL RIFTING

The Big Creek–Beaverhead belt lies along the previously inferred trace of initial miogeoclinal deposits extending between British Columbia and southeastern Idaho (Stewart, 1972). Our study provides a geographic link for Cryogenian through Ordovician extensional tectonics along the Cordillera. Furthermore, our data help constrain the geometry and timing of rifting and formation of a trailing western Laurentian margin during breakup of Rodinia. The data particularly pertain to interpretations for number and extent of continental extension and sequential rift-drift events related to initiation of the Cordilleran miogeocline.

Cordilleran Extensional History

There is controversy over whether the Windermere Supergroup formed as initial deposits of the early Paleozoic rift-drift event (Stewart, 1972; Stewart and Suczek, 1977; Armin and Mayer, 1983; Bond and Kominz, 1984; Bond, 1997; Fedo and Cooper, 2001) or whether there were two Cordilleran rift-drift events, in which the Windermere formed during a separate early continental separation and miogeoclinal subsidence cycle before a latest Ediacaran–early Paleozoic cycle (Thompson et al., 1987; Ross, 1991; Colpron et al., 2002). This has been a difficult question to answer because the strata of both potential cycles are geographically superimposed, and, in both, stratigraphic thickening progressed basinward to the west (present coordinates; Stewart and Poole, 1974; Stewart, 1980; Turner et al., 1989; Ross, 1991; Stevens, 1991; Poole et al., 1992; Link et al., 1993; Cecile et al., 1997; Price and Sears, 2000). Further, the Neoproterozoic rocks are poorly dated and correlations are based primarily on lithostratigraphy and inferred synchronicity of glaciogenic deposits.

Cryogenian extensional mafic magmatism began at ca. 780 Ma in the northern Canadian and northern U.S. Cordillera (Fig. 8; Jefferson and Parrish, 1989; Heaman et al., 1992; LeCheminant and Heaman, 1994; Harlan et al., 2003; Burtis, 2007). The 750–723 Ma mafic volcanic rocks are associated with local sedimentary accumulations in northern Canada, British Columbia, and Arizona (Parrish and Scammell, 1988; Roots and Parrish, 1988; McDonough and Parrish, 1991; Heaman et al., 1992; Karlstrom

et al., 2000; possibly including mafic volcanic rocks at the base of the Windermere Supergroup in northeastern Washington, 762 ± 44 Ma, Sm–Nd age—Devlin et al., 1988). In southwestern and south-central Idaho, orthogneisses isolated within younger plutonic rocks and not associated with sedimentary units are dated at 726 ± 5 Ma (Alexander et al., 2006) and ca. 695 Ma (Durk et al., 2007). In southeastern Idaho, volcanic clasts in Scout Mountain diamictite are dated at 701 ± 4 Ma and 717 ± 4 Ma and are interpreted to be derived from and to date the unconformably underlying, extension-related Bannock Volcanic Member (Fig. 5; Fanning and Link, 2008). The presently available data indicate widely distributed Cordilleran igneous activity at 780, 750, 725, and 700 Ma (Thompson et al., 1987; Timmons et al., 2001; Harlan et al., 2003; although there is little information from the southern U.S. Cordillera) and provide general evidence of synchronous extension along the margin. Because these 780–700 Ma igneous rocks are generally not associated or paired with an interconnected interregional subsiding sedimentary basin, they are interpreted as manifestations of restricted local extension (Roots and Parrish, 1988; Sears, 1990; Ross, 1991; Rainbird et al., 1996; Timmons et al., 2001). However, they are closely linked geographically to the trace of the younger Neoproterozoic and early Paleozoic sedimentary basins (Fig. 8). Although locally expressed, these igneous events and basins signaled onset of rifting that resulted in the more continuous Windermere Supergroup.

Along much of the Cordillera, the 750–723 Ma igneous rocks provide maximum ages for overlying glaciogenic rocks of the Windermere Supergroup, which are the oldest Cordilleran-wide correlated Cryogenian deposits (Crittenden et al., 1972, 1983; Gabrielse, 1972; Stewart, 1972; Eisbacher, 1985; Christie-Blick and Levy, 1989; Aitken, 1991; Link et al., 1993; Miller, 1994; Prave, 1999; Lund et al., 2003a). Direct dating of those widely distributed Windermere interlayered glacial diamictite and subalkalic volcanic rocks at ca. 685 Ma (Ferri et al., 1999; Lund et al., 2003a; Fanning and Link, 2004, 2008) generally substantiates the interregional correlations of associated sedimentary rocks. A younger pulse of igneous activity and associated sedimentary rocks that further date Neoproterozoic extension and sedimentation include (1) a U–Pb zircon date of 667 ± 5 Ma for a tuffaceous layer in the Scout Mountain Member of the Pocatello Formation in southeastern Idaho (Fig. 5; Fanning and Link, 2004), (2) the 665–650 Ma alkalic plutons and 664 Ma volcanic rocks of the Big Creek–Beaverhead belt, and (3) a date of 651 ± 3 Ma for alkalic plutonic rocks that are related to early

extension along the Liard line in the southeastern Yukon Territory (Pigage and Mortensen, 2004). Based on these few dates, thick, widely distributed, deep-water terrigenous detrital (carbonaceous shale and carbonate) rocks equivalent to the middle parts of the Windermere Supergroup (Horsethief Creek, Monk, Moores Station units; Fig. 5) followed the 665–650 Ma igneous activity and substantiate synchronous extension along a connected subsiding sedimentary basin system. Succeeding subsidence-phase rocks include (1) coarser detritus related to an Ediacaran glacial episode that have a minimum date of ca. 608 Ma (Kendall et al., 2004), (2) deeper-water interglacial strata, and (3) late Ediacaran sandstones (Ross, 1991; Link et al., 1993; Rainbird et al., 1996; Lund et al., 2003a), spanning from ca. 570 Ma to the Early Cambrian (Colpron et al., 2002; Fig. 5). Based on the traditional correlations along the Cordillera, the available geochronology, and accepted criteria for miogeoclines, an interconnected, subsiding, margin-long basin was established by ca. 700–685 Ma and lasted until near 570 Ma.

Onset of rifting associated with the Paleozoic miogeocline, as determined from subsidence calculations and dating of volcanic rocks, occurred at 570 Ma (Armin and Mayer, 1983; Bond and Kominz, 1984; Bond, 1997; Colpron et al., 2002). An unconformity in many sections in the latest Ediacaran (between ca. 570 and 540 Ma; Fig. 5) is the result of the continental breakup (Devlin, 1989; Warren, 1997; Colpron et al., 2002). Subsidence accelerated along the length of the margin in the Early Cambrian to Early Silurian and resulted in drift-phase sedimentation, forming the early Paleozoic miogeocline (Cecile, 1982; Armin and Mayer, 1983; Bond and Kominz, 1984; Turner et al., 1989; Gabrielse and Yorath, 1991; Levy and Christie-Blick, 1991; Stevens, 1991; Poole et al., 1992; Bond, 1997; Price and Sears, 2000; Pyle and Barnes, 2003). Late Cambrian and late Early Ordovician alkalic and ultrapotassic igneous rocks (Fig. 8) are distributed along the Canadian miogeocline and provide evidence of recurrent, long-lived crustal extension (Norford and Cecile, 1994a, 1994b; Goodfellow et al., 1995; Cecile et al., 1997; Pyle and Barnes, 2003; Pigage and Mortensen, 2004). The 530–485 Ma rift-related magmatism across central Idaho (Gillerman et al., 2008; this study) extends this activity along the U.S. Cordillera. Subsequent short-lived Devonian to Mississippian rejuvenation of extension was accompanied by igneous activity (Fig. 8) and a possible change in extension direction (Turner et al., 1989; Stevens, 1991; Root, 2001; Pyle and Barnes, 2003); it is interpreted as back-arc extension during slab rollback (Nelson et al., 2006).

Pre-Windermere (780–720 Ma) igneous rocks signaled onset of extension, and, together with Windermere extension, Cordilleran rift trends were further developed during successive latest Neoproterozoic (post-570 Ma) to early Paleozoic extension (Fig. 8). The geographically superposed pulses of igneous activity along the Big Creek–Beaverhead belt enhance the coincidence of multiple rift-related igneous activities. With the Idaho data, the more complete stratigraphic and chronologic connection between northern and southern Cordilleran segments supports the model of two overprinted miogeoclinal along the Cordilleran margin (Ross, 1991; Ross et al., 1995; Colpron et al., 2002), with the proviso that complete development of the Windermere system was interrupted by renewed rifting in the late Ediacaran.

Geometry of Cordilleran Rift System

Despite the heterogeneous amounts of younger thrust telescoping and crustal extension along the Cordillera, numerous palinspastic reconstructions of individual miogeoclinal segments demonstrate that facies trends are generally preserved (Stewart and Poole, 1974; Stevens, 1981; Abbott et al., 1986; Levy and Christie-Blick, 1989) and that rift geometry of the Cordilleran miogeoclinal can be examined as a single tectonic feature (Fig. 8). Studies of several segments suggest that relative differences in the general present-day proportions (including width, depth, or abruptness of sedimentary facies) generally reflect the original proportions (Stewart, 1972; Aitken, 1993; Speed, 1994; Sears and Price, 2003). By including evidence from the northeastern Washington–southeastern Idaho segment (Turner and Otto, 1988; Link et al., 1993; Lund et al., 2003a; this study), along-margin restorations suggest that the present expression of a zigzag-geometry miogeoclinal, with segments having different apparent width, is essentially inherited from the original rift-margin geometry (Hansen et al., 1993; Cecile et al., 1997; Lund, 2008).

The nature and controls of continental extension and the variety of processes and geometries are the subjects of many studies and controversies (Rosenbaum et al., 2008). However, zigzag-shaped rift margins are commonly modeled as the result of margin asymmetry inherent in pure-shear extension (Wernicke, 1985; Lister et al., 1986, 1991; Etheridge et al., 1990; Thomas, 1993, 2006) where opposing upper- and lower-plate extending segments develop (Fig. 7A). Lower-plate margins in such asymmetric systems are characterized by rotated crustal blocks, marked thinning of continental crust, and subsidence across a broad conti-

mental shelf. In contrast, upper-plate margins are characterized by limited crustal thinning, relatively narrow continental shelf margins, and proximal continental drainage divides. Asymmetric geometries may reverse across transfer zones or be offset across transform zones (Fig. 7B) characterized by transverse zone-parallel sedimentary belts and abrupt changes in crustal thickness (Thomas, 1993). Segment types are discriminated by basement thinning, amount of subsidence, and the type and thickness of fill deposited in the developing basin segments (Lister et al., 1986, 1991; Wernicke, 1985; Etheridge et al., 1990; Thomas, 1993, 2006). Models are based on general orthogonal extension (relative to the rift and to parallel facies belts) that persists through time, although oblique, transtension, and polyphase extension would result in complications (Corti et al., 2003; Ziegler and Cloetingh, 2004).

Detailed models for asymmetric extensional geometries in the Canadian Cordillera are presented in Hansen et al. (1993) and Cecile et al. (1997). The broad miogeoclinal basin of the paired Selwyn Basin and Mackenzie Platform in northern Canada formed on a northwest-striking lower-plate extensional margin (Fig. 8). In British Columbia, the markedly narrower system of the Ketchika Trough and the MacDonald Platform formed on a northwest-striking upper-plate margin (Cecile, 1982; Gabrielse and Yorath, 1991; Ross, 1991; Cecile et al., 1997). The northeast-striking Liard transfer zone separates these adjacent, opposite-polarity segments (Cecile et al., 1997). In central British Columbia, the McDonald–Hay River transform offsets miogeoclinal facies belts (Cecile et al., 1997). Related features are margin-parallel arches (e.g., West Alberta Arch) and Neoproterozoic and early Paleozoic within-plate alkalic magmatism (Okulitch, 1985; Sevigny, 1988; Norford and Cecile, 1994a, 1994b).

The Kootenay Arc, southern British Columbia–northeastern Washington, joins the northwest- and northeast-trending Cryogenian to early Paleozoic structures and facies trends of two intersecting margin segments (Fig. 8). The northern part is characterized by Cryogenian–Ordovician north- to north-northwest-trending Purcell, Dogtooth, and Windermere Arches and normal-fault-bounded, westward-deepening depositional basins (Kubli and Simony, 1992; Norris and Price, 1966; Lis and Price, 1976; Lickorish and Simony, 1995; Warren, 1997; Price and Sears, 2000), which underwent dominantly east-west to northeast-southwest extension (Warren, 1997). Geographically overlapping these on the south, there are concurrently active northeast-trending normal faults, linked to northwestward-deepening syntectonic

deposits on the north flank of the recurrently emergent Montania Arch (Deiss, 1941; Lis and Price, 1976; Benvenuto and Price, 1979; Bush, 1991; Colpron and Price, 1995; Warren, 1997; Price and Sears, 2000). Structures related to the northeast-trending St. Mary–Moyie zone underwent complex reactivation in the Mesozoic (Warren, 1997; Larson and Price, 2006) but originated as a Cryogenian to early Paleozoic transform between the upper-plate extensional margin in the southern Canadian Cordillera and an upper-plate margin in Washington–Idaho (Lund, 2008).

Using criteria for asymmetric extension and the extensional model for the Canadian margin, the original tectonic setting for the northeastern Washington–southeastern Idaho segment is interpreted based on (1) relatively thin miogeoclinal strata and relatively narrow continental-shelf facies belt (Sloss, 1950, 1954; Scholten, 1957; Stewart, 1972; Armstrong, 1975; Turner and Otto, 1988, 1995; Lund et al., 2003a), (2) the Cryogenian and Late Cambrian–Early Ordovician Big Creek–Beaverhead belt marking a rift-border fault, (3) the Lemhi Arch, a continental drainage divide, that paralleled but lay inboard of the miogeoclinal rocks and that underwent recurrent Cryogenian to Devonian uplift (Sloss, 1954; Scholten, 1957; Armstrong, 1975; Ruppel, 1986), and (4) aeromagnetic and isotopic data indicating that North American basement is present and extended, but not offset, across central Idaho beneath the miogeoclinal (Armstrong et al., 1977; O'Neill and Lopez, 1985; Sims et al., 2005). These features are characteristic of an upper-plate margin that underwent northeast-southwest extension.

In southern Idaho, the facies data indicate a 400-km-long, narrow zone of northwest-deepening, Cambrian to Devonian slope facies rocks that extend northeast between southeastern Idaho and north-central Nevada (James and Oaks, 1977; Poole et al., 1977, 1992; Stewart and Suczek, 1977; Stewart, 1980; Stevens, 1981, 1991; Little, 1987; Miller et al., 1991). A coincident narrow belt of northeast-trending, Neoproterozoic carbonate and volcanoclastic, slope-facies rocks is preserved in north-central Nevada (Stevens, 1981; Ehman and Clark, 1985, 1990; Little, 1987; Lund, 2008). Further, the isotopically defined Precambrian continental margin (Speed, 1982; Elison et al., 1990; Tosdal et al., 2000) parallels the facies trend. The northeast-striking deviation of facies belts and the Precambrian basement boundary were portrayed previously as an offset caused by tearing of Sevier thrust faults (1) along a proposed Snake River fault obscured by Neogene rocks in the Snake River Plain (Poole et al., 1977, 1992) or (2) along the Wells fault on the south

side of the Snake River Plain (Thorman, 1970). However, a retrodeformation study of the Wells fault refutes the model of facies offsets along that fault, indicating instead that the offset of the facies belts was due to original geometry of the miogeoclinal margin (Stevens, 1981). This northeast-striking facies trend also is interpreted as part of a more fundamental feature including (1) basement structures of unspecified origin that controlled northeast-striking sedimentary facies (Stevens, 1981; Coats and Riva, 1983), (2) a northeast-striking, right-lateral offset of the Cordilleran rift (Tosdal et al., 2000), or (3) the boundary of an embayment in the Laurentian margin (Little, 1987). Based on the stratigraphic, structural, and isotopic data in context of Cordilleran rift tectonics, the Snake River Plain is underlain by a recurrently reactivated fundamental structure that was caused by a Cryogenian to early Paleozoic, northeast-trending transfer zone (left-lateral relative motion) between opposite-polarity central Idaho and Great Basin rift segments (Lund, 2008).

The Great Basin was formed by several miogeoclinal segments. The northern (main) segment is characterized by wide, thick, northwest-trending Neoproterozoic to early Paleozoic miogeoclinal facies belts, including carbonate-shelf facies rocks that deepen westward toward structurally telescoped basin facies rocks of the Antler allochthon (Stewart, 1980, 1991; Poole et al., 1992). The early Paleozoic basin facies strata contain intermittent within-plate and basinal origin alkalic basalts (Turner et al., 1989). Isotopic signatures from younger plutonic rocks document progressive westward thinning of Laurentian basement toward its edge in central Nevada by means of northwest-trending rift-related normal faults (Tosdal et al., 2000; Crafford and Grauch, 2002; Grauch et al., 2003). These characteristics suggest that the northern Great Basin margin formed as a northwest-striking, lower-plate rift segment (Lund, 2008). Along the northern edge of the Great Basin (southern side of the Snake River zone), parallel northeast-trending early Paleozoic structurally controlled Tooele and lesser Arco Arches and the Ibex and other subbasins are superimposed on the regional northwestward deepening of this segment (Ross, 1964; James and Oaks, 1977; Hurst et al., 1985; Hintze, 1988; Cooper and Keller, 2001) and reflect intersection of the Snake River and Great Basin segments.

In southwestern Nevada, northeast-trending miogeoclinal facies belts (Stewart, 1980, 1991; Poole et al., 1992) form the Mina deflection (Wetterauer, 1977; Tosdal et al., 2000). This change in trend is interpreted as an oroclinal bend of the margin (Speed, 1978; Stewart,

1985) or an original feature of the Laurentian margin (Oldow, 1984; Stewart, 1988b; Saleeby et al., 1994; Tosdal et al., 2000). The Mina deflection was probably a short transform-margin segment (Lund, 2008). Exposures of Neoproterozoic and Paleozoic miogeoclinal strata of southeastern California, although offset by the Walker Lane fault zone and possibly by the San Andreas fault system and the Mojave-Sonora megashear, include northwest- and northeast-trending segments (Stewart, 1980, 2005; Stevens et al., 1992; Cooper and Keller, 2001; Fedo and Cooper, 2001). These faulted remnants may have formed short rift and transfer segments.

Inheritance and Reactivation

The general northwest trends of the rift margin and of individual rift segments are orthogonal to most older basement domains and domain boundaries. In Idaho, the buried Great Falls tectonic zone (as expressed by aeromagnetic signature, porphyry Mo deposits, etc.; Sims et al., 2005; Klein and Sims, 2007) is orthogonal to the northeastern Washington-southeastern Idaho rift segment, but features related to it continue at depth southwest of the Big Creek-Beaverhead belt, across central Idaho, toward the western edge of Laurentia.

Domain boundaries in the basement may have exerted control on the location of the transform-transfer zones (compare Figs. 1 and 9). For example, the Liard transfer overlies a preexisting basement structure, and the MacDonald-Hay River transfer overlies an aeromagnetic anomaly along the southwestern extent of the Great Slave Lake basement shear zone (Cecile et al., 1997; Ross et al., 2001). The Vulcan Low, an aeromagnetic feature of the Albertan subsurface that is interpreted as a Paleoproterozoic domain boundary or suture (Ross et al., 1991), underlies the St. Mary-Moyie transform. This basement feature was reactivated by Neoproterozoic to Paleozoic syndepositional block faulting (the St. Mary-Moyie transform; Lund, 2008) and by Mesozoic deformation (Price and Sears, 2000; Sears and Price, 2003). The Snake River transfer may have been controlled by the underlying Madison mylonite zone (Lund, 2008), where that mylonite is the northwestern margin of coherent Wyoming province rocks (Karlstrom et al., 2002; Sims et al., 2005). Similar control of transfer and transform segments by preexisting basement structures has been demonstrated for other continental rifts (Khalil and McClay, 2001; Mohriak and Rosendahl, 2003) and in experimental data (Corti et al., 2003).

Extension-related alkalic magmatism is closely associated with fault-controlled arches

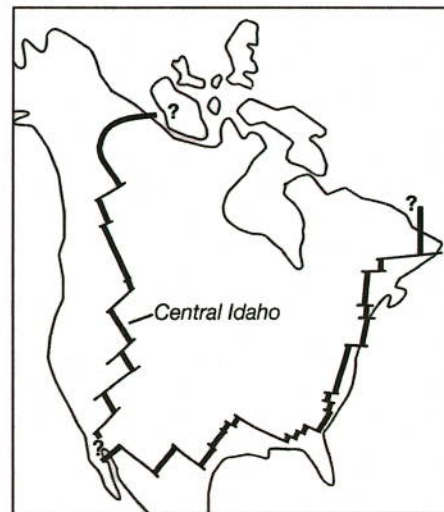


Figure 9. Traces of Neoproterozoic and early Paleozoic rift margins of eastern and western Laurentia. Wide line—rift segment, thin line—transform or transfer zone. Data were compiled from Cecile et al. (1997), Poole et al. (2005), Thomas (2006), and this study.

and particularly with intersections between them and basement-controlled transform-transfer zones (i.e., Redstone Arch and Fort Norman line; Cecile et al., 1997). This pattern is consistent with location of magmatic activity at the intersections of the Purcell Arch and St. Mary-Moyie zone, the Lemhi Arch (Big Creek-Beaverhead belt) and Great Falls tectonic zone, and the Lemhi Arch and Snake River zone in the northern U.S. Cordillera.

Rift-margin structures may reactivate older structures and may control subsequent deformation (Thomas, 2006). The Cordilleran zigzag reentrants and promontories were precursors to the salients and recesses in the successive thrust belts (Fig. 1). For example, deformed rocks of the Antler orogen extend farther east in Idaho (north of the Snake River Plain) than in southern Idaho and Nevada (south of the Snake River Plain), mirroring the eastward offset of the miogeocline and the relatively thinned underlying basement north of the Snake River Plain. Similarly, the Mesozoic Cordilleran thrust sheets parallel the zigzag rift margin, constituting “a best-fit curve around that framework” (Thomas, 1977, p. 1270; Oldow, 1984). This geometry is somewhat distorted in the Montana fold-and-thrust belt, where the major thrust slabs (Sears, 1988) diverge from the trend of the miogeocline, perhaps because deformation was transferred into the thick Mesoproterozoic Belt Supergroup strata underlying and east of the miogeocline (Fig. 1).

Geometry of Laurentian Rift Systems

A compilation of previous studies (Hansen et al., 1993; Cecile et al., 1997; Tosdal et al., 2000), in conjunction with data presented herein for the Washington to southeastern Idaho segment and interpretation of the St. Mary–Moyie and Snake River zones, provides a consistent geometric model for formation of the Cordilleran margin of western Laurentia (Lund, 2008). Along the entire margin, there is continuity in northwest-trending rift segments offset by northeast-trending transfer and transform zones as proposed for the Canadian Cordillera (Fig. 8).

Growing geochronologic evidence shows that during the late Neoproterozoic to early Paleozoic, Laurentia (as the center of Rodinia) was bounded by three major rift margins (Bond et al., 1984; Thomas, 1993; Speed, 1994; Sears, 2001; Poole et al., 2005). Along these margins, alkalic and subalkalic magmas help constrain the timing of extension (Hanson and Al-Shaieb, 1980; Myers et al., 1981; Tollo et al., 2004). The northern Franklinian margin (Cecile and Norford, 1993), western Cordilleran (Cecile et al., 1997), and eastern Appalachian (to Sonoran) (Thomas, 1993, 2006; Poole et al., 2005) define a triangular Laurentia-bounding margin geometry (Fig. 9). Along the Franklinian margin, voluminous 723 Ma mafic magmatism dates initial extension followed by early Paleozoic trailing-margin sedimentary rocks (Heaman et al., 1992; Shellnutt et al., 2004; Cecile et al., 2001). Along the Appalachian margin, two main phases of extension occurred at 760–700 Ma and 620–535 Ma, in which initial separation occurred at ca. 570 Ma, and final rifting and breakup occurred at ca. 540–535 Ma (Aleinikoff et al., 1995; Cawood et al., 2001). For the Cordillera, localized extensional events occurred at 780–720 Ma (described in Cordilleran Extensional History section). Two main rift-drift events occurred as early margin-wide extension from ca. 700 to 685 Ma and initiated sedimentation in a deepening basin until ca. 570 Ma. Renewed rifting and accelerated extension occurred from latest Ediacaran through Ordovician. Thus, extension along Laurentian margins significantly overlapped.

There are two proposed scenarios for the connection between the southern end of the western Laurentian Cordilleran rift system and the eastern Laurentian Appalachian rift system: (1) offset of a continuous Appalachian–Cordilleran rift system by the younger San Andreas fault and Mojave–Sonora megashear (Stevens et al., 1992; Stewart, 1988a, 2005) or (2) a poorly preserved wrap-around Appalachian–Cordilleran system (Speed, 1994; Poole et al., 2005). These scenarios assume northeast-trending (northwest-southeast extending) rift and northwest-trending transform

segments for both margins (e.g., Speed, 1994; Sears and Price, 2003; Poole et al., 2005; Dickinson, 2006). However, studies of segments of the western Laurentian margin suggest northeast-southwest extension (Hansen et al., 1993; Cecile et al., 1997; Tosdal et al., 2000; this study), whereas the extension direction for the Atlantic margin was northwest-southeast (Thomas, 2006). Furthermore, the miogeoclinal shelf and slope deposits in Sonora face south-southeast along the Gulf of Mexico margin and are a part of the Appalachian margin (Poole et al., 2005). Thus, a change in extension direction must have occurred between western Sonora and southeastern California (Fig. 9). Based on this, complicated younger structural offsets in southern California probably obscure a right-angle junction (part of a triple junction?) between miogeoclinal rocks that formed along orthogonally extending Cordilleran and Appalachian rift systems.

Additional zones of extension as manifested by Cambrian alkalic plutonism in the southern U.S. Cordillera may bear on the evolution of the Laurentian rift margins. Late Cambrian (536 Ma) alkalic mafic complexes and dikes in central and western Colorado (Bickford et al., 1989) are related to a larger west-northwest-oriented zone of widely dispersed Cambrian to Ordovician magmatic centers and uplifted blocks from Oklahoma to Utah (Larson et al., 1985; Allen, 2004). This structurally controlled magmatic belt was interpreted to be part of a Cambrian to Early Ordovician continental rift system (Hoffman et al., 1974; Larson et al., 1985) that was a multiply reactivated, late-stage rift structure associated with the eastern margin of Laurentia (Thomas, 1993). On another structural trend, Late Cambrian (ca. 500 Ma) alkalic dikes and plutons are documented over a broad area in New Mexico and have been related to continental extension (Evans and Clemons, 1988; McLemore et al., 1999; McMillan et al., 2000). The New Mexico igneous rocks are explained by any or all of the following models: (1) a northeast-trending aulacogen (McMillan et al., 2000), (2) distal extension along projections of the southern Oklahoma and Delaware aulacogens (Evans and Clemons, 1988; McMillan et al., 2000), (3) tapping of mantle magma by a transform fault formed along the rifted southern margin of Laurentia (Thomas, 1977; Evans and Clemons, 1988), and/or (4) reactivation of long-lived basement features inland from the rifted or transform-faulted southwestern margin of Laurentia (McMillan et al., 2000). These two Cambrian alkalic igneous belts (Oklahoma–Utah and New Mexico) are not related to trailing-margin (miogeoclinal) sedimentation, and they probably originated as failed rift systems related to the Appalachian rift margin, as previously suggested.

Rifting along the Cordilleran margin was coeval with Appalachian and Franklinian margins, and it is interpreted to reflect rifting of the Rodinia supercontinent at ca. 750 Ma and continental breakup along Cordilleran and Appalachian margins in Early Cambrian (Meert and Torsvik, 2003). Formation of interconnected extensional basins signaling a continuous subsiding trailing margin in the Cordillera occurred after ca. 725–685 Ma (when dated igneous rocks formed part of regionally correlated sedimentary successions). Although the timing is becoming fairly well established, the tectonic setting of Cryogenian to early Paleozoic extensional events, with respect to what was being rifted and the age(s) of continental separation, is debated (see Meert and Powell, 2001; Meert and Torsvik, 2003; Bogdanova et al., 2008). The Cryogenian and Late Cambrian–Early Ordovician extension-related magmatic belt of central Idaho has important implications for Neoproterozoic to early Paleozoic continental reconstructions and paleogeographic models.

CONCLUSIONS

Seven composite alkalic syenite-diorite suites and a tuffaceous diamicite, preserved as roof pendants and inliers in the Idaho batholith and Challis volcanic-plutonic complex of central Idaho, provide new data on the timing and geometry of the Cordilleran rift margin of western Laurentia. SHRIMP U–Pb zircon dates define two discrete magmatic events with ages of ca. 665–650 Ma and 500–485 Ma. Together with previously documented 685 Ma bimodal alkali-calcic volcanic rocks of the Edwardsburg Formation, the new data document recurrent alkalic igneous activity for ~200 Ma along the more than 200-km-long northwest-oriented Big Creek–Beaverhead belt. Geochemical data provide evidence of alkali-calcic and alkalic, within-plate magmas that originated during progressive crustal extension. Comparison of evidence indicates that the magmatic pulses are related to uplift events that are recorded by gaps in the stratigraphic record of adjacent miogeoclinal and cratonic rocks. The magmatic belt defines a multiply reactivated border-fault and uplift system (Lemhi Arch) that was geographically paired with an elongate depocenter (miogeocline).

The Big Creek–Beaverhead belt of Cryogenian and Cambrian–Early Ordovician extensional magmatism is coextensive with the recurrent uplifts of the Lemhi Arch and is parallel to outboard, progressively southwest-thinned, continental crust and to narrow, deepening, water, miogeoclinal facies belts. The Big Creek–Beaverhead belt defines a major

northwest-trending segment of the Neoproterozoic Windermere and early Paleozoic Cordilleran miogeoclinal belts, extending across northeastern Washington and southeastern Idaho. This mostly obscured segment connects to other poorly understood northeast-trending segments: (1) the St. Mary–Moyie transform (defining the southern boundary of the Kootenay Arc) in British Columbia–northeastern Washington and (2) the Snake River transfer in southeastern Idaho–northern Nevada. Thus, western Laurentian rift geometry was a systematic zigzag pattern formed by northwest-trending (northeast-southwest extending) rift zones offset by northeast-trending transfer and transform zones.

The dated extensional igneous rocks in central Idaho better define and substantiate models for geographically superimposed Windermere and early Paleozoic rift and basin subsidence cycles. The continuity of dated interlayered volcanic and sedimentary rocks and the subsequent deepening water subsidence-phase terrigenous detrital sequence indicate that active 700–650 Ma extension was succeeded by an interconnected rift-subsidence basin. The basin was disrupted by renewed latest Ediacaran rifting that resulted in more significant subsidence and a complete trailing-margin section.

Extension on the Cordilleran and Appalachian margins was generally synchronous. However, they represent opposing Laurentian margins having divergent extensional geometries and with a junction now obscured by complicated faulting in present-day Southern California–Sonora.

ACKNOWLEDGMENTS

Access into the isolated plutonic suites was skillfully provided by G. Wilson and R. Tucker by helicopter and R. Gillihan by horseback. J. Wurster, Payette National Forest, and T. Cookro (U.S. Geological Survey [USGS]) helped collect samples of Ramey Ridge and Acorn Butte suites. S. Miller and P. Barbarick (Bureau of Land Management) collected a new sample from the Arnett Creek suite. R.S. Lewis (Idaho Geological Survey) provided photographs of syenite-diorite textural relations. R.G. Tysdal (USGS) provided insights into regional stratigraphy and structure and sedimentary features in the core. A.K. Khudoley discussed Cryogenian to Ordovician alkalic magmatism in Siberia with the authors. Initial discussions with the late S.W. Hobbs (USGS) and R. Hargraves (Princeton University) spurred our interest in the diamictite of Daugherty Gulch. S.W. Hobbs provided a copy of an unpublished report on the Daugherty Gulch core by B.F. Leonard (USGS). Many thanks to E.H. Bennett (Idaho Geological Survey) for having stored the Daugherty Gulch core and making it and the accompanying well logs available for study. Thoughtful reviews of the manuscript by F.G. Poole, D.B. Stoesser, and P.K. Link improved presentation of data and interpretations. Comments by Associate Editor P.A. Cawood helped

clarify interpretations. Encouragement and insights into geology of the Canadian Cordillera by reviewer M. Colpron are particularly appreciated.

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MANUSCRIPT RECEIVED 20 OCTOBER 2008
 REVISED MANUSCRIPT RECEIVED 8 APRIL 2009
 MANUSCRIPT ACCEPTED 10 MAY 2009

Printed in the USA